Predicting future climate change: lessons from palaeoclimatology

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Abstract
Sophisticated General Circulation Models (GCMs) are used to model the climate and are the principal tools used to predict future climate change. The outputs of these models have important policy implications but the uncertainties associated with the predictions and model inadequacy are not always appreciated by user-groups. This paper assesses some of the limitations associated with GCMs, including their limited ability in resolving rapid climate change in the past. The paper suggests that the variability of past climates gives us clues to the potential for rapid climate change in the future and outlines several locations in the Earth's natural systems where such rapid change might be identified and propagated.

Introduction

Nonlinearity is an important characteristic of all global environmental systems, occurs at a range of temporal and spatial scales and controls the climate, and human responses to it. Implicit in such non-linear systems are the existence of multiple equilibria and thresholds (so-called tipping points) which result in the system exhibiting rapid and unpredictable change. The interactions between the non-linear atmosphere, hydrosphere, biosphere and geosphere, and society are complex and form one of the main sources of uncertainty in our predictions of future climate and environmental change. It is clear that such uncertainty is of intense interest to the insurance industry (and business more generally) as rapid change would impose a very severe burden on the financial viability of economic systems. There are plausible system thresholds that could overwhelm the capabilities of the industry to respond. However, the insurance industry is only partly aware of the nature of threshold responses in the climate, the non-linear feedbacks that these create and the probabilities of rapid change.

Reconstruction of past climate change has identified a number of threshold responses of the climate system where climate changed rapidly in response to internal and external forcings.

There is compelling evidence that the climate system in the past (and perhaps in the present) displays emergent behaviour and General Circulation Models (GCMs) are not always able to mimic such behaviour. Emergence is a characteristic of dynamic systems where the large-scale behaviour of the system is effectively independent of the behaviour of the small-scale components of that system. Scientific attempts to unravel the complexity of multivariate systems have tended to follow reductionist paths and the use of GCMs is a typical example of this. However, the presence of emergence in the climate system means that reductionism may not be a valid response to complexity in natural systems.

Two related questions therefore present themselves: First, does this mean that our GCMs are unable to account for the likely future dynamic evolution of the climate system since the system is likely to display emergence? Second, is their inability to mimic the rapid climate shifts in the past (and some broad scale elements of the present climate system) mean that they are not preparing us for threshold responses?

Clearly, these questions have important policy implications as climate change can be both rapid and unpredictable. As a result, this paper highlights the nature of these changes; assesses the degree to which computer modeling of climate can provide realistic prediction of future climate for end-users; highlight some of the technical and practical problems in predicting the evolution of complex systems like the global climate and discusses the lessons from palaeoclimate reconstructions in providing a base from which future climate change can be understood.

Climate Change

There is a scientific consensus (Oreskes 2004.) that the mean surface temperature of the Earth has warmed in recent decades, and that the warming amounts to around 0.8 C since the beginning of the 20th century (IPCC 2007). From this, GISS estimate that 2005 was the warmest year since reliable instrumental measurements become available, although the WMO and CRU place it just behind 1998. Detection and attribution studies show that there is high probability (at least 90%) that this warming is largely the result of anthropogenic emissions of greenhouse gases (mainly CO_2) in the troposphere and that the amount and rate of warming are outside of the range of natural variation and unprecedented for the Holocene (the last 11,000 years). Continued warming is expected to have important consequences for a range of Earth systems (including the atmosphere, cryosphere, oceans, hydrological systems and the biosphere) and there are compelling reasons to expect increases in the magnitude and frequency of some natural hazards such as floods (Huntington 2006), droughts (Mason and Goddard 2001) and landslides (eg Fischer et al. 2006), and increases in the intensity of tropical cyclones (Emanuel 2005). There are also concerns about the stability of several of the large ice sheets on Earth (e.g. Overpeck et al. 2006) as these have the ability to impact upon global sea levels and regulate ocean currents.

The pattern and extent of future warming has enormously important policy implications for governments and business. The only way in which such predictions can be made is through the use of General Circulation Models (GCMs).

Predictions of future climate change

Prediction problems can be broken down into two types: initial value and boundary value problems. In climate prediction, the initial value problem tries to determine the evolution of the climate system once we know its initial state; put simply the system may display chaos and successful prediction is not possible a short time after the system has begun to evolve. This is largely why weather forecast errors saturate within a few days. The boundary value problem tries to assess climate change after the system has been forced by an external parameter, such as changes in radiative forcing associated with rising Greenhouse Gas (GHG) concentrations, and certain levels of prediction are possible for future climate states.

Climate models are used to make predictions and projections of future climate change. These have developed from the simple energy balance models of the 1960s to today's Atmosphere Ocean General Circulation Models (AOGCM) and Earth System Models (ESM) which incorporate sophisticated understandings of the coupled behaviour of the atmosphere, oceans and cryosphere, and terrestrial processes such as carbon cycling and land-use changes. Parts of the models have been used to develop weather forecasting tools, but climate change is a more difficult problem as verification and development of the models from real-world observation is restricted by the long timeframe over which the models are forecasting. In addition, the complex nature of the global climate system means that even the most sophisticated GCMs are unable to always model successfully important elements of the climate including the location of certain storm tracks, regional responses and ice-sheet dynamics. These limitations are not always considered by policy makers. As a result, it is timely to consider some of the uncertainties and limitations associated with climate predictions using GCMs, and suggest some ways forward.

Problems with GCMs

There are a number of sources of uncertainty in modeling attempts to explore future climate change (Collins 2002; Stainforth et al. submitted) and these can be sub-divided into two major categories: theoretical limitations and practical limitations in model predictability .

Theoretical limitations

The first major uncertainty associated with the operation and output of GCMs is ***Initial Condition Uncertainty.*** At macroscopic scales IC uncertainty is produced by our incomplete understanding of the large-scale state of the system and is partly a function of inadequate observations of relevant parts of the climate system. For instance, precipitation patterns in Western Europe are linked to changes in the North Atlantic Oscillation which is a ratio of the pressure difference between the Azores High and the Icelandic Low pressure systems. These systems also probably interact with, and are affected by, the thermohaline circulation in the North Atlantic and different macroscopic initial conditions in this system will produce differences in model predictions at timescales of decades and longer. A full understanding of the operation of such large-scale oceanic systems is not presently achievable. At smaller scales, microscopic IC produces an irreducible element to uncertainty, which is not solvable by increasing the power of the model or its resolution (Nicolis and Prigogine 1989; Svozil 1993; Nicolis 1996). Although such uncertainty derives from the imprecision of our knowledge of initial conditions at small scales, it may not be a barrier to successful prediction of the climate system if such initial condition uncertainties are not magnified during system evolution to affect the large-scale behaviour of the system. However, if the macroscopic behaviour of the system is sensitively dependent upon the microscopic, then the whole system may be expected to display non-linearity and may be chaotic. There is a vigorous debate concerning the extent to which the large-scale climate system is chaotic, but there are also suggestions that the large-scale behaviour of elements of the climate system are effectively isolated from the microscopic behaviour, although this may be less true at regional scales. This has practical implications for the success of climate modeling which are discussed later.

Exploring IC uncertainty in climate prediction is made difficult because we only have one climate system which we can observe and our knowledge of how it has behaved in the past in response to forcings (which are themselves subject to observational uncertainties) are only partly known. More seriously, we have no knowledge of how it responds to future forcings (Stainforth et al. in submission) although we can estimate this using our knowledge of the past.

In essence, uncertainties in predicting future climate change come down to the tension between reductionist and emergent responses to highly complex, coupled and dynamic systems. Such issues have been discussed in a number of sciences including physics (eg Anderson 1972), ecology (Willis and Whittaker 2002) and geomorphology (Harrison 2001). Reductionism argues that determinist approaches to science and positivist views of causation are the appropriate methodologies for exploring complex, many-body systems. It suggests that causal and effect relations are bound by linearity and that such 'one-to-one' relationships (Bohm 1957) thus allows perfect prediction and retrodiction. This paradigm suggests that everything can be reduced to a set of quantitative laws governing the behaviour of basic forces and a few basic elementary particles.

However, this view tends to break down when the system under study displays non-linearity. Non-linear systems are those whose causal powers are not derivable from the aggregations of lower level behaviour, since these cannot be known (Kim 1992). In this case complex systems may display emergent behaviour (Goldenfeld and Kadanoff 1999). Emergence can be seen as the ability of complex (often non-linear) systems to create dissipative structures (areas where the system exchanges matter and energy at far-from equilibrium situations), at scales which may not be understood by reductionist methodologies. These structures will exist in any open systems characterised by irreversible processes and emergence is possible because the system is insensitive to changes in initial conditions and is therefore stable with respect to changes at the microscopic scale and to changes in extrinsic conditions (and is therefore in equilibrium). As Schweber (1993, 36) put it "Emergence refers to properties of the solutions - in particular, the properties that are not readily apparent from the equations". We can see that microscopic variations in a system are inconsequential for understanding the large-scale system. Thus, by integrating out the high-frequency and short-wavelength parameters of a many-body system (associated with the microscale variability) we are able to describe the dynamics of the large-scale system. The practical implications of this suggest that there exists system behaviour and structures which are not amenable to explanation or prediction by reductionist methodologies (the latter which are embedded in the operation of GCMs). Our ability to predict and analyse the future evolution of such large-scale emergent structures (perhaps including El Nino Southern Oscillations (ENSO) and the Thermohaline Circulation (THC) in the climate system may therefore not be increased simply by increasing the sophistication and power of our computer models. Indeed, we can demonstrate this; in some cases ENSO events are not predictable beyond 8-12 months using an ensemble of coupled atmosphere-ocean models (see Collins et al. 2002).

From this, it is clear that the scale of enquiry is central to our future understanding of the climate system with certain elements of the climate system showing reasonably predictable responses to greenhouse gas forcings (eg the atmosphere and the upper ocean), while other, large-scale, parts of the system (eg ENSO) display non-linearity and predictability is lost on short timescales.

Practical limitations

Forcing uncertainty is the uncertainty associated with the climate response to future atmospheric levels of greenhouse gases. Since the future levels of these depend on how economic and political systems develop, there is considerable uncertainty concerning the amount of warming that will occur in the future. Despite this, there is general agreement that anthropogenic greenhouse gases will be the dominant driver of the global climate for the rest of the 21st century (IPCC 2007) and the influences of these may have recently accelerated. Rahmstorf et al. (2007) have argued that global warming since 1990 has accelerated in line with the upper limits predicted by the IPCC TAR (2001) although there are opposing views on this.

Model Inadequacy is a more serious problem, as it may mask future rapid shifts in climate which have the potential to overwhelm ecological, political, economic and social systems. There are three major limitations with the models (Stainforth et al. in submission). First, our understanding of all the processes that drive climate change are themselves inadequate, and this is reflected in limitations in the models. Gaps in our understanding include ice sheet dynamics under conditions of warming; the response of permafrost to warming, the response of oceanic circulation to changes in temperature and salinity. However, this is a significant failing as it is in these systems that climate change may produce rapid, non-linear and essentially irreversible responses. Second, a number of relevant climate processes occur at smaller scales than the model grid (including cloud processes, bulk surface fluxes, land-use changes) and these must be parameterized. This necessarily introduces uncertainties into the model, as we don't always have good understanding of the behaviour of such sub-grid processes, in particular their response to feedbacks. Finally, the coarse-grained structure of GCMs and parameterization means that there are important climatic processes whose behaviour is not well-modelled (including ENSO and the formation of tropical cyclones).

Model Uncertainty is the final practical limitation to predictions derived from GCMs and is represented by the range of plausible responses of the climate system to the forcings and under differing scenarios, and driven by uncertainties in our understanding of the operation of the system. To counteract this, there have been a large number of attempts to use ensembles of model runs to assess the range of likely future climate change (eg Allen and Stainforth 2002; Stainforth et al. 2005). However, as Stainforth et al. (submitted) caution, the ensembles

> "are not independent samples of the range of possible models, thanks to the substantial collaborations between modelling centres, the publication of results in the literature, understanding based on the same text books and educational facilities and even influences of the same computer hardware".

As a result, the shape of the probability distribution functions may not represent accurately the likely probabilities of real-world responses to climate forcings.

Lessons from palaeoclimatology

Given the range of uncertainties outlined above in the predictions and projections of future climate change from the outputs of GCMs, what opportunities are left to us to obtain information about future change? An important philosophical lesson from geology is that of uniformitarianism: which states that the key to the past lies in the present (Hutton 1795). We can overturn this doctrine and state that as far as the climate system is involved, a crucial philosophical viewpoint must be that the key to the present (and future) lies in the past. Consequently, understanding how past climates have changed in response to various forcings and the response of other systems may give us insight into future climate change. Unfortunately, the pattern that emerges does not give us grounds for optimism.

By using a number of proxies for climate change we can (with varying degrees of success) reconstruct past climates. Our reconstructions employ a range of techniques including dendrochronology, lichenometry, analysis of ice, marine and lake cores, coral records and reconstructions based upon cave, and loess sediments. In certain ways these become necessarily less accurate the further back in time we travel. Our most accurate reconstructions are from the Late Quaternary, which can be sub-divided into the late Pleistocene (the time period covering the Last Glacial Maximum around 26-18ka BP) and the Holocene from around 11.4ka BP to the present day.

The Pleistocene

The Last Glacial Maximum (LGM) of the late Pleistocene is a period when large areas of the high and mid-latitudes were covered by continental-scale ice sheets. After 18ka BP the ice sheets started to melt, and this process continued into the early part of the Holocene. The climates of this time have been reconstructed using a number of proxy records, including oxygen isotopes from ice cores (eg Grootes et al. 1993**)**, foraminifera from marine cores (eg Dowsett 1991**)** and mollusk records from thick terrestrial loess deposits (eg Keen 1995). From these, it is clear that climate change during the LGM in the period leading up to the Holocene was rapid and significant (Alley et al. 2003). The largest shift in temperature at this time occurred during the Younger Dryas stadial between 11.6 to 12.9 ka BP; the start of which was recorded by a reduction in Greenland temperatures of 15°C (Alley et al. 1993). At the end of the Younger Dryas temperatures rose as sharply, with most of the temperature change occurring over a decade or less. The likely cause of this cooling event appears to be rapid drainage of glacier-dammed lakes into the North Atlantic from the St Lawrence, which caused a rapid shutdown in the THC (Broecker 1997; 1998), although other mechanisms have been proposed (e.g. Renssen et al. 2000). This is only the most severe of the rapid climate changes that occurred during the deglaciation of the ice sheets; and forms one of a number of ice-rafting events from ice masses located over eastern Canada called Heinrich Events (Bond et al. 1992; Andrews 1998). The extent to which the Younger Dryas was a global event is debated. There are glacier advances dated to this period from various parts of the Southern Hemisphere including New Zealand (eg Denton and Hendy 1994) and Patagonia (eg Glasser et al. 2004; Glasser et al 2006) but these may also have been a response to a regional cooling event at this time known as the Antarctic Cold Reversal.

The Holocene

Until quite recently, the Holocene was seen as a time of relatively stable climate, which allowed complex civilizations to develop. However, since the mid-1990s a number of reconstructions of late Holocene climate have shown that the climate varied considerably regionally at decadal and centennial timescales (eg Mann and Jones 2003; Moberg et al. 2005). Two important periods of climate change occurred during the last 1000 years: the Medieval Warm Period, and the Little Ice Age. The Medieval Warm Period (MWP) occurred from around AD 950-1200 and was most probably restricted to parts of the Northern hemisphere (Jones and Mann 2004) and is mirrored by Viking occupation of southern and eastern Greenland. Regionally, temperatures were probably as high as during the latter part of the twentieth century, although there is no evidence to show that this warming was global, nor that warming was regionally synchronous (Bradley et al. 2003). The Little Ice Age, which lasted from about AD 1300 until AD 1900 formed the largest glacial advances of the Holocene and was marked by periods of famine and disease in Europe and North America (Ruddiman 2003; Grove 2004). Many palaeoclimatologists argue that cooling was largely restricted to parts of the Northern Hemisphere (eg Mann et al. 1999) although there is evidence to show that glaciers advanced and receded in the Southern Hemisphere during this time (eg Grove 2004; Koch and Kilian 2005; Harrison et al. 2007).

Rapid change

The lessons from palaeoclimatology are:

1. the models do not always recreate the rapid shifts in climate that we know occurred during the Pleistocene/Holocene transition.

2. the climate appears to be regionally and globally liable to rapid shifts when the forcings have increased rapidly; the implications of this are that recent rapid atmospheric forcing by greenhouse gas emissions make more likely rapid climate change which will be abrupt on the timescale of human economic, cultural and political systems and global ecosystems (NAS 2002).

3. even during the relatively stable climate of the Holocene change has been regionally variable and this has had important impacts on human systems.

4. the climate in the past has undergone rapid change and significant climate system elements have become reorganized in response to feedbacks and thresholds in the system, which are poorly understood and poorly modeled.

5. many instances in the past of rapid climate change have probably been the result of interactions between ice sheets and oceans during deglaciation and under conditions of positive climate forcing, conditions which are beginning to operate today. This is not well understood, and modeling ice sheet dynamics is a significant failing in ice sheet models and in GCMs.

Implications for the ice sheets

The response of the major ice sheets on earth to continued warming is a major uncertainty, but it has the potential to create rapid and catastrophic sea level change and interrupt major oceanic and atmospheric circulations (Oppenheimer 1998). It is therefore worth discussing this in more detail. Our understanding of the dynamic evolution of the ice sheets under conditions of present and future AGW is limited by the low level of sophistication of numerical ice sheet models (Alley et al. 2005). These models are inadequate largely because the empirical data with which to tune them are limited, and our physical understanding of the processes operating at the junction between ice sheets and the ocean (the grounding line) are low (Vieli and Payne 2005). We also have limited understanding of basal conditions under large parts of the ice sheets and under the outlet glaciers and ice streams that drain them (eg De Angelis and Skvarca 2003). More research on subglacial rheology, debris concentrations and basal temperature gradients are required as these affect basal shear stresses, basal sliding, water availability and the likelihood of rapid sediment deformation. All of these variables have the potential to affect the dynamic response of the ice sheets to warming.

Consequently, we have been forced to assess ice sheet stability largely on the basis of current observations and assessments of their behaviour in the past.

Thanks to recent advances in Quaternary science we have detailed information on the nature and timing of deglacial events at the end of the Pleistocene which may inform us to the likely response of the ice sheets under conditions of present and future AGW. From palaeoglaciological investigations, we can suggest that the ice sheets with major terrestrial margins (such as large parts of the Laurentide and Fenno-Scandian ice sheet) were largely stable and melted slowly following the LGM. Those ice margins which were significantly coupled to the ocean (such as that which drained through the Hudson Strait) disintegrated rapidly and this resulted in rapid discharge of icebergs and significant sea level rise from displacement.

During the deglaciation between the LGM and the early Holocene, large sections of the Northern Hemisphere ice sheets underwent periods of dynamic change and at this time sea levels rose rapidly. Two elements of this change were Dansgaard-Oeschger (D-O) events and Heinrich events. D-O events were periods of repeated regional rapid warming and gradual cooling which occurred with an underlying periodicity of 1,500 years or so (Rahmstorf 2003). At least 20 D-O events occurred during the last glaciation and, although evidence for these is found in many regions of the world (Voelker 2002), their strongest effect is seen in the Greenland ice cores where temperatures rose abruptly between 10 and 15° C . While these events are therefore more rapid and of higher magnitude than anything predicted by GCMs over the next century, Rial et al. (2004) ask: "Could present global warming be just the beginning of one of those natural, abrupt warming episodes, perhaps exacerbated (or triggered) by anthropogenic CO_2 emissions? Since there is no reliable mechanism that explains or predicts the D/O, it is not clear whether the warming events occur only during an ice age or can also occur during an interglacial, such as the present".

How are the ice sheets behaving now?

Over the last few years there have been signs that the West Antarctic and Greenland ice sheets have been displaying anomalous behaviour.

The stability of the West Antarctic Ice Sheet (WAIS) is seen as one of the key issues in deciding whether 'dangerous climate change' (Oppenheimer and Alley 2004; 2005) is occurring or likely. Until recently, there has been little evidence to show the nature of climate change from the centre of the Antarctic continent, and the data sets that did exist are of short duration. More recently, however, gravitational studies by Velicogna and Wahr (2006) have shown that the Antarctic ice sheet decreased its mass at a rate of 152 ± 80 km3/year of ice (equivalent to 0.4 ± 0.2 mm/year of global sea level rise) between 2002 and 2005, and most of this ice loss came from the WAIS.

From the Antarctic peninsula and ice shelves around the margins of the West Antarctic ice sheet we now have evidence that significant warming is occurring (eg Turner et al. 2005). In this region, much work has concentrated on the ice shelves which buttress the ice sheet and on the behaviour of its outlet glaciers, especially Thwaites and Pine Island Glaciers. These large glaciers drain the WAIS into the Amundsen Sea and their behaviour is closely linked with the overall stability of the WAIS (Shepherd et al 2004). In 2001 Rignot showed that the grounding line on Thwaites Glacier had retreated 1.4km between 1992 and 1996, and the lower part of the glacier had downwasted at 1.4m/year. Rignot et al. (2002) further reported that Pine Island Glacier had increased its velocity over 150km of the glacier by 18% between 1992 and 2000, downwasted at 1.6m/year and seen a 5km grounding line retreat between 1992 and 1996. We can suggest that the recent behaviour of Pine Island Glacier may be the initiation of a shift in the long-term stability of the WAIS (eg Payne et al. 2004) and this leads glaciologists to suggest that there is a strong coupling between the interior of the ice sheet and the surrounding ocean.

The large ice shelves that buttress the outlet glaciers have also undergone recent change. In 2002 the Larsen B ice shelf collapsed and this is probably the first time that this ice shelf has collapsed during the Holocene (Domack et al. 2005). With the removal of horizontal compressive forces by ice shelf collapse, the glaciers draining the Antarctic Peninsula have increased their velocities by factors of two to eight with downwasting rates in the order of 10s metres/year (Scambos et al. 2004). Consequently, overall ice loss is now greater than 27 cubic km/year (Rignot et al. 2004). Understanding and assessing the future behaviour of these ice masses is difficult, but their dynamic evolution is likely to be sensitively dependent upon the configuration of their subglacial bedrock channels through which they flow, and it appears that the glaciers flow into ice shelves several hundreds of metres thicker than was previously believed (Thomas et al. 2004). These subglacial conduits are thus the routes by which ice can be discharged from the central parts of the WAIS if ice sheet collapse is initiated and this was suggested by Mercer (1978).

Unlike the WAIS , the Greenland ice sheet is not at or below sea level and, as a result, is unlikely to be collapse quickly. However, parts of the ice sheet are clearly unstable; a number of the large outlet glaciers draining the ice sheet have recently increased their velocities and ice discharge rates and this is helping to drain large areas of inland ice. The most dynamic of these is the tidewater Jakobshavn Isbrae which is the fastest moving glacier in the world. The glacier has doubled its velocity (up to 40m/day) and has undergone a 4km terminus recession (Dietrich et al. 2005; Mayer and Herzfeld 2005). Glaciological work shows that downwasting and acceleration of the glacier are linked and self-sustaining and is caused by positive feedback mechanisms triggering rapid ice discharge, deep crevassing and basal uncoupling to produce high calving fluxes (see Zwally et al. 2002).

Recent research (Chen et al. 2006) between April 2002 and November 2005 shows a five times increase in melting from southeast Greenland over the last two years of the study compared to the first 18 months with estimated total ice melting rate over Greenland as -239 ± 23 cubic kilometers per year. This increase in melting is likely to add over 0.5mm per year of global sea level rise and may affect North Atlantic oceanic currents and climates. One of the authors, (B.D.Tapley) said "This melting process may be approaching a point where it won't be centuries before Greenland's ice melts, but a much shorter time-frame," but also noted that it is not possible to suggest a timetable for this.

Likely future evolution?

In the last few years there have been a number of attempts to assess the future stability of the ice sheets (eg Gregory et al. 2004; Hansen 2005; Oppenheimer and Alley 2005; Dowdeswell 2005; Overpeck et al. 2006). All suggest that melting will occur with a rise in global mean surface temperatures of between 1ºC and 3ºC or polar warming of less than 5ºC. The IPCC AR4 (2007) indicate that such temperatures may be reached during this century.

Policy implications

With atmospheric levels of C02 being higher than for at least 650 ka BP (Siegenthaler et al. 2005) the human influence on the global climate is now profound. The perturbation we are imposing on the global climate system is significantly larger than any plausible natural variability. Although the Holocene, up to now, has been a period of relative climatic stability, there has been no change in the forcings comparable to the likely doubling of atmospheric C02 concentrations which are likely to have been achieved by the middle of this century. Predicting the likely response of this, beyond saying that climate sensitivity to doubling C02 concentrations is likely to be around 3° C (IPCC AR4) is difficult. As Rial et al 2004 caution: "since the climate system is complex, occasionally chaotic, dominated by abrupt changes and driven by competing feedbacks with largely unknown thresholds, climate prediction is difficult, if not impracticable".

It seems likely that the models will underestimate rather than overestimate the climate sensitivity over the long run, because they omit relevant variables. It is also likely that the regional response to GHG forcing will be beyond the current capability of GCMs and Regional Climate Models (RCMs) to resolve.

Broecker writing in 1999 bemoaned the limitations of the climate models to recreate some aspects of past climates and wrote: "No one understands what is required to cool Greenland by 16 °C and the tropics by 4 ± 1 °C, to lower mountain snowlines by 900 m, to create an ice sheet covering much of North America, to reduce the atmosphere's CO2 content by 30%, or to raise the dust rain in many parts of Earth by an order of magnitude. If these changes were not documented in the climate record, they would never enter the minds of the climate dynamics community".

Conclusions

The inference from the analysis presented in this paper is that policy-makers have tended to ignore, or not be aware of, the limitations and uncertainties associated with climate predictions based upon climate and ice-sheet models (see Oppenheimer and Alley 2004). It would appear that end-users of the models have largely failed to recognize the lack of predictive skill at the regional level and the inability of models to produce rapid climate change and sea level rise in their predictions. While much progress has been made in refining the models against the palaeoclimate record, significant model uncertainties and limitations remain, and understanding these must be an important goal for policy analysts.

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Biography Dr Harrison is a Quaternary scientist with over 20 years experience in palaeoenvironmental reconstruction and palaeoclimatology. He has worked for 10 field seasons in the mountain of Patagonia on climate change research; 4 field seasons in Kazakhstan; 2 field seasons in Iceland and also in the Arctic and Himalayas. Over the last 5 years he has developed research interests in the response of business to climate change and provided much of the scientific input to the Lloyd's 360 project which looked at the commercial risks from climate change. He is Associate Professor in Quaternary Science at Exeter University and a Senior Research Associate at Oxford University.

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