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Climate sensitivity: What is it, and why is it important?

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Headlines

- Climate sensitivity is defined as *the global mean increase in surface air temperature resulting from a doubling of atmospheric CO_2 concentration, once the system has reached equilibrium.*
- Its exact value is uncertain and depends mainly on the strength of climate feedbacks that can amplify or dampen the effect of CO_2 forcing. The most uncertain of these feedbacks is that related to clouds.
- Although based on a simple benchmark experiment, climate sensitivity is a good indicator of the rate of future global warming for a given scenario of greenhouse gas emissions. In turn, many societally relevant regional impacts, such as local changes in temperature and aridity, scale approximately linearly with global warming.
- All other things being equal, a higher climate sensitivity implies a lower remaining carbon budget for a given global warming limit. An understanding of climate sensitivity is thus central to carbon budget estimates for the Paris Agreement goals.

Introduction

Human activities have driven carbon dioxide (CO_2) concentrations in the atmosphere to levels unprecedented in human history, reaching 415 parts per million (ppm) in May 2019 – the highest value in 14 million years. If current emission levels are sustained, the concentrations of CO_2 and other greenhouse gases are expected to continue rising rapidly. It is therefore urgently necessary to accurately quantify how increasing greenhouse gas concentrations will affect the global climate system.

The most obvious and direct manifestation of increasing greenhouse gas levels is the rise in global temperatures – "global warming". This temperature increase results from the radiative forcing induced by greenhouse gases (Box 1), which acts like a blanket to increase the amount of heat in the climate system. However, this warming will unfold at different rates in different regions of the world: for example, theory and observations agree that land masses warm faster than the ocean surface[1].

Contents

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Furthermore, temperature changes also affect other essential components of the climate system: most notably the hydrological cycle (precipitation and evaporation)[2], the mean sea level[3], and the circulation of the atmosphere, including the jet streams, mid-latitude storms, and tropical cyclones[4,5].

To simplify these complex interactions, it is common practice to quantify climate change in terms of the global mean surface temperature response – a single number summarising the many facets of global climate change. The most commonly used metric of future changes in global temperature is the **climate sensitivity.**

What is climate sensitivity and why is it uncertain?

Climate sensitivity is defined as *the global mean increase in surface air temperature resulting from a sustained doubling of atmospheric CO_2 concentration, once the system has reached equilibrium.* Its value is determined by two factors: the radiative forcing due to CO_2 doubling, and the climate feedbacks that can amplify or dampen the forcing (Box 1). The assumption that the system is at equilibrium means it has reached a stable average temperature, but this takes longer than a thousand years: this is because the oceans warm very slowly, owing to their large volume and slow circulation. Climate sensitivity is therefore a theoretical value that would only be achieved after several millennia.

In practice, when using model or real-world data, the two conditions required to assess climate sensitivity (forcing equal to a doubling of CO_2, climate at equilibrium) are typically not met. However, the sensitivity can still be accurately quantified if both radiative forcing and the climate feedbacks are known precisely. Climate scientists are currently working to reliably estimate these numbers in situations where the climate is not in equilibrium.

Climate sensitivity is mainly estimated in three ways[6]: using direct historical observations of temperature and the radiative budget; using palaeoclimate proxy data; and using global climate models. It is also possible to use a combination of these methods – for example, constraining model projections with observations (Fig. 1). While the results are consistent across methods overall, they are also subject to large uncertainties (bars in Fig. 1). The causes of these uncertainties are reviewed below for each of the methods.

Instrumental historical observations

Observational estimates of global temperature and the radiative budget can be used to calculate climate feedbacks, and therefore climate sensitivity[6], if the radiative forcing is known sufficiently accurately. Reliable estimates of global mean temperature are available from about 1880 onwards. The radiative budget is quantified using data from global satellite observations, but these exist only from the 1980s. More uncertain estimates can be obtained from measurements of the global ocean heat content, whose time evolution closely follows that of the radiative budget, going back to around 1960[7].

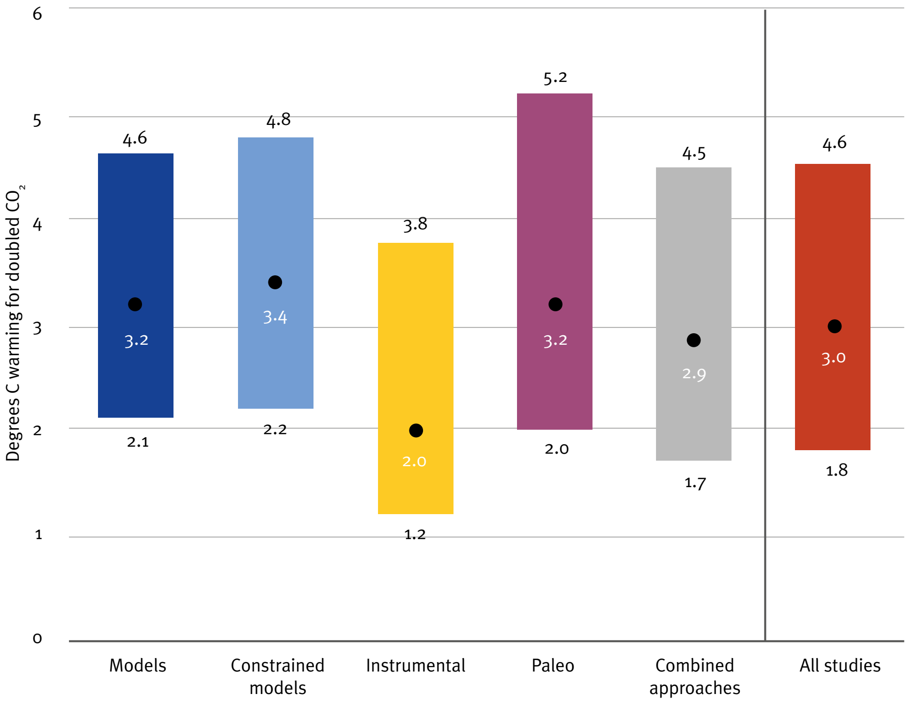

Figure 1: Range of climate sensitivity values from different estimation methods (global climate models, palaeoclimate proxies, instrumental observations). Note that because the values of climate sensitivity are bounded on the lower end (the sensitivity must be positive) but not on the upper end, the probability distributions are asymmetric, with a long tail of high values with low probability. Adapted from https://www.carbonbrief.org/explainer-how-scientists-estimate-climate-sensitivity

Box 1: Radiative forcing and climate feedbacks

In an equilibrium climate, the system is in a state of approximate *radiative balance:* it absorbs as much energy in the form of sunlight as it re-emits into space in the form of infrared radiation. Any external perturbation to this radiative balance (such as by increasing CO_2 concentration, or decreasing the intensity of the Sun) is termed a *radiative forcing,* measured in watts per square metre (W/m^2). The forcing quantifies the change in global mean energy flux into (or out of) the climate system, i.e. at the top of the atmosphere. For example, our best estimate for the forcing due to a doubling of atmospheric CO_2 is 3.7 W/m^2.

If CO_2 forcing was the only process affecting the Earth's radiative budget (with radiative emission into space proportional to temperature following Planck's law), climate sensitivity would be close to 1° C, with fairly little uncertainty. However, global warming triggers *climate feedbacks,* i.e. climate responses to the warming which can modulate the original radiative perturbation. The overall effect of feedbacks is to amplify the radiative perturbation and so increase climate sensitivity. The strongest feedback involves increases in atmospheric water vapour (a greenhouse gas) following increases in warming, an amplifying feedback which alone doubles the climate sensitivity. Further (weaker) feedbacks amplifying global warming involve decreasing ice and snow cover and changes in cloud properties.

Additional climate feedbacks exist that act on very long timescales (millennia or longer), and which mainly involve changes in ice sheets and vegetation. Such feedbacks – sometimes called *Earth system feedbacks* – remain poorly understood and are typically not represented in global climate models, so they are not included in the definition of climate sensitivity. Because of their long timescales, these slow feedbacks are of limited relevance to climate projections over the coming decades to centuries. However, they must be accounted for when estimating climate sensitivity from palaeoproxy data[12].

Estimating climate feedbacks and sensitivity from direct observations is challenging for a number of reasons. Firstly, radiative forcing is not an observable quantity, and estimates of it are subject to substantial uncertainty. This is particularly the case for the forcing due to anthropogenic aerosols (suspended particles that cool the climate by interfering with sunlight). Complex interactions between aerosols and clouds can enhance the radiative forcing of aerosols, but these interactions remain poorly understood[8]. To make matters worse, aerosol concentrations themselves are difficult to measure and therefore also uncertain.

Secondly, even if the historical radiative forcing and climate feedbacks were perfectly known, data from the historical climate can yield a biased estimate of climate sensitivity. This is because the forcing agents at play during the industrial era include not only CO_2, but also other greenhouse gases (e.g., methane), aerosols, and volcanic eruptions. Different forcing agents cause different feedbacks and therefore different sensitivities[9], which means the feedbacks observed in recent decades may not reflect the feedbacks that will be operating in the future, when CO_2 will become the dominant forcing agent. Research suggests that both volcanic and aerosol forcing have contributed to limiting the amplifying effect of feedbacks (making climate sensitivity lower) in past measurements[10].

A third issue is that feedbacks (particularly by clouds) depend strongly on the spatial pattern of surface warming. The warming pattern observed in recent decades is substantially different from the future long-term warming pattern predicted by climate models, likely resulting from the effect of natural variability. By running climate model simulations in which we impose the observed historical sea surface warming pattern, we can see that historical feedbacks were less amplifying than the future long-term feedbacks – that is to say, historical climate sensitivity calculations are biased towards a low estimate[11]. This low bias is commonly referred to in the climate literature as the "pattern effect". It will likely take more than another decade for a warming pattern to emerge in the real world that is representative of long-term forced climate change.

Palaeoclimate

Using indirect ("proxy") estimates of past climate conditions, we can delve considerably further into the past and exploit the large climate variations that have occurred naturally to estimate the sensitivity of the Earth's climate. The past few million years have been characterised by glacial-interglacial cycles, where the Earth's climate oscillated between conditions similar to the present day (interglacials), and ice ages with extensive continental ice sheets and considerably lower atmospheric CO_2 concentrations.

Proxy records of past climate variables, such as temperature and CO_2, can be obtained from various sources including ice cores, marine sediments and tree rings. Such records are typically point measurements. CO_2 is well-mixed in the atmosphere, but temperature is highly variable in space, causing uncertainty in proxy-based reconstructions of global temperature. However, challenges remain even when assuming perfect knowledge of past CO_2 and temperature conditions. Colder past climates were considerably different from present and future warm climates due to the presence of large continental ice sheets. As temperatures increased, the melting of these ice sheets triggered strong amplifying feedbacks which are not operating in the current climate. Other feedbacks involved large-scale changes in vegetation cover that modified the ability of land surfaces to absorb sunlight, and changes in the dust content of the atmosphere. Therefore, it is necessary to separately account for the (relatively uncertain) ice sheet, vegetation and dust feedbacks in order to quantify climate sensitivity from past climate changes[12].

Global climate models

Compared with inferences based on the direct or palaeo observational records, quantifying climate sensitivity from global climate model experiments is considerably more straightforward. However, climate models suffer from errors in the representation of both radiative forcing and climate feedbacks. This is because these quantities depend on physical processes that cannot be explicitly represented in the models, either because they occur on very small spatial or temporal scales, or because the physics are too complex or poorly understood (e.g. interactions between vegetation and climate). Such processes are instead “parameterised”, i.e. simulated in simplified ways, and this approach introduces inaccuracies. Examples of processes that must be parameterised include radiative transfer, cloud microphysical processes (i.e. the formation, growth and precipitation of liquid droplets and ice crystals that form clouds), and atmospheric convection.

In current global climate models, the single most important contributor to uncertainty in climate sensitivity is the cloud feedback – that is, the effect on the radiative budget of changes in the area, thickness and altitude of clouds. Estimates of the net global cloud feedback range from near-neutral (no feedback, favouring low climate sensitivity) to strongly amplifying (high climate sensitivity)[13]. This reflects the fact that the future cloud response to warming is uncertain, because it strongly depends on parameterised processes in climate models.

Is climate sensitivity higher than we thought?

Climate model simulations are performed by modelling centres around the world as part of the Coupled Model Intercomparison Project (CMIP). The CMIP simulations follow a common protocol to ensure that results from different models are comparable. Results from CMIP phase 5 (CMIP5) yielded climate sensitivity values ranging from 2.1° to 4.7° C (Fig. 2), supporting the IPCC Fifth Assessment Report’s “likely” range of 1.5° – 4.5° C for climate sensitivity. However, early results from the latest generation of climate models (developed for CMIP6) suggest a substantially higher range (Fig. 2), with several models producing sensitivity values exceeding 5° C.

The increase in climate sensitivity has been traced to a more amplifying cloud feedback, but the causes for this enhanced feedback are a matter of investigation. It appears that, for at least some of the models, the strongly amplifying cloud feedback is accompanied by an unrealistically negative aerosol forcing during the historical period[14]. This trade-off between warming by cloud feedback and cooling by aerosol forcing allows for a reasonably realistic simulation of historical warming, but implies much stronger future long-term warming once CO_2 becomes the dominant forcing agent[i]. It is unlikely that this trade-off affects all of the high-sensitivity models, however, and it remains entirely possible that these models are realistic in their representation of a strongly amplifying cloud feedback. This possibility would have important implications for carbon budget calculations for global warming limits, as discussed in the next section.

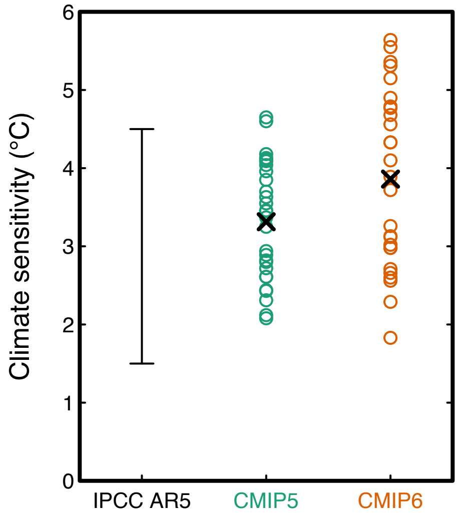

Figure 2: Climate sensitivity in individual CMIP5 and CMIP6 models (28 and 27 models, respectively) and the IPCC Fifth Assessment Report’s “likely” range. Black crosses denote the multi-model means. Values taken from Zelinka et al.[15].

Why is climate sensitivity important?

Temperature projections and regional climate change

In climate models, forcing, feedbacks and climate sensitivity are typically estimated in idealised simulations with constant CO_2 forcing, which are run for 150 years or longer. Although equilibrium is not reached, the length of the simulations and the constant forcing make it possible to relatively reliably estimate the equilibrium warming. In more realistic simulations of the future, with forcing agent concentrations increasing gradually in time, the climate is in a state of strong disequilibrium; furthermore, a variety of forcing agents are at play, such as non-CO_2 greenhouse gases, aerosols and volcanoes, which can elicit different feedbacks compared with CO_2.

Despite the differences in these simulations, climate sensitivity provides a good prediction of the rate of global warming in realistic scenarios of future climate change. As illustrated by the case of Representative Concentration Pathway 8.5 (RCP8.5), a high-emissions scenario, models with high climate sensitivity systematically predict more 21st century warming than the low-sensitivity models (Fig. 3a).

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i. CO_2 forcing is generally projected to rise much faster than aerosol forcing in the future owing to controls on air pollution, which are currently a major source of anthropogenic aerosols. Furthermore, CO_2 accumulates in the atmosphere over long time-periods, whereas aerosols are short-lived.

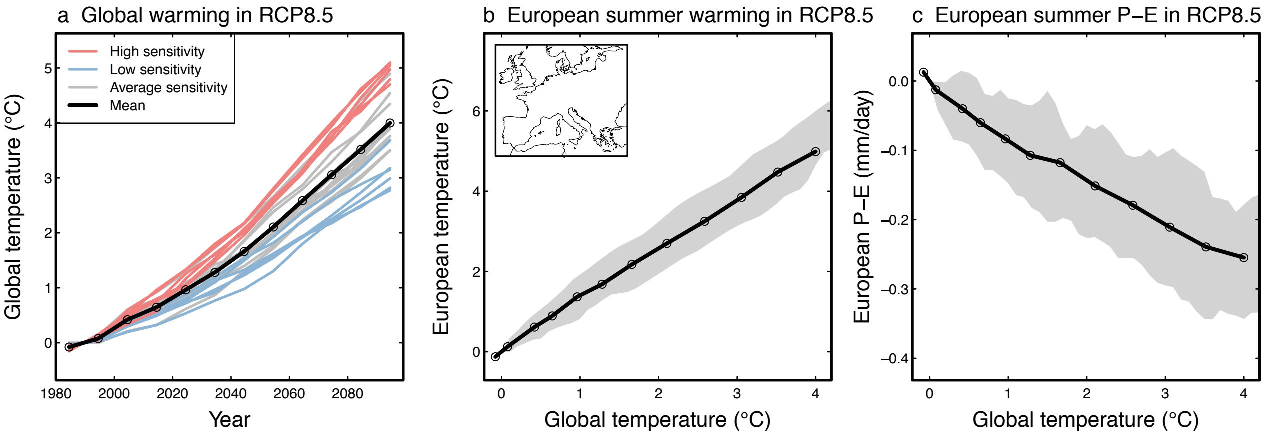

Figure 3: (a) Global warming in the RCP8.5 scenario as a function of time in 24 CMIP5 models (thin blue, grey and red curves). Temperature anomalies are relative to the period 1980-1999. Red denotes models with above average climate sensitivity (top 25% of the model distribution), blue denotes below average sensitivity (bottom 25%), grey denotes average sensitivity. The thick black line is the multi-model mean, and circles denote decadal averages. (b) European mean temperature (10°W to 30°E, 35°N to 60°N, inset map) during the summer half-year (May to October), versus global warming in the RCP8.5 simulations. The grey shading denotes the 90% range of the models. (c) As in (b) but for precipitation minus evaporation (P-E). More negative P-E indicates increasing aridity.

One might argue that climate sensitivity is of limited relevance to the economy or society, since impacts depend on regional changes rather than on global warming. However, many indicators of regional climate change are known to scale linearly with global warming, to a good approximation. Examples are shown in Fig. 3b-c for European summertime temperature and precipitation minus evaporation (P-E), a measure of hydrological changes. Decreasing P-E, as in Fig. 3c, means drier conditions and decreasing moisture availability. The results in Fig. 3b-c show how, on regional scales and considering a single season, changes in temperature and the hydrological cycle are tied to global warming. Since higher climate sensitivity implies more global warming, it also tends to favour larger regional climate changes. Similar relationships have been found for extreme temperature and rainfall[16].

It should be noted, as a caveat, that regional changes also depend on other factors, especially the spatial patterns of sea surface temperature change. In particular, atmospheric circulation changes (e.g. shifts in the position of the jet streams) do not scale well with global warming and climate sensitivity in models[17], as they are more closely linked to the spatial pattern of warming. These circulation changes can then further modify temperature and the hydrological cycle regionally, in addition to the effect of global warming.

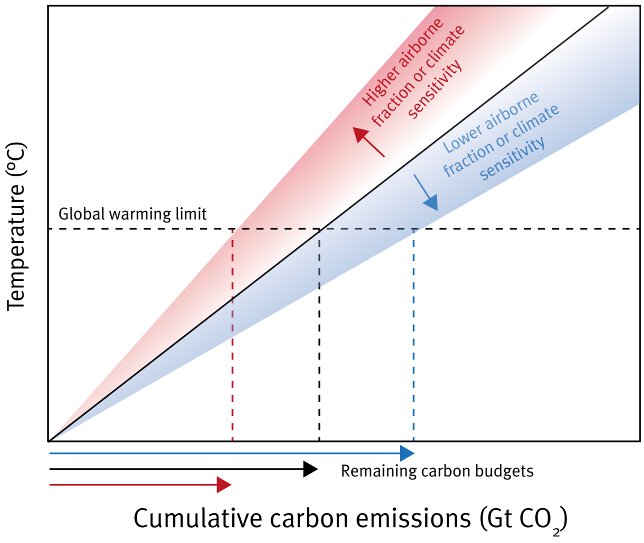

Figure 4: Schematic of the relationship between global mean temperature, cumulative carbon emissions, and the remaining carbon budget for an arbitrary global warming limit. The thick black line is the best estimate of the relationship between temperature and cumulative emissions, while the grey shading represents the uncertainty. Thick coloured arrows indicate the qualitative impact of an increase (red) or decrease (blue) in airborne fraction or climate sensitivity. Thin coloured dashed lines and arrows represent the remaining carbon budgets for cases of high (red), intermediate (black) and low (blue) airborne fraction or climate sensitivity.

Carbon budgets

Current climate policy is shaped by the limits set out in the 2015 Paris Agreement, whereby the international community aims to hold the increase in global temperature to well below 2° C, and aspires to stay below 1.5° C of global warming.
To inform policy decisions consistent with these goals, we need to know how much global warming would result from a given amount of carbon emissions. Insights from climate model experiments suggest a nearly linear relationship between global warming and cumulative carbon emissions (Fig. 4). This linear relationship means that we can easily estimate how much more carbon can be emitted before a given global warming limit is exceeded. This amount of allowable future emissions is called a **carbon budget.**

The relationship between global warming and cumulative carbon emissions depends mainly on two factors[18,19]: (1) how much of the emitted carbon remains in the atmosphere as CO_2 (as opposed to being stored in the ocean or in the land); (2) how much warming occurs for a given increase in CO_2. Point (1) is sometimes termed the "airborne fraction", while point (2) is related to the climate sensitivity[ii]. High airborne fraction implies that for a given amount of emissions, more CO_2 remains in the atmosphere to cause warming. High climate sensitivity means that for a given amount of atmospheric CO_2, the planet warms more. This means that high values of either airborne fraction or climate sensitivity lead to a smaller carbon budget for a given warming limit, and vice versa (Fig. 4, arrows).

Uncertainty in the relationship between cumulative emissions and global warming means the carbon budget is actually a probability distribution, and published values refer to a specific probability (typically 50% or 66%) of exceeding a given global warming value. While the probability distribution of the carbon budget reflects the uncertainties in airborne fraction and climate sensitivity, with the latter dominating[20], it also accounts for a range of additional factors[21]. These factors include uncertain non-CO_2 forcings, possible unrepresented carbon feedbacks (e.g. the release of methane by melting permafrost), and uncertain historical warming and carbon emissions.

[ii] Strictly speaking, point (2) is quantified by the *transient climate response*, a measure of near-term global warming, before the climate system has reached equilibrium. For simplicity we ignore this distinction in the discussion below.

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