ON THE DETERMINATION OF CLIMATE FEEDBACKS FROM ERBE DATA

by Richard S. Lindzen and Yong-Sang Choi

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Abstract

Climate feedbacks are estimated from fluctuations in the outgoing radiation budget from the latest version of Earth Radiation Budget Experiment (ERBE) nonscanner data. It appears, for the entire tropics, the observed outgoing radiation fluxes increase with the increase in sea surface temperatures (SSTs). The observed behavior of radiation fluxes implies negative feedback processes associated with relatively low climate sensitivity. This is the opposite of the behavior of 11 atmospheric models forced by the same SSTs. Therefore, the models display much higher climate sensitivity than is inferred from ERBE, though it is difficult to pin down such high sensitivities with any precision. Results also show, the feedback in ERBE is mostly from shortwave radiation while the feedback in the models is mostly from longwave radiation. Although such a test does not distinguish the mechanisms, this is important since the inconsistency of climate feedbacks constitutes a very fundamental problem in climate prediction.

Introduction

The purpose of the present note is to inquire whether observations of the earth's radiation imbalance can be used to infer feedbacks and climate sensitivity. Such an approach has, as we will see, some difficulties, but it appears that they can be overcome. This is important since most current estimates of climate sensitivity are based on global climate model (GCM) results, and these obviously need observational testing.

To see what one particular difficulty is, consider the following conceptual situation: We instantaneously double CO_2. This will cause the characteristic emission level to rise to a colder level with an associated diminution of outgoing longwave radiation (OLR). The resulting radiative imbalance is what is generally referred to as radiative forcing. However, the resulting warming will eventually eliminate the radiative imbalance as the system approaches equilibrium. The actual amount of warming associated with equilibration as well as the response time will depend on the climate feedbacks in the system. These feedbacks arise from the dependence of radiatively important substances like water vapor (which is a powerful greenhouse gas) and clouds (which are important for both infrared and visible radiation) on the temperature. If the feedbacks are positive, then both the equilibrium warming and the response time will increase; if they are negative, both will decrease. Simple calculations as well as GCM results suggest response times on the order of decades for positive feedbacks and years or less for negative feedbacks [*Lindzen and Giannitsis*, 1998, and references therein]. The main point of this example is to illustrate that the climate system tends to eliminate radiative imbalances with characteristic response times.

Now, in 2002–2004 several papers noted that there was interdecadal change in the top-of-atmosphere (TOA) radiative balance associated with a warming between the 1980's and 1990's [*Chen et al.*, 2002; *Wang et al*., 2002; *Wielicki et al.*, 2002a, b; *Cess and Udelhofen*, 2003; *Hatzidimitriou et al.*, 2004; *Lin et al*., 2004]. *Chou and Lindzen* [2005] inferred from the interdecadal changes in OLR and temperature that there was a strong negative feedback. However, this result was internally inconsistent since the persistence of the imbalance over a decade implied a positive feedback. A subsequent correction to the satellite data eliminated much of the decadal variation in the radiative balance [*Wong et al*., 2006].

However, it also made clear that one could not readily use decadal variability in surface temperature to infer feedbacks from ERBE data. Rather one needs to look at temperature variations that are long compared to the time scales associated with the feedback processes, but short compared to the response time over which the system equilibrates. This is also important so as to unambiguously observe changes in the radiative budget that are responses to fluctuations in SST as opposed to changes in SST resulting from changes in the radiative budget; the latter will occur on the response time of the system. The primary feedbacks involving water vapor and clouds occur on time scales of days [*Lindzen et al.*, 2001; *Rodwell and Palmer*, 2007], while response times for relatively strong negative feedbacks remain on the order of a year [*Lindzen and Giannitsis*, 1998, and references therein]. That said, it is evident that, because the system attempts to restore equilibrium, there will be a tendency to underestimate negative feedbacks relative to positive feedbacks that are associated with longer response times.

Data and Analysis

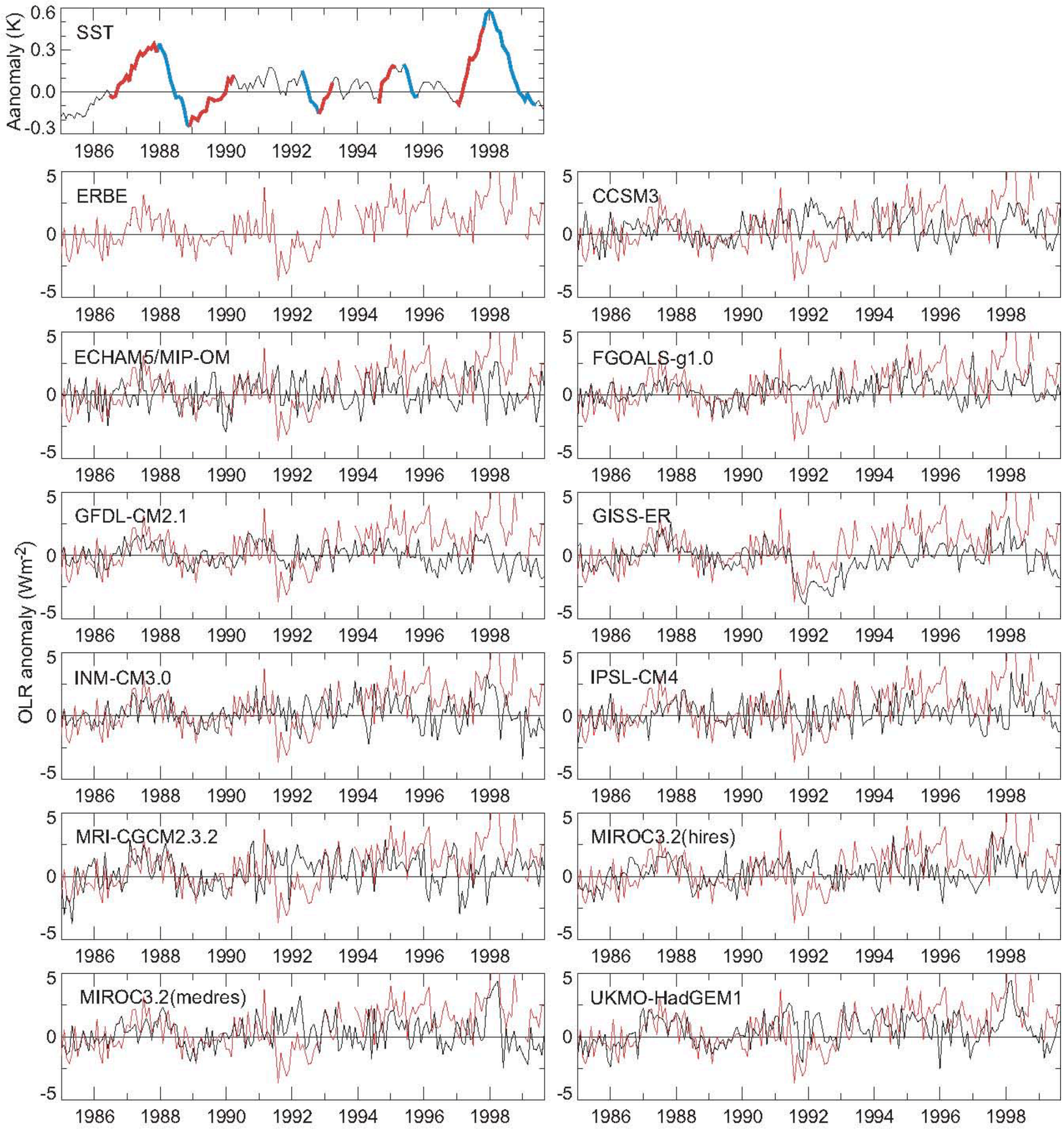
The observed data used in this study are the 16-year (1985–1999) monthly record of the sea surface temperatures (SSTs) from the National Centers for Environmental Prediction, and the earth radiation budget from the Earth Radiation Budget Experiment (ERBE) [*Barkstrom*, 1984] nonscanner edition 3 dataset. Note that the ERBE nonscanner data are the only stable long-term climate dataset based on broadband flux measurements, and they were recently altitude-corrected [*Wong et al*., 2006]. The data can provide reasonably reliable evidence of fluctuations in the anomalies of SST, OLR, and reflected shortwave radiation (SWR) from the tropical means (20°S–20°N); the anomalies are deseasonalized by the monthly means for the period of 1985 through 1989 for the purpose of comparison with climate models [*Wielicki et al.*, 2002a, b]. The effect of land temperature (22% of the whole tropics) on the tropical radiation budget could not be taken into account in this study, due to limited satellite retrievals of surface temperature over the land [*Chou and Lindzen*, 2005].

The anomalies include a semiannual signal due to the temporal aliasing effect that needs to be eliminated [*Trenberth*, 2002]. The relevant sampling error of the tropical monthly ERBE data is about 1.7 W m–2 for SWR and 0.4 W m–2 for OLR [*Wielicki et al*., 2002a, b]. This spurious signal, particularly in the SWR, can be removed in a 36-day average, reducing the SWR error to the order of 0.3 W m–2. However, in this study, the 36-day average was not applied because we wish to relate monthly SSTs to monthly ERBE TOA fluxes. Instead, the moving average with a 7-month smoother was used for the SWR anomalies alone; however, we will see that the smoothing does not much affect the main results. With respect to instrumental stability, the nonscanner records agree relatively well with the scanner records for the period from 1985 to 1989, but no longer agree with them as well for the later period (difference of up to 3 W m–2) [*Wong et al*., 2006]. The fundamental difference between the two types of radiometers comes from the fact that, while the nonsanner views the entire hemisphere of radiation, the scanner views radiance from a single direction and estimates the hemispheric emission or reflection [*Wielicki et al*., 2002a]. It is difficult to quantify possible influences due to this difference, but the present study requires only short term stability and this may be less affected.

The analysis was also made for the model TOA fluxes. The atmospheric model intercomparison projects (AMIP) program for the 4th Assessment Report of the Intergovernmental Panel on Climate Change (IPCC-AR4) provides model results for atmospheric GCMs forced by observed SSTs. AMIP also provides the equilibrium climate sensitivity for the models included [*IPCC*, 2007].

The next obvious question is whether fluctuations with the time scales associated with feedback processes exist in the observed data and models. Figure 1 shows that such fluctuations (ΔFlux) are amply available in OLR and SWR, although data are not currently available in some periods in 1993 and 1999. However, it is possible that many of the very small fluctuations are simply noise. Restricting oneself to fluctuations in SST (ΔT) which exceed 0.2 K still leaves nine cases in the available data (red and blue lines in Fig. 1a):

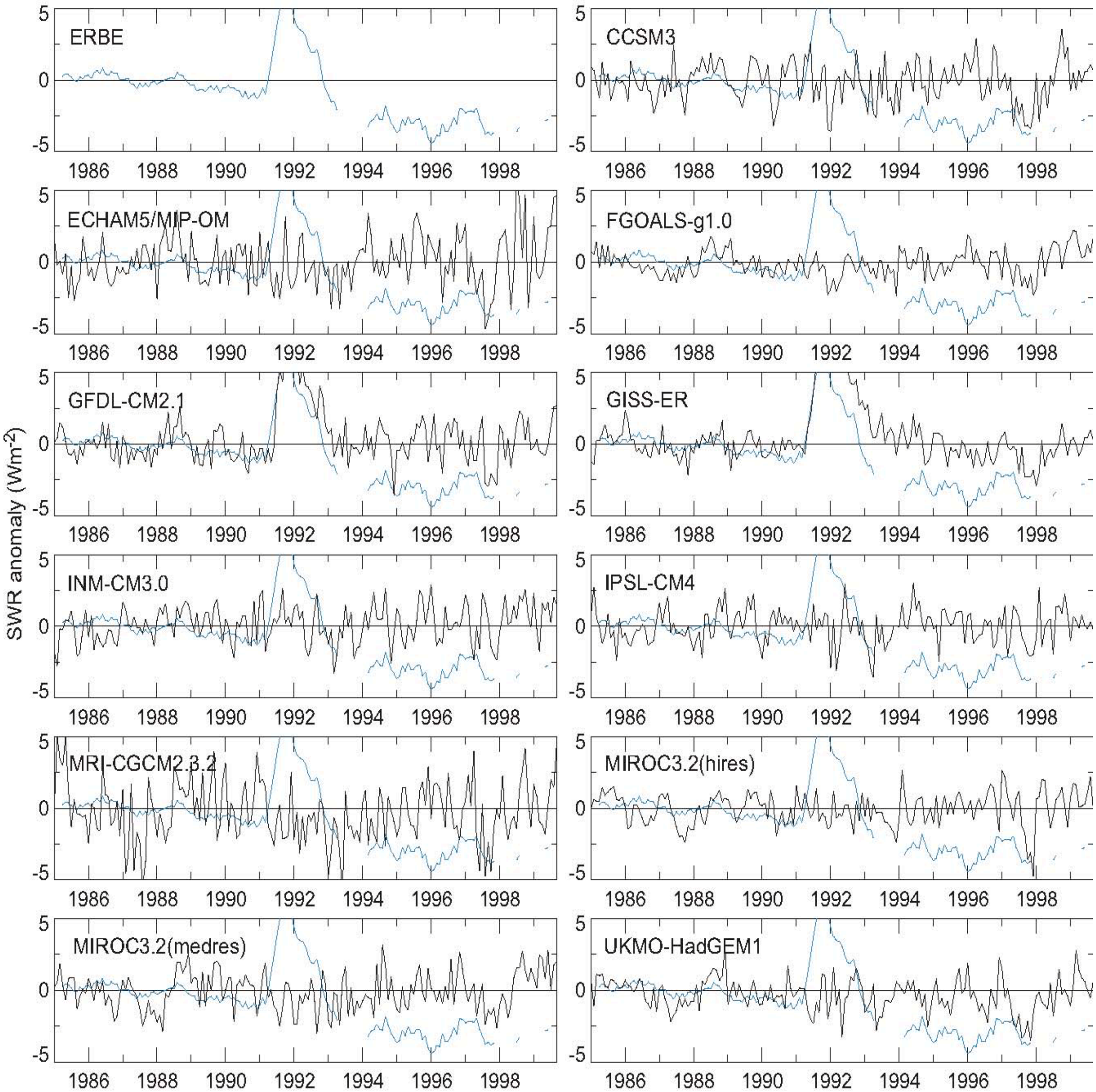
Figure 1a



Monthly SST, and TOA OLR from ERBE (red) and AMIP models (black) for 20°S–20°N. The major SST intervals for which ΔT exceeds 0.2°C are indicated by red and blue colors.

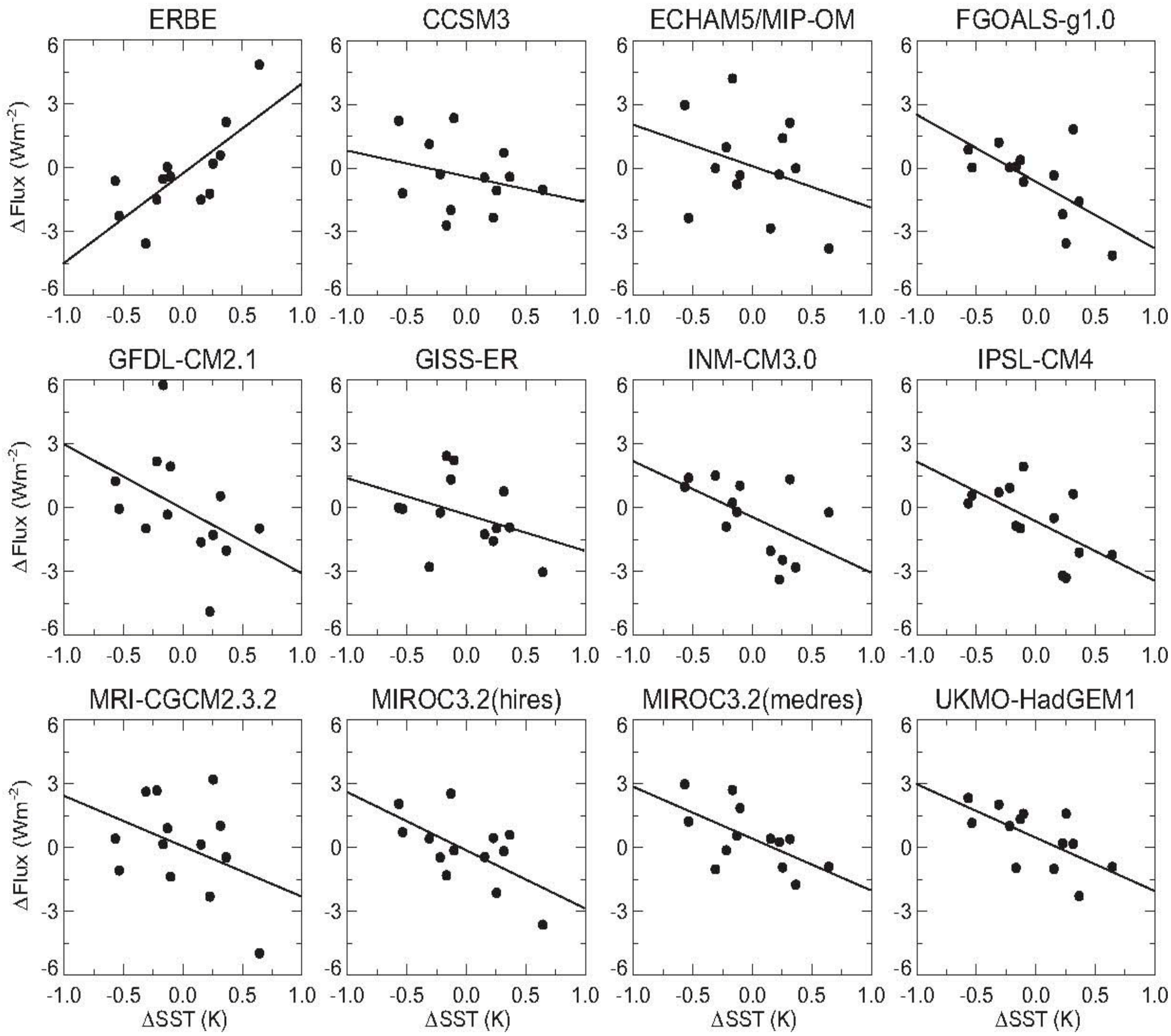
Note that appreciable fluctuations of the anomalies are due to El Niño events (in 1982/83 1986/87, 1991/92, and 1997/98), La Niña events (in 1988/90), and Pinatubo eruption (in 1991) [*Wielicki et al*., 2002a; *Wong et al*., 2006].

Figure 1b



The same as Fig. 1a but for reflected shortwave radiation from ERBE (blue) and AMIP models (black).

Figure 2



Scatterplots of net ΔFlux against ΔT for ERBE and models. Plots for ΔT > 0.1°C are displayed. b. TOA shortwave radiation from ERBE.

Figure 2 compares estimates of net ΔFlux/ΔT for intervals for which ΔT exceeded 0.1 K; the net flux is calculated for OLR+SWR. Results are shown both for 11 AMIP models, and for the ERBE data. ERBE has a positive ΔFlux/ΔT, whereas all models have a negative ΔFlux/ΔT.

Table 1

Case	Number	ERBE			Models		
		Slope	R	SE	Slope	R	SE
0.1 K, Unfiltered	13	5.1	0.81	1.87	–2.41	–0.49	2.4
0.2 K, Unfiltered	9	4.79	0.91	1.00	–2.33	–0.53	1.76
0.3 K, Unfiltered	6	6.16	0.96	0.79	–2.00	–0.58	1.49
0.4 K, Unfiltered	3	6.14	0.97	1.37	–2.62	–0.88	1.54
0.1 K, 3 months	5	5.73	0.92	1.59	–2.14	–0.64	1.46
0.1 K, 5 months	4	6.73	0.87	3.16	–1.60	–0.34	2.02
0.1 K, 7 months	4	9.04	0.93	2.12	–1.50	–0.20	3.39
0.2 K, 3 months	4	6.19	0.97	1.67	–1.99	–0.61	1.33
0.2 K, 5 months	3	7.94	0.9	4.45	–2.73	–0.49	1.82
Monthly interval	176	–0.77	–0.02	32.27	–1.66	–0.06	14.81
0.05 K, Monthly	54	–0.79	–0.03	3.85	–2.10	–0.11	2.76
0.1 K, Monthly	6	5.07	0.34	5.6	–4.64	–0.30	5.03

Regression statistics between net ΔFlux and ΔT, and standard errors of ΔFlux/ΔT for ERBE and models. SW is filtered with 7-month smoother in all cases.

Table 1 compares net ΔFlux/ΔT for intervals for which ΔT exceeded 0.1, 0.2 K,… , for 3, 5, and 7 month time smoothing, for all monthly intervals. We see that all provide essentially the same result, but that scatter is significantly reduced by using threshold 0.2 K without time smoothing. One may take ΔFlux/ΔT with one month intervals, and secure more than hundred cases (Table 1). However, unless we confine ΔT to exceed 0.1 K, the inclusion of what is essentially noise leads to an increase in scatter, and statistically insignificant ΔFlux/ΔT. In addition, based on the known uncertainty of ERBE data, it is expected that uncertainty in ΔFlux/ΔT for ΔT ≥ 0.2 K is up to 1.5 and 2 W m–2 K–1 for the SW and LW fluxes, respectively. That said, the opposite signal between ERBE and the models is hardly attributable to observational errors. Note that we will next show that ΔFlux/ΔT is a measure of the feedback factor for the climate system.

Following *Chou and Lindzen* [2005] and *Lindzen et al.* [2001], we use the following equation to relate ΔFlux/ΔT to equilibrium climate sensitivity. In the nonfeedback climate, climate sensitivity is defined as the response of temperature ΔT0 to an external forcing ΔQ:

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

where G0 is a nonfeedback gain. The mean outgoing longwave (LW) radiation in the whole tropics is approximately 255 W m–2 [*Barkstrom*, 1984], and is equivalent to an effective emitting temperature of 259 K. Thus G0 is calculated by the inverse of the derivative of the

Planck function with respect to the temperature at 259 K; $G_0 \approx 0.25$ W–1 m2 K. For a doubling of CO_2 ($\Delta Q \approx 3.7$ W m–2), ΔT_0 is ~0.925 K (= 0.25 × 3.7).

In the presence of feedback processes, an additional forcing proportional to the response ΔT (i.e., FΔT) is provided to ΔQ in Eq. (1). The response is now

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

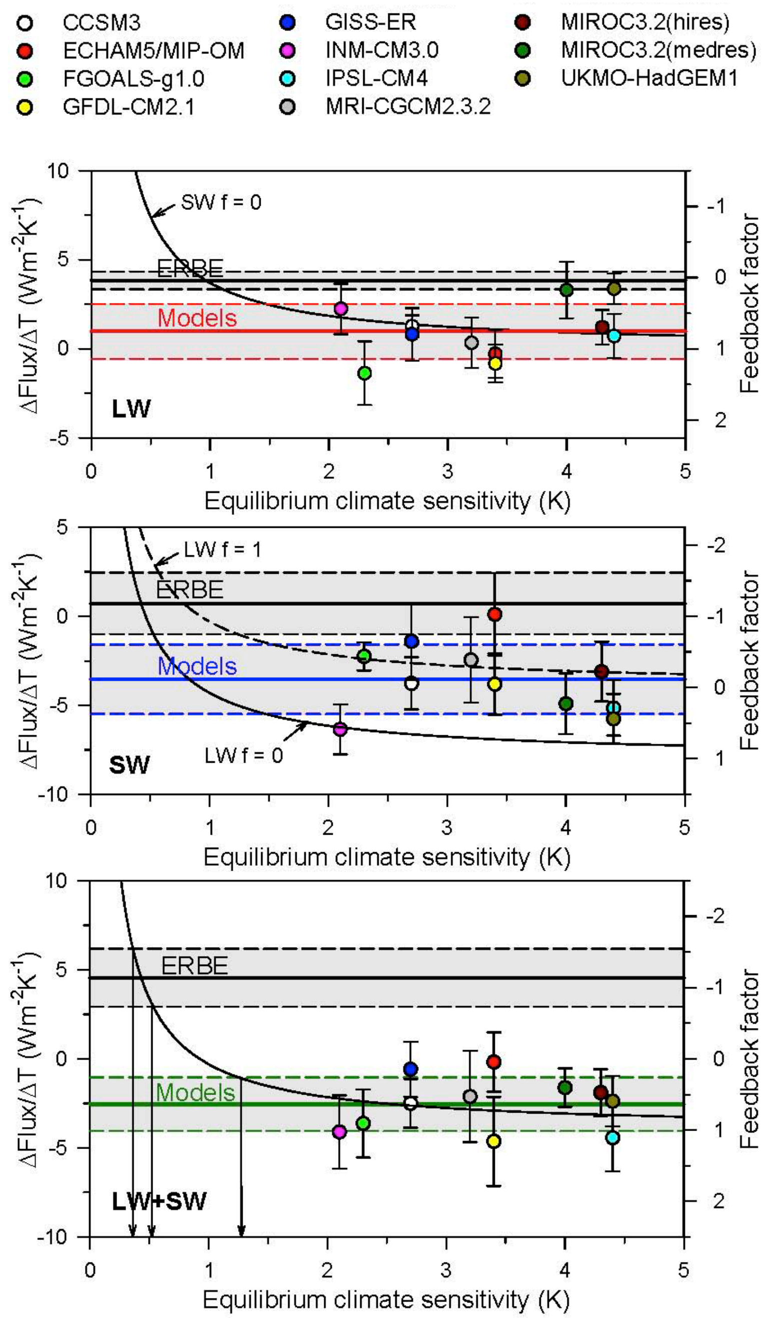
and

$$\Delta T = \Delta T_0 / (1 - f) \tag{3}$$

where $f = G_0F$ is the feedback factor. The net feedback is positive for $0 < f < 1$, and negative for $f < 0$. The feedback parameter F is –ΔFlux/ΔT, assuming the same incoming radiation in the system. The negative sign pertains because increased outgoing flux means energy loss. For example, with ΔT = 0.2 K and ΔFlux = 0.9 W m–2, F is –4.5 W m–2 K–1 (= –0.9/0.2) that is equivalent to f = –1.1, resulting in ΔT of ~0.5 K for a doubling of CO_2 in Eq. (3). Namely, given F = –4.5 W m–2 K–1, climate sensitivity is about a half of that for the nonfeedback condition. On the other hand, negative ΔFlux/ΔT is equivalent to climate sensitivity for a doubling of CO_2 higher than 1 K. All models agree as to positive feedback, and all models disagree very sharply with the observations. However, it is difficult to accurately determine sensitivity from ΔFlux/ΔT from the models. Varying ΔFlux/ΔT values even slightly by 1 W m–2 K–1, which can simply be a measurement error [*Wong et al.*, 2006], and climate sensitivity for a doubling of CO_2 may have any value higher than 1 K. For example, the 2 K to 4.5 K is the likelihood range of climate sensitivity in IPCC-AR4, which corresponds to ΔFlux/ΔT = –2.3 to –3.3 W m–2 K–1. Similar explanation on why climate sensitivity is so unpredictable can be also found in *Roe and Baker* [2007].

When considering LW and SW fluxes separately, F is replaced by $F_{LW} + F_{SW}$. In the observed ΔOLR/ΔT, the nonfeedback change of 4 W m–2 K–1 is included. Also ΔSWR/ΔT needs to be balanced with ΔOLR/ΔT. From the consideration, F_{LW} = –ΔOLR/ΔT + 4 and F_{SW} = –ΔSWR/ΔT – 4. In the case of no SW feedback (F_{SW} = 0), ΔOLR/ΔT less than 4 W m–2 K–1 represents positive feedback; ΔOLR/ΔT more than 4 W m–2 K–1 represents negative feedback; ΔOLR/ΔT less than 0 W m–2 K–1 represents infinite feedback, which is physically unreal.

Figure 3



ERBE-observed and AMIP-simulated ratios of LW (a), SW (b), and total (LW+SW) (c) radiative flux changes to temperature changes (ΔFlux/ΔT) with respect to the equilibrium climate sensitivity. The horizontal solid and dashed lines are the mean and the standard error of ΔFlux/ΔT. The solid curves are theoretical estimate of climate sensitivity for LW feedback under assumption of no SW feedback (a), for SW feedback under assumption of no LW feedback (b), and for total feedback (c).

Concluding Remarks

In Figure 3, we show 3 panels. We see that ERBE and model results differ substantially. In panels a and b, we evaluate Equation (3) using ΔFlux for only OLR and only SWR. The curves are for the condition assuming no SW feedback and assuming no LW feedback in panels a and b, respectively. In panel a, model results fall on the curve given by Equation (3), because the model average of SW feedbacks is almost zero. In panel b, models with smaller LW feedbacks are closer to the curve for no LW feedback; the model results would lie on the curve assuming positive LW feedback. When in panel c we consider the total flux (i.e., LW + SW), model results do lie on the theoretically expected curve. Looking at Figure 3, we note several important features:

1) The models display much higher climate sensitivity than is inferred from ERBE.

2) The (negative) feedback in ERBE is mostly from SW while the (positive) feedback in the models is mostly from OLR.

3) The theoretical relation between ΔF/ΔT and sensitivity is very flat for sensitivities greater than 2°C. Thus, the data does not readily pin down such sensitivities. This was the basis for the assertion by *Roe and Baker* [2007] that determination of climate sensitivity was almost impossible [*Allen and Frame*, 2007]. However, this assertion assumes a large positive feedback. Indeed, Fig. 3c suggests that models should have a range of sensitivities extending from about 1.5°C to infinite sensitivity (rather than 5°C as commonly asserted), given the presence of spurious positive feedback. However, response time increases with increasing sensitivity [*Lindzen and Giannitsis*, 1998], and models were probably not run sufficiently long to realize their full sensitivity. For sensitivities less than 2°C, the data readily distinguish different sensitivities, and ERBE data appear to demonstrate a climate sensitivity of about 0.5°C which is easily distinguished from sensitivities given by models.

Note that while TOA flux data from ERBE are sufficient to determine feedback factors, this data do not specifically identify mechanisms. Thus, the small OLR feedback from ERBE might represent the absence of any OLR feedback; it might also result from the cancellation of a possible positive water vapor feedback due to increased water vapor in the upper troposphere [*Soden et al*., 2005] and a possible negative iris cloud feedback involving reduced upper level cirrus clouds [*Lindzen et al*., 2001]. With respect to SW feedbacks, it is currently claimed that model SW feedbacks are largely associated with the behavior of low level clouds [*Bony et al*., 2006, and references therein]. Whether this is the case in nature cannot be determined from ERBE TOA observations. However, more recent data from CALIOP do offer height resolution, and we are currently studying such data to resolve the issue of what, in fact, is determining SW feedbacks. Finally, it should be noted that our analysis has only considered the tropics. Following *Lindzen et al.* [2001], allowing for sharing this tropical feedback with neutral higher latitudes could reduce the negative feedback factor by about a factor of two. This would lead to an equilibrium sensitivity that is 2/3 rather than 1/2 of the non-feedback value. This, of course, is still a small sensitivity.

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Richard S. Lindzen and Yong-Sang Choi
Program in Atmospheres, Oceans, and Climate
Massachusetts Institute of Technology

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