23. RECENT DEVELOPMENTS IN HOLOCENE CLIMATE MODELLING

HANS RENSSEN (hans.renssen@geo.falw.vu.nl)

Faculty of Earth and Life Sciences
Vrije Universiteit Amsterdam
De Boelelaan 1085, NL-1081 HV Amsterdam
The Netherlands

PASCALE BRACONNOT (pasb@lsce.saclay.cea.fr)

Laboratoire des Sciences du Climat et de l' Environnement
UMR CEA-CNRS 1572
CEA Saclay bat 709
91191 Gif sur Yvette cedex
France

SIMON F.B. TETT (simon.tett@metoffice.com)

Hadley Centre for Climate Prediction and Research
(Reading Unit)
Met Office
Meteorology Building, University of Reading
Reading, RG6 5615
UK

HANS VON STORCH (hans.von.storch@gkss.de)

GKSS-Forschungszentrum Geesthacht GmbH
Max-Planck-Straße
D-21502 Geesthacht
Germany

S.L. WEBER (weber@knmi.nl)

Royal Netherlands Meteorological Institute KNMI
P.O. Box 201
NL-3730 AE De Bilt
The Netherlands

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Introduction

To improve our understanding of climate variability on decadal to centennial time-scales, it is crucial to use a hierarchy of climate models in addition to palaeoclimate reconstructions based on proxy data. Climate models give a physically consistent overview of the global climate on all time-scales. They are useful tools in palaeoclimatology, since: (i) they can be used to test hypotheses that have been inferred from palaeo-data; and (ii) they can provide plausible explanations of observed phenomena (e.g., Isarin and Renssen (1999), Kohfeld and Harrison (2000)). In recent years, considerable progress in palaeoclimate modelling has been made with the extensive use of models that consider the coupling of the different components of the climate system (atmosphere, ocean, sea-ice, vegetation). The aim of this paper is to inform the palaeo-data community on recent developments in palaeoclimate modelling, with special reference to the Holocene climate. In the first section, different model types and experiments are discussed, together with a short overview of Holocene climate modelling studies and differences between models and palaeo-data. In the second section, three important issues are further illustrated by discussing in detail three studies that use state-of-the-art models.

Different types of models

Different types of numerical models of the coupled atmosphere-ocean system vary in complexity. For an extensive overview of the history of numerical climate models, the reader is referred to McGuffie and Henderson-Sellers (2001). It suffices here to summarise the main current model types. It is convenient to make a distinction between complex climate models, simple climate models, and models of intermediate complexity.

Complex models: General Circulation Models

The standard complex models that are currently operational, are so-called atmosphere-ocean general circulation models (or AOGCMs, see e.g., McAvaney et al. (2001), IPCC (2001)). These models have been developed in the last 15 years, based on the combination of atmospheric and oceanic general circulation models (AGCMs and OGCMs, respectively) that have been developed since the 1970s. AOGCMs can be thought of consisting of a three-dimensional matrix with a global coverage, in which the dynamics of atmosphere and ocean are simulated based on physical laws. In the atmosphere, the models simulate the weather, including climatic parameters such as temperature, wind and humidity. In the oceanic part, ocean currents, temperature and salinity are computed. Furthermore, a sea-ice model is included that simulates sea-ice thermodynamics and dynamics. At the atmosphere-ocean interface, AOGCMs calculate the vertical exchange of momentum, heat and fresh water. Most AOGCMs include the simulation of both an annual and diurnal cycle. The typical horizontal resolution is 125 to 250 km, whereas the vertical resolution is generally about 20 layers for both the atmosphere and ocean (IPCC 2001). Processes that take place on smaller spatial scales (so-called sub-grid scale processes, e.g., cloud formation) cannot be computed, so that their effects have to be parameterized (i.e., described in a simple way based on the relationship with large-scale variables).

AOGCMs require the definition of some upper and lower boundary conditions. The upper boundary condition is the amount of solar radiation received at the top of the atmosphere, which varies as a function of the seasonal cycle. Together with changes in the concentration of atmospheric trace gases (such as CO_2 and CH_4), this variation in solar radiation can be considered as the driving force of the model. Lower boundary conditions are related to the characteristics of the earth's surface, such as bathymetry and land topography, but also land surface characteristics such as albedo and soil type. Recently developed AOGCMs incorporate complex land-surface schemes, some of which include a dynamical vegetation component (i.e., AOVGCMs, e.g., Cox et al. (2000)). In these AOVGCMs, the potential vegetation types (or biomes) are calculated based on bioclimatic parameters that determine their distribution, such as precipitation and the length of the growing season. Subsequently, the relevant characteristics of the biomes, such as surface albedo and roughness, are fed back into the atmospheric component. Consequently, in AOVGCMs it is no longer necessary to prescribe the vegetation characteristics as in AOGCMs. Similarly, in AOGCMs the temperatures of the ocean surface are calculated and not prescribed as in AGCMs. Thus, the lower boundary conditions that have to be prescribed depend on the components of the climate system that are included in the climate model. A drawback of AOGCMs is that their complexity requires time-consuming calculations on super-computers, which makes their application expensive, and transient experiments are limited to 500 to 1000 years around specific time periods in the past.

Simple models: Energy Balance Models

At the other end of the complexity spectrum are simple climate models. In this family of models, energy balance models (EBMs) are the most widely applied category in palaeoclimatology. Generally, latitude is the only dimension that is varying in EBMs. As their name suggests, EBMs calculate the energy balance for latitudinal zones, followed by a prediction of the surface temperature. Atmospheric dynamics are not included in EBMs, as heat transport is calculated on the basis of diffusion. Typically, the grids used are much coarser that applied in GCMs. As a consequence of the simplified calculation and lower resolution, EBMs run on simpler computers and can be applied on much longer time-scales. EBMs have been widely applied to study climate change at a hemispheric-to-global scale (e.g., Crowley (2000)).

Earth System Models of Intermediate Complexity

A recent line in model development is focused on building earth system models of intermediate complexity (or EMICs, e.g., Claussen et al. (2002)). EMICs incorporate climatic sub-systems (i.e., atmosphere, ocean, biosphere, land ice) in a simplified but efficient manner, while retaining those properties of the climate system that are relevant for specific research purposes (Pethoukov et al. 2000). Within the family of EMICs, large differences in complexity exist. Some EMICs include a two dimensional description of the climate system, with an EBM as atmospheric component. On the other hand, some EMICs are very close to GCMs in the sense that they contain a three-dimensional description of atmospheric dynamics on time-scales ranging from synoptic to millennial (e.g., Opsteegh et al.

(1998)). EMICs have in common that they incorporate a large number of parameterizations compared to GCMs. As a result of the reduced complexity, EMICs are relatively fast and are ideal to study the interaction between various components on centennial-to-millennial time-scales. GCMs and EMICs have been recently expanded with process-based forward models (transfer equations) that simulate the behaviour of proxy indicators themselves (e.g., biomes, glaciers, isotopes; see example in this chapter). With the inexorable increase in computing power through Moore's law, an EMIC may be an earlier and lower resolution version of a current AOGCM.

Different types of experiments

Equilibrium experiments

For a proper understanding of palaeoclimate modelling studies, it is important to know the difference between equilibrium and transient experiments. In equilibrium experiments, the boundary conditions are changed and kept time invariant (except for the annual and diurnal cycles). Where the focus is on the equilibrium response, the normal experimental design is to use a simplified and non-dynamical ocean. It is possible to perform such an equilibrium experiment with forcings for the present-day, and also for a period in the past (e.g., a snapshot experiment for 6 kyr BP). Thus an equilibrium experiment shows the equilibrium response of the climate system to the forcings at a given moment in time, giving insight into the 'background' internal variability of the model at various time-scales. Often an equilibrium experiment with present-day boundary conditions is called a "control" experiment, which is used as a reference experiment to which simulations with different boundary conditions are compared to analyse the model's sensitivity to a particular forcing. Later in this chapter an example of a control run is discussed.

Transient experiments

In transient experiments, time-dependant forcings are used. These forcings may be changes in solar radiation and greenhouse-gas concentrations, but also "slow" factors like orbital parameters if the interest is in climate change at relatively long time-scales (centuries to millennia). This implies that the model adjusts constantly to the 'external' changes. As will be shown in an example later on, a comparison of a forced run with the control run provides valuable information on the effect of forcings on the dynamics of the coupled system. Transient experiments on "palaeoclimatic" time-scales (multi-millennial) are only feasible with simple models and EMICs, as coupled GCMs are still too expensive to make such long runs.

Data assimilation

A third type of model experiment exists that is likely to play an important role in palaeoclimate modelling in the near future. In this type of simulation, the model is driven by data, as is commonly applied in modern weather forecasting. In so-called data assimilation, meteorological observations are used from selected sites (i.e., stations) to drive (or 'nudge') the climate model toward the state presented by the data. This method is not yet commonly applied in palaeoclimatology, but in the near future it will be used for past climates by introducing proxy data into the model (see von Storch et al. (2000)).

Holocene climate modelling studies

Climate model experiments on Holocene climate have been performed since the 1980's (e.g., Kutzbach (1981), Kutzbach and Otto-Bliesner (1982)). In the early stages, AGCMs were used to study the effects of orbital variations on the climate of the early-to-mid Holocene by performing snapshot experiments for 9, 6 and 3 kyr BP. The latter simulations were part of a series that started at the last glacial maximum (21 kyr BP) and went forward at 3 kyr intervals. They were done primarily within the Cooperative Holocene Mapping Project (COHMAP, e.g., Kutzbach and Street-Perrott (1985), Kutzbach and Guetter (1986), COHMAP members (1988), Wright et al. (1993)). The simulations for the early Holocene showed clearly the strengthening of the summer monsoons in response to orbital forcing, which is in agreement with reconstructed high lake levels in North Africa and Southeast Asia (e.g., Kutzbach and Street-Perrott (1985)).

A few years later, AGCMs were coupled to simplified ocean models (so-called mixed layer models) to simulate the 9 kyr BP climate, thereby taking the exchange between atmosphere and ocean into consideration (Mitchell et al. 1988; Kutzbach and Gallimore 1988). In the 1990's, the Palaeoclimate Modelling Intercomparison Project (PMIP, e.g., Joussaume and Taylor (1995)) aimed at comparing the performance of GCMs for climatic conditions different from today, by setting up equilibrium palaeoclimate simulations with identical sets of boundary conditions. They focussed at the 21 and 6 kyr BP time-slices. For the 6 kyr BP experiments it was assumed that the orbital forcing was the primary forcing factor affecting the climate system. Therefore, all other boundary conditions were kept at present-day values. The results of PMIP made it clear that all AGCMs capture the large scale features of the 6 kyr BP climate. However, some systematic discrepancies were found compared to data reconstructions. In Northern Africa, for instance, AGCMs clearly underestimated the strengthening of the summer monsoons as indicated by lake-level studies and palaeo-vegetation reconstructions (Joussaume et al. 1999). Further studies showed that part of this model-data mismatch can be attributed to the lack of oceanic and land-surface feedbacks in the atmospheric models (e.g., Kutzbach and Liu (1997), Broström et al. (1998), Harrison et al. (1998), Braconnot et al. (1999, 2000a, 2000b), DeNoblet et al. (2000), Texier et al. (1997, 2000)). However, even fully coupled AOGCMs and EMICs (including ocean and vegetation components) underestimate the African monsoon amplification at 6 kyr BP (Ganopolski et al. 1998; Hewitt and Mitchell 1998; McAvaney et al. 2001).

The 6 kyr BP simulation results for Europe showed that most PMIP models were able to reproduce the summer warming reconstructed using pollen data (Masson et al. 1999). Also, the reconstructed drier conditions in Northwestern Europe and wetter conditions in Southern Europe were simulated by some models. However, the characteristic winter temperature pattern of warming in Northeastern Europe was poorly reproduced, probaly due to different sea surface temperature patterns in the Atlantic Ocean at 6 kyr BP that were not included in the PMIP simulations (Masson et al. 1999). The reader is referred to Braconnot et al. (this volume) for an overview of the PMIP 6 kyr BP simulations carried out with coupled atmosphere-ocean models.

In addition to the PMIP-type equilibrium simulations, transient experiments have recently been completed, in which time-dependent forcings are applied to simulate the evolution of the Holocene climate. Fully coupled AOGCMs have been used to look at the last few hundred years (e.g., Cubasch et al. (1997), Tett et al. (1999)), whereas EMICs have been used to take the entire Holocene into account (Claussen et al. 1999; Weber 2001; Brovkin et al. 2002; Crucifix et al. 2002). Furthermore, EMICs are now used to study specific Holocene climate events in transient experiments, such as the 8.2 kyr BP cooling event (Renssen et al. 2001a; 2002) and the termination of the African Humid Period at ~6 kyr BP (Claussen et al. 1999; Renssen et al. 2003).

Scale differences between models and palaeo-data

If we want to make a useful comparison of palaeo-data with palaeoclimate model simulations, it is important that both represent similar spatial scales. In AOGCMs, the skill with which mean climate and climate variability is reliably simulated depends on the spatial scale (von Storch 1995). Large-scale features, reflecting global differential heating, the impact of continental-scale patterns of mountain ranges and land-sea contrasts are usually well reproduced, whereas smaller scales are not well described because of insufficient resolution of the physiographic details, in particular secondary mountain ranges (like the Alps), marginal seas (like the North Sea) or specifics of land use. The Third Assessment Report of the IPCC (IPCC 2001) found that AOGCMs fare realistically on spatial scales on 10^7 km^2 and more, while smaller scales should be considered less reliably simulated in global climate models (Giorgi et al. 2001). This limitation of climate models represents a significant obstacle for co-operation between palaeoclimatologists and climate modelers, as the former collect data and generate knowledge mostly on scales much smaller than 10^7 km^2, although there are exceptions to this. Thus comparison of climate model data with palaeoclimatic evidence is methodically difficult.

A solution to this dilemma is given by the "downscaling" concept. According to this concept, the state and statistics of smaller scales are a function of the state and statistics of the larger scales. The details of the "downscaling" function are determined by the physiographic details of the considered region or locality.

Two main classes of downscaling are in use (for an overview, refer to Giorgi et al. (2001)). First, for the "dynamical method" (Giorgi 1990) a regional climate model with a high spatial resolution is "nested" in a GCM with a relatively low resolution. The regional model is driven at the boundary, and possibly in the interior, by large-scale information taken from the global model. The skill of this approach was recently demonstrated by the "Big Brother" experiment of Denis et al. (2002). This technique has matured in the past few years (e.g., Whetton et al. (2001)) and has been applied to palaeoclimates (e.g., Hostetler et al. (1994), Renssen et al. (2001b)). Second, the "empirical method" (Wigley et al. 1990) assumes that an empirically determined link between the large-scales and the regional or local scales remains valid also under changing conditions. Then the link is used to interpret the large-scale output of a climate model. The method has been applied to various parameters, both meteorological (like precipitation) and proxy (like phenological dates) data.

The inverse of these downscaling methods is used for the reconstruction of relevant atmospheric patterns with the help of local proxy-data (e.g., Appenzeller et al. (1998), Crueger and von Storch (2001)). This method, named "upscaling", is also a key tool in attempts to force climate models to incorporate specific proxy-data based information (data-driven experiments, see above, von Storch et al. (2000)).

Recent examples

In this section, some recent developments in climate modelling are discussed that are relevant to palaeoclimatology. First, natural variability, as simulated in an AOGCM control run, is discussed. Second, the potential of performing a series of transient experiments with varying forcings is considered, using an example of AOGCM simulations for the 20th century. Third, forward modelling is further explained by discussing the simulation of Holocene glacier length variations driven by a transient EMIC simulation.

Example 1: internal climate variability in a control run

As already mentioned, the purpose of control runs is to simulate an equilibrium climate, given the repeated cycles of radiation and the constant values of greenhouse-gas concentrations, and its variability. Control runs have demonstrated the skill of state-of the-art AOGCMs to reproduce reliably the large-scale aspects of contemporary climate. Control runs prepared with different climate models produce rather similar results on large scales, whereas on regional scales there may be marked differences (McAvaney et al. 2001). The important point, often overlooked by geoscientists, is that the climate is varying for internal reasons independent of any time dependent forcing (except for the prescribed but constant annual insolation cycle). In fact, the proper way to conceptualise climate as a process is to think about it as a random process, whose characteristics are conditioned by certain external factors. When the latter are fixed, the characteristics are fixed, when they are time dependent (or transient), the characteristics are time dependent as well.

The random character of the climate system in state-of-the-art climate models is demonstrated in Figure 1, which displays a 1000-years time series of air temperature at the ground averaged over different regions. The model used is ECHO-G, an AOGCM developed in Hamburg (Legutke and Voss 1999). Apart from a slight downward trend, the time-series is characterised by irregular variations with significant deviations from the "normal" (the long-term mean). For instance, in the 6th century, prolonged warming appears in Europe, with maximum anomalies of more than 0.5 K. Comparison of the time-series for Europe and the Northern Hemisphere also shows that the amplitude of climate variability decreases when averaging over a larger area.

The dynamical background of these seemingly spontaneous variations lies mainly in the chaotic dynamics of the atmosphere. That non-linear systems can develop such irregular variations has been known for many decades. The difference from, for instance, Lorenz' famous 3-component system (Lorenz 1976) is that the dynamics in the real atmosphere and, to a lesser degree, in state-of-the-art climate models has very many degrees of freedom, with many of them behaving chaotically. The sum of these many chaotic processes is not a "simple" pattern of variations, as non-linear dynamical theorists like to show, but the truly "chaotic" variations as shown in Figure 1. Being in character multiple non-linear, the

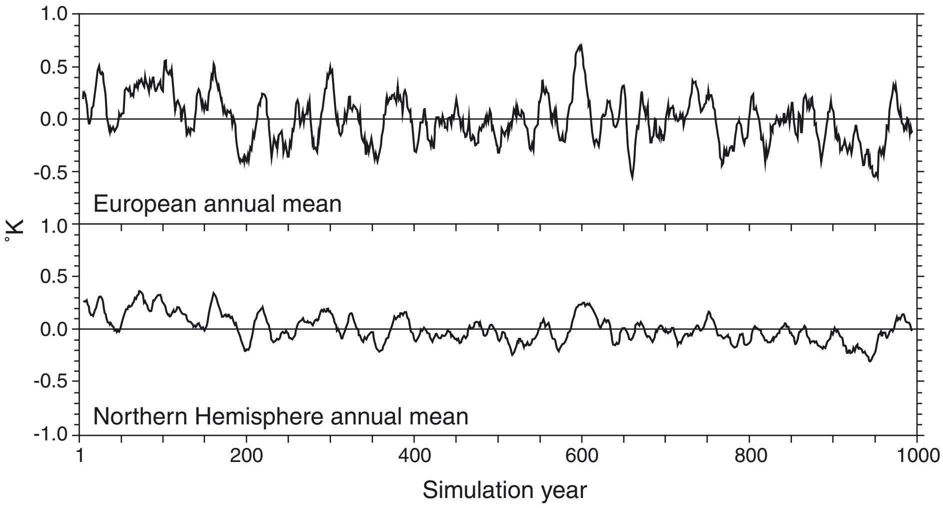

Figure 1. Time-series of area-averaged air temperature near the ground in a 1000 year control run (Zorita, pers. comm.) simulated by an AOGCM.

phenomena can consistently and efficiently be described with the mathematical concept of randomness (von Storch et al. 2001).

The "internally generated" variability, as simulated in control runs, makes the analysis of climate change signals difficult, as these signals are often masked by the noise. This is one of the problems that are faced in the detection of natural and anthropogenic signals in climate (Hasselmann 1979), in particular for assessing the reality of global warming (e.g., Hegerl et al. (1997)). For palaeoclimatology it is equally important to obtain a good signal-to-noise ratio in palaeoclimate simulations and climate reconstructions based on proxy records.

Example 2: simulating the 20th century using transient ensemble experiments

In this section we give an example of transient simulations carried out with a state-of-the art AOGCM. It is important for the palaeo-data community to recognise the difference between this type of experiment compared to the 'standard' PMIP-type equilibrium experiment, as it is expected that transient simulations will become more important in palaeoclimatology as computer power increases.

The numerical experiments were carried out at the Hadley Centre with the aim of understanding 20th century temperature change. The model used for these simulations was HadCM3 (Pope et al. 2000; Gordon et al. 2000); the third generation Hadley Centre coupled model (AOGCM). Four ensembles of experiments were run. Each ensemble used the same forcing but started from different initial conditions. These initial conditions represent the state of the coupled ocean-atmosphere system in the model at the start of the simulation. It is necessary to perform multiple experiments (i.e., ensembles) with slight changes in these initial conditions, because it is useful to consider the range of internally generated year-to-year variability in the model (see example 1). As a coupled climate model is sensitive to the initial state, each member of the ensemble gives a different, but equally valid, solution. Thus ensembles are needed to average out the noise

The average of each ensemble is used to determine the response to the forcing. Each ensemble was for the period 1860 to 1999. The four ensembles are:

- **GHG** – forcing with well-mixed greenhouse gases alone. Changes in CO_2, CH_4, NO and several (H)CFC's are included in these simulations.
- **Anthro** – forcing with changes in well-mixed greenhouse gases, anthropogenic sulphate aerosols and ozone. Both the direct scattering effect of aerosols and their indirect effect of increasing cloud albedo are included in the simulations. Note that the poorly understood indirect effect is the dominant negative forcing in these simulations. Ozone changes include both tropospheric increases and, since 1974, stratospheric decreases.
- **Natural** – forcing with changes in volcanic aerosol and solar irradiance changes.
- **All** – forcing with both natural and anthropogenic forcing.

More details of the experiments and their analysis can be found in Stott et al. (2000) and Tett et al. (2002). These experiments are compared with estimates of near-surface temperature changes from Parker et al. (1994, updated to 1999). The **All** ensemble average performs well in reproducing observed global-mean temperature changes (Fig. 2). None of the other simulations manages to reproduce the observed changes. The **All** simulation also captures the large-scale patterns of temperature change (not shown but see Stott et al. (2000)). Thus both natural and anthropogenic forcings are required to explain changes in 20th century temperatures. Closer examination of the **Anthro** simulations shows that very little net anthropogenic warming happened prior to the mid 1960s (some warming then cooling). By comparing this with the results from the **GHG** ensemble this seems to be due to the impacts of the other anthropogenic forcings and in particular the cooling effect of sulphate aerosol on climate. In the **Natural** ensemble the climate warms until the 1960s when it cools again. Over the 20th century as a whole natural forcings cause no net warming.

We extend the analysis by making the simple assumption that the observations are a linear combination of the simulated signals plus climate noise:

$$\mathbf{y} = \beta\mathbf{X} + \mathbf{u}$$

where $\mathbf{y}$ is the observations, $\mathbf{X}$ a matrix of simulated signals, $\mathbf{u}$ a realisation of climate noise and β a vector of amplitudes applied to the simulated signals. Climate noise is estimated from simulated climate variability and β is estimated using the observations and the ensemble average signals. We carry out an analysis of temperature change for the period 1897 to 1997 using decadal mean observed and simulated temperature anomalies. We considered only three signals — **GHG**, **Anthro** and **Natural**. The **GHG** and **Anthro** signals were transformed to give **G** (the effect of greenhouse gases alone) and **SO** (the effect of sulphates and ozone). We then estimate amplitudes and uncertainty ranges (5–95%) for these signals. Uncertainty estimates are computed from a control simulation (see above). From the amplitudes and uncertainty estimates we can estimate the linear-trend in temperature from different forcings. Figure 3 shows the results of this analysis for the entire century and for two fifty-year periods. Note that all trends, including their uncertainties,

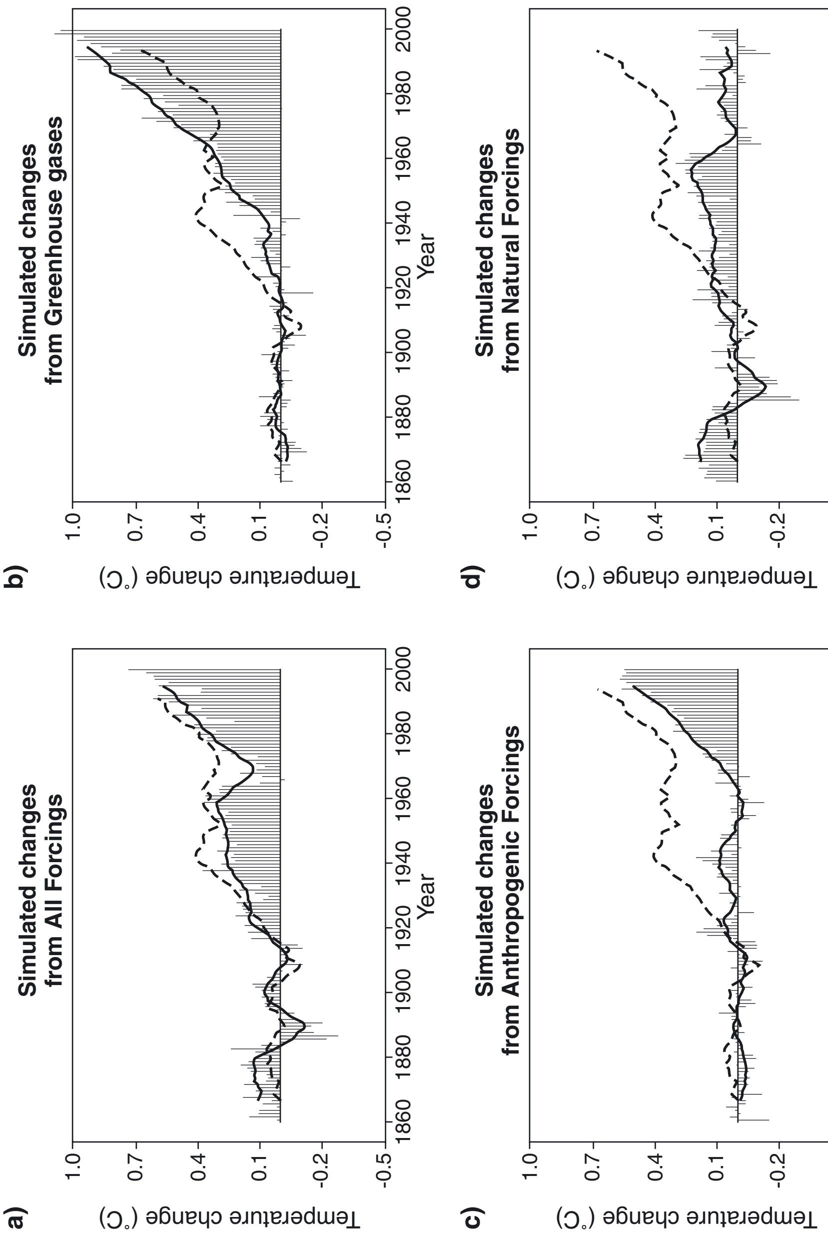

Figure 2. Time-series of simulated and observed global-mean temperatures. a) **All**, b) **GHG**, c) **Anthro**, d) **Natural** ensembles. All values are anomalies relative to 1890–1919 and the simulated values had data discarded where there were no observed data. The bars are annual-averages while the thick line shows a 10-year running mean. The observations are shown as a dashed line.

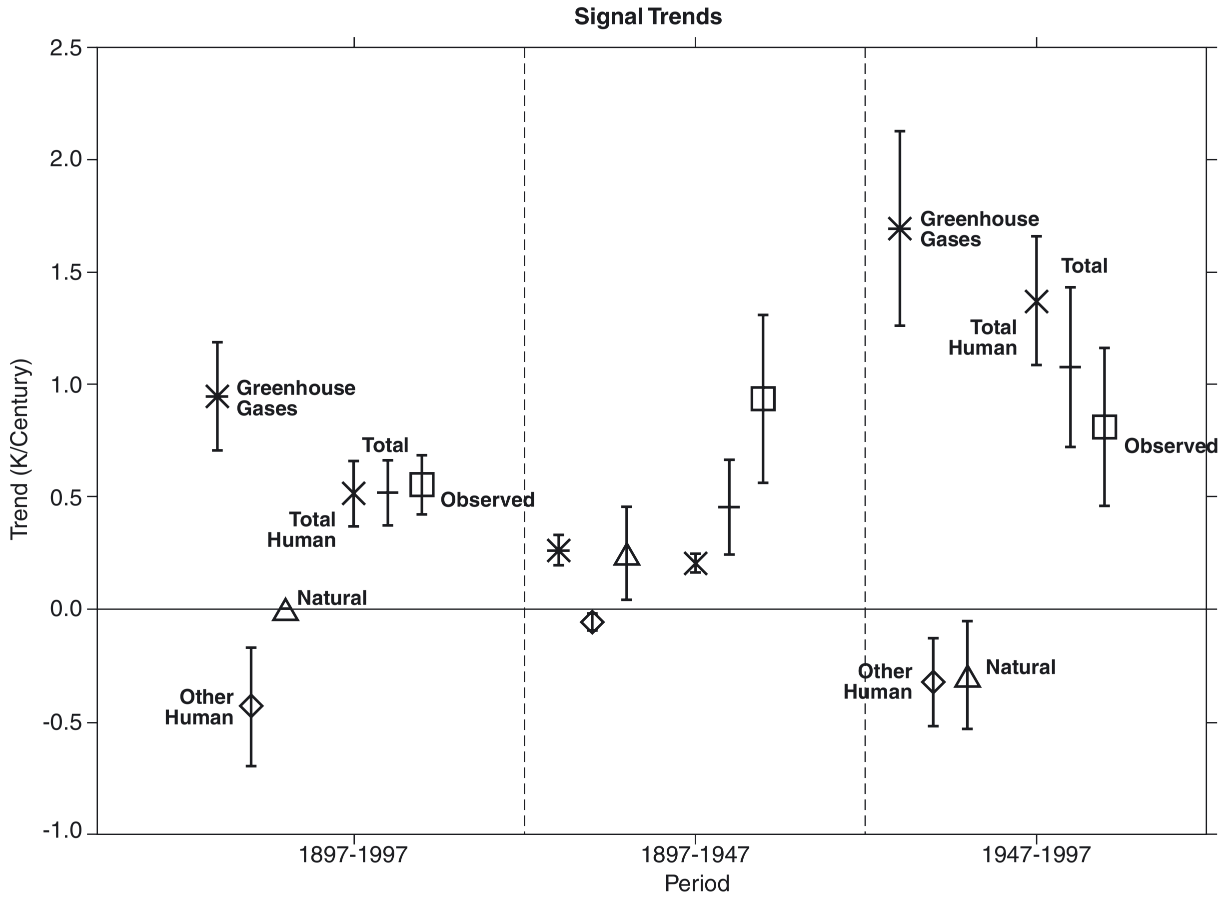

Figure 3. Best-estimate linear trend and uncertainty ranges (K/century) for G (left error bar with asterisk), other human (centre error bar with diamond), Natural (right error bar with triangle). The error bars show the 5 to 95% uncertainty ranges. The best-estimate trend is shown as a symbol at the centre of the bar. Also shown is the total anthropogenic trend (x), total trend (+) and observed trends (square). Uncertainty in the observations is estimated from internal climate variability simulated by the **Control** simulation.

exclude zero and are of the expected sign — all signals are detected. Thus we have positive evidence that all the signals we consider are present in the observations.

Over the 20th century anthropogenic forcings cause a warming trend of 0.5 ± 0.15 K/century ("total human" in Fig. 3). The trend due to greenhouse gases is 0.9 ± 0.24 K/century while remaining anthropogenic factors cool at a rate of 0.4 ± 0.26 K/century ("other human" in Fig. 3). The uncertainty in the total anthropogenic warming trend is less than the uncertainties in the individual trends as they are correlated with one another. Over the century natural forcings contribute little to the observed trend ("natural" in Fig. 3).

Our analysis considers only uncertainty in the amplitude of the simulated response and neglects uncertainty in the time-dependence of the forcing and in the spatial patterns of response, as well as uncertainties in the observations. However our best estimates are consistent with the observations. Furthermore in a single ensemble of simulations forced with both natural and anthropogenic forcings changes in simulated near-surface temperature are consistent with those observed (see Fig. 2) suggesting that those uncertainties may not be too great.

During the first half of the 20th century, greenhouse gases and natural forcings cause warming trends of about 0.2 to 0.3 K/century, while other anthropogenic factors produce negligible cooling trends (Fig. 3). Over the last half of the century, greenhouse gases warm the climate at a rate of 1.7 ± 0.43 K/century with natural forcings (largely volcanic aerosol) and other anthropogenic factors (mainly the indirect effect of sulphate aerosols) both causing an estimated cooling trend of about 0.3 ± 0.2 K/century. Thus, since 1947, changes in aerosol concentrations (anthropogenic and natural) have offset about a third of the greenhouse gas warming. Results presented here all rely on simulated internal variability to assess consistency or to estimate uncertainty ranges. We find that amplifying simulated internal variability by a factor of two still leads to a detection of the effects of greenhouse gases but no detection of other anthropogenic effects or those of natural forcings.

The above analysis shows how multiple transient simulations can be used to quantify the contribution of various forcing factors to observed climate change. It is expected that similar transient simulations will be performed in the near future for past centuries, e.g., the last 1000 years, including the Little Ice Age, or even earlier millennia. Consequently, this type of analysis will enable palaeoclimatologists to obtain a better understanding of the mechanisms behind climatic changes observed in proxy data.

Example 3: forward modelling

Clearly it is crucial to develop an understanding of proxy data in order to be able to interpret past climatic fluctuations registered by proxy data. To achieve this aim, models of proxy data are being developed and tested. These so-called "forward models" are process-based environmental models (physical, biological, chemical, or empirical), which are driven by climate model output to simulate a synthetic proxy record or time series, which can be directly compared with the actual proxy data (Weber and Von Storch 1999). Some forward models are dynamically incorporated within climate models, so that their results feed back into the atmospheric component and thus can have an effect on the simulated climate. Examples are the simulation of biomes in AOVGCMs, (Harrison et al. 1998), the inclusion of large lakes (Hostetler et al. 1993) and the incorporation of dust uptake, transport and deposition (e.g., Mahowald et al. (1999)). Other forward models operate off-line and are only fed by the climate model output (one-way procedure). Recent applications of this type are e.g., wetlands, lakes and rivers (Coe 1995; 1997), stable water isotopes (Jouzel et al. 2000; Werner et al. 2000), ocean sediment cores (Heinze 2001), glaciers (Reichert et al. 2001) and local sea level (van der Schrier et al. 2002). Such forward models consist of two components: one, a process-based model of the proxy parameter itself and two, a component transferring the climatic conditions to a forcing term driving the first component. Here we give an example where synthetic glacier length records are generated for the Holocene epoch using a glacier model coupled to the intermediate-complexity climate model ECBilt (Weber and Oerlemans 2003). The glacier model consists of a mass-balance component and an ice-flow component.

The climate model ECBilt is a dynamic, three-dimensional EMIC. It is computationally fast because its atmospheric component has a coarse vertical resolution (i.e., 3 layers) and simplified parameterisations. The Holocene climate is simulated in a 10,000-year transient experiment driven by orbital insolation changes (Weber 2001). All other boundary conditions (orography, concentration of trace gases and surface characteristics) are set to their present-day values. It is thus an idealised experiment, aimed at a better understanding of the glacier response to orbital forcing. The simulated climatic signal can be compared directly to proxy-based reconstructions of the climate during, for example, the mid-Holocene climate at 6 kyr BP. In addition, the implied glacier length changes provide an indirect validation of the simulated climatic signal.

We consider three glaciers, ranging from maritime to continental. The simulated glacier lengths are shown in Figure 4. The long-term trends in glacier length are associated with the orbital forcing. Nigardsbreen (Southern Norway) shows a phase of rapid expansion during the mid-Holocene, followed by more gradual growth. Here, the long-term trend in the annual mass balance is primarily determined by high summer temperatures in the early Holocene. There is a smaller contribution from enhanced winter precipitation. Both the temperature signal (Cheddadi et al. 1997) and the precipitation signal (Dahl and Nesje 1996) seem consistent with proxy data. At Rhonegletscher (the Swiss Alps) there is a maximum glacier extent at 3–5 kyr BP, which seems unlikely. ECBilt simulates warmer and wetter summers in the early Holocene. Proxy data confirm the temperature signal (Cheddadi et al. 1997), but they are not conclusive with respect to the precipitation signal. Earlier time-slice model experiments also differ widely in the simulated water budget response (Masson et al. 1999). The present results suggest that the simulated precipitation response is not realistic.

The climatic response at the Abramov glacier (Kirghizia) is characterised by enhanced summer precipitation, associated with a northward extension of the Indian monsoon reaching its maximum at 6 kyr BP. This signal is well known (Joussaume et al. 1999). The timing of the maximum, which is consistent with lake level data (Harrison et al. 1996), can be understood from the lagged response of the monsoon system to the orbital forcing. September precipitation, which dominates the transient signal, peaks around 6 kyr BP following August insolation. The simulated glacier length reaches a pronounced post-glacial maximum at that time. This result indicates that the northward extension of the monsoon is coincident with glacial advance, which should be recognisable in the field by the remains of end moraines.

There is considerable variability on time-scales shorter than millennial in the simulated glacier length records. These length variations are due to internal climatic variability. They

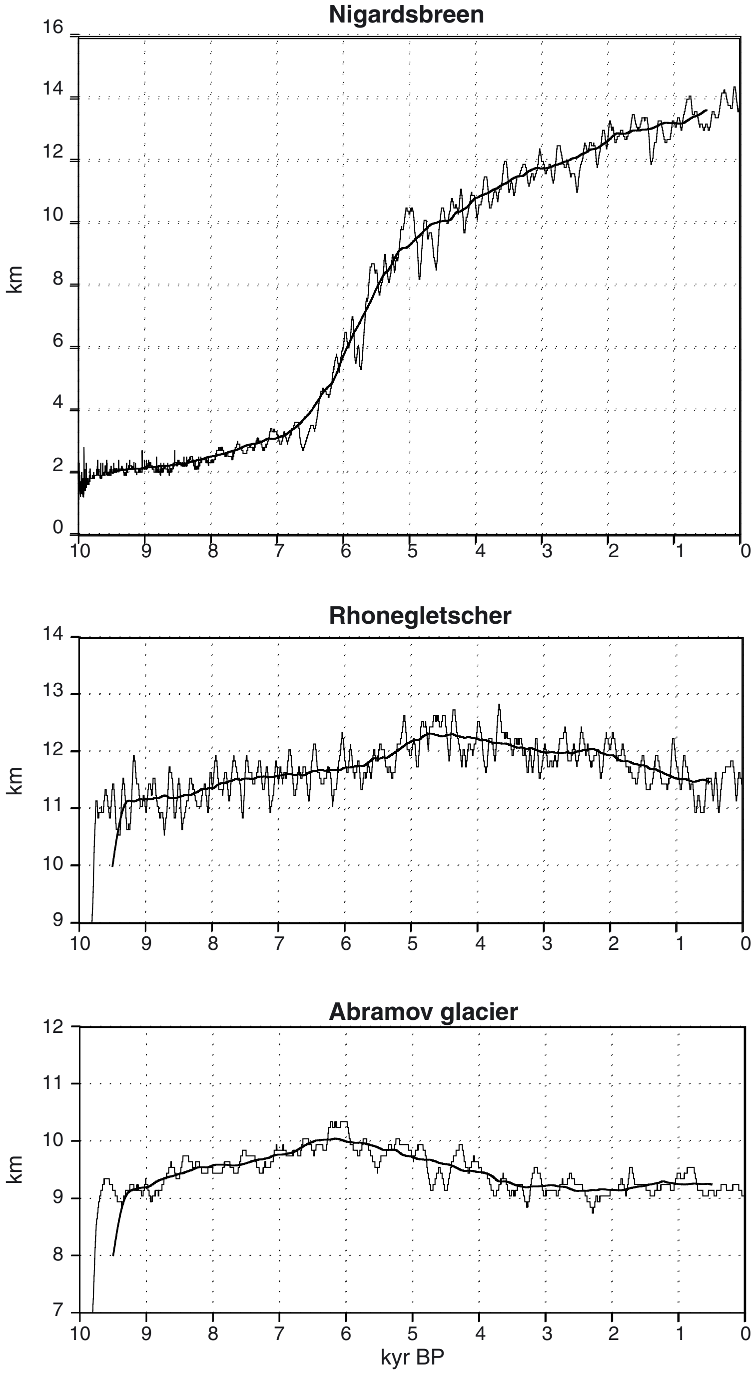

Figure 4. The glacier length (in km) as simulated by ECBilt for the three glaciers Nigardsbreen, Rhonegletscher and Abramov glacier as a function of time (in kyr BP). The long-term trends, as given by the 999-yr running mean, are also shown.

are typically asynchronous among the three different glaciers. Length variations can be shown to behave as a lagged moving-average process, with a glacier-specific memory.

Conclusion

This paper has discussed recent developments in Holocene climate modelling that are important for narrowing the gap between climate modelling and palaeoclimatology. The evolution of climate models is rapid; a process that is partly governed by the development of increasingly fast computers and partly by the needs of the climate community. Standard general circulation models are now describing the coupled atmosphere-ocean system, and dynamical vegetation models are presently being incorporated. This means that, at present, the dynamics of the climate system can be simulated far more completely than by atmospheric GCMs which were the standard models until recently. A more complete incorporation of the interaction of several components of the climate system is also the aim of recently developed EMICs. The use of coupled models also enables the performance of more realistic time-dependent (or transient) experiments that are driven by time-varying forcings, as opposed to the equilibrium experiments with AGCMs that have been dominating palaeoclimate research until recently. Transient experiments produce time-varying climate variability that can be compared with observations. Another development relevant to palaeoclimatologists is forward modelling. Forward models use climatic input from climate models to calculate the dynamics of the proxy data, so that it becomes possible to make a more direct data-model comparison.

Summary

The aim of this paper is to inform the palaeo-data community about recent developments in palaeoclimate modelling, with special reference to the Holocene climate. First, these developments are discussed in some detail, and subsequently the main issues are illustrated with the aid of examples. In a first example, natural variability in a coupled climate model simulation without a change in forcing is discussed. A second example discusses the response of a coupled climate model to transient forcings, thus clarifying the difference between natural "background" climatic noise and forced climate variability. A third example discusses the forward modelling approach. In particular, the simulation of glacier length variations for the last 10,000 years is shown in detail.

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