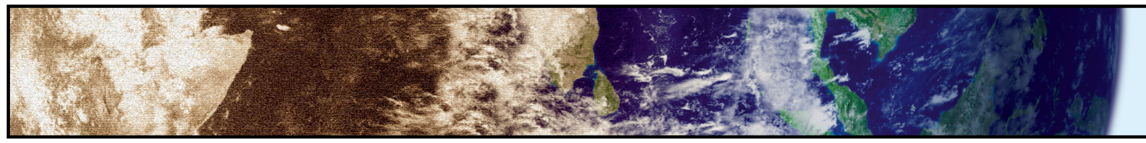

Climate sensitivity - How sensitive is Earth's climate to CO_2?

PRESENT PAST

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The Climate Sensitivity (CS) is a key parameter for assessing future climate change, as is its counterpart the Transient Climate Response (TCR – see figure 1). The TCR is defined as the 20-year global, annual mean temperature change averaged around the time of CO_2 doubling, under a forcing scenario of CO_2 increasing at a rate of 1% per year compounded. Just like the CS, the TCR quantifies physical feedbacks in the climate system associated with the surface, clouds, water vapor, sea ice, etc., but it is more relevant for transient climate change in the near future and does not suffer from the "long tail" evident in estimates of the CS (Frame et al. 2005).

A notable feature of the latest version of the Coupled Model Intercomparison Project (CMIP5) is that atmosphere models coupled to simple slab or mixed-layer oceans are not included in the design, limiting a direct comparison of the range of CSs with previous versions of CMIP (although the CS and TCR tend to be well correlated in models and the effective CS can be calculated from experiments included in CMIP5).

We may use modern observations to aid in building complex models of the climate system from "first principles" i.e. by solving the dynamical equations of the atmosphere and ocean and parameterizing sub-grid-scale processes in as much detail as possible. Multiple data sources may be used to evaluate both the individual building blocks and the emergent properties of the model; data from process-based observations, possibly gathered during dedicated field campaigns, historical in situ measurements, remotely sensed data, etc. We can then interpret measures such as the CS and TCR computed from complex modes as estimates that integrate our understanding of climate (embodied in the laws of physics) and modern day observations.

The range of CS and TCR has not changed much in successive generations of models. The example in the figure shows a range of 1.2-2.6ºC for the CMIP3 models and 1.3-2.4ºC for the CMIP5 models available at the time of writing.

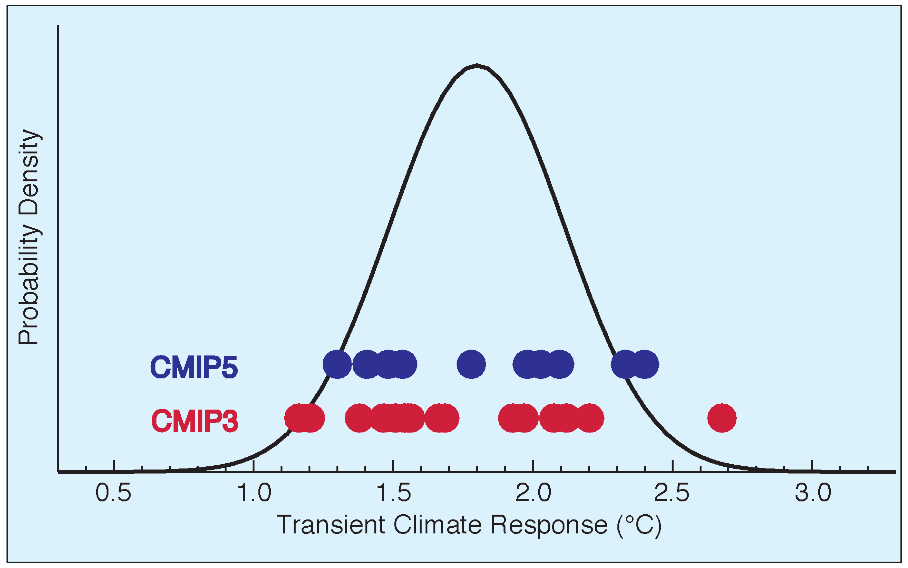

Figure 1: A comparison of estimates of the Transient Climate Response from complex models evaluated against modern observations from versions 3 and 5 of the Coupled Model Intercomparison Project (Meehl et al. 2007; Taylor et al. 2011) shown as red and blue dots respectively. A PDF of the TCR computed from a simple model with parameters constrained by observations is also shown (Gregory and Forster 2008).

An alternative approach comes from using simple climate models that may only simulate aggregate variables such as global mean temperature. Simple models can be run many times and statistical approaches can be used to formally estimate the parameters of the model based on constraints from observations/estimates of e.g. recent ocean heat uptake and radiative forcing. Measures such as CS and TCR then come with likelihood estimates and the uncertainty may be expressed as a probability density function (PDF – see Fig. 1).

Unfortunately, using different observational data sources from different modern (and paleo) time periods, have not produced tight constraints on variables such as the TCR. The 5-95% range in the example from the figure from (Gregory and Forster 2008) is 1.3-2.3ºC, comparable with the ad hoc range from CMIPs. CMIP ranges of CS are also comparable with observationally constrained PDFs (Knutti and Hegerl 2008).

As the signal of climate change emerges from the noise of natural variability, PDFs based on simple-model constraints should narrow. Collection of new and more detailed modern observations, particularly of climate processes such as clouds, should allow us to better improve and evaluate our complex models. One recent approach combines complex modeling with formal parameter estimation to produce PDFs of global and regional change (Sexton et al., in press). This allows multiple modern observational records to be used to constrain projections, although the cost of implementation is high. There is still scope for much research in quantifying how sensitive Earth's climate is to CO_2 change using modern models and data.

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The climate record definitively shows that the Earth's climate is sensitive to various drivers, including greenhouse gases, orbital variations and continental shifts. Unfortunately, there are no past analogs for the anticipated 21st Century climate changes, and so a principal challenge in applying these constraints to the future is to interpret these changes quantitatively.

At the global scale, the framework of radiative forcing and response is a powerful method to constrain sensitivity, however, there are many nuances. First, the system being described needs to be defined - what are the forcings, and what are the responses? This might seem clear at first glance, but actually depends on the availability of data and what timescales are being considered (Fig. 1). Second, there needs to be clarity in how the calculated sensitivity relates to either the "climate sensitivity" determined by models, or the related concept of the transient climate response.

The commonly used "Charney sensitivity" - the equilibrium surface temperature response to $2xCO_2$ allowing most atmospheric processes to react, but holding ice sheets, vegetation, atmospheric composition and ocean circulation constant - is a useful climate model metric. Constraining this from paleo-data requires information on all the "constant" components, most notably for the Last Glacial Maximum (LGM), where many (though not all) the elements are available (Köhler et al. 2010; Schmittner et al. 2011), and perhaps the last millennium, where many aspects are not fundamentally different from today (Hegerl et al. 2006). However, while the Charney sensitivity is a useful characterization of the any particular atmospheric model, it is not the same as what would actually occur if $2xCO_2$ were reached and maintained for a long time.

There are important nuances: climate sensitivity to cooler conditions might not be equivalent to climate sensitivity to warmer ones (indeed evidence suggests it is 80 to 90% smaller; Hargreaves et al. 2007; Crucifix 2006; Hansen et al. 2005) and some forcings just can't be fitted into a global forcing/response framework at all (such as orbital variations). Furthermore, there is often substantial uncertainty in the forcings - whether it is the size of ice sheets at the LGM, or solar forcing in medieval times, that must be taken into account in assessing the uncertainties in any estimates.

Constraining any sensitivities from the paleo-record is thus still a work in progress. Predominantly data-driven approaches (like Köhler et al. 2010 or Lorius et al. 1990, for the LGM) suggest a Charney sensitivity of around 3ºC (with a 2σ range of ~1-5ºC). Synthesis estimates that use a combination of intermediate models constrained by LGM paleo-data have given ranges of 1.2-4.3ºC (5-95%) (Schneider von Deimling et al. 2006) and 1.7-2.6ºC (17-83% range) (Schmittner et al. 2011). Note however, that the latter estimate includes a vegetation feedback, not included in the standard definition of the Charney sensitivity. A correction for this reduces the estimated sensitivity by about 0.2ºC. There is a large (and as yet barely quantified) sensitivity to model structure in these calculations since the models used to date do not give a very good fit to the regional details of the proxy data.

By expanding the framework to incorporate excluded fast and slow feedback elements, it is possible to estimate the long-term "Earth System Sensitivity" (ESS) (Lunt et al. 2010; Hansen et al. 2008), i.e. the temperature realized after all the feedbacks have worked themselves out. For instance, Lunt et al. (2010) found that the addition of ice sheet and vegetation responses (derived from Pliocene proxy data), increased their model sensitivity to CO_2 by ~50%. However, this will apply only at very long timescales (many tens of thousands of years or even longer). Intermediate definitions of the sensitivity might also be calculated - for instance, taking dust, aerosol and ozone changes (fast atmospheric responses) or ocean circulation changes as feedbacks as well, but still holding ice sheets and vegetation constant.

Linking estimates of climate drivers in the past, estimates of the climate response, and the prospects for future change is however a crucial task (Schmidt 2010). To a large extent it requires the use of climate models, and the incorporation of a paleo-climate modeling component in CMIP5 will serve as a good testbed for using the paleo-record to assess the credibility of many aspects of the future projections (not simply the global mean temperature sensitivity).

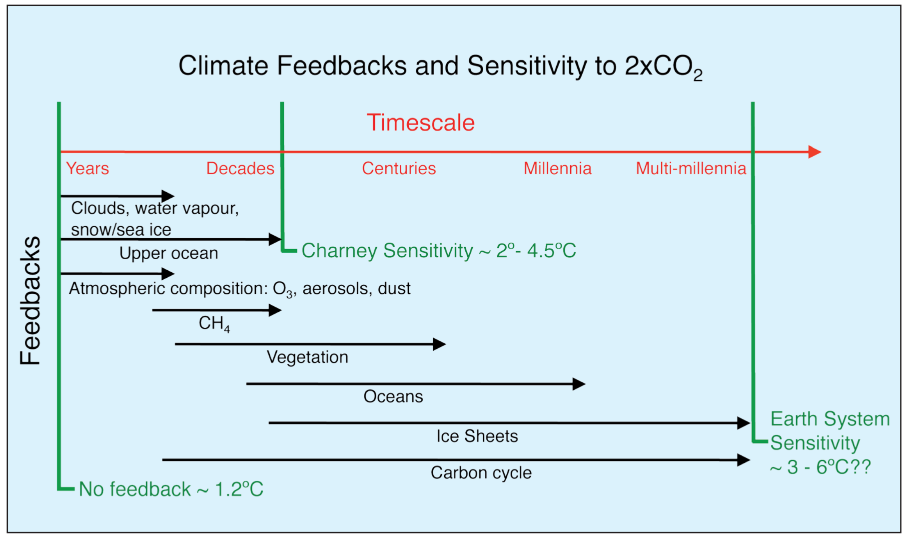

Figure 1: Climate sensitivities are a function of what feedbacks are included and what timescales are being considered.

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