and carbon cycles. Charlson *et al.*[4] have hypothesized that enhanced DMS emissions from warmer ocean surface waters may produce new cloud condensation nuclei as the DMS converts to n.s.s. SO_4^{2-} in fine, accumulation-mode particles. The resulting sulphate haze would produce a direct radiative cooling influence and the enhanced cloud cover would produce indirect cooling. The DMS that converts to SO_2 (as opposed to methanesulphonate, MSA) may, however, convert further to n.s.s. SO_4^{2-} produced in SSAW, which, being predominantly associated with >2 μm diameter particles, will dry-deposit at a rapid rate[24,25]. From the observed n.s.s. SO_4^{2-} distributions during Bermuda-area CASE[7] or Barbados[8] sampling, we estimate that $>90\%$ of the n.s.s. SO_4^{2-} deposition comes from >1.2 μm diameter particles. A 20-30 hr residence time, given a MBL height of 1-2 km, implies a dry deposition velocity $>1\ \mathrm{cm\ s^{-1}}$ (roughly twice that for SO_2), similar to the n.s.s. SO_4^{2-} dry deposition velocity needed for mass balance of sulphur species in data sets obtained in the Atlantic[10,26] MBL.

Heterogeneous SO_2 conversion in sea-salt particles, combined with the large dry deposition rates of these particles, defines a rapid recycling pathway for sulphur from the oceans (as DMS) and back to the oceans (as n.s.s. SO_4^{2-} in SSAW). This SO_2 removal pathway is highlighted in Fig. 1. Estimated DMS[26] and sea-salt particle[7] emission rates for the summertime western North Atlantic Ocean near Bermuda during CASE were such that at least half of the DMS emitted may have been converted to n.s.s. SO_4^{2-} in sea-salt particles.

Worldwide, the oceanic DMS emission rate has been estimated to be $<10^{12}\ \mathrm{mol\ S\ yr^{-1}}$ (see for example ref. 27) whereas the sea-salt particle emission rate is estimated to be 10^{15}–$10^{16}\ \mathrm{g\ yr^{-1}}$ (see for example ref. 28). If all the carbonate in sea-salt particles is released in the process of producing n.s.s. SO_4^{2-} by O_3-oxidation of SO_2 in SSAW, then $>10^{11}\ \mathrm{mol\ S\ yr^{-1}}$ (between 10% and nearly 100%) of global DMS emissions that convert to SO_2 (MSA excluded) may be rapidly returned to surface ocean waters by the pathway highlighted in Fig. 1. (Additional conversion in the SSAW, for example through metal catalysis[29], will increase the amount of sulphur rapidly recycled through this pathway.) Instead of enhanced haze and cloud albedo resulting from increased DMS emissions, it may be that sulphur is more rapidly exchanged between the surface waters of the global oceans and the MBL. Heterogeneous conversion of sulphur in SSAW and subsequent dry removal from the MBL might be viewed as a 'self-cleansing' process which helps to maintain a steady state (homeostasis) for the global sulphur cycle. □

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Deriving global climate sensitivity from palaeoclimate reconstructions

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To assess the future impact of anthropogenic greenhouse gases on global climate, we need a reliable estimate of the sensitivity of the Earth's climate to changes in radiative forcing. Climate sensitivity is conventionally defined as the equilibrium surface temperature increase for carbon dioxide doubling, $\Delta T_{2\times}$. Uncertainties in cloud processes spread general circulation model (GCM) estimates of this parameter over the range $1.5<\Delta T_{2\times}<4.5$ °C (refs 1, 2). An alternative to model-based estimates is in principle available from the reconstruction of past climates[3–6], which implicitly includes cloud feedback. Here we retrieve the sensitivity of two palaeoclimates, one colder and one warmer than present, by independently reconstructing both the equilibrium surface temperature change and the radiative forcing. Our results yield $\Delta T_{2\times}$ = 2.3 ± 0.9 °C. This range is comparable with estimates from GCMs and inferences from recent temperature observations and ocean models[7,8]. Future application of the method to additional climates in the geological record might constrain climate sensitivity enough to narrow the model uncertainties of global warming predictions.

A change in the global mean surface temperature $\Delta\bar{T}$ is normally produced by a change in net radiative forcing[9],

$$\Delta\bar{Q}=\Delta Q_{\mathrm{Sun}}+\Delta Q_{\mathrm{albedo}}+\Delta Q_{\mathrm{green}}+\Delta Q_{\mathrm{aerosol}} \qquad (1)$$

where $\Delta Q_{\mathrm{Sun}}=(S_0/4)(1-\alpha)(\Delta S/S_0)$ is the forcing from solar irradiance changes ΔS, $\Delta Q_{\mathrm{albedo}}=-(S_0/4)\Delta\alpha$ the forcing from surface albedo changes $\Delta\alpha$, $\Delta Q_{\mathrm{green}}$ $(\Delta CO_2, \Delta CH_4, \ldots)$ the forcing from greenhouse gas concentration changes and $\Delta Q_{\mathrm{aerosol}}(\Delta SO_4, \ldots)$ the forcing from changes in atmospheric aerosols. Present-day solar constant and planetary albedo values are $S_0\approx 1{,}370\ \mathrm{W\ m^{-2}}$ and $\alpha\approx 0.30$. A greenhouse gas reference forcing commonly used in global warming analyses is that from atmospheric CO_2 doubling[9], $\Delta Q_{2\times}\approx 4.4\ \mathrm{W\ m^{-2}}$.

The simplest energy balance for an applied radiative perturbation is $\Delta\bar{Q}=\lambda_b\Delta\bar{T}$, where $\lambda_b\approx 3.8\ \mathrm{W\ m^{-2}\ K^{-1}}$ is the radiative damping coefficient for black-body cooling[10]. In that case, $\Delta T_{2\times}=\Delta Q_{2\times}/\lambda_b\approx 1.2$ °C.

If feedbacks over cloud-covered parts of the Earth were neutral, an imposed radiative forcing would still induce a feedback flux per unit surface area of Earth, $\Delta Q_{\mathrm{clear}}$, by water vapour feedbacks in clear-sky regions. The energy balance including clear-sky feedbacks, $\Delta\bar{Q}+\Delta Q_{\mathrm{clear}}=\lambda_b\Delta\bar{T}$, yields the clear-sky radiative damping, $\lambda_{\mathrm{clear}}\equiv\Delta\bar{Q}/\Delta\bar{T}=\lambda_b-\Delta Q_{\mathrm{clear}}/\Delta T$. Radiative-convective models[11] supported by satellite observations[12] indicate positive feedback $(\Delta Q_{\mathrm{clear}}/\Delta T>0)$ from increased

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infrared opacity of the troposphere as more water evaporates into air columns over warmer surfaces. This reduces radiative damping and increases climate sensitivity relative to black-body cooling. GCMs show positive water vapour feedback, and agree to within ~10% on clear-sky damping and climate sensitivity[1]; $\lambda_{clear} = 2.2 \pm 0.2$ W m^{-2} K^{-1}, $\Delta T_{2\times} = \Delta Q_{2\times}/\lambda_{clear} \approx 2.0 \pm 0.2$ °C.

Lindzen[13] holds that GCMs err on this point and that water vapour feedback is in fact negative ($\Delta Q_{clear}/\Delta T < 0$, because tropical cumulus towers enhanced by global warming dry the upper troposphere enough to reduce infrared opacity worldwide). In that case, $\lambda_{clear} > \lambda_b$; perhaps[8] $\lambda_{clear} \approx 8.8$ W m^{-2} K^{-1} ($\Delta T_{2\times} \sim 0.5$ °C).

In GCMs, the main uncertainty in $\Delta T_{2\times}$ is associated with radiative fluxes, ΔQ_{cloud}, induced by cloud formation and cloud radiative processes when the climate changes. The energy balance $\Delta\bar{Q} + \Delta Q_{clear} + \Delta Q_{cloud} = \lambda_b \Delta\bar{T}$ gives $\lambda \equiv \Delta\bar{Q}/\Delta\bar{T} = \lambda_b - \Delta Q_{clear}/\Delta T - \Delta Q_{cloud}/\Delta T$, or

$$\lambda = \lambda_{clear} - (\Delta Q_{cloud}/\Delta T) \qquad (2)$$

Figure 1 shows that λ and $\Delta T_{2\times}$ of GCMs are uncertain by a factor of three because of uncertainties in cloud radiative feedback. Even the sign of $\Delta Q_{cloud}/\Delta T$ is uncertain.

The Intergovernmental Panel on Climate Change (IPCC) in 1990 projected global warming from a transient ocean–climate model[14] for a range of GCM-derived sensitivities[2], $\Delta T_{2\times} = 1.5$, 2.5 and 4.5 °C. Four future radiative forcing scenarios, A, B, C and D, were examined, representing progressively more constrained greenhouse gas emissions. All the scenarios produced some global warming, but unconstrained emissions (Business as Usual, A; "SA90" in IPCC updates) in combination with $\Delta T_{2\times} = 1.5$ °C produced about the same warming as the most highly constrained emissions (D) with $\Delta T_{2\times} = 4.5$ °C. As D would be much more costly to implement than A, $\Delta T_{2\times}$ uncertainties can severely affect the cost projections of emission controls.

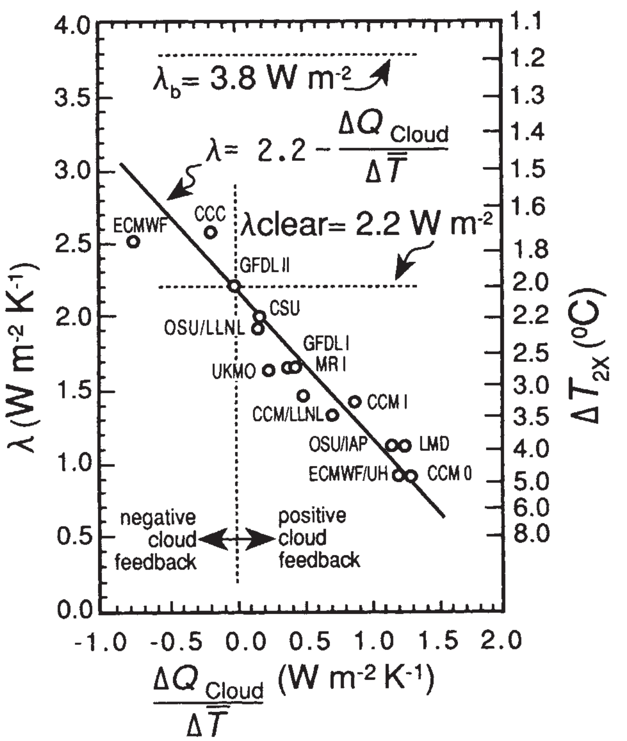

FIG. 1 Radiative damping coefficient, λ, and CO_2 doubling climate sensitivity $\Delta T_{2\times}$ plotted against the cloud radiative feedback parameter of 14 GCMs. Data points are derived from Cess *et al.*[1] where the GCM acronyms are defined (but note the different definition of the symbol λ in ref. 1). The linear correlation (solid line) implies that the differences in GCM-derived climate sensitivity arise mainly from differences in the way that cloud formation and cloud radiation physics are represented in simplified form (parameterized) in different GCMs.

TABLE 1 Sensitivity estimates of two palaeoclimates

Last Glacial Maximum (LGM) cooling ~21.5 kyr BP

Radiative forcing (W m^{-2})	Component mean	Component r.m.s.	Cumulative mean	Cumulative r.m.s.
Component	$\langle \Delta Q_i \rangle$	σ_{Q_i}	$\langle \Delta Q \rangle$	$\sigma_Q = \sqrt{\sum \sigma_{Q_i}^2}$
Sun	0.0	0.2	0.0	0.2
Albedo	−3.0	0.5	−3.0	0.5
Greenhouse	−2.8	0.3	−5.8	0.6
Aerosol	−0.9	0.7	−6.7	0.9
Temperature response (°C)			$\langle \Delta T \rangle$	σ_T
			−3.0	0.6
Climate sensitivity (°C)			$\langle \Delta T_{2\times} \rangle$	$\sigma_{T_{2\times}}$
			2.0	0.5

Middle Cretaceous Maximum (MCM) warming ~100 Myr BP

Radiative forcing (W m^{-2})	Component mean	Component r.m.s.	Cumulative mean	Cumulative r.m.s.
Component	$\langle \Delta Q_i \rangle$	σ_{Q_i}	$\langle \Delta Q \rangle$	$\sigma_Q = \sqrt{\sum \sigma_{Q_i}^2}$
Sun	−1.2	0.2	−1.2	0.2
Albedo	5.8	0.9	4.6	0.9
Greenhouse	11.1	6.7	15.7	6.8
Temperature response (°C)			$\langle \Delta T \rangle$	σ_T
			9.0	2.0
Climate sensitivity (°C)			$\langle \Delta T_{2\times} \rangle$	$\sigma_{T_{2\times}}$
			2.5	1.2

Are more accurate $\Delta T_{2\times}$ estimates possible that are independent of GCMs? One approach is to vary $\Delta T_{2\times}$ in an ocean-climate model[14] to fit the historical temperature record[15]. The radiative forcing history from greenhouse gas buildup (~2 W m^{-2} by 1990 relative to pre-industrial) yields a relatively low sensitivity by this technique[7,8], $\Delta T_{2\times} \approx 1.5$ °C. But the calculations omit effects of increasing sulphate aerosols from fossil-fuel burning which could have produced partially compensating negative forcing (~−1 W m^{-2} by 1990 from direct scattering of sunlight and/or cloud condensation nuclei increasing marine stratus albedo[16]). Wigley and Raper[17] found the historical temperature record could be fitted when anthropogenic sulphate aerosol forcing was included with $\Delta T_{2\times} \approx 3.4$ °C, whereas Schlesinger *et al.*[18] estimate $\Delta T_{2\times} = 2.2 \pm 0.8$ °C with aerosol forcing. Apart from uncertainties in transient forcing and ocean mixing, natural (unforced) variability during the historical temperature period can affect climate sensitivity derived by this method. This is minimized by working with the long-term average palaeoclimates whose forcings and responses are large.

We sought to derive $\Delta T_{2\times}$ for the Last Glacial Maximum (LGM) about twenty thousand years before present (21.5 kyr BP) and the Mid-Cretaceous Maximum (MGM), about 100 Myr BP, periods of large negative and positive climatic perturbation relative to today[19]. Besides the fast feedbacks included in $\Delta T_{2\times}$ (timescales < 1 yr), 'slow' feedbacks (timescales of decade or more) can come into play involving greenhouse gas concentrations, ocean circulation changes[5] and icecap–bedrock dynamics[20]. Following equilibrium climate models[21,22], we treat fast feedbacks as part of the response, slow feedbacks as direct radiative forcing.

For the LGM, we estimated a mean solar radiative forcing of zero with some residual uncertainty based on 'solar-cycle' variations of Sun-like stars[23]. Surface albedo forcing was estimated from changes in glacial ice, vegetation and topography[3,22]. Greenhouse forcing was from IPCC radiative parameterizations for pre-industrial concentrations[9], $(CO_2)_0 = 279$ p.p.m.v.,

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$(CH_4)_0 = 790$ p.p.b.v., and LGM concentrations from the Vostok ice core[3,24], $CO_2 = 195$ p.p.m.v., $CH_4 = 350$ p.p.b.v. Aerosol forcing was based on LGM sulphate particulate loading in ice cores[25] with numerical values from Harvey's[26] LGM base case reduced by a factor of 2 to 10 (ref. 27). For the MCM, a slightly negative solar forcing was estimated based on the Sun brightening by 0.5% per 100 Myr (ref. 28; ±10% uncertainty). Surface albedo forcing was based on calculations of the present Earth relative to the darker (ice-free and more ocean-covered) Earth 100 Myr BP with r.m.s. deviations from uncertainties in surface vegetation and continental boundaries[29]. MCM greenhouse forcing was computed from the parameterization of Kiehl and Dickenson[30] for atmospheric palaeo-CO_2 concentrations 2–11 times present levels estimated from several independent reconstructions[31–35]. Values are given in Table 1. Component uncertainties are treated as r.m.s. deviations σ_{Q_i} about the means $\langle \Delta Q_i \rangle$ of independent random variables ΔQ_i. Also in Table 1 are the cumulative forcing mean and r.m.s. uncertainty computed from[36] $\langle \Delta Q \rangle = \sum \langle \Delta Q_i \rangle$ and $\sigma_Q = (\sum \sigma_{Q_i}^2)^{1/2}$. Our net forcing estimates are -6.7 ± 0.9 W m^{-2} (LGM) and 15.7 ± 6.8 W m^{-2} (MCM).

We estimated the statistical mean and r.m.s. global surface temperature changes, $\langle \Delta T \rangle$ and σ_T, from LGM and MCM palaeotemperature reconstructions as follows. Figure 2*a* shows distributions of zonal mean temperature change reconstructed for the Mid-Holocene (~5–6 kyr BP), Eemian (~125 kyr BP) and Mid-Pliocene (~3.3–4.3 Myr BP) periods normalized to their global mean temperature changes[5]. Budyko and Izrael[6] hypothesized that such normalized surface temperature changes are 'universal'. The least-squares curve fit

$$t(x) = \Delta T(x)/\Delta \bar{T} = 0.36 + 3.21x^4 \qquad (3)$$

where $x = \sin$ (latitude) fits the data of Fig. 2*a* well (correlation coefficient $r^2 = 0.89$), although it is not obvious that it describes temperature distributions of all periods[37]. A counterexample (not shown) may have occurred in the Eocene when sea surface temperatures were apparently cooler in the tropics and warmer at high latitudes[38].

Figure 2*b* shows that Equator-to-pole temperature changes relative to present of both the LGM and MCM are described well by equation (3). The LGM temperature change distribution was synthesized from CLIMAP sea surface temperature (SST) changes equatorward of sea ice[21,22,39] and high-latitude air temperature changes derived from ice cores. The surface air temperature change at Vostok station on the Antarctic plateau is from deuterium isotope fractionation (δD)[24]. Surface cooling relative to present at Dye 3 and Camp Century, Greenland, was derived from oxygen isotope fractionation (δ^{18}O)[40] assuming a change of ~1.0‰ δ^{18}O per degree centigrade[41]. Equator-to-pole temperature distributions in Fig. 2*b* are meant to represent long-term equilibrium climates. The ~9 °C cooling predicted by

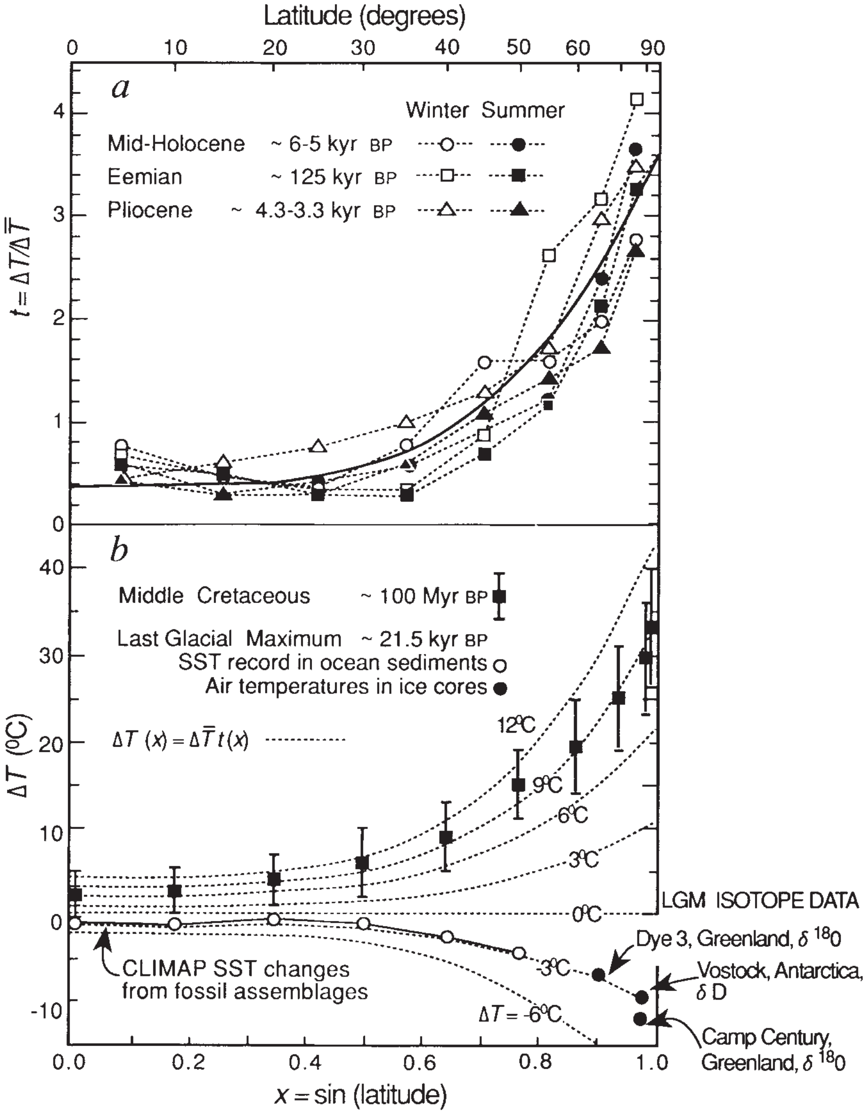

FIG. 2 *a*, Longitudinally averaged surface temperature changes from the present-day climate, normalized by dividing by globally averaged temperature changes. Points connected by dotted lines are based on palaeodata from the mid-Holocene, Eemian and Pliocene warm periods. The solid line is a least-squares fit to this data for a fourth-order polynomial in the sine of latitude (equation (3)). *b*, Longitudinally averaged surface temperature changes from the present-day climate (average of the northern and southern hemispheres). Points are derived from palaeodata for the mid-Cretaceous maximum warming and Last Glacial Maximum cooling (see text). Lines show the best-fit normalized temperature curve of *a* multiplied by the indicated selection of positive and negative global mean temperature changes. Comparison of the lines with a set of points gives the range of global mean temperature change that best fits the data.

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the $\Delta\bar{T} = -3$ °C curve at the latitude of Summit, Greenland (72° N) is a reasonable time-average of alternating mild (−7 °C) and severe (−12 °C) LGM glacial climates observed at this station[42]. MCM data in Fig. 2*b* are surface temperature reconstructions of Barron and Washington[43] minus the present surface air temperature distributions. The points shown are an average response of the two hemispheres, with the error bars showing upper and lower bound (not r.m.s.) uncertainties. Comparison of this data with the $\Delta T(x) = \Delta\bar{T}\,t(x)$ curves, also shown in Fig. 2b for different global mean temperature changes, yielded our $\Delta\bar{T}$ estimates, -3.0 ± 0.6 °C (LGM) and 9.0 ± 2.0 °C (MCM).

The climate sensitivity mean and standard deviation in Table 1 were estimated from

$$\langle \Delta T_{2\times} \rangle \approx \frac{\Delta Q_{2\times} \langle \Delta T \rangle}{\langle \Delta Q \rangle}$$
$$\sigma T_{2\times} \approx \langle \Delta T_{2\times} \rangle \sqrt{\left(\frac{\sigma_Q}{\langle \Delta Q \rangle}\right)^2 + \left(\frac{\sigma_T}{\langle \Delta T \rangle}\right)^2} \qquad (4)$$

The approximation assumes statistically independent radiative forcing and temperature estimates and $(\sigma_Q/\langle \Delta Q \rangle)^2$, $(\sigma_T/\langle \Delta T \rangle)^2 \ll 1$. (Small errors associated with nonlinearities and uncertainties in $\Delta Q_{2\times}$ are neglected.) Our $\Delta T_{2\times}$ climate sensitivities are then 2.0 ± 0.5 °C (LGM) and 2.5 ± 1.2 °C (MCM).

Although our method retrieves a Pleistocene $\Delta\bar{T} \approx -3$ °C, Lorius *et al.*[3] cite LGM global cooling of ∼−4 to −5 °C. If the polar cooling estimated from ice cores is roughly correct, a global mean cooling of −4 to −5 °C implies that the weak (<1 °C) CLIMAP sea surface cooling from the Equator to 30 degrees latitude is a considerable underestimate (see Fig. 2*b*). Preliminary data from palaeothermometers in groundwater[44] and isotopic and micropalaeontological records in ocean sediment cores[45] admit the possibility of LGM tropical and mid-latitude cooling >3 °C. A recalibrated, colder CLIMAP data set along with consistent land data could increase our LGM sensitivity estimate—perhaps to the mid-point of the IPCC range[2], $\langle \Delta T_{2\times} \rangle \approx 3.0$ °C.

Still, it is remarkable that independently estimated sensitivities of the LGM and MCM are so close to each other and to the $\Delta T_{2\times} \approx 2.2$ °C that Schlesinger *et al.*[18] obtained by fitting the historical temperature record to the output of an ocean-climate model forced by anthropogenic greenhouse gases and aerosols. The 'best guess' $\Delta T_{2\times} \approx 3.4$ °C of Wigley and Raper[17], obtained with slightly different assumptions about the net radiative forcing history, is closer to the 'colder' LGM. In general, $\Delta T_{2\times}$ can change as new palaeoclimate estimates are added to the statistical data base. More focused analysis of palaeodata along with palaeo-reconstructions of additional periods could narrow the uncertainties further.

The numbers are already good enough to rule out a very low sensitivity implied by Lindzen[13]. For example, $\Delta T_{2\times} \approx 0.5$ °C requires relatively large, and nonphysical, forcings (∼−26 W m^{-2} to chill the LGM, ∼+80 W m^{-2} to heat the MCM). The climate fluctuations imprinted in the geological record could not have occurred if sensitivity was so sluggish. Interpolation of the IPCC's Business-as-Usual scenario curves[2] to our empirical sensitivity estimates for a variety of choices of the ocean model's adjustable parameters, and our own calculations with the Hoffert *et al.* ocean model[14], indicate that global warming by the end of the next century will be 3-4 °C. Such a warming is unprecedented in the past million years, and represents a secular climatic change much faster than previously experienced by natural ecosystems during glacial-interglacial transitions. □

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Patterns of species diversity in the deep sea as a function of sediment particle size diversity

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UNDERSTANDING the processes that generate and maintain patterns of species diversity is a major focus of contemporary ecological and evolutionary research. In the deep sea, species diversity varies geographically and bathymetrically[1–3], and may attain levels that rival tropical communities[4]. Many hypotheses have been proposed concerning the forces that shape patterns of species diversity in the deep sea[5], but so far it has not been possible to relate these patterns to potential causes in a direct quantitative way. The nature of sediments should be important in structuring deep-sea communities because deposit feeders rely on the sediments for nutrition and comprise most of the organisms in the deep sea[6]. The composition of soft sediment communities is influenced by sediment particle

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