MIT Joint Program on the Science and Policy of Global Change

Quantifying Uncertainties in Climate System Properties using Recent Climate Observations

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To inform processes of policy development and implementation, climate change research needs to focus on improving the prediction of those variables that are most relevant to economic, social, and environmental effects. In turn, the greenhouse gas and atmospheric aerosol assumptions underlying climate analysis need to be related to the economic, technological, and political forces that drive emissions, and to the results of international agreements and mitigation. Further, assessments of possible societal and ecosystem impacts, and analysis of mitigation strategies, need to be based on realistic evaluation of the uncertainties of climate science.

This report is one of a series intended to communicate research results and improve public understanding of climate issues, thereby contributing to informed debate about the climate issue, the uncertainties, and the economic and social implications of policy alternatives. Titles in the Report Series to date are listed on the inside back cover.

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Abstract

We apply the optimal fingerprint detection algorithm to three independent diagnostics of the recent climate record and derive joint probability density distributions for three uncertain properties of the climate system. The three properties are climate sensitivity, the rate of heat uptake by the deep ocean, and the strength of the net aerosol forcing. Knowing the probability distribution for these properties is essential for quantifying uncertainty in projections of climate change. We briefly describe each diagnostic and indicate its role in constraining these properties. Based on the marginal probability distributions, the 5-95% confidence intervals are 1.4-7.7 K for climate sensitivity and 0.30-0.95 W/m^2 for the net aerosol forcing using uniform priors; and 1.3-4.2 K and 0.26-0.88 W/m^2 using an expert prior for climate sensitivity. The oceanic heat uptake is not so well constrained. The uncertainty in the *net* aerosol forcing in either case is much less than the uncertainty range usually quoted for the indirect aerosol forcing alone.

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Estimation of uncertainty for long-term climate change requires estimates for the probability distribution functions (pdf) of key properties of the climate system. Attempts thus far [1, 2] have used pdfs based on expert judgment to analyze such uncertainty. For near-term climate change, recent studies [3] have applied the uncertainty estimates derived from the climate change detection algorithm for particular models to climate projections based on these models. A key restriction in this approach is that both forcing and response do not change qualitatively in the transition from observed to forecast periods. Hence it is not applicable to modelled climate change under scenarios which differ significantly from the recent past (e.g., stabilization cases or severe changes in sulfur emissions). Given the political priority to establish what constitutes a "safe" stabilization level for greenhouse gases, an objective means of quantifying uncertainty in the long-term response, despite uncertainty in other forcings, is clearly desirable.

To get around these problems, one must determine both the range of climate system properties and the range of forcings that produce simulations consistent with twentieth century climate change [4, 5]. To determine such ranges, we use the MIT 2D (zonal mean) statistical-dynamical model [6] to simulate the twentieth century climate record and systematically vary the uncertain model properties and forcings to assess which simulations "match" the observed climate record. The two most uncertain properties that control the climate system's decadal to century response to radiative forcings are climate sensitivity (S) and the rate of heat uptake by the deep ocean (K_v). [1] Simulations by modern atmosphere-ocean general circulation models (AOGCMs) reveal significant differences in these properties between models [5]. Previous estimates of the uncertainties in these quantities have generally been based only on expert judgment and/or on the range of values found in AOGCMs. A recent exception [8] has estimated the range of climate sensitivity from observations, but without considering the uncertain heat uptake. Although positive AOGCM climate change detection results place a lower bound on climate sensitivity [5, 3],

[1] Here we define climate sensitivity (S) as the equilibrium global mean temperature change in response to a doubling of CO_2 concentration. We measure the rate of heat uptake by the deep ocean by an effective diffusivity of heat anomalies (K_v) into the ocean below the climatological mixed layer. For AOGCMs, this effective diffusivity can be derived from transient climate change experiments and should not be confused with the model's sub-grid scale diffusion coefficient. We note that S and K_v are closely related to the climate sensitivity and transient climate response factors used in ref. [7] to represent both the short and long term behavior of AOGCMs.

the upper bound on S depends strongly on the rate of ocean heat uptake [5, 3]. The MIT climate model has the flexibility to vary both S and K_v, unlike AOGCMs (see ref. [6]).

The primary uncertainty in the radiative forcing is the total uncertainty in anthropogenic aerosol forcing which arises from the uncertainty in aerosol radiative properties and cloud effects as well as in their concentrations over the industrial period [9]. [Uncertainty in the natural forcings (primarily solar and volcanic forcings) exists although the estimated changes during the 20th century appear to be small compared to the uncertainty in the aerosol forcing [9].] Here we extend the analysis in refs. [4, 5] by including the strength of the anthropogenic aerosol forcing as a third major uncertainty. We measure this forcing by the net forcing (both direct and indirect) for the decade of the 1980s (F_{aer}).

We apply the optimal fingerprint detection algorithm [10, 11, 12] to three independent diagnostics of the recent climate record and to ensembles of climate simulations with the MIT climate model covering the period 1860-1995. By comparing them, we derive a joint probability density function (pdf) for the three uncertain properties of the climate system defined above: S, K_v, and F_{aer}. These properties jointly determine the large-scale response of the climate system to changes in the radiative forcing. Knowing the probability distribution for these properties, we can then determine the likelihood of the climate system response to individual forcing scenarios [13].

The three diagnostics are derived from the upper-air temperature record [14], the surface temperature record [15, 16], and the record of ocean temperatures [17]. As in refs. [4, 5], we use the same upper-air temperature diagnostic as in refs. [18, 12]. The temperature changes are computed for points on a latitude-height grid as the difference in the 1986-1995 and 1961-80 zonal means. The years 1963-4 and 1992 were removed to limit the effect of the Mt. Agung and Mt. Pinatubo eruptions on the estimated temperature changes.

We construct a surface temperature diagnostic by computing decadal mean temperature anomalies for the 1946-1995 periods with respect to a 1906-1995 climatology from the surface temperature record of [15, 16]. Using an observational data mask, we compute area-weighted zonal averages for the temperature anomalies over 4 zonal bands (90S-30S, 30S-0, 0-30N, 30N-90N) and create a latitude-time pattern of temperature change. By using a longer climatology, we make use of the additional information that the most recent 50 years were warmer than the previous 40 (see refs. [3, 19]). For both the surface and upper-air temperature diagnostics, the observational errors are small on such spatial scales [20] and are neglected in this analysis.

Our third climate change diagnostic is the trend in global-average ocean temperatures down to 3000m depth computed from data in ref. [17]. We compute a trend from a 44-point time series of 5-year averages using the 1948-1995 period. Ordinary least squares regression is used to compute the trend with the observational errors providing uncertainty in the estimated trend. Because the data are sparsely distributed across ocean basins, the global average will have uncertainty due to this sampling error. One attempt to include this is provided [17] but the true uncertainty remains unknown. This observational error was then added to the climate noise estimate discussed next.

Each diagnostic is used to compute a goodness-of-fit statistic, r^2, which is inversely weighted by the size of the deviations that we should expect from climate noise [4, 5]. For the surface and upper-air diagnostics, the climate noise estimates were taken from successive segments of the control simulation of the Hadley Centre's second generation coupled atmosphere-ocean general circulation model (HadCM2) [21]. For the ocean diagnostic, we used a control simulation of the Geophysical Fluid Dynamics Laboratory (GFDL) R30 model [22] in addition to the observational error. From the r^2 statistics, a likelihood is computed to provide a joint probability distribution for the model parameter space given the independent observations [4, 5]. This likelihood estimate represents the probability that a particular choice of model parameters is correct given the observed record of climate change. More formally, we reject a choice of model parameters, Π ($\Pi = \{S, K_v, F_{aer}\}$) as producing a simulation of the twentieth century that is inconsistent with observed climate given the unforced variability of the climate as estimated by the HadCM2 and GFDL R30 models at some level of significance.

The distribution of the r^2 statistic for a given diagnostic indicates the diagnostic's importance for rejecting particular regions of parameter space (Fig. 1). To interpret the r^2 figures, we first choose a level of significance and then calculate the Δr^2 value for rejection where Δr^2 is the difference from the minimum r^2 value. This means that if Δr^2 is larger than this cutoff value, we reject the hypothesis that this Δr^2 value could arise from unforced variability of the climate system. This criterion designates a region of parameter space as being rejected at the given significance level. The rejection regions were estimated using $\Delta r^2 \sim mF_{m,n}$ where $F_{m,n}$ is the F-statistic with m and n degrees of freedom, m is the number of constrained model properties, and n is the degrees of freedom in the control simulations.

In general, the combination of lower oceanic heat uptake, higher climate sensitivity, and weaker aerosol cooling will provide a simulation with a larger change in surface and upper-air temperatures. For the deep ocean, however, a larger warming will occur for stronger oceanic heat uptake. For low climate sensitivity, the r^2 statistics for both S and K_v for a given aerosol forcing show little variation with oceanic heat uptake. Thus, different diagnostics provide constraints in the different regions of parameter space. The surface and upper-air diagnostics reject similar regions of parameter space, namely low K_v and high S, while the ocean diagnostic shows a rejection of the high K_v and high S region. When aerosol cooling is increased (decreased) (not shown for each diagnostic), the rejection regions shift toward the higher (lower) response regions indicating that a higher sensitivity is required to reproduce the observed temperature changes.

Because each individual diagnostic provides the likelihood that the modeled temperature change is correct given a set of model parameters (or in Bayesian notation, $p(\Delta T|\Pi)$), we use results from each diagnostic to update the probability distribution $p(\Pi|\Delta T)$ by applying Bayes' Theorem [23].[2] By taking one of the distributions as the initial prior, two sequential posterior distributions are computed with the final distribution representing the combined uncertainty from the three climate change diagnostics. If desired, a prior distribution based on expert judgment can be used initially. In the absence of an explicit expert prior, we assume a uniform probability distribution as the first prior. [3]

The combined probability distribution (Fig. 2) resulting from the Bayesian updating procedure shows cross-sections of the three-dimensional pdf at six aerosol forcing strengths. The constraints on climate sensitivity and aerosol forcing are fairly strong, while

[2]To compute the updated distributions for $p(\Pi|\Delta T)$, we first interpolate the r^2 values onto a finer grid (ΔS = 0.1K, $\Delta(K_v^{1/2})$ = 0.1 cm/s$^{1/2}$) using a thin plate spline algorithm over the range: S = 0.5–10. K and K_v = 0.2–64. cm^2/s. After the data were interpolated in the S-K_v plane, a cubic spline interpolation was used to interpolate between the aerosol forcing levels to a resolution of ΔF_{aer} = 0.05 W/m^2 over the range 0-1.5 W/m^2. After the r^2 values were interpolated, the probability of rejection was estimated to generate the probability distribution on the finer grid spacing. For all integral estimates of total probability or marginal probability distributions, the finer grid spacing was used over the ranges defined.

[3]The ranges explored for the model parameters set the limits of the uniform priors: S = 0.5-10.K, K_v = 0.2-64 cm^2/s, and F_{aer} = 0-1.5 W/m^2. We can assess the impact of these priors by examining the posterior distributions.

oceanic heat uptake is less constrained. Recent papers [24, 25] have examined the ocean temperature trends as a function of basin which may better constrain the oceanic heat uptake. We also integrate the three dimensional pdf to obtain marginal or one-dimensional pdfs for each model parameter (Fig. 3 and Table 1). For the marginal pdfs, we also show the result if an expert prior from ref. [26] is used for climate sensitivity while keeping the uniform priors on K_v and F_{aer}. This demonstrates that including expert judgment will alter the shape of the marginal pdf for climate sensitivity and changes the net aerosol forcing pdf to compensate for the reduced climate sensitivities. While expert judgments are in part subjective, they can take into account information not included in our analysis, such as paleoclimate data.

Summary of Results.

1. The joint probability distribution for S, K_v, and F_{aer} shows stronger constraints on model properties than our earlier results [4, 5]. This is because of two improvements in our analysis: aerosol-forcing uncertainty is now considered explicitly and the additional diagnostics place limits on previously unconstrained regions of parameter space (high climate sensitivity and high ocean heat uptake).

2. With uniform priors, the strength of the net aerosol forcing lies within 0.25–0.98 W/m^2 for the 5-95% confidence range. The main diagnostic constraining this forcing is the surface temperature record. We stress that the constrained quantity in this case is the *net* non-greenhouse gas forcing. We note that this uncertainty range is much smaller than the uncertainty range given by the IPCC [9] for the indirect aerosol forcing alone.

3. For the effective ocean diffusivity, the 5-95% confidence range is 0.15-56.0 cm^2/s. Although the lower bound is well constrained by the observations (see Fig. 1), we note that the estimated probabilities will be affected by the assumed prior. This large uncertainty in K_v is much greater than that usually assumed [9, 2]. Ocean heat uptake remains one of the least understood large-scale processes in climate change studies. Although many mechanisms are known to affect heat uptake in the ocean, the sensitivity of the global heat uptake to changes in model parameterizations for these mechanisms is poorly understood [27]. Our result suggests that more research is required.

4. With uniform priors, the 5% to 95% confidence range for the climate sensitivity is estimated to be 1.2-8.5 K. We estimate the probability of S being outside the IPCC's range of 1.5-4.5 K [9] to be 30% with a 23% chance of exceeding 4.5 K. With an expert prior

applied to climate sensitivity and uniform priors elsewhere, climate sensitivity is estimated to be within 1.3-4.2 K for the 5-95% range and the probability of S being outside the IPCC's range increases to 12% with a 3.5% chance of exceeding 4.5 K.

Although the estimated pdf is consistent with the r^2 distribution, there will always be a non-zero probability outside the explored region (as indicated by the non-zero pdf of climate sensitivity at 10.0 K). We could fit specific theoretical distributions and use these to calculate the probability of the tail regions. For example, a range of theoretical distribution families fit to the climate sensitivity pdf indicate roughly a 3.5% chance that $S > 10$ K. However, these distributions all have infinite tails. Alternatively, we could assume that the probability outside the region explored is negligible.

Our estimates of probability are independent of the MIT 2D model to the extent that the MIT 2D model represents the large-scale behavior of different 3D models (or the climate system). Based on comparisons of the transient behavior (including the diagnostics presented here) under various forcing scenarios [6], the model behavior matches AOGCMs well for 100-150 year simulations. The MIT model cannot simulate some kinds of strong nonlinearity (e.g., the shutdown of the thermohaline circulation) but there is no indication of such behavior over the last 150 years. As was shown in ref. [6], the dependence of changes in different characteristics on surface warming for different versions of the MIT model is similar to that for the different AOGCMs. Because we have chosen to explore ranges of model parameters which extend beyond typical values of properties of existing AOGCMs, it is important to note that the MIT model produces similar dependencies for the range of climate sensitivity used in this study.

As described in the IPCC TAR [9], a long list of forcings can be identified for the industrial period (1750-present). We have included the three forcings (greenhouse gases, sulfate aerosols, and stratospheric ozone) which we expect to be most important for the diagnostics we have used. When considering the implications of the pdf for the aerosol forcing, we note that neglected forcings which have patterns similar to those of the sulfate aerosols, are implicitly included in the constraint we find for the aerosol forcing. This pattern is nearly invariant with longitude [28] adding further support for our using a zonal mean model. The most serious omission would be a forcing with a unique spatial pattern.

Among the forcing factors listed by the IPCC [9], the changes attributed to biomass burning, mineral dust, land-use change, and solar activity particularly have spatial distributions different from sulfate aerosols. These forcings combine to produce an estimated forcing of only -0.1 W/m^2. The forcings with patterns similar to sulfate aerosols (tropospheric ozone, sulfate, BC, OC, aerosol indirect effect) total an estimated -0.95 W/m^2. Thus, we doubt that the additional factors would significantly alter our results. We also note that the aerosol forcing is estimated for the 1980s and that this is with respect to equilibrium conditions (pre-1860 in our simulations). We have assumed that the pattern of the forcing [9] has not changed. Because we use the temperature record for 1906-1995, it is really only the forcings for this period which matter .

As discussed previously [5] the estimated natural variability, which is used to compute the noise covariance matrix in the detection algorithm, is obtained from two models, HadCM2 and GFDL R30, depending on the diagnostic. Thus, we have implicitly neglected the expected dependence of natural variability on climate sensitivity or ocean heat uptake [29].

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Table 1: Probability fractiles for climate system properties. Probabilities with uniform and expert priors on climate sensitivity based on ref. [26]. In both cases, uniform priors are used on K_v and F_{aer}.

Fractile	Uniform Priors			Expert Prior on S only		
	S (K)	K_v (cm^2/s)	F_{aer} (W/m^2)	S (K)	K_v (cm^2/s)	F_{aer} (W/m^2)
0.025	1.2	0.8	0.25	1.2	0.6	0.20
0.05	1.4	1.8	0.30	1.3	1.3	0.26
0.167	1.9	5.3	0.45	1.8	4.0	0.41
0.5	2.9	17.6	0.67	2.4	16.0	0.63
0.833	5.3	43.6	0.83	3.3	43.6	0.77
0.95	7.7	56.0	0.96	4.2	56.0	0.88
0.975	8.5	60.0	0.98	4.7	60.0	0.93
Mean	3.5	19.4	0.67	2.5	18.5	0.62

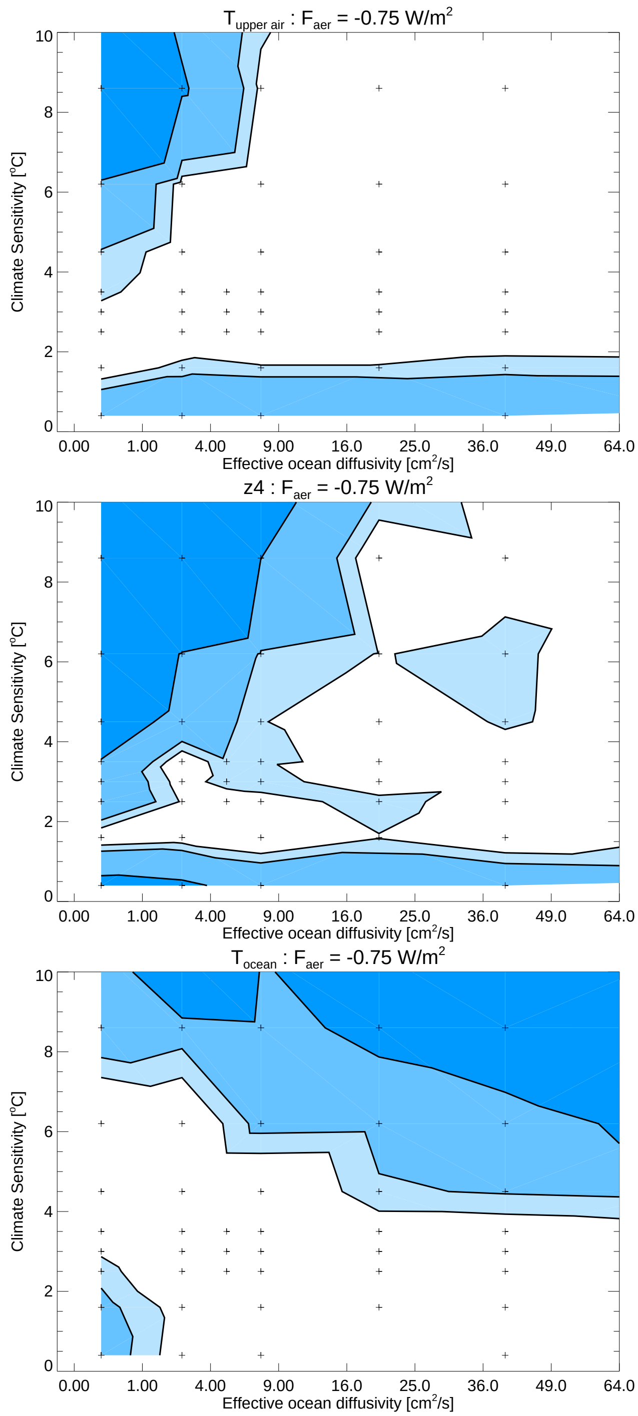

Figure 1: The distribution of r^2 for different diagnostics given a net aerosol forcing of -0.75 W/m^2: upper air (a), surface air (b), and deep ocean temperatures (c). The thick lines and shading represent the rejection region boundaries at the 20, 10, and 1 percent levels of significance (from light to dark, respectively.)

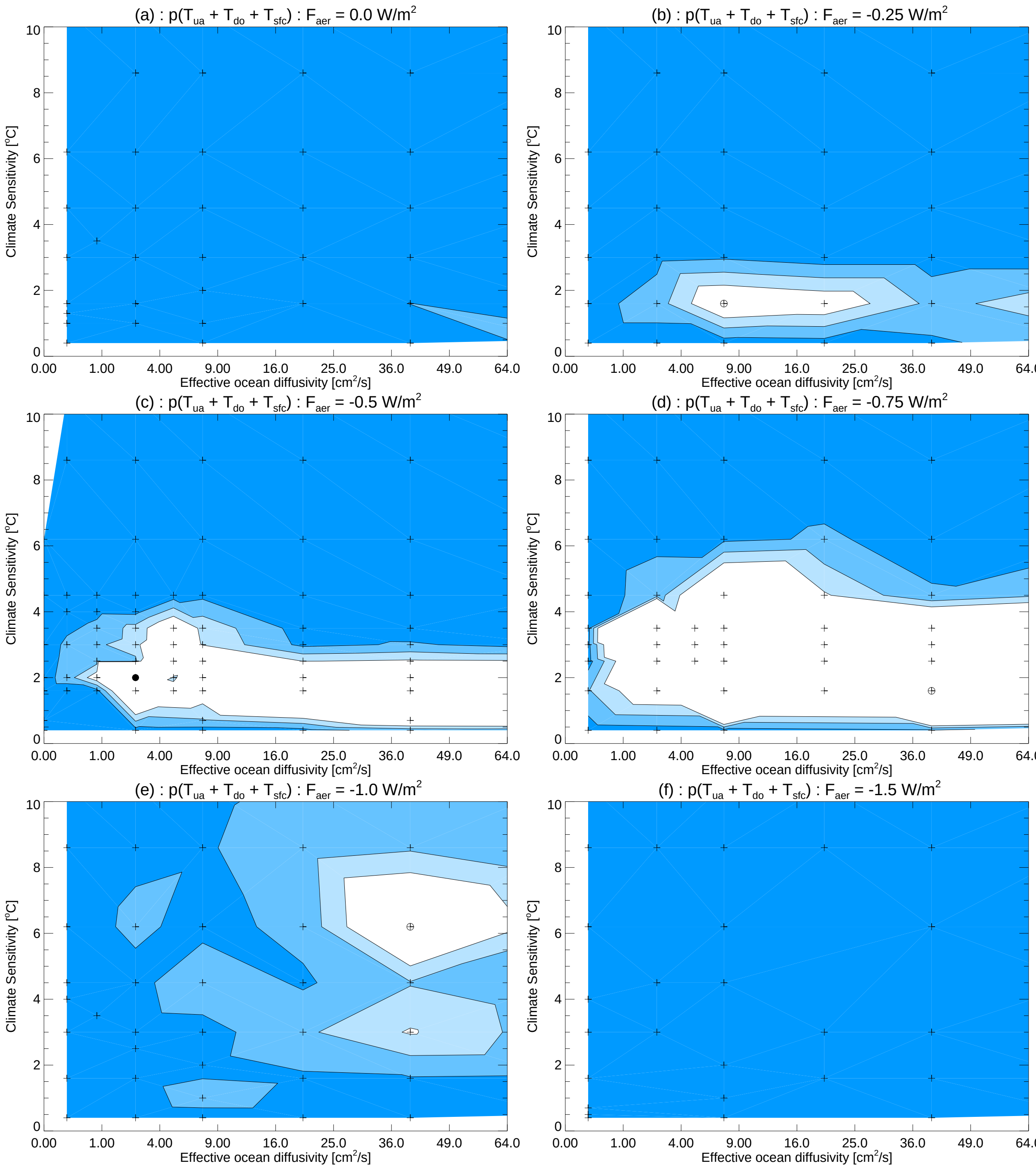

Figure 2: The dependence of the final updated distribution of r^2 on S and K_v (vertical and horizontal axes) and on net aerosol forcing (panels a-f). Shading and contours are as defined in Figure 1.

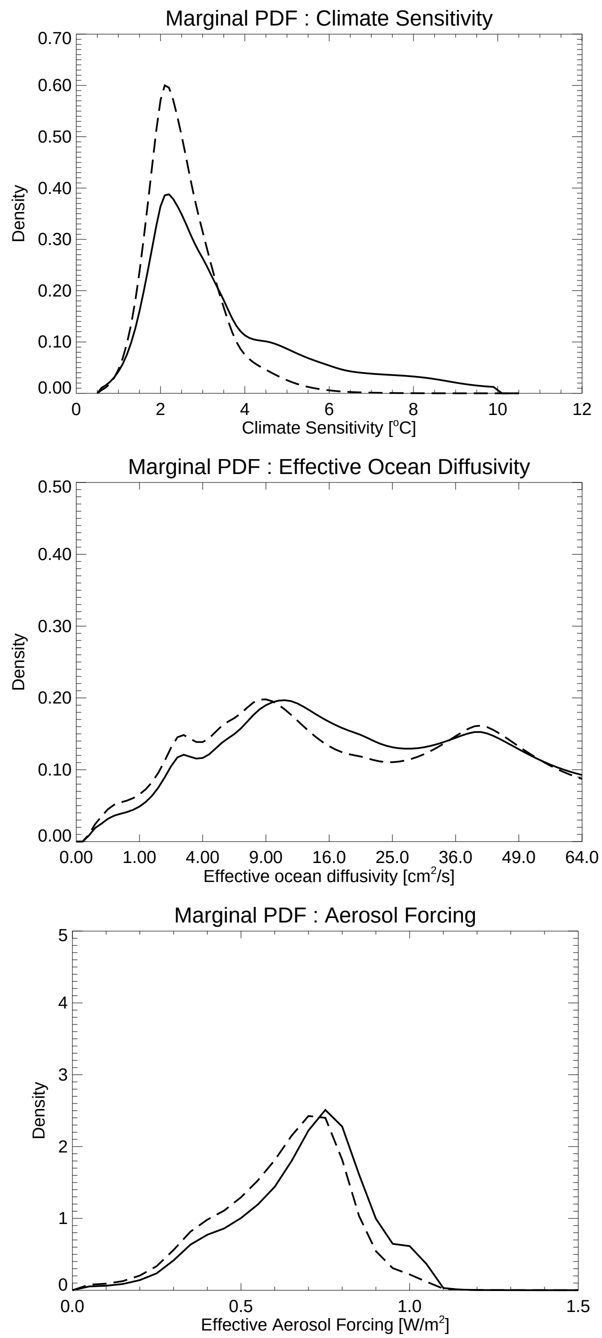

Figure 3: Marginal probability density functions for three climate model parameters with (dashed) and without (solid) the use of an expert prior for climate sensitivity from ref. [26].

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