

1 **The Biogeophysical Climatic Impacts of Anthropogenic**
2 **Land Use Change during the Holocene**

3

4 **M. Clare Smith¹, Joy S. Singarayer¹, Paul J.Valdes², Jed O. Kaplan³, Nicholas P.**
5 **Branch⁴**

6 ¹ Centre for Past Climate Change and Department of Meteorology, University of Reading,
7 Reading, United Kingdom

8 ² School of Geographical Sciences, University of Bristol, Bristol, United Kingdom

9 ³ Institute of Earth Surface Dynamics, University of Lausanne, Lausanne, Switzerland

10 ⁴ School of Archaeology, Geography and Environmental Science, University of Reading,
11 Reading, United Kingdom

12

13 Correspondence to: M. Clare Smith (mcsmith@pgr.reading.ac.uk)

14

15 **Abstract**

16 The first agricultural societies were established around 10kaBP and had spread across much
17 of Europe and southern Asia by 5.5kaBP with resultant anthropogenic deforestation for crop
18 and pasture land. Various studies (e.g. Joos et al., 2004, Kaplan et al., 2011, Mitchell et al.,
19 2013) have attempted to assess the biogeochemical implications for Holocene climate in
20 terms of increased carbon dioxide and methane emissions. However, less work has been done
21 to examine the biogeophysical impacts of this early land use change. In this study, global
22 climate model simulations with HadCM3 were used to examine the biogeophysical effects of
23 Holocene land cover change on climate, both globally and regionally, from the early
24 Holocene (8 kaBP) to the early industrial era (1850 CE).

25 Two experiments were performed with alternative descriptions of past vegetation: (i) potential
26 natural vegetation simulated by TRIFFID but no land-use changes, and (ii) where the
27 anthropogenic land use model, KK10 (Kaplan et al., 2009, 2011) has been used to set the
28 HadCM3 crop regions. Snapshot simulations have been run at 1000 year intervals to examine

1 when the first signature of anthropogenic climate change can be detected both regionally, in
2 the areas of land use change, and globally. Results from our model simulations indicate that in
3 regions of early land disturbance such as Europe and South-east Asia detectable temperature
4 changes, outside the normal range of variability, are encountered in the model as early as
5 7kaBP in the June/July/August (JJA) season and throughout the entire annual cycle by 2-
6 3kaBP. Areas outside the regions of land disturbance are also affected, with virtually the
7 whole globe experiencing significant temperature changes (predominantly cooling) by the
8 early industrial period. The global annual mean temperature anomalies found in our single
9 model simulations were -0.22°C at 1850 CE, -0.11°C at 2kaBP and -0.03°C at 7kaBP.
10 Regionally, the largest temperature changes were in Europe with anomalies of -0.83°C at
11 1850 CE, -0.58°C at 2kaBP and -0.24°C at 7kaBP. Large-scale precipitation features such as
12 the Indian monsoon, the intertropical convergence zone (ITCZ), and the North Atlantic storm
13 track are also impacted by local land use and remote teleconnections. We investigated how
14 advection by surface winds, mean sea level pressure (MSLP) anomalies, and tropospheric
15 stationary wave train disturbances in the mid- to high-latitudes led to remote teleconnections.

16

17 **1 Introduction**

18 The first agricultural societies were established in the Near East around 10kaBP and had
19 spread across most of Europe by 5.7kaBP (Zohary et al., 2012) and to India by 9kaBP
20 (Tauger, 2013). In China domestication of millet and rice began about 8.5 kaBP initially
21 spreading more slowly than in Europe but reaching S.E. Asia by 5.5kaBP (Roberts, 2013;
22 Tauger, 2013). Agriculture was also independently developed in Mesoamerica with maize
23 possibly being cultivated as far back as 9kaBP (Piperno et al., 2009) but, as in China, it spread
24 slowly to other areas.

25 The most important anthropogenic alteration of the natural environment was the clearing of
26 forests to establish cropland and pasture, and the exploitation of forests for fuel and
27 construction materials (Darby, 1956). This long history of anthropogenic land cover change
28 (ALCC) has implications for regional hydrology and climate, and possibly for global climate.
29 Deforestation results in both biogeochemical and biogeophysical changes. The
30 biogeochemical changes tend to increase temperature by the emission of greenhouse gases
31 such as CO_2 and CH_4 (CH_4 emissions are influenced not just directly by deforestation but by
32 irrigation in rice agriculture and by emissions from livestock and humans). The impacts of

1 biogeophysical changes are many and varied, being dependent on the local climate, soil, and
2 the natural vegetation that is being replaced, e.g. if natural savannah or grassland is replaced
3 by crops the impact will not be as great as if woodland is replaced.

4 There are several mechanisms by which biogeophysical changes due to deforestation can
5 affect regional climate. A combination of reduction in aerodynamic roughness, in the root
6 extraction of moisture and in the capture of precipitation on the canopy leads to reduced
7 evaporation and thus decreases the fluxes of moisture and latent heat from the surface to the
8 atmosphere. These changes work to increase the local surface temperature (Lean and
9 Rowntree, 1993). Conversely, the increase in surface albedo due to deforestation acts to
10 decrease surface temperature by increasing the reflection of shortwave radiation. This is
11 particularly true at high latitudes where lying snow is a factor for some of the year and the
12 snow covered ground is no longer masked by the canopy of the forest. Generally, in mid to
13 high latitudes the albedo increase is considered to be the dominant effect; leading to a net
14 cooling of the regional surface temperature, whereas in the moist tropics the evaporation is
15 more important and, therefore, a localised overall warming may result (Betts et al., 2006).

16 During the Holocene the climate has been influenced by natural forcings. Orbital variations
17 have caused a decline in summer solar insolation in the Northern Hemisphere over the last
18 6000 years. During the same period concentrations of greenhouse gases such as CO₂ and CH₄
19 have been increasing. On decadal to centennial timescales fluctuations in solar and volcanic
20 activity have also had a climatic impact. (Wanner et al., 2008; Schmidt et al., 2011) The
21 impact of ALCC is superimposed on these natural forcings. The extent and timing of these
22 early anthropogenic land surface changes is the subject of much debate, as is their role in
23 changing Holocene climate. Ruddiman (2003) proposed the idea that anthropogenic impacts
24 on greenhouse gases, and consequently climate change, began thousands of years ago as a
25 consequence of early agriculture and have been increasing in amplitude ever since, which he
26 termed 'the early anthropogenic hypothesis'. The idea has been hotly debated in the literature
27 (e.g. Broecker and Stocker, 2006; Joos et al., 2004; Singarayer et al., 2011; Mitchell et al.,
28 2013; Kaplan et al., 2011). Whilst the early anthropogenic hypothesis may likely not account
29 entirely for the pre-industrial rises in CO₂ and CH₄ there is no doubt that land use changes do
30 have climatic impact on both regional and global scales. The real debate is the scale of these
31 effects of early agriculture.

1 Whilst paleoecological and archaeological evidence of anthropogenic land use changes exists,
2 there are not enough sites to comprehensively determine continental scale impacts of
3 deforestation (Kaplan, 2009). Therefore, in order to better estimate impacts of anthropogenic
4 land use, several databases of land use change have been developed. Examples of these
5 include the HYDE 3.1 (History Database of the Global Environment, Goldewijk et al., 2011),
6 KK10 (Kaplan et al., 2009 and 2011) and Pongratz et al. (2008) models. Although the
7 methodologies differ in the details, the basic premise of these models is that from an
8 estimated database of historical population trends, anthropogenic deforestation is calculated
9 based on population density and the suitability of land for crops or pasture.

10 To quantify the impact of ALCC on climate, datasets of past ALCC can be used in
11 conjunction with climate models. Several studies have estimated the influence of pre-
12 industrial ALCC on global climate (He et al., 2014; Kutzbach, 2011; Pongratz et al., 2010).
13 Globally, the biogeophysical effects of anthropogenic land use change have been estimated to
14 cause a slight cooling that is offset by the biogeochemical warming, giving a net global
15 warming (He et al., 2014; Pongratz et al., 2010). At the local to regional scale, in the most
16 intensively altered landscapes of Europe, Asia, and North America, the biogeophysical effects
17 can be comparable with the biogeochemical (He et al., 2014; Pongratz et al., 2010). In
18 addition, Strandberg et al. (2014) used a regional climate model to evaluate the climatic effect
19 of anthropogenic deforestation in Europe at 6kaBP and 0.2kaBP with both the HYDE 3.1 and
20 KK10 ALCC scenarios. For the KK10 scenario at 6kaBP small but significant temperature
21 differences were found in summer and, at 0.2kaBP, changes up to ± 1 °C were found over
22 widespread areas in both summer and winter. Other authors (e.g. Oglesby et al., 2010; Cook
23 et al., 2012) have modelled a decrease in precipitation in response to deforestation in
24 Mesoamerica.

25 These existing studies are, however, limited in either temporal or spatial extent and do not
26 address the question of when anthropogenically induced climate change first occurs. In this
27 study global climate model simulations are used to provide a comprehensive evaluation of the
28 influence that the biogeophysical effects of regional human-induced land cover change have
29 had on the climate both globally and regionally throughout much of the Holocene. As
30 described in detail in Sect. 2, the period under consideration is from 8kaBP to pre-industrial
31 (1850) and snapshot simulations with HadCM3 were run at 1000-year intervals. The results
32 are highlighted in Sect. 3. For evaluation purposes palaeoclimate reconstructions from

1 Bartlein et al .(2011) and Marcott et al .(2013) have been compared with the results from the
2 model runs (Sect. 4). The implications are discussed in Sect. 5.

3

4 **2 Methodology**

5 **2.1 Model Description**

6 The climate simulations in this study were performed with the UK Hadley Centre coupled
7 global climate model, HadCM3 (Gordon et al., 2000; Pope et al., 2000) with the Met Office
8 Surface Exchange Scheme (MOSES2.1) (Essery, 2003) and TRIFFID (Top-down
9 Representation of Interactive Foliage and Flora Including Dynamics) (Cox, 2001) dynamic
10 vegetation. The experimental set-up is summarised in Table 1.

11 HadCM3, is a coupled atmospheric, ocean and sea ice model. The atmospheric component
12 has a horizontal resolution of 2.5° latitude and 3.75° longitude with 19 unequally spaced
13 levels in the vertical and a 30 minute time step. It has an Eulerian advection scheme and
14 includes effects of CO_2 , N_2O , CH_4 , CFC11 and CFC12. The spatial resolution over the ocean
15 is 1.25° by 1.25° with 20 unequally spaced layers extending to a depth of 5,200m. It also
16 includes a 1.25° by 1.25° resolution model for the formation of sea-ice with simple dynamics
17 whereby the sea-ice drifts on the ocean currents (Cattle and Crossley 1995).

18 TRIFFID is coupled to the GCM (General Circulation Model) via MOSES every 10 days of
19 the model run. Within TRIFFID nine surface types are specified: 5 plant functional types
20 (PFTs) and 4 non-vegetation types.

21 HadCM3 was widely used in both the third and fourth assessment reports of the
22 Intergovernmental Panel on Climate Change (IPCC, 2001, 2007) and still performs well in a
23 number of tests relative to other global GCMs (Covey et al., 2003; IPCC, 2007). For the fifth
24 IPCC assessment it has been superseded by HadGEM2 (Collins et al., 2011), but being
25 relatively computationally efficient HadCM3 can be the better choice for some palaeoclimate
26 modelling applications as it allows more and/or longer runs to be conducted than would be
27 possible with a higher-resolution model.

28

1 **2.2 Project-Specific Model Configuration**

2 The version of HadCM3 used does not include interactive ice, carbon cycle, or methane and
3 thus must be forced with prescribed changes in orbit, greenhouse gases and ice-sheet
4 evolution. Orbital parameters are taken from Berger and Loutre (1991), atmospheric
5 concentrations of gases are determined from ice cores (CO₂ from Vostok (Petit et al., 1999;
6 Loulergue et al., 2008) and CH₄, and N₂O from EPICA (Spahni et al., 2005)) and the ice-
7 sheet evolution is estimated using the ICE5G model of Peltier (2004). For further details of
8 these natural forcings of the climate model readers are referred to Singarayer et al. (2011).

9 To prescribe Holocene ALCC the KK10 dataset of Kaplan et al. (2009 and 2011) was used.
10 The original dataset is on a 5' spatial resolution and has modelled crop and pasture land use
11 for every year from 8 kaBP to present. For this study the data at 1000-year intervals were
12 taken (8kaBP, 7kaBP, etc.) for both crop and pasture combined and upscaled to the spatial
13 resolution of HadCM3 to formulate a time series of cropland masks (Fig. 1). Within TRIFFID
14 the global crop area is designated by a cropland mask, which can only be occupied by
15 agricultural-type vegetation (i.e. C₃ and C₄ grasses) or bare soil (Betts et al., 2007). Hence, the
16 actual cropland is equivalent to the mask area, less inland water, urban and ice tiles, and less
17 the area covered by non-grassland vegetation or bare soil. The crop mask area is not
18 dynamically updated by climate data. The cropland area incorporates the natural C₃/C₄ grass
19 fractional areas before converting tree fractions.

20 For each simulation, the boundary condition forcings (orbit, greenhouse gases and ice sheets)
21 were specified, and in all simulations the initial conditions were the same, based on a spun-up
22 early industrial simulation. Simulations were run for 1000 years. By the final 500 years of the
23 simulation the climate system has adjusted to a new surface equilibrium and thus these final
24 500 years were averaged to result in the mean altered climatic conditions. The relatively long
25 averaging period increases the signal-to-noise ratio between the modified and control
26 climates, and thus distinguishes differences that are statistically significant, but which can be
27 hidden by decadal/multidecadal variability in shorter averaging periods. This is especially
28 important for assessing the impact of agriculture in the earlier time slices of the Holocene
29 when the land use change is small and localised.

30

1 **3 Results**

2 **3.1 Surface Air Temperature**

3 **3.1.1 Local Impacts of Land Use**

4 In the regions where ALCC was significant, surface air temperature changes can be seen in all
5 the time slice simulations (Figs. 2, 3 and 4) and for all time slices except 8kaBP (not shown)
6 the temperature anomalies in most regions are outside the normal range of variability, which
7 is considered to be within 2 standard deviations of the mean. The anomalies are more
8 pronounced in the JJA season (Fig. 2) than DJF (Fig. 3). This is due to a combination of the
9 land imbalance between the northern and southern hemispheres, the lack of land surface
10 changes in the extra tropical southern hemisphere and the enhanced effect of land surface
11 changes during the season of greatest solar insolation and plant growth (which is JJA in the
12 northern hemisphere).

13 The direct temperature response to ALCC varies with the degree of latitude but the
14 relationship is not straightforward as it depends on local climate, soil, and the natural
15 vegetation. In the extratropics, where the albedo effect is generally dominant, there is a trend
16 towards increasing (negative) anomalies with an increase in disturbance fraction (Fig. 5a for
17 P.I.). At 7kaBP the range of extra tropical temperature anomalies within the areas of land
18 disturbance is $+0.1/-1.2^{\circ}\text{C}$ (JJA) and $+0.6/-0.5^{\circ}\text{C}$ (DJF), by 4kaBP it has increased to $+0.4/-$
19 2°C (JJA) and $+0.9/-1^{\circ}\text{C}$ (DJF) and by the pre-industrial period (PI; 1850) it had reached
20 $+0.7/-4^{\circ}\text{C}$ (JJA) and $+0.3/-2^{\circ}\text{C}$ (DJF). Although there are some positive temperature
21 anomalies, the vast majority of grid points show a negative temperature trend. Regions with
22 the highest ALCC intensity show the largest negative temperature anomalies, in particular
23 Europe and E. Asia/China, where the agricultural land use occurs earliest and has the highest
24 concentration of land conversion (Figs. 2, 3, 4 and 5a).

25 The tropical response shows less of a trend because the impact of reduced evaporation is more
26 significant and thus there are conflicting signals between the cooling effect of increased
27 albedo and the warming effect of reduced evaporation. In some tropical areas ALCC leads to
28 net cooling while in other areas net warming is simulated (Figs. 2, 3, 4 and 5a), partly
29 dependent on the availability of moisture at the surface and partly on cloud cover changes
30 (not shown). The main areas that show a warm anomaly response are Southern Africa and
31 India in JJA from 5kaBP and in DJF the area bordering the Bay of Bengal where E. India is

1 warmer from 6kaBP extending to the east coast of the Bay of Bengal by 2kaBP. The Indian
2 JJA warming is enhanced by cloud feedbacks; a decrease in monsoon circulation leads to
3 decreased cloudiness, thus increasing the shortwave radiation reaching the surface and
4 warming the lower atmosphere. In contrast, tropical South America generally shows a net
5 cooling in response to ALCC. Around the mid to late Holocene the tropical temperature
6 anomaly range within the areas of land disturbance is $\pm 0.5^{\circ}\text{C}$ and by the early industrial era
7 (1850CE) it had reached $\pm 1^{\circ}\text{C}$ (Fig. 5a).

8 Analysis of the standard deviation of both the KK10 and Control simulations indicated no
9 significant changes in the amplitude of interannual variability of surface temperature or
10 precipitation (using the F-test statistic).

11 **3.1.2 Remote Impacts of Land Use**

12 In addition to the local temperature changes described above, cooling can also be observed in
13 regions remote from the areas of major ALCC, particularly in the Northern Hemisphere. The
14 most intense cooling is always in the regions of ALCC but even as early as 7kaBP in the JJA
15 season our model simulations show a band of cooling that stretches across much of the extra-
16 tropical Northern Hemisphere and the North Atlantic (Fig. 2). This cooling starts influencing
17 the northern Pacific regions by 5kaBP and by the early industrial era the surface air
18 temperature over most of the world's land masses and much of the ocean is cooler due to the
19 effects of ALCC in remote areas.

20 In the DJF season in the Northern Hemisphere ALCC leads to cooling both locally and
21 regionally starting at 7kaBP (Fig. 3). The model simulations also show cooling in the Arctic
22 and warming in Siberia. Cooling remote from the areas of major ALCC becomes more
23 extensive by 3kaBP and most landmasses of the Northern Hemisphere are cooler than the
24 control simulation by 2kaBP. The Siberian warm anomaly has ceased by 3kaBP but it remains
25 less affected by the cooling than other regions. In the Southern Hemisphere, cooling remains
26 more localised until 2kaBP by which time the majority of the land surface is cooler than the
27 control.

28 There is an increased temperature anomaly response for the same level of disturbance fraction
29 in the later timeslices (Fig. 5b) implying that the responses to the land use changes are not just
30 due to the local effects. Some possible mechanisms for these remote impacts are large-scale
31 circulation changes (such as stationary waves in the upper troposphere at mid-high latitudes

1 and monsoonal circulation changes), near-surface advection, and the amplifying factors of
2 snow cover. These mechanisms will be discussed in more detail in Sect. 3.2 for atmospheric
3 dynamics and Sect. 3.3.2 for snow cover changes. Changes to the natural vegetation cover
4 (outside the regions of land use) due to the climatic impacts of land use were also investigated
5 as a potential mechanism of further feedbacks but were not found to be significant.

6 **3.2 Atmospheric Dynamics**

7 **3.2.1 Upper Tropospheric Dynamics**

8 There are several factors that can impact the dynamics of the upper troposphere in response to
9 ALCC. Cooler surface air temperature means that the density of the air is greater and,
10 therefore, the geopotential height in cooler regions will be lower (See Fig. 7c and d). The
11 changes to Mean Sea Level Pressure (MSLP) described in Sect. 3.2.2 can change the areas of
12 convergence and divergence leading to changes in regions of vertical motions. The unequal
13 latitudinal distribution of the temperature anomalies, with the regions of greatest cooling in
14 the mid-latitudes of the Northern Hemisphere, affects the meridional temperature gradient
15 leading to a change in baroclinicity, which has been shown to impact storm tracks (Yin,
16 2005).

17 In the JJA season from 7kaBP there is a reduction of the 500hPa geopotential height over the
18 extra-tropical Northern Hemisphere in a pattern similar to the temperature pattern but more
19 extensive, completely encircling the globe. From 3kaBP onwards the height reduction
20 expands southwards so the geopotential height is lowered almost everywhere by pre-industrial
21 times. The most intense reduction is always in a zonal belt across Europe and E. Asia. The
22 Southern Hemisphere response from 5kaBP appears to show a standing wave pattern affecting
23 the subtropical highs.

24 In the DJF season a stationary wave in the anomaly field in both the Northern and Southern
25 Hemispheres is apparent at all timeslices but most pronounced at 4kaBP (Fig. 7d). This is a
26 recognised response to surface temperature anomalies as described in Hoskins and Karoly
27 (1981) although, in this case, there are multiple thermal anomalies caused by ALCC. By
28 2kaBP there is a reduction in 500hPa geopotential height over most of the globe. Note that
29 there are several areas that are not statistically significant in Fig. 7c and d implying a large
30 amount of variability. These geopotential height changes contribute to the simulated remote
31 temperature changes by altering the regions of vorticity, which in turn influence the regions of

1 ascent and descent and thus the MSLP and surface climatic conditions. The geopotential
2 height anomalies can also alter the pattern of the upper level winds thus influencing surface
3 storm tracks.

4 In particular, the positioning of the geopotential height anomalies in the earlier timeslices (up
5 to 4kaBP, Fig.7d) indicate an increased tendency towards a positive Tropical/ Northern
6 Hemisphere (TNH) pattern (Barnston et al., 1991) with above average heights over the
7 Bering Strait/ Gulf of Alaska and northeastward of the Gulf of Mexico and below average
8 heights over eastern Canada. This would be expected to cause cooler temperatures over the
9 continental United States by increasing the transport of cold polar air into the United States.
10 In several time slices it can be seen that DJF temperature anomalies over Bering Strait/Alaska
11 (e.g. Fig. 3) show a warming pattern where there are positive geopotential height anomalies in
12 Fig. 7c and d. The DJF temperature anomalies (warming) over Siberia up to 4ka (Fig. 3) also
13 relate to the stationary wave pattern. The decreased heights over the polar regions in most of
14 the time slices are indicative of a positive Arctic Oscillation (AO) (Fig. 7b) which has been
15 shown to be correlated with milder winters in Siberia (Tubi and Dayan, 2012).

16 **3.2.2 Mean Sea Level Pressure**

17 Changes to MSLP can also have an effect on the climate system. The colder surface air
18 temperature in the region of disturbance means reduced ascent in those regions and thus
19 higher MSLP and as discussed in Sect. 3.2.2 the changes in the upper troposphere can
20 influence the surface conditions. The MSLP changes can be seen quite clearly in China from
21 6ka and in Europe by 4kaBP (Fig. 6 for PI JJA), although the DJF situation in Europe is less
22 coherent probably due to more remote influences such as the North Atlantic storm track.
23 These MSLP changes could play a part in steering weather systems and thus influencing the
24 climate in remote regions. For example, the JJA MSLP anomaly pattern (Fig. 7a and b) over
25 the North and Central Atlantic is indicative of a negative phase of the North Atlantic
26 Oscillation (NAO) with above-normal pressure over the North Atlantic and below-normal
27 pressure over the central Atlantic. This pattern is apparent from 5kaBP. The NAO alters the
28 intensity and location of the North Atlantic jet stream and storm track and thus the patterns of
29 heat and moisture transport (Hurrell, 1995). A negative NAO would contribute to a tendency
30 to wetter summers in all but the southernmost regions of Europe (Folland et al., 2009) and this
31 was seen in the results described in Sect. 3.3.1. The DJF NAO shows a trend towards a
32 positive NAO which could result in drier winters over the Mediterranean region (Hurrell et

1 al., 2003) although it is difficult to ascertain whether this is the case as the pattern is not as
2 consistent as JJA.

3 **3.2.3 Surface Advection**

4 Some of the cooling in regions adjacent to the areas of land disturbance is due to advection by
5 low-level winds. In regions with a prevailing wind direction an advection pattern can be
6 clearly seen. Fig. 6 shows advection of cold air from the areas of land use change in East Asia
7 to the east across the West Pacific and from the region of land disturbance in Mexico to the
8 west across the East Pacific, this also affects the sea surface temperatures (SSTs) in these
9 regions of the Pacific (not shown). In other areas where the surface wind direction is more
10 variable the effect is more difficult to detect but probably does contribute to the spread of the
11 cold anomaly outward from the region of land disturbance.

12 **3.3 Hydroclimate**

13 Our model simulations show precipitation changes in response to ALCC but it should be
14 noted that there are larger uncertainties in climate model simulated precipitation and other
15 variables related to model dynamics than for temperature, which is primarily controlled by
16 thermodynamics (Shepherd, 2014). Different climate models show a wide range of responses
17 in their dynamics to palaeo and future climate change scenarios and we acknowledge that
18 aspects of the precipitation anomaly patterns in this study may be less robust than that for
19 other climate variables.

20 The precipitation responses to ALCC (Fig. 8 for JJA and Fig. 9 for DJF) tend to be caused by
21 a response to large-scale circulation changes rather than being directly attributable to local
22 land use. The European precipitation response in the DJF season is not entirely consistent
23 throughout the time slices but the general response is a slight decrease in precipitation around
24 the western and Mediterranean coasts with this dryness extending further into the continent by
25 1 kaBP. The simulations show that Europe in the JJA season has an increase in precipitation
26 compared to the control from 7kaBP onwards, which gradually increases in extent, possibly
27 influenced by the increased tendency to a negative North Atlantic Oscillation (NAO). Positive
28 anomalies begin in the warm pool of the Gulf of Mexico and extend across the North Atlantic
29 following the track of positive anomalies to the 850hPa wind field (Fig. 10). In addition, as
30 the cooler temperature anomalies extend quite high in the troposphere over Europe this

1 increases relative humidity throughout the low to mid troposphere (not shown) and thus the
2 likelihood of large-scale precipitation.

3 In India there is a decrease in monsoon precipitation from 5kaBP, which then gradually
4 increases in intensity. This is partly driven by the slightly cooler Indian sub-continent
5 temperatures (Fig. 11) in the critical months for monsoon development and also by cooler
6 temperatures in Europe and East Asia and increased snow cover on the Tibetan plateau. This
7 leads to decreased monsoonal circulation and decreased cloudiness. There are also changes to
8 the East Asian monsoon with wetter conditions to the north and south of the region and drier
9 conditions in the centre. This pattern is seen reasonably consistently from 4kaBP onwards.

10

11 **3.3.1 Inter Tropical Convergence Zone (ITCZ)**

12 Analysis of the precipitation fields (Figs. 9 and 10) shows an overall southward migration of
13 the ITCZ. These changes are most obvious in the Atlantic and Pacific Oceans and over the
14 continent of Africa. In the DJF season there are changes to the ITCZ in the Western Pacific
15 but a consistent pattern is not seen until 3ka when there is southward shift in the ITCZ over
16 the Atlantic and Atlantic coasts leading by 2kaBP to a decrease in precipitation in the interior
17 of southern Africa and wetter conditions on the coasts. There is generally increased
18 precipitation over the Indian Ocean and the Amazonian region of South America and a
19 reduction over the Bay of Bengal but this pattern is not entirely consistent throughout the
20 timeslices. Similarly, in the JJA season there are changes to the ITCZ throughout all the
21 timeslices but these do not all show a consistent pattern. The most persistent changes are
22 increased precipitation over the Pacific from 5kaBP, over Central America from 7kaBP and a
23 southward shift from 2kaBP. This southward shift in the ITCZ in the JJA season impacts the
24 West African monsoon with lower precipitation in a belt across the monsoon region by 2kaBP
25 although the west coast of North Africa is wetter.

26 The generally cooler temperatures in the Northern Hemisphere may influence the latitudinal
27 position of the Hadley cell and thus the location of the ITCZ via the influence on the inter-
28 hemispheric temperature gradient resulting in the strengthening of the northward cross-
29 equatorial energy transport (e.g. Kang et al., 2008). This shift south of the ITCZ to transport
30 heat to the cooler northern hemisphere is seen in both the DJF and JJA seasons.

1 **3.3.2 Snow Cover**

2 Lower surface air temperatures in the ALCC scenario relative to the control lead to an
3 increase in winter snow accumulation (Fig. 12). This increase is seen by 5kaBP mostly in
4 northern and mountainous regions. The areas affected gradually increase so that by 3kaBP
5 more temperate and lower lying areas see increases in snow depth. The effects are most
6 pronounced in North America and Europe. In regions outside the areas of permanent snow
7 cover increases in snow depth will delay the melting of the snow pack and thus result in a
8 longer period of snow cover. The increased snow cover due to the cold temperature anomalies
9 will cause additional cooling due to the increased albedo. This will be greatest in regions of
10 deforestation where the snow-covered ground is no longer masked by the canopy of the
11 forest. This increased snow cover would also lead to decreases in precipitation due to lower
12 rates of moisture recycling over land. It should be noted that the modelling of snow fall is
13 subject to the same uncertainties as described in Sect. 3.3.

14

15 **4 Temporal Evolution of Holocene Climate**

16 In the control experiment the changes in orbital configuration, greenhouse gases(GHG), and
17 icesheets/sea level lead to monotonically increasing global temperatures through the Holocene
18 (Fig. 13a). Analysis of previous experiments to assess the sensitivity to different natural
19 forcings (data from Singarayer and Valdes, 2010; Singarayer et al., 2011) suggest that while
20 the changes to orbital configuration effect a cooling in global temperature over the Holocene,
21 this is outweighed by increases in greenhouse gases (~17ppm CO₂ from 8kaBP to late pre-
22 industrial time), which result in overall warming. Cooling through the Holocene occurs in
23 northern hemisphere summer, when forced with orbit and GHG variations, but is not as
24 pronounced as when only forced by orbital variations. In winter, when HadCM3 is forced
25 with orbit-only variation there is little change in temperature, but when GHG increases are
26 included this becomes a warming over the Holocene, which then outweighs the reduced
27 summer cooling. Whilst this contrasts with recent data compilations that suggest a general
28 decline in global temperatures since the mid-Holocene (Marcott et al., 2013) it is within the
29 range of other climate model responses when compared with the Paleoclimate Model
30 Intercomparison Project 3 (PMIP3) Mid-Holocene (MH) minus late Pre-Industrial (PI)
31 temperature anomalies. Although palaeodata syntheses may suggest a cooling of northern
32 hemisphere temperatures, there are regional and seasonal variations in the data such as that

1 from the Bartlein et al.(2013) (Fig. 14c) and Mauri et al (2015) data compilations. In both
2 these compilations and the combined proxy reconstructions of Wanner et al. (2008) the
3 cooling is most evident over the higher latitude northern hemisphere.

4 For our specific configuration of HadCM3 the inclusion of land use changes through the
5 Holocene has a significant impact on the progression of modelled global average
6 temperatures, such as to alter the direction of the multi-millennial trend described above thus
7 reducing the mismatch between the model simulations and the proxy reconstructions which
8 show an extra-tropical cooling trend throughout the Holocene. The increasing magnitude and
9 spread of ALCC through the Holocene reconstructed in KK10 counteracts the influence of
10 increasing greenhouse gases, so that temperatures are effectively steady from 3kaBP in
11 HadCM3 (Fig. 14a).

12 These global trends are composed of considerable heterogeneity at the regional scale. When
13 broken down into zonal regions it can be clearly seen that the difference in trends
14 with/without land use is greatest in the Northern extratropics (Fig. 13b), where the Holocene
15 trend is modified from increasing temperatures to decreasing temperatures in the late
16 Holocene by the addition of ALCC. There are impacts on mean temperatures in the tropics
17 (Fig. 13c) and southern extratropics (Fig. 13d) but not sufficient to influence the direction of
18 the Holocene trend.

19 MH minus PI anomalies of annual mean surface air temperature (Fig. 14a) show near global
20 distribution of cooling, except over high latitude sea-ice regions, which are particularly
21 influenced by changes in obliquity (higher in the MH than PI). The PMIP3 suite of models
22 show a similar pattern of surface air temperature anomalies. The cooling is most dramatic
23 over the tropics and monsoonal regions, where changes in the seasonality of insolation (due to
24 orbital precession variation) intensify monsoon circulation in the early and mid-Holocene and
25 the resulting additional cloud cover reduces incoming shortwave radiation as well as
26 increased surface water altering the balance of sensible to latent heat fluxes. The inclusion of
27 land use change in the KK10 experiment reduces the magnitude of MH cooling, especially in
28 the mid-latitudes. Over Europe and eastern North America the anomaly is reversed to a
29 warming (i.e. over these regions the cooling from deforestation shown in Fig. 2, increases and
30 outstrips the warming from greenhouse gases). These are also regions where there is the
31 highest concentration of pollen reconstructions (Fig. 14c; Bartlein et al., 2011; Mauri et al.,
32 2014). The influence of land use change improves the data-model comparison with the

1 reconstruction by Bartlein et al. (2011) over these key areas (Fig. 14a-c). Likewise, when
2 simulated top-level ocean temperatures are compared with SST data used within the Marcott
3 et al. (2013) compilation, the inclusion of land use improves the data-model comparison (Fig.
4 14d and e). However, the largest MH warming in the model is in the summer months,
5 whereas, recent seasonal temperature reconstructions (also using pollen; Mauri et al., 2014)
6 suggest the largest and most widespread MH warming may have occurred in winter in the
7 MH.

8 In contrast, using the KK10 ALCC scenario as a boundary condition to the climate model
9 does not improve the agreement in annual mean MH - PI precipitation anomalies when
10 compared to the palaeoclimate reconstruction of Bartlein et al. (2011) (Fig. 15). Model and
11 data are in reasonable agreement in most regions except for Europe and the temperate regions
12 of Asia where the data implies a wetter MH than PI, which is not seen in the model runs. The
13 difference over Europe is exacerbated by the inclusion of land use, which results in a drier
14 MH.

15

16 **5 Discussion**

17 Anthropogenic land cover change leads to climate change well beyond the core regions of
18 land use early in the Holocene. These results suggest that regional ALCC has an effect on the
19 atmospheric circulation, e.g. the ITCZ shift is a remote response on global scale. The
20 implications of this finding are that regional models or atmospheric-only models would not
21 simulate these atmospheric circulation changes as well as a global coupled model. In this
22 study we observed multiple thermal anomalies (from intense regions of cooling directly over
23 anthropogenic land use change), but the standing wave response of the geopotential height
24 field would likely also be seen even for a single thermal source from just one region (Hoskins
25 and Karoly, 1981). Regional models have the advantages of higher spatial resolution and
26 more detailed orography but they may not include these potential remote atmospheric changes
27 possibly resulting in different impacts from the same land cover forcing in regional and global
28 model simulations. The positioning of the major temperature anomalies in the mid-latitudes
29 and at similar latitudes may be particularly significant in producing the stationary wave
30 pattern.

31 Whilst these are the results from only one model there are many similarities in the distribution
32 of the temperature anomalies with those found by He et al. (2014) for 1850 CE and Pongratz

1 et al. (2010) for the 20th century although the temperature changes found in this study were
2 greater e.g. a pre-industrial global annual mean temperature anomaly of -0.23°C as opposed to
3 the -0.17°C estimated by He et al. (2014). Running similar simulations with a greater number
4 of models would improve the robustness of the results particularly with respect to
5 hydroclimate due to the high uncertainties involved. The variability in the results from
6 different models can be greater than the variability of the property that is being assessed
7 (Pitman et al., 2009; de Noblet-Ducoudre et al., 2012; Brovkin et al., 2013). These
8 inconsistencies have been attributed to disagreements in how land use change is implemented,
9 the parameterisation of albedo, the representation of crop phenology and evapotranspiration
10 and the partitioning of available energy between latent and sensible heat fluxes (Pitman et al.,
11 2009; de Noblet-Ducoudre et al., 2012; Boisier et al., 2012). The albedo and turbulent heat
12 fluxes from our model simulations for the North America/Eurasia region (Fig. 16) are
13 within the range of other climate model responses when compared with those from the Land-
14 Use and Climate, IDentification of robust impacts (LUCID) set of experiments (Boisier,
15 2012). The negative turbulent and latent heat fluxes would offset some of the cooling due to
16 the increased albedo. Although the largest albedo changes are in the DJF season the impact of
17 this will be lessened due to the lower levels of incoming solar radiation in this season.
18 Results could be further improved by the running of transient simulations that could capture
19 events such as the Maunder minimum. A transient simulation response could result in
20 different biogeophysical impacts to the ones achieved when using a long equilibrium
21 simulation where the ocean-land-atmosphere system can reach more of a steady state and the
22 climate sensitivity is different.

23 From late preindustrial era simulations, one using observed atmospheric greenhouse gas
24 concentrations and the other using greenhouse gas concentrations in a world with no
25 anthropogenic emissions (based on linear projection from earlier Holocene trends from
26 Kutzbach et al., 2011), He et al. (2014) estimated a net global warming of 0.9°C due to the
27 biogeochemical effects of ALCC, with between 0.5 and 1.5°C warming in the areas of most
28 intense land use changes. Incorporating this degree of warming into our early industrial era
29 (1850 CE) simulations there would still be a net cooling in Europe, E. Asia and N.E. America
30 with, e.g., a net cooling of up to 2°C in parts of Europe. To put this in perspective the IPCC
31 (IPCC WG II, 2014a) consider a temperature rise of more than 2°C to be undesirable and that
32 changes of 1°C could have an impact on vulnerable ecosystems. However, the temperature
33 changes in this study took place over a much longer period than the timeframe considered in

1 the IPCC and ecosystems and human societies would have had more time to adapt. The
2 consequences of these changes for agricultural societies would vary depending on the pre-
3 existing conditions. For example, in drier regions, where crops are more likely to be water
4 limited, cooler, wetter summer conditions may have been beneficial to the agricultural output
5 although the risk of erosion would be increased. The generally lower temperatures might also
6 make societies more vulnerable to further transient cooling effects such as volcanic activity.

7 There are discrepancies between simulations of the mid-Holocene climate and the
8 independent data-based reconstructions. Both the simulations from this study and virtually all
9 the PMIP3 (Palaeoclimate Model Intercomparison 3; <https://pmip3.lsce.ipsl.fr>) models show a
10 temperature increase from the mid-Holocene to the PI whereas the Marcott et al. (2013) (and
11 Mann et al. (2008) on a shorter timescale) reconstructions show a decrease. It is interesting to
12 note that ALCC reduces this mismatch for HadCM3, especially in key areas such as Europe.
13 Other factors that could lead to this discrepancy are uncertainties in the proxy reconstructions
14 and deficiencies in climate models (e.g Lohmann et al., 2013). These climate model
15 deficiencies include low resolution and sensitivity and, importantly, their dependence on soil
16 moisture whereby energy is utilised for evaporation rather than for temperature increase.

17 This study shows a significant increase in precipitation over Europe with increasing land use
18 which means that the PI becomes wetter than the mid-Holocene, which leads to increases in
19 soil moisture and changes in the sensible to latent heat flux balance and, in combination with
20 increased albedo caused by deforestation, this results in cooler temperatures for PI than MH.
21 If the land-atmosphere coupling strength was different and soil moisture was strongly reduced
22 with deforestation it is likely that the cooling effect would be smaller (cf. Strandberg, 2014).

23 Further uncertainties arise from the robustness of the land use reconstructions, which is
24 difficult to evaluate due to the lack of global-scale evidence for human impact on the Earth's
25 land surface. Much of the uncertainty comes from the lack of knowledge about the magnitude
26 and distribution of the global human population and the rate of technological evolution and
27 intensification through time. As part of our initial investigations simulations were also run
28 using an alternative land use scenario (the HYDE 3.1 dataset; Goldewijk et al., 2011). The
29 HYDE 3.1 reconstruction has substantially lower levels of land use early in the Holocene (as
30 compared with KK10), which resulted in a later development of consistent temperature
31 anomalies at 4kaBP (not shown) in comparison with the KK10 land use scenario. The
32 decision was taken to proceed with the KK10 data due to its assumptions of a larger per capita

1 land use earlier in the Holocene when agricultural methods were less efficient. Several
2 ongoing international initiatives that aim to synthesise palaeoecological and archaeological
3 data promise to lead to more robust reconstructions of Holocene ALCC in the future (e.g.
4 PAGES LandCover6k project; <http://www.pages-igbp.org/workinggroups/landcover6k/intro>).

5 By the early industrial period simulated biogeophysical temperature changes in the regions of
6 land disturbance are of the same order of magnitude (e.g. 0.83°C annual anomaly in the main
7 agricultural areas of Europe) as the changes seen due to CO₂ increases during the industrial
8 period (0.85°C, IPCC synthesis report, 2014b). Part of Ruddiman's original hypothesis
9 (Ruddiman, 2003) is that pre-industrial global warming caused by anthropogenic CO₂/CH₄
10 emissions should have been ~2 °C at higher latitudes, but there was no evidence for this
11 warming. Ruddiman (2003) attributed this to a natural cooling trend caused by decreasing
12 summer insolation. This study suggests that biogeophysical effects of the land use changes
13 may also have played a part in counteracting the warming due to anthropogenic greenhouse
14 gas emissions as acknowledged in Ruddiman (2013). The precipitation changes might also
15 have an impact on the availability of water for rice irrigation and on natural wetlands thus
16 affecting the production of methane.

17

18 **6 Conclusions**

19 In our global model simulations that use a Holocene ALCC scenario as a boundary condition,
20 a surface temperature response to the biogeophysical effects of ALCC is seen in regions of
21 early land use such as Europe and S.E. Asia as early as 7kaBP in the JJA season and
22 throughout the entire annual cycle by 2-3kaBP. Areas outside the major regions of ALCC are
23 also affected, with virtually the whole globe experiencing significant temperature changes
24 with a net global cooling of 0.22°C by the pre-industrial period. Although the temperature
25 changes are predominantly cooling some regions such as India, Southern Africa and Siberia
26 show warming as a response to ALCC. The greatest changes are generally seen in the JJA
27 season with a mean regional cooling of 1.4°C experienced in Europe and 1°C in E. Asia in the
28 early industrial period (1850 CE). Much of the precipitation response to the land use tends to
29 be due to large-scale circulation changes such as a decrease in the intensity of the Indian
30 monsoon, the southward movement of the ITCZ and changes to the North Atlantic storm
31 track. In Europe there is a slight decrease in precipitation in the DJF season and a more
32 substantial increase in the JJA season. Some causal factors for the teleconnections are

1 advection by surface winds, MSLP anomalies, and tropospheric stationary wave train
2 disturbances in the mid- to high-latitudes.

3 The potential for an early global impact of ALCC on climate strongly implied by this study
4 suggests that due consideration of this should be taken in simulations covering the Holocene.
5 The inclusion of ALCC in the model improves the model comparison for surface air
6 temperature with the data-driven palaeoclimate reconstructions especially in key areas such as
7 Europe. The remote teleconnections seen in this study have implications for the regional
8 modelling of land use change due to circulation changes that occur outside the domain of the
9 regional model.

10 Overall, our model simulations indicate an increase in global surface air temperatures through
11 the Holocene. Globally, the inclusion of ALCC data reduces the magnitude of this warming
12 especially in the late Holocene when the temperatures remain relatively constant. Regionally,
13 in the Northern extratropics, this warming is reversed in the late Holocene. It should be noted
14 that in this study it is not possible to distinguish the anthropogenic component of the
15 biogeochemical changes as the same atmospheric CO₂ and CH₄ concentrations (from ice core
16 measurements) are prescribed for both the KK10 and control simulations. However, the level
17 of early industrial warming due to the biogeochemical impacts of ALCC predicted by He et
18 al. (2014) would negate the early industrial biogeophysical cooling seen in this study in all
19 regions except for the most intensively altered landscapes of Europe, E. Asia and N.E.
20 America.

21 Other caveats are the large uncertainties in the land use data and, therefore, in our
22 understanding of the Holocene evolution of land surface-climate interactions as well as our
23 ability to evaluate climate models. To reduce these uncertainties there is an urgent need to
24 extend land cover reconstructions and prehistory of land use globally (cf. LandCover6k
25 PAGES initiative). It must also be stressed that these are the results from only one model and
26 the use of equilibrium simulations may produce different results from a transient simulation.
27 However, these uncertainties are more likely to impact the timing, intensity and regional
28 details of the changes rather than the overall trends and patterns.

29

30 **Data Availability**

1 Data is available from the Bristol Research Initiative for the Dynamic Global Environment
2 website: <http://www.bridge.bris.ac.uk/resources/simulations>

3

4 **Acknowledgements**

5 MCS is supported by a University of Reading doctoral scholarship. MCS and JSS would like
6 to thank Sandy Harrison and Beni Stocker for useful discussions on land cover
7 reconstructions. JOK was supported by the European Research Council (313797
8 COEVOLVE)

9

1 **References**

- 2 Barnston, A., Livezy, R., and Halpert, M.: Modulation of southern oscillation northern-
3 hemisphere midwinter climate relationships by the QBO. *J. Climate*, 4, 203-217, 1991.
- 4 Bartlein, P.J., Harrison, S.P., Brewer, S., Connor, S., Davis, B.A.S., Gajewski, K., Guiot, J.,
5 Harrison-Prentice, T.I., Henderson, A., Peyron, O., Prentice, I.C., Scholze, M., Seppä, H.,
6 Shuman, B., Sugita, S., Thompson, R.S., Viau, A.E., Williams, J. and Wu, H.: Pollen-based
7 continental climate reconstructions at 6 and 21 ka: a global synthesis, *Clim. Dynam.*, 37, 775-
8 802, doi:10.1007/s00382-010-0904-1, 2011.
- 9 Berger, A., and Loutre, M. F.: Insolation values for the climate of the last 10 million years,
10 *Quaternary Sci. Rev.*, 10, 297-317, doi:10.1016/0277-3791(91)90033-Q, 1991.
- 11 Betts, R. A., Falloon, P. D., Goldewijk, K. K., and Ramankutty, N.: Biogeophysical effects of
12 land use on climate: Model simulations of radiative forcing and large-scale temperature
13 change, *Agr. Forest Meteorol.*, 142, 216-233, doi:10.1016/j.agrformet.2006.08.021, 2007.
- 14 Boisier, J. P., de Noblet-Ducoudré, N., Pitman, A.J., Cruz, F.T., Delire, C., den Hurk,
15 B.J.J.M., Molen, M.K., Müller, C. and Voltaire, A.: Attributing the impacts of land-cover
16 changes in temperate regions on surface temperature and heat fluxes to specific causes:
17 Results from the first LUCID set of simulations, *J. Geophys. Res.-Atmos.*, 117,
18 doi:http://dx.doi.org/10.1029/2011JD017106, 2012 Broecker, W.S. and Stocker, T.F.: The
19 Holocene CO₂ rise: Anthropogenic or natural?, *Eos Trans. AGU*, 87, 27–27,
20 doi:10.1029/2006EO030002, 2006.
- 21 Brovkin, V., Boysen, L., Arora, V. K., Boisier, J. P., Cadule, P., Chini, L., Claussen, M.,
22 Friedlingstein, P., Gayler, V., Hurk, v.d., B.J.J.M, Hurtt, G.C., Jones, C.D., Kato, E., Noblet-
23 Ducoudré, d., N., Pacifico, F., Pongratz, J. and Weiss, M.: Effect of anthropogenic land-use
24 and land-cover changes on climate and land carbon storage in CMIP5 projections for the
25 twenty-first century. *J. Climate*, 26, 6859, 2013.
- 26 Collins, W.J., Bellouin, N., Doutriaux-Boucher, M., Gedney, N., Halloran, P., Hinton, T.,
27 Hughes, J., Jones, C.D., Joshi, M., Liddicoat, S., Martin, G., O'Connor, F., Rae, J., Senior, C.,
28 Sitch, S., Totterdell, I., Wiltshire, A. and Woodward, S.: Development and evaluation of an
29 Earth-system model – HadGEM2, *Geosci. Model Dev.*, 4, 1051-1075, doi:10.5194/gmdd-4-
30 1051-2011, 2011.

1 Cook, B. I., Anchukaitis, K. J., Kaplan, J. O., Puma, M. J., Kelley, M., and Gueyffier, D.:
2 Pre-Columbian deforestation as an amplifier of drought in Mesoamerica, *Geophys. Res. Lett.*,
3 39, L16706, doi:10.1029/2012GL052565, 2012.

4 Covey, C., AchutaRao, K.M., Cubasch, U., Jones, P., Lambert, S.J., Mann, M.E., Phillips,
5 T.J. and Taylor, K.E.: An overview of results from the Coupled Model Intercomparison
6 Project, *Global Planet. Change*, 37, 103-133, doi:10.1016/S0921-8181(02)00193-5, 2003.

7 Cox, P.M.: Description of the TRIFFID dynamic global vegetation model. Cox, P. M., Tech.
8 Note 24, Hadley Centre, Met Office, 16 pp, available at <http://www.metoffice.gov.uk>, (last
9 access 15 September 2015), 2001.

10 Darby, H.C.: The clearing of the woodland in Europe. In: *Man's Role in Changing the Face of*
11 *the Earth*, edited by Thomas Jr., W.L., University of Chicago Press, Chicago, 183-216, 1956.

12 de Noblet-Ducoudré, N., Pitman, A., Delire, C., Hurk, v. d., B.J.J.M., Boisier, J. P., Brovkin,
13 V., Cruz, F., Voldoire, A., Molen, v.d., M.K., Müller, C., Bonan, G.B., Gayler, V., Reick,
14 C.H., Strengers, B.J. and Lawrence, P. J.: Determining robust impacts of land-use induced
15 land-cover changes on surface climate over North America and Eurasia; results from the first
16 set of LUCID experiments. *J. Climate*, 25, 3261-3281, 2012.

17 Essery, R. L. H., Best, M. J., Betts, R. A., Cox, P. M., and Taylor, C. M.: Explicit
18 representation of subgrid heterogeneity in a GCM land surface scheme, *J. Hydrometeorol.*, 4,
19 530-543, doi:10.1175/1525-7541(2003)004<0530:EROSHI>2.0.CO;2, 2003.

20 Folland, C.K., Knight, J., Linderholm, H.W., Fereday, D., Ineson, S., and Hurrell, J.W.: The
21 Summer North Atlantic Oscillation: Past, Present, and Future, *J. Climate*, 22, 1082-1103.
22 doi:10.1175/2008JCLI2459.1, 2009.

23 Goldewijk, K. K., Beusen, A., Van Drecht, G., and De Vos, M.: The HYDE 3.1 spatially
24 explicit database of human-induced global land-use change over the past 12,000 years, *Global*
25 *Ecol. Biogeogr.*, 20, 73-86, doi:10.1111/j.1466-8238.2010.00587, 2011.

26 Gordon, C., Cooper, C., Senior, C.A., Banks, H., Gregory, J.M., Johns, T.C., Mitchell, J.F.B.
27 and Wood, R.A.: The simulation of SST, sea ice extents and ocean heat transports in a version
28 of the Hadley Centre coupled model without flux adjustments, *Clim. Dynam.*, 16, 147-168,
29 doi:10.1007/s003820050010, 2000.

1 He, F., Vavrus, S. J., Kutzbach, J. E., Ruddiman, W. F., Kaplan, J. O. and Krumhardt, K. M.:
2 Simulating global and local surface temperature changes due to Holocene anthropogenic land
3 cover change, *Geophys. Res. Lett.*, 41, 623-631, 2014.

4 Hoskins, B. J., and Karoly, D. J.: The steady linear response of a spherical atmosphere to
5 thermal and orographic forcing, *J. Atmos. Sci.*, 38, 1179-1196, 1981.

6 Hurrell, J. W. : Decadal trends in the north Atlantic oscillation: Regional temperatures and
7 precipitation. *Science*, 269, 676-679. doi:10.1126/science.269.5224.676, 1995.

8 Hurrell, J. W., Kushnir, Y., Ottersen, G., and Visbeck, M.: An Overview of the North Atlantic
9 Oscillation, in *The North Atlantic Oscillation: Climatic Significance and Environmental*
10 *Impact* edited by Hurrell, J. W., Kushnir, Y., Ottersen, G., and Visbeck, M., Am. Geophys.
11 Union, Washington, D. C., doi: 10.1029/134GM01, 2003 Intergovernmental Panel on Climate
12 Change (IPCC): *Impacts, Adaptation and Vulnerability – Contribution of Working Group II*
13 *to the Third Assessment Report of IPCC*, Cambridge University Press, 2001.

14 Intergovernmental Panel on Climate Change (IPCC): *The Physical Science Basis –*
15 *Contribution of Working Group I to the Fourth Assessment Report of the IPCC*, Cambridge
16 University Press, 2007.

17 IPCC, 2014: *Climate Change 2014: Impacts, Adaptation, and Vulnerability. Part A: Global*
18 *and Sectoral Aspects. Contribution of Working Group II to the Fifth Assessment Report of*
19 *the Intergovernmental Panel on Climate Change* edited by: Field, C.B., Barros, V.R.,
20 Dokken, D.J., Mach, K.J., Mastrandrea, M.D., Bilir, T.E., Chatterjee, M., Ebi, K.L., Estrada,
21 Y.O., Genova, R.C, Girma, B., Kissel, E.S., Levy, A.N., MacCracken, S., Mastrandrea, P.R.,
22 and White, L.L., Cambridge University Press, Cambridge, United Kingdom and New York,
23 NY, USA, 1132 pp., 2014a.

24 IPCC, 2014: *Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II*
25 *and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*,
26 edited by Core Writing Team, Pachauri, R.K. and Meyer, L.A. , IPCC, Geneva, Switzerland,
27 151 pp., 2014b.

28 Joos, F., Gerber, S., Prentice, I. C., Otto-Bliesner, B. L., and Valdes, P. J.: Transient
29 simulations of Holocene atmospheric carbon dioxide and terrestrial carbon since the last
30 glacial maximum, *Global Biogeochem. Cy.*, 18, GB2002. doi:10.1029/2003GB002156, 2004.

1 Kang, S. M., Held, I. M., Frierson, D. M., and Zhao, M.: The response of the ITCZ to
2 extratropical thermal forcing: Idealized slab-ocean experiments with a GCM, *J. Climate*, 21,
3 3521-3532, 2008.

4 Kaplan, J. O., Krumhardt, K. M., and Zimmermann, N.: The prehistoric and preindustrial
5 deforestation of Europe, *Quaternary Sci. Rev.*, 28, 3016-3034.
6 doi:10.1016/j.quascirev.2009.09.028, 2009.

7 Kaplan, J. O., Krumhardt, K. M., Ellis, E. C., Ruddiman, W. F., Lemmen, C., and Goldewijk,
8 K. K.: Holocene carbon emissions as a result of anthropogenic land cover change, *Holocene*,
9 21, 775-791, doi:10.1177/0959683610386983, 2011.

10 Kutzbach, J. E., Vavrus, S. J., Ruddiman, W. F., and Philippon-Berthier, G.: Comparisons of
11 atmosphere–ocean simulations of greenhouse gas-induced climate change for pre-industrial
12 and hypothetical ‘no-anthropogenic’ radiative forcing, relative to present day, *Holocene*, 21,
13 793-801, doi:10.1177/0959683611400200, 2011.

14 Lean, J., and Rowntree, P. R.: A GCM simulation of the impact of Amazonian deforestation
15 on climate using an improved canopy representation, *Q. J. Roy. Meteor. Soc.*, 119, 509-530.
16 doi:10.1002/qj.49711951109, 1993.

17 Lohmann, G., Pfeiffer, M., Laepple, T., Leduc, G. and Kim, J.-H., : A model-data comparison
18 of the Holocene global sea surface temperature evolution. *Clim. Past*, 9, 1807-1839,
19 doi:10.5194/cp-9-1807-2013, 2013.

20 Louergue, L., Schilt, A., Spahni, R., Masson-Delmotte, V., Blunier, T., Lemieux, B.,
21 Barnola, J., Raynaud, D., Stocker, T.F. and Chappellaz, J.: Orbital and millennial-scale
22 features of atmospheric CH₄ over the past 800000 years, *Nature*, 453, 383-386,
23 doi:10.1038/nature06950, 2008.

24 Mann, M. E., Zhang, Z., Hughes, M. K., Bradley, R. S., Miller, S. K., Rutherford, S., and Ni,
25 F.: Proxy-based reconstructions of hemispheric and global surface temperature variations over
26 the past two millennia, *P. Natl. Acad. Sci. USA*, 105, 13252-13257,
27 doi:10.1073/pnas.0805721105, 2008.

28 Marcott, S. A., Shakun, J. D., Clark, P. U., and Mix, A. C.: A reconstruction of regional and
29 global temperature for the past 11300 years, *Science*, 339, 1198-1201,
30 doi:10.1126/science.1228026, 2013.

1 Mauri, A., Davis, B. A. S., Collins, P. M., and Kaplan, J. O.: The climate of Europe during
2 the Holocene: A gridded pollen-based reconstruction and its multi-proxy evaluation,
3 *Quaternary Sci. Rev.*, 112, 109-127, doi:10.1016/j.quascirev.2015.01.013, 2015.

4 Mitchell, L., Brook, E., Lee, J., Buizert, C. and Sowers, T.: Constraints on the late Holocene
5 anthropogenic contribution to the atmospheric methane budget, *Science*, 342, 964-966.
6 doi:10.1126/science.1238920, 2013.

7 Oglesby, R. J., Sever, T. L., Saturno, W., Erickson III, D. J. and Srikishen, J.: Collapse of
8 the Maya: Could deforestation have contributed?, *J. Geophys. Res.*, 115, D12106,
9 doi:10.1029/2009JD011942, 2010.

10 Peltier, W. R.: Global glacial isostasy and the surface of the ice-age Earth: the ICE-5G (VM2)
11 model and GRACE, *Annu. Rev. Earth Pl. Sc.*, 32, 111-149,
12 doi:10.1146/annurev.earth.32.082503.144359, 2004.

13 Petit, J. R., Lipenkov, V. Y., Kotlyakov, V. M., Barkov, N. I., Bender, M., Davis, M.,
14 Delaygue, G., Barnola, J.-., Stievenard, M., Lorius, C., Delmotte, M., Pepin, L., Chappellaz,
15 J., Saltzman, E., Legrand, M., Basile, I., Jouzel, J., Raynaud, D. and Ritz, C.: Climate and
16 atmospheric history of the past 420000 years from the Vostok ice core, Antarctica, *Nature*,
17 399, 429-436. doi:10.1038/20859, 1999.

18 Piperno, D. R., Ranere, A. J., Holst, I., Iriarte, J., and Dickau, R.: Starch grain and phytolith
19 evidence for early ninth millennium BP maize from the Central Balsas River Valley,
20 Mexico, *P. Natl. Acad. Sci.*, 106, 5019-5024, doi: 10.1073/pnas.0812525106, 2009.

21 Pitman, A. J., de Noblet-Ducoudré, N., Cruz, F. T., Davin, E. L., Bonan, G. B., Brovkin, V.,
22 Claussen, M., Delire, C., Ganzeveld, L., Gayler, V., Hurk, v.d., B. J. J. M., Lawrence, P.J.,
23 Molen, v.d., M. K., Müller, C., Reick, C.H., Seneviratne, S.I., Strengers, B.J. and Voldoire, A.:
24 Uncertainties in climate responses to past land cover change: First results from the LUCID
25 intercomparison study, *Geophys. Res. Lett.*, 36, L14814. doi:10.1029/2009GL039076, 2009.

26 Pongratz, J., Reick, C., Raddatz, T., and Claussen, M.: A reconstruction of global agricultural
27 areas and land cover for the last millennium, *Global Biogeochem. Cy.*, 22, GB3018.
28 doi:10.1029/2007GB003153, 2008.

29 Pongratz, J., Reick, C. H., Raddatz, T., and Claussen, M.: Biogeophysical versus
30 biogeochemical climate response to historical anthropogenic land cover change, *Geophys.*
31 *Res. Lett.*, 37, L08702, doi:10.1029/2010GL043010, 2010.

1 Pope, V. D., Gallani, M. L., Rowntree, P. R., and Stratton, R. A.: The impact of new physical
2 parametrizations in the Hadley centre climate model: HadAM3, *Clim. Dynam.*, 16, 123-146.
3 doi:10.1007/s003820050009, 2000.

4 Roberts, N.: *The Holocene: an Environmental History*, John Wiley and Sons, 2013.

5 Ruddiman, W. F.: The anthropogenic greenhouse era began thousands of years ago, *Climatic*
6 *Change*, 61, 261-293. doi:10.1023/B:CLIM.00000004577.17928.fa, 2003.

7 Ruddiman, W. F.: The early anthropogenic hypothesis: Challenges and responses, *Rev.*
8 *Geophys.*, 45, RG4001. doi:10.1029/2006RG000207, 2007.

9 Ruddiman, W. F.: The Anthropocene, *Annu. Rev. Earth Pl. Sc.*, 41, 45-68.
10 doi:10.1146/annurev-earth-050212-123944, 2013.

11 Schmidt, G. A., Shindell, D. T., and Harder, S.: A note on the relationship between ice core
12 methane concentrations and insolation, *Geophys. Res. Lett.*, 31, doi:10.1029/2004GL021083,
13 2004.

14 Shepherd, T. G.: Atmospheric circulation as a source of uncertainty in climate change
15 projections, *Nat. Geosci.*, 7, 703-708, doi:10.1038/NGEO2253, 2014.

16 Singarayer, J. S., and Valdes, P. J.: High-latitude climate sensitivity to ice-sheet forcing over
17 the last 120kyr. *Quaternary Sci. Rev.*, 29, 43-55. doi:10.1016/j.quascirev.2009.10.011, 2010.

18 Singarayer, J. S., Valdes, P. J., Friedlingstein, P., Nelson, S., and Beerling, D. J.: Late
19 Holocene methane rise caused by orbitally controlled increase in tropical sources, *Nature*,
20 470, 82-85. doi:10.1038/nature09739, 2011.

21 Spahni, R., Chappellaz, J., Stocker, T.F., Loulergue, L., Hausammann, G., Kawamura, K.,
22 Flückiger, J., Schwander, J., Raynaud, D., Masson-Delmotte, V. and Jouzel, J.: Atmospheric
23 methane and nitrous oxide of the late Pleistocene from Antarctic ice cores, *Science*, 310,
24 1317-1321. doi:10.1126/science.1120132, 2005.

25 Strandberg, G., Kjellström, E., Poska, A., Wagner, S., Gaillard, M.J., Trondman, A.K., Mauri,
26 A., Davis, B.A.S., Kaplan, J.O., Birks, H.J.B., Bjune, A.E., Fyfe, R., Giesecke, T., Kalnina,
27 L., Kangur, M., van der Knaap, W.O. , Kokfelt, U., Kuneš, P., Latalowa, M., Marquer, L.,
28 Mazier, F., Nielsen, A.B., Smith, B., Seppä, H. and Sugita, S.: Regional climate model
29 simulations for Europe at 6 and 0.2 k BP: sensitivity to changes in anthropogenic
30 deforestation, *Clim. Past*, 10, 661-680. doi:10.5194/cp-10-661-2014, 2014.

1 Tauger, M.B: Agriculture in world history, Routledge, 2013.

2 Tubi, A., and Dayan, U.: The Siberian high: Teleconnections, extremes and association with
3 the Icelandic low, *Int. J. of Climatol.*, 33, 1357-1366, doi:10.1002/joc.3517, 2013.

4 Wanner, H., Beer, J., Bütikofer, J., Crowley, T.J., Cubasch, U., Flückiger, J., Goosse, H.,
5 Grosjean, M., Joos, F., Kaplan, J.O., Küttel, M., Müller, S.A., Prentice, I.C., Solomina, O.,
6 Stocker, T.F., Tarasov, P., Wagner, M. and Widmann, M.: Mid- to Late Holocene climate
7 change: an overview, *Quaternary Sci. Rev.*, 27, 1791-1828,
8 doi:10.1016/j.quascirev.2008.06.013, 2008.

9 Zohary, D., Hopf, M., and Weiss, E.: Domestication of plants in the old world: The origin and
10 spread of domesticated plants in south-west Asia, Europe, and the Mediterranean basin, 4th
11 edn., Oxford University Press, Oxford, 2012.

12

13

14

15

16

17

18

19

20

21

22

23

24

25

26

27

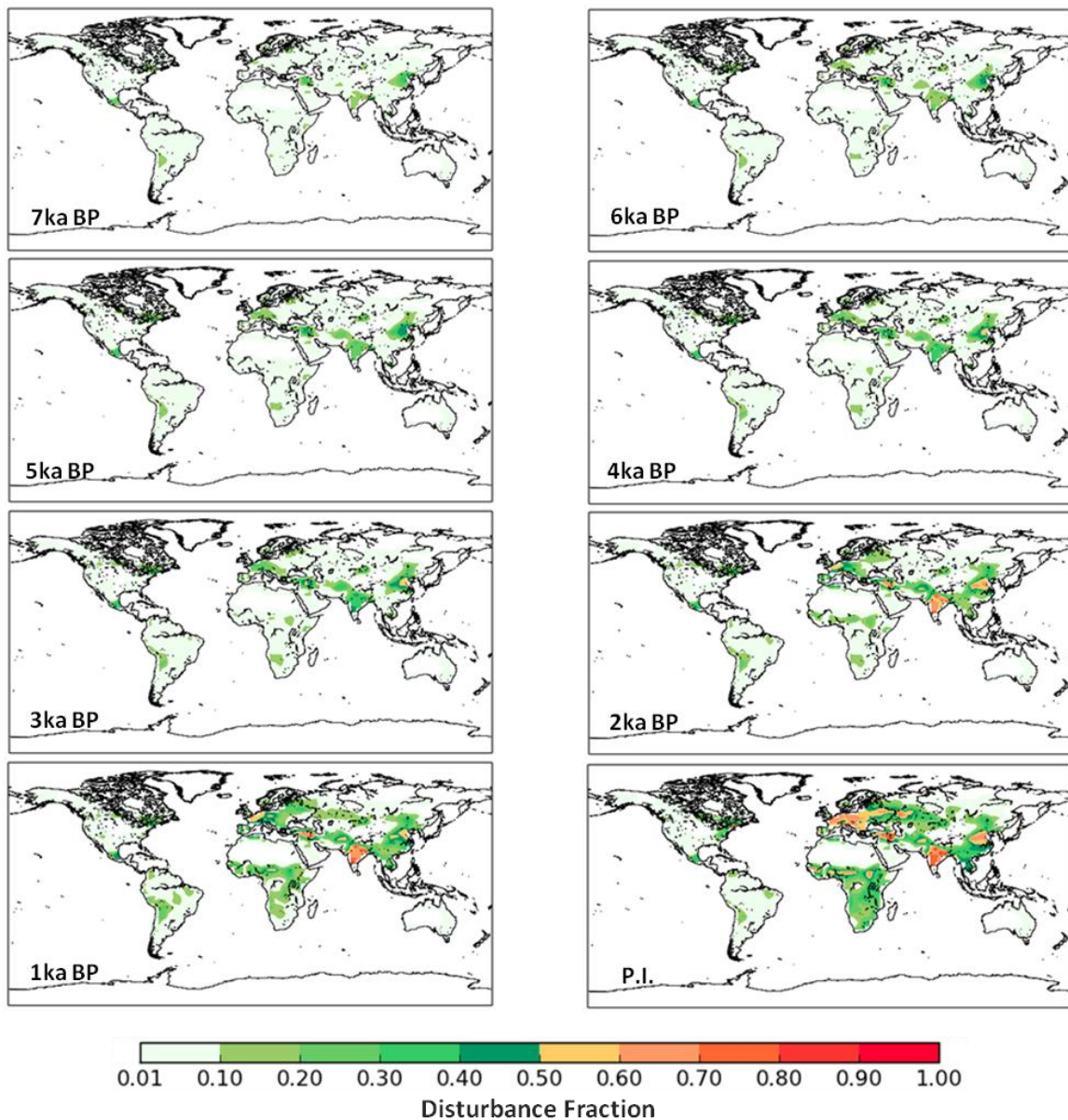
1
2
3
4

Component	Model Name	Details		
GCM	HadCM3	Description		Coupled atmospheric, ocean and sea ice model
		Atmosphere	Horizontal Resolution	2.5° latitude and 3.75° longitude.
			Vertical	19 unequally spaced layers
			Time Step	30 minutes
		Physics	Advection Scheme	Eulerian
			Chemistry	CO ₂ , N ₂ O, CH ₄ , CFC11 and CFC12
		Ocean	Horizontal Resolution	1.25° by 1.25°
			Vertical	20 unequally spaced layers extending to depth of 5200m
		Ice	Horizontal Resolution	1.25° by 1.25°
Land Surface Scheme	MOSES 2	Coupled to GCM every 10 days of model run.		
Dynamic Vegetation	TRIFFID	Nine surface types including 5 plant functional types		
Boundary conditions		Orbit (Berger and Loutre, 1991), greenhouse gases (from ice cores) and ice sheets (Peltier, 2004)		
Initial Conditions		Based on a spun-up early industrial simulation.		
Simulations	KK10	Dynamic vegetation from TRIFFID with crop mask based on KK10 land use data set (Kaplan et al., 2009 and 2011)		
	Control	Natural dynamic vegetation from TRIFFID		
Simulation dates		8 kaBP, 7 kaBP, 6 kaBP, 5 kaBP, 4kaBP, 3kaBP, 2kaBP, 1kaBP, 1850CE		
Length of simulation		1000 years		

Averaging Period		500 years
------------------	--	-----------

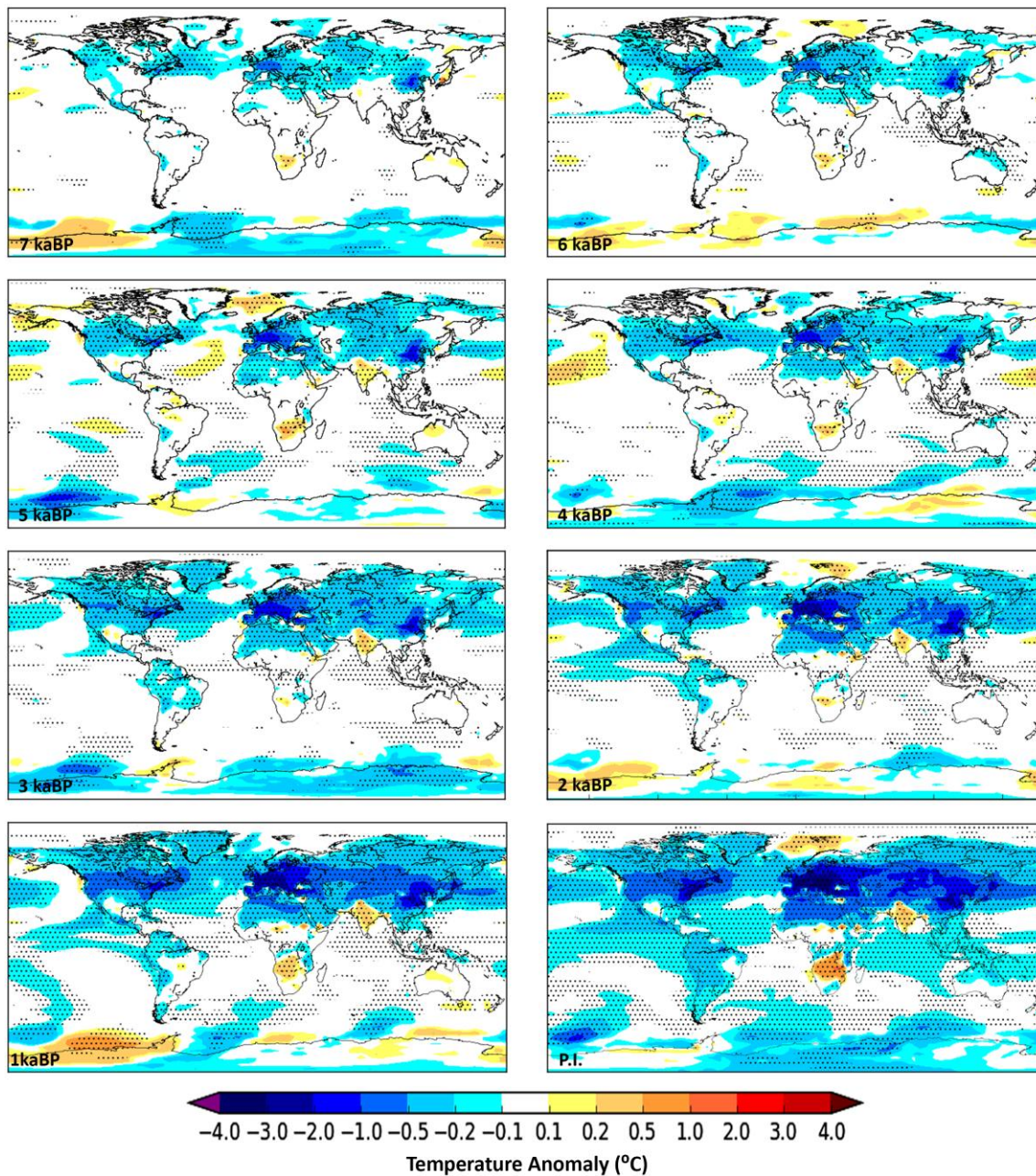
1

2 **Table 1.** Summary of the experimental set-up.



1
2
3
4
5
6
7
8
9

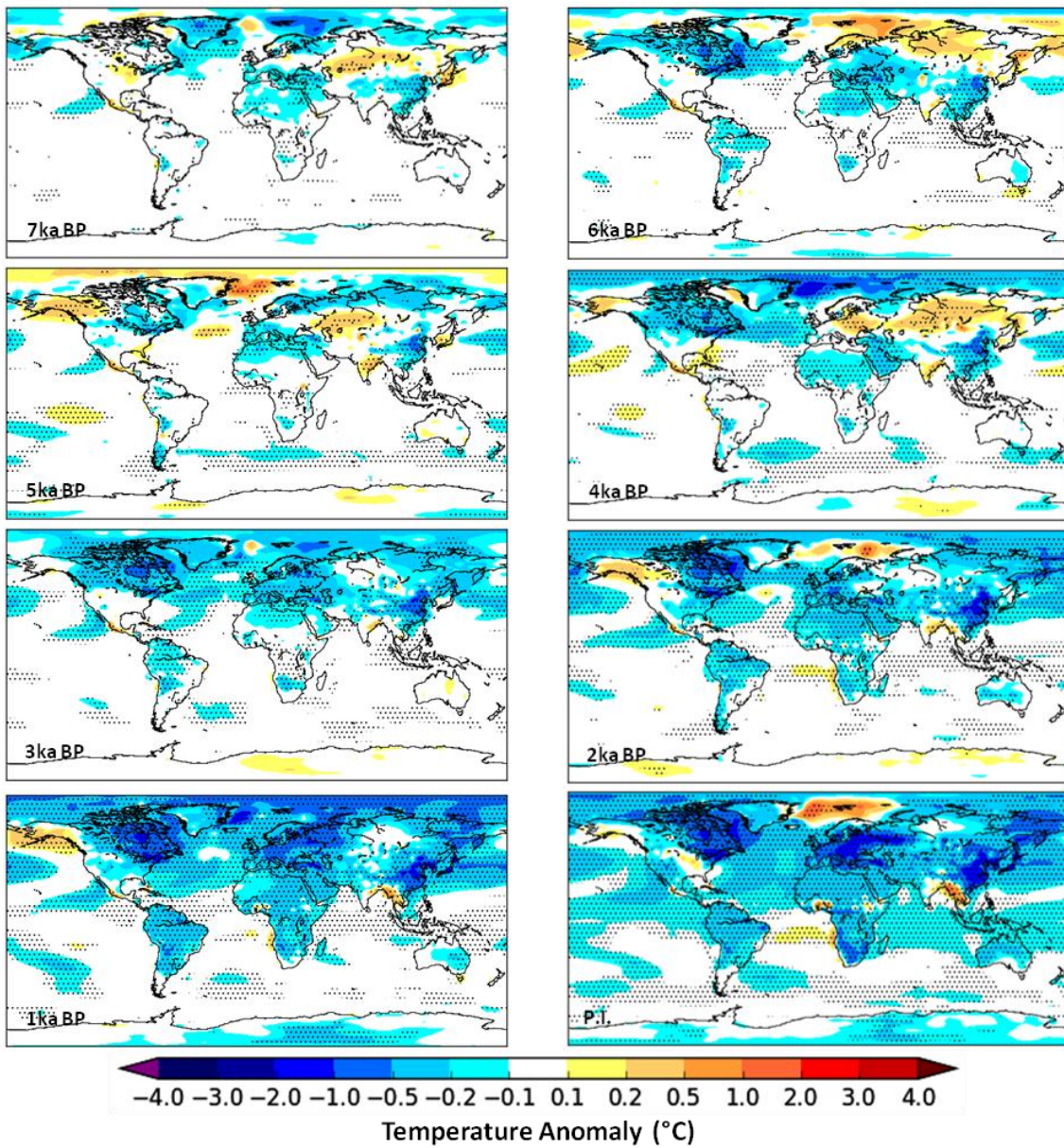
Figure 1. Fraction of anthropogenically disturbed land at 1000 year intervals from the late pre-industrial (1850 CE) period to 7kaBP. The land disturbance data is based on the anthropogenic land-use scenario KK10 (Kaplan et al., 2009, 2011).



1

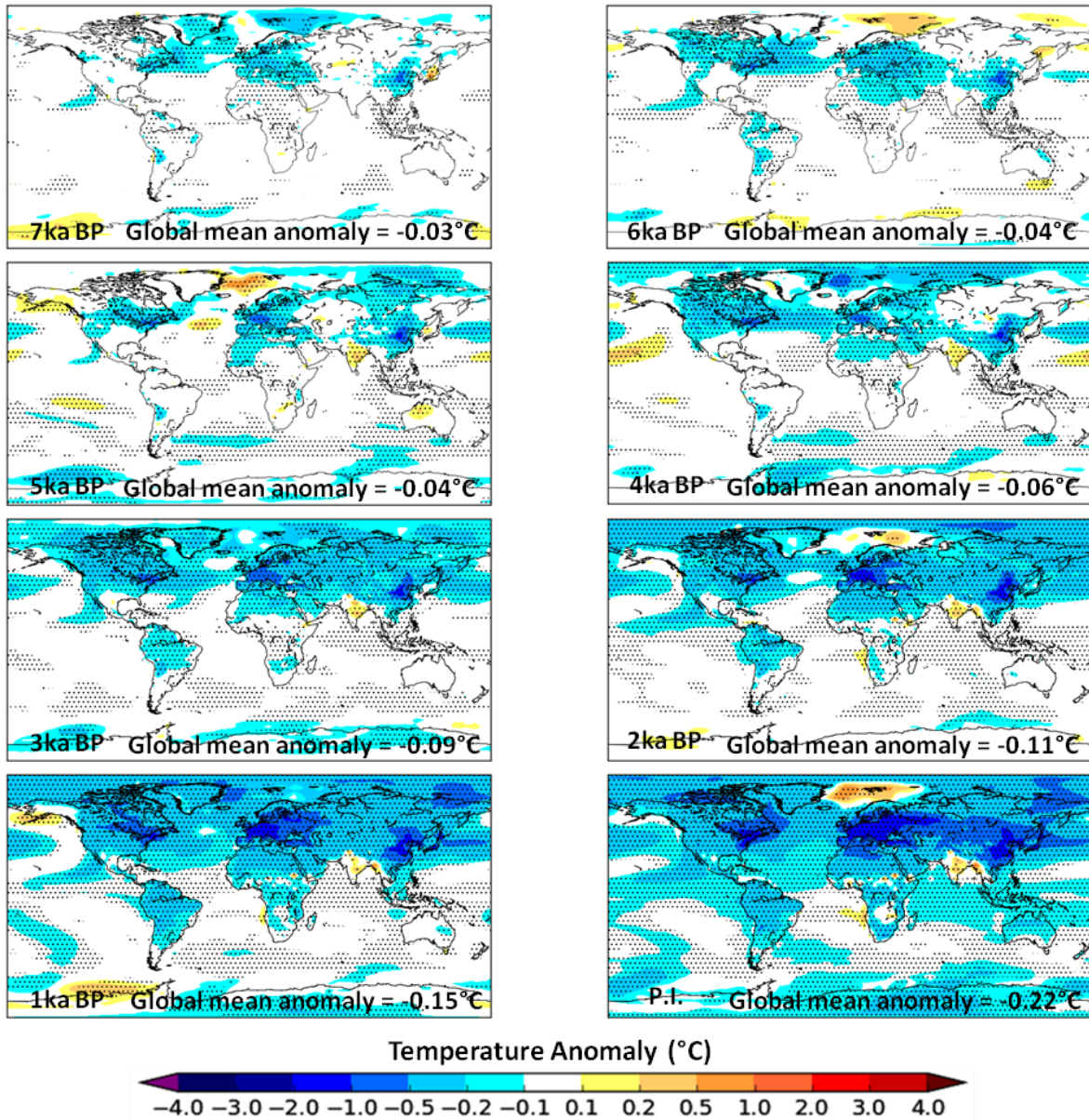
2

3 **Figure 2.** JJA Temperature Anomalies (°C) for KK10 minus Control at timeslices from the
 4 late pre-industrial period (1850 CE) to 7kaBP. The stippling indicates grid boxes where the
 5 anomalies are significant at the 95% level using Wilcoxon ranksum statistical analysis.



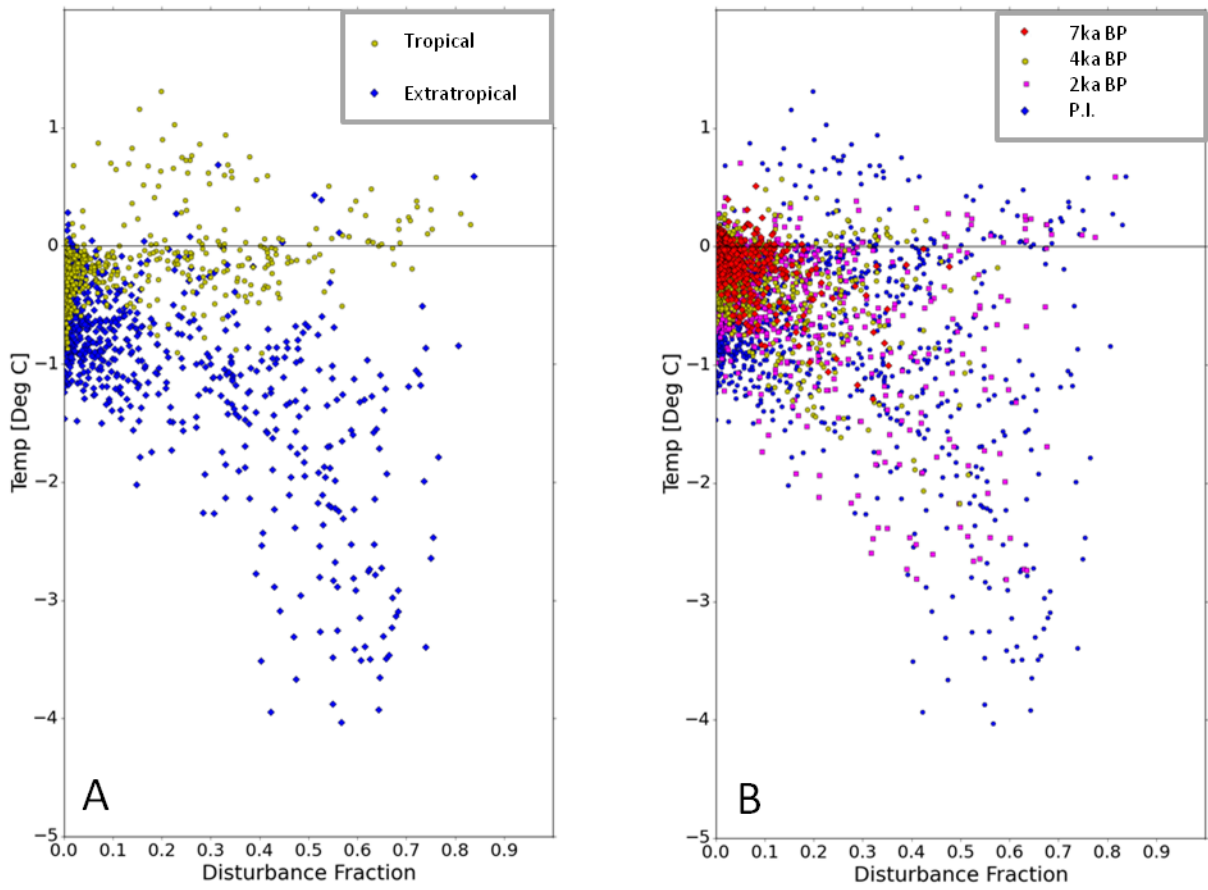
1
2
3
4
5
6

Figure 3. DJF Temperature Anomalies (°C) for KK10 minus Control at timeslices from the late pre-industrial period (1850 CE) to 7kaBP. The stippling indicates grid boxes where the anomalies are significant at the 95% level using Wilcoxon ranksum statistical analysis.



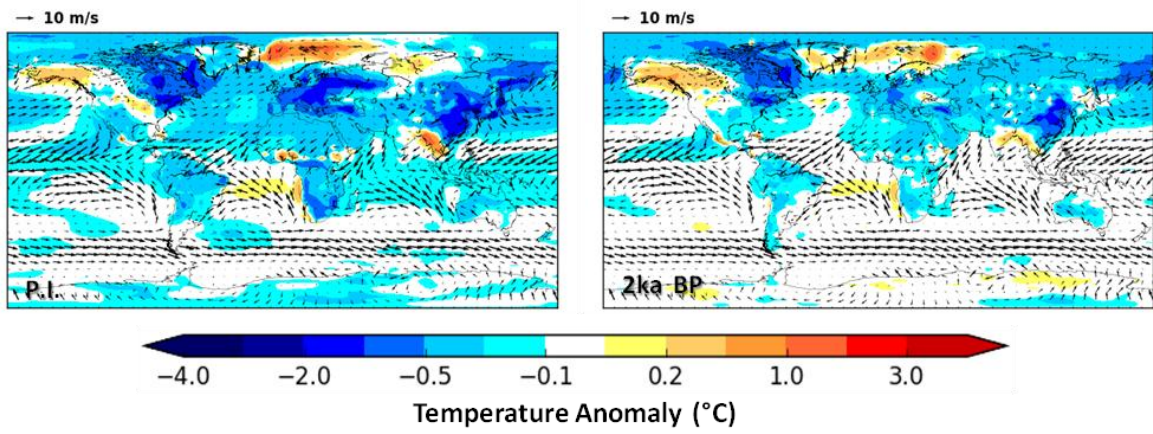
1
2
3
4
5
6

Figure 4. Annual Temperature Anomalies (°C) for KK10 minus Control at timeslices from the late pre-industrial period (1850 CE) to 7kaBP. The stippling indicates grid boxes where the anomalies are significant at the 95% level using Wilcoxon ranksum statistical analysis.



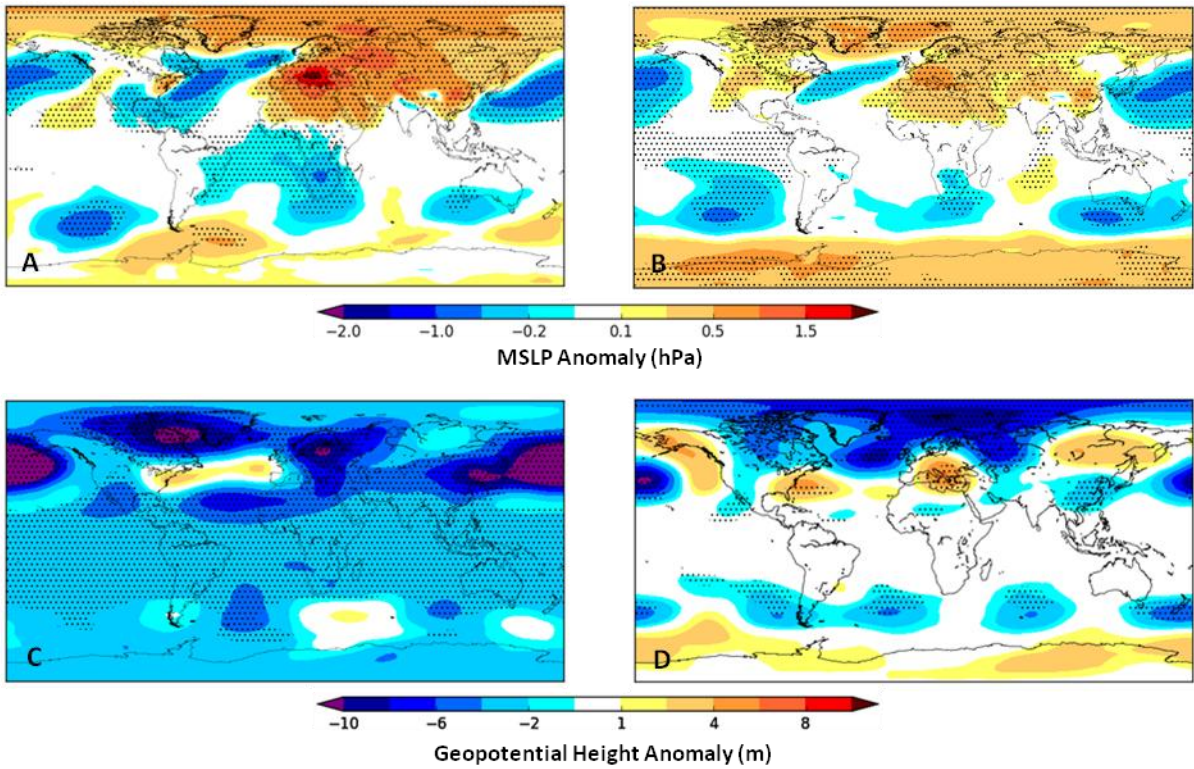
1
2
3
4
5
6
7
8
9

Figure 5. Relationship between the fraction of anthropogenically disturbed land (from KK10) and the resultant JJA temperature anomaly ($^{\circ}\text{C}$). **(a)** For the late pre-industrial (1850 CE) period for extratropical and tropical grid cells. **(b)** For 7kaBP, 4kaBP, 2kaBP and late pre-industrial (1850 CE) timeslices.



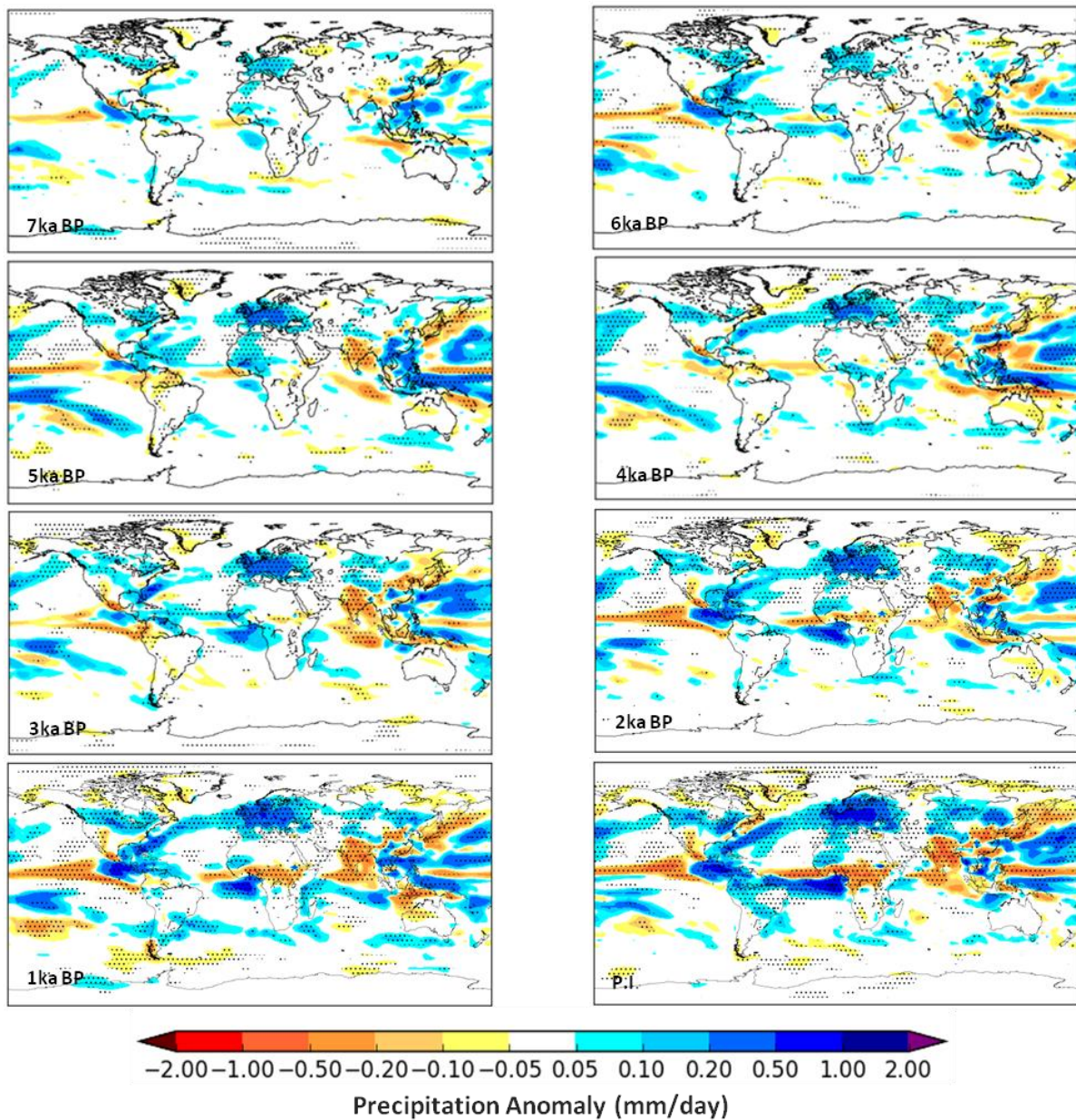
1
2
3
4
5
6
7
8
9
10

Figure 6. DJF surface temperature anomalies for KK10 - Control and KK10 surface winds for the late pre-industrial (1850 CE) and 2kaBP.



1
2
3
4
5
6
7
8
9
10
11
12
13

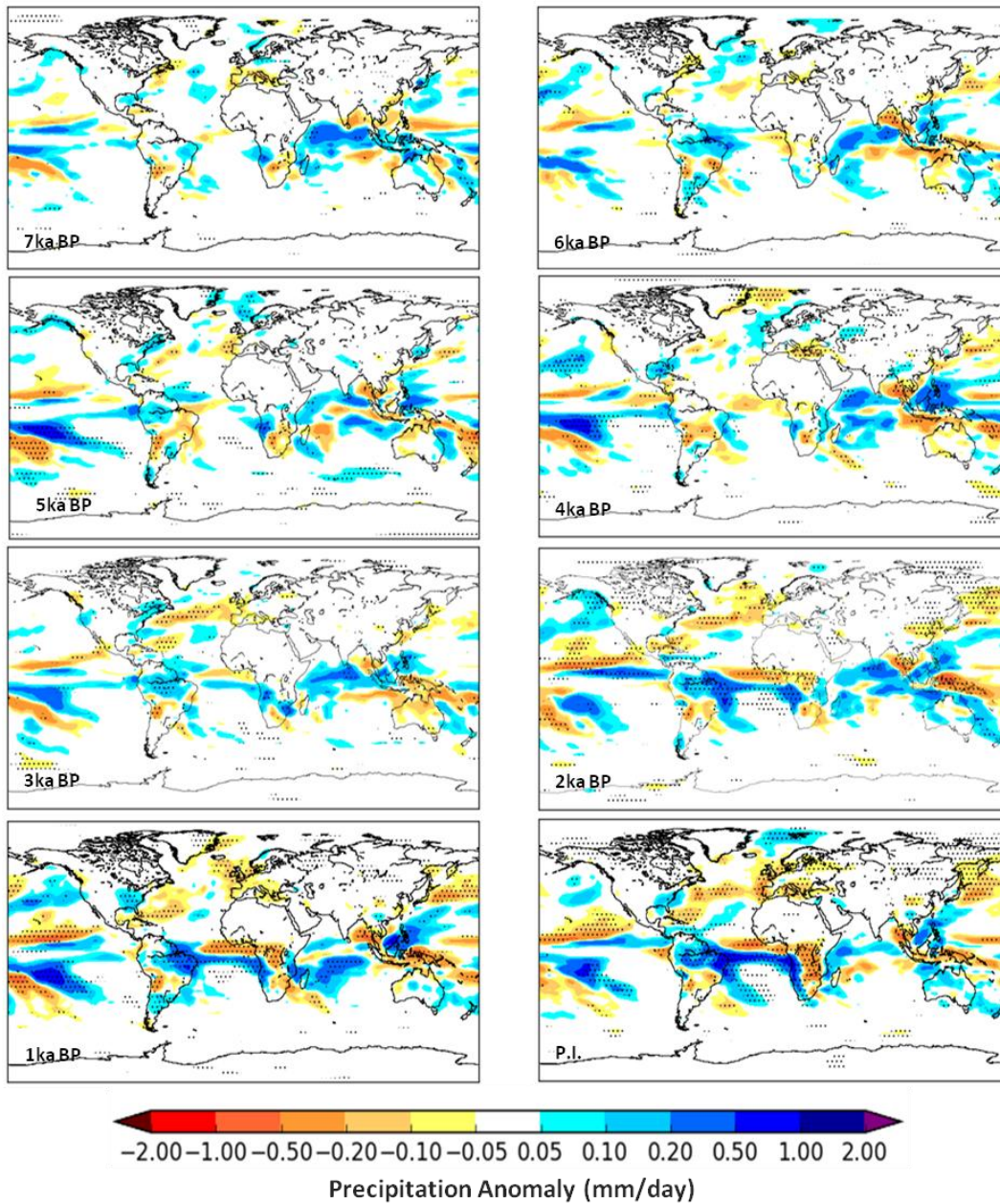
Figure 7. Modifiers of climate in regions outside the areas of anthropogenic land use change. All anomalies are for KK10 - Control: (a) JJA MSLP changes at 1850 CE, the stippling indicates grid boxes where the anomalies are significant at the 95% level using Wilcoxon ranksum statistical analysis; (b) as (a) but for 4kaBP; (c) DJF 500Pa geopotential height anomalies at 1850 CE demonstrating stationary wave pattern. The stippling indicates grid boxes where the anomalies are significant at the 95% level using Wilcoxon ranksum statistical analysis. (d) as (c) but for 4kaBP.



1

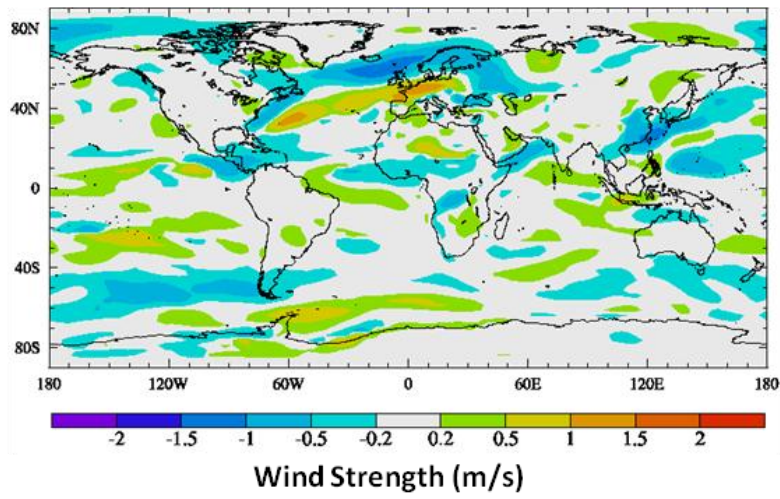
2

3 **Figure 8.** JJA Precipitation Anomalies (mm day⁻¹) for KK10 minus Control at timeslices from
 4 the late pre-industrial period (1850 CE) to 7kaBP. The stippling indicates grid boxes where
 5 the anomalies are significant at the 95% level using Wilcoxon ranksum statistical analysis.



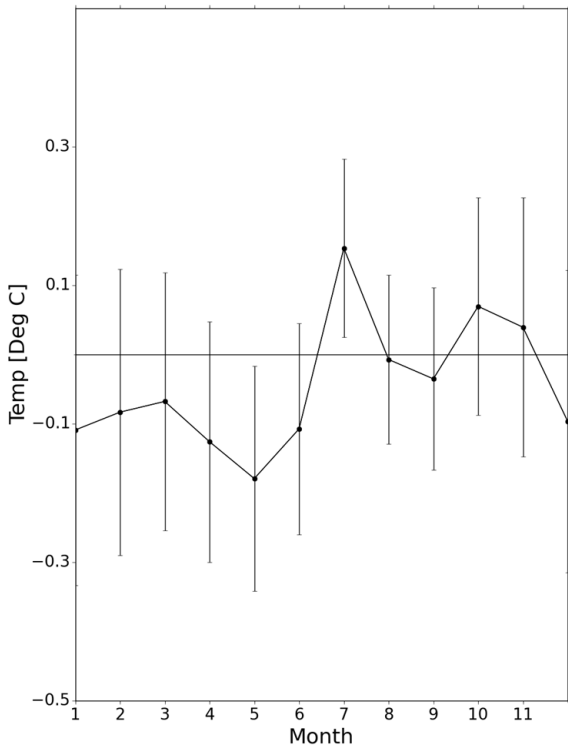
1
2
3
4
5
6
7
8

Figure 9. DJF Precipitation Anomalies (mm day^{-1}) for KK10 minus Control at timeslices from the late pre-industrial period (1850 CE) to 7kaBP. The stippling indicates grid boxes where the anomalies are significant at the 95% level using Wilcoxon ranksum statistical analysis.



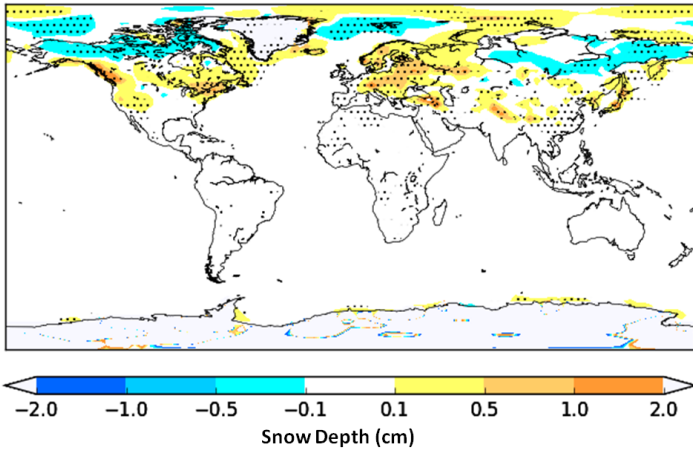
1
2
3
4
5
6
7
8

Figure 10. JJA 850hPa Wind Strength Anomalies (ms^{-1}) for KK10 minus Control for the late pre-industrial period (1850).



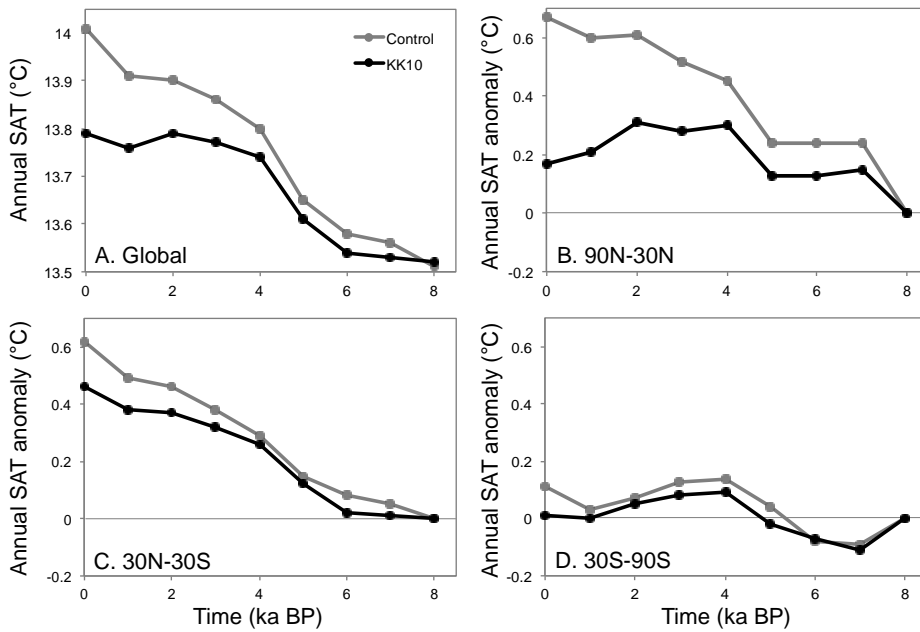
1
2
3
4
5
6
7
8
9

Figure 11. Indian temperature seasonality; KK10 - Control surface temperature anomalies for the late pre-industrial (1850CE) simulation; the vertical bars indicate the normal range of variability which is considered to be within 2 standard deviations of the mean.



1
2

3 **Figure 12.** DJF Snowdepth Anomalies (cm) for KK10 minus Control for the late pre-
4 industrial period (1850 CE). The stippling indicates grid boxes where the anomalies are
5 significant at the 95% level using Wilcoxon ranksum statistical analysis.



1

2 **Figure 13.** Time series plots for the Holocene simulations from HadCM3. (a) annual mean
 3 global Surface Air Temperature (SAT) for the Control simulation in grey, and KK10
 4 simulation in black; (b) anomaly in northern extratropical (30-90N) annual mean temperature
 5 from the equivalent simulation at 8kaBP; (c) same as (b) but for the tropics (30N-30S); (d)
 6 same as (b) but for the southern extratropics (30-90S).

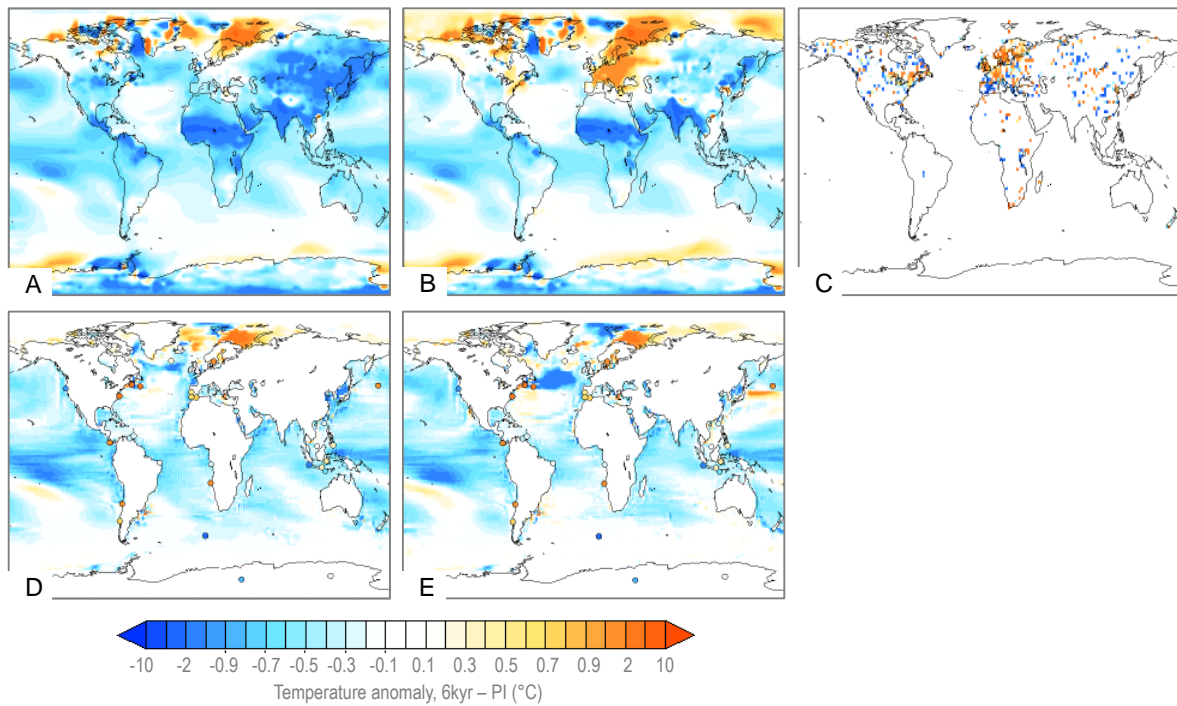
7

8

9

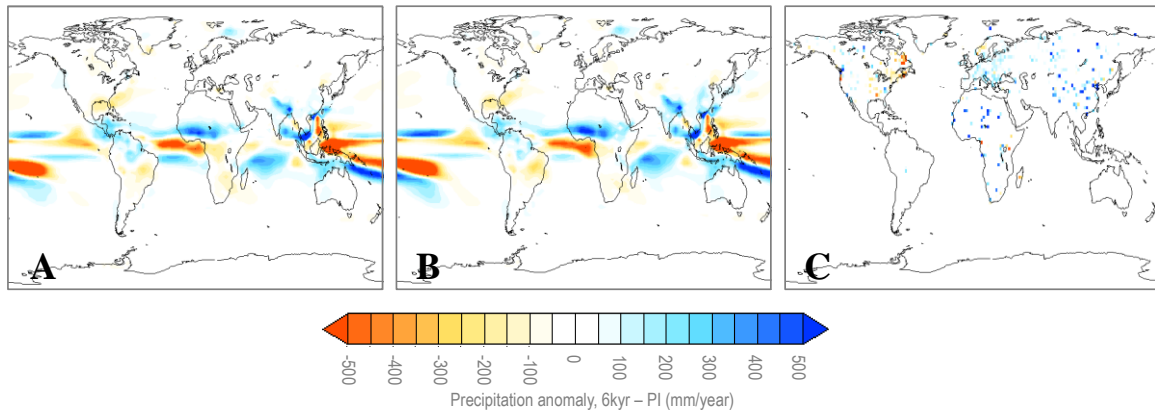
10

11



1
 2 **Figure 14.** Mid-Holocene minus Pre-Industrial temperature anomalies. **(a)** annual mean
 3 surface air temperature anomalies from the control experiment; **(b)** annual mean surface air
 4 temperature anomalies from the KK10 experiment; **(c)** Bartlein et al. (2011) pollen-based
 5 reconstructions of the mean annual air temperature anomaly; **(d)** ocean annual mean
 6 temperature for the top two layers (25m) from the control experiment with palaeo-proxy
 7 reconstructions (Marcott et al., 2013) overlain in coloured circles ; **(e)** as in **(d)** but for the
 8 KK10 experiment.

9
 10
 11
 12



1

2 **Figure 15.** Mid-Holocene minus Pre-Industrial precipitation anomalies. **(a)** annual mean
 3 precipitation anomalies from the control experiment; **(b)** annual precipitation anomalies from
 4 the KK10 experiment; **(c)** Bartlein et al. (2011) pollen-based reconstructions of the mean
 5 annual precipitation anomaly.

6

7

8

9

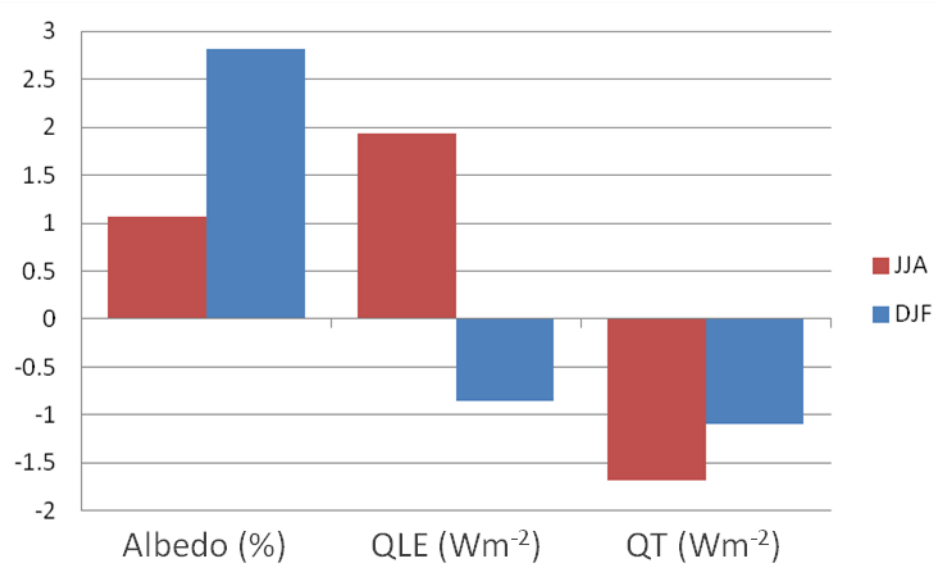
10

11

12

13

14



1
2
3
4
5
6
7

Figure 16. JJA and DJF albedo, latent heat flux (QLE) and turbulent heat flux (QT) anomalies for KK10 minus Control for the late pre-industrial period (1850 CE) for the North America/Eurasia land surface.