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‘Study the past, if you would divine the future’: a retrospective on measuring and understanding Quaternary climate change.

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Abstract

Research on Quaternary climate change is reviewed from the perspective of the real or potential contribution to improving our ability to predict the climate of the future. For convenience the literature is divided into four timescales: orbital, sub-Milankovitch , Holocene and the last two thousand years. Four ‘challenges’ provide a framework for discussion: better understanding of the way the climate system works, better forecasting of the drivers of climate change, improved estimates of climate sensitivity (change in global mean annual temperature per unit increase in forcing) and evaluation of the models used to predict the climate of the future. Although a great deal of progress has been made, it is concluded that there are some aspects of our scientific culture that limit our potential. These include our tradition of storytelling rather than critical hypothesis testing, an over-emphasis on the role of surface water sinking in the far north Atlantic as a driver of ocean circulation and an attendant under-emphasis on the critical importance of changes in atmospheric circulation, and a lack of rigour in testing the hypothesis that changes in solar irradiance are an important driver of climate change.

Key words: climate change, solar forcing, climate sensitivity, AMOC, GCM model evaluation

Introduction

When I was invited to write this review paper for the 50th anniversary celebrations of the British Quaternary Research Association I was both honoured and daunted. Measuring and understanding the climate changes of the past are at the heart of much of Quaternary science, covering timescales from orbital cycles that extend to and beyond the last 2.6Myrs, to historical accounts and instrumental measurements covering just the last few centuries. The range of archives that Quaternary scientists have accessed includes marine sediments, ice cores, lake sediments, speleothems, loess, peat, sea shells, corals, tree rings, historical documents and many others (Lowe and Walker 1997, Bradley 2015). Given the truly multidisciplinary nature of Quaternary science, it is no surprise that the range of proxies used to extract information on past climate is now truly vast, including the physical and chemical characteristics of sediments, the isotopic composition of organic and inorganic materials, and a huge array of fossil evidence. Even the most cursory review of 50-years of development in each part of such a vast field of endeavour would produce a substantial book rather than one paper.

Rather than try to review everything, and to track the advances that have been made in their historical context, I have opted rather to look at the major advances in measuring and understanding the climate of the past from the perspective of where we now stand. Fifty years ago the study of past climate change could have been viewed as an academic diversion of little direct relevance to society. Times have changed, and we are now confident that human disturbance of the carbon cycle has upset the natural order. The carbon dioxide content of the atmosphere is higher than it has been for at least 800,000 years and the result is that the long-term decline of temperature over the last 8000 years, forced by changes in Earth's axis, has been reversed and the climate is warming. It has long been known that adding carbon dioxide and methane to the atmosphere traps outgoing long-wave radiation (the greenhouse effect), warming both the atmosphere and the oceans. The direct effect of increased concentrations of greenhouse gases on the radiation budget is known with near certainty, but there are so many feedbacks in the Earth system that the net effect on the average air temperature is very difficult to predict. Despite the best efforts of those who have tried to model the way that the planet will respond to rising greenhouse gases, we face a very uncertain future.

Intentionally or not, Quaternary researchers have now become part of a much wider and even more multi-disciplinary research community whose aim is to understand Earth's climate system and to predict the consequences of our actions. The key tools are computer models, with a range of spatial scales and internal complexities, which try to encapsulate what we know about the movements of mass and energy around the globe and aim to predict the effects of changes due to human activities. Important changes include land use, deforestation, air pollution and of course the changing chemistry of the atmosphere and increase in greenhouse gases. Improving predictions of future climate is important because living with climate change is going to be difficult and expensive, and better predictions will allow us to plan more effectively and perhaps avoid some of the worst potential environmental disasters. Presently, although there is general agreement among scientists that anthropogenic climate change is real and dangerous, our ability to predict the climate of the future is inadequate.

Quaternary scientists have already made many important contributions to our understanding of the climate system and to improving predictions of the climate of the future and we are well placed to play a central role in this critical endeavour in future. However, much of our work, in the past and at present, was never intended for this purpose; it was like most good research driven simply by curiosity. The result is that the importance and potential of some of our work is not immediately obvious to those working purely on future climate change, or indeed to some within our own community. What I have tried to do in this paper is to review major advances in Quaternary science from the perspective of their real or potential contribution to improving our ability to predict the climate of the future.

To provide some structure, I have arbitrarily divided the research on the basis of the timescales of interest into: orbital, sub-Milankovitch , Holocene and the last 2ka. In terms of contribution to improving our ability to predict the climate of the future I identify four 'challenges'.

> ***Challenge 1: Better understanding of how the climate system works.*** The challenge here is to identify and where possible quantify key processes that will help to improve the models that are used for climate prediction.
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> ***Challenge 2: Better forecasting of the drivers of climate change.*** This includes direct, external drivers of the climate system, such as orbital and solar forcing estimates, slower feedbacks such as the response of vegetation and better predictions of the magnitude and frequency of internal components of the climate system, such as El Niño/Southern Oscillation.
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> ***Challenge 3: Improved estimates of climate sensitivity***. The equilibrium change in global mean annual temperature per unit increase in radiative forcing is the key metric that describes how the climate system will respond to rising greenhouse gas emissions. Uncertainty in climate sensitivity effectively translates into uncertainty in every other aspect of the climate system.
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> ***Challenge 4: Testing General Circulation and Earth System Models.*** These are immensely complicated and contain many parameters that are not quantified with certainty. Even within a single model slight changes to the physics, within reasonable bounds of uncertainty, can result in very different predictions for the future. Different models and model variants can be tested and ranked according to how well they reproduce the observed climate, but also by how well they reconstruct the climate of the past.

Orbital-scale research on Quaternary climate change

For students entering university in the 21st century, the concept of multiple glaciations and interglacials, driven by changes in Earth's orbital parameters is already axiomatic. In the 1960s that was far from true. Although the idea that subtle changes in Earth's orbit might change the spatial and seasonal distribution of energy enough to cause ice ages had been proposed as early as the 19th century, and the three cycles had been fully quantified by Milutin Milankovitch by the 1920s, the theory was not widely accepted. In the first edition of 'Pleistocene Geology and Biology', published in 1968, Richard West was able to include some of the earliest oxygen isotope records from foraminifera in deep sea cores from the Caribbean (Emiliani 1955), which showed cyclical changes, and noted that "there is no reason to suppose that the climatic conditions which lead to repeated glaciations in the Pleistocene have ceased and it is probable that we are now living in a temperate interglacial period. The cause of the climate changes remains very much unknown, though there are a number of hypotheses" (p. 1). However, he may have known rather more than he was at liberty to reveal, for he also notes that "it is to be expected that there is a relation between orbit and climate" and that "Palaeotemperature determinations from the ocean basin sediments promise to give a general scheme of climatic change during the Pleistocene which may eventually become the basis for a world-wide correlation of the Pleistocene succession" (p. 223).

Richard West was correct in his predictions of course, for also in Cambridge at that time Nick Shackleton was pushing the boundaries of mass spectrometry to produce a very long oxygen isotope record based on foraminifera from marine sediment cores taken in the deep ocean, far from the disturbing influence of ice sheets. That work was to completely revolutionise our understanding of the climate of the past and, more widely of the functioning of the Earth system. When calcium carbonate minerals form in sea water the ratios of the stable isotopes of oxygen are partly determined by temperature, and the early papers interpreted the isotopic record in that way, primarily as a palaeotemperature record. However, in their classic paper of 1973, Nick Shackleton and Neil Opdyke reinterpreted the data as primarily a record of the isotopic ratios of the water in which the foraminifera lived, and therefore of the volume of water in the global ocean, and thus also of the volume of ice on land. The marine evidence demonstrated multiple glaciations, each one terminating with an abrupt collapse in ice volume and presumably substantial warming. The rhythmic changes in isotope ratios were found in many marine sediment cores providing, as predicted, a basis for world-wide correlation of events, and that event stratigraphy has since been confirmed in many other archives. Some of the excitement of those great discoveries is conveyed by the Imbries in their classic 'Ice Ages: solving the mystery' (Imbrie and Imbrie, 1986).

The isotopic records from 57 globally distributed benthic marine cores have now been combined (Figure 1) to provide a stacked record extending to 5.3Myr, far beyond the 2.6 Myr that now defines the Quaternary (Lisiecki and Raymo 2005). Although this provides a clear event stratigraphy, and confirms the orbital forcing of glacial cycles, it should not be viewed directly as a record of past changes in temperature; it is simply the average of the aligned records. The timescale of the stack is based largely on tuning to the orbital obliquity (tilt, c. 41k yr) cycle with a variable lag that is determined using a simple ice sheet model constrained to some extent by assumptions about sedimentation rate. Changes in the temporal coherence between the stack and the precessional cycle suggest that the relative importance of temperature and ice volume in controlling the variations in isotope ratios may not be constant through time and that changes within the system may be abrupt. Also analysis of sedimentation rates suggests that changes in isotope ratios may not be synchronous, with changes in the Atlantic leading those in the Pacific (Skinner and Shackleton 2005; Lisiecki and Raymo 2009).

Figure 1 about here $$

The important point in the context of this paper is that the oxygen isotopic ratios in benthic foraminifera are not a measure of changes in global average air temperature; they record a mixture of deep water temperature, salinity and the volume of the global ocean. The variations in $\delta^{18}O$ must of course be related in some way to variations in global mean air temperature, but there is no reason to assume that the appropriate scaling used for translation should be either linear or constant through time.

Fortunately, for the late Pleistocene the marine record is matched by even richer information contained within ice cores from Greenland and Antarctica. The highest temporal resolution is obtained in Greenland, where accumulation rate is highest, but those records only extend back to the last interglacial (Andersen et al. 2004). Antarctica has a much longer record, extending back over 800ka (Augustin et al. 2004), and as well as a continuous record of changes in $\delta^{18}O$ of precipitation provides a series of other proxies for past climate and environmental change, including variations in the carbon dioxide content of the atmosphere from bubbles trapped in the ice.

The Last Glacial Maximum (LGM), informally defined as the part of Marine Isotope Stage 2 (centred around 18 ^{14}C ka BP or 21 cal. ka BP) when global water volume was at a minimum and ice volume at a maximum (Lambeck et al. 2000, Yokoyama et al. 2000, Stokes et al. 2009, Hughes et al. 2014), provides an exceptionally useful time interval for using the climate of the past to help predict the climate of the future. It is more useful than earlier intervals because the boundary conditions in terms of the position of oceans and continents and the height of mountain ranges and plateaux, were essentially the same as today. Also, since the formation of large ice sheets takes a long time, ice sheet maxima lag the Milankovitch cycles, so that when the ice sheets were at their maximum size about 21,000 years ago the conditions that initiated glaciation had already changed and the spatial and seasonal patterns of radiation were very similar to today. There is also the great advantage that the LGM lies within the range of radiocarbon dating, as well as many other techniques including luminescence and cosmogenic exposure dating (Walker 2005), so that both organic and inorganic sediments and even landforms can be dated.

The great potential of the LGM for checking the ability of models to faithfully reconstruct the climate of the past was realised almost as soon as there were computers capable of running such models. In the 1970s this took the form of the first large interdisciplinary project aimed at 'Climate: Long Range Investigation, Mapping and Prediction'. The CLIMAP project was led by John Imbrie, Jim Hays and Nick Shackleton and brought together a large interdisciplinary team. It was a bold endeavour, to try to map temperatures across the whole globe at the LGM, but they produced such maps in 1976 and revised them in 1981 (CLIMAP project members 1976, 1981). Various projects have continued with different aspects of the work, including the 'spectral mapping project' SPECMAP, the Cooperative Holocene Mapping Project (COHMAP: 1988, Kutzbach et al. 1993, 1998), Environmental Processes of the Ice Age: Land, Oceans and Glaciers (EPILOG: Mix et al. 2001,), Glacial Atlantic Ocean Mapping (GLAMAP: Pflaumann et al. 2003, Sarnthein et al. 2003) and Multiproxy Approach for the Reconstruction of the Glacial Ocean Surface (MARGO: 2009, Gersonde et al. 2005).

The CLIMAP results for high latitudes have been refined rather than rejected by later research. The picture of large ice sheets, covering as much as 25% of the continents (Dyke et al. 2002, Svendsen et al. 2004, Hughes et al. 2014) and vast areas of tundra and steppe in the northern hemisphere (Kutzbach et al. 1998) is still accepted. High pressure cells with clockwise circulation developed over the American and Fennoscandian ice sheets, changing the hydrological conditions well beyond their limits, and ocean circulation also changed, with the Arctic Front in the North Atlantic moving as far south as the coast of Portugal and the area of winter sea ice expanding around Antarctica. The amount and seasonality of Arctic sea ice cover have been more contested (Sarnthein et al. 2003). Much more controversial, however, were the results for the tropics, where CLIMAP suggested very small increases in sea surface temperature, of generally just 1 to 2 degrees C, and with some areas even slightly warmer than today. These estimates, centred on about 1.5˚C were based on marine proxies (e.g. Moore et al. 1980), and they contrasted markedly with estimates of air temperature obtained on land, based on groundwater noble gases, coral geochemistry and changes in altitude of tropical glaciers, which suggested a reduction in tropical mean annual temperature of about 5˚C (Stute et al. 1995, Beck et al. 1992, Broecker et al. 1997). Such a large difference in estimated temperature change far from the influence of the large northern hemisphere ice sheets is critically important, because the dominant driver of change is likely to be the greenhouse effect. A small change in tropical temperatures would suggest global equilibrium sensitivity towards the low end of the accepted range whereas a 5˚C cooling would put it at the upper end.

The contradictory results for tropical temperature change derived on land and at sea were reviewed by Crowley (2000) who argued that both were wrong. CLIMAP tropical sea surface temperature estimates were based on total species assemblages, mainly comprising foraminifera, and since the assemblages do not change much down-core the reconstructed temperature changes are small. However, alternative methods of data analysis and different proxies, including Mg/Ca ratios, yield lower temperatures. He further argued that tropical cooling of 5˚C is unrealistic because it would cause a much larger change in tropical biota, including corals, than is actually observed. Pollen-derived estimates of temperature change (e.g. Colinvaux et al. 1996) are considered flawed because they are poorly calibrated and also because under LGM conditions the very low levels of atmospheric CO_2 would have a direct influence on plant distribution, giving an advantage to C4 grasses (Street-Perrot et al. 1997), so that a transfer function approach is not appropriate. The estimates of temperature based on the altitude of glaciers were flawed partly because the lapse rate calculations did not take account of changes in sea level and the moisture content of the air. Palaeotemperature estimates based on oxygen isotope ratios in coral, although very strongly correlated with the seasonal cycle over recent decades fail verification tests because they severely overestimate tropical sea surface temperatures in the first half of the 20th century. Crowley reconciles the evidence by arguing for a temperature reduction in the tropical warm pool of about 2.5˚C, which would translate into a mid-range equilibrium sensitivity of about 3˚C for a doubling of atmospheric CO_2.

Orbital scale and understanding climate

The contribution of orbital scale climate change studies to our understanding of the climate system has been profound. They have demonstrated, beyond doubt, that Earth's climate is part of a complex system, held far from equilibrium and with more than one stable state. Small changes that are not even in the total amount of solar energy that Earth receives, only in the spatial and seasonal distribution of that energy, result in huge changes in climate. The changes in energy supply are sinusoidal but the climate response is very nonlinear, characterised by generally slow declines into long periods with large northern hemisphere ice sheets followed by very abrupt terminations characterized by rapid melting and warming (Ruddiman 2006). Above all, studies on these timescales demonstrate the critical importance of feedbacks to the climate system. Climate models that are designed to run over decades to centuries focus on the 'fast feedbacks' (water vapour, lapse rate, cloud cover, snow and ice albedo), but at orbital scales it is the slow feedbacks that dominate, including ocean heat uptake, ice sheet growth and decay and associated changes in ocean and atmospheric circulation, changes in atmospheric dust/aerosols, in vegetation and in the movement of carbon in and out of the atmospheric store (PALEOSENS 2012).

Perhaps the most critical and illuminating of the slow feedbacks is the role of natural changes in the long-lived greenhouse gases; carbon dioxide, methane and nitrous oxide. Evidence from the ice cores demonstrates that atmospheric concentrations of CO_2 are tuned not to the sinusoidal changes in orbital forcing, but to the non-linear response of the ice sheets. Ice sheet growth is accentuated by a declining greenhouse effect due to falling CO_2 levels, implying a massive transfer of carbon from the atmosphere into the oceans, perhaps due to dust borne nutrients stimulating planktonic productivity (Mahowald et al. 1999, Jickells et al. 2005, Moore and Braucher 2008, Maher et al. 2010).

The rapid rises of atmospheric CO_2 that accompany the major termination events are more difficult to explain. Some of the CO_2 may represent increased volcanism in response to changes in crustal loading during deglaciation (Huybers and Langmuir 2009; Kutterolf et al. 2013), but model experiments (Roth and Joos 2012) and changes in carbonate ion concentrations and stable carbon isotope ratios in marine archives suggest that it is more likely released from the deep ocean store (Yu et al. 2010). Cheng et al. (2009) have suggested that terminations occur because large isostatically compensated ice sheets are predisposed to collapse, which is triggered by rapidly rising insolation and accentuated by rising CO_2 generated with 'multiple positive feedbacks'. Proposed feedbacks include a shift in atmospheric circulation resulting in enhanced wind-driven upwelling around Antarctica (Cheng et al. 2009) and release of carbon from the Indo Pacific Ocean in response to a weakening of Atlantic meridional overturning circulation (Brovkin et al. 2012). Any release of deep-ocean carbon associated with anthropogenic climate changes would produce a dangerous magnification of greenhouse warming, so the mechanisms that control exchanges between the atmosphere and deep ocean need to be better understood.

Orbital scale and better forecasting

In terms of better forecasting the drivers of climate change, orbital-scale studies are useful because the regular cyclicity allows direct prediction of forcings, and their attendant feedbacks, into the future. In the early Pleistocene the marine records demonstrate that the pacing of ice sheet growth and collapse was dictated mainly by the tilt cycle (41k), but throughout the Quaternary there is a slow background decline in temperature, that is not explained by orbital parameters, and in the late Quaternary glacial cycles are much longer, apparently dominated by eccentricity (100k). The switch in periodicity may represent growth of the Antarctic ice sheet to the extent that it terminated off-shore during interglacials (Raymo et al. 2006). Thus the trend has been towards longer glaciations and short interglacials that last for about 10,000 years. The Holocene is already more than 10,000 years old, so based on comparison with previous glacial cycles we should be descending into another long glaciation. Since global temperatures are rising, and glaciers are melting, that is clearly not the case.

Ten thousand years ago the tilt and precessional cycles were in phase and near maxima, with both providing high summer insolation in the northern hemisphere (Figure 2). Since then they have both been declining. We are now close to the minimum of a precessional cycle, so over the next ten thousand years precession will drive an increase in insolation whilst the longer tilt component continues to decline (Loutre and Berger 2000, Ruddiman 2008). The result is that summer insolation in the north will remain fairly stable for the next ten thousand years and then rise as the tilt cycle reverses. The current low levels of insolation will not be repeated for another 50,000 years.

Figure 2 about here $$

Whilst it is reassuring to know that we are unlikely to slip into another glaciation in the near future, the prediction that the present interglacial is likely to last for more than 50,000 years is startling because over the whole of the Quaternary that has never happened. William Ruddiman has suggested that we have escaped glaciation because early agriculture released both carbon dioxide and methane into the atmosphere, enhancing the greenhouse effect (Ruddiman 2003, 2007, 2013). He argues that in previous interglacials both CO_2 and methane start high but then decline, providing a positive feedback to cooling and the slow decline into a glaciation. In the present interglacial, however, CO_2 started to rise at about 8ka, coincident with the spread of agriculture across Eurasia and widespread deforestation, and methane started to rise at about 5ka, coincident with the rapid expansion of rice production. The ‘early anthropocene’ hypothesis, although focussed on the Holocene, is a direct product of research on measuring and understanding climate change at orbital timescales.

Research on orbital timescales can also help us to predict the future consequences of anthropogenic warming because they provide some constraints on the level of warming that the Earth system can accommodate without disastrous consequences, such as methane hydrate release and runaway melting of the polar ice sheets. Marine Oxygen Isotope stages 5e, 9 and 11 are thought to have been the warmest of the Late Pleistocene (Masson-Delmotte et al. 2010), and since sea levels indicate that the polar ice sheets survived we can assume that global mean temperatures did not reach dangerous levels. The global mean temperatures of Marine Isotope Stages 5e, 9 and 11 must therefore represent some minimum threshold for relatively ‘safe’ global warming. Unfortunately, the estimates of temperature during those stages are conflicting. Oxygen isotope ratios from Dome C in Antarctica, assuming a constant polar amplification, suggest mean global temperatures as much as 2˚C higher than the Holocene maximum, whereas evidence from ocean cores suggest a much smaller difference, of just a fraction of a degree (Hansen and Sato 2012). James Hansen and colleagues have argued that the ice core data from warm interglacials are unreliable because when it is warm enough to reduce summer sea ice cover around Greenland and Antarctica there is more heat flow from ocean to atmosphere and the polar warming is greatly enhanced, so that the oxygen isotope ratios from ice reflect high temperatures locally, but not globally. They argue that the ocean cores give the more reliable indication of global mean temperatures and conclude that there have been no Pleistocene interglacials with mean average temperatures more than 1˚C higher than the Holocene maximum, which we have likely reached or even exceeded as a result of anthropogenic warming. Given this interpretation of the warmth of previous interglacials, the 'safe target' of 2˚C warming, relative to preindustrial temperatures might be disastrous and Earth may be "poised to experience strong amplifying polar feedbacks in response to moderate additional global warming" (Hansen and Sato 2012 p. 25).

Orbital scale and calculating equilibrium sensitivity

Calculating the equilibrium sensitivity of Earth's climate using palaeoclimate data is a deceptively simple procedure. All that is required is an estimate of the change in global mean average annual temperature (in ˚C or K) between two points in time, and an estimate of the change in forcing (in W/m^2) averaged across the Earth's surface.

Sensitivity = Change in temperature ÷ change in forcing

At first glance the glacial cycles revealed in the marine cores, and the last glacial cycle in particular, might seem ideal for this purpose, since the changes in forcing are known from Milankovitch theory and the changes in temperature can be inferred from changes in the $\delta^{18}O$ of foraminifera or, for the Late Quaternary, from other archives including ice cores. In fact that is not true at all, because the concept of climate sensitivity refers to the energy budget of the whole globe, and the two Milankovitch cycles that dominate the spatial and seasonal changes in energy (precession and tilt) do not change the global forcing at all. Changes in eccentricity (100k and 400k cycles) produce a very slight change in the distance from Earth to the Sun, but the changes in total energy received are extremely small. Using the glacial cycles to estimate sensitivity requires the slow responses of the climate system to be treated as the forcing factors (PALAEOSENS 2012).

At orbital timescales, the last glacial cycle provides the best opportunity to use the climate of the distant past to estimate climate sensitivity, because of the large temperature difference between the LGM and the Holocene thermal maximum. There are only three ways to change the temperature of the Earth's surface: change the amount of energy received from the Sun, change Earth's albedo, and thus the amount of solar energy reflected back to space, or change the strength of the greenhouse effect. Although it seems rather counterintuitive, changes in the amount of energy received from the Sun cannot be used on the forcing side of the equilibrium sensitivity equation. This is firstly because the orbital changes do not alter the total amount of energy that Earth receives, so the change averaged across Earth's surface is necessarily zero, but also because ice sheet growth and decay lag the orbital changes (Ruddiman 2006, 2008) so that by the LGM the spatial and seasonal distribution of energy was very similar to today. It is changes in albedo and in the strength of the greenhouse effect that are used as the forcing parameters. The changes in albedo are caused by changes in the extent of ice, sea level, vegetation and in the dust content of the atmosphere (aerosols). The changes in greenhouse effect are caused by changes in the three long-lived greenhouse gases: CO_2, CH_4 and N_2O.

There have been many attempts to use the temperature difference of the LGM/Holocene to estimate climate sensitivity, but comparing and interpreting the results is complicated because of the choices about what to include as a forcing and what to leave as part of the feedbacks. If the results are to be compared directly with the sensitivity estimates produced by model experiments, then it is useful to separate feedbacks into 'slow' and 'fast', with a timescale of 100 years typically defining the separation between them (Charney et al. 1979; PALAEOSENS 2012). The fast, or 'Charney' feedbacks simulated in model experiments are changes in water vapour content, lapse rate, cloud cover and snow and sea-ice albedo. Those are the feedbacks that will determine the fast response of the climate system to future increases in greenhouse gases. Ideally, palaeoclimate studies that aim to calculate a sensitivity value that is directly comparable with model results should aim to include only these feedbacks, so that anything operating over longer timescales is included in the forcing side of the equation. For the LGM the most important slow feedbacks are changes in the long-lived greenhouse gases (CO_2, CH_4 and N_2O), changes in the extent of land ice, changes in vegetation distribution and changes in aerosols (dust). In practice, most palaeoclimate studies, including those using the LGM, do not calculate the forcing effect (in W/m^2) of all of the slow feedbacks. The greenhouse gas concentrations can be taken from the Antarctic ice cores, so they are the easiest to account for and the magnitude of forcing that they exert is well understood. Over orbital timescales the extent of land ice can be estimated using the change in sea level inferred from the long benthic foraminifera $\delta^{18}O$ records, but for the LGM the estimate can be based on the mapped extent of the ice sheets. Changes in vegetation, and their effect on albedo, are more difficult again, but we have much more information for the LGM than for any earlier time-period. The most difficult parameter to quantify is the change in forcing due to changes in atmospheric aerosols. Unfortunately, the LGM was very dusty (Mahowald 1999; Maher et al. 2010), so aerosols could account for 20% of the difference in forcing (Köhler et al. 2010; Rohling et al. 2012; PALAEOSENS 2012).

Very few estimates of sensitivity using the LGM have explicitly included all of these slow feedbacks in the forcing side of the equation, with the result that the sensitivity estimates include both the Charney sensitivity and the missing slow feedbacks. This is one of the reasons that palaeoclimate estimates of sensitivity have varied so widely (PALAEOSENS 2012). Hansen and Sato (2012) for example, treat aerosols as one of the fast feedbacks, rather than as an albedo forcing. Ignoring one or more of the slow feedbacks is convenient of course, because it greatly reduces the uncertainty on the forcing side of the equation and therefore the uncertainty bounds of the estimate of sensitivity, but that does not help if the aim is to compare with the results of model experiments because the uncertainty just reappears in the difference between the palaeoclimate estimate of sensitivity and the Charney sensitivity that is required for the comparison.

A recent attempt at fully quantifying the slow feedback 'forcings' for the LGM, and using them to calculate the Charney sensitivity is Köhler et al. (2010). The best estimates for the forcing parameters (Figure 3) amount to -9.5 Wm^{-2} with an assumed normally distributed uncertainty with a standard deviation of between 0.9 Wm^{-2} and 1.9 Wm^{-2}. The range of estimates for global cooling at the LGM is taken as 5.8 ± 1.4 K, based on the results of Schneider von Deimling et al. (2006), which is somewhat larger than the range used in the PMIP2 (Masson-Delmotte et al. 2006, 2010) experiments (4.6 ± 0.9 K). Rather than calculate sensitivity directly, Köhler et al. (2010) also take into account the accumulating evidence that climate sensitivity is not independent of the climate state by including an uncertain scaling factor of 0.85 with a standard deviation of 0.2. Charney sensitivity is thus estimated as the ratio of temperature change and radiative forcing change divided by the scaling factor. By including the uncertainties on both sides of the equation, and in the scaling factor, they obtain a best estimate for $2xCO_2$ of 2.4 K with a 95% confidence range of 1.4 to 5.2 K. The asymmetry of the uncertainty, despite symmetrical errors in the components of the calculation, is an inevitable consequence of calculation by division with large uncertainty in the denominator (forcing).

Figure 3 about here $$

It has to be concluded that the results of this very detailed and careful analysis of climate sensitivity using the large difference in temperature between the LGM and pre-industrial conditions are rather disappointing. Rather than reducing the unhelpfully large range of uncertainties provided by model estimates, the reasonable range of estimates is even larger than the consensus range of 2 to 4.5 K (IPCC 2007, Knutti and Hegerl 2008) and in fact larger than the range of uncertainties first proposed in the 1970s. Köhler et al. (2010 p.141) conclude that "uncertainties are large if all forcings and uncertainties are considered properly, and the LGM does not provide a strong constraint on sensitivity".

Rohling et al. (2012) used a compilation of global sea surface temperature records spanning at least one Late Pleistocene glacial cycle to calculate sensitivity using global average values for temperature and forcing changes, producing similar results to Köhler et al. (2010). However, they also divide the data into 10˚ latitudinal bands, allowing them to consider spatial variations in the magnitude of forcing and temperature response. The results reveal low sensitivity near the equator but a sharp increase at 20˚ – 30˚ latitude. Sensitivity of the Arctic, at these long timescales, appears to be much lower than that of the Antarctic, perhaps suggesting that the current high sensitivity is due to sea-ice feedbacks, which will cease to operate when all of the Arctic sea ice has melted. This conflicts with the prediction that loss of sea ice will increase the polar amplification (Hanson and Sato 2012).

Whether current and future research on the last glacial cycle, and more specifically on the LGM can help to greatly reduce the uncertainty in estimates of sensitivity remains moot. Through the efforts of several large interdisciplinary projects, as well as individual and group endeavour, we already know much more about the climate and environmental changes of the LGM than any earlier period. Despite that effort, both sides of the sensitivity equation remain very uncertain.

Although we now have a plethora of methods available to measure the climate of the past, on land and in the ocean, estimates using different archives do not necessarily agree and many have some bias in terms of seasonality, so that it remains very difficult to tightly constrain the mean annual temperature of the whole of Earth's surface at the LGM. There have been several recent attempts to estimate the global mean air temperature anomaly of the LGM, but despite being based on similar sources of evidence they not only disagree, in some cases their confidence limits do not overlap (Table 1). Although reconstructing the temperature difference between the Last Glacial Maximum and the Holocene has been a research priority since the 1970s, the real uncertainties remain very large.

Table 1 about here $$

On the forcing side of the equation the problem is even more acute. The changes in greenhouse gas forcing are not the problem, it is the different components of the change in albedo that are most difficult to quantify. Estimates of ice sheet extent at the LGM are unlikely to improve enough to make much difference, and the same may be true of the effects of changes in sea level and vegetation distribution. The largest uncertainty is likely to remain the aerosol effect of dust, and it is difficult to see how that can be resolved. In addition, it now seems that the equilibrium sensitivity of Earth's climate is not stable through time (Caballero and Huber 2013; Kutzbach et al. 2013), so that sensitivity estimates based on the LGM may be an unreliable indication of the real sensitivity that we will face in a warming future.

Orbital scale and testing models

In theory, reconstructions of temperature or other climatic changes on orbital timescales could be used to estimate the relative reliability of climate models on the basis that those that reconstruct most closely the known magnitude and/or pattern of climate change in the past might be most reliable for forecasting future changes. This might provide a way of determining whether models that produce a high equilibrium sensitivity are more or less reliable than those that produce a low equilibrium sensitivity. An obvious target interval is the LGM, because of the wealth of palaeoclimate reconstructions, large difference in temperatures, and estimated changes in forcing due to slow feedbacks. The simple logic is that those models that get closest to the estimated cooling for the LGM will produce the best estimates for future warming under elevated greenhouse gas forcing. However this logic only holds if there is a strong relationship between the magnitude of LGM cooling and the magnitude of future warming produced by the range of models under consideration. In fact, Crucifix (2006) found that for four models included in the PMIP2 LGM experiment there was no correlation between equilibrium climate sensitivity and LGM temperature anomaly. The differences were attributed largely to cloud feedbacks, particularly the non-linear response of subtropical shallow-convective clouds to changes in temperature. He concludes that "global estimates of the LGM temperature only weakly constrain climate sensitivity..." (p. 4).

Hargreaves et al. (2013) have recently used a larger set of seven models, also from the PMIP2 experiment, and confirm no correlation between equilibrium climate sensitivities and global LGM temperature anomalies. They do, however, find a significant relationship in the tropics (20˚S to 30˚N) and use this to estimate sensitivity using two approaches; one based directly on the regression of equilibrium sensitivity on LGM tropical temperature change and one based on a Bayesian Model Averaging approach. Both methods yield rather low estimates of sensitivity and they conclude a best estimate of 2.5˚C for a doubling of CO_2 with a high probability of lying below 4˚C. A similarly low sensitivity value is proposed by Schmittner et al. (2011) who also used temperature differences between the Holocene and LGM to test models, but in this case they used 47 versions of a model of intermediate complexity, each of which produced a different equilibrium sensitivity, within the range 0.3 to 8.3˚C for a doubling of CO_2. The models were tested according to how well they fitted a SST (Margo project) and SAT (Bartlein et al. 2011) data set using a Bayesian framework which takes account of uncertainties in both the models and the proxy data. Their estimated equilibrium sensitivity is low (2.3˚C for $2xCO_2$) and has a very narrow uncertainty range (90% = <1.5˚C).

Considerable caution is required in accepting such low estimates of equilibrium sensitivity and particularly the statement that values of >4˚C are very unlikely. The approach used by Hargreaves et al. (2013) is based on a regression relationship that is effectively defined on the basis of seven points, and the target region was not defined entirely *a priori*, but chosen because it produced the strongest correlation. Schmidt et al. (2014) have combined more models and they also find that significant but weaker correlations are restricted to this region. The models are then tested against an estimated LGM temperature anomaly that is not only much lower than many recent estimates but has confidence limits that fall well short of the best estimates calculated using other methods (Table $). Underestimating the real uncertainty of the climate reconstructions will of course result in underestimates of the uncertainty in the estimates of equilibrium sensitivity. The conclusions of these studies are in direct contradiction to those of Sherwood et al. (2014) who argue that the main reason that different climate models have very different sensitivities is the way that lower tropospheric mixing affects low level cloud production and that such mixing is unrealistically weak in models that have low climate sensitivity. They conclude a most likely value for $2xCO_2$ sensitivity of 4˚C with a lower limit of about 3˚C.

There is clearly potential for using climate changes over orbital timescales, including the temperature difference between the LGM and Holocene, to test the performance of different models or variants of the same model using a perturbed physics approach. However, uncertainties in the proxy data need to be considered very carefully, because underestimating them risks rejection of reasonable models and also translates into underestimates of the uncertainty bounds of equilibrium sensitivity. The range of recent estimates for the temperature difference between the LGM and Holocene (Table $), and often small overlap in the quoted uncertainties, suggest that the real uncertainties are routinely underestimated. Considerable caution is also required when testing models using the climate of the distant past, with very different boundary conditions. It is not necessarily true that the models that perform best in replicating, for example, estimated SST at the LGM will be the best at predicting changes in future air temperatures in the most populated and/or vulnerable parts of the world over the next century.

Sub-Milankovitch scale research on Quaternary climate change

The shortest, and strongest of the orbital cycles influencing past climate is precession, with a cycle of 23k yrs. However, it has long been known that Earth's climate has changed dramatically over shorter timescales than this. The best known events are the rapid and extreme fluctuations in climate that accompanied the transition from the Last Glacial Maximum, with widespread northern hemisphere glaciation and an atmospheric CO_2 concentration of only 190ppm, to the relatively stable warmth of the Holocene.

Early work on pollen suggested that deglacial warming was interrupted by a period of very cold conditions. The early warm phase is generally known as the Bølling/Allerød and the cold phase as the Younger Dryas (pollen of the mountain avens, *Dryas octopetala*, is a key indicator of cold conditions). The Older Dryas is a shorter cold interval that, in some pollen sequences, separated the Bølling and Allerød. In Britain the warm phase is sometimes known as the Windermere interstadial, after a lake in northern England, and the cold phase as the Loch Lomond stadial, since glaciers re-advanced into the Loch Lomond area of Scotland at that time (Pennington 1977). Abrupt warming at the end of the Younger Dryas heralded the beginning of the Holocene interglacial, although some instability remained until the final melting of the large North American ice sheet, the most spectacular event occurring around 8.2ka, when there was an abrupt but short-lived cooling (Alley et al. 1997 Thomas et al. 2007).

The Lateglacial events are so recent, and so extreme, that they are still evident in the landscapes of much of northern Europe. In Wales, for example, cirque glaciers re-formed in the mountains of Snowdonia (Bendle and Glasser 2012) and the Brecon Beacons (Shakesby and Matthews 1993), and the same was true in the English Lake district (McDougall 2001, 2013, Wilson 2002), and in the hills of Ireland. In Scotland it was cold enough for a substantial ice sheet to reform over the western Grampians (Sissons 1979, Golledge 2010, Ballantyne 2012). The remaining Scandinavian ice sheet also re-advanced (Dahl and Nesje 1992, Nesje and Dahl 1993) as did mountain glaciers throughout mainland Europe and elsewhere (Ivy-Ochs et al. 1999, 2007). Countless lakes and peat bogs, throughout Europe, have been cored to reveal the basal transition from relatively inorganic Marine Isotope Stage 2 (MIS2) sediments into the warm organic rich sediments of the Interstadial, the return of cold conditions and onset of the Holocene. No QRA field trip is complete unless the pale layer of the Younger Dryas has been dragged from the depths and laid open to view.

Many of the great characters of the QRA, and of the Quaternary research community more generally, have worked on the Lateglacial. Brian Sissons, of the University of Edinburgh, was a driving force in the early days of the QRA, not just through his own research but also through his influence on students and young colleagues. A list of those who passed through Edinburgh in the 1970s and early 80s reads like a who's-who of the British Quaternary, including Colin Ballantyne, Alastair Dawson, Murray Gray, John Matthews, Rick Shakesby, David Smith and of course the dynamic duo of John Lowe and Mike Walker. Research on the Lateglacial has been a strong theme of Quaternary science for more than forty years and enthusiasm never seems to wane.

One of the best loved characters, and the person who made the most unexpected discovery, however, must be the late Russell Coope, of the University of Birmingham. Russell gave many wonderful talks on the use of beetles (coleoptera) to reconstruct past temperatures, always with a twinkle in his eye as he explained that it is the genitalia that are most useful for identification (Coope 1979). With the help of Martin Atkinson and Keith Briffa, Russell pioneered the 'mutual climatic range' technique (Atkinson et al. 1987) and used it to show that rather than rising slowly, as suggested by the many pollen records, temperatures during warming events increased extremely rapidly, with the transition from glacial to fully interglacial conditions occurring within the span of a single lifetime (Figure 4). That was a brave and bold assertion, but many other lines of evidence now show that he was right.

Figure 4 about here $$

The clearest evidence for very rapid climate changes in the Lateglacial comes from ice cores. The first long core was obtained from Camp Century (Greenland) in 1966 and more recent campaigns include the Greenland Ice-Core Project and the Greenland Ice Sheet Project (Jouzel 2013). The relatively rapid accumulation of ice in Greenland means that only the last glacial cycle is preserved, and the resolution of the record declines with depth, as the annual layers become compressed and distorted (Dahl-Jensen et al. 2013). Accumulation rates are very much slower in Antarctica and deep cores there now extend back for more than 800k yrs. As well as confirming the spectacular nature of the Lateglacial climate changes, the evidence from the ice cores also placed those changes in a longer term context, demonstrating that far from being unique they represent the end of a long sequence of rapid climate fluctuations.

Figure 5 about here $$

The evidence for rapid climate change is particularly clear in the Greenland ice cores where it can be tracked in a wide range of proxies, including snow accumulation rates, methane concentration, dust concentration, salt (Na^+), and water isotopes ($\delta^{18}O$ and δD). Abrupt changes in $\delta^{18}O$ in the ice show 25 rapid warming events since the last interglacial. They were first described by Willi Daansgard and Hans Oeschger and have become known as 'Dansgaard-Oeschger oscillations', or more formally as Greenland interstadials (GI-1 to GI-25) with intervening Greenland stadials (Figure 5). The sub-annual resolution of continuous flow analysis of the ice cores allows even finer division into sub-stages, identified by letters, providing an unrivalled 'event stratigraphy' (Lowe et al. 2008, Blockley et al. 2012). At high latitudes there is a strong linear relationship between local temperature and the water isotopic ratios in snowfall (Dansgaard 1964), and this was originally used to reconstruct temperatures in both Greenland and Antarctica. It has since been shown that although this method works well in Antarctica, changes in the seasonality of snowfall in Greenland are also important, to the extent that temperature changes were underestimated by a factor of 2. A range of other methods are now used to estimate temperature changes in Greenland and to calibrate the water isotope record (Table 2).

Table 2 about here $$.

North Atlantic marine cores, from sites with relatively rapid deposition, also display clear evidence of sub-Milankovitch , millennial scale events recorded in a wide range of proxies. For example, in the Norwegian Sea, the Greenland stadials and interstadials are reflected in changes in magnetic susceptibility, and in the oxygen isotope ratios in both planktonic and benthic foraminifera (Dokken and Jansen 1999). Changing water temperatures are also indicated by changes in species distribution, and the percentage of *Neogloboquadrina pachyderma* (sinistral), a planktonic foram which indicates cold water, has been used to demonstrate the slow cooling during stadials and the aburpt warming of the terminations (Bond et al. 1993). North Atlantic marine cores also contain varying amounts of ice-rafted detritus (IRD), deposited from icebergs, including six episodes of exceptionally high deposition known as Heinrich events, indicating either rapid surging or collapse of ice sheets terminating in the ocean and the release of huge armadas of icebergs into the North Atlantic. These occur at the end of some Greenland stadials and herald the onset of rapid warming.

Although the clearest records of the rapid sub-Milankovitch climate fluctuations come from the high-resolution Greenland ice cores, and their impact on the North Atlantic is clear, where there are archives with sufficiently high resolution they can also be seen very far from these latitudes (Voelker 2002). They are clearly seen in colour changes of marine sediments from the Cariaco Basin and Arabian sea and in sea surface temperature estimates for the Iberian margin based on alkenones (Deplazes et al. 2013). On land, changes in the $\delta^{18}O$ of speleothems (stalagmites) indicate that Greenland interstadials are associated with wetter conditions in China, indicating a stronger East Asian monsoon (Zhao et al. 2010) but drier conditions in the Peruvian Andes, indicating a weak South American summer monsoon and northward shift of the Inter Tropical Convergence Zone (ITCZ: Kanner et al. 2012). Hydrological changes are also suggested by median grain-size fluctuations in some high resolution loess sequences (Liu and Ding1998).

The clearest indication of the global significance of the sub-Milankovitch climate changes seen so clearly in the Greenland ice cores comes from Antarctica. Although snow accumulation rates are much lower in Antarctica, and so the temporal resolution of the records is not as high as in Greenland, there is less dust contamination so the air trapped between the ice crystals can be analysed to produce a continuous record of changes in air chemistry (Luethi et al. 2008), including the long lived greenhouse gases CO_2, Methane and

N_2O. The trapped air is not exactly the same age as the ice because it takes time for it to become completely sealed-off from the atmosphere, and so the temporal resolution of the gas record is lower than that of the ice. Methane is less affected by contamination and can also be measured in Greenland ice, so that rapid changes in methane provide one way to link the records from the two poles. Carefully synchronising the records in this way has revealed that the rapid warming events of the northern hemisphere are mirrored in the south by more muted changes in temperature that are in the opposite direction. During Greenland stadials the evidence from Antarctica suggests slow warming, and the abrupt termination (warming) events in Greenland mark a change to slow cooling in Antarctica (Jouzel and Masson-Delmotte 2010). The 'bi-polar seesaw' suggests very large-scale changes in heat transport between the hemispheres.

Very detailed analysis of the warming events at the start of the Bølling (GS-2 into GI-1e) and Holocene (end of GS1) reveal a clear sequence of events, with decreasing dust concentrations occurring about a decade in advance of a very abrupt (1 to 3 years) change in deuterium excess (based on the difference between the two water isotopes), indicating an abrupt change in the source region for water vapour reaching Greenland. Changes in $\delta^{18}O$ were remarkably abrupt in the earlier event (1-3yrs) but took several decades (~60yrs) at the start of the Holocene. Steffenson et al. (2008) argue that the fall in dust levels represents a northward shift in the ITCZ and abrupt intensification of the Pacific monsoon, wetting the Asian deserts and reducing the dust supply, and then a "complete reorganization of the mid- to high-latitude atmospheric circulation almost from one year to the next" (p. 681). As northern sea ice retreats the rate of snow accumulation increases and local temperatures rise.

The most popular mechanism used to explain the rapid millennial scale changes is abrupt changes in the oceanic heat transport in the North Atlantic; a hypothesis proposed by Wallace Broecker who championed the idea of the 'great ocean conveyor' (Figure 6) which can be switched on and off (Broecker et al. 1985; Broecker 1991; Stocker et al. 1992; Broecker 2010). Loose terminology has caused some problems in the discussion of the mechanisms involved, and the term 'thermo-haline circulation' in particular has been criticized, since the movement and mixing of water is driven mainly by the wind, not directly by the sinking of cold salty water (Wunsch 2002). The critical part of the system is the Atlantic Meridional Overturning Circulation (AMOC), with warm near surface waters moving northwards and then sinking to flow back south at depth as nutrient-poor North Atlantic Deep Water (NADW). When NADW production is low, much colder and more nutrient rich deep waters formed off Antarctica (Antarctic Bottom Water) extend further north. Since the AMOC includes transport of water and thus heat across the equator, weakening it allows southern hemisphere warming, providing a mechanism for the 'bipolar see-saw'.

Figure 6 about here $$

A range of proxies has been used to try to reconstruct changes in the strength of the AMOC and disentangle the link between ocean circulation changes and temperatures in Greenland and Antarctica. The difference in the carbon isotope ratios in planktic and benthic forams, for example, is linked to nutrient content and so changes is response to the amount of mixing, so that when the AMOC is strong the difference is small and it increases as the amount of overturning declines (Duplessy et al. 1989). The Cd/Ca ratio of foram tests is also linked to nutrient content, allowing changes in the dominance of the different bottom waters to be reconstructed (Boyle 1988). Perhaps the most direct link to the strength of ocean currents, however, is the ratio of ^{231}Pa to ^{230}Th. Both isotopes are produced in sea water by the decay of Uranium but Palladium is less easily adsorbed onto settling particles, and so is more likely to be carried away by currents. As the strength of the AMOC declines, therefore, the ratio of ^{231}Pa to ^{230}Th approaches the mean value of seawater (Siddall et al. 2007, Gherardi et al. 2009, Lippold et al. 2011).

There is no doubt that over the last glacial cycle there were large changes in temperature over Greenland, with slow cooling followed by extremely abrupt warming, and that these changes are linked in some way to the behaviour of glaciers and/or sea ice and to the strength of the AMOC. More muted temperature changes in the opposite direction occur in Antarctica. The driving mechanism for these changes, however, remains contentious. In recent years there has been great emphasis on the role of freshwater incursions into the North Atlantic, producing a fresh, low-density surface layer that restricts or prevents deepwater formation and this slows or even stops the AMOC.

Recent work on monitoring the ocean currents in the North Atlantic, however, casts some doubt on the concept of NADW formation as the driver of global climate changes. The upper limb of the AMOC is generally depicted as a surface flow of warm water carrying heat from the tropics which cools, evaporates and becomes increasingly saline en route; eventually sinking off the coast of Greenland (Figure 6). In fact, the movement of surface drifters (Figure 7) reveals that tropical surface waters very rarely enter the North Atlantic; they rotate clockwise and remain within the subtropical gyre (Lozier 2012). Contrary to popular belief, the 'Gulf Stream' of warm surface water does not continue to flow northward along the coast of Europe and Scandinavia. In fact, at 55˚N "poleward heat transport is negligibly small" (Riser and Lozier 2013). Surface waters clearly do not form the upper limb of the AMOC.

Figure 7 about here $$

There is very little evidence to support the idea that local sinking of surface water at high latitudes drives the overturning circulation and the proposed link between the rate of overturning in the North Atlantic and the rate of water and heat transport by the AMOC is tenuous at best. The properties of NADW are not simply a function of the properties of the water at overturning sites. Mixing powered by the wind, by tides, and by eddies produced by the interaction of water bodies with different properties (such as density and velocity) are likely the dominant controls (Wunsch and Ferrari 2004; Lozier 2012).

Sub-Milankovitch scale and understanding climate

In terms of understanding and modelling the way that the climate system works, perhaps the greatest contribution from work on sub-Milankovitch, millennial-scale climate change, and on the Lateglacial in particular, has been the recognition that warming can be extremely rapid. When viewed from the perspective of the Holocene, the events of the Lateglacial appear extreme and since landforms and sediments of this age are abundant it is no surprise that the events have received so much attention. When viewed from the longer term perspective of the last glacial cycle, however, the events of the Lateglacial do not look unusual at all; in fact rapid warming events are common. The critical point, however, is that the large-magnitude excursions are restricted to periods when there are still large continental ice sheets in the northern hemisphere. They do not occur under true interglacial conditions. The last event that fits into this category is the so-called '8.2 event' which was an abrupt cooling of the North Atlantic region driven by the final release of melt-water into Hudson Bay.

The records of orbital-scale climate change clearly demonstrate that Earth's climate system has more than one stable mode, and it swings very rapidly from the glacial state to the interglacial state. The ice core and other records of millennial scale events demonstrate that in the glacial state the ocean-atmosphere-cryosphere system is inherently unstable, and displays rapid and extreme reorganisations of energy that extend across both hemispheres, but with opposite signs. Irrespective of the exact mechanism, the critical point is that these extreme events do not occur during full interglacial conditions without large continental ice sheets. They are not a feature of the world we now live in. A better understanding of the millennial-scale sub-Milankovitch climate changes will undoubtedly lead to better modelling of the Earth system as it functions during glaciations. However, that is not the system that we need to model in order to better predict the climate of the future.

A huge amount of effort has gone into characterising and dating the rapid climate changes of the last glacial cycle, providing a beautifully detailed event stratigraphy. From the perspective of a Quaternary scientist, interested in the climate of the past, this is a fascinating pursuit. From the perspective of this paper, however, dating and characterising those events does not really help us to 'divine the future'. It might even be argued that by focussing so strongly on the ocean as the driver of climate change, and in particular on the sinking of surface waters in the North Atlantic, the research has had something of distorting influence on our understanding of the relative importance of changes in atmospheric and oceanic circulation.

Sub-Milankovitch scale and prediction

Research on sub-Milankovitch climate change has contributed little to our ability to predict future drivers of climate change. The evidence for regular cyclicity is poor, and the argument that the cycles continue into the Holocene, beyond the 8.2 event, is unconvincing. Bond et al. (1997) argued for a pervasive cycle centred around 1470 years extending throughout the last glaciation and the Holocene. However, Wunsch (2000) has demonstrated that the spectral peak centred on 1470 years is a simple alias of the seasonal cycle that arises inevitably when the length of the true year (365.2422 days) is interpolated to an integer multiple of 365 days. The problem occurs even if there are dating errors and natural 'filters' such as diffusion and bioturbation. When the energy associated with this peak is removed, the original records appear virtually unchanged. There is no cyclicity, no signal, just red noise.

If the temperature excursions are driven partly by the influx of cold fresh water into the North Atlantic then there may be some implications for the effects of enhanced melting of Greenland ice or increased river runoff into the Arctic Ocean. However, the mechanistic link between freshwater inputs and the strength of the AMOC is not clear and the nature and likely magnitude of perturbations is so different from those operating under glacial conditions that they are unlikely to provide a realistic analogue. Perhaps the most useful aspect of research on sub-Milankovitch scales, in terms of predicting future climate, is the insights they give into the feedbacks associated with rapid climate changes and the mechanisms that propagate the effects of locally-driven climate changes across large areas. Such information may help us to predict the consequences of the extremely rapid rise in high latitude temperatures that may occur if the loss of sea ice leads to greatly enhanced polar amplification of warming.

Sub-Milankovitch scale, sensitivity and model testing

Despite the high resolution of the records, and the large magnitudes of change, it would be very difficult to use the sub-Milankovitch perturbations to estimate climate sensitivity. The large changes in temperature shown in the Greenland ice cores are at least partially balanced by perturbations in the opposite direction in Antarctica, so the changes in global mean temperature are difficult to estimate. Without a clear driving mechanism the change in forcing cannot be estimated either.

There have been many model experiments attempting to reproduce the large and rapid sub-Milanovitch climate changes, but the system being modelled is that operating under glacial conditions, with large northern hemisphere continental ice sheets. From a modelling perspective this is interesting, but there is little reason to suspect that the models that best reproduce, for example, the response of sea surface temperatures to catastrophic flooding of the North Atlantic with fresh water will necessarily be the best at forecasting the response of air temperature to rising CO_2.

Holocene-scale research on Quaternary climate change

In comparison with the great cycles of glaciations, or the rapid fluctuations of the Lateglacial, the Holocene can appear rather dull. Until about 8000 years ago there was still a substantial remnant of glacial ice in North America and so Earth's climate was still operating in the very unstable 'ice-age mode' of activity. The '8.2 event', caused by the final outburst of meltwater, can be seen as the last of the rapid fluctuations and heralds the beginning of the much more settled 'interglacial mode'. Although they are small in comparison with the more distant past, however, the climate changes that have occurred during the last 8000 years have been large enough to shift vegetation zones and contribute to the collapse of civilizations.

One of the clearest indications that Holocene climate has not remained stable is the past behaviour of glaciers. The size of glaciers is controlled by the balance between winter accumulation and summer ablation, so they can record past changes in both temperature and atmospheric circulation (Nesje and Dahl 1993, Dahl et al. 2003, Nesje et al. 2008). The study of glaciers and their deposits has a long history and there have been many attempts to assess the degree of synchrony in response (Grove 2004, Denton and Karlén 1973), but attempts have been hampered by generally poor dating control, particularly for events before the last millennium. Mayewski et al. (2004) used the available glacial records as a basis for identifying periods of significant rapid climate change and compared those, qualitatively, with globally distributed palaeoclimate records from a wide variety of archives. They concluded that after 8ka there were five periods of rapid climate change, four of which involved polar cooling and tropical drying. Solar variability is invoked as the most likely forcing mechanism. In a much more comprehensive and quantitative review of the evidence, however, Wanner et al. (2008) concluded that “we cannot find any time period for which a rapid or dramatic climatic transition appears even in a majority of the timeseries” (p.1818).

The dominant control on the climate evolution of the Holocene has undoubtedly been the slow orbitally-driven decline in northern hemisphere insolation. Radiation received at 60˚N was at a maximum about 11,000 years ago, and has been declining over the whole of the Holocene. Northern hemisphere temperatures lag the orbital forcing because of the regional cooling effect of remnant ice, and the timing of peak warmth varies spatially (Kaufman et al. 2004). In Northern Scandinavia it reached a maximum at about 6000 BP in the so-called Hypsithermal or Holocene Climatic Optimum. High summer insolation results in more intense monsoons, so that rains were able to penetrate much further inland, resulting for example in an early Holocene ‘green Sahara’ with many large lakes supporting an abundance of wildlife and a human population based around fishing (Drake et al. 2011). As insolation declined the ITCZ, and with it the monsoon rains, shifted southward resulting in more arid conditions in both Africa and Asia (Kutzbach and Otto-Bliesner 1982, Claussen and Gayler 1997, Arbuszewski et al. 2013), with severe impact on human populations (Cullen et al, 2000, de Menocal 2001). Vegetation loss resulted in a positive feedback driven by rising albedo, changes in soil moisture and in East Africa perhaps by convection feedbacks associated with Indian Ocean sea-surface temperatures (Kutzback et al. 1996, deMenocal et al. 2000, Tierney and deMenocal 2013), so that although the underlying forcing was slow, the transition to desert was rapid. At high latitudes the slow cooling is reflected in a retreat of latitudinal and altitudinal tree lines (Barnekow 1999, MacDonald et al. 2000, Gervais et al. 2002), and an increase in sea ice. The effects on the strength of ENSO (Cobb et al. 2013) and on the North Atlantic Oscillation are less certain.

Holocene research and understanding the climate system

The great advantage of the mid to late Holocene relative to earlier periods is that the climate system has settled into the ‘interglacial mode’ and the boundary conditions are essentially the same as those under which we now live. A better understanding of the way that the climate system has operated over this timescale is directly relevant to modelling the way it is likely to behave in future. We also have access to a huge variety of Holocene palaeoclimate archives and proxies, so that we can potentially reconstruct many different aspects of past climate.

The study of Holocene-scale palaeoclimate has a very long history, but the pace of research has accelerated enormously in recent decades, partly reflected in the success of a dedicated journal (*The Holocene*, established and still edited by John Matthews) but also by a great proliferation of Holocene palaeoenvironmental reconstructions published elsewhere. The number of papers returned from a Web of Science search of the topics 'Holocene' and 'climate' has more than doubled over the last two decades (1994-2003, <4,500 and 2004 - 2013 >9,500). Some of the traditions of Holocene research, however, limit our ability to use many published studies to improve the models that are used to characterise the climate system. One of the problems is our tradition of storytelling rather than critical hypothesis testing, or the 'curse of the plausible narrative'.

The aim of many Holocene-scale studies is simply to describe the events of the past at a specific location. This results in many papers that are largely descriptive, in that they present a set of proxy evidence, describe it, and then use it to tell a story about local changes in climate or environment. However, reconstructing past environmental change using proxy archives is often a very complex problem, with more than one way to interpret results and thus more than one potential story to be told. If the story that is chosen is treated as a hypothesis to be tested, or even better if alternative and competing hypotheses are proposed based on the same evidence, then there is potential to make real progress. Too often, however, the critical step of actually finding a way to test reasonable hypotheses is never taken, so that the literature is crammed with plausible but untested narratives, many of which are incompatible with each other. Many climate reconstructions at this temporal scale are also effectively qualitative, with axes indicating 'warmer' or 'drier' rather than providing numbers that can be critically compared and contrasted with other approaches or with models. Where numbers are supplied, they are often based on calibrations based on spatial patterns rather than on measured changes over time, which involves necessary but potentially dangerous assumptions. It is very difficult, for example to find anywhere in the world where there is a spatial gradient in the temperature of one season which is not paralleled by changes in several other parameters, so that isolating individual environmental controls on potential proxies using space-for-time substitution is very difficult, as is realistically defining uncertainty.

There is certainly great potential for Holocene scale studies to contribute to a better and more quantitative understanding of the way our climate system works, and substantial progress is already being made. However, realising our full potential will require a change in emphasis away from storytelling. Posing very different interpretations of past events, based on different proxies, and just leaving the different alternative explanations hanging in the literature as competing narratives impedes progress. Palaeoclimate data cannot, for example, be used to test models if they do not provide a clear and unambiguous picture of past events.

Rather than passively accepting many different narratives, a more productive approach might be to embrace the conflict between competing hypotheses and critically test them. Conflict between hypotheses need not of course equate to conflict between individuals or research groups; good scientists are as critical of their own work as they are of that of others. There is nothing to be gained from protecting hypotheses, no matter how nice the story. In science, you must seek to 'murder your darlings' (Haines-Young and Petch 1986).

Despite the difficulties of summarizing the available Holocene palaeoclimate literature, there have been some excellent regional syntheses that provide a coherent picture of changes that can be used to test and improve model performance. The Arctic provides a particularly useful region for such studies because of the wealth of palaeoclimate data, both on and off-shore, and because of the importance of positive feedbacks that are likely to amplify future warming. The palaeoclimate evidence for the Arctic has been critically summarized by Bigelow et al. (2003) and Kaufman et al. (2004). Previous attempts at modelling the climate of the region, however, have either concentrated on a particular time-slice or used low-resolution simplified models (Earth Models of Intermediate Complexity (EMICs), e.g Weber 2001, Crucifix et al. 2002).

Applying models to a single time-slice, commonly 6000 BP (e.g. An et al. 2014) , is problematic because it assumes equilibrium with the boundary conditions at that time, which may be unrealistic under conditions when the forcing is constantly changing. EMICs allow the transient response to be explored but there are important limitations of spatial scale and/or over-simplification of important components of the atmosphere and oceans as well as usually lacking realistic treatment of sea ice or vegetation feedbacks. The transient Holocene simulation of Renssen et al. (2005), however, used a much more sophisticated model that included improved simulation of Hadley circulation, sophisticated sea ice behaviour, a land surface scheme that takes account of the heat capacity of the soil and a dynamic vegetation model simulating the behaviour of trees, grass and bare soil. Although the model was run using only the changes in orbital forcing and greenhouse gases, with no perturbations of solar irradiance or major volcanic eruptions, the results compared very well with the best available proxy climate evidence, suggesting that "most of the long-term trends at high northern latitudes can be explained by the response of the coupled system to orbital and greenhouse forcing" (p.39). One of the striking outcomes of this research is that although the forcing is smooth, there is a marked increase in inter-annual variability, especially in winter over the last 4ka, interpreted as a response to interaction between sea ice and deep convection. This large internally generated variability, as a response to a smooth decline in forcing, has important implications for interpreting high resolution records of regional climate change and also for understanding how the combination of large internally generated perturbations and short term external (e.g. volcanic) forcing can push the system beyond critical thresholds. Studies like this have much to offer both the modelling and palaeoclimate communities.

Holocene climate and better forecasting

Were it true, as Bond proposed, that sub-orbital scale changes in climate extended through the last cold stage and the Holocene, including the 8.2 event and the 'Little Ice Age', it would furnish us with a powerful tool for predicting changes in the future. The reality is, however, that there are no such regular cycles and the spectral peak on which the theory rests has been shown to be an artefact of the method used to collate the data. Rather than accept this, however, the literature continues to be peppered with attempts to explain the Holocene ‘Bond cycles’. Undeterred by the clear refutation of the argument for regular cyclicity, many authors have simply amended the terminology so that the cycles are described as ‘quasi-periodic’.

The term ‘quasi-periodic’ is so vague that it effectively becomes a protection strategy used to prevent a hypothesis from being critically tested. The climate of the Earth is a complex system and so even in the absence of any external forcing it would not be expected to evolve over time as pure white noise. Some persistence (redness) is inevitable and so the time-series behaviour of natural climate is characterised by a broad band of variability rather than by sharp peaks. The continuing search for a unifying explanation for a ‘quasi-periodic’ sequence of events over the Holocene is a distraction, ignoring the fact that most ‘events’ are not synchronous or uniform across space. On the contrary, it is <u>differences</u> in the nature and magnitude of response across space and over time that will provide the clues to the drivers of Holocene climate changes.

Perhaps the most common proposed explanation for fluctuations in climate seen in proxy records over Holocene timescales is solar variability. Determining whether there is a strong link between the variable output of energy from the Sun and Earth’s climate is critically important. Rising Total Solar Irradiance (TSI) since the Maunder Minimum of the late 17th and early 18th centuries (Luterbacher et al. 2001) could explain some of the warming seen in instrumental records, and if that is true it implies that the impact of changes in greenhouse gases has been exaggerated. Current models used to predict future climate do not include any mechanism that would allow very small changes in TSI to be magnified sufficiently to produce the proposed links (Feulner and Rahmstorf 2010). If such a mechanism is missing then the models are seriously flawed and since solar irradiance is predicted to fall over the next few decades (Roth and Joos 2012), predicted rates of temperature rise may be alarmist.

Quantifying the impact of past changes in TSI on climate is surely one of the most critical questions facing Quaternary scientists and the implications for understanding and modelling the likely impact of natural variability and greenhouse gas forcing on future climate change are profound. However, progress has been hampered to some extent by a lack of rigour in comparing palaeoclimate records with estimates of TSI.

One critical problem is that we do not actually have long records of changes in TSI. Measurements of total energy received at the top of the atmosphere are only available for the satellite era. For the last 400 years, since the invention of the telescope, we have some records of sunspot activity and it is generally assumed that fewer sunspots implies lower TSI, on the basis that the reduced energy caused by the darker spots is outweighed by the brighter faculae that surround them. However, the satellite data cast doubt on the assumed strong correlation of sunspot numbers and total irradiance. During the last solar cycle irradiance was as high as the previous two cycles, even though the sunspot number was 20-30% lower (Fröhlich and Lean 2004; Lean 2005; National Research Council 2006).

Over Holocene timescales we have to rely on two proxies that are linked only indirectly to Total Solar Irradiance. Changes in the strength of the solar wind influence the degree to which Earth is shielded from cosmic radiation and so there is an indirect negative link between solar activity and the production of cosmogenic nuclides, including ^{10}Be and ^{14}C in Earth's atmosphere. The ^{10}Be tends to adhere to aerosols and is thus deposited by rain or snow and radiocarbon is easily oxidised to CO_2 and joins the carbon cycle. Continuous records of ^{10}Be concentrations can be obtained from ice cores and shorter local records from lake sediments. Variations in ^{14}C are available from independently (dendrochronologically) dated tree rings. Given the mode of deposition, neither record can be considered completely independent of climate. Changes in circulation can influence aerosol trajectories and precipitation patterns so that ^{10}Be from a single location cannot be considered a reliable indication of global production rates, as evidenced by differences between ice cores. Fluxes in carbon between the biosphere, soils and the shallow ocean also respond to changes in climate. Although records of the two isotopes are not identical, they are similar enough to suggest that the dominant common signal is the production rate. However, the link between the abundance of the cosmogenic isotopes and the magnitude of change in forcing at the Earth surface is effectively un-quantified. This means that we have a good indication of times in the past when cosmogenic isotope production was unusually high or low, inferring differences in solar activity, but no widely agreed way to convert isotope abundance into changes in forcing.

Too often the proposed link between palaeoclimate records and solar forcing on Holocene timescales is based on nothing more than speculation and wishful thinking. The problem is most acute when the palaeoclimate reconstruction is based on low-resolution proxy data and the dating control is sufficiently insecure to allow considerable uncertainty in the age of peaks and troughs. In such cases a bit of stretching and squeezing, within the realistic boundaries of uncertainty, will certainly provide a reasonable visual match with at least some of the peaks and troughs in the smoothed isotopic records. Trying a range of band-pass filters also helps.

A related problem arises where the link to solar variability is based on comparing time-series of radiocarbon dated events with one of the isotopic records (^{10}Be or ^{14}C). Perhaps the best example is the work of Magny (1993, 2004, 2013) who has shown a remarkably good match between a record of lake level changes from France and Switzerland and the ^{14}C curve, with higher lake levels coinciding with periods of low solar activity (Figure 8). However, the lake level record consists mainly of a histogram of scores based on the number of times that the uncertainty ranges of a set of radiocarbon dates falls into any given 50yr period, and so has much in common with a cumulative probability density function or CPF (Bleicher 2013a, 2013b). The problem is that the uncertainty range of a radiocarbon date is strongly controlled by where it falls on the calibration curve, and the calibration curve is identical to the ^{14}C curve used as a proxy for solar activity. In times of low solar activity, radiocarbon dates have larger uncertainty ranges and will therefore produce a wider CPF and also 'score' over more 50yr time slots. The result is that any collection of radiocarbon dates, even if they are completely randomly or perfectly evenly spaced over real time, will produce a CPF that looks a bit like the ^{14}C record. This is clearly demonstrated by Bleicher (2013 p.758) who generated CPFs based on even and randomly spaced dates and produced better correlations with the ^{14}C curve that those obtained using the real data of Magny (2004), concluding that "in fact every CPF from any data set is likely to produce significant correlations with the ^{14}C residual curve, because this record is in fact at the heart of a CPF's construction". Essentially the same argument has been used by Chiverrell et al. (2011a, 2011b) to question the reconstruction of UK fluvial activity produced by Macklin et al. (2010, 2011). Substituting ^{10}Be records for the ^{14}C curve does not help because the two are linked via production rates.

Figure 8 about here $$

The role of solar forcing in Holocene climate change remains very uncertain, but that is not to say that there is no supporting evidence. On the contrary there are several high resolution well dated proxy data series, particularly those that record changes in circulation rather than temperature, where the comparison with solar forcing proxies is difficult to ignore (Bond et al. 2001, Neff et al. 2001, Blaauw et al. 2004, Hughes et al. 2006, Zhang et al. 2008, Nichols and Huang 2012, Engels and van Geel 2012). Rather than focussing so much on apparent correlations between proxies for palaeoclimate and solar forcing, a more profitable approach might be to set out to specifically test hypotheses that include some mechanism that might magnify the very small changes in TSI. This approach is most powerful when the underlying hypothesis makes clear predictions in terms of the spatial pattern and seasonality of specific climate parameters. A recent example is the work of Martin-Puertas et al. (2012) who tested the hypothesis that it is changes in the UV part of the spectrum, which are much larger than the changes in TSI, that indirectly influence climate through changing tropospheric ozone production, resulting in shifts of the mid-latitude storm tracks (Haigh 1994, 1996, Gray et al. 2010, van Geel and Ziegler 2013). In the northern hemisphere low solar activity is predicted to cause a southward shift, with pressure patterns similar to the negative phase of the winter NAO. They studied annually laminated sediments in Lake Meerfelder Maar, Germany, where diatom productivity is limited by phosphorous availability. Strengthened winds in late winter and early spring encourage upward mixing of nutrients so diatomaceous varve thickness provides a proxy for seasonal wind strength. They show that the abrupt increase in varve thickness at 2800BP, which lasts for 200 years, is synchronous with the start of the 'Homeric minimum' in solar activity.

Holocene climate, global Sensitivity and testing GCM models

If the large change in global temperature between the LGM and Holocene cannot improve current estimates of global climate sensitivity then it seems very unlikely that the much smaller changes that have occurred over the Holocene will be useful. As with the LGM example, the orbitally-driven decline in northern hemisphere insolation cannot be used on the forcing side of the equation because averaged over the whole year and the whole globe the forcing change sums to zero.

The mid Holocene (6ka) has long been used as one of the time-slices in the PMIP experiments. The main difference in forcing, relative to preindustrial conditions, is higher summer insolation in the northern hemisphere, which palaeoclimate evidence suggests resulted in a decline in Arctic sea ice and a shifted ITCZ and thus strengthened monsoons. In PMIP2, models were compared with a range of palaeoenvironmental reconstructions and although they reproduced the large-scale patterns of climate change quite well, they tended to underestimate the magnitude of regional changes (Braconnot et al. 2012). For example, the summer monsoon precipitation over Africa and the Asian monsoon regions are underestimated by 20-50% and the ability of the models to quantify vegetation feedbacks is also uncertain (Braconnot et al. 2007, 2012, Zhao and Harrison 2012).

Although the PMIP experiments have doubtless been useful for comparing the ability of models to perform beyond the range of climate over which they were calibrated, it seems that little has so far been achieved in terms of improving the ability of the models to predict the future with less uncertainty. This is partly because the models used to reconstruct the past are rarely identical to those used to predict the future (Schmidt et al. 2014). The mid Holocene time slice is also being used in PMIP3, however, and for the first time also as part of CMIP5. This means that precisely the same models will be used to reconstruct the past and to predict the future under a range of scenarios. If it can be shown that the behaviour of models run for the mid Holocene is strongly correlated to the way they behave when run into the future then there is potential for using the mid Holocene fit with palaeoclimate to select a sub-set of models that are most realistic. Few complete model runs are available to date, but Schmidt et al. (2014) report some promising results for reconstructions and future predictions of South American rainfall and Arctic sea ice changes.

Climate change of the last two thousand years

From the perspective of using the climate of the past to constrain predictions of the climate of the future, the last one to two thousand years holds perhaps the most promise. At these timescales we have access to many annually-resolved proxies, recording a wide variety of climatic parameters, and they can be calibrated and verified by direct comparison with measured instrumental climate data (Jones et al. 2004, 2009). This allows us to reconstruct not just the direction of changes in climate in the past, but to provide numerical estimates with realistically defined uncertainties. There are also many other proxies that are not annually resolved and which are not usually calibrated using instrumental data, relying for example on training sets based on spatial patterns rather than changes over time (Birks 1998, Lotter et al. 1997, Brooks and Birks 2001, Seppa et al. 2004). These need to be treated with more caution, and the tradition of storytelling certainly extends into this time period, particularly for the more complex proxies with a variety of potential influencing factors. However, for all proxies over this timescale there is strong potential for real critical testing of hypotheses by, for example, comparison of proposed climate reconstructions with long instrumental data series, historical evidence or with annually resolved proxies where strong calibration and verification criteria can be demonstrated.

The availability of long instrumental data series, that have been properly checked and homogenized, underpins all of the work that is done over this timescale. Just a few years ago much of that data was difficult to access, but through the efforts of many groups and individuals, the data are now much more widely and freely available. In the United States they have always embraced the concept of public ownership of data collected using the public purse (Smith et al. 2008), but in Europe and elsewhere obtaining free access to the data that society paid to collect has been a real battle. The Climate Research Unit at the University of East Anglia have been at the forefront of those efforts and they were able to circumvent some of the legal issues by converting the raw station data into global gridded products, which are now very widely used (Jones and Moberg 2003, Mitchell and Jones 2005, Harris et al. 2014). The Dutch meteorological organisation have also played their part, particularly by hosting the powerful 'KNMI Climate Explorer' web application (Trouet and van Oldenborgh 2013).

Although most instrumental series only cover the last 100 to 150 years, over which time measurement protocols are reasonably uniform, there are much longer series available from many locations extending right back to the invention of the relevant instruments (Camuffo et al. 2014). The thermometer, for example, was invented and further developed in Padua, (Camuffo and Bertolin 2012a, 2012b) and the Celsius scale proposed by Anders Celsius working in Uppsala, so there are very long early instrumental series available for Italy and Sweden, as well as the UK, The Netherlands and elsewhere (Manley 1974, Parker et al. 1992, Bergstrom and Moberg 2002, Camuffo and Jones 2002, Maugeri et al. 2002, Yan et al. 2002, Alcoforado et al. 2012). Measurements taken by early instruments cannot be compared directly with more recent measurements, however, because of differences in design and in measurement procedures. Early thermometers, for example were not sealed and so the measurements need to be corrected for changes in atmospheric pressure. Even much later they were sufficiently precious to either be kept indoors, in an unheated room, or mounted high on walls without proper shielding from the Sun (Boehm et al. 2010, Camuffo 2002). Calibrating and checking early instrumental data is a very difficult task, requiring the skills of both historians and physical scientists, but several projects have focused on this and although early instrumental series still need to be treated with some caution, the long series now available for several regions are an enormous asset to those working with other proxies.

For the last few hundred years there is little doubt that the most powerful proxies of all are historical documents (Shabalova and van Engelen 2003, Brázdil et al. 2005). These include diaries containing specific climate-related information recorded by individuals, but even more useful are the rather dull accounts of production and trade that are often recorded over generations using a standardized practice. Recent examples include the use of tax records of ship sailing dates to reconstruct winter/spring temperature, based on the number of days when Stockholm harbour was frozen (Leijonhufvud et al. 2010), and a reconstruction of April/May temperatures back to 1693 based on the dates of ice melt on a river in northern Finland (Loader et al. 2011). Similar studies have used information on phenology or crop quantity, quality or harvest dates (de Cortazar-Atauri et al. 2010, Kiss et al. 2011, Garnier et al. 2011, Wetter and Pfister 2011, Mozny et al. 2012, Ge et al. 2014). Ship's logs are a particularly rich source of information (Kuettel et al. 2010, Brohan et al. 2012, Ward and Wheeler 2012). Where historical documents indicate the extremity of past climate conditions without quantifying it a quantitative reconstruction can nevertheless be produced by scoring the evidence on a 7-point scale of -3 to +3 (e.g. very cold or wet to very warm or dry). This simple but powerful method has been used to reconstruct the Central European temperature of every month and season for the last 500 years (Dobrovolný et al. 2010), complete with calibration and verification statistics and clearly defined uncertainties. The same method could undoubtedly be used to reconstruct other climate parameters such as rainfall amount, sunshine parameters and perhaps to reconstruct past changes in the dominant patterns of circulation. Applying the methods of historical climatology to the first half of the last millennium is more challenging and data sources are much less abundant, but progress is being made (Pfister et al. 1996, 1998, Glaser and Riemann 2009, Pribyl et al. 2012).

Of the annually-resolved proxies, tree rings have been most widely used. Where there are clear seasons, the width of tree rings can be measured and cross-dated between living trees, standing or fallen dead wood, building timbers and logs dragged from lakes, mires or river sediments (Fritts 1976). When the dating is done properly it provides a precise annual chronology (Baillie and Pilcher 1973). Unfortunately ring widths only respond strongly to climate when it becomes growth limiting, so the most useful trees come from high latitudes or altitudes, which is an important source of potential bias in large-scale reconstructions. Tree ring density, measured by X-ray analysis of chemically-treated wood samples cut to a known thickness (Schweingruber et al. 1979, 1988) provides a more powerful proxy in many conifers, but the high cost of specialised equipment limits its application. A related but much less expensive method based on scanning wood samples using reflected light is showing promise (McCarroll et al. 2002, 2011, Campbell et al. 2007, Bjorklund et al. 2014). A problem with these tree growth proxies is that they display long-term trends related to tree age rather than climate, and removing these risks also removing some of the low-frequency climate signal (Cook et al. 1995). There are some sophisticated methods available to deal with this so called 'segment length curse' (Briffa et al. 2004, Melvin and Briffa 2008) but an alternative approach is to use the stable isotopic composition of the wood cellulose as a set of climate proxies which generally do not appear to suffer from difficult age effects (Gagen et al. 2007, 2008, Young et al. 2011a). Isotope ratio mass spectrometers are now largely automated and recent advances in sample preparation and measurement (Young et al. 2011b, Gagen et al. 2012, Woodley et al. 2012a 2012b) make tree ring stable isotope measurement relatively fast and cost effective. It may even be possible to extract very strong climate signals (Figure 9) from trees that are not growing under climatic stress (Young et al. 2012, Young et al. In revision), which would allow tree ring chronologies originally compiled for archaeological dating to be re-sampled to produce very long climate reconstructions for mid-latitude regions that are highly populated.

Figure 9 about here $$

In the tropics many trees do not have clear annual rings, so that dendroclimatology is much more challenging, though not impossible (Robertson et al. 2004, D'Arrigo et al. 2006, Buckley et al. 2007). Large tropical corals produce annual layers and SST reconstructions can be produced using oxygen isotopes or Sr/Ca ratios which are less sensitive to salinity changes. The records produced so far generally extend for only a few hundred years and are poorly replicated, although fragmentary $\delta^{18}O$ records extending back to AD 900 from Palmyra atoll in the mid tropical Pacific have been used to reconstruct changes in ENSO activity (Cobb et al. 2003). Annually banded mollusc shells are also being used to produce long chronologies, using methods borrowed from dendrochronology (Hudson et al. 1976, Scourse et al. 2006, Butler et al. 2009, 2013, Schoene 2013), and climate reconstructions based on isotopic and chemical proxies from this archive can be expected to become increasingly important.

Many of the proxies obtained from peat cores, lakes and shallow marine sediments to reconstruct past climate over Holocene timescales can also be used at higher resolution over the last one or two millennia, providing the possibility to calibrate or at least verify them using instrumental climate data. When used to reconstruct summer temperature at high latitudes or altitudes these proxies cannot realistically compete with well replicated, calibrated and verified tree ring studies and efforts might be more profitably aimed at proxies that record other seasons or other climate variables. Proxies that record temperature outside of the summer season are particularly valuable (Blass et al. 2007, De Jong et al. 2013), as are those that can point to past changes in precipitation, storminess and circulation patterns. Speleothems, which have produced spectacular results over sub-Milankovitch and orbital timescales (Fairchild et al. 2006, Fleitmann et al. 2007, Wang et al. 2008), generally perform less well in reconstructing the high resolution climate changes of the last millennium, though there are exceptions (Mangini et al. 2005). Water isotopes in high latitude ice caps can be used to reconstruct past temperature, but changes in seasonality of snowfall are a serious source of error (Sodemann et al. 2008). Isotope records from low latitude ice cores are poorly calibrated and the climate signals are difficult to define (Vimeux et al. 2009, Jones et al. 2009).

The large and growing volume of work on reconstructing the climate of the last one or two millennia (Jones et al. 2009, Bradley 2014) still provides only a partial picture of climate evolution that is strongly biased by season (towards the summer) and by location (N Hemisphere, high latitudes and altitudes). Despite this limitation there has been a huge emphasis in recent years on drawing together the available evidence to produce either Northern Hemisphere or global reconstructions of mean annual temperature (Mann et al. 1998, 1999, 2008, Moberg et al. 2005, Juckes et al. 2007, Ljungqvist 2010, Christiansen and Ljungqvist 2011, 2012). There has been intense debate over the statistical methods that are used to achieve this and different authors, often using very similar data sets, have produced very different reconstructions, particularly with regard to the amplitude of past climate changes. Despite continued efforts we seem little closer to agreement on reconstructing the evolution or amplitude of mean annual temperature change at a hemispheric or global scale over the last millennium.

Of course it would be interesting and useful to reconstruct the mean annual temperature of the whole globe for the last one or two millennia, but our ability to do that with sufficient confidence using the available proxy evidence is seriously in doubt. A major problem is the seasonal and spatial bias and very variable strength of the available proxy data. Whatever statistical method is used to combine them it is inevitable that the signals will be weakened just because of the smoothing effect of averaging over all seasons and then over space. There is also the added problem that averaging across space assumes that the signal of interest is synchronous, which may not be true for many of the most important climate variations over the last millennium. The Little Ice Age, for example is certainly not characterized by a synchronous change in temperature even across Europe. If important climate changes are driven by changes in circulation, then we would not expect the nature or magnitude of response to be the same in all places or to occur at the same time. On the contrary changes in circulation, whether externally forced or not, might be expected to produce changes that propagate rather as a wave, moving through time and over space.

A useful analogy is a tsunami, which if recorded at a single place, as you cling to your house perhaps, would be seen as first a dramatic fall in sea level, then an enormous wave, followed by a devastating backwash. If the same event is viewed as the synchronous average of water level at several places on a transect along a tidal river then it becomes averaged out and appears as no more than a minor event, simply because the wave hits different places at different times. If averaged over space and time by taking average water levels over that week or month, then the tsunamis is lost in the noise of the time-series. By averaging over space and time we risk missing the most important events in every region just because they did not involve changes in the same climate parameters everywhere and at the same time.

The importance of capturing regional differences in the past behaviour of climate, and of moving beyond mean annual temperature, has been recognized in recent years by the PAGES community who have launched a major programme of research aimed at reconstructing temperature and precipitation, for whatever seasons are appropriate given the available evidence, for continental-scale regions (Figure 10). The first major outcome is a set of 'PAGES 2K' temperature reconstructions, of varying length, for the Arctic, N. America, Europe, Asia, South America, Australasia and Antarctica (PAGES 2k 2013). The available evidence for Africa (Nicholson 2013) was considered too sparse. The records confirm the general picture of slowly declining temperatures until the late 19th century and modern warming. Variability at shorter timescales shows little coherence globally apart from cool conditions between AD 1580 and 1880. There is little evidence to support the concept of a globally synchronous 'Medieval Warm Period' or 'Little Ice Age'.

Figure 10 about here $$

Although these are the best large–scale reconstructions available to date, they can only be considered as a starting point since they combine only a tiny proportion of the proxy evidence that is available. The European reconstruction, for example, comprises only ten data sets: one documentary and nine tree ring density and there is a strong high latitude/altitude bias. The

Asian reconstruction is based entirely on tree rings, both ring widths and densities. It is also questionable whether the continental scale reconstructions are the most appropriate. In Europe, for example, summer temperatures in Fennoscandia and southern Europe do not vary in parallel and indeed are often anti-correlated, so averaging proxy data from these regions may simply remove climatic information. Reconstructions for large regions where climate is more homogeneous would perhaps be more valuable, particularly if the results are to be used to test the performance of GCMs.

Last 2ka and understanding climate

Climate reconstructions over this timescale are certainly useful for improving understanding of the way the climate system works and for improving models. By effectively extending instrumental series using annually or near annually resolved proxies, albeit with larger uncertainties, we can provide a longer term perspective on natural climate variability that can be contrasted with the behaviour of the system after industrialisation. The wide and growing variety of proxies available allows us to produce a wide range of climate parameters (Jones et al. 2009). Summer temperatures are perhaps easiest to reproduce, using tree ring widths and densities for example, though the statistical treatment of these proxies often leads to a loss of low-frequency information that needs to be considered very carefully (Cook et al. 1995, Esper et al. 2002). In areas where trees experience moisture stress they also allow reconstructions of drought conditions (Cook et al. 2007), which in many areas are far more important that changes in temperature. In the moist mid-latitudes reconstructions of past rainfall based on tree rings can be based on rather weak calibration and verification statistics and they need to interpreted with caution (McCarroll et al. In Press).

One of the largest sources of uncertainty in current GCMs, and a major reason for the large differences in model sensitivity to rising greenhouse gas emissions is the way that they deal with clouds (Sherwood 2014). The relationship between temperature change, cloud type and amount of radiation received at the ground surface is currently difficult to model and existing instrumental data on sunshine hours or cloud cover are inadequate for model testing. It has recently been argued that carbon isotopes in tree rings carry a sunshine signal because in areas with limited moisture stress the dominant control on isotopic fractionation in the leaf is the rate of photosynthesis, which is controlled mainly by sunshine (photon flux) rather than air temperature (McCarroll and Loader 2004, McCarroll et al. 2011). In northern Norway, for example it has been shown that when summer temperatures and sunshine hours diverge, the tree ring isotope values track sunshine rather than temperature (Young et al. 2010). Long sunshine/cloud cover reconstructions have now been produced for northern Fennoscandia (Gagen et al. 2011, Young et al. 2012, Loader et al. 2013) and the eastern Alps (Hafner et al. 2014) and there is potential for testing the method in many other regions, though the scarcity of reliable instrumental data for local calibration is a problem.

Perhaps the greatest potential for using proxy data to improve our understanding of the climate system is by reconstructing past changes in atmospheric circulation, at least for some regions of the world (Luterbacher et al. 2010). This is much more challenging than reconstructing summer temperature and will require collaboration between many branches of palaeoclimatology. The historical climatologists, as usual, have the richest sources of information, at least for the last few hundred years, and can produce reconstructions of, for example, air pressure, wind direction and speed, rainfall amount and seasonality, and storm frequency. These can be combined to reproduce pressure patterns and changes in the position of seasonal storm tracks. Many archives, including lakes, peat, speleothems and trees yield proxies that record changes in the water isotopes and these could carry a much longer record of past changes in the isotopic ratios of precipitation, which are linked to changes in circulation (Young et al. In revision). The proxy-based reconstructions could then be compared with models that predict the impact of, for example, changes in circulation forced by periods of low solar radiation or by large volcanic eruptions. Some 'isotope enabled' models now have the ability to reconstruct directly the changes in water isotopes in precipitation, though the accuracy with which they do this is difficult to assess because of the very short and discontinuous instrumental records (Loader et al. In press).

Last 2ka and better prediction

In terms of improving predictions of the future drivers of climate change, one of the most challenging problems for both palaeoclimatologists and for future climate modelling is understanding the natural 'modes of variability' of the coupled ocean-atmosphere system. In particular, the fluctuations in pressure and sea surface temperature across the equatorial Pacific, which characterize the El Niño Southern Oscillation (ENSO) are responsible for shifts in atmospheric circulation that have disastrous consequences regionally and which influence climate globally. There is serious concern that anthropogenic climate change may have caused an increase in ENSO variability over recent decades (Masson-Delmotte et al. 2013), but the instrumental records are too short to place this in the context of longer-term changes in frequency, magnitude or duration. McGregor et al. (2013) identify 14 attempts to use proxy evidence to extend the record of ENSO variability, mostly just for the last few hundred years, using either coral or lake records from the tropical Pacific or using proxies from more distant locations that are teleconnected to ENSO via, for example, changes in precipitation. Combining estimates of ENSO is complicated by time-varying dating uncertainties, because even small age offsets will cause an averaged record to lose variance, and since uncertainty increases back in time that would result in a spurious increase in variance over time. McGregor solved this very simply by combining estimates of variance over a moving 30yr window rather than combing raw data and they conclude that "ENSO variance for any 30yr period during the interval 1590-1880 was considerably lower than that observed during 1979-2009. Using a much longer, albeit fragmentary record of oxygen isotope ratios from modern and fossil corals from tropical Pacific islands Cobb et al. (2013) also conclude that twentieth century ENSO variability is anomalously high.

From a European perspective the most important mode of climatic variability is usually considered to be the North Atlantic Oscillation, defined as the pressure difference between Iceland and the Azores. The winter expression of this mode is much stronger than that in summer. The NAO index has been extended back to 1692 using early instrumental measurements of air pressure (Cornes et al. 2012), and relevant pressure fields back to 1750 using information from ship's logs. Attempts at reconstructing longer series using natural proxy evidence have proven problematic, however, mainly because few natural archives record conditions in winter. The reconstruction by Trouet et al. (2009), for example, which suggests a persistent positive phase during the Mediaeval Climate Anomaly is based only on the difference between two proxy series from a Scottish speleothem and Moroccan trees. They propose that "The persistent positive phase reconstructed for the MCA appears to be associated with prevailing La Niña–like conditions possibly initiated by enhanced solar irradiance and/or reduced volcanic activity and amplified and prolonged by enhanced AMOC. The relaxation from this particular ocean-atmosphere state into the LIA appears to be globally contemporaneous and suggests a notable and persistent reorganization of large-scale oceanic and atmospheric circulation patterns."

This view of the transition from an 'anomalous' Mediaeval into the European Little Ice Age being a product of solar irradiance driven changes in the Pacific being teleconnected north by changes in the strength of the AMOC owes more to popular paradigm than hard evidence. An alternative explanation is that there was nothing particularly unusual about Medieval climate, it just appears warm from our perspective because we view it across the unusually cold conditions of the European Little Ice Age. There is growing evidence that it was volcanism that triggered the cold conditions of the Little Ice Age and that the feedbacks, just as in previous cold episodes, are dominated by changes in albedo, particularly sea ice, at high latitudes (Zhong et al. 2011, Miller et al. 2012, McCarroll et al. 2013).

Last 2ka, climate sensitivity and model testing

This is the only time period where we have very high quality, annually resolved data for a variety of different climate parameters for many areas of the globe. To use these rich data to mathematically estimate global climate sensitivity, however, we would have to combine them into an estimate of mean annual global temperature. That would effectively mean degrading them by smearing out the information on seasonal and spatial differences in temperature and either ignoring other climatic variables or mixing them in a multivariate analysis in the hope that some temperature information could be squeezed out. We would still be left with the problem of calculating the changes in forcing, again averaged over the whole globe. The result of such an analysis will be an estimate of sensitivity that has a very large uncertainty. Attacking the problem from a mathematical and therefore global perspective is, however, unnecessary.

The reason that estimating climate sensitivity is so important is that it is a defining property of the models that are used to predict the climate of the future. Sensitivity is not built in to these models as an input variable; it is an emergent property of their complexity. Even under identical forcing conditions, different models, or variants of the same models with small differences in the uncertain input parameters, will produce different levels of temperature change because of the interaction of the feedback parameters. The differences in temperature response are likely to be reflected in differences in the response of other climate parameters.

The great advantage of the last one or two millennia is that it is becoming feasible to use the high quality but spatially and seasonally biased early instrumental, documentary and natural archive palaeoclimate data directly to test the models by comparing the reconstructed climate parameters, whatever they may be, with the same parameters for the same places produced by the models. The models that give the best fit with observed climate changes in the past can reasonably be considered the most reliable. Since the sensitivity of the models will strongly constrain their ability to reconstruct realist changes in climate in the recent past, we might expect the 'best' models to represent a narrow range of sensitivities. Since the boundary conditions for the recent past are as close as we can ever come to those of the near future, we can assume that the sensitivity estimates reflect the real sensitivity of the world that we are going to have to live in. That assumption cannot be made when sensitivity is estimated using any period in the distant past.

Although the potential is there to test climate models and constrain sensitivity estimates using the best palaeoclimate data from the last 1 or 2ka, and some progress has been made at a hemispheric scale (Hegerl et al. 2006, Hegerl and Russon 2011), regional data provide a much stronger constraint. Some initial results are presented in the IPCC report (Masson-Delmotte et al. 2013) using the PAGES2k syntheses (Figure 5,12, p.419) but there are still some very significant barriers to progress. The obvious barrier is the computing time needed to run complex simulators such as GCM models for even a few hundred years. Computer power increases every year, but if the complexity of the models continues to increase to keep pace then we will never be able to use this approach. A possible solution is to use faster, simplified versions such as EMICS, at least to narrow the range of reasonable model variants, though we risk losing critical information on, for example, ENSO or the NAO. An even faster approach that might circumvent these problems is to use an 'emulator' model (Crucifix 2012) calibrated to a specific simulator (e.g. GCM). Many technical problems remain to be fully solved, including how to constrain the potentially vast number of model variants that are produced when just a few uncertain parameters are varied, how to deal with the varying uncertainty in climate reconstructions, and how to quantify the degree of fit between model output and both palaeoclimate reconstructions and modern climate.

Summary

I set out to review progress in 'measuring and understanding climate change' from a deliberately narrow perspective: specifically to explore how studies of Quaternary palaeoclimatology have contributed, or could contribute, to improving our ability to predict the climate of the future. The four 'timescales' used to divide the literature are purely arbitrary and others may disagree about the importance of the four 'challenges' that I have identified. They do at least provide a focus for discussion.

Challenge 1: Better understanding of how the climate system works.

Quaternary science has been absolutely central to providing a long term perspective on the variability of Earth's climate. In particular the work on ocean cores, and later ice cores, has demonstrated the clear link between orbital cycles and ice ages and shows the critical role of the greenhouse effect even at these long timescales. That work demonstrates unequivocally that the climate of the Earth is a complex system, constantly charged with energy from the Sun and kept far from equilibrium, and that as with other complex systems it can exhibit more than one state of operation. In the cold state, when there are large northern hemisphere ice sheets, the system displays incredible instability, with very large and rapid swings in temperature at high northern latitudes that are teleconnected around the globe resulting in more muted and opposing changes in the southern hemisphere. In the warm state, when large northern hemisphere ice sheets are restricted to Greenland, these rapid fluctuations do not occur. If we are to use the climate changes of the past to constrain predictions of the future, we need to accept that it is the climate system as it operates in the warm state that needs to be understood and modelled.

It is when it is operating in the cold state, however, that Earth's climate is most interesting, so it is not surprising that so much effort has gone into studying the rapid and extreme 'sub-Milankovitch ' scale perturbations of the last glacial cycle. As long as the difference in behaviour of the two climate states is acknowledged, this should not matter. In reality, however, the complex 2-state behaviour of Earth's climate if often ignored and the behaviour during the cold state is treated as if it applies directly to the way the system operates at present.

One of the exciting things that happens when there are large northern hemisphere continental ice sheets is that huge pulses of meltwater escape into the North Atlantic (Bond et al. 1992), capping it with a layer of cold fresh water and with armadas of icebergs. Such events undoubtedly have profound impacts on both atmospheric and oceanic circulation and the impacts on climate are captured in many proxy archives (Nesje et al. 2004). The critical role of the oceans in transporting heat from equatorial to high latitude regions, and across the hemispheres, is not in doubt, and the simple diagrams portraying the 'thermohaline circulation' have helped to convey that concept to a generation of students. This concept has been used to explain the global consequences of meltwater incursions into the North Atlantic by changing the strength of the north/south flows of Atlantic water, or the Atlantic Meridional Overturning Circulation (AMOC). The essence of the argument is that changing the temperature and/or salinity of surface waters in the far northern Atlantic influences the rate of sinking and that it is the rate of sinking that controls the strength of the AMOC circulation and therefore the transport of heat both across the equator and from the equator northwards.

The concept that it is changes in temperature and salinity of surface waters in the far north Atlantic that determines the strength of the AMOC and therefore that changes in ocean circulation are a driving force of natural climate change has become the dominant paradigm in Quaternary science. However there is scant evidence to support it. On the contrary, monitoring experiments demonstrate unequivocally that surface waters do not form the upper limb of the AMOC and that there is virtually no movement of surface water from the sub-tropical gyre into the north Atlantic. Also it has long been argued that it is simply not physically possible for the sinking of salty water to generate enough energy to do the work necessary to move the ocean currents. The dominant source of energy for the circulation of Earth's oceans is the wind. Changes in ocean circulation do not drive changes in atmospheric circulation. It is changes in atmospheric circulation that change the direction and strength of the winds and thus the movement of water and heat in the ocean. Of course the atmosphere and oceans are a coupled system, with the ocean acting as the dominant heat store and so there are strong feedbacks in that system, but the concept of the ocean as the dominant driver of climate change is just not supported by theory or evidence. This applies not just to the Atlantic, but also to the Pacific, where ENSO, although quantified in terms of changes in sea surface temperature is predominantly a wind-driven phenomenon. It is the strength of winds flowing along the west coast of South America that determines the amount of upwelling of cold water and so the east-west temperature gradient across the basin.

The strong focus of Quaternary science on sub-Milankovitch scale rapid climate changes, and on the strength of the AMOC as an explanation for them, has led to a paradigm that distorts the relative importance of oceanic and atmospheric circulation as drivers of climate change in the past and in the future. The atmosphere and ocean are coupled, but the ocean is not the dominant driver of change. We need to refocus on the importance of changes in atmospheric circulation.

Challenge 2: Better forecasting of the drivers of climate change

The strongest driver of climate change over the Quaternary has been cyclical changes in the spatial and seasonal distribution of solar energy caused by changes in Earths's orbit and position relative to the Sun. The so-called Milankovitch cycles are well understood and their behaviour can be predicted with certainty (Berger and Loutre 1991). The amount of energy received at 60˚N was at a maximum about 11,000 years ago and has been declining ever since. The result has been the final melting of the remnants of the large continental ice sheets and a slow decline in temperature over the last 8,000 years. Although we are approaching a minima in solar input to the critical northern latitudes, the climate is not continuing to cool and there is no prospect of us entering another ice age for about 50,000 years. Such a long interglacial is completely unprecedented over the whole of the Quaternary and reverses a trend towards longer glaciations and shorter interglacials.

The likely reason for escaping the slow decline into another glaciation is the greenhouse effect. In previous interglacials the long-lived greenhouse gases have risen initially but then declined as carbon is removed from the atmosphere and stored in the deep ocean. During the late Holocene, in contrast, both carbon dioxide and methane have increased, probably as a response to anthropogenic changes in land cover. It is difficult to explain how such early human activity could have such a large effect on the behaviour of the carbon cycle and in particular to explain why we have not experienced the expected transfer of carbon into the deep oceans. The 'early anthropocene' hypothesis is an emerging paradigm in Quaternary science and if it is true it is a powerful example of the consequences of human intervention in the carbon cycle. One of the consequences is that we, as the generation that has perhaps the only chance to make a difference, need to look well beyond the next 100 years that seems to be the horizon of interest to most of the modelling community.

As well as changing in response to orbital cycles, the solar energy that Earth receives changes over shorter time scales in response to the internal dynamics of the Sun. The role of changes in total solar irradiance as a driver of past climate change has been under discussion for many years in Quaternary science, but it is emerging as a really critical question for the whole of society. It is beyond dispute that the temperature of the Earth has been rising over recent decades and that there has been a measurable increase since pre-industrial times. However, from a palaeoclimatological perspective placing the benchmark for 'pre-industrial temperature' at AD1750 makes little sense, because in many areas of the northern Hemisphere that is part of the Little Ice Age. Indeed the whole concept of a pre-industrial benchmark for global mean temperature is unhelpful because we know that regional temperatures have always changed and global mean annual temperatures are notoriously difficult to estimate because much of our evidence has a seasonal and spatial bias. One legitimate reading of measured climate change is that most of it is a response to changes in energy received from the Sun.

The Little Ice Age corresponds to a time of low solar activity as evidenced by very low sunspot numbers (the Maunder Minimum), so part of the industrial-age increase in temperatures could simply reflect the recovery of solar activity to normal levels. The relatively rapid warming of the late 20th century also corresponds to a so-called 'grand maximum' of solar activity which has recently ended and solar activity is predicted to decline over the next few decades. If there is a direct link between solar activity and Earth surface temperature then it is completely missing from the climate models. In that case some of the increase caused by solar activity has been erroneously attributed to rising greenhouse gases over the same period and future temperature changes driven by the greenhouse effect may be greatly exaggerated.

From a personal perspective, I doubt that there is a strong direct link between solar activity and surface temperature and although I have looked for it in my own research I have not found it. I am sceptical because there is no clear physical mechanism to explain it and also because the accurate measurements of TSI over the satellite era suggest that changes in energy received at the top of the atmosphere are extremely small and not well correlated with sunspot numbers. However, there is a huge volume of material in the Quaternary literature that claims a strong link between solar activity and climate. Some of those claims are not based on solid evidence, and in some cases the link has been shown to be an artefact of the methods used to compile the proxy data, but others are hard to refute.

If there is a direct link between solar irradiance and climate, that is not included in the models that we use to predict future climate, then the models are severely compromised. Direct measurements of TSI are far too short to solve this problem, so the Quaternary community, with access to long proxy records that can be used to reconstruct both TSI and climate parameters is uniquely equipped to address this critical question. Efforts are hampered, however, by our tradition of storytelling rather than critical hypothesis testing and by an abundance of studies that lack the necessary rigour. If we are to solve this problem we need to first pose clear testable hypotheses, based on realistic physics, and then use our resources to critically test them. Repeatedly claiming to see a visual match between proxy climate series and proxy TSI forcing series is really not good enough.

Challenge 3: Improved estimates of climate sensitivity

The sensitivity of the climate system, defined in terms of change in temperature per unit change in forcing is an essential metric for understanding and predicting future climate change. Greenhouse gases are well mixed in the atmosphere and will exert a global forcing which will stimulate a series of fast feedbacks and we need to be able to model those feedbacks properly if we are to estimate the eventual degree of warming.

However, from a palaeoclimate perspective the whole concept of 'global climate sensitivity' is fundamentally flawed for the simple reason that it operates at the scale of the whole planet. The calculation requires a single value for the change in mean annual temperature for the whole Earth surface and a single value for the change in forcing, again averaged over the whole of the Earth's surface. This makes sense for computer models of the future impact of greenhouse gases, because they are well mixed in the atmosphere and they thus exert a global forcing. Models of varying degrees of complexity can predict the fast feedback response to that forcing, which will differ seasonally and across space, and average it to give a change in mean annual temperature for the whole globe.

Although it has been argued many times that the climate of the past provides an ideal way to calculate sensitivity, the reality is that it has not really worked. Even the large temperature difference between the LGM and Holocene has not reduced the uncertainty defined in the 1970s.

The heart of the problem is that natural climate changes have not been driven by forcing changes that are averaged across the planet surface. The dominant controls on Earth's climate throughout the Quaternary have been the Milankovitch changes in tilt and precession. The changes in energy received due to changes in eccentricity are too small to directly influence climate and must operate through modulating the amplitude of the precessional cycle. The critical point is that on a global scale the net forcing effect of the tilt and precessional cycles is zero, and yet they drag Earth into long cold glaciations and then throw it into brief warm interglacials. Changes in the distribution of energy during the Holocene forced changes in direct insolation over Africa, changing the strength of the monsoon and transforming parts of the Sahara desert into huge lakes supporting populations of fisherfolk, but again there was no change in forcing at a global scale.

The natural changes in Earth's climate have been driven by changes in the distribution of energy across the Earth's surface, both spatially and seasonally. The future consequences of human induced climate change will similarly depend on the spatial and seasonal patterns of those changes. Ignoring the geography of climate is a convenient way of comparing the average behaviour of computer models driven by the same change in atmospheric chemistry; it is not however a sensible way to view the climate of the past or the future.

By using the slow feedbacks of the climate system as the forcing side of the sensitivity equation, and attempting to combine seasonally and geographically biased temperature proxies to provide an estimated global mean annual temperature, it has been possible to use the difference between the LGM and Holocene to demonstrate that model-based estimates of climate sensitivity are not unreasonable, but not to reduce the uncertainty. There is no other time slice where we have a sufficiently large temperature difference and sufficient spatial information on slow feedbacks (forcing) and temperature to improve this estimate. We may have to accept that the ‘challenge’ of reducing the uncertainty in estimates of climate sensitivity is simply beyond the ability of the Quaternary research community for the simple reason that ‘global sensitivity’ is not a useful concept over the timescales of the Quaternary.

Approaches that use the spatial differences in forcing and response (Rohling et al. 2012) , rather than mean global values, have perhaps more promise, but are still hampered by the uncertainties in estimates of both temperature change and forcing change, which are accentuated when the available data have to be spread between different regions or latitudinal bands. Even if these estimates could be greatly improved, and we could produce a near perfect estimate of the behaviour of the fast feedbacks at some time in the past, such as the LGM, it might not actually help because it seems that the behaviour of those feedbacks depends on the climate state (Caballero and Huber 2013, Kutzbach et al. 2013). We cannot be confident that a sensitivity estimate based on time windows that were much colder or much warmer than near future conditions will provide a realistic estimate of the future evolution of climate.

Although we may not be able to use our data to reduce the uncertainty in estimates of climate sensitivity mathematically, we may be able to solve the problem indirectly. The concept of ‘climate sensitivity’ is important because it provides a simple metric that can be used to compare different computer models and those models are the only tools we have to predict the climate of the future. At present we do not know whether the models that predict large temperature changes (high sensitivity) are better or worse than those that produce smaller rises (low sensitivity). However, the sensitivity of the models is not defined in advance, as an input parameter, it is an emergent property of the complexity of the model. That means that if we can test the models directly, by how well they can reconstruct the climate of the past, and then exclude those that perform poorly, the remaining models might cover a smaller and more realistic range of sensitivities.

Challenge 4: testing models

The concept of using the climate of the past to test the ability of complex models to reconstruct climate conditions beyond the range used for calibration is well established. The PMIP projects have focussed especially on the LGM and mid Holocene time slices as periods that were colder and warmer than pre-industrial. A range of ‘palaeo-data synthesis products’ have been produced for these periods, allowing the fit between model output and spatial variability in key climate and environmental parameters to be assessed, qualitatively or quantitatively. There is little doubt that PMIP have achieved their stated aim of informing model development. However, in the context of the focus of this paper, it is difficult to argue that the palaeoclimate-model comparisons have contributed to reducing the uncertainty in predicting the climate of the future. The models get bigger and more complicated but the range of sensitivities, and therefore the range of predicted future climate scenarios, does not decline. In fact, in terms of the uncertainty of temperature rise over the next century, almost 40 years of model development have hardly reduced the uncertainty at all.

One of the limitations of previous PMIP programmes has been that the models used to reconstruct the past were rarely exactly the same as those used to predict the future. This makes it very difficult to directly use skill in reconstructing the past as an indication of likely skill in predicting the future. However, the PMIP3 programme that is currently underway is very different because it is linked directly to CMIP5. This means that precisely the same models are being used to reconstruct the past and to predict the future under a range of different scenarios. The three time periods chosen for comparison are the LGM (21ka), the mid Holocene (6ka) and the last millennium (AD850 to 1850) and it is hoped that all of the models (>30) will be run for all three (Schmidt et al. 2014).

The combination of CMIP5 and PMIP3 provides a real opportunity to at last use the climate of the past to constrain the uncertainties in predicting the climate of the future. Even if we accept that there are large uncertainties in the environmental reconstructions for the LGM and mid-Holocene, it is inevitable that some models will perform much better than others. Model simulations of global LGM cooling, for example, range from -3 to -6˚C (Braconnot et al. 2012) and different models predict very different shifts in the ITCZ during the mid Holocene (Schmidt et al. 2014). Few model runs are available for the last millennium but it seems that there are clear differences in the frequency domain that could be used to test their ability to produce realistic internal variability at appropriate scales, as well as in response to volcanic forcing (Schmidt et al. 2014).

The question remains, however, how will the results of PMIP3 be used? Is the aim simply to compare the models or are they being critically tested? The models in question here are of course enormously complex and sophisticated and huge resources have been invested in them. However, from a general scientific perspective a model is effectively a hypothesis. It is a way of summarizing one view of the way that the system works. Science does not make progress by posing alternative hypotheses and then simply admiring them, or gathering evidence to support them. On the contrary, progress is made by seeking conflict between competing hypotheses and firm evidence. If the aim of PMIP3 is to critically test these hypotheses then they will have to be prepared to reject some of them.

A call to reject some of the models that are available may seem harsh, but if we are to really make progress at last in reducing uncertainty it is an essential step. If there are firm grounds for believing that some models are simply better than others at simulating reality then we should focus our limited resources on the best ones. Limiting the number of models is only the first step, however, for each of these models has essentially been tuned to modern climate. For every model there are many parameters that are not known with certainty and which therefore have to be estimated. If the parameters are chosen to ensure that the model behaves properly under current climate conditions it is not surprising that most models make very similar predictions for conditions very similar to today. That is one reason why models generally agree on predictions for the next few decades. However, this simply masks some of the real uncertainty in the physics of the models and if we are to constrain that uncertainty then we will need to run not just one version of each model but many versions with key parameters varied within reasonable bounds of uncertainty. In such a 'perturbed physics ensemble' the parameters that influence clouds are likely to be a particularly large source of uncertainty.

Using the climate of the past to filter the best model variants from a large perturbed physics ensemble of even a single state of the art model is a daunting task, requiring enormous computing resources. There are also a number of challenges in terms of how we quantify the degree of agreement between palaeoclimate reconstructions and model outputs, particularly since our reconstructions always have uncertainties and those can be difficult to define precisely.

If the LGM cooling is going to be used as a way of narrowing the reasonable range of model sensitivities then an essential first step is to demonstrate a strong and unequivocal correlation between the degree of cooling that models or model variants reconstruct and the amount of future warming that they predict. So far, no such robust relationship has been demonstrated. In such circumstances using the LGM cooling to rank models is not justified. Similarly, if changes in hydrological systems in response to changes in radiative forcing are not strongly correlated in models or model variants used to reconstruct the mid Holocene and the future then the mid-Holocene climate is not a useful tool for model ranking.

The climate of the last one or two thousand years has not been used in previous PMIP experiments, but it is the period for which we have the best data. Many reconstructions for this period are annually resolved and are calibrated and verified in such a way that the uncertainties can be quantified. If we can bring the relevant data together to produce a range of regionally-resolved 'synthesis products', which is the aim of PAGES2, then there is a real chance of using them to critically test the ability of models to reconstruct both the spatial pattern and time evolution of externally forced climate changes and also to use their behaviour in the frequency domain to test their ability to produce realistic internal variability. This may be our most realistic chance of reducing the range of modelled sensitivity values and therefore narrowing the predicted range of possible futures.

Conclusion

There is no doubt that Quaternary research has contributed enormously to our current understanding of the patterns and magnitude of past climate changes and of the major factors that have controlled that change. The state of the art models that are our only tool for predicting future climate change have benefited from that research. However, it is difficult to argue that our efforts have actually improved the level of confidence that we can place in predictions of the climate of the future. On the contrary, after almost 40 years of research the wide range of climate sensitivities predicted by those models, defined as the equilibrium change in global mean annual temperature per unit increase in greenhouse forcing, has hardly changed. If we cannot predict future changes in global mean temperature, then predictions of regional changes in temperature or of other climate and environmental changes are also compromised. In reality we still do not know whether, even over the next century, we are going to experience changes that can be managed, albeit at great expense, or changes that are so extreme that we cross irreversible thresholds leading to catastrophic warming.

The complex models used for climate prediction cannot be critically tested using the short period for which widespread meteorological measurements are available, partly because they are effectively tuned using those same data. Quaternary scientists are uniquely placed to help to reduce the uncertainty by providing the means to critically test the models using the climate of the past. However there are some obstacles that currently limit our ability to do this effectively, in particular:

1. *The curse of the plausible narrative.* We have a tradition of storytelling rather than critical hypothesis testing. The result is a literature with many conflicting narratives that are difficult to reconcile. The modellers cannot use our reconstructions for model testing if we cannot agree which ones are most realistic. It is our responsibility to critically test competing hypotheses, not theirs.

2. *Sea-sickness.* The concept that climate change is dominated by changes in ocean circulation, and that it is changes in the sinking of surface waters in the far north Atlantic that controls the amount of heat transport in the oceans, has had a distorting influence on our treatment of the relative importance of atmospheric and oceanic circulation. The main source of energy for ocean circulation is wind and changes in atmospheric circulation drive changes in the ocean. We need to refocus on the atmosphere because it is changes in atmospheric circulation, and the attendant changes in the hydrological system, that are going to have the greatest impacts on nature and society in a warmer world.

3. *Solar glare.* The climatic impacts of changes in solar irradiance are strongly contested and if models do not treat this potential forcing properly they are severely compromised. Quaternary scientists could potentially solve this problem by using quantified comparisons of past climate and past changes in solar output to test physically-based hypotheses for solar forcing. However, efforts are hampered by a lack of rigour and too much emphasis on visual pattern matching.

Time is running out for improving predictions of future climate change. As the scientific community continues to produce increasingly complicated models, which get no nearer to agreement on the sensitivity of the system to greenhouse forcing, the developed and developing economies around the world continue to pump greenhouse gases into the air at an accelerating rate. Our continued lack of ability to restrict the very wide range of predictions for the future consequences of our actions limits our ability to plan for the future and the uncertainty encourages those who either genuinely do not believe that the climate is changing, or who have a vested interest in making that argument.

After 50 years the QRA has much to celebrate. However, we have the misfortune to live in interesting times. Earth's climate is changing rapidly and we urgently need to improve our ability to predict the climate of the future. This is not the time to rest on our laurels, and admire our achievements, we need to act quickly to use our expertise to critically test the state of the art models and reject those that are not realistic. The best models will then need to be tested more thoroughly using perturbed physics ensembles to access the underlying uncertainties that are hidden by tuning the models to recent climate. The role of Quaternary scientists is to provide rigorously tested and quantified reconstructions of climate and environmental changes in the past, at appropriately large spatial scales, with which to test the models. Many colleagues are now working towards this goal, not least through the PMIP3 and PAGES2 initiatives. We have certainly made progress in this critical endeavour, but we have a very long way to go. The title of this paper is based on a saying attributed to Confucius and it seems fitting to end with another from the same source: "real knowledge is to know the extent of one's ignorance".

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Reference	Temperature anomaly (˚C)	Confidence level
Schmittner et al. (2011)	1.7 – 3.7	90%
Annan and Hargreaves (2013)	3.1 – 4.7	95%
Hanson and Sato (2012)	4.0 – 6.0	1σ
Schneider von Deimling et al. (2006)	4.4 – 7.2	95%
Holden et al. (2010)	4.6 – 8.3	90%
Shakun and Carlson (2010)	≥ 4.9	--

Table 1. Recently published temperature anomalies for the LGM. Note that in some cases the statistically-defined uncertainties do not overlap, suggesting that real uncertainty is routinely underestimated.

Table 2. Proxy evidence that has been used to extract palaeoclimate and palaeoenvironmental information from ice cores. Based largely on reviews by Jouzel et al. (2013) and Bradley (2014).

Proxy	Interpretation	Examples
$\delta^{18}O$ and δD of ice	Vary (almost) in parallel. In Antarctica a strong temperature proxy, in Greenland originally misinterpreted as temperature but is also influenced by seasonality of snow.	Dansgaard et al. 1964, 1993; Johnsen et al. 1972; Jouzel et al. 1987, 2003; Werner et al. 2000
Deuterium excess: d	Based on the difference between the two water isotopes. Reflects changes in the source area for the moisture, including ocean surface temperature, relative humidity and wind speed. Also influenced by extent of sea ice.	Merlivat and Jouzel 1979; Vimeux et al. 2001; Masson-Delmotte et al. 2005; Stenni et al. 2010; Uemura et al. 2012
Dust concentration	Reflects wetness in source areas including central Asia and perhaps changes in atmospheric circulation. Shows seasonal cycle.	Ram and Illing (1995) Ruth et al. (2007) Lambert et al. (2008, 2012)
Electrical conductivity	Measures ice acidity, influenced by volcanic eruptions and by dust content.	Rasmussen et al. 2008
Volcanic events	Identified by sulphate spikes or glass shards (tephra). Important records of past climate forcing and provides stratigraphic markers to link ice sheets and other archives.	Hammer 1977; Abbot and Davies 2012. Svensson et al. 2012.
N_2O	Long-lived greenhouse gas, only reliable where dust content is low (Antarctica not Greenland). Linked to production in oceans and soils, lower when terrestrial biomass is reduced (glacials).	Leuenberger and Siegenthaler 1992; Wolf and Spahni 2007.
CO_2	Long-lived greenhouse gas, measurements only reliable where dust content is low (Antarctica not Greenland). Lead/lag relationship with temperature is complicated because it takes time for the gas bubbles to be isolated from the air. Recent work suggests temperature and CO_2 can change simultaneously (Parennin et al. 2013).	Raynaud et al. (1993, 2003) Blunier et al. (1995, 2007)
Methane CH_4	Long-lived greenhouse gas, mainly produced in wetlands or released from peat. Less influenced by dust, so used to cross-date the Greenland and Antarctic chronologies.	Stauffer et al. 1985, 1988; Loulerge et al. 2008
N_2/O_2 ratios	Linked to insolation perhaps. Used for dating.	Bender 2002; Landais et al. 2012
^{10}Be	Proxy for total solar irradiance	Steinhilber et al. (2012)
Air content	Related to air pressure and so altitude where the snow accumulated.	Lorius et al. 1968
^{40}Ar and ^{15}N	Fractionation of these gases in firn records abrupt temperature changes. Used to calibrate the $\delta^{18}O$	Severinghaus et al. 1998; Landais et al.

	record in Greenland to temperature and applied to one rapid event in Antarctica.	2004; Huber et al. 2006.
Noble gases	Neon, krypton and xenon. Solubility linked to temperature.	Severinghaus and Battle 2006; Headly and Severinghaus 2007.
Annual layer thickness	Can be translated into snow accumulation rate using models of compaction and flow. Annual layers not visible when accumulation is very slow.	Alley et al. 1993; Svensson et al. 2006
Borehole temperature	Temperature profiles retain a record of heat diffusion and suggest large temperature difference (>20C) between LGM and Holocene	Werner et al. 2000
$\delta^{17}O$ in ice	Potential for separating the effects of temperature and humidity on deuterium excess	Landais et al. 2008; Winkler at al. 2012
^{18}O in O_2 gas	Linked to primary productivity	Bender et al. 1985; Landais et al. 2010
^{17}O in air bubbles	Linked to biological oxygen productivity	Blunier et al. 2002

(note to reviewer: these references have not yet been added to the bibliography, I will add them if the table is included DMcC) $$

Figure captions

Figure 1. Stack of 57 $\delta^{18}O$ records from benthic foraminifera on a horizontal scale of thousands of years, including the palaeomagnetic timescale. Note the underlying long-term decline in $\delta^{18}O$ that is not driven by orbital forcing but may represent a slow decline in the strength of the greenhouse effect. The onset of glacial cycles, with large northern hemisphere ice sheets, marks the onset of the Pleistocene at 2.6m yr. The early Pleistocene is dominated by the c.41ka tilt cycle whereas the late Pleistocene is dominated by much longer glaciations and short interglacials apparently dominated by the very weak c. 100 ka eccentricity cycle. Note the change in vertical scale. From Lisiecki and Raymo (2005)

Figure 2. Evolution of the Precessional and tilt cycles over the last and next 50,000 years and the impact on radiation received at 65˚N. Note that 10,000 years ago both cycles were at a maximum so that radiation received was high and has been declining throughout the Holocene. We are now close to the bottom of the precessional cycle but the rise in energy received over the next 10,000 years will be matched by continuing decline of the longer tilt cycle. We seem to have escaped the expected decline into another long glaciation, perhaps due to an enhanced greenhouse effect caused by the early spread of agriculture, and there is now no prospect of substantial orbitally-driven cooling for at least another 40,000 years. The current interglacial is set to be the longest of the entire Quaternary.

Figure 3. Results of an attempt by Kohler et al. (2010) to use the temperature difference between the LGM and Holocene to calculate equilibrium climate sensitivity. The cooling estimate is based on Schneider von Deimling et al. (2006) and the slow feedbacks of the climate system, which change both the albedo and the greenhouse effect, are used on the forcing side of the equation. The resulting sensitivity estimate represents the fast feedback (or Charney) response that is required for comparison with model predictions. (A) Probability distribution of global cooling at the LGM. (B) Probability distribution of the LGM global radiative forcing relative to today from greenhouse gases, orbital forcing, ice sheets, vegetation and dust. (C) Uncertainty distribution in the scaling factor used to translate LGM climate sensitivity to present day sensitivity. (D) Probability distribution for present day equilibrium climate sensitivity for atmospheric CO_2 doubling resulting from panels A to C. The dashed lines represent uncertainty.

Figure 4. Mean annual temperature for Central England reconstructed using the mutual climatic range technique applied to beetle (coleopteran) remains. Note that warming is extremely rapid. Based on Atkinson, Briffa and Coope (1987).

Figure 5. The INTIMATE event stratigraphy 8000–48,000 b2k from Blockley et al. (2012). NGRIP and GRIP $\partial^{18}O$ profiles are shown against both age and depth. The warm 'Greenland interstadials' are marked in grey and numbered.

Figure 6. The great ocean conveyor model championed by Wallace Broecker. Note that warm surface waters carry heat across the equator and up to the far north Atlantic where it sinks to form North Atlantic Deep Water which flows back south. Together these ocean currents form the Atlantic Meridional Overturning Circulation.

Figure 7. Trajectories of 273surface drifters deployed in the North Atlantic from 1990 to 2002, passing through (a) the Gulf Stream box and (b) the Iceland Basin. Drifters entering and within the boxes are shown in green; drifters that have exited are shown in blue. These results show that surface waters do not form the upper limb of the Atlantic Meridional Overturning Circulation and that there is little if any movement of surface water northwards out of the sub-tropical gyre or southwards out of the subarctic gyre. From Lozier (2012).

Figure 8. High lake level episodes for the Jura Mountains, Swiss Plateau and French pre-Alps compared with variations in the atmospheric residual ^{14}C from Magny (2013), suggesting that

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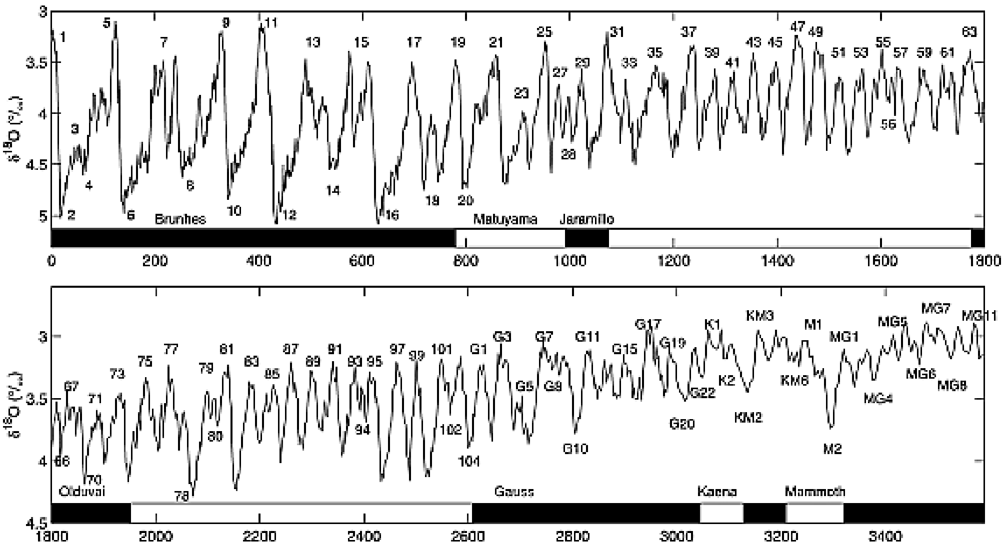
Figure 10 The PAGES 2k (2013) Network. Boxes show the continental-scale regions and the pie charts represent the fraction of proxy data types used for each regional reconstruction. Note the strong bias towards tree rings.



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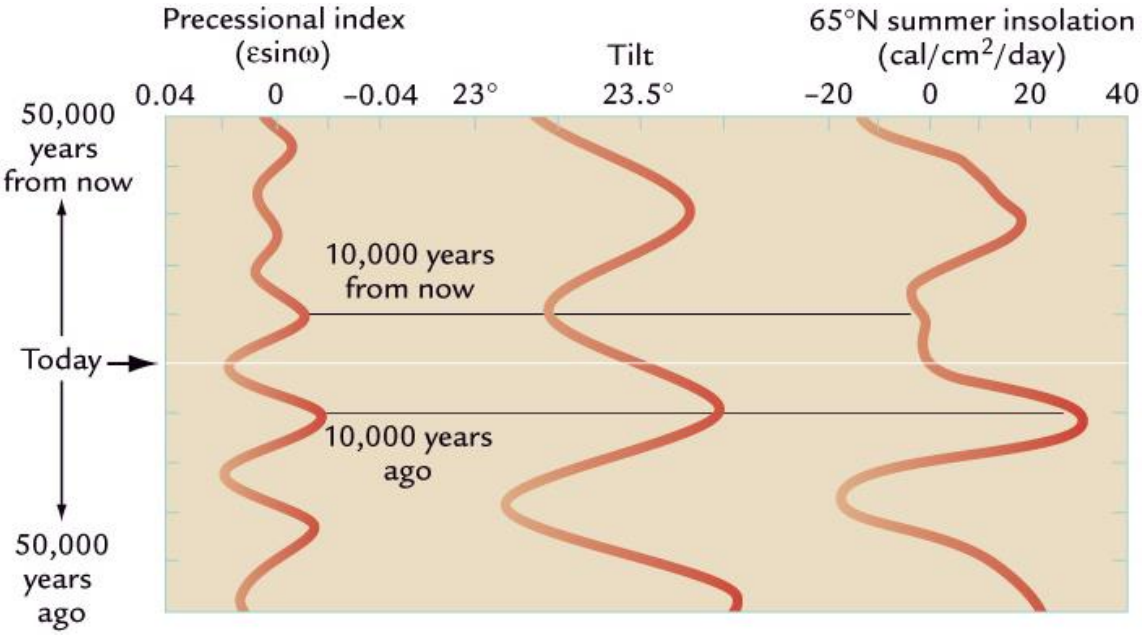

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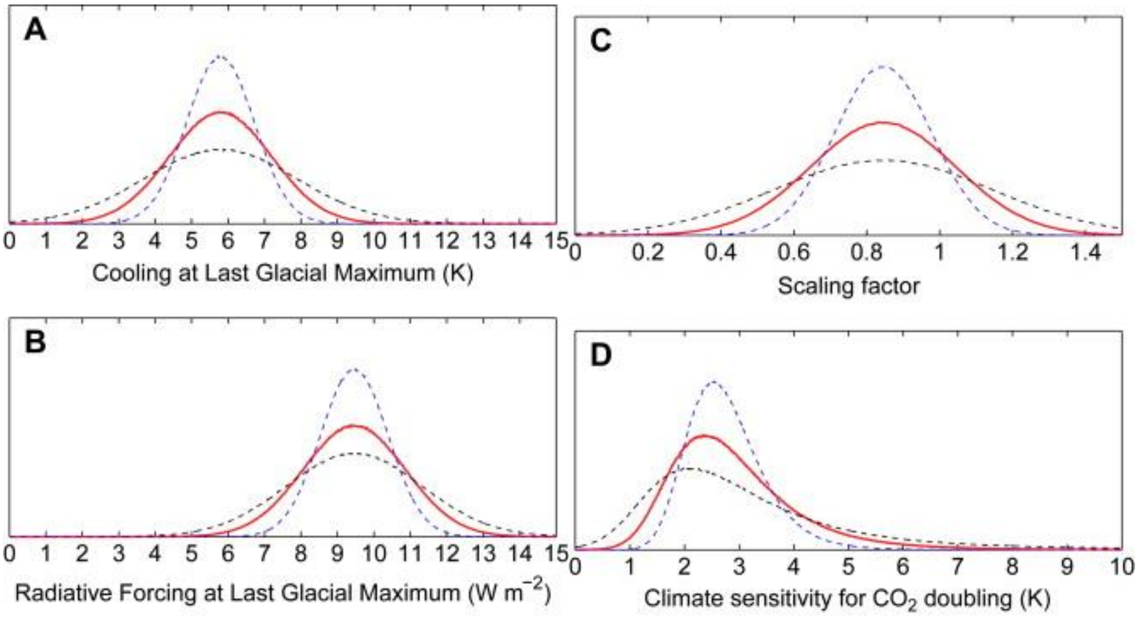

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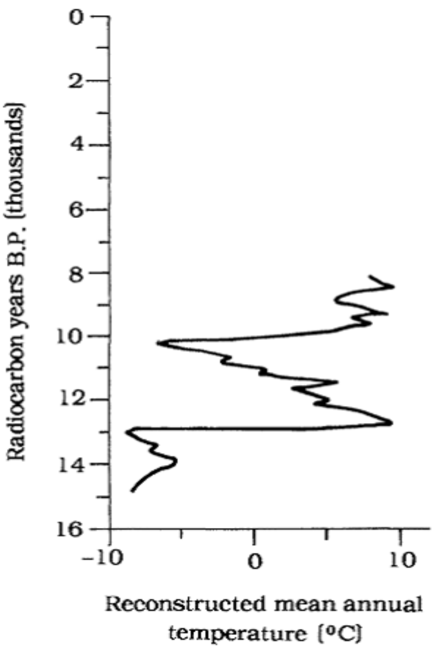

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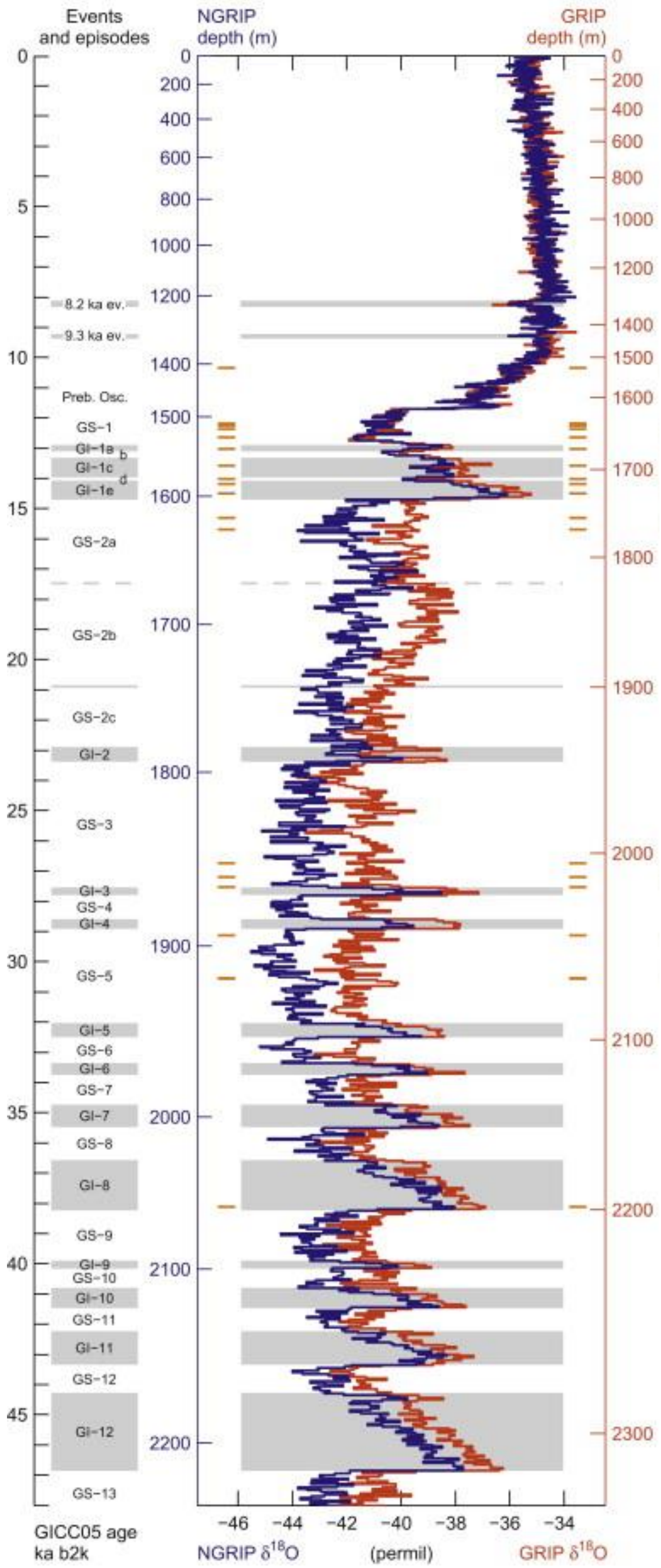

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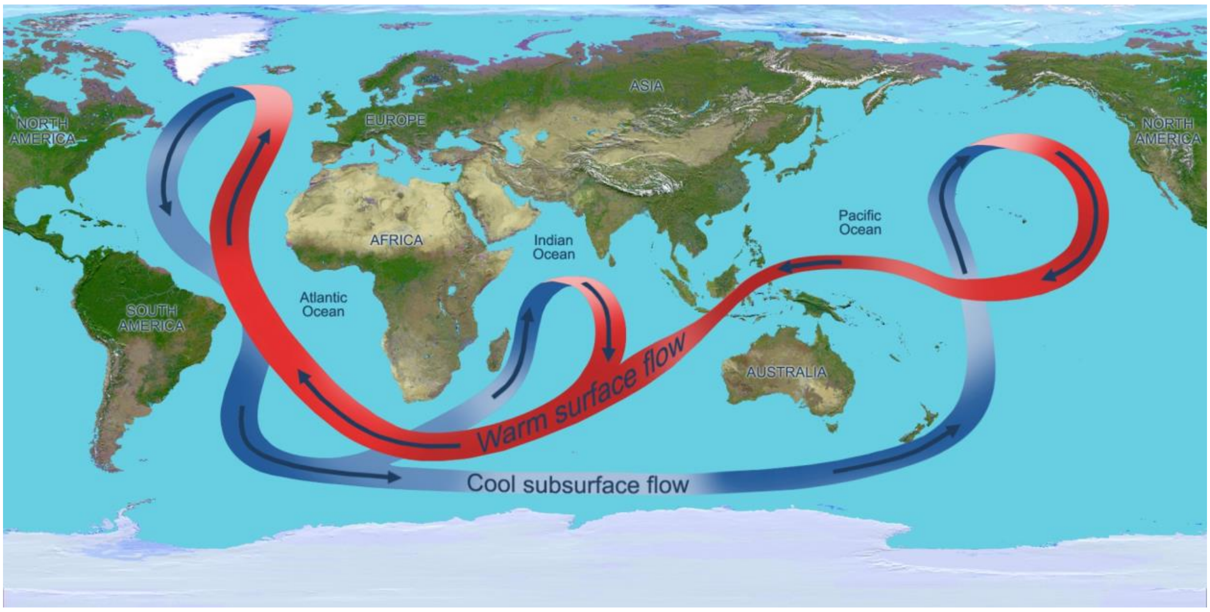

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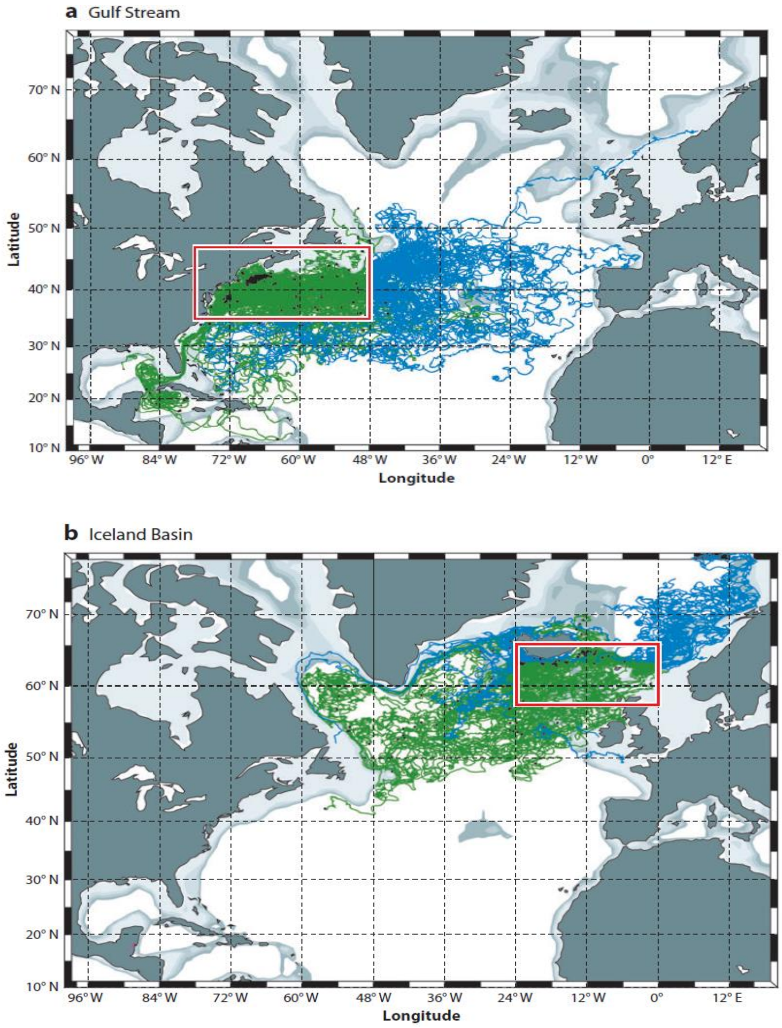

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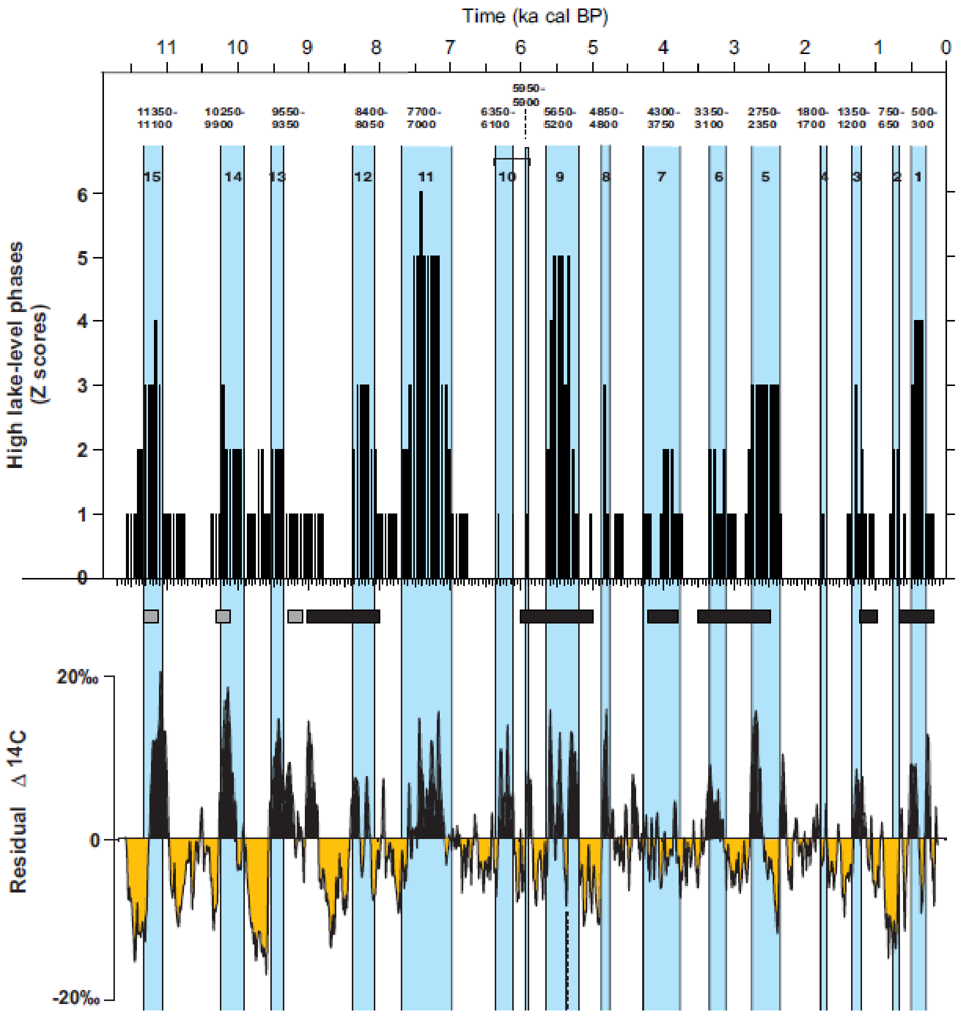

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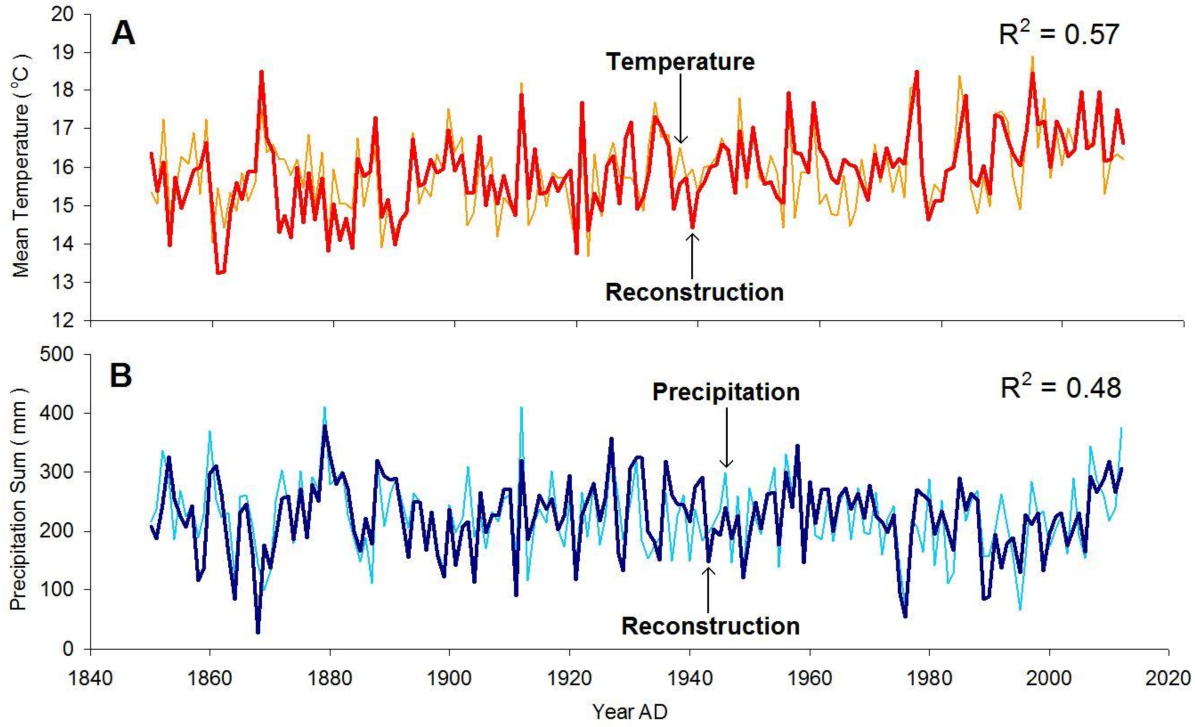

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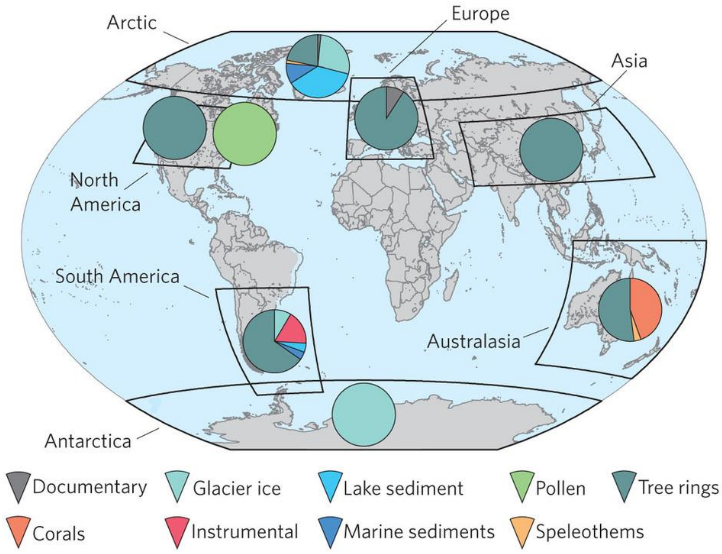

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