On the observational determination of climate sensitivity and its implications

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Abstract

To estimate climate sensitivity from observations, *Lindzen and Choi* [2009] used the deseasonalized fluctuations in sea surface temperatures (SSTs) and the concurrent responses in the top-of-atmosphere outgoing radiation from the ERBE satellite instrument. Distinct periods of warming and cooling in the SST were used to evaluate feedbacks. This work was subject to significant criticism by *Trenberth et al*. [2009], much of which was appropriate. The present paper is an expansion of the earlier paper in which the various criticisms are addressed and corrected. In this paper we supplement the ERBE data for 1985-1999 with data from CERES for 2000-2008. Our present analysis accounts for the 36 day precession period for the ERBE satellite in a more appropriate manner than in the earlier paper which simply used what may have been undue smoothing. The present analysis also distinguishes noise in the outgoing radiation as well as radiation changes that are forcing SST changes from those radiation changes that constitute feedbacks to changes in SST. Finally, a more reasonable approach to the zero-feedback flux is taken here. We argue that feedbacks are largely concentrated in the tropics and extend the effect of these feedbacks to the global climate. We again find that the outgoing radiation resulting from SST fluctuations exceeds the zero-feedback fluxes thus implying negative feedback. In contrast to this, the calculated outgoing radiation fluxes from 11 atmospheric GCMs forced by the observed SST are less than the zero-feedback fluxes consistent with the positive feedbacks that characterize these models. The observational analysis implies that the models are exaggerating climate sensitivity.

1. Introduction

It is usually claimed that the heart of the global warming issue is so-called greenhouse warming. This simply refers to the fact that the earth balances the heat received from the sun (mostly in the visible spectrum) by radiating in the infrared portion of the spectrum back to space. Gases that are relatively transparent to visible light but strongly absorbent in the infrared (greenhouse gases) will interfere with the cooling of the planet, thus forcing it to become warmer in order to emit sufficient infrared radiation to balance the net incoming sunlight. By the net incoming sunlight, we mean that portion of the sun's radiation that is not reflected back to space by clouds and the earth's surface. The issue then focuses on a particular greenhouse gas, carbon dioxide. Although carbon dioxide is a relatively minor greenhouse gas, it has increased significantly since the beginning of the industrial age from about 280 ppmv to about 390 ppmv, and it is widely accepted that this increase is primarily due to man's emissions. However, it is also widely accepted that the warming from a doubling of carbon dioxide would only be about 1°C (based on simple Planck black body calculations; it is also the case that a doubling of any concentration in ppmv produces the same warming because of the logarithmic dependence of carbon dioxide's absorption on the amount of carbon dioxide).

This amount of warming is not considered catastrophic, and, more importantly, this is much less than current climate models suggest the warming from a doubling of carbon dioxide will be. The usual claim from the models is that a doubling of carbon dioxide will lead to warming of from 1.5°C to 5°C and even more. What then is really fundamental to 'alarming' predictions? It is the 'feedback' within models from the more important greenhouse substances, water vapor and clouds. Within all current climate models, water vapor increases with increasing temperature so as to further inhibit infrared cooling. Clouds also change so that their net effect resulting from both their infrared absorptivity and their visible reflectivity is to further reduce the net cooling of the earth. These feedbacks are still acknowledged to be highly uncertain, but the fact that these feedbacks are strongly positive in most models is considered to be a significant indication that the result has to be basically correct. Methodologically, this is an unsatisfactory approach to such an important issue. Ideally, one would seek an observational test of the issue. As it turns out, it may be possible to test the issue with existing data from satellites and there has recently been a paper [*Lindzen and Choi*, 2009] that has attempted this though, as we will show in this paper, the details of that paper were, in important ways, incorrect. The present paper attempts to correct the approach and arrives at similar conclusions.

2. Feedback formalism

A little bit of simple theory shows how one can go about doing this. In the absence of feedbacks, the behavior of the climate system can be described by Fig. 1. ΔQ is the radiative forcing, G_0 is the zero-feedback response function of the climate system, and ΔT_0 is the response of the climate system in the absence of feedbacks. The checkered circle is a node. Figure 1 symbolizes the temperature increment, ΔT_0, that a forcing increment, ΔQ, would produce with no feedback,

$$\Delta T_0 = G_0 \Delta Q \qquad (1)$$

It is generally accepted [*Hartmann*, 1994] that without feedback, doubling of carbon dioxide will cause a forcing of $\Delta Q \approx 3.7\ \mathrm{Wm^{-2}}$ and will increase the temperature by $\Delta T_0 \approx 1.1$°C (due to the black body response) [*Schwartz*, 2007]. We therefore take the zero-feedback response function of (1) to be $G_0 \approx 0.3$ (=1.1/3.7) K $\mathrm{W^{-1}m^2}$ for the earth as a whole.

With feedback, Figure 1 is modified to Fig. 2. The response is now

$$\Delta T = G_0(\Delta Q + F\Delta T) \qquad (2)$$

Here F is a feedback function that represents all changes in the climate system (for example, changes in cloud cover or humidity) that act to increase or decrease feedback-free effects. Thus, F should not include the response to ΔT that is already incorporated into G_0. The choice of zero feedback flux for the tropics in *Lindzen and Choi* [2009] is certainly incorrect in this respect. At present, the best choice seems to remain $1/G_0$ (3.3 W m^{-2} K^{-1}) [*Colman*, 2003; *Schwarz*, 2007], though a lower value than this might be appropriate due to the high opacity of greenhouse gases.

Solving (2) for the temperature increment ΔT we find

$$\Delta T = \frac{\Delta T_0}{1-f}. \qquad (3)$$

The dimensionless feedback fraction is $f = F\,G_0$.

From Fig. 2, the relation of the change in flux, ΔFlux, to the change in temperature is given by

$$\Delta\text{Flux} - \text{ZFB} = -\frac{f}{G_0}\Delta T \qquad (4)$$

The quantities on the left side of the equation indicate the amount by which feedbacks supplement the zero-feedback response (ZFB) to ΔQ. At this point, it is crucial to recognize that our equations, thus far, are predicated on the assumption that the ΔT to which the feedbacks are responding is that produced by ΔQ. Physically, however, any fluctuation in ΔT should elicit the same flux regardless of the origin of ΔT. When looking at the observations, we emphasize this by rewriting (4) as

$$\Delta\text{Flux} - \text{ZFB} = -\frac{f}{G_0}\Delta\text{SST} \qquad (5)$$

where SST is the observed sea surface temperature.

When restricting ourselves to tropical feedbacks, equation (5) is replaced by

$$-G_0\left(\frac{\Delta\text{Flux} - \text{ZFB}}{\Delta\text{SST}}\right)_{tropics} \approx 2f \qquad (6)$$

where the factor 2 results from the sharing of the tropical feedbacks over the globe following the methodology of *Lindzen, Chou and Hou* [2001] (See Appendix 2 for more explanation). The longwave (LW) and shortwave (SW) contributions to f are given by

$$f_{LW} = -\frac{G_0}{2}\left(\frac{\Delta\text{OLR} - \text{ZFB}}{\Delta\text{SST}}\right)_{tropics} \qquad (7a)$$

$$f_{SW} = -\frac{G_0}{2}\left(\frac{\Delta\text{SWR}}{\Delta\text{SST}}\right)_{tropics} \qquad (7b)$$

Here we can identify ΔFlux as the change in outgoing longwave radiation (OLR) and shortwave radiation (SWR) measured by satellites associated with the measured ΔSST, the change of the sea-surface temperature. Since we know the value of G_0, the experimentally determined slope allows us to evaluate the magnitude and sign of the feedback factor f provided that we also know the value of the zero-feedback flux. Note that the natural forcing, ΔSST, that can be observed, is different from the equilibrium response temperature ΔT in Eq. (3). The latter cannot be observed since, for the short intervals considered, the system cannot be in equilibrium, and over the longer periods needed for equilibration of the whole climate system, ΔFlux at the top of the atmosphere is restored to zero. Indeed, as explained in *Lindzen and Choi* [2009], it is, in fact, essential, that the time intervals considered, be short compared to the time it takes for the system to equilibrate, while long compared to the time scale on which the feedback processes operate (which are essentially the time scales associated with cumulonimbus convection). The latter is on the order of days, while the former depends on the climate sensitivity, and ranges from years for sensitivities of 0.5°C for a doubling of CO_2 to many decades for higher sensitivities [*Lindzen and Giannitsis*, 1998]. Finally, for observed variations, there is the fact that changes in radiation (as for example associated with volcanoes) can cause changes in SST as well as respond to changes in SST, and there is a need to distinguish these two possibilities. This is not an issue with model results from the AMIP program where observed variations in SST are specified. Of course, there is always the problem of noise arising from the fact that clouds depend on factors other than surface temperature. Note that this study deals with observed outgoing fluxes, but does not specifically identify the origin of the changes. This is discussed in Appendix 1.

3. The data and their problems

Now, it turns out that SST is measured [*Kanamitsu et al.*, 2002], and is always fluctuating as we see from Fig. 3. High frequency fluctuations, however, make it difficult to objectively identify the beginning and end of warming and cooling intervals [*Trenberth et al.*, 2010]. This ambiguity is eliminated with a 3 point centered smoother. (A two point lagged smoother works as well.) In addition, the net outgoing radiative flux from the earth has been monitored since 1985 by the ERBE instrument [*Barkstrom*, 1984] (nonscanner edition 3) aboard ERBS satellite, and since 2000 by the CERES instrument (ES4 FM1 edition 2) aboard the Terra satellite [*Wielicki et al.*, 1998]. The results for both LW (infrared) radiation and SW (visible) radiation are shown in Fig. 4. The sum is the net flux.

With ERBE data, there is, however, the problem of satellite precession with a period of 36 days. In *Lindzen and Choi* [2009] that used ERBE data, we attempted to avoid this problem (which is primarily of concern for the short wave radiation) by smoothing data over 7 months. It has been suggested (Takmeng Wong, personal communication) that this is excessive smoothing. In the present paper, we start by taking 36 day means rather than monthly means. The CERES instrument is flown on a sun-synchronous satellite for which there is no problem with precession. Thus for the CERES instrument we use the conventional months. However, here too we examine the effect of modest smoothing.

The discontinuity between the two datasets needs some discussion. There is the long-term discrepancy of the average which is generally acknowledged to be due to the absolute calibration problem (up to 3 W m^{-2}) [*Wong et al*., 2006]. With CERES, the spectral darkening was resolved by multiplying SW flux by the scale factor (up to 1.011) from *Matthews* et al. [2005]. However, this long-term stability should not matter for our analysis which focuses on short-term fluctuations. One major concern to be considered in this study is the higher seasonal fluctuation of CERES SW radiation than ERBE. The bias is up to 6.0 W m^{-2} as estimated by *Young et al*. [1998]. This is attributed to different sampling patterns, that is, ERBS observes all local times over a period of 72 days, while Terra observes the region only twice per day (around 10:30 AM and 10:30 PM). To avoid this problem, the anomalies for radiative flux are separately referenced to the monthly means for the period of 1985 through 1989 for ERBE, and for the period of 2000 through 2004 for CERES. However, the issue of the reference period is also insignificant in this study that uses enough segments to cancel out this seasonality.

Both ERBE and CERES data are best for the tropics. The ERBE field-of-view is between 60°S and 60°N. For latitudes 40° to 60°, 72 days are required instead of 36 days to reduce the precession effect [*Wong et al*., 2006]. Both datasets have no/negligible shortwave radiation in winter hemispheric high latitudes, which would compromise our analysis. Moreover, our analysis involves relating changes in outgoing flux to changes in SST. This is appropriate to regions that are mostly ocean covered like the tropics or the southern hemisphere, but distinctly inappropriate to the northern extratropics. However, as we will argue in Appendix 2, the water vapor feedback is almost certainly restricted primarily to the tropics, and there are reasons to suppose that this is also the case for cloud feedbacks. The methodology developed in *Lindzen, Chou, and Hou* [2001] permits the easy extension of the tropical processes to global values.

Finally, there will be a serious issue concerning distinguishing atmospheric phenomena involving changes in outgoing radiation that result from processes other than feedbacks (the Pinatubo eruption for example) and which cause changes in sea surface temperature, from those that are caused by changes in sea surface temperature (namely the feedbacks we wish to evaluate). Our admittedly crude approach to this is to examine the effect of considering fluxes with a time lags and leads relative to temperature changes. The lags and leads examined are from one to five months. Our procedure will be to choose lags that maximize R (the correlation). This is discussed in Section 4.

Turning to the models, AMIP (Atmospheric Model Intercomparison Program) is responsible for intercomparing models used by the IPCC (the Intergovernmental Panel on Climate Change), has obtained the calculated changes in both short and long wave radiation from models forced by the observed sea surface temperatures shown in Fig. 3. These results are shown in Figs. 5 and 6 where the observed results are also plotted for comparison. We can already see that there are significant differences. Note that it is important to use the AMIP results rather than those from the coupled atmosphere-ocean models (CMIP). Only for the former can we see the results for the same SST as applies to the ERBE/CERES observations. Moreover, in the AMIP results, we are confident that the temperatures are forcing the changes in outgoing radiation; in the coupled models it is more difficult to be sure that we are calculating outgoing fluxes that are responding to SST forcing rather than temperature perturbations resulting from independent fluctuations in radiation.

4. Calculations

With all the above readily available, it is now possible to directly test the ability of models to adequately simulate the sensitivity of climate. The procedure is simply to identify intervals of change for ΔSST in Fig. 3 (for reasons we will discuss at the end, it is advisable, but not essential, to restrict oneself to changes greater than 0.1°C), and for each such interval, to find the change in flux. Let us define i_1, i_2,…i_m as selected time steps that correspond to the starting and the ending points of intervals. ΔFlux/ΔSST can be basically obtained by Flux(i_1)–Flux(i_2) divided by SST(i_1) –SST(i_2). As there are many intervals, ΔFlux/ΔSST is a regression slope for the plots (ΔFlux, ΔSST) for a linear regression model. Here we use a zero y-intercept model (y = ax) because the presence of the y-intercept is related to noise other than feedbacks. Thus, a zero y-intercept model may be more appropriate for the purpose of our feedback analysis; however, the choice of regression model turns out to be minor. As already noted, the data need to be smoothed to minimize noise, and it is also crucial to distinguish ΔSST that are forcing changes in ΔFlux, and not responses to ΔFlux. Otherwise, ΔFlux/ΔSST can vary [*Trenberth et al.*, 2010] and/or may not represent feedbacks that we wish to determine. As an attempt to avoid such problems, though imperfectly, we need to consider smoothing (i.e., use of Flux’(i) and SST’(i), where the prime designates the smoothed value) and lag-lead methods (e.g., use of Flux’(i+lag) and SST’(i)) for ERBE 36-day and CERES monthly data. For a stable estimate of ΔFlux/ΔSST, the time step i should be also selected based on the maximum and minimum of the smoothed SST (i.e., SST'). As shown in Fig. 3, this study selected $SST'(i_1) - SST'(i_2)$ that exceeds 0.1 K. The impact of thresholds for ΔSST on the statistics of the results is minor [*Lindzen and Choi*, 2009].

Figure 7 shows the impact of smoothing and leads and lags on the determination of the slope as well as on the correlation, R, of the linear regression. In general, the use of leads for flux will emphasize forcing by the fluxes, and the use of lags will emphasize responses by the fluxes to changes in SST. For LW radiation, the situation is fairly simple. Smoothing increases R somewhat, and for 3 point symmetric smoothing, R maximizes for slight lag or zero – consistent with the fact that feedbacks are expected to result from fast processes. Maximum slope is found for a lag of 1 'month', though it should be remembered that the relevant feedback processes may operate on a time scale shorter than we resolve. The situation for SW radiation is, not surprisingly, more complex since phenomena like the Pinatubo eruption lead to increased light reflection and associated cooling of the surface (There is also the obvious fact that many things can cause fluctuations in clouds, which leads to noise). We see two extremes associated with changing lead/lag. There is a maximum negative slope associated with a brief lead, and a relatively large positive slope associated with a 3–4 month lag. It seems reasonable to suppose that the effect of forcing extends into the results at small lags because it takes time for the ocean surface to respond, and is only overcome for larger lags where the change in flux associated with feedback dominates. Indeed, excluding the case of Pinatubo volcano for larger lags does little to change the results (less than 0.3 W m^{-2}/K). Under such circumstances, we expect the maximum slope for SW radiation in Fig. 7 to be an underestimate of the actual feedback. We also consider the standard error of the slope to show data uncertainty. The results for the lags associated with maximum R are shown in Table 1. We take LW and SW radiation for lag = 1 and lag = 3, respectively, and measure the slope ΔFlux/ΔSST for the sum of these fluxes. The standard error of the slope in total radiation for the appropriate lags comes from the regression for scatter plots of (ΔSST, Δ(OLR+SWR)). As we see in Table 1, model sensitivities indicated by the IPCC AR4 (Fig. 8) are likely greater than the possibilities estimated in the observations.

We next wish to see whether the outgoing fluxes from the AMIP models are consistent with the sensitivities in Fig. 8. For the AMIP results, for which there was no ambiguity as to whether fluxes constituted a response, there was little dependence on smoothing or lag, so we simply used the AMIP fluxes without smoothing or lag. The results are shown in Table 2. In contrast to the observed fluxes, the implied feedbacks in the models are all positive, and in one case, marginally unstable. Given the uncertainties, however, one should not take that too seriously.

Table 3 compares the sensitivities implied by Table 2 with those in Fig. 8. The agreement does not seem notable; however, even 90% confidence levels are consistent with the independently derived sensitivities (obtained by running models to near equilibrium with increased CO_2). For positive feedbacks, sensitivity is strongly affected by small changes in f that are associated standard errors in Table 2. Consequently, the range of sensitivity estimated from standard errors of f includes “infinity”. This is seen in Fig. 9 in the pink region. It has, in fact, been suggested by *Roe and Baker* [2007], that this sensitivity of the climate sensitivity to uncertainty in the feedback factor is why there has been no change in the range of climate sensitivities indicated by GCMs since the 1979 Charney Report. By contrast, in the green region, which corresponds to the observed feedback factors, sensitivity is much better constrained.

5. Discussion and conclusions

Since our analysis of the data only demands relative instrumental stability over short periods, it is difficult to see what data problems might change our results significantly. A major concern is the different sampling from the ERBE and CERES instruments. The addition of CERES data to the ERBE data used by *Lindzen and Choi* [2009] certainly does little to change their results concerning ΔFlux/ΔSST – except that its value is raised a little (This is also true for the case that CERES data only is used.).

The conclusion appears to be that all current models exaggerate climate sensitivity (some greatly). It also suggests, incidentally, that in current coupled atmosphere-ocean models, that the atmosphere and ocean are too weakly coupled since thermal coupling is inversely proportional to sensitivity [*Lindzen and Giannitsis*, 1998]. It has been noted by *Newman et al*. [2009] that coupling is crucial to the simulation of phenomena like El Niño. Thus, corrections of the sensitivity of current climate models might well improve the behavior of coupled models. It should also be noted that there have been independent tests that also suggest sensitivities less than predicted by current models (*Lindzen and Giannitsis* [1998], based on response to sequences of volcanic eruptions — they also noted that the response to individual volcanoes in the two years following eruption were largely independent of sensitivity, and, hence, of little use for distinguishing different sensitivities; *Lindzen* [2007], and *Douglass et al*. [2007], both based on the vertical structure of observed versus modeled temperature increase; and *Schwartz* [2007, 2008], based on ocean heating). Most claims of greater sensitivity are based on the models that we have just shown can be highly misleading on this matter. There have also been attempts to infer sensitivity from paleoclimate data [*Hansen*, 1993], but these are not really tests since the forcing is essentially unknown and may be adjusted to produce any sensitivity one wishes. It is important to realize that climate sensitivity is essentially a single number. Economists who treat climate sensitivity as a probability distribution function [*Weitzman*, 2009; *Stern*, 2008; *Sokolov et al.*, 2009] are mistakenly confusing model uncertainty concerning this particular number with the existence of a real range of possibility. The high sensitivity results that these studies rely on for claiming that catastrophes are possible are almost totally incompatible with the present results – despite the uncertainty of the present results.

One final point needs to be made. Low sensitivity of global mean temperature anomaly to global scale forcing does not imply that major climate change cannot occur. The earth has, of course, experienced major cool periods such as those associated with ice ages and warm periods such as the Eocene [*Crowley and North*, 1991]. As noted, however, in *Lindzen* [1993], these episodes were primarily associated with changes in the equator-to-pole temperature difference and spatially heterogeneous forcing. Changes in global mean temperature were simply the residue of such changes and not the cause. It is worth noting that current climate GCMs have not been very successful in simulating these *changes* in past climate.

Appendices

Appendix 1. Origin of Feedbacks

While the present analysis is a direct test of feedback factors, it does not provide much insight into detailed mechanism. Nevertheless, separating the contributions to f from long wave and short wave fluxes provides some interesting insights. The results are shown in Tables 1 and 2. It should be noted that the consideration of the zero-feedback response, and the tropical feedback factor to be half of the global feedback factor is actually necessary for our measurements from the Tropics; however, these were not considered in *Lindzen and Choi* [2009]. Accordingly, with respect to separating longwave and shortwave feedbacks, the interpretation by *Lindzen and Choi* [2009] needs to be corrected. These tables show recalculated feedback factors in the presence of the zero-feedback Planck response. The negative feedback from observations is from both longwave and shortwave radiation, while the positive feedback from models is usually but not always from longwave feedback.

As concerns the infrared, there is, indeed, evidence for a positive water vapor feedback [*Soden et al*., 2005], but, if this is true, this feedback is presumably cancelled by a negative infrared feedback such as that proposed by *Lindzen et al*. [2001] in their paper on the iris effect. In the models, on the contrary, the long wave feedback appear to be positive (except for two models), but it is not as great as expected for the water vapor feedback [*Colman*, 2003; *Soden et al*., 2005]. This is possible because the so-called lapse rate feedback as well as negative longwave cloud feedback serves to cancel the TOA OLR feedback in current models. Table 2 implies that TOA longwave and shortwave contributions are coupled in models (the correlation coefficient between f_{LW} and f_{SW} from models is about -0.5.). This coupling most likely is associated with the primary clouds in models — optically thick high-top clouds [*Webb et al*., 2006]. In most climate models, the feedbacks from these clouds are simulated to be negative in longwave and strongly positive in shortwave, and dominate the entire cloud feedback [*Webb et al*., 2006]. Therefore, the cloud feedbacks may also serve to contribute to the negative OLR feedback and the positive SWR feedback. New spaceborne data from the CALIPSO lidar (CALIOP; *Winker et al.* [2007]) and the CloudSat radar (CPR; *Im et al*. [2005]) should provide a breakdown of cloud behavior with altitude which may give some insight into what exactly is contributing to the radiation.

Appendix 2. Concentration of climate feedbacks in the tropics

Although, in principle, climate feedbacks may arise from any latitude, there are substantive reasons for supposing that they are, indeed, concentrated in the tropics. The most prominent model feedback is that due to water vapor, where it is commonly noted that models behave as though relative humidity were fixed. Pierrehumbert [2009] examined outgoing radiation as a function of surface temperature theoretically for atmospheres with constant relative humidity. His results are shown in Fig. 10.

We see that for extratropical conditions, outgoing radiation closely approximates the Planck black body radiation (leading to small feedback). However, for tropical conditions, increases in outgoing radiation are suppressed, implying substantial positive feedback. There are also good reasons to suppose that cloud feedbacks are largely confined to the tropics. In the extratropics, clouds are mostly stratiform clouds that are associated with ascending air while descending regions are cloud-free. Ascent and descent are largely determined by the large scale wave motions that dominate the meteorology of the extratropics, and for these waves, we expect approximately 50% cloud cover regardless of temperature. On the other hand, in the tropics, upper level clouds, at least, are mostly determined by detrainment from cumulonimbus towers, and cloud coverage is observed to depend significantly on temperature [*Rondanelli and Lindzen*, 2008]. As noted by *Lindzen et al*. [2001], with feedbacks restricted to the tropics, their contribution to global sensitivity results from sharing the feedback fluxes with the extratropics. This leads to the factor of 2 in Eq. (6).

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Table legends

Table 1. Mean±standard error of the variables for the likely lag for the observations. The units for the slope are W m^{-2} K^{-1}. Also shown are the estimated mean and range of equilibrium climate sensitivity (in degrees C) for a doubling of CO_2 for 90%, 95%, and 99% confidence levels.

	Variables		Comments
a	Slope, LW	5.3±1.3	Lag = 1
b	Slope, SW	1.9±2.6	Lag = 3
c	Slope, Total	6.9±1.8	= a+b for the same SST interval
d	f_{LW}	–0.3±0.2	Calculated from a
e	f_{SW}	–0.3±0.4	Calculated from b
f	f_{Total}	–0.5±0.3	Calculated from c
g	Sensitivity, mean	0.7	Calculated from f
h	Sensitivity, 90%	0.6–1.0	Calculated from f
i	Sensitivity, 95%	0.5–1.1	Calculated from f
j	Sensitivity, 99%	0.5–1.3	Calculated from f

Table 2. Regression statistics between ΔFlux and ΔSST and the estimated feedback factors (*f*) for LW, SW, and total radiation in AMIP models; the slope is ΔFlux/ΔSST, N is the number of the points or inttervals, R is the correlation coefficient, and SE is the standard error of ΔFlux/ΔSST.

		LW				SW				LW+SW			
	N	Slope	R	SE	f_{LW}	Slope	R	SE	f_{SW}	Slope	R	SE	f
CCSM3	19	1.5	0.4	1.8	0.3	–3.1	–0.5	2.2	0.5	–1.6	–0.3	2.7	0.7
ECHAM5/MPI-OM	18	2.8	0.6	1.7	0.1	–1.1	–0.2	3.1	0.2	1.7	0.3	3.0	0.2
FGOALS-g1.0	18	–0.2	–0.1	1.6	0.5	–2.8	–0.7	1.3	0.4	–3.0	–0.7	1.6	1.0
GFDL-CM2.1	18	1.5	0.6	1.0	0.3	–0.4	–0.1	2.8	0.1	1.1	0.2	2.5	0.3
GISS-ER	22	2.9	0.6	1.4	0.1	–3.3	–0.5	2.3	0.5	–0.5	–0.1	1.8	0.6
INM-CM3.0	24	2.9	0.6	1.5	0.1	–3.1	–0.6	1.7	0.5	–0.3	–0.1	1.9	0.5
IPSL-CM4	22	–0.4	–0.1	2.1	0.6	–2.6	–0.5	2.0	0.4	–3.0	–0.5	2.1	0.9
MRI-CGCM2.3.2	22	–1.1	–0.2	2.2	0.7	–3.9	–0.4	3.1	0.6	–5.0	–0.6	2.6	1.2
MIROC3.2(hires)	22	0.7	0.1	2.2	0.4	–2.1	–0.5	1.6	0.3	–1.4	–0.3	2.5	0.7
MIROC3.2(medres)	22	4.4	0.7	1.8	-0.2	–5.3	–0.7	2.3	0.8	–0.9	–0.2	1.9	0.6
UKMO-HadGEM1	19	5.2	0.7	2.2	-0.3	–5.9	–0.7	2.1	0.9	–0.8	–0.1	2.2	0.6

Table 3. Comparison of model equilibrium climate sensitivities for a doubling of CO_2 defined from IPCC AR4 and estimated from feedback factors estimated in this study. The ranges of climate sensitivities for models are also estimated from the standard errors in Table 2 for 90%, 95%, and 99% confidence levels.

Models	AR4 sensitivity	Sensitivity, mean	Sensitivity, 90%	Sensitivity, 95%	Sensitivity, 99%
CCSM3	2.7	4.2	1.2 – Infinity	1.0 – Infinity	0.8 – Infinity
ECHAM5/MPI-OM	3.4	1.4	0.7 – 28.9	0.7 – Infinity	0.6 – Infinity
FGOALS-g1.0	2.3	22.4	2.4 – Infinity	2.1 – Infinity	1.6 – Infinity
GFDL-CM2.1	3.4	1.6	0.9 – 15.4	0.8 – Infinity	0.7 – Infinity
GISS-ER	2.7	2.5	1.2 – Infinity	1.1 – Infinity	1.0 – Infinity
INM-CM3.0	2.1	2.4	1.2 – Infinity	1.1 – Infinity	0.9 – Infinity
IPSL-CM4	4.4	19.5	1.9 – Infinity	1.6 – Infinity	1.3 – Infinity
MRI-CGCM2.3.2	3.2	Infinity	2.8 – Infinity	2.2 – Infinity	1.5 – Infinity
MIROC3.2(hires)	4.3	3.8	1.2 – Infinity	1.1 – Infinity	0.9 – Infinity
MIROC3.2(medres)	4	3.0	1.3 – Infinity	1.2 – Infinity	1.0 – Infinity
UKMO-HadGEM1	4.4	2.8	1.2 – Infinity	1.1 – Infinity	0.9 – Infinity

Figure legends

Figure 1. A schematic for the behavior of the climate system in the absence of feedbacks.

Figure 2. A schematic for the behavior of the climate system in the presence of feedbacks.

Figure 3. Tropical mean (20°S to 20°N latitude) 36-day averaged and monthly sea surface temperature anomalies with the centered 3-point smoothing; the anomalies are referenced to the monthly means for the period of 1985 through 1989. The SST anomaly was scaled by a factor of 0.78 (the area fraction of the ocean to the tropics) to relate with the flux. Red and blue colors indicate the major temperature fluctuations exceeding 0.1°C used in this study. The cooling after 1998 El Niño is not included because of no flux data is available for this period (viz Fig. 4).

Figure 4. The same as Fig. 3 but for outgoing longwave (red) and reflected shortwave (blue) radiation from ERBE and CERES satellite instruments. 36-day averages are used to compensate for the ERBE precession. The anomalies are referenced to the monthly means for the period of 1985 through 1989 for ERBE, and 2000 through 2004 for CERES. Missing periods are the same as reported in Wong et al. (2006).

Figure 5. Comparison of outgoing longwave radiations from AMIP models (black) and the observations (red) as found in Fig. 4.

Figure 6. Comparison of reflected shortwave radiations from AMIP models (black) and the observations (blue) shown in Fig. 4.

Figure 7. The impact of smoothing and leads and lags on the determination of the slope (top) as well as on the correlation coefficient, R, of the linear regression (bottom).

Figure 8. Equilibrium climate sensitivity of 11 AMIP models (from *IPCC* [2007]).

Figure 9. Sensitivity vs. feedback factor.

Figure 10. OLR vs. surface temperature for water vapor in air, with relative humidity held

fixed. The surface air pressure is 1bar, and Earth gravity is assumed. The temperature profile is the water/air moist adiabat. Calculations were carried out with the Community Climate Model radiation code (from Pierrehumbert [2009]).

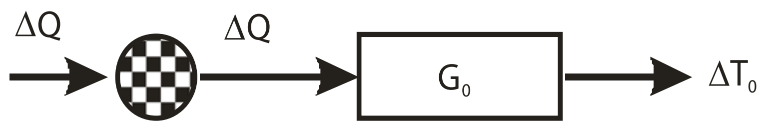


Figure 1

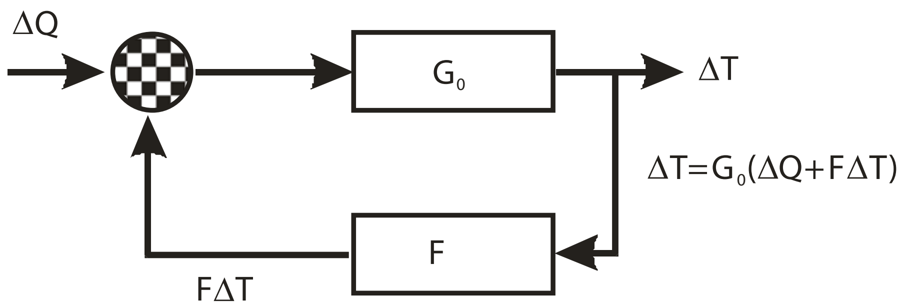


Figure 2

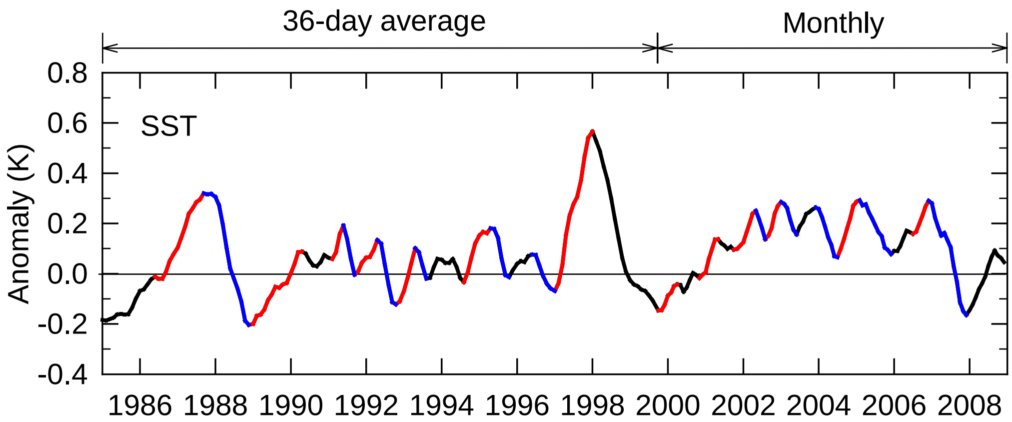
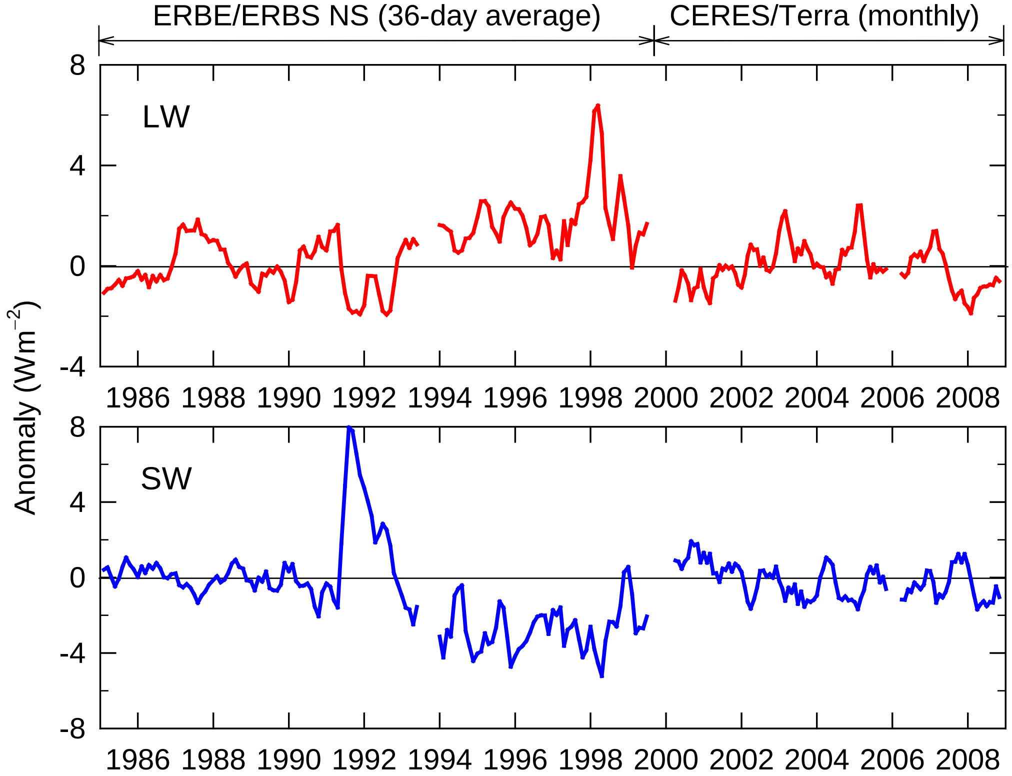

36-day average
Monthly
0.8
0.6
0.4
0.2
0.0
-0.2
-0.4
Anomaly (K)
SST
1986
1988
1990
1992
1994
1996
1998
2000
2002
2004
2006
2008

Figure 3



488

489 Figure 4

490

491

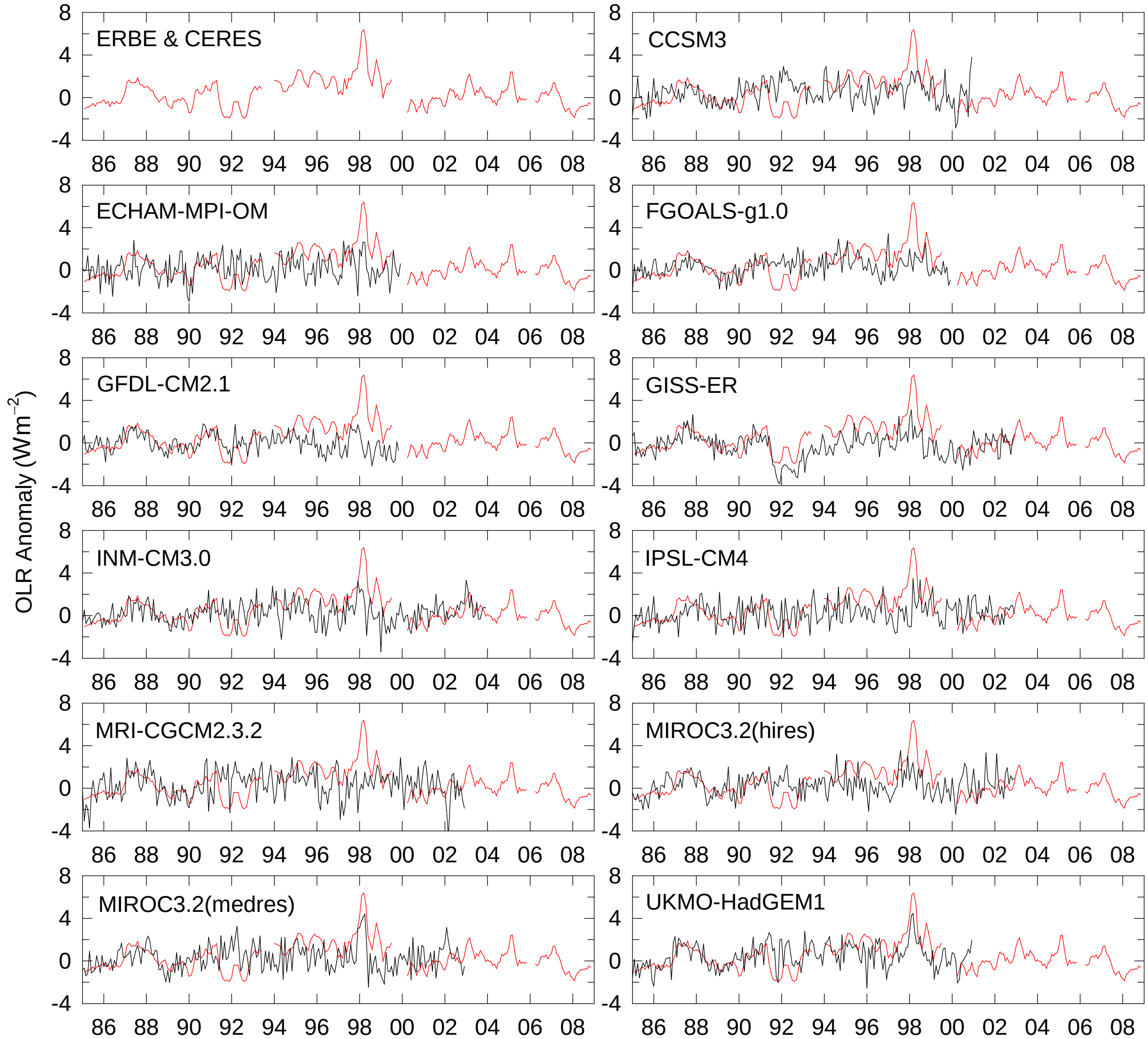
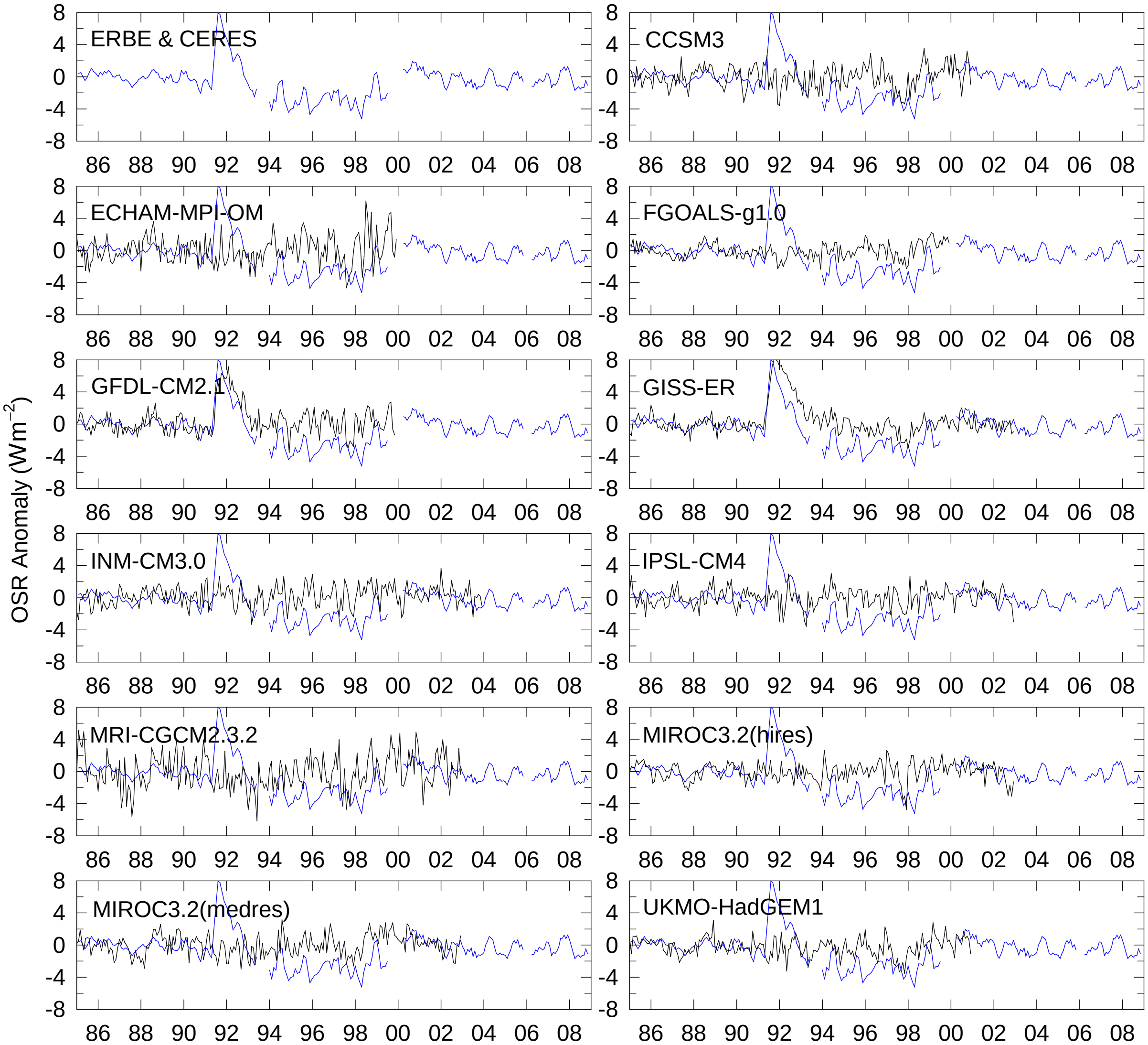

OLR Anomaly (Wm^{-2})
ERBE & CERES
CCSM3
ECHAM-MPI-OM
FGOALS-g1.0
GFDL-CM2.1
GISS-ER
INM-CM3.0
IPSL-CM4
MRI-CGCM2.3.2
MIROC3.2(hires)
MIROC3.2(medres)
UKMO-HadGEM1
86 88 90 92 94 96 98 00 02 04 06 08
8 4 0 -4

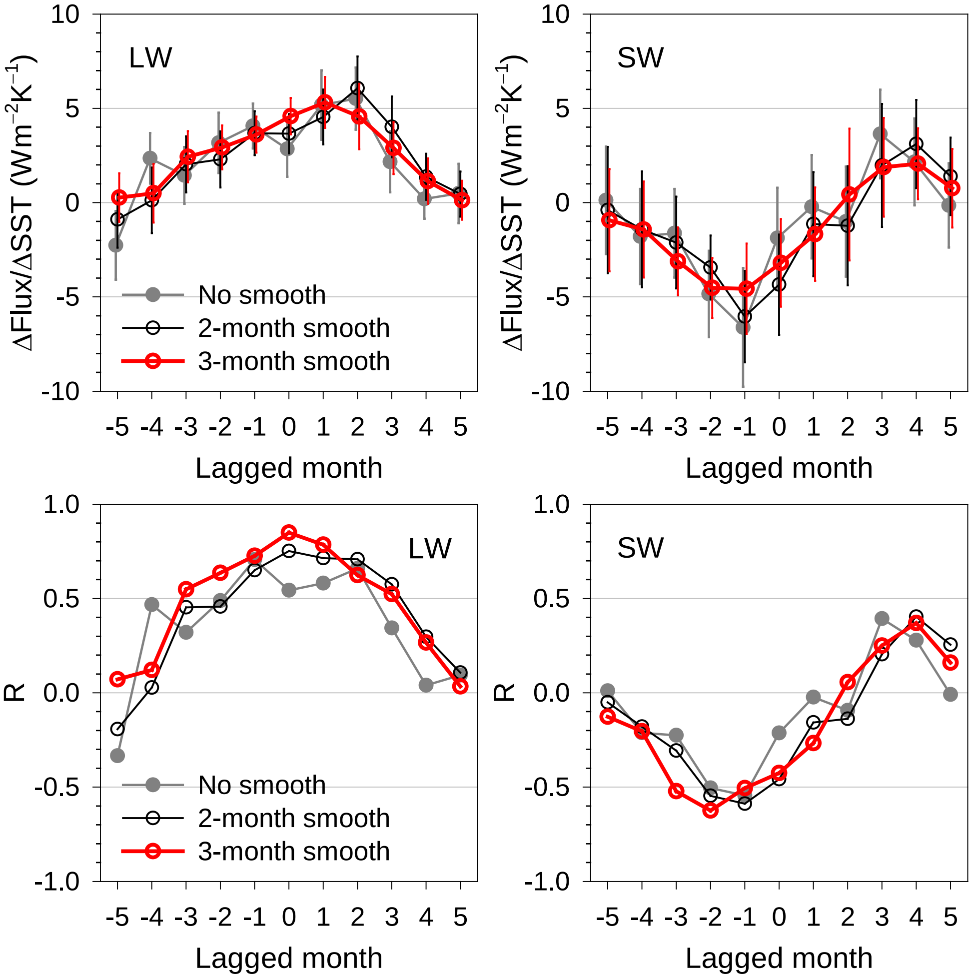
Figure 5



496

497 Figure 6

498



499

500 Figure 7

501

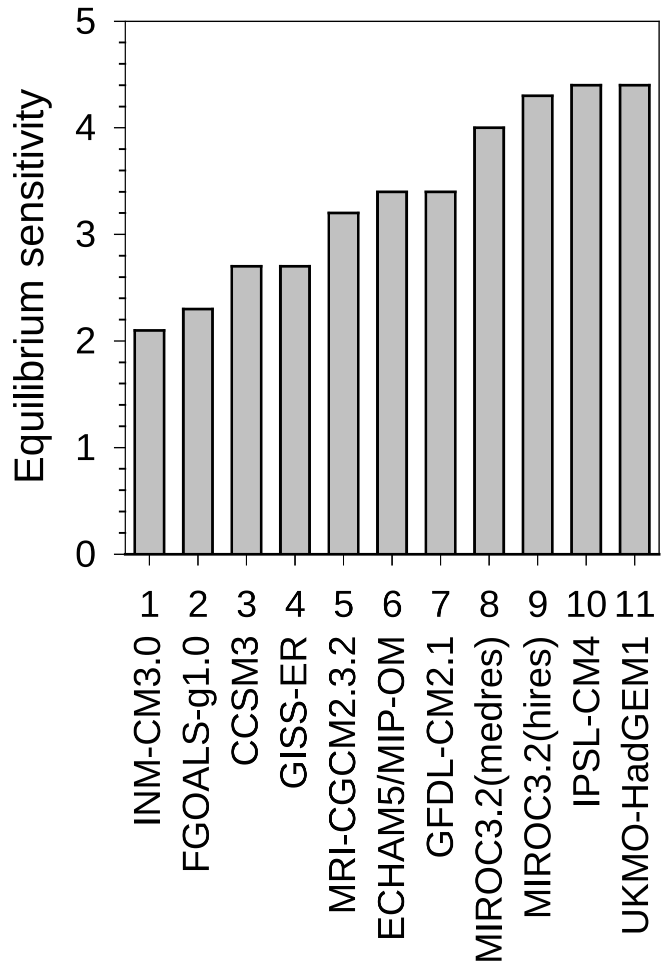
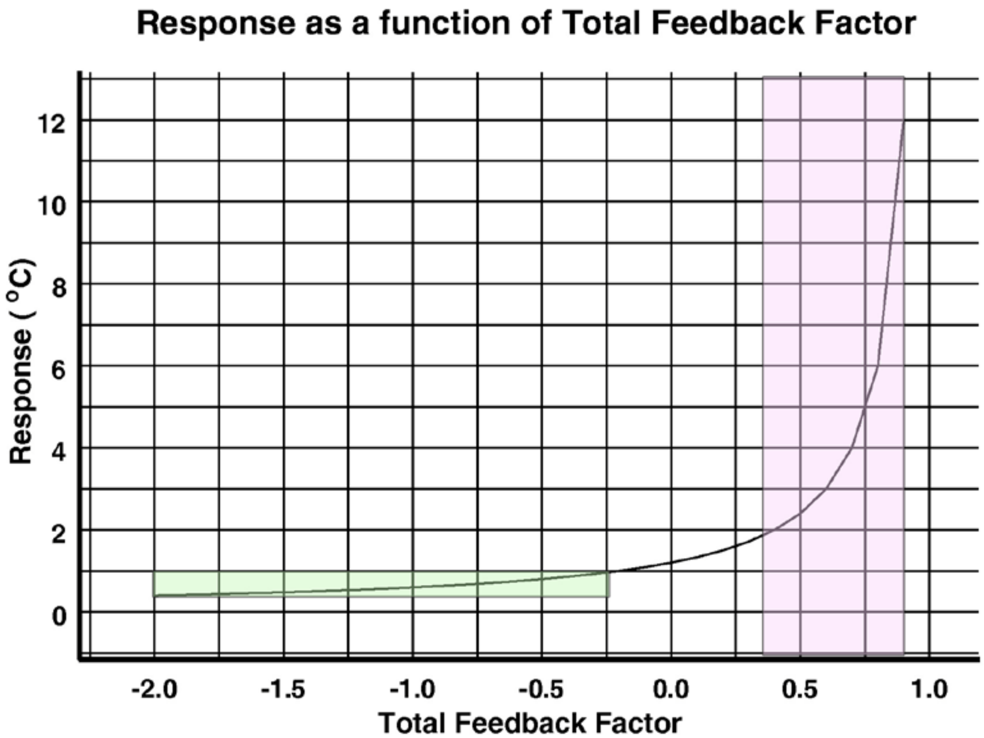

Equilibrium sensitivity
0
1
2
3
4
5
1 2 3 4 5 6 7 8 9 10 11
INM-CM3.0
FGOALS-g1.0
CCSM3
GISS-ER
MRI-CGCM2.3.2
ECHAM5/MIP-OM
GFDL-CM2.1
MIROC3.2(medres)
MIROC3.2(hires)
IPSL-CM4
UKMO-HadGEM1

Figure 8



505

506 Figure 9

507

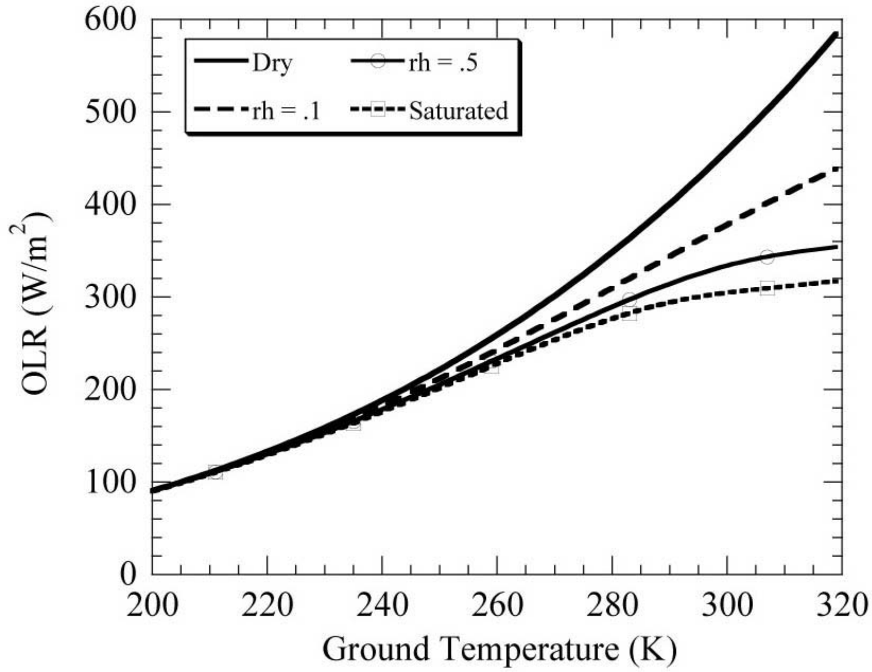

Dry
rh = .5
rh = .1
Saturated
OLR (W/m2)
Ground Temperature (K)
0
100
200
300
400
500
600
200
220
240
260
280
300
320

Figure 10