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Palaeoclimatic insights into future climate challenges

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Palaeoclimatic data document a sensitive climate system subject to large and perhaps difficult-to-predict abrupt changes. These data suggest that neither the sensitivity nor the variability of the climate are fully captured in some climate-change projections, such as the Intergovernmental Panel on Climate Change (IPCC) Summary for Policymakers. Because larger, faster and less-expected climate changes can cause more problems for economies and ecosystems, the palaeoclimatic data suggest the hypothesis that the future may be more challenging than anticipated in ongoing policy making. Large changes have occurred repeatedly with little net forcing. Increasing carbon dioxide concentration appears to have globalized deglacial warming, with climate sensitivity near the upper end of values from general circulation models (GCMs) used to project human-enhanced greenhouse warming; data from the warm Cretaceous period suggest a similarly high climate sensitivity to CO_2. Abrupt climate changes of the most recent glacial–interglacial cycle occurred during warm as well as cold times, linked especially to changing North Atlantic freshwater fluxes. GCMs typically project greenhouse-gas-induced North Atlantic freshening and circulation changes with notable but not extreme consequences; however, such models often underestimate the magnitude, speed or extent of past changes. Targeted research to assess model uncertainties would help to test these hypotheses.

Keywords: abrupt climate change; global warming; climate sensitivity; palaeoclimatology; Younger Dryas; Last Glacial Maximum

1. Introduction

Humanity is likely to make decisions about fossil-fuel use and other issues related to climate based in part on scientific efforts such as the Intergovernmental Panel on Climate Change (IPCC 2001). The IPCC in turn relies heavily on model projections of future climate changes, and of likely impacts of those climate changes (see also Nordhaus & Boyer 2000; NRC 2002).

Palaeoclimatic research influences climate studies in many ways. Proxy data on past climates are used to demonstrate possibilities (something that happened must be possible), to illuminate processes, to initialize models and to test models.

One contribution of 14 to a Discussion Meeting 'Abrupt climate change: evidence, mechanisms and implications'.

Coordinated studies such as the Palaeoclimate Modelling Intercomparison Project (PMIP), a leading effort endorsed by the World Climate Research Programme and the International Geosphere–Biosphere Programme (Harrison *et al*. 2002) to assess the performance of world-class models against palaeoclimatic data, provide important insights on model success. Here, I review a range of palaeoclimatic studies, including some PMIP papers, indicating that the general circulation models (GCMs) used to project future climate changes are reasonably accurate in reconstructing past climate changes or are undersensitive to changing CO_2 levels or other forcing, but generally are not overly sensitive. Simple analogy in turn suggests that projections of future climate changes are unlikely to have overestimated changes, and may have underestimated those changes. Palaeoclimatic data almost always show much variability, but climate-change projections prepared for policy makers are often presented with smoothed curves to avoid the appearance of undue confidence in any single projection. Together, these suggest the hypothesis that the future may be more challenging for economies and ecosystems than is indicated by materials on which policy makers are relying.

I first consider palaeoclimatic versus modelled sensitivity to CO_2 changes, including the relatively well-characterized Last Glacial Maximum and the less certain conditions of the mid-Cretaceous warm climate. I then review the evidence for larger climate changes associated with millennial oscillations than have been modelled in response to likely forcing. It is important to remember that there are no conclusive results yet, and alternative interpretations certainly exist for the topics presented here. But, based on my reading of the literature, the most likely interpretation in each case considered is of a more sensitive climate system than simulated by the average GCM used in the IPCC, and a climate system with quite large variability.

2. Modelled and reconstructed sensitivity to CO_2 changes

Ice-age cycles, including the most recent global glaciation, which peaked *ca.* 20 000–25 000 years ago (see Clark & Mix 2002), had large climatic effects despite minimal global forcing. Global synchronization appears to have been achieved primarily by CO_2 changes, and the implied climate sensitivity to CO_2 falls towards the upper end of values reported from GCMs used to project future climate changes in response to human-produced CO_2.

During the most recent few million years, and especially the most recent 0.75 million years, the Earth has experienced repeated global coolings and warmings associated with the growth and shrinkage of land ice (e.g. Imbrie *et al*. 1993; Zachos *et al*. 2001). The observed frequencies of variation clearly tie these ice-age cycles to features of the Earth's orbit (Berger 1988; Imbrie *et al*. 1993). These orbital variations had little effect on globally and annually averaged incoming solar radiation (insolation: 0.5 W m^{-2} or 0.2% variations over the 100 000 year eccentricity cycle (Crowley & North 1991)). Instead, orbital variations served primarily to redistribute insolation in time and space (variations at a site and season of typically 10% to more than 20%; Berger 1988). Because some of the forcing shows a north–south hemispheric asymmetry, one early hypothesis by Croll (see review in Berger 1988) suggested that ice ages also should have alternated between the north and south.

To the contrary, ice ages have been found to be globally nearly synchronous (Imbrie *et al*. 1993) (figure 1). The extremum of the most recent global ice age was closely

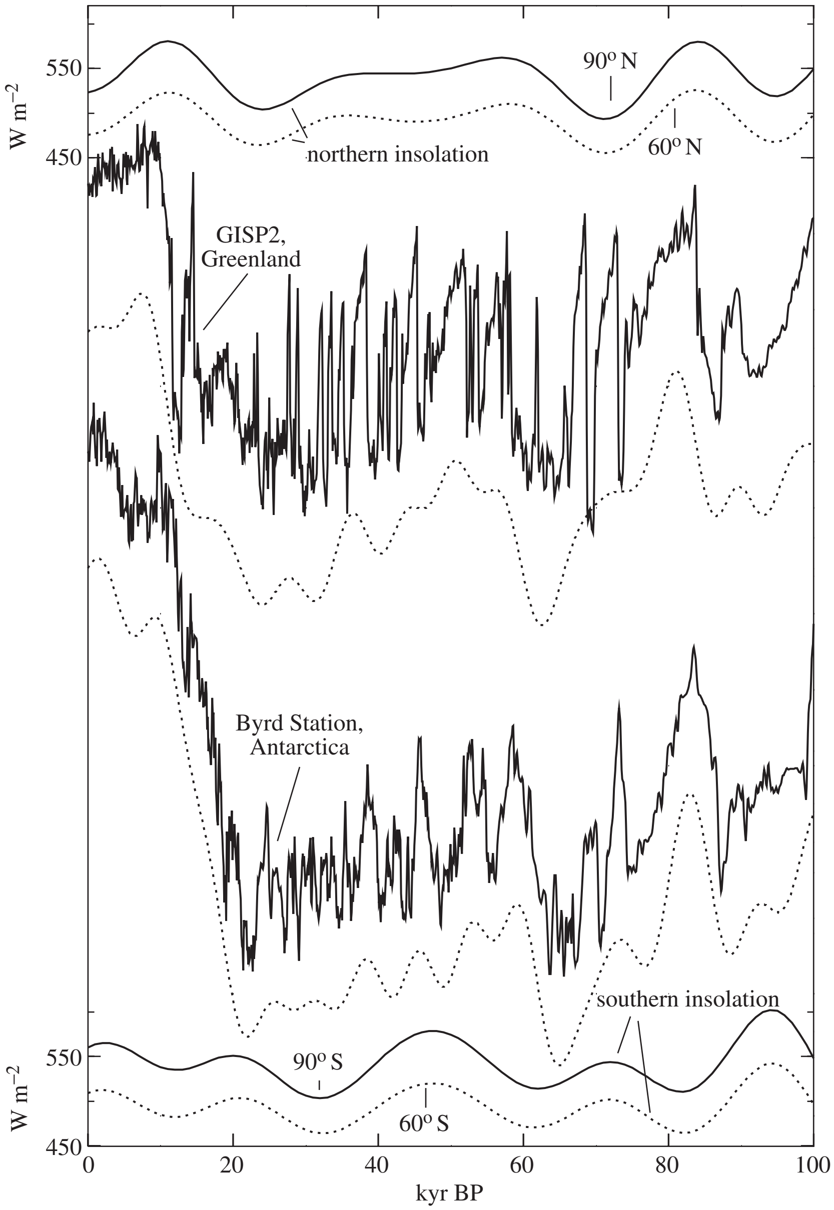

Figure 1. History of the isotopic ratio of ice (a proxy for temperature) from northern (GISP2; central Greenland) and southern (Byrd, West Antarctica) sites, synchronized by Blunier & Brook (2001), both in raw data and smoothed using a lowpass filter, compared with northern (top) and southern (bottom) midsummer insolation, following Alley *et al*. (2002). The coldest time of the most recent ice age in central Greenland was *ca.* 24 kyr BP and corresponded closely to the minimum in local midsummer insolation; the coldest time in West Antarctica was only slightly younger, and corresponded approximately to the maximum in local midsummer insolation. The rescaling used in the figure obscures the much larger temperature changes in the northern record than in the southern record.

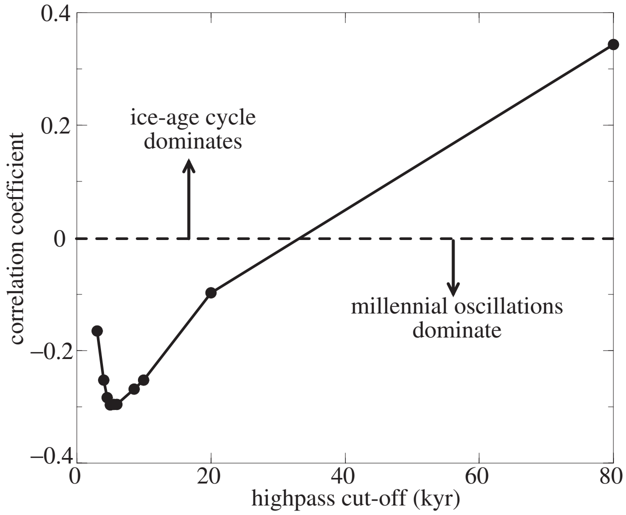

Figure 2. Correlation coefficient of the ice-core data shown in figure 1, for the complete dataset (indicated as highpass cut-off of 80 000 years), and the dataset after lower-frequency components were removed, following Alley *et al*. (2002). The positive correlation coefficient from the raw data results from the global Ice Age; the negative correlation coefficients for higher-frequency data reflect the broad antiphasing between millennial-scale events in north and south. The trend towards zero correlation coefficients at the highest frequencies considered may reflect small dating errors or 'noise' in the climate system.

linked in time to the minimum in midsummer insolation at high northern latitudes and to the maximum in midsummer insolation at high southern latitudes, indicating a clear northern control on the ice-age cycle (Imbrie *et al*. 1993; Alley *et al*. 2002).

Abundant evidence shows that the ice-age changes in insolation and ice coverage are in themselves insufficient to explain the large global temperature changes synchronous with northern changes (e.g. Suarez & Held 1979; Weaver *et al*. 1998). Leading hypotheses have been that oceanic circulation or greenhouse-gas concentrations were responsible for amplification and global synchronization of the effects of the orbital insolation changes.

Of these, it is increasingly evident that greenhouse-gas concentrations provide the answer (e.g. Genthon *et al*. 1987; Lorius *et al*. 1990). As discussed below (see Broecker 1998), oceanic circulation changes were probably primarily responsible for the large and hemispherically asymmetric signature of abrupt climate changes (Broecker 1998).

The correlation coefficient between important northern (Greenland Ice Sheet Project II (GISP2), central Greenland) and southern (Byrd Station, West Antarctica) temperature records between 90 000 and 10 000 years ago (as shown in figure 1) is plotted in figure 2 for the full dataset and with the low-frequency components removed. These are reliable, high-resolution temperature records and probably have the best relative dating of any bipolar records available (Blunier & Brook 2001). For the full datasets, the significantly positive correlation coefficient shows the bipolar synchrony of the ice-age cycle. However, using a highpass filter to remove the orbital signals from the records causes a switch to a negative correlation coefficient.

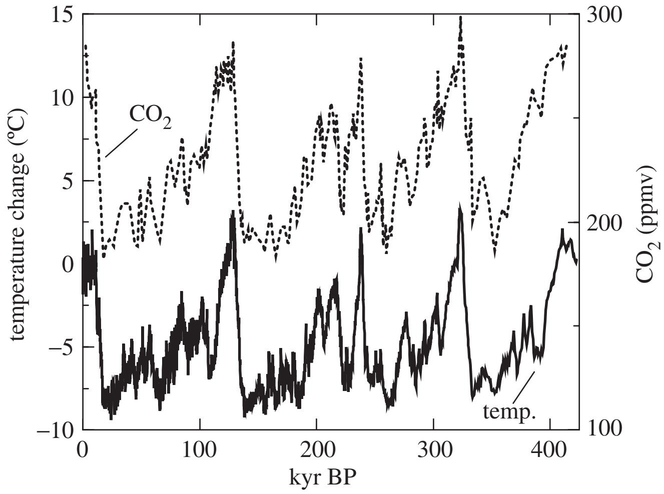

Figure 3. History of temperature in central East Antarctica at Vostok Station (from the isotopic ratios of the ice, calibrated following Salamatin *et al*. (1998)), and CO_2 concentration in the atmosphere, from Petit *et al*. (1999). The small differences in timing of events evident by inspection largely disappear when the temperatures are adjusted for changing source-water temperatures following Cuffey & Vimeux (2001). As shown by Genthon *et al*. (1987) and Lorius *et al*. (1990), among others, the temperature record is closely related to the CO_2 record, and to records with northern influence, including ice-volume history and northern insolation, but shows relatively weak dependence on southern insolation.

Inspection of the climate records before and after filtering shows that the high-frequency, anticorrelated records are associated with the Dansgaard–Oeschger and Heinrich–Bond cycles (see, for example, Broecker 1998; Alley & Clark 1999). Several of the cold intervals of this oscillation immediately followed freshening of the North Atlantic (e.g. Broecker *et al*. 1989; Barber *et al*. 1999; Broecker & Hemming 2001; Clark *et al*. 2001; Teller *et al*. 2002; Fisher *et al*. 2002) and, as discussed below, the climate anomalies associated with these coolings are consistent with freshening of the North Atlantic. Importantly, in many models (e.g. Crowley 1992; Stocker *et al*. 1992; Broecker 1998), freshening of the North Atlantic produces a northern cooling and southern warming in opposition, with the connection made through the effects of North Atlantic deep water as an important link in the global oceanic thermohaline circulation.

It thus appears highly likely that changes in the oceanic circulation were responsible for causing the millennial see-saw behaviour between the high northern and southern latitudes. In turn, this indicates that the oceanic circulation is unlikely to be responsible for synchronizing the polar regions over ice-age cycles.

Atmospheric CO_2 concentrations (as well as methane and nitrous oxide concentrations (Raynaud *et al*. 1992)) have varied in concert with the ice-age cycle and global temperatures (figure 3; data from Petit *et al*. (1999). The Vostok, East Antarctica temperature record, especially when corrected for the effects on ice-isotopic ratios of changing source-water temperatures (Cuffey & Vimeux 2001), exhibits a very tight link between CO_2 concentration and southern temperature (cf. Shackleton 2000).

A possible slight lag of CO_2 concentration change behind temperature change (Fischer *et al*. 1999) has received much attention (there were 31 citations listed in the ISI Science Citation Index database in April 2003) (see also Caillon *et al*. 2003). Certainly, temperature is affected by more than just CO_2 concentration, so small decoupled changes in temperature or CO_2 concentration are expected. It remains, as shown by Cuffey & Vimeux (2001), that all large changes in CO_2 concentration and temperature in East Antarctica have been correlated.

The data thus indicate that the globalization of the orbital ice-age signal was achieved by CO_2. The climatic sensitivity to this forcing can then provide a test of the sensitivity of climate models.

A simple way to do this was presented by Hoffert & Covey (1992), who compared the global-mean temperature change and the change in forcing from greenhouse gases, albedo, etc., to obtain a climate sensitivity to CO_2 for both glacial-maximum conditions and conditions from the warm mid-Cretaceous of *ca.* 100 million years ago. Hoffert & Covey's (1992) sensitivities to CO_2 doubling were $2.0 \pm 0.5\,^{\circ}C$ for the comparison with ice-age conditions, and $2.5 \pm 1.2\,^{\circ}C$ for the comparison with mid-Cretaceous conditions, or an average sensitivity of *ca.* 2.3 °C. Models used in the IPCC (2001) produce estimates that range from 1.5 to 4.5 °C for the warming from a doubling of CO_2 from recent values, so the initially reassuring result from Hoffert & Covey (1992) was that palaeoclimatic sensitivities were slightly less than the midpoint of those models.

However, it now appears that higher sensitivities are indicated by data from both warm and cold climates. Starting with the cold climates, the Hoffert & Covey (1992) ice-age estimate was based on early palaeoclimatic data that rather clearly underestimated changes (specifically, the CLIMAP sea-surface temperatures (SSTs)). As updated by Cuffey & Brook (2000), improved data yield a sensitivity of *ca.* 3.9 °C warming for a CO_2 doubling, which falls towards the upper end of the IPCC sensitivities.

Recent PMIP papers provide spatial context to this globally lumped sensitivity indicated by Cuffey & Brook (2000) based on Hoffert & Covey (1992). Pinot *et al*. (1999) compared results of PMIP simulations for the Last Glacial Maximum with a new, improved dataset for tropical land-surface temperatures (Farrera *et al*. 1999). Results of nine models were matched against data in distinct regions, using simulations with specified (CLIMAP) SSTs and with SSTs calculated within the models.

Most simply, 'the LGM simulations with prescribed SSTs underestimate the observed temperature changes' except in one region, where agreement is obtained (Pinot *et al*. 1999, p. 857). Using computed SSTs, 'four of the nine simulations produce a cooling in good agreement with terrestrial data' (Pinot *et al*. 1999, p. 857), whereas the other five models tended to underestimate the reconstructed warming from the Last Glacial Maximum to today.

Similarly, Kageyama *et al*. (2001) compared PMIP simulations from 17 climate models (16 of which were GCMs) with new databases for Eurasian warming from the last Ice Age. Kageyama *et al*. (2001, p. 23) found

> overall the Last Glacial Maximum climate simulated by the models over western Europe is warmer, especially in winter, and wetter than the one depicted by the reconstructions. . . . The same disagreement, but of smaller amplitude, is found over Central Europe and the eastern Mediterranean Basin, while models and data are in broad agreement over western Siberia.

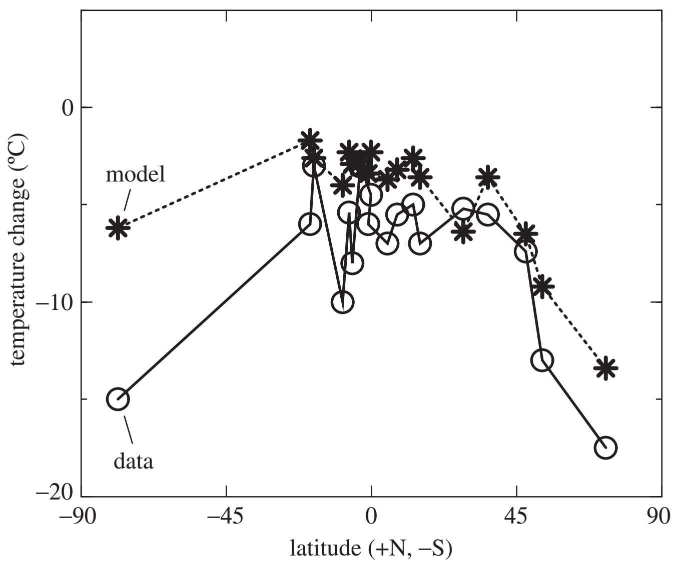

Figure 4. Comparison of reconstructed ('data') and calculated ('model') mean annual temperature difference between recent and glacial-maximum land-surface conditions, plotted against latitude for sites tabulated by Pollard & Thompson (1997). The Cuffey *et al*. (1995) temperature changes have been used for central Greenland, the Salamatin *et al*. (1998) calibration for Vostok, East Antarctica, and the mean has been adopted where Pollard & Thompson (1997) listed a range of reconstructed temperatures. Model results are from the GENESIS GCM as run using the mixed-layer ocean for the PMIP intercomparison (generally, smaller changes are modelled using specified CLIMAP SST data, clearly inconsistent with the reconstructions). The model is skilful in simulating the pattern of change with polar amplification, but appears to be less sensitive than the actual climate.

Kageyama *et al*. (2001, table 1) also summarized work of predecessors, including four additional studies reaching essentially the same result: that models were accurate or less sensitive than indicated by the data.

A more geographically extensive comparison using a single model from among the PMIP suite is given by Pollard & Thompson (1997), and summarized in figure 4. Again, the model underestimated the reconstructed changes almost everywhere, but simulated the pattern of changes rather well.

Looking across all of these PMIP intercomparisons, one finds in general that the models are skilful, matching many patterns of climate change quite well, and that there is a tendency for the models either to be accurate in simulating the magnitudes of changes or to underestimate the changes that occurred. Using a sensitivity near the high end of those exhibited by the models would match the data reasonably well (although with a few remaining problems, such as in western Europe); using a sensitivity in the middle or towards the low end of those exhibited by the models underestimates the changes indicated by the proxy records. Rarely are these changes overestimated by models.

The Cretaceous-to-modern climate sensitivity is also of interest. Certainly, because of their greater age, indicators are less clear for the climates notably warmer than recently. Problems include lack of direct air samples from ice cores to measure CO_2 concentrations, possible evolutionary change affecting biotic indicators of climate, and diagenetic effects that may have shifted original climate records. Nonetheless, it appears from available data that the climate sensitivity to CO_2 indicated by changes from mid-Cretaceous warmth to today is higher than estimated by Hoffert & Covey (1992), and toward the high end of IPCC model estimates.

The ice ages of the last million years culminated a long but variable cooling trend over *ca.* 100 million years from mid-Cretaceous warmth (Zachos *et al.* 2001). Many processes, including changes in continental positions and oceanic heat transport, may have contributed to the cooling. However, model results (see, for example, Barron *et al.* 1995; DeConto & Pollard 2003) indicate that long-term reduction in the CO_2 concentration of the atmosphere played a central role in the cooling trend.

As noted above, Hoffert & Covey (1992) found a climate sensitivity to CO_2 of 2.5 ± 1.2 °C in comparing Cretaceous with modern climate, in the low-to-middle range of the IPCC sensitivities. Hoffert & Covey (1992) used a mid-range estimate for Cretaceous CO_2 concentration of four times recent values, and climate data from Barron & Washington (1985), which in turn are based on Barron (1983). Subsequently, using a new suite of model experiments to assess more accurately the effects of continental positions and oceanic heat transport, Barron *et al.* (1995) found that the most likely result is a somewhat higher sensitivity of 2.5–4.0 °C for the warming from doubled CO_2, in the middle-to-upper range of the IPCC sensitivities.

The Barron (1983) dataset for mid-Cretaceous climate is based on many sources, including the oxygen-isotopic composition of carbonate shells of planktonic foraminifera. Barron (1983) noted the possibility of diagenesis having affected these ratios, with the likelihood that the estimated temperatures would have been lowered diagenetically. Recent data from Pearson *et al.* (2001), Wilson & Norris (2001) and Wilson *et al.* (2002) do indicate higher tropical SSTs during warm periods than previously reported, although based on only local reconstructions.

Pearson *et al.* (2001) analysed selected samples from warm climates of the Late Cretaceous, Middle Eocene, and Late Eocene (but not from the mid-Cretaceous). The new data indicate warmer temperatures than from previous results for equivalent latitudes and ages, by *ca.* 15 °C.

The 15 °C increase in estimated warm-period temperatures may be too large, because the Pearson *et al.* (2001) sites were in shallow-water locations that may have been anomalously warm, and for other reasons (Zachos *et al.* 2002), but a non-zero diagenetic overprint that has acted to reduce estimates of warm-period tropical SSTs still seems likely (Pearson *et al.* 2002). In light of the importance of the issue, additional work seems warranted.

Wilson & Norris (2001) and Wilson *et al.* (2002) studied mid-Cretaceous samples from the western tropical Atlantic, concentrating on especially well-preserved, glassy foraminiferal shells from clay-rich samples. These studies found considerable variability in SSTs over time, with peak temperatures well above both modern SSTs and the values assumed for broader regions by Barron *et al.* (1995). In the western tropical Atlantic, SSTs are 27–29 °C today. Wilson & Norris (2001) found peak temperatures of $32–33 \pm 3$ °C for the mid-Cretaceous. Wilson *et al.* (2002) used a slightly different approach to estimate temperatures between 30 and 33 °C. However, they then noted that, if account were taken of the likely effect of higher Cretaceous CO_2 concentration on oceanic pH, and the resulting effect on the oxygen-isotopic composition of carbonate shells, the SST estimate would be increased to 32–36 °C.

The 6.2 °C globally averaged mid-Cretaceous warming above recent temperatures used by Barron *et al.* (1995) in obtaining a mid-to-high-range climate sensitivity included SSTs less than 30 °C, although Barron *et al.* (1995) noted the possibility of warmer temperatures to at least 32 °C. Any underestimate of SSTs used by Barron *et al.* (1995) would require either fortuitously offsetting errors elsewhere in the calculation (such as underestimation of Cretaceous CO_2 concentrations) or else a higher climate sensitivity. (And, if the pH effect proposed by Wilson *et al.* (2002) is accurate and higher CO_2 concentration would cause uncorrected isotopic ratios to give greater underestimates of SSTs, the influence of higher Cretaceous CO_2 on the calculated climate sensitivity could be partly offset by even warmer SSTs.) This in turn strengthens the evidence that palaeoclimatic sensitivity for both warm and cold times has been towards the upper end of values exhibited by GCMs used by the IPCC.

3. Sensitivity to internal forcings

Switching next to abrupt climate changes of the most recent glacial cycle, it again appears that the models are relatively skilful, but show a tendency to underestimate changes. There was little CO_2 forcing for these abrupt climate changes (Monnin *et al.* 2001), so implications of model insensitivity are less directly linked to future CO_2 impacts. And yet, a tendency for the models to exhibit smaller anomalies in response to perturbations than does the real world seems of interest in assessing response to future perturbations.

I follow a widespread, but certainly not universally accepted, procedure here. I assume that the millennial-spaced Dansgaard–Oeschger and Heinrich–Bond oscillations and their associated abrupt climate changes include northern coolings caused by freshening of the North Atlantic disrupting the North Atlantic thermohaline overturning circulation, and northern warmings related to re-establishment of that circulation.

Many events, especially the cold event *ca.* 8200 years ago, the Younger Dryas cold event centred on *ca.* 12 000 years ago, and the Preboreal Oscillation just younger than the Younger Dryas, involved cooling immediately after large freshwater inputs to the North Atlantic from outburst floods of ice-dammed lakes (see, for example, Broecker *et al.* 1989; Barber *et al.* 1999; Clark *et al.* 2001; Teller *et al.* 2002; Fisher *et al.* 2002). Also, Atlantic freshening from Heinrich-event surging of Laurentide ice earlier in the Ice Age lengthened and strengthened cold intervals that marked the minima of Bond cycles (Broecker & Hemming 2001). As discussed below, freshening of the North Atlantic surface is modelled to cause climatic anomalies that closely match those changes reconstructed from proxy data, including the see-saw behaviour with the South Atlantic.

North Atlantic freshenings and coolings could possibly have been forced independently in some other way, perhaps from the tropical Pacific Ocean (Clement *et al.* 2001) or elsewhere. However, ice-dynamical processes involved in ice-dam and ice-surge behaviour have long (centuries, millennia) response times, such as the interval required for advance and retreat to block and unblock ice-marginal river channels, or for major subglacial thermal changes associated with surging (see, for example, Alley & MacAyeal 1994). Yet, the North Atlantic freshening events occurred at the onsets of cooling steps, within one to a few centuries dating uncertainties. Together, these make highly unlikely any model of separate triggering from elsewhere of both freshening and cooling. Repeated close correspondence of freshening and cooling also argues against the hypothesis that the events are physically unrelated but fortuitously close in time. More reasonable is the inference that the freshenings did cause the coolings.

Perhaps the main argument against a North Atlantic model for these millennial changes (see, for example, Seager *et al*. 2002) is the tendency for modelled responses to be smaller in area and magnitude than reconstructed responses. This could indicate a lack of understanding of causation, but more likely indicates a lack of model sensitivity. We indeed have argued (Alley *et al*. 2001) that forcing in addition to North Atlantic freshening was active, perhaps linked to solar processes (cf. Bond *et al*. 2001), but it appears that the additional forcing was quite small (based on amplitudes of non-abrupt changes including those in the Holocene) and will not be considered further here. Thus, I interpret data-model mismatches as resulting largely from model insensitivity.

Briefly (see Broecker (1997) for additional discussion), the North Atlantic is the only part of the world's ocean in which warm waters reach high latitudes. The high-salinity waters of the North Atlantic become sufficiently dense (when cooled during winters) to sink into the deep ocean, to be replaced by additional high-salinity, warm waters, preventing sea-ice formation and associated cooling. The waters that sink flow southward in the Atlantic, into the Southern Ocean, and eventually upwell in various regions and return along the surface of the Atlantic in a great 'conveyor-belt' circulation. Sufficient freshening of the surface of the North Atlantic can allow sea-ice formation, with large climatic effects. Some models (especially simple ones; reviewed in NRC (2002)) simulate step-like resumption of the circulation, and this is believed to explain the special abruptness of warmings from millennial cool events.

Numerous models have been used to address abrupt climate changes associated with the millennial Dansgaard–Oeschger and Heinrich–Bond cycles (see review in NRC (2002)). Here, I am primarily interested in GCMs of the type given increasing prominence in the IPCC exercise, rather than the simpler models on which much understanding of processes rests.

PMIP is conducting a meltwater–pulse intercomparison project, but the results are not yet available. Thus, our ability to evaluate model performance on abrupt climate changes will soon improve. Until then, I provide an incomplete review of the vast amount of literature on this topic.

Fawcett *et al*. (1997) and Ágústsdóttir (1998) (see also Ágústsdóttir *et al*. 1999) conducted simulations with the GENESIS GCM with a mixed-layer ocean. To avoid unnatural sea-ice build-up in the North Atlantic under modern conditions, the model uses a special heat-flux convergence in the Nordic seas. This mimics the local effects of the currents that transport salt and heat to the high-latitude North Atlantic, suppressing sea-ice formation in the winter. By comparing simulations with and without this Nordic-seas correction, with boundary conditions appropriate to the times studied, the authors assessed the effects of past freshening of the North Atlantic allowing sea-ice formation.

For the Younger Dryas interval, Fawcett *et al*. (1997) found numerous points of agreement between the simulated and reconstructed climate changes, including shifts in Greenland of precipitation seasonality and storm tracks. Ágústsdóttir *et al*. (1999) additionally demonstrated changes in Ekman divergence affecting upwelling off Venezuela and elsewhere that match climate changes reconstructed from sediment records. The degree of agreement in many regions supports the inference that North Atlantic freshening and cooling produced the event (Fawcett *et al.* 1997).

However, Fawcett *et al.* (1997) also found that the modelled response to North Atlantic cooling under Younger Dryas boundary conditions underestimated the reconstructed climate changes in key ways. For example, the model underestimated reconstructed changes in temperature and snow accumulation in central Greenland, and the cooling in regions including New Zealand and the northeastern United States (Peteet *et al.* 1993; Denton & Hendy 1994; Ivy-Ochs *et al.* 1999).

Ágústsdóttir (1998) undertook a similar modelling exercise for the cold event *ca.* 8200 years ago. Eliminating the excess heat convergence in the Nordic seas under the appropriate boundary conditions for that interval had rather limited impact on the modelled climate, in strong contrast to the observations (see, for example, Alley *et al.* 1997). However, by specifying an initial heat convergence 60% larger than that used in the model for today and then eliminating that larger anomalous heat convergence, an anomaly pattern was obtained that matched proxy data rather well. Although we have presented circumstantial evidence for slightly more vigorous North Atlantic overturning circulation during the mid-Holocene than recently (Alley *et al.* 1999), Ágústsdóttir (1998) considered the perturbation used in the model to be rather large; Ágústsdóttir's (1998) results are thus suggestive of a model that is somewhat undersensitive compared with the natural climate system.

An extensive programme, intercomparing model output with climate reconstructions, focused especially on European conditions associated with abrupt climate changes but clearly looking more broadly, has been undertaken by Renssen and co-workers using the Hamburg or ECHAM model. Renssen (1997) specified ice sheets, CO_2, orbital parameters and SSTs for Younger Dryas times in the Hamburg atmospheric model (see also Isarin & Renssen 1999). Specification in the model of North Atlantic Younger Dryas SST cooling inferred from changes in foraminiferal assemblages did not create climate anomalies large enough to match the reconstructed changes over Europe or elsewhere. Use of model-derived SST fields and sea-ice cover in the North Atlantic corresponding to a complete shutdown of the thermohaline circulation (a perturbation that Renssen (1997) indicated to be larger than that supported by available datasets) created climate anomalies broadly matching those reconstructed in many parts of Europe, Greenland, Atlantic Canada and the region around the North Pacific, but underestimated changes elsewhere. This mirrored the earlier results of Rind *et al.* (1986) in a similar experiment with an earlier-generation model.

Experiments by Renssen *et al.* (2001) (see also Renssen & Isarin 2001*b*) using the ECHAM4/T42 version of the Hamburg model, and the boundary conditions that in Renssen (1997) were indicated to be more extreme than likely, found that nesting a mesoscale model in the global model improved data–model agreement in Europe for the Younger Dryas anomalies, but that wintertime temperatures simulated for western Europe still exceeded reconstructions by 5–25 °C; the authors suggested that specification of even more extensive sea-ice cover might improve data-model agreement.

A full review of the literature on effects of North Atlantic freshening with boundary conditions corresponding to various times in the past (and the present and future) is beyond the scope of this report, especially as the experiments conducted to date are not fully comparable in boundary conditions or in datasets used for comparisons.

My understanding of Manabe & Stouffer (1997, 1999), Schiller *et al.* (1997), Rutter *et al.* (2000), Renssen & Isarin (2001*a*), Ganopolski & Rahmstorf (2001), Rind *et al.* (2001) and Vellinga & Wood (2002), among other papers, is that

(i) models forced by North Atlantic cooling or freshening actually show considerable skill in simulating many aspects of the coolings associated with the abrupt climate changes of the past, including strong cooling around the North Atlantic and downwind especially in the wintertime, southward shifts in storm tracks to a more zonal configuration forced by expanded sea-ice coverage in the North Atlantic and see-saw behaviour with southern warming; however,

(ii) there is a tendency for the models to simulate smaller-amplitude, reduced-area, reduced-duration, or reduced-rate changes compared with those observed, although the strongest forcings of the models can produce model anomalies that sometimes match or even exceed observed anomalies in some regions.

4. Discussion

A literature review such as this highlights at once the amazing richness of the available results, the strong indications from these results about behaviour in the climate system, and the large gaps in understanding of models and data. The exercise is appropriate for hypothesis generation, but cannot yet support strong conclusions.

In assessing climate changes from the Last Glacial Maximum to today, for example, significant uncertainties remain in reconstructing temperatures in key areas, including oceanic regions. One must be mindful of the differing time resolutions of records, and thus the possibility that the 'Last Glacial Maximum' or 'the coldest time of the last glaciation' averages over a longer interval and so gives a higher temperature in some records than in others, and that the reconstructed cooling would thus underestimate the actual cooling. This is balanced against the clearly time-transgressive nature of the coldest time for at least some records (see, for example, Alley *et al.* 2002), such that choosing the coldest time in many different records could serve to overestimate the total cooling achieved at any one time. Although some palaeothermometers (e.g. borehole thermometry in ice sheets (Cuffey *et al.* 1995) and noble-gas palaeothermometry of groundwaters (Stute *et al.* 1995)) record little more than temperature, others may be sensitive to additional factors and may rely on a uniformitarian hypothesis that is in some ways untestable (Alley 2001).

For the mid-Cretaceous, additional uncertainties arise in estimating climate, and in learning the CO_2 concentration associated with the climate. For abrupt climate changes, no really secure global-anomaly maps exist, although great progress is being made especially for the Younger Dryas (see, for example, Renssen 1997; Peteet 2000; Rutter *et al.* 2000).

An additional complicating factor arises from uncertainty about causes. Although orbital control on insolation with CO_2 globalization is widely accepted as the explanation for the ice-age cycles, it is not universally accepted, and hypotheses for additional or alternative forcings appear frequently (see, for example, Sharma 2002). More uncertainties exist in deeper time. For abrupt climate changes, I believe that freshwater forcing of North Atlantic thermohaline-circulation change is the leading hypothesis. Simulation of response to this forcing does produce anomaly maps with great similarity to many aspects of the reconstructed changes, but competing ideas exist. Clearly, if the actual forcing is not used in simulations, mismatch between reconstructed and modelled changes will not provide quantitatively accurate information on model sensitivity.

And yet, the available data and models represent very good science. Extensive, careful efforts have been devoted to dating, determining transfer functions between proxies and climate, and developing climate histories for diverse places across the globe. Physical insight, a hierarchy of models, and model–data intercomparisons have been used to develop strong hypotheses for the climate changes of the past. Errors are real, alternate interpretations are possible, but there is no good basis for dismissing the most likely results because of the associated uncertainties. And, if one accepts at face value our leading understanding of the causes of the climate changes, of the magnitudes of the changes, and of the modelled response to such changes, a picture emerges of skilful models that are a little undersensitive.

The undersensitivity of models is clearest for the warming from the Ice Age to today. Strong evidence indicates globalization of climate anomalies by CO_2 changes. The changes in CO_2 are well established (Petit *et al*. 1999), as are the changes in temperature. Well-organized, rigorous intercomparisons have been run under PMIP, as summarized above. The general-circulation climate models involved in such studies exhibit skill but tend to underestimate changes.

A re-examination of changes to warm climates with higher CO_2 is needed in light of new data suggesting that tropical SSTs were higher during warm climates than previously estimated (Poulsen *et al*. 1999; Wilson & Norris 2001; Norris *et al*. 2002; Wilson *et al*. 2002; Pearson *et al*. 2001; cf. Zachos *et al*. (2002) and Pearson *et al*. (2002)). The best previous results (Barron *et al*. 1995) placed the actual climate sensitivity to CO_2 near or above the midpoint of sensitivities of models used by the IPCC, and very preliminary results suggest that the actual sensitivity to CO_2 may be yet higher.

In looking at the freshening of the North Atlantic, the modelled response to known forcings is sufficiently similar to observed anomaly patterns to support the contention that North Atlantic changes caused abrupt climate changes. However, the models again seem to underestimate the changes, suggesting a lack of model sensitivity to this forcing as well as to CO_2. (Also, Harrison *et al*. (2002) noted a similar slight model undersensitivity in some regions to the orbital changes that caused mid-Holocene warmth.)

One additional point, which may be the most important one, should be made here. In the Summary for Policymakers and the technical summary of the IPCC Working Group I Third Assessment Report (IPCC 2001), projections of future changes in climate and sea level are given as smooth curves or bands. The reason for doing so is clear: no particular projection with more realistic variability would have a high probability of being correct, so one should not mislead policy makers with a spuriously precise projection. Nonetheless, variability is inevitable, abrupt climate shifts affecting large regions are possible (NRC 2002), and faster and less-expected changes are harder to handle. The full IPCC document includes extensive consideration of abrupt climate changes and variability, but the enduring images given to policy makers represent a world that is less variable and hence more favourable to humans and ecosystems than is likely to occur.

5. Summary

Decisions about human response to climate change will be based on projected costs and benefits, broadly defined, with much of the information being presented to populations and policy makers through the IPCC (2001) and similar efforts. I have played only a very small role in the IPCC effort, but have stated in print that 'I am favorably impressed with the IPCC process and product' (Alley 2000, p. 220). Nonetheless, because of the great importance of the IPCC, careful assessment of the results, and consideration of whether they could be improved, is vital.

Based on a palaeoclimatic perspective, the future could prove more challenging climatically than indicated by IPCC (2001), and especially by the Summary for Policymakers. An assessment of the effect of CO_2, particularly during deglacial warming but also in looking at the warm climate of the mid-Cretaceous, indicates that the average behaviour of the models underpinning the IPCC somewhat underestimates climate sensitivity, although the more sensitive of the models are rather accurate. Similarly, models often exhibit some skill in simulating abrupt climate changes of the past, but with a tendency to underestimate the size, extent, or rate of the changes.

From these results, we should seriously consider the hypothesis that future changes will on average be larger than indicated by the mean of the major climate-model projections, and may involve abrupt climate changes not projected accurately by the models (Alley *et al*. 2003). Additionally, the IPCC (2001) Working Group I Summary for Policymakers and Technical Summary show future projections as smooth bands or curves (again, for good reasons), whereas more challenging variability, perhaps with abrupt climate shifts, is to be expected. Taken together, these results may recommend additional urgency in the study of climate change.

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References

Ágústsdóttir, A. M. 1998 Abrupt climate changes and the effects of North Atlantic deepwater formation: results from the GENESIS global climate model and comparison with data from the Younger Dryas event and the event at 8200 years BP and the present. PhD thesis, The Pennsylvania State University, PA, USA.

Ágústsdóttir, A. M., Alley, R. B., Pollard, D. & Peterson, W. 1999 Ekman transport and upwelling from wind stress from GENESIS climate model experiments with variable North Atlantic heat convergence. *Geophys. Res. Lett.* **26**, 1333–1336.

Alley, R. B. 2000 *The two-mile time machine: ice cores, abrupt climate change, and our future.* Princeton University Press.

Alley, R. B. 2001 The key to the past? *Nature* **409**, 289.

Alley, R. B. & Clark, P. U. 1999 The deglaciation of the Northern Hemisphere: a global perspective. *A. Rev. Earth Planet. Sci.* **27**, 149–182.

Alley, R. B. & MacAyeal, D. R. 1994 Ice-rafted debris associated with binge/purge oscillations of the Laurentide ice sheet. *Paleoceanography* **9**, 503–511.

Alley, R. B., Mayewski, P. A., Sowers, T., Stuiver, M., Taylor, K. C. & Clark, P. U. 1997 Holocene climatic instability: a prominent, widespread event 8200 years ago. *Geology* **25**, 483–486.

Alley, R. B., Ágústsdóttir, A. M. & Fawcett, P. J. 1999 Ice-core evidence of Late-Holocene reduction in North Atlantic Ocean heat transport. In *Mechanisms of global climate change at millennial time scales* (ed. P. U. Clark, R. S. Webb & L. D. Keigwin). Geophysical Monograph vol. 112, pp. 301–312. Washington, DC: American Geophysical Union.

Alley, R. B., Anandakrishnan, S. & Jung, P. 2001 Stochastic resonance in the North Atlantic. *Paleoceanography* **16**, 190–198.

Alley, R. B., Brook, E. J. & Anandakrishnan, S. 2002 A northern lead in the orbital band: north–south phasing of ice-age events. *Quat. Sci. Rev.* **21**, 431–441.

Alley, R. B. (and 10 others) 2003 Abrupt climate change. *Science* **299**, 2005–2010.

Barber, D. C. (and 10 others) 1999 Forcing of the cold event of 8200 years ago by catastrophic drainage of Laurentide lakes. *Nature* **400**, 344–348.

Barron, E. J. 1983 A warm, equable Cretaceous: the nature of the problem. *Earth Sci. Rev.* **19**, 305–338.

Barron, E. J. & Washington, W. M. 1985 Warm Cretaceous climates: high atmospheric CO_2 a plausible mechanism. In *The carbon cycle and atmospheric* CO_2*: natural variations archean to present* (ed. E. T. Sundquist & W. S. Broecker), pp. 546–553. Washington, DC: American Geophysical Union.

Barron, E. J., Fawcett, P. J., Peterson, W. H., Pollard, D. & Thompson, S. L. 1995 A 'simulation' of mid-Cretaceous climate. *Paleoceanography* **10**, 953–962.

Berger, A. 1988 Milankovitch theory and climate. *Rev. Geophys.* **26**, 624–657.

Blunier, T. & Brook, E. J. 2001 Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* **291**, 109–112.

Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M. N., Showers, W., Hoffmann, S., Lotti-Bond, R., Hajdas, I. & Bonani, G. 2001 Persistent solar influence on North Atlantic climate during the Holocene. *Science* **294**, 2130–2136.

Broecker, W. S. 1997 Thermohaline circulation, the Achilles heel of our climate system: will man-made CO_2 upset the current balance? *Science* **278**, 1582–1588.

Broecker, W. S. 1998 Paleocean circulation during the last deglaciation: a bipolar seesaw? *Paleoceanography* **13**, 119–121.

Broecker, W. S. & Hemming, S. 2001 Climate swings come into focus. *Science* **294**, 2308–2309.

Broecker, W. S., Kennett, J. P., Flower, B. P., Teller, J. T., Trumbore, S., Bonani, G. & Wolfli, W. 1989 Routing of meltwater from the Laurentide ice-sheet during the Younger Dryas cold episode. *Nature* **341**, 318–321.

Caillon, N., Severinghaus, J. P., Jouzel, J., Barnola, J. M., Kang, J. C. & Lipenkov, V. Y. 2003 Timing of atmospheric CO_2 and Antarctic temperature changes across termination. III. *Science* **299**, 1728–1731.

Clark, P. U. & Mix, A. C. 2002 Ice sheets and sea level of the Last Glacial Maximum. *Quat. Sci. Rev.* **21**, 1–7.

Clark, P. U., Marshall, S. J., Clarke, G. K. C., Hostetler, S. W., Licciardi, J. M. & Teller, J. T. 2001 Freshwater forcing of abrupt climate change during the last glaciation. *Science* **293**, 283–287.

Clement, A. C., Cane, M. A. & Seager, R. 2001 An orbitally driven tropical source for abrupt climate change. *J. Clim.* **14**, 2369–2375.

Crowley, T. J. 1992 North Atlantic deep water cools the Southern Hemisphere. *Paleoceanography* **7**, 489–497.

Crowley, T. J. & North, G. R. 1991 *Paleoclimatology*. Oxford University Press.

Cuffey, K. M. & Brook, E. J. 2000 Ice sheets and the ice-core record of climate change. In *Earth system science: from biogeochemical cycles to global change* (ed. M. C. Jacobson, R. J. Charlson, H. Rodhe & G. H. Orians), pp. 459–497. Academic.

Cuffey, K. M. & Vimeux, F. 2001 Covariation of carbon dioxide and temperature from the Vostok ice core after deuterium-excess correction. *Nature* **412**, 523–527.

Cuffey, K. M., Clow, G. D., Alley, R. B., Stuiver, M., Waddington, E. D. & Saltus, R. W. 1995 Large Arctic temperature change at the glacial–Holocene transition. *Science* **270**, 455–458.

DeConto, R. M. & Pollard, D. 2003 Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. *Nature* **421**, 245–249.

Denton, G. H. & Hendy, C. H. 1994 Younger Dryas age advance of Franz-Josef glacier in the Southern Alps of New Zealand. *Science* **264**, 1434–1437.

Farrera, I. (and 18 others) 1999 Tropical climates at the Last Glacial Maximum: a new synthesis of terrestrial palaeoclimate data. I. Vegetation, lake-levels and geochemistry. *Climate Dynam.* **15**, 823–856.

Fawcett, P. J., Ágústsdóttir, A. M., Alley, R. B. & Shuman, C. A. 1997 The Younger Dryas termination and North Atlantic deepwater formation: insights from climate model simulations and Greenland ice core data. *Paleoceanography* **12**, 23–38.

Fischer, H., Wahlen, M., Smith, J., Mastroianni, D. & Deck, B. 1999 Ice core records of atmospheric CO_2 around the last three glacial terminations. *Science* **283**, 1712–1714.

Fisher, T. G., Smith, D. G. & Andrews, J. T. 2002 Preboreal oscillation caused by a glacial Lake Agassiz flood. *Quat. Sci. Rev.* **21**, 873–878.

Ganopolski, A. & Rahmstorf, S. 2001 Rapid changes of glacial climate simulated in a coupled climate model. *Nature* **409**, 153–158.

Genthon, C., Barnola, J. M., Raynaud, D., Lorius, C., Jouzel, J., Barkov, N. I., Korotkevich, Y. S. & Kotlyakov, V. M. 1987 Vostok ice core—climatic response to CO_2 and orbital forcing changes over the last climatic cycle. *Nature* **329**, 414–418.

Hoffert, M. I. & Covey, C. 1992 Deriving global climate sensitivity from paleoclimate reconstructions. *Nature* **360**, 573–576.

Imbrie, J. (and 18 others) 1993 On the structure and origin of major glaciation cycles. 2. The 100 000-year cycle. *Paleoceanography* **8**, 699–735.

IPCC 2001 *Climate change 2001: the scientific basis. Contribution of Working Group I to the Third Assessment Report of the International Panel on Climate Change.* (ed. J. T. Houghton, Y. Ding, D. J. Griggs, M. Noguer, P. J. van der Linden, X. Dai, K. Maskell & C. A. Johnson). (Available at http://www.ipcc.ch.) Cambridge University Press.

Isarin, R. F. B. & Renssen, H. 1999 Reconstructing and modelling late Weichselian climates: the Younger Dryas in Europe as a case study. *Earth Sci. Rev.* **48**, 1–38.

Ivy-Ochs, S., Schluchter, C., Kubik, P. W. & Denton, G. H. 1999 Moraine exposure dates imply synchronous Younger Dryas glacier advances in the European Alps and in the Southern Alps of New Zealand. *Geografiska Annaler* A **81**, 313–323.

Kageyama, M., Peyron, O., Pinot, S., Tarasov, P., Guiot, J., Joussaume, S. & Ramstein, G. 2001 The Last Glacial Maximum climate over Europe and western Siberia: a PMIP comparison between models and data. *Climate Dynam.* **17**, 23–43.

Lorius, C., Jouzel, J., Raynaud, D., Hansen, J. & Le Treut, H. 1990 The ice-core record: climate sensitivity and future greenhouse warming. *Nature* **347**, 139–145.

Manabe, S. & Stouffer, R. J. 1997 Coupled ocean–atmosphere model response to freshwater input: comparison with Younger Dryas event. *Paleoceanography* **12**, 321–336.

Manabe, S. & Stouffer, R. J. 1999 Are two modes of thermohaline circulation stable? *Tellus* A **51**, 400–411.

Monnin, E., Indermuhle, A., Dallenbach, A., Fluckiger, J., Stauffer, B., Stocker, T. F., Raynaud, D. & Barnola, J. M. 2001 Atmospheric CO_2 concentrations over the last glacial termination. *Science* **291**, 112–114.

Nordhaus, W. D. & Boyer, J. 2000 *Warming the world: economic modelling of global warming.* MIT Press.

Norris, R. D., Bice, K. L., Magno, E. A. & Wilson, P. A. 2002 Jiggling the tropical thermostat in the Cretaceous hothouse. *Geology* **30**, 299–302.

Harrison, S. P., Braconnot, P., Joussaume, S., Hewitt, C. & Stouffer, R. J. 2002 Comparison of palaeoclimate simulations enhances confidence in models. *Eos* **83**, 447.

NRC 2002 *Abrupt climate change: inevitable surprises*. Washington, DC: National Academy Press.

Pearson, P. N., Ditchfield, P. W., Singan, J., Harcourt-Brown, K. G., Nicholas, C. J., Olsson, R. K., Shackleton, N. J. & Hall, M. A. 2001 Warm tropical sea surface temperatures in the late Cretaceous and Eocene epochs. *Nature* **413**, 481–487.

Pearson, P. N., Ditchfield, P. & Shackleton, N. J. 2002 Tropical temperatures in greenhouse episodes-reply. *Nature* **419**, 898.

Peteet, D. M. 2000 Sensitivity and rapidity of vegetational response to abrupt climate change. *Proc. Natl Acad. Sci. USA* **97**, 1359–1361.

Peteet, D. M., Daniels, R. A., Heusser, L. E., Vogel, J. S., Southon, J. R. & Nelson, D. E. 1993 Late-glacial pollen, macrofossils, and fish remains in northeastern USA: the Younger Dryas oscillation. *Quat. Sci. Rev.* **12**, 597–612.

Petit, J. R. (and 18 others) 1999 Climate and atmospheric history of the past 420 000 years from the Vostok ice core, Antarctica. *Nature* **399**, 429–436.

Pinot, S., Ramstein, G., Harrison, S. P., Prentice, I. C., Guiot, J., Stute. M. & Joussaume, S. 1999 Tropical paleoclimates of the Last Glacial Maximum: comparison of Paleoclimate Modelling Intercomparison Project (PMIP) simulations and paleodata. *Climate Dynam.* **15**, 857–874.

Pollard, D. & Thompson, S. L. 1997 Climate and ice-sheet mass balance at the Last Glacial Maximum from the GENESIS Version 2 global climate model. *Quat. Sci. Rev.* **16**, 841–864.

Poulsen, C. J., Barron, E. J., Peterson, W. H. & Wilson, P. A. 1999 A reinterpretation of mid-Cretaceous shallow marine temperatures through model-data comparison. *Paleoceanography* **14**, 679–697.

Raynaud, D., Barnola, J. M., Chappellaz, J., Zardini, D., Jouzel, J. & Lorius, C. 1992 Glacial interglacial evolution of greenhouse gases as inferred from ice core analysis—a review of recent results. *Quat. Sci. Rev.* **11**, 381–386.

Renssen, H. 1997 The global response to Younger Dryas boundary conditions in an AGCM simulation. *Climate Dynam.* **13**, 587–599.

Renssen, H. & Isarin, R. F. B. 2001*a* Data-model comparison of the Younger Dryas event: discussion. *Can. J. Earth Sci.* **38**, 477–478.

Renssen, H. & Isarin, R. F. B. 2001*b* The two major warming phases of the last deglaciation at ~ 14.7 and ~ 11.5 ka cal BP in Europe: climate reconstructions and AGCM experiments. *Global Planet. Change* **30**, 117–153.

Renssen, H., Isarin, R. F. B., Jacob, D., Podzun, R. & Vandenberghe, J. 2001 Simulation of the Younger Dryas climate in Europe using a regional climate model nested in an AGCM: preliminary results. *Global Planet. Change* **30**, 41–57.

Rind, D., Peteet, D., Broecker, W., McIntyre, A. & Ruddiman, W. 1986 The impact of cold North Atlantic sea surface temperatures on climate: implications for the Younger Dryas cooling (11–10 k). *Climate Dynam.* **1**, 3–33.

Rind, D., deMenocal, P., Russell, G., Sheth, S., Collins, D., Schmidt, G. & Teller, J. 2001 Effects of glacial meltwater in the GISS coupled atmosphere–ocean model. 1. North Atlantic deep water response. *J. Geophys. Res. Atmos.* **106**, 27 335–27 353.

Rutter, J. W., Weaver, A. J., Rokosh, D., Fanning, A. F. & Wright, D. G. 2000 Data–model comparison of the Younger Dryas event. *Can. J. Earth Sci.* **37**, 811–830.

Salamatin, A. N., Lipenkov, V. Y., Barkov, N. I., Jouzel, J., Petit, J. R. & Raynaud, D. 1998 Ice core age dating and paleothermometer calibration based on isotope and temperature profiles from deep boreholes at Vostok Station (East Antarctica). *J. Geophys. Res. Atmos.* **103**, 8963–8977.

Schiller, A., Mikolajewicz, U. & Voss, R. 1997 The stability of the North Atlantic thermohaline circulation in a coupled ocean–atmosphere general circulation model. *Climate Dynam.* **13**, 325–347.

Seager, R., Battisti, D. S., Yin, J., Gordon, N., Naik, N., Clement, A. C. & Cane, M. 2002 Is the Gulf Stream responsible for Europe's mild winters? *Q. J. R. Meteorol. Soc.* B **128**, 2563–2586.

Shackleton, N. J. 2000 The 100 000-year ice-age cycle identified and found to lag temperature, carbon dioxide, and orbital eccentricity. *Science* **289**, 1897–1902.

Sharma, M. 2002 Variations in solar magnetic activity during the last 200 000 years: is there a Sun–climate connection? *Earth Planet. Sci. Lett.* **199**, 459–472.

Stocker, T. F., Wright, D. G. & Broecker, W. S. 1992 The influence of high-latitude surface forcing on the global thermohaline circulation. *Paleoceanography* **7**, 529–541.

Stute, M., Forster, M., Frischkorn, H., Serejo, A., Clark, J. F., Schlosser, P., Broecker, W. S. & Bonani, G. 1995 Cooling of tropical Brazil (5 °C) during the Last Glacial Maximum. *Science* **269**, 379–383.

Suarez, M. & Held, I. 1979 The sensitivity of an energy balance climate model to variations in the orbital parameters. *J. Geophys. Res. Oceans* **84**, 4825–4836.

Teller, J. T., Leverington, D. W. & Mann, J. D. 2002 Freshwater outbursts to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation. *Quat. Sci. Rev.* **21**, 879–887.

Vellinga, M. & Wood, R. A. 2002 Global climatic impacts of a collapse of the Atlantic thermohaline circulation. *Climatic Change* **54**, 251–267.

Weaver, A. J., Eby, M., Fanning, A. F. & Wiebe, E. C. 1998 Simulated influence of carbon dioxide, orbital forcing and ice sheets on the climate of the Last Glacial Maximum. *Nature* **394**, 847–853.

Wilson, P. A. & Norris, R. D. 2001 Warm tropical ocean surface and global anoxia during the mid-Cretaceous period. *Nature* **412**, 425–429.

Wilson, P. A., Norris, R. D. & Cooper, M. J. 2002 Testing the Cretaceous greenhouse hypothesis using glassy foraminiferal calcite from the core of the Turonian tropics on Demerara Rise. *Geology* **30**, 607–610.

Zachos, J. C., Pagani, M., Sloan, L., Thomas, E. & Billups, K. 2001 Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* **292**, 686–693.

Zachos, J. C., Arthur, M. A., Bralower, T. J. & Spero, H. J. 2002 Tropical temperatures in greenhouse episodes. *Nature* **419**, 897–898.

Discussion

J. Marotzke (*School of Ocean & Earth Science, University of Southampton, UK*). Only some cold events during the last 100 kyr were preceded by iceberg surge (Heinrich events), whereas many cold events were not. Does this not cast doubt on Heinrich events as triggers of cold spells?

R. B. Alley. Among possible explanations, we have presented one in which a primary millennial oscillation entrains the slower multi-millennial Heinrich oscillation (Alley & Clark 1999). Heinrich events 0, 1, 2, 4 and 5 at least seem to have involved increased ice-rafted-debris transport from the Laurentide ice sheet (Hemming 2003), probably linked to thermally modulated surging (MacAyeal 1993) into an ocean already cooled as part of the higher-frequency Dansgaard–Oeschger oscillation. Following a surge, the MacAyeal (1993) model indicates that the Laurentide ice sheet was frozen to its bed and could not discharge icebergs rapidly. Inland ice-sheet thickening over several millennia is modelled to have eventually thawed the bed; however, faster-response-time marginal processes not explicitly considered in the original MacAyeal (1993) model may have prevented surge initiation until triggered by a cooling. Several plausible triggers exist but none is yet well established (Alley & Clark 1999). The additional meltwater of Heinrich-event icebergs added to an already cold and fresh North Atlantic may then have further interrupted northward heat transfer by the thermohaline circulation, causing climate anomalies associated with Heinrich events to be larger and longer lasting than those associated with Dansgaard–Oeschger oscillations, as observed in some regions (Alley & Clark 1999; Hemming 2003). This clearly does not explain the Dansgaard–Oeschger oscillation. Despite a voluminous literature (see, for example, Alley *et al*. 2001), I doubt that there is yet an explanation that is both complete and widely accepted, but I believe that rapid progress is being made.

M. Harrison (*Hadley Centre, Bracknell, UK*). How do we handle possible abrupt changes in future, given insensitive models not producing such changes in the next 100 years?

R. B. Alley. I have been helping our elder daughter, Janet, learn to drive. Most of her efforts are devoted to staying in lane, regulating speed, making safe turns, parking, and other activities that she must deal with continually. However, as a fast learner and a bright young lady, she is also interested in appropriate responses if a tyre fails at high speed or if an impaired driver crosses the centre line into her lane: events that are possible, that she may never experience, but that are potentially very important. I believe that the IPCC is offering high-quality 'driver education' to policy makers on highly likely greenhouse-gas-induced climate change and its impacts. But policy makers are also interested in those events that have occurred, that may not occur in our lifetimes or in Janet's, but that are potentially very important, and I do not believe that the climate-change community has done well in developing and providing to policy makers the information that they need to apply their skills to these low-probability/high-impact events. As sketched by the Committee on Abrupt Climate Change of the US National Research Council (NRC 2002), an integrated research programme to learn better the events that are possible, how likely such events are, what their impacts might be, and how to avoid such events or enhance resiliency and adaptability to them, could provide the information that would allow policy makers to make wise decisions.

Additional references

Hemming, S. R. 2003 Heinrich events. *Rev. Geophys.* (In the press.)

MacAyeal, D. R. 1993 A low-order model of growth/purge oscillations of the Laurentide ice sheet. *Palaeoceanography* **8**, 767–773.