

# The relative roles of CO<sub>2</sub> and palaeogeography in determining ~~late~~Late Miocene climate: results from a terrestrial model-data comparison.

Catherine D. Bradshaw<sup>1\*</sup>, Daniel J. Lunt<sup>1</sup>, Rachel Flecker<sup>1</sup>, Ulrich Salzmann<sup>2</sup>, Matthew J. Pound<sup>2,3,4</sup>, Alan M. Haywood<sup>3</sup>, Jussi T. Eronen<sup>5,6</sup>

[1] Bristol Research Initiative for the Dynamic Global Environment (BRIDGE), School of Geographical Sciences, University of Bristol, University Road, Bristol BS8 1SS, UK

[2] School of the Built and Natural Environment, Northumbria University, Newcastle upon Tyne, NE1 8ST, UK

[3] School of Earth and Environment, University of Leeds, Leeds, LS2 9JT, UK

[4] British Geological Survey, Kingsley Dunham Centre, Keyworth, Nottingham NG12 5GG, UK

[5] Department of Geosciences and Geography, P.O. Box 64, FI-00014, University of Helsinki, Finland

[6] Biodiversity and Climate Research Centre LOEWE BiK-F, Senckenberganlage 25, D-60325 Frankfurt am Main, Germany

Correspondence to: C.D. Bradshaw (c.bradshaw@bristol.ac.uk)

## Abstract

The ~~late~~Late Miocene (~~~11.6—5.3 Ma~~) palaeorecord provides evidence for a warmer and wetter climate than that of today and there is uncertainty in the palaeo-CO<sub>2</sub> record of at least ~~2045~~0ppmv. We present results from fully coupled atmosphere-ocean-vegetation simulations for the ~~late~~Late Miocene that examine the relative roles of palaeogeography (topography and ice sheet geometry) and CO<sub>2</sub> concentration in the determination of ~~late~~Late Miocene climate through comprehensive terrestrial model-data comparisons. Assuming that the ~~se~~ data accurately reflects the ~~late~~Late Miocene climate, and that the ~~late~~Late Miocene palaeogeographic reconstruction used in the model is robust, then results indicate that:

1. Both palaeogeography and atmospheric CO<sub>2</sub> contribute to the proxy-derived precipitation differences between the ~~late~~Late Miocene and modern. However these

contributions exhibit synergy and so do not add linearly. ~~can be largely accounted for by the palaeogeographic changes alone.~~ However, t

2. The vast majority of the proxy-derived ~~temperature~~temperatures differences between the ~~lateLate~~ Miocene and modern can only ~~begin to~~ be accounted for if we assume a palaeo-CO<sub>2</sub> concentration towards the higher end of the range of estimates.

## 1 Introduction

The terrestrial palaeorecord contains evidence that the ~~lateLate~~ Miocene (~~11.616~~ - 5.333 Ma; [Gradstein et al., 2004](#); [Hilgen et al., 2005](#)) climate was, in many regions, much warmer and/or wetter than today (e.g. [Pound et al., 2011](#); [2012](#)). For example, warm temperate forests thrived in what is now the Circumboreal Region, ([Worobiec and Lesiak, 1998](#); [Denk et al., 2005](#)), and grasslands existed in modern desert regions (e.g. the Arabian peninsula, [Kingston and Hill, 1999](#); [the Sahara Desert, Vignaud et al., 2002](#)). Moreover, although it is widely suggested that a large-scale Antarctic ice sheet has existed throughout the ~~lateLate~~ Miocene (e.g. [Shackleton and Kennett, 1975](#); [Lewis et al., 2008](#)), Northern Hemisphere glaciation is thought to have been limited ([Moran et al., 2006](#); ~~Kamikuri et al., 2007~~; ~~Jakobsson et al., 2007~~).

The ~~lateLate~~ Miocene is also a period in which significant tectonic reorganisation occurred including major tectonic uplift of the Himalayas ([Harrison et al., 1992](#); [Molnar et al., 1993](#); [Rowley and Currie, 2006a](#); [Fang et al., 2005](#)), the Andes ([Gregory-Wodzicki, 2002](#); [Garzione et al., 2008](#)), the North American Rockies ([Morgan and Swanberg, 1985](#)), the East African Plateaus ([Saggerson and Baker, 1965](#); [Yemane et al., 1985](#)), and the Alps ([Spiegel et al., 2001](#); [Kuhleemann, 2007](#)). ~~Significant differences existed in the oceans too); and ocean gateways~~, e.g. there is evidence for an open Panama Gateway between the Atlantic and Pacific Oceans ([Keigwin, 1982](#); [Duque-Caro, 1990](#)) and an unrestricted Indonesian Seaway between the Pacific and Indian Oceans ([van Andel et al., 1975](#); [Edwards, 1975](#); [Kennett et al., 1985](#); [Cane and Molnar, 2001](#)) during the ~~lateLate~~ Miocene.

Modelling studies have shown that ~~lateLate~~ Miocene-like differences in palaeogeography and ice sheet extents are likely to have resulted in large changes to both atmospheric and oceanic circulation compared to the present day (e.g. Andes uplift: ([Takahashi and Battisti, 2007](#)); Rockies uplift: ([Seager et al., 2002](#); [Foster et al., 2010](#)); uplift of the Tibetan Plateau: ([Ramstein et al., 1997](#); [Lunt et al., 2010](#); ~~Zhongshi-Zhang et al., 2007a~~; ~~Zhongshi-Zhang et al., 2007b~~; [Ruddiman et al., 1997](#); [Kutzbach and Behling, 2004](#)); Northern Hemispheric

glaciation: (Crowley and Baum, 1995; Lunt et al., 2004; Tonazzio et al., 2004); Panama Gateway closure: (Lohmann et al., 2006; Nisancioglu et al., 2003; Schneider and Schmittner, 2006; Lunt et al., 2008b); and closure of the Paratethys: (Ramstein et al., 1997; Zhang et al., 2011).

~~Although there is considerable scatter in the Late Miocene atmospheric CO<sub>2</sub> reconstructions deduced from a variety of proxies, the estimates are generally quite low as compared to older time periods such as the Eocene (Beerling and Royer, 2011); (Freeman and Hayes, 1992; Pagani et al., 1999a; Pagani et al., 1999b; Pearson and Palmer, 2000; Figure 1: Demiecco et al., 2003; Kurschner et al., 2008; Kurschner et al., 1996; Retallack, 2001; Tripathi et al., 2009). Late Miocene atmospheric CO<sub>2</sub> levels are therefore uncertain, but it is likely that the concentration fluctuated over this period.~~

It is therefore reasonable to expect that the combination of all of these changes ~~(Figure 2)~~ would result in a climate that was significantly different to that which we experience today, and this is supported by GCM modelling simulations for the ~~late~~Late Miocene (e.g. ~~(~~Steppuhn et al., 2006; Gladstone et al., 2007; Lunt et al., 2008a; Micheels et al., 2011; Knorr et al., 2011). However, in order to evaluate the simulated ~~late~~Late Miocene climate from a GCM it is necessary to make quantitative comparisons with the palaeorecord.

Previous climate modelling has generally struggled to simulate the warm conditions inferred from the palaeorecord, when imposing relatively low CO<sub>2</sub> concentrations in the models (e.g. Steppuhn et al., 2006; e.g. Micheels et al., 2007), although a recent study which also included prescribed vegetation has met with some success, highlighting the importance of vegetation (Knorr et al., 2011). For computational reasons, most of the ~~late Miocene modelling to date has not been carried out with a coupled atmosphere-ocean climate model and recent work~~Late Miocene modelling to date has been carried out with atmosphere-only climate models (Barron, 1985; Kutzbach et al., 1989; Ruddiman and Kutzbach, 1989; Fluteau et al., 1999; Zhongshi et al., 2007a; Zhongshi et al., 2007b; Lunt et al., 2008a), atmosphere-slab ocean climate models which are incapable of simulating dynamic changes to ocean currents or overturning circulation patterns (Prell and Kutzbach, 1992; Dutton and Barron, 1997; Ramstein et al., 1997; Ruddiman et al., 1997; Kutzbach and Behling, 2004; Steppuhn et al., 2006; Steppuhn et al., 2007; Micheels et al., 2007; Micheels et al., 2009a) or uncoupled atmosphere and ocean models (Lohmann et al., 2006). Recent work using a fully coupled ~~atmosphere-ocean climate model~~ has highlighted the disparity between the results obtained

from atmosphere-only or atmosphere-slab ocean models and ~~a-coupled atmosphere-ocean~~ models capable of incorporating important ocean circulation feedbacks (Micheels et al., 2011; Knorr et al., 2011). This is particularly important for simulating the ~~lateLate~~ Miocene, since major reorganisation of Atlantic, Pacific and Indian Ocean circulation patterns occurred at this time (Collins et al., 1996; Kennett et al., 1985; Duque-Caro, 1990; Kameo and Sato, 2000; Cane and Molnar, 2001). Many of the most advanced GCM's now incorporate a dynamic land surface component which both responds, and feeds back to, the atmospheric state, therefore allowing a more stringent test of the GCM than using a fixed late Miocene vegetation reconstruction.

These lines of evidence - warm/wet ~~lateLate~~ Miocene climate, uncertain CO<sub>2</sub>, and significant tectonic uplift and ice sheet change - demand the question, "Can modelling inform what the likely relative roles of palaeogeographic changes and CO<sub>2</sub> concentrations are in determining ~~lateLate~~ Miocene climate". Here we present the model results from a fully coupled atmosphere-ocean-vegetation GCM for the ~~lateLate~~ Miocene, and compare our simulations with a synthesis of available climate proxies from the terrestrial realm. ~~Notable differences are found to occur in the oceanographic realm for our Late Miocene simulations as compared to our control simulation, such as warming in the Indian and Pacific Oceans, reductions in overturning circulation in the North Atlantic and widespread cooling of the waters across the Panama Gateway.~~ Most marine proxy data available for the ~~lateLate~~ Miocene comes from the  $\delta^{18}\text{O}_c$  of foraminifera, much of which has not been converted to palaeotemperature estimates. Furthermore, the treatment of the uncertainties associated with that data such as  $\delta^{18}\text{O}_{sw}$  reconstruction, depth habitat and seasonality considerations, and preservation requires in-depth discussion which is outside the scope of this manuscript. Consequently, although we do show and discuss near-surface oceanographic differences between ~~our lateout-Late~~ Miocene and CTRL simulations, these differences will be described in more depth and compared with data in a forthcoming paper (Bradshaw et al., in prep.)

As the timing and magnitude of the tectonic changes during the ~~lateLate~~ Miocene are uncertain, GCM simulations for this period are representative of a large timeslab. But this also allows us to examine the impact of large changes in the paleogeography and to assemble as large a number of palaeodata as possible. We therefore examine the compiled palaeodata for any marked differences in the climate between the Tortonian (11.616-7.25 Ma) and the

Messinian (7.25-5.333 Ma) of the ~~late~~Late Miocene and compare the model simulations to those sub-datasets separately.

## 2 Data and Model Descriptions

In this section we describe the terrestrial palaeodata and the GCM that has been used, and discuss some of the uncertainties that are inherent in both.

### 2.1 Description of the Quantitative ~~late~~Late Miocene Palaeodata

The palaeorecord of the ~~late~~Late Miocene is sparse in comparison with more recent periods such as the LGM, (Farrera et al., 1999); ~~Waelbroeck et al., 2009~~). We have compiled a total of ~~1030~~1409 terrestrial climate reconstructions from the literature for the climatic variables of mean annual temperature (MAT), ~~and~~ mean annual precipitation (MAP), ~~mean temperature of the coldest month (CMT) and mean temperature of the warmest month (WMT). We include in our database those data whose age uncertainty overlaps with our criteria for the late Miocene (5.33-11.61 Ma), see Appendix A for more information.~~ As comparison is made to the dataset separately for the Messinian and the Tortonian stages, any data that had age uncertainty placing it between the two stages was duplicated, one datapoint residing in each stage. The total number of datapoints considered therefore increased to ~~1166~~1671.

The data synthesis largely consists of the data from the NECLIME database (Bruch et al., 2007; ~~Utescher et al., 2011~~) and the NOW Neogene Mammal Database (Eronen et al., 2010; Fortelius, 2012) but also includes data from a variety of other sources including the synthesis of (Steppuhn et al., 2007), as detailed in the Supplementary Information. These reconstructions are derived from microfauna, macrofauna, microflora and macroflora proxies using a variety of different reconstruction methods including the co-existence approach (Mosbrugger and Utescher, 1997), the climatic amplitude method (Fauquette et al., 2006), nearest living relative methods (Wolfe, 1993; ~~Spicer, 2007~~; ~~Spicer et al., 2009~~), species characteristic approaches (Eronen et al., 2010; ~~Jacobs and Deino, 1996~~), and species composition approaches (Montuire et al., 2006; ~~Bohme et al., 2008~~; ~~van Dam, 2006~~). ~~A discussion of the uncertainties in proxy reconstructions is given in 0 and all—AH~~ of these terrestrial data are provided in the Supplementary Information. Comparison is also made to the ~~Tortonian~~-biome reconstruction ~~datasets~~dataset of Pound et al. (2011; ~~2012~~).

### ~~2.1.1 Uncertainty in Late Miocene Proxy Reconstructions~~

~~As a multi-proxy dataset is used, there is considerable variability in the age constraint due to uncertainty in the dating. Although the data from small mammals and microfloras are generally very well dated, large mammals and macrofloras have much looser age determination and many macroflora datapoints are only consigned to the Messinian or the Tortonian. Therefore the data is considered in 2 timeslabs: the Messinian (5.3–7.25 Ma) and the Tortonian (7.25–11.6 Ma). We note that the proxy data used to is likely to have been generated under a range of different orbital configurations and other climatic cycles, and individual datapoints could therefore represent extremes in these cycles, but this is unlikely to have biased the entire dataset.~~

~~For a comparison study between fossil data and GCM simulations where a palaeontological assemblage has been used to reconstruct climatic conditions, such as the mean annual temperature or precipitation, two main uncertainties exist: taphonomical (the death, decay, deposition and preservation of fossils) and the uncertainties associated with the data reconstruction techniques. Discussion of these are outside the scope of this paper, and the reader is referred to the source references for those details: (Taphonomy: Western and Behrensmeyer, 2009; Martin, 1999; Behrensmeyer et al., 2000); (Mosbrugger and Utescher, 1997; Fauquette et al., 2006; Wolfe, 1993; Spicer, 2007; Spicer et al., 2009; Eronen et al., 2010; Jacobs and Deino, 1996; Montuire et al., 2006; Data reconstruction techniques: Bohme et al., 2008; van Dam, 2006).~~

## **2.2 Description of the Modern Reference Climate (Modern Potential Natural Climate)**

For the purpose of this study, we are interested in the ~~late~~**Late** Miocene climate because we wish to understand how and why it was different to today. In terms of understanding these differences, we are only interested in the changes that occur under natural forcing (a modern potential natural climate), however the modern instrumental~~al~~ record contains both natural and anthropogenic forcing. In order to make an estimate of the extent of anthropogenic influence on the modern instrumental record ( $IR_{20thC}$ ), we derive grid-based correction factors using the anomaly between the “Preindustrial” simulations ( $Model_{preind}$ ) and the “20<sup>th</sup> Century” simulations ( $Model_{20thC}$ ) of the CMIP-4 GCM ensemble (IPCC, 2007) as detailed in Equation 1. Through application of the correction factors, it is assumed that the modern instrumental record is a suitable proxy for the potential natural climate state.

$$\text{Modern potential natural climate} = IR_{20thC} - \left( \frac{1}{n} \sum_{i=1}^n Model_{20thC} - Model_{Preind} \right) \quad (1)$$

The modern instrumental record used is the CRU-TS 3.0 Climate Database (Mitchell and Jones, 2005), an interpolation of mean monthly surface weather station data to a 0.5 x 0.5 degree grid, spanning the period 1900-2006. The modern climate estimates for temperature and precipitation at any grid point will be least certain in areas of largest interpolation where no weather stations exist, but in other areas there is uncertainty in the records too due to limited temporal operation of some stations, station movement, urbanisation and other land use changes. The observational datasets are at a higher spatial resolution than the HadCM3L model, therefore when aggregating the observations to the HadCM3L model resolution, the minimum and maximum values within each grid box are retained in order to account for within-grid box uncertainty.

## 2.3 Description of the Models

### 2.3.1 HadCM3L GCM

The Hadley Centre Coupled Model, version 3 (HadCM3) is a fully coupled atmosphere-ocean general circulation model comprising the atmospheric model HadAM3 (resolution 2.5° latitude by 3.75° longitude), and the ocean model HadOM3 (resolution 1.25° latitude by 1.25° longitude) (Gordon et al., 2000). It is computationally demanding to run HadCM3 simulations to equilibrium so we use a reduced ocean resolution version, HadCM3L, (resolution 2.5° latitude by 3.75° longitude) (Cox et al., 2000). In order to run the model without the use of flux corrections, we removed the two grid cells representing Iceland (Jones, 2003). The model contains many parameterizations of atmospheric and oceanic processes, including atmospheric convection (Gregory and Rowntree, 1990), boundary layer mixing (Smith, 1993), ocean layer mixing (Gent and McWilliams, 1990), and radiation (Edwards and Slingo, 1996). HadCM3L has been used in studies of the future (e.g. Cox et al., 2000) and the past (e.g. Lunt et al., 2007; Lunt et al., 2010). As HadCM3L has a reduced ocean resolution, model performance is degraded compared to HadCM3. For example, the ocean resolution of HadCM3L is such that the Denmark Straits are unresolved and therefore the two terrestrial grid points representing Iceland are removed to prevent an unacceptable build up of sea ice in the Nordic Sea (Jones, 2003). Full details of the model are given in Appendix B Section 1.1.) The version of HadCM3L used here does not therefore require the

1 ~~inclusion of flux corrections to maintain a stable overturning circulation for the modern ocean~~  
2 ~~(Jones, 2003).~~

3 ~~The model land surface processes are simulated using the MOSES 2.1 (Met Office Surface~~  
4 ~~Exchange Scheme) land surface scheme (Essery and Clark, 2003). There are nine surface~~  
5 ~~types, five of which correspond to plant functional types (PFTs): broadleaf tree, needleleaf~~  
6 ~~tree, C<sub>3</sub> grass, C<sub>4</sub> grass and shrub; the remainder representing bare soil, water bodies, ice and~~  
7 ~~urban surfaces.~~

### 8 2.3.2 TRIFFID and BIOME4 Vegetation Models

9 The dynamic global vegetation model coupled to HadCM3L is TRIFFID (Top-down  
10 Representation of Interactive Foliage and Flora Including Dynamics), a full description of  
11 which is given in (Cox, 2001) and (Hughes et al., ~~2004~~). ~~TRIFFID calculates areal coverage,~~  
12 ~~leaf area index and canopy height for the five PFTs, which each respond differently to~~  
13 ~~climate and CO<sub>2</sub> forcing, and all can co-exist within the same gridbox. 2004).~~ Refer to  
14 Appendix B, Section 2.1 for more details about the TRIFFID model and its uncertainties.

15 ~~Two modes of coupling between the TRIFFID and the GCM are used: a spinup equilibrium~~  
16 ~~mode in which the fluxes between the land and the atmosphere are calculated by the GCM~~  
17 ~~and averaged over ~5 years, and a dynamic mode, used for the last 100 years of the~~  
18 ~~simulations, in which the fluxes are averaged over 10 days. The averaged fluxes are then~~  
19 ~~passed to TRIFFID which calculates the growth and expansion of the existing vegetation, and~~  
20 ~~updates the land surface parameters based on the new vegetation distribution and structure.~~

### 21 ~~2.3.3 BIOME4~~

22 In order to make model-data comparisons with a recently published biome reconstruction  
23 dataset for the ~~late~~Late Miocene (Pound et al., 2011; 2012), the HadCM3L model  
24 climatologies are used to drive the BIOME4 vegetation model offline (Kaplan, 2001; Kaplan  
25 et al., 2003). The 28 different vegetation classes used by Pound et al. (2011; 2012) match  
26 the vegetation classifications of the BIOME4 model, which has successfully been used for  
27 previous pre-Quaternary vegetation model-data comparisons (e.g. Salzmann et al., 2009).  
28 See Appendix B, Section 3.1 for a discussion of the uncertainties in the BIOME4 model.  
29 2009).

#### 2.3.42.3.3 Model Uncertainties

No GCM is able to reproduce exactly the modern climate, and a discussion of the biases in the HadCM3 version of our GCM is given in Gordon et al. (2000). The HadCM3L version of the model suffers from temperature biases, particularly for too cool ocean temperatures at high latitudes. Temperatures over the land surface are generally within the range of uncertainty of the CRU-TS 3.0 modern instrumental data (Mitchell and Jones, 2005), but there is still a marked cold bias at high northern latitudes of up to 10°C above 70°N. Similar biases exist for precipitation, with the mean annual precipitation being underestimated by up to a third in parts of northern South America and southern India, and overestimated by the same amount in some tropical regions of South America, Africa and Australia.

Both TRIFFID and BIOME4 are based on physiological constraints that influence vegetation distributions, and are therefore attempting to model the potential natural vegetation for a given climatology. Estimates of anthropogenic land surfaces (urban and agricultural) amount to some 16% of the total land surface area (CIESIN, 2004; Matthews, 1983; Ramankutty and Foley, 1999; Schneider et al., 2009) and therefore discrepancies will exist in those areas between reference maps of reconstructed natural vegetation (the vegetation that might exist if human influence were to cease, based on extrapolation of remnants of natural vegetation) and model-produced maps of potential natural vegetation (the vegetation that might exist if human influence had never existed), and these may be significant in areas of deforestation (Prentice et al., 1992).

The TRIFFID model has been compared to IGBP-DIS land cover dataset, which represents the modern distribution of vegetation as derived from satellite image interpretation (Loveland and Belward, 1997; Betts et al., 2004). This suggests that the shrub PFT is overestimated at high latitudes, that the broadleaf tree PFT is overestimated in equatorial regions, and that grasses tend to be globally slightly underestimated. The discrepancies between the satellite imagery and the TRIFFID model are suggested to be a combination of orographic representation leading to underestimation of precipitation, differences between the anthropogenic masks used in their version of the model and that found on the satellite imagery, and the inadequate treatment of natural disturbance mechanisms such as fire (Betts et al., 2004). The version of the model used here aims to reproduce the vegetation of a modern potential natural climate, and therefore evaluation of the predicted PFTs is difficult. It is also hard to identify, when the modern predicted PFTs are likely wrong (e.g. needleleaf

~~forests in south-east Asia and extensive broadleaf forests in Australia), whether those inadequacies are due to the climate model or the vegetation model.~~

~~The BIOME4 model has been shown to produce a fair comparison to the vegetation data of Olson et al. (1983), with most of the areas of discrepancy being attributed to differences between potential natural vegetation and actual modern vegetation as a result of human influence. Other suggested sources of error include incorrect climatological forcing due to the difficulty in representing orographic extremes, and missing parameterisations in the model, particularly related to seasonality (Prentice et al., 1992).~~

A discussion of model uncertainties is given in 0. The GCM simulations were integrated over 2100 model years. Particular attention is given to the question of equilibrium in the climate system, particularly for the deep ocean as it has been shown that the global mean air temperature may not be a good indicator of model equilibrium (Brandefelt and Otto-Bliesner, 2009). The trend in the global mean ocean temperatures is very small for the ~~late~~Late Miocene simulations ( $<0.0008^{\circ}\text{C}/\text{century}$ ) and our control simulation ( $0.0005^{\circ}\text{C}/\text{century}$ ), although it is not possible to rule out some remaining disequilibrium. The analysis in this paper is carried out using the climatological means of the last 50 years of the simulations, after the TRIFFID dynamic mode ~~washad~~ used for 100 years to ensure full consideration of interannual variability in the vegetation component.

### 3 Experiment Design

#### 3.1 Model Setup

For the ~~late~~Late Miocene, the palaeogeography (including ice sheet geometry) derived by Markwick (2007) is used, which is representative of the timeslab 11.6 to 5.3 Ma. The length of this paleogeographic timeslab is necessarily long due to difficulties associated with the poorly constrained chronology and age estimates uncertainty for the various tectonic and ice sheet changes (Markwick and Valdes, 2004).

The orography, ice sheet configuration and continental positions assumed for the model simulations are shown in Figure 2, and are discussed in detail in 0 and 0, together with a discussion of the associated uncertainties in the late Miocene boundary conditions.~~now discussed in more detail. As we perform both changes in palaeogeography and  $\text{CO}_2$  concentration, there is the possibility that non-linearity in the climate system could produce a different result to that which we present here (i.e. if we were to consider the impact of  $\text{CO}_2$  changes independently of paleogeography changes).~~ However, recent work for the Pliocene,

using the HadCM3 version of the model used here, suggests that any non-linearity would be small (refer to Figure 6C of Lunt et al., in press).

### **Modern Configuration**

For the modern control (CTRL) simulation we use the standard UK Meteorological Office configurations for modern continental positions and both terrestrial and ice sheet elevations as defined in the ETOPO5 dataset (NOAA, 1988). Land ice grid boxes and soil type distributions used in CTRL are based on the identification of Wilson and Henderson-Sellers (1985). The two land grid boxes that represent Iceland are converted to ocean grid boxes as discussed in Section 2.3.4.

We also use modern ETOPO5 bathymetry (NOAA, 1988). The two new ocean grid boxes replacing Iceland are assigned the average depth of the surrounding ocean grid boxes. We note that the land-sea mask generation algorithms that map the high-resolution data onto the HadCM3L grid also results in the non-representation of many other small islands.

For the purpose of this study we wish to model a potential natural modern climate, and therefore assume a preindustrial atmospheric CO<sub>2</sub> concentration in the range 275–284 ppm (Etheridge et al., 1996) and use a value of 280ppm in CTRL, consistent with the Climate Model Intercomparison Project standard (Meehl et al., 2007). For the modern control simulation (hereafter referred to as CTRL), we use the standard UK Meteorological Office configuration for the model boundary conditions as described in Appendix C. The late

## **3.1 — Late Miocene boundary conditions significantly reduce the Palaeoenvironmental Configuration**

### **3.1.1 — Orography**

The orography for the Late Miocene has been reconstructed by (Markwick, 2007) using a methodology of establishing relationships between modern elevations of most and their tectonophysiographic settings and applying those same relationships to the geological record; (a full description of the technique is described in Markwick, 2007; Markwick and Valdes, 2004). The methodology results in significant reductions in the Miocene elevations of most of the world's highest regions compared to modern. We now describe those reductions in relation to the palaeoelevation estimates of other authors; however, given the uncertainties associated with reconstructing past mountain elevation (e.g. Ehlers and Poulsen, 2009), any inconsistencies with that of Markwick (2007) would require further sensitivity experiments.

~~The Tibetan Plateau in the Late Miocene simulations is an average 30% lower than in CTRL. The timing of Tibetan uplift is debated, with many studies indicating an elevation during the Late Miocene similar to today (e.g. Currie et al., 2005; DeCelles et al., 2007; Spicer et al., 2003; Garzzone et al., 2000; Rowley et al., 2001; Rowley and Currie, 2006b). There is evidence that the uplift did not occur in a spatially uniform pattern (England and Searle, 1986; Clark et al., 2005; Rowley and Garzzone, 2007), but a recent study has shown that modelled SSTs in the Indian and Pacific Oceans were very similar for three simulations assuming different uplift histories (Lunt et al., 2010).~~

~~The Andes are 45% lower in the Late Miocene simulations than the CTRL. This is consistent with palaeobotanical and geomorphological data which suggests that the elevation of the Andes was no more than half of the modern elevation at ~11–10 Ma (Gregory-Wodzicki, 2000).~~

~~The Alps are on average 30% lower in the Late Miocene simulations than the CTRL. Palaeoaltitude estimates for the Eastern Alps are between 500 and 2000m (Jimenez-Moreno et al., 2008; Kuhlemann, 2007). These estimates are consistent with the Markwick (2007) reconstruction.~~

~~The Western Cordillera of North America, is on average 25% higher in the Late Miocene simulations than in the CTRL. The evolution of topographic changes in the Western Cordillera are debated. There is evidence for high elevation in the Western Cordillera of North America since the Eocene (Sjostrom et al., 2006) or even the Cretaceous (Chase et al., 1998), with the elevation in Nevada during the Middle Miocene being some 1–1.5km higher than it is today. The dating of the collapse to present-day elevation in Nevada is put at ~13 Ma by Wolfe et al. (1997) but the palaeoelevation reconstruction from Markwick (2007) differs in that Nevada is ~9% lower in LM than CTRL. Orographic lowering in Wyoming is believed to have occurred during the Eocene (Chase et al., 1998), and the methodology used by Markwick (2007) puts Wyoming 27% lower in the Late Miocene than in the CTRL. A recent study of the Western Cordillera of North America also considers that elevations were generally lower during the Late Miocene than today, but argues for considerable evidence of surface uplift during the Late Miocene in the westernmost regions north of 45°N (Foster et al., 2010).~~

## **1.1 — Ice Sheet Configuration**

The extent of Late Miocene glaciation is much debated (Shackleton and Kennett, 1975; Webb and Harwood, 1991; Wilson, 1995; Pekar and DeConto, 2006; Huybrechts, 1993; Marchant et al., 1996) and perhaps highly variable (Lear et al., 2000). For the Late Miocene, the Markwick (2007) reconstruction assumes the East and West Antarctic ice sheets cover the whole Antarctic continent, although ice thickness is generally less than CTRL. The sensitivity experiments of DeConto et al. (2008) and Micheels et al. (2009a) support the evidence for the onset of some glaciation in the Northern Hemisphere during the Miocene (Moran et al., 2006; highest regions as compared to CTRL, and Kamikuri et al., 2007; Jakobsson et al., 2007; Denton and Armstrong, 1969; Thiede and Myhre, 1995 and references therein; Talwani and Udintsev, 1976; Warnke and Hansen, 1977). Markwick (2007) assumes much reduced Northern Hemisphere ice for the Late Miocene compared with the extent of the present day Greenland ice sheet. Future work will investigate the impact of ice sheet extent assumptions on the simulated Late Miocene climate.

The high latitude topography is lowered in the Late Miocene simulations compared to the CTRL, due to the removal of continental ice rather than post-Miocene uplift. Greenland is 60% lower and Antarctica is very spatially variable, but is on average 4% lower.

All non-ice covered land grid boxes were initialised with the TRIFFID shrub PFT before the TRIFFID model was switched on as the simulations were continuations of pre-existing simulations with global homogenous shrub coverage.

## **1.2—Continental Positions**

extents and thicknesses are also altered. We use the Late Miocene land-sea definitions of Markwick (2007), aggregated up to the GCM resolution. Notable changes to the land-sea mask for the late Miocene includein continental positions are the more southerly position of the Australian continent, the closed Bering Straits, the extension of Eurasia in to the Arctic, and the open Panama Gateway, the open and Indonesian Seaway for the Late Miocene.

## **1.3—Soil Type**

Some of the palaeoenvironmental conditions are largely unknown for the Miocene period, such as the soil type which is typically unknown outside of some regions in North America and Kenya (Retallack, 2004; Retallack et al., 1995; Retallack et al., 1990; Retallack et al., 2002a; Retallack et al., 2002b), and so for these we use an average homogenous value derived

from the CTRL simulation. Soil parameters have been shown to have significant influence on soil hydrology (Osborne et al., 2004), and therefore our choice of soil type could bias the simulated vegetation distribution and modify the simulated climate. Future experiments will be conducted to assess the extent to which the soil type could affect the results presented here.

#### **1.4 Bathymetry**

Late Miocene palaeobathymetry is assumed to be the same as for the CTRL, except for the land-sea mask itself, which results in a wider Indonesian Seaway and an open Panama Gateway. The algorithm used to generate the Late Miocene bathymetry assumes that the depth of all Late Miocene ocean grid boxes (that would be land today in the CTRL simulation) is the average of all neighbouring ocean grid boxes.

As palaeomagnetic studies for plate tectonic reconstruction are difficult to apply to the Indonesian islands (Gourlan et al., 2008), the evolution of their palaeogeography remains uncertain. It is generally believed that the restriction of the Indonesian Seaway has occurred in the past 20 Ma, with timing estimates for different water mass restrictions from Early Miocene to early Late Miocene (van Andel et al., 1975; Edwards, 1975), or even Pliocene (Kennett et al., 1985; Cane and Molnar, 2001; Srinivasan and Sinha, 1998). The Late Miocene seaway in our simulations is >2000m at its deepest, as for the CTRL.

The closure of the Panama gateway in the Pliocene is the most recent major modification to oceanic gateways (Keigwin, 1982; Duque Caro, 1990). The shoaling of the region is believed to have commenced in the Middle Miocene (Keller and Barron, 1983; Duque Caro, 1990; Droxler et al., 1998) but determining potential depth of the sill during the Late Miocene is problematic; a depth of 200–500m at 6Ma is suggested by Collins et al. (1996) and <100m depth by the early Pliocene by Keigwin (1982). Duque Caro (1990) suggests that in the earliest Late Miocene sill depth was ~1000m, and 150–500m between 8.6–7 Ma, <150m at 7–6.3 Ma and <50m at the latest Miocene. Given our large timeslab, possible sill depths for the Late Miocene range from 1000m to <50m; our bathymetric algorithm produces a depth of 995m which lies within the range of estimates.

The choice of bathymetry has been shown to have large regional impacts (e.g. the Greenland-Scotland Ridge: Robinson et al., 2011), but the uncertainty associated with palaeobathymetric

reconstruction is large and therefore future work will seek to assess the climatic uncertainty associated with the choice of palaeobathymetry through climate sensitivity studies.

, the closed Bering Strait, the extension of Eurasia into the Arctic and the more southerly position of the Australia.

### 3.1.2 Atmospheric CO<sub>2</sub> Concentration

We carried out two ~~lateLate~~ Miocene simulations, with two different values of atmospheric CO<sub>2</sub> concentration. Firstly, a value of 280 ppm was used, the same as CTRL, the difference in climate between these two simulations indicates the role of the palaeogeographic changes that we have made on the simulated climate. Secondly, a value of 400ppm was used in order to test the sensitivity of the results to uncertainties in the CO<sub>2</sub> concentration; the difference between the 400ppm simulation and the 280ppm simulation indicates the role of CO<sub>2</sub> in determining the simulated ~~lateLate~~ Miocene climate. These values are well within the range of uncertainty of ~~lateLate~~ Miocene CO<sub>2</sub> concentrations based on the available proxy data, shown in Figure 1. (~~Freeman and Hayes, 1992; Pagani et al., 1999a; Pagani et al., 1999b; Pearson and Palmer, 2000; Demieco et al., 2003; Kurschner et al., 2008; Kurschner et al., 1996; Retallack, 2001; Tripathi et al., 2009~~). The ~~lateLate~~ Miocene simulation with a 280ppm CO<sub>2</sub> concentration is hereafter referred to as LM280 and the simulation with a CO<sub>2</sub> concentration of 400ppm is referred to as LM400. We test the non-linearity of our results with an additional simulation which is identical to the CTRL simulation except that an atmospheric CO<sub>2</sub> concentration of 400 ppm is used.

Full access to the model output for these simulations is provided at <http://www.bridge.bris.ac.uk/resources/simulations>.

## 3.2 Model-Data Comparison Methodology

There are many difficulties associated with developing a methodology for a model-data comparison and in the interpretation of any results. We have considered uncertainties associated with model and data reconstructions that are often overlooked. These include uncertainties associated with the reference climatologies; poor temporal constraint leading to data location uncertainty; data transportation issues and palaeorotation uncertainty.

The methodology for estimating error ranges for each data record is as follows. First, we estimate the modern calibration error for each proxy type, if this is not already given in the original source, by assuming the same calibration error as the next most similar proxy

reconstruction method (for example, by assuming that the ~~physiognomic~~physiognometrie/morphology reconstructions have the same uncertainty as defined for CLAMP; see Table [12](#) for full details).

Secondly, we translate the individual proxy data locations back to their estimated palaeolocations as dictated by their age control, with the uncertainty in age resulting in a number of possible locations for each data point. We use the same palaeorotation as Markwick (2007) to ensure consistency with the ~~late~~Late Miocene palaeogeography used in HadCM3L. As we do not include Iceland as land in our model, we exclude the Icelandic data from our analysis, ~~reducing the original number of datapoints from 1109 to 1104.~~

Thirdly, we consider it unreasonable to expect our climate model to be able to reconstruct the exact climate at a single grid cell given the model uncertainties outlined in [Appendix B](#)~~Section 2.3.4~~, and because comparisons made between the BIOME4 model and maps of potential natural vegetation have shown that greater reliance should be placed on broad-scale patterns rather than individual grid cells (Prentice et al., 1992). ~~We, and~~ therefore ~~we~~ assume that all ~~adjacent~~ grid cells adjacent to the cell containing~~that underlying the~~ palaeodata could potentially be consistent with that data. Taking this approach to the model-data comparisons also allows for uncertainties in the true location of the data due to either ~~data~~ transportation or incorrect assumptions in the palaeorotations applied.

Because HadCM3L does not completely reproduce the modern reference climatology in the CTRL simulation it is assumed that the relative differences simulated by the model are more robust than the absolute values, and results are presented as the anomaly between the ~~late~~Late Miocene simulations and the modern control simulation. When comparisons are made between the model results and the palaeorecord, bias corrections are applied to the simulation results (the anomalies between the modern control simulation and the modern reference climate). As the CTRL simulation is designed to represent the modern potential natural climate and not the climate of the 20<sup>th</sup> century, we are unable to use the ideal bias correction factors given in Equation 2,

$$Model_{HadCM3Lc} = Model_{HadCM3L} + (Observations_{20thC} - Model_{HadCM3L20thC})$$

(2)

Therefore, the modern potential climate estimates from Equation 1 are used, as detailed in Equation 3,

$$Model_{HadCM3Lc} = Model_{HadCM3L} + (Modern\ potential\ natural\ climate - CTRL)$$

(3)

Where bias corrections have been applied to the model simulations, these are denoted by the subscript ‘c’, e.g. LM280c. When these corrections are applied to the modern control simulation CTRL, Equation 3 reduces to,

$$CTRLc = Modern\ potential\ natural\ climate \quad (4)$$

Whilst it is possible to make bias corrections to the climatology, we are unable to do the same for TRIFFID modelled vegetation cover because of the interactive coupling within the climate model. The standard anomaly method is however applied when forcing the BIOME4 model (e.g. Haxeltine and Prentice, 1996; Texier et al., 1997), applying a correction factor equal to the anomaly between the CTRL model climate and the modern climate, although it should be noted that this correction is based on the modern instrumental record and does not include our estimates of uncertainty of that record for a modern potential natural climate. In order to simplify the model-data comparison, the 28 biomes of the BIOME4 model are combined into 9 megabiomes: Tropical Forest, Temperate Forest, Warm-Temperate Forest, Grassland and Dry Shrubland, Savannah and Dry Woodland, Tundra, Boreal Forest, Desert and Land Ice.

~~In order to acknowledge~~To reflect the uncertainties in the data reconstructions and model results, both are treated as a range of possible values rather than the mean of possible values. The term ‘overlap’ is defined as consistency between model and data if their uncertainty ranges overlap (see Figure 3A). Overlap between model and data does not necessarily imply ‘good’ agreement because the uncertainty ranges may be large, but rather that where model and data overlap it is not possible to determine that they are different. In the figures and discussion that follow, the differences between model and data uncertainties are defined as the minimum possible difference given the range of possible values for both; if the true values for each were at the extremes of their respective uncertainties, then the differences between the two would be much larger (see Figure 3B).

#### 4 Description of the simulated ~~late~~Late Miocene climate

Here, model results are described in terms of anomalies; the difference between the ~~late~~Late Miocene (LM280 or LM400) and the modern (CTRL) climates. We also conduct a model-

model comparison of the present work with respect to previous Miocene GCM simulations, although direct comparison is difficult where different palaeogeographies, model complexity and ocean model setup have been used. ~~(Table 1)~~. Comparison between results from different vegetation models is even more difficult because of the different plant functional type and biome classification systems used. The ~~lateLate~~ Miocene simulations have been used to drive the interactive vegetation model TRIFFID which is coupled to the climate in the GCM, but also offline ~~to drive~~ by the BIOME4 model, and therefore it is possible to make some general comparison between the results of the two vegetation models we have used.

#### 4.1 Global

The annual and seasonal (DJF – December/January/February, JJA – June/July/August) mean air temperature and precipitation anomalies between our ~~lateLate~~ Miocene and CTRL experiments are shown in Figures 4(A-C) and 5(A-C). The magnitudes of the global anomalies are modest for LM280-CTRL (the impact of changing paleogeography alone), with the ~~mean annual temperature (MAT)~~ only 0.3°C higher and the ~~mean annual precipitation (MAP)~~ only 4.5 mm/a (0.4%) higher. The Micheels et al. (2011) model results obtain a larger MAT difference of 1.5°C between their same-CO<sub>2</sub> Tortonian and control simulations. However these results are not really comparable to ours because of the palaeogeographic differences between our studies, and the fact that they used a prescribed ~~lateLate~~ Miocene vegetation in their Tortonian simulations. It could also be that the influence of palaeogeographic change is dependent on the background CO<sub>2</sub> (Micheels et al., 2011 assume a CO<sub>2</sub> concentration of 360 ppmv for both simulations compared to our 280 ppmv). The magnitudes of the global anomalies are higher for LM400-CTRL (the impact of changing paleogeography and CO<sub>2</sub> concentration), with the MAT 2.88°C higher and the MAP 46.6 mm/a (4%) higher.

#### 4.2 Regional

Regions of palaeogeographic change in the ~~lateLate~~ Miocene simulations, either due to ice removal or lower mountains, generally experience significant increases in temperature (Figure 4A-C). These changes are consistent with the lapse rate cooling effect and have been found in other modelling studies of ice removal (Lunt et al., 2004; Tonazzio et al., 2004) and mountain uplift (Foster et al., 2010). However, since we have performed a number of palaeoenvironmental changes simultaneously, these simulations represent a combination of all of these changes.

#### 4.2.1 High Latitudes

Over Greenland we find a mean annual warming of the interior due to palaeogeographic change of more than 10°C as shown in Figure 4A, with a maximum warming of 8°C and 21°C for DJF and JJA respectively (Figures 4B-C). Lapse rate corrected temperature calculations reveal these temperatures to be ~3-5°C cooler than would be expected due solely to topographic lowering. In LM280, the DJF precipitation over most of Greenland is very low (<0.5 mm/day; Figure 5C) and the ice-free land surface remains largely snow-free resulting in a lower land surface albedo for LM280 than CTRL in both DJF and JJA, consistent with the small imposed ~~late~~Late Miocene Greenland ice sheet in the GCM (Figure 2). A reduction in the albedo should also lead to warming of this region. Several modelling studies report warming across the whole of Greenland, and further afield, as a result of the complete melting of the Greenland ice sheet (Crowley and Baum, 1995; Tonazzio et al., 2004; Lunt et al., 2004), whereas our study shows large areas of cooling in the North Atlantic. This could be due to the impact of an open Panama gateway (e.g. Lunt et al., 2008b), differences in runoff from a less-glaciated Greenland and from the ~~late~~Late Miocene Arctic Eurasian landmass, or it could also be caused by vegetation-climate feedback mechanisms, potentially affected by the high latitude cold bias in our GCM. Although there is little significant difference in the high latitude vegetation cover predicted by the TRIFFID model for the palaeogeographic changes we have made (Figure 6A compared to Figure 6B), there are large changes in vegetation predicted for the CO<sub>2</sub> increase (Figure 6B compared to Figure 6C). The unglaciated areas of Greenland and the high latitudes of North America and Eurasia are modelled as vegetated by TRIFFID with the dominant PFT being C3 grasses for LM280 (Figure 6B), altering to shrub for LM400 (Figure 6C). Figures 4F and 5F show that the largest climatic changes as a result of higher CO<sub>2</sub> are seen in the winter months of the high northern latitudes. Both vegetation models (TRIFFID and BIOME4) agree that the higher CO<sub>2</sub> LM400 simulation has more trees at the high northern latitudes, in Asia and in North America, than the lower CO<sub>2</sub> LM280 simulation, for which grasses are predicted.

The Antarctic orographic differences result in temperature anomalies between LM280 and CTRL that are broadly consistent with the lapse rate cooling effect, but, opposite to the Greenland case in that the changes are slightly greater than lapse rate corrected temperature calculations suggest. The high southern latitudes are generally simulated as being significantly warmer and having significantly more precipitation in LM280 than CTRL, and

these changes are more pronounced in JJA (Southern Hemisphere winter) than DJF (Southern Hemisphere summer). The Southern Ocean has significantly less sea ice in both LM280 and LM400 than in CTRL (not shown), and we infer that the reduction in the ice-albedo feedback mechanism amplifies the warming associated with topographic lowering.

#### 4.2.2 Eurasia

The model results suggest that palaeogeographic changes alter the seasonality of the ~~late~~Late Miocene climate of the south-western European countries, which is generally simulated to be warmer (with some grid boxes up to 6 degrees warmer) and drier in the summer months (Figure 4B and 5B) and cooler and wetter in the winter months (Figure 4C and 5C) for LM280 compared to the CTRL. Conversely, a reduction in seasonal temperature range is seen in LM280 compared to CTRL in the region from the Paratethys across into central-eastern Asia, and both DJF and JJA are modelled as wetter in LM280 than in CTRL. These changes are consistent with simulations of the closure of the Paratethys (e.g. Ramstein et al., 1997) and with the evidence in this region for dynamic changes in vegetation and mammal communities across the late Miocene, with the rise and fall of the Pikermian paleobiome (see Eronen et al., 2009 and references therein). There is a shift in the distribution of modelled PFTs in Central Asia from a mix of shrubs, trees and grasses in LM280 to bare soil in CTRL, causing large changes in surface albedo (not shown). There are large differences in the predicted vegetation in this locality for LM400 compared to LM280 with the majority of Eurasia being forested under the LM400 scenario (Figure 6C). ~~—There is evidence in this region for dynamic changes in vegetation and mammal communities across the Late Miocene, with the rise and fall of the Pikermian paleobiome (see Eronen et al., 2009 and references therein).~~

We calculate lapse rate corrected temperatures for this region and suggest that topographic changes in the Tibetan Plateau are largely responsible for the LM280 warming seen here in DJF. In JJA however, the temperatures are ~3°C cooler than lapse rate corrections would suggest. Our study indicates a decrease in JJA precipitation across the northern Indian Ocean and the Indian Subcontinent, indicating a weakened monsoon, and this is consistent with many modelling studies (Ramstein et al., 1997; Steppuhn et al., 2006; Lunt et al., 2008a; Zhongshi-Zhang et al., 2007a; Zhongshi-Zhang et al., 2007b). However, the ~~Tortonian~~ vegetation reconstructions for the Indian subcontinent are composed of tropical forests and savanna; which suggests an increase in precipitation relative to today (Pound et al., 2011;

2012). These reconstructions are consistent with some modelling studies (e.g. Lunt et al., 2008a; Micheels et al., 2011), which simulate large increases in precipitation in this region. It is not clear what the probable cause of these differences in our results for this region might be.

Away from the Tibetan Plateau, Southern Asia is cooler in LM280 than CTRL, in disagreement with many modelling uplift studies that suggest this region could have been ~ 5°C warmer with lower orography (Ruddiman et al., 1997; Kutzbach and Behling, 2004) and is also in disagreement with other lateLate Miocene climate simulations (Knorr et al., 2011; Micheels et al., 2011). We suggest that either the changes we have made to the Indonesian Seaway may play an important role in the climate of Southern Asia, or that this region is very sensitive to other palaeoenvironmental differences. However, our modelled Southern Asia vegetation changes are generally consistent with other modelling work (Kutzbach and Behling, 2004; Lunt et al., 2010).

#### 4.2.3 Americas

In Northwestern America, there are cooler temperatures and an increase in precipitation simulated for LM280 compared to CTRL (Figure 4 and 5). Correspondingly the PFT distribution changes from wooded vegetation in the lateLate Miocene simulations to deserts in CTRL, and this result is robust to the CO<sub>2</sub> concentration assumed (Figure 6). Alaska and northwest Canada are warmer and wetter in the lateLate Miocene simulations than CTRL. Vegetation changes here are small if only palaeogeographic changes are considered (Figure 6A and 6B), but much larger under a higher lateLate Miocene CO<sub>2</sub> assumption (Figure 6A and 6C).

Over Central America the lateLate Miocene configuration simulates colder and drier air than CTRL, with a large reduction in JJA precipitation (Figure 5). In this region C4 grasses and bare soil are simulated for the lateLate Miocene, replaced by trees in CTRL, and this finding is robust to the CO<sub>2</sub> concentration assumed for the lateLate Miocene (Figure 6). Both vegetation models agree that in modern Central America, an open Panama Gateway results in the desertification of the southern tip of North America and the north tip of South America, for both CO<sub>2</sub> concentration assumptions, and therefore we are confident that this vegetation change is due to the palaeogeographic changes made.

In South America, temperature changes between the ~~late~~Late Miocene simulations and the CTRL are small, and those seen in Figure 4C are generally associated with the differences in the land-sea mask of the two simulations. There is some warming in the region of Andean uplift for the ~~late~~Late Miocene compared to CTRL. The largest climatic changes in this region though are for precipitation, which shows an increase in seasonality with wetter winters (JJA) and drier summers (DJF). Vegetation changes here are also notable (Figure 6). The Patagonian region is simulated as dominated by the needleleaf tree PFT in LM280 and LM400, changing to desert in CTRL. The Amazon Rainforest is much reduced in our ~~late~~Late Miocene simulations compared to CTRL, replaced by C4 grass and bare soil PFTs. As similar vegetation distributions occur in LM280 and LM400, we are confident that the vegetation changes are due to the palaeogeographic alterations made. There is very little difference in the vegetation predictions for the tropics between the two CO<sub>2</sub> scenarios by TRIFFID, however, BIOME4 predicts quite large differences between the two CO<sub>2</sub> scenarios, such that the tropical forests of the lower CO<sub>2</sub> scenario are replaced by grasslands and dry shrublands, and this trend is seen in South America, but also in Africa and in SE Asia.

#### 4.2.4 Africa / Middle East

Widespread year-round ~~reductions in precipitation~~drying, an increase in seasonality and the expansion of the Sahara Desert occurs in North Africa in the ~~late~~Late Miocene simulations compared to CTRL (Figures 4, 5 and 6). The presence of the Sahara desert itself can contribute to cooling at high latitudes and the aridification / cooling of North Africa (Micheels et al., 2009b). Evidence of rooting systems in the Saharan region during the ~~late~~Late Miocene suggests the presence of some vegetation (Düringer et al., 2007) but it is suggested that this vegetation was probably seasonal in nature, with the landscape alternating between being sand-covered and vegetated due to the presence of large lakes (Düringer et al., 2007; Vignaud et al., 2002). The large area of ~~reduced precipitation that extends~~drying we ~~see extending~~ from the mid-Pacific Ocean into North Africa, are in direct contrast to the Lunt et al. (2008a) study which shows large increases in precipitation in these regions. In the Micheels et al. (2007) work, the prescribed vegetation assumed in the Sahara (grasslands and savannahs) in their TORT simulations is responsible for up to 6°C of warming, and increases in precipitation in North Africa.

All of the areas predicted to be bare soil ~~are~~ by TRIFFID are predicted to be slightly more vegetated by BIOME4.

#### 4.2.5 Australia

Australia is drier in winter for the ~~late~~Late Miocene simulations compared to CTRL; the band of increased precipitation for DJF, Figure 5F, is most likely an artefact of the different land-sea masks in the two simulations. The palaeorecord provides evidence for the ~~drying~~ aridification of central Australia as a result of continental drift (Truswell, 1993; McGowran et al., 2004) which is perhaps consistent with our simulated precipitation results.

The TRIFFID model results clearly show vegetation changes as Australia shifts northwards between the ~~late~~Late Miocene and CTRL simulations. In the ~~late~~Late Miocene experiments, Australia is dominated by C4 grass and bare soil PFTs, largely replaced by the broadleaf tree PFT in the CTRL (Figure 6).. However the interpretation of the TRIFFID results, and analysis of the role of palaeogeography or CO<sub>2</sub> in this region are hampered by the inconsistencies between the vegetation modelled by TRIFFID and the modern Australian vegetation distribution.

### 5 Model-Data Comparison Results and Discussion

In this section we compare our simulations with the ~~late~~Late Miocene terrestrial proxy data reconstructions (hereafter referred to as LMdata), to assess the extent to which solely changing the paleogeography to that appropriate for the ~~late~~Late Miocene results in a better consistency ~~to~~-with this dataset, and to determine the sensitivity of those results to the uncertainty in the paleoCO<sub>2</sub> record.

We compare LMdata, categorised into the two stages of the Messinian and the Tortonian, with the climatologies of: a) the bias corrected CTRL simulation, equivalent to our modern potential climate estimates; b) the bias corrected LM280 simulation; and c) the bias corrected LM400 simulation. The extent to which comparison b) is better than comparison a) indicates the importance of palaeogeographic changes in determining ~~late~~Late Miocene climate. The extent to which comparison c) is better than comparison b) indicates the importance of CO<sub>2</sub> uncertainties in determining ~~late~~Late Miocene climate.

The numerical results from the model-data comparisons are given in Tables ~~23~~ and ~~34~~, and shown in Figures 7 to 16.

## 5.1 Mean Annual Temperature

### 5.1.1 Late Miocene Data Compared to Modern Potential Natural Climate Estimates

The differences in MAT we find between the ~~late~~Late Miocene paleodata and our modern climate estimates are shown in Figure 7. The comparison shows that, where we are confident that the MATs suggested by the palaeodata are different to our modern climate MAT estimates, they are significantly warmer; no datapoint is suggesting cooler temperatures in the ~~late~~Late Miocene. There are ~~129~~456 overlaps from a total of ~~429~~479 datapoints (see Table ~~23~~). There is a clear difference in the microfaunal MAT reconstructions between the Tortonian and the Messinian (Figure 7F compared to Figure 7C). These results suggest that the Mediterranean Basin was much warmer in the Tortonian than today, but by the Messinian we are unable to confidently say that the MAT was any different to today. The same signal is perhaps apparent in the macroflora data in south-western Europe and in South America (Figure 7E compared to Figure 7B) and in the microflora data in south-western Europe (Figure 7D compared to Figure 7A).

### 5.1.2 Late Miocene Data Compared to the ~~late~~Late Miocene simulations

Figure 8 and Table ~~23~~ show how well our LM280c simulation compares to the ~~late~~Late Miocene MAT data reconstructions. Figure 8 clearly demonstrates that we are unable to generate the warm MATs suggested by the data for either the Messinian or the Tortonian by just changing the palaeogeography alone, and this finding is robust to the choice of proxy with microflora, macroflora and microfaunal proxies indicating the same results (Table ~~2~~: ~~1173~~:436 overlaps from ~~429~~479 datapoints).

The LM400c simulation, which includes both a palaeogeographic change and a CO<sub>2</sub> change, provides a better match to the data overall (Figure 9, Table ~~3~~: ~~2714~~:297 overlaps from ~~429~~479 datapoints), with overlap between the modelled MATs and all but one of the Tortonian faunal-based MAT reconstructions (Figure 9F). It is also noticeable that the model-data comparison for the Messinian aged microflora datapoints results in many more overlaps than the similar comparison with Tortonian aged data (Figure 9A compared to Figure 9D). Despite these improvements in the model-data comparison, there are still many datapoints which indicate warmer MATs than the LM400c simulation is able to model.

~~In~~The summary, Figures 10A and 10B show that there are regions where the palaeogeographic changes have resulted in improvements in the model-data comparison (e.g. north-western America, mid-western Europe) and regions where the changes have resulted in

deteriorations (e.g. northern and southern extents of the European data, south-east Asia, Australia). Table [23](#) details that the palaeogeographic changes result in a simulation that overlaps less with the MAT data than the modern potential natural climate estimates overlap (a reduction of [1220](#) overlaps, or [34%](#)), but that overall the modelled MATs are closer to the data reconstructions (13% better) due to improvements in the distance to overlap, mostly seen in the model-data comparison of Tortonian-aged datapoints.

Figures 10C and 10D demonstrate the impact that CO<sub>2</sub> uncertainty has on the model-data comparison. They clearly show that the MATs of LM400c are an improvement in their match to the data reconstructions than the MATs of LM280c (Table [3: 2254-232](#) datapoints are closer to model-data overlap in the LM400c than the LM280c - an improvement of [5248%](#) overall, and a [3634%](#) improvement in the number of overlaps). However, the data are very spatially biased towards Europe, and Figures 10C and 10D also show that outside of Europe and mid-eastern Asia, the model-data comparison actually deteriorates as a result of the increased CO<sub>2</sub> concentration.

These results suggest that changing the palaeogeography to that appropriate for the [lateLate](#) Miocene and increasing the CO<sub>2</sub> concentration from that of the modern potential natural climate together results in a model simulation that is able to reproduce much of the warm European MATs (Figure 9), but that it is unable to entirely reproduce the warm MATs indicated by the [lateLate](#) Miocene data elsewhere.

## 5.2 Mean Annual Precipitation

### 5.2.1 Late Miocene Data Compared to Modern Potential Natural Climate Estimates

The comparison between the [lateLate](#) Miocene data and the modern climate estimates for precipitation are shown in Figure 11. The MAP data reconstructions suggest that where we are confident that the climate of the [lateLate](#) Miocene is different to a potential natural modern one, the MAPs are wetter; none of the datapoints suggest drier MAPs in the [lateLate](#) Miocene. There are a large number of overlaps between the [lateLate](#) Miocene MAP data and the modern MAP estimates (Table [2: 8523-1037](#) overlaps from a total of [10524253](#) datapoints). The faunal Tortonian data reconstruct a much wetter climate in Europe (Figure 11F), whereas the Messinian data reconstructions in this region are largely similar to today (Figure 11C). A similar pattern is perhaps seen in the microflora and macroflora MAP reconstructions in this area (Figure 11D compared to Figure 11A; Figure 11E compared to Figure 11B).

## 5.2.2 Late Miocene Data Compared to the ~~late~~Late Miocene simulations

The LM280c simulation compares well to the reconstructed MAPs (Figure 12; Table ~~2: 9643: 1152~~ overlaps from ~~1052+253~~ datapoints), and of the macroflora reconstructions, just ~~52~~ datapoints do not overlap (Figure 12B and 12E; Table ~~23~~). The ~~88+01~~ datapoints which do not overlap with the modelled MAPs are wetter than the model predicts.

There is a significant improvement in the model-data comparison for MAPs when compared with the modern climate comparison (Figure 12 compared to Figure 11; Table ~~2: 1923: 240~~ improvements versus just ~~2327~~ deteriorations) suggesting that changing the palaeogeography alone takes the modelled MAPs closer to the reconstructed MAPs.

Although the LM400 simulation predicts generally an increase in the MAP as compared to the LM280 simulation (Figure 5D-F), both the simulations result in MAP predictions that are within the uncertainty of the ~~late~~Late Miocene MAP reconstructions, and therefore little change is seen in the model-data comparison for this variable between the two simulations (Figure 13 compared to Figure 12; Table ~~3: 9644: 1152~~ overlaps for LM280c compared with ~~965+158~~ for LM400c). This suggests that the palaeogeographic changes we have made are the driving force behind the modelled precipitation changes observed, and that the CO<sub>2</sub> concentration plays only a minor role. However, Figure 14 demonstrates that many of the datapoints overlap both the LM280c modelled MAPs and those of our modern potential natural climate. In fact, Figures 14C and 14D demonstrates that many of the datapoints in western Europe are further from overlap with the LM400c modelled MAPs than they are with the LM280c modelled MAPs, indicating that higher CO<sub>2</sub> may actually worsen model-data comparison in this region.

## **5.3 Mean Cold Month Temperature**

### **5.3.1 Late Miocene Data Compared to Modern Potential Natural Climate Estimates**

The comparison between the late Miocene data and the modern climate estimates for the CMT are shown in Supplementary Figure S1. The CMT data reconstructions suggest that where we are confident that the climate of the late Miocene is different to a potential natural modern one, the CMTs are warmer; none of the datapoints suggest cooler CMTs in the late Miocene. There are many overlaps between the late Miocene CMT data and the modern CMT estimates (Table 2; 235 overlaps from a total of 332 datapoints).

### 5.3.2 Late Miocene Data Compared to the late Miocene simulations

The LM280c simulation compares reasonably well to late Miocene CMT data (Supplementary Figure S2; Table 2; 289 overlaps from a total of 332 datapoints) and the LM400c simulation compares even better, especially to the microflora reconstructions (Supplementary Figure S3; Table 3; 312 overlaps from a total of 332 datapoints). The palaeogeographic changes made account for a 16% improvement in the number of overlaps and the CO<sub>2</sub> concentration changes account for an additional 7% improvement as compared to the modern climate estimates (Supplementary Figure S4; Tables 2 and 3).

## 5.4 Mean Warm Month Temperature

### 5.4.1 Late Miocene Data Compared to Modern Potential Natural Climate Estimates

The comparison between the late Miocene data and the modern climate estimates for the WMT are shown in Supplementary Figure S5. With the exception of south-east Asia, the WMT data reconstructions suggest that the WMTs were significantly warmer in the late Miocene than in the modern. There are few overlaps between the late Miocene WMT data and the modern WMT estimates (Table 2; just 30 overlaps from a total of 332 datapoints).

### 5.4.2 Late Miocene Data Compared to the late Miocene simulations

The LM280c simulation fails to reproduce the warmer temperatures reconstructed by the WMT data for the late Miocene (Supplementary Figure S6; Table 2; 48 overlaps from a total of 332 datapoints) but the LM400c simulation compares better, particularly for the western Mediterranean (Supplementary Figure S7). The palaeogeographic changes made account for a 5% improvement in the number of overlaps and the CO<sub>2</sub> concentration changes made account for an additional 14% improvement (Supplementary Figure S8; Tables 2 and 3).

## 5.5 Megabiomes

### 5.5.1 Late Miocene-Tortonian Data Compared to Modern Potential Natural Control Simulation

Figures 15A and 15D show the comparison between the megabiomes derived from the BIOME4 model driven by the modern control simulation CTRLc and the Messinian data and Figures 15G and 15J show the comparison between the megabiomes derived from the BIOME4 model driven by the modern control simulation CTRLc and the Tortonian data. There are some large vegetation shifts are documented at the biome level (Pound et al., 2011; 2012), and when aggregated to megabiome level, the vegetation shifts between the late

~~MioceneTortonian~~ and today are still noticeable, and in fact there are only ~~22996~~ overlaps from a total of ~~556225~~ datapoints (Table ~~23~~). There are many regions where there are significant differences in the megabiomes predicted for the ~~late MioceneTortonian~~ and today, including north-west America and north-east Asia, Greenland and northern Europe, the Mediterranean Basin, central Asia, southern and south-eastern Asia, South America, South Africa and in Australia/~~New Zealand~~.

#### ~~5.3.25.5.2~~ ~~Late MioceneTortonian~~ Data Compared to the ~~lateLate~~ Miocene simulations

Figures 15B and 15E show the comparison between the ~~Messinian~~ megabiomes ~~of the Tortonian data~~ and those predicted by ~~BIOME4 model driven by~~ the LM280c simulation ~~and~~ Figures 15H and 15K show the comparison between the Tortonian megabiomes and those predicted by BIOME4 model driven by the LM280c simulation. Little difference is seen between the comparison with the Messinian age data and the Tortonian age data. As discussed in Section 4.2.3, the BIOME4 model predicts quite a significant change in the vegetation of Central America, ~~however the megabiome data do not support this change (Figure 15B compared to 15A and 15H compared to 15J), suggesting that perhaps the response of the model to the open Panama Gateway is too strong, or that our simulated Panama Gateway is too wide, but unfortunately there are no datapoints in this region to verify that change.~~ The LM280c simulates a change from the modern tropical forests of central Africa to grasslands and dry shrubland, but, ~~unfortunately again,~~ there are no datapoints in this region in order to verify this change. The LM280c simulation also results in a shift from the modern ~~borealBoreal~~ forest to tundra in north-eastern Asia but the datapoints in this region do not support this change ~~(Figure 15B/15E and 15H/15K).~~ Compared to the CTRLc simulation, the LM280c simulation does not improve the model-data comparison for most datapoints, and in fact overall the comparison worsens very slightly (Figure 16A ~~and 16C~~; Table ~~2: 343-45~~ improvements versus ~~4149~~ deteriorations, a worsening of ~~12~~%).

Figures 15C and 15F show the comparison between the ~~Messinian~~ megabiomes ~~and the LM400c simulation and Figures 15I and 15L show the comparison between~~ of the Tortonian ~~megabiomes data~~ and the LM400c simulation. The megabiomes are in agreement for many of the datapoints of northwest America and north and east Asia that disagreed with the CTRLc simulation (Figure 15F, ~~and to a lesser extent Figure 15L).~~ There are ~~277424~~ agreements between the megabiomes reconstructed by the data and the LM400c simulation, from ~~556225~~ datapoints in total. The ~~279404~~ datapoints that disagree with this simulation are located

mainly in central, southern and south-eastern Asia, and the land surrounding the Mediterranean Basin. When compared to the LM280c simulation, the model-data comparison worsens in South America in particular (Figure 16B and 16D). Overall though, the LM400c simulation offers a ~~1013~~% improvement over the LM280c (Table 34).

#### 5.45.6 Discussion

We ask the question to what extent the different palaeogeography of the ~~lateLate~~ Miocene can explain the differences in the climate documented in the palaeorecord, and how important ~~is~~ the uncertainty in the atmospheric CO<sub>2</sub> concentration?~~-is-~~

The results from the model-data comparison for the ~~MATmean-annual-temperature~~ are difficult to interpret. Table ~~23~~ details that a large number of improvements and a large number of deteriorations exist as a result of the palaeogeographic changes made. The Messinian data generally suggests that the palaeogeographic changes are not able to improve the model-data comparison as compared to the modern potential climate estimates but the Tortonian data generally suggests the opposite. However, overall there are more datapoints closer to overlap with the ~~lateLate~~ Miocene simulation LM280c than the modern potential natural climate estimates, but ~~that~~ the total number of overlaps reduces as a result of palaeogeographic change (Figure ~~1718~~). However, Figure ~~1718~~ demonstrates that the atmospheric CO<sub>2</sub> concentration is very important in determining the ~~MATsmean-annual-temperatures~~ of the ~~lateLate~~ Miocene, with large improvements seen for both the Messinian and Tortonian model-data comparison as a result of increasing the assumed CO<sub>2</sub> concentration. Furthermore, the higher CO<sub>2</sub> assumption combined with the palaeogeographic changes made in the LM400c simulation offers a significant improvement in the model-data comparison when compared to the modern potential natural climate estimates with nearly twice as many overlaps with the datapoints (Table 3). This finding is consistent with similar model-data comparison work for the preceding middle Miocene (Krapp and Jungclaus, 2011), and is also consistent with the model-data comparison for the CMT and WMT (Figure 184). In order to test the robustness of this result, an additional comparison is made with the CTRL simulation with 400 ppm atmospheric CO<sub>2</sub> (hereafter referred to as CTRL400c). Supplementary Figure S9 shows that the CTRL400c simulation has a comparable (and, in fact, slightly higher) number of overlaps with the late Miocene data to the LM400c simulation, and 172 more overlaps than the modern potential natural climate estimates, further supporting the importance of atmospheric CO<sub>2</sub> in determining the MATs.

Figure 1718 shows that the different ~~lateLate~~ Miocene palaeogeography is important in determining the different ~~MAPmean annual precipitation~~ for both the Messinian and the Tortonian stages, because all of the proxy reconstructions considered show an improvement in the model-data comparison for the ~~lateLate~~ Miocene simulation as compared to the modern potential climate estimates (Table 23). Figure 1718 however suggests that the uncertainty in the atmospheric CO<sub>2</sub> concentration is not important in determining the different ~~MAPmean annual precipitation~~ documented for the ~~lateLate~~ Miocene, as only very small differences are found between the results of LM400c and LM280c (Table 34). The locations where LM280c compares better to the MAP reconstructions than the modern potential climate estimates are spread across the spatial range of the dataset (N. America, Europe, Asia, Africa), and therefore it is difficult, without the use of further sensitivity studies, to relate these improvements definitively to a particular paleogeographic change. Although an open Panama Gateway can be linked to reduced precipitation across most of Europe and Asia, it can also be linked to a pocket of increased precipitation centred over modern day Myanmar (see Figure 9 of Lunt et al., 2008b), which is consistent with our LM280c simulation and is the locality for some of the improvements obtained in the MAP model-data comparison. We can, however, probably rule out both the retreat of the Paratethys Sea and the uplift of the Himalayas for the southern and eastern Asian improvements in the MAP model-data comparison, as these paleogeographic changes have been shown to result in increased precipitation in India and China (e.g. Ramstein et al., 1997; Lunt et al., 2010; Zhang et al., 2007a; Zhang et al., 2007b). These changes could, however, be related to some of the improvements seen in western Asia (refer to Figure 6 of Ramstein et al., 1997 and Figure 2b and 2c of Fluteau et al., 1999) and in Europe (refer to Figure 2b and 2c of Fluteau et al., 1999). The further improvements in the MAP model-data comparison across Asia could therefore be related to changes made to the Indonesian Seaway, and the improvements across Europe related to the elevation changes made in Europe itself (e.g. Henrot et al., 2010) and also perhaps related to the additional late Miocene eurasian landmass. Further sensitivity studies are required to test these hypotheses.

The ~~Tortonian~~ megabiome results show very little difference in the model-data comparison with the ~~lateLate~~ Miocene simulation LM280c as compared to the model-data comparison with the control simulation CTRLc. However, thereThere is ~~also~~ an improvement in the megabiome model-data comparison for the higher CO<sub>2</sub> simulation as compared to the lower

CO<sub>2</sub> simulation, again suggesting that atmospheric CO<sub>2</sub> concentration is an important consideration.

Our modelling results are in general agreement with those of other similar studies (Knorr et al., 2011; Micheels et al., 2011) in that palaeogeographic changes for the ~~late~~Late Miocene causes regions of warming in the areas of ~~tectonic orographic~~ change, and regions of cooling, particularly in the North Atlantic. A recent study has found that vegetation changes are very important in determining the ~~late~~Late Miocene climate, perhaps three times more important than the palaeogeographic changes (Knorr et al., 2011). They find reasonable agreement between their ~~late~~Late Miocene model simulation with preindustrial atmospheric CO<sub>2</sub> concentrations and the palaeorecord when the vegetation used in the model is prescribed, thereby reconciling a warmer climate with low CO<sub>2</sub> concentrations. ~~We do not find the same results with our GCM. Although there are many differences, the main difference between our study and that of Knorr et al. (2011) is that we use an interactive vegetation model dynamically linked to our GCM, whereas a Miocene vegetation reconstruction based on palaeodata is imposed in their work.~~ The benefit of a coupled GCM-vegetation model is that the simulated vegetation distribution will be in equilibrium with the modelled climate, but the disadvantage is that the simulated vegetation will not match the palaeodata if the climate is incorrectly modelled. ~~Based on our results, we hypothesise that the vegetation reconstruction imposed in the Knorr et al. (2011) study is not in equilibrium with their simulated climate and that a higher than preindustrial level of CO<sub>2</sub> is required in order for the vegetation distributions recorded in the palaeorecord to have occurred.~~ The relative impacts of vegetation changes versus CO<sub>2</sub> changes from this model, and the implications for the findings presented here, will be explored in future work.

The megabiome model-data comparison results for LM400c show that some of the documented extreme changes in vegetation are captured by the BIOME4 model when forced by this simulation. BIOME4/LM400c predicts temperate forests in what is now the circumboreal region, consistent with Denk et al. (2005) and Worobiec and Lesiak (1998); and grasslands in the Arabian peninsula, consistent with Kingston and Hill (1999). However, despite the improvements that we find in the model-data comparison for changing palaeogeography and for changing the CO<sub>2</sub> assumption, significant model-data inconsistencies remain~~there are still significant model data inconsistencies, particularly in eastern Asia for the mean annual temperature in both the Messinian and the Tortonian (as~~

with the model data comparison of Knorr et al. (2011)). Further verification of this discrepancy is provided by the inconsistency in both the Tortonian mean annual temperature and the Tortonian megabiome reconstructions in this region, however there is consistency between the Late Miocene simulations and the data reconstructions for the Asian MAPs. We therefore suggest that future work with this GCM should concentrate on the sensitivity of the climate in this region to the various tectonic changes independently in order to isolate the impact that each change has.

## 6 Summary and Conclusions

### 6.1 Summary

We have simulated two potential ~~lateLate~~ Miocene climates using the fully coupled atmosphere-ocean-vegetation GCM model HadCM3L-TRIFFID. Both simulations assume a paleogeography appropriate for the ~~lateLate~~ Miocene, but one has preindustrial levels of CO<sub>2</sub> at 280ppm and the other a slightly higher CO<sub>2</sub> concentration at 400ppm. We compare the 280ppm ~~lateLate~~ Miocene simulation against a modern control simulation with the same preindustrial level of CO<sub>2</sub> in order to investigate the potential role of palaeogeography in determining the simulated ~~lateLate~~ Miocene climate. We also compare the two ~~lateLate~~ Miocene climate simulations to investigate the role of CO<sub>2</sub> in determining the simulated climates.

The global anomaly between the two 280ppm climatologies is close to zero for both ~~MAT~~mean annual temperature and ~~MAP~~precipitation, but there are significant regional differences. Our ~~lateLate~~ Miocene palaeogeographic changes have been found to alter the climate: for example, there is a large area of cooler drier air that spreads across the North Atlantic Ocean into Eurasia for the ~~lateLate~~ Miocene, the lapse rate effect on temperature results in regional warming in areas of topographic lowering for the ~~lateLate~~ Miocene, and significant reductions in ~~lateLate~~ Miocene monsoonal precipitation are modelled.

A database containing ~~10304,109~~ terrestrial proxy records for the ~~lateLate~~ Miocene has been compiled from the literature. From these data we infer a substantially warmer and wetter ~~lateLate~~ Miocene climate as compared to a potential modern natural climate. We have made quantitative comparisons of these data and ~~theTortonian~~ biome data (Pound et al., 2011; 2012), to both our ~~lateLate~~ Miocene simulations and our potential modern natural climate estimates, in order to investigate the extent to which palaeogeographic changes alone can

1 | explain ~~late~~Late Miocene climate and to examine the importance of the uncertainty in the  
2 | CO<sub>2</sub> paleorecord.

3 | We have used proxy data from a variety of different reconstruction methods, and overlap  
4 | between different methods of climate reconstruction used on the same proxy data will  
5 | indicate reliability, but ideally overlap between entirely independent proxies is desirable. We  
6 | therefore also recommend, where possible, the expansion and better temporal constraint of  
7 | the multiproxy palaeorecord for the ~~late~~Late Miocene, in particular for the regions where the  
8 | model-data inconsistencies are largest, and a more robust and consistent assessment of  
9 | uncertainties in all proxy records. We have tried to account as much as possible for  
10 | uncertainties in the proxy reconstructions so we have some confidence that the reconstructed  
11 | signals are robust. However, we cannot rule out significant errors in our palaeogeographic  
12 | reconstruction and/or the model itself. We will conduct future modelling with the same  
13 | coupled atmosphere-ocean-vegetation GCM in order to establish sensitivity to our choice of  
14 | palaeogeographic configuration on the simulated climate, and also the relative roles of CO<sub>2</sub>  
15 | concentration versus vegetation distribution.

## 17 | 6.2 Conclusions

18 | Despite the uncertainties in proxy data reconstructions and the limited distribution and  
19 | spatial/temporal bias of the palaeorecord, we do find significant evidence to suggest that the  
20 | paleogeography ~~does~~can have an important role in determining aspects of the ~~late~~Late  
21 | Miocene climate, particularly for the ~~MAP~~mean annual precipitation for both the Messinian  
22 | and the Tortonian. However, the uncertainty in the palaeo-CO<sub>2</sub> record is also important  
23 | because the sensitivity model simulation, CTRL400c suggests that both the palaeogeography  
24 | and the atmospheric CO<sub>2</sub> contribute to determining the MAPs, but that they do not add  
25 | linearly. The megabiome model-data comparisons show some improvements as a result of  
26 | the palaeogeographic alterations on the west coast of North America, in Australasia, south-  
27 | east Europe and eastern Asia. We find significantly better model-data consistency for the  
28 | higher CO<sub>2</sub> ~~late~~Late Miocene simulation than for the lower CO<sub>2</sub> simulation for the both  
29 | Messinian and Tortonian ~~MAT~~mean annual temperatures but only modest differences for  
30 | the ~~MAP~~mean annual precipitation of both stages. We therefore conclude that the ~~MAP~~mean  
31 | ~~annual precipitation~~ differences ~~are~~can be driven by either the palaeogeographic changes ~~that~~  
32 | ~~we have made,~~ or by the atmospheric CO<sub>2</sub> changes made. ~~–However, given~~but that the

temperature differences are very strongly influenced by the CO<sub>2</sub> concentration it is perhaps likely that late Miocene atmospheric CO<sub>2</sub> concentrations were generally nearer the higher end of the range of reconstructions, and that palaeogeography plays a much more localised role in determining late Miocene temperatures.

## **Appendix A. Uncertainty in late Miocene Proxy Reconstructions**

~~As a multi-proxy dataset is used, there is considerable variability in the age constraint due to uncertainty in the dating. Although the data from small mammals and microfloras are generally very well dated, large mammals and macrofloras have much looser age-determination and many macroflora datapoints are only consigned to the Messinian or the Tortonian. Therefore the data is considered in 2 timeslabs: the Messinian (5.3-7.25 Ma) and the Tortonian (7.25-11.6 Ma). We note that the proxy data used is likely to have been generated under a range of different orbital configurations and other climatic cycles, and individual datapoints could therefore represent extremes in these cycles, but this is unlikely to have biased the entire dataset. There is some evidence that vegetation proxies give slightly higher estimates than palaeozoological climate reconstructions (See Eronen et al. 2012). In the dry parts of climate cycles, the preservation potential of plant remains and organic matter in general is low (for discussion see Bruch et al. 2011, Eronen et al. 2012), and even palynomorphs consisting of comparatively chemically resistant sporopollenine may not outlast highly oxidizing conditions. Hence, the fossil plant record might represent the wetter part of a climate cycle. On the other hand, deposits like coarser clastic successions favor preservation of fossil mammals in arid habitats over closed forest.~~

~~For a comparison study between fossil data and GCM simulations where a palaeontological assemblage has been used to reconstruct climatic conditions, such as the MAT or MAP, two main uncertainties exist in the reconstructions: taphonomical (the death, decay, deposition and preservation of fossils) and the uncertainties associated with the data reconstruction techniques. Below we discuss comparisons between different techniques. For a full details on each technique and detailed taphonomical biases in each case, the reader is referred to the original publications: (Taphonomy: Western and Behrensmeyer, 2009; Martin, 1999; Behrensmeyer et al., 2000); (Data reconstruction techniques: Mosbrugger and Utescher, 1997; Fauquette et al., 2006; Wolfe, 1993; Spicer, 2007; Spicer et al., 2009; Eronen et al., 2010; Jacobs and Deino, 1996; Montuire et al., 2006; Böhme et al., 2008; van Dam, 2006).~~

~~The majority of palaeobotanical climate reconstructions used in this study come from two techniques, the co-existence approach and CLAMP. The co-existence approach uses a Nearest Living Relative (NLR) philosophy; the climate tolerance of a fossil taxons closest extant relative is likely to be comparable to that of the fossil (Mosbrugger and Utescher, 1997). From this idea an assemblage of fossils can be compared to find the overlapping climate space that all taxa could have survived in, based on the climate tolerances of each fossils nearest living relative (Mosbrugger and Utescher, 1997). CLAMP is the latest iteration of the plant physiognomy philosophy that has been known for~~

1 ~~almost a century; plants vary their structure with climate, in a non-random way (Bailey and Sinnott,~~  
2 ~~1915; Holdridge, 1947). CLAMP uses leaf morphology to reconstruct quantitative climate information~~  
3 ~~using "training sets" from undisturbed modern day floras (Spicer et al., 2009; Yang et al., 2011). It is~~  
4 ~~known that the leaf physiognomic techniques (e.g. CLAMP) produce lower MAT estimates than the~~  
5 ~~coexistence approach (Mosbrugger & Utescher, 1997; Uhl et al., 2003, 2006, 2007). Reasons for this~~  
6 ~~are still unknown (but see Uhl 2007 for discussion). On the other hand, the coexistence approach~~  
7 ~~produces estimates that are comparable to each other, which limits the bias in our dataset. For more~~  
8 ~~extensive comparison between different palaeobotanical proxy methods, see Thiel et al. (2012).~~

9 ~~The palaeozoological climates reconstructions come from very different techniques. For small~~  
10 ~~mammals the basis is taxonomy: Species richness is related to climate parameters for small~~  
11 ~~mammals (Montuire et al., 2006; van Dam, 2006; Böhme et al., 2008). For large mammals the~~  
12 ~~reconstruction technique is based on traits that are not determined by taxonomy, but by morphology:~~  
13 ~~The tooth morphology of herbivores is related to the vegetation abrasiveness and other wearing~~  
14 ~~agents. These are mainly controlled by water stress and climate, and therefore these characteristics~~  
15 ~~can be used to derive palaeoprecipitation estimates from large herbivores (Eronen et al., 2010).~~

16 ~~As the climate reconstruction techniques are derived from different sources, there are both age and~~  
17 ~~location uncertainties to consider in making a comparison between the estimates of each technique.~~  
18 ~~It is fairly rare to have cases where multiple evidence sources are available. There is some indication~~  
19 ~~that at the locality level that these techniques produce overall agreement (see discussion in Van Dam~~  
20 ~~2006, p. 200). The initial results from a larger comparison between large mammal estimates and~~  
21 ~~vegetation proxies using pooled regional data also show overall agreement (Eronen et al. 2012, but~~  
22 ~~see discussion above).~~

23 As a multi-proxy dataset is used, there is considerable variability in the age constraint due to  
24 uncertainty in the dating and many datapoints are only consigned to the Messinian or the  
25 Tortonian. Ages used were from Pangaea (<http://www.pangaea.de>) or the original paper  
26 where available, otherwise we used ages for mammal zones or geological stages as given in  
27 the Paleobiology Database (<http://paleodb.org>) or in Bernor et al. (1993). Age data for large  
28 mammal data that derive from the NOW database follows the NOW database dating scheme,  
29 based on Steininger et al. 1999. The data is considered in 2 timeslabs: the Messinian (5.3-  
30 7.25 Ma) and the Tortonian (7.25-11.6 Ma), and any datapoints that overlap both stages in  
31 their age uncertainty were duplicated, one datapoint placed in each stage. Likewise, any  
32 datapoints with age uncertainty ranges spanning the Messinian into the Pliocene were placed  
33 in our Messinian database and any datapoints with age uncertainty ranges spanning the  
34 Tortonian into the middle Miocene were placed in our Tortonian database. We note that the  
35 proxy data used is likely to have been generated under a range of different orbital  
36 configurations and other climatic cycles, and individual datapoints could therefore represent

1 extremes in these cycles, but this is unlikely to have biased the entire dataset. There is some  
2 evidence that vegetation proxies give slightly higher estimates than palaeozoological climate  
3 reconstructions (see Eronen et al. 2012). In the dry parts of climate cycles, the preservation  
4 potential of plant remains and organic matter in general is low (for discussion see Bruch et al.  
5 2011, Eronen et al. 2012), and even palynomorphs consisting of comparatively chemically  
6 resistant sporopollenine may not outlast highly oxidizing conditions. Hence, the fossil plant  
7 record might represent the wetter part of a climate cycle. On the other hand, deposits like  
8 coarser clastic successions favor preservation of fossil mammals in arid habitats over closed  
9 forest.

10 For a comparison study between fossil data and GCM simulations where a palaeontological  
11 assemblage has been used to reconstruct climatic conditions, such as the MAT or MAP, two  
12 main uncertainties exist in the reconstructions: taphonomical (the death, decay, deposition  
13 and preservation of fossils) and the uncertainties associated with the data reconstruction  
14 techniques. Below we discuss comparisons between different techniques. For a full details on  
15 each technique and detailed taphonomical biases in each case, the reader is referred to the  
16 original publications: (Taphonomy: Western and Behrensmeyer, 2009; Martin, 1999;  
17 Behrensmeyer et al., 2000); (Data reconstruction techniques: Mosbrugger and Utescher,  
18 1997; Fauquette et al., 2006; Wolfe, 1993; Spicer, 2007; Spicer et al., 2009; Eronen et al.,  
19 2010; Jacobs and Deino, 1996; Montuire et al., 2006; Böhme et al., 2008; van Dam, 2006).

20 The majority of palaeobotanical climate reconstructions used in this study come from two  
21 techniques, the co-existence approach and CLAMP. The co-existence approach uses a  
22 Nearest Living Relative (NLR) philosophy; the climate tolerance of a fossil taxons closest  
23 extant relative is likely to be comparable to that of the fossil (Mosbrugger and Utescher,  
24 1997). From this idea an assemblage of fossils can be compared to find the overlapping  
25 climate space that all taxa could have survived in, based on the climate tolerances of each  
26 fossils nearest living relative (Mosbrugger and Utescher, 1997). CLAMP is the latest iteration  
27 of the plant physiognomy philosophy that has been known for almost a century; plants vary  
28 their structure with climate, in a non-random way (Bailey and Sinnott, 1915; Holdridge,  
29 1947). CLAMP uses leaf morphology to reconstruct quantitative climate information using  
30 "training sets" from undisturbed modern day floras (Spicer et al., 2009; Yang et al., 2011). It  
31 is known that the leaf physiognomic techniques (e.g. CLAMP) produce lower MAT estimates  
32 than the coexistence approach (Mosbrugger & Utescher, 1997; Uhl et al., 2003, 2006, 2007).

Reasons for this are still unknown (but see Uhl 2007 for discussion). On the other hand, the coexistence approach produces estimates that are comparable to each other, which limits the bias in our dataset. For more extensive comparison between different palaeobotanical proxy methods, see Thiel et al. (2012)

The palaeozoological climates reconstructions come from very different techniques. For small mammals the basis is taxonomy: Species richness is related to climate parameters for small mammals (Montuire et al., 2006; van Dam, 2006; Böhme et al., 2008). For large mammals the reconstruction technique is based on traits that are not determined by taxonomy, but by morphology: The tooth morphology of herbivores is related to the vegetation abrasiveness and other wearing agents. These are mainly controlled by water-stress and climate, and therefore these characteristics can be used to derive palaeoprecipitation estimates from large herbivores (Eronen et al., 2010).

As the climate reconstruction techniques are derived from different sources, there are both age and location uncertainties to consider in making a comparison between the estimates of each technique. It is fairly rare to have cases where multiple evidence sources are available. There is some indication that at the locality level that these techniques produce overall agreement (see discussion in Van Dam 2006, p. 200). The initial results from a larger comparison between large mammal estimates and vegetation proxies using pooled regional data also show overall agreement Eronen et al. 2012, but see discussion above).

## **Appendix B. Model Details**

### **1.1 HadCM3L GCM**

#### **1.1.1 Description**

The model contains many parameterizations of atmospheric and oceanic processes, including atmospheric convection (Gregory and Rowntree, 1990), boundary layer mixing (Smith, 1993), ocean layer mixing (Gent and McWilliams, 1990), and radiation (Edwards and Slingo, 1996). HadCM3L has been used in studies of the future (e.g. Cox et al., 2000) and the past (e.g. Lunt et al., 2007; Lunt et al., 2010). As HadCM3L has a reduced ocean resolution, model performance is degraded compared to HadCM3. For example, the ocean resolution of HadCM3L is such that the Denmark Straits are unresolved and therefore the two terrestrial grid points representing Iceland are removed to prevent an unacceptable build up of sea ice in

1 the Nordic Sea (Jones, 2003). The version of HadCM3L used here does not therefore require  
2 the inclusion of flux corrections to maintain a stable overturning circulation for the modern  
3 ocean (Jones, 2003).

4 The model land surface processes are simulated using the MOSES-2.1 (Met Office Surface  
5 Exchange Scheme) land surface scheme (Essery and Clark, 2003). There are nine surface  
6 types, five of which correspond to plant functional types (PFTs): broadleaf tree, needleleaf  
7 tree, C<sub>3</sub> grass, C<sub>4</sub> grass and shrub; the remainder representing bare soil, water bodies, ice and  
8 urban surfaces.

### 9 1.1.2 Uncertainties

10 No GCM is able to reproduce exactly the modern climate, and a discussion of the biases in  
11 the HadCM3 version of our GCM is given in Gordon et al. (2000). The HadCM3L version of  
12 the model suffers from temperature biases, particularly for too-cool ocean temperatures at  
13 high latitudes. Temperatures over the land surface are generally within the 1-2°C of the  
14 uncertainty of the CRU-TS 3.0 modern instrumental data (Mitchell and Jones, 2005), but  
15 there is still a marked cold bias at high northern latitudes of up to 10°C above 70°N and the  
16 regions of highest orography (Himalayas and Andes) are too warm by >3°C. Similar biases  
17 exist for precipitation, with the MAP being underestimated by up to a third in parts of  
18 northern South America and southern India, and overestimated by the same amount in some  
19 tropical regions of South America, Africa and Australia.

## 20 2.1 TRIFFID Vegetation Model

### 21 2.1.1 Description

22 TRIFFID calculates areal coverage, leaf area index and canopy height for the five PFTs,  
23 which each respond differently to climate and CO<sub>2</sub> forcing, and all can co-exist within the  
24 same gridbox.

25 Two modes of coupling between the TRIFFID and the GCM are used: an equilibrium mode  
26 used during model spin-up, in which the fluxes between the land and the atmosphere are  
27 calculated by the GCM and averaged over ~5 years, and a dynamic mode, used for the last  
28 100 years of the simulations, in which the fluxes are averaged over 10 days to include  
29 interannual variability. The averaged fluxes are then passed to TRIFFID which calculates the

1 growth and expansion of the existing vegetation, and updates the land surface parameters  
2 based on the new vegetation distribution and structure.

### 3 2.1.2 Uncertainties

4 TRIFFID is based on physiological constraints that influence vegetation distributions, and is  
5 therefore attempting to model the potential natural vegetation for a given climatology.  
6 Estimates of anthropogenic land surfaces (urban and agricultural) amount to some 16% of the  
7 total land surface area (CIESIN, 2004; Matthews, 1983; Ramankutty and Foley, 1999;  
8 Schneider et al., 2009) and therefore discrepancies will exist in those areas between reference  
9 maps of reconstructed natural vegetation (the vegetation that might exist if human influence  
10 were to cease, based on extrapolation of remnants of natural vegetation) and model-produced  
11 maps of potential natural vegetation (the vegetation that might exist if human influence had  
12 never existed), and these may be significant in areas of deforestation (Prentice et al., 1992).

13 The TRIFFID model has been compared to IGBP-DIS land cover dataset, which represents  
14 the modern distribution of vegetation as derived from satellite image interpretation (Loveland  
15 and Belward, 1997; Betts et al., 2004). This suggests that the shrub PFT is overestimated at  
16 high latitudes, that the broadleaf tree PFT is overestimated in equatorial regions, and that  
17 grasses tend to be globally slightly underestimated. The discrepancies between the satellite  
18 imagery and the TRIFFID model are suggested to be a combination of orographic  
19 representation leading to underestimation of precipitation, differences between the  
20 anthropogenic masks used in their version of the model and that found on the satellite  
21 imagery, and the inadequate treatment of natural disturbance mechanisms such as fire (Betts  
22 et al., 2004). The version of the model used here aims to reproduce the vegetation of a  
23 modern potential natural climate, and therefore evaluation of the predicted PFTs is difficult.  
24 It is also hard to identify, when the modern predicted PFTs are likely wrong (e.g. needleleaf  
25 forests in south-east Asia and extensive broadleaf forests in Australia), whether those  
26 inadequacies are due to the climate model or the vegetation model, and it is likely that due to  
27 the interactive coupling between TRIFFID and HadCM3L that the existing GCM biases will  
28 be amplified in the vegetation predictions.

### 29 3.1 BIOME4 Vegetation Model Uncertainties

30 As with the TRIFFID model, the BIOME4 model is also based on physiological constraints  
31 that influence vegetation distributions, and therefore attempts to model the potential natural

1 vegetation for a given climatology. The BIOME4 model has been shown to produce a fair  
2 comparison to the vegetation data of Olson et al. (1983), with most of the areas of  
3 discrepancy being attributed to differences between potential natural vegetation and actual  
4 modern vegetation as a result of human influence. Other suggested sources of error include  
5 incorrect climatological forcing due to the difficulty in representing orographic extremes, and  
6 missing parameterisations in the model, particularly related to seasonality (Prentice et al.,  
7 1992).

## 8 **Appendix C. Model Boundary Conditions: Modern Configuration**

9 For the modern control (CTRL) simulation we use the standard UK Meteorological Office  
10 configurations for modern continental positions and both terrestrial and ice sheet elevations  
11 as defined in the ETOPO5 dataset (NOAA, 1988). Land ice grid boxes and soil type  
12 distributions used in CTRL are based on the identification of Wilson and Henderson-Sellers  
13 (1985). The two land grid boxes that represent Iceland are converted to ocean grid boxes as  
14 discussed in Section 2.3.4.

15 We also use modern ETOPO5 bathymetry (NOAA, 1988). The two new ocean grid boxes  
16 replacing Iceland are assigned the average depth of the surrounding ocean grid boxes. We  
17 note that the land-sea mask generation algorithms that map the high resolution data onto the  
18 HadCM3L grid also results in the non-representation of many other small islands.

19 For the purpose of this study we wish to model a potential natural modern climate, and  
20 therefore assume a preindustrial atmospheric CO<sub>2</sub> concentration in the range 275-284 ppm  
21 (Etheridge et al., 1996) and use a value of 280ppm in CTRL, consistent with the Climate  
22 Model Intercomparison Project standard (Meehl et al., 2007). Other greenhouse gases are  
23 also set to preindustrial level atmospheric gases (CH<sub>4</sub>=760 ppb, N<sub>2</sub>O=270 ppb, CFC's=0 ppt).

## 24 **Appendix D. Late Miocene Configuration**

### 25 **1.2 Orography**

26 The orography for the late Miocene has been reconstructed by (Markwick, 2007) using a  
27 methodology of establishing relationships between modern elevations and their  
28 tectonophysiographic settings and applying those same relationships to the geological record;  
29 (a full description of the technique is described in Markwick, 2007 and Markwick and  
30 Valdes, 2004). The methodology results in significant reductions in the Miocene elevations

1 of most of the world's highest regions compared to modern. We now describe those  
2 reductions in relation to the palaeoelevation estimates of other authors; however, given the  
3 uncertainties associated with reconstructing past mountain elevation (e.g. Ehlers and Poulsen,  
4 2009), any inconsistencies with that of Markwick (2007) would require further sensitivity  
5 experiments.

6 The Tibetan Plateau in the late Miocene simulations is an average 30% lower than in CTRL.  
7 The timing of Tibetan uplift is debated, with many studies indicating an elevation during the  
8 late Miocene similar to today (e.g. Currie et al., 2005; DeCelles et al., 2007; Spicer et al.,  
9 2003; Garzione et al., 2000; Rowley et al., 2001; Rowley and Currie, 2006b). There is  
10 evidence that the uplift did not occur in a spatially uniform pattern (England and Searle,  
11 1986; Clark et al., 2005; Rowley and Garzione, 2007), but a recent study has shown that  
12 modelled SSTs in the Indian and Pacific Oceans were very similar for three simulations  
13 assuming different uplift histories (Lunt et al., 2010).

14 The Andes are 45% lower in the late Miocene simulations than the CTRL. This is consistent  
15 with palaeobotanical and geomorphological data which suggests that the elevation of the  
16 Andes was no more than half of the modern elevation at ~11-10 Ma (Gregory-Wodzicki,  
17 2000).

18 The Alps are on average 30% lower in the late Miocene simulations than the CTRL.  
19 Palaeoaltitude estimates for the Eastern Alps are between 500 and 2000m (Jimenez-Moreno  
20 et al., 2008; Kuhlemann, 2007). These estimates are consistent with the Markwick (2007)  
21 reconstruction.

22 The Western Cordillera of North America, is on average 25% higher in the late Miocene  
23 simulations than in the CTRL. The evolution of topographic changes in the Western  
24 Cordillera are debated. There is evidence for high elevation in the Western Cordillera of  
25 North America since the Eocene (Sjostrom et al., 2006) or even the Cretaceous (Chase et al.,  
26 1998), with the elevation in Nevada during the Middle Miocene being some 1-1.5km higher  
27 than it is today. The dating of the collapse to present day elevation in Nevada is put at ~13  
28 Ma by Wolfe et al. (1997) but the palaeoelevation reconstruction from Markwick (2007)  
29 differs in that Nevada is ~9% lower in LM than CTRL. Orographic lowering in Wyoming is  
30 believed to have occurred during the Eocene (Chase et al., 1998), and the methodology used  
31 by Markwick (2007) puts Wyoming 27% lower in the late Miocene than in the CTRL. A  
32 recent study of the Western Cordillera of North America also considers that elevations were

generally lower during the late Miocene than today, but argues for considerable evidence of surface uplift during the late Miocene in the westernmost regions north of 45°N (Foster et al., 2010).

### **1.3 Ice Sheet Configuration**

The extent of late Miocene glaciation is much debated (Shackleton and Kennett, 1975; Webb and Harwood, 1991; Wilson, 1995; Pekar and DeConto, 2006; Huybrechts, 1993; Marchant et al., 1996) and perhaps highly variable (Lear et al., 2000). For the late Miocene, the Markwick (2007) reconstruction assumes the East and West Antarctic ice sheets cover the whole Antarctic continent, although ice thickness is generally less than CTRL. The sensitivity experiments of DeConto et al. (2008) and Micheels et al. (2009a) support the evidence for the onset of some glaciation in the Northern Hemisphere during the Miocene (Moran et al., 2006; Kamikuri et al., 2007; Denton and Armstrong, 1969; Thiede and Myhre, 1995 and references therein; Talwani and Udintsev, 1976; Warnke and Hansen, 1977). Markwick (2007) assumes much reduced Northern Hemisphere ice for the late Miocene compared with the extent of the present day Greenland ice sheet. Future work will investigate the impact of ice sheet extent assumptions on the simulated late Miocene climate.

The high latitude topography is lowered in the late Miocene simulations compared to the CTRL, due to the removal of continental ice rather than post-Miocene uplift: Greenland is 60% lower and Antarctica is very spatially variable, but is on average 4% lower.

All non-ice covered land grid boxes were initialised with the TRIFFID shrub PFT before the TRIFFID model was switched on as the simulations were continuations of pre-existing simulations with global homogenous shrub coverage.

### **1.4 Continental Positions**

We use the late Miocene land-sea definitions of Markwick (2007), aggregated up to the GCM resolution. Notable changes in continental positions are the more southerly position of the Australian continent, the closed Bering Strait, the extension of Eurasia in to the Arctic, and the open Panama Gateway and Indonesian Seaway for the late Miocene.

### **1.5 Soil Type**

Some of the palaeoenvironmental conditions are largely unknown for the Miocene period, such as the soil type which is typically unknown outside of some regions in North America and Kenya (Retallack, 2004; Retallack et al., 1995; Retallack et al., 1990; Retallack et al., 2002a; Retallack et al., 2002b), and so for these we use an average homogenous value derived from the CTRL simulation. Soil parameters have been shown to have significant influence on soil hydrology (Osborne et al., 2004), and therefore our choice of soil type could bias the simulated vegetation distribution and modify the simulated climate. Future experiments will be conducted to assess the extent to which the soil type could affect the results presented here.

## **1.6 Bathymetry**

Late Miocene palaeobathymetry is assumed to be the same as for the CTRL, except for the land-sea mask itself, which results in a wider Indonesian Seaway and an open Panama Gateway. The algorithm used to generate the late Miocene bathymetry assumes that the depth of all late Miocene ocean grid boxes (that would be land today in the CTRL simulation) is the average of all neighbouring ocean grid boxes.

As palaeomagnetic studies for plate tectonic reconstruction are difficult to apply to the Indonesian islands (Gourlan et al., 2008), the evolution of their palaeogeography remains uncertain. It is generally believed that the restriction of the Indonesian Seaway has occurred in the past 20 Ma, with timing estimates for different water mass restrictions from Early Miocene to early late Miocene (van Andel et al., 1975; Edwards, 1975), or even Pliocene (Kennett et al., 1985; Cane and Molnar, 2001; Srinivasan and Sinha, 1998). The late Miocene seaway in our simulations is >2000m at its deepest, as for the CTRL.

The closure of the Panama gateway in the Pliocene is the most recent major modification to oceanic gateways (Keigwin, 1982; Duque-Caro, 1990). The shoaling of the region is believed to have commenced in the Middle Miocene (Keller and Barron, 1983; Duque-Caro, 1990; Droxler et al., 1998) but determining potential depth of the sill during the late Miocene is problematic; a depth of 200-500m at 6Ma is suggested by Collins et al. (1996) and <100m depth by the early Pliocene by Keigwin (1982). Duque-Caro (1990) suggests that in the earliest late Miocene sill depth was ~1000m, and 150-500m between 8.6-7 Ma, <150m at 7-6.3 Ma and <50m at the latest Miocene. Given our large timeslab, possible sill depths for the

late Miocene range from 1000m to <50m; our bathymetric algorithm produces a depth of 995m which lies within the range of estimates.

The choice of bathymetry has been shown to have large regional impacts (e.g. the Greenland-Scotland Ridge: Robinson et al., 2011), but the uncertainty associated with palaeobathymetric reconstruction is large and therefore future work will seek to assess the climatic uncertainty associated with the choice of palaeobathymetry through climate sensitivity studies.

## 7 Acknowledgements

This work was carried out using the computing facilities of the Advanced Computing Research Centre at the University of Bristol, <http://www.bris.ac.uk/acrc>. Catherine Bradshaw is funded by a NERC PhD studentship. ~~Dan Lunt~~ is funded by RCUK. ~~Matthew M.~~ Pound is funded by the Natural Environment Research Council (UK) and the British Geological Survey University Funding Initiative (PhD studentship NE/G523563/1). We thank the two reviewers and Gregor Knorr for their comments on an earlier version of this manuscript.

## 8 References

Bailey, I., and Sinnott, E., 1915. A botanical index of Cretaceous and Tertiary climates. *Science* 41, 831-834.

Barron, E. J.: Explanations of the Tertiary Global Cooling Trend, *Palaeogeography Palaeoclimatology Palaeoecology*, 50, 45-61, 1985.

Beerling, D. J., and Royer, D. L.: Convergent Cenozoic CO<sub>2</sub> history, *Nature Geosci*, 4, 418-420, <http://www.nature.com/ngeo/journal/v4/n7/abs/ngeo1186.html#supplementary-information>, 2011.

Behrensmeyer, A. K., Kidwell, S. M., and Gastaldo, R. A.: Taphonomy and paleobiology, *Paleobiology*, 26, 103-147, 10.1666/0094-8373(2000)26[103:tap]2.0.co;2, 2000.

Bernor R. L., Mittmann H.-W. & Rögl F. 1993 Systematics and Chronology of the Götzendorf "Hipparion" (Late Miocene, Pannonian F, Vienna Basin). *Ann. Naturhist. Mus. Wien* 95(A), 101-120

Betts, R. A., Cox, P. M., Collins, M., Harris, P. P., Huntingford, C., and Jones, C. D.: The role of ecosystem-atmosphere interactions in simulated Amazonian precipitation decrease and forest dieback under global climate warming, *Theoretical and Applied Climatology*, 78, 157-175, 10.1007/s00704-004-0050-y, 2004.

Bohme, M., Ilg, A., and Winklhofer, M.: ~~late~~Late Miocene "washhouse" climate in Europe, *Earth and Planetary Science Letters*, 275, 393-401, DOI 10.1016/j.epsl.2008.09.011, 2008.

Brandefelt, J., and Otto-Bliesner, B. L.: Equilibration and variability in a last glacial maximum climate simulation with CCSM3, *Geophysical Research Letters*, 36, L19712, 2009.

1 Bruch, A. A., Uhl, D., and Mosbrugger, V.: Miocene climate in Europe - Patterns and  
2 evolution - A first synthesis of NECLIME, *Palaeogeography Palaeoclimatology*  
3 *Palaeoecology*, 253, 1-7, DOI 10.1016/j.palaeo.2007.03.030, 2007.

4 Bruch, A.A., Utescher, T., Mosbrugger, V., and NECLIME members, 2011. Precipitation  
5 pat-terns in the Miocene of Central Europe and the development of continentality. *Palaeoge-*  
6 *ography, Palaeoclimatology, Palaeoecology*, 304: 202-211.

7 Cane, M. A., and Molnar, P.: Closing of the Indonesian seaway as a precursor to east African  
8 aridification around 3-4 million years ago, *Nature*, 411, 157-162, 2001.

9 Chase, C. G., Gregory-Wodzicki, K. M., Parrish-Jones, J. T., and DeCelles, P. G.:  
10 Topographic history of the western Cordillera of North America and controls on climate, in:  
11 Tectonic boundary conditions for climate model simulations, edited by: Crowley, T. J., and  
12 Burke, K., *Oxford Monographs on Geology and Geophysics*, 39, Oxford University Press,  
13 Oxford UK, 73-99, 1998.

14 Clark, M. K., House, M. A., Royden, L. H., Whipple, K. X., Burchfiel, B. C., Zhang, X., and  
15 Tang, W.: Late Cenozoic uplift of southeastern Tibet, *Geology*, 33, 525-528, Doi  
16 10.1130/G21265.1, 2005.

17 Collins, L. S., Coates, A. G., Berggren, W. A., Aubry, M. P., and Zhang, J. J.: The late  
18 Miocene Panama isthmian strait, *Geology*, 24, 687-690, 1996.

19 Cox, P.: Description on the TRIFFID Dynamic Global Vegetation Model. , 2001.

20 Cox, P. M., Betts, R. A., Jones, C. D., Spall, S. A., and Totterdell, I. J.: Acceleration of global  
21 warming due to carbon-cycle feedbacks in a coupled climate model, *Nature*, 408, 184-187,  
22 2000.

23 Crowley, T. J., and Baum, S. K.: Reconciling Late Ordovician (440 Ma) glaciation with very  
24 high (14X) CO<sub>2</sub> levels, *J. Geophys. Res.*, 100, 1093-1101, 10.1029/94jd02521, 1995.

25 Currie, B. S., Rowley, D. B., and Tabor, N. J.: Middle Miocene paleoaltimetry of southern  
26 Tibet: Implications for the role of mantle thickening and delamination in the Himalayan  
27 orogen, *Geology*, 33, 181-184, 2005.

28 DeCelles, P. G., Kapp, P., Ding, L., and Gehrels, G. E.: Late Cretaceous to middle Tertiary  
29 basin evolution in the central Tibetan Plateau: Changing environments in response to tectonic  
30 partitioning, aridification, and regional elevation gain, *Geological Society of America*  
31 *Bulletin*, 119, 654-680, 10.1130/b26074.1, 2007.

32 DeConto, R. M., Pollard, D., Wilson, P. A., Palike, H., Lear, C. H., and Pagani, M.:  
33 Thresholds for Cenozoic bipolar glaciation, *Nature*, 455, 652-U652, Doi  
34 10.1038/Nature07337, 2008.

35 Demicco, R. V., Lowenstein, T. K., and Hardie, L. A.: Atmospheric pCO<sub>2</sub> since 60 Ma  
36 from records of seawater pH, calcium, and primary carbonate mineralogy, *Geology*, 31, 793-  
37 796, 2003.

38 Denk, T., Grimsson, F., and Kvacek, Z.: The Miocene floras of Iceland and their significance  
39 for late Cainozoic North Atlantic biogeography, *Bot J Linn Soc*, 149, 369-417, 2005.

40 Denton, G. H., and Armstrong, R.I.: Miocene-Pliocene Glaciations in Southern Alaska, *Am J*  
41 *Sci*, 267, 1121-&, 1969.

- 1 Droxler, A. W., Burke, K., Cunningham, A. D., Hine, A. C., Rosencrantz, D., and Duncan,  
2 D.: Caribbean constraints on circulation between Atlantic and Pacific oceans over the past 40  
3 million years, in: Tectonic Boundary Conditions for Climate Reconstruction, edited by: K.,  
4 B., Oxford Monographs Geology and Geophysics, 169– 191, 1998.
- 5 Duque-Caro, H.: Neogene stratigraphy, paleocenography and paleobiogeography in  
6 northwest South-America and the evolution of the Panama Seaway~~NEOGENE~~  
7 ~~STRATIGRAPHY, PALEOCEANOGRAPHY AND PALEOBIOGEOGRAPHY IN~~  
8 ~~NORTHWEST SOUTH-AMERICA AND THE EVOLUTION OF THE PANAMA~~  
9 ~~SEAWAY~~, Palaeogeography Palaeoclimatology Palaeoecology, 77, 203-234, 1990.
- 10 Düringer, P., Schuster, M., Genise, J. F., Mackaye, H. T., Vignaud, P., and Brunet, M.: New  
11 termite trace fossils: Galleries, nests and fungus combs from the Chad basin of Africa (Upper  
12 Miocene-Lower Pliocene), Palaeogeography Palaeoclimatology Palaeoecology, 251, 323-  
13 353, DOI 10.1016/j.palaeo.2007.03.029, 2007.
- 14 Dutton, J. F., and Barron, E. J.: Miocene to present vegetation changes: A possible piece of  
15 the Cenozoic cooling puzzle, Geology, 25, 39-41, 1997.
- 16 Edwards, A. R.: Southwest Pacific Cenozoic paleogeography and an integrated Neogene  
17 paleocirculation model., Init. Rep. DSDP, 30, 667–684, 1975.
- 18 Edwards, J. M., and Slingo, A.: Studies with a flexible new radiation code .1. Choosing a  
19 configuration for a large-scale model, Quarterly Journal of the Royal Meteorological Society,  
20 122, 689-719, 1996.
- 21 Ehlers, T. A., and Poulsen, C. J.: Large paleoclimate influence the ~~interpretation~~~~intepretation~~  
22 of Andean Plateau paleoaltimetry, Earth and Planetary Science Letters, 281, 238-248, 2009.
- 23 England, P., and Searle, M. P.: The Cretaceous–Tertiary deformation of the Lhasa Block and  
24 its implications for crustal thickening in Tibet, Tectonics, 5, 1-14, 1986.
- 25 Eronen, J. T., Atabadian, M. M., Micheels, A., Karme, A., Bernor, R. L., and Fortelius, M.:  
26 Distribution history and climatic controls of the ~~late~~~~Late~~ Miocene Pikermian chronofauna,  
27 Proceedings of the National Academy of Sciences of the United States of America, 106,  
28 11867-11871, DOI 10.1073/pnas.0902598106, 2009.
- 29 Eronen, J. T., Puolamäki, K., Liu, L., Lintulaakso, K., Damuth, J., Janis, C., and Fortelius,  
30 M.: Precipitation and large herbivorous mammals , part II: Application to fossil data., Evol  
31 Ecol Res, 12, 235-248, 2010.
- 32 Eronen, J.T., Micheels, A., Utescher, T. 2012. Comparison of estimates for Mean Annual  
33 Precipitation from different proxies. A Pilot Study for European Neogene. Evolutionary  
34 Ecology Research. (available online 16.4.2012 at [http://www.evolutionary-](http://www.evolutionary-ecology.com/forthcoming.html)  
35 [ecology.com/forthcoming.html](http://www.evolutionary-ecology.com/forthcoming.html))
- 36 Essery, R., and Clark, D. B.: Developments in the MOSES 2 land-surface model for PILPS  
37 2e, Global Planet Change, 38, 161-164, Doi 10.1016/S0921-8181(03)00026-2, 2003.
- 38 Etheridge, D. M., Steele, L. P., Langenfelds, R. L., Francey, R. J., Barnola, J. M., and  
39 Morgan, V. I.: Natural and anthropogenic changes in atmospheric CO<sub>2</sub> over the last 1000  
40 years from air in Antarctic ice and firn, J Geophys Res-Atmos, 101, 4115-4128, 1996.
- 41 Fang, X. M., Yan, M. D., Van der Voo, R., Rea, D. K., Song, C. H., Pares, J. M., Gao, J. P.,  
42 Nie, J. S., and Dai, S.: Late Cenozoic deformation and uplift of the NE Tibetan plateau:  
43 Evidence from high-resolution magneto stratigraphy of the Guide Basin, Qinghai Province,

- China, Geological Society of America Bulletin, 117, 1208-1225, Doi 10.1130/B25727.1, 2005.
- Farrera, I., Harrison, S. P., Prentice, I. C., Ramstein, G., Guiot, J., Bartlein, P. J., Bonnefille, R., Bush, M., Cramer, W., von Grafenstein, U., Holmgren, K., Hooghiemstra, H., Hope, G., Jolly, D., Lauritzen, S.-E., Ono, Y., Pinot, S., Stute, M., and Yu, G.: Tropical climates at the Last Glacial Maximum: a new synthesis of terrestrial palaeoclimate data. I. Vegetation, lake-levels and geochemistry. , Climate Dynamics, 15, 823-856, 1999.
- Fauquette, S., Suc, J.-P., Bertini, A., Popescu, S.-M., Warny, S., Bachiri Taoufiq, N., Perez Villa, M.-J., Chikhi, H., Feddi, N., Subally, D., Clauzon, G., and Ferrier, J.: How much did climate force the Messinian salinity crisis? Quantified climatic conditions from pollen records in the Mediterranean region, Palaeogeography, Palaeoclimatology, Palaeoecology, 238, 281-301, 10.1016/j.palaeo.2006.03.029, 2006.
- Fluteau, F., Ramstein, G., and Besse, J.: Simulating the evolution of the Asian and African monsoons during the past 30 Myr using an atmospheric general circulation model, Journal of Geophysical Research-Atmospheres, 104, 11995-12018, 1999.
- Foster, G. L., Lunt, D. J., and Parrish, R. R.: Mountain uplift and the glaciation of North America - a sensitivity study, Climate of the Past, 6, 707-717, DOI 10.5194/cp-6-707-2010, 2010.
- Freeman, K. H., and Hayes, J. M.: Fractionation of carbon isotopes by phytoplankton and estimates of ancient CO<sub>2</sub> levels, Global Biogeochemical Cycles, 6, 185-198, 1992.
- Garzzone, C. N., Dettman, D. L., Quade, J., DeCelles, P. G., and Butler, R. F.: High times on the Tibetan Plateau: Paleoelevation of the Thakkhola graben, Nepal, Geology, 28, 339-342, 2000.
- Garzzone, C. N., Hoke, G. D., Libarkin, J. C., Withers, S., MacFadden, B., Eiler, J., Ghosh, P., and Mulch, A.: Rise of the Andes, Science, 320, 1304-1307, DOI 10.1126/science.1148615, 2008.
- Gent, P. R., and McWilliams, J. C.: Isopycnal mixing in ocean circulation models, Journal of Physical Oceanography, 20, 150-155, 1990.
- [Gladstone, G., Flecker, R., Valdes, P., Lunt, D., Markwick, P. \(2007\) The Mediterranean hydrologic budget from a Late Miocene global climate simulation. Palaeogeography, Palaeoclimatology, Palaeoecology, 251, 254-267](#)
- Gordon, C., Cooper, C., Senior, C. A., Banks, H., Gregory, J. M., Johns, T. C., Mitchell, J. F. B., and Wood, R. A.: The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments, Climate Dynamics, 16, 147-168, 2000.
- Gourlan, A. T., Meynadier, L. M., and Allègre, C. J.: Tectonically driven changes in the Indian Ocean circulation over the last 25 Ma: Neodymium isotope evidence, Earth and Planetary Science Letters, 267, 353-364, 2008.
- [Gradstein, F.M., Ogg, J.G., Smith, A.G., Bleeker, W., Lourens, L.J., 2004. A new Geological Time Scale, with special reference to Precambrian and Neogene. Episodes 27, 83-100](#)
- Gregory-Wodzicki, K. M.: Andean paleoelevation estimates: A review and critique, Geological Society of America Bulletin, 112, 1091-1105, 2000.

- Gregory-Wodzicki, K. M.: A late Miocene subtropical-dry flora from the northern Altiplano, Bolivia, *Palaeogeography Palaeoclimatology Palaeoecology*, 180, 331-348, Pii S0031-0182(01)00434-5, 2002.
- Gregory, D., and Rowntree, P. R.: A mass flux convection scheme with representation of cloud ensemble characteristics and stability-dependent closure, *Monthly Weather Review*, 118, 1483-1506, 1990.
- Harrison, T. M., Copeland, P., Kidd, W. S. F., and Yin, A.: Raising Tibet, *Science*, 255, 1663-1670, 1992.
- Haxeltine, A., and Prentice, I. C.: BIOME3: An equilibrium terrestrial biosphere model based on ecophysiological constraints, resource availability, and competition among plant functional types, *Global Biogeochemical Cycles*, 10, 693-709, 1996.
- [Henrot, A.-J., François, L., Favre, E., Butzin, M., Ouberdous, M., and Munhoven, G.: Effects of CO<sub>2</sub>, continental distribution, topography and vegetation changes on the climate at the Middle Miocene: a model study, \*Clim. Past\*, 6, 675-694, doi:10.5194/cp-6-675-2010, 2010](#)
- [Hilgen, F., Aziz, H.A., Bice, D., Iaccarino, S., Krijgsman, W., Kuiper, K., Montanari, A., Raffi, I., Turco, E., Zachariasse, W.-J., 2005. The Global boundary Stratotype Section and Point \(GSSP\) of the Tortonian Stage \(Upper Miocene\) at Monte Dei Corvi. \*Episodes\* 28, 6-17](#)
- [Holdridge, L.R. 1947. Determination of world formations from simple climatic data. \*Science\* 105, 367-368](#)
- Hughes, J. K., Valdes, P. J., and Betts, R. A.: Dynamical properties of the TRIFFID dynamic global vegetation model., 2004.
- Huybrechts, P.: Glaciological modeling of the late Cenozoic East Antarctic ice sheet: stability or dynamism?, *Geographical Annals*, 75A, 221-238, 1993.
- IPCC: Climate Change 2007 – The Physical Science Basis, in: Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, 2007 edited by: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M., and Miller, H. L., Cambridge University Press, Cambridge, 2007.
- Jacobs, B. F., and Deino, A. L.: Test of climate-leaf physiognomy regression models, their application to two Miocene floras from Kenya, and Ar-40/Ar-39 dating of the late Miocene Kapturo site, *Palaeogeography Palaeoclimatology Palaeoecology*, 123, 259-271, 1996.
- ~~[Jakobsson, M., Backman, J., Rudels, B., Nycander, J., Frank, M., Mayer, L., Jokati, W., Sangiorgi, F., O'Regan, M., Brinkhuis, H., King, J., and Moran, K.: The early Miocene onset of a ventilated circulation regime in the Arctic Ocean, \*Nature\*, 447, 986-990, Doi 10.1038/Nature05924, 2007.](#)~~
- Jimenez-Moreno, G., Fauquette, S., and Suc, J. P.: Vegetation, climate and palaeoaltitude reconstructions of the Eastern Alps during the Miocene based on pollen records from Austria, Central Europe, *J Biogeogr*, 35, 1638-1649, DOI 10.1111/j.1365-2699.2008.01911.x, 2008.
- Jones, C.: A fast ocean GCM without flux adjustments, *Journal of Atmospheric and Oceanic Technology*, 20, 1857-1868, 2003.
- Kameo, K., and Sato, T.: Biogeography of Neogene calcareous nannofossils in the Caribbean and the eastern equatorial Pacific - floral response to the emergence of the Isthmus of Panama, *Marine Micropaleontology*, 39, 201-218, 2000.

- Kamikuri, S.-i., Nishi, H., and Motoyama, I.: Effects of late Neogene climatic cooling on North Pacific radiolarian assemblages and oceanographic conditions, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 249, 370-392, 2007.
- [Kaplan, J.O. 2001. Geophysical Applications of Vegetation Modeling. Ph.D. Thesis, Lund University, Lund.](#)
- [Kaplan, J.O., Bigelow, N.H., Bartlein, P.J., Christensen, T.R., Cramer, W., Harrison, S.P., Matveyeva, N.V., McGuire, A.D., Murray, D.F., Prentice, I.C., Razzhivin, V.Y., Smith, B. and Walker, D.A., Anderson, P.M., Andreev, A.A., Brubaker, L.B., Edwards, M.E., Lozhkin, A.V. and Ritchie, J. \(2003\). Climate change and Arctic ecosystems II: Modeling, palaeodata-model comparisons, and future projections. \*Journal of Geophysical Research-Atmosphere\*, 108\(D19\), 8171](#)
- Keigwin, L.: Isotopic paleoceanography of the Caribbean and East Pacific: role of Panama uplift in late Neogene time., *Science*, 217, 350-353, 1982.
- Keller, G., and Barron, J. A.: Paleooceanographic Implications of Miocene Deep-Sea Hiatuses, *Geological Society of America Bulletin*, 94, 590-613, 1983.
- Kennett, J. P., Keller, G., and Srinivasan, M. S.: Miocene Planktonic Foraminiferal Biogeography and Paleooceanographic Development of the Indo-Pacific Region, *Geol Soc Am Mem*, 163, 197-236, 1985.
- Kingston, J. D., and Hill, A.: ~~late~~Late Miocene palaeoenvironments in Arabia: A synthesis., in: *Fossil vertebrates of Arabia*, edited by: Whybrow, P. J., and Hill, A., Yale University Press, 1999.
- Knorr, G., Butzin, M., Micheels, A., and Lohmann, G.: A warm Miocene climate at low atmospheric CO<sub>2</sub> levels, *Geophys Res Lett*, 38, 10.1029/2011gl048873, 2011.
- [Krapp, M., and Jungclauss, J.H.: The Middle Miocene climates as modelled in an atmosphere-ocean-biosphere model, \*Climates of the Past\* 7\(4\), 1169-1188](#)
- Kuhlemann, J.: Paleogeographic and paleotopographic evolution of the Swiss and Eastern Alps since the Oligocene, *Global Planet Change*, 58, 224-236, DOI 10.1016/j.gloplacha.2007.03.007, 2007.
- Kurschner, W. M., vanderBurgh, J., Visscher, H., and Dilcher, D. L.: Oak leaves as biosensors of late Neogene and early Pleistocene paleoatmospheric CO<sub>2</sub> concentrations, *Marine Micropaleontology*, 27, 299-312, 1996.
- Kurschner, W. M., Kvacek, Z., and Dilcher, D. L.: The impact of Miocene atmospheric carbon dioxide fluctuations on climate and the evolution of terrestrial ecosystems, *Proceedings of the National Academy of Sciences of the United States of America*, 105, 449-453, DOI 10.1073/pnas.0708588105, 2008.
- Kutzbach, J. E., Guetter, P. J., Ruddiman, W. F., and Prell, W. L.: Sensitivity of Climate to Late Cenozoic Uplift in Southern Asia and the American West - Numerical Experiments, *Journal of Geophysical Research-Atmospheres*, 94, 18393-18407, 1989.
- Kutzbach, J. E., and Behling, P.: Comparison of simulated changes of climate in Asia for two scenarios: Early Miocene to present, and present to future enhanced greenhouse, *Global Planet Change*, 41, 157-165, DOI 10.1016/j.gloplacha.2004.01.015, 2004.
- Lear, C. H., Elderfield, H., and Wilson, P. A.: Cenozoic deep-sea temperatures and global ice volumes from Mg/Ca in benthic foraminiferal calcite, *Science*, 287, 269-272, 2000.

- 1 Lewis, A. R., Marchant, D. R., Ashworth, A. C., Hedenas, L., Hemming, S. R., Johnson, J.  
2 V., Leng, M. J., Machlus, M. L., Newton, A. E., Raine, J. I., Willenbring, J. K., Williams, M.,  
3 and Wolfe, A. P.: Mid-Miocene cooling and the extinction of tundra in continental  
4 Antarctica, *Proceedings of the National Academy of Sciences of the United States of*  
5 *America*, 105, 10676-10680, DOI 10.1073/pnas.0802501105, 2008.
- 6 Lohmann, G., Butzin, M., Micheels, A., Bickert, T., and Mosbrugger, V.: Effect of vegetation  
7 on the ~~late~~Late Miocene ocean circulation, *Climate of the Past Discussions*, 2, 605-631, 2006.
- 8 Loveland, T. R., and Belward, A. S.: The IGBP-DIS global 1 km land cover data set,  
9 DISCover: First results, *International Journal of Remote Sensing*, 18, 3289-3295, 1997.
- 10 Lunt, D. J., de Noblet-Ducoudre, N., and Charbit, S.: Effects of a melted Greenland ice sheet  
11 on climate, vegetation, and the cryosphere., *Climate Dynamics*, 23, 679-694, 2004.
- 12 Lunt, D. J., Ross, I., Hopley, P. J., and Valdes, P. J.: Modelling late Oligocene C-4 grasses  
13 and climate, *Palaeogeography Palaeoclimatology Palaeoecology*, 251, 239-253, DOI  
14 10.1016/j.palaeo.2007.04.004, 2007.
- 15 Lunt, D. J., Flecker, R., Valdes, P. J., Salzmann, U., Gladstone, R., and Haywood, A. M.: A  
16 methodology for targeting palaeo proxy data acquisition: A case study for the terrestrial late  
17 Miocene, *Earth and Planetary Science Letters*, 271, 53-62, DOI 10.1016/j.epsl.2008.03.035,  
18 2008a.
- 19 Lunt, D. J., Valdes, P. J., Haywood, A., and Rutt, I. C.: Closure of the Panama Seaway during  
20 the Pliocene: implications for climate and Northern Hemisphere glaciation, *Climate*  
21 *Dynamics*, 30, 1-18, DOI 10.1007/s00382-007-0265-6, 2008b.
- 22 Lunt, D. J., Flecker, R., and Clift, P. D.: The impacts of Tibetan uplift on palaeoclimate  
23 proxies, *Geological Society of London, Special Publications*, 342, 279-291, 2010.
- 24 ~~Lunt, D. J., Haywood, A. M., Schmidt, G. A., Salzmann, S., Valdes, P. J., and Dowsett, H. J.:~~  
25 ~~On the causes of mid-Pliocene warmth and polar amplification., *Earth and Planetary Science*~~  
26 ~~*Letters*, in press.~~
- 27 Marchant, D. R., Denton, G. H., Swisher, C. C., and Potter, N.: Late Cenozoic Antarctic  
28 paleoclimate reconstructed from volcanic ashes in the Dry Valleys region of southern  
29 Victoria Land, *Geological Society of America Bulletin*, 108, 181-194, 1996.
- 30 Markwick, P. J., and Valdes, P. J.: Palaeo-digital elevation models for use as boundary  
31 conditions in coupled ocean-atmo sphere GCM experiments: a Maastrichtian (late  
32 Cretaceous) example, *Palaeogeography Palaeoclimatology Palaeoecology*, 213, 37-63,  
33 10.1016/j.palaeo.2004.06.015, 2004.
- 34 Markwick, P. J.: The palaeogeographic and palaeoclimatic significance of climate proxies for  
35 data-model comparisons, *Deep-Time Perspectives on Climate Change: Marrying the Signal*  
36 *from Computer Models and Biological Proxies*, 251-312, 2007.
- 37 Martin, R. E.: *Taphonomy. A Process Approach*, Cambridge University Press, Cambridge,  
38 508 pp., 1999.
- 39 Matthews, E.: Global vegetation and land use: New high-resolution data bases for climate  
40 studies. , *Journal of Climate and Applied Meteorology*, 22, 474-487, 1983.
- 41 McGowran, B., Holdgate, G. R., Li, Q., and Gallagher, S. J.: Cenozoic stratigraphic  
42 succession in southeastern Australia, *Aust J Earth Sci*, 51, 459-496, 2004.

1 Meehl, G. A., Covey, C., Taylor, K. E., Delworth, T., Stouffer, R. J., Latif, M., McAvaney,  
2 B., and Mitchell, J. F. B.: The WCRP CMIP3 multimodel dataset: a new era in climate  
3 change research, *Bulletin of the American Meteorological Society*, 88, 1383-1394, 2007.

4 Micheels, A., Bruch, A. A., Uhl, D., Utescher, T., and Mosbrugger, V.: A ~~late~~Late Miocene  
5 climate model simulation with ECHAM4/ML and its quantitative validation with terrestrial  
6 proxy data, *Palaeogeography Palaeoclimatology Palaeoecology*, 253, 251-270, DOI  
7 10.1016/j.palaeo.2007.03.042, 2007.

8 Micheels, A., Bruch, A., and Mosbrugger, V.: Miocene Climate Modelling Sensitivity  
9 Experiments for Different Co<sub>2</sub> Concentrations, *Palaeontol Electron*, 12, -, Artn 5a, 2009a.

10 Micheels, A., Eronen, J., and Mosbrugger, V.: The ~~late~~Late Miocene climate response to a  
11 modern Sahara desert, *Global Planet Change*, 67, 193-204, DOI  
12 10.1016/j.gloplacha.2009.02.005, 2009b.

13 Micheels, A., Bruch, A. A., Eronen, J., Fortelius, M., Harzhauser, M., Utescher, T., and  
14 Mosbrugger, V.: Analysis of heat transport mechanisms from a ~~late~~Late Miocene model  
15 experiment with a fully-coupled atmosphere-ocean general circulation model,  
16 *Palaeogeography, Palaeoclimatology, Palaeoecology*, In Press, Corrected Proof, 2011.

17 Mitchell, T. D., and Jones, P. D.: An improved method of constructing a database of monthly  
18 climate observations and associated high-resolution grids. , *International Journal of*  
19 *Climatology*, 25, 693-712, 2005.

20 Molnar, P., England, P., and Martinod, J.: Mantle Dynamics, Uplift of the Tibetan Plateau,  
21 and the Indian Monsoon, *Rev Geophys*, 31, 357-396, 1993.

22 Montuire, S., Maridet, O., and Legendre, S.: ~~late~~Late Miocene-Early Pliocene temperature  
23 estimates in Europe using rodents, *Palaeogeography Palaeoclimatology Palaeoecology*, 238,  
24 247-262, DOI 10.1016/j.palaeo.2006.03.026, 2006.

25 Moran, K., Backman, J., Brinkhuis, H., Clemens, S. C., Cronin, T., Dickens, G. R., Eynaud,  
26 F., Gattacceca, J., Jakobsson, M., Jordan, R. W., Kaminski, M., King, J., Koc, N., Krylov, A.,  
27 Martinez, N., Matthiessen, J., McInroy, D., Moore, T. C., Onodera, J., O'Regan, M., Palike,  
28 H., Rea, B., Rio, D., Sakamoto, T., Smith, D. C., Stein, R., St John, K., Suto, I., Suzuki, N.,  
29 Takahashi, K., Watanabe, M., Yamamoto, M., Farrell, J., Frank, M., Kubik, P., Jokat, W.,  
30 and Kristoffersen, Y.: The Cenozoic palaeoenvironment of the Arctic Ocean, *Nature*, 441,  
31 601-605, Doi 10.1038/Nature04800, 2006.

32 Morgan, P., and Swanberg, C. A.: On the Cenozoic Uplift and Tectonic Stability of the  
33 Colorado Plateau, *J Geodyn*, 3, 39-63, 1985.

34 Mosbrugger, V., and Utescher, T.: The coexistence approach — a method for quantitative  
35 reconstructions of Tertiary terrestrial palaeoclimate data using plant fossils, *Palaeogeography,*  
36 *Palaeoclimatology, Palaeoecology*, 134, 61-86, 10.1016/s0031-0182(96)00154-x, 1997.

37 Nisancioglu, K. H., Raymo, M. E., and Stone, P. H.: Reorganization of Miocene deep water  
38 circulation in response to the shoaling of the Central American Seaway, *Paleoceanography*,  
39 18, -, Artn 1006 Doi 10.1029/2002pa000767, 2003.

40 Olson, J. S., Watts, J. A., and Allison, L. J.: Carbon in live vegetation of major world  
41 ecosystems., *Oak Ridge National Laboratory*, Oak Ridge, 1983.

1 Osborne, T. M., Lawrence, D. M., Slingo, J. M., Challinor, A. J., and Wheeler, T. R.:  
2 Influence of vegetation on the local climate and hydrology in the Tropics: Sensitivity to soil  
3 parameters. , Climate Dynamics, 23, 45–61, 2004.

4 Pagani, M., Arthur, M. A., and Freeman, K. H.: Miocene evolution of atmospheric carbon  
5 dioxide, Paleoceanography, 14, 273-292, 1999a.

6 | Pagani, M., Freeman, K. H., and Arthur, M. A.: ~~late~~Late Miocene atmospheric CO<sub>2</sub>  
7 concentrations and the expansion of C-4 grasses, Science, 285, 876-879, 1999b.

8 Pearson, P. N., and Palmer, M. R.: Atmospheric carbon dioxide concentrations over the past  
9 60 million years, Nature, 406, 695-699, 2000.

10 Pekar, S. F., and DeConto, R. M.: High-resolution ice-volume estimates for the early  
11 Miocene: Evidence for a dynamic ice sheet in Antarctica, Palaeogeography  
12 Palaeoclimatology Palaeoecology, 231, 101-109, DOI 10.1016/j.palaeo.2005.07.027, 2006.

13 Pound, M. J., Haywood, A. M., Salzmann, U., Riding, J. B., Lunt, D. J., and Hunter, S. J.: A  
14 | Tortonian (~~late~~Late Miocene, 11.61-7.25 Ma) global vegetation reconstruction,  
15 Palaeogeography Palaeoclimatology Palaeoecology, 300, 29-45, 2011.

16 | Pound, M.J., Haywood, A.M., Salzmann, U., Riding, J.B. Global vegetation dynamics and  
17 latitudinal temperature gradients during the mid to Late Miocene (15.97 - 5.33 Ma). Earth  
18 Science Reviews 112, 1-22, 2012

19 Prell, W. L., and Kutzbach, J. E.: Sensitivity of the Indian Monsoon to Forcing Parameters  
20 and Implications for Its Evolution, Nature, 360, 647-652, 1992.

21 Prentice, I. C., Cramer, W., Harrison, S. P., Leemans, R., Monserud, R. A., and Solomon, A.  
22 M.: A global biome model based on plant physiology and dominance, soil properties and  
23 climate, J Biogeogr, 19, 117-134, 10.2307/2845499, 1992.

24 Ramankutty, N., and Foley, J. A.: Estimating historical changes in global land cover:  
25 Croplands from 1700 to 1992., Global Biogeochemical Cycles, 13, 997-1027, 1999.

26 Ramstein, G., Fluteau, F., Besse, J., and Joussaume, S.: Effect of orogeny, plate motion and  
27 land sea distribution on Eurasian climate change over the past 30 million years, Nature, 386,  
28 788-795, 1997.

29 Retallack, G. J., Dugas, D. P., and Bestland, E. A.: Fossil Soils and Grasses of a Middle  
30 Miocene East-African Grassland, Science, 247, 1325-1328, 1990.

31 Retallack, G. J., Bestland, E. A., and Dugas, D. P.: Miocene Paleosols and Habitats of  
32 Proconsul on Rusinga Island, Kenya, J Hum Evol, 29, 53-91, 1995.

33 Retallack, G. J.: A 300-million-year record of atmospheric carbon dioxide from fossil plant  
34 cuticles, Nature, 411, 287-290, 10.1038/35077041, 2001.

35 | Retallack, G. J., Tanaka, S., and Tate, T.: ~~late~~Late Miocene advent of tall grassland paleosols  
36 in Oregon, Palaeogeography Palaeoclimatology Palaeoecology, 183, 329-354, Pii S0031-  
37 0182(02)00250-X, 2002a.

38 Retallack, G. J., Wynn, J. G., Benefit, B. R., and McCrossin, M. L.: Paleosols and  
39 paleoenvironments of the middle Miocene, Maboko Formation, Kenya, J Hum Evol, 42, 659-  
40 703, DOI 10.1006/jhev.2002.0553, 2002b.

- 1 Retallack, G. J.: Late Oligocene bunch grassland and early Miocene sod grassland paleosols  
2 from central Oregon, USA, *Palaeogeography Palaeoclimatology Palaeoecology*, 207, 203-  
3 237, DOI 10.1016/j.palaeo.2003.09.027, 2004.
- 4 Robinson, M. M., Valdes, P. J., Haywood, A. M., Dowsett, H. J., Hill, D. J., and Jones, S. M.:  
5 Bathymetric controls on Pliocene North Atlantic and Arctic sea surface temperature and  
6 deepwater production, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 309, 92-97,  
7 10.1016/j.palaeo.2011.01.004, 2011.
- 8 Rowley, D. B., Pierrehumbert, R. T., and Currie, B. S.: A new approach to stable isotope-  
9 based paleoaltimetry: implications for paleoaltimetry and paleohypsometry of the High  
10 Himalaya since the ~~late~~Late Miocene, *Earth and Planetary Science Letters*, 188, 253-268,  
11 2001.
- 12 Rowley, D. B., and Currie, B. S.: Palaeo-altimetry of the late Eocene to Miocene Lunpola  
13 basin, central Tibet, *Nature*, 439, 677-681, Doi 10.1038/Nature04506, 2006a.
- 14 ~~Rowley, D. B., and Currie, B. S.: Palaeo-altimetry of the late Eocene to Miocene Lunpola~~  
15 ~~basin, central Tibet, 439, 677-681, 2006b.~~
- 16 Rowley, D. B., and Garzione, C. N.: Stable isotope-based paleoaltimetry, *Annu Rev Earth Pl*  
17 *Sc*, 35, 463-508, 2007.
- 18 Ruddiman, W. F., and Kutzbach, J. E.: Forcing of Late Cenozoic Northern Hemisphere  
19 Climate by Plateau Uplift in Southern Asia and the American West, *Journal of Geophysical*  
20 *Research-Atmospheres*, 94, 18409-18427, 1989.
- 21 Ruddiman, W. F., Kutzbach, J. E., and Prentice, I. C.: Testing the climatic effects of  
22 orography and CO<sub>2</sub> with general circulation and biome models., *Tectonic Uplift and Climate*  
23 *Change*, edited by: Ruddiman, W. F., Plenum Publishing Corporation, New York, 1997.
- 24 Saggerson, E. P., and Baker, B. H.: Post-Jurassic erosion-surfaces in eastern Kenya and their  
25 deformation in relation to rift structure, *Quarterly Journal of the Geological Society of*  
26 *London*, 121, 51-72, 1965.
- 27 Salzmann, U., Haywood, A. M., and Lunt, D. J.: The past is a guide to the future? Comparing  
28 Middle Pliocene vegetation with predicted biome distributions for the twenty-first century,  
29 *Philosophical Transaction of the Royal Society A*, 367, 189-204, 2009.
- 30 Schneider, A., Friedl, M. A., and Potere, D.: A new map of global urban extent from MODIS  
31 satellite data., *Environ Res Lett*, 4, 044003, 2009.
- 32 Schneider, B., and Schmittner, A.: Simulating the impact of the Panamanian seaway closure  
33 on ocean circulation, marine productivity and nutrient cycling, *Earth and Planetary Science*  
34 *Letters*, 246, 367-380, 2006.
- 35 Seager, R., Battisti, D. S., Yin, J., Naik, N., Gordon, N., Clement, A. C., and Cane, M. A.: Is  
36 the Gulf Stream responsible for Europe's mild winters?, *Quarterly Journal of the Royal*  
37 *Meteorological Society*, 128, 2563-2586, 2002.
- 38 Shackleton, N. J., and Kennett, J. P.: Paleotemperature history of the Cenozoic and the  
39 initiation of Antarctic glaciation: oxygen and carbon isotope analyses in DSDP Sites 277,  
40 279, and 281. p. 743-755, U.S. Government Printing Office, Washington, 743-755, 1975.
- 41 Sjostrom, D. J., Hren, M. T., Horton, T. W., Waldbauer, J. R., and Chamberlain, C. P.: Stable  
42 isotopic evidence for a pre-late Miocene elevation gradient in the Great Plains-Rocky  
43 Mountain region, USA, *Geol S Am S*, 398, 309-319 2006

- Smith, R. N. B.: Experience and developments with the layer cloud and boundary layer mixing schemes in the UK Meteorological Office Unified Model. , ECMWF/GCSS workshop on parameterisation of the cloud-topped boundary layer, Reading, England, 1993,
- Spicer, R. A., Harris, N. B. W., Widdowson, M., Herman, A. B., Guo, S. X., Valdes, P. J., Wolfe, J. A., and Kelley, S. P.: Constant elevation of southern Tibet over the past 15 million years, *Nature*, 421, 622-624, 10.1038/nature01356, 2003.
- Spicer, R. A.: Recent and Future Developments of CLAMP: Building on the Lagacy of Jack A. Wolfe., *Cour. Forsch. -Inst. Senckenberg*, 258, 109-118, 2007.
- Spicer, R. A., Valdes, P. J., Spicer, T. E. V., Craggs, H. J., Srivastava, G., Mehrotra, R. C., and Yang, J.: New Developments in CLAMP: Calibration using global gridded mteorological data., *Palaeogeography, Palaeoclimatology, Palaeoecology*, 283, 91-98, 2009.
- Spiegel, C., Kuhlemann, J., Dunkl, I., and Frisch, W.: Paleogeography and catchment evolution in a mobile orogenic belt: the Central Alps in Oligo-Miocene times, *Tectonophysics*, 341, 33-47, 2001.
- Srinivasan, M. S., and Sinha, D. K.: Early Pliocene closing of the Indonesian Seaway: evidence from north-east Indian Ocean and tropical Pacific deep sea cores, *J Asian Earth Sci*, 16, 29-44, 1998.
- Steininger, F.F., 1999. Chronostratigraphy, Geochronology and Biochronology of the Miocene European Land Mammal Mega-Zones (ELMMZ) and the Miocene Mammal-Zones. In: Rössner, G.E., Heissig, K. (Eds.), The Miocene Land Mammals of Europe, Verlag Dr. Friedrich Pfeil, pp. 9-24
- Steppuhn, A., Micheels, A., Geiger, G., and Mosbrugger, V.: Reconstructing the ~~late~~Late Miocene climate and oceanic heat flux using the AGCM ECHAM4 coupled to a mixed-layer ocean model with adjusted flux correction, *Palaeogeography Palaeoclimatology Palaeoecology*, 238, 399-423, DOI 10.1016/j.palaeo.2006.03.037, 2006.
- Steppuhn, A., Micheels, A., Bruch, A. A., Uhl, D., Utescher, T., and Mosbrugger, V.: The sensitivity of ECHAM4/ML to a double CO2 scenario for the ~~late~~Late Miocene and the comparison to terrestrial proxy data, *Global Planet Change*, 57, 189-212, DOI 10.1016/j.gloplacha.2006.09.003, 2007.
- Takahashi, K., and Battisti, D. S.: Processes controlling the mean tropical Pacific Precipitation Pattern: I. The Andes and the Eastern Pacific ITCZ., *Journal of Climate*, 20, 3434-3451, 2007.
- Talwani, M., and Udintsev, G. e. a.: in: Initial Reports Deep Sea Drilling Project US Government Printing Office, Washington, DC, 1976.
- Texier, D., de Noblet, N., Harrison, S. P., Haxeltine, A., Jolly, D., Joussaume, S., Laarif, F., Prentice, I. C., and Tarasov, P.: Quantifying the role of biosphere-atmosphere feedbacks in climate change: coupled model simulations for 6000 years BP and comparison with palaeodata for northern Eurasia and northern Africa, *Climate Dynamics*, 13, 865-882, 1997.
- Thiede, J., and Myhre, A. M.: Non-steady behaviour in the Cenozoic Polar North atlantic System: The Onset and Variability of Northern Hemisphere Glaciations, *Philosophical Transactions: Physical Sciences and Engineering*, 352, 373-385, 1995.

- 1 [Thiel, C., Klotz, S. & Uhl, D., 2012. Palaeoclimate estimates for selected leaf-floras from the](#)
- 2 [Middle Pliocene \(Reuverian\) of Central Europe based on different palaeobotanical](#)
- 3 [techniques. Turkish Journal of Earth Sciences 21, 263-287](#)
- 4 Tonazzio, T., Gregory, J. M., and Huybrechts, P.: Climatic Impact of a Greenland
- 5 Deglaciation and its Possible Irreversibility', Journal of Climate, 17, 21-33, 2004.
- 6 Tripathi, A., Roberts, C., and Eagle, R.: Coupling of CO<sub>2</sub> and ice sheet stability over major
- 7 climate transitions of the last 20 million years, Science, 326, 1394-1397, 2009.
- 8 Truswell, E. M.: Vegetation changes in the Australian tertiary in response to climatic and
- 9 phytogeographic forcing factors, Aust Syst Bot, 6, 533-557, 10.1071/sb9930533, 1993.
- 10 [Uhl, D., Mosbrugger, V., Bruch, A. & Utescher, T. 2003. Reconstructing palaeotemperatures](#)
- 11 [using leaf floras – case studies for a comparison of leaf margin analysis and the coexistence](#)
- 12 [approach. Review of Palaeobotany and Palynology 126, 49–64](#)
- 13 [Uhl, D., Bruch, A.A., Traiser, C. & Klotz, S. 2006. Palaeoclimate estimates for the Middle](#)
- 14 [Miocene Schrotzburg flora \(S-Germany\) – a multi-method approach. International Journal of](#)
- 15 [Earth Science 95, 1071–1085.](#)
- 16 [Uhl, D., Klotz, S., Traiser, C., Thiel, C., Utescher, T., Kowalski, E.A. & Dilcher, D.L. 2007.](#)
- 17 [Paleotemperatures from fossil leaves – a European perspective. Palaeogeography,](#)
- 18 [Palaeoclimatology, Palaeoecology 248, 24–31](#)
- 19 Utescher, T., Böhme, M., and Mosbrugger, V.: The Neogene of Eurasia: Spatial gradients
- 20 and temporal trends — The second synthesis of NECLIME, Palaeogeography,
- 21 Palaeoclimatology, Palaeoecology, 304, 196-201, 10.1016/j.palaeo.2011.03.012, 2011.
- 22 van Andel, T. H., Heath, G. R., and Moore, T. C., Jr.: Cenozoic history and
- 23 paleoceanography of the Central Equatorial Pacific Ocean, Geological Society of America
- 24 Memoir, 143, 134p, 1975.
- 25 van Dam, J. A.: Geographic and temporal patterns in the late Neogene (12-3 Ma) aridification
- 26 of Europe: The use of small mammals paleoprecipitation proxies, Palaeogeography
- 27 Palaeoclimatology Palaeoecology, 238, 190-218, 10.1016/j.palaeo.2006.03.025, 2006.
- 28 Vignaud, P., Düringer, P., Mackaye, H. T., Likies, A., Blondel, C., Boissérie, J. R., de Bonis,
- 29 L., Eisenmann, V., Etienne, M. E., Geraads, D., Guy, F., Lehmann, T., Lihoreau, F., Lopez-
- 30 Martinez, N., Mourer-Chauvire, C., Otero, O., Rage, J. C., Schuster, M., Viriot, L., Zazzo, A.,
- 31 and Brunet, M.: Geology and palaeontology of the Upper Miocene Toros-Menalla hominid
- 32 locality, Chad, Nature, 418, 152-155, Doi 10.1038/Nature00880, 2002.
- 33 ~~Waelbroeck, C., A., Kucera, P. M., Rosell Melé, A., Weinelt, M., R., S., Mix, A. C., and al.,~~
- 34 ~~e.: Constraints on the magnitude and patterns of ocean cooling at the Last Glacial Maximum,~~
- 35 ~~Nature Geoscience, 2, 127–132,~~
- 36 ~~[http://www.nature.com/ng/journal/v2/n2/supinfo/ng411\\_S1.html](http://www.nature.com/ng/journal/v2/n2/supinfo/ng411_S1.html), 2009.~~
- 37 Warnke, D. A., and Hansen, M. E.: Sediments of glacial origins in the area of DSDP leg 38
- 38 (Norwegian Greenland seas): Preliminary results from Sites 336 and 344, Naturforsch. Ges.
- 39 Freib. Breisgau Ber., 67, 371-392, 1977.
- 40 Webb, P. N., and Harwood, D. M.: Late Cenozoic Glacial History of the Ross Embayment,
- 41 Antarctica, Quaternary Sci Rev, 10, 215-223, 1991.
- 42 Western, D., and Behrensmeyer, A. K.: Bones track community structure over four decades
- 43 of ecological change, Science, 324, 1061-1064, 2009.

1 Wilson, M. F., and Henderson-Sellers, A.: A global archive of land cover and soils data for  
2 use in general-circulation climate models, *Journal of Climatology*, 5, 119-143, 1985.

3 Wilson, T. J.: Cenozoic Transtension Along the Transantarctic Mountains West Antarctic  
4 Rift Boundary, Southern Victoria-Land, Antarctica, *Tectonics*, 14, 531-545, 1995.

5 Wolfe, J. A.: A method of obtaining climatic parameters from leaf assemblages., U.S.  
6 Geological Survey Bulletin, 2040, 73pp, 1993.

7 Wolfe, J. A., Schorn, H. E., Forest, C. E., and Molnar, P.: Paleobotanical evidence for high  
8 altitudes in Nevada during the Miocene, *Science*, 276, 1672-1675, 1997.

9 Worobiec, G., and Lesiak, M. A.: Plant megafossils from the Neogene deposits of Stawek-1A  
10 (Belchatow, Middle Poland), *Rev Palaeobot Palyno*, 101, 179-208, 1998.

11 [Yang, J., Spicer, R., Spicer, T., Li, C.-S., 2011. 'CLAMP Online': a new web-based](#)  
12 [palaeoclimate tool and its application to the terrestrial Paleogene and Neogene of North](#)  
13 [America. \*Palaeobiodiversity and Palaeoenvironments\* 91, 163-183](#)

14 Yemane, K., Bonnefille, R., and Faure, H.: Paleoclimatic and Tectonic Implications of  
15 Neogene Microflora from the Northwestern Ethiopian Highlands, *Nature*, 318, 653-656,  
16 1985.

17 ~~Zhang, Z., Nisancioglu, K. H., FlatÃy, F., Bentsen, M., Bethke, I., and Wang, H.: Tropical~~  
18 ~~seaways played a more important role than high latitude seaways in Cenozoic cooling, *Clim.*~~  
19 ~~*Past*, 7, 801-813, 10.5194/cp-7-801-2011, 2011.~~

20 ~~ZhongshiZhang, Z., Huijun, W., Zhengtang, G., and Dabang, J.: Impacts of tectonic changes~~  
21 ~~on the reorganization of the Cenozoic paleoclimatic patterns in China, *Earth and Planetary*~~  
22 ~~*Science Letters*, 257, 622-634, 2007a.~~

23 ~~ZhongshiZhang, Z., Wang, H., Guo, Z., and Jiang, D.: What triggers the transition of~~  
24 ~~palaeoenvironmental patterns in China, the Tibetan Plateau uplift or the Paratethys Sea~~  
25 ~~retreat? *Palaeogeogr. Palaeoclimatol., Palaeogeography, Palaeoclimatology, Palaeoecology*, 245,~~  
26 ~~317-331, 2007b.~~

27 ~~Zhang, Z., Nisancioglu, K. H., FlatÃy, F., Bentsen, M., Bethke, I., and Wang, H.: Tropical~~  
28 ~~seaways played a more important role than high latitude seaways in Cenozoic cooling, *Clim.*~~  
29 ~~*Past*, 7, 801-813, 10.5194/cp-7-801-2011, 2011.~~

Ref	Miocene Period	Model Name	Atmospheric Model Resolution	Ocean Model Type	Oceanic Model Resolution	Land-sea-Mask	Ice-Sheet Configuration	Orography	Vegetation	Modern-CO <sub>2</sub>	Miocene CO <sub>2</sub>	Same CO <sub>2</sub> for Modern and Miocene?
1	Early	CCM1	4.5°x7.5°	None		Miocene	All-ice-free	Miocene	N/A	Mo	Mo	Yes
2	Early	GENESIS	7.5°x4.5°	Slab		Miocene	All-ice-free	Miocene	Modern+Miocene	Mo	Mo	Yes
3	Late	LDM-5.3	2°x5.6°	None		Miocene	No-Greenland	Miocene	Modern	345	345	Yes
4	Late	ECHAM4	3.75°x3.75°	Slab		Modern	No-Greenland	Miocene	Miocene	353	353	Yes
5	Early	GENESIS	3.7°x3.7°	Slab		Modern	No-Greenland	Miocene	N/A	330,990	990	Yes (future)
6	Late	CCM0	4.5°x7.5°	None		Modern	Modern	Half+No-mountains	N/A	330	330	Yes
7	Late	HadAM3	2.5°x3.75°	None		Miocene	Reduced-Modern-Greenland	Miocene	Homogenous	280	395	No
8	Late	E-CHAM4	3.75°x3.75°	Slab		Modern	No-Greenland	Miocene	Miocene	353	353	Yes
9	Late	Planet Simulator	5.6°x5.6°	Slab		Modern	No-Greenland	Miocene	Miocene	280,360,700	280,360,700+	Yes
10	Late	COSMOS	3.75°x3.75°	Dynamic	3°x3°	Miocene	No-Greenland	Miocene	Miocene	360	360	Yes
11	All	CCM0 + CCM1	4.5°x7.5°	Slab		Modern	Modern+Last Glacial	No+Half-Himalayas	Modern	330	600, 1000	No
12	Late	LDM-5.3	2°x5.6°	Slab		Miocene	No-Greenland	Miocene	N/A	Modern	Modern	Yes
13	Late	CCM0	4.5°x7.5°	None		Modern	Modern	Half+No-mountains	N/A	330	330	Yes
14	Early	CCM1	4.5°x7.5°	Slab		Modern	Modern	Half+No-mountains	N/A	330	330,660	Yes
15	Late	E-CHAM4	3.75°x3.75°	Slab		Modern	No-Greenland	Miocene	Modern	353	353	Yes
16	Late	ECHAM4	3.75°x3.75°	Slab		Modern	No-Greenland	Miocene	Modern	353	700	No
17	Middle	CAM-v3.1/CLM-v3.0	3.75°x3.75°	Slab		Miocene	No-Greenland	Miocene	Miocene	379	180,355,700	No
18	Middle	CAM-v3.1/CLM-v3.0	3.75°x3.75°	Slab		Miocene	No-Greenland+West-Antarctic	Miocene	Miocene	355	180,350,700	No
19	All	IAP	4°x5°	None		Modern	Middle-Pliocene	Miocene	Modern	345	345	Yes
20	All	IAP	4°x5°	None		Modern	Middle-Pliocene	Miocene	Modern	345	345	Yes
21	Middle	Planet Simulator	5.6°x5.6°	Slab		Miocene	All-ice-free	Miocene+Modern	Miocene+Modern	280	200,280,500	Yes
22	Late	ECHAM5-MPIOM	3.75°x3.75°	Dynamic	3°x3°	Miocene	No-Greenland	Miocene	Miocene	280	280	Yes
23	Late	HadCM3L+TRIFFID	2.5°x3.75°	Dynamic	2.5°x3.75°	Miocene	Reduced-Modern Greenland+Antarctic	Miocene	Dynamic	280	280	Yes

References: 1: Barron, 1985; 2: Dutton and Barron, 1997; 3: Fluteau *et al.*, 1999; 4: Francois *et al.*, 2006; 5: Kutzbach and Behling, 2004; 6: Kutzbach *et al.*, 1989; 7: Lunt *et al.*, 2008a; 8: Micheels *et al.*, 2007a; 9: Micheels *et al.*, 2009; 10: Micheels *et al.*, 2011; 11: Prell and Kutzbach, 1992; 12: Ramstein *et al.*, 1997; 13: Ruddiman and Kutzbach, 1989; 14: Ruddiman *et al.*, 1997; 15: Stepphun *et al.*, 2006; 16: Stepphun *et al.*, 2007; 17: Tong *et al.*, 2009; 18: You *et al.*, 2009; 19: Zhongshi *et al.*, 2007a; 20: Zhongshi *et al.*, 2007b; 21: Henrot *et al.*, 2010; 22: Knorr *et al.*, 2011; 23: This study

Table 1. Details of published Miocene simulations.

Method	Calibration-range					Reference
Method	MAT (°C)	MAP (mm/a)	CMT (°C)	WMT (°C)	MAMLT (°C)	Reference
Climatic Amplitude Approach	3.4	257	<u>5.5</u>	<u>2.9</u>		Fauquette et al., 1998; <del>2006</del>
Coexistence Approach	2	200	as MAT	as MAT		Mosbrugger and Utescher, 1997; <i>This study</i>
Small mammal species composition		350-400				van Dam, 2006
Herpetofaunal composition		250-275				Bohme et al 2008
Mammal tooth-crown height		388				Eronen et al 2010
CLAMP	1-3		1.5-3	1-4		Wolfe 1994; Teodorides et al., (2011— <del>in press</del> )

Leaf morphology	1.98	57.97	3.55	2.92	Jacobs and Deino
Rodent species composition	0.004-4.8				Montuire et al 2006
Palaeosol Analysis		141			Retallack et al 2002
CLAMP Multiple regression analysis	0.7	180			Gregory-Wodzicki, 2000
<del>Physiognomic</del> Physiognometric/morphology reconstructions	as CLAMP		as CLAMP	as CLAMP	This study
Nearest living relative, autecology	as Climatic Amplitude Approach	as Climatic Amplitude Approach	<u>as Climatic Amplitude Approach</u>	<u>as Climatic Amplitude Approach</u>	This study

Table ~~12~~. Calibration errors for climate reconstructions of proxy data.

Variable	Epoch	CTRLc-LMdata datapoints in agreement	LM280c-LMdata datapoints in agreement	Total number of LMdata datapoints	Number of LM280c improvements over CTRLc All data (overlaps)	Number of LM280c deteriorations over CTRLc All data (overlaps)	<del>NetGlobal</del> LM280c improvement over CTRLc All data (overlaps)	Percentage Improvement LM280c over CTRLc All data (overlaps)
MAT (Macroflora)	Messinian	<u>1312</u>	<u>89</u>	<u>5540</u>	<u>20(211(3))</u>	<u>27(723(6))</u>	<u>-7(-512(-3))</u>	<u>-13(-930(-8))</u>
MAT (Microflora)		<u>5274</u>	<u>4153</u>	<u>159187</u>	<u>88(3392(34))</u>	<u>62(4476(52))</u>	<u>26(-1116(-18))</u>	<u>16(-79(-10))</u>
MAT (Fauna)		<u>69</u>	<u>79</u>	<u>79</u>	<u>0(10)</u>	<u>0(0)</u>	<u>0(10)</u>	<u>0(140)</u>
MAT (Macroflora)	Tortonian	<u>1748</u>	<u>1849</u>	<u>90404</u>	<u>4359(8)</u>	<u>3649(7)</u>	<u>740(1)</u>	<u>840(1)</u>
MAT (Microflora)		<u>2833</u>	<u>2932</u>	<u>98422</u>	<u>53(1473(13))</u>	<u>29(1330(14))</u>	<u>24(43(-1))</u>	<u>24(35(-1))</u>
MAT (Fauna)		<u>13</u>	<u>14</u>	<u>20</u>	<u>6(1)</u>	<u>1(0)</u>	<u>5(1)</u>	<u>25(5)</u>
<b>MAT (All)</b>	<b>lateLate Miocene</b>	<b><u>129456</u></b>	<b><u>117436</u></b>	<b><u>429479</u></b>	<b><u>210232(59)</u></b>	<b><u>155(71170(79))</u></b>	<b><u>55(-1262(-20))</u></b>	<b><u>13(-34)</u></b>
MAP (Macroflora)	Messinian	<u>4735</u>	<u>5038</u>	<u>5138</u>	<u>4(43(3))</u>	<u>1(10(0))</u>	<u>3(3)</u>	<u>6(68(8))</u>
MAP (Microflora)		<u>102428</u>	<u>146172</u>	<u>154184</u>	<u>50(4652(47))</u>	<u>4(3(2))</u>	<u>4748(44)</u>	<u>31(2927(24))</u>
MAP (Fauna)		<u>129262</u>	<u>135276</u>	<u>143298</u>	<u>14(736(17))</u>	<u>1(13(3))</u>	<u>13(633(14))</u>	<u>9(411(5))</u>
MAP (Macroflora)	Tortonian	<u>6979</u>	<u>7585</u>	<u>7989</u>	<u>9(740(8))</u>	<u>2(12)</u>	<u>78(6)</u>	<u>9(87)</u>
MAP (Microflora)		<u>7398</u>	<u>7697</u>	<u>94415</u>	<u>19(846(5))</u>	<u>7(6(5))</u>	<u>13(39(-1))</u>	<u>14(38(-1))</u>
MAP (Fauna)		<u>432435</u>	<u>482484</u>	<u>531532</u>	<u>96(5793(56))</u>	<u>1044(7)</u>	<u>86(5082(49))</u>	<u>1645(9)</u>
<b>MAP (All)</b>	<b>late Miocene</b>	<b><u>852</u></b>	<b><u>964</u></b>	<b><u>1052</u></b>	<b><u>192(129)</u></b>	<b><u>23(17)</u></b>	<b><u>169(112)</u></b>	<b><u>16(11)</u></b>
CMT (Macroflora)	Messinian	<u>31</u>	<u>38</u>	<u>48</u>	<u>16(11)</u>	<u>5(4)</u>	<u>11(7)</u>	<u>23(15)</u>
CMT (Microflora)		<u>106</u>	<u>135</u>	<u>142</u>	<u>36(34)</u>	<u>5(5)</u>	<u>31(29)</u>	<u>22(20)</u>
CMT (Fauna)		<u>53</u>	<u>57</u>	<u>74</u>	<u>18(13)</u>	<u>12(9)</u>	<u>6(4)</u>	<u>8(5)</u>
CMT (Macroflora)	Tortonian	<u>45</u>	<u>59</u>	<u>68</u>	<u>23(18)</u>	<u>4(4)</u>	<u>19(14)</u>	<u>28(21)</u>
CMT (Microflora)		<u>45</u>	<u>59</u>	<u>68</u>	<u>23(18)</u>	<u>4(4)</u>	<u>19(14)</u>	<u>28(21)</u>
CMT (Fauna)		<u>45</u>	<u>59</u>	<u>68</u>	<u>23(18)</u>	<u>4(4)</u>	<u>19(14)</u>	<u>28(21)</u>
<b>CMT (All)</b>	<b>late Miocene</b>	<b><u>235</u></b>	<b><u>289</u></b>	<b><u>332</u></b>	<b><u>93(76)</u></b>	<b><u>26(22)</u></b>	<b><u>67(54)</u></b>	<b><u>20(16)</u></b>
WMT (Macroflora)	Messinian	<u>0</u>	<u>2</u>	<u>48</u>	<u>27(2)</u>	<u>21(0)</u>	<u>6(2)</u>	<u>13(4)</u>
WMT (Microflora)		<u>5</u>	<u>9</u>	<u>143</u>	<u>15(5)</u>	<u>124(1)</u>	<u>-109(4)</u>	<u>-76(3)</u>
WMT (Fauna)		<u>2</u>	<u>6</u>	<u>72</u>	<u>47(4)</u>	<u>23(0)</u>	<u>24(4)</u>	<u>33(6)</u>
WMT (Macroflora)	Tortonian	<u>23</u>	<u>31</u>	<u>69</u>	<u>31(8)</u>	<u>15(0)</u>	<u>16(8)</u>	<u>23(12)</u>
WMT (Microflora)		<u>23</u>	<u>31</u>	<u>69</u>	<u>31(8)</u>	<u>15(0)</u>	<u>16(8)</u>	<u>23(12)</u>
WMT (Fauna)		<u>23</u>	<u>31</u>	<u>69</u>	<u>31(8)</u>	<u>15(0)</u>	<u>16(8)</u>	<u>23(12)</u>
<b>WMTMAP (All)</b>	<b>lateLate Miocene</b>	<b><u>304037</u></b>	<b><u>481452</u></b>	<b><u>3324253</u></b>	<b><u>120(19210(136))</u></b>	<b><u>183(127(24))</u></b>	<b><u>-63(18183(115))</u></b>	<b><u>-19(515(9))</u></b>
Megabiome	Messinian	<u>405</u>	<u>99</u>	<u>242</u>	<u>N/A(15)</u>	<u>N/A(21)</u>	<u>N/A(-6)</u>	<u>N/A(-2)</u>
Megabiome	Tortonian	<u>12496</u>	<u>12392</u>	<u>314225</u>	<u>N/A(1915)</u>	<u>N/A(2019)</u>	<u>N/A(-14)</u>	<u>N/A(0(-2))</u>
<b>Megabiome</b>	<b>late Miocene</b>	<b><u>229</u></b>	<b><u>222</u></b>	<b><u>556</u></b>	<b><u>N/A(34)</u></b>	<b><u>N/A(41)</u></b>	<b><u>N/A(-7)</u></b>	<b><u>N/A(-1)</u></b>

Table 23. Results from the model-data comparison for modern and lateLate Miocene palaeogeography

Variable	Epoch	LM280c-LMdata datapoints in agreement	LM400c-LMdata datapoints in agreement	Total number of LMdata datapoints	Number of LM400c improvements over LM280c All data (overlaps)	Number of LM400c deteriorations over LM280c All data (overlaps)	<del>Net</del> Global LM400c improvement over LM280c All data (overlaps)	Percentage Improvement LM400c over LM280c All data (overlaps)
MAT (Macroflora)	Messinian	89	3424	5540	39(2925(17)	9(38(2)	30(2617(15)	55(4743(38)
MAT (Microflora)		4153	101405	159487	116423(0)	3(223(12)	113(-2400(-12)	71(-153(-6)
MAT (Fauna)		79	79	79	0(0)	0(0)	0(0)	0(0)
MAT (Macroflora)	Tortonian	1849	5163	90404	63(4374(53)	18(1047(9)	45(3357(44)	50(3756(44)
MAT (Microflora)		2932	5977	98422	56(4177(58)	25(1143)	31(3052(45)	32(3143(37)
MAT (Fauna)		14	19	20	6(5)	0(0)	6(5)	30(25)
<b>MAT (All)</b>	<del>late</del> <b>Late</b> <b>Miocene</b>	<b>117436</b>	<b>271297</b>	<b>429479</b>	<b>280(180305(197)</b>	<b>55(2673(36)</b>	<b>225(154232(164)</b>	<b>52(3648(34)</b>
MAP (Macroflora)	Messinian	5038	4938	5138	10(0)	1(10(0)	0(-1(0)	0(-2(0)
MAP (Microflora)		146172	147174	154484	5(16(2)	23(0)	3(12)	2(1)
MAP (Fauna)		135276	135277	143298	7(047(3)	1(07(2)	6(040(4)	43(0)
MAP (Macroflora)	Tortonian	7585	7687	7989	4(2)	1(10(0)	3(14(2)	4(12)
MAP (Microflora)		7697	7597	94445	15(1)	4(24)	11(-1(0)	12(-140(0)
MAP (Fauna)		482484	483485	531532	24(5)	2928(4)	-54(1)	-1(0)
<b>MAP (All)</b>	<del>late</del> <b>Late</b> <b>Miocene</b>	<b>9641152</b>	<b>9654158</b>	<b>10524253</b>	<b>56(966(13)</b>	<b>38(842(7)</b>	<b>18(124(6)</b>	<b>2(0)</b>
CMT (Macroflora)	<del>Messinian</del>	38	42	48	10(7)	3(3)	7(4)	15(8)
CMT (Microflora)		135	141	142	7(7)	1(1)	6(6)	4(4)
CMT (Fauna)		57	65	74	17(11)	3(3)	14(8)	19(11)
CMT (Macroflora)	<del>Tortonian</del>	59	64	68	7(7)	4(2)	3(5)	4(7)
CMT (Microflora)		57	65	74	17(11)	3(3)	14(8)	19(11)
CMT (Fauna)		59	64	68	7(7)	4(2)	3(5)	4(7)
<b>CMT (All)</b>	<del>late</del> <b>Miocene</b>	<b>289</b>	<b>312</b>	<b>332</b>	<b>41(32)</b>	<b>11(9)</b>	<b>30(23)</b>	<b>9(7)</b>
WMT (Macroflora)	<del>Messinian</del>	2	13	48	46(11)	0(0)	46(11)	96(23)
WMT (Microflora)		9	14	143	134(5)	0(0)	134(5)	94(3)
WMT (Fauna)		6	23	72	66(17)	0(0)	66(17)	92(24)
WMT (Macroflora)	<del>Tortonian</del>	31	43	69	38(12)	0(0)	38(12)	55(17)
WMT (Microflora)		31	43	69	38(12)	0(0)	38(12)	55(17)
WMT (Fauna)		31	43	69	38(12)	0(0)	38(12)	55(17)
<b>WMT (All)</b>	<del>late</del> <b>Miocene</b>	<b>48</b>	<b>93</b>	<b>332</b>	<b>284(45)</b>	<b>0(0)</b>	<b>284(45)</b>	<b>86(14)</b>
Megabiome	<del>Messinian</del>	99	119	242	N/A(39)	N/A(19)	N/A(20)	N/A(8)
Megabiome	<del>Tortonian</del>	12392	158424	314225	N/A(5838)	N/A(239)	N/A(3529)	N/A(1143)
<b>Megabiome</b>	<del>late</del> <b>Miocene</b>	<b>222</b>	<b>277</b>	<b>556</b>	<b>N/A(97)</b>	<b>N/A(42)</b>	<b>N/A(55)</b>	<b>N/A(10)</b>

Table 34. Results from the model-data comparison for high and low CO<sub>2</sub> concentration assumptions

Figure 1. Atmospheric CO<sub>2</sub> reconstructions for the ~~late~~Late Miocene.

Figure 2. Model orographic and ice sheet configurations for the Preindustrial (CTRL) and the ~~late~~Late Miocene (LM). The LM panel shows the percentage differences in high topography between CTRL and LM.

Figure 3. Model-data comparison definitions. Panel A shows the definition for ‘overlap’ used in the model-data comparison: all four instances shown are considered as an overlap. Panel B shows that although we define model-data mismatch as the minimum possible distance to overlap, the maximum possible differences could be much greater if the true values for both the model and the data were to lie at the extremes of the uncertainty ranges.

Figure 4. Difference between the ~~late~~Late Miocene and Preindustrial simulations (LM280-CTRL: left side) and between the ~~late~~Late Miocene CO<sub>2</sub> simulations (LM400-LM280: right side) for mean air temperature. Only significant differences are shown using a 95% confidence interval Student’s T-Test, white areas are not significant.

Figure 5. Difference between the ~~late~~Late Miocene and Preindustrial simulations (LM280-CTRL: left side) and between the ~~late~~Late Miocene CO<sub>2</sub> simulations (LM400-LM280: right side) for mean precipitation. Only significant differences are shown using a 95% confidence interval Student’s T-Test, white areas are not significant.

Figure 6. Dominant PFTs of the ~~late~~Late Miocene and Preindustrial simulations, as predicted by the TRIFFID dynamic vegetation model.

Figure 7. Results from the model-data comparison for mean annual temperature, ~~late~~Late Miocene data – modern potential natural climate estimates.

Figure 8. Results from the model-data comparison for mean annual temperature, ~~late~~Late Miocene data – LM280c.

Figure 9. Results from the model-data comparison for mean annual temperature, ~~late~~Late Miocene data – LM400c.

Figure 10. Improvements in the model-data comparison for mean annual temperature. The lefthand column (A,B) shows the improvement that the ~~late~~Late Miocene palaeogeography makes to the model-data comparison. The righthand column (C, D) shows the improvement that higher CO<sub>2</sub> makes to the model-data comparison. Green circles indicate an improvement, red circles indicate a deterioration. The datapoints showing ‘no difference’ are plotted underneath the other datapoints in order to highlight the differences.

Figure 11. Results from the model-data comparison for mean annual precipitation, ~~late~~Late Miocene data – modern potential natural climate estimates

Figure 12. Results from the model-data comparison for mean annual precipitation, ~~late~~Late Miocene data – LM280c

Figure 13. Results from the model-data comparison for mean annual precipitation, ~~late~~Late Miocene data – LM400c

Figure 14. Improvements in the model-data comparison for mean annual precipitation. The lefthand column (A, B) shows the improvement that the ~~late~~Late Miocene palaeogeography makes to the model-data comparison. The righthand column (C, D) shows the improvement that higher CO<sub>2</sub> makes to the model-data comparison. Green circles indicate an improvement, red circles indicate a deterioration. The datapoints showing ‘no difference’ are plotted underneath the other datapoints in order to highlight the differences.

Figure 15. Results from the model-data comparison for ~~Tortonian~~-megabiomes: ~~late~~Late Miocene data on CTRLc, LM280c, and LM400c

Figure 16. Improvements in the model-data comparison for ~~Tortonian~~-megabiomes. ~~Panels~~Panel A and C shows the improvement that the ~~late~~Late Miocene palaeogeography makes to the model-data comparison. Panel B and D shows the improvement that higher CO<sub>2</sub> makes to the model-data comparison. Green circles indicate an improvement, red circles indicate a deterioration.

Figure 17. Model-~~data~~Data comparison summary for MAT and MAP. Shown is the percentage of the total number of datapoints that overlap with the model results.

Figure 18. Model-data comparison summary for CMT and WMT. Shown is the percentage of the total number of datapoints that overlap with the model results.

Figure S1. Results from the model-data comparison for mean cold month temperature, late Miocene data – modern potential natural climate estimates.

Figure S2. Results from the model-data comparison for mean cold month temperature, late Miocene data – LM280c.

Figure S3. Results from the model-data comparison for mean cold month temperature, late Miocene data – LM400c.

Figure S4. Improvements in the model-data comparison for mean cold month temperature. The lefthand column (A,B) shows the improvement that the late Miocene palaeogeography makes to the model-data comparison. The righthand colum (C, D) shows the improvement that higher CO<sub>2</sub> makes to the model-data comparison. Green circles indicate an improvement, red circles indicate a deterioration. The datapoints showing ‘no difference’ are plotted underneath the other datapoints in order to highlight the differences.

Figure S5. Results from the model-data comparison for mean warm month temperature, late Miocene data – modern potential natural climate estimates

Figure S6. Results from the model-data comparison for mean warm month temperature, late Miocene data – LM280c

Figure S7. Results from the model-data comparison for mean warm month temperature, late Miocene data – LM400c

Figure S8. Improvements in the model-data comparison for mean warm month temperature. The lefthand column (A,B) shows the improvement that the late Miocene palaeogeography makes to the model-data comparison. The righthand colum (C, D) shows the improvement that higher CO<sub>2</sub> makes to the model-data comparison. Green circles indicate an improvement, red circles indicate a deterioration. The datapoints showing ‘no difference’ are plotted underneath the other datapoints in order to highlight the differences.

Figure S9. Model-data comparison summary for MAT and MAP including CTRL400c. Shown is the percentage of the total number of datapoints that overlap with the model results.