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# The early Eocene equable climate problem revisited

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241

## Abstract

The early Eocene “equable climate problem”, i.e. warm extratropical annual mean and above-freezing winter temperatures evidenced by proxy records, has remained as one of the great unsolved problems in paleoclimate. Recent progress in modeling and paleoclimate proxy development provides an opportunity to revisit this problem to ascertain if the current generation of models can reproduce the past climate features without extensive modification. Here we have compiled early Eocene terrestrial temperature data and compared with climate model results with a consistent and rigorous methodology. We test the hypothesis that equable climates can be explained simply as a response to increased greenhouse gas forcing within the framework of the atmospheric component of the Community Climate System Model (version 3), a climate model in common use for predicting future climate change. We find that, with suitably large radiative forcing, the model and data are in general agreement for annual mean and cold month mean temperatures, and that the pattern of high latitude amplification recorded by proxies can be reproduced.

## 1 Introduction

The early Eocene (56–48 Ma) encompasses the warmest climates of the past 65 million years. Annual-mean and cold-season continental temperatures were substantially warmer than modern, while meridional temperature gradients were greatly reduced (Wolfe, 1995; Greenwood and Wing, 1995; Barron, 1987). Reconstructions of warm climates on land are confirmed in the marine realm, with bottom water temperatures ~10 °C higher than modern values (Miller et al., 1987; Zachos et al., 2001; Lear et al., 2000). This implies that early Eocene winter temperatures in deep water formation regions, located at the surface in high latitudes, could not have dropped much below 10 °C, consistent with the high-latitude occurrence of frost-intolerant flora and fauna (Greenwood and Wing, 1995; Spicer and Parrish, 1990; Hutchison, 1982; Wing and Greenwood, 1993; Markwick, 1994, 1998).

242

Modelling studies of the early Eocene over the past decades have consistently failed to reproduce the warm continental interior temperatures inferred from paleoclimate proxies, an issue that has come to be known as the “equable climate problem” (Sloan and Barron, 1990, 1992; Sloan, 1994). The model-data mismatch is typically  $\sim 20^{\circ}\text{C}$  for winter temperatures; for mean annual temperature (MAT) the error is typically less, but reaches  $10\text{--}20^{\circ}\text{C}$  near the poles (Shellito et al., 2003; Huber et al., 2003; Winguth et al., 2010; Roberts et al., 2009; Shellito et al., 2009). An apparently similar model-data discrepancy exists for the Cretaceous (Spicer et al., 2008; Donnadieu et al., 2006), but we restrict ourselves here to the early Eocene. On the other hand, models have had reasonable success at simulating the cooler intervals of the early Paleogene, such as the middle-to-late Eocene (Roberts et al., 2009; Liu et al., 2009; Eldrett et al., 2009) and Paleocene (Huber, 2009).

The early Eocene equable climate problem, with its suggestion that climate models may lack or misrepresent crucial processes responsible for the warm continental temperatures, has stimulated much innovative thinking in climate modelling (Valdes, 2000). Routes to generating warm early Eocene winter continental interior and polar temperatures that have been explored include: large lakes (Sloan, 1994; Morrill et al., 2001); polar stratospheric clouds (Sloan and Barron, 1992; Sloan and Pollard, 1998; Sloan et al., 1999; Peters and Sloan, 2000; Kirk-Davidoff et al., 2002; Kirk-Davidoff and Lamarque, 2008); increased ocean heat transport (Berry, 1922; Covey and Barron, 1988; Sloan et al., 1995); finer resolution simulations of continental interiors (Sewall and Sloan, 2006; Thrasher and Sloan, 2009); a permanent, positive phase of the Arctic Oscillation (Sewall and Sloan, 2001); altered orbital parameters (Sloan and Morrill, 1998; Lawrence et al., 2003; Sewall et al., 2004); altered topography and ocean gateways (Sewall et al., 2000); radiative convective feedbacks (Abbot et al., 2009a); altered vegetation (Sewall et al., 2000; Shellito and Sloan, 2006a,b), and changes in sea surface temperature (SST) distributions (Sloan et al., 2001; Huber and Sloan, 1999; Sewall and Sloan, 2004).

243

Recent work includes interactive ocean-atmosphere coupling; these studies have either failed to reproduce warm winter season temperatures (Roberts et al., 2009; Huber et al., 2003; Shellito et al., 2009) or not presented a comparison of these terrestrial winter temperatures (Tindall et al., 2010; Lunt et al., 2010; Huber and Sloan, 2001; Heinemann et al., 2009; Winguth et al., 2010) with proxies, so it seems important at this point to revisit the current status of the early Eocene equable climate problem.

While many of these studies have shown regional improvement in the model-data mismatch, the general outcome of the investigations cited above has been a failure to provide a single, parsimonious, global solution to the equable climate problem. One mechanism (e.g. cloud feedbacks or ocean gateway opening) might explain warmth at polar latitudes but leave temperatures in the western interior of North America unexplained, or vice versa. The failure of these various hypothesized resolutions to the equable climate problem – whether tested individually or in concert – has fostered a sense that they are not the leading order solution and that something major is missing from our understanding of the climate system and its representation in conventional climate models (Zeebe et al., 2009). It is therefore important to revisit this problem with the latest generation of models and with an up-to-date proxy data compilation.

One ostensibly straightforward route to resolving the equable climate problem that has not been fully explored to date is simply to raise greenhouse gas forcing sufficiently to yield above-freezing continental temperatures year round. It is known that  $\text{CO}_2$  concentrations were above modern during the Eocene (Pearson and Palmer, 2000; Pearson et al., 2009; Pagani et al., 2005; Henderiks and Pagani, 2008; Lowenstein and Demicco, 2006), and estimates are as high as  $\sim 4400$  ppmv (Pearson and Palmer, 2000). In addition to its direct warming effect, increased  $p\text{CO}_2$  may have primed the climate system to be sensitive to forcing by alterations in other boundary conditions, such as ocean gateways (Sijp et al., 2009) or insolation, or it may have enhanced nonlinear sensitivity through feedbacks, e.g. due to wetland methane emissions (Sloan et al., 1992; Beerling et al., 2009a).

244

The reluctance to pursue the avenue of enhanced greenhouse gas forcing has clear origins: it is very difficult to simultaneously achieve warm continental interiors without overheating the tropics. Until recently, tropical surface temperature reconstructions indicated Early Eocene tropical surface temperatures comparable or even lower than modern (Zachos et al., 1994; Crowley and Zachos, 2000), and climate models easily exceed these temperatures even at CO<sub>2</sub> levels insufficient to give the required mid- to high-latitude terrestrial warming (e.g. Shellito et al., 2003). Thus, the equable climate problem is intimately related to – though distinct from – the “cool tropics” or “low gradient problem” (Barron, 1987; Adams et al., 1990; Huber et al., 2003).

However, the strictures imposed by the low gradient problem have been considerably relaxed recently by the realization that older tropical temperature reconstructions were subject to diagenetic cold bias, and more recent reconstructions using a range of proxies indicate that tropical temperatures may actually have been as high as ~35 °C (see discussion in Sect. 2.1). As suggested by Huber et al. (2003), polar annual mean temperatures of ~15 °C could potentially exist in equilibrium with such warm tropical temperatures without invoking novel climate mechanisms. This opens up the possibility of resolving the equable climate problem by simply raising greenhouse forcing, without recourse to novel mechanisms.

In this paper we test the hypothesis that the equable climate problem, i.e. warm extratropical annual mean and above-freezing winter temperatures, can be explained simply as a response to increased greenhouse gas forcing within the framework of the Community Climate System Model version 3 (CCSM3). CCSM3 is a climate model in common use for predicting future climate change, used here with standard physics (Collins et al., 2006). The focus is on validation of the model through model-data comparison.

The study is structured as follows. First, Sect. 2 reviews the relevant temperature and CO<sub>2</sub> reconstructions, which guide decisions on the methodology presented in Sect. 3. In Sect. 4, we present a set of Eocene atmospheric general circulation model experiments compared with early Eocene proxy records. We discuss the robustness and

245

limitations of this study in Sect. 5. Finally, Sect. 6 summarizes our conclusions and discusses some implications of the model-data agreement.

## 2 Interpreting proxy data constraints

Some preliminary considerations must be dealt with before the model-data comparison can be carried out, because proxy records are unavoidably subject to varying interpretations. While there may not be universal agreement on the best interpretative approach, we aim here to at least be clear what our framework is. In this section, we review characterizations of the early Eocene paleotemperature and greenhouse gas records, with an emphasis on uncertainties and potential biases. Since this paper focuses on terrestrial records, the discussion of ocean proxies is brief, but it is important for context because SSTs are being interpreted to much warmer values previously thought. We briefly discuss the ocean temperature proxy record in the next section, but given the uncertainties that remain in proxy calibrations and assumptions that go into those proxies, they deserve their own comprehensive treatment and model-data comparison. As described in Sect. 2.2 a factor that has been poorly appreciated by the modelling community is that it is now well acknowledged in the paleobotanical literature that most published terrestrial paleoclimate records have also been significantly biased to overly cool values. In this study we focus on annual mean and winter season terrestrial temperatures in order to assess progress in the early Eocene equable climate problem *sensu strictu*.

### 2.1 Sea surface temperature records

A new characterization of Eocene temperature has recently developed as the product of both a better understanding of diagenetic contamination of older tropical SST records (Schrag et al., 1995; Schrag, 1999; Huber and Sloan, 2000; Pearson et al.,

246

2001b, 2007, 2008) and the development of new proxies such as TEX<sub>86</sub> and Mg/Ca (Schouten et al., 2002, 2003; Pearson et al., 2007; Lear et al., 2008; Sexton et al., 2006; Sluijs et al., 2006, 2007, 2008; Liu et al., 2009) for ocean near-surface temperatures. As summarized in Huber (2008), SSTs of ~35 °C are now reconstructed in the  
5 early Eocene tropics (Pearson et al., 2001b, 2007; Tripathi et al., 2003; Tripathi and Elderfield, 2004; Zachos et al., 2003). Extratropical SSTs are also reconstructed to values hotter than previously thought (Bijl et al., 2009; Sluijs et al., 2006, 2009; Zachos et al., 2006; Hollis et al., 2009; Creech et al., 2010; Liu et al., 2009; Eldrett et al., 2009).

These hot temperatures have major implications for our understanding of past climate dynamics, and of the equable climate problem in particular. Unfortunately, large and difficult-to-quantify uncertainties persist in proxy interpretations (Shah et al., 2008; Ingalls et al., 2006; Herfort et al., 2006; Kim et al., 2010; Liu et al., 2009; Pearson et al., 2001a; Hugué et al., 2006, 2007; Wuchter et al., 2004, 2005; Hugué et al., 2009; Trommer et al., 2009; Turich et al., 2007). Of particular concern is the need to  
15 extrapolate calibrations out to temperatures and environmental conditions far beyond modern values. Furthermore, calibrations have proliferated: for example, 6 different calibrations are now in use for the TEX<sub>86</sub> proxy, and a site specific calibration has been frequently recommended (Hugué et al., 2007; Kim et al., 2010), but this is difficult to implement in deep paleoclimates.

20 If the different proxies told consistent stories, these uncertainties might be less consequential. Unfortunately, different SST proxies and records present apparently irreconcilable visions. For example, the presence of 11 °C South Atlantic temperatures (Ivany et al., 2008) in the same latitude band as 30 °C temperatures in the South Pacific (Bijl et al., 2009; Hollis et al., 2009) raises questions about the proxy interpretations.  
25 The occurrence of ~10 °C deep ocean temperatures (Zachos et al., 2001) requires that some regions see temperatures fall to this value at least in winter, in agreement with the results of (Ivany et al., 2008), but South Pacific records seem to preclude temperatures this cold. Significant work remains to be done to resolve these discrepancies. Reconciling the different approaches and narrowing uncertainty in SSTs is beyond the

scope of this paper. Indeed, we believe that by focusing on terrestrial temperatures this will create a solid benchmark for comparison with SST records in future work.

## 2.2 Terrestrial temperature records

Prior studies have discussed ways in which existing methodologies and sampling approaches introduce warm biases. These include undersampling large regions that were  
5 likely cooler than the mean, such as Antarctica, Siberia and northeastern North America, as well as preferential temporal sampling, e.g. clipping seasonal or orbital scale cycles (Sloan and Barron, 1990; Sloan, 1994; Valdes, 2000). These factors are probably important considerations for developing a better understanding of regional scale  
10 climate patterns, but are unlikely to undermine the widespread evidence of continuous warmth in the early Eocene or overwhelm the large cool biases in existing terrestrial reconstructions. On the contrary, many factors conspire to introduce an overall cool bias in the terrestrial paleotemperature record.

15 Firstly, almost no terrestrial records from the Eocene have been obtained for 30° N to 30° S band, effectively clipping the the warmest climatic end-member. Second, where records have been derived, various other factors contribute toward a cool bias. Taphonomic and ecological factors in leaf physiognomic techniques (Burnham, 1989; Burnham et al., 2001; Boyd, 1994; Greenwood, 2005; Spicer et al., 2005; Kowalski, 2002; Kowalski and Dilcher, 2003) have been shown to lead to systematic cold biases of  
20 2–8 °C. Further cool biases are introduced by under-sampling the flora, particularly in the early, pioneering studies (Wilf, 1997; Burnham et al., 2005; Wilf et al., 2003). Third, at high latitudes, polar deciduous habits skew high latitude “toothiness” based MAT interpretations to low values (Boyd, 1990, 1994). Fourth, at low latitudes, floral physiognomic techniques are biased to cool values because the proxy becomes insensitive at temperatures much warmer than modern (Head et al., 2009). And finally,  
25 many records come from regions with significant paleo-elevation (Wyoming, Okanagan Highlands Wolfe et al., 1998; Smith et al., 2009) and hence may record temperatures cooler than sparsely-sampled low elevation, low relief areas, though this effect may be

partially offset by the fact that records are frequently derived from basins within these high-relief regions (Sewall et al., 2000).

If the early Eocene was even warmer than previously thought – as this review of terrestrial and ocean proxies suggest (their large uncertainties notwithstanding) – the mystery of Eocene equable climate deepens unless either the radiative forcing or sensitivity were greater than has been typically explored. It also implies that it is very difficult to achieve early Eocene conditions in models and that studies purporting to simulate the early Eocene may instead only be warm enough to match late or middle Eocene conditions. From this perspective the approach that we utilize in this study, in which various efforts are made to overcome the cold bias of previous proxy compilations, makes the goal of reproducing equable climates harder to reach.

### 2.3 Greenhouse gases and radiative forcing

Increased radiative forcing, usually ascribed to greenhouse gases, is part of every feasible solution to the equable climate problem tried so far. But, the upper range of likely greenhouse gas forcing has expanded from prior work. Modeling studies have predominantly explored the low end (520–2000 ppm) of paleo-CO<sub>2</sub> proxy estimates even though the range extends easily up to 4400 ppm in the early Eocene (Pearson and Palmer, 2000). This partly reflects attention to the lower end of CO<sub>2</sub> estimates typically derived from leaf stomatal indices (Royer et al., 2001; Beerling and Royer, 2002a). Yet the calibration of this proxy at high CO<sub>2</sub> is weakly constrained (Beerling and Royer, 2002b; Beerling et al., 2009b) and recent evidence that leaves adapt the size of the stomata and not their density at high CO<sub>2</sub> values (Franks and Beerling, 2009) raises questions about the validity of the proxy at high CO<sub>2</sub>. Indeed, as recently affirmed by Smith et al. (2010) the stomatal methods should probably be considered semi-quantitative under high CO<sub>2</sub> conditions.

Boron and alkenone approaches also have increasing error bars at high CO<sub>2</sub>, because their calibrations lose sensitivity at high values (Pagani, 2002; Pearson and Palmer, 2000). The nahcolite approach of Lowenstein and Demicco (2006) only

249

constrains the minimum early Eocene CO<sub>2</sub>, and any value above 1125–2985 ppm is possible.

Overall, existing proxy records have much greater accuracy at low CO<sub>2</sub> and once values are significantly higher than modern (somewhere above 560 to 1220 ppm, depending on the proxy) there is little certainty in the actual value, although constraints can be inferred from carbon mass balance models (Zeebe et al., 2009; Panchuk et al., 2008). These constraints currently allow for a wide range of potential early Eocene CO<sub>2</sub> values.

Carbon dioxide is also not the only important source of radiative forcing. Methane concentrations, for example, could have been much higher in the early Eocene (Sloan and Barron, 1992; Sloan et al., 1999; Beerling et al., 2009a). Alternatively clouds may have functioned differently (Sloan and Pollard, 1998; Kump and Pollard, 2008). We have no proxies for either of these factors, which is one reason why relatively few studies have incorporated them. They represent arguably reasonable, but unconstrained conjectures.

Furthermore, strong radiative forcing in simulations (Shellito et al., 2003; Kump and Pollard, 2008) produce tropical SSTs warmer than traditional reconstructions (Zachos et al., 1994; Crowley and Zachos, 2000) and hotter even than revised reconstructions (Pearson et al., 2007; Huber, 2008). So while it is understood that stronger radiative forcing in models warms extratropical continental interiors, Eocene tropical temperatures constraints provided a primary motivation for not forcing models with greenhouse gas forcing sufficient to maintain extratropical winter warmth (Barron, 1987; Huber and Sloan, 2000).

Finally, it is perhaps an accident of history that much of the Eocene modelling work was performed with the NCAR atmospheric models (CCM, CAM, GENESIS), which have consistently had relatively weak sensitivity of global mean temperature to increased greenhouse gas concentrations (2–3 °C warming per doubling of pCO<sub>2</sub>) (Kothavala et al., 1999; Otto-Bliesner et al., 2006; Kiehl et al., 2006). Other models, such as the UK models and ECHAM5/MPI models, have higher sensitivities and have

consistently produced temperatures as warm as high-CO<sub>2</sub> CAM simulations with lower radiative forcing (Heinemann et al., 2009; Jones et al., 2010). These more sensitive models run into the same problems with tropical SSTs, though, because they have similar (Holland and Bitz, 2003) although not identical (Abbot et al., 2009a) amounts of high-latitude amplification of global warming after normalizing by climate sensitivity. These Eocene studies in models with high sensitivity have not quantitatively examined the issue of continental winter temperatures, which leaves an important gap in evaluating how close we are to solving this long-standing climate problem.

### 3 Methods

#### 3.1 Proxy records

Various factors must be considered in interpreting climate from proxy records and ambiguity exists in any paleoclimate model-data comparison. Here we describe the methodologies we developed for the model data comparison and the inclusion of random and biased sources of uncertainty and error. The details of the proxy data used in this study, including original references and calibration information are summarized in Table 1.

##### 3.1.1 MAT

Macrofloral assemblage data provide many of the quantitative paleotemperature estimates for the Eocene. Different approaches to estimating paleotemperature with macroflora, such as CLAMP (Climate-Leaf Analysis Multivariate Program) (Wolfe, 1995) and LMA (Leaf Margin Analysis) (Wilf, 1997), usually lead to qualitatively similar but quantitatively different MAT estimates. Even within these methodologies, the impacts of differing calibration data sets and approaches can lead to significantly different estimates (Greenwood et al., 2004, 2003; Adams et al., 2008; Spicer et al., 2009). Comparison of different approaches often leads to estimates that differ beyond

251

the stated error estimates of the underlying calibration studies (Uhl et al., 2007). Trends are normally more robust, but since model-data comparison requires aggregating data into “snapshots”, uncertainty in absolute values, rather than trends, is crucial to this study.

For these reasons it is more important, for the purposes of this study, to properly account for systematic biases and spatio-temporal sampling uncertainty in proxy records rather than be primarily concerned with the stated random accuracy of the methodologies, which likely substantially underestimate the true uncertainty. Wherever possible in this study, we chose to minimize systematic errors while maintaining as consistent an approach in estimating temperature as possible.

To that end, we primarily rely on LMA to estimate MAT, as this approach has a long history and widespread usage in the literature, and it is arguably less sensitive to subjectivity in scoring, sampling, and calibration than other approaches such as CLAMP (Wilf, 1997; Peppe et al., 2010). To offset the well established cool biases in the LMA approach we use the Kowalski and Dilcher (2003) calibration wherever possible or unless a compelling reason exists not to. In some cases the only published data available are from pollen-derived transfer functions or CLAMP-derived temperatures. In another important special case, e.g. Australian flora, there are well-established systematic differences in the relationship between toothiness and temperature that arise from the long isolation of Australia’s flora. Consequently we use the Australian calibration (Greenwood et al., 2004) for Australian and New Zealand flora. See Table 1 for more details.

More fundamentally, it is unclear how to apply this proxy, or any proxy, outside of its calibration regime, in particular at temperatures much hotter – and with higher  $p\text{CO}_2$  and potentially greater rainfall – than those observed in modern vegetated regions (Utescher et al., 2009). Today, MAT above 28 °C does not occur in well-vegetated regions and such techniques are suspect and likely underestimate MAT in regions that are likely to be above that range (Head et al., 2009), since leaves can not be more than 100% entire margined. We have every reason to think that mean MAT in the tropics

252

was above 30 °C in the early Eocene, but just how much hotter is an unanswered question (Huber, 2008; Jaramillo et al., 2010; Kobashi et al., 2004). We will not be able to address with great rigor the true warmth of the tropical to subtropical regions in this study because of a paucity of terrestrial proxies and the ambiguity of interpreting such records under non-analogue conditions. The only relevant tropical terrestrial MAT constraint we were able to identify from published records and it was middle Eocene in age, but we include it in this study for lack of any other available data (See Table 1).

In some regions, where few macrofloral records exist or where LMA or CLAMP have not been applied, other proxies are available and so we rely on them to fill in the large spatial gaps. Especially crucial to constrain are high and low latitude temperatures. Such additional sources of quantitative information include: nearest Living Relative (NLR) transfer functions, isotopic estimates, and organic geochemical indicators. MAT can be estimated from palynoflora (Eldrett et al., 2009; Greenwood et al., 2010) and from various Nearest Living Relative based empirical correlations with MAT (Utescher and Mosbrugger, 2007; Utescher et al., 2009; Mosbrugger et al., 2005; Poole et al., 2005) and allometric considerations (Head et al., 2009). Oxygen and hydrogen isotopes provide paleotemperature estimates provided certain parameters are well constrained (Eberle et al., 2010; Koch, 1998; Fricke and Wing, 2004; Jahren and Sternberg, 2003; Jahren, 2007; Jahren and Sternberg, 2008; Jahren et al., 2009).

The MBT-CBT proxy is an organic geochemical proxy for annual mean air temperature derived ultimately from soil bacteria (Weijers et al., 2007b). We use this proxy to provide additional information in the early Eocene, although it is possible that the proxy is biased to summer temperatures at extreme high latitudes (Weijers et al., 2007a; Eberle et al., 2010), or worse not reflect surface temperature at all (Peterse et al., 2009). Attempts to include other information from kaolinite-isotope MAT proxy records (Sheldon and Tabor, 2009) were unsuccessful because suitable age constraints were not available.

253

### 3.1.2 Seasonal temperatures

MAT proxies are not the only, or even most relevant variable to consider, given that seasonality, or the lack thereof, is the defining characteristic of the equable climate problem (Sloan and Barron, 1990, 1992). The most vexing difficulty in prior work has been in explaining warm winter temperatures. Various approaches to constraining winter temperatures exist. These include using: CLAMP-based cold month mean (CMM) estimates (Wolfe, 1995; Spicer et al., 2009), the palm/cycad line (Greenwood and Wing, 1995; Wing and Greenwood, 1993; Eldrett et al., 2009), crown crocodylian presence (Hutchinson, 1982; Markwick, 1994, 1998, 2007), or isotopic analysis (Eberle et al., 2010). These methodologies provide different kinds of information, e.g., CLAMP provides quantitative, explicit estimates of CMM, whereas the palm/cycad line and crocodylian indicators constrain temperatures to be higher than a threshold value. Differing comparison methodologies are required to the complementary information provided by these different proxies.

The abovementioned studies and others demonstrate that the early Eocene did not experience below freezing temperatures over a huge expanse of the land's surface, except perhaps in intermontane Canadian Rockies (Wing, 1987; Spicer and Parrish, 1990; McIver and Basinger, 1999; Greenwood et al., 2005; Smith et al., 2009). For some regions (central Antarctica, north central Canada and parts of Eurasia), data coverage is too sparse to state this definitively, but the existing high latitude data indicate temperatures were so warm it is difficult to physically justify large zones substantially cooler than freezing even in most of the regions with missing data. This is especially true given that inclusion of other potential factors such as higher  $p\text{CO}_2$  decreases the tolerance of frost-sensitive flora to cold temperatures (Royer et al., 2002). The presence of crown crocodylians in Kazakhstan and Mongolia also fills in the gaps and argue against temperatures below freezing in the regions we might expect to be the coldest (Markwick, 1998, 2007). The important exceptions being in inland Antarctica and at high paleoelevations, where proxy data to constrain temperatures are sparse

254

to non-existent and physical considerations indicate that temperatures may have been cold.

This interpretation does not rely overly on any one proxy being correct as they are independently corroborated by lack of sea ice diatoms Stickleby et al. (2009), high latitude occurrence of tropical clays (Robert and Chamley, 1991), warm polar SSTs (Bijl et al., 2009; Sluijs et al., 2006), and 11 °C bottom water temperatures (Zachos et al., 2001; Lear et al., 2000). These disparate lines of evidence all which argue for temperatures remaining above freezing year-round in the regions and seasons expected to be cold.

The published error bars around winter season temperature reconstructions are non-negligible and those around summer temperatures are even larger. More of concern are the conceptual issues involved in trying to use modern floral/climate relationships to estimate past climate regimes with no clear analogue (Utescher et al., 2009) such as subtropical climates in polar night. When strongly restrictive and conserved traits are involved, such is clearly the case with freezing temperatures and palm trees and crocodiles, the proxy data provide a stronger constraint than in regions of climate regime space where no clear strong biological constraints exist or which no flora currently occupy.

Based on these considerations, we focus attention on the winter season aspect of the equable climate problem and do not make a quantitative comparison with summer temperature proxy reconstructions. In this study we find that winter temperatures were above freezing everywhere except for interior Antarctic and small regions of high paleoelevation, primarily restricted to western North America. We adopt a two-pronged approach to model-data comparison for winter temperatures.

First, we show that seasonal-mean modelled temperatures everywhere (with the exceptions mentioned) were above freezing. This broad-strokes approach enables comparison with many of the crucial, but qualitative, indicators and characterizations of warm winters that are ubiquitous in the early Eocene geological record, including inferences from floral and faunal assemblages and soil and clay types, (Greenwood and Wing, 1995; Markwick, 2007; Pigg and Devore, 2010; Collinson and Hooker, 2003;

255

Valdes, 2000). Such indicators provide important constraints but are difficult to directly compare quantitatively with model output (although see Sellwood and Valdes, 2006 for one excellent approach).

The second approach we employ is to compare CMM temperatures predicted by the models, point by point with those inferred from paleoclimate proxies. This approach is quantitative and provides explicit error bars, but estimates may not be altogether accurate for all the reasons describe previously for MAT. Furthermore, CMM estimates are probably more likely to accurately delimit lower bounds on CMM because those involve strong biological/physical constraints (i.e. palms do not tolerate freezing) whereas the upper error bounds on CMM are probably less constrained. Such CMM estimates are not as broadly available as the qualitative records but are roughly comparable in their extent as the MAT records and often derive from the same analyses. In this study we compile CMM estimates, primarily derived from CLAMP, but supplemented by other sources, such as from isotopic analysis and from the co-existence approach. The error estimates are derived from the primary sources and do not account for temporal variation, unlike the MAT error bars. The CMM estimate sources are detailed in Table 1.

### 3.1.3 Uncertainties in topography, timing, and paleolocation

A major source of ambiguity in model-data comparison is the difference between the modelled elevation and the actual elevation of the proxy locality. Errors of ~6 °C can be introduced by a 1 km difference in elevation between model and data (Sewall et al., 2000). Every effort has been made to minimize this error by making the most accurate reconstruction of paleoelevation utilizing the available information at the time (Sewall et al., 2000). But, such estimates are controversial and have uncertainty associated with them of at least ±500 m in mountainous regions where much of the proxy data are found (Peppe et al., 2010; Hren et al., 2010; Rowley, 2007; Forest, 2007; Meyer, 2007; Molnar, 2010). Paleoelevation scholarship has progressed in the decade since our topography dataset was created and many important details have changed. Nevertheless, little consensus exists on those details (Davis et al., 2009; Mulch et al., 2007;

256





weak sensitivity to achieve climate conditions close to those of the early Eocene. For context and to give an indication of the robustness of these results we include results from another case, EOCENE-2240, carried out with 2240 ppm CO<sub>2</sub> and which we have previously shown is a better fit to mid-to-late Eocene climate (Eldrett et al., 2009).

5 In Sect. 4 we present a comparison of zonal mean model results for modern and Eocene conditions, maps of modelled Eocene surface temperatures, and point-wise comparison of proxy data and model results. For this comparison, all proxy records were rotated back to their paleo-locations at approximately 55 mya utilizing the GPLATES software ([www.gplates.org](http://www.gplates.org)) and plate model of Mueller et al. (2008) with  
10 some slight adjustments made to accomodate georeferencing differences between the model plate locations and the GPLATES reconstruction.

First, the modern location of the paleo-localities was introduced into GPLATES and then the localities were rotated back to their positions at 55 mya (Fig. 1a). The plate locations determined by GPLATES (<http://www.gplates.org>) were then compared with the  
15 paleogeography utilized in the modelling, derived originally from Sewall et al. (2000) in order to determine that they were generally comparable (Fig. 1b,c). In almost all cases the paleogeography in the model was consistent with the GPLATES 55 mya reconstruction and where adjustments were necessary they were objectively determined by comparing continental boundaries and making uniform, small adjustments to GPLATES  
20 reconstructed latitude and longitude to align correctly with geographic features in the model paleogeography. When this was done, model predictions were compared with proxies at these adjusted locations and inspection of the results was used to evaluate that the potential errors introduced by differences in true and modelled paleo-location are small (Fig. 1d).

25 Furthermore a comparison of the model paleogeography we used and the early Eocene paleotopography independently derived by Paul Markwick (Markwick, 2007) revealed that our reconstructions were in general agreement, indicating robustness of general features, but differing in important details, for example in the intermontane western North America (Fig. 1e,f). In the subsequent model-data comparison

261

we accounted for uncertainty in the true paleolocation, its temporal variation over the early Eocene, and the discretization introduced by the model grid by including error bars of  $\pm 2.5^\circ$  latitude on both proxy records and the model results. A spreadsheet including the references and values used in the paleotemperature reconstruction are  
5 included as supplemental material. The GPLATES Markup Language (GPML) file used to rotate modern localities is also available as supplemental material.

## 4 Results

### 4.1 Modeled zonal-mean surface temperatures

As an overview of the climate changes occurring in the Eocene simulation, Fig. 2 compares modeled zonal-mean surface temperatures from the EOCENE-4480 run with a  
10 CAM3 simulation using modern boundary conditions as specified by the Atmospheric Model Intercomparison Project (AMIP). The zonal mean includes averaging over both land and ocean. At high latitudes, the Eocene case is 30–50°C warmer than modern in both MAT and seasonal means. Differences are more muted in the tropics, ranging from  
15 6–10°C in all seasons. These results qualitatively capture the salient features noted in the proxy temperature record (Barron, 1987): annual mean temperatures much warmer than modern, especially at high latitudes, winter season temperatures generally above freezing, and a much reduced equator-to-pole temperature gradient. The hemispheric asymmetry in temperature increase is consistent with the removal of the Antarctic Ice  
20 Sheet and the associated 15-20°C change associated with the decrease in elevation.

### 4.2 Maps of modeled surface temperatures

To evaluate the robustness of these climate change patterns to the chosen level of greenhouse gas forcing, we compare the EOCENE-2240 and EOCENE-4480 simulations in Fig. 3. MAT in the EOCENE-4480 case (Fig. 3) is above freezing everywhere.

262

With the exception of inland Antarctica and a small region in high latitude, mountainous western North America, it is everywhere warmer than 10 °C as required by proxy records (see discussion in Sect. 3.1.2). Maximum SSTs are about 35 °C, in general agreement with proxies (Pearson et al., 2007; Huber, 2008; Jaramillo et al., 2010; Schouten et al., 2007). The hottest temperatures are on land in the subtropics, where annual mean values are 45 °C. No proxy records currently exist to support or refute this result.

Winter temperatures in the EOCENE-4480 case remain above freezing everywhere except for inland Antarctica. Temperatures in the intermontane western North America dip near zero in agreement with the existence of microthermic flora – though with an absence of frost intolerant macroflora – in various upland localities in the region (Smith et al., 2009). Maximum terrestrial temperatures in summer of up to 50 °C are reached in northern and southwestern Africa, southern Central South America, southwest North America, central Europe and western central Asia. The interiors of North America and Amazonia are also very hot (>40 °C) in summer. It should be noted that these are seasonal (3-month) means and that temperature extrema on subseasonal time scales may be substantially higher; we present seasonal means here since they are likely to best reflect the processes that shape the long-term distribution of proxies, flora and fauna. A comparison with the coldest climatological monthly mean temperature is carried out below for strict comparison with CMM proxy records.

Comparison of the two Eocene cases allows the identification of certain regions that are especially sensitive to a globally uniform radiative forcing. The land masses respond roughly homogeneously in the annual mean, though with some enhanced warming in eastern and central North America, central Asia and central South America. Seasonal differences are more pronounced. In boreal winter, the mid-to-high latitude interiors of North America and Asia are loci of focused warming in response to the increase in  $p\text{CO}_2$ . Northern and central North American and Asian MATs are between –10° and 0 °C in winter in the EOCENE-2240 case, but above freezing everywhere in the EOCENE-4480 case. Localized warming is also evident in austral

winter in the interior of Antarctica and in the boreal summer of west central Asia, and in midlatitude North America and Europe. These results show that localized temperature changes of up to 10 °C can occur seasonally even with globally homogeneous forcing and annual-mean SST change of less than 3 °C everywhere. Though the EOCENE-2240 case is very warm and has no sea ice, the additional greenhouse gas forcing in the EOCENE-4480 case yields a polar-amplified warming, with most of the amplification occurring in the winter season. The overall pattern is quite similar to the effect of doubling  $p\text{CO}_2$  in modern CAM3/CCSM3 simulations, despite the large differences in continental configuration, background greenhouse gas values, and other boundary conditions. Polar amplification in the absence of sea-ice has been attributed in various studies to increased latent heat transport (Langen and Alexeev, 2007; Caballero and Langen, 2005) and to high-latitude cloud feedbacks (Abbot et al., 2009a,b); the latter have been shown to be especially important in Eocene CAM3 simulations. Much of the terrestrial winter response seen here (between the two Eocene cases) correlates well with the near complete loss of snow cover and associated albedo decreases in the EOCENE-4480 case.

Most crucially for this study, the fact that winter season temperatures are above freezing in all regions for which quantitative and qualitative proxy data indicate frost intolerance (Greenwood and Wing, 1995; Markwick, 1998; Collinson and Hooker, 2003; Markwick, 2007; Kvacek, 2010) suggests that the EOCENE-4480 simulation does not suffer from the equable climate problem. This comes at the expense of a very large radiative forcing, causing temperatures >40 °C on land over significant regions. However, tropical SSTs in the EOCENE-4480 simulation are in good agreement with what is currently the best tropical proxy temperature records we have, from Tanzania (Pearson et al., 2001b, 2007) and off the coast of Colombia (Jaramillo et al., 2010), though the absolute temperatures inferred from this record are subject to significant uncertainty (Huber, 2008).



case and the modern CAM3 AMIP simulation discussed in Sect. 4.1. This comparison (Fig. 7) reveals a generally very good correspondence between the modelled and reconstructed warming at all latitudes. The zonal-mean modelled temperature anomaly with respect to modern conditions gives a good overall agreement with the observational anomalies, though regional and local details introduce significant scatter. Overall, it appears that the model is capable of quantitatively reproducing the polar amplification of terrestrial warming in passing from modern to early Eocene conditions.

## 5 Discussion

New estimates emerging from improved data coverage, the introduction of new proxies, and reinterpretation of older proxy records, show that temperatures in the early Eocene were much warmer than previously reconstructed (Covey et al., 1996). Thus, revisiting the equable climate problem requires rethinking how to frame the problem. Most climate model simulations have not investigated temperatures – including tropical temperatures – nor greenhouse gas forcing, in the ranges now considered likely. Our perspective in this paper is that most previous attempts to solve the equable climate problem have suffered from the question being ill-posed. Climate was even warmer than previously reconstructed and the forcing or climate sensitivity were also probably larger. We have opted here to incorporate very high values of greenhouse gas radiative forcing. Note that this approach is equivalent to “tuning” climate sensitivity to a higher value, but is much simpler in practice. The 4480 ppm CO<sub>2</sub> concentration used here should not be construed literally: it is merely a means to increase global mean warmth. Utilizing these new reconstructions for comparison makes the model-data comparison more challenging than prior attempts given the historical tendency of the models to underestimate extratropical temperatures. Yet our results clearly show that the model makes credible predictions for both the winter season warmth, mean terrestrial temperature change, and the polar amplification of warming in the early Eocene, to our knowledge, the first time.

267

Discrepancies remain, however. These do not appear to be of a magnitude that a compelling case for completely “missing physics” can be made, but they do point to many areas that need improvement. There are several factors that should be considered in further refining the model data comparison. These include some concrete refinements: improvements to paleotopography and their representation in the model, exploration of sensitivity to orbital forcing, and improvements to the vegetation boundary conditions.

### 5.1 Further improvements

#### 5.1.1 Resolution

As discussed by Sewall et al. (2000) the low resolution paleotopography used in this study is probably not only a source of random temperature error (due to random errors in topography) but also probably a source of cold bias. This is because paleoclimate proxies are likely to be warmer than model estimates since they reflect conditions at the bottom of valleys in high relief areas that have a high mean elevation. This systematic bias was not corrected for in this study.

One approach to solving this problem is to use a limited domain atmospheric model, at very high resolution (i.e. ~50 km) which might enable a full representation of the basin and peak scale orographic details that plague lower resolution, global climate model. This has been attempted over Eocene North America by Sewall and Sloan (2006), and Thrasher and Sloan (2009, 2010) with varying success. These studies clearly demonstrate that modelled temperatures are sensitive to resolving fine scale topographic detail, but little reduction of the model-data mismatch common to lower-resolution global model simulations was noted. This is probably due to two factors. First, knowledge of the true paleotopographic variation on those length scales is still rudimentary, hence random scatter is introduced. Second, and more importantly, limited area models use dynamics to downscale global model output to finer scales, but they are unlikely to generate temperature results substantially different than the global

268

model used to drive them. In other words, limited domain models such as this, typically produce the same aggregate temperatures as the global models, but with better representations of fine scale variations, and consequently if the global model is biased to too cold values, the limited area model will produce the same result, but at higher resolution (Sloan, 2006; Thrasher and Sloan, 2009, 2010).

Limited domain modeling is likely to be a very effective approach once global models produce results roughly in agreement with proxy data and once high resolution topographic reconstructions are available and are trustworthy. Our results here suggest that this high resolution approach in conjunction with high resolution, accurate paleotopography (and vegetation) may lead to a substantial reduction in the midlatitude scatter between proxy data reconstructions and model output. Building model boundary conditions directly from a properly geo-referenced DEM, such as implemented by Markwick and Valdes (2004) is a clearly better approach than the one originally used in Sewall et al. (2000) and subsequently re-used in the coupled modeling studies of Huber and Sloan (2001), Shellito et al. (2009); Winguth et al. (2010) and in this study.

### 5.1.2 Orbit

All of these simulations were carried out assuming a modern orbital configuration. Clearly, orbital variations caused fluctuations in temperature patterns in the early Eocene and inclusion of this feature in simulations would alter the seasonal and spatial patterns in detail. Cold summer orbital and hot summer orbit configurations would clearly impact the seasonality in these simulations (Morrill et al., 2001; Sewall et al., 2004; Lawrence et al., 2003). Nevertheless, it is revealing that the relatively crude approach used here of increasing the global radiative forcing due to  $p\text{CO}_2$  had a more pronounced impact on reconciling models with data, including seasonally sensitive data, than any orbital approach as so far.

### 5.1.3 Vegetation

The simulations are likely to be sensitive in their regional details – especially in terms of the localized seasonal temperature extrema – to the vegetation distributions and their representation in land surface models (Sewall et al., 2000; Shellito and Sloan, 2006a,b). Cold season, extratropical temperatures are likely to be especially sensitive to vegetation albedo and leaf area index (LAI), whereas peak tropical temperatures are likely to be sensitive to evapotranspirational fluxes, which in turn are controlled by a variety of parameters including LAI, stomatal resistance, leaf phenology, vegetation type fractional coverage, and soil hydrology (Bonan, 2008).

Likely regimes of vegetation albedo are relatively easy to estimate and to test sensitivity to (Upchurch et al., 1999; Otto-Bliesner and Upchurch, 1997; DeConto et al., 1999; Sewall et al., 2000), but the issue of how much latent flux tropical vegetation is capable of in very hot scenarios is a largely unconstrained problem, but very sensitive to the treatment of angiosperm evapotranspiration (Boyce and Lee, 2010). In our view, the least constrained aspect of the modeled temperature distributions is peak subtropical-to-tropical terrestrial warmth, as evidenced by the troubling prediction of warmth at the upper limit of proxy error bars in the subtropics by the model. The large, 15 °C differences between localized terrestrial summer maxima surface temperature and the warmest open ocean SSTs is good evidence that a better understanding of the controls on evapotranspiration is necessary to accurately predict tropical-to-subtropical terrestrial temperatures.

### 5.1.4 Sampling of warm regions

Additional challenges to estimating temperatures in the warmest regions derive from the filter of existing reconstruction methodologies and geographic sampling. At present there are no quantitative terrestrial temperature records for the early Eocene from 27° N to 27° S latitude. We have some hints from the late Paleocene (Head et al., 2009; Huber, 2009) and earliest Eocene (Jaramillo et al., 2010) that tropical conditions were

significantly hotter than modern, but in general the early Eocene terrestrial tropics are almost terra incognita (Burnham and Johnson, 2004; Jacobs, 2004; Jacobs and Herendeen, 2004; Kaiser et al., 2006). For our analysis we included one flora from Africa of middle Eocene age, to have at least some tropical information, but given that the early  
5 Eocene, as indicated by marine records, was likely somewhat warmer than the middle Eocene, the one tropical estimate shown here is likely underestimates the true temperature. If we had also included the tropical SST records of Pearson et al. (2007) or Jaramillo et al. (2010) it would not have altered our conclusions, given the fact that those studies reconstructed values are substantially warmer than modern and they  
10 have large error bars.

Furthermore, floristic approaches to temperature reconstruction are obviously biased to regions with vegetation. We have no idea how hot the arid and semi-arid regions were. Today arid-to-semi-arid regions account for 30% of terrestrial surface area and this was approximately true in the Eocene as well (Ziegler et al., 2003). So in terms of  
15 even approximately reconstructing the mean terrestrial temperature in the early Eocene about one third of the surface area is thoroughly unsampled and it is likely to be the hottest third. It may be that we must await proxy records from these regions utilizing different techniques, such as “clumped isotopes” (Eiler, 2007), to place meaningful constraints on temperatures in these areas.

## 20 6 Summary and conclusions

In prior work, robust model-data differences have suggested to many that climate models fail to reproduce the leading order feedbacks in a warmer world. Acknowledging the large error bars in the proxies and the various layers of uncertainty in the model data-comparison we have performed, it appears that the model reproduces the recon-  
25 structed MAT, CMM, and the proper degree of high latitude amplification. The congruence of models and data in the simulations here and the conditions under which the

congruence occurs suggest that the leading order physics are well represented, but perhaps either radiative forcing or climate sensitivity was near the upper range of plausible values, or some combination of the two. We have shown that increased radiative forcing in the form of very high  $p\text{CO}_2$  seems to resolve the equable climate problem  
5 without running too far afield of other constraints, but this does not necessarily mean that  $p\text{CO}_2$  was the only major forcing factor. We have not addressed whether the enhanced radiative forcing was due to  $p\text{CO}_2$ , methane, other greenhouse gases, novel cloud feedbacks, or other “missing” factors. We have also not established whether large forcing is actually necessary, the alternative being high values of climate sensitivity as  
10 in the study of Heinemann et al. (2009) and only moderate increases in forcing.

Regardless of which explanation is correct, the model’s ability to capture the reduced seasonality signal and spatial gradients derived terrestrial paleoclimate proxies provides a validation of CAM3/CCSM3 for climates vastly different from, and warmer than modern. With the overall modern patterns in good agreement with the proxy data  
15 significant future progress should be possible in understanding the large-scale dynamics of past greenhouse climates within the framework of a validated model. Significant work remains to resolve remaining model-data differences. The most troubling result is extremely hot temperatures in the tropics, but sufficient data do not exist to rule this result out at this time. Provided compelling evidence becomes available ruling out terrestrial temperatures much above 35 °C in the subtropics (at this time it does not) then it is likely that improvements in the model’s representation of tropical evapotranspiration and vegetation type can alleviate the problem in future work. In mid-to-high latitudes it seems likely to us that the remaining model-data discrepancies are due to regional and local scale features, such as unresolved topographic and vegetation details, which can be resolved utilizing better boundary condition data sets and higher resolution models.  
20 Our results do not exclude the existence of more exotic forcings and feedbacks (e.g. Kump and Pollard, 2008; Sloan and Pollard, 1998; Emanuel, 2002; Korty et al., 2008), but instead provide a baseline for comparison involving simple and well understood forcing mechanisms. From a reductionist point of view, such novel processes may not



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287

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288

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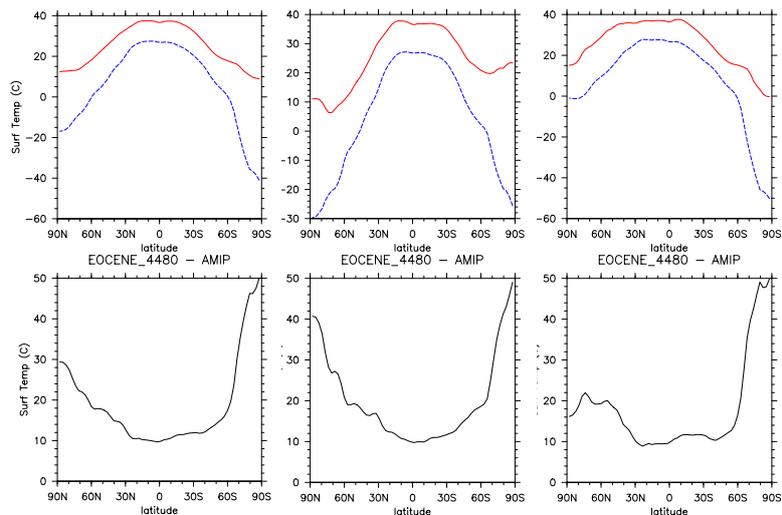
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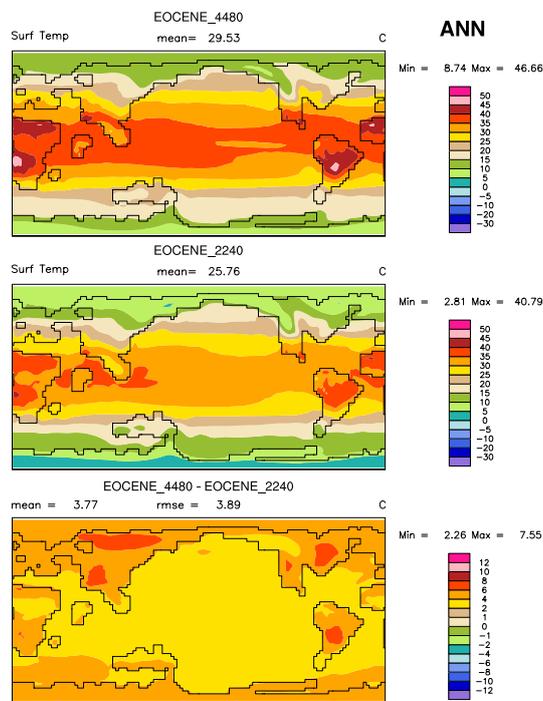
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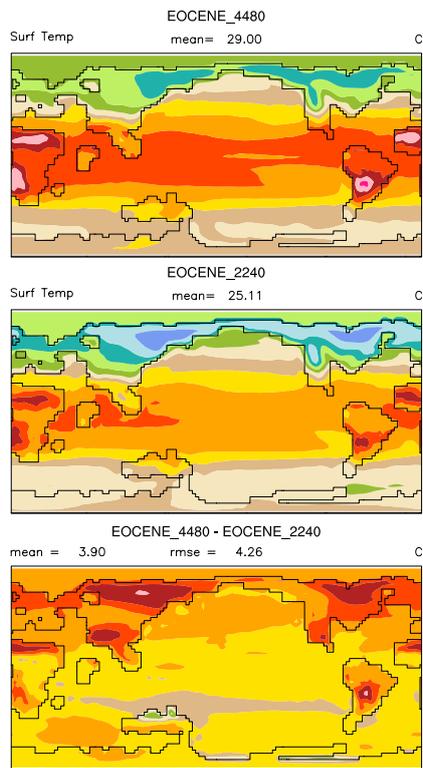




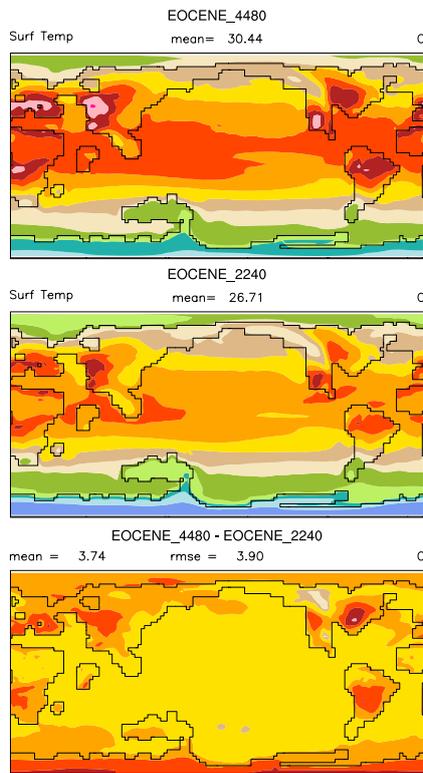
**Fig. 2.** Zonal mean surface temperature from the EOCENE-4480 (red) and modern (AMIP) simulations as described in the text. In the upper row, the annual (left) boreal winter (middle) and boreal summer (right) means are shown. In the lower row, the anomaly, EOCENE-4480 minus modern is shown.



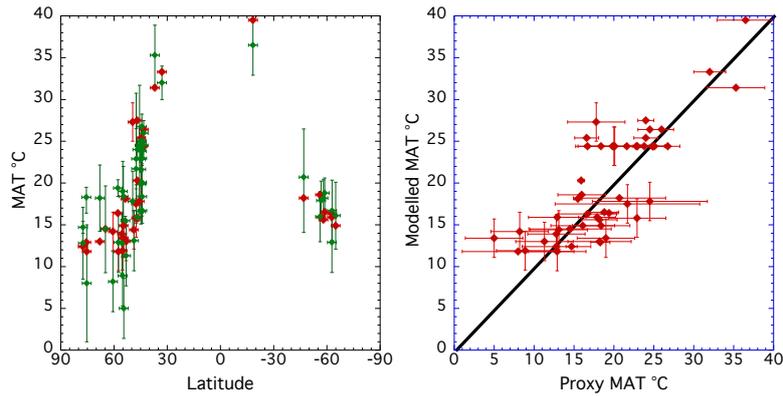
**Fig. 3a.** Maps of time averaged surface temperature from the EOCENE-4480 and EOCENE-2240 simulations. The annual (a), boreal winter (b), and boreal summer (c) means are shown. The upper row is the EOCENE-4480 case, the middle row is the EOCENE-2240 case, and the bottom row is the anomaly. The range is values is indicated in the color bar in (a) and is the same in (a-c). Units are in °C.



**Fig. 3b.** Figure 3 continued, see Fig. 3a for description.

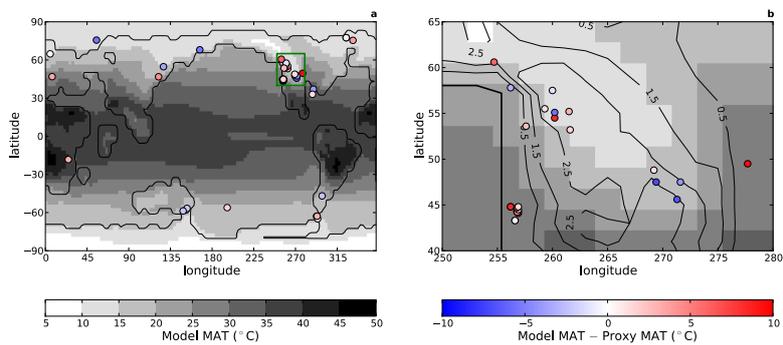


**Fig. 3c.** Figure 3 continued, see Fig. 3a for description.



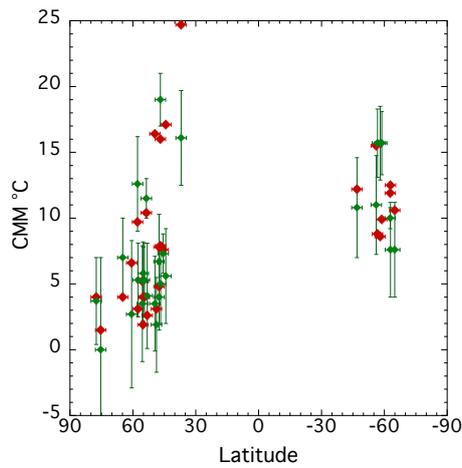
**Fig. 4.** On the left: a pointwise comparison of annual mean terrestrial surface temperature (MAT) from EOCENE-4480 (red) and proxy data estimates (green) versus latitude. The vertical error bars on the proxies represent the methodological error plus the temporal variation (if a time series) as further described in the text. The model vertical error bars are only included in regions considered to have high elevation and represent topographic uncertainty as described in the text. Horizontal error bars indicate the uncertainty in the position of the paleolocality and the propagation of that uncertainty onto the discrete model grid (as described in the text). On the right, the temperatures reconstructed from proxies are plotted versus modelled and the 1-to-1 line is plotted in black. The error bars are the same as previously described.

301



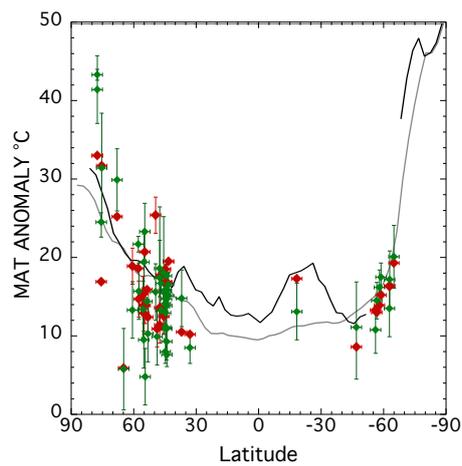
**Fig. 5.** Difference between modelled and reconstructed MAT as indicated by the color bar (on the right) overlain on the the model predicted temperatures (grey scale color bar on left).

302



**Fig. 6.** A pointwise comparison of cold month mean temperature (CMM) from the EOCENE-4480 (red) and proxy data estimates (green) versus latitude. The error bars are as described in the text

303



**Fig. 7.** A pointwise comparison of anomaly of annual mean terrestrial surface temperature (MAT) from the EOCENE-4480 (red) and proxy data estimates (green) minus the modern temperature value at those locations versus latitude. The error bars are the same as described in Fig. 3. The zonal mean anomaly of the EOCENE-4480 case minus the modern AMIP case plotted originally in Fig. 2 is shown in grey on this figure for reference. The land-only zonal mean anomaly is also shown in black.

304