

# Complementary observational constraints on climate sensitivity

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A persistent feature of empirical climate sensitivity estimates is their heavy tailed probability distribution indicating a sizeable probability of high sensitivities. Previous studies make general claims that this upper heavy tail is an unavoidable feature of (i) the Earth system, or of (ii) limitations in our observational capabilities.

Here we show that reducing the uncertainty about (i) oceanic heat uptake and (ii) aerosol climate forcing can — in principle — cut off this heavy upper tail of climate sensitivity estimates. Observations of oceanic heat uptake result in a negatively correlated joint likelihood function of climate sensitivity and ocean vertical diffusivity. This correlation is opposite to the positive correlation resulting from observations of surface air temperatures. As a result, the two observational constraints can rule out complementary regions in the climate sensitivity-vertical diffusivity space, and cut off the heavy upper tail of the marginal climate sensitivity estimate.

## 1. Introduction

Most current observational estimates of climate sensitivity (the projected equilibrium temperature change for a doubling of carbon dioxide concentration) produce positively skewed probability density functions (pdfs) [Meehl *et al.*, 2007]. This form of pdf implies that rather high climate sensitivity values (e.g., above 9 K) still have a sizeable probability. Whether the true climate sensitivity is indeed located in this “heavy upper tail” [Forest *et al.*, 2002; Tol and de Vos, 1998; Royer *et al.*, 2007] of current estimates is a question of considerable policy relevance [Keller *et al.*, 2004; Weitzman, 2007]. This is because the projected climate change impacts typically rise in a highly nonlinear function with the climate sensitivity [Yohe *et al.*, 2004; Nordhaus, 2008].

The uncertainty in climate sensitivity arises primarily from uncertainty in the dynamical feedbacks which act on external radiative forcing, such as water vapor, ice albedo, and cloud feedbacks [Bony *et al.*, 2006]. Estimates based on surface temperature alone are confounded by the uncertain transient response due to oceanic thermal inertia [Hansen *et al.*, 1985]. Several studies have shown how combining different types of observational constraints (e.g., global surface temperature and ocean heat time series) can reduce the uncertainty about the climate sensitivity [Forest *et al.*, 2002; Knutti *et al.*, 2002; Gregory *et al.*, 2002]. The inclusion of more observations is generally expected to reduce uncertainty, but as we will discuss, surface temperature and ocean heat in particular are complementary to each other when estimating the uncertainty in climate sensitivity.

Roe and Baker [2007] claims that the heavy tail of climate sensitivity is “an inevitable and general consequence of the nature of the climate system”. The logic behind this reasoning is that a combination of many uncertain feedback factors approaches, via the central limit theorem, a normally distributed total feedback factor. This normally distributed prior for the feedback results in, via the nonlinear relationship between feedback and climate sensitivity, a heavy upper tail of climate sensitivity estimates. However, this reasoning neglects the additional constraint on the total feedback provided by observations of the overall system response. Allen *et al.* [2006] reviews the problem from an observational perspective, citing uncertainties in the observed forcing and energy imbalance as limiting factors, and concludes that high values of equilibrium warming “are not and, for many data sources, cannot be excluded by the comparison of models with observations”.

Here we analyze (i) the mechanisms causing the heavy upper tail and (ii) how observations could test whether the climate sensitivity is indeed located in this region. We show that combining high precision ocean heat observations with surface temperature can — in principle — cut off the heavy tail of climate sensitivity estimates. This is because the two constraints produce a joint likelihood function of climate sensitivity and vertical diffusivity which is mutually exclusive of either very low or very high climate sensitivity estimates. In a nutshell, a tail of the climate sensitivity estimate which is allowed by one observational constraint is excluded by the other constraint.

## 2. Methodology

We use perfect model experiments to demonstrate that the heavy tail of climate sensitivity estimates depends on the nature and quality of the considered observational constraints. We perform two experiments. In a first experiment, we map the relationship between the transient warming (which is the observational constraint) and the equilibrium climate sensitivity (which is the inferred parameter). Specifically, we calculate the response of surface temperature and ocean heat anomalies as a function of climate sensitivity, using a range of assumptions about vertical ocean diffusivity. In a second experiment, we perform a perfect model assimilation to estimate the climate sensitivity from hypothetical observations of changes in surface air temperatures and oceanic heat anomalies. Following Forest *et al.* [2006], we consider uncertainty in the climate sensitivity and vertical diffusivity. To demonstrate our general point about the effect of parameter correlations on the heavy upper tail of climate sensitivity, we neglect uncertainty in aerosol forcing. We return to this assumption in the Discussion section.

Using the DOECLIM climate model [Kriegler, 2005; Tanaka *et al.*, 2007], a zero dimensional energy balance model coupled to a one dimensional diffusive ocean, we calculate the global surface temperature anomaly and ocean heat anomaly using historical forcings from 1750 to 2000 [Tanaka *et al.*, 2007]. We assume a climate sensitivity of

3.26 K and a vertical diffusivity of  $0.55 \text{ cm}^2 \text{ s}^{-1}$ , obtained by the fit to observational data reported in *Kriegler* [2005]. We consider a surface temperature observation to be the year 2000 temperature relative to the 1961–1990 average, and an ocean heat observation to be the 1957–1996 global aggregate oceanic heat uptake.

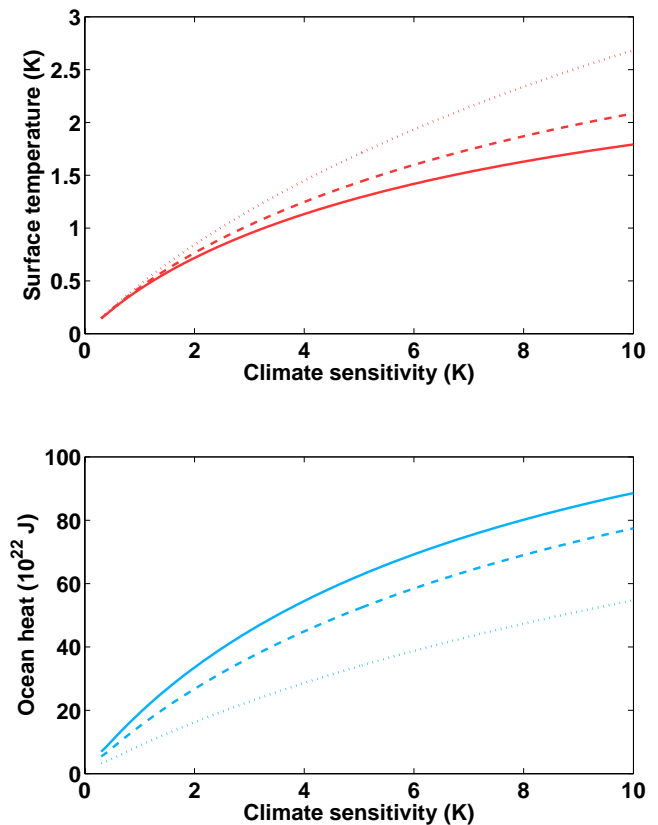
To simulate observational error we superimpose Gaussian noise with standard deviations 0.05 K and  $4 \times 10^{22} \text{ J}$ , respectively. (For visual clarity, we did not superimpose this noise in Figure 2, so that the simulated observations are the model predictions and the joint likelihood is centered on the assumed parameter values. Assimilation experiments with different noise realizations indicate that our conclusions are robust to observational errors.) We calculate the likelihood of the simulated observations over a regular grid of climate sensitivities and vertical diffusivities. The observation errors of the two considered quantities are assumed to be independent, so their joint likelihood is the product of their individual likelihoods.

To obtain the joint posterior probability density of the estimated parameters, we multiply the joint likelihood by priors uniform between 0 and 10 K for climate sensitivity and between 0 and  $4 \text{ cm}^2 \text{ s}^{-1}$  for vertical diffusivity, and normalize the probability density function. Integrating this over vertical diffusivity gives the marginal posterior probability density of climate sensitivity.

### 3. Analysis

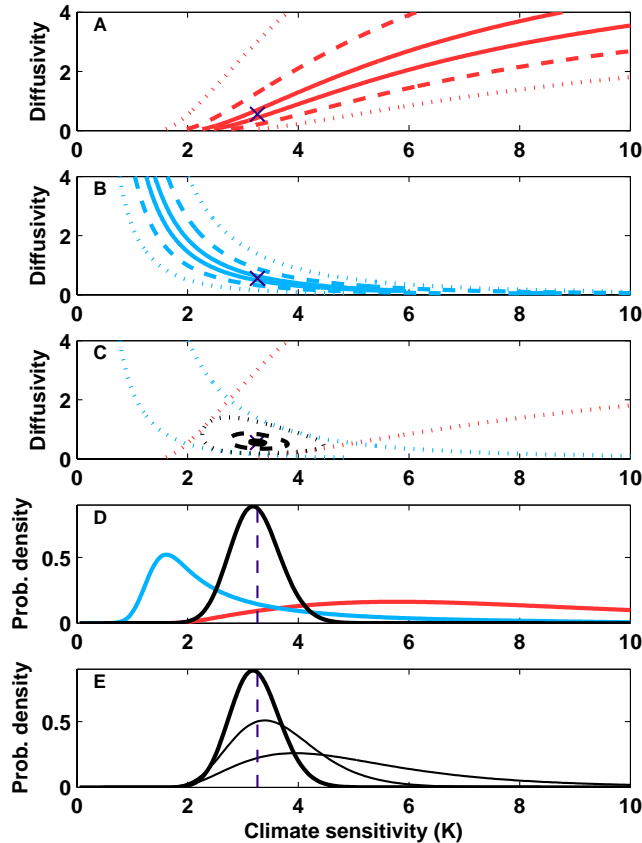
From an observational perspective, the heavy tail of climate sensitivity arises because of nonlinearity between the transient and equilibrium temperature responses (Fig. 1) [cf. *Hansen et al.*, 1985, which makes this point for the surface temperature constraint]. The transient response is what is observed, while the equilibrium response is what is estimated. If the two are nonlinearly related, a Gaussian uncertainty in the observed response produces a non-Gaussian uncertainty in the implied equilibrium climate sensitivity. The heavy upper tail of the climate sensitivity pdf appears because a small increase in the transient response corresponds to a large increase in the inferred climate sensitivity. A

small error in measurement is thereby amplified into a large uncertainty in estimated sensitivity.



**Figure 1.** Transient climate response as a function of climate sensitivity, at fixed vertical ocean diffusivities  $K_V = 0.1$  (dotted),  $0.5$  (dashed), and  $1 \text{ cm}^2 \text{ s}^{-1}$  (solid). *Upper:* Global mean surface temperature anomaly in the year 2000. *Lower:* Global ocean heat anomaly (1957–1996). Note that the response has negative curvature for either type of observation, but that surface temperature and ocean heat have opposite responses to changes in diffusivity.

The nonlinearity between observations and climate sensitivity occurs for both surface temperature and ocean heat observations (note the negative curvature of all curves in Fig. 1), leading to a heavy upper tail in the climate sensitivity pdf for either type of observation considered alone (Fig. 2, panel D, red and blue curves). A possibly surprising conclusion, however, is that these two heavy-tailed distributions can combine to produce a climate sensitivity pdf without the heavy tail (panel D, black curve). The resolution of this paradox lies in the fact that uncertainty in climate sensitivity is also influenced by the uncertain rate of ocean heat uptake, and that surface temperature and ocean heat observations rule out complementary regions of the sensitivity-diffusivity parameter space.



**Figure 2.** *Panels A, B, and C:* Joint likelihood functions for climate sensitivity (CS) and vertical ocean diffusivity ( $K_V$ ). The likelihood in panel A assumes a hypothetical observation of the global mean surface temperature anomaly in the year 2000 (SAT). The likelihood in panel B assumes a hypothetical observation of global oceanic heat anomaly from 1957–1996 ( $\Delta H$ ). Panel C assumes both observations, with likelihoods from A and B superimposed for reference. The solid, dashed, and dotted lines are 5, 50, and 95 % of the maximum likelihood (occurring at the true parameter values, crosses). *Panel D:* The marginal likelihood for climate sensitivity assuming both observations. The blue, red, and black lines are for  $\Delta H$ , SAT, and the combined  $\Delta H$  and SAT observations, respectively. The dashed vertical line shows the true CS value. *Panel E:* The marginal likelihood for climate sensitivity due to increasing  $\Delta H$  observation errors from 4 (thick curve), 12, to  $24 \times 10^{22}$  J (thin curves).

Observations of changes in surface air temperatures result in positively correlated estimates of climate sensitivity and ocean diffusivity (Fig. 2, panel A). This is because the climate sensitivity signal in the observed warming is confounded by the amount of warming “in the pipeline” due to the lagged response of the oceans [Hansen *et al.*, 1985, 1997]. A given surface air temperature change is consistent with either a relatively large heating which is penetrating rapidly into the oceans and delaying some of the surface warming (i.e., a high climate sensitivity and a high ocean diffusivity), or a relatively small heating which is penetrating slowly into the oceans so the surface warming is quickly experienced (i.e., a low climate sensitivity and a low ocean diffusivity). This relation can also be seen in the upper panel of Fig. 1,

where for a fixed transient response, larger diffusivities correspond to larger sensitivities.

In contrast, observations of changes in oceanic heat content result in negatively correlated estimates of climate sensitivity and ocean diffusivity (panel B). This is because a given change in oceanic heat content is consistent with either a relatively large heating which is penetrating slowly into the oceans (i.e., a high climate sensitivity and a low ocean diffusivity) or a relatively small heating which is penetrating rapidly into the oceans (i.e., a low climate sensitivity and a high ocean diffusivity). This can also be seen in the lower panel of Fig. 1, where for a fixed transient response, larger diffusivities correspond to smaller sensitivities.

Considered separately, the information contained in (i) the changes in surface air temperatures and (ii) the changes in oceanic heat content result in right skewed estimates of climate sensitivity (panel D, red and blue lines, respectively). However, combining these two lines of evidence results in an approximately symmetric pdf for the climate sensitivity (panel D, black line). Combining the information contained in atmospheric and oceanic observations can hence, in principle, cut off the heavy upper tail of climate sensitivity estimates. This effect is due to the interactions between the two constraints. Specifically, high climate sensitivity estimates that would be consistent with the surface air temperature constraint (panel A) are inconsistent with the oceanic heat constraint (panel B). Likewise, low climate sensitivity estimates that would be consistent with the oceanic heat constraint (panel B) are inconsistent with the surface air temperature constraint (panel A). The joint posterior for the combined observations lies in the mutually consistent region where the posteriors for each individual observation overlap (panel C, black curves).

## 4. Discussion

The heavy upper tail of climate sensitivity can be reduced when both observational constraints are combined. This tail thinning occurs because the high sensitivities allowed by the temperature constraint are disfavored by the ocean heat constraint, and vice versa. Surface temperature observations permit high climate sensitivities if there is substantial unrealized “warming in the pipeline” from the oceans. However, complementary ocean heat observations can be used to test this and can potentially rule out large ocean warming. Ocean heat observations are compatible with high sensitivities if there is substantial surface warming which is penetrating poorly into the oceans. Again, complementary surface temperature observations can test this, and can potentially rule out large surface warming. High climate sensitivities can thus be excluded when both observational constraints are jointly applied. In parameter space, these complementary constraints take the form of (i) a positive correlation between climate sensitivity and ocean vertical diffusivity implied by surface temperature observations, and (ii) an opposite, negative correlation implied by ocean heat observations (Fig. 1; Fig. 2, panels A & B).

## 5. Caveats

It is important to stress that our simple analysis neglects several likely important processes and sources of uncertainty. For example, our analysis neglects uncertainties about historic climate forcings [Forest *et al.*, 2002]. Since direct and indirect aerosol forcings are significant and their combined uncertainty is comparable in size to the forcing itself [Alley *et al.*, 2007], treatment of this uncertainty will widen the tails of the climate sensitivity pdf [Allen *et al.*, 2006]. The

specific conclusions shown in Figs. 1 and 2 are hence contingent on (i) the chosen model structure and (ii) a case where the forcing uncertainty has been reduced to negligible levels. Future research with more realistic models (e.g., Weaver *et al.* [2001]) and considering forcing uncertainty (cf. Forest *et al.* [2002]) is needed to translate our general findings into specific recommendations for the design of observation systems.

## 6. Conclusions

The observational constraints considered in published studies [Meehl *et al.*, 2007] provide little guidance to decide the question whether the real climate sensitivity is indeed located in the heavy upper tail. In our analysis, cutting off the heavy tail required (i) neglecting model structural uncertainty, (ii) a reduction of the aerosol forcing uncertainty to negligible levels, and (iii) reduced uncertainty in oceanic heat uptake. Our analysis suggests that one promising avenue to decide whether the true climate sensitivity is indeed located in the heavy upper tail of current estimates is through improving the skill of the existing ocean observation system to estimate the anthropogenic heat uptake [AchutaRao *et al.*, 2006; Levitus *et al.*, 2005]. The skewness of the climate sensitivity estimate in our analysis hinges critically on the observation error of the oceanic heat content observations: reducing the uncertainty about the oceanic heat uptake shortens the upper tail of the resulting climate sensitivity estimates (Fig. 2, panel E).

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