Constraints on the transient climate response from observed global temperature and ocean heat uptake

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Abstract

Projections of future transient global temperature increase in climate models for a known forcing depend on the strength of the atmospheric feedbacks and the rate of transient ocean heat uptake. A Bayesian framework and an intermediate complexity climate model are used to calculate a probability density function (PDF) of the transient climate response (TCR), constrained by observations of global surface warming and ocean heat uptake. The PDF constrained by observations is wider than the TCR range of current climate models, and has a slightly lower mean. Uncertainties in the observed ocean warming are shown to potentially affect the TCR. It is proposed, however, that even if models were found to overestimate ocean heat uptake, correcting that bias would lead to revisions in surface temperature projections over the twenty-first century that are smaller than the uncertainties introduced by poorly quantified atmospheric feedbacks.

1. Introduction

Uncertainties in the projected increase in global surface temperature for a given emission scenario on timescales of a century depend on uncertainties in climate sensitivity (the global mean equilibrium surface temperature response to a doubling of atmospheric carbon dioxide), uncertainties in the transient ocean heat uptake, in the carbon cycle, and in radiative forcing [*Knutti, et al.*, 2002]. Probability density functions (PDFs) for climate sensitivity mostly show skewed distributions with a non-zero probability for high values [e.g. *Knutti, et al.*, 2002; *Forest, et al.*, 2006], caused by the nonlinear relation between observable quantities (or the transient model response) and climate sensitivity. While the likelihood of high climate sensitivities is an issue of current debate [*Annan and Hargreaves*, 2006], projections on a timescale of a few decades are more linearly related to the observed surface temperature trends and thus better constrained [e.g. *Frame, et al.*, 2006; *Stott, et al.*, 2006]. Thus a more relevant measure to consider for century timescale projections is the transient climate response (TCR, the global mean temperature response at the time of CO_2 doubling (year 60-80) in a 1%/yr CO_2 increases scenario. Note that TCR is a benchmark to compare model responses for an idealized scenario, not a projection. However, since the radiative forcing increases relatively linearly in non-intervention scenarios and since the relative uncertainty (i.e. the spread divided by the mean) is approximately constant in those scenarios [*Knutti, et al.*, 2007], changes in the PDF of TCR transfer linearly to SRES scenario projections. Direct estimates of uncertainties in scenario projections include carbon cycle feedback and radiative forcing uncertainties [*Wigley and Raper*, 2001; *Knutti, et al.*, 2003] in addition to TCR, making it difficult to disentangle the different uncertainty contributions. TCR is thus the most useful quantity to separate the uncertainties in the physical response from those in radiative forcing and the carbon cycle.

2. A PDF of the transient climate response

We use a very large ensemble of simulations with a climate model of intermediate complexity (EMIC) to derive a probability density function for TCR. PDFs of model parameters like climate sensitivity, ocean diffusivity and radiative forcing scaling factors are constrained by the observed global surface warming and global ocean heat uptake. A uniform prior for climate sensitivity of 1 - 10°C is assumed in the standard case. The method is identical to the one used by *Tomassini et al.* [2007] (see supplementary material). The resulting PDF of TCR, conditional on that model and the observations used, is shown as a thick solid line in Fig. 1. It has a most likely value of 1.6°C, a median of 1.58°C and a 5-95% percent range of 1.11-2.34°C, and is only slightly narrower when using an expert prior on climate sensitivity (thick dashed). The individual values for TCR of the WCRP CMIP3 (World Climate Research Programme Coupled Model Intercomparison Project phase 3) coupled atmosphere ocean general circulation models (AOGCMs) are shown as dots for comparison. The mean TCR of all AOGCMs is 1.8°C, close to the most likely value determined from this probabilistic method. *Stott et al.* [2006] derived PDFs for TCR for three AOGCMs using the detection analysis [*Stott and Kettleborough*, 2002]. They use spatial and temporal patterns to estimate the warming attributable to greenhouse gases (GHG), the cooling caused by aerosol forcing, and the component of natural variability. Based on simple model results, the GHG attributable warming and TCR are related linearly [*Allen, et al.*, 2000; *Frame, et al.*, 2006], such that the Stott method does not use observations of ocean heat uptake. The PDFs are shifted towards higher TCR values by about half a degree and are also higher than most of the raw TCR values of the AOGCMs, suggesting that many models might underestimate projected future warming. The TCR range derived from the *Forest et al.* [2006] method [*Stott and Forest*, 2007] lies in between the two and is more consistent with our results.

The TCR ranges estimated here are consistent with previous estimates but somewhat lower for three reasons. First, the PDF of climate sensitivity from our method tends to be on the low side; therefore the TCR values are also lower. Second, a high warming (high TCR) can lead to a strong reduction or shutdown of the Atlantic meridional overturning circulation (MOC) in some simulations, resulting in a higher transient ocean heat uptake and sea level rise and an associated smaller transient surface warming [*Knutti and Stocker*, 2000] compared to a case where the MOC changes only little. Two simulations with identical parameters are shown in Fig. 2a/b to illustrate that effect. The one that was additionally perturbed with freshwater in the North Atlantic, which reduces the MOC, warms much slower, although their equilibrium warming is the same. Therefore, the essentially linear scaling between the GHG attributable warming and TCR used in *Stott et al.* [2006] (see Fig. 2 in [*Allen, et al.*, 2000]) may slightly overestimate the probability for high TCR. Our results suggest that there is more spread in that scaling relationship (Fig. 2c), and a negative feedback caused by the MOC in dynamic ocean models that can lead to TCR values up to half a degree lower. GCMs have not explored this hypothesis so far, but additional simulations with a zero dimensional energy balance model are consistent with it and show that TCR is reduced by a up to half a degree C if the upwelling velocity is chosen to be temperature dependent to account for changes in deepwater formation. Third, differences arise from the use of different priors. A more detailed discussion of the various methods is given in the supplementary material. In summary, it is important to recognize that different methods, assumptions, datasets and types of models can lead to results that are significantly different, although consistent within their uncertainties. Accounting for structural uncertainty and separating the effects of different dataset would be an important but non-trivial next step to further isolate the differences.

3. The effect of ocean heat uptake on TCR

The TCR depends on climate sensitivity and the transient heat uptake of the ocean. In contrast to the atmosphere, where observations are abundant, the evaluation of transient changes in the ocean is more difficult. While warming is evident in all ocean basins [*Levitus, et al.*, 2005], data coverage is poor at greater depth, in certain areas and earlier in time [*Harrison and Carson*, 2007]. The record is short to evaluate trends, short-term variability is poorly understood, and trends and variability are found to be sensitive to spatial infilling of data [*Gregory, et al.*, 2004; *AchutaRao, et al.*, 2006]. A recent study suggested that ocean warming might be overestimated due to instrument related biases [*Gouretski and Koltermann*, 2007]. Despite the sensitivity of estimates of heat uptake to these issues, recent reconstructions [e.g. *Ishii, et al.*, 2006] agree with the newer *Levitus et al.* [2005] estimates for the upper 700 m depth range. The Levitus data therefore remain at least among the best current estimates of ocean heat uptake, but future revisions may still be significant.

Some ocean models agree well with observed global trends and patterns in ocean heat uptake [*Barnett, et al.*, 2001] at least in the upper 700m. Other models tend to overestimate anthropogenic heat uptake [*Gent, et al.*, 2006], but the spread between different ensemble members can be large [e.g. *Gregory, et al.*, 2004]. *Forest et al.* [2006] suggested that an earlier generation of AOGCMs transported heat too effectively below the mixed layer. The discrepancies were found to be smaller after the newer MIT model (*Forest et al.* [2006]) was calibrated to the newer generation of AOGCMs (see *Stott and Forest* [2007]), but still indicate too effective mixing in the models. The raw heat uptake in the latest generation also tends to be higher [*Hegerl, et al.*, 2007] than observed by *Levitus et al.* [2005], but could be consistent within the uncertainties. Whether and by how much ocean models overestimate the observed anthropogenic ocean heat uptake is not the focus of this study. Indeed, it might be too early to pin this down given the uncertainties in the observations, interpolation issues and problems with instrument biases. It is, however, instructive to estimate the effect of a potential bias in ocean heat uptake on TCR, in order to quantify the robustness of current temperature projections for scenarios [*Knutti, et al.*, 2007; *Meehl, et al.*, 2007].

The dependence of TCR (from a 1%/yr CO_2 increase scenario) on global ocean heat uptake to 700m and 3000m (from a 20th century scenario) is shown in Fig. 3a and 3b, respectively. For each value of S (different symbols), values of ocean vertical diffusivity (K) from 2 to 15 $\cdot$ 10^{-5} m^2s^{-1} (increments of 0.5 $\cdot$ 10^{-5} m^2s^{-1}) are connected by a line. Standard values for all forcings are used in each simulation, no observational constraints are used. As expected, a higher K increases the overall ocean heat uptake (Fig. 3b) for a given S, because the heat is more effectively mixed into the deep ocean. For the upper 700 m (Fig. 3a), the dependence is small for low S, and even slightly reversed for high S, resulting in the somewhat counterintuitive increased heat uptake for low diffusivities. The explanation is that while reducing diffusivity reduces the total heat uptake, it may increase the heat uptake near the surface if the atmospheric warming is strong (high S), because the transport of heat below the mixed layer is strongly reduced.

Note that Fig. 3 shows linear trends in heat content for simplicity and to be comparable to *Levitus et al.* [2005]. While this figure might suggest that a climate sensitivity of 1.5°C gives the best match to the observed ocean heat uptake, this is not true if the agreement of model and data is estimated over the full time series and if surface warming data is included. The best agreement with both surface warming and ocean heat uptake, including the decadal variations, is found for S = 2.49°C (see Fig. 5 of *Tomassini et al.* [2007]). Fig. 3 should be interpreted as a sensitivity analysis of TCR with regard to ocean heat uptake rather than an attempt to constrain S or K, which was done previously by *Tomassini et al.* [2007].

Five main conclusions emerge from this paper. First, ocean heat uptake is not very sensitive to the value of K, at least in our model. Heat uptake is more strongly influenced by S (i.e. the atmospheric warming), in agreement with earlier results [*Raper, et al.*, 2002]. This also seems consistent with the TCR range by *Forest et al.* [2006] which, despite their claim that most of the AOGCMs mix heat too effectively in the deep ocean, is remarkably similar to the raw AOGCM TCR values. Second, in our model, reducing heat uptake in the upper layers for a constant S would require an *increase* in K in order to mix heat more efficiently into the deep ocean, which would *decrease* TCR, at least for high S. Reducing heat uptake in the whole ocean, however, would require a *decrease* in K, resulting in an *increase* of TCR. Since the uncertainties are largest in the deep ocean, there is more concern that models may overestimate deep ocean heat uptake. On the other hand, confidence in the observed heat uptake for the deep ocean is much lower than for the upper ocean. But it is important to note that the conclusions may depend on the dataset that is used. Third, as shown in the schematic illustration in Fig. 3c, for a given reduction in total heat uptake (‘high Lev’ to ‘low Lev’), the absolute change in TCR would be larger if S is large (shift from open circle to open square) than if S is small (shift from filled circle to filled square). For medium values of S (e.g. 2.5°C), the expected change in TCR caused by a reduction in the observed heat uptake of 20% for example is relatively small, probably less than half a degree. For constant S, this might increase the most likely projected warming for high SRES scenarios by about a degree, which is significant, but well within the range of uncertainties estimated by various methods [*Knutti, et al.*, 2007]. Fourth, however, if in a Bayesian framework the evolution of surface warming is used additionally to constrain the model, the effect would be smaller. The reduction in ocean heat uptake would then likely suggest a reduction of climate sensitivity at the same time to remain consistent with the observed surface warming. The filled square in Fig. 3c would therefore shift downward to a line of lower sensitivity (dashed), thus reducing TCR to near its original value (cross). This is similar to the argument used in the detection method [*Frame, et al.*, 2006; *Stott, et al.*, 2006], in which TCR is linearly related to the observed GHG attributable warming, independent of the observed heat uptake. Fifth, however, previous methods based on ocean models that diffuse temperature anomalies into depth (without temperature dependent deepwater formation rates) [*Forest, et al.*, 2006] or based on scaling AOGCMs based on observed surface warming trends [*Stott, et al.*, 2006] may not fully capture the negative feedback of the ocean MOC slowdown, which can act to reduce the transient surface warming. This effect is particularly effective for high warming and on timescale beyond a few decades, but may also be strongly model dependent.

4. Conclusions

A PDF of TCR derived from a very large ensemble of model simulations and constrained by global surface and ocean warming is presented. It is wider than the distribution of TCR values in the latest generation of AOGCMs. Some AOGCMs show a larger heat uptake [*Hegerl, et al.*, 2007] or heat uptake efficiency [*Forest, et al.*, 2006] than measured by *Levitus et al.* [2005]. In our view, however, limitations and large uncertainties in the observational dataset and limitations in the ocean models prevent a definitive conclusion as to whether current AOGCMs are inconsistent with the observed ocean heat uptake. Reducing ocean model heat uptake would likely cause projections over the twenty-first century to be revised upwards, if all other things were kept equal. The sensitivity of TCR to ocean heat uptake, however, suggests that those changes in scenario projections should be relatively small compared to the overall uncertainty range. Other constraints (e.g. the patterns of surface warming) will tend to counteract these changes, such that the current uncertainty range of projections based on a variety of methods and models and constrained by both oceanic and surface warming should be robust [*Knutti, et al.*, 2007; *Meehl, et al.*, 2007] against future small revisions in the ocean heat budget.

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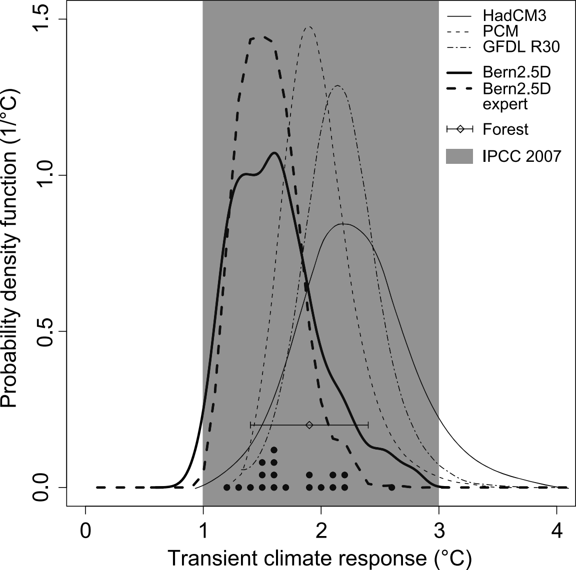
Fig. 1: Probability density function (PDF) of the transient climate response from a large ensemble of model simulations with the Bern2.5D model constrained by observed surface warming and ocean heat uptake, for a uniform (thick solid) and expert prior (thick dashed) distribution of climate sensitivity. PDFs for three AOGCMs (HadCM3, PCM and GFDL R30) obtained with the detection algorithm [*Stott, et al.*, 2006], the range from another EMIC[*Forest, et al.*, 2006; *Stott and Forest*, 2007], individual values from the CMIP3 models (symbols) and the IPCC AR4 range 1 – 3°C (10 – 90% confidence, grey band) [*Meehl, et al.*, 2007] are given for comparison.

Fig. 2: a) Atlantic meridional overturning circulation (MOC) and b) global temperature in two TCR simulations with identical model parameters. Radiative forcing increases linearly to 3.71 Wm^{-2} between year 0 an 70 and is constant thereafter. For the dashed case, freshwater discharge into the North Atlantic increases linearly to 0.2 between year 0 and 70, and is zero thereafter. No freshwater perturbation is added in the solid case. The MOC reduction results in an increased ocean heat uptake and a reduced transient surface response. c) Relation of linear warming trend 1950-2000 induced by CO_2 only and TCR for different ocean diffusivities, climate sensitivities and mixing parameterizations. Crosses mark cases with freshwater perturbations to reduce the MOC (as in panel a, dashed), circles mark those without perturbations.

Fig. 3: a) Relation between TCR and the observed ocean heat uptake to 700m (linear trend 1955 to 2003, annual values), b) same but heat uptake to 3000m (linear trend 1955-59 to 1993-98, pentadal values). Each value of climate sensitivity (S) is indicated by a

different symbol, different values of the ocean vertical diffusivity (K) are connected by a line. Linear trends from *Levitus et al.* [2005] are given as solid vertical lines, with plus minus two standard deviations uncertainties as dashed vertical lines. c) schematic illustration of the effect of a decrease in ocean heat uptake (‘high Lev’ to ‘low Lev’) for low S (filled circle to filled square) and high S (open circle to open square), and the effect of changing K and S at the same time (filled circle to cross). See text for details.



HadCM3
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Probability density function (1/°C)
Transient climate response (°C)
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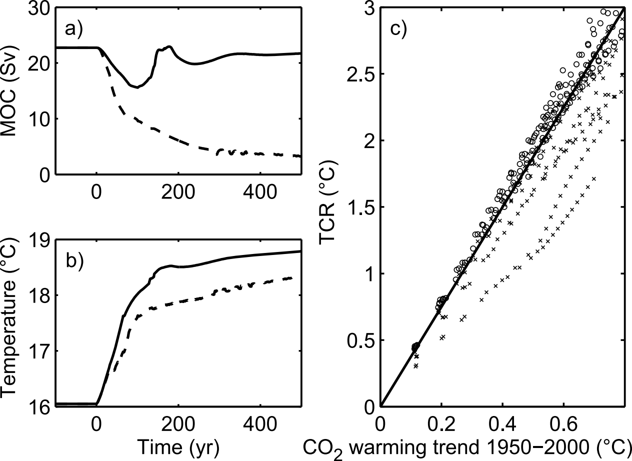
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Temperature (°C)
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Time (yr)
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TCR (°C)
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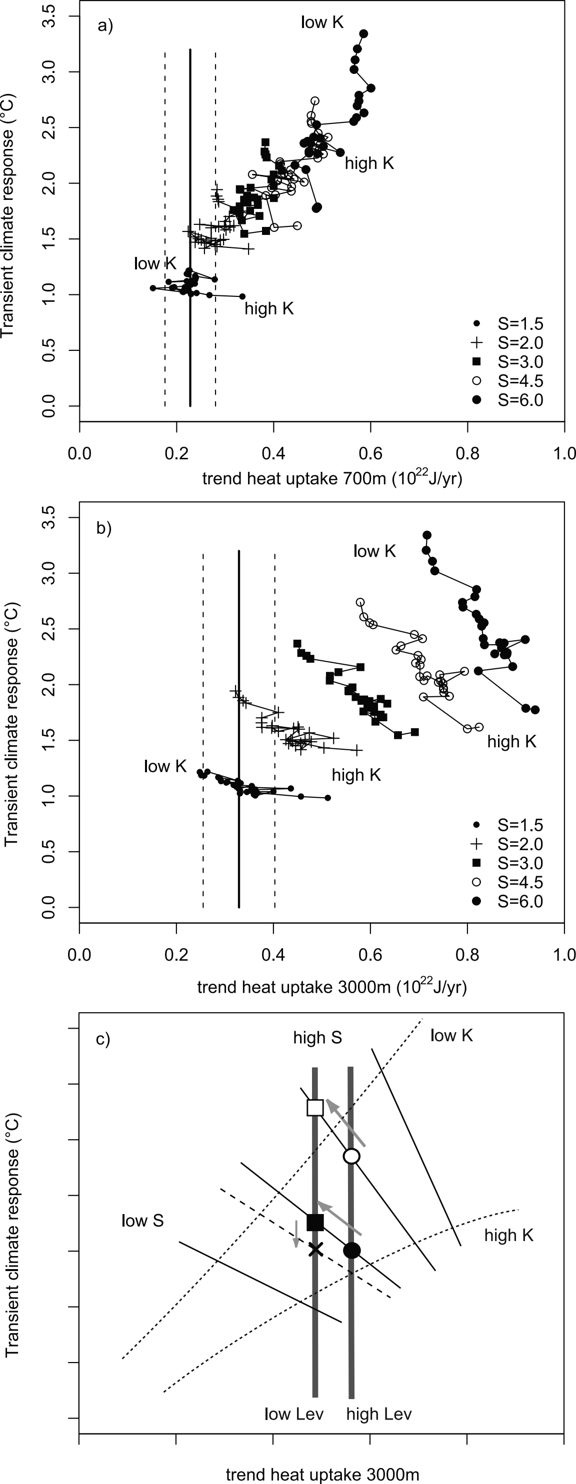
CO_2 warming trend 1950−2000 (°C)


a)
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b)
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Transient climate response (°C)
trend heat uptake 3000m (10^{22}J/yr)
c)
high S
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high Lev
Transient climate response (°C)
trend heat uptake 3000m