The equilibrium sensitivity of the Earth's temperature to radiation changes

The Earth's climate is changing rapidly as a result of anthropogenic carbon emissions, and damaging impacts are expected to increase with warming. To prevent these and limit long-term global surface warming to, for example, 2 °C, a level of stabilization or of peak atmospheric CO_2 concentrations needs to be set. Climate sensitivity, the global equilibrium surface warming after a doubling of atmospheric CO_2 concentration, can help with the translation of atmospheric CO_2 levels to warming. Various observations favour a climate sensitivity value of about 3 °C, with a likely range of about 2–4.5 °C. However, the physics of the response and uncertainties in forcing lead to fundamental difficulties in ruling out higher values. The quest to determine climate sensitivity has now been going on for decades, with disturbingly little progress in narrowing the large uncertainty range. However, in the process, fascinating new insights into the climate system and into policy aspects regarding mitigation have been gained. The well-constrained lower limit of climate sensitivity and the transient rate of warming already provide useful information for policy makers. But the upper limit of climate sensitivity will be more difficult to quantify.

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When the radiation balance of the Earth is perturbed, the global surface temperature will warm and adjust to a new equilibrium state. But by how much? The answer to this seemingly basic but important question turns out to be a tricky one. It is determined by a number termed equilibrium climate sensitivity, the global mean surface warming in response to a doubling of the atmospheric CO_2 concentration after the system has reached a new steady state. Climate sensitivity cannot be measured directly, but it can be estimated from comprehensive climate models. It can also be estimated from climate change over the twentieth century or from short-term climate variations such as volcanic eruptions, both of which were observed instrumentally, and from climate changes over the Earth's history that have been reconstructed from palaeoclimatic data. Many model-simulated aspects of climate change scale approximately linearly with climate sensitivity, which is therefore sometimes seen as the 'magic number' of a model. This view is too simplistic and misses many important spatial and temporal aspects of climate change. Nevertheless, climate sensitivity is the largest source of uncertainty in projections of climate change beyond a few decades[1–3] and is therefore an important diagnostic in climate modelling[4,5].

THE CONCEPT OF FORCING, FEEDBACK AND CLIMATE SENSITIVITY

The concept of radiative forcing, feedbacks and temperature response is illustrated in Fig. 1. Anthropogenic emissions of greenhouse gases, aerosol precursors and other substances, as well as natural changes in solar irradiance and volcanic eruptions, affect the amount of radiation that is reflected, transmitted and absorbed by the atmosphere. This externally imposed (naturally or human-induced) energy imbalance on the system, such as the increased long-wave absorption caused by the emission of anthropogenic CO_2, is termed radiative forcing (ΔF). In a simple global energy balance model, the difference between these (positive) radiative perturbations ΔF and the increased outgoing long-wave radiation that is assumed to be proportional to the surface warming ΔT leads to an increased heat flux ΔQ in the system, such that

$$\Delta Q = \Delta F - \lambda \Delta T \qquad (1)$$

Heat is taken up largely by the ocean, which leads to increasing ocean temperatures[6]. The changes in outgoing long-wave radiation that balance the change in forcing are influenced by climate feedbacks. For a constant forcing, the system eventually approaches a new equilibrium where the heat uptake ΔQ is zero and the radiative forcing is balanced by additional emitted long-wave radiation. Terminology varies, but commonly the ratio of forcing and equilibrium temperature change $\lambda = \Delta F/\Delta T$ is defined as the climate feedback parameter (in W m^{-2} °C^{-1}), its inverse $S' = 1/\lambda = \Delta T/\Delta F$ the climate sensitivity parameter

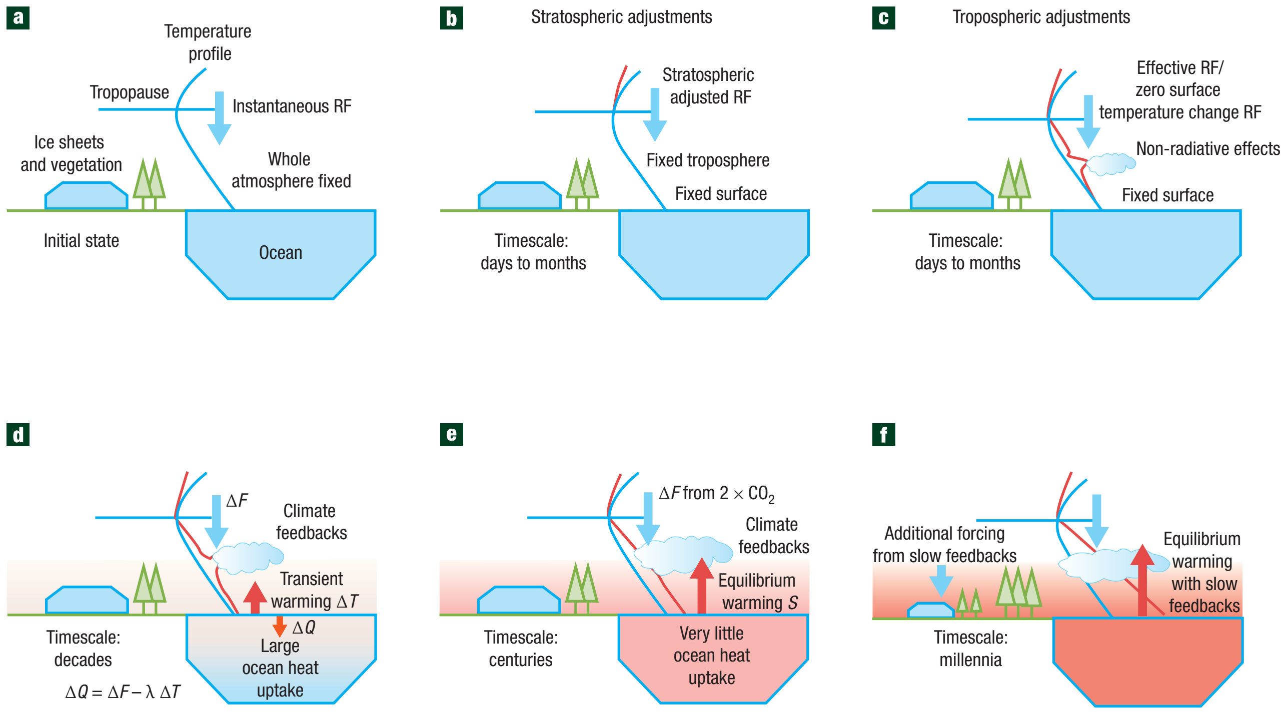

Figure 1 The concept of radiative forcing, feedbacks and climate sensitivity. **a**, A change in a radiatively active agent causes an instantaneous radiative forcing (RF). **b**, The standard definition of RF includes the relatively fast stratospheric adjustments, with the troposphere kept fixed. **c**, Non-radiative effects in the troposphere (for example of CO_2 heating rates on clouds and aerosol semi-direct and indirect effects) occurring on similar timescales can be considered as fast feedbacks or as a forcing. **d–f**, During the transient climate change phase (**d**), the forcing is balanced by ocean heat uptake and increased long-wave radiation emitted from a warmer surface, with feedbacks determining the temperature response until equilibrium is reached with a constant forcing (**e**, **f**). The equilibrium depends on whether additional slow feedbacks (for example ice sheets or vegetation) with their own intrinsic timescale are kept fixed (**e**) or are allowed to change (**f**).

(in °C W^{-1} m^2) and $S = \Delta T_{2\times CO_2}$ the equilibrium climate sensitivity, the equilibrium global average temperature change for a doubling (usually relative to pre-industrial) of the atmospheric CO_2 concentration, which corresponds to a long-wave forcing of about 3.7 W m^{-2} (ref. 7). The beauty of this simple conceptual model of radiative forcing and climate sensitivity (equation (1)) is that the equilibrium warming is proportional to the radiative forcing and is readily computed as a function of the current CO_2 relative to the pre-industrial CO_2: $\Delta T = S \ln(CO_2/CO_{2(t=1750)})/\ln 2$. The total forcing is assumed to be the sum of all individual forcings. The sensitivity S can also be phrased as[8–10]

$$S = \Delta T_0/(1 - f) \qquad (2)$$

where f is the feedback factor amplifying (if $0 < f < 1$) or damping the initial blackbody response of $\Delta T_0 = 1.2$ °C for a CO_2 doubling. The total feedback can be phrased as the sum of all individual feedbacks[9] (see Fig. 2; examples of feedbacks are increases in the greenhouse gas water vapour with warming; other feedbacks are associated with changes in lapse rate, albedo and clouds). To first order, the feedbacks are independent of T, yielding a climate sensitivity that is constant over time and similar between many forcings. The global temperature response from different forcings is therefore approximately additive[11]. However, detailed studies find that some feedbacks will change with the climate state[12–14], which means that the assumption of a linear feedback term $\lambda\Delta T$ is valid only for perturbations of a few degrees. There is a difference in the sensitivity to radiative forcing for different forcing mechanisms, which has been phrased as their 'efficacy'[7,15]. These effects are represented poorly or not at all in simple climate models[16]. A more detailed discussion of the concepts and the history is given in refs 5, 7, 17–20.

Note that the concept of climate sensitivity does not quantify carbon-cycle feedbacks; it measures only the equilibrium surface response to a specified CO_2 forcing. The timescale for reaching equilibrium is a few decades to centuries and increases strongly with sensitivity[21]. The transient climate response (TCR, defined as the warming at the point of CO_2 doubling in a model simulation in which CO_2 increases at 1% yr^{-1}) is a measure of the rate of warming while climate change is evolving, and it therefore depends on the ocean heat uptake ΔQ. The dependence of TCR on sensitivity decreases for high sensitivities[9,22,23].

ESTIMATES FROM COMPREHENSIVE MODELS AND PROCESS STUDIES

Ever since concern arose about increases of CO_2 in the atmosphere causing warming, scientists have attempted to estimate how much warming will result from, for example, a doubling of the atmospheric CO_2 concentration. Even the earliest estimates ranged remarkably close to our present estimate of a likely increase of between 2 and 4.5 °C (ref. 24). For example, Arrhenius[25] and Callendar[26], in the years 1896 and 1938, respectively, estimated that a doubling of CO_2 would result in a global temperature increase of 5.5 and 2 °C. Half a century later, the first energy-balance models, radiative convective models and general circulation models (GCMs) were used to quantify forcings and feedbacks, and with it the climate sensitivity S (refs 9, 21, 27–31). Climate sensitivity is not a directly tunable quantity in GCMs and depends on many parameters related mainly to atmospheric processes. Different sensitivities in GCMs can be obtained by perturbing parameters affecting clouds, precipitation, convection, radiation, land surface and other processes. Two decades ago, the largest uncertainty in these feedbacks was attributed to clouds[32]. Process-based studies now find a stronger constraint on the combined feedbacks from increases in water vapour and changes in the lapse rate. These studies still identify low-level clouds as the dominant uncertainty in feedback[4,5,33].

Requiring that climate models reproduce the observed present-day climatology (spatial structure of the mean climate and its variability) provides some constraint on model climate sensitivity. Starting in the 1960s (ref. 27), climate sensitivities in early GCMs were mostly in the range 1.5–4.5 °C. That range has changed very little since then, with the current models covering the range 2.1–4.4 °C (ref. 5), although higher values are possible[34]. This can be interpreted as disturbingly little progress or as a confirmation that model simulations of atmospheric feedbacks are quite robust to the details of the models. Three studies have calculated probability density functions (PDFs) of climate sensitivity by comparing different variables of the present-day climate against observations in a perturbed physics ensemble of an atmospheric GCM coupled to a slab ocean model[35–37]. These distributions reflect the uncertainty in our knowledge of sensitivity, not a distribution from which future climate change is sampled. The estimates are in good agreement with other estimates (Fig. 3). The main caveat is that all three studies are based on a version of the same climate model and may be similarly influenced by biases in the underlying model.

CONSTRAINTS FROM THE INSTRUMENTAL PERIOD

Many recent estimates of the equilibrium climate sensitivity are based on climate change that has been observed over the instrumental period (that is, about the past 150 years). Wigley *et al.*[38] pointed out that uncertainties in forcing and response made it impossible to use observed global temperature changes during that period to constrain S more tightly than the range explored by climate models (1.5–4.5 °C at the time), and that the upper end of the range was particularly difficult to estimate, although qualitatively similar conclusions appear in earlier pioneering work[9,10,21,39]. Several studies subsequently used the transient evolution of surface temperature, upper air temperature, ocean temperature or radiation in the past, or a combination of these, to constrain climate sensitivity. An overview of ranges and PDFs of climate sensitivity from those methods is shown in Fig. 3. Several studies used the observed surface and ocean warming over the twentieth century and an estimate of the radiative forcing to estimate sensitivity, either by running large ensembles with different parameter settings in simple or intermediate-complexity models[3,38,40–46], by using a statistical model[47] or in an energy balance calculation[48]. Satellite data for the radiation budget were also used to infer climate sensitivity[49]. The advantage of these methods is that they consider a state of the climate similar to today's and use similar timescales of observations to the projections we are interested in, thus providing constraints on the overall feedbacks operating today. However, the distributions are wide and cannot exclude high sensitivities. The main reason is that it cannot be excluded that a strong aerosol forcing or a large ocean heat uptake might have hidden a strong greenhouse warming.

Some recent analyses have used the well-observed forcing and response to major volcanic eruptions during the twentieth century, notably the eruption of Mount Pinatubo. The constraint

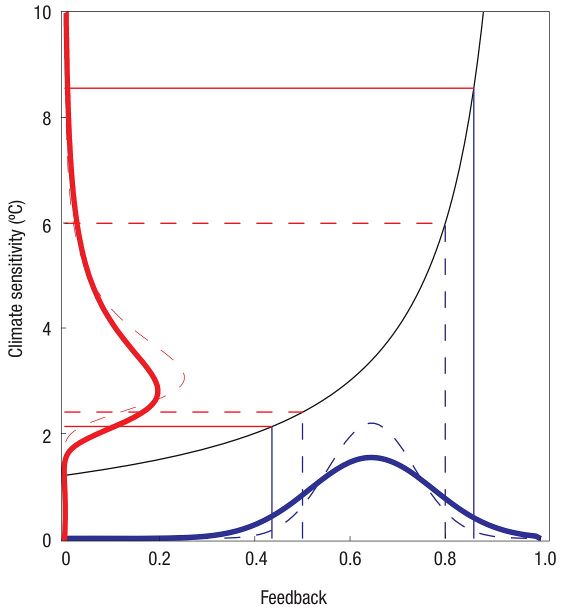

Figure 2 Relation between amplifying feedbacks f and climate sensitivity S. A truncated normal distribution with a mean of 0.65 and standard deviation of 0.13 for the feedback f (solid blue line) is assumed here for illustration. These values are typical for the current set of GCMs[8,33]. Because f is substantially positive and the relation between f and S is nonlinear (black line, equation (2)), this leads to a skewed distribution in S (solid red line) with the characteristic long tail seen in most studies. Horizontal and vertical lines mark 5–95% ranges. A decrease in the uncertainty of f by 30% (dashed blue line) decreases the range of S, but the skewness remains (dashed red line). The uncertainty in the tail of S depends not only on the uncertainty in f but also on the mean value of f. Note that the assumption of a linear feedback (equation (1)) is not valid for f near unity. Feedbacks of 1 or more would imply unphysical, catastrophic runaway effects. (Modified from ref. 8.)

is fairly weak because the peak response to short-term volcanic forcing has a nonlinear dependence on equilibrium sensitivity, yielding only slightly enhanced peak cooling for higher values of S (refs 42, 50, 51). Nevertheless, models with climate sensitivity in the range 1.5–4.5 °C generally perform well in simulating the climate response to individual volcanic eruptions and provide an opportunity to test the fast feedbacks in climate models[5,52,53].

PALAEOCLIMATIC EVIDENCE

Some early estimates of climate sensitivity drew on palaeoclimate information. For example, the climate of the Last Glacial Maximum (LGM) is a quasi-equilibrium response to substantially altered boundary conditions (such as large ice sheets over landmasses of the Northern Hemisphere, and different vegetation) and different atmospheric CO_2 levels. Simple calculations relating the peak cooling to changes in radiative forcing yielded estimates mostly between 1 and 6 °C, which turned out to be close to Arrhenius's estimates[9,54–56]. Simulations of the LGM are still an important testbed for the response of climate models to radiative forcing[57]. In some recent studies, parameters in climate models have been perturbed systematically to estimate S (refs 14, 58, 59). The idea is to estimate the sensitivity of a perturbed model by running it to equilibrium with doubled CO_2 and then evaluate whether the same model yields realistic simulations of the LGM conditions. This method avoids directly estimating the relationship between

Figure 3 Distributions and ranges for climate sensitivity from different lines of evidence. **a**, The most likely values (circles), likely (bars, more than 66% probability) and very likely (lines, more than 90% probability) ranges are subjective estimates by the authors based on the available distributions and uncertainty estimates from individual studies, taking into account the model structure, observations and statistical methods used. Values are typically uncertain by 0.5 °C. Dashed lines indicate no robust constraint on an upper bound. Distributions are truncated in the range 0–10 °C; most studies use uniform priors in climate sensitivity. Details are discussed in refs 18, 24, 75 and in the text. Single extreme estimates or outliers (some not credible) are marked with crosses. The IPCC[24] likely range and most likely value are indicated by the vertical grey bar and black line, respectively. **b**, A partly subjective classification of the different lines of evidence for some important criteria. The overall level of scientific understanding (LOSU) indicates the confidence, understanding and robustness of an uncertainty estimate towards assumptions, data and models. Expert elicitation[90] and combined constraints are difficult to assess; both should have a higher LOSU than single lines of evidence, but experts tend to be overconfident and the assumptions are often not clear.

forcing and response, and thus avoids the assumption that the feedback factor is invariant for this very different climatic state. Instead, the assumption is that the change in feedbacks with climate state is simulated well in a climate model. The resulting estimates of climate sensitivity are quite different for two such attempts[58,59], illustrating the crucial importance of the assumptions in forcings (dust, vegetation or ice sheets) and of differences in the structure of the models used[60].

A few people have used palaeoclimate reconstructions from the past millennium to gain insight into climate sensitivity on the basis of a large sample of decadal climate variations that were influenced by natural forcing, and particularly volcanic eruptions[61,62]. Because of a weak signal and large uncertainties in reconstructions and forcing data (particularly solar and volcanic forcing), the long time horizon yielded a weak constraint on S (ref. 62) (see Fig. 3), arising mainly from low-frequency temperature variations associated with changes in the frequency and intensity of volcanism. Direct estimates of the equilibrium sensitivity from forcing between the Maunder Minimum period of low solar forcing and the present are also broadly consistent with other estimates[63].

Some studies of other, more distant palaeoclimate periods[64,65] seem to be consistent with the estimates from the more recent past. For example, the relationship between temperature over the past 420 million years[64] supports sensitivities that are larger than 1.5 °C, but the upper tail is poorly constrained and uncertainties in the models that are used are significant and difficult to quantify.

There are few studies that yield estimates of S that deviate substantially from the consensus range, mostly towards very low values. These results can usually be attributed to erroneous forcing assumptions (for example hypothesized external processes such as cosmic rays driving climate[66]), neglect of internal climate variability[67], overly simplified assumptions, neglected uncertainties, errors in the analysis or dataset, or a combination of these[68–71]. These results are typically inconsistent with comprehensive models. In some cases they were refuted by testing the method of estimation with a climate model with known sensitivity[50,72–74].

Several studies and assessments have discussed the available estimates for climate sensitivity in greater detail[4,5,17,18,23,24,75]. In summary, most studies find a lower 5% limit between 1 and 2 °C (Fig. 3). The combined evidence indicates that the net feedbacks f to radiative forcing (equation (2)) are significantly positive and emphasizes that the greenhouse warming problem will not be small. Figure 3 further shows that studies that use information in a relatively complete manner generally find a most likely value between 2 and 3.5 °C and that there is no credible line of evidence that yields very high or very low climate sensitivity as a best estimate. However, the figure also quite dramatically illustrates that the upper limit for S is uncertain and exceeds 6 °C or more in many studies. The reasons for this, and the caveats and limitations, are discussed below.

On the basis of the available evidence, the IPCC Fourth Assessment Report concluded that constraints from observed recent climate change[18] support the overall assessment that climate sensitivity is very likely (more than 90% probability) to be larger than 1.5 °C and likely (more than 66% probability) to be between 2 and 4.5 °C, with a most likely value of about 3 °C (ref. 24). More recent studies support these conclusions[8,45,51,64], with the exception of estimates based on problematic assumptions discussed above[67,69,71].

A LACK OF PROGRESS?

The large uncertainty in climate sensitivity seems disturbing to many. Have we not made any progress? Or are scientists just anchored on a consensus range[76]? Indeed, observations have not strongly constrained climate sensitivity so far. The latest generation of GCMs, despite clear progress in simulating past and present climate[5,18,24,77], covers a range of S of 2.1–4.4 °C (ref. 5), which is very similar to earlier models and not much different from the canonical range of 1.5–4.5 °C first put forward in 1979 by Charney[78] and later adopted in several IPCC reports[20,79].

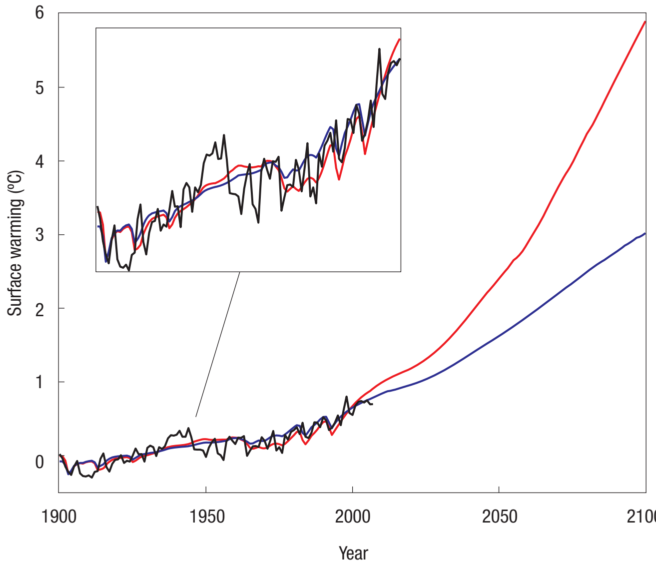

Figure 4 The observed global warming provides only a weak constraint on climate sensitivity. A climate model of intermediate complexity[3], forced with anthropogenic and natural radiative forcing, is used to simulate global temperature with a low climate sensitivity and a high total forcing over the twentieth century (2 °C, 2.5 W m^{-2} in the year 2000; blue line) and with a high climate sensitivity and low total forcing (6 °C, 1.4 W m^{-2}; red line). Both cases (selected for illustration from a large ensemble) agree similarly well with the observed warming (HadCRUT3v; black line) over the instrumental period (inset), but show very different long-term warming for SRES scenario A2 (ref. 101). For simplicity, ocean parameters are kept constant here.

The fact that high sensitivities are difficult to rule out was recognized more than two decades ago[9,21,39,80]. One reason is that the observed transient warming relates approximately linearly to S only for small values but becomes increasingly insensitive to S for shorter timescales and higher S, largely because ocean heat uptake prevents a linear response in S (equation (1))[21–23,81]. In addition, the uncertainty in aerosol forcing prevents the conclusions that the total forcing ΔF is strongly positive[7]; if ΔF were close to zero, S would have to be large to explain the observed warming. This is illustrated in Fig. 4, showing that both a low sensitivity combined with a high forcing and a high sensitivity with a low forcing can reproduce the forced component of the observed warming[3,40–45,48]. A high sensitivity can also be compensated for by a high value of ocean heat uptake. Different combinations of forcing, sensitivity, ocean heat uptake and surface warming (all of which are uncertain) can therefore satisfy the global energy balance (equation (1)).

A further fundamental reason for the fat tail of S is that S is proportional to $1/(1-f)$ (equation (2))[8–10]. This relation goes remarkably far in explaining the PDFs of S on the basis of the range of the feedback f estimated in GCMs[33], if the uncertainty in f is assumed to be Gaussian. Reducing the uncertainty in f reduces the range of S, in particular the upper bound, but the skewness remains[8] (see Fig. 2). Recent work on constraining individual feedbacks[4] is promising and helps in isolating model uncertainties and deficiencies, but it has not yet narrowed the range of f

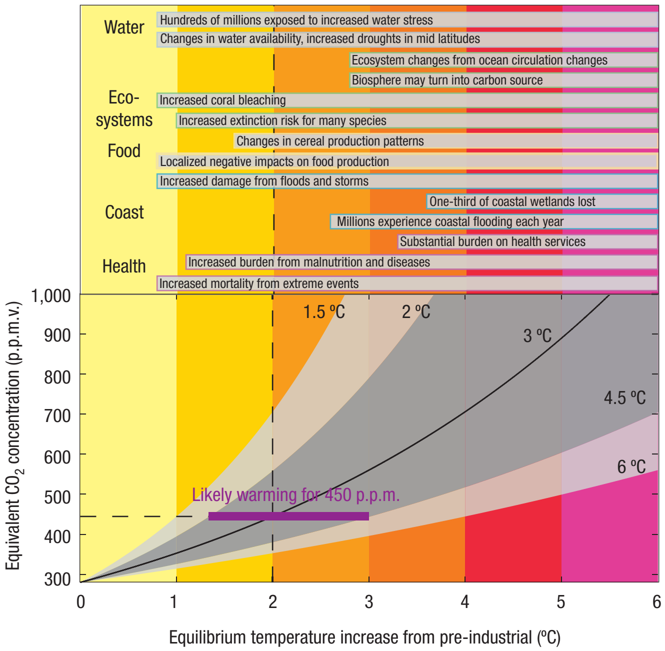

Figure 5 Relation between atmospheric equivalent CO_2 concentration chosen for stabilization and key impacts associated with equilibrium global temperature increase. According to the concept of climate sensitivity, equilibrium temperature change depends only on climate sensitivity S and on the logarithm of CO_2. The most likely warming is indicated for S = 3 °C (black solid), the likely range (dark grey) is for S = 2–4.5 °C (ref. 24) (see Fig. 3). The 2 °C warming above the pre-industrial temperature, often assumed to be an approximate threshold for dangerous interference with the climate system, is indicated by the black vertical dashed line for illustration. Stabilization at 450 p.p.m. by volume (p.p.m.v.) equivalent CO_2 concentration (horizontal dashed line) has a probability of less than 50% of meeting the 2 °C target, whereas 400 p.p.m. would probably meet it[22]. Selected key impacts (some delayed) for several sectors and different temperatures are indicated in the top part of the figure, based on the recent IPCC report (Fig. SPM.2 in ref. 100). For high CO_2 levels, limitations in the climate sensitivity concept introduce further uncertainties in the CO_2–temperature relationship not considered here (see the text).

substantially. In the example of Fig. 2, reducing the 95th centile of S from about 8.5 °C to 6 °C requires a decrease in the total feedback uncertainty of about 30%.

Although uncertainties remain large, it would be presumptuous to say that science has made no progress, given the improvements in our ability to understand and simulate past climate variability and change[18] as well as in our understanding of key feedbacks[4,5]. Support for the current consensus range on S now comes from many different lines of evidence, the ranges of which are consistent within the uncertainties, relatively robust towards methodological assumptions (except for the assumed prior distributions; see below) and similar for different types and generations of models. The processes contributing to the uncertainty are now better understood.

LIMITATIONS AND WAYS FORWARD

There are known limitations to the concept of forcing and feedback that are important to keep in mind. The concept of radiative forcing is of rather limited use for forcings with strongly varying vertical or spatial distributions[7,19]. In addition, the equilibrium response depends on the type of forcing[15,82,83]. As mentioned above, climate sensitivity may also be time-dependent or state-dependent[12,14,16,84]; for example, in a much warmer world with little snow and ice, the surface albedo feedback would be different from today's. Some models indicate that sensitivity depends on the magnitude of the forcing or warming[84,85]. These effects are poorly understood and are mostly ignored in simpler models that prescribe climate sensitivity. They are likely to be particularly important when estimating climate sensitivity directly from climate states very different from today's (for example palaeoclimate), for forcings other than CO_2, and in simple models in which climate sensitivity is a prescribed fixed number and all radiative forcings are treated equally as a change in the flux at the top of the atmosphere. Structural problems in the models, for example in the representation of cloud feedback processes or the physics of ocean mixing[5], in particular in cases in which all models make similar simplifications, will also affect results for climate sensitivity and are very difficult to quantify.

The classical 'Charney' sensitivity that results from doubling CO_2 in an atmospheric GCM coupled to a slab ocean model includes the feedbacks that occur on a timescale similar to that of the surface warming (namely mainly water vapour, lapse rate, clouds and albedo feedbacks). There is an unclear separation between forcing and fast feedbacks (for example clouds changing as a result of CO_2-induced heating rates rather than the slower surface warming[86,87]). Additionally, slow feedbacks with their own intrinsic timescale, for example changes in vegetation or the retreat of ice sheets and their effect on the ocean circulation, could increase or decrease sensitivity on long timescales[88,89] but are kept fixed in models (see Fig. 1). Currently, the climate sensitivity parameter (the response to 1 W m^{-2} of any forcing) times the forcing at the time of CO_2 doubling, the equilibrium climate sensitivity for CO_2 doubling in a fully coupled model, the 'Charney' sensitivity of a slab model and the effective climate sensitivity determined from a transient imbalance are all mostly assumed to be the same number and are all termed 'climate sensitivity'. Because few coupled models have been run to equilibrium and the validity of these concepts for high forcings is not well established, care should be taken in extrapolating observationally constrained effective sensitivities or slab model sensitivities to long-term projections for CO_2 levels beyond doubling, because feedbacks should be quite different in a substantially warmer climate.

Despite these limitations, S is a quantity that is useful in estimating the level of CO_2 concentrations consistent with an equilibrium temperature change below some 'dangerous' threshold, as shown in Fig. 5, although the lack of a clear upper limit on S makes it difficult to estimate a safe CO_2 stabilization level for a given temperature target. What are the options for learning more about climate sensitivity? Before discussing this, a methodological point affecting estimates of S needs to be mentioned: results from methods estimating a PDF of climate sensitivity depend strongly on their assumptions of a prior distribution from which climate models with different S are sampled[42]. Studies that start with climate sensitivity being equally distributed in some interval (for example 1–10 °C) yield PDFs of S with longer tails than those that sample models that are uniformly distributed in feedbacks (that is, the inverse of S (refs 35, 49)). Truly uninformative prior assumptions do not exist, because the sampling of a model space is ambiguous (that is, there is no single metric of distance between two models). Subjective choices are part of Bayesian methods, but because the data constraint is weak here, the implications are profound. An alternative prior distribution that has been used occasionally is an estimate of the PDF of S based on expert opinion[43,44,90] (Fig. 3). However, experts almost invariably know about climate change in different periods (for example the observed warming, or the temperature at the LGM), which introduces concern about the independence of prior and posterior information.

Another option that makes use of this Bayesian framework is to combine some of the individually derived distributions to yield a better constraint[62,91]. Combining pieces of information about *S* that are independent of each other and arise from different time periods or climatic states should provide a tighter distribution. The similarity of the PDFs arising from various lines of evidence shown in Fig. 3 substantially increases confidence in an overall estimate. However, the difficulty in formally combining lines of evidence lies in the fact that every single line of evidence needs to be entirely independent of the others, unless dependence is explicitly taken into account. Additionally, if several climate properties are estimated simultaneously that are not independent, such as *S* and ocean heat uptake, then combining evidence requires combining joint probabilities rather than multiplying marginal posterior PDFs[62]. Neglected uncertainties will become increasingly important as combining multiple lines of evidence reduces other uncertainties, and the assumption that the climate models simulate changes in feedbacks correctly between the different climate states may be too strong, particularly for simpler models. All of this may lead to unduly confident assessments, which is a reason that results combining multiple lines of evidence are still treated with caution. Figure 3b is a partly subjective evaluation of the different lines of evidence for several criteria that need to be considered when combining lines of evidence in an assessment. The prospect for the success of these combined constraints may be better than that of arriving at a tight constraint from a single line of observations. Additionally, rather than evaluating models by using what is readily observed (but may be weakly related to climate sensitivity)[34], ensembles of models could help to identify which observables are related to climate sensitivity and could thus provide a better constraint[36,92]. Future observations of continued warming of atmosphere and ocean, along with better estimates of radiative forcing, will eventually provide tighter estimates. New data may open additional opportunities for evaluating climate models. Finally, for the particular purpose of understanding climate sensitivity and characterizing uncertainty, large ensembles of models with different parameter settings[34] probably provide more insight than a small set of very complex models.

POLICY IMPLICATIONS

Whether the uncertainty in climate sensitivity matters depends strongly on the perspective. There is no consensus on whether the goal of the United Nations Framework Convention on Climate Change of 'stabilization of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate' is a useful target to inform policy. But if certain levels of warming are to be prevented even in the long run (for example to prevent the Greenland ice sheet from melting), then climate sensitivity, particularly the upper bound[93], is critical. For example, if the damage function is assumed to increase exponentially with temperature and the tail of climate sensitivity is fat (that is, the damage with temperature increases faster than the probability of such an event), then the expected damage could be infinite, entirely dominated by the tiny probability of a disastrous event[94]. In contrast, if a cost–benefit framework with sufficiently large discounting is adopted, climate change beyond a century is essentially irrelevant; if the exponential discounting dominates the increasing damage, then climate sensitivity is unimportant simply because the discounted damage is insensitive to the stabilization level. Thus, the policy relevance of climate sensitivity for mitigation depends on an assumed economic framework, discount rate and the timescale of interest.

For short-term scenario projections, the transient climate response and peak warming in a CO_2 overshoot scenario are better constrained than equilibrium changes, because they are linearly related to observations and show much less skewed distributions[81,95–98]. The prospects for well-constrained projections on the timescales of a few decades are thus brighter[1], and these may be more useful for decision makers in the short term. Furthermore, for a stabilization at, for example, 450 p.p.m. CO_2 equivalent forcing, which is a level that would avoid a long-term warming of 2 °C above pre-industrial temperatures with a probability of rather less than 50% (Fig. 5), the necessary emission reductions are large and not strongly affected by the uncertainty in *S*. The uncertainties in such an emission pathway are shown in Fig. 6, considering only CO_2. Taking non-CO_2 forcings into account requires even lower emissions.

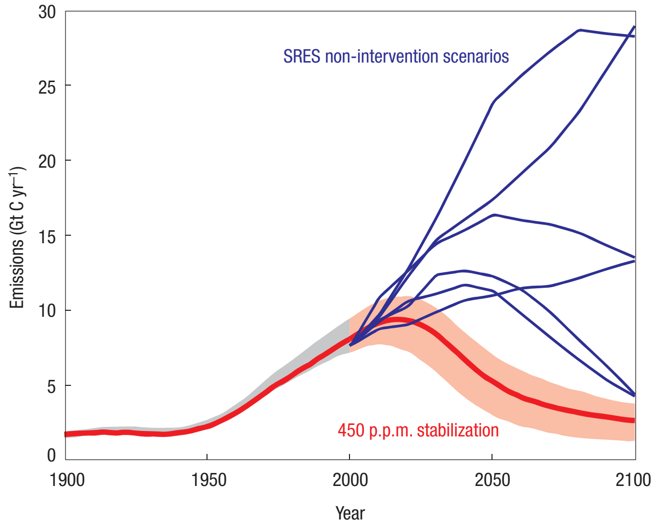

Figure 6 Allowed emissions for a stabilization of atmospheric CO_2 at 450 p.p.m. as shown in Fig. 5. Emission reductions needed for stabilization at 450 p.p.m. (red) must be much larger than in any of the illustrative SRES scenarios (blue lines). The best guess (red line) is based on a climate sensitivity of 3.2 °C and standard carbon cycle settings in a climate model of intermediate complexity[99]. Uncertainties in emission reduction (red band) are quantified by combining a low climate sensitivity (1.5 °C) with a fast carbon cycle and a high climate sensitivity (4.5 °C) with a slow carbon cycle. The emission pathway over the next few decades for stabilization at low levels is not strongly affected by the uncertainty in climate sensitivity. Accounting for non-CO_2 forcings requires even lower emissions than shown here to reach the same equivalent radiative forcing target. (Modified from ref. 99.)

Long-term stabilization targets depend on climate sensitivity and on carbon-cycle–climate feedbacks[99]. The uncertainty in both of these, if the past is indicative of the future, may not decrease quickly. However, the tight constraint on the lower limit of sensitivity indicates a need for strong and immediate mitigation efforts if the world decides that large climate change should be avoided (Figs 5 and 6). The uncertainty in short-term targets is quite small, and as scientists continue to narrow the estimates of the climate sensitivity, and as the feasibility of emission reductions is explored, long-term emission targets can be adjusted on the basis of future insight.

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