Making sense of palaeoclimate sensitivity

PALAEOSENS Project Members ‡

Many palaeoclimate studies have quantified climate changes of the pre-anthropogenic past to calculate climate sensitivity (temperature change in response to change in the radiative forcing of climate), but a lack of consistent methodologies produces a wide range of estimates and hinders comparability of results. Here we present a stricter approach to improve inter-comparison of estimates from palaeoclimate data for the past 65 million years in a manner compatible with equilibrium projections for future climate change. Following this approach, we find a first estimate of 0.4-1.4 $K(Wm^{-2})^{-1}$, similar to the 0.6-1.2 $K(Wm^{-2})^{-1}$ compiled by the IPCC[Ref.1].

Motivation

Characterising the complex responses of climate to changes in the radiation budget requires the definition of consistent measurement indices. One such index, climate sensitivity, represents the global equilibrium surface temperature response to the radiative forcing caused by a doubling of atmospheric CO_2. Despite progress in modelling and data acquisition, uncertainties remain regarding the exact value of climate sensitivity and its potential variability through time. The range of climate sensitivities in climate models used for Intergovernmental Panel for Climate Change (IPCC) Assessment Report 4 is 2.1-4.4 K [Ref.1], or 0.6-1.2 K warming per Wm^{-2} of forcing. The new class of Coupled Model Intercomparison Project 5 (CMIP-5) models suggests a similar range[2]. Observational studies have not narrowed this range[3], and indicate that the upper limit is particularly difficult to estimate.

Large palaeoclimate changes can be used to estimate climate sensitivity on centennial to multi-millennial timescales, when estimates of both global mean temperature and radiative perturbations linked with slow components of the climate system (e.g., carbon cycle, land ice) are available (Figure 1). Here we evaluate published estimates for climate sensitivity from a variety of geological episodes (Table 1). However, we find that intercomparison is hindered by major differences in the definition of climate sensitivity among the various studies (Table 1). Improvements in quantifying climate sensitivity from palaeodata clearly require a consistent definition of which processes are included and excluded in the estimated sensitivity, much like the need for strict taxonomy in biology. The definition must agree as closely as possible with that used in modelling studies of past and future climate, while remaining sufficiently pragmatic (operational) to be applicable within the context of the limitations and challenges of extracting quantitative environmental data from different climate states in the geological past.

Here we propose a consistent operational definition for palaeoclimate sensitivity, and we illustrate how a tighter definition narrows the range of reported estimates. Consistent intercomparison is crucial to detect systematic differences in sensitivity values, for example due to changing continental configurations, different climate background states, and the types of radiative perturbations considered. These differences may then be evaluated in terms of additional controls on sensitivity to radiative changes, such as those arising from plate tectonics, weathering cycles,

‡ A full list of authors and their affiliations appears at the end of the paper

changes in ocean circulation, non-CO_2 greenhouse gases, enhanced water-vapour and cloud feedbacks under warm climate states, etc. The palaeoclimate record of the last 65 million years allows such investigations across geological episodes with very different climates, both warmer and colder than today. Clarifying the dependence of feedbacks, and therefore climate sensitivity, on the background climate state is a top priority, because it is central to the utility of past climate sensitivity estimates in assessing the credibility of future climate projections[1,4].

Quantifying climate sensitivity

In the climate modelling community, 'equilibrium climate sensitivity' is classically defined as the simulated global mean surface air temperature increase (ΔT, in K) in response to a doubling of atmospheric CO_2, starting from pre-industrial conditions (which corresponds to a radiative perturbation, ΔR, of 3.7 Wm^{-2})[Refs.1,4]. This definition is valid only in a pure modelling framework, and we introduce the less restrictive definition of the 'climate sensitivity parameter' as the mean surface temperature response to any radiative perturbation ($S = \Delta T/\Delta R$; where T and R are centennial to multi-millennial averages). For brevity and simplicity, we refer here to S as 'climate sensitivity', with a definition most suitable for palaeoclimatic studies on geological timescales.

In the definition of S an initial perturbation ΔR_0 leads to a temperature response ΔT_0 following the Stefan-Bolzmann Law, which is the temperature-dependent blackbody radiation response. This is often referred to as the Planck reponse[5], with a value S_0 of about 0.3 $K(Wm^{-2})^{-1}$ for the present-day climate[6,7]. The radiative perturbation of the climate system is increased (weakened) by various positive (negative) feedback processes, which operate at a range of different timescales (Figure 1). Because the net effect of positive feedbacks is found to be greater than that of negative feedbacks, the end result is an increased climate sensitivity relative to the Planck response.

Importantly, all feedbacks, and thus the calculated climate sensitivity, may depend in a – largely unknown – nonlinear manner on the state of the system prior to perturbation; the 'background climate state' [8-18]. The relationship of S with background climate state differs among climate models[15,19-21]. Regardless, a suggestion of asymmetry is found in a data-comparison of climate sensitivity for the last 800,000 years with that for the Last Glacial Maximum (LGM) (*Supplementary Table 1*, based on Ref.7). Climate sensitivity for the last 800,000 years, calculated relative to CO_2 forcing and corrected for the radiative impacts of ice-sheet variations (see below)[7], clearly shows fluctuations through time by almost 0.5 $K(Wm^{-2})^{-1}$ around a mean of about 1.1 $K(Wm^{-2})^{-1}$, while its range for the LGM alone occupies only the lower half of this distribution (Figure 2). Still, no simple relationship with the general climate state is apparent.

'Fast' versus 'slow' processes

Climate sensitivity depends on processes that operate on many different timescales, from seconds to millions of years, due to both direct response to external radiative forcing, and internal feedback processes (Figure 1). Hence, the timescale over which climate sensitivity is considered is critical. A somewhat artificial, yet operationally pragmatic decision is needed to categorise and process as 'slow' or 'fast', depending on the timescale of interest (see *Supplementary Information*), the resolution of the (palaeo-)records considered, and the character of changes therein. If a process results in temperature changes that reach steady state slower than the timescale of the underlying radiative perturbation, then it is considered 'slow'; if it is faster/coincident, then it is ‘fast’. As such, a process may be categorised 'fast' or 'slow' depending on the particular underlying perturbation considered. Although further distinctions are possible within the ‘fast’ category[3], these are not relevant to the concepts developed here.

In the present-day context, the atmospheric greenhouse gas (GHG) concentrations and the radiative perturbation due to anthropogenic emissions increase much faster than observed for any natural process within the Cenozoic[22-24]. For the present, therefore, the relevant timescale (τ) is the emission timescale, which can be taken as 100 years[Ref.25]. Processes can then be distinguished as faster or slower than τ. Ocean heat uptake plays out over multiple centuries. Combined with further ‘slow’ processes, it causes climate change over the next few decades to centuries to be dominated by the so-called ‘transient climate response (TCR)’ to radiative changes that result from changing GHG concentrations and aerosols[6,26]. After about 100 years, this TCR is thought to amount to roughly two-thirds of the equilibrium (see below) climate sensitivity[6,27]. Climate models account for the feedbacks from changes in water-vapour content, lapse rate, cloud cover, snow and sea-ice albedo[28]. These are the so-called fast feedbacks that equilibrate within a few years following a radiative perturbation, and the resulting response is often referred to as the ‘fast-feedback’ or ‘Charney’ sensitivity[25]. To approximate the ‘equilibrium’ value of that climate sensitivity, accounting for ocean heat uptake and further slow processes, models might be run over centuries with all the associated computational difficulties of doing so[29-32], or alternative approaches may be used that exploit the energy balance of the system for known forcing or extrapolation to equilibrium[33].

The long-term palaeoclimate record cannot be used to constrain the decadal to centennial response of the climate system to ongoing and future GHG and aerosol emissions. This is because the past perturbations typically were slower and more gradual, and because the temporal resolution and dating accuracy of palaeoclimate proxy records are rarely better than centennial. In palaeoclimate studies, not only the well-recognised fast feedbacks, but also other (slower) changes in the climate system need to be addressed to quantify the full climate system response to a radiative perturbation; the so-called slow feedback processes. In palaeoclimate studies, a pragmatic distinction has therefore emerged to distinguish ‘fast’ and ‘slow’ processes relative to the timescales of temperature responses measured in palaeodata, where ‘fast’ is taken to apply to processes up to centennial scales, and ‘slow’ to processes with timescales close to millennial or longer. Thus, changes in natural GHG concentrations are governed by ‘slow’ feedbacks related to global biogeochemical cycles with timescales of centuries and longer (Figure 1). Similarly slow are the radiative influences of vegetation-albedo feedbacks that depend on centennial-scale changes in global vegetation cover, and in global ice area/volume (continental ice sheets, with centennial to millennial timescales).

Other processes have both fast and slow components, which presents additional complications. For example, palaeorecords of atmospheric dust deposition show that important aerosol variations have happened on decadal to astronomical timescales[34-38], reflecting both slow controlling processes related to ice-volume and land-surface changes, and fast processes related to changes in atmospheric circulation. A further complication arises from the lack of adequate global atmospheric dust data for any geological episode except the LGM [e.g.39,40], even though that is essential because the spatial distribution of dust in the atmosphere tends to be very inhomogeneous and because temporal variations in some locations tend to take place over several orders of magnitude[34-38]. Moreover, palaeoclimate modells generally struggle to account for aerosols, with experiments neither prescribing nor implicitly resolving aerosol influences. So far, understanding of aerosol/dust feedbacks remains weak and in need of improvements in both data coverage and process modelling, especially given that dust forcing may account for some 20% of the glacial-interglacial change in the radiative budget [e.g.7,41].

In summary, the concept of 'fast' and 'slow' feedbacks may be difficult to apply, depending on the context of each individual study. In order to compare results between a variety of studies, it is therefore most effective to consider only the classical 'Charney' water-vapour, cloud, lapse rate, and snow and sea-ice feedbacks[25] as 'fast', and all other feedbacks as 'slow'. In addition, palaeoclimate studies generally do not address the TCR that dominates present-day changes, but do capture a more complete longer-term system response. This response may be characterised by the term 'quasi-equilibrium' because climate is never fully in equilibrium on all timescales, but for brevity we follow many studies in referring to this concept as 'equilibrium' (steady-state) climate sensitivity. An appropriate comparison would compare values from palaeodata with equilibrium climate sensitivity values in climate models, following operational decisions to distinguish between forcing and slow feedbacks.

Forcing and slow feedbacks

The external drivers of past natural climate changes mainly resulted from changes in the sun's intensity over time[42,43], from temporal and spatial variations in insolation due to changes in astronomical parameters[44-46], from changes in continental configurations[17,47], and from geological processes that directly affect the carbon cycle (e.g., volcanic outgassing). However, the complete Earth system response to such forcings as recorded by palaeodata cannot be immediately deduced from the (equilibrium) 'fast feedback' sensitivity of climate models, because of the inclusion of slow feedback contributions in the full Earth system response. When making estimates of palaeoclimate sensitivity, agreement is therefore needed about which of the slower feedback processes are viewed as feedbacks (implicitly accounted for in S), and which are best considered as radiative forcings (explicitly accounted for in R).

We use a convenient operational distinction[33,48] in which a process is considered as a radiative forcing if its radiative influence is not changing with temperature on the timescale considered, and as a feedback if its impact on the radiation balance is affected by temperature changes on that timescale. For example, the radiative impacts of GHG changes over the last 800,000 years may be derived from concentration measurements of CO_2, CH_4, and N_2O in ice cores[49-51], and the radiative impacts of land-ice albedo changes may be calculated from continental ice-sheet estimates, mainly based on sea-level records[52-54]. Thus, the impacts of the slow biogeochemical and land-ice albedo feedbacks can be explicitly accounted for before climate sensitivity is calculated. In other words, these slow feedbacks are effectively considered as forcings, leaving only fast feedbacks to be considered implicitly in the calculated climate sensitivity, which thus approximates the (equilibrium) 'fast-feedback' sensitivity concept from modelling studies [e.g.7,41,55].

Operational challenges

All palaeoclimate sensitivity studies are affected by limitations of data availability. Below we discuss such limitations to reconstructions of forcings and feedbacks, and of global surface temperature responses. First, however, we highlight a critical caveat, namely that the climate response depends to some degree on the type of forcing (e.g., shortwave versus longwave, surface versus top-of-atmosphere, and local versus global). In other words, various radiative forcings with similar absolute magnitudes have different spatial distributions and physics. Consequently, the concept of global mean radiative forcing is a simplification that adds a (difficult to quantify) level of uncertainty to the results.

Astronomical forcing is a key driver of climate change. In global annual mean calculations of radiative change, astronomical forcing is very small and often ignored[41,55]. However, this obscures its importance, related to seasonal changes in the spatial distribution of insolation over the planet[44,45,56,57]. Global numerical models explicitly account for the latitudinal and temporal intricacies of astronomical forcing, but sensitivity has not yet been reported from highly resolved models with an interactive carbon cycle and ice sheets, which were integrated long enough to explore the influences of slowly evolving astronomical changes (timescales of 10^4 years and longer). Analytical studies suffer from uncertainty about which aspect(s) of astronomical forcing are most critical; spatial variability in temperature due to obliquity changes may be readily accounted for[41], but seasonal aspects may also be important. There also remains discussion about the relative importances of instantaneous insolation changes[57] and integrated summer-insolation energy[58] at specific latitudes. Given these complexities, we propose that the contribution of the astronomical forcing to ΔR may be neglected or included using its annual mean variation proportional to $(1/(1-e^2)^{0.5})$, where e is the eccentricity factor, which in the Quaternary accounts for up to about 0.4 Wm^{-2} at the Earth surface[45]. When other components of the system respond to the seasonal aspects of forcing, such as Quaternary ice-sheet variations, these may be accounted for as forcings themselves.

Next regarding forcings and feedbacks, GHG concentrations from ice cores are not available for times prior to 800 thousand years ago (ka), when CO_2 levels instead have to be estimated from indirect methods. These so-called 'proxy data' are based on pCO_2-dependent physico-chemical or biological processes, such as the abundance of stomata on fossil leaves[59], fractionation of stable carbon isotopes by marine phytoplankton[60], boron speciation and isotopic fractionation in sea water as a function of pH and preserved in biogenic calcite[61], and the stability fields of minerals precipitated from waters in contact with the atmosphere[62]. Considerable uncertainties are involved in these reconstructions of GHG concentrations, although progress is being made in improving the methods, their temporal coverage, and mutual consistency[63]. Although recent work has aimed to determine a synthesis high-resolution CO_2 record based on available data for the last 20 million years[Ref.64], there remains an urgent need for new data and updated syntheses, particularly for warmer climate states. Also, proxies are needed for reconstruction of CH_4 and N_2O concentrations in periods pre-dating the ice-core records.

Another example regarding the reconstruction of forcings/feedbacks concerns the assessment of land-ice albedo changes. Although good methods exist for the generation of continuous centennial to millennial scale sea-level (ice-volume) records over the last 500,000 years[Refs.52-54], such detailed information remains scarce for older periods. One approach that addresses this deficiency is a model-based deconvolution of deep-sea stable oxygen isotope records[54], which has recently been extended to provide a first-order estimate of sea-level variability back to 35 Million years ago (Ma)[65]. However, this urgently requires independent validation from new sea-level data, especially to address uncertainties about the volume-to-area relationships that would be different for incipient ice sheets (with a large surface area relative to limited height/volume) than for mature ice sheets[66,67]. Before 35 Ma, there is thought to have been (virtually) no significant land-ice volume[68], but this does not exclude the potential existence of major semi-permanent snow/ice-fields[69,70], and there remain questions whether these would constitute 'fast' (snow) or 'slow' (land-ice) feedbacks. The contribution of sea ice to albedo feedback also remains uncertain, with little quantitative information on past sea-ice extent beyond the LGM.

Similar examples of uncertainties and limited data availability could be listed for all feedbacks. In spite of such limitations, a "deep-time" (pre-1 Ma) geological perspective must be maintained, because: (1) they offer the most reasonable natural equivalents, in terms of cumulative effect, to the current rate and magnitude of GHG emissions[71-72]; and (2) only ancient records offer insight into climate states globally warmer than the present. A critical caveat is the assumption that all temperature change may be attributed to the forcings considered. This is important, because overlooked/unknown forcings could also have important influences. Any palaeodata-derived value for S should therefore be accompanied by careful documentation of the considered forcings, and the potential for overlooked forcings must be a prime target in further investigations of any differences between (identically defined) values of S from different geological episodes.

For many reasons, no past perturbation will ever present a perfect analogue for the ongoing anthropogenic perturbation. It is more useful to consider past warm climate states as test-beds for evaluating processes, responses, and to challenge/validate model simulations of those past climate states. Such data-model comparisons will drive model skill and understanding of processes, improving confidence in future multi-century projections. This raises the need for an 'experimental design' in palaeo-studies that minimises the impacts of very long-term influences on temperature sensitivity to radiative forcing, for example due to changes in continental configuration, orography, biological evolution of vegetation, etc. This can be achieved by focussing on highly resolved documentation of specific perturbations, superimposed upon distinctly different long-term background climate states. An example is the pronounced transient global warming and carbon-cycle perturbation during the Paleocene-Eocene Thermal Maximum (PETM) anomaly[73-74], which punctuated an already warm climate state[75].

When deep-time case studies are developed, one further complication must be considered when calculating radiative perturbations. The radiative forcing due to a doubling of CO_2 concentrations is estimated to be about 3.7 Wm^{-2} when starting from pre-industrial concentrations, but at higher CO_2 levels, this value per CO_2 doubling becomes larger[e.g.14], which would imply a lower value for S than would be estimated using 3.7 Wm^{-2}. Data-led studies may help with a first-order documentation of this dependence, following an approach in which S is calculated from CO_2 and temperature measurements under the assumption of a constant 3.7 Wm^{-2} per CO_2 doubling. That would (knowingly) overestimate S for high-CO_2 episodes, and the difference with other, identically defined, S values for different climate background states may then be used to gauge the magnitude of any deviation from 3.7 Wm^{-2} per CO_2 doubling.

We now turn to issues regarding the reconstruction of past global surface temperature responses (i.e. ΔT in eq. 1 below), where again much remains to be improved. Most existing studies on palaeoclimate sensitivity (see Table 1) have used one or more of the following: polar temperature variations from Antarctic ice cores (since 800 ka) with a correction for 'polar amplification' (usually estimated at 1.5-2.0 [Refs.76,77]); deep-sea temperature variations from marine sediment-core data with a correction for the ratio between global surface temperature and deep-sea temperature changes (often estimated at 1.5); single-site sea surface temperature (SST) records from marine sediment-cores; or compilations of SST data of varying geographic coverage from marine sediment-cores[7,41,55,78-80]. So far, few studies have included terrestrial temperature proxy records other than those from ice cores[81], yet better control on land-surface data is crucial because of land-sea contrasts and seasonal contrasts. Overall, there is a lack of spatial coverage in records of temperature response as much as there is for the radiative changes. Continued development is needed of independently validated (multi-proxy) and spatially representative (global) datasets of high temporal resolution relative to the climate perturbations studied.

Uncertainties in individual reconstructions of temperature change may in exceptional cases be reported to ±0.5 K, but more comprehensive uncertainty assessments normally find them to be considerably larger[82,83]. Compilation of such records to represent changes in global mean surface temperature changes involves the propagation of further assumptions/uncertainties, for example due to interpolation from limited spatial coverage, so that the end-result is unlikely to be constrained within narrower limits than ±1°C even for well-studied intervals. Finally, intercomparison of independent reconstructions for the same episode reveals 'hidden' uncertainties that arise from differences between each study's methodological choices, uncertainty determination, and data-quality criteria, which are hard to quantify and often ignored or poorly elucidated. Take the LGM for example, which for temperature is among the best-studied intervals in geological history. The MARGO compilation[83] inferred a global SST reduction of –1.9 ± 1.8 K relative to the present. This was used in another study to infer a global mean surface air temperature anomaly of –3 +1.3/–0.7 K [Ref.81]. The latter contrasts with a previous estimate of –5.8 ± 1.4 K [Ref.84], which is consistent with tropical (30°S to 30°N) SST anomalies of –2.7 ± 1.4 K [Ref.85], but those in turn are contested. MARGO[83] for example suggested such cooling for the Atlantic tropics, but less for the Indian and Pacific tropics, giving a global tropical cooling of only –1.7 ± 1.0 K. Clearly, even a well-studied interval gives rise to a range of temperature estimates, which translates to a broad range of climate sensitivity estimates.

It is evident that progress in quantifying palaeoclimate sensitivity will not only rely on a common concept and terminology that allows like-for-like comparisons (see below). It will also rely on an objective, transparent, and hence reproducible discussion in each study of the assumptions and uncertainties that affect the values for change in both temperature and radiative forcing.

Ways forward

Here we propose a new terminology to help palaeoclimate sensitivity studies adopt common concepts and approaches, and thus improve the potential for like-for-like comparisons between different studies. First we outline how our concept of 'equilibrium' S for palaeo-studies relates to 'equilibrium' S for modern studies. Then, we present a notation system that is primarily of value to palaeodata-based studies to clarify which slow feedbacks are explicitly accounted for.

When the ΔT response to an applied GHG radiative forcing ΔR is small relative to 'pre-perturbation' reference temperature $\bar{T}$, then 'equilibrium' climate sensitivity S^a (where a indicates *actuo*) is given by (see details in *Supplementary Material*):

$$S^a = \frac{\Delta T}{\Delta R} = \frac{-1}{\lambda_P + \sum_{i=1}^{N} \lambda_i^f} \qquad (1)$$

Here λ_P is the Planck feedback parameter (–3.2 $Wm^{-2}K^{-1}$) and λ_i^f (in $Wm^{-2}K^{-1}$) represents the feedback parameters of any number (*N*) of fast (*f*) feedbacks; i.e., those acting faster than timescale τ (see footnote§).

S^a is the 'Charney' sensitivity calculated by most climate models in 'double-CO_2' equilibrium simulations. As stated earlier, the range reported in the IPCC AR4 for S^a is 0.6-1.2 $K(Wm^{-2})^{-1}$. However, the Earth system in reality responds to a perturbation according to an equilibrium climate sensitivity parameter S^p (where *p* indicates *palaeo*), but the timescales to reach this equilibrium are very long, so that the forcing normally changes before equilibrium is reached. When attempting to determine S^a from palaeostudies, slow processes therefore need to be considered. To obtain S^a from past changes in CO_2 and T, i.e. from the palaeoclimate sensitivity S^p (again ΔT due to an applied GHG radiative forcing ΔR), a correction is needed for the slow feedback influences. Again under the small ΔT assumption, and using λ_j^s to represent any number (*M*) of slow feedbacks, this leads to the general expression (see *Supplementary Information*):

$$S^a = S^p \left(1 + \frac{\sum_{j=1}^{M} \lambda_j^s}{\lambda_P + \sum_{i=1}^{N} \lambda_i^f}\right) \qquad (2)$$

A recent study[47] defined the term 'Earth system sensitivity' (*ESS*) to represent the long-term climate response of Earth's climate system to a given CO_2 forcing, including both fast and slow processes. In our notation, $ESS = \Delta R_{2\times CO2}\, S^p$, where $\Delta R_{2\times CO2}$ is the forcing due to a CO_2-doubling (3.7 Wm^{-2}).

Here we introduce a more explicit notation regarding what was (not) included in the climate sensitivity diagnosis. It is the 'specific climate sensitivity' $S_{[A,B\ldots]}$, expressed in $K(Wm^{-2})^{-1}$), where slow feedback processes A, B, etc., are explicitly accounted for (see *Supplementary Information*). In other words, processes A, B, etc., are included in the forcing term, ΔR, rather than implicitly within *S*. This requires from the outset that a comprehensive view is taken of the various causes of change in the radiative balance. Table 2 summarises the various common permutations of *S* that may be encountered in palaeostudies, using 'LI' for albedo changes due to land-ice volume/area changes, 'VG' for vegetation-albedo feedback, 'AE' for aerosol feedback, and 'CO2' for carbon cycle feedbacks (see also case studies in Table 1).

The most practical version of *S* to be estimated from palaeodata is $S_{[CO2,LI]}$, because $S_{[CO2,LI]} = S_{[CO2]}$ during times (pre-35 Ma) without ice volume, and because the global vegetation cover changes that underlie the vegetation-albedo feedback, the atmospheric dust fluctuations that underlie the aerosol feedback, and both CH_4 and N_2O fluctuations generally remain poorly constrained by proxy data. Common reporting of $S_{[CO2,LI]}$ would bring results closer in line with the model-based concept of 'equilibrium' fast-feedback sensitivity, which ignores slow feedback processes and implicitly resolves fast feedbacks. The aforementioned issues with aerosol influences

§ The feedback parameters are defined here in the form $\lambda = \Delta R/\Delta T$, see *Supplementary Information*.

mean that it would currently be best (where possible) for estimates from palaeodata to present a range of climate sensitivity values based on both implicit and explicit consideration of the aerosol feedback (i.e., to report both $S_{[CO2,LI]}$ and $S_{[CO2,LI,AE]}$).

Table 2 shows example estimates for S following the various possible definitions. The first example uses records of palaeodata since 800 ka to unravel the strength of the various fast and slow feedbacks[7]. The second example lists estimates for $S_{[CO2]}$, $S_{[CO2,LI]}$, and $S_{[CO2,LI,VG]}$ from a more model-led exercise for the Middle Pliocene (~3 to 3.3 Ma)[Ref.16], with $\Delta T = 3.3$ K relative to the present and $\Delta R_{CO2} = 1.9$ Wm^{-2} due to CO_2 increase from 280 to 400 ppmv[Ref.47]. The calculations for both examples are detailed in *Supplementary Information*. In both cases, a broad range of S values is found, depending on which feedbacks are included (Table 2), which highlights the importance of reporting sensitivity estimates from different studies using a strict common definition. Comparison across different definitions unrealistically widens the range of values reported, notably towards the high end of the range because omission of 'forcing' due to the action of any slow feedbacks will cause overestimation of S (Figure 3).

For a first-order estimate of the range of S from palaeodata that approximates compatibility with the centennial-scale 'equilibrium' values of the IPCC[1], values need to be used that account for 'CO_2' or 'GHG' as well as 'LI', and preferably also 'AE' and/or 'VG' (Tables 1,2; Figure 3). This yields about 0.8 –0.4/+0.6 K(Wm^{-2})$^{-1}$ at 95% confidence limits (Figure 3). This includes uncertainties outlined in the source studies as well as any unaccounted-for dependence on different background climate states, but excludes potential additional uncertainties highlighted in this study. The long tail at the high end extends the total range to 2.2 K(Wm^{-2})$^{-1}$ based on data for the PETM (Table 1; Figure 3). Including the Earth System Sensitivity values, gauged from $S_{[CO2]}$, further extends the upper limit to more than 3 K(Wm^{-2})$^{-1}$ (Figure 3).

Outlook

We have demonstrated the need for standardisation in the approach for determining palaeoclimate S directly from data for ΔT and ΔR. However, we see this as one approach among several. A further approach optimally calibrates climate models to palaeodata fields, and then explores model sensitivities to perturbations[18,81]. Climate perturbations due to different types of forcing may then be studied using a diversity of geological time-slices, to understand the role of the climate background state. Yet another approach comprises hypothetical scenarios with global circulation models that are initialised for different background climate conditions, which will clarify the impacts of each radiative perturbation term in isolation and in different combinations. Only a combination of these diverse approaches will provide the richness of information and potential for intercomparison and independent validation that is needed for a fundamental understanding of how climate sensitivity changed through time, and why. Such robust foundations of process understanding will drive better projections of future climate change.

Improvements are clearly needed in palaeoclimate studies concerning the reconstructions of both forcings/feedbacks and temperature responses. Improved estimates of global temperature change require increased spatial densities of records for targeted time intervals. Uncertainties in temperature quantifications need to be challenged using multiple different proxies, with careful estimates of error propagation in both proxy representation (e.g., seasonal bias)[e.g.83] and when upscaling from regional records to global means[e.g.41]. A deep-time view that includes past warm climate states is relevant because of the future trajectory of climate. Improved quantification of past CO_2 levels remains essential, and there is an urgent need for methods to estimate past CH_4 and N_2O levels as well as atmospheric dust/aerosol concentrations. Similarly, detailed information remains essential for the other main processes, such as land-ice cover and vegetation changes.

Complications arise in comparisons between different long-term climate states that are widely separated in time, due to the impacts of long-term changes in the Earth System (including plate tectonics, evolution of biological systems, etc.) and solar evolution. We propose that climate sensitivity as a function of background climate state may be more successfully investigated using highly resolved documentation of shorter-term events superimposed upon different background climate states. We propose Late Palaeocene warming, the various transient Palaeocene-Eocene carbon cycle perturbations (including the PETM) that were superimposed on slightly different background states, the Middle Eocene Climatic Optimum, the late Oligocene warming event, the Middle Miocene Climatic Optimum, and the Pliocene warm period as excellent targets for focussed international efforts. For such episodes, the pre-perturbation background climate state needs to be characterised, as do the main radiative forcing/feedback changes through the perturbation and the (global) temperature response. For both the forcing/feedbacks and the temperature response, sufficient spatial coverage must be developed to obtain sensible global mean estimates as well as the spatial distributions. These reconstructions must be supported by comprehensive evaluation of uncertainties and their propagation into the end-results. Reporting of results should follow clear definitions, such as those proposed here, to allow like-for-like comparisons between studies.

References

1. Solomon, S. et al. (eds.) Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (Cambridge University Press, Cambridge, U.K., 2007).
2. Collins, M. & Long, D. Climate sensitivity – how sensitive is Earth's climate to CO_2; present. PAGES News 20, 10 (2012).
3. Knutti, R. & Hegerl, G.C. The equilibrium sensitivity of the earth's temperature to radiation changes. Nature Geoscience 1, 735–743 (2008).
4. Houghton, J.T. et al. (eds.) Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change (Cambridge University Press, Cambridge, U.K., 2001)
5. Roe, G.H. Feedbacks, timescales and seeing red. Annual Review of Earth and Planetary Sciences 37, 93–115 (2009).
6. Dufresne, J.-L. & Bony, S. An assessment of the primary sources of spread of global warming estimates from coupled atmosphere-ocean models. Journal of Climate 21, 5135–5144 (2008).
7. Köhler, P. et al. What caused Earth's temperature variations during the last 800,000 years? Data-based evidences on radiative forcing and constraints on climate sensitivity. Quaternary Science Reviews 29, 129–145 (2010).
8. Roe, G.H. & Baker, M.B. Why is climate sensitivity so unpredictable? Science 318, 629-632 (2007).
9. Baker, M.B. & Roe, G.H. The shape of things to come: why is climate change so predictable? Journal of Climate 22, 4574–4589 (2009).

10. Hannart, A., Dufresne, J.-L. & Naveau, P. Why climate sensitivity may not be so unpredictable. Geophysical Research Letters 36, L16707 (2009).
11. Zaliapin, I. & Ghil, M. Another look at climate sensitivity. Nonlinear Processes in Geophysics 17, 113–122 (2010).
12. Roe, G.H. & Baker, M.B. Comment on ”Another look at climate sensitivity” by Zaliapin and Ghil (2010). Nonlinear Processes in Geophysics 18, 125–127 (2011).
13. Zaliapin, I. & Ghil, M. Reply to Roe and Baker's comment on ”Another look at climate sensitivity” by Zaliapin and Ghil (2010). Nonlinear Processes in Geophysics 18, 129–131 (2011).
14. Colman, R. & McAvaney, B. Climate feedbacks under a very broad range of forcing. Geophysical Research Letters 36, L01702, doi:10.1029/2008GL036268 (2009).
15. Hargreaves, J.C., Abe-Ouchi, A. & Annan, J.D. Linking glacial and future climates through an ensemble of GCM simulations. Climate of the Past 3, 77–87 (2007).
16. Lunt, D.J. et al. On the causes of mid-Pliocene warmth and polar amplification. Earth and Planetary Science Letters 321-322, 128–138 (2012).
17. Haywood, A.M et al. Are there pre-Quaternary geological analogues for a future greenhouse warming? Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences 369, 933–956 (2011).
18. Edwards, T.L., Crucifix, M. & Harrison, S.P. Using the past to constrain the future: how the palaeorecord can improve estimates of global warming. Progress in Physical Geography, 31, 481–500 (2007).
19. Crucifix, M. Does the Last Glacial Maximum constrain climate sensitivity? Geophysical Research Letters 33, L18701, doi: 10.1029/2006GL027137 (2006).
20. Laîné, A., Kageyama, M., Braconnot, P. & Alkama, R. Impact of greenhouse gas concentration changes on the surface energetics in the IPSL-CM4 model: regional warming patterns, land/sea warming ratio, glacial/interglacial differences. Journal of Climate 22, 4621–4635 (2009).
21. CCSM-Paleo-CMIP: http://www.cesm.ucar.edu/events/ws.2011/Presentations/Paleo/bette.pdf
22. Joos, F. & Spahni, R. Rates of change in natural and anthropogenic radiative forcing over the past 20,000 years. Proceedings of the National Academy of Science 105, 1425–1430 (2008).
23. Köhler, P., Knorr, G., Buiron, D., Lourantou, A. & Chapellaz, J. Abrupt rise in atmospheric CO_2 at the onset of the Bølling/Allerød: in-situ ice core data versus true atmospheric signals. Climate of the Past 7, 473–486 (2011).
24. Hönisch, B. et al. The geological record of ocean acidification. Science 335, 1058–1063 (2012).
25. Charney, J.G. et al. Carbon dioxide and climate: a scientific assessment (National Academy of Sciences, Washington D.C., 1979).
26. Knutti, R. & Tomassini, L. Constraints on the transient climate response from observed global temperature and ocean heat uptake, Geophysical Research Letters 35, L09701, doi:10.1029/2007GL032904 (2008).
27. Gregory, J.M. & Forster, P.M. Transient climate response estimated from radiative forcing and observed temperature change. Journal of Geophysical Research 113, D23105, doi: 10.1029/2008JD010405 (2008).
28. Soden, B.J. & Held, I.M. An assessment of climate feedbacks in coupled ocean-atmosphere models. Journal of Climate 19, 3354–33360 (2006).
29. Huber, M., Mahlstein, I., Wild, M., Fasullo, J. & Knutti, R. Constraints on climate sensitivity from radiation patterns in climate models. Journal of Climate 24, 1034–1052 (2011).

30. Huybers, P. Compensation between model feedbacks and curtailment of climate sensitivity. Journal of Climate 23, 3009–3018 (2010).
31. Lemoine, D.M. Climate sensitivity distributions dependence on the possibility that models share biases. Journal of Climate 23, 4395–4415 (2010).
32. Hansen, J., Sato, M., Kharecha, P & von Schuckmann, K. Earth's energy imbalance and implications. Atmospheric Chemistry and Physics 11, 13421–13449 (2011).
33. Gregory, J.M. et al. A new method for diagnosing radiative forcing and climate sensitivity. Geophysical Research Letters 31, L03205, doi:10.1029/2003GL 018747 (2004).
34. Lambert, F. et al. Dust-climate couplings over the past 800,000 years from the EPICA Dome C ice core. Nature 425, 616–619 (2008).
35. Winckler, G., Anderson, R.F., Fleisher, M.Q., McGee, D. & Mahowald, N. Covariant glacial-interglacial dust fluxes in the equatorial Pacific and Antarctica. Science 320, 93–96.
36. Roberts, A.P., Rohling, E.J., Grant, K.M., Larrasoaña, J.C. & Liu, Q. Atmospheric dust variability from major global source regions over the last 500,000 years. Quaternary Science Reviews 30, 3537-3541 (2011).
37. Ruth, U., Wagenbach, D., Steffensen, J. P. & Bigler, M. Continuous record of microparticle concentration and size distribution in the central Greenland NGRIP ice core during the last glacial period. Journal of Geophysical Research 108, 4098, doi: 10,1029/2002JD002376 (2003).
38. Naafs, B.D.A. et al. Strengthening of North American dust sources during the late Pliocene (2.7 Ma). Earth and Planetary Science Letters 317-318, 8–19 (2012).
39. Kohfeld, K.E. & Harrison, S.P. DIRTMAP: the geological record of dust. Earth Science Reviews 54, 81–114 (2001).
40. Mahowald, N., Albani, S., Engelstaedter, S., Winckler, G. & Goman, M. Model insight into glacial-interglacial paleodust records. Quaternary Science Reviews 30, 832–854 (2011).
41. Rohling, E.J., Medina-Elizalde, M., Shepherd, J.G., Siddall, M. & Stanford, J.D. Sea surface and high-latitude temperature sensitivity to radiative forcing of climate over several glacial cycles. Journal of Climate 25, 1635–1656 (2012).
42. Gough, D.O. Solar interior structure and luminosity variations. Solar Physics 74, 21–34 (1981)
43. Gray, L.J. et al. Solar influences on climate. Reviews of Geophysics 48, RG4001, doi: 10.1029/2009RG000282 (2010).
44. Milankovitch, M. Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem, Special Publications 133, Section Mathematics and Natural Sciences (Royal Serbian Academy, Belgrade), 633pp (1941).
45. Berger A. Support for the astronomical theory of climatic change. Nature 269, 44–45 (1977).
46. Laskar, J. et al. A long-term numerical solution for the insolation quantities of the Earth. Astronomy and Astrophysics 428, 261–285 (2004).
47. Lunt, D.J. et al. Earth system sensitivity inferred from Pliocene modelling and data. Nature Geoscience 3, 60–64 (2010).
48. Gregory, J. & Webb, M. Tropospheric adjustment induces a cloud component in CO2 forcing. Journal of Climate 21, 58–71 (2008).
49. Lüthi, D. et al. High-resolution CO_2 concentration record 650,000−800,000 years before present. Nature 453, 379–382 (2008).
50. Loulergue, L. et al. Orbital and millennial-scale features of atmospheric CH_4 over the past 800,000 years. Nature 453, 383–386 (2008).

51. Schilt, A. et al. Glacial-interglacial and millennial-scale variations in the atmospheric nitrous oxide concentration during the last 800,000 years. Quaternary Science Reviews 29, 182–192 (2010).
52. Waelbroeck, C. et al. Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. Quaternary Science Reviews 21, 295–305 (2002).
53. Rohling, E. J. et al. Antarctic temperature and global sea level closely coupled over the past five glacial cycles. Nature Geoscience 2, 500–504 (2009).
54. Bintanja, R., van de Wal, R. & Oerlemans, J. Modelled atmospheric temperatures and global sea levels over the past million years. Nature 437, 125–128 (2005).
55. Hansen, J. et al. Target atmospheric CO_2: Where should humanity aim? The Open Atmospheric Science Journal 2, 217–231, doi: 10.2174/1874282300802010217 (2008).
56. Imbrie, J. & Imbrie, J.Z. Modeling the climatic response to orbital variations. Science 207, 943–953 (1980).
57. Huybers, P. & Denton, G.H. Antarctic temperature at orbital timescales controlled by local summer duration. Nature Geoscience 1, 787–792 (2009).
58. Huybers, P. Early Pleistocene glacial cycles and the integrated summer insolation forcing. Science 313, 508–511 (2006).
59. Beerling, D.J. & Royer, D.L. Fossil plants as indicators of the phanerozoic global carbon cycle. Annual Review of Earth and Planetary Sciences 30, 527–556 (2002).
60. Pagani, M., Zachos, J.C., Freeman, K.H., Tipple, B. & Bohaty, S. Marked decline in atmospheric carbon dioxide concentrations during the paleogene. Science 309, 600–603 (2005).
61. Hönisch, B., Hemming, N.G., Archer, D., Siddall, M. & McManus, J.F. Atmospheric carbon dioxide concentration across the mid-Pleistocene transition. Science 324, 1551–1554 (2009).
62. Lowenstein, T.K. & Demicco, R.V. Elevated Eocene atmospheric CO_2 and its subsequent decline. Science 313, p.1928 (2006).
63. Beerling, D.J. & Royer D.L. Convergent Cenozoic CO_2 history. Nature Geoscience 4, 418–420 (2011).
64. van de Wal, R.S.W., de Boer, B., Lourens, L.J., Köhler, P. & Bintanja, R. Reconstruction of a continuous high-resolution CO_2 record over the past 20 million years. Climate of the Past 7, 1459–1469 (2011).
65. de Boer, B., van de Wal, R.S.W., Lourens, L.J. & Bintanja, R. Transient nature of the Earth's climate and the implications for the interpretation of benthic records. Palaeogeography, Palaeoclimatology, Palaeoecology, in press, doi: 10.1016/ j.palaeo.2011.02.001 (2011).
66. Cramer, B.S., Miller, K.G., Barrett, P.J. & Wright, J.D. Late Cretaceous-Neogene trends in deep ocean temperature and continental icevolume: Reconciling records of benthic foraminiferal geochemistry ($\delta^{18}O$ and Mg/Ca) with sea level history. Journal of Geophysical Research 116, C12023 (2011).
67. Gasson, E. et al. Exploring uncertainties in the relationship between temperature, ice volume and sea level over the past 50 million years. Reviews of Geophysics 50, RG1005, doi: 10.1029/2011RG000358 (2012).
68. Zachos, J., Pagani, M., Sloan, L., Thomas, E. & Billups, K. Trends, rhythms, and aberrations in global climate 65 Ma to Present. Science 292, 686–693 (2001).
69. Miller, K.G., Wright, J.D. & Browning, J.V. Visions of ice sheets in a greenhouse world. Marine Geology 217, 215–231 (2005).
70. Sluijs, A. et al. Eustatic variations during the Paleocene-Eocene greenhouse world. Paleoceanography 23, PA4216, doi: 10.1029/2008PA001615 (2008).

71. Dickens, G.R., Castillo, M.M. & Walker, J.C.G. A blast of gas in the latest Paleocene: Simulating first-order effects of massive dissociation of oceanic methane hydrate. Geology 25, 259-262 (1997).
72. Lourens, L.J. et al. Astronomical pacing of late Palaeocene to early Eocene global warming events. Nature 435, 1083–1087 (2005).
73. Zachos, J.C., Dickens, G.R. & Zeebe, R.E. An early Cenozoic perspective on greenhouse warming and carbon-cycle dynamics. Nature 451, 279–283 (2008).
74. Zeebe, R.E., Zachos, J.C. & Dickens, G.R. Carbon dioxide forcing alone insufficient to explain Palaeocene-Eocene Thermal Maximum warming. Nature Geoscience 2, 576–580 (2009).
75. Huber, M. & Caballero, R. The early Eocene equable climate problem revisited. Climate of the Past 7, 603–633 (2011).
76. Lorius, C., Jouzel, J., Raynaud, D., Hansen, J. & Le Treut, H. The ice-core record: climate sensitivity and future greenhouse warming. Nature 347, 139–145 (1990).
77. Masson-Delmotte, V. et al. Past and future polar amplification of climate change: Climate model intercomparisons and ice-core constraints. Climate Dynamics 26, 513–529 (2006).
78. Lea, D. The 100000-yr cycle in tropical SST, greenhouse gas forcing, and climate sensitivity. Journal of Climate 17, 2170–2179 (2004).
79. Hansen, J. et al. Climate change and trace gases. Philosophical Transactions of the Royal Society London A 365, 1925–1954 (2007).
80. Bijl, P.K. et al. Transient Middle Eocene atmospheric CO_2 and temperature variations. Science 330, 819–821 (2010).
81. Schmittner, A. et al. Climate sensitivity estimated from temperature reconstructions of the Last Glacial Maximum. Science 334, 1385–1388 (2011).
82. Rohling, E.J. Progress in palaeosalinity: overview and presentation of a new approach. Paleoceanography 22, PA3215, doi:10.1029/2007PA001437 (2007).
83. MARGO project members. Constraints on the magnitude and patterns of ocean cooling at the Last Glacial Maximum. Nature Geoscience 2, 127–132 (2009).
84. Schneider von Deimling, T., Ganopolski, A., Held, H. & Rahmstorf, S. How cold was the Last Glacial Maximum? Geophysical Research Letters 33, L14709, doi: 10.1029/2006GL026484 (2006).
85. Ballantyne, A.P., Lavine, M., Crowley, T.J., Liu, J. & Baker, P.B. Meta-analysis of tropical surface temperatures during the Last Glacial Maximum. Geophysical Research Letters 32, L05712, doi: 10.1029/2004GL021217 (2005).
86. Hoffert, M.I. & Covey, C. Deriving global climate sensitivity from palaeoclimate reconstructions. Nature 360, 573–576 (1992).
87. Hansen, J.E. & Sato, M. Paleoclimate implications for human-made climate change. In: Berger, A., Mesinger, F. & Šijački, D. (eds.) *Climate change at the eve of the second decade of the century: inferences from paleoclimate and regional aspects: Proceedings of Milutin Milankovitch 130th Anniversary Symposium*, in press (2011).
88. Pagani, M., Liu, Z., LaRiviere, J. & Ravelo, A.C. High Earth-system climate sensitivity determined from Pliocene carbon dioxide concentrations. Nature Geoscience 3, 27–30 (2009).
89. Covey, C., Sloan, L.C., & Hoffert, M.I. Paleoclimate data constraints on climate sensitivity: the paleocalibration method. Climatic Change 32, 165–184 (1996).
90. Zachos, J.C., Stott, L.D. & Lohmann, K.C. Evolution of early Cenozoic marine temperatures. Paleoceanography 9, 353–387 (1994).
91. Sloan, L.C. & Barron, E.J. A comparison of Eocene climate model results to quantified paleoclimatic interpretations. Palaeogeography, Palaeoclimatology, Palaeoecology 93, 183–202 (1992).

92. Berner, R.A. A model for atmospheric CO_2 over Phanerozoic time. Am J Sci 291, 339-376 (1991).
93. Freeman, K.H. & Hayes, J.M. Fractionation of carbon isotopes by phytoplankton and estimates of ancient CO_2 levels. Global Biogeochemical Cycles 6, 185–198 (1992).
94. Cerling, T.E. Carbon dioxide in the atmosphere; evidence from Cenozoic and Mesozoic Paleosols. Am J Sci 291, 377-400 (1991).
95. Royer, D.L., Pagani, M., & Beerling, D.J. Geobiological constraints on Earth system sensitivity to CO_2 during the Cretaceous and Cenozoic. Geobiology, doi: 10.1111/j.1472-4669.2012.00320.x (2012).
96. Panchuk, K., Ridgwell, A., & Kump, L.R. Sedimentary response to Paleocene-Eocene Thermal Maximum carbon release: A model-data comparison. Geology 36, 315–318 (2008).
97. Borzenkova, I.I. Determination of global climate sensitivity to the gas composition of the atmosphere from paleoclimatic data. Izvestiya, Atmospheric and Oceanic Physics 39, 197–202 (2003).
98. Park, J. & Royer, D.L. Geologic constraints on the glacial amplification of Phanerozoic climate sensitivity. American Journal of Science 311, 1–26 (2011).
99. Schmidt, G.A. Climate sensitivity – how sensitive is Earth's climate to CO_2; past. PAGES News 20, 11 (2012).
100. Wunsch, C. & Heimbach, P. How long to oceanic tracer and proxy equilibrium? Quaternary Science reviews 27, 637–651 (2008).

Acknowledgements
This review is the outcome of the first 'PALAEOSENS' workshop, which took place in March 2011 in Amsterdam. We thank the Royal Netherlands Academy of Arts and Sciences (KNAW) for funding and hosting this workshop, PAGES for their support, and Jonathan Gregory for discussions. This study contributes to UK-NERC consortium NE/I009906/1 and a Royal Society Wolfson Research Merit Award (EJR). AS thanks the European Research Council for ERC starting grant # 259627. Some of the work was supported by grant 243908 'Past4Future' of the EU's seventh framework programme; this is Past4Future contribution no.X.

Author contributions
EJR, AS, and HAD initiated the PALAEOSENS workshop, and led the drafting of this study together with PK, ASvdH, and RvdW. The other authors contributed specialist insights, discussions and feedback.

Additional Information
Supplementary Information accompanies this paper on www.xxxxxxxxx. Reprints and permissions information is available online at www.xxxxxxxx. Correspondence and requests for materials should be addressed to EJR.

PALAEOSENS Project Members: E.J. Rohling*, A. Sluijs, H.A. Dijkstra, P. Köhler, R.S.W. van de Wal, A.S. von der Heydt
with D. Beerling, A. Berger, P.K. Bijl, M. Crucifix, R. deConto, S.S. Drijfhout, A. Fedorov, G. Foster, A. Ganopolski, J. Hansen, B. Hönisch, H. Hooghiemstra, M. Huber, P. Huybers, R. Knutti, D.W. Lea, L.J. Lourens, D. Lunt, V. Masson-Demotte, M. Medina-Elizalde, B. Otto-Bliesner, M. Pagani, H. Pälike, H. Renssen, D.L. Royer, M. Siddall, P. Valdes, J.C. Zachos, R.E. Zeebe.

Rohling: School of Ocean and Earth Science, University of Southampton, National Oceanography Centre, Southampton, SO14 3ZH, UK e.rohling@noc.soton.ac.uk
Sluijs: Department of Earth Sciences, Faculty of Geosciences, Utrecht University, Budapestlaan 4, 3584 CD Utrecht, the Netherlands A.Sluijs@uu.nl
Dijkstra: Institute of Marine and Atmospheric Research Utrecht, Utrecht University, NL-3584 CC Utrecht, Netherlands H.A.Dijkstra@uu.nl
Köhler: Alfred Wegener Institute for Polar and Marine Research (AWI), PO Box 12 01 61, D-27515 Bremerhaven, Germany Peter.Koehler@awi.de
Van de Wal: Institute of Marine and Atmospheric Research Utrecht, Utrecht University, NL-3584 CC Utrecht, Netherlands R.S.W.vandeWal@uu.nl
Von der Heydt: Institute for Marine and Atmospheric Sciences, Utrecht University, Princetonplein 5, 3584CC Utrecht, The Netherlands A.S.vonderHeydt@uu.nl
Beerling: Department of Animal and Plant Sciences, University of Sheffield S10 2TN, UK d.j.beerling@sheffield.ac.uk
Berger: Georges Lemaitre Centre for Earth and Climate Research, Earth and Life Institute -- Université catholique de Louvain, Chemin du Cyclotron 2, box L7.01.11, 1348 Louvain-la-Neuve, Belgium andre.berger@uclouvain.be
Bijl: Department of Earth Sciences, Faculty of Geosciences, Utrecht University, Budapestlaan 4, 3584 CD Utrecht, the Netherlands P.K.Bijl@uu.nl
Boulder, CO, USA ottobli@ucar.edu
Crucifix: Georges Lemaitre Centre for Earth and Climate Research, Earth and Life Institute -- Université catholique de Louvain, Chemin du Cyclotron 2, box L7.01.11, 1348 Louvain-la-Neuve, Belgium michel.crucifix@uclouvain.be
deConto: Department of Geosciences, 611 North Pleasant St., 233 Morrill Science Center, University of Massachusetts, Amherst MA 01003-9297, USA deconto@geo.umass.edu
Drijfhout: Royal Netherlands Meteorological Institute, PO Box 201, 3730 AE De Bilt, the Netherlands drijfhou@knmi.nl
Fedorov: Department of Geology and Geophysics, Yale University, P.O. Box 208109, New Haven, CT 06520-8109, USA alexey.fedorov@yale.edu
Foster: School of Ocean and Earth Science, University of Southampton, National Oceanography Centre, Southampton, SO14 3ZH, UK gavin.foster@noc.soton.ac.uk
Ganopolski: PIK, PO Box 601203, 14412 Potsdam, Germany andrey@pik-potsdam.de
Hansen: NASA Goddard Institute for Space Studies, 2880 Broadway, New York, NY 10025, USA James.E.Hansen@nasa.gov
Hönisch: Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY, USA hoenisch@ldeo.columbia.edu
Hooghiemstra: Institute for Biodiversity and Ecosystem Dynamics, University of Amsterdam, Science Park 904, 1098 XH Amsterdam, Netherlands H.Hooghiemstra@uva.nl
Huber: Earth and Atmospheric Sciences Department, Purdue University, West Lafayette, Indiana, USA huberm@purdue.edu
Huybers: Department of Earth and Planetary Sciences, Harvard University, 20 Oxford St. Cambridge MA 02138, USA phuybers@fas.harvard.edu
Knutti: Institute for Atmospheric and Climate Science, ETH Zurich, Universitätstrasse 16, CH-8092 Zurich, Switzerland reto.knutti@env.ethz.ch
Lea: Department of Earth Science, University of California, Santa Barbara, CA 93106-9630, USA lea@geol.ucsb.edu
Lourens: Department of Earth Sciences, Faculty of Geosciences, Utrecht University, Budapestlaan 4, 3584 CD Utrecht, the Netherlands L.J.Lourens@uu.nl
Lunt: School of Geographical Sciences, University of Bristol University Road, Bristol, BS8 1SS, UK d.j.lunt@bristol.ac.uk
Masson-Delmotte: LSCE (IPSL/CEA-CNRS-UVSQ), UMR 8212), LCEA Saclay, 91 191 Gif sur Yvette cédex, France valerie.masson@lsce.ipsl.fr
Medina-Elizalde: Centro de Investigación Científica de Yucatán, Unidad Ciencias del Agua, Cancún, México martin.medina@cicy.mx
Otto-Bliesner: National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO, USA ottobli@ucar.edu
Pagani: Department of Geology and Geophysics, Yale University, P.O. Box 208109, New Haven, CT 06520-8109, USA mark.pagani@yale.edu
Pälike: School of Ocean and Earth Science, University of Southampton, National Oceanography Centre, Southampton, SO14 3ZH, UK heiko@noc.soton.ac.uk
Renssen: Dept. of Earth Sciences, Faculty of Earth & Life Sciences, VU University Amsterdam, De Boelelaan 1085, NL1081HV Amsterdam, The Netherlands h.renssen@vu.nl

Royer: Department of Earth and Environmental Sciences, Wesleyan University, Middletown, CT 06459, USA droyer@wesleyan.edu
Siddall: Department of Earth Sciences, University of Bristol, Wills Memorial Building, Queen's Road, Bristol BS8 1RJ, UK mark.siddall@bristol.ac.uk
Valdes: School of Geographical Sciences, University of Bristol, University Road, Bristol BS8 1SS, UK P.J.Valdes@bristol.ac.uk
Zachos: Earth & Planetary Sciences, Univ. of Calif., Santa Cruz CA, USA jzachos@ucsc.edu
Zeebe: School of Ocean and Earth Science and Technology, Department of Oceanography, University of Hawaii at Manoa, 1000 Pope Road, MSB 629 Honolulu, HI 96822, USA zeebe@hawaii.edu

*communicating author. e-mail: e.rohling@noc.soton.ac.uk

Source and ***Number for Fig.3***	**Time window**	**Explicitly considered forcings**	**Temperature data used**	***S*** **(K/(Wm^{-2})$^{-1}$)**	**Notes**
Ref.76 ***1***	GC	GHG (CO_2, CH_4) LI AE	ΔT_{aa} *(using 50% polar amplification)*	0.75 to 1.00	
Ref.86 ***2***	LGM	GHG (CO_2, CH_4) LI AE VG	CLIMAP and $\Delta T_{aa\ \&\ gld}$	0.80 ± 0.14	Value after authors' suggested correction of CLIMAP temperatures
Ref.78 ***3***	GC	GHG (CO_2, CH_4)	ΔT_{trop}	1.1 ± 0.1	Author's linear regression case
Ref.55 ***4***	GC	GHG (CO_2, CH_4, N_2O) LI	ΔT_{aa} *(using 2x polar amplification)*	0.75 ± 0.25	
Ref.55 ***5***	GC	GHG (CO_2, CH_4, N_2O)	ΔT_{aa} *(using 2x polar amplification)*	1.5 ± 0.5	
Ref.7 ***6***	LGM	GHG (CO_2, CH_4, N_2O) LI AE VG insolation	ΔT_{global} = –5.8 ± 1.4 K; GLAMAP extrapolated with model (Ref.84)	0.72 + 0.33 – 0.23	Includes a scaling factor (0.85) for smaller *S* during LGM compared to 2 x CO_2 based on GCM model-output (Refs.15,19)
This paper, based on Ref.7 ***7***	GC (< 800 ka)	GHG (CO_2, CH_4, N_2O) LI AE VG insolation	ΔT_{NH} = model-based deconvolution of benthic $\delta^{18}O$ (Ref.54), scaled to global ΔT using a NH polar amplification on land of 2.75 ± 0.25	0.68 to 2.32	This covers the range of $S_{[GHG,X]}$ given in Table S2. For details see *Supplementary Information*
Ref.87 ***8***	GC	GHG (CO_2, CH_4, N_2O) LI	ΔT_{aa} *(using 2x polar amplification)* and 1.5x ΔT_{ds}	0.75 ± 0.13	
Ref.41 ***9***	GC	GHG (CO_2, CH_4, N_2O) LI AE insolation	36-record global synthesis of sea surface temperature changes along with $\Delta T_{aa\ \&\ gld}$. Polar amplification diagnosed, not imposed.	0.85 +0.5 –0.4	Total range uncertainties
Ref.41 ***10***	GC	GHG (CO_2, CH_4, N_2O) LI insolation	36-record global synthesis of sea surface temperature changes along with $\Delta T_{aa\ \&\ gld}$. Polar amplification diagnosed, not imposed.	1.05 ± 0.5	Total range uncertainties
Ref.88 ***11***	Early to Middle Pliocene (4.2-3.3 Ma)	CO_2. Earth System Sensitivity (sensu Ref.47)	Using model-based ΔT for Middle and Early Pliocene of 2.4-2.9˚C and 4˚C. ΔCO_2 alkenone. (sources in Ref.88)	(3.3 Ma) 1.92 ± 0.27 to 2.35 ± 0.35 (4.2 Ma) 2.60 ±0.38	
Ref.64 ***12***	Miocene optimum to Present-day	Slow feedbacks	Deconvolution of benthic $\delta^{18}O$ (Ref.65)	0.78 ± 10%	f=0.71, β=5.35, γ=1.3 details in *Supplementary Information*
This paper (compilation) ***13***	Eocene-Oligocene Transition (~34 Ma)	CO_2. Earth System Sensitivity (sensu Ref.47)	Model-based ΔT, with range of CO_2 values.	1.72 +1.79 –1.07	Details in *Supplementary Information.*
This paper (compilation) ***14***	Late Eocene vs. Present	CO_2. Earth System Sensitivity (sensu Ref.47)	Model-based ΔT, with range of CO_2 values.	1.82 +0.53 –0.97	Details in *Supplementary Information.*
Ref.80 ***15***	Middle Eocene Climatic Optimum (~40 Ma)	CO_2. Ice-free world. Event study (not affected by plate tectonics and evolution effects).	ΔT_{ds} (2 records) and ΔT_{mg} (7 records; subtropics to high lats.; no tropical data). ΔCO_2 alkenone	0.95 ± 0.6	500 kyr timescale. Biased to high-latitude sensitivity. ΔT_{ds} = ΔT_{mg}
Ref.80 ***16***	Mid to Late Eocene transition (41-35 Ma)	CO_2. Largely ice-free world. Event study (not affected by plate tectonics and evolution effects).	ΔT_{ds} (Ref.73) and ΔT_{mg}. ΔCO_2 = difference mid Eocene alkenone and late Eocene $\delta^{11}B$	0.95 ± 0.6	multi-million year timescale.
Ref.89 ***17***	Early Eocene (~55-50 Ma)	CO_2. Ice-free world. (Potential influences of plate tectonics and biological evolution not considered).	ΔT_{mg} (Refs.90,91). ΔCO_2 based on modelling (Ref.92) marine organic carbon isotope fractionation (Ref.93) and soil nodules (Ref.94)	0.65	Recalculated in Ref.95. NB. Ref.90 underestimated tropical SST.
This paper (compilation) ***18***	PETM (~56 Ma)	CO_2. Ice-free world. Event study (not affected by plate tectonics and evolution effects).	ΔT_{ds} (>6 records) and ΔT_{mg} (>11 records; equatorial to polar). ΔCO_2 based on deep ocean carbonate chemistry (Refs.74,96)	0.88 to 2.16	Details in *Supplementary Information.* Assumes all warming due to C input, and spread in *S* represents various background CO_2 and C-injection scenarios. ΔT_{ds} = ΔT_{mg}

Ref.97 ***19***	Cretaceous and early Palaeogene	CO_2. Largely ice-free world. (Potential influences of plate tectonics and biological evolution not considered).		1	Recalculated in Ref.95.
Ref.95 ***20***	Cretaceous and early Palaeogene	CO_2. Largely ice-free world. Earth System Sensitivity (sensu Ref.47)	ΔT after Refs.55,73. ΔCO_2 based on Ref.63.	>0.8	
Ref.98 ***21***	Phanerozoic	CO_2. Ice-free situation. (Potential influences of plate tectonics and biological evolution not considered).	$\Delta T_{mg,}$ ΔCO_2 based on GEOCARBSULF	0.8 to 1.08	

Table 1. Summary of key studies that have empirically determined *S* for the Pleistocene and some deep time periods from comparison between data-derived timeseries for temperature and for radiative change. Comparison of results between studies is greatly hindered by the different 'versions' of *S* used, as related to different notions of which processes should be explicitly accounted for, and by the different approaches taken to approximate global mean surface temperature. GC = Glacial Cycles; LGM = Last Glacial Maximum; PETM = Palaeocene-Eocene Thermal Maximum. In the subscripts, aa = Antarctica; gld = Greenland; trop = tropical; ds = deep sea; global = global mean; mg = Mg/Ca. All values are reported as in the source study. When no uncertainties are listed, this does not mean that there is no uncertainty, but only that it was not specified. All values for *S* are reported in $K/(Wm^{-2})^{-1}$, where necessary after transformation using 3.7 Wm^{-2} per doubling of CO_2, bearing in mind the caveats for this at high CO_2 concentrations as elaborated in the main text.

Number for Fig.3	***S***	**Explicitly considered radiative perturbation**	**Period in which it is practical to use the definition**	**Value after Ref.7 for LGM ($K(Wm^{-2})^{-1}$)**	**Value after Refs.16,47 for Pliocene ($K(Wm^{-2})^{-1}$)**
22	$S_{[CO2]}$	$\Delta R_{[CO2]}$	All (esp. pre-35Ma when LI = ~0)	2.63 ± 0.57	1.2
23	$S_{[CO2,\ LI]}$	$\Delta R_{[CO2,\ LI]}$	<35 Ma	0.95 ± 0.26	0.97
24	$S_{[CO2,\ LI,\ VG]}$	$\Delta R_{[CO2,\ LI,\ VG]}$	<35 Ma	0.8 ± 0.25	0.82
25	$S_{[CO2,\ LI,\ AE]}$	$\Delta R_{[CO2,\ LI,\ AE]}$	<35 Ma but mainly <800 ka	0.72 ± 0.24	
26	$S_{[CO2,\ LI,\ AE,\ VG]}$	$\Delta R_{[CO2,\ LI,\ AE,\ VG]}$	<35 Ma but mainly <800 ka	0.63 ± 0.22	
27	$S_{[GHG]}$	$\Delta R_{[GHG]}$	<800 ka	1.97 ± 0.46	
28	$S_{[GHG,\ LI]}$	$\Delta R_{[GHG,\ LI]}$	<800 ka	0.85 ± 0.23	
29	$S_{[GHG,\ LI,\ VG]}$	$\Delta R_{[GHG,\ LI,\ VG]}$	<800 ka	0.73 ± 0.23	
30	$S_{[GHG,\ LI,\ AE]}$	$\Delta R_{[GHG,\ LI,\ AE]}$	<800 ka	0.66 ± 0.22	
31	$S_{[GHG,\ LI,\ AE,\ VG]}$	$\Delta R_{[GHG,\ LI,\ AE,\ VG]}$	<800 ka	0.58 ± 0.20	

Table 2. Common permutations of *S* that may be encountered in palaeostudies. *S* is presented with a subscript that identifies the explicitly considered radiative perturbations (third column); all other processes are implicitly resolved as feedbacks within *S*. The period in which the various definitions of *S* are practical are determined by the availability of data for the explicitly considered processes. Subscript CO_2 indicates the radiative impact of atmospheric CO_2 concentration changes; LI represents the radiative impact of global Land Ice-volume changes; VG stands for the radiative impact of global vegetation cover changes; AE indicates the radiative impact of aerosol changes; GHG stands for the impact of changes in all non-water natural greenhouse gases (notably CO_2, CH_4, and N_2O). Column 4 gives calculated values for all suggested permutations of *S* for the LGM, based on a previous data compilation of ΔR (Ref.7). See *Supplementary Information* for details and error analysis of *S*. Column 5 gives examples for the Pliocene, with details in the *Supplementary Information* (Refs.16,47).

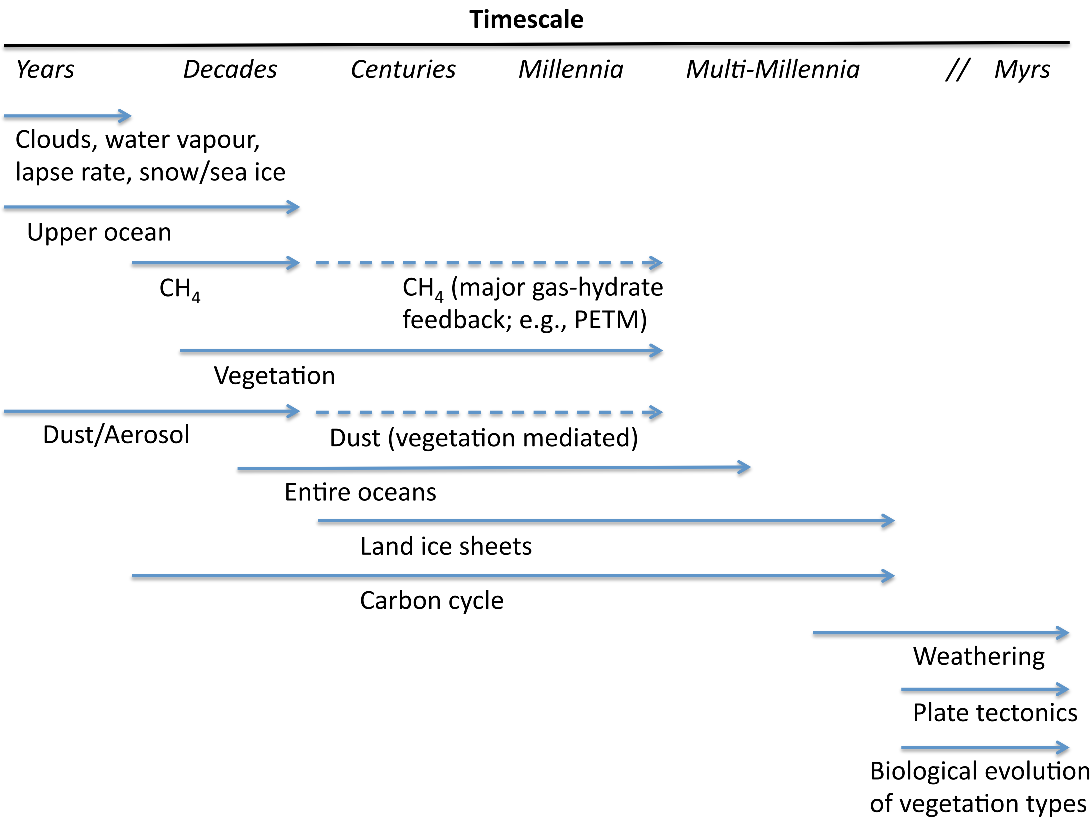

Figure 1. Typical timescales of different feedbacks relevant to equilibrium climate sensitivity, as discussed in this study. Modified and extended after Ref.99. Ocean timescales were extended to multi-millennial timescales, after Ref.100.

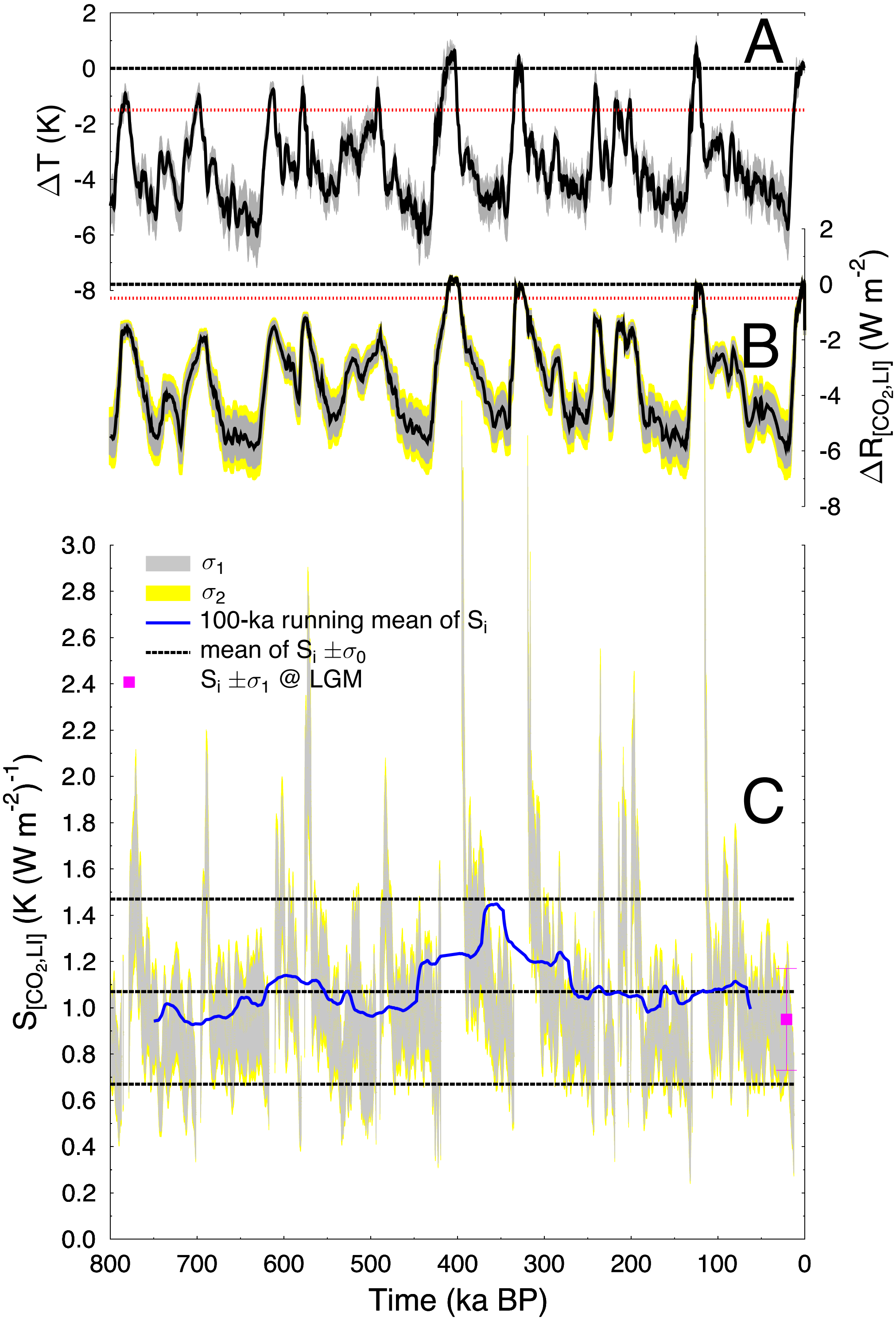

Figure 2. Illustration of variability of climate sensitivity using a calculation of $S_{[CO2,LI]}$, as defined in this study, for the last 800,000 years (for details, see *Supplementary Information*). (**A**) Changes in global temperature. (**B**) Changes in radiative forcing due to changes in CO_2 and surface albedo due to land ice. (**C**) Calculated $S_{[CO2,LI]}$, which is only considered robust and calculated when $\Delta T < -1.5$ K and $\Delta R_{[CO2,LI]} < -0.5$ Wm^{-2}, as indicated by the dotted red lines in (A) and (B). Mean of $S_i \pm \sigma_0$ and 100-kyr running mean are shown together with individual results for single points. Magenta marker denotes S $\pm\sigma_1$ for the LGM only (23−19 ka). The grey and yellow areas in A,B,C denote σ_1 (standard deviation) and σ_2 (upper estimate) uncertainties, respectively. See Supplementary Information for a more detailed figure and further details including an in-depth description how uncertainties were calculated.

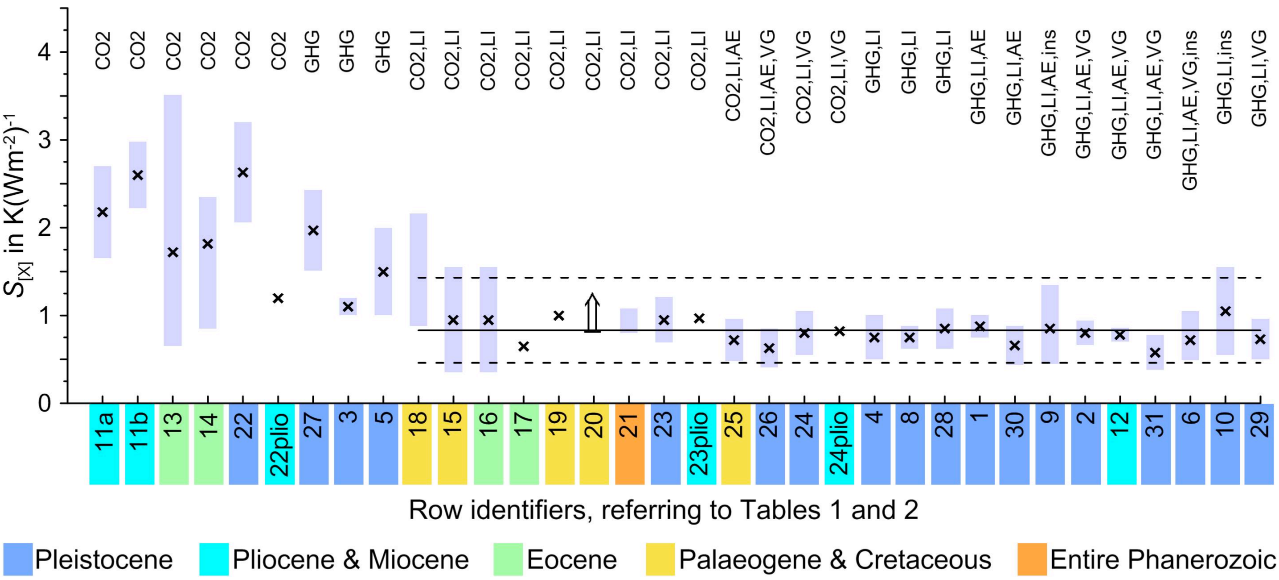

Figure 3. Summary of Tables 1 and 2 (x-axis labels refer to numbered rows). The Pliocene values from Table 2 are identified with 'plio'. Colour tabs refer to broad geological intervals as shown in the legend. Codes at the top indicate which conditions were explicitly accounted for; i.e., as 'forcings'. Asterisks refer to the fact that, in an ice-free world, the influence of LI is effectively accounted for with a value of 0 Wm^{-2}. Bars show ranges for estimates where ranges are reported, and crosses show central values where reported. Arrow indicates the value that was reported as >0.8 $K(Wm^{-2})^{-1}$. Black lines show mean (solid) and 95% confidence limits (dashed) for all estimates that account for at least ‘CO2’ and ‘LI’ $K(Wm^{-2})^{-1}$).