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Comment on "Heat capacity, time constant, and sensitivity of Earth's climate system" by S. E. Schwartz

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1. Introduction

[1] *Schwartz* [2007] (hereinafter referred to as SES) recently suggested a method to calculate equilibrium climate sensitivity (the equilibrium global surface warming for a doubling of the atmospheric CO_2 concentration), the effective heat capacity of the Earth's climate system, the temperature response time scale relevant to climate change, an estimate of total radiative forcing as well as the magnitude of the aerosol forcing over the 20th century. The main results are that the characteristic response time scale of global temperature is 5 ± 1 years and climate sensitivity is 0.30 ± 0.14 K/(W m^{-2}), corresponding to an equilibrium temperature increase for doubling atmospheric CO_2 of 1.1 ± 0.5 K. In practical terms, this means that global surface temperature is nearly in equilibrium with radiative forcing, and that the sum of all feedbacks (water vapor, lapse rate, clouds, albedo) is close to zero. These results are at odds with most of the current literature on climate sensitivity, the idea of commitment warming and with the magnitudes of climate feedbacks quantified in models and observations. If true, the low climate sensitivity would allow for much higher atmospheric greenhouse gas concentrations to be consistent with a given stabilization temperature compared to the current consensus. It would also imply that stabilization of atmospheric CO_2 would lead to stabilization of global temperatures within a few years.

[2] In simple terms, SES uses an energy balance argument to claim that climate sensitivity S is given by $S = \tau/C$, where C is the effective heat capacity of the Earth and τ is a characteristic time scale. The effective heat capacity is obtained by a regression of observed global ocean heat uptake versus global surface temperature (as done in several previous studies), while τ is obtained from the autocorrelation of the observed linearly detrended global surface temperature time series. Details are discussed by *Schwartz* [2007].

2. Time Scales in the Climate System

[3] *Foster et al.* [2007] focus on several statistical problems in the analysis of SES and show that the observed global temperature in fact does not behave like a first-order autoregressive (AR(1)) process. They also demonstrate that the method proposed by SES would result in a biased estimate of the time scale even for a perfect AR(1) process because of the limited length of the time series and the presence of a trend. From a physical point of view, the critical assumption made by SES is that the Earth has a single time scale which characterizes interannual variability as well as the response to radiative forcing imposed for decades to centuries. Given the large number of processes which affect temperature in the climate system and which operate on time scales of days to centuries, this seems a priori implausible. Interannual variations in global temperature, which dominate the autocorrelation estimate by SES are determined mostly by atmospheric processes and, e.g., patterns like ENSO, with typical time scales shorter than a few years, i.e., exchanges of heat between the upper ocean and space. They are also influenced by short-term variations in the radiative forcing, e.g., by volcanic eruptions. The response time scale of global temperature to sustained radiative forcing, on the other hand, has several components. The response of the atmosphere is fast (order years or less), the land and sea ice components react slower, and the long time scales of the response are dominated by the time it takes for the ocean to equilibrate with the forcing [e.g., *Hansen et al.*, 1985]. While the ocean mixed layer reacts relatively quickly, both models and observations indicate that the typical time scales for diffusion and advection of heat and other tracers into the deep ocean are from decades to centuries [e.g., *Sabine et al.*, 2002; *Stouffer*, 2004]. These time scales can be estimated from the distributions of temperature and salinity in the oceans themselves, as well as from measurements of how quickly the anthropogenic perturbations of heat, carbon dioxide, carbon isotopes and CFCs are mixed into the deep ocean.

[4] Three dimensional coupled atmosphere ocean general circulation models (AOGCMs) provide the most comprehensive description of the climate system. Here we use 19 of these models from the recent World Climate Research Programme's (WCRP's) Coupled Model Intercomparison

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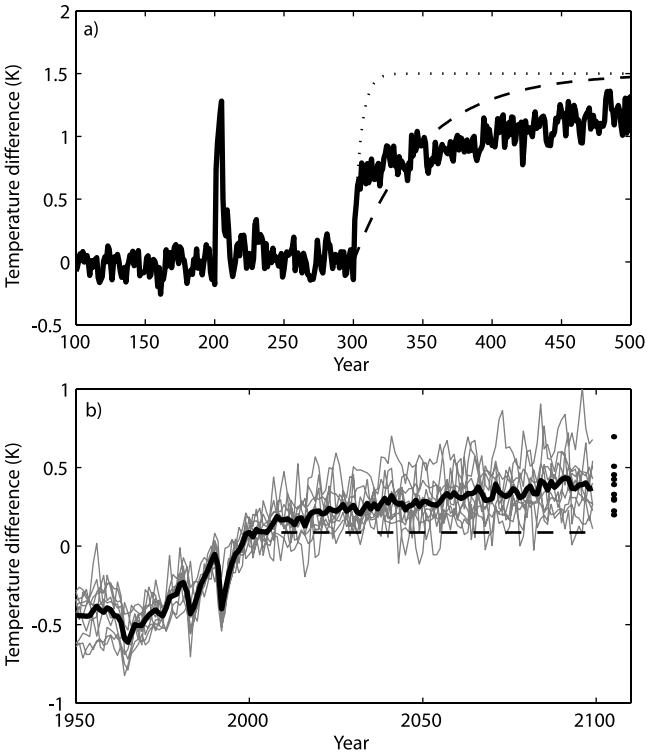

Figure 1. (a) Global temperature anomaly from control in the Ecbilt Clio model (solid). Atmospheric CO_2 is quadrupled for 5 years after year 200 and set back to its preindustrial value in year 205 and is doubled and kept constant after year 300. While the response to the short forcing and the initial response to the step forcing are quick, temperature is still not in equilibrium 200 years after CO_2 doubling. Two simple relaxation profiles proportional to $1 - \exp(-t/\tau)$ as assumed by SES (his equation (6)) with a time scale τ of 5 and 50 years, respectively, are given for illustration. (b) Global temperature anomaly relative to the 1990–2009 period for ten of the CMIP3 AOGCM models that included volcanic forcing in the 20th century and did simulate the commitment scenario in which atmospheric concentrations are kept constant at year 2000 values (thin gray lines, average as black thick line). The dotted line indicates the year 2000 temperature level. Dots at the right indicate the means over the last 20 years for each model. While temperature recovers within a few years after the Pinatubo eruption in 1991, the time scale to reach equilibrium with constant forcing after year 2000 is many decades or more.

Project phase 3 (CMIP3) multimodel data set, and a model of intermediate complexity to illustrate the point of different time scales.

[5] The clearest picture is seen in Figure 1a where the intermediate complexity climate model Ecbilt-Clio is used in a highly idealized scenario. Ecbilt [*Opsteegh et al.*, 1998] is a T21 quasi-geostrophic three layer atmosphere with simplified physics, but which does resolve synoptic variability of weather patterns. Clio [*Goosse and Fichefet*, 1999] is a standard 3-D general circulation ocean model with a resolution of three degrees and twenty unevenly spaced layers. The coupled model includes a dynamic and thermodynamic sea ice component. A simulation is performed where the preindustrial CO_2 concentration is instantly quadrupled (corresponding to a forcing of about 8 W/m^2) in year 200 for only 5 years and set back to preindustrial levels, and then doubled and kept constant after year 300 (about 4 W/m^2). While the responses to the short-term forcing (an idealized "inverse" Pinatubo) as well as the initial warming to the step forcing at year 300 are fast (a few years), the time it takes to reach a new equilibrium for a sustained forcing is more than a century. Two simple relaxation profiles proportional to $1 - \exp(-t/\tau)$ with a time scale τ of 5 and 50 years are shown, respectively. They illustrate that no single time scale can describe the temperature response. Rather, the temperature response contains short time scales (less than a few years), e.g., atmospheric adjustments and land surface processes, medium time scales, e.g., the melting of sea ice, as well as long time scales. The latter are dominated by the time it takes for the deep ocean to equilibrate with the surface, which is on the order of hundreds of years.

[6] SES suggests that because of the short response time scale, the surface temperature today is in near equilibrium with the forcing. Model simulations to test that do already exist, and Figure 1b shows global temperature for the time period 1950 to 2100 from the simulation of the 20th century and the so-called commitment simulation where the radiative forcing is kept constant after the year 2000 [*Meehl et al.*, 2007]. Individual models with volcanic forcing available in CMIP3 are shown as gray thin lines, the average is shown as a thick black line. The warming realized after 2000 is called "constant composition warming commitment" [*Meehl et al.*, 2005, 2007; *Wigley*, 2005]. Despite the fact that the time scale determined from interannual variability with the AR(1) method on average is 5–7 years (depending on how it is calculated, see section 3), the time scale for global temperature to equilibrate with the constant year 2000 forcing is several decades to centuries. In none of the models the temperature is close to equilibrium with the forcing in year 2000, as argued by SES. The warming commitments in 2080–2099 (shown as dots at the right of Figure 1) relative to 2000 (taken as the 1990–2009 average here to reduce noise) is 0.30 K on average (0.11 K for the lowest model), i.e., a factor of ten (three for the lowest) larger than the 0.03 K given by SES. Another argument used by SES is that a short response time scale is supported by a relatively quick response and recovery to volcanic eruptions. The black solid in Figure 1b demonstrates that the response to Pinatubo is indeed short in those models, but it provides little information about the long time scale in the commitment warming. The reasons are the short and abrupt nature of the forcing, and the different processes involved in the response.

[7] For the climate change problem, in order to achieve stabilization of global temperature, the relevant response time scales are those of the deep ocean, and the short time scales found by SES are therefore irrelevant to the problem of estimating climate sensitivity. The argument of abrupt temperature shifts in glacial periods is misleading, because these were local or regional warming events caused by a change in the ocean thermohaline circulation and sea ice, with little or no signal in global temperature. The signal of volcanic eruptions is also known to provide little constraint on the time scale of interest and on climate sensitivity

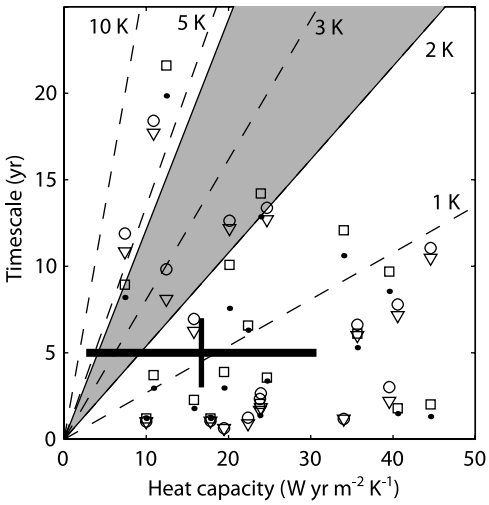

Figure 2. Time scales and heat capacities calculated for 17 CMIP3 AOGCMs following the method proposed by SES. Heat capacities are computed from linear trends in surface temperature divided by linear trends in total Earth energy content (calculated from top of atmosphere imbalances). Time scales are estimated as τ_{max}, the maximum of the autocorrelation time constant (squares), and the mean of the five highest τ values (denoted τ_5 here, dots) both from the transient 20th century simulation (1880–2000) (see text for details). Additionally, τ_{max} (circles) and τ_5 (triangles) are calculated from long unforced control simulations of the corresponding models (the first 50 years of the controls are not used because of spin up effects, linear detrending is used afterward). Constant lines of climate sensitivity (for doubling of CO_2) are given as dashed lines. The "likely" range of climate sensitivity 2–4.5 K, which at the same time is the range all models should fall into, is shaded in gray. The ± 2 standard deviation ranges for the heat capacity and time scale estimate by SES are shown as thick solid lines for comparison.

[*Frame et al.*, 2005; *Wigley et al.*, 2005], because the dependence of the short-term response on sensitivity becomes weaker as shorter time scales are considered, and insensitive to high climate sensitivity values [*Wigley and Raper*, 1990; *Knutti et al.*, 2005].

3. Implications for Estimating Climate Sensitivity

[8] An obvious test for the method proposed by SES is to apply it to the simulation of the 20th century by AOGCMs and compare the estimated sensitivity with the known sensitivity of the models, determined from simulations in which atmospheric CO_2 is doubled. The 20th century simulations used here include most forcings (greenhouse gases, aerosols, ozone, volcanic and solar), but the forcing magnitudes vary across models. We follow the method by SES but use the integrated top of atmosphere net radiative flux instead of ocean temperature to determine the change in the total heat content of the Earth. This provides a better estimate of total heat content change, as it avoids the additional step of correcting for warming of the land and cryosphere, but otherwise it is irrelevant to the conclusions, since the problem lies in the time scale rather than the heat capacity. Seventeen models provide the necessary data, and the results are summarized in Figure 2. The estimated effective heat capacities C of the models are in the range of 7 to 45 W a m^{-2} K^{-1} (mean: 24 W a m^{-2} K^{-1}, standard deviation 11 W a m^{-2} K^{-1}), in reasonable agreement with the estimate of SES from observations (16.7 ± 7 W a m^{-2} K^{-1}). These figures are based on linear trends over the period 1950–2000: it is important to stress that the effective heat capacity of the climate system is likely to depend on the details of the time history of the forcing over the period of interest, which introduces a further source of uncertainty into the analysis. A few models are somewhat on the high end, in agreement with previous analysis that on average, models tend to overestimate observed ocean heat uptake [*Forest et al.*, 2006] compared to *Levitus et al.* [2005]. However, as discussed in section 4, the uncertainties in observed ocean heat uptake are likely to be larger than indicated by *Levitus et al.* [2005].

[9] Following SES, we calculate the time scale τ from the autocorrelation of the linearly detrended time series of the transient temperature 1880–2000. However, this is found to be rather difficult in some models. In a few cases, the time scale τ (Δt) increases with the lag Δt, and in two cases the autocorrelation at lag 2 years is already negative. *Foster et al.* [2007] point out that the assumption of global temperature being an AR(1) process is not supported by the observations, and the models analyzed here support that conclusion. In Figure 2, we show the maximum τ_{max} of the time scale τ (taken from the series of τ (Δt) versus lag Δt up to the first negative autocorrelation, as in SES) as well as the mean of the five highest values of τ (Δt) (denoted τ_5 here) to show that the results are sensitive to how the time scale τ is calculated.

[10] Linear detrending of global temperature is problematic, since the external radiative forcing was not linear over the 20th century. The time scales might also be influenced by volcanic eruptions. Also, the time series is short, which leads to an uncertain and biased estimate of the time scale [*Foster et al.*, 2007]. To partly circumvent these problems, long unforced control simulations from the corresponding models were therefore used additionally to estimate the time scale τ. Again, τ_{max} and τ_5 were calculated. The first 50 years of the control simulations were not used, since many models show drift in the early phase of the simulation. Linear detrending was used for the rest of the control simulation to remove remaining small drifts. Control simulations on average are 350 years long after removing the first 50 years.

[11] From Figure 2 it is already evident that most model estimates are outside the "likely" range of climate sensitivity of 2.0–4.5 K [*Meehl et al.*, 2007] (shaded in gray), which is close to the range of 2.1–4.4 K of the true sensitivities spanned by the AOGCMs. Lines of constant climate sensitivity (1, 2, 3, 5 and 10 K, for CO_2 doubling) are indicated for comparison.

[12] The predicted climate sensitivity S = τ/C of the AOGCMs versus true sensitivity are shown in Figure 3, using the 20th century (top) and control (bottom) simulations. This demonstrates that the method proposed by SES to calculate climate sensitivity underestimates the model climate sensitivities by at least a factor of two, no matter which time scale τ and which simulation are used. Corre-

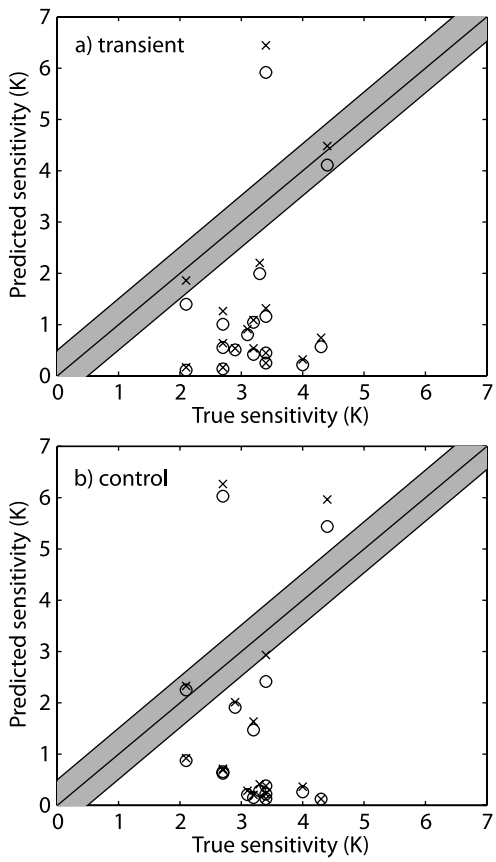

Figure 3. Predicted climate sensitivity versus true climate sensitivity from 17 CMIP3 AOGCMs using the method proposed by SES. (a) Results using the transient 20th century simulations (1880–2000) using the maximum of the autocorrelation time constant τ_{max} (marked with crosses) and the mean of the five highest τ_5 values (circles). (b) Same but for long unforced control simulations from the corresponding simulations rather than the 20th century to estimate the time scales. According to the uncertainty estimated by SES (0.5°C, one standard deviation), and if the method had skill in predicting climate sensitivity, two thirds of the models should fall into the gray area.

lation between the predicted and true sensitivities is lower than 0.1 (r^2) in all cases, demonstrating that the method neither has skill in predicting the true values of climate sensitivity of the models nor is able to discriminate between high and low sensitivities.

[13] The method is shown to fail even in the idealized model situation where the time series of heat uptake and surface warming are perfectly known (no observational uncertainties), and where a control time series of surface temperature of several hundred years without forcing (volcanoes) or trends is available. As explained in section 2, the problem is that the time scale estimated from autocorrelation is too short, and has little if any relevance to the response time scale of interest.

[14] So the obvious question to be asked is the following: Does the time scale τ, even if too short, provide any information at all about climate sensitivity? Could there be processes or parameterizations in the models that affect both the interannual and the deep ocean time scale, such that even if the two are different, we could learn about the deep ocean time scale from observing the interannual time scale? One may argue that the number of CMIP3 models is small, and their sensitivity range is rather narrow to answer that. In Figure 4, the time scales τ_{max} and τ_5 are shown for 203 models from the climate*prediction*.net (http://www.climateprediction.net) "BBC Climate Change Experiment" coupled ensemble [*Allen*, 1999; N. E. Faull et al., The climateprediction.net BBC climate change experiment, part 1: Design of the coupled model ensemble, submitted to *Proceedings of the Royal Society A*, 2008]. Each one of these models is a version of the coupled HadCM3 AOGCM, with different parameter settings in the ocean and atmosphere, and they cover a very wide range of sensitivities between zero and 9 K [*Stainforth et al.*, 2005]. Spin-ups of the ocean and atmosphere are done separately, and the models are coupled using flux corrections. The time scales are calculated as before, from 160-year unforced control simulations. The first 20 years of the simulation are ignored to avoid large biases related to drift, linear detrending is used for the remaining 140 years. Irrespective of how the time scale is calculated, the correlation is less than 0.01 (r^2), suggesting that the time scale τ estimated from interannual variations provides no information about the model's climate sensitivity, and contradicting the linear relationship between the time scale τ and climate sensitivity, $S = \tau/C$, assumed by *Schwartz* [2007].

[15] The complete lack of any relationship between autocorrelation time and sensitivity in Figure 4 is surprising. It may simply arise from noise in the correlation time scale estimates, so further extension of the climate*prediction*.net ensemble, which is ongoing, may eventually reveal some relationship. While we would not expect a simple relationship as assumed by SES, we may expect some link to

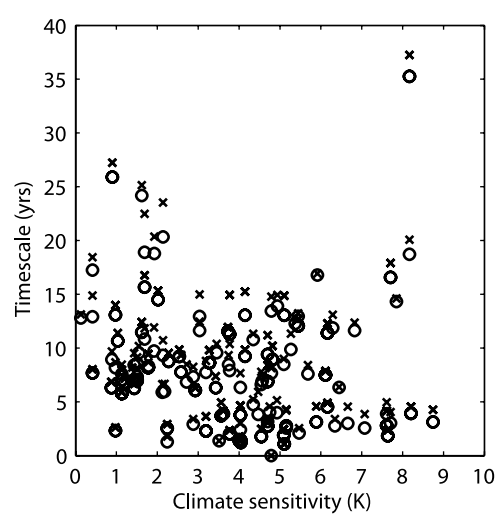

Figure 4. Characteristic time scale calculated from interannual variability as proposed by *Schwartz* [2007] versus climate sensitivity for 203 climateprediction.net control simulation. A wide range of climate sensitivities is covered by varying parameters in the atmosphere in each model version. Circles denote the maximum time scale τ_{max}, and crosses denote the mean of the five highest time scale values τ_5 for each model (see text). In both cases, correlation between the time scales and climate sensitivities is negligible.

emerge on physical grounds. If, however, the link is so weak that it takes many thousands of simulated years for it to emerge at all, then the prospects for using this to constrain sensitivity in the real world are bleak. An additional analysis of 100 year segments of long AOGCM control simulations show that the estimate of the autocorrelation time scale can vary by a factor of three depending on what segment is chosen within the control, suggesting that the estimate of the autocorrelation time scale is not robust even within one model, in agreement with the results of *Foster et al.* [2007].

4. Treatment of Uncertainties

[16] SES quote a symmetric range of 1.1 ± 0.5 K for climate sensitivity, calculated from S = τ/C. However, if the quoted values for τ = 5 ± 1 years and C = 16.7 ± 7 W a m^{-2} K^{-1} are assumed to be normal distributions, as implied by the analysis of SES, then S, the ratio of the two is not a normal distribution but skewed, with a long tail toward high values. While the ratio of the means of τ and C suggest a climate sensitivity of 1.11 K (assuming a forcing of 3.71 W m^{-2} for doubling CO_2), the most likely value of the distribution of S is in fact 0.88 K, with a 5–95% range ("very likely" in the IPCC terminology) of 0.55–3.32 K. SES ignores that his estimate of climate sensitivity, although biased toward low values because of the erroneously short time scale, yields a skewed distribution that extends well into the "likely 2–4.5 K" range quoted by IPCC [*Meehl et al.*, 2007], as shown by the thick solid lines in Figure 2 overlapping the gray area. The reason for this large effect here is that the uncertainty of the numerator τ is small, but is large for the denominator C, leading essentially to an inverse Gaussian distribution [*Roe and Baker*, 2007].

[17] A number of recent studies have pointed out that the uncertainty in the ocean heat uptake estimates may be larger than previously thought. These are related to problems in instrument calibration, changes in instrument types over time, poor sampling coverage in many areas as well as interpolation schemes [*Gregory et al.*, 2004; *AchutaRao et al.*, 2006; *Gouretski and Koltermann*, 2007; *Harrison and Carson*, 2007]. The decadal variations in ocean heat uptake are poorly understood, not well simulated in models, and may be partly caused by interpolation of the sparse data [*Gregory et al.*, 2004; *AchutaRao et al.*, 2006]. This suggests that the uncertainty in the effective heat capacity C, and therefore the uncertainty in climate sensitivity, may be larger than assumed by SES. While the sensitivity of a climate model at equilibrium is independent of the ocean heat uptake, the effective sensitivity estimated from the transient behavior of the model does depend on the ocean heat uptake.

[18] Figure 2 also shows that the uncertainty in the estimated time scale τ is larger (indicated by the spread of the four different symbols for a given heat capacity of one model) than estimated by SES. On average, the standard deviation of the four estimates of τ used here is about 3 years for a given model, i.e., a factor of three larger than estimated by SES from the observed warming. As noted above, additional tests performed on nonoverlapping segments taken from the long control runs of the AOGCMs also show that the estimated time scale can vary by a factor of three between different segments of the same model simulation.

[19] Taken together, we conclude that the uncertainties in both the heat capacity and the time scale, as well as the skewed distribution are not adequately captured in the analysis of SES.

[20] Finally, while quoting the IPCC "likely" range of 2–4.5 K [*Meehl et al.*, 2007], SES fails to mention the fact that many earlier studies have used the observed warming of ocean and atmosphere to constrain the range of climate sensitivity [*Andronova and Schlesinger*, 2001; *Forest et al.*, 2002; *Knutti et al.*, 2002, 2003; *Frame et al.*, 2005; *Forest et al.*, 2006; *Tomassini et al.*, 2007] and, in some of those, the value of total as well as aerosol forcing. While some of these studies do not completely rule out a climate sensitivity of 1 K, they show that such a low value compares poorly with observed trends in ocean heat uptake and surface temperature, and is therefore very unlikely. Many of these studies are based on more realistic climate models, and most attempt to account for uncertainties in forcing, climate feedbacks and ocean heat uptake, and give distributions of climate sensitivity that are consistent with those estimated from AOGCMs and paleoclimate data [*Murphy et al.*, 2004; *Annan et al.*, 2005; *Piani et al.*, 2005; *Knutti et al.*, 2006; *Schneider von Deimling et al.*, 2006]. SES fails to discuss this large body of literature and to provide evidence why all existing methods should overestimate climate sensitivity by about a factor of three.

5. Conclusions

[21] We have demonstrated that the existence of a single time scale in global temperature, which at the same time characterizes internal variability as well as the response to various types of forcing, is not supported by physical climate models. Both physical understanding of the climate system as well as models support the existence of a series of time scales related to different processes. The single time scale determined from the autocorrelation analysis by SES is biased toward short-term atmospheric and oceanic mixed layer processes and, therefore, has little relevance to the much longer term climate change problem (Figure 1). For that reason, the method proposed by SES to estimate climate sensitivity, when applied to a series of AOGCMs, is shown to underestimate the known climate sensitivities of the models by at least a factor of two (even when using the assumptions most favorable for SES), and is shown to have no skill in discriminating between high and low sensitivities (Figure 3). The analysis of a large ensemble of AOGCMs from the climateprediction.net project suggests that the time scale as estimated by *Schwartz* [2007] does not relate to climate sensitivity at all (Figure 4). While there is definitely value in using observed trends to constrain climate sensitivity, we conclude that the method proposed by SES fails to yield any quantitative information about the Earth's climate sensitivity and the response time scales relevant to the climate change problem.

[22] In his reply, *Schwartz* [2008] proposes revised methods to estimate the response time scale of the system (his equations (5) and (6)), but based on essentially the same arguments. We applied that method to all GCM control simulations and find that correlation is still insignificant, and that the revised method has no more skill in predicting climate sensitivity than the original one, even in the optimal situation of a long control simulation without forcing and

trend, and in the absence of sampling uncertainty or measurement biases. Since the method fails in an ideal case of a GCM control, it is unclear why it should work either in more realistic model studies, or in the real world. Note that Schwartz's method ought not depend on the details of the actual geophysical system under study, such as the rate of ocean heat uptake or the degree of interannual or seasonal variability: the methods proposed by Schwartz make no assumptions about the real physical system, and (if they were indeed applicable) should be able to determine the climate sensitivity of a climate model with any set of model parameters, or even for example a world without ENSO, without sea ice, or for an aquaplanet without continents. Thus from the four possible reasons to explain the "major discrepancies between application of the approach of SES to observed and modeled climate data" identified in the reply [*Schwartz*, 2008], three are irrelevant. To estimate a model sensitivity from model output, (1) errors and uncertainties in observations, (2) shortness of the observed record and (4) inaccuracy of the modeled quantities are not of interest. The only remaining reason is point 3 identified in the reply [*Schwartz*, 2008]: there are indeed "Inherent flaws in the approach to the inference of climate system time constant from autocorrelation analysis": there is no such thing as a single climate system time constant.

[23] **Acknowledgments.** We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI), and the WCRP's Working Group on Coupled Modeling (WGCM) for their roles in making available the WCRP CMIP3 multimodel data set. Support of this data set is provided by the Office of Science, U.S. Department of Energy. Participants of the climate*prediction*.net projects are acknowledged for donating their computer time to make that project happen.

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