MIT Joint Program on the Science and Policy of Global Change

Constraining Climate Model Parameters from Observed 20th Century Changes

Chris E. Forest, Peter H. Stone and Andrei P. Sokolov

Report No. 157
April 2008

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Henry D. Jacoby and Ronald G. Prinn,
Program Co-Directors

For more information, please contact the Joint Program Office

Postal Address: Joint Program on the Science and Policy of Global Change
77 Massachusetts Avenue
MIT E40-428
Cambridge MA 02139-4307 (USA)

Location: One Amherst Street, Cambridge
Building E40, Room 428
Massachusetts Institute of Technology

Access: Phone: (617) 253-7492
Fax: (617) 253-9845
E-mail: globalchange@mit.edu
Web site: http://mit.edu/globalchange/

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CONSTRAINING CLIMATE MODEL PARAMETERS FROM OBSERVED 20TH CENTURY CHANGES

Chris E. Forest,* Peter H. Stone, and Andrei P. Sokolov

Abstract

We present revised probability density functions for climate model parameters (effective climate sensitivity, the rate of deep-ocean heat uptake, and the strength of the net aerosol forcing) that are based on climate change observations from the 20th century. First, we compare observed changes in surface, upper-air, and deep-ocean temperature changes against simulations of 20th century climate in which the climate model parameters were systematically varied. The estimated 90% range of climate sensitivity is 2. to 5. K. The net aerosol forcing strength for the 1980s has 90% bounds of -0.70 to -0.27 W/m^2*. The rate of deep-ocean heat uptake corresponds to an effective diffusivity,* K_v *, with a 90% range of 0.04 to 4.1 cm*^{2}*/s. Second, we estimate the effective climate sensitivity and rate of deep-ocean heat uptake for 11 of the IPCC AR4 AOGCMs. By comparing against the acceptable combinations inferred by the observations, we conclude that the rate of deep-ocean heat uptake for the majority of AOGCMs lie above the observationally based median value. This implies a bias in the predictions inferred from the IPCC models alone. This bias can be seen in the range of transient climate response from the AOGCMs as compared to that from the observational constraints.*

Contents

*MIT Joint Program on the Science and Policy of Global Change (E-mail: ceforest@mit.edu)

1 INTRODUCTION

The recognition that anthropogenic activity is causing global warming (Houghton *et al.*, 2001; Solomon *et al.*, 2007) has emphasized the importance of developing climate models with predictive capability. In recent decades considerable effort has been devoted to evaluating state-of-the-art climate models from this point of view. A good summary of this work is given in Chapter 8 of the latest IPCC report (Randall *et al.*, 2007). Much of the work has focused on evaluating the models' ability to simulate the annual mean state, the seasonal cycle, and the inter-annual variability of the climate system, since good data is available for evaluating these aspects of the climate system. However good simulations of these aspects do not guarantee a good prediction. For example, Stainforth *et al.* (2005) have shown that many different combinations of uncertain model sub-grid scale parameters can lead to good simulations of global mean surface temperature, but do not lead to a robust result for the model's climate sensitivity.

A different test of a climate model's capabilities that comes closer to actually testing its predictive capability on the century time scale is to compare its simulation of changes in the 20th century with observed changes. A particularly common test has been to compare observed changes in global mean surface temperature with model simulations using estimates of the changes in the 20th century forcings. The comparison often looks good, and this has led to statements such as: "...the global temperature trend over the past century .... can be modelled with high skill when both human and natural factors that influence climate are included" (Randall *et al.*, 2007). However the great uncertainties that affect the simulated trend (e.g., climate sensitivity, rate of heat uptake by the deep-ocean, and aerosol forcing strength) make this a highly dubious statement. For example, a model with a relatively high climate sensitivity can simulate the 20th century climate changes reasonably well if it also has a strong aerosol cooling and/or too much ocean heat uptake. Depending on the forcing scenario in the future, such models would generally give very different projections from one that had all those factors correct.

There have in recent years been a number of studies using the observed 20th century temperature to calculate probability density functions (PDFs) for the above mentioned uncertain parameters (Andronova & Schlesinger, 2001; Forest *et al.*, 2002, 2006; Knutti *et al.*, 2003). A meaningful test of a model's capabilities can be provided by comparing properties of different state-of-the-art models with their values, as implied by 20th century changes. Forest *et al.* (2006) have presented such a comparison for the models used in the IPCC TAR but not for those models used in the IPCC AR4. Here, we present an update of the Forest *et al.* (2006) results, in which we use the 20th century observations to constrain the effective climate sensitivity rather than the equilibrium climate sensitivity, while simultaneously constraining the ocean heat uptake and aerosol forcing; and we also now analyze 11 of the IPCC 4AR models for which the necessary data is available. Recent improvements made in the climate model have caused the model's effective and equilibrium sensitivities to differ significantly from each other when the climate sensitivity is large. The effective sensitivity is obviously more relevant for describing 20th century changes. Section 2 describes the version of the MIT climate model used in the present study, Section 3 describes the method for constraining climate model parameters, Section 4 gives the results, and Section 5 summarizes and discusses the results.

2 MIT 2D CLIMATE MODEL

The model used in this study is the climate component of the MIT Integrated Global System Model, Version 2 (Sokolov *et al.*, 2005). This model is an updated version of the model described in Sokolov & Stone (1998). Here we give a brief summary of the model and of the changes made since Forest *et al.* (2006)

The model consists of a zonally averaged atmospheric model coupled to a mixed layer Q-flux ocean model, with heat anomalies diffused below the mixed layer. The atmospheric model is derived from the Goddard Institute for Space Studies (GISS) Model II general circulation model (GCM) (Hansen *et al.*, 1983) and uses parameterizations of the eddy transports of momentum, heat and moisture by baroclinic eddies (Stone & Yao, 1987, 1990). The model uses the GISS radiative transfer code which contains all radiatively important trace gases as well as aerosols. The surface area of each latitude band is divided into fractions of land, ocean, land-ice and sea-ice, with the surface fluxes computed separately for each surface type. The version used here has 4 degree latitudinal resolution and 11 layers in the vertical. The zonal averaging and the relatively low meridional and vertical resolution are necessary to make the model computationally efficient enough so that we can carry out simulations totalling hundreds of thousands of years, as required by our methodology (see next section). The ocean mixed layer model and the thermodynamic sea-ice model have 4 degree by 5 degree latitude-longitude resolution and are described by Hansen *et al.* (1984).

The climate sensitivity of the MIT model can be varied by changing the strength of the cloud feedback (Sokolov, 2006), differences in which have been shown to be the main reason for the differences in model climate sensitivity between different AOGCMs (e.g., Cess *et al.*, 1990; Colman, 2003). The rate of mixing thermal heat anomalies into the deep ocean is controlled by the global mean value of the vertical diffusivity coefficient for mixing anomalies (K_v). Sokolov & Stone (1998) and Sokolov *et al.* (2003) have shown that the large-scale response of a given coupled atmosphere-ocean GCM (AOGCM) to forcings typical of the 20th and 21st century can be duplicated by the MIT 2D model with an appropriate choice of these two parameters for any scenario. This ability to mimic the AOGCMs is what allows us to use the MIT 2D model to explore how consistent different AOGCMs are with observed 20th century temperature changes.

The method for changing cloud feedback in the model has been changed from the method used previously. In the earlier versions of the model the cloud cover at all levels was changed by a fixed fraction, which depended on the changes in global mean surface temperature (Sokolov & Stone, 1998). In the present version high cloud covers and low cloud covers are changed in opposite directions by a constant factor, which is again dependent on changes in the global mean surface temperature. The new method is described by Sokolov (2006), who shows that this method is in better agreement with changes simulated by AOGCMs, and does not change the 2D model's ability to mimic global scale temperature changes simulated by AOGCMs.

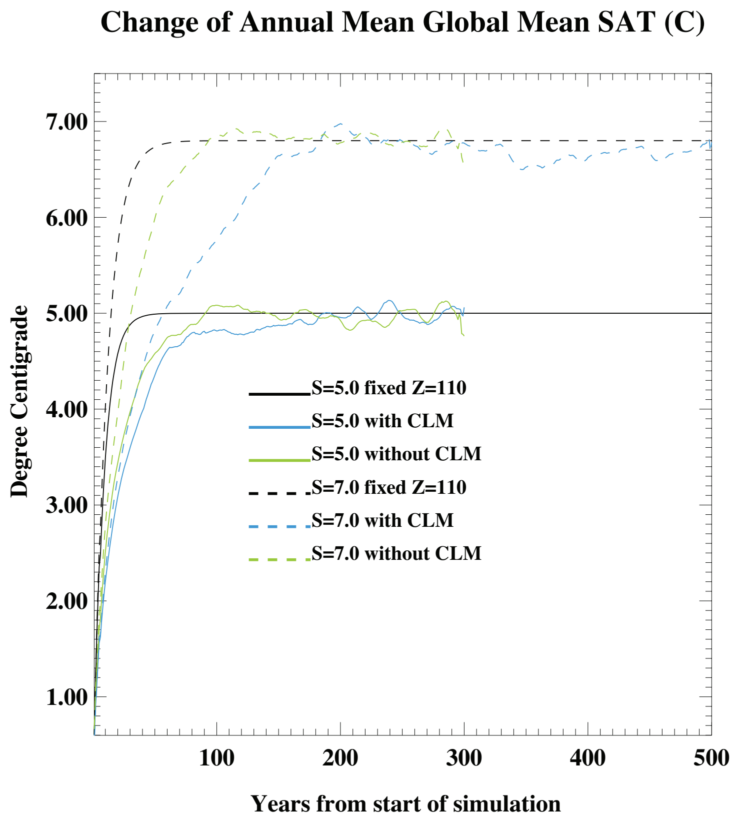

Figure 1: Global-mean surface air temperature in simulations by MIT 2D climate model with (blue) and without (green) the CLM land-surface model for an instantaneous doubling of CO2 concentration. The response by an energy balance model with a 110 meter deep ocean mixed-layer is shown by thin black line.

The most significant change that has been made in the current version of the MIT 2D climate model is the replacement of the old GISS land surface scheme by the Community Land Model (CLM2.1) described by Bonan *et al.* (2002). (See Schlosser *et al.* (2007) for the description of the coupling to the 2D model.) This improved the simulation of evaporation and removed the tendency of the land model to be too hot in summer, due to excessive evaporation in spring causing the land to dry out. This was also a problem in the parent GCM. However the slower response of the land evaporation to warming in the new model significantly altered the transient response of the IGSM to an external forcing. **Figure 1** shows changes in surface air temperature in simulations with an instantaneous doubling of CO2 concentration. While evaporation from land is too small to directly affect the global surface energy budget in a significant way, a small rate of land evaporation response to surface warming leads to a delay in the increase of atmospheric water vapor. This, in turn, causes slower warming by reducing the incoming longwave radiation at the surface.

The differences in the response to an external forcing between the two versions of the 2D model result in different relations between equilibrium (S_{eq}) and effective (S_{eff}) climate sensitivities.

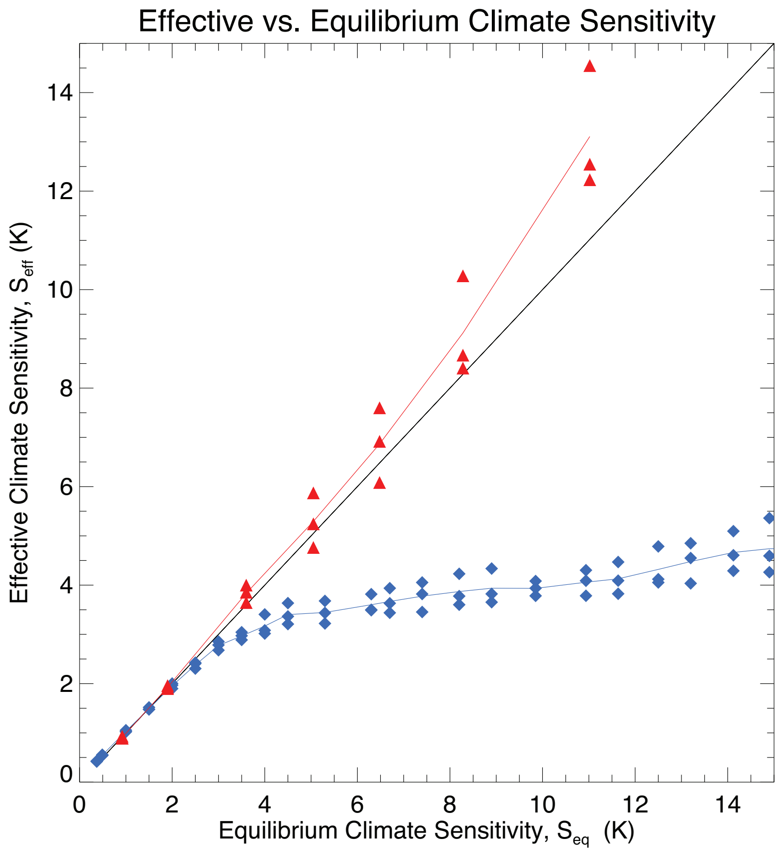

Figure 2: Comparison of effective and equilibrium climate sensitivities for the MIT 2D climate model with (blue diamonds) and without (red triangles) the CLM land-surface model. The three values for each equilibrium sensitivity correspond to K_v equal to 0.25, 2.25, and 6.25 cm^2/s. The red and blue lines are the averages of these three values. The black line indicates where the two sensitivities are equal.

Effective sensitivity is defined as $S_{eff} = \frac{F_{2x}}{\lambda_{eff}}$, where F_{2x} is the forcing due to CO2 doubling and λ_{eff} is the climate feedback parameter estimated at the time of CO2 doubling in a scenario where CO2 increases by 1% per year (Murphy, 1995). In effect the slower increase of evaporation when the climate warms delays the onset of the positive water vapor feedback in the simulation with the new model, and reduces S_{eff} relative to S_{eq} . In the earlier versions of the model the two sensitivities were essentially equal. Since the 20th century changes are transient, it is clearly preferable to use them to constrain S_{eff} rather than S_{eq} . **Figure 2** shows the relationship between S_{eff} and S_{eq} in the new model. The two sensitivities are virtually equal for $S_{eq} < 3$ degrees but S_{eff} is considerably less than S_{eq} for large values of S_{eq} .

3 METHODS

3.1 Estimation of probability distributions

The methodology for quantifying uncertainty in climate system properties follows the basic method in Forest *et al.* (2001, 2002, 2006) with the modifications required to use the climate model. This can be summarized as consisting of two parts: simulations of the 20th century climate record and the comparison of the simulations with observations using optimal fingerprint diagnostics. First, we require a large sample of simulated records of climate change in which climate parameters have been systematically varied. Second, we employ a method of comparing model data to observations that appropriately filters “noise” from the pattern of climate change. The variant of optimal fingerprinting proposed by Allen & Tett (1999) provides this tool and yields detection diagnostics that are objective estimates of model-data goodness-of-fit. In the use of the temperature change diagnostics and the estimation of the posterior probability distribution, the methodology is identical to that in Forest *et al.* (2006). The three temperature change diagnostics that we use are: (i) the decadal mean surface temperature changes over 4 equal-area latitude bands for the period 1946-1995 referenced to the 1905-1995 climatology; (ii) the trend in the global mean ocean temperature (down to 3 km depth) during 1948-1995; and (iii) the latitude-height pattern of the zonal mean upper air temperature difference between the 1961-1980 and 1986-1995 periods. The likelihood functions based on each diagnostic are combined using Bayes’ Theorem.

The description of the climate model experiments, the ensemble design, and the algorithm for estimating the joint PDFs are in Forest *et al.* (2001, 2002, 2006). There are two major differences from Forest *et al.* (2006) that were required when using the new model. First, a land-use change data set for the twentieth century was not included in these simulations because none was available in the new model’s format. However, the contribution of this forcing to the total 20th century forcing is very small (Solomon *et al.*, 2007). Thus, the set of applied climate forcings was reduced to: greenhouse gas concentrations, sulfate aerosol loadings, tropospheric and stratospheric ozone concentrations, solar irradiance changes, and stratospheric aerosols from volcanic eruptions. We refer to this set of forcings as GSOSV. (Details on these forcings are in the auxiliary material in Forest *et al.* (2006).)

The second change was required to accommodate the change from equilibrium to effective climate sensitivity. Because S_{eff} has an upper bound at about 8 K in the new climate model, we truncate the distribution at 8 K rather than 10 K as was done in our previous studies. Thus, for the uniform prior cases, the cumulative probability above 8 K will differ from the results in Forest *et al.* (2006). In the case where an expert prior is used for S_{eff} , the prior has near zero probability above 8 K and the results are basically unaffected.

When conducting 20th century simulations, we use different values of the strength of the cloud feedback which changes both S_{eff} and S_{eq} . While S_{eff} is a more appropriate measure of transient climate response, results from our previous studies were presented in terms of S_{eq} because, first, S_{eq} and S_{eff} were virtually the same for older versions of the model and second, there is a one to one correspondence between S_{eq} and the strength of the cloud feedback. For the new model,

Table 1: Values of S_{eff} and K_v for AOGCMs used in the IPCC AR4 (top) and TAR (bottom).

Index	Model	K_v	S_{eff}
1	CGCM3.1(T47)	2.9	3.4
2	ECHO-G	1.3	2.8
3	GFDL-CM2.0	4.	2.2
4	GFDL-CM2.1	4.	2.2
5	INM-CM3.0	0.7	2.0
6	MIROC3.2(medres)	4.0	4.8
7	GISS-EH	1.7	2.2
8	CCSM3	3.4	2.2
9	GISS-ER	3.1	2.2
10	HadCM3	1.9	3.6
11	PCM	2.1	1.9
12	HadCM2	3.0	2.8
13	ECHAM3	1.6	2.4
14	MRI (old)	7.5	3.2
15	CSM	3.8	1.9

S_{eff} is very different from S_{eq} (for high values), so results of the present study are presented in terms of S_{eff} . The value of S_{eff} for a given strength of cloud feedback depends slightly on K_v . On figures below, we use values of S_{eff} averaged over K_v .

3.2 Matching procedure for AOGCMs.

As discussed earlier, the large-scale response of the MIT model is controlled by the parameters, S_{eff} (or S_{eq}) and K_v . This flexibility provides the ability to match the large-scale response of AOGCMs by choosing appropriate combinations of these two parameters. Fits for the models were obtained based on the data for surface air temperature (SAT) and thermosteric sea level rise from the simulations with 1% per year increase in CO2 concentration. Unfortunately the required data are available for only nine (9) AR4 models as part of the CMIP3 dataset (Meehl *et al.*, 2007). Fits for the HadCM3 and five TAR models are based on the results from CMIP2 simulations. In **Figure 3**, the values of S_{eff} required to match models' responses are compared with values of S_{eff} published for the corresponding models. Effective sensitivities for the AR4 models were estimated from the data on "top of the atmosphere fluxes" from the archived CMIP3 dataset (Meehl *et al.*, 2007) using values of the adjusted radiative forcing due to CO2 doubling (F_{2X}) given in Table 8S.1 from the IPCC AR4. Values of S_{eff} for CMIP2 models were taken from the literature. It should be noted that for some models F_{2X} is not available and in these cases, a forcing of 3.71 Wm^{-2} was used. **Table 1** gives the 2D model's values of S_{eff} and K_v that match the performance of the listed models.

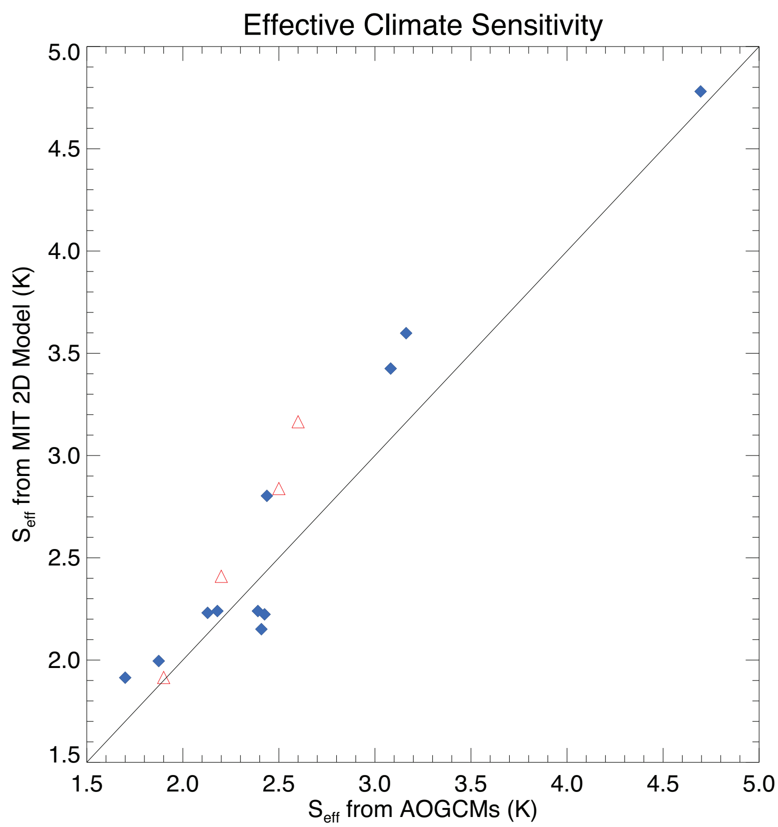

Figure 3: Comparison of effective climate sensitivity estimated from AOGCM simulations vs effective climate sensitivity required to fit the AOGCM transient response. Blue diamonds refer to AR4 models and red triangles refer to TAR models.

Table 2: Fractiles for marginal PDFs with and without expert prior on S_{eff} .

		0.05	0.50	0.95	Mean	Mode
Expert Prior	S_{eff}	2.0	2.8	5.0	2.9	2.4
	K_v	0.04	0.78	4.1	0.97	0.49
	F_{aer}	-0.27	-0.50	-0.70	-0.50	-0.55
Uniform Prior	S_{eff}	2.1	4.0	7.4	4.1	3.0
	K_v	0.12	1.7	6.1	1.7	2.2
	F_{aer}	-0.32	-0.58	-0.77	-0.56	-0.65

Heat uptake in the oceans is sometimes measured by a coupled model's heat uptake efficiency, E, (Gregory & Mitchell, 1997). We have compared our measure, K_v , with E estimated for the nine coupled AOGCMs using the CMIP3 datasets. They are correlated, with a correlation of 0.837.

4 RESULTS

4.1 Posterior distributions using the new model, IGSM2

The one-dimensional marginal distributions from the current analysis (**Figure 4**) and the Forest *et al.* (2006) estimates (their Figure 2) are very similar and indicate the model responses are nearly identical. The fractiles for S_{eff} , K_v , and F_{aer} are in **Table 2**. The aerosol forcing remains well constrained. The distribution for climate sensitivity with the expert prior, as before, has a well-defined mode at 2.8 K while the upper tail remains significant. The expert prior on climate sensitivity remains an important feature of the results with a reduction in the likelihood above 4.5 from 42 to 8 percent in the new results. As before, K_v is well constrained by the three diagnostics with the surface temperature providing a strong constraint on the upper bound. The two-dimensional marginal distributions are shown in **Figure 5** for the S-$\sqrt{K_v}$ parameter space. The positions of the climate models' heat uptake generally remain significantly to the right of the median and mode for the distribution. Given that the mode is an estimate of the most likely value, the AR4 models appear to have a positive bias in their ocean heat uptake, although we have not been able to obtain the data necessary to calibrate 10 of the AR4 models.

We can also explore the possible bias in the AR4 models' predictions from our distributions. We show the distributions for TCR and SLR (respectively, changes in SAT and thermosteric sealevel rise averaged over years 61-80 in simulations with 1% per year increases in CO2 concentration) as estimated from our new distribution and also as estimated for the AOGCMs (**Figure 6**). Taking the means of the PDFs and the AOGCM distributions, we find that the AOGCMs appear biased low for TCR. There is also a high bias in the AOGCMs for SLR, but this is partly compensated by their low warming bias.

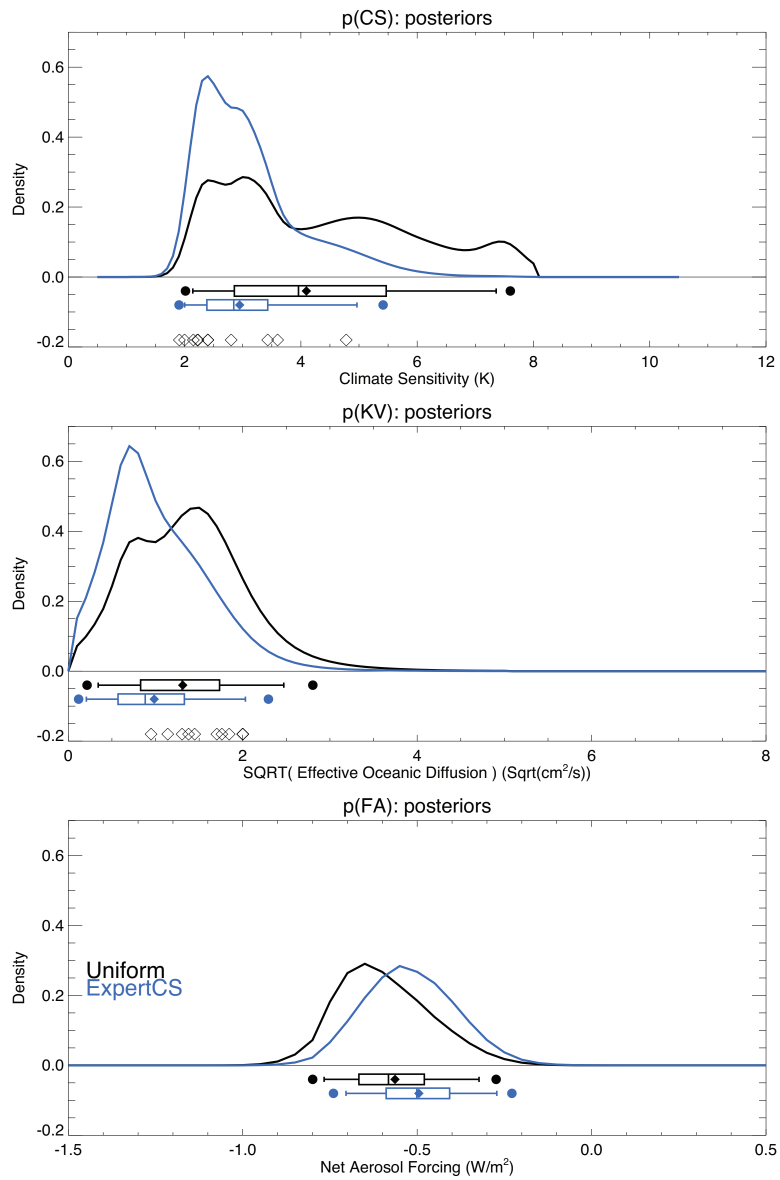

Figure 4: The marginal *posterior* probability density function for the three climate system properties for two cases with the new model. In each panel, the marginal pdfs are shown for the GSOSV forcings. In one case (black), uniform priors are used on all parameters and in the second case (blue), an expert prior on climate sensitivity (Webster & Sokolov, 2000) is used with uniform priors elsewhere. Marginal distributions are estimated by integrating the density function over the remaining two parameters and renormalizing. The whisker plots indicate boundaries for the percentiles 2.5-97.5 (dots), 5-95 (vertical bar at ends), 25-75 (box ends), and 50 (vertical bar in box). The mean is indicated with the diamond and the mode is the peak in the distribution. The values for S_{eff} and K_v for the AR4 AOGCMs are shown as diamonds below the whisker plots.

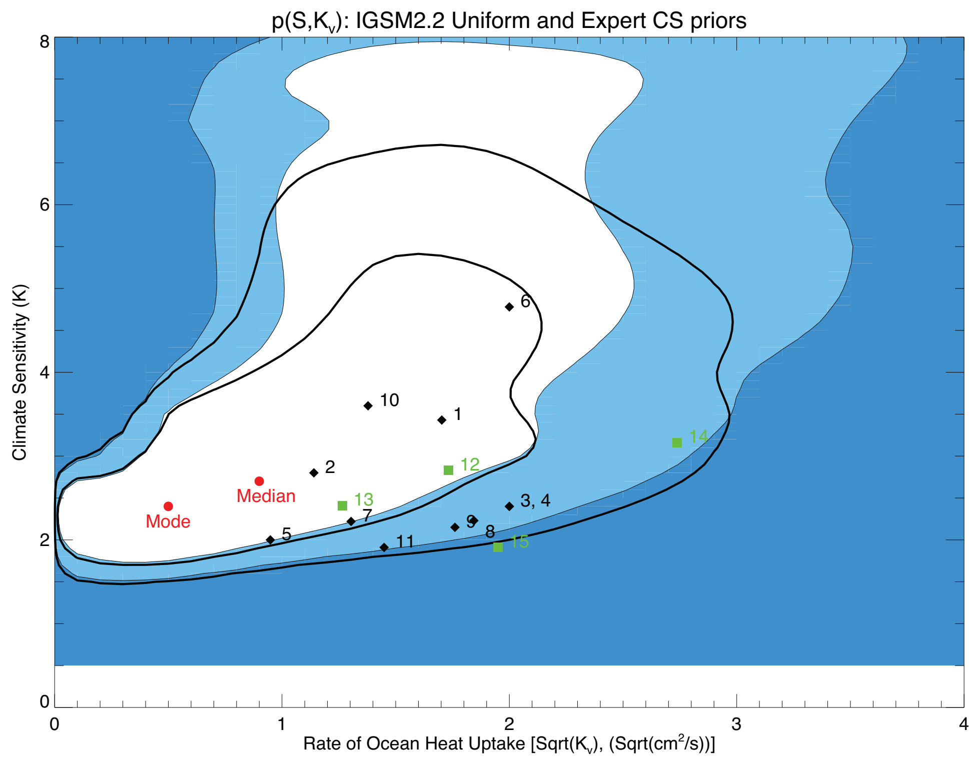

Figure 5: The marginal *posterior* probability density function for GSOSV results with uniform (shading) and expert prior on S (thick contours) for the S -K_v parameter space. The shading denotes rejection regions for a given significance level — 10%, and 1%, light to dark, respectively. The positions of AOGCMs (diamonds and squares) represent the parameters in the MIT IGSM2 model that match the transient response in surface temperature and thermal expansion component of sea-level rise under a common forcing scenario. Model names and parameter values are listed in Table 1. Lower K_v values imply less deep-ocean heat uptake and hence, a smaller effective heat capacity of the ocean. Eleven AOGCMs used in the IPCC AR4 report (black diamonds) and four used only in the IPCC TAR (from Sokolov *et al.* (2003)) (green squares) represent the models with sufficient information available. The median and mode (red circles) are shown for the case with the expert prior.

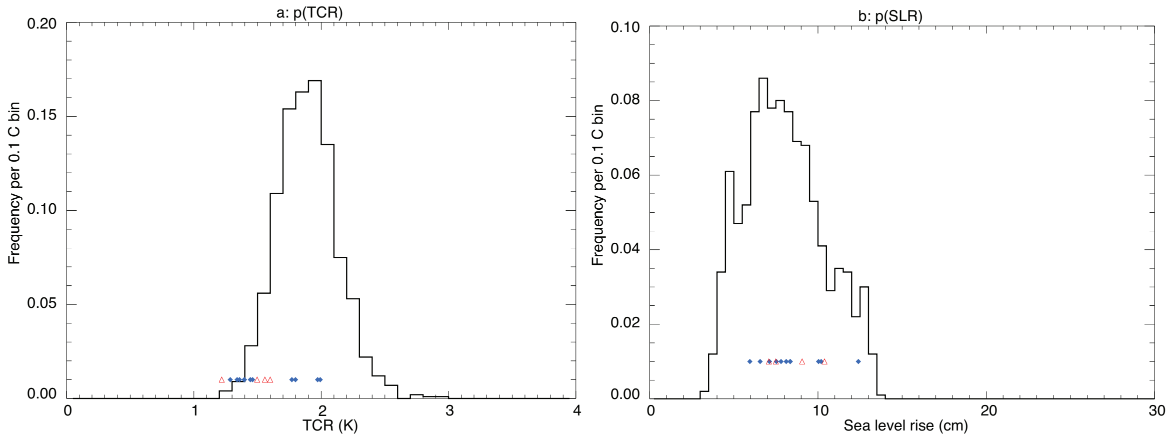

Figure 6: (a) $p(TCR|data)$ and (b) $p(SLR|data)$ from 2D model results based on 1000 member Latin Hypercube sample. The values of TCR and SLR for the AOGCMs in Table 1 are shown with blue diamonds for the AR4 models and red triangles for the TAR models.

4.2 Robustness of the ocean heat-uptake results

Since there is a significant discrepancy between the AOGCMs' simulations of ocean heat uptake and the uptake we estimate from observations, we have explored the sensitivity of our posterior distribution for K_v to various diagnostics. First we show in **Figure 7** how our 1D marginal distributions change when we remove information associated with the three different diagnostics. In particular, we compare our standard results based on all 3 diagnostics with what happens if: (i) we leave out the upper-air diagnostic, (ii) we leave out the deep-ocean temperature change diagnostic, and (iii) we replace the surface temperature change diagnostic using 4 latitude bands, z4, by one using only hemispheric averages, z2, but still retaining the decadal time series. In the last case we retain the contrast in hemispheric temperature averages that reflects the aerosol forcing being larger in the Northern Hemisphere, but remove the polar amplification component in the z4 diagnostic. In all cases we see that the PDFs for S_{eff} and F_{aer} are not much affected and we conclude that these PDFs are relatively robust. However in the case of the K_v distribution we see that removing any of the diagnostics weakens the constraint on K_v , with the removal of the deep-ocean temperature diagnostic showing the most effect. Nevertheless the mode for K_v is relatively robust, and indeed it is smallest when the deep-ocean temperature diagnostic is removed. Thus all the diagnostics contribute to the discrepancy between our estimate of the deep-ocean heat uptake and the uptake simulated by the AOGCMs, although the discrepancy is most significant when the deep-ocean temperature diagnostic is included.

Second, we looked at how sensitive our estimate of the ocean heat uptake is to newly discovered errors in the observed ocean temperature trend which were not taken into account in our results given above. In our standard analysis we used the ocean trend and error estimates given by Levitus *et al.* (2005). It has recently come to light that the XBT data that they used in their analysis contained systematic errors (Gouretski & Koltermann, 2007). The Gouretski and Koltermann analysis indicates that the Levitus et al. trend should be reduced by 37%, while a more recent analysis re-

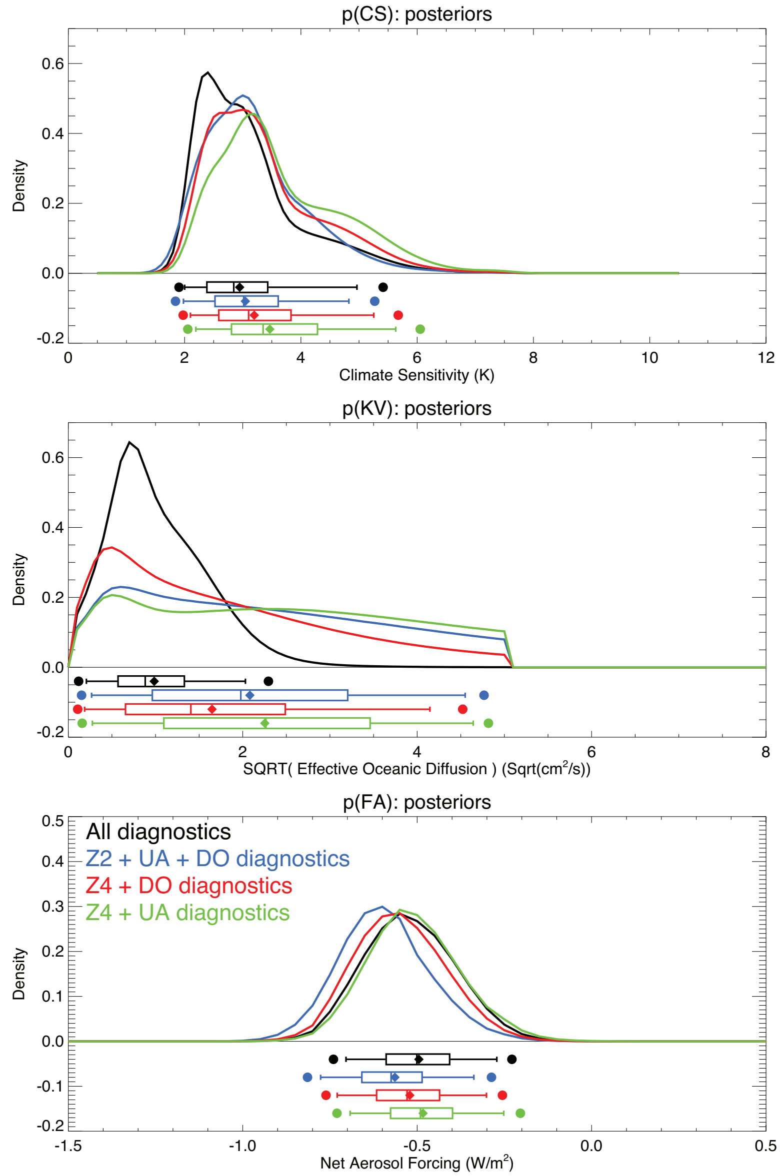

Figure 7: Posterior distributions using alternative combinations of climate change diagnostics. Standard diagnostics (black) using surface (z4), upper-air (UA), and deep-ocean (DO) temperatures; z2 + UA + DO (blue), hemispheric averages replace four equal-area zonal bands; z4 + DO (red), no upper-air temperatures; and z4 + UA (green), no deep-ocean temperatures.

ported at the AGU meeting in December, 2007, indicates it should be reduced by 24% (J. Antonov, personal communication). We repeated our analysis with the trend reduced by 37% and a (larger) error estimate taken from Gouretski and Koltermann. The results (not shown) had the mode for K_v reduced to a value consistent with that when the ocean diagnostic was removed (Figure 7) and the distribution was somewhat broadened, but not as much as when the ocean diagnostic was removed. Thus the discrepancy in the models' heat uptake remains.

Finally we note two recent studies based on the Levitus *et al.* (2005) analysis of the ocean heat uptake that also indicate that AOGCMs are overestimating the 20th century heat uptake. Pierce *et al.* (2006) compared 20th century simulations of the heat uptake using the PCM and HadCM3 models with the Levitus *et al.* (2005) results using the observational data mask. Their Figure 11 shows that both models are overestimating the ocean heat uptake, particularly below the mixed layer. Andrews & Allen (2007) compared the performance of the AR4 AOGCMs with 20th century changes in surface temperature and ocean heat uptake, and found that the AOGCMs were generally overestimating the effective heat capacity of the climate system, which is of course equivalent to mixing heat into the ocean too efficiently.

5 DISCUSSION AND CONCLUSIONS

We present two new results in this paper. First, we have estimated the S_{eff} and K_v values that correspond to eleven (11) of the AR4 AGOCMs models. This serves to characterize the "ensemble of opportunity" (EOP) in terms of both equilibrium and transient responses. Together, these two properties provide a good metric for comparing the behavior of different AOGCMS with one another and with respect to the distributions for these properties as estimated from climate change observations. Second, we present the updated probability distribution for the three climate system properties, $\theta = \{S_{eff}, K_v, F_{aer}\}$, with S_{eq} replaced by S_{eff} . These distributions are similar to those from Forest *et al.* (2006), because the forcings are almost identical (no land-use change in the present case) and the climate change diagnostics were identical.

From the positions of the AOGCMs within this distribution, we can estimate the AOGCMs' projections under specific forcing scenarios. As noted by many (e.g., Prinn *et al.*, 1999), the total uncertainty in the climate change projections is a combination of the uncertainties in both the forcings and the climate system response. By considering the AOGCM positions within the context of the $p(\theta|\Delta T, C_N)$ distributions, one can infer the range of uncertainty in the climate system response that is represented by their projections. Furthermore, we can track the change in the uncertainty implied by the projections in the various IPCC reports. As shown in Figure 5, the projections from both the TAR and AR4 indicate a significant shift in the climate model response as estimated by the AOGCMs and their means, medians, and ranges. Although the complete set of models is not available, we still find a clear indication that the AOGCMs, as a whole, overestimate the rate of deep-ocean heat uptake as implied by the observations. We quantify this by considering the distributions of TCR and SLR (Figure 6) obtained by using a Latin Hypercube sample from the posterior distribution with an expert prior on S_{eff} . The range of both TCR and SLR implied by the AR4 AOGCMs is narrower than that based on observational constraints, while the latter is still narrower than the IPCC AR4's official projections. We also see that the AR4 results appear biased low for temperature change while biased high for sea level rise. This is expected given the positions of the AOGCMs in the joint distribution of S_{eff} -K_v .

Acknowledgments

This work was supported in part by the Office of Science (BER), U.S. Dept. of Energy Grant No. DE-FG02-93ER61677, NSF, and by the MIT Joint Program on the Science and Policy of Global Change. We thank Adam Schlosser for his work on coupling CLM in the climate model. We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI) and the WCRP's Working Group on Coupled Modelling (WGCM) for their roles in making available the WCRP CMIP3 multi-model dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy.

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