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General circulation modelling of Holocene climate variability

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Abstract

Results from a series of Goddard Institute for Space Studies (GISS) General Circulation Models (GCMs) are used to assess climate variability in the pre-anthropogenic Holocene, the interval following the end of the last glacial beginning roughly 11.5 kyr BP. In particular, we focus on the forced aspects of this variability. The principle forcings are orbital, solar, volcanic and events (such as the 8.2 kyr BP event). Land use and greenhouse gases also play a small role. We discuss suitable comparisons to paleo-data and the appropriateness of model experimental design for single and multiple forcing runs using time-slices and transient simulations. As an example, we focus on the response to solar and volcanic forcing in the context of the Maunder Minimum, a cool period of the late 17th century and demonstrate that although (northern) hemispheric mean temperature changes can be reasonably simulated in most models, the details of regional patterns depend more heavily on the included physics. In particular, we highlight the role of the stratosphere as well as the importance of ocean and vegetation feedbacks.

1. Introduction

There is an increasing community focus on Holocene climate variability that is driven partly by the increasing number of quality proxy paleo-records, but also by the obvious relevance for assessing future climate change (Renssen and Osborn, 2003). Climate models are becoming more sophisticated in how they can be forced and the quality of the physics contained therein. Better estimates of potential forcings are being produced (Bard et al., 2000; Crowley, 2000; Hansen et al., 2002) and better chronologies are allowing more complete and integrated snapshots of past climate than were available previously (e.g. Harrison et al., 2003). It is therefore unsurprising that more climate model simulations are now being performed with relevance to Holocene climate variability (Gonzalez-Rouco et al., 2003; Shindell et al., 2003; Widmann and Tett, 2003; Goosse et al., 2004). However, due to inherent limitations in knowledge of past boundary conditions, the appropriate initial state of the ocean and the limited simulation lengths possible with state-of-the-art models, the experiments being performed are often abstractions, rather than a reconstruction of a particular time period. Analysis of such experiments can often be problematic for paleoclimatologists. We will therefore attempt to address the question: What can climate modelling tell us about Holocene variability?

Firstly, we discuss the principal forcings that might have played significant roles, the time history of these forcings that can be used by models, and what the modelled response to these forcings generally imply.

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Modellers have an inbuilt bias toward forced climate change because the cause and effect are clear. Intrinsic variability in models or in the climate record is much more problematic due to the tight coupling of the physical (ocean, atmosphere, sea ice) and bio-geochemical cycles and the inevitable ambiguity regarding cause and effect that results. However, the characterisation of internal (or unforced) variability inevitably arises when trying to identify forced variability in the observed record. There are global mean, regional and local climate changes that will have occurred even in the absence of forcing, due for instance to variability in the ocean circulation on decadal or even centennial time-scales (e.g. Delworth et al., 1993). Distinguishing this kind of variability from the forced (and therefore more predictable) kind is a key research topic.

Secondly, we discuss time-slice experiments that have been, or can be, performed, and the kinds of paleo-data that are useful for validating model simulations. These kinds of experiments will focus on particular intervals where conditions are presumed to have been relatively stable (for timescales relevant for the particular model configuration) and attempt to model the impact of all particular forcings and relevant changes in boundary conditions. Due to the difficulty in producing time-slice paleo-reconstructions and the observed variability in the climate, only a few time-slices are likely to be examined in any detail; chiefly the mid-Holocene (8–6 kyr BP), the Maunder Minimum (late 17th century) and the pre-industrial (mid 19th century). Finally we discuss the simulations of transient climate change where coupled models of varying complexity are run with time-varying forcings to try and estimate the transient response of climate. These are principally being done for the last 500–1000 years.

2. Climate models

We will restrict the discussion to general circulation models (GCMs) of the climate. In particular, we will highlight model results using various flavours of the GISS GCM. These models, at minimum, have a reasonable simulation of the atmospheric circulation, radiative transfer and hydrologic cycle, but can differ markedly in the complexity of the ocean component and any associated bio-geochemical modules. Models of intermediate complexity have also been used to address these issues and are particularly useful for conceptual studies and for longer time-scales than can be accommodated in a GCM. For instance, these models have shown changes in the overturning circulation during the Holocene (Renssen et al., 2001; Goosse et al., 2004) and have been able to explore vegetation feedbacks more efficiently than GCMs (Claussen et al., 1999). However, many of the interesting issues for the Holocene relate to spatial or seasonal patterns that cannot be resolved in such models, for instance, relating to the winter atmospheric circulation, El Niño/Southern Oscillation (ENSO) or monsoonal changes. GCMs too, have difficulty with some important aspects of climate variability (i.e. ENSO in coupled models, the stratospheric connection to the North Atlantic Oscillation (NAO)/Arctic Oscillation (AO) etc.), but many aspects of atmospheric variability are reasonably simulated.

Climate models can be run in two ways, in equilibrium experiments and in transient mode. For the first case, a change is generally made in the boundary conditions and the model is run long enough so that it reaches a statistical steady state. This takes between 5 and 10 years of simulation in an atmosphere-only model (AGCM), around 20–30 years in models with some kind of upper ocean component (such as an ocean mixed layer (AGCM-ML or Qflux model)) and around 500–1000 years for models with a deep ocean (a fully coupled ocean-atmosphere GCM) due to the enormous heat capacity and slow mixing of the oceans. Intermediate configurations sometimes include a mixed layer model with some diffusion into the deeper ocean. Depending on the question to be addressed and the time-scales involved, a suitable equilibrium experiment can usually be defined. Note however, that there is an implicit assumption that longer term processes that are not explicitly considered (ice sheets, carbon cycle, sea surface temperatures (for an AGCM), the ocean heat transports (for an AGCM-ML) etc.) are constant. This can be problematic, particularly for the ocean and land surface (including the vegetation), both of which influence, and are influenced by, the atmospheric state. However, the advantage of solid statistics (due to the averaging out of short-term 'weather' noise) is frequently paramount. These kinds of experiments are generally validated using time-slices and spatial patterns of change.

The second kind of experiment is a transient run, where starting from an initial condition, the forcings change in time and a climate history is simulated. This kind of experiment most resembles paleo-record time series but is plagued with difficulties for the modeller. Firstly, defining suitable initial conditions is extremely tricky and laborious. In particular, the oceans retain a 'memory' of the past few hundred years which can play a role in the subsequent climate evolution. This is likely to be most important in the early Holocene where the ocean is still reacting to deglaciation and also in subsequent periods when the ocean may not be in equilibrium and observations may place only a weak constraint on oceanic changes. Secondly, any one experiment will have a great deal of chaotic behaviour unique to that realization and so many simulations (i.e. a 5–10 member ensemble) need to be done to get good a signal-to-noise ratio, particularly for assessing regional changes. Finally, accurate estimates of changes in forcing through time are extremely sparse and come with high uncertainties. Thus, transient climate simulations are only now starting to appear for periods earlier than about 1850.

Some transient experiments can be performed using transient sea surface temperatures (SST) and sea ice as a 'forcing' to the atmospheric model. This is possible for periods where good estimates of global SST and sea ice are available (since about 1870 so far (i.e. Rayner et al., 2003)). Conceivably, the SST record could be extended back using calibrated networks of proxies (Rutherford et al., 2003), which could avoid the need for expensive ocean model calculations in these transient runs, but sea ice reconstructions will remain problematic prior to direct observations. However, the response of the atmosphere to varying SST does not necessarily reproduce the variations in the coupled system that gave rise to the SST changes in the first place (i.e. Kushnir et al., 2002).

The latest generation of GCMs come with optional interactive components for many bio-geochemical subsystems. Due to the natural delay in using state-of-the-art models for paleo-climate, very few of these have been applied to Holocene climate variability as yet. Preliminary results with dynamic vegetation (Braconnot et al., 1999) and interactive stratospheric ozone (Shindell et al., 2001a; Rind et al., 2004), are however available and are discussed briefly below. We discuss in the final section likely future experiments.

3. Forcings and response

For climate models, forcings can be simply defined as changes in the background conditions that are external to the model calculations. For instance, changes to insolation either due to solar variability or orbital changes are clearly an external forcing. Changes in atmospheric composition can be considered to be forcing, as long as that particular component is not a prognostic output from the model. For example, in a model that does not calculate atmospheric chemistry, methane concentrations can be considered a forcing, while in a model with a prognostic methane cycle, it would be a feedback to changes in emissions or temperature. We discuss all the principle forcings (including some more speculative ideas) and the possibility of having reasonably accurate time-histories of them. We consider particularly the changes in forcing prior to around 1850 (the pre-industrial) to avoid complications due to recent anthropogenic perturbations to the climate system.

Forcings can be usefully characterised by the instantaneous change in the radiation budget at the tropopause (the instantaneous forcing). The eventual equilibrium global temperature change is roughly proportional to the forcing, with the climate sensitivity as the constant of proportionality. Another quantity, the adjusted forcing where the stratosphere is allowed to adjust radiatively while the troposphere is fixed, is a slightly better predictor (Hansen et al., 1997a), but the typically small distinction will not be addressed here. The concept of instantaneous forcing though is not very useful either for forcings which are extremely regional in effects, or for predicting regional climate change. For instance, orbital radiative forcing is almost zero in the global mean, but regionally and seasonally is extremely large. For each forcing, we discuss some of the most useful paleo-records or analyses for validating climate models and some of the key results and continuing problems.

Climate sensitivity is calculated as a matter of course by the global mean response to a doubling of CO_2 using an AGCM with an ocean mixed layer. The instantaneous forcing from this is around 4 W/m^2 and the range of climate response to that is around $3\pm1\,^{\circ}$C (Houghton et al., 2001). In estimates that follow we will assume a canonical sensitivity of 0.75 °C per W/m^2 unless otherwise stated. Some degree of polar amplification usually occurs (especially in winter), though the extent of this is dependent on the base state (e.g. ice extent) (Rind et al., 1995). Some forcings can have a greater impact on global mean temperature than an equivalent amount of CO_2, for instance, the black carbon (soot) impact on sea ice albedo (Hansen and Nazarenko, 2004), but the concept of equivalent radiative forcings is still a useful metric. Over the Holocene there is no strong evidence that climate sensitivity has changed substantially.

3.1. Well-mixed greenhouse gases

The most important greenhouse gas is water vapour. However, this has such a short lifetime in the troposphere (around 10 days) it can be considered a feedback to changes in other radiative forcings and is always modelled as such. The other principal natural greenhouse gases CO_2, CH_4 and N_2O have been measured in gas bubbles trapped in ice cores and have a past history that is rather well known. In fact the history of these gases over the pre-industrial Holocene is remarkably stable, with CO_2 variations of 20 ppmv around the pre-industrial value of 280 ppmv and N_2O values of around 260 ± 18 ppbv (Sowers et al., 2003). There is a small dip of around 4–10 ppmv in CO_2 from 1600–1800 CE in the Law Dome analysis (Etheridge et al., 1996) (although the timing of this dip is 300 years earlier in the Taylor Dome ice core (Indermuhle et al., 1999)). CH_4 is slightly more variable, with a peak value of 700 ppbv at 11 kyr BP, a minimum of around 575 ppbv at 5.5 kyr BP and up to 700 ppbv again by the pre-industrial period (Chappellaz et al., 1997; Houghton et al., 2001). There is a notable short term dip in CH_4 of at least 60 ppbv at around 8.2 kyr BP.

The radiative forcing from GHGs over the Holocene is therefore small, a maximum of 0.5 W/m^2 from CO_2 and around 0.09 W/m^2 from 5.5 kyr BP minimum to about 1850 for methane. The 17th to 18th century decrease in CO_2 is consistent with a 0.1–0.2 W/m^2 negative forcing at that time. The 8.2 kyr dip in CH_4 gives a small 0.05 W/m^2 and was probably not radiatively significant. The short lifetime of CH_4 in the atmosphere (8–10 years) and the relatively coarse sampling in the ice cores does not preclude shorter term excursions, but any such excursions are similarly unlikely to have had much climatic impact. There are additional indirect effects from CH_4 changes on stratospheric water vapour and on ozone, but these are smaller than the direct impact.

CO_2 and methane have effects on the global temperature proportional to their forcing and thus for the Holocene changes described above, the global annual mean changes will have been within the range of $\approx$0–0.5 °C.

3.2. Orbital changes

The largest perturbation in radiative forcing since the disappearance of the Laurentian and Fenno-Scandinavian ice sheets is undoubtedly the change in insolation (incoming solar radiation) from the mid-Holocene to the present due principally to the change in precession (Fig. 1). At around 6 kyr BP Northern Hemisphere

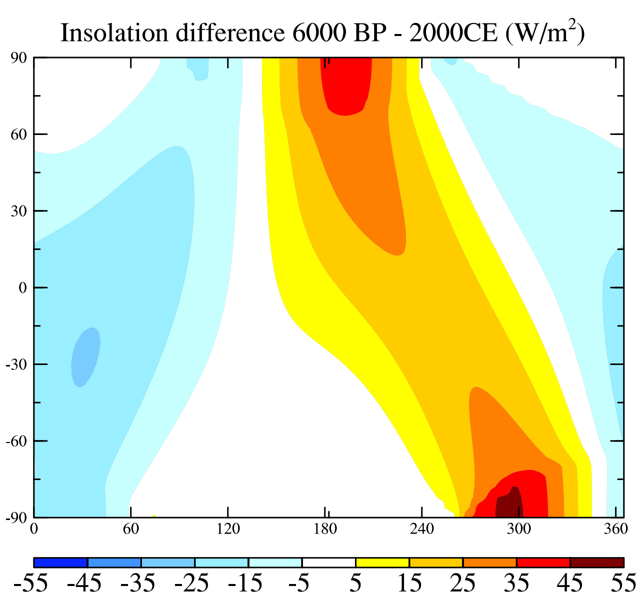

Fig. 1. The difference in orbital forcing at 6 kyr BP compared to 2000CE as a function of latitude and time of year (Julian days). The annual mean change in total global irradiance is extremely small, but in the high latitudes averages to about 5 W/m^2 increase and about a 1 W/m^2 decrease in the tropics (Berger, 1978).

summers had a significantly larger insolation than at present ($>$25 W/m^2 over the season), while tropical areas received slightly less in the annual mean. The high accuracy to which this forcing is known (Berger, 1978) and the similarity of the ice sheet configuration to today, has made the 6 kyr BP period a favorite target for climate models, particularly through the Paleoclimate Modelling Intercomparison Project (PMIP) (Hewitt and Mitchell, 1996; Joussaume et al., 1999; Braconnot et al., 1999; Otto-Bleisner, 1999; Kitoh and Murakami, 2002; Liu et al., 2003a).

The most useful observational targets for this period are the records of Northern Hemisphere (NH) high latitude warmth and maps of hydrological changes across N. America, Africa and Asia (Cheddadi et al., 1997; Texier et al., 1997; Viau and Gajewski, 2001; Harrison et al., 2003) (Fig. 2). Some fossil coral analyses have indicated that there may have been reductions in the ENSO variability (Gagan et al., 1998; Liu et al., 2003b). No good estimates of NH mean or global temperature are available and so this period is principally a test of the regional climate change, particularly in the hydrologic cycle, produced by the models.

Fig. 2 illustrates the importance of allowing ocean feedbacks for forcings that have time-scales longer than the ocean response time (see also Braconnot et al., 1999; Otto-Bleisner, 1999). These results are from recent simulations with the GISS model where only orbital forcing is changed while using pre-industrial atmospheric composition. For the first experiment, the ocean temperatures are kept fixed at pre-industrial values (1876–1885 mean) (Rayner et al., 2003), while in the second experiment the Qflux ocean was used which allows for a thermodynamic response to the forcing (results shown are 5 year annual means after the models have equilibrated). An increase in Sahel rainfall is clear in both cases, consistent with the lake level records, but the effect is doubled in the Qflux experiment. In prescribed SST experiments the surface energy budget is out of balance and the evaporation and hydrologic cycle are not fully adjusted to the forcing (Miller and Tegen, 1998).

Other feedbacks (principally vegetation, see Section 3.5 also) are likely to have also been important for this period (Braconnot et al., 1999). Similarly, Eurasian warming is clearly enhanced when the ocean temperatures and sea ice are allowed to respond to the high latitude increase in insolation. Comparisons to the northern European reconstruction of coldest month temperature (Cheddadi et al., 1997) (not shown) indicate that the west-to-east transition from cooling to warming is not well captured by either experiment. Overall there is no significant global mean temperature change (<-0.02 °C). The reconstructed pattern of drying over North America is similar to that seen in during La Niña phases of ENSO (Liu et al., 2003b), and as in previous

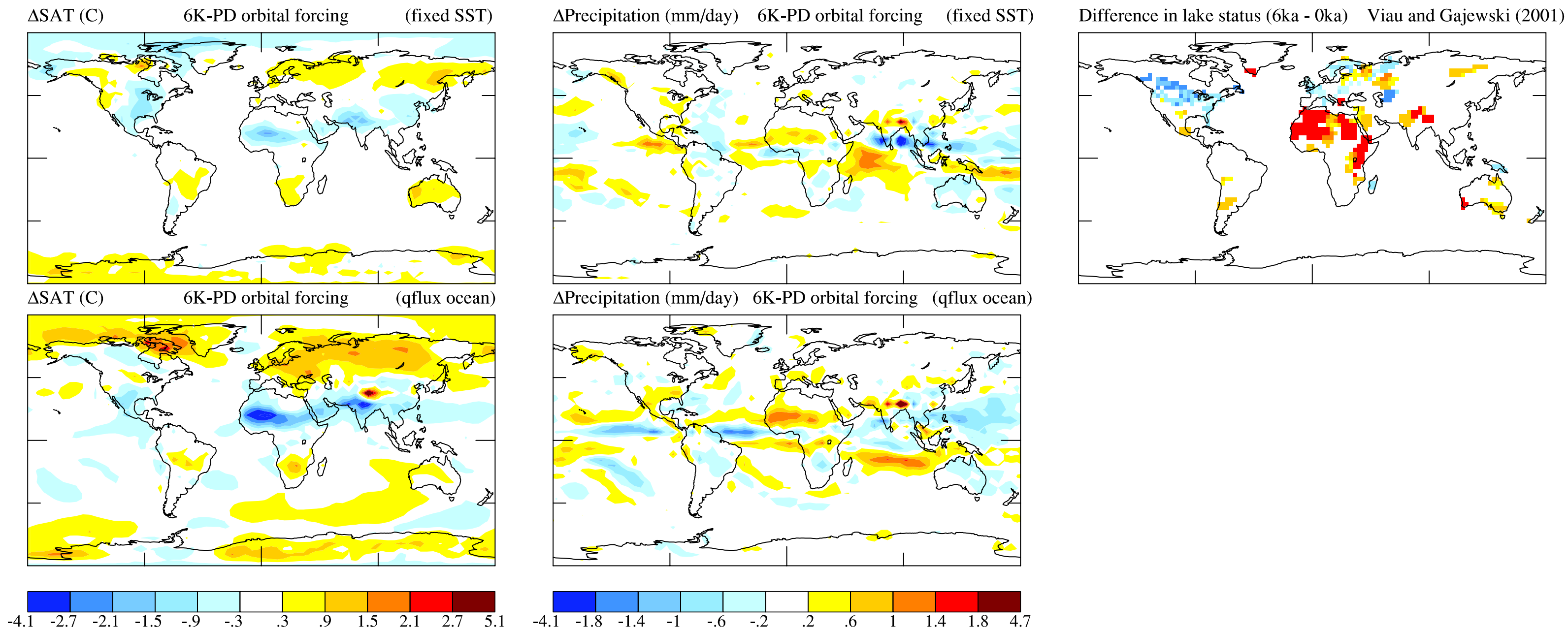

Fig. 2. The difference in annual mean surface temperature and precipitation patterns forced by the 6 kyr BP to present day change of insolation for two experiments (top row) with present day SST, (bottom row) with a Qflux ocean. The top right-hand figure shows reconstructed lake level differences that can be qualitatively compared to the changes in precipitation (Harrison et al., 2003; Viau and Gajewski, 2001) (red implies higher lake levels, blue lower).

studies, this pattern is not well simulated, conceivably because of the lack of change in ENSO with a purely thermodynamic ocean.

3.3. Volcanic aerosols

Explosive volcanic eruptions can deposit enormous amounts of sulphate aerosols into the stratosphere where they can linger long enough to cause significant radiative change. The aerosols are reflective and so increase the Earth's albedo (cooling the planet), while at the same time are locally absorbing (heating the lower stratosphere). Volcanic aerosols in the troposphere are swiftly rained out and are of less climatic importance due to the much shorter residence time (weeks rather than years). The global radiative forcing for an event such as the eruption of Mt. Pinatubo in 1991 reached a peak of -3 to $-4\,\mathrm{W/m^2}$ (Hansen et al., 1996, 1997b). Note that the lower stratospheric warming impact of volcanic aerosols can only be simulated with a relatively sophisticated radiative transfer code and has been shown to be climatically important (see below) (Kirchner et al., 1999; Stenchikov et al., 2002; Shindell et al., 2001b, 2004).

Determining the time history of the volcanic forcing involves the timing of eruptions, the amount of sulphate released and the composition, particle size and the fraction of sulphate entering the stratosphere. Satellite and proxy measurements have shown that it is principally tropical volcanoes that have the most lasting effects (Pinatubo, Tambora, Toba, etc.) and that these lead to increased sulphate deposition in ice cores in both Antarctica and Greenland. These ice core records have therefore been used to estimate sulphate forcing back through the Holocene (Hammer et al., 1997; Udisti et al., 2000). The most reliable records are for the last 1000 years or so (Crowley, 2000; Robertson et al., 2001) where many of the peaks can be reliably calibrated to known eruptions (e.g. Tambora in 1815, Huaynaputina in 1600). Note that the radiative forcing from even well-observed eruptions can differ significantly depending on the data set used.

While the impact of a single eruption lasts at most a few years (the residence time for aerosols in the stratosphere), there are apparent changes in volcanic frequency on decadal and even centennial time scales in the Holocene. Thus, changes in volcanic forcing can impact longer term climate. Experiments with Energy Balance Models (EBMs) (i.e. Crowley, 2000) have shown that the hemispheric mean cooling in response to volcanic forcing (Mt. Pinatubo for instance) is well simulated by such models and is proportional to the climate sensitivity. This result is borne out in decadal-scale simulations with GCMs as well (Shindell et al., 2003). There is some evidence too for volcanic impacts on ENSO frequency (Adams et al., 2003), though this has yet to be reproduced in models.

An interesting picture emerges when we consider the seasonal data. Based upon the instrumental period, Robock and Mao (1995) showed that there is a significant tendency toward winter warming over specific portions of the NH continents, particularly Eurasia, in the first and second winters following an eruption. Work with long historical archives (Fischer et al., 2004) and with seasonally resolved paleo-reconstructions (Shindell et al., 2004) has shown that this appears to be true over the last 500 years as well.

Volcanic aerosols from El Chichon (1982) and especially Pinatubo (1991) have been thoroughly observed and assuming that previous large eruptions had similar spatial and size distributions of aerosol and time decay, model simulations of the effects can easily be done. Fig. 3 shows a GCM study of the effects of the Krakatau. Santa Maria and Pinatubo eruptions from five ensemble members (so a total of 15 eruption cycles) (mean radiative forcing $-3.7\,\mathrm{W/m^2}$ in the year following the eruption based on the Crowley (2000) dataset). Examining the mean temperature for the winter following the eruption demonstrates quite clearly the Eurasian warming/Mediterranean cooling pattern seen in the paleo-reconstructions and modern observations. This pattern of winter warmth is associated with an enhanced wintertime circulation, represented by a positive phase of the NAO/AO (Hurrell, 1995; Thompson and Wallace, 1998). Note that the wintertime NAO phase is highly variable and the changes seen in the figure in the mean are much smaller than the interannual variability and so are only statistically significant over many realizations. However, these regional patterns are almost completely absent in the decadal mean (Shindell et al., 2003). Over this longer term the patterns are of generalised cooling, with an enhancement toward the poles. This pattern of 'winter warming' is at least partially related to changes in temperature gradients and dynamics in the stratosphere, and models with only poor stratospheric resolution are less able to capture this phenomena (Robock et al., 1999).

3.4. Solar irradiance

Observations during the satellite era have shown definitively that solar irradiance can vary during the roughly 11-year solar cycle (e.g. Willson, 1997). During the solar maximum, the sun is particularly active and dark sunspots as well as bright faculae are more evident. Total solar irradiance (TSI) is about 0.1% stronger than during a solar minimum when there are very few sunspots. While the integrated solar irradiance varies little, the higher frequency bands (such as for UV radiation) can vary by 1–10% (Lean et al., 1995). The direct forcing from solar minimum to solar maximum is

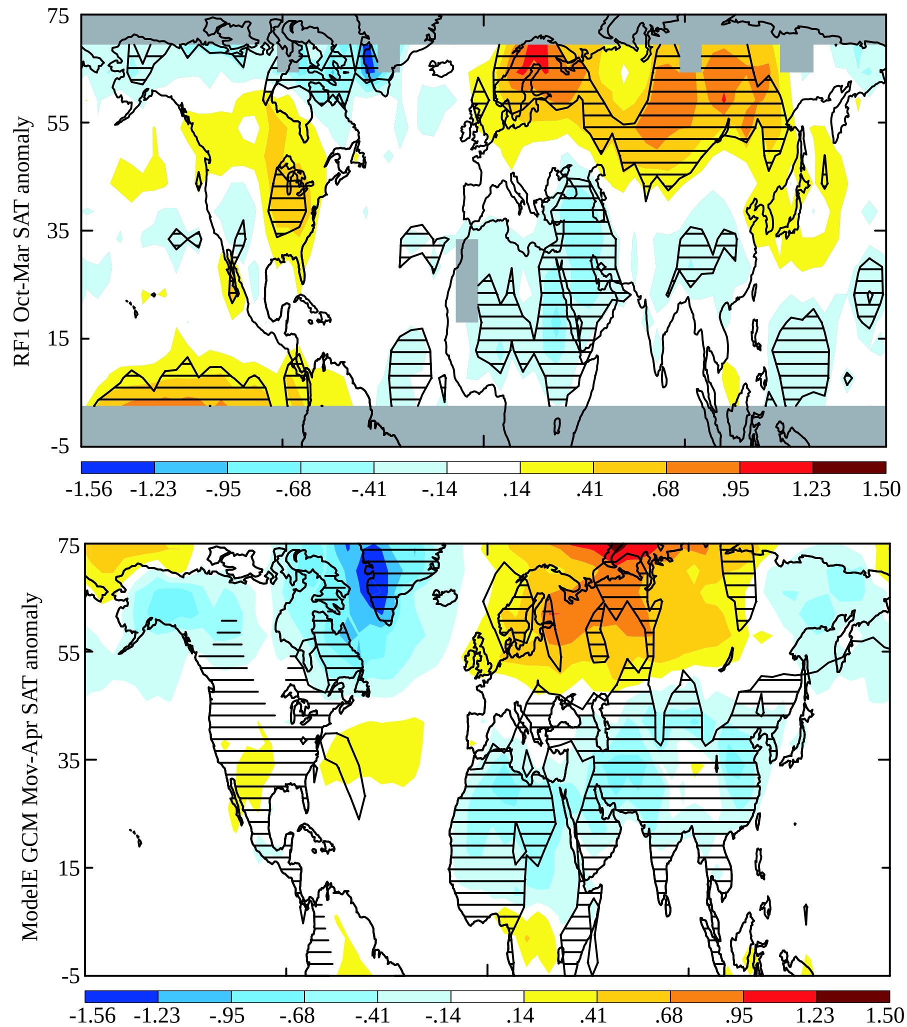

Fig. 3. (Top) The regional pattern of warming and cooling (°C) seen in seasonally resolved paleo-reconstructions in the winter following a major tropical eruption based on an average of 11 large and well identified historical eruptions (mean forcing $-3.8\,W/m^2$) and (bottom) the ensemble mean impact of increased volcanic aerosols seen in GCM simulations with a mean forcing of $-3.7\,W/m^2$ (Shindell et al., 2004). Significant (95% confidence) areas of change (with respect to winters with no volcanic forcing) are hatched.

about $0.24\,W/m^2$, after accounting for geometry and the Earth's albedo. It remains uncertain how longer-term solar variability relates to changes over the solar cycle.

Approximately in phase with the solar irradiance, the solar magnetic field also varies and this in turn modulates the amount of cosmogenic radiation hitting the terrestrial atmosphere. These high energy cosmic rays are the source for cosmogenic nuclides such as ^{14}C, ^{10}Be, ^{7}Be, ^{36}Cl and ^{3}H which thus have production rates that vary on solar cycle timescales (less production during a solar maximum) (Lal and Peters, 1967; Masarik and Beer, 1999). Paleo-records of these isotopes have therefore been used to estimate solar irradiance in the past (Stuiver and Quay, 1980; Bard et al., 2000). These records have generally been calibrated to the estimates of solar forcing at the Maunder Minimum, themselves highly uncertain (Bard et al., 1997; Lean et al., 2002) and so do not provide an independent estimate of the forcing. Additionally, the amount of contamination in records of ^{10}Be due to climate impacts on its deposition have not yet been satisfactorily quantified (Salyk et al., 2003). However, millennial time series combining reconstructions based on the observed sunspot cycle with the cosmogenic isotopes do exist, with the caveat that significant uncertainties exist in the magnitude of long-term solar change (Bard et al., 1997).

Isolating the effects of solar forcing in the climate record is difficult, particularly in the instrumental record because of the number of other factors that change at the same time and the shortness of the record. Paleoclimate reconstructions (Mann et al., 1998) are, however, long enough to be used. Waple et al. (2002) showed that the presumed long-term solar variability appears to have a significant response in surface temperature patterns when filtered (>40 years) and lagged (by

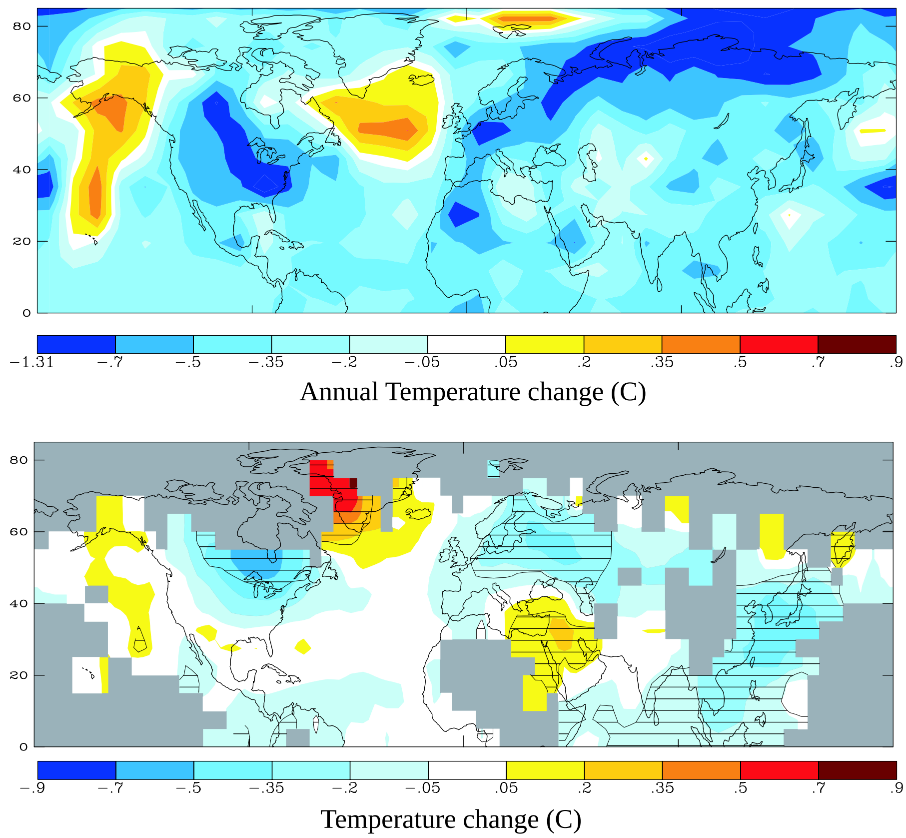

Fig. 4. (a) The temperature pattern seen in a model with interactive stratospheric ozone when forced by spectrally resolved solar forcing comparable to the scaling in lower panel (Shindell et al., 2001b). (b) The annual temperature pattern correlated to the solar irradiance reconstruction when smoothed to leave only variability > 40 years and lagged by 20 years (Waple et al., 2002).

10–30 years) (Fig. 4b). The lag is consistent with an ocean thermal response to the forcing. Note the approximate anti-correlation to the winter warming pattern for volcanic effects seen above, underscoring the spatial patterns of temperature change over oceanic and terrestrial regions that appears to be superposed on the global mean responses to certain climate forcings.

Models that are run with a TSI change generally produce minimal global mean temperature changes (around 0.2 °C at equilibrium for a 0.1% change), in line with their sensitivity to CO_2 increases (Hansen et al., 1997a). This is generally not sufficient to match observed variability over the solar cycle, particularly in the stratosphere. Stratospheric ozone, however, is particularly sensitive to changes in the UV portion of the spectrum and models that include a realistic stratosphere and interactive ozone calculation tend to see much more realistic variability over the solar cycle (Haigh, 1996; Shindell et al., 1999). Over the solar cycle the ocean does not react significantly. However, for longer term variability, the ocean thermal response (at minimum) implied in the observations must be considered.

The AGCM-ML results described in Shindell et al. (2001b) demonstrated a significant shift in regional temperatures (Fig. 4a) which corresponded well with the paleo-data. These results are equilibrium results, i.e. the mixed layer ocean has had time to thermally adjust to the change in surface fluxes, while the ocean heat transport remained constant. The change of temperature seen here is mostly related to a shift in sea level pressure from mid- to high latitudes, consistent with a more negative phase of the NAO/AO. Interannual variability in these indices does not change appreciably. The comparison with the Waple et al. (2002) analysis is appropriate because both cases are looking only at the solar component of change (over a long time period) and the lag is appropriate for an ocean thermal response. Given the uncertainties in the magnitude of solar forcing, the climate sensitivity and the ability of the proxy data to capture long-term variability, the match is surprisingly good. As in the response to volcanic forcing, a large part ($\approx 50\%$) of this result was due to the stratospheric temperature gradient changes induced by the interactive ozone response.

The purported impacts of longer term (millennial) solar variability in the Holocene (Bond et al., 2001; Hu et al., 2003) will require both better forcing functions and fully dynamic ocean models to even start to address the issues adequately. Initial results suggest that ocean dynamics might well lead to hemispheric differences in response (Broecker et al., 1999; Goosse et al., 2004), as well as potentially important responses to any changes in the NAO/AO (Visbeck et al., 1998; Delworth and Dixon, 2000).

3.5. Land surface changes

Changes to the land surface impact the climate in a number of direct ways through changes in the albedo, evapo-transpiration and runoff, but also indirectly through changes to biogenic emissions of volatile organic compounds (which impact atmospheric oxidation and hence methane and tropospheric ozone), organic aerosol formation (which interact directly with the radiation and indirectly with cloud formation) and the potential for dust emission (which decreases as vegetation expands). Other impacts are possible through changes in biomass burning (a source of CO, black carbon etc.) linked to aridity and changing vegetation types, and the effects of wetland expansion or contraction on methane emissions. Vegetation change estimates exist for the last 3 centuries (i.e. Ramankutty and Foley, 1999) and through projects such as BIOME6000 for the mid-Holocene (Prentice and Webb III, 1998).

At present, no climate models have included the full range of effects. However, preliminary results with changes to vegetation distributions (Braconnot et al., 1999) have shown that this is probably a key factor in modelling the mid-Holocene and that feedbacks between climate and vegetation are likely to be important in more recent periods (Bauer et al., 2003).

3.6. Other forcings

There are a large number of additional radiative forcings that may have been of some importance during the Holocene. These include dust, tropospheric and stratospheric ozone, natural sulphates, organic aerosols, stratospheric water vapour and minor greenhouse gases (such as SO_2). Unfortunately, these fields are not well-mixed (i.e. there is significant spatial heterogeneity) and there are very few direct measurements.

The dust record is somewhat constrained due to a few ocean sediment records downwind of the major source regions. During the African Humid Period (8 kyr BP to 5.5 kyr BP), Atlantic dust deposition was reduced by about one-third, compared to the present day (deMenocal et al., 2000). Currently, African and Arabian sources contribute $-0.12\,W/m^2$ to the global forcing at the top of the atmosphere (Miller et al., 2004), corresponding to a forcing of $0.04\,W/m^2$ during the African Humid Period. However, this forcing could be up to four times higher since the present-day dust load calculated by Miller et al. (2004) is at the low end of estimates and the assumed particle absorption is possibly too high. On the other hand, Prospero and Lamb (2003) suggest that transport of Sahelian dust across the Atlantic may have been unusually large during the satellite era used to constrain dust models, compared to the pre-industrial value reflected in the sediment cores. The potential for African dust to contribute significantly to the climate forcing at 6 kyr BP is thus still uncertain.

Another example is the potential for greenhouse gas forcing by volcanic emissions of SO_2 that has been hypothesised to have been a factor in the response to non-explosive eruptions such as Laki in 1783 (Highwood and Stevenson, 2003).

In order to assess the global importance of these ancillary forcings, models must be run to produce fields that are consistent with the appropriate boundary conditions and available emission histories. This is an active research area and given the uncertainties involved and the sparsity of good validating data, results from such simulations are likely to remain speculative for some time to come.

4. Time-slices

Time-slice experiments can be very useful in matching model results to multiple sources of paleo-data with reasonable time control (particularly ice cores, tree rings, etc.). However, a number of points need to be borne in mind. Firstly, the period of time for which the forcings are assumed fixed determines which model configurations are most appropriate. For example, the response of a short-lived forcing (up to a couple of decades) may be calculated with a mixed layer ocean whereas the ocean dynamic response would be crucial to longer forcings. Secondly, all known relevant forcings must be included and thirdly, the data to which the models are compared must be tightly controlled chronologically and with as widespread coverage as possible. The response will depend on the duration of the forcing as in the example above distinguishing the ocean response to a single volcano versus several decades of with unusually high volcanic activity. Similarly, the dynamic ocean response can depend on the time-scale of the forcing (Visbeck et al., 1998). One must therefore be extremely vary of extrapolating results to longer periods.

Below, we highlight the Maunder Minimum ($\approx$1650–1710) period. We refer the readers interested in the 6 kyr BP time slice to the more comprehensive discussion of the PMIP experiments in Joussaume et al. (1999) and references cited therein. The PMIP2 project is slated to coordinate simulations for the early Holocene (10 kyr BP), but we are not aware of any further organised efforts to look at other Holocene time-slices.

4.1. Maunder minimum

The Maunder Minimum (1650–1710) has become the standard period for assessing variability in pre-industrial climate. It was a period during which very few sunspots were observed and has been recognised as one of the coolest intervals of the so-called Little Ice Age (LIA) (Eddy, 1976). Unlike the LIA, the Maunder Minimum is a fairly well defined chronological period and since it is well within the historical and instrumental period (for a few long records), there is a relative wealth of data available for comparison. The time period involved is certainly long enough for the ocean thermal response to be important and possibly for ocean dynamics, particularly in the North Atlantic, to start to respond. Therefore the experiments described below using a mixed layer ocean are at the margin of appropriateness for this period.

The data comparison that we use is the difference in the 1660–1680 mean (which allows time for the ocean thermal response to be felt) compared to a century later (1770–1790). There is a significant difference in both the estimated solar forcing and volcanic forcing for these two periods and yet sizable anthropogenic modification of atmospheric composition has not yet taken place. Land use (probably) and greenhouse gas changes are minimal between these two periods. Fig. 5 shows the comparison in the multi-decadal mean temperatures. The spatial pattern is consistent with the results from the solar-only cases (Fig. 4), although the scale of change is larger in the model (hemispheric mean change of 0.5 °C, compared to 0.2–0.4 °C in the observations) (Shindell et al., 2003). This could be because the sensitivity of this particular model (1 °C per W/m^2) is a little high (compared to the canonical 0.75 °C per W/m^2), or because the forcing is too large (i.e. the solar change is over-estimated in the reconstruction of Lean et al. (2002)), or due to uncertainties in the amplitude of low-frequency variability in paleo-climate reconstructions (Esper et al., 2002; Mann and Hughes, 2002). Given these uncertainties the comparison is reasonable, but this does highlight some of the problems in performing time-slice experiments as compared to single factor experiments described above.

5. Transient simulations

For a successful transient simulation, good records for all the relevant forcings, a reasonable initial condition and computer resources to run multiple long runs must be available. These restrictions have meant that simulations have generally been done only for the last 500–1000 years with coupled models of any

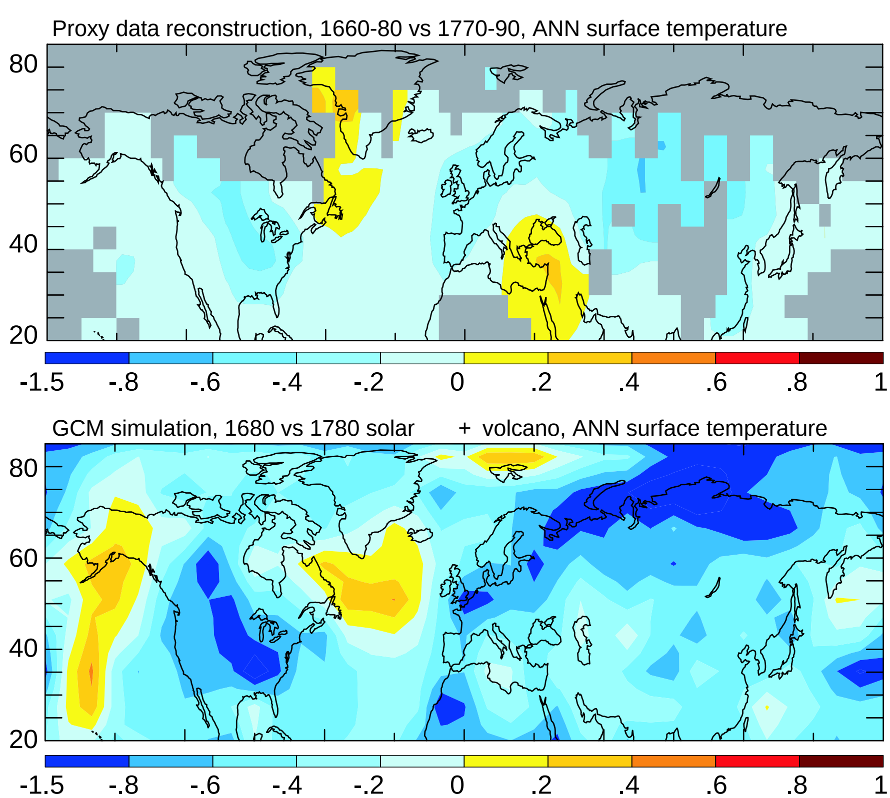

Fig. 5. Difference in annual mean temperature (°C) between 1660 and 1680 and a century later (a) from Mann et al. (1998) spatial reconstructions and (b) from a combination of the solar and volcanic forcing results from a GCM (Shindell et al., 2003).

complexity (Gonzalez-Rouco et al., 2003; Widmann and Tett, 2003; Goosse et al., 2004; Rind et al., 2004). The length of the simulations implies that the runs are almost always done with a relatively old version of a GCM, and so some components of the forcing or the physics are somewhat less than state-of-the-art. The target data set for comparison are predominantly the various NH temperature reconstructions (Briffa et al., 1998; Mann et al., 1999; Esper et al., 2002; Luterbacher et al., 2004, etc.).

Other Holocene transient climate events that have been modelled thus far are the 8.2 kyr BP event (Renssen et al., 2001) and the termination of the African Humid Period (Claussen et al., 1999; Renssen et al., 2003) albeit with simplified models. The 8.2 kyr event is likely to be a good target for future modelling attempts because the signal is quite strong and there is a relatively well constrained forcing function (from the final draining of Lake Agassiz or possibly the final collapse of the Hudson Bay Dome, around 5–15 Sv yrs of freshwater) (von Grafenstein et al., 1998; Barber et al., 1999).

The principle modelling issue is finding a suitable (ocean) initial condition to perturb. However, we will not focus on these events here.

5.1. The last 500 years

Given the emphasis in this review on climate changes during the past few centuries, we focus on transient simulations of a period (1550–1800) that encompasses the Maunder Minimum but that does not include either the spin-up phase of the simulation or the very large radiative perturbations that start in the 19th century. We calculate the anomalies in the NH mean surface temperature so that the mean over this period is zero for both the target paleo-reconstruction and the model results. This gives comparable results for the natural, unforced variability for this period and minimises distortions introduced from varying climate sensitivities to increasing GHG toward the end of the simulation.

Fig. 6 shows the NH mean surface temperature anomaly for a meta-ensemble of 6 runs (Waple et al., 2002) reconstruction. We use the term meta-ensemble to distinguish from a standard ensemble where only the initial conditions are different. In these cases, the depth of the deep ocean layer, the ocean initial conditions (cold or warm starts) and magnitude of the volcanic forcing vary. These runs are relatively coarse resolution ($8^\circ \times 10^\circ$), 9 layers in the vertical and have an ocean mixed layer with diffusion into a deep ocean with solar, volcanic and GHG forcing (Robertson et al., 2001). A number of points can be made: (i) The short term impact of volcanic eruptions is clear in the model ensemble mean and in the observations (despite some uncertainty over timing and importance). The magnitude of the model response is dependent on the radiative forcing function and the model sensitivity, thus a match between the reconstruction and the models could be

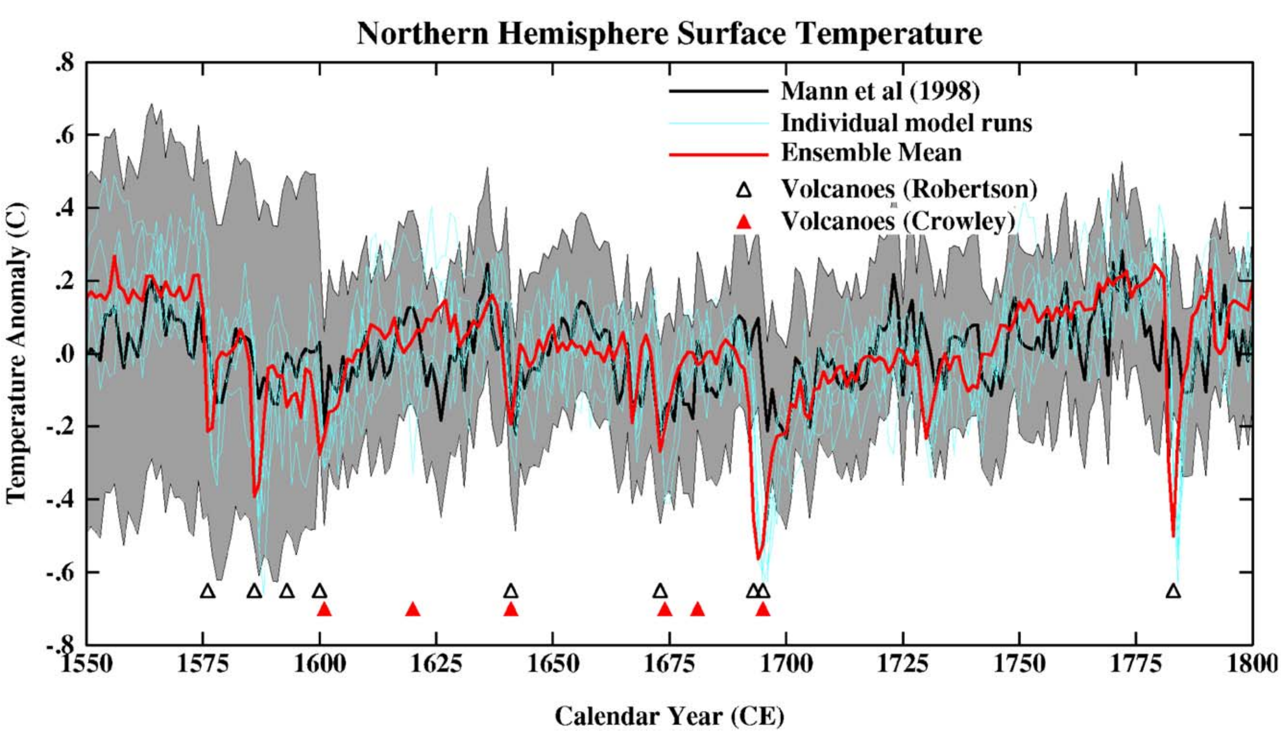

Fig. 6. 1550–1800 comparison between a proxy reconstruction (with error bars) and a "meta-ensemble" of six model runs. The years with significant volcanic forcing in the simulation are highlighted, together with significant volcanoes from the latest Crowley (2000) and Robertson et al. (2001) compilations.

fortuitous. It seems clear however, that the response to Laki (1783–1784) is over-estimated, as is the response to Tambora in 1816 (not shown), (ii) the background variability in NH mean SAT is similar in the proxy record ($\sigma = 0.1\,^{\circ}$C) and individual model results ($\sigma = 0.14$–$0.2\,^{\circ}$C), the higher values are associated with two runs with relatively shallow oceans (1300 m) compared to 5000 m in the other runs and (iii) the long-term (decadal and longer) variability is reasonably captured in the models. However, while the NH mean temperature series are well correlated to the proxy reconstruction ($r^2 = 0.1-0.26$ (individual runs), 0.23 (for the ensemble mean) ± 0.12 for 95% significance levels), the spatial patterns of variability (not shown) are not. This is a clear demonstration that while it appears to be relatively easy to get a reasonable estimate of the forced changes in hemispheric mean temperature (as with energy balance or intermediate models (Crowley, 2000; Bauer et al., 2003; Goosse et al., 2004)), the regional patterns of change are much more sensitive to the quality of the simulation and the physics contained therein (i.e. compare Shindell et al., 2001a; Waple et al., 2002). In particular, the improvements to the horizontal resolution, the inclusion of a reasonably resolved stratosphere, ocean dynamics, and interactive ozone are all likely to impact the regional changes.

6. Future directions

Over the next few years, computing power will increase, the physics included in models will become more sophisticated and longer simulations will be performed. At the same time, estimates of potential climate forcings will likely improve and more complete records of paleo-climate will become available. Forward modelling of proxy data (such as ice core isotopes or biomes) has the potential to further improve the match to the data (and/or highlight continuing problems in the models).

It can be seen from the results discussed in this paper that databases of paleo-climate change that can be used to isolate forcing mechanisms (such as for long term solar variability, or volcanic forcing), or provide time-slice reconstructions are extremely useful for modelling purposes. Since most paleo-records are single location time-series, this implies that more work will be needed to integrate different records into spatial networks such as the Mann et al. (1998) and Luterbacher et al. (2004) reconstructions, or the hydrological databases of Harrison et al. (2003). This task is quite difficult, due in part to problems of chronology, differences in time resolution, confounding non-climatic influences, biases in spatial sampling etc. However, this will be increasingly necessary in order to provide a solid framework for validating climate models and testing hypotheses. Some individual records (for instance, from ice cores) are also very useful targets, but matches to these are not sufficient to validate global climate changes without support from the spatial networks discussed above.

It is clear that the successes in modelling Holocene variability have relied on extending the range of feedbacks that are permitted in the models. For instance, ozone for the solar case, SST for almost all cases, vegetation dynamics for the mid-Holocene, and well-resolved stratospheric dynamics for capturing some of the atmospheric circulation changes. Further feedbacks will continue to be added into the models, such as interactive dust and other aerosols, carbon cycles and tropospheric chemistry. The addition of these biogeochemical modules in fully coupled models should allow for a fuller assessment, for instance, of whether GHG changes from around 8 kyr BP can conceivably be anthropogenic (as recently speculated (Ruddiman, 2003)), or whether they were likely just responding to changes in climate (Gerber et al., 2003).

For quasi-steady states longer than a few decades, but shorter than a few millennia, ocean dynamics cannot be expected to be in equilibrium with the forcing. Thus for multi-decadal to centennial changes, coupled transient simulations are required with all the extra work that this requires (Widmann and Tett, 2003; Gonzalez-Rouco et al., 2003). This is the timescale most frequently seen in Holocene paleo-data, and there is still some way to go in being able to validate models at this frequency band.

One thing that it is extremely unlikely is that models will suddenly start to provide exact matches to the time-series of any particular climate record. There is such a degree of weather 'noise' or intrinsic variability that no particular climate simulation, however sophisticated, will follow the same semi-random path that the actual climate took. Ensembles over many such simulations may average over much of this 'noise' but these averages by necessity smooth over many short term or spatial variations, which are none the less recorded in any particular proxy. Therefore while we are likely to have reasonable estimates of the changes in global or hemispheric mean temperature, accurate simulations of regional climate are likely to remain elusive. Simulations of hydrological variables (precipitation, drought) or extreme events are more problematic given their much greater variability in time and space. Some information can be gained from down-scaling large scale circulation behaviour (such as the NAO/AO, or ENSO variability) to the regional or even local scale (i.e. Thompson and Wallace, 2001), but that is likely to be the exception. Regional modelling, or zoomed models (with locally higher resolution) may play some role as well.

So to answer the question posed at the beginning of this paper; Climate modelling has hinted that, for the Holocene (i) although hemispheric mean temperature changes appear to be robust responses to forcings, the spatial patterns of response can be quite complicated and dependent on the physics contained within the models, (ii) modes of variability such as the NAO/AO (and potentially ENSO) can be affected by forcings, and (iii) the kinds of changes that have occurred are still posing challenges to our ability to model them. By further focusing on the variability of this period, we may therefore be able to increase our understanding of Holocene climate and improve our confidence that models are capable of modelling more recent and future climate changes.

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