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On the observational determination of climate sensitivity and its implications

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24 **Abstract**

25 To estimate climate sensitivity from observations, *Lindzen and Choi* [2009] used the
26 deseasonalized fluctuations in sea surface temperatures (SSTs) and the concurrent responses in
27 the top-of-atmosphere outgoing radiation from the ERBE satellite instrument. Distinct periods of
28 warming and cooling in the SST were used to evaluate feedbacks. This work was subject to
29 significant criticism by *Trenberth et al.* [2009], much of which was appropriate. The present
30 paper is an expansion of the earlier paper in which the various criticisms are addressed and
31 corrected. In this paper we supplement the ERBE data for 1985-1999 with data from CERES for
32 2000-2008. Our present analysis accounts for the 36 day precession period for the ERBE satellite
33 in a more appropriate manner than in the earlier paper which simply used what may have been
34 undue smoothing. The present analysis also distinguishes noise in the outgoing radiation as well
35 as radiation changes that are forcing SST changes from those radiation changes that constitute
36 feedbacks to changes in SST. Finally, a more reasonable approach to the zero-feedback flux is
37 taken here. We argue that feedbacks are largely concentrated in the tropics and extend the effect
38 of these feedbacks to the global climate. We again find that the outgoing radiation resulting from
39 SST fluctuations exceeds the zero-feedback fluxes thus implying negative feedback. In contrast
40 to this, the calculated outgoing radiation fluxes from 11 atmospheric GCMs forced by the
41 observed SST are less than the zero-feedback fluxes consistent with the positive feedbacks that
42 characterize these models. The observational analysis implies that the models are exaggerating
43 climate sensitivity.

44

45 **1. Introduction**

46 It is usually claimed that the heart of the global warming issue is so-called greenhouse
47 warming. This simply refers to the fact that the earth balances the heat received from the sun
48 (mostly in the visible spectrum) by radiating in the infrared portion of the spectrum back to
49 space. Gases that are relatively transparent to visible light but strongly absorbent in the infrared
50 (greenhouse gases) will interfere with the cooling of the planet, thus forcing it to become warmer
51 in order to emit sufficient infrared radiation to balance the net incoming sunlight. By the net
52 incoming sunlight, we mean that portion of the sun's radiation that is not reflected back to space
53 by clouds and the earth's surface. The issue then focuses on a particular greenhouse gas, carbon
54 dioxide. Although carbon dioxide is a relatively minor greenhouse gas, it has increased
55 significantly since the beginning of the industrial age from about 280 ppmv to about 390 ppmv,
56 and it is widely accepted that this increase is primarily due to man's emissions. However, it is
57 also widely accepted that the warming from a doubling of carbon dioxide would only be about
58 1°C (based on simple Planck black body calculations; it is also the case that a doubling of any
59 concentration in ppmv produces the same warming because of the logarithmic dependence of
60 carbon dioxide's absorption on the amount of carbon dioxide).

61 This amount of warming is not considered catastrophic, and, more importantly, this is much
62 less than current climate models suggest the warming from a doubling of carbon dioxide will be.
63 The usual claim from the models is that a doubling of carbon dioxide will lead to warming of
64 from 1.5°C to 5°C and even more. What then is really fundamental to 'alarming' predictions? It
65 is the 'feedback' within models from the more important greenhouse substances, water vapor and
66 clouds. Within all current climate models, water vapor increases with increasing temperature so
67 as to further inhibit infrared cooling. Clouds also change so that their net effect resulting from

68 both their infrared absorptivity and their visible reflectivity is to further reduce the net cooling of
69 the earth. These feedbacks are still acknowledged to be highly uncertain, but the fact that these
70 feedbacks are strongly positive in most models is considered to be a significant indication that
71 the result has to be basically correct. Methodologically, this is an unsatisfactory approach to
72 such an important issue. Ideally, one would seek an observational test of the issue. As it turns
73 out, it may be possible to test the issue with existing data from satellites and there has recently
74 been a paper [*Lindzen and Choi, 2009*] that has attempted this though, as we will show in this
75 paper, the details of that paper were, in important ways, incorrect. The present paper attempts to
76 correct the approach and arrives at similar conclusions.

77

78 **2. Feedback formalism**

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

$$85 \quad \Delta T_0 = G_0 \Delta Q \quad (1)$$

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

90

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

92

$$93 \quad \Delta T = G_0(\Delta Q + F\Delta T) \quad (2)$$

94

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

101 Solving (2) for the temperature increment ΔT we find

$$102 \quad \Delta T = \frac{\Delta T_0}{1-f}. \quad (3)$$

103 The dimensionless feedback fraction is $f = F G_0$.

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

$$105 \quad \Delta \text{Flux} - \text{ZFB} = -\frac{f}{G_0} \Delta T \quad (4)$$

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

112
$$\Delta\text{Flux} - \text{ZFB} = -\frac{f}{G_0} \Delta\text{SST} \quad (5)$$

113 where SST is the observed sea surface temperature.

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

115
$$-G_0 \left(\frac{\Delta\text{Flux} - \text{ZFB}}{\Delta\text{SST}} \right)_{tropicals} \approx 2f \quad (6)$$

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

119
$$f_{LW} = -\frac{G_0}{2} \left(\frac{\Delta\text{OLR} - \text{ZFB}}{\Delta\text{SST}} \right)_{tropicals} \quad (7a)$$

120
$$f_{SW} = -\frac{G_0}{2} \left(\frac{\Delta\text{SWR}}{\Delta\text{SST}} \right)_{tropicals} \quad (7b)$$

121 Here we can identify ΔFlux as the change in outgoing longwave radiation (OLR) and
 122 shortwave radiation (SWR) measured by satellites associated with the measured ΔSST , the
 123 change of the sea-surface temperature. Since we know the value of G_0 , the experimentally
 124 determined slope allows us to evaluate the magnitude and sign of the feedback factor f provided
 125 that we also know the value of the zero-feedback flux. Note that the natural forcing, ΔSST , that
 126 can be observed, is different from the equilibrium response temperature ΔT in Eq. (3). The latter
 127 cannot be observed since, for the short intervals considered, the system cannot be in equilibrium,
 128 and over the longer periods needed for equilibration of the whole climate system, ΔFlux at the
 129 top of the atmosphere is restored to zero. Indeed, as explained in *Lindzen and Choi* [2009], it is,
 130 in fact, essential, that the time intervals considered, be short compared to the time it takes for the
 131 system to equilibrate, while long compared to the time scale on which the feedback processes

132 operate (which are essentially the time scales associated with cumulonimbus convection). The
133 latter is on the order of days, while the former depends on the climate sensitivity, and ranges
134 from years for sensitivities of 0.5°C for a doubling of CO₂ to many decades for higher
135 sensitivities [Lindzen and Giannitsis, 1998]. Finally, for observed variations, there is the fact that
136 changes in radiation (as for example associated with volcanoes) can cause changes in SST as
137 well as respond to changes in SST, and there is a need to distinguish these two possibilities. This
138 is not an issue with model results from the AMIP program where observed variations in SST are
139 specified. Of course, there is always the problem of noise arising from the fact that clouds
140 depend on factors other than surface temperature. Note that this study deals with observed
141 outgoing fluxes, but does not specifically identify the origin of the changes. This is discussed in
142 Appendix 1.

143

144 **3. The data and their problems**

145 Now, it turns out that SST is measured [Kanamitsu *et al.*, 2002], and is always fluctuating as
146 we see from Fig. 3. High frequency fluctuations, however, make it difficult to objectively
147 identify the beginning and end of warming and cooling intervals [Trenberth *et al.*, 2010]. This
148 ambiguity is eliminated with a 3 point centered smoother. (A two point lagged smoother works
149 as well.) In addition, the net outgoing radiative flux from the earth has been monitored since
150 1985 by the ERBE instrument [Barkstrom, 1984] (nonscanner edition 3) aboard ERBS satellite,
151 and since 2000 by the CERES instrument (ES4 FM1 edition 2) aboard the Terra satellite
152 [Wielicki *et al.*, 1998]. The results for both LW (infrared) radiation and SW (visible) radiation
153 are shown in Fig. 4. The sum is the net flux.

154 With ERBE data, there is, however, the problem of satellite precession with a period of 36

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

162 The discontinuity between the two datasets needs some discussion. There is the long-term
163 discrepancy of the average which is generally acknowledged to be due to the absolute calibration
164 problem (up to 3 W m^{-2}) [*Wong et al.*, 2006]. With CERES, the spectral darkening was resolved
165 by multiplying SW flux by the scale factor (up to 1.011) from *Matthews et al.* [2005]. However,
166 this long-term stability should not matter for our analysis which focuses on short-term
167 fluctuations. One major concern to be considered in this study is the higher seasonal fluctuation
168 of CERES SW radiation than ERBE. The bias is up to 6.0 W m^{-2} as estimated by *Young et al.*
169 [1998]. This is attributed to different sampling patterns, that is, ERBS observes all local times
170 over a period of 72 days, while Terra observes the region only twice per day (around 10:30 AM
171 and 10:30 PM). To avoid this problem, the anomalies for radiative flux are separately referenced
172 to the monthly means for the period of 1985 through 1989 for ERBE, and for the period of 2000
173 through 2004 for CERES. However, the issue of the reference period is also insignificant in this
174 study that uses enough segments to cancel out this seasonality.

175 Both ERBE and CERES data are best for the tropics. The ERBE field-of-view is between
176 60°S and 60°N . For latitudes 40° to 60° , 72 days are required instead of 36 days to reduce the
177 precession effect [*Wong et al.*, 2006]. Both datasets have no/negligible shortwave radiation in

178 winter hemispheric high latitudes, which would compromise our analysis. Moreover, our
179 analysis involves relating changes in outgoing flux to changes in SST. This is appropriate to
180 regions that are mostly ocean covered like the tropics or the southern hemisphere, but distinctly
181 inappropriate to the northern extratropics. However, as we will argue in Appendix 2, the water
182 vapor feedback is almost certainly restricted primarily to the tropics, and there are reasons to
183 suppose that this is also the case for cloud feedbacks. The methodology developed in *Lindzen,*
184 *Chou, and Hou* [2001] permits the easy extension of the tropical processes to global values.

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

193 Turning to the models, AMIP (Atmospheric Model Intercomparison Program) is responsible
194 for intercomparing models used by the IPCC (the Intergovernmental Panel on Climate Change),
195 has obtained the calculated changes in both short and long wave radiation from models forced by
196 the observed sea surface temperatures shown in Fig. 3. These results are shown in Figs. 5 and 6
197 where the observed results are also plotted for comparison. We can already see that there are
198 significant differences. Note that it is important to use the AMIP results rather than those from
199 the coupled atmosphere-ocean models (CMIP). Only for the former can we see the results for the
200 same SST as applies to the ERBE/CERES observations. Moreover, in the AMIP results, we are

201 confident that the temperatures are forcing the changes in outgoing radiation; in the coupled
202 models it is more difficult to be sure that we are calculating outgoing fluxes that are responding
203 to SST forcing rather than temperature perturbations resulting from independent fluctuations in
204 radiation.

205

206 **4. Calculations**

207 With all the above readily available, it is now possible to directly test the ability of models to
208 adequately simulate the sensitivity of climate. The procedure is simply to identify intervals of
209 change for ΔSST in Fig. 3 (for reasons we will discuss at the end, it is advisable, but not
210 essential, to restrict oneself to changes greater than 0.1°C), and for each such interval, to find the
211 change in flux. Let us define i_1, i_2, \dots, i_m as selected time steps that correspond to the starting and
212 the ending points of intervals. $\Delta\text{Flux}/\Delta\text{SST}$ can be basically obtained by $\text{Flux}(i_1) - \text{Flux}(i_2)$
213 divided by $\text{SST}(i_1) - \text{SST}(i_2)$. As there are many intervals, $\Delta\text{Flux}/\Delta\text{SST}$ is a regression slope for
214 the plots $(\Delta\text{Flux}, \Delta\text{SST})$ for a linear regression model. Here we use a zero y-intercept model ($y =$
215 ax) because the presence of the y-intercept is related to noise other than feedbacks. Thus, a zero
216 y-intercept model may be more appropriate for the purpose of our feedback analysis; however,
217 the choice of regression model turns out to be minor. As already noted, the data need to be
218 smoothed to minimize noise, and it is also crucial to distinguish ΔSST that are forcing changes in
219 ΔFlux , and not responses to ΔFlux . Otherwise, $\Delta\text{Flux}/\Delta\text{SST}$ can vary [Trenberth *et al.*, 2010]
220 and/or may not represent feedbacks that we wish to determine. As an attempt to avoid such
221 problems, though imperfectly, we need to consider smoothing (i.e., use of $\text{Flux}'(i)$ and $\text{SST}'(i)$,
222 where the prime designates the smoothed value) and lag-lead methods (e.g., use of $\text{Flux}'(i+\text{lag})$
223 and $\text{SST}'(i)$) for ERBE 36-day and CERES monthly data. For a stable estimate of $\Delta\text{Flux}/\Delta\text{SST}$,

224 the time step i should be also selected based on the maximum and minimum of the smoothed
225 SST (i.e., SST'). As shown in Fig. 3, this study selected $SST'(i_1) - SST'(i_2)$ that exceeds 0.1 K.
226 The impact of thresholds for ΔSST on the statistics of the results is minor [*Lindzen and Choi,*
227 2009].

228 Figure 7 shows the impact of smoothing and leads and lags on the determination of the slope
229 as well as on the correlation, R , of the linear regression. In general, the use of leads for flux will
230 emphasize forcing by the fluxes, and the use of lags will emphasize responses by the fluxes to
231 changes in SST. For LW radiation, the situation is fairly simple. Smoothing increases R
232 somewhat, and for 3 point symmetric smoothing, R maximizes for slight lag or zero – consistent
233 with the fact that feedbacks are expected to result from fast processes. Maximum slope is found
234 for a lag of 1 ‘month’, though it should be remembered that the relevant feedback processes may
235 operate on a time scale shorter than we resolve. The situation for SW radiation is, not
236 surprisingly, more complex since phenomena like the Pinatubo eruption lead to increased light
237 reflection and associated cooling of the surface (There is also the obvious fact that many things
238 can cause fluctuations in clouds, which leads to noise). We see two extremes associated with
239 changing lead/lag. There is a maximum negative slope associated with a brief lead, and a
240 relatively large positive slope associated with a 3–4 month lag. It seems reasonable to suppose
241 that the effect of forcing extends into the results at small lags because it takes time for the ocean
242 surface to respond, and is only overcome for larger lags where the change in flux associated with
243 feedback dominates. Indeed, excluding the case of Pinatubo volcano for larger lags does little to
244 change the results (less than $0.3 \text{ W m}^{-2}/\text{K}$). Under such circumstances, we expect the maximum
245 slope for SW radiation in Fig. 7 to be an underestimate of the actual feedback. We also consider
246 the standard error of the slope to show data uncertainty. The results for the lags associated with

247 maximum R are shown in Table 1. We take LW and SW radiation for lag = 1 and lag = 3,
248 respectively, and measure the slope $\Delta\text{Flux}/\Delta\text{SST}$ for the sum of these fluxes. The standard error
249 of the slope in total radiation for the appropriate lags comes from the regression for scatter plots
250 of $(\Delta\text{SST}, \Delta(\text{OLR}+\text{SWR}))$. As we see in Table 1, model sensitivities indicated by the IPCC AR4
251 (Fig. 8) are likely greater than the possibilities estimated in the observations.

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

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

268

269

270 **5. Discussion and conclusions**

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

277 The conclusion appears to be that all current models exaggerate climate sensitivity (some
278 greatly). It also suggests, incidentally, that in current coupled atmosphere-ocean models, that the
279 atmosphere and ocean are too weakly coupled since thermal coupling is inversely proportional to
280 sensitivity [*Lindzen and Giannitsis*, 1998]. It has been noted by *Newman et al.* [2009] that
281 coupling is crucial to the simulation of phenomena like El Niño. Thus, corrections of the
282 sensitivity of current climate models might well improve the behavior of coupled models. It
283 should also be noted that there have been independent tests that also suggest sensitivities less
284 than predicted by current models (*Lindzen and Giannitsis* [1998], based on response to
285 sequences of volcanic eruptions — they also noted that the response to individual volcanoes in
286 the two years following eruption were largely independent of sensitivity, and, hence, of little use
287 for distinguishing different sensitivities; *Lindzen* [2007], and *Douglass et al.* [2007], both based
288 on the vertical structure of observed versus modeled temperature increase; and *Schwartz* [2007,
289 2008], based on ocean heating). Most claims of greater sensitivity are based on the models that
290 we have just shown can be highly misleading on this matter. There have also been attempts to
291 infer sensitivity from paleoclimate data [*Hansen*, 1993], but these are not really tests since the
292 forcing is essentially unknown and may be adjusted to produce any sensitivity one wishes. It is

293 important to realize that climate sensitivity is essentially a single number. Economists who treat
294 climate sensitivity as a probability distribution function [*Weitzman, 2009; Stern, 2008; Sokolov et*
295 *al., 2009*] are mistakenly confusing model uncertainty concerning this particular number with the
296 existence of a real range of possibility. The high sensitivity results that these studies rely on for
297 claiming that catastrophes are possible are almost totally incompatible with the present results –
298 despite the uncertainty of the present results.

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

307

308 **Appendices**

309 **Appendix 1. Origin of Feedbacks**

310 While the present analysis is a direct test of feedback factors, it does not provide much insight
311 into detailed mechanism. Nevertheless, separating the contributions to f from long wave and
312 short wave fluxes provides some interesting insights. The results are shown in Tables 1 and 2. It
313 should be noted that the consideration of the zero-feedback response, and the tropical feedback
314 factor to be half of the global feedback factor is actually necessary for our measurements from
315 the Tropics; however, these were not considered in *Lindzen and Choi* [2009]. Accordingly, with
316 respect to separating longwave and shortwave feedbacks, the interpretation by *Lindzen and Choi*
317 [2009] needs to be corrected. These tables show recalculated feedback factors in the presence of
318 the zero-feedback Planck response. The negative feedback from observations is from both
319 longwave and shortwave radiation, while the positive feedback from models is usually but not
320 always from longwave feedback.

321 As concerns the infrared, there is, indeed, evidence for a positive water vapor feedback [*Soden*
322 *et al.*, 2005], but, if this is true, this feedback is presumably cancelled by a negative infrared
323 feedback such as that proposed by *Lindzen et al.* [2001] in their paper on the iris effect. In the
324 models, on the contrary, the long wave feedback appear to be positive (except for two models),
325 but it is not as great as expected for the water vapor feedback [*Colman*, 2003; *Soden et al.*, 2005].
326 This is possible because the so-called lapse rate feedback as well as negative longwave cloud
327 feedback serves to cancel the TOA OLR feedback in current models. Table 2 implies that TOA
328 longwave and shortwave contributions are coupled in models (the correlation coefficient
329 between f_{LW} and f_{SW} from models is about -0.5). This coupling most likely is associated with
330 the primary clouds in models — optically thick high-top clouds [*Webb et al.*, 2006]. In most

331 climate models, the feedbacks from these clouds are simulated to be negative in longwave and
332 strongly positive in shortwave, and dominate the entire cloud feedback [Webb *et al.*, 2006].
333 Therefore, the cloud feedbacks may also serve to contribute to the negative OLR feedback and
334 the positive SWR feedback. New spaceborne data from the CALIPSO lidar (CALIOP; Winker *et*
335 *al.* [2007]) and the CloudSat radar (CPR; Im *et al.* [2005]) should provide a breakdown of cloud
336 behavior with altitude which may give some insight into what exactly is contributing to the
337 radiation.

338

339 **Appendix 2. Concentration of climate feedbacks in the tropics**

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

346 We see that for extratropical conditions, outgoing radiation closely approximates the Planck
347 black body radiation (leading to small feedback). However, for tropical conditions, increases in
348 outgoing radiation are suppressed, implying substantial positive feedback. There are also good
349 reasons to suppose that cloud feedbacks are largely confined to the tropics. In the extratropics,
350 clouds are mostly stratiform clouds that are associated with ascending air while descending
351 regions are cloud-free. Ascent and descent are largely determined by the large scale wave
352 motions that dominate the meteorology of the extratropics, and for these waves, we expect
353 approximately 50% cloud cover regardless of temperature. On the other hand, in the tropics,

354 upper level clouds, at least, are mostly determined by detrainment from cumulonimbus towers,
355 and cloud coverage is observed to depend significantly on temperature [*Rondanelli and Lindzen,*
356 2008]. As noted by *Lindzen et al.* [2001], with feedbacks restricted to the tropics, their
357 contribution to global sensitivity results from sharing the feedback fluxes with the extratropics.
358 This leads to the factor of 2 in Eq. (6).

359

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366

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434

435 **Table legends**

436 **Table 1.** Mean±standard error of the variables for the likely lag for the observations. The units
 437 for the slope are $W\ m^{-2}\ K^{-1}$. Also shown are the estimated mean and range of equilibrium
 438 climate sensitivity (in degrees C) for a doubling of CO_2 for 90%, 95%, and 99% confidence
 439 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

440

441 **Table 2.** Regression statistics between $\Delta Flux$ and ΔSST and the estimated feedback factors (f)
 442 for LW, SW, and total radiation in AMIP models; the slope is $\Delta Flux/\Delta SST$, N is the number of
 443 the points or intervals, R is the correlation coefficient, and SE is the standard error of
 444 $\Delta Flux/\Delta SST$.

	N	LW				SW				LW+SW			
		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

445

446 **Table 3.** Comparison of model equilibrium climate sensitivities for a doubling of CO₂ defined
 447 from IPCC AR4 and estimated from feedback factors estimated in this study. The ranges of
 448 climate sensitivities for models are also estimated from the standard errors in Table 2 for 90%,
 449 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

450

451

452 **Figure legends**

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

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

455 **Figure 3.** Tropical mean (20°S to 20°N latitude) 36-day averaged and monthly sea surface
456 temperature anomalies with the centered 3-point smoothing; the anomalies are referenced to the
457 monthly means for the period of 1985 through 1989. The SST anomaly was scaled by a factor of
458 0.78 (the area fraction of the ocean to the tropics) to relate with the flux. Red and blue colors
459 indicate the major temperature fluctuations exceeding 0.1°C used in this study. The cooling after
460 1998 El Niño is not included because of no flux data is available for this period (viz Fig. 4).

461 **Figure 4.** The same as Fig. 3 but for outgoing longwave (red) and reflected shortwave (blue)
462 radiation from ERBE and CERES satellite instruments. 36-day averages are used to compensate
463 for the ERBE precession. The anomalies are referenced to the monthly means for the period of
464 1985 through 1989 for ERBE, and 2000 through 2004 for CERES. Missing periods are the same
465 as reported in Wong et al. (2006).

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

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

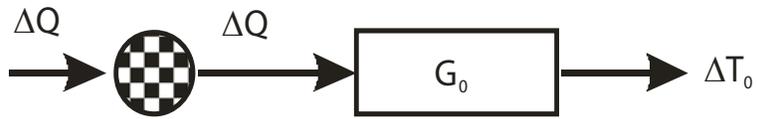
470 **Figure 7.** The impact of smoothing and leads and lags on the determination of the slope (top) as
471 well as on the correlation coefficient, R , of the linear regression (bottom).

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

473 **Figure 9.** Sensitivity vs. feedback factor.

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

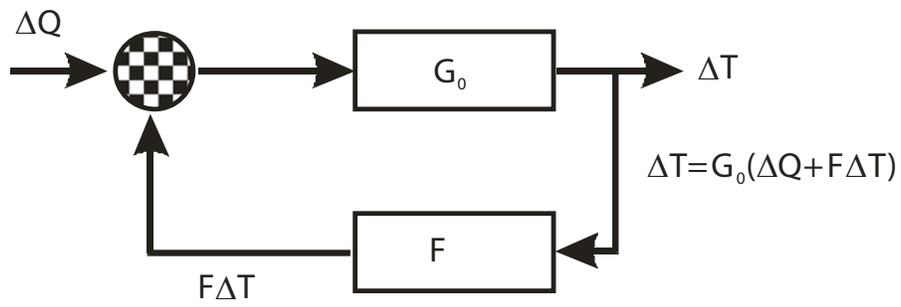
475 fixed. The surface air pressure is 1bar, and Earth gravity is assumed. The temperature profile is
476 the water/air moist adiabat. Calculations were carried out with the Community Climate Model
477 radiation code (from Pierrehumbert [2009]).
478



479

480 Figure 1

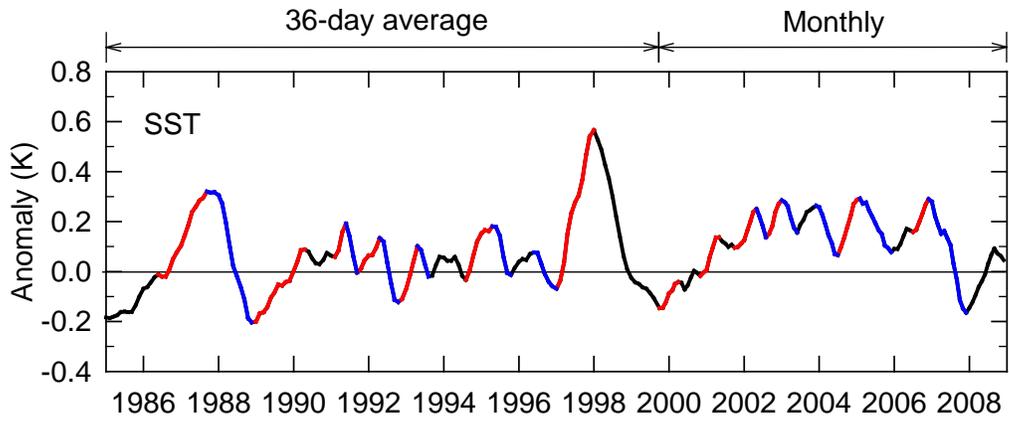
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483 Figure 2

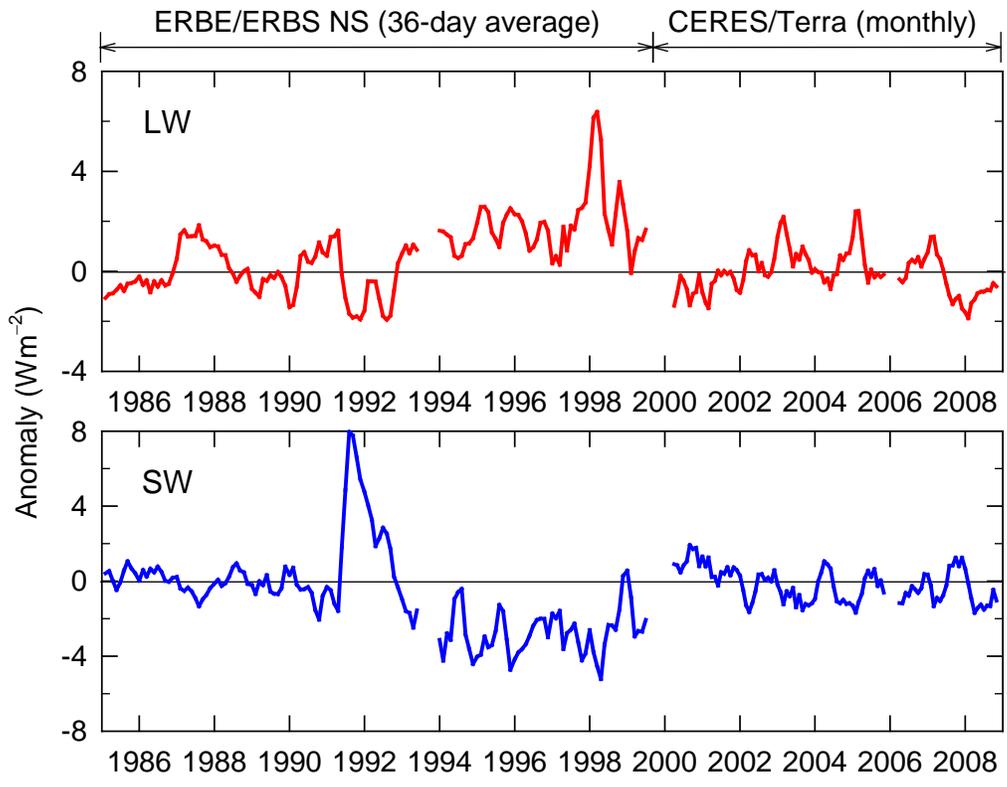
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486 Figure 3

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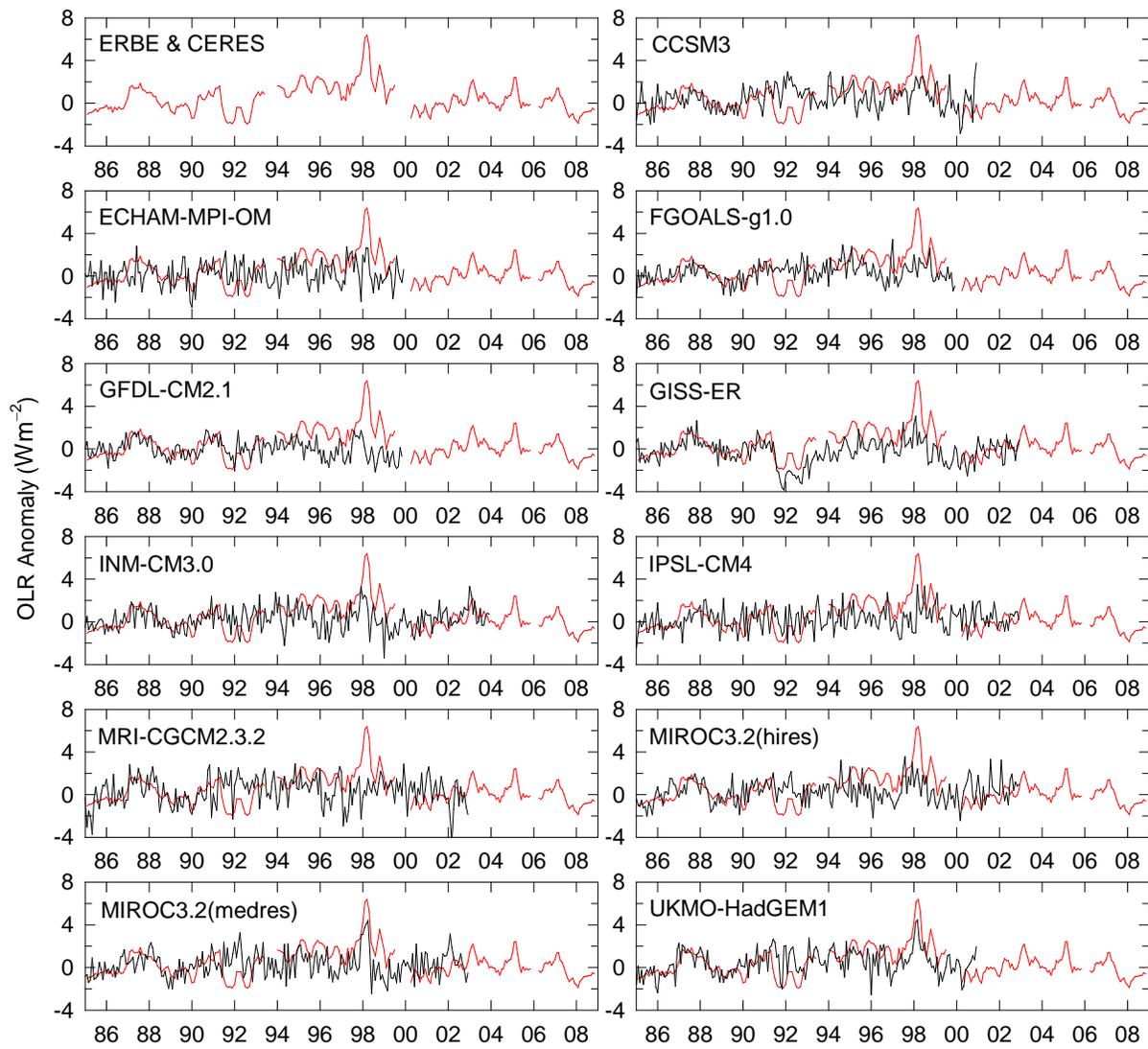


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489 Figure 4

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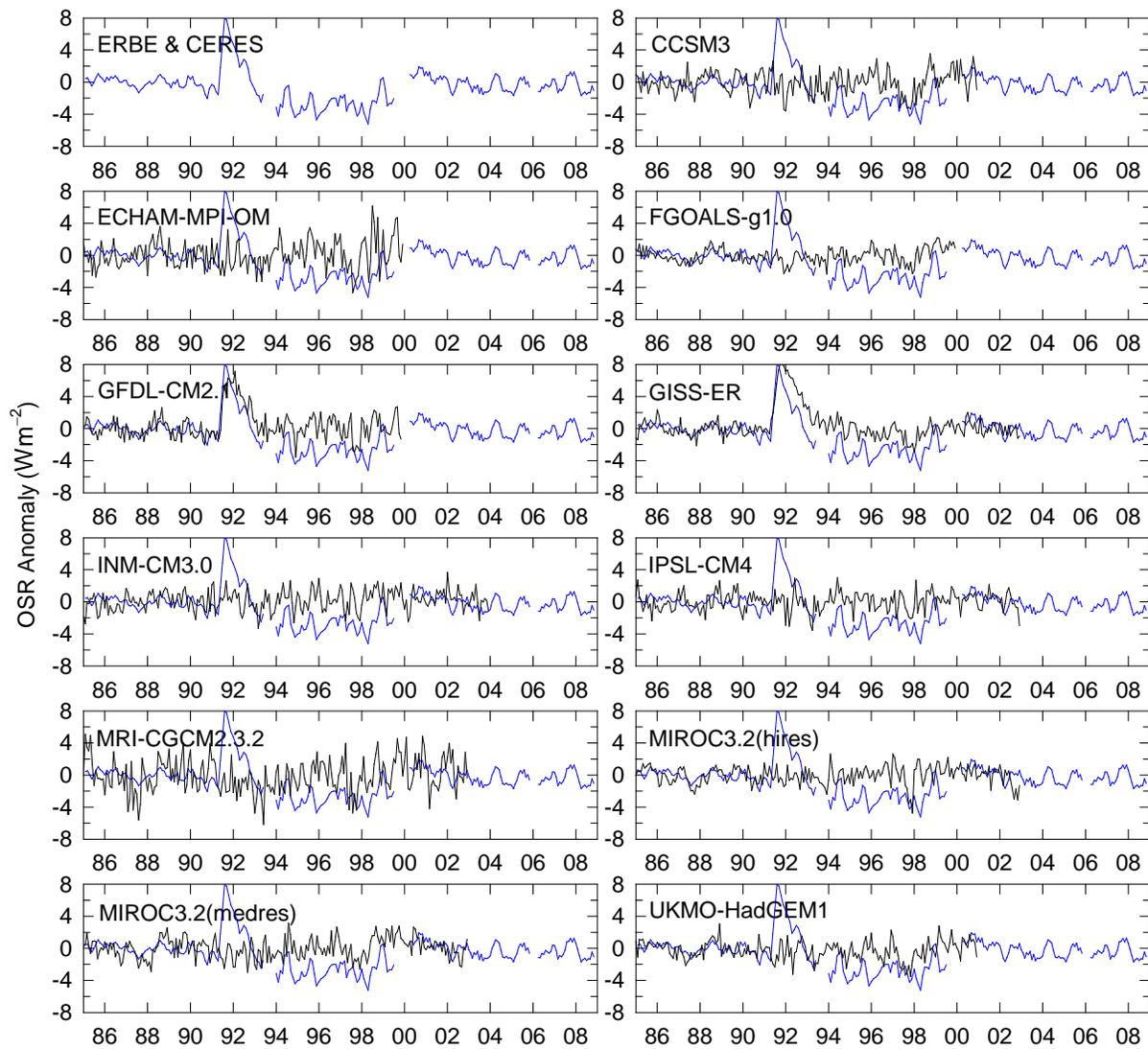


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493 Figure 5

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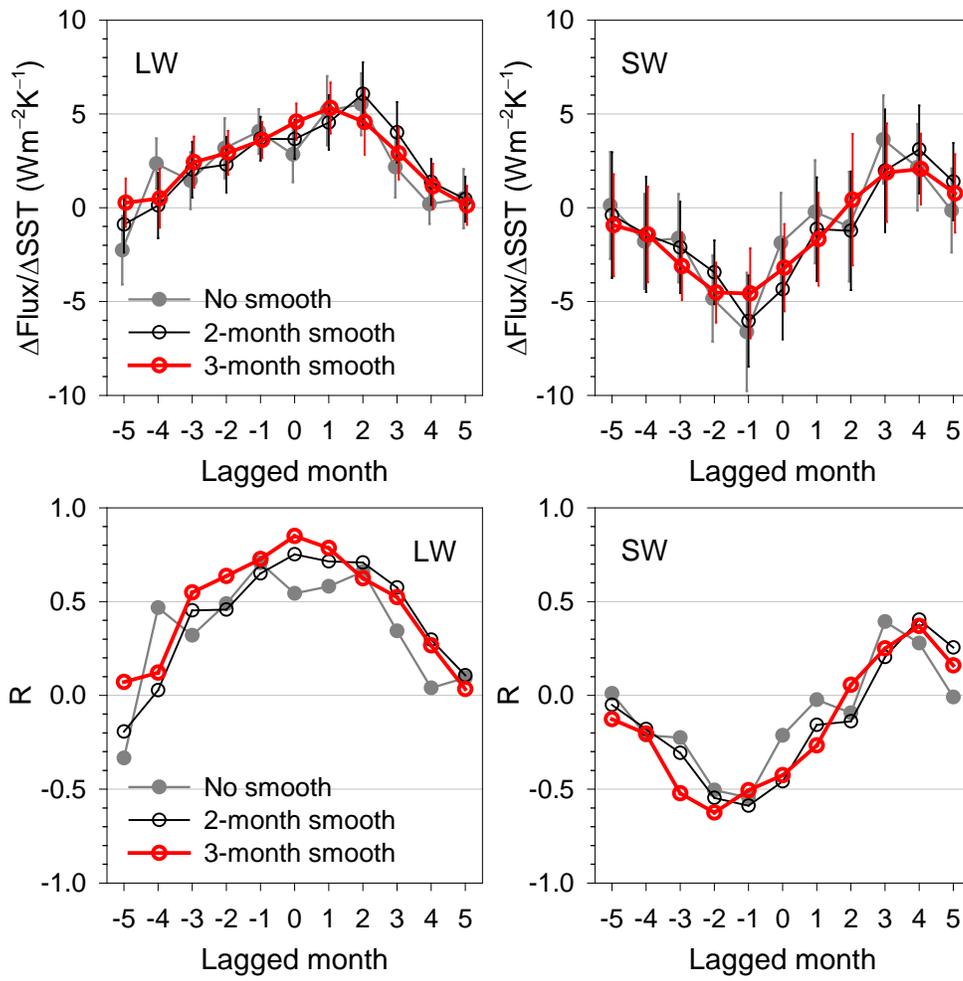
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497 Figure 6

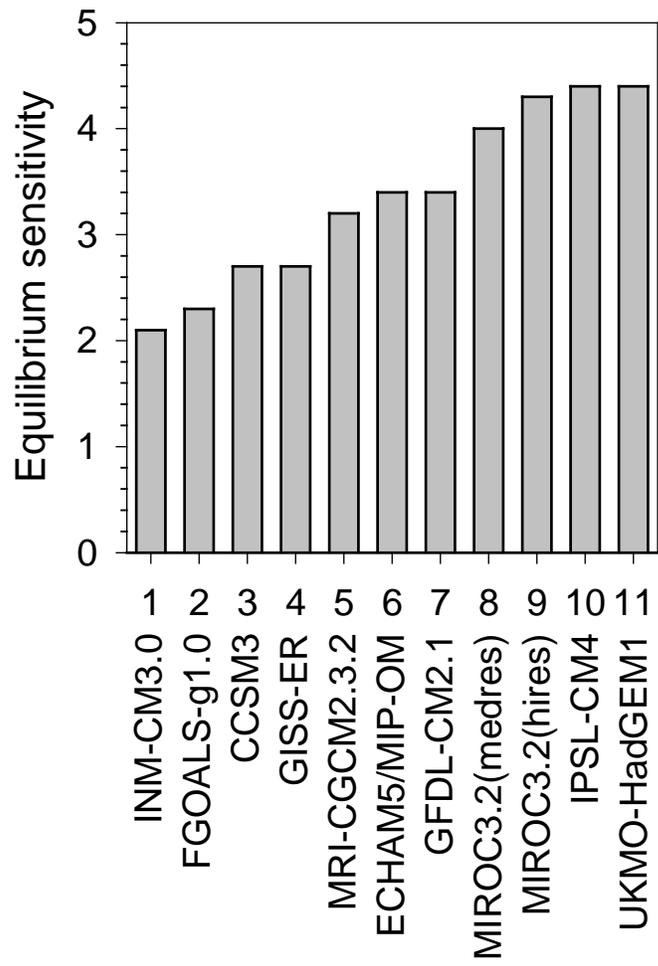
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500 Figure 7

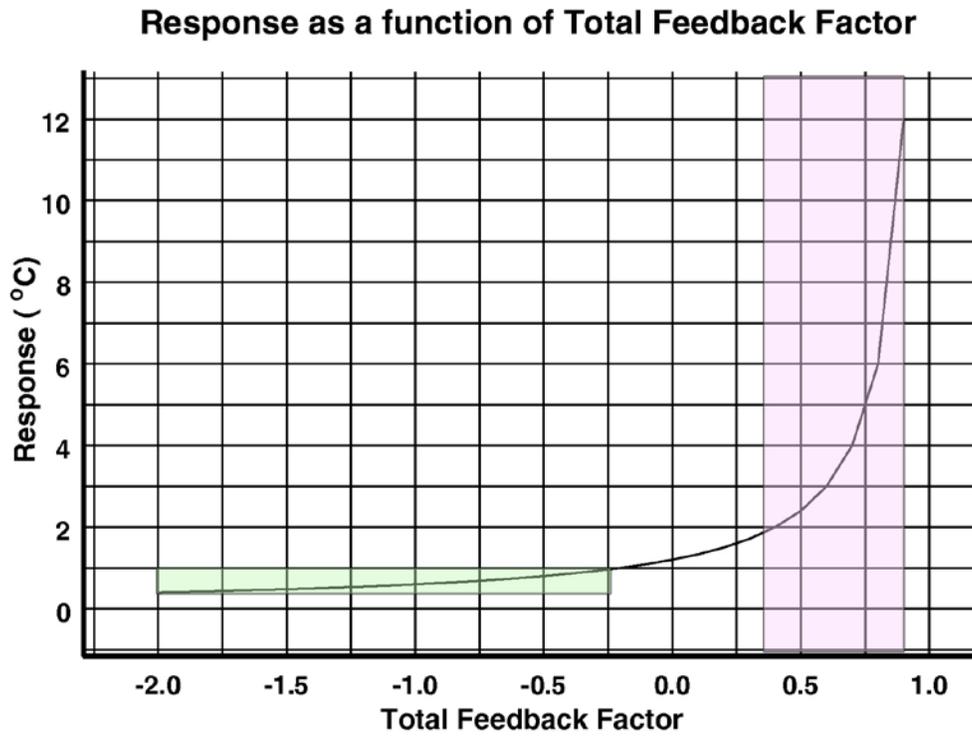
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503 Figure 8

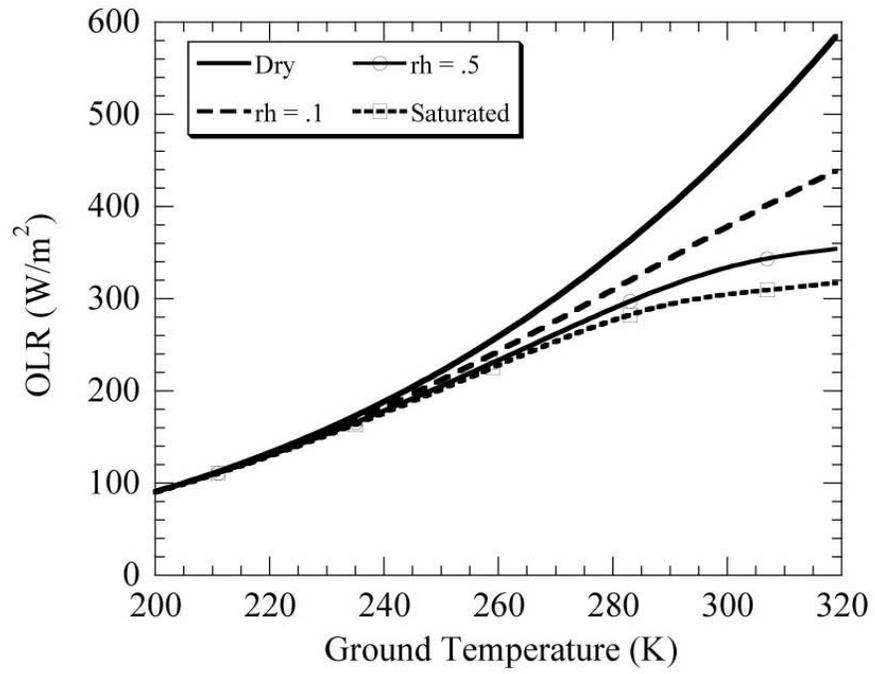
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506 Figure 9

507



508

509 Figure 10