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What have we learnt from palaeoclimate simulations?

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Abstract

There has been a gradual evolution in the way that palaeoclimate modeling and palaeoenvironmental data are used together to understand how the Earth System works, from an initial and largely descriptive phase through explicit hypothesis testing to diagnosis of underlying mechanisms. Analyses of past climate states are now regarded as integral to the evaluation of climate models, and have become part of the toolkit used to assess the likely realism of future projections. Palaeoclimate assessment has demonstrated that changes in large-scale features of climate that are governed by the energy and water balance show consistent responses to changes in forcing in different climate states, and these consistent responses are reproduced by climate models. However, state-of-the-art models are still largely unable to reproduce observed changes in climate at a regional scale reliably. While palaeoclimate analyses of state-of-the-art climate models suggest an urgent need for model improvement, much work is also needed on extending and improving palaeoclimate reconstructions and quantifying and reducing both numerical and interpretative uncertainties.

Keywords: palaeoclimate modelling, palaeoenvironmental data synthesis, climate reconstruction, forward modelling, CMIP5

Introduction

Climate has varied continuously through Earth's history. There are several styles of climate variability, associated with different drivers and operating on characteristic time scales. For example, there are periodic climate changes, resulting from astronomical or 'orbital' forcing on seasonal and multi-millennial timescales (Berger, 1978). Examples of progressive changes include the long-term cooling through the Cenozoic in response to changes in land-sea configuration and atmospheric composition (Zachos et al., 2001; Fletcher et al., 2007), or the cooling trend of the last two millennia caused by orbitally-driven changes in incoming solar radiation (insolation). Finally, there are rapid climate shifts such as those that were caused by the re-organisation of the coupled atmosphere-ocean circulation during the Dansgaard-Oeschger cycles (Bond et al., 1993; Kageyama et al., 2010). The combination of these styles of variability gives rise to a large and diverse set of examples of the response of regional and global climates to changes in climate forcing.

The impacts of past climate change are recorded by a variety of geological, isotopic and biological records (Bradley, 2014). These records can be interpreted, either using qualitative inference or explicit statistical approaches, to provide reconstructions of past climate variables. Such reconstructions document how the climate system behaves in response to different kinds of forcing – this illustration of what responses are physically possible is the basis for the idea that "the past is the key to the future" (Masson-Delmotte et al., 2013). However, there has been an increasing emphasis in recent years on the importance of palaeoclimatic and palaeoenvironmental reconstructions for climate-model evaluation (e.g. Izumi et al., 2013; Li et al., 2013; Perez Sanz et al., 2014). This arises from recognition that meteorological records from recent decades sample a range of climate variability that is too limited to provide a robust test of how well a numerical climate model can simulate a large climate change. Past climates provide a unique opportunity for "out-of-sample" evaluation of model performance, and thus a measure of the reliability of model predictions of the future (Braconnot et al., 2012; Harrison et al., 2014; Schmidt et al., 2014; Harrison et al., 2015).

Palaeoclimate simulations and data-model comparisons have been made for many iconic events in the past, including the early Holocene (ca 9 ka: Marzin and Braconnot, 2009; Marzin et al., 2013), Younger Dryas (ca 12.9–11.7 ka: Renssen et al., 2015), Last Interglacial (ca 125 ka: Bakker et al., 2013), the mid-Pliocene warm period (ca 4.2 Ma: Haywood et al., 2010a, 2010b), and the Eocene (ca 55-50 Ma: Lunt et al., 2012). In this paper, however, we only focus on the three periods that were included in the current phase of the Coupled Modelling Intercomparison project (CMIP5), the Last Millennium (850-1850 CE), the mid-Holocene (6 ka) and the Last Glacial Maximum (21 ka). These are the experiments that were used as part of climate-model evaluation reported in the Working Group 1 report to the Intergovernmental Panel on Climate Change (Flato et al., 2013). We summarise what has been learnt from the evaluation of these three simulations and the future challenges that face palaeoclimatology.

A Brief History of Palaeoclimate Simulations

The initial focus for palaeoclimate modeling was the Last Glacial Maximum (LGM, ca 21 ka), a time when there was a large change in forcing due to the presence of large ice sheets over North America (Laurentide ice sheet) and northern Europe (Eurasian ice sheet) and a radical change in atmospheric composition compared to the pre-industrial period. The earliest experiments (Alyea, 1972; Williams et al., 1974; Gates, 1976; Manabe and Hahn, 1977; Kutzbach and Guetter, 1986) were made with atmospheric general circulation models (AGCMs), and required changes in sea-surface temperature (SST) to be specified from observations (CLIMAP, 1976, 1981). In some cases (e.g. Alyea, 1972; Williams et al., 1974), simulations were confined to a single season because of the limitations in computing power. Nevertheless, these equilibrium simulations established that the presence of large ice sheets had a major impact on northern hemisphere climates, both through the direct effect of replacing vegetated land surfaces with highly-reflecting ice on albedo, and through the displacement of atmospheric circulation patterns caused by the increase in regional elevation by the mountain-like ice masses.

The Cooperative Holocene Mapping Project: COHMAP Members, 1988; Wright et al., 1993) subsequently broadened the focus to encompass simulations of the whole of the period from the LGM to present, in order to examine the impact of changing orbital configuration on radiative forcing and climate. However, these simulations were still equilibrium simulations made with an atmosphere-only model, thus requiring SSTs to be prescribed along with changes in the ice sheet height and extent, land-sea geography, atmospheric composition, and insolation. The COHMAP experiments were particularly important because they demonstrated the role of orbital changes in the evolution of the northern hemisphere monsoon systems (Kutzbach and Street-Perrott, 1985). A key aspect of the COHMAP project was the creation of large-scale syntheses of palaeoenvironmental and palaeoclimate data in order to document regional climate changes over the last glacial-interglacial, thus creating the basis for systematic comparisons of simulated and observed regional climates (Wright et al., 1993).

The availability of large-scale data syntheses, as well as the identification of mechanisms underpinning large-scale regional climate changes, was a motivation for the choice of the mid-Holocene (MH, 6ka) and the LGM as the experimental foci for the Palaeoclimate Modelling Intercomparison Project (PMIP). The goal of PMIP is to compare the behaviour of different climate models when run using the same forcing. The first phase of PMIP (PMIP1: Joussaume and Taylor, 2000) focused on comparison of AGCMs. By the second phase of the project (PMIP2: Crucifix et al., 2005), climate models routinely included an explicit simulation of ocean circulation (coupled ocean-atmosphere models: OAGCMs) and some models also included dynamic vegetation (coupled ocean-atmosphere-vegetation models: OAVGCMs). The evaluations of MH and LGM simulations carried out by PMIP have established unequivocally that climate models can reproduce observed, first-order global or hemispheric changes in climate in response to changes in forcing (Joussaume et al., 1999; Braconnot et al., 2007a, b; Zheng et al., 2008; Otto-Bliesner et al., 2009). However, they have also shown that models differ, often quite substantially, in their predictions, and comparison with palaeoclimate reconstructions shows that models often fail to capture regional changes accurately (e.g. Joussaume et al., 1999; Coe and Harrison, 2002; Brewer et al., 2007; Perez Sanz et al., 2014). Understanding the reasons for inter-model differences, and for model-data discrepancies has become the major focus of the third phase of the PMIP project (PMIP3: Braconnot et al., 2011; Braconnot et al., 2012) – and the reason that palaeoclimate experiments were included for the first time in CMIP5 (Taylor et al., 2011), the core international project that assembled model runs for the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC).

Palaeo-simulations in CMIP5

Three palaeoclimate-simulations are included in the CMIP5 set of simulations: LGM, MH, and the Last Millennium (LM: 850-1850 CE). The LGM and MH simulations are equilibrium simulations. Both the LGM and the MH represent substantially different climate states from the present day and from each other, and have large natural forcings that are relatively well known (Braconnot et al., 2012; Harrison et al., 2015). The LM is a transient simulation to examine natural climate variability under conditions more similar to those of the present day (Schmidt et al., 2011).

At the LGM, the orbital parameters were nearly the same as they are today (Table 1) so that the differences in insolation were small. The major differences in forcing were caused by the presence of large ice sheets in the northern hemisphere (and concomitant changes in sea level and palaeogeography) and the lower atmospheric concentration of greenhouse gases. The changes in greenhouse gas concentrations (CO_2: 185 ppm, CH_4: 350 ppb, N_2O: 200 ppb) are well known from ice core records (EPICA Community Members, 2004). The decrease in the greenhouse gases relative to pre-industrial alone results in a radiative forcing of the troposphere of −2.8 W m^{-2} (Braconnot et al., 2007a). The expansion of the ice sheets at the LGM resulted in a sea-level lowering of ca 130m, and associated changes in albedo had an important effect on climate, particularly in the northern hemisphere. The marginal limits of the North American (Laurentide), Greenland and European (Eurasian) ice sheets are well known (e.g. Dyke and Prest, 1987; Mickelson and Colgan, 2003; Dyke, 2004; Gyllencreutz et al., 2007; Simpson et al., 2009; Ehlers et al., 2011; Mangerud et al., 2013) but there is no direct evidence for the distribution of ice mass. The form and height of the ice sheets are therefore inferred through a combination of physical modelling and indirect observational constraints (e.g. information on relative sea-level changes). A composite ice sheet was created for the CMIP5 experiments (Abe-Ouchi et al., 2015) by combining information from three reconstructions of the distribution of ice mass (ICE-6G v2.0: Argus and Peltier, 2010; GLAC-1a: Tarasov et al., 2012; ANU: Lambeck et al., 2010). In the CMIP5 LGM simulations, calculations using a simplified shortwave radiative model of the atmosphere, perturbed by changes in individual boundary conditions separately, show that the change in the ice sheets results in an implied forcing of between −1.85 and −3.49 Wm^{-2} depending on the climate model (Abe-Ouchi et al., 2015) while the overall change in forcing varied between –3.62 and –5.20 W m^{-2}. Thus, the change in forcing due to changes in atmospheric composition and expansion of the ice sheets at the LGM is of a similar magnitude to that projected for the next century.

The CMIP5 LGM experiments do not include the additional climate forcing that results from changes in vegetation distribution (Prentice et al., 2000; Harrison and Bartlein, 2012) because the observations of LGM vegetation are too sparse (in many regions) to provide a global gridded data set to use as a model input. LGM vegetation is therefore either computed (in models which include dynamic vegetation) or prescribed to be the same as the pre-industrial control simulation in the CMIP5 simulations (Table 1). The CMIP5 simulations also ignore the potential impact of known changes in atmospheric dust loading (Kohfeld and Harrison, 2001).

The MH provides an opportunity to evaluate simulations at a time of changed seasonality, when the influence of changes in ice sheet extent and land-sea geography on global climate was negligible. The seasonal and latitudinal distribution of MH insolation was different from present because of the difference in orbital configuration (Table 1). Seasonal contrast in the northern hemisphere was enhanced (by about 60 Wm^{-2}), through an increase in summer insolation and a decrease in winter insolation, and correspondingly reduced by decreased summer and increased winter insolation in the southern hemisphere. Greenhouse gas concentrations were similar to levels in the pre-industrial era (CO_2: 280 ppm, CH_4: 650 ppb, N_2O: 270 ppb. Although there were changes in vegetation distribution (Prentice et al., 2000; Harrison and Bartlein, 2012), these were not taken into account in the CMIP5 experiments (Table 1). As is the case for the LGM, the main focus of analyses of the MH experiments is on the impact of a large change in forcing on the mean climate response.

The LM is a transient simulation, included in CMIP5 to examine natural climate variability in a climate state close to that of the present day (Schmidt et al., 2011) and as a reference for detecting and attributing observed twentieth-century changes in climate patterns and trends resulting from human activities (Hegerl et al., 2011). The LM also provides opportunities to investigate the link between volcanism and climate, including ENSO variability (Emile-Geay et al., 2008; Wilson et al., 2010), to test the stability of atmospheric modes (e.g. Yiou et al., 2012), to analyse the interaction between short-term variability and land-surface feedbacks (e.g. Acosta Navarro et al., 2014), and to explore changes in recurrence or intensity of extreme events (e.g. Fallah and Cubasch, 2015). The LM simulation is characterized by changes in orbital, solar, volcanic and land-use forcing. With the exception of the orbital changes, there are large uncertainties associated with each of these forcings (Schmidt et al., 2011). The CMIP5 protocol therefore defines a number of alternative forcing histories to take account of these large uncertainties (Table 1). There are, for example, two reconstructions of the volcanic forcing (Crowley et al., 2008; Gao et al., 2008), five reconstructions of solar forcing (Wang et al., 2005; Muscheler et al., 2007; Steinhilber et al., 2009; Delaygue and Bard, 2011; Vieira et al., 2011) and two land-use scenarios (Hurtt et al., 2006; Pongratz et al., 2008). Modelling groups have been allowed to choose which forcing “scenarios” they use. While this makes comparison between models more difficult, some modelling groups have run ensembles of simulations using different forcing scenarios (e.g. Goosse et al., 2005; Bothe et al., 2013; Otto-Bliesner et al., 2015) thus allowing the effects of uncertainty in forcing to be assessed.

Many modelling groups have run palaeoclimate simulations as part of CMIP5 (Table 2). The MH experiment is relatively simple and has a smaller perturbation than the LGM, thus requiring less time to reach equilibrium. Thus many more groups have performed the MH experiment than have performed the LGM experiment. Only a few groups have performed the LM simulation – in part because multiple forced and unforced runs are required for a complete diagnosis. Nevertheless, there are sufficient simulations for all three periods to allow comparisons of the reaction of different climate models to the same change in forcing and evaluation of the realism of the simulations through comparison with palaeodata.

A Brief History of Palaeodata Synthesis

With the exception of ice-core records of the well-mixed trace gases, individual palaeoenvironmental records document local or regional changes – although the spatial sampling scale may vary from metres up to some tens or hundreds of kilometres. The synthesis of records at a regional scale provides a way of documenting robust responses to past climate changes. Regional data syntheses are the appropriate tool for extracting information that is comparable to simulated climates, given the spatial resolution of current climate models. The highest resolution of the CMIP5 models used for palaeoclimate experiments, for example, is ca 1° x 1° latitude/longitude.

The comparison of individual records from a region, and identification of similarities in their response, is standard practice. Data synthesis, however, requires that the individual records are interpreted using a common approach. One of the earliest examples of this was the synthesis of lake records from northern Africa (Street and Grove, 1976) that led to the creation of the Global Lake Status Database (GLSDB: Street and Grove, 1979; Street-Perrott et al., 1989; Kohfeld and Harrison, 2000; Fig. 1), one of the databases used by the COHMAP project. The GLSDB had transparent rules for site selection, and used an explicit method to categorise individual records into status classes (high, intermediate, low) so that they were easily compared both within and between regions. This focus on lake status also facilitated direct comparison with model output, because lake status is sensitive to changes in the balance between precipitation and evaporation (Street-Perrott and Harrison, 1984; Cheddadi et al., 1987).

The COHMAP project used pollen data as a source of information about regional vegetation and climate, but it was not until the creation of the Palaeovegetation Mapping Project (BIOME 6000: Prentice and Webb, 1998) as part of the International Geosphere-Biosphere Programme that these data were treated in a systematic and consistent way. BIOME 6000 developed an approach to translate pollen assemblages into vegetation reconstructions, quaintly termed biomisation, which involved classification of individual pollen taxa into plant functional types (PFTs), the characterization of major vegetation types (biomes) according to their characteristic or defining PFTs, and the application of an algorithm to select the most likely biome represented at a site (Prentice et al., 1996). BIOME 6000 produced vegetation maps for the MH and LGM (Prentice et al., 2000; Bigelow et al., 2003; Pickett et al., 2004; Marchant et al., 2009), explicitly for comparison with vegetation simulations made either using OAVGCMs or by running a biogeography model driven by outputs from e.g. OAGCMs (e.g. Harrison et al., 1998; Wohlfahrt et al., 2004). The biomisation approach has also been used to produce maps for other time intervals for certain regions (Marchant et al., 2001; Williams et al., 2004).

There have been other efforts to create datasets comparable to model outputs. The Dust Indicators and Records from Terrestrial and MArine Palaeoenvironments (DIRTMAP: Kohfeld and Harrison, 2001; Maher et al., 2014) database contains estimates of aeolian accumulation rates at key time periods measured in ice cores, marine cores and at terrestrial locations. The modern and LGM dust deposition estimates from DIRTMAP have been used for evaluation of dust-cycle simulations (e.g. Werner et al., 2003; Bauer and Ganopolski, 2014). The Global Palaeofire Working Group (website) has created a global synthesis of charcoal records (Power et al., 2010), which provides a qualitative record of changes in biomass burning of the last glacial-interglacial cycle. Much of the focus on this group has been on documenting regional changes (Power et al., 2008; Marlon et al, 2008; Daniau et al., 2010; Marlon et al., 2013) or investigating the controls on fire (Daniau et al., 2012), but the data set has potential to be used for model evaluation (e.g. Brücher et al., 2014).

Palaeoenvironmental data have long been used to reconstruct climate variables quantitatively (e.g. Grichuk, 1969; Imbrie and Kipp, 1971; McIntyre et al., 1976; Hutson and Prell, 1980; Bartlein et al., 1984; Atkinson et al., 1987; Guiot, 1987; Huntley and Prentice, 1988; Guiot, 1990). The development of well-documented, quantitative global palaeodata sets portraying the spatial climatic patterns of the LGM and MH time periods is a central objective of the PMIP research programme.

The palaeoceanographic community has provided a global reconstruction of LGM SSTs (MARGO Project Members 2009), which supersedes the CLIMAP data set that was developed in the 1980s. MARGO (Multiproxy Approach for the Reconstruction of the Glacial Ocean surface) defined the LGM as the interval between 19 and 23 ka. The project compiled 696 SST reconstructions from this interval. The data set includes all available microfossil-based (transfer functions based on planktonic foraminifera, diatom, dinoflagellate cyst and radiolarian abundances) and geochemical (alkenones and planktonic foraminiferal Mg/Ca ratio) reconstructions. Each type of sensor has a different geographical coverage – the reconstructions from the Southern Ocean, for example, are largely based on diatom records, whereas most of the tropical records are derived from foraminiferal assemblages. Nevertheless, there are some regions of the world where reconstructions based on multiple sensors are available and could be compared to provide an estimate of robustness.

In the global reconstruction, the data were gridded at $5^\circ \times 5^\circ$ resolution, where each grid cell was assigned an SST estimate by averaging individual reconstructions that fall into the same cell, weighted by a mean reliability index. The resulting SST anomalies show robust spatial and seasonal changes (Fig. 2), and there is first-order agreement on the magnitude of latitudinal anomalies between geochemical and microfossil-based reconstructions with the strongest mean annual cooling in the mid-latitude North Atlantic – a feature confirmed by reconstructions from four different types of sensor.

For the MH, the only global SST product available is the Global database for alkenone-derived HOlocene Sea-surface Temperature (GHOST), which includes reconstructions based on Mg/Ca and alkenones (Kim, 2004; Leduc et al., 2010). Model comparisons using the GHOST data set have shown significant mismatches between the modelled and reconstructed SST anomalies (Schneider et al., 2010; Hargreaves et al., 2013; Lohmann et al., 2013). An attempt has been made to produce a more comprehensive data set, including reconstructions from Mg/Ca and alkenone palaeothermometry and statistical estimates obtained using planktonic foraminifera and organic-walled dinoflagellate cyst census counts (Hessler et al., 2014). However, analyses of these data show that the MH change in SST is small compared to the magnitude of known methodological uncertainties associated with SST reconstructions, and also compared to the differences between the observed modern ocean temperature datasets used as the baseline to determine the MH change in SST. Hessler et al. (2014) concluded that, unlike the LGM, where robust changes in SST patterns emerge despite the methodological uncertainties (MARGO Project Members, 2009), MH SSTs do not provide a reliable benchmark for model simulations.

Terrestrial environments are diverse and many types of geochemical, isotopic and biological data have been used to provide quantitative reconstructions for specific areas and ecosystems (see e.g. Atkinson et al. 1987; Stute et al. 1992; Heiri et al. 2003; Jones et al. 2004). The most widespread source of quantitative reconstructions is palaeovegetation (fossil pollen and plant macrofossil) records. Palaeovegetation records provide a unique combination of near-global coverage of information on several distinct aspects of climate (seasonal temperature, rainfall, soil moisture), combined with robust and well-documented methodologies to derive reconstruction uncertainties.

The terrestrial palaeoecology community has produced a unified gridded data set for the MH and the LGM based on combining all existing quantitative reconstructions, subject to availability of the primary data (i.e. the reconstructions) and a transparent screening procedure (Bartlein et al., 2011). Although the reconstructions were produced using different techniques, ranging from simple regression through analogue techniques to inverse modelling, analyses for regions with multiple reconstructions made using different methods show that the choice of method has little impact on the results. Thus, compositing reconstructions made with different methods provides robust and coherent reconstructions of the large regional climate changes at the MH and LGM (Fig. 2). Although this synthesis represents the state-of-the-art target for model evaluation and benchmarking, the coverage is poor for many important regions including Australia and South America. There are pollen records from both regions (Fig. 2) that could be used to make statistical reconstructions of climate variables; even in regions that are relatively well represented in the gridded data set, there is the potential for a much-expanded set of climate reconstructions.

Pre-industrial climate provides a baseline for the detection and attribution of recent anthropogenic impacts on the Earth system (Hegerl et al., 2011), and this provides the major motivation for the inclusion of LM simulations in CMIP5. Reconstruction of annual climate before the pre-instrumental period relies on the use of natural archives, including isotopic records from laminated sediments or corals, ice core records and tree rings. However, statistical reconstructions from tree rings provide by far the largest number of pre-instrumental records. The major focus of data synthesis to date has been on seasonal (e.g. Briffa et al., 2002; Luterbacher et al., 2004; Guiot et al., 2005; Xoplaki et al., 2005) or annual temperature. Reconstructions of regional or hemispheric temperature changes over the last millennium (e.g. Jones et al., 1998; Briffa et al., 2002; Esper et al., 2002; Moberg et al., 2005; Rutherford et al., 2005; Mann et al., 2007, 2008; Ljungqvist et al., 2012; PAGES 2k Consortium, 2013; Shi et al., 2013; Neukom et al., 2014) generally use several of these palaeodata sources, combined with historical and instrumental records when available. While there are regional reconstructions of precipitation (Pauling et al., 2006; Steinman et al., 2012; Wilson et al., 2012; Feng et al., 2013), there is currently no global synthesis of precipitation data. Although there is broad agreement on multidecadal to centennial time scales, there is considerable variability among individual hemispheric temperature reconstructions on short time scales over the last millennium (Fig. 3), depending on methodology and the selection of site-based reconstructions included in the reconstructions (Juckes et al., 2007; Fernandez-Donado et al., 2013). The range in reconstructed change in northern hemisphere average temperature between the Mediaeval Warm Anomaly and the Little Ice Age, for example, encompasses the simulated range of temperature change across different models using different combinations of forcings and including simulations made with and without volcanic forcing (Fernandez-Donado et al., 2013). Thus, the large uncertainties in the reconstructions coupled with similarly large uncertainties in the forcing currently limits the usefulness of the last millennium as a target for model evaluation *sensu stricto*.

Confronting models with observations

Palaeodata document what has actually happened in the past, but explanation of observed changes is dependent on a conceptual model of how the climate system works and is therefore rarely unequivocal. Models that incorporate current understanding of physical climate processes provide a way of making the conceptual model explicit. Thus, one of the most fruitful approaches to understanding the mechanisms of climate change is through confronting observations and model experiments in hypothesis-testing mode, where the ability of the model to reproduce observed patterns in space or time indicates the plausibility of the underlying conceptual explanation while disagreement indicates that alternative explanations are required.

The demonstration by Kutzbach and Street-Perrott (1985) that the evolution of the African monsoon over the last glacial-interglacial cycle was a direct response to orbital forcing (Kutzbach and Street-Perrott, 1985) provides the classic example of this hypothesis-testing approach. In this paper, the water balance (precipitation minus evaporation) over northern Africa (8.9 and 26.6 °N) was calculated based on a sequence of only January or only July climates (known as perpetual January or perpetual July simulations, where the mean annual climate is then calculated as the average of the two monthly simulations and the seasonal contrast as the difference between these two months) with an atmospheric general circulation model forced by changes in insolation, ice sheet extent, and sea-surface temperatures. The simulations predicted the observed temporal evolution of lake status. Analyses of the simulations (including additional sensitivity experiments) confirmed that the primary driver of the observed changes in lake status was changes in orbital forcing. Changes in boundary conditions changes associated with northern-hemisphere ice sheets or atmospheric composition had little impact on the regional water balance. The primary importance of orbitally-induced changes in insolation as a driver of the waxing and waning of the northern hemisphere monsoons has subsequently been confirmed with more advanced models and modeling protocols (Zhao et al., 2005; Braconnot et al., 2007; Marzin and Braconnot, 2009; Dallmayer et al., 2015). However, it is clear that there is considerably more complexity in the seasonal evolution of monsoon rainfall than originally thought (Fig. 4) and considerable millennial- and sub-millennial scale variability is superimposed on the orbitally-driven evolution (Otto-Bleisner et al., 2014). It is also clear that feedbacks associated with the ice sheets, ocean conditions and climate-induced changes in land-surface conditions are necessary to produce the observed temporal evolution of the northern hemisphere monsoons (e.g. Clausen and Gayler, 1977; Ganopolski et al., 1998; de Noblet-Ducoudré et al., 2000; Zhao et al., 2005; Patricola and Cook, 2007; Zhao et al., 2007; Marzin and Braconnot, 2009; Ohgaito and Abe-Ouchi, 2009; Dallmeyer et al., 2010; Zhao and Harrison, 2012; Dallmeyer et al., 2015).

This hypothesis-testing approach underpins data-model comparison of regional climate changes during the MH and LGM conducted as during the first phase of PMIP, which focus on demonstrating how far large-scale patterns are a consequence of changes in orbital and/or glacial boundary conditions. For example, comparisons with model simulations driven by the combined influence of known changes in orbital, ice sheet and greenhouse gas forcing have been used to explain observed differences in the temporal evolution of fire regimes between tropical and extratropical regions of the northern and southern hemisphere over the last glacial-interglacial cycle (Daniau et al., 2012; Fig. 5). However, the hypothesis-testing approach is much more powerful when it is used to test potential mechanisms explicitly through experiments that separate out the potential influence of individual forcings. For example, Harrison and Prentice (2003) used a simple biogeography-biogeochemistry model driven by climate-model simulations of the LGM to demonstrate the necessity of including the direct impacts of low CO_2 on productivity and water-use efficiency to explain observed changes in tropical vegetation distribution. They showed that the area of tropical forests would have increased in response to climate changes at the LGM, whereas the observed reduction of tropical forest and increase in grassland could only be achieved when CO_2 was lowered to glacial levels. A similar conclusion was reached by Bragg et al. (2013), comparing simulated and observed glacial-interglacial changes in leaf-wax $\delta^{13}C$ of terrestrial origin from a transect of marine cores recording vegetation shifts in southern Africa. Bragg et al. (2013) also discussed the general importance of atmospheric CO_2 concentration as a driver of vegetation changes, and the relative roles of climate and CO_2 changes in glacial-interglacial vegetation shifts – a topic that has suffered from some misconceptions, as the two kinds of effect are neither mutually exclusive, nor independent. The importance of the direct effects of changing CO_2 on vegetation productivity, and hence fuel load, has subsequently been demonstrated as an important control on LGM fire regimes (Martin Calvo et al., 2014).

Model evaluation and benchmarking

The importance of assessing how well state-of-the-art climate models can simulate large climate changes has led to the increasing use of palaeodata for the purposes of model evaluation. At its simplest, model evaluation can involve qualitative comparisons of spatial patterning. Such map-map comparisons can be powerful. For example, the inability of climate models to capture the spatial expansion of the northern African monsoon during the MH is readily apparent by comparing maps of observed and simulated water balance (see e.g. Perez Sanz et al., 2014). However, when the discrepancies are in the magnitude of a signal rather than spatial pattern (or sign) then quantitative comparisons are necessary.

There are many potential sources of uncertainty in using palaeodata to make climate reconstructions. Some of these uncertainties are strictly numerical but others are associated with dating, methodologies, baseline choice, or interpretation – and are much more difficult to deal with when making quantitative comparisons. Numerical uncertainties (e.g. root mean-squared errors on statistically-based climate reconstructions) are easily factored into data-model comparisons (see e.g. Hargreaves et al., 2013). The other sources of uncertainty, even when quantifiable, are often ignored.

An absolute chronology is fundamental to comparisons of palaeo-records and the construction of palaeodata syntheses. The development of reliable techniques to construct age models has been a major focus for the community (e.g. Bennett, 1994; Boreux et al., 1997; Bennett and Fuller, 2002; Blaauw et al., 2003; Heegaard, 2003; Blaauw and Christen, 2005; Bronk Ramsey, 2009; Blaauw, 2010; Blaauw and Christen, 2011 Werner and Tingley, 2015). Age models are only meaningful when created using calibrated radiocarbon dates (Bartlein et al., 1995) because of the variability in the radiocarbon calibration curve. However, the gradual refinement of the radiocarbon calibration curve (Reimer et al., 2009: Reimer et al., 2013), and increasing understanding of the need to account for reservoir ages in deriving calibrated ages on both marine (Craig, 1957; Stuiver et al., 1998; Reimer and Reimer, 1991; Franke et al., 2008) and freshwater (Godwin, 1951; Philippsen, 2013) sediments, means that even calibrated age models may need to be revisited during the construction of data syntheses. Although there has been an awareness of chronological uncertainties, most approaches to dealing with these in the context of data-model comparison have been in terms of either selecting sites with chronologies that are believed to be most reliable or through assigning some kind of quality control index (e.g. Street-Perrott et al., 1989; Wright et al., 1993; MARGO Project Members, 2009; Giesecke et al., 2014) – an approach that is difficult to combine with numerical estimates of uncertainty.

There are uncertainties caused by the different techniques used in different laboratories for measuring particular variables. Differences in the protocols used for sample cleaning, types of machine used for measurement and machine calibration, for example, have been shown to yield differences of up to 3°C in the sea-surface temperature estimates derived from Mg/Ca measurements of planktonic foraminifera (Roesenthal et al., 2004; Greaves et al., 2008). Similarly large inter-laboratory differences have been found for stable isotope analyses on bone collagen, again deriving from differences in sample cleaning and instrumentation (Pestle et al., 2014). While inter-laboratory differences in other types of measurement appear to be smaller than the actual measurement uncertainty (e.g. Foster et al., 2013), the fact that there are differences between measurements made by different groups poses difficulties for data synthesis. Again, it is difficult to know how to incorporate these uncertainties within a traditional data-model comparison framework.

The general approach to using palaeo-reconstructions for model evaluation is to use the estimated change in reconstructed climate and compare this with the simulated change between a palaeo-experiment and a control, usually a pre-industrial control. Little thought has been given to the choice of baseline climate, either for the reconstructions or for the simulations. Hessler et al. (2014) showed that the choice of baseline climate, in this case SST anomalies based on either the WOA98, WOA09 or HadiSST data sets, made an average absolute difference of 0.3-0.4°C to mid-Holocene SST reconstructions with differences of >1°C in the Mediterranean and eastern Pacific. Although the MARGO Project used a standard baseline climatology (MARGO Project Members, 2009), other SST data sets (e.g. Ruddiman and Mix, 1993; Leduc et al., 2010; Marcott et al., 2013) have different definitions of the baseline climate, and this needs to be taken into account when combining these sources to create data sets for model evaluation. Differences in temperature between the pre-industrial (PI) control and the mid-20th century in the historical simulations (1961-1990 CE, i.e. the interval most nearly corresponding to the modern observational data sets used for statistical calibrations) can also be of the order of 0.5–1.0°C locally and this may also contribute to mis-matches between simulated and reconstructed climates (e.g. Wagner et al., 2012).

The largest source of unquantifiable uncertainty in palaeoclimate reconstructions is associated with the climate interpretation of a given record. Biological assemblages contain a wealth of information, and this underpins their use to make reconstructions of multiple climate variables (e.g. Webb et al., 1993; Jackson et al., 2000; Davis et al., 2003; Cheddadi et al., 2007; Fréchette et al., 2008; Guiot et al., 2008). Derivation of a statistical relationship with a specific climate variable under modern climate conditions is based on the fact that this variable either controls or is correlated with something that controls the growth of the organism. July temperature, for example, has a limited direct impact on plant growth, but it is generally correlated in the northern hemisphere with the length and warmth of the growing season, which is the major determinant of whether plants can accumulate sufficient carbon to survive and reproduce. Very high July temperatures also tend to be associated with heat and/or moisture stress that can impact photosynthesis and strongly determine the composition and structure of vegetation (Kohfeld and Harrison, 2000; Harrison et al., 2010). Similarly, mean January temperature in the northern hemisphere is usually highly correlated with daily extreme low temperatures in winter, which determine whether a plant is killed by frost and therefore exert a strong selective pressure that differentiates plants with different overwintering mechanisms (Woodward, 1987; Harrison et al., 2010). The definition and adoption of "bioclimatic" variables, such as mean temperature of the coldest month, accumulated growing season warmth, and indices of plant-available soil moisture, for climate reconstruction was an attempt to move closer to the actual controls on plant growth (e.g. Cheddadi et al., 1997; Tarasov et al., 1999; Peyron et al., 2000) and which could therefore be expected to be invariant through time. Nevertheless, the palaeoclimate record is characterized by changes in seasonality, interannual variability and the frequency of extremes – all of which have the potential to invalidate modern-day correlations even between bioclimatic variables and species abundance. Furthermore, at least as far as terrestrial plants are concerned, statistical relationships with climate are modulated by the fact that plants respond directly to changes in atmospheric CO_2 concentration through changes in productivity and water-use efficiency (Street-Perrott et al., 1997; Cowling and Sykes, 1999; Harrison and Prentice, 2003: Prentice and Harrison, 2009). It is not possible to take this into account using statistical techniques, and this is likely a contributory cause of the breakdown of statistical relationships between climate and tree-ring width in recent years (D'Arrigo et al., 2008; Gagen et al., 2011) and could also impact reconstructions of high-CO_2 intervals such as the mid-Pliocene (e.g. Salzmann et al., 2013) and low-CO_2 intervals such as the LGM (Jolly and Haxeltine, 1997; Cowling and Sykes, 1999; Prentice and Harrison, 2009). The effect of changing CO_2 will also impact on palaeo-reconstructions of other plant properties, including leaf area index and tree cover (e.g. Gonzalez et al., 2008; Williams et al., 2011).

An alternative way of exploiting palaeoenvironmental data for climate-model evaluation is to use models that explicitly simulate the sensor – for example, vegetation (Kaplan et al., 2003; Prentice et al., 2011a), tree rings (Evans et al., 2006; Li et al., 2014), fire (Prentice et al., 2011b; Martin Calvo et al., 2014), the dust cycle (Werner et al., 2003; Mahowald et al., 2006; Takemura et al., 2009), peat growth (Charman et al., 2013), glacier mass balance (Michelmayer et al., 2008), marine biogeochemistry (Aumont et al., 2003; Bopp et al., 2003) or phytoplankton abundance (Le Quéré et al., 2005), and stable isotopes in corals (Thompson et al., 2011). This type of "forward" modelling using a simple biogeography model (BIOME4: Kaplan et al., 2003), for example, makes direct comparisons with observations possible and discriminates between the performance of different climate models (Fig. 6). Many climate models now explicitly simulate isotopic tracers (e.g. Schmidt et al., 2007; Sturm et al., 2010; Holloway et al., 2016) to facilitate model evaluation and diagnosis. Similarly, there are an increasing number of models that simulate vegetation, fire and the dust cycle (e.g. Lawrence et al., 2011; Reick et al., 2013; Kok et al., 2014), primarily in order to account for feedbacks to climate. The explicit simulation of these components of the climate system greatly facilitates comparisons with natural records (see e.g. Wasson and Claussen, 2002; Ohgaito et al., 2013).

A natural extension of the forward modelling approach is to use inversion techniques to derive quantitative climate reconstructions that are consistent with a process-based model (Guiot et al., 2000; Wu et al., 2007; Hatté et al., 2009; Garretta et al., 2010; Boucher et al., 2014). The use of process-based models in palaeoclimate reconstruction sidesteps many of the potential problems with correlation-based statistical methods. One caveat to the use of process-based modelling is the assumption that the model used is correct. Projections of future changes in vegetation (Piao et al., 2013; Friedlingstein et al., 2014) and fire (Harrison et al., 2010; Kloster et al., 2012; Kelley and Harrison, 2014) show that different process-based models produce radically different simulations for the 21st century, despite being equally good at reproducing modern day vegetation patterns and fire regimes. Thus, as with climate models, it is imperative either to use an ensemble of models or to demonstrate that the forward-model selected is reliable. It should also be noted that process-based modelling does not overcome the problem of equifinality (different climates generating similar effects), although it is possible to use this modelling approach to determine the range of potential climates and the probabilities associated with each (Garreta et al., 2010).

A particular motivation for the use of process-based models for reconstructing climate from palaeovegetation records is the strong effect of changes in atmospheric CO_2 concentration on vegetation composition (as discussed above). Manifest today in worldwide "woody thickening" (increase of tree density especially in tropical savannas), this same effect also accounts for much of the extreme reduction in forest area in the tropics and subtropics during glacial periods; and the globally lower than present terrestrial carbon storage at the LGM (Harrison et al., 2010), which was only partly counteracted by greater than present storage of inert carbon in permafrost (Ciais et al., 2013). There is no obvious way to build the CO_2 effect into statistical climate reconstruction methods because at any one time there is very limited variation in CO_2 concentration across the globe. Process-based models, including BIOME4, include a CO_2 effect on vegetation composition (a consequence of the effect of CO_2 on photosynthesis, and the differential effects on plants with the C_3 and C_4 pathways) and so inversion of such models can take known changes in CO_2 concentration into account (Guiot et al., 2000; Wu et al., 2007; Hatté et al., 2009; Garretta et al., 2010; Boucher et al., 2014). An alternative approach, which decouples the consideration of CO_2 effects from the use of a specific process-based model, involves defining a bioclimatic index that reflects "apparent" plant-available moisture, as sensed by plants responding to changes in atmospheric CO_2. Wang et al. (2013) used this approach to modify results of a statistical model to predict vegetational responses to future climate change. It could potentially be adapted to "correct" palaeoclimate reconstructions made from palaeovegetation data by any method (Prentice, 2015). The correction would generally be to increase palaeoprecipitation estimates for periods of low CO_2 and to decrease them for periods of high CO_2. The use of such a physically-based correction factor could provide a rapid method of modifying existing statistical reconstructions of palaeoprecipitation to account for the direct impact of CO_2 on plant growth.

Despite various sources of uncertainty, quantitative reconstructions can be used for model evaluation and benchmarking as long as the climate signal being examined is larger than the potential uncertainties (Lohmann et al., 2013; Harrison et al., 2014) and especially when reconstructions derived using different methods (e.g. statistical techniques, forward modelling, model inversion) show similar, spatially coherent patterns that are consistent with a single climatic explanation (Bartlein et al., 2011).

The terms 'evaluation' and 'benchmarking' are not synonymous. Benchmarking is a measurement tool, whereby model outputs are compared to a pre-defined set of observations using appropriate metrics to define the degree of agreement quantitatively (Taylor, 2001; Gleckler et al., 2008). Benchmarking serves multiple functions. It allows the performance of different models to be compared, but it can also be used to identify processes that require improvement in a particular model or to evaluate parameter choices, including ensuring that improvements to one component of a model do not compromise performance in another. Benchmarking is routinely used to assess climate-model performance under modern conditions, including investigation of parameter uncertainties (e.g. Murphy et al., 2004) and multi-model comparison (e.g. Reichler and Kim, 2008). It has been used to inform model development (e.g. Jackson et al., 2008) and to assess the reliability of projections of future climate (e.g. Hall and Qu, 2006). Globally comprehensive syntheses that include multiple climate variables now make routine benchmarking of palaeosimulations possible (e.g. Flato et al., 2013; Hargreaves et al., 2013; Harrison et al., 2014).

Evaluation of the CMIP5 simulations: what have we learnt?

There are now many papers presenting analyses and evaluations of the CMIP5 palaeosimulations. The PAGES 2k–PMIP3 group (2015) have made initial analyses of the LM simulations. A preliminary summary of the analyses related to the MH and LGM simulations was presented by Harrison et al. (2015). Here we draw on the Harrison et al. (2015) paper to outline some of the major lessons that have been learnt from comparing simulated and reconstructed climates for these two periods.

Large-scale features of climate that are governed by the energy- and water-balance show remarkably consistent simulated responses to changes in forcing in different climate states. For example, the magnitude of the temperature change over land compared to ocean is consistent in both warm and cold climate states: depending on the sign of the forcing, the land warms or cools by ca 2.36 times more than the ocean (Fig. 7). The ratio of the land-sea temperature contrast is constant over a wide range of climates, including climates with higher-than-present CO_2 levels (Izumi et al., 2013; Lunt et al., 2013; Hill et al., 2014; Schmidt et al., 2014). Land-ocean contrast is primarily driven by changes in surface downward clear-sky longwave radiation, which includes the effect of changes in CO_2, water vapour, and atmospheric energy transport (Izumi et al., 2014). The relative change in tropical temperature compared to high latitudes (often referred to as polar or high-latitude amplification) is also consistent across different climate states. Again, the major driver of this contrast is surface downward clear-sky longwave radiation, with surface albedo playing a significant but secondary role in promoting high-latitude amplification in both cold and warm climates (Izumi et al., 2014). The simulated magnitude of the relative changes in land-sea contrast and in high-latitude amplification is supported by historical and LGM observations, confirming that the simulated changes are realistic (Izumi et al., 2013). Thus, palaeo-evaluation of the CMIP5 simulations does confirm that the large-scale patterns of temperature change in future projections are believable.

Large-scale changes in precipitation scale with temperature, increasing as temperature increases and decreasing in cold climates. The change in precipitation per degree change in temperature is approximately the same in palaeoclimate, historical, and increased CO_2 simulations. The rate of change is consistently smaller than the rate of change in saturation vapour pressure (i.e. it is much less than predicted by the Clausius-Clapyeron relationship), partly because of energetic constraints on evaporation, and partly because of constraints in water availability over land (Trenberth and Shea, 2005; Allan, 2009). Geographical differences in the strength of these constraints means there are larger changes in precipitation per degree temperature change over the ocean than over land, and in extratropical than tropical land areas (Li et al., 2013). All of these large-scale features are consistent with palaeoclimate and historical observations (Li et al., 2013). Again, the palaeoclimate diagnosis of the CMIP5 simulations confirms that the large-scale patterns of precipitation change in future projections are believable.

The CMIP5 simulations of MH and LGM climates show only moderate skill in predicting observed patterns of climate change overall (Hargreaves et al., 2013; Hargreaves and Annan, 2014; Harrison et al., 2014; Harrison et al., 2015) and this arises because of persistent problems in simulating regional climates. In the MH, for example, models predict an increase in the northern hemisphere monsoons in response to the orbitally-driven increase in summer insolation. This increase in monsoons is amplified by ocean and land-surface feedbacks. Nevertheless, the models do not produce as large a change in either the amount of rainfall or the extent of the area influenced by monsoon precipitation as indicated by observations. The discrepancy between observed and CMIP5 simulated changes in MH precipitation over northern Africa between 15°-30° N is at least 50% (Perez-Sanz et al., 2014). The mismatch between simulated and observed monsoon climates has not been reduced in the CMIP5 simulations compared to simulations made with earlier generations of models (Harrison et al., 2015).

Failure to capture the magnitude of an observed change suggests that there are feedback processes that are either not included or are poorly treated in the current generation of models. However, differences in the sign of regional changes between observations and simulations are likely to indicate more fundamental problems. The CMIP5 MH simulations show drier conditions in the Eurasian mid-continent, particularly between 45°-60° N, whereas observations systematically show the region was wetter than today. The simulated drying leads to a significant warm temperature bias in this region, whereas observations indicate that the mid-continent had cooler summers than today. Discrepancies in the sign of regional climate change are also found in other extratropical regions, most notably in southern Europe where the models show warmer summers and the observations indicate cooler summers during the MH (Mauri et al., 2014). Mauri et al. (2014) suggested this mismatch was due to poor simulation of the short-term variability in atmospheric circulation, specifically the prevalence of anticyclonic blocking in summer and increased dominance of the positive phase of the North Atlantic Oscillation in winter during the MH.

There has been little assessment of how well models reproduce changes in short-term climate variability during the MH and LGM, in part because of the lack of large-scale syntheses of high-resolution palaeodata. In general, models underestimate interannual variability under modern climate conditions (Flato et al., 2013). The direct observational record is too short to know how well they capture decadal to centennial-scale variability. Oxygen isotope measurements on marine carbonates (corals, molluscs) from the tropical Pacific Ocean show a substantial reduction in the strength of the El Niño-Southern Oscillation (ENSO) through most of the Holocene, culminating between 5 and 3 ka (Emile-Geay et al., 2015). Most of the CMIP5 simulations show a reduction in ENSO in the MH compared to the pre-industrial climate, but it is very much smaller than the reduction shown by the palaeo-observations and only marginally significant. This raises the possibility that an important component of the observed changes in ENSO may result from internal variability. The contribution of internal variability to projected future climate generally decreases through the 21st century, but nevertheless remains an important contribution to the uncertainties in projections of regional precipitation e.g. in Asia and Europe throughout the century (Kirtman et al., 2013). It is also clear that a considerable part of the differences in the simulated response to forcing during the last millennium can be explained in terms of internal variability (Goosse et al., 2012; Masson-Delmotte et al., 2013).

Model responses to forcing at a regional scale are not always consistent. Various CMIP5 models show opposite changes in the location of the southern hemisphere westerlies during the LGM, for example, with half showing a equatorward shift and half showing a poleward shift in mean position compared to the pre-industrial control (Chavaillez et al., 2013; Rojas, 2013). The models that unexpectedly simulate a poleward shift of the jet stream at LGM show a strong cooling in the lower troposphere at high latitudes, which suggests that inter-model differences in the position of the westerlies may reflect different sensitivity to prescribed changes in the Antarctic ice sheet (Chavaillez et al., 2013). In the CMIP5 MH simulations, there is a consistent reduction of summer sea-ice cover in response to increases in summer insolation but some models show increased and some decreased ice thickness in winter (Berger et al., 2013). These inter-model differences appear to be related to differences in the cloud feedback. Differences in the response between models are potentially helpful, providing that the actual response is well-constrained by observations, because they offer the possibility of determining the correct sensitivity to different feedbacks.

The systematic biases in the simulation of regional climates means that models are generally better at simulating mean values of any climate variable than at simulating the spatial variability or the geographical patterning in that variable (Harrison et al., 2014: Fig. 8). Nevertheless, the benchmarking of the CMIP5 MH and LGM shows that some models consistently perform better than others, even in the prediction of spatial patterning (Harrison et al., 2014). Unfortunately, better performance in palaeo-simulations is not related to better performance under modern conditions (Harrison et al., 2015). The ability to simulate modern climate regimes and processes does not mean that a model will be good at simulating climate changes. This emphasises how important it is to test models against the palaeorecord if we are to have any confidence in their projections of future climate (Braconnot et al., 2012; Hargreaves and Annan, 2014; Schmidt et al., 2014).

The future

Evaluation of the CMIP5 palaeo-simulations demonstrates the value of using past climates as data targets in model intercomparisons. It has been shown that the broad-scale temperature and precipitation simulated responses to past changes in forcing are correctly represented, and this suggests they are features of the actual response of the climate system to changes in forcing rather than model artefacts. Projected changes in land-sea temperature contrast, high-latitude amplification, temperature seasonality and the scaling of precipitation with temperature are therefore likely to be reliable. But models are much less reliable at predicting regional climate changes. The palaeo-record has the ability to discriminate between models where they show differences in the response to forcing, and again this provides a way of determining which models are more or less reliable. Efforts to improve the skill of climate models based on evaluation using modern climate states are having a declining impact (Knutti, 2010; Rausser et al., 2014), pointing to a need for innovation (Stevens and Bony, 2013; Palmer, 2014). We are therefore at a key moment for the climate modelling enterprise to benefit from insights gained from the study of past climates. There are a number of areas that have been identified as potential sources of error in the simulation of regional palaeoclimates, including the balance between deep and shallow convection in monsoon regimes (Zheng and Braconnot, 2013), incorrect representation of water- and energy-exchanges between the land and the atmosphere (Harrison et al., 2015), poor understanding of the relationship between mean climate state and short-term climate variability (Emile-Geay et al., 2015) and failure to capture the short-term variability in atmospheric circulation (Mauri et al., 2014). Further investigation of these issues in the radically different climate regimes of the past could provide clues to improve state-of-the-art models.

The need for renewed effort is not confined to the modeling community. For example, our ability to evaluate model performance in the southern hemisphere is currently limited by a lack of coherent and consistent syntheses of the available palaeoenvironmental data. There is an urgent need for quantitative climate reconstructions covering South America and Australia. The use of existing quantitative reconstructions could also be improved, in particular through the development of standardized measures of uncertainty and exploitation of probabilistic approaches to comparison. Our ability to explore the linkages between forced changes in the mean climate and short-term climate variability is limited by the lack of global-scale syntheses of high-resolution records that extends beyond the past two millennia. Again, a community focus on producing such syntheses would be worthwhile. But the likely complexity of the seasonal changes in climate in the geological past, coupled with the known complexity of the controls on biological systems, means there will always be large uncertainties associated with statistical reconstructions. An emphasis on developing and using process-based models of a range of palaeoenvironmental sensors is also required to improve climate-model evaluation. The application of process-based models will facilitate a more systematic exploitation of existing syntheses of qualitative data. Many of the synthetic products are out-of-date and do not include sites published in the last decade or so, and thus a community effort to update these data sets would be useful.

Equilibrium time-slice simulations have been the focus of climate modeling for many decades, and this type of simulation will still be a focus of the next phase of the Climate Model Intercomparison Project (CMIP6: Meehl et al., 2014). However, many features of the climate system cannot be examined using equilibrium simulations. As the CMIP5 Last Millennium experiment has demonstrated, long transient simulations are now possible; and indeed transient simulations of both the deglaciation and Holocene using the same models that are used for future climate projections are planned within PMIP. Evaluation of transient simulations poses new and as yet unexplored issues for data synthesis and data-model comparison.

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Figure and Table Captions

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Figure 3: Reconstructed northern hemisphere global annual temperatures during the last 2000 years, redrawn from Masson-Delmotte et al. (2013). All series are anomalies from the 1881–1980 CE mean (horizontal dashed line) and have been smoothed with a filter that reduces variations on time scales less than about 50 years. Curves from instrumental records are plotted in blue, and the purple lines show a locally-weighted regression curve with a 25-yr window half-width fit to the original unsmoothed series, and the 95-percent bootstrap confidence intervals for that curve that show the impact of the individual series to the overall curve.

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Figure 6: Simulated and observed vegetation changes across North America during the mid-Holocene (MH, 6 ka). The simulations were made using the BIOME4 biogeography model (Kaplan et al., 2003) driven by long-term averages of monthly mean temperature, sunshine and precipitation derived from Palaeoclimate Modelling Intercomparison Project (PMIP2) simulations made with the (a) CSIRO-Mk3L-1.0 coupled ocean-atmosphere (OA) and (c) ocean-atmosphere-vegetation (OAV) models. The observed vegetation during the MH (b) is derived from the BIOME6000 dataset (Prentice et al., 2000; Bigelow et al., 2003; data available from the PMIP2 website: https://pmip2.lsce.ipsl.fr/). The OAV model does not shows appreciably greater agreement with the observed vegetation then the less complicated OA model.

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Figure 8: Comparison of median and interquartile ranges (IQR) of observed and simulated growing season temperatures (as measured by growing degree days above a threshold of 5°C: GDD5) in (a) the mid-Holocene and (b) the Last Glacial Maximum. The comparisons are made using only the model land grid cells where there are observations. The reconstructed GDD5 is from the Bartlein et al. (2011) data set. The models are colour-coded to show whether they are CMIP5 simulations or from the previous generation of simulations made by the Palaeoclimate Modelling Intercomparison Project (PMIP2), and whether they are ocean–atmosphere (OA), ocean–atmosphere-vegetation (OAV) or OA carbon-cycle (OAC) models. The simulated median for each model is shown by a vertical line, the box represents the IQR.

Table 1: Description of the palaeosimulations included in the fifth phase of the Coupled Model Intercomparison Project (CMIP5) and of the boundary conditions specified for these experiments.

Table 2: List of models and institutions contributing palaeoclimate simulations to the fifth phase of the Coupled Model Intercomparison Project (CMIP5). The model names are the codes used to identify each model in the CMIP5 archive.

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Abbreviation (in this paper)	**Name of Experiment (in ESGF database)**	**Description**	**Boundary conditions**
PiControl	*piControl*	Equilibrium simulation of 1850 CE, used as control for MH and LGM simulations (also used as a baseline for historical simulations by groups that did not run the palaeosimulations)	**Orbital parameters**: eccentricity = 0.016724, obliquity = 23.446°, perihelion-180° = 102.04° **Trace gases**: CO_2 = 280 ppm, CH_4 = 650 ppb, N_2O = 270 ppb, CFC = 0, O_3 = modern-10 DU **Ice sheet**: modern **Land surface**: modern or computed with dynamical vegetation model **Carbon cycle:** Interactive, with atmospheric concentration prescribed and ocean and land carbon fluxes diagnosed as recommended in CMIP5 *Note*: modelling groups that did not run palaeosimulations could have used a slightly different configuration for the PiControl
LM		Transient simulation of the last millennium, 850-1850 CE	
MH	*midHolocene*	Equilibrium simulation of 6 ka	**Orbital parameters**: eccentricity = 0.018682, obliquity = 24.105°, perihelion-180° = 0.87° **Trace gases**: CO_2 = 280 ppm, CH_4 = 650 ppb, N_2O = 270 ppb, CFC = 0, O_3 = same as in CMIP5 PI **Ice sheet**: as in CMIP5 PiControl **Land surface**: Computed using a dynamical vegetation module or prescribed as in PiControl, with phenology computed for models with active carbon cycle or prescribed from data **Carbon cycle**: Interactive, with atmospheric concentration prescribed and ocean and land carbon fluxes diagnosed as recommended in CMIP5
LGM	*lgm*	Equilibrium simulation of the Last Glacial Maximum, 21 ka	**Orbital parameters**: eccentricity = 0.018994, obliquity = 22.949°, perihelion-180° = 114.42° **Trace gases**: CO_2 = 185 ppm, CH_4 = 350

			ppb, N_2O = 200 ppb, CFC =0, O_3 = as in CMIP5 PI **Ice sheet**: Prescribed consensus ice sheet as described on PMIP3 website, with consistent changes to land-sea mask and sea level **Land surface**: Computed using a dynamical vegetation module or prescribed as in PiControl, with phenology computed for models with active carbon cycle or prescribed from data **Carbon cycle**: Interactive, with atmospheric concentration prescribed and ocean and land carbon fluxes diagnosed as recommended in CMIP5

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Model name	Institution	PI control	Last Millennium	Mid-Holocene	Last Glacial Maximum
BCC-CSM1	Beijing Climate Center, China Meteorological Administration, China	✔		✔	
CNRM-CM5	Centre National de Recherches Météorologiques/Centre Européen de Recherche et Formation Avancée en Calcul Scientifique, France	✔		✔	✔
CSIRO-Mk3-6-0	Commonwealth Scientific and Industrial Research Organisation in collaboration with the Queensland Climate Change Centre of Excellence, Australia	✔		✔	
EC-EARTH	EC-Earth consortium	✔		✔	
FGOALS-g2	LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences; and CESS, Tsinghua University, China	✔		✔	
FGOALS-g2	LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences; and CESS, Tsinghua University, China	✔		✔	
GFDL-ESM2G	NOAA Geophysical Fluid Dynamics Laboratory, US	✔		✔	
GFDL-ESM2M	NOAA Geophysical Fluid Dynamics Laboratory, US	✔		✔	
GISS-E2-R	NASA Goddard Institute for Space Studies, US	✔	✔	✔	✔
HadGEM2-CC	Hadley Center, UK Met. Office, UK	✔		✔	
HadGEM2-ES	Hadley Center, UK Met. Office, UK	✔		✔	
INM-CM4	Institute for Numerical Mathematics, Russia	✔		✔	
IPSL-CM5A-LR	Institut Pierre-Simon Laplace, France	✔		✔	✔
IPSL-CM5A-MR	Institut Pierre-Simon Laplace, France	✔		✔	
MIROC-ESM	Japan Agency for Marine-Earth	✔	✔	✔	✔

	Science and Technology, Atmosphere and Ocean Research Institute (The University of Tokyo), and National Institute for Environmental Studies, Japan				
MIROC5	Japan Agency for Marine-Earth Science and Technology, Atmosphere and Ocean Research Institute (The University of Tokyo), and National Institute for Environmental Studies, Japan	✔		✔	
MPI-ESM-P	Max Planck Institute for Meteorology, Hamburg, Germany	✔	✔	✔	✔
MRI-CGCM3	Meteorological Research Institute, Tsukuba, Japan	✔		✔	✔
NCAR-CCSM4	National Center for Atmospheric Research, US/Dept. of Energy/NSF	✔		✔	✔
NorESM1-M	Norwegian Climate Centre, Norway	✔		✔	

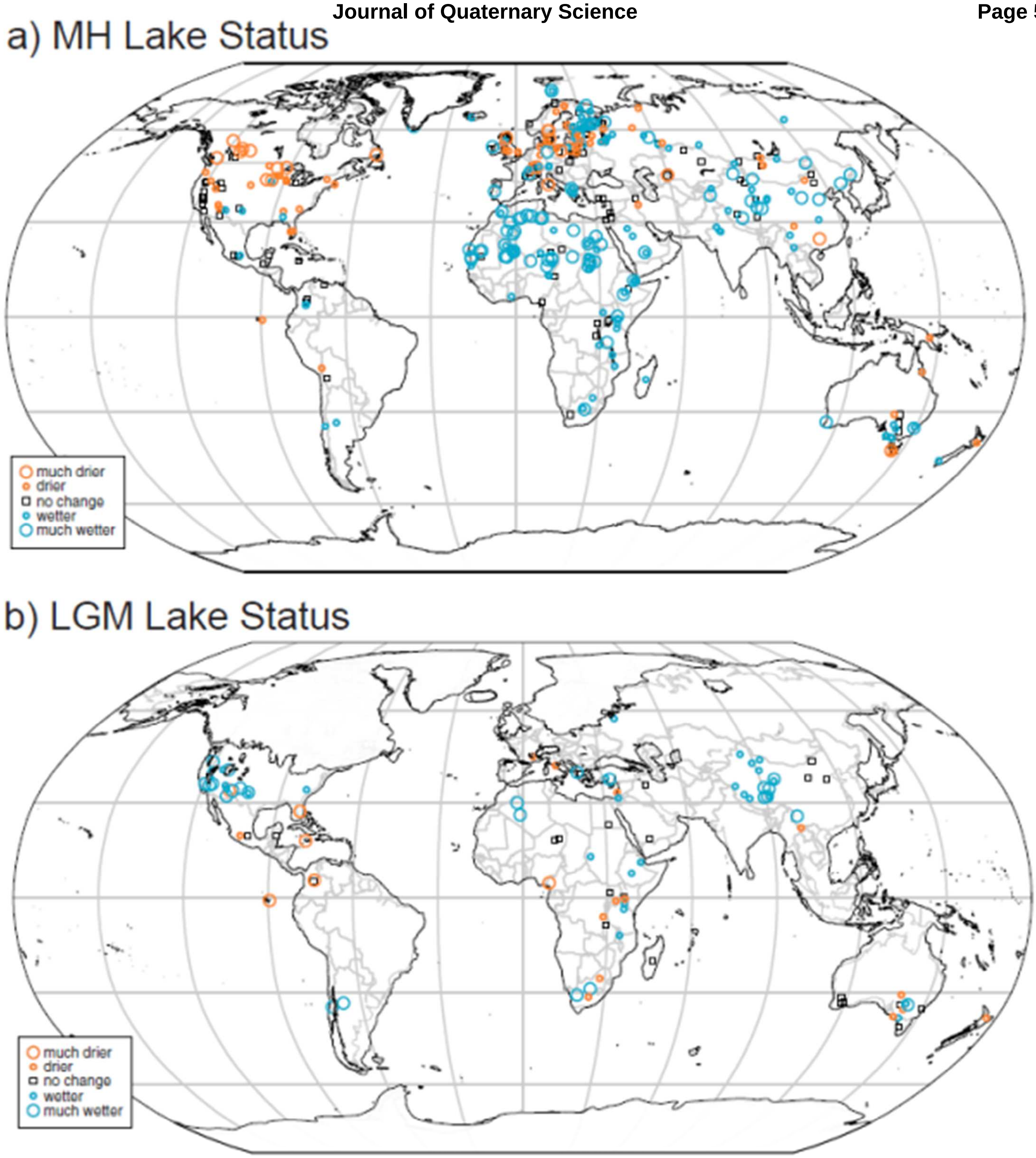

Figure 1: Changes in lake status (a) in the mid-Holocene (MH, ca 6000 yr B.P.) and (b) at the Last Glacial Maximum (LGM, ca 21000 yr B.P.) compared to present day. Data from the Global Lake Status Database (Kohfeld and Harrison, 2000).

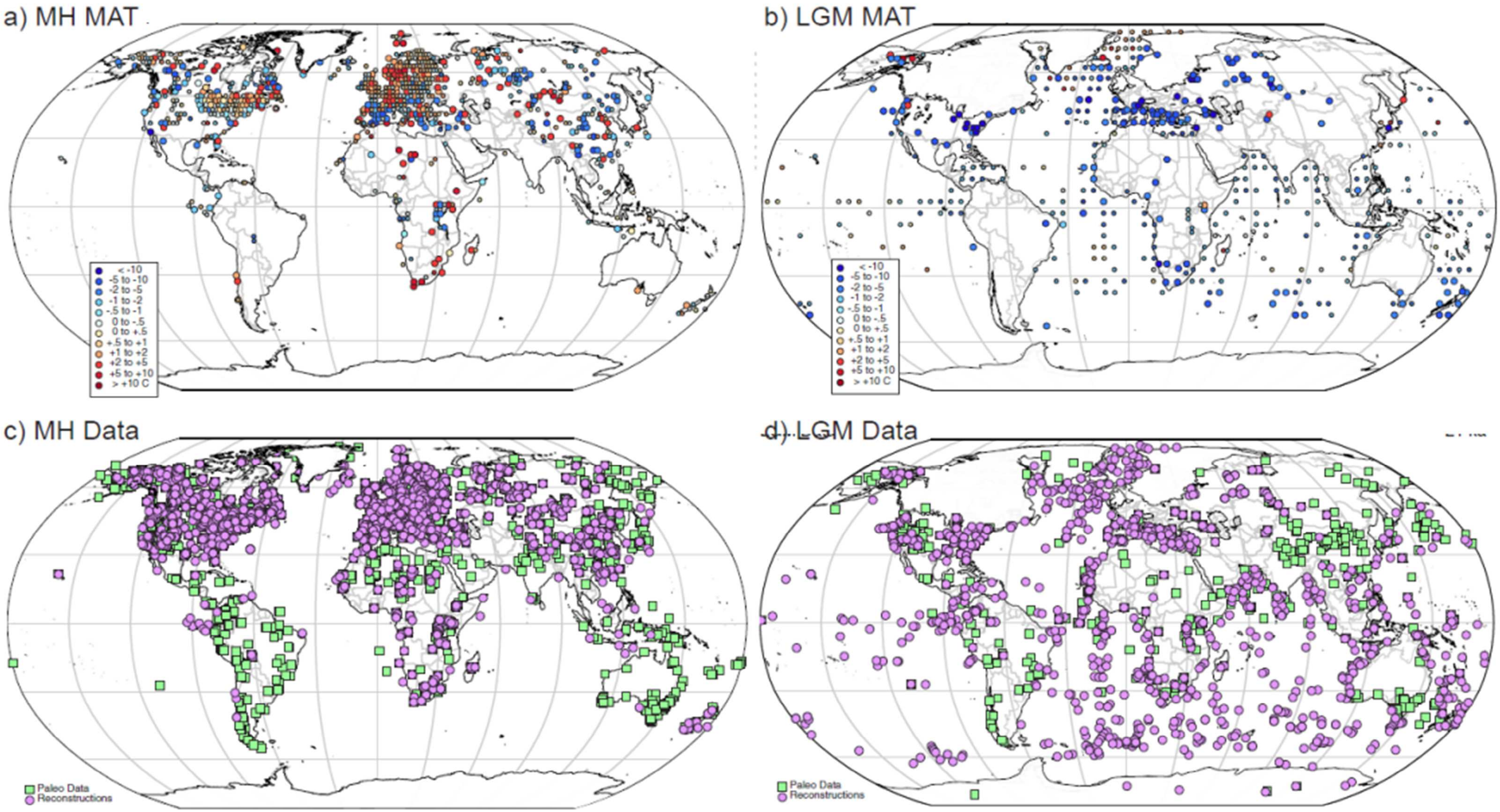

Figure 2: Reconstructed changes in mean annual temperature (MAT) (a) in the mid-Holocene (MH, ca 6000 yr B.P.) and (b) at the Last Glacial Maximum (LGM, ca 21000 yr B.P.) compared to present day. The reconstructions of ocean temperature are from the MARGO database (MARGO Project Members

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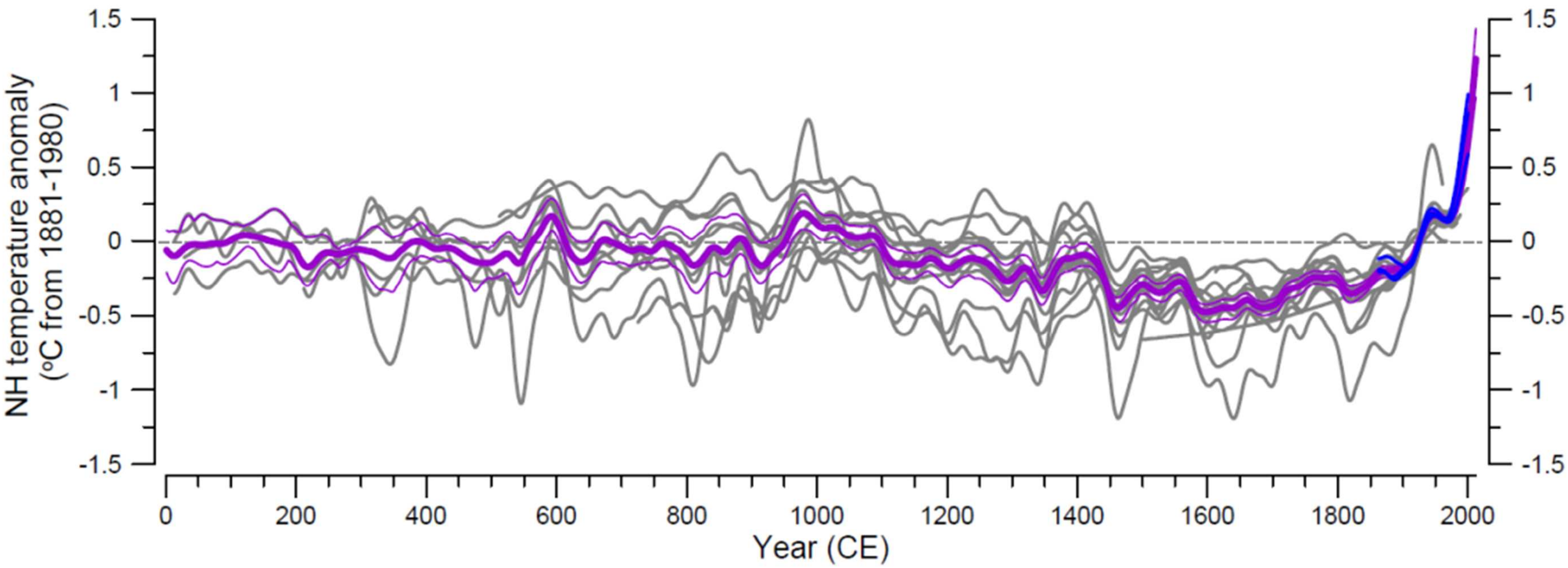

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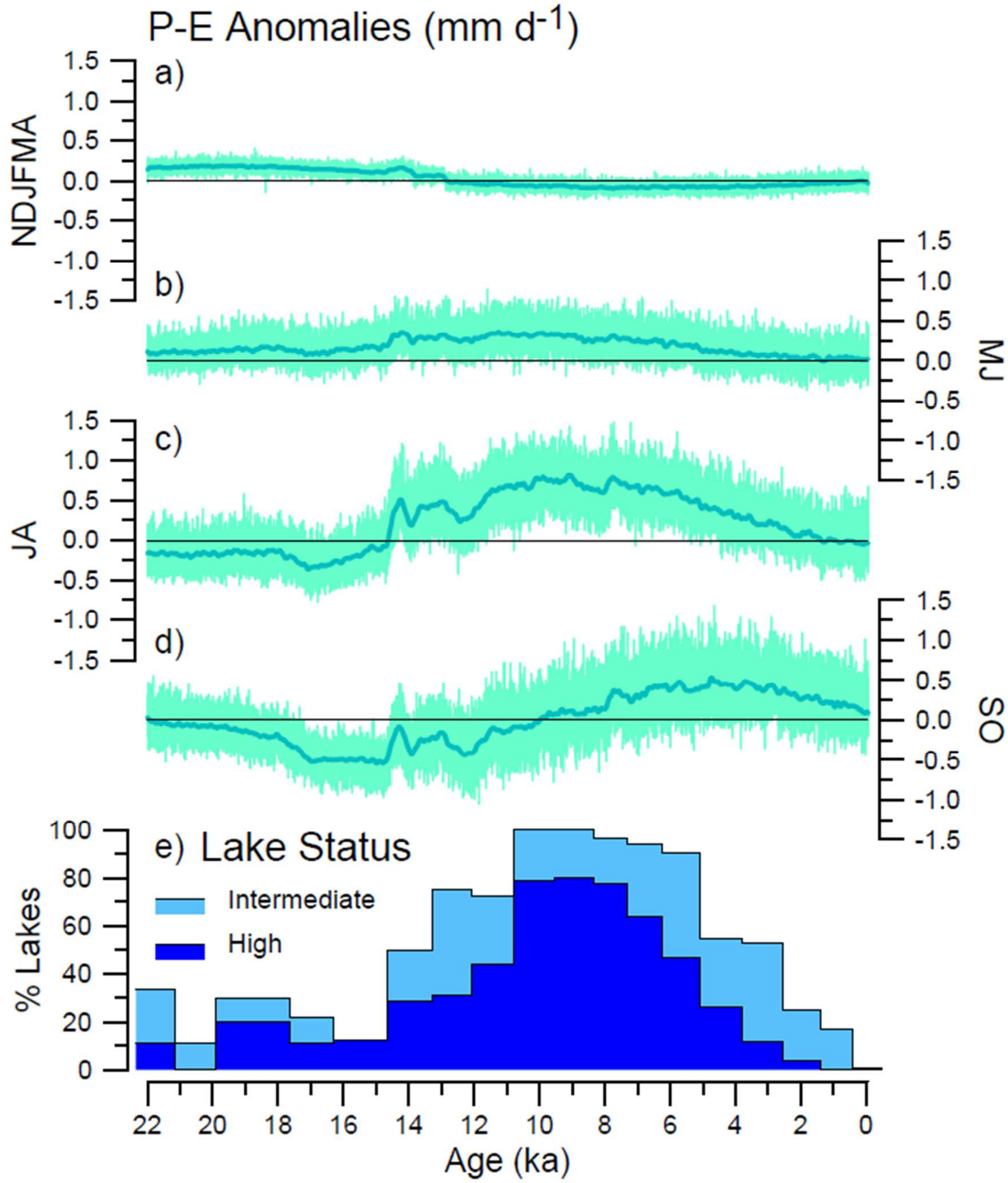

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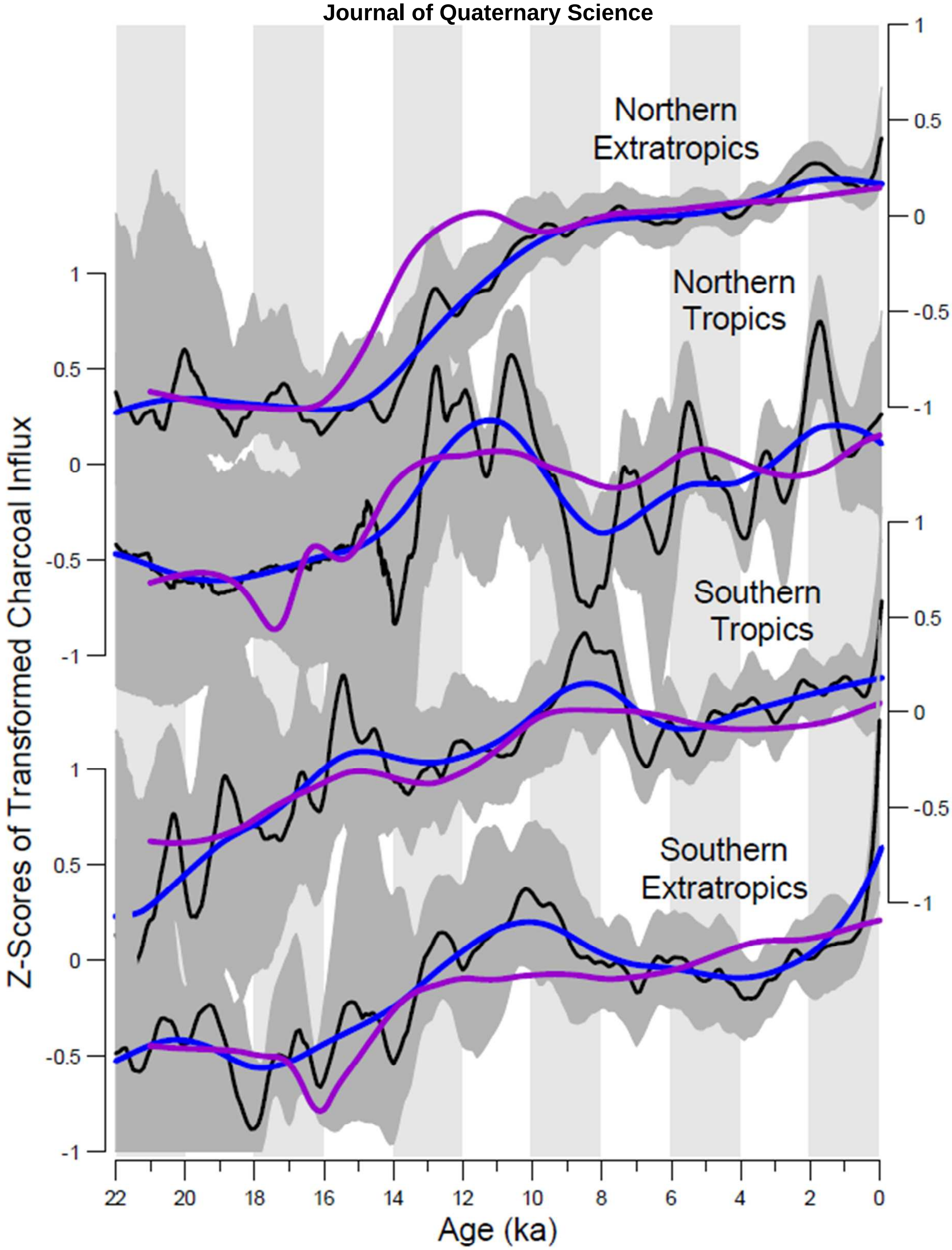

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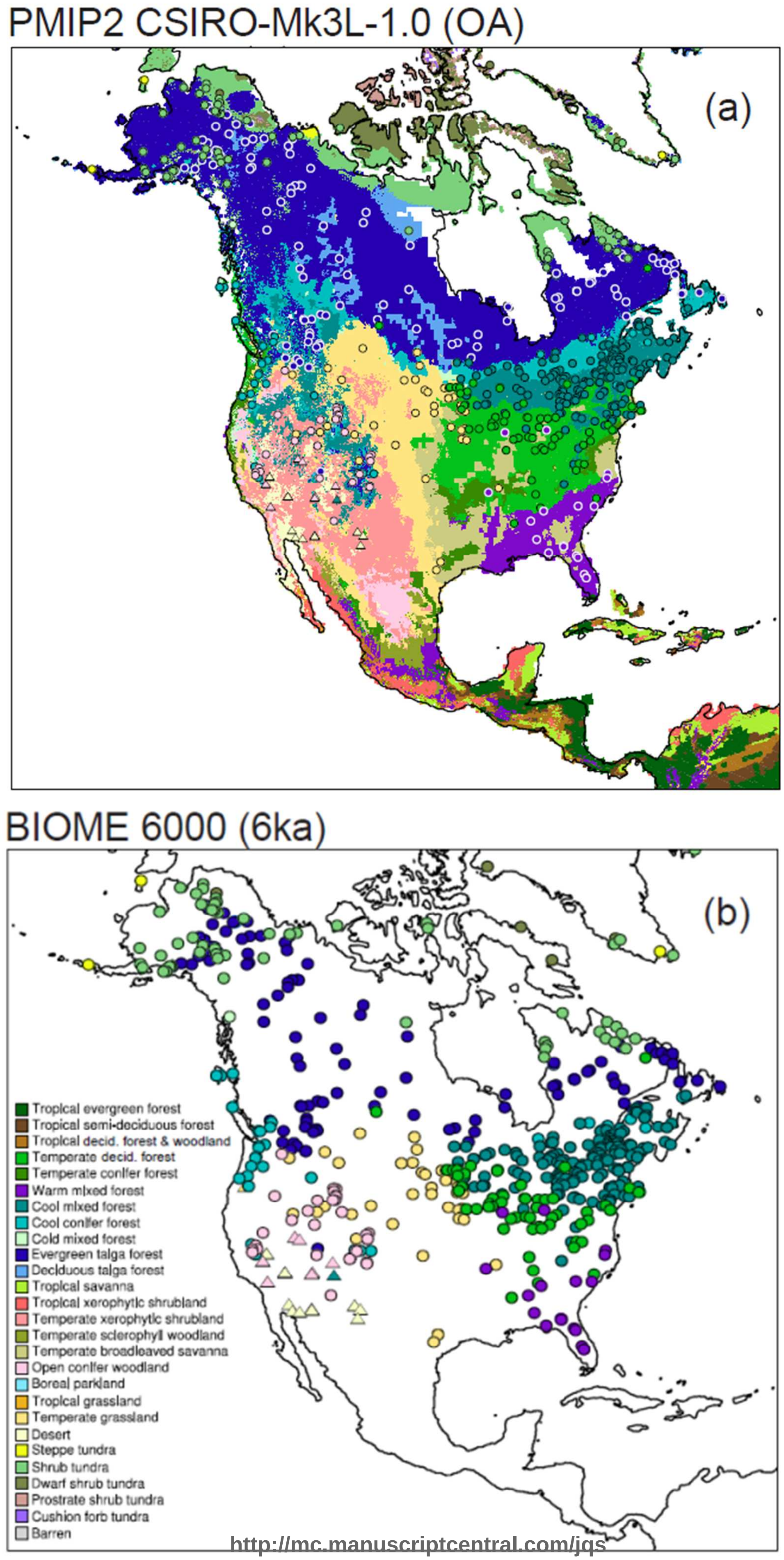
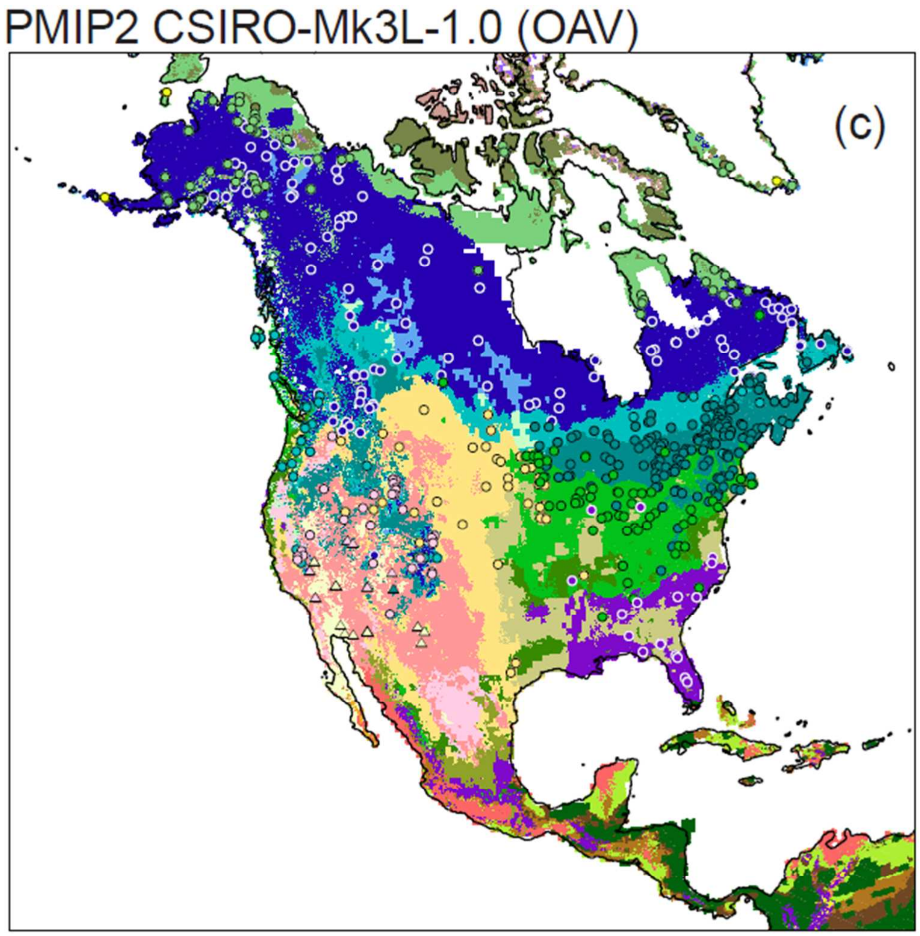

PMIP2 CSIRO-Mk3L-1.0 (OA)
(a)
BIOME 6000 (6ka)
(b)
Tropical evergreen forest
Tropical semi-deciduous forest
Tropical decid. forest & woodland
Temperate decid. forest
Temperate conifer forest
Warm mixed forest
Cool mixed forest
Cool conifer forest
Cold mixed forest
Evergreen taiga forest
Deciduous taiga forest
Tropical savanna
Tropical xerophytic shrubland
Temperate xerophytic shrubland
Temperate sclerophyll woodland
Temperate broadleaved savanna
Open conifer woodland
Boreal parkland
Tropical grassland
Temperate grassland
Desert
Steppe tundra
Shrub tundra
Dwarf shrub tundra
Prostrate shrub tundra
Cushion forb tundra
Barren



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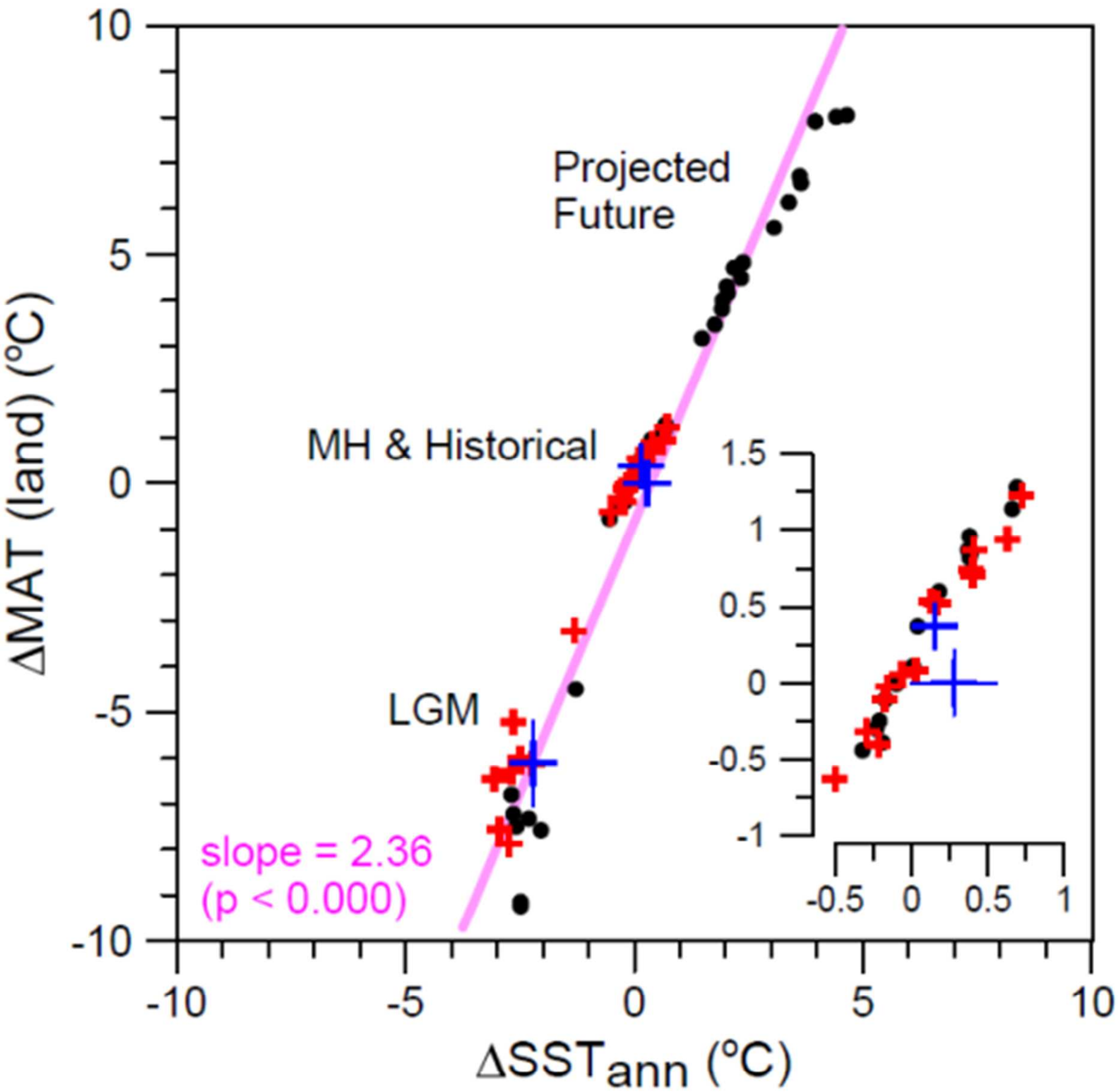

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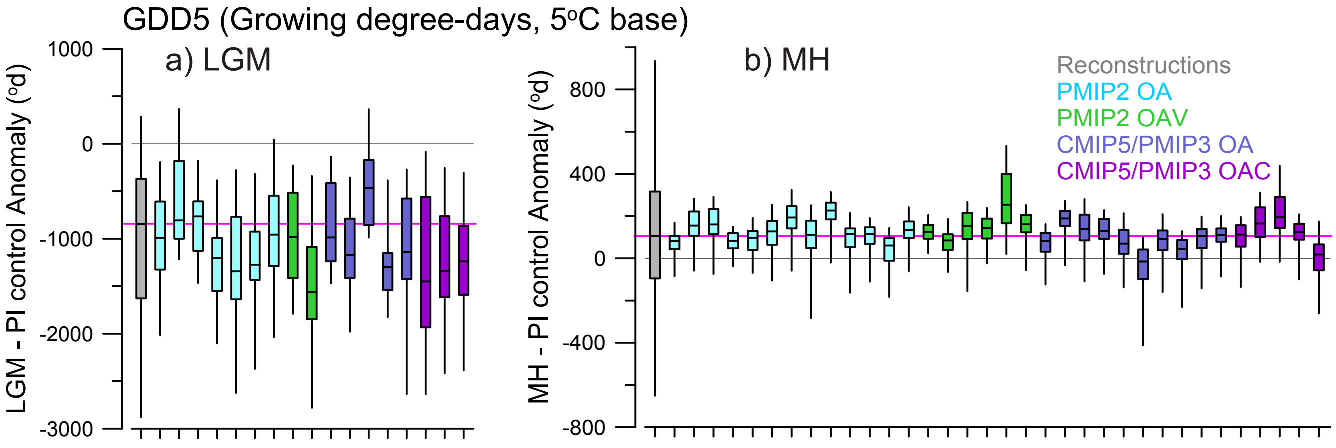
GDD5 (Growing degree-days, 5°C base)
a) LGM
LGM - PI control Anomaly (°d)
1000
0
-1000
-2000
-3000
b) MH
MH - PI control Anomaly (°d)
800
400
0
-400
-800
Reconstructions
PMIP2 OA
PMIP2 OAV
CMIP5/PMIP3 OA
CMIP5/PMIP3 OAC