Constraining the Sensitivity of Regional Climate with the Use of Historical Observations

APOSTOLOS VOULGARAKIS AND DREW T. SHINDELL

NASA Goddard Institute for Space Studies, and Center for Climate Systems Research, Columbia University, New York, New York

(Manuscript received 15 January 2010, in final form 26 July 2010)

ABSTRACT

A novel method is presented for calculating how sensitive regional climate is to radiative forcings, based on global surface temperature observations. Forcings that originate in both the region of interest and outside of it are taken into account. It is found that the transient temperature sensitivity parameter (β, defined as the observed temperature response per unit forcing) can be better constrained for 50°S–25°N than for the rest of the globe. The average β in this region is 0.35°C (W m^{-2})$^{-1}$. The models used in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) 1% yr^{-1} CO_2 increase experiment exhibit a β in this region that, on average, is higher by 35%. The results show that for 50°S–25°N β may provide a more valuable constraint for model evaluation than global mean climate sensitivity.

1. Introduction

One of the aspects of the climate system that is not well constrained is climate sensitivity: the degree to which surface temperature responds to radiative forcings imposed by human activities or due to natural changes. The optimum way to calculate climate sensitivity is to use observational data, though such data have substantial limitations. Evaluation against those results provides an important test for climate models. Most studies so far have examined the global equilibrium climate sensitivity (ECS) or the transient climate response (TCR). The former refers to the "global annual mean surface air temperature change experienced by the climate system after it has attained equilibrium in response to a doubling of atmospheric CO_2" (Randall et al. 2007). The latter is again the temperature change in a model at the time of CO_2 doubling but following a gradual (1% yr^{-1}) increase in CO_2 concentrations.

The recent efforts of scientists have concentrated on narrowing the range of the estimates of global ECS. Its calculation remains highly uncertain, with observationally based studies (e.g., Andronova and Schlesinger 2001; Gregory et al. 2002; Forest et al. 2002; Frame et al. 2005; Hegerl et al. 2006; Meehl et al. 2007; Gregory and Forster 2008; Knutti and Hegerl 2008) finding the low end of the most likely range being 1.5°C and the high end extending above 4.5°C, although such high values are less likely (Meehl et al. 2007). Uncertainty in observational estimates of ECS or TCR is comparable to estimates using models (Stott et al. 2006; Knutti and Tomassini 2008; Forest et al. 2008; Knutti and Hegerl 2008).

Although there have been many studies looking at global climate sensitivity, there has been little focus on constraining the sensitivity of regional climate to radiative forcings. Here, we use global surface temperature observation datasets to extract useful information on the temperature sensitivity for different latitudinal zones of the globe. Though the influence of variability becomes larger at smaller spatial scales, the influence of the most poorly known historical forcings can be small in some regions, making it worthwhile to explore how well regional sensitivity can be constrained compared with global sensitivity. The topic is rather complex, as regional climate can respond to both local and nonlocal forcings (Boer and Yu 2003). Our approach takes advantage of the recently published analysis by Shindell and Faluvegi (2009, hereafter SF09) of the response of zonally averaged temperatures to local and nonlocal zonally perturbed forcings. The main aim of this manuscript is to describe the method and the way in which

Corresponding author address: Apostolos Voulgarakis, NASA Goddard Institute for Space Studies, 2880 Broadway, New York, NY 10025.
E-mail: avoulgarakis@giss.nasa.gov

DOI: 10.1175/2010JCLI3623.1

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uncertainty is assessed (section 2) and to make an initial attempt to find regions of the globe where sensitivity is better constrained (section 3). We extend the study by comparing the results of our method to estimates from GCM experiments.

2. Methods

We focus on calculating the *regional transient temperature sensitivity parameter*, which we symbolize as β and define as the temperature response per unit forcing for each 4°–wide latitude band examined. Our parameter is similar to the *climate sensitivity parameter* defined in Ramaswamy et al. (2001), except that in our case we refer to a regional and transient climate state instead of a global and equilibrium climate state. We avoid using the term "climate sensitivity," as it is widely associated with changes following the global climate system reaching equilibrium. We also avoid using the term "transient climate response," as the latter is associated to a $2 \times CO_2$ perturbation. Note that β inherently includes the ocean heat uptake changes that occurred during the period of interest, so that we do not need to separately estimate this quantity.

To obtain β for various parts of the globe, we use global surface temperature data from the Goddard Institute for Space Studies (GISS) analysis (Hansen et al. 2006), updated for southern polar latitudes (Steig et al. 2009). For comparison, we also calculate β using the global surface temperature dataset (HadCRUT3) provided by the Climatic Research Unit (CRU) United Kingdom (Brohan et al. 2006). Linear regression analysis is applied to calculate zonal mean temperature trends for 1957–2003, the period for which data coverage is largely spatially complete, and for 1880–2003, the period for which we have comparable climate model simulations. We use zonal mean adjusted forcings ("adjusted" meaning that the stratosphere is allowed to adjust radiatively to the presence of the forcing agent) from well-mixed greenhouse gases (WMGHGs), aerosols (direct and indirect effect), ozone, and land use change. The forcings are calculated in the GISS model (Hansen et al. 2007) and are based on the underlying simulated distribution for aerosols or ozone and on the observed WMGHG concentrations and land use.

The formula used to calculate β is

$$\beta = \frac{\Delta T^{\mathrm{OBS}}}{F_W + F_D + F_I + F_O + F_L}, \tag{1}$$

where ΔT is the observed temperature response and the F components are the separate radiative forcings from changes in WMGHGs (W), aerosols (direct effect D and indirect effect I), ozone (O), and land use (L). In the WMGHGs, aerosol, and ozone forcings, we also include the influence of forcings from regions outside of the latitude band that we are examining. This approach is based on calculations of the temperature response per unit forcing to forcings applied locally and nonlocally in various latitude bands (SF09). The formula describing the calculation for a given latitude band L becomes

$$\beta^{L}_{\mathrm{LOC+NON_LOC}} = \Delta T^{L}_{\mathrm{OBS}} \Bigg/ \left[\frac{\sum_{i=1}^{4}(F_W_i \times r_W_i^L)}{\sum_{i=1}^{4}(r_W_i^L)} + \frac{\sum_{i=1}^{4}(F_D_i \times r_D_i^L)}{\sum_{i=1}^{4}(r_D_i^L)} + \frac{\sum_{i=1}^{4}(F_I_i \times r_I_i^L)}{\sum_{i=1}^{4}(r_I_i^L)} + \frac{\sum_{i=1}^{4}(F_O_i \times r_O_i^L)}{\sum_{i=1}^{4}(r_O_i^L)} + F_L_L \right]. \tag{2}$$

The subscripts in the terms of the equation denote the origin of the forcing. We consider four broad geographical origins of forcing: the southern extratropics (90°–28°S), the tropics (28°S–28°N), the northern midlatitudes (28°–60°N), and the northern high latitudes (60°–90°N). Note that the influence of southern extratropical aerosols and ozone to other regions is set to zero, since the nonlocal influence of aerosols/ozone from this region is small because of the actual forcings being small (SF09). For land use changes, we only consider local forcings (from the latitude zone of interest L).

The r terms are the responses at latitude L per unit forcing in area i taken from Figs. 1a,c,e,g,i of SF09, which are used to weight the actual forcings from each region

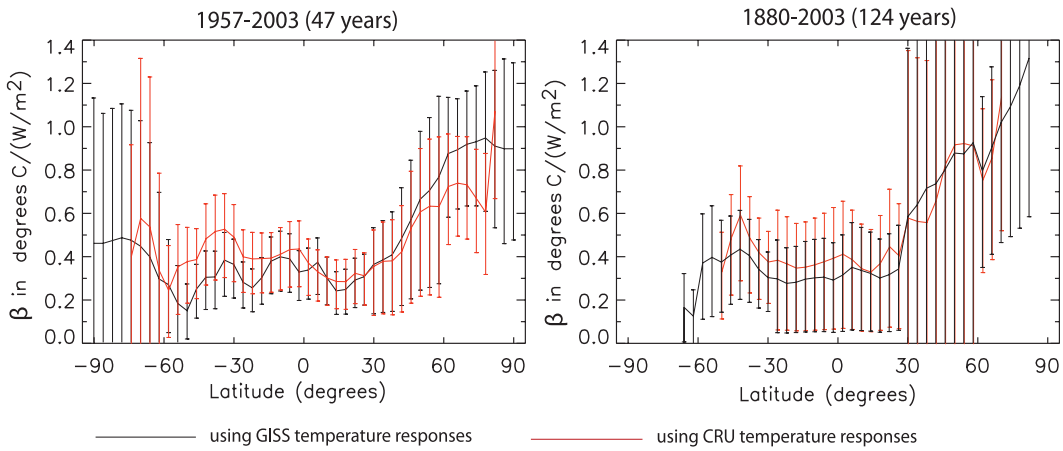

FIG. 1. Parameter β for 46 latitude bands. Calculated using Eqs. (1) and (2) with observed surface temperatures (GISS, CRU; 1957–2003, 1880–2003) and with estimated radiative forcings for WMGHGs (Forster and Taylor 2006), aerosols (direct and indirect), ozone, and land-use change [the latter three are taken as employed in GISS model simulations (Hansen et al. 2007)]. Error bars reflect uncertainty in the temperature regression analysis, the magnitude of internal variability, and current uncertainty in radiative forcing estimates. See section 2 for more details.

to reflect their influence in each latitude zone examined. The use of partial sensitivities is similar to the radiative kernels developed in Soden et al. (2008) and Shell et al. (2008), though those were used to examine climate feedbacks rather than sensitivity.

Note that in the southern extratropics (90°–28°S), we do not have information for the relative influence of local aerosol/ozone forcings from Fig. 1 of SF09, and the uncertainty of the nonlocal r terms is large, so we decide to use the r terms of WMGHGs in 60°–28°S for all agents (except land use change), as they are similar to those of sulfate and have smaller uncertainty. For 90°–60°S, we treat aerosol forcing as for 60°–28°S; however, for ozone, we apply only the local forcing, as in this region the local stratospheric ozone component is much more dominant. Equation (2) is modified accordingly for those latitudes.

We take into account the 2σ uncertainty in the observed temperatures related to the error of the linear regression and to internal variability. The latter is estimated using data from the GISS 1150-yr control run, which is also used in SF09, by calculating trends for each latitude band in 47- or 124-yr segments. Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) estimates of global forcings' percentage 2σ uncertainties relative to the mean are applied to the various latitude bands: ±5% for WMGHGs (Forster and Taylor 2006), ±80% for aerosol direct effect, +57%/−157% for aerosol indirect effect, ±100% for land use, +85%/−28% for tropospheric ozone, and ±100% for high-latitude stratospheric ozone (Fig. 2.20 of Forster et al. (2007)). Note that ideally we should have applied different percentage forcing uncertainties to different regions because there may be latitudinal variations in how uncertain the forcings are. However, so far there have been no multimodel studies/assessments reporting latitudinally varying uncertainty, and hence we rely on global estimates.

The uncertainty in the r terms is taken from Fig. 1 of SF09. Note that because of the lack of other studies evaluating the r terms, we rely on the values obtained from one model, which is currently a limitation of our approach. Similarly, the r terms could depend on the resolution of the model, and this is an issue that should be explored in future work.

Note that the AR4 uncertainty for tropospheric ozone is applied for latitudes north of 60°S, while for south of 60°S, we apply the AR4 uncertainty for stratospheric ozone, based on the relative importance of tropospheric/stratospheric ozone forcings. The total uncertainty in the inferred climate sensitivity is calculated using standard analytic error propagation methods. We did not use a probabilistic method but a local perturbative approach, which could be inappropriate in some regions, especially where the forcings are associated with very large uncertainties. However, for a probabilistic approach, we would have needed probability distribution functions of the forcings for each different latitude band, which, to our knowledge, is not available in present-day literature.

3. Results and discussion

In Fig. 1 we show the results of our calculation of β for the four different surface temperature datasets. There are polar latitude bands with missing values, especially for the earliest part of the 1880–2003 period. Trends

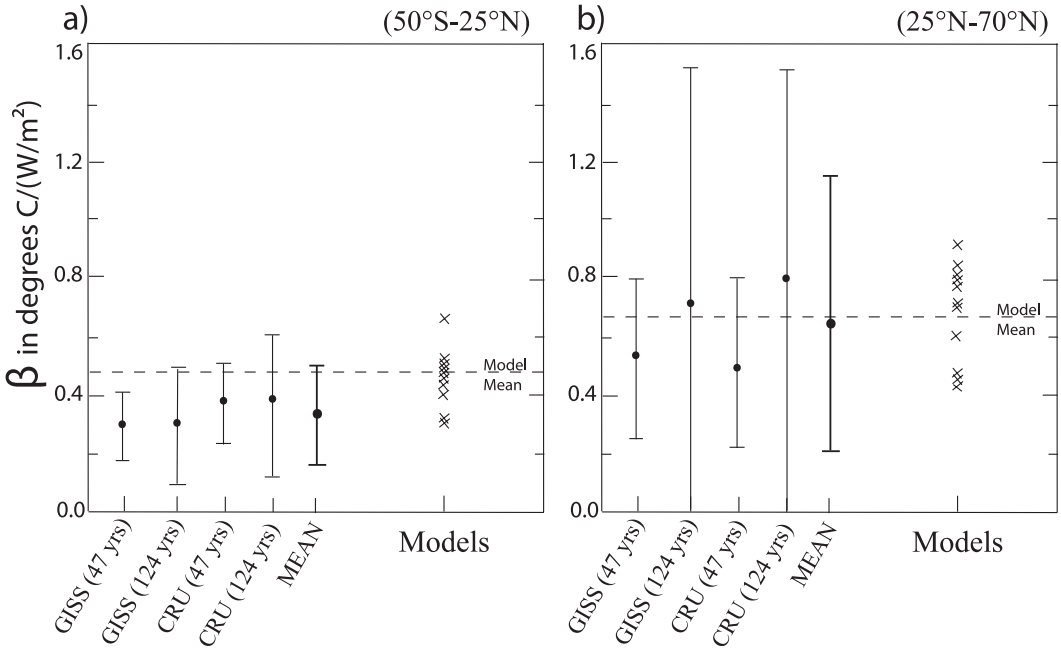

FIG. 2. Parameter β averaged in the region between (a) 50°S and 25°N and (b) 25° and 75°N. Results using all four different global temperature datasets, as well as the mean from all estimates, are presented. Crosses show β for these regions as estimated from the models participating in the Coupled Model Intercomparison Project phase 3 (CMIP3) 1% yr^{-1} experiment. The dashed horizontal line shows the model mean.

were calculated from annual mean data for latitudes where data from at least one station cover the whole extent of the period examined.

Parameter β is smallest in the tropics and southern midlatitudes, where values range mostly between 0.25° and 0.4°C $(W\ m^{-2})^{-1}$. At high latitudes, the sensitivity, as expected, is significantly larger, similar to the findings of Boer and Yu (2003). Also, the relative uncertainty in the tropics and southern midlatitudes is smaller than in the rest of the globe. The error bars can get very large over northern midlatitudes (especially for the estimate using the 1880–2003 temperature time series) because the aerosol direct and indirect effects, which are highly uncertain, are most important in this region. Similarly, uncertainty in the stratospheric ozone forcing makes the error bars large at southern high latitudes. The best agreement between the four estimates is found in the tropics. Note that the nonnegligible (though smaller) uncertainty in the tropics and southern midlatitudes is mainly driven by the uncertainty in the aerosol and ozone forcings, as found in our calculations (not shown).

The global mean β estimated using the only dataset with full global data coverage (GISS 1957–2003) is 0.39°C $(W\ m^{-2})^{-1}$, ranging from 0.18 to 0.56°C $(W\ m^{-2})^{-1}$. The magnitude of uncertainty (factor of 3) is very similar to the range of current estimates of transient climate response (1°–3.5°C; see Hegerl et al. 2007), which supports our methodology and characterization of uncertainty. The average β estimated by the IPCC AR4 models would be equal to 0.46°C $(W\ m^{-2})^{-1}$ (TCR = 1.76°C; Randall et al. 2007; radiative forcing resulting from the doubling of CO_2 = 3.80 W m^{-2}). This figure is larger than our estimate but well within the uncertainty range that we find.

Figure 2a shows the average values of β for 50°S–25°N, a region with smaller uncertainty compared to the rest of the globe, but one that still covers more than half of the earth's surface. The mean range (5%–95%) from all the estimates is 0.15°–0.51°C $(W\ m^{-2})^{-1}$ and the mean β is 0.35°C $(W\ m^{-2})^{-1}$. The different analyses reveal values that are quite similar. By performing the same calculations for the northern extratropics (25°–75°N; Fig. 2b), we find that β in this region cannot be constrained as well as for 50°S–25°N. The use of temperature datasets covering different periods makes our estimate very uncertain for 25°–75°N.

The range of β values that we obtain from the AR4 1% yr^{-1} CO_2 increase experiment for 50°S–25°N (Fig. 2a) overlaps with the observational estimate, but the average of the former [0.47°C $(W\ m^{-2})^{-1}$] is 35% higher than that of the latter. While the multimodel mean β is within the 95% confidence interval of the observationally based estimates, it is only marginally so, and

TABLE 1. Twentieth-century temperature change for 50°S–25°N as simulated by IPCC models, which also participated in the 1% yr^{-1} CO_2 experiment and in the GISS and CRU observational analyses. Italic numbers indicate the averages.

Dataset	Full dataset name	Temperature change (°C)
CGCM3.1(T47)	Canadian Centre for Climate Modelling and Analysis Coupled General Circulation Model, version 3.1	1.11
GISS-EH	GISS Model E with the HYCOM ocean model	0.44
GISS-ER	GISS Model E with Russell ocean model	0.52
MIROC3.2(medres)	Model for Interdisciplinary Research on Climate 3.2, medium-resolution version	0.44
ECHAM5		0.51
NCAR CCSM3	National Center for Atmospheric Research Community Climate System Model, version 3	0.70
HadCM3	Third climate configuration of the Met Office Unified Model	0.52
HadGEM1	Hadley Centre Global Environmental Model, version 1	0.43
Model average		*0.58*
Model average (without CGCM3.1)		*0.51*
Observed GISS		0.48
Observed CRU		0.56
Observational average		*0.52*

using IPCC terminology, our results show that some of the models very likely overestimate sensitivity in this region. For the northern extratropics, model mean β lies within the uncertainty of our estimate, but the uncertainty itself is so large that we cannot extract safe conclusions. Note that comparison with the better-constrained β over the 1957–2003 period shows that more than half of the models have larger values for 50°S–25°N and around one-quarter of them for 25°–75°N.

The way that β is defined, it theoretically depends on the underlying time-dependent forcing. However, by comparing 50°S–25°N results from the same model (GISS) for the historical (1880–2003) run and for the 1% yr^{-1} CO_2 experiment (two different cases of forcing), we find that β in the latter is only 8% higher than in the former (in 25°–75°N, this difference is 3 times larger—32%—which is consistent with the larger role of aerosols in this region, whose temporal evolution has been less smooth with time). This indicates that our parameter in 50°S–25°N should be fairly robust to differences between 1% yr^{-1} increasing CO_2 and historical forcing.

To examine if some of the models have indeed been too sensitive when simulating historical periods, we calculated the observed (GISS and CRU) twentieth-century change of 50°S–25°N average surface temperature and compared it with the temperature change simulated by the AR4 models that participated in the Climate of the 20th Century experiment as well as the 1% yr^{-1} experiment (see Table 1). In the observations, the average temperature change is 0.52°C, while in models it is 0.58°C. There is a slight overestimation of the temperature increase by the average GCM—around 12%. This is less than the 35% difference between our observationally constrained β and the one using the model output from the 1% yr^{-1} experiment. Also, note that if we ignore the CGCM3.1(T47) (refer to Table 1 for expansion), which is an outlier, modeled β equals 0.51°C, which is almost identical to the observations.

If some of the models are indeed too sensitive in this region, then they may still have captured the temperature change fairly well because the forcing by other agents is over- or underestimated. An overestimation of the aerosol forcing would support this argument. Kiehl (2007) pointed out that models with very diverse climate sensitivities can simulate observed temperature changes quite well, mainly because models use different radiative forcing values, especially for anthropogenic aerosols. Stott et al. (2006) also discussed the possibility of the agreement of model output with twentieth-century temperature records being a result of a fortuitous cancellation of errors in forcings. A recent study from Myhre (2009) suggested that the direct aerosol radiative forcing in current estimates is most likely too high (mean forcing is around -0.5 W m^{-2}, while they find a forcing of -0.3 W m^{-2}).

4. Conclusions

We use global surface temperature observations and estimated forcings from various agents to examine an aspect of the climate system not thoroughly examined previously: regional climate sensitivity. In the tropics and southern midlatitudes, we find that β (a parameter defined here as the temperature response per unit forcing for each latitude band) has its lowest values, that it is most robust to changes in the surface temperature dataset analyzed (especially in the tropics), and that its uncertainty is smallest (though not very small).

Our results show that 50°S–25°N β may provide a more valuable constraint for model evaluation than global mean climate sensitivity. Some of the AR4 models reveal a higher β than what we find using observations for this region. The reason for this is not clear. However, we find that although the models do, on average, slightly overestimate the twentieth-century temperature change for this region, the difference is relatively smaller than for β. The models may be simulating temperature changes fairly well because they overestimate the aerosol forcing.

We note that the r terms that we used to take the nonlocal forcings into account could be model and resolution dependent, so that our results for β could have been different had we used r terms from another model/resolution. Assessing this extra uncertainty would require a multimodel study focusing on the determination of the transient temperature sensitivity parameter (β), and such a study has not been performed so far. We suggest that the evaluation of the r terms and, subsequently, β could be a useful focus of further multimodel intercomparison exercises.

Acknowledgments. The authors wish to thank Greg Faluvegi for his help in analyzing the data, Reto Ruedy for providing the radiative forcing data for the period examined, and NASA ACMAP for its support.

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