# On the observational determination of climate sensitivity and its implications

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**Abstract**

We estimate climate sensitivity from observations, using the deseasonalized fluctuations in sea surface temperatures (SSTs) and the concurrent fluctuations in the top-of-atmosphere (TOA) outgoing radiation from the ERBE (1985-1999) and CERES (2000-2008) satellite instruments. Distinct periods of warming and cooling in the SSTs were used to evaluate feedbacks. An earlier study (Lindzen RS, Choi Y-S (2009) Geophys Res Lett 36:L16705) was subject to significant criticisms. The present paper is an expansion of the earlier paper where the various criticisms are taken into account. The present analysis accounts for the 72 day precession period for the ERBE satellite in a more appropriate manner than in the earlier paper. We also attempt to distinguish 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. We argue that feedbacks are largely concentrated in the tropics; however, the tropical feedbacks are adjusted to account for their impact on the globe as a whole. We again find that the outgoing radiation resulting from SST fluctuations exceeds the zero-feedback response thus implying negative feedback. In contrast to this, the calculated TOA outgoing radiation fluxes from 11 atmospheric models forced by the observed SST are less than the zero-feedback response, consistent with the positive feedbacks that characterize these models. The results imply that the models are exaggerating climate sensitivity.

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## 1. Introduction

The heart of the global warming issue is so-called greenhouse warming. This 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) interfere with the cooling of the planet, forcing it to become warmer in order to emit sufficient infrared radiation to balance the net incoming sunlight. By net incoming sunlight, we mean that portion of the sun's radiation that is not reflected back to space by clouds, aerosols and the earth's surface. $CO_2$, a relatively minor greenhouse gas, has increased significantly since the beginning of the industrial age from about 280 ppmv to about 390 ppmv, presumably due mostly to man's emissions. This is the focus of current concerns. However, warming from a doubling of $CO_2$ would only be about 1°C (based on simple calculations where the radiation altitude and the Planck temperature depend on wavelength in accordance with the attenuation coefficients of well-mixed $CO_2$ molecules; a doubling of any concentration in ppmv produces the same warming because of the logarithmic dependence of $CO_2$'s absorption on the amount of $CO_2$) (1).

This modest warming is much less than current climate models suggest for a doubling of $CO_2$. Models predict warming of from 1.5°C to 5°C and even more for a doubling of CO2. Model predictions depend on 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 visible reflectivity decreases, causing increased solar absorption and warming of the earth.

Cloud feedbacks are still considered to be highly uncertain (1), but the fact that these feedbacks are strongly positive in most models is considered to be an indication that the result is basically correct. Methodologically, this is unsatisfactory. Ideally, one would seek an observational test of the issue. Here we suggest that it may be possible to test the issue with existing data from satellites.

Indeed, an earlier study by Forster and Gregory (2) examined the anomaly of the annual mean temperature and radiative flux observed from a satellite. However, with the annual time scale, the signal of short-term feedback associated with water vapor and clouds can be contaminated by unknown time-varying radiative forcing in nature, and the accurate feedbacks cannot be diagnosed (3). In a recent paper (4) we attempted to resolve these issues though, as we will show in this paper, the details of that paper were, in important ways, also incorrect (5-7). There were four major criticisms to Lindzen and Choi (4): (i) statistical insignificance of the results, (ii) misinterpretation of air-sea interaction in the Tropics, (iii) misuse of uncoupled atmospheric models, and (iv) incorrect computation of climate sensitivity. The present paper responds to the criticism, and corrects the earlier approach. The earlier results are not significantly altered.

## 2. Feedback formalism

In the absence of feedbacks, the behavior of the climate system can be described by Fig. 1a $\Delta Q$ is the radiative forcing, $G_0$ is the zero-feedback response function of the climate system, and $\Delta T_0$ is the response of the climate system in the absence of feedbacks. The checkered circle is a node. Fig. 1a symbolically shows the temperature increment, $\Delta T_0$, that a forcing increment, $\Delta Q$, would produce with no feedback,

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

It is generally accepted that in the absence of feedback, a doubling of $CO_2$ will cause a forcing of

$\Delta Q \approx 3.7\ \mathrm{Wm}^{-2}$ and will increase the temperature by $\Delta T_0 \approx 1.1$ K (8, 9). We therefore take the zero-feedback response function of Eq. (1) to be $G_0 \approx 0.3$ (=1.1/3.7) K W$^{-1}$ m$^{2}$ for the earth as a whole.

With feedback, Fig. 1a is modified to Fig. 1b. The response is now

$$\Delta T = G_0(\Delta Q + F\Delta T) \tag{2}$$

Here $F$ is a feedback function that represents all changes in the climate system (for example, changes in cloud cover and humidity) that act to increase or decrease feedback-free effects. Thus, $F$ should not include the zero-feedback (ZFB) response to $\Delta T$ that is already incorporated into $G_0$. The choice of ZFB response for the tropics in Lindzen and Choi (4) is certainly incorrect in this respect (5, 6). At present, the best choice seems to remain $1/G_0$ (3.3 W m$^{-2}$ K$^{-1}$) (9, 10).

Solving Eq. (2) for the temperature increment $\Delta T$ and inserting Eq. (1) into Eq. (2) we find

$$\Delta T = \frac{\Delta T_0}{1-f} \tag{3}$$

The dimensionless feedback fraction is $f = F\,G_0$. Also, dividing Eq. (2) by $G_0$, we obtain

$$-\frac{f}{G_0}\Delta T = \Delta Q - \frac{\Delta T}{G_0} \tag{4}$$

When looking at the observations, $\Delta Q$ and $\Delta T$ in Eq. (4) may be replaced by the change in outgoing net radiative flux, ΔFlux, and the change in sea surface temperature, ΔSST, respectively, leading to

$$-\frac{f}{G_0}\Delta\mathrm{SST} = \Delta\mathrm{Flux} - \mathrm{ZFB} \tag{5}$$

where ZFB indicates the zero-feedback response to ΔSST, i.e., $\Delta\mathrm{SST}/G_0$. The quantities on the right side of the equation indicate the amount by which feedbacks supplement ZFB response to

ΔFlux. At this point, it is crucial to recognize that our equations are predicated on the assumption that the ΔSST to which the feedbacks are responding is produced by ΔFlux. Physically, however, we expect that any fluctuation in temperature should elicit the same flux regardless of the origin of temperature change. Note that the natural forcing, ΔSST, that can be observed, is actually not the same as the equilibrium response temperature $\Delta T$ in Eq. (4). 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 (TOA) is restored to zero. The choice of the short intervals may serve to remove some natural time-varying radiative forcing that contaminates the feedback signal (3). As explained in Lindzen and Choi (4), it is 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, in the tropics, 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 (11).

Recent studies argued that quantification of feedback based on Eq. (5) is inadequate with our available tropical domain due to the exchange of energy between the tropics and the extratropics (5, 7). However, there are good reasons to consider the tropics; for example, concentration of water vapor in the tropics (see supporting information (SI) for more explanation). However, when restricting ourselves to tropical feedbacks, Eq. (5) must be replaced by

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

where the factor 2 results from the sharing of the tropical feedbacks over the globe following the methodology of Lindzen, Chou and Hou (12), (hereafter LCH01) and Lindzen, Hou and Farrell (13); that is to say that the contribution of the tropical feedback to the global feedback is only about half of the tropical feedback. The precise choice of this factor does not affect the major conclusion of this study (SI for more details).

From Eq. (6), the longwave (LW) and shortwave (SW) contributions to $f$ are given by

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

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

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. Since we know the value of $G_0$, the experimentally determined slope (the quantity on the right side of Eq. (7)) allows us to evaluate the magnitude and sign of the feedback factor $f$ provided that we also know the value of the ZFB response (ΔSST/$G_0$ in this study). For observed variations, the changes in radiation (associated for example with volcanoes or non-feedback changes in clouds) 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 less of an issue with model results from AMIP (Atmospheric Model Intercomparison Project) 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, and this is true for AMIP as well as for nature. Note that this study deals with observed outgoing fluxes, but does not specifically identify the origin of the changes (see SI for more details).

**3. The data and their problems**

SST is measured (14), and is always fluctuating (viz. Fig. 2). To relate this SST with the flux in the entire tropics, the SST anomaly was scaled by a factor of 0.78 (the area fraction of the ocean to the tropics). High frequency fluctuations, however, make it difficult to objectively identify the beginning and end of warming and cooling intervals (5). This ambiguity is eliminated with a 3 point centered smoother. (A two point lagged smoother works too.) In addition, the net outgoing radiative flux from the earth has been monitored since 1985 by the ERBE (Earth Radiation Budget Experiment) instrument (15) (nonscanner edition 3) aboard ERBS (Earth Radiation Budget Satellite) satellite, and since 2000 by the CERES (Clouds and the Earth's Radiant Energy System) instrument (ES4 FM1 edition 2) aboard the Terra satellite (16). The results for both LW radiation and SW radiation are shown in Fig. 3. The sum is the net outgoing flux.

With ERBE data, there is the problem of satellite precession with a period of 72 days, although in the deep tropics all clock hours are covered in 36 days. In Lindzen and Choi (4) 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 (7) 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 take a 3 point smoothing in the flux data to minimize the effect of noise. This is also why we use the 36-day averaged SST for 1985–1999 and monthly SST for 2000–2009 in Fig. 2.

The discontinuity between the two datasets requires comment. There is the long-term discrepancy of the average which is believed to be due to the absolute calibration problem (up to 3 W $m^{-2}$) (17). With CERES, we attempt to resolve the spectral darkening problem by multiplying SW flux by the scale factor (up to 1.011) from Matthews et al. (18). However, this long-term stability should not matter for our analysis which considers fluctuations over a few months for which the drift is insignificant. There is also the higher seasonal fluctuation in CERES SW radiation than in ERBE. The bias is up to 6.0 W $m^{-2}$ as estimated by Young et al. (19). This is attributed to different sampling patterns; ie, 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, we reference the anomalies for radiative flux separately 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.

The quality of ERBE and CERES data are best in 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 (17). 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, we believe that the water vapor feedback is primarily restricted to the tropics, and there are reasons to suppose that this is also the case for cloud feedbacks (SI). The methodology developed in LCH01 permits the easy evaluation of the contribution of tropical processes to global values. As noted by LCH01, this does not preclude there being extratropical contributions as well, but these are not considered in the present paper.

Finally, there is the serious issue of distinguishing atmospheric phenomena involving changes in outgoing radiation that result from processes other than feedbacks (Pinatubo and non-feedback cloud variations for example) and which cause changes in SST, from those that are caused by changes in SST (namely the feedbacks we wish to evaluate) (5, 6). Our crude approach to this is to examine the effect of fluxes with 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 Materials and Methods. To be sure, Fourier transform methods wherein one investigates phase leads and lags might normally be cleaner, but, given the gaps in the radiation data as well as the incompatibilities between ERBE and CERES, the present approach which focuses on individual warming and cooling events seems more appropriate.

Turning to the models, AMIP is responsible for intercomparing atmospheric models used by the IPCC (the Intergovernmental Panel on Climate Change); the AMIP models are forced by the same observed SSTs shown in Fig. 2. We have obtained the calculated changes in both SW and LW radiation from the AMIP models. These results are shown in Figs. S4 and S5 where the observed results are also superimposed for comparison. We can already see that there are significant differences. In addition, we will also consider results from CMIP (the Coupled Model Intercomparison Project), where coupled ocean-atmosphere models were intercompared.

## 4. Results

### 4.1. Climate sensitivity in observation and comparison to AMIP models

Given the above, it is now be possible to directly test the ability of models to adequately simulate the sensitivity of climate (see Materials and Methods). Fig. 4 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. 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 and non-feedback cloud fluctuations lead to changes in SW reflection and associated fluctuations in surface temperature.

We see two extrema 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 anomalous 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 $m^{-2}$ $K^{-1}$). Under such circumstances, we expect the maximum slope for SW radiation in Fig. 4 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)). With the slope and its standard error, the feedback fractions for LW, SW, and total radiation ($f_{SW}$, $f_{LW}$, and $f_{Total}$) are obtained via Eqs. (6) and (7). Finally, with $f_{Total}$, the equilibrium climate sensitivity for a doubling of $CO_2$ is obtained via Eq. (3). Here the statistical confidence intervals of the sensitivity estimate at 90%, 95%, and 99% levels are also calculated by the standard error of the feedback fraction $f_{Total}$. This interval would prevent any problems arising from limited sampling. As a result, the climate sensitivity for a doubling of $CO_2$ is estimated to be 0.7K (with the confidence interval 0.5K–1.3K at 99% levels). This observational result shows that model sensitivities indicated by the IPCC AR4 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 IPCC AR4. To the AMIP results, for which there was less ambiguity as to whether fluxes constituted a response (noise still exists due to autonomous cloud fluctuations), the same approach as that for the observations was applied. Maximum R occurs at zero lag in both LW and SW radiation, so we simply chose the AMIP fluxes without 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 climate sensitivities in K for a doubling of $CO_2$ implied by feedback factors $f$ in Table 2 with those in IPCC AR4. To indicate statistical significance of our results obtained from limited sampling, we also calculated the confidence intervals of the climate sensitivity using the standard errors of $f$ in Table 2. All the sensitivities in IPCC AR4 are within the 90% confidence intervals of our sensitivity estimates. The agreement does not seem notable, but this is because, for positive feedbacks, sensitivity is strongly affected by small changes in $f$ that are associated standard errors in Table 2. Consequently, the confidence intervals include “infinity”. This is seen in Fig. 5 in the pink region. It has, in fact, been suggested by Roe and Baker (20), 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 (21). By contrast, in the green region, which corresponds to the observed feedback factors, sensitivity is much better constrained.

### 4.2. Comparison to CMIP models and their limitations

It has been argued that CMIP models are more appropriate for the present purpose since the uncoupled AMIP modes are prescribed with incomplete forcings of SST (5). However, it is precisely for this reason that AMIP models are preferred for our purpose. Note that we are considering atmospheric feedbacks to SST fluctuations. As already seen, in analyzing observed behavior, the presence of SST variations that are primarily caused by atmospheric changes (from volcanoes, non-feedback cloud variations, etc.) leads to difficulty in distinguishing SST variations that are primarily forcing atmospheric changes (i.e., feedbacks). This situation is much simpler with AMIP results since we can be sure that SST variations (which are forced to be the same as observed SST) cannot respond to atmospheric changes. The fact that CMIP SST variations are significantly different from observed SST variations further makes it unlikely that the model atmospheric processes are implicitly forcing the SST's used for AMIP. Note that important ocean phenomena such as El Niño-Southern Oscillation and Pacific Decadal Oscillation are generally misrepresented by CMIP models. As noted, AMIP results are still subject to noise since outgoing radiation includes changes associated with non-feedback cloud variations.

In applying our methodology to CMIP, we see that coupled models differ in the behaviors of SST, and the intervals of SST must be selected differently. Some models have much smaller variability of SST than nature and only a few intervals of SST could be selected. As we see in Fig. 6 at a glance, the CMIP results (black dots) display behavior somewhat similar to ERBE and CERES results (red open circles) with respect to lags. However, when identifying each number, we found that the results are quantitatively ambiguous. The slope $\Delta$OLR/$\Delta$SST for lag = 1 is between 0.6 and 5.8 though it remains robust that LW feedbacks in most models are higher than nature. Not surprisingly, the inconsistent LW feedback was also shown in previous studies by Forster and Gregory (2) and Forster and Taylor (22). The slope $\Delta$SWR/$\Delta$SST for lag = 3 is between −3.4 and 3.9 so that one cannot precisely determine the feedback in the models. These values, moreover, do not correspond well to the independently known model climate sensitivities in IPCC AR4. Based on our simple model (Materials and Methods), this ambiguity results mainly from non-feedback internal radiative (cloud-induced) change that changes SST (see Fig. S3 and SI for more information). Also, such cloud-induced radiative change can generate the anomalous sinusoidal shape of the slopes $\Delta$SWR/$\Delta$SST with respect to lags as shown in Fig. 6. Therefore, previous studies that use the slopes $\Delta$SWR/$\Delta$SST at zero lag (2, 5) may misinterpret SW feedback. This confirms that for more accurate estimation of 'model' feedbacks, AMIP models are more appropriate than CMIP models. Furthermore, nature is better than CMIP because nature properly displays the real magnitude of SST forcing and the associated atmospheric changes, even though it also includes SST response to radiative forcing.

## 5. Conclusions and discussions

We have corrected the approach of Lindzen and Choi (4), based on all the criticisms made of the earlier work (5-7). First of all, to improve the statistical significance of the results, we supplemented ERBE data with CERES data, filtered out data noise with 3-month smoothing, objectively chose the intervals based on the smoothed data, and provided confidence intervals for all sensitivity estimates. These constraints helped us to more accurately obtain climate feedback factors than with the original use of monthly data. Next, our new formulas for climate feedback and sensitivity reflected sharing of tropical feedback with the globe, so that the tropical region was identified as an open system. Last, the feedback factors inferred from the atmospheric models are more consistent with IPCC-defined climate sensitivity than those from the coupled models. This is because, in the presence of cloud-induced radiative changes altering SST, the climate feedback estimates by the present approach tends to be inaccurate. With all corrections, the conclusion appears to be that all current models seem to exaggerate climate sensitivity (some greatly).

Our analysis of the data only demands relative instrumental stability over short periods, and is largely independent of long term drift. Concerning the different sampling from the ERBE and CERES instruments, Murphy et al. (23) repeated the Forster and Gregory (2) analysis for the CERES and found very different values than those from the ERBE. However, in this study, the addition of CERES data to the ERBE data does little to change the results for $\Delta$Flux/$\Delta$SST – except that its value is raised a little (as is also true when only CERES data is used.).

Our study also suggests that, in current coupled atmosphere-ocean models, the atmosphere and ocean are too weakly coupled since thermal coupling is inversely proportional to sensitivity (11). It has been noted by Newman et al. (24) 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 be noted that there have been independent tests that also suggest sensitivities less than predicted by current models. These tests are based on response to sequences of volcanic eruptions (11), on the vertical structure of observed versus modeled temperature increase (25, 26), on ocean heating (9, 27), and on satellite observations (3). 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 (28), but these are not really tests since the forcing is essentially unknown given major uncertainties in clouds and dust loading.

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 (29). As noted, however, in Lindzen (30), 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.

## Materials and Methods

### a. Simple model analysis

Following Spencer and Braswell (3), we assume an hypothetical climate system with uniform temperature and heat capacity, for which SST and forcing are time-varying. Then the model equation of the system is

$$C_p \left[ \frac{d\Delta T}{dt} \right] = Q(t) - F \cdot \Delta T(t) \tag{8}$$

where $C_p$ is the bulk heat capacity of the system (14 yr W m$^{-2}$ K$^{-1}$ in this study, from ref. 9); $\Delta T$ is SST deviation away from an equilibrium state of energy balance; $F$ is the feedback function that is the same as the definition in Eq. (2); $Q$ is any forcing that changes SST (2, 3). $Q$ is composed of three sources of forcing: (i) external radiative forcing (from anthropogenic greenhouse gas emission, e.g.), (ii) internal non-radiative forcing (from heat transfer from ocean, e.g.), and (iii) internal radiative forcing (from water vapor or clouds, e.g.). Among the three forcings, the two external and internal 'radiative' forcings, and $F{\cdot}\Delta T$(t) constitute TOA net radiative flux anomaly, i.e., $\Delta$Flux. This simple model is used, in order to investigate sensitivity of our approach to feedback function and to radiative forcing. Results are shown in SI.

**b. Feedback estimation method**

As already noted, the data need to be smoothed first to minimize noise. Then the procedure is simply to identify intervals of maximum change in ΔSST (red and blue in Fig. 2), and for each such interval, to find the change in flux. The reasoning for this is that, by definition, a temperature change is required to produce radiative feedback, and so the greatest signal (and least noise) in the estimation of feedback should be associated with the largest temperature changes. Thus, it is advisable, but not essential, to restrict oneself to changes greater than 0.1°C; in fact, the impact of thresholds for ΔSST on the statistics of the results is minor (1). Let us define $t_1$, $t_2$,…, $t_\mathrm{m}$ as selected time steps that correspond to the starting and the ending points of intervals. Again, for stable estimation of ΔFlux/ΔSST, the time steps should be selected based on the maximum and minimum of the smoothed SST. ΔFlux/ΔSST can be basically obtained by Flux($t_{i+1}$) − Flux($t_i$) divided by SST($t_{i+1}$) − SST($t_i$) where $t_i$ is $i$th selected time steps ($i$ = 1, 2, …, m−1). As there are many intervals, the final 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 though the choice of regression model turns out to also be minor.

One must also distinguish ΔSST's that are forcing changes in ΔFlux, from responses to ΔFlux. Otherwise, ΔFlux/ΔSST can vary (5) and/or may not represent feedbacks that we wish to determine. To avoid such a problem, we use lag-lead methods (e.g., use of Flux($t$+lag) and SST($t$)) for ERBE 36-day and CERES monthly data). 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.

The above procedures help to obtain a more accurate climate feedback factor than the use of original monthly data. This was tested by a Monte-Carlo test of the above simple feedback-forcing model. With minimal cloud-induced radiative changes, our method always gives the feedback factor close to the true value (Fig. S2), whereas the conventional regression method with monthly data tends to underestimate the feedback particularly in the presence of increasing external radiative forcing (e.g., increasing $CO_2$ forcing) (3).

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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 K) for a doubling of $CO_2$ for 90%, 95%, and 99% confidence levels.

**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 intervals, R is the correlation coefficient, and SE is the standard error of ΔFlux/ΔSST.

**Table 3**. Comparison of model equilibrium climate sensitivities (in K) for a doubling of $CO_2$ defined from IPCC AR4 and estimated from feedback factors in this study. The obvious difference between two columns labeled 'sensitivity' is discussed in more detail in the last paragraph of section 3.1. The estimated climate sensitivities for models as well as their confidence intervals are given for 90%, 95%, and 99% confidence levels.

## Figure Legends

**Fig. 1**. A schematic for the behavior of the climate system in the absence of feedbacks (a), in the presence of feedbacks (b).

**Fig. 2**. 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. 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. 3).

**Fig. 3**. The same as Fig. 2 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 ref. 17.

**Fig. 4**. 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).

**Fig. 5**. Sensitivity vs. feedback factor.

**Fig. 6.** Same as Fig. 4, but for the 10 CMIP models (black dots); GISS model was excluded because only few intervals of SST are obtained. The values for the 3-month smoothing in Fig. 4 are superimposed by red dots.

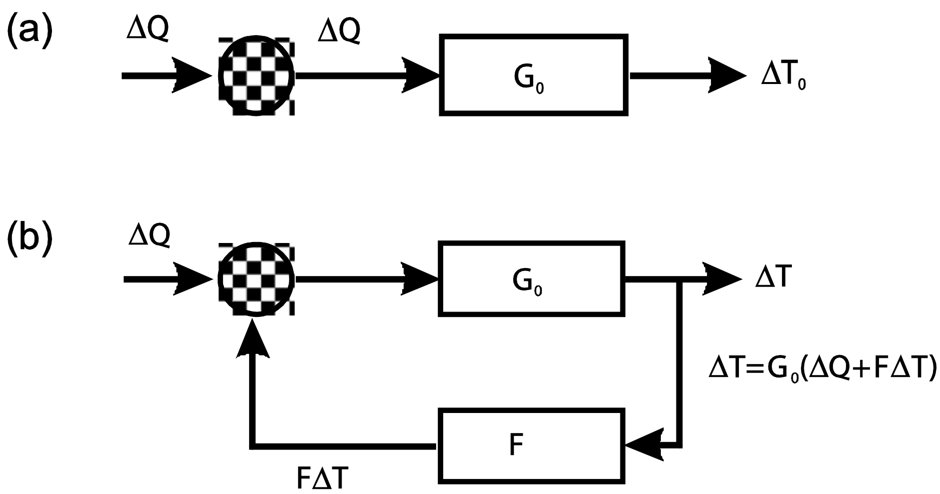


Figure 1

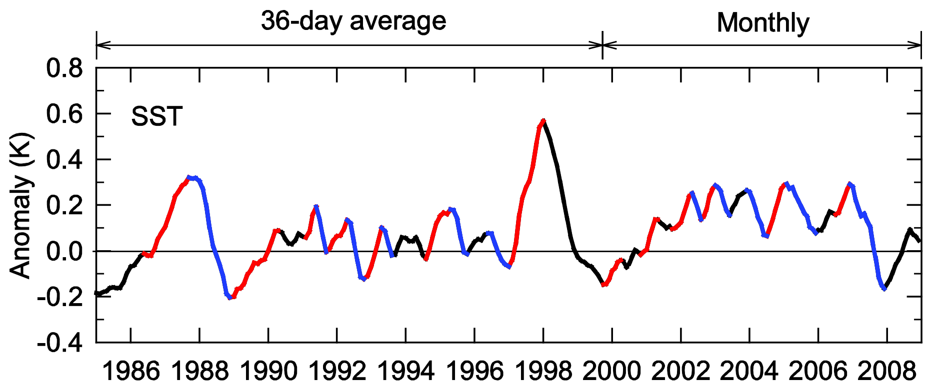
36-day average
Monthly
SST
Anomaly (K)
0.8
0.6
0.4
0.2
0.0
-0.2
-0.4
1986 1988 1990 1992 1994 1996 1998 2000 2002 2004 2006 2008

Figure 2

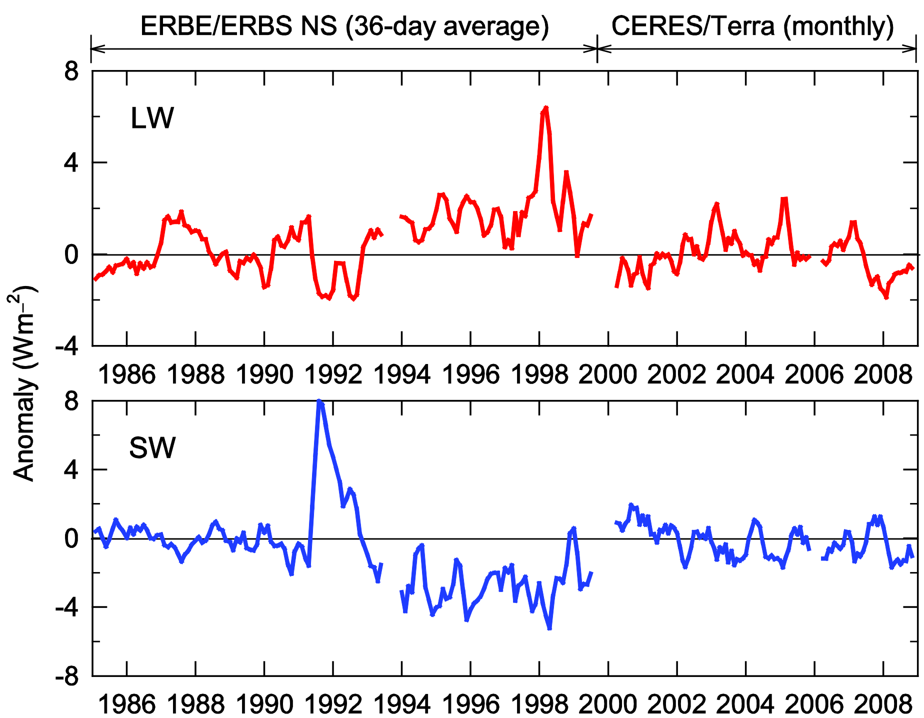
ERBE/ERBS NS (36-day average)
CERES/Terra (monthly)
LW
SW
Anomaly (Wm$^{-2}$)
8
4
0
-4
-8
1986 1988 1990 1992 1994 1996 1998 2000 2002 2004 2006 2008

Figure 3

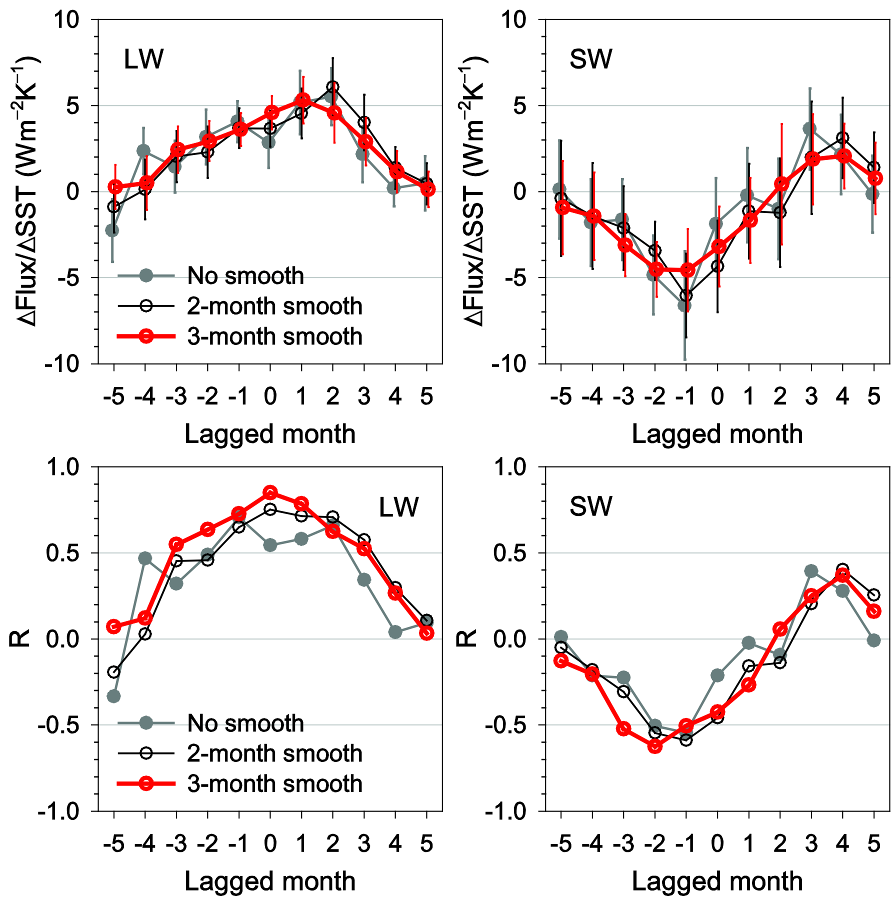
LW
SW
ΔFlux/ΔSST (Wm$^{-2}$K$^{-1}$)
R
No smooth
2-month smooth
3-month smooth
Lagged month
10
5
0
-5
-10
1.0
0.5
0.0
-0.5
-1.0
-5 -4 -3 -2 -1 0 1 2 3 4 5

Figure 4

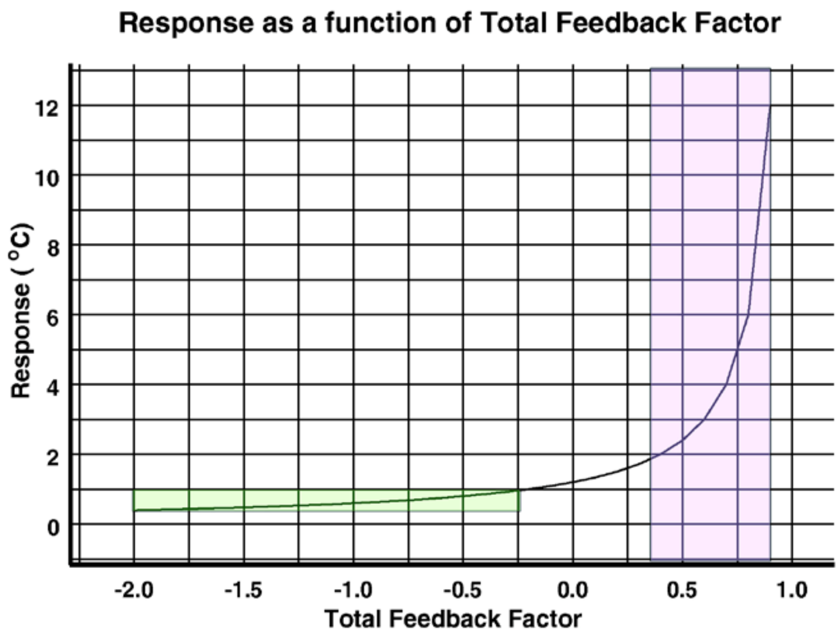
Response as a function of Total Feedback Factor
Response ( °C)
12
10
8
6
4
2
0
-2.0
-1.5
-1.0
-0.5
0.0
0.5
1.0
Total Feedback Factor

Figure 5

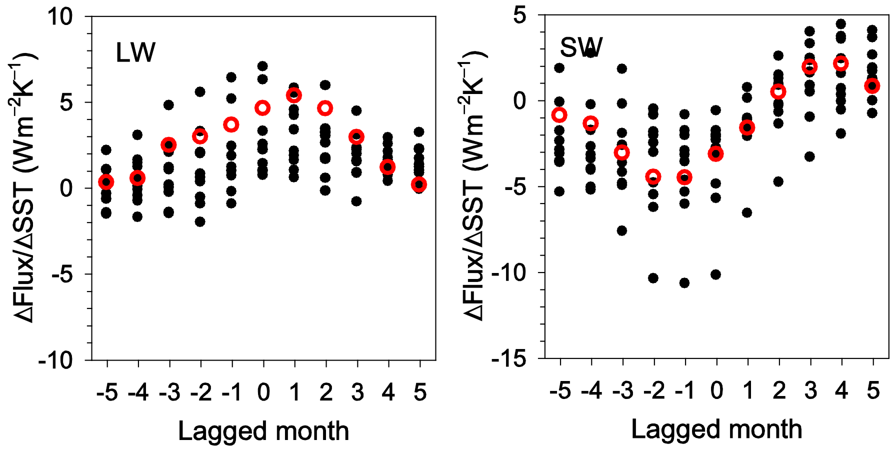


Figure 6

**Supporting Information**

# On the observational determination of climate sensitivity and its implications

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## 1. 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 mostly 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 (31) examined outgoing radiation as a function of surface temperature theoretically for atmospheres with constant relative humidity. His results are shown in Fig. S1.

Relative humidity is low in the extratropics, while it is high in the extratropics. 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 (32). As noted by LCH01, 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). The choice of larger factor leads to smaller contribution of tropical feedback to global sensitivity, but the effect on the climate sensitivity estimated from the observation is minor. For example, with the factor of 3, climate sensitivity from the observation and the models is 0.8 K and a higher value (between 1.3 K and 6.4 K), respectively. With the factor of 1.5, global equilibrium sensitivity from the observation and the models is 0.6 K and any value higher than 1.6 K, respectively. Note that, as in LCH01 (12), we are not discounting the possibility of feedbacks in the extratropics, but rather we are focusing on the tropical contribution to global feedbacks.

## 2. 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 (4). Accordingly, with respect to separating longwave and shortwave feedbacks, the interpretation by Lindzen and Choi (4) 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 (33), but, if this is true, this feedback is presumably cancelled by a negative infrared feedback such as that proposed by LCH01 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 (10, 33). 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 (34). 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 (34). 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; 35) and the CloudSat radar (CPR; 36) should provide a breakdown of cloud behavior with altitude which may give some insight into what exactly is contributing to the radiation.

**3. Simple model analysis**

The model system was basically forced by random internal non-radiative forcing changing

SST. Integration is done at monthly time steps. Fig. S2 compares the simple regression method and our method; given the feedback function $F$ = 6 W m$^{-2}$ K$^{-1}$ (it indicates negative feedback as it is larger than Planck response 3.3 W m$^{-2}$ K$^{-1}$), the system was additionally forced by random internal radiative forcing (the root mean square (RMS) is set to be 10% of RMS of internal non-radiative forcing, considering the observed magnitude of cloud forcing over the tropics), and transient external radiative forcing (0.4 W m$^{-2}$ per decades by increasing $CO_2$) (3). The maximum R occurs at small (zero or a month) lag and the corresponding ΔFlux/Δ$T$ (5.7 W m$^{-2}$ K$^{-1}$) is close to the assumed $F$, whereas the simple regression method underestimates ΔFlux/Δ$T$ (3.2 W m$^{-2}$ K$^{-1}$); the difference is statistically significant by a Monte-Carlo test (with 100 repetitions).

We now attempt to confine the simulation to SW radiation. This requires separation of a feedback function $F$ to those for SW and LW radiation (i.e., $F = F_{SW} + F_{LW}$). For convenience, $F_{LW}$ is set to zero. In SW, a positive feedback function indicates negative feedback. In addition, the transient external forcing originates from LW radiation and can be removed for the simulation of SW radiation. Sensitivity to the feedback function $F$ for SW radiation is shown in Fig. S3. A smaller feedback function turns out to have maximum R at larger lag, and the estimated climate feedbacks are the lagged response though it is somewhat less reliable than those at zero lag; the uncertainty of the feedback estimate from the lagged response is within ±0.3 for a small feedback function between −2 and 2. However, it is also clearly found that the smoother with a time window longer than three months effectively reduces the uncertainty and gives a much more accurate estimate of feedback. This indicates the necessity of stronger smoothing to minimize SST variations that are primarily forced by non-feedback atmospheric changes, but to retain SST variations that are primarily forcing atmospheric changes.

While the present SW simulation is for the absence of $F_{LW}$, many climate models as well as nature appear to have a positive $F_{LW}$, as shown by the slopes for LW in Tables 1 and 2. With a positive $F_{LW}$, the system with $F_{SW} < -2$ does not have to be very unstable and can also generate the sinusoidal shape of the slopes with respect to lags. In any of the cases, with either no internal cloud-induced radiative change or the prescribed temperature variation, $\Delta$Flux/$\Delta T$ at zero lag (with maximum R) is always identical to the assumed $F$. This explains why AMIP systematically shows maximum R at zero lag, while CMIP does not.

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**Figure Legends**

**Fig. S1**. OLR vs. surface temperature for water vapor in air, with relative humidity held fixed. The surface air pressure is 1bar. The temperature profile is the water/air moist adiabat. Calculations were carried out with the Community Climate Model radiation code (31).

**Fig. S2**. Comparison between simple regression method and the method used in this study, based on simple model results.

**Fig. S3**. Sensitivity of the method used in this study to feedback functions, based on simple model results.

**Fig. S4** Comparison of outgoing longwave radiation from AMIP models (black) and the observations (red) shown in Fig. 3.

**Fig. S5** Comparison of reflected shortwave radiation from AMIP models (black) and the observations (blue) shown in Fig. 3.

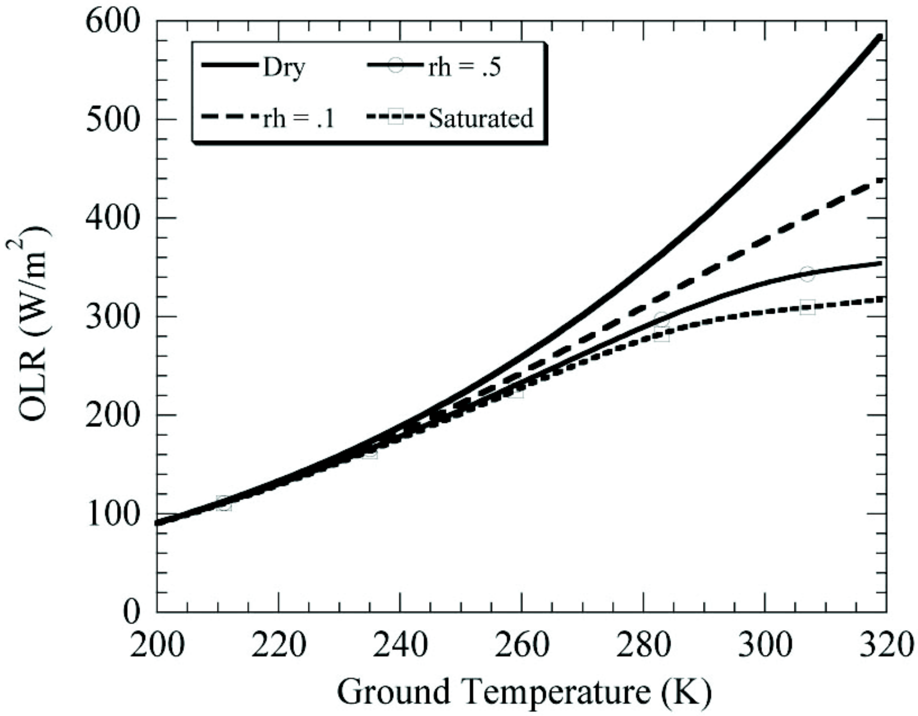

Dry
rh = .5
rh = .1
Saturated
OLR (W/m²)
600
500
400
300
200
100
0
200
220
240
260
280
300
320
Ground Temperature (K)

Figure S1

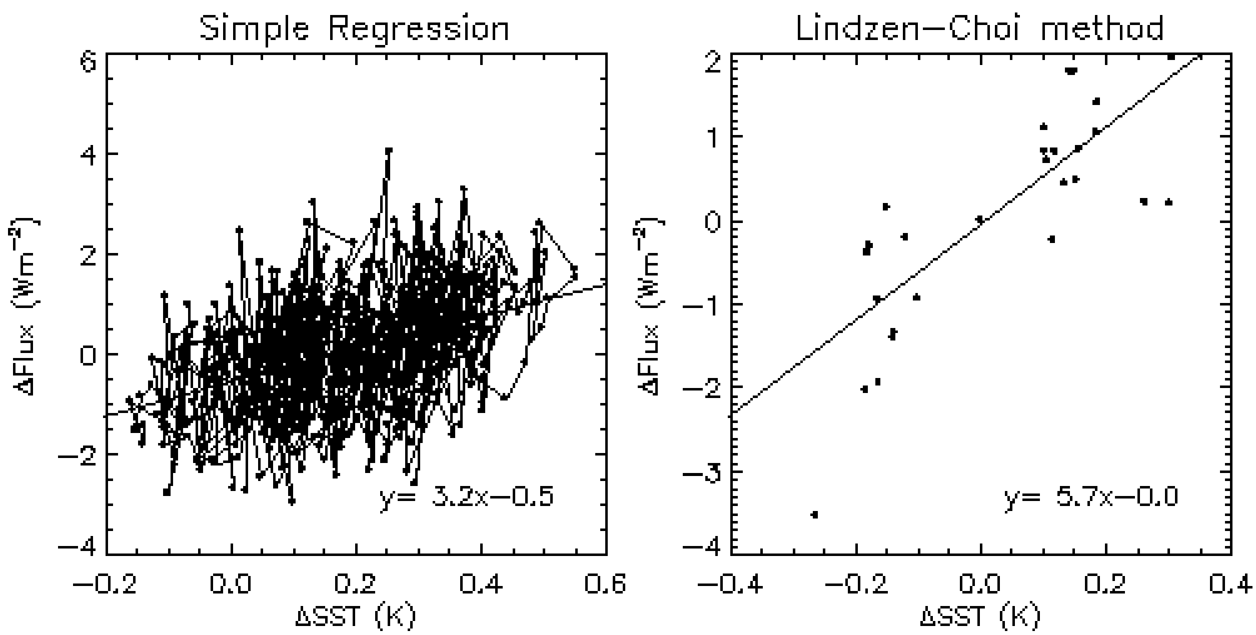
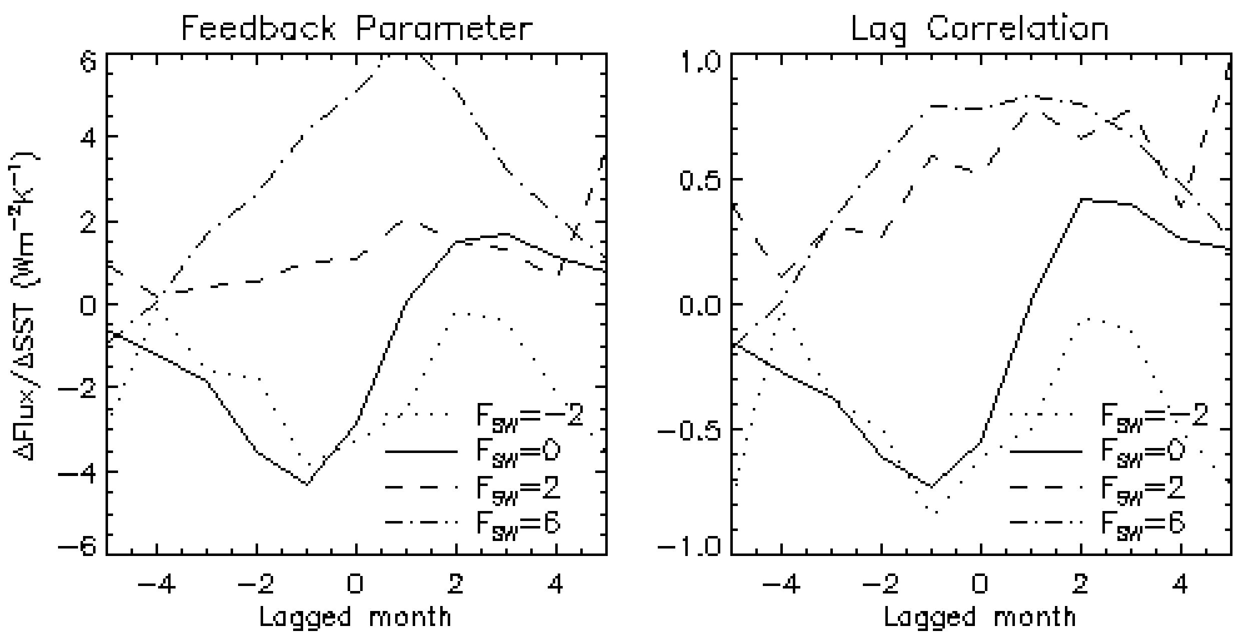
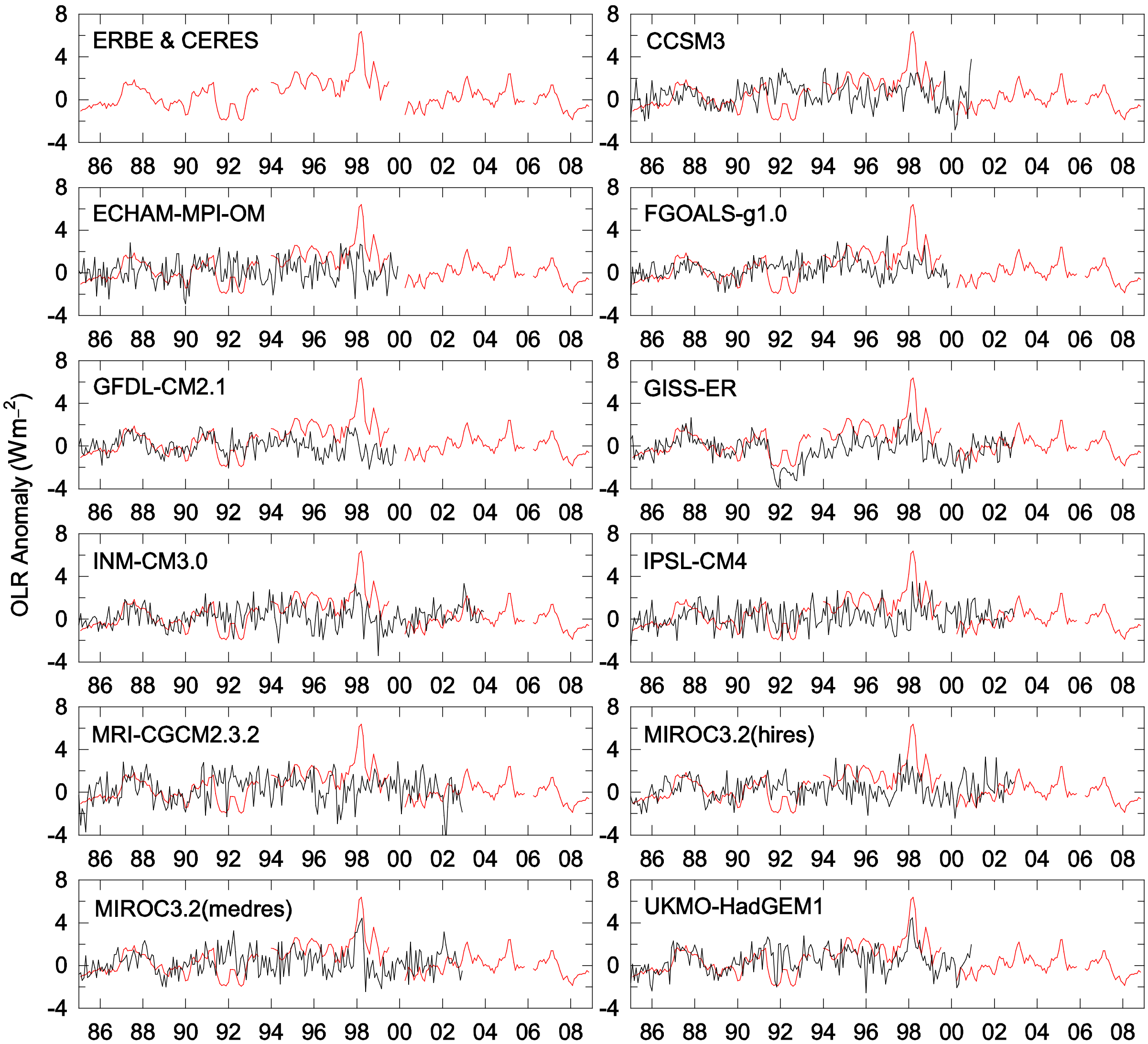

Simple Regression
Lindzen−Choi method
ΔFlux (Wm$^{-2}$)
ΔSST (K)
y= 3.2x−0.5
y= 5.7x−0.0

Figure S2



Feedback Parameter
ΔFlux/ΔSST (Wm^-2K^-1)
Lagged month
Lag Correlation
Lagged month
F_SW=−2
F_SW=0
F_SW=2
F_SW=6


Figure S3



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
8
4
0
-4
86 88 90 92 94 96 98 00 02 04 06 08


Figure S4

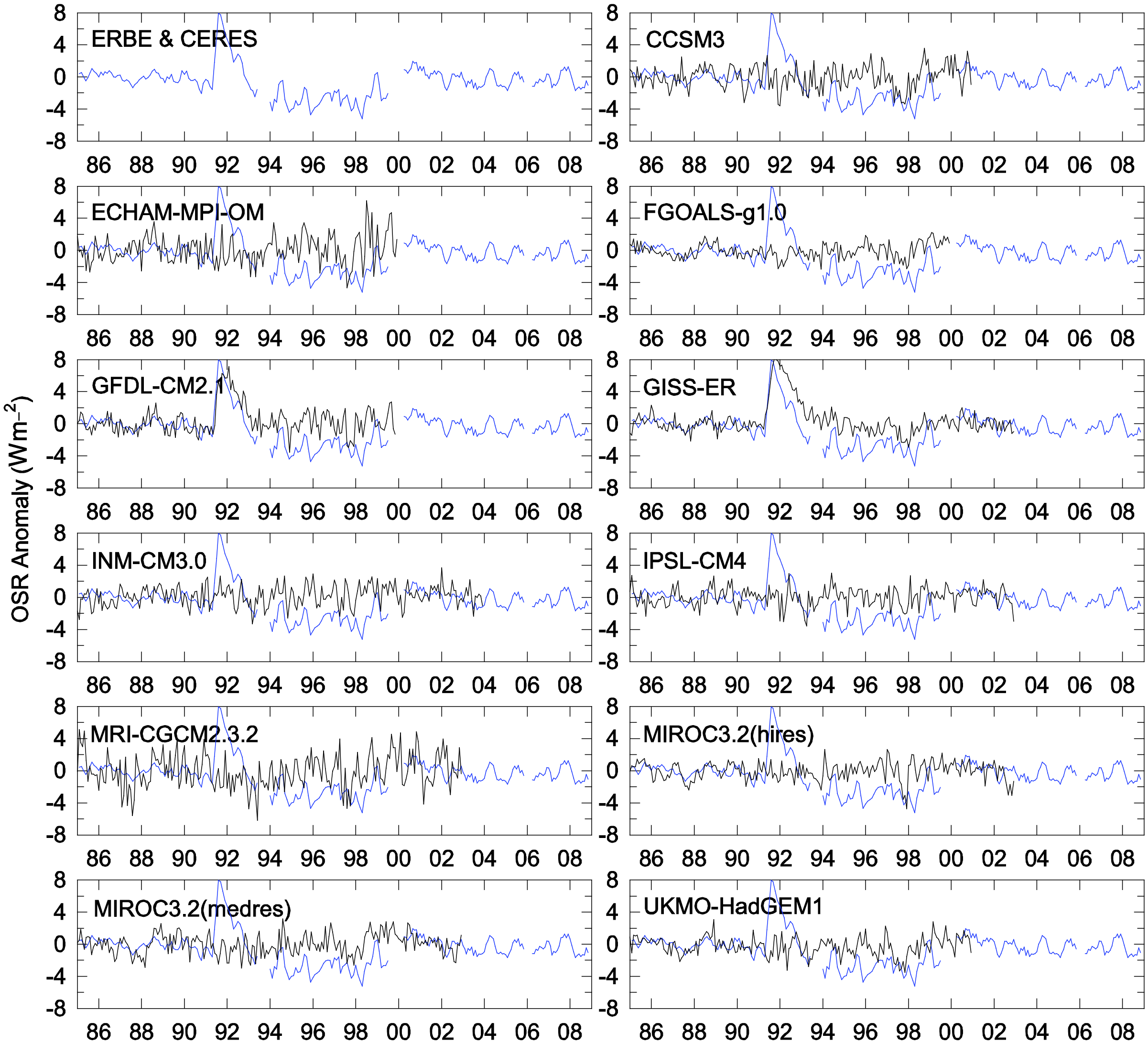

ERBE & CERES
CCSM3
ECHAM-MPI-OM
FGOALS-g1.0
GFDL-CM2.1
GISS-ER
INM-CM3.0
IPSL-CM4
MRI-CGCM2.3.2
MIROC3.2(hires)
MIROC3.2(medres)
UKMO-HadGEM1
OSR Anomaly (Wm-2)
86 88 90 92 94 96 98 00 02 04 06 08
8 4 0 -4 -8

Figure S5