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Earth system sensitivity inferred from Pliocene modelling and data

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Quantifying the equilibrium response of global temperatures to an increase in atmospheric carbon dioxide concentrations is one of the cornerstones of climate research. Components of the Earth's climate system that vary over long timescales, such as ice sheets and vegetation, could have an important effect on this temperature sensitivity, but have often been neglected. Here we use a coupled atmosphere–ocean general circulation model to simulate the climate of the mid-Pliocene warm period (about three million years ago), and analyse the forcings and feedbacks that contributed to the relatively warm temperatures. Furthermore, we compare our simulation with proxy records of mid-Pliocene sea surface temperature. Taking these lines of evidence together, we estimate that the response of the Earth system to elevated atmospheric carbon dioxide concentrations is 30–50% greater than the response based on those fast-adjusting components of the climate system that are used traditionally to estimate climate sensitivity. We conclude that targets for the long-term stabilization of atmospheric greenhouse-gas concentrations aimed at preventing a dangerous human interference with the climate system should take into account this higher sensitivity of the Earth system.

Since the 1979 National Research Council report[1], the concept of climate sensitivity has been discussed extensively (see, for example, refs 2–4). It is usually defined as the increase in global mean temperature owing to a doubling of CO_2 after the 'fast' short-term feedbacks, typically acting on timescales of years to decades, in the atmosphere and upper ocean have had time to equilibrate[5]. These fast feedbacks correspond to the physics available in climate models *circa* 1980 (see, for example, ref. 6), and include, for example, water vapour, snow albedo, sea-ice albedo and clouds. This sensitivity (described hereafter as the 'Charney' sensitivity) remains a useful benchmark for comparing different climate models in idealized circumstances, and has been one of the central concepts used by the Intergovernmental Panel on Climate Change in their assessments of future climate change[7,8].

However, there are many other processes operating, over a variety of timescales, that have a role in determining the ultimate response of the climate system to a rise in greenhouse gases. Specifically, changes in dust and other aerosols, vegetation, ice sheets and ocean circulation will all modify the eventual equilibrium surface temperature response to a given CO_2 forcing[9,10]. We term this temperature response the 'Earth system sensitivity'—the long-term equilibrium surface temperature change given an increase in CO_2, including all Earth system feedbacks, but neglecting processes associated with the carbon cycle itself, such as marine productivity[11] or weathering[12]. The Earth system sensitivity may be significantly different to Charney sensitivity because the feedbacks related to, for example, ice sheets—principally through albedo and topographic effects[13], and vegetation—through albedo and hydrological feedbacks[14], are likely to be significant.

Estimating Earth system sensitivity

In theory, the relationship between Charney sensitivity and Earth system sensitivity could be investigated in a pure modelling framework, with the integration of ice sheet, vegetation, aerosol and other components into state-of-the-art climate models. However, such Earth system modelling is in its infancy—ice sheet models in particular are undergoing a period of rapid development at present (see, for example, refs 15 and 16). Furthermore, owing to the long timescales of the processes involved, the equilibrium state of these Earth system models could take tens of thousands of years of model integration to obtain, making them impractical with current computing power.

It has been suggested[9] that Earth history and the palaeo proxy record provides a potentially unique opportunity to estimate Earth system sensitivity, given multi-millennial records of palaeo temperature and CO_2 forcing, for example from the Quaternary ice core record. Using this data approach, Earth system sensitivity has been estimated[9] to be approximately double that of Charney sensitivity. However, for such an approach to be reliable, an abundance of evenly distributed and highly accurate/precise proxy data is required to determine past global mean surface temperature, as are robust measurements of atmospheric CO_2 concentration and variation over long timescales. However, proxy temperature reconstructions are geographically biased and of variable quality. Where the proxy data are most abundant, in the Quaternary period, atmospheric CO_2 concentrations were never significantly higher than pre-industrial[17], reducing the relevance of this period for future climate change considerations.

However, a combined palaeoclimate modelling and data approach can be used to investigate the concept and significance of Earth system sensitivity. Providing that the palaeoenvironmental boundary conditions (for example, CO_2, vegetation, ice sheets) are sufficiently known, they can be prescribed in a climate model, which can be used to provide a truly global estimate of surface temperature for the given CO_2 forcing. This estimate can be assessed by proxy data wherever it is available. Furthermore, imposing the

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Table 1 | Various estimates of ESS (°C), CS (°C) and the ratio ESS/CS.

		CS	ESS	CS_{2*CO_2}	ESS_{2*CO_2}	ESS/CS
1	Control	1.57	2.28	3.04	4.42	1.45
	Nonlinearity					
2	λ depends on orography	1.57	2.26	3.04	4.39	1.44
3	Forcings add nonlinearly	1.57	2.28	3.04	4.42	1.45
	Methodology					
4	Ice/vegetation respond to local temperature forcing	1.57	2.31	3.04	4.48	1.47
	Uncertainty					
5	Error in F_c	1.57	2.28	3.04	4.42	1.45
6	Error in λ_l	1.57	2.09	3.04	4.06	1.33
7	Error in F_o	1.57	2.28	3.04	4.42	1.45
8	Error in λ_s	1.48	2.09	2.87	4.06	1.41
9	Ratio of SSTs	1.57	2.08	3.04	4.03	1.32

The control assumes a linear system, and no errors in the mid-Pliocene simulation. Estimates 2–3 test the assumption of linearity. Estimate 4 makes a more physically based assumption that the ice and vegetation feedbacks respond to the local temperature, rather than the global mean forcing. Estimates 5–9 make various assumptions about the nature of the error in the mid-Pliocene simulation. The values of CS_{2*CO_2} and ESS_{2*CO_2} are the implied Charney and Earth system sensitivities for a doubling of CO_2, calculated from CS and ESS by multiplying by log(560/280)/log(400/280) = 1.94. For more details of all of the estimates, see Supplementary Information.

long-term feedbacks directly from palaeo observations is more robust than calculating them using the pure modelling approach, the results of which are likely to be highly model dependent. The mid-Pliocene warm period (about 3.3–3 million years ago) is a unique time slab to test the concept of Earth system sensitivity because atmospheric CO_2 concentrations[18,19] and temperatures[20] were higher than pre-industrial, climatic fluctuations on orbital timescales were much reduced relative to the Quaternary[21] and data sets exist[22] that allow at least two of the important longer term feedbacks, vegetation and ice sheet extent, to be addressed. For this period, changes in continental configuration are negligible, and the main external forcings are orographic changes[22], and the elevated CO_2 (probably driven by very long-term shifts in the balance between tectonic-related emissions and weathering[23]). The climatic response induced by these forcings will include vegetation and ice sheet changes, and as both the forcings and vegetation and ice sheet changes are reasonably constrained by the geological record, they can be imposed independently in a model. The Pliocene Research, Interpretation and Synoptic Mapping (PRISM) project[20,22,24] has produced data sets of orography, vegetation, and ice sheet extent and elevation for the mid-Pliocene. These are shown and discussed in the Supplementary Information. Mid-Pliocene atmospheric CO_2 has been reconstructed by a variety of proxies[18,19], and a value of 400 ppmv, including a likely contribution from non-CO_2 greenhouse forcing, is within the uncertainties of the reconstructions (typically[18] a mean value of 380 ppmv with maxima as high as 425 ppmv), and has been used in previous modelling studies of the mid-Pliocene climate (see, for example, ref. 25). As for the current climate, there were probably other feedbacks, for example, aerosols and atmospheric chemistry, that had a role, and so our analysis will not be complete, but will provide a closer approximation of Earth system sensitivity than has been achieved[9] thus far.

Given that mid-Pliocene PRISM orography was substantially lower than modern in some regions (especially the Rocky Mountains in North America), it is likely that the reconstructed mid-Pliocene vegetation and ice sheets include a contribution that is due to the modified orographic forcing, and not directly to the CO_2 forcing. To calculate the Earth system sensitivity, which does not depend on the orographic forcing, we therefore need to take account of this. We carry out an ensemble of general circulation model (GCM) simulations that include various combinations of CO_2 and orographic forcings and vegetation and ice feedbacks, and make use of standard forcing/feedback analysis techniques (see, for example, refs 5 and 26). We initially carry out four GCM simulations (see Supplementary Information for more details). A simulation that has boundary conditions j and k modified from pre-industrial to mid-Pliocene we name E_{jk}. The four boundary conditions considered are atmospheric CO_2 (c), orography (o), vegetation (v) and ice sheets (i). Thus, a pre-industrial simulation is E, a mid-Pliocene simulation is E_{ociv} and, for example, a simulation with pre-industrial ice and vegetation but mid-Pliocene orography and CO_2 is E_{oc}. The corresponding surface air temperature distribution in these simulations we name T, T_{ociv} and T_{oc}, respectively. The temperature change distributions relative to pre-industrial in the last two simulations are ΔT_{ociv} and ΔT_{oc}, with global means denoted as $\langle \Delta T_{\mathrm{ociv}} \rangle$ and $\langle \Delta T_{\mathrm{oc}} \rangle$. The four simulations we initially carry out are $E, E_{\mathrm{c}}, E_{\mathrm{oc}}$ and E_{ociv}. We also define CS as the Charney temperature response to a CO_2 forcing from 280 to 400 ppmv, and similarly for the Earth system response, ESS. Given the logarithmic dependence of forcing on CO_2 concentration, traditional climate sensitivity, usually defined as the Charney temperature response to a doubling of CO_2, is given by $CS_{2*CO_2} \approx 1.9CS$.

If a climate system in equilibrium is perturbed by a radiative forcing at the top of the atmosphere, F ($\mathrm{Wm^{-2}}$), the system will eventually reach a new equilibrium, with a global mean surface temperature changed by $\langle \Delta T \rangle$ (K) relative to the unperturbed state. A climate feedback parameter can be defined, $\lambda(\mathrm{W\,m^{-2}K^{-1}})$, such that $\langle \Delta T \rangle = F/\lambda$ (ref. 26). λ represents the combination of many feedback processes, such as sea ice albedo and clouds, as well as emission of long-wave radiation at the surface. It is often assumed that independent radiative forcings add linearly, that different components of λ add linearly and that λ is independent of the type of forcing. Under these assumptions, we have

$$\langle \Delta T_{\mathrm{c}} \rangle = \frac{F_{\mathrm{c}}}{\lambda_{\mathrm{s}}} \tag{1}$$

$$\langle \Delta T_{\mathrm{oc}} \rangle = \frac{F_{\mathrm{c}} + F_{\mathrm{o}}}{\lambda_{\mathrm{s}}} \tag{2}$$

$$\langle \Delta T_{\mathrm{ociv}} \rangle = \frac{F_{\mathrm{c}} + F_{\mathrm{o}}}{\lambda_{\mathrm{s}} + \lambda_{\mathrm{l}}} \tag{3}$$

where F_{c} is the forcing owing to a CO_2 increase from 280 to 400 ppmv, F_{o} is the forcing owing to a decrease in orography from

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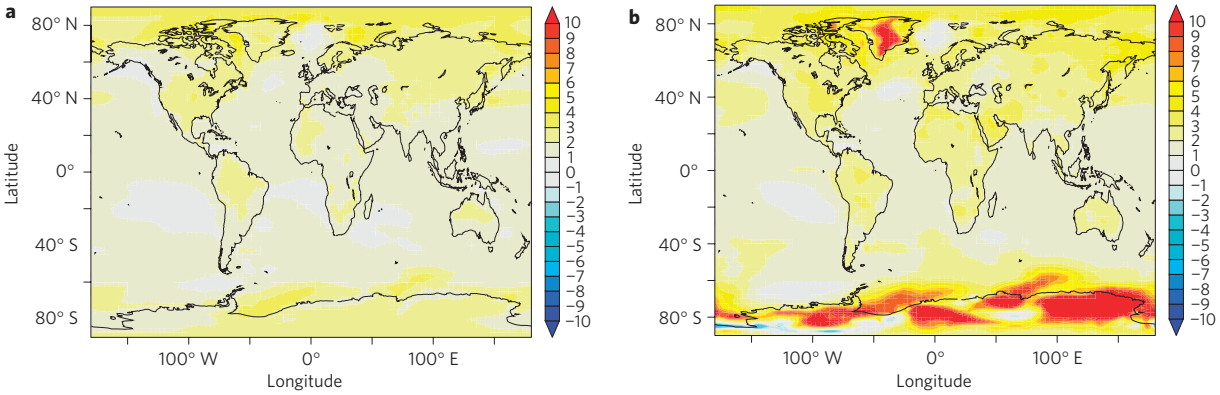

Figure 1 | Charney sensitivity and Earth system sensitivity. a, CS = ΔT_c (°C). **b**, ESS (°C) calculated from Supplementary Equations S16 and S17.

modern to mid-Pliocene, λ_s is the climate feedback parameter for long-wave emission and the snow, sea ice and other short-term feedbacks combined and λ_l is the climate feedback parameter for the vegetation and ice sheet long-term feedbacks. From equation (1), the Charney sensitivity $CS = F_c/\lambda_s$ is equal to $\langle \Delta T_c \rangle = 1.6\,°C$, and in equation (3) the total mid-Pliocene temperature change is $\langle \Delta T_{ociv} \rangle = 3.3\,°C$. The Earth system sensitivity, $ESS = F_c/(\lambda_s + \lambda_l)$, can also be calculated from equations (1)–(3) as

$$ESS = \langle \Delta T_{ociv} \rangle \frac{\langle \Delta T_c \rangle}{\langle \Delta T_{oc} \rangle} = 2.3\,°C \qquad (4)$$

Equation (4) gives the global mean temperature change expected for a stabilized future climate at 400 ppmv (about half the radiative forcing of a CO_2 doubling from pre-industrial), with equilibrated ice sheets and vegetation. In this case, the ratio ESS/CS = 1.45, meaning that the Earth system sensitivity is about 45% greater than the equivalent Charney sensitivity (Table 1).

The above analysis assumes that the climate system is linear; however, studies have highlighted the existence of significant nonlinearities (see, for example, ref. 27). In Supplementary Section S4, we test several assumptions about the nature of such nonlinearities. By making use of a further GCM simulation, E_o, we show that our calculated value of the ratio ESS/CS varies between 1.44 and 1.45 depending on whether we assume a linear system, a system in which the climate feedback parameter depends on the forcing or a system in which forcings add nonlinearly. Furthermore, in Supplementary Section S5, we present an alternative analysis, in which we take a more physically based approach, by assuming that the vegetation and ice sheets respond to the local temperature change induced by the CO_2 and orography forcing, rather than the global mean forcing. In this case, the ratio ESS/CS is slightly higher (1.47), and we can calculate the geographical distribution of Earth system sensitivity and compare it with the Charney sensitivity (Fig. 1). This clearly shows the impact on temperature of the reduction in Antarctic and Greenland ice volume and vegetation changes in the tropics. Table 1 shows the values of CS, ESS and the ratio ESS/CS for our various approaches and assumptions.

Model evaluation relative to mid-Pliocene SST data

To have confidence in our predictions of Earth system sensitivity, it is essential to evaluate the model performance. Two of our simulations can be evaluated relative to observations—the control E, and the mid-Pliocene E_{ociv}. The control climate of the GCM has previously been extensively assessed[28], and it performs very well in comparison with other GCMs according to a variety of metrics[8,29].

The simulated mid-Pliocene surface air temperature change, ΔT_{ociv}, is shown in Fig. 2. The global mean change,

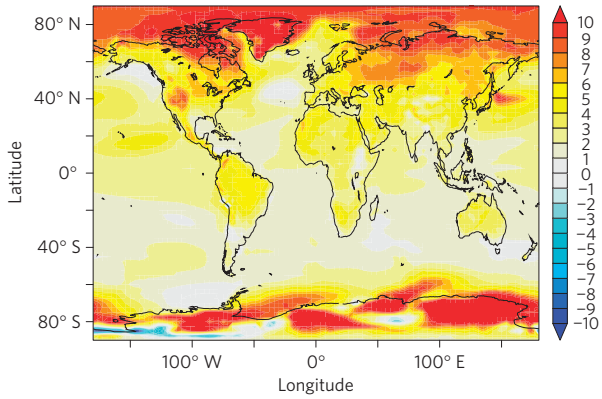

Figure 2 | Simulated mid-Pliocene temperature change. Modelled mid-Pliocene minus pre-industrial surface air temperature, ΔT_{ociv} (°C).

$\langle \Delta T_{ociv} \rangle = 3.3\,°C$, is consistent with previous independent modelling work[30,31]. Sea surface temperatures (SSTs) and continental climate from an older version of our mid-Pliocene model simulation have been assessed relative to reconstructions of SST from the PRISM2 data set[25,32], and palaeobotanical data[33] (this terrestrial analysis is presented for our more fully spun-up simulations in Supplementary Section S7). However, these assessments were carried out at a limited number of locations, whereas to quantify the uncertainty in our estimate of Earth system sensitivity we require the error in the global mean. Owing to limited data coverage, this is impossible to obtain directly, but we can make use of the recently produced PRISM3 global SST data set[34] to provide an estimate. This data set is underpinned by data from 86 sites[20,34], which provide estimates of SST based on a combination of faunal analysis of planktic foraminifera, Mg/Ca and alkenones. The global SST data set is produced by interpolation and extrapolation into data-sparse regions, based primarily on modern SST zonal gradients, and informed by expert palaeoceanographic knowledge. The data set is not infallible, but it does represent our best current estimate of global mid-Pliocene SSTs. The SSTs in the global data set, as well as the sites from which it is derived, are shown in Fig. 3a relative to pre-industrial SST for the period 1901–1920 from the HadISST data set[35]. The global mean observed SST change, mid-Pliocene minus pre-industrial, ΔSST_{obs} is 1.67 °C. This compares very favourably with our modelled SST change, ΔSST, calculated from E_{ociv} and E, of 1.83 °C. However, the spatial distribution of SST change in the model (Fig. 3b) is not in such good agreement with the data. In particular, the model does not reproduce the large increases in SST in the PRISM3 data set in the Atlantic sector of the Arctic. However, in the zonal mean there are some similarities between the model and data, such as enhanced warming

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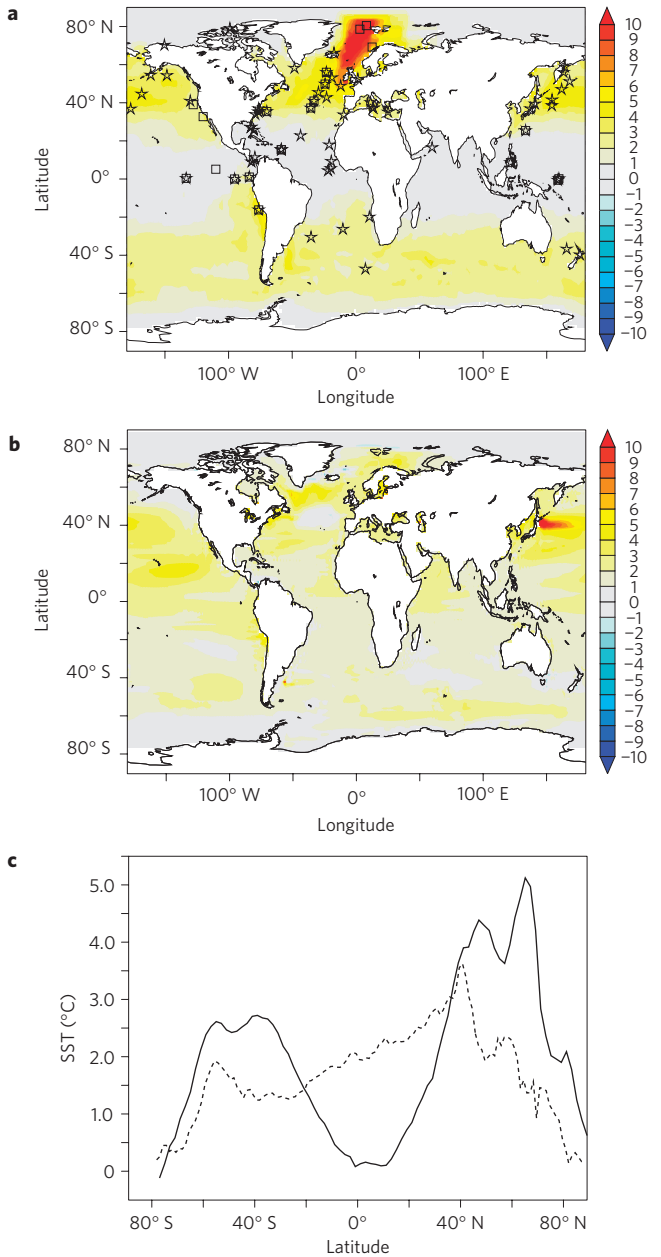

Figure 3 | Model–data comparison. a, Annual mean SST difference (°C), PRISM3 SST minus HadISST (1901–1920). The symbols show the location of the individual sites in the PRISM3 data set. Stars are from faunal analysis, squares are from Mg/Ca and alkenone data. **b**, Annual mean SST difference (°C) between model simulations E_{ociv} and E. **c**, Zonal mean of **a** (solid line) and **b** (dashed line).

at mid-latitudes (Fig. 3c). The difference between modelled and observed SST change, $\delta\mathrm{SST} = 0.15\,^{\circ}\mathrm{C}$, is probably an underestimate of the error in $\Delta T_{\mathrm{ociv}}, \delta T$, because temperature changes on land and in regions of sea ice are in general larger than changes in open ocean. Assuming that $\delta\mathrm{SST}/\delta\mathrm{T}$ is proportional to $\Delta\mathrm{SST}/\Delta T_{\mathrm{ociv}}$, we can infer that the error in our model estimate of ΔT_{ociv} is about 0.3 °C (the model overestimating the mid-Pliocene temperature change by this amount).

Estimating uncertainty in the Earth system sensitivity

The formulation of equations (1)–(3), coupled with our comparison with mid-Pliocene data, allows us to investigate the uncertainty in our estimate of Earth system sensitivity. For this uncertainty analysis, presented in detail in Supplementary Section S6, we make various assumptions about the nature of the error in our estimate of mid-Pliocene temperature change, $\delta T = 0.3\,^{\circ}\mathrm{C}$. Assuming an error in F_{c} implies that our imposed value of 400 ppmv for the mid-Pliocene is wrong (in our case, too high, given that our mid-Pliocene simulation is too warm). Assuming an error in λ_{l} implies that there are either errors in the PRISM vegetation and/or ice boundary conditions (for example, the prescribed mid-Pliocene Antarctic ice sheet is too small), or errors in the way we have implemented these boundary conditions (for example, by assigning an albedo to boreal forest that is too low). Assuming an error in F_{o} implies either that the PRISM3 orography is wrong (in our case, too low), or that there is an error in the way the model responds to the orography (for example, if the lapse rate in the model is too large). Assuming an error in λ_{s} implies that the short-term climate sensitivity of the GCM is wrong (too high in our case). The final assumption we make is that our estimate of Charney sensitivity is robust, but that our estimate of Earth system sensitivity is overestimated by a factor equal to the ratio of our predicted mid-Pliocene SST change to observed mid-Pliocene SST change, $\Delta\mathrm{SST}/\Delta\mathrm{SST}_{\mathrm{obs}}$. This in effect uses the model to convert observed mid-Pliocene SSTs to an estimate of Earth system sensitivity.

As shown in Table 1, none of these assumptions greatly changes our estimate of ESS/CS—across all of the analyses presented in this article, the smallest value of ESS/CS we obtain is 1.3, and the largest is 1.5.

Our combined modelling and data approach results in a smaller response (ESS/CS ~ 1.4) than has recently been estimated[9] using palaeo data from the Last Glacial Maximum, 21,000 years ago (ESS/CS ~ 2). This is probably due to the fact that transitions from glacial to interglacial conditions in the Quaternary involve large changes in the Laurentide and Eurasian ice sheets (see, for example, ref. 36), which result in a significant large-scale albedo feedback in these regions that is irrelevant for climates warmer than present. Furthermore, the main driver of Quaternary climate change is ultimately orbital forcing, which is close to zero in the global mean, and is therefore difficult to reconcile with a traditional climate sensitivity analysis.

Implications and outlook

Traditionally, the Intergovernmental Panel on Climate Change has focused on Charney sensitivity[7,8], and groups have used Charney equilibrium scenarios to determine the degree of emissions likely to lead to 'dangerous' climate change (see, for example, refs 37 and 38). Our work argues that the equilibrium climate change associated with an increase of CO_2 is likely to be significantly larger than has traditionally been estimated.

How long the Earth system takes to reach this equilibrium cannot be addressed in this modelling framework. Although vegetation may take several centuries to reach close to equilibrium[39], ice sheets may take millennia[40]. However, observations[41] and modelling[42] indicate that ice sheets may equilibrate much faster, in part owing to surface melt water entering crevasses and decreasing basal friction. Given the uncertainties in the timescale for vegetation and ice-sheet responses, estimates of the impacts of long-term greenhouse-gas stabilization scenarios should focus on the Earth system sensitivity rather than the traditional Charney sensitivity.

Future work in this field should explore the uncertainty in the mid-Pliocene boundary conditions and their influence on estimates of Earth system sensitivity, as well assessing the relationship between the Earth system and Charney sensitivities with more than one set of model parameters, and with more than one model. Furthermore, considerable effort should be applied to developing more complete and efficient high-resolution Earth system models that can be run to equilibrium under future climate scenarios, and evaluated relative to the palaeo record, including the mid-Pliocene.

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Methods

We use the UK Met Office fully coupled atmosphere–ocean GCM, HadCM3[28], to carry out the five GCM simulations: E, E_c, E_{oc}, E_o and E_{ociv}. Both the E and E_{ociv} simulations are over 1,100 years long, and E_c is over 500 years long. These three simulations are all continuations of simulations presented by Lunt *et al.*[43]. The other two simulations, E and E_{oc}, are 200 years long. The shorter integration is appropriate for these simulations, as they are initialized from longer simulations with the same atmospheric CO_2 concentration, resulting in a faster spin-up.

For a full description of the model, the simulations, the boundary conditions used and further information on the analysis, see Supplementary Information.

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Author contributions

D.J.L. carried out the GCM simulations and analysis. D.J.L., A.M.H., G.A.S. and P.J.V. were involved in the study design. H.J.D. developed the PRISM mid-Pliocene boundary conditions and the PRISM3 SST data set. U.S. carried out the model–data comparison with the vegetation data set. All authors discussed the results and commented on the manuscript.

Additional information

The authors declare no competing financial interests. Supplementary information accompanies this paper on www.nature.com/naturegeoscience. Reprints and permissions information is available online at http://npg.nature.com/reprintsandpermissions. Correspondence and requests for materials should be addressed to D.J.L.

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