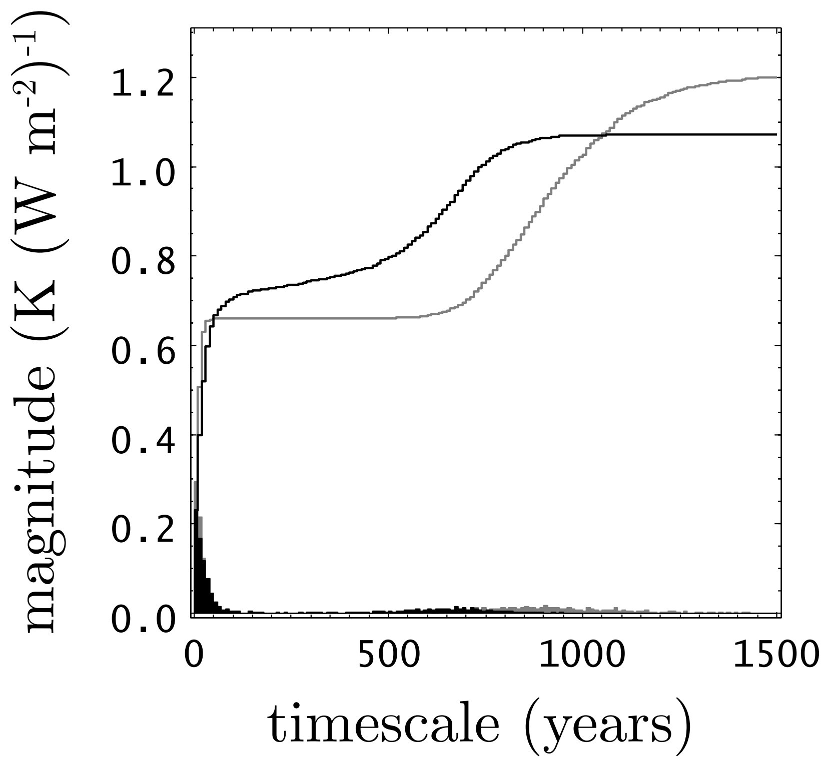
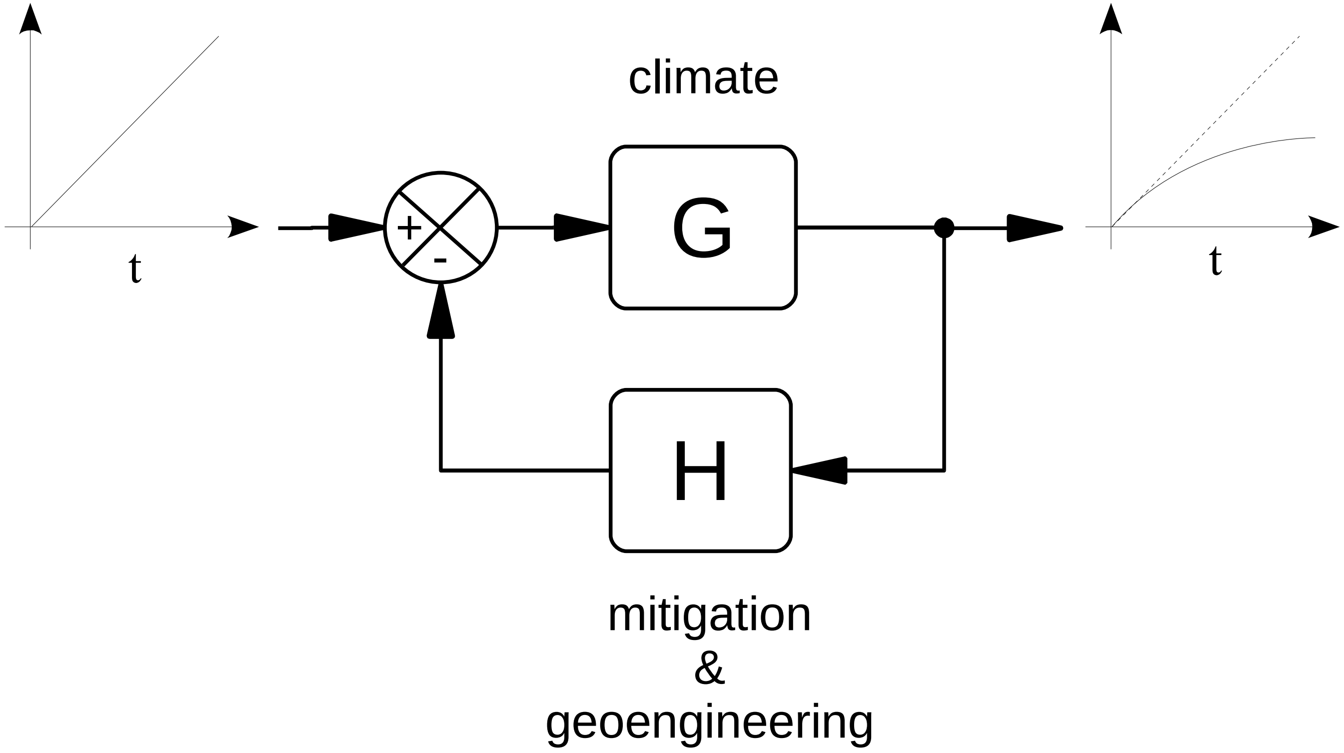
the timescales of climate models



climate and society



Reports

Climate Response Times: Dependence on Climate Sensitivity and Ocean Mixing

Abstract. *The factors that determine climate response times were investigated with simple models and scaling statements. The response times are particularly sensitive to (i) the amount that the climate response is amplified by feedbacks and (ii) the representation of ocean mixing. If equilibrium climate sensitivity is 3°C or greater for a doubling of the carbon dioxide concentration, then most of the expected warming attributable to trace gases added to the atmosphere by man probably has not yet occurred. This yet to be realized warming calls into question a policy of "wait and see" regarding the issue of how to deal with increasing atmospheric carbon dioxide and other trace gases.*

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The abundances of CO_2 and other trace gases in the atmosphere are changing, and it is believed that this change will affect global climate (*1–4*). The global mean surface air warming at equilibrium ($t \rightarrow \infty$) expected to result from doubling of CO_2 (from, say, 300 to 600 ppm) has been estimated (*1, 2*) as

$$\Delta T_{eq}(2 * CO_2) = 3.0° \pm 1.5°C \qquad (1)$$

with the range being a subjective estimate of the uncertainty based on climate modeling studies and empirical evidence for climate sensitivity.

If there were no climate feedbacks (that is, if the atmospheric temperature gradient and all other factors were fixed), the planet would have to warm by

$$\Delta T_0(2 * CO_2) = 1.2° \text{ to } 1.3°C \qquad (2)$$

to restore the radiation balance with space after CO_2 is doubled (*5, 6*). Thus the climate sensitivity (Eq. 1) implies a net climate feedback factor

$$f = 2.4 \pm 1.2 \qquad (3)$$

where f is the ratio of the equilibrium surface air warming to the warming that would have occurred in the absence of any feedbacks.

An important point in evaluating the transient response to a change in CO_2 is that the response time depends on f. This is illustrated by some simple but progressively more realistic systems. First, consider an atmosphereless blackbody planet ($f \equiv 1$). If the equilibrium temperature of this planet suddenly changes a small amount (say, because of a change in the solar constant), it will approach its new temperature exponentially with the blackbody e-folding time (*3*)

$$\tau_b = c/4\sigma T_i^3 \qquad (4)$$

where c is a time-invariant heat capacity per unit of area, σ the Stefan-Boltzmann constant, and T_i the initial temperature.

Second, consider a planet with climate feedback factor f and fixed heat capacity c. This system has the e-folding time (*6*)

$$\tau = f\tau_b \qquad (5)$$

because most climate feedbacks come into play only in response to the climate change (not the change in climate forcing). For example, if CO_2 is doubled, the initial heating is ~4 W m^{-2}, independent of f or $\Delta T_{eq}(2 * CO_2)$. However, if positive feedbacks come into play, such as the water vapor feedback that responds to temperature, the heating decreases more gradually than in the absence of the positive feedback, and the full response is delayed.

In these examples the heat capacity c is in immediate thermal contact with the atmosphere. This is relevant to a planet totally covered by a mixed-layer ocean, if there is negligible heat exchange between the mixed layer and the deeper ocean. For mixed-layer depth d_0 = 100 m and T_i = 255 K, the effective temperature of the earth, τ_b is ~3.5 years and thus $f \simeq 3$ yields $\tau \simeq 10$ years.

The earth is more complex than this, principally because there is significant heat exchange between the mixed layer and deeper ocean, as recognized in the CO_2 assessment reports (*1, 2, 4*) and earlier (*7*). Also, we must account for the fact that oceans cover only 70 percent of the earth.

The box diffusion ocean model of Oeschger *et al.* (*8*) provides insight into the effect of the deeper ocean on climate sensitivity. This model has a well-mixed upper layer connected to the deeper ocean by Fickian diffusion. The diffusion coefficient k is specified from observed behavior of transient tracers, such as tritium sprinkled on the ocean surface during atomic testing in the 1960's. The d_0 appropriate for time scales greater than 1 year is the global mean annual maximum, ~100 m (*6*). With this d_0, transient tracers imply an effective global k of 1 to 2 cm^2 sec^{-1} (*6, 8, 9*).

The relation between climate response time, τ, and f can be demonstrated by a scale analysis. Let $d \simeq (k\tau)^{1/2}$ represent depth of penetration of temperature change into the diffusive layer and $D = d_0 + d$ represent total depth of penetration. The surface response time for the box diffusion model is proportional to the depth of penetration of the temperature change. Thus

$$\tau \simeq \frac{D}{d_0}\tau_0 = \frac{D}{d_0} f \tau_b \qquad (6)$$

where τ_0 is the (isolated) mixed-layer response time and τ_b is the blackbody (no feedback) mixed-layer response time. In the limit $k \rightarrow 0$, $D \rightarrow d_0$ and Eq. 6 reduce to the isolated mixed-layer result (Eq. 5). For large k, $d \simeq (k\tau)^{1/2} >> d_0$ and

$$\tau \simeq \frac{(k\tau)^{1/2}}{d_0} f \tau_b \qquad (7)$$

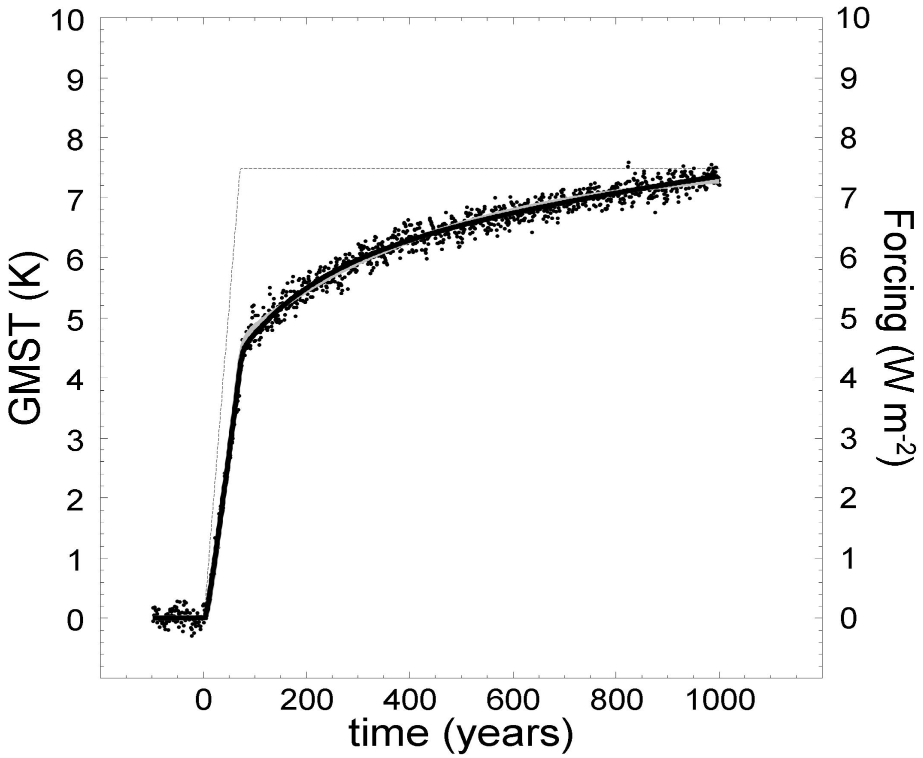
or, solving for τ

$$\tau \simeq kf^2 \left(\frac{\tau_b}{d_0}\right)^2 \propto f^2 \text{ (large } k) \qquad (8)$$

Typical values for k (1 cm^2 sec^{-1}) and τ_0 (10 years) yield $d \simeq 170$ m, large enough to be nearer the diffusive regime than the isolated mixed-layer regime. More quantitative calculations, given below, indicate that for these values τ is nearly proportional to f^2 for the box diffusion model.

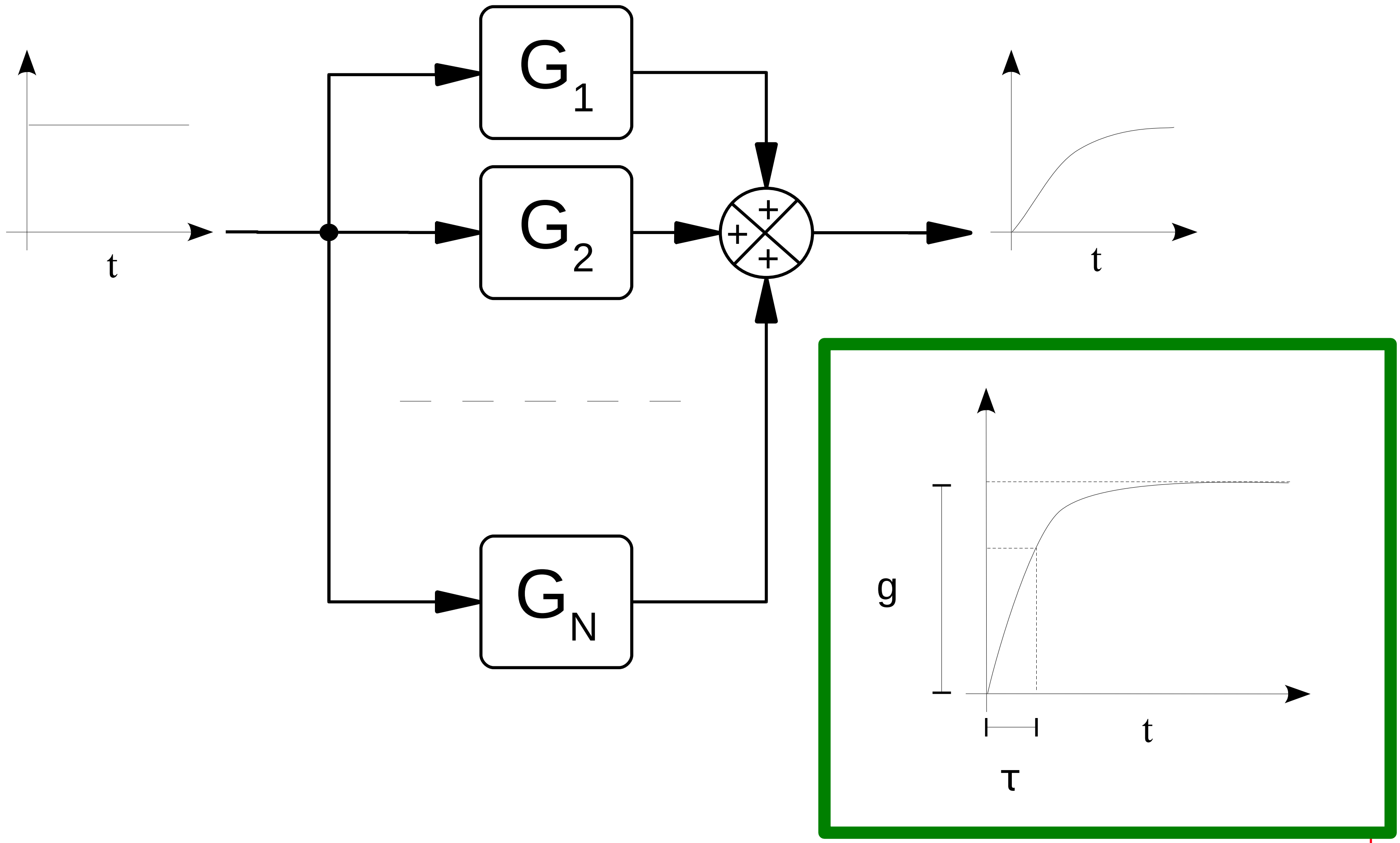
(Hansen et al. – 1985)

climate model response functions

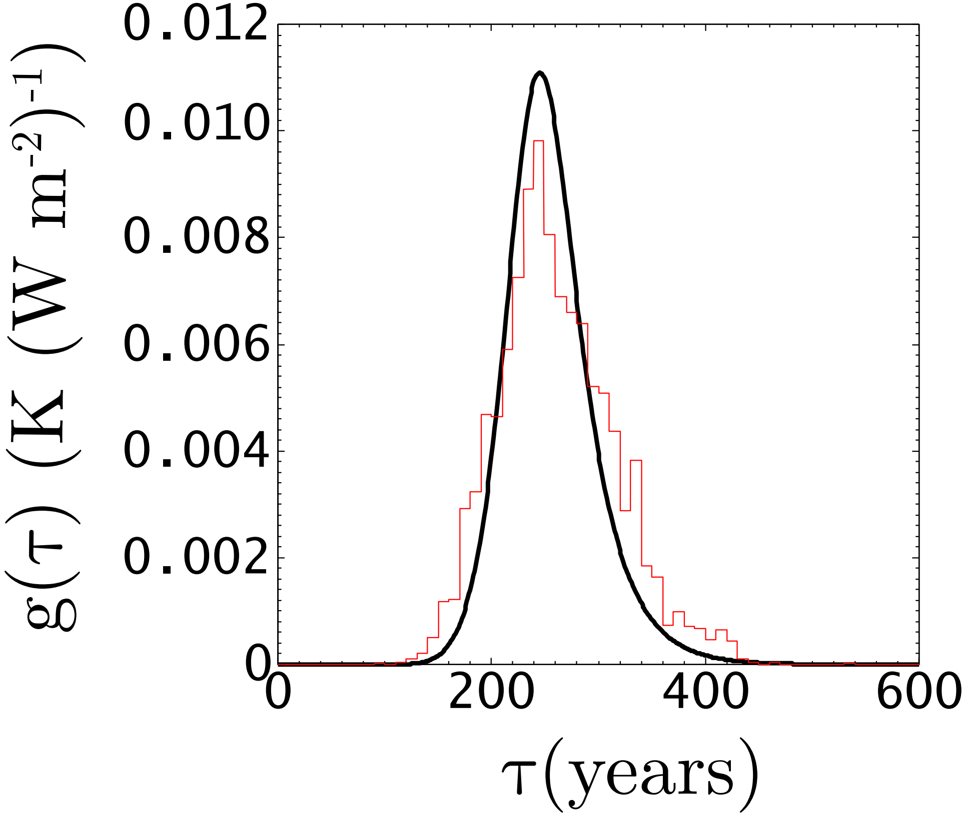


(Li et al. – 2008)

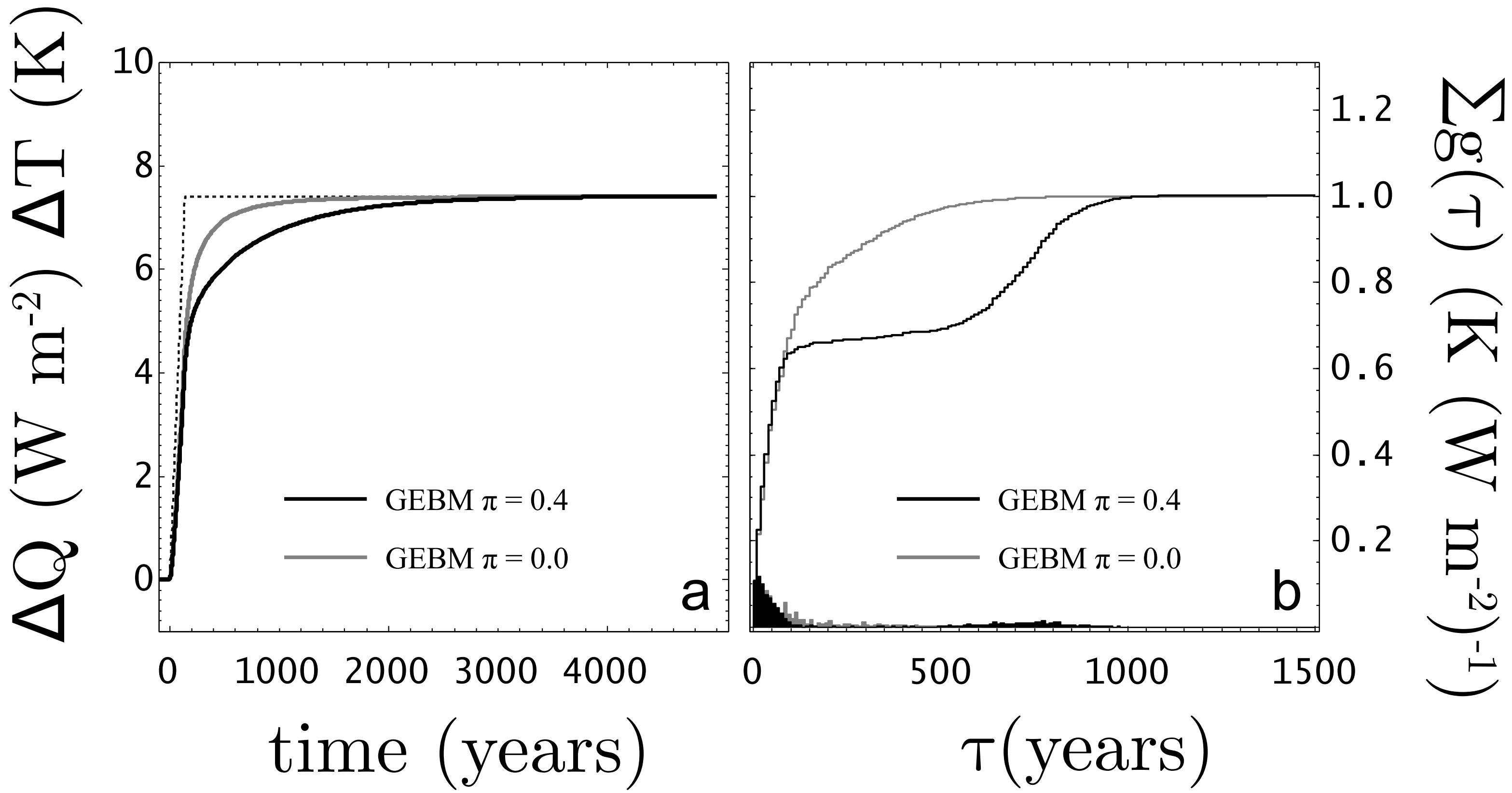
multiple response functions



timescale – gain relationship

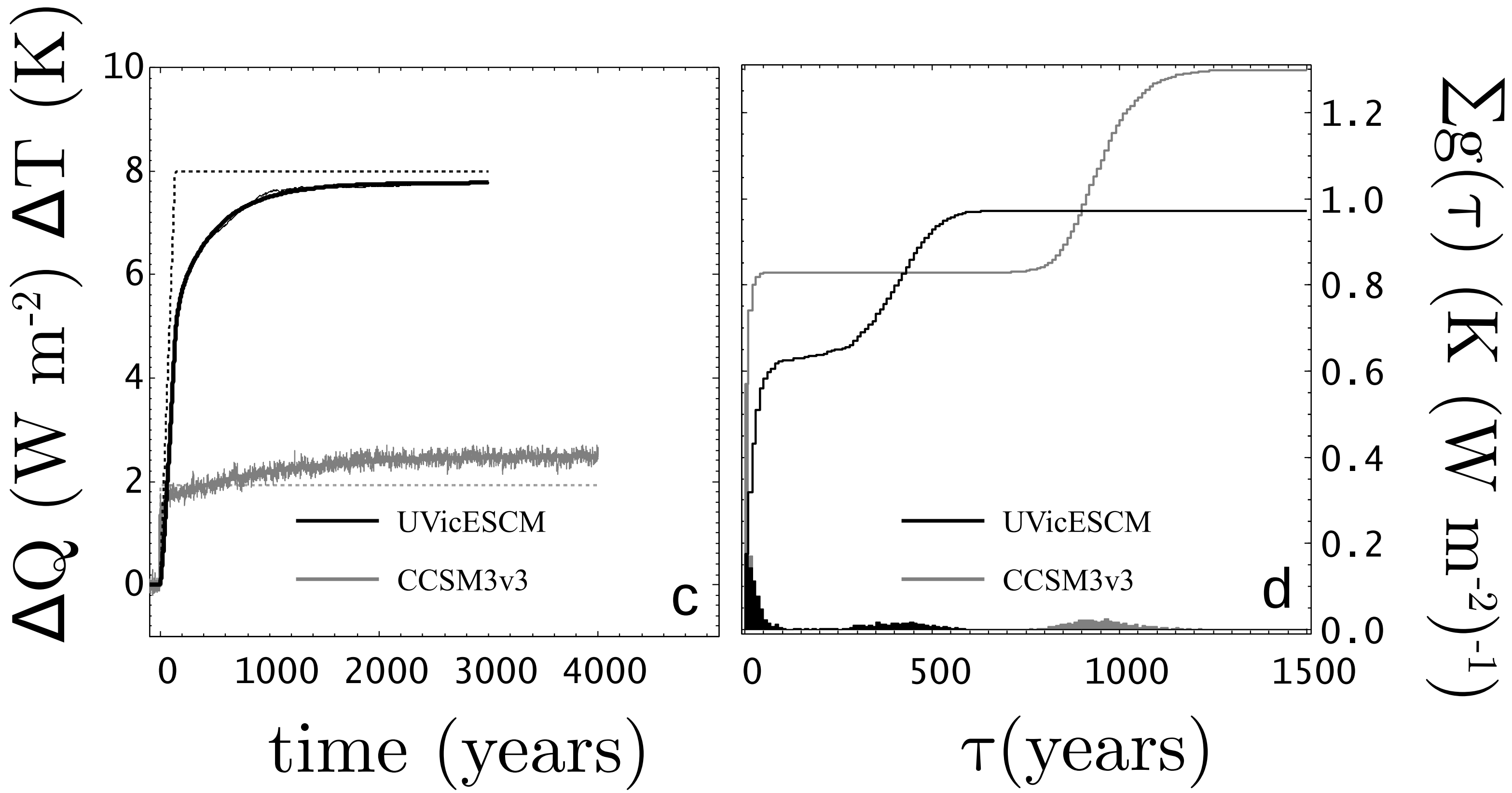


MAGICC



(Clim. Dynamics – in press)

CCSM3, UVic



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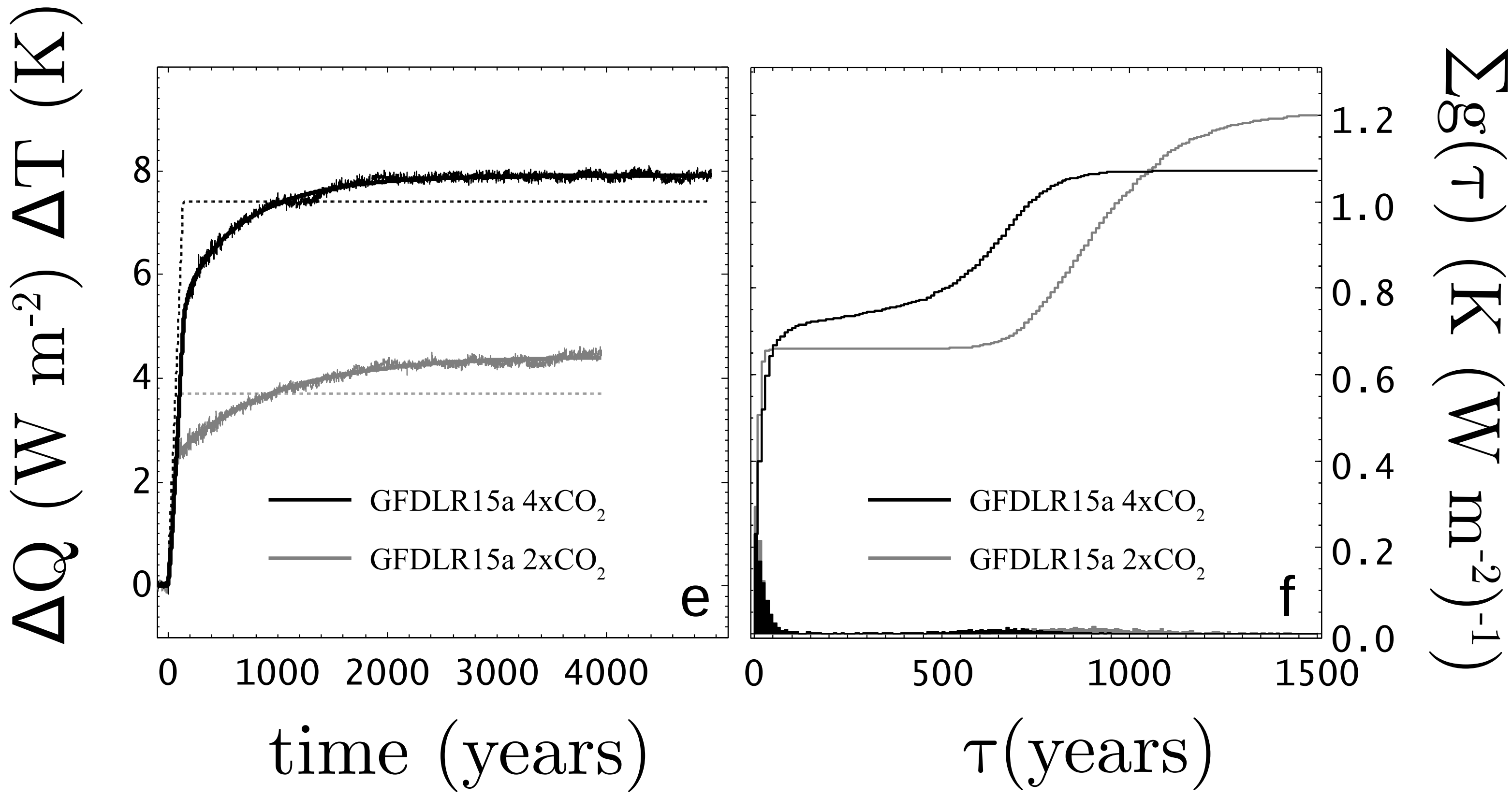
GFDL



(Clim. Dynamics – in press)

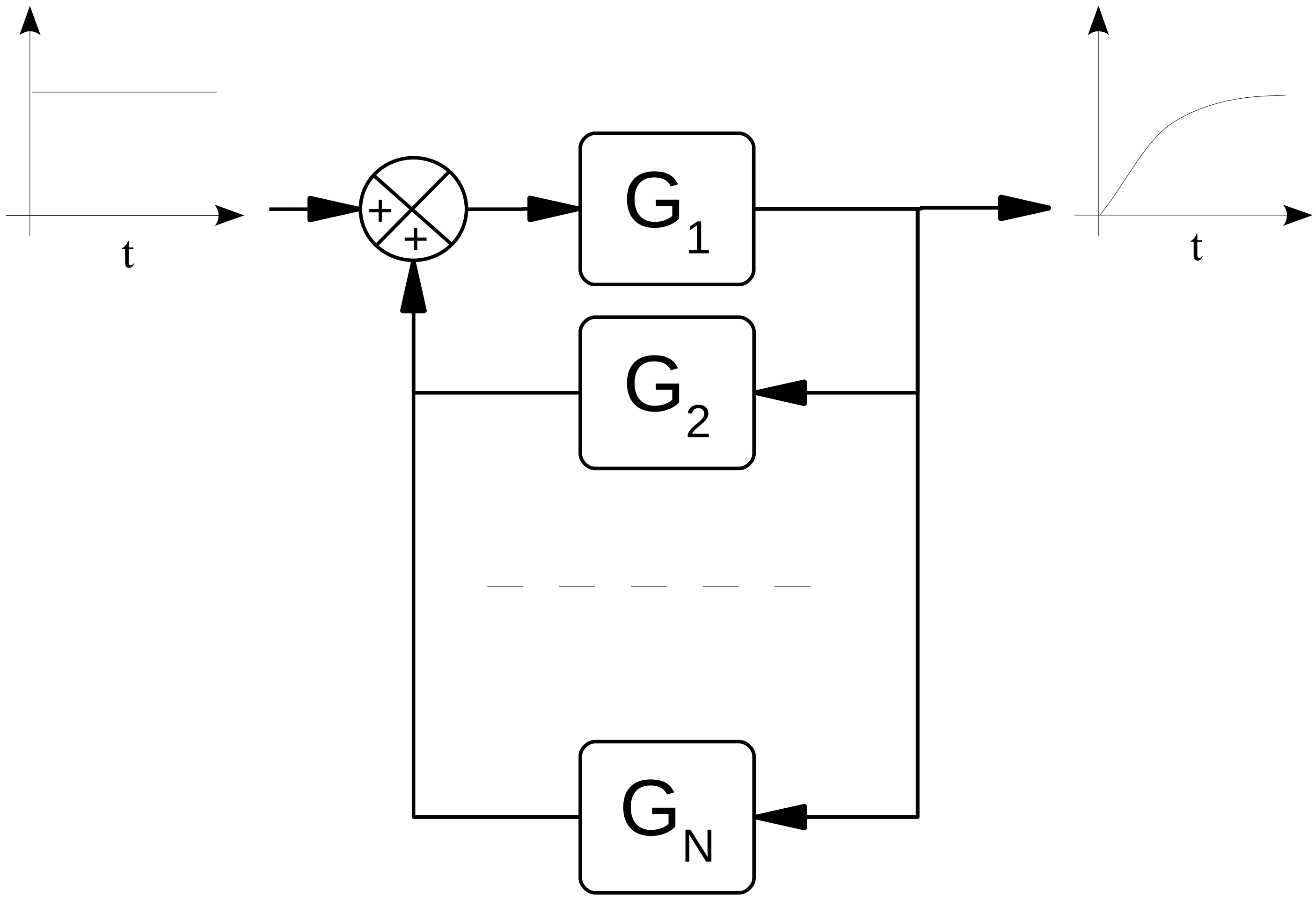
Supplementary Table for Section 8.8

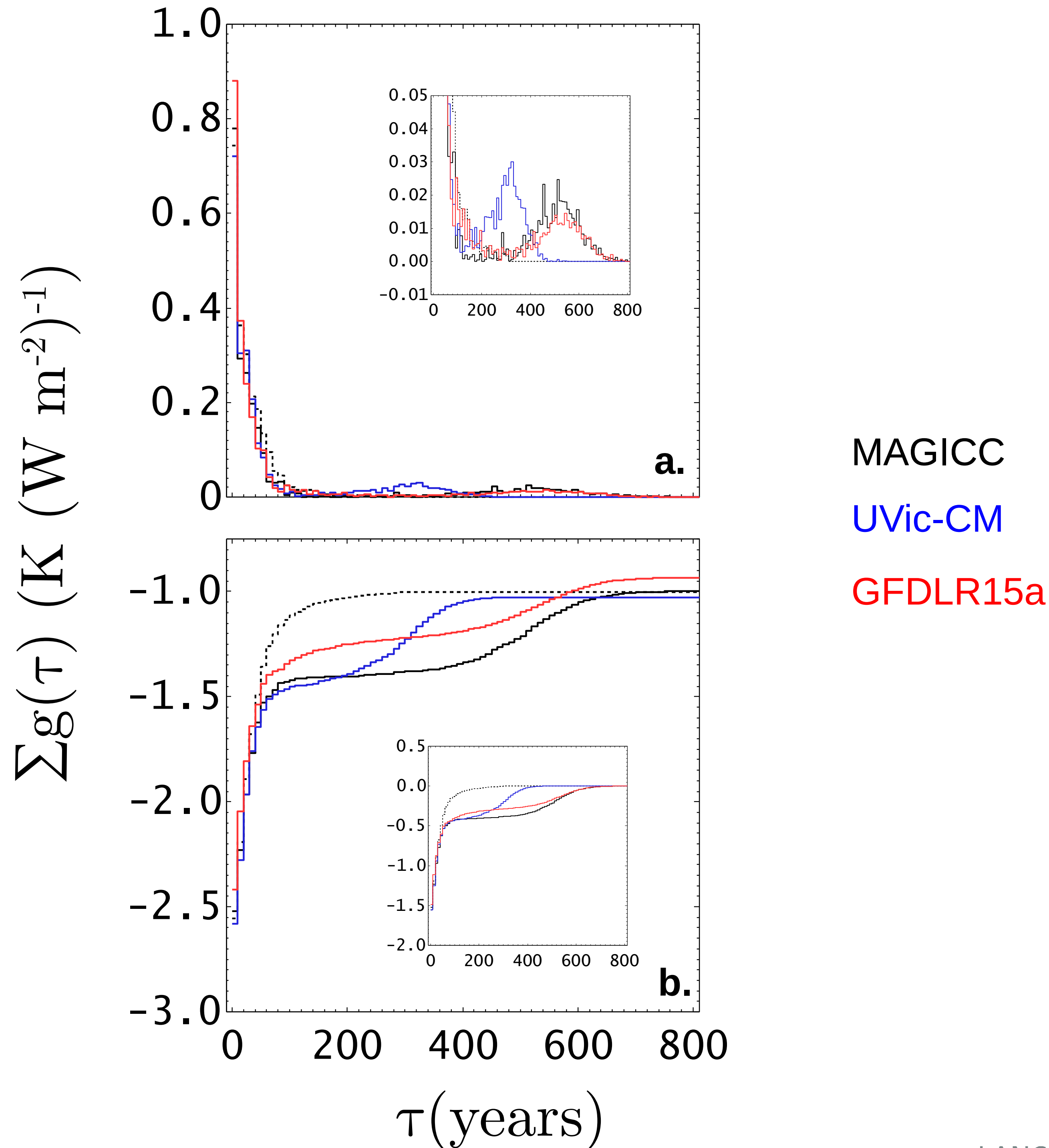
Table S8.1 Parameter values used in a simple climate model (MAGICC) to approximately reproduce results from the AOGCM multi-model dataset at PCMDI.

AOGCM	F_{2x} ($W\ m^{-2}$)	ΔT_{eff} (°C)	k ($cm^2\ s^{-1}$)	RLO
3: CCSM3	3.95	2.37	1.73	1.18
4: CGCM3.1(T47)	3.32	3.02	1.57	1.58
6: CNRM-CM3	3.71	2.45	1.21	1.10
7: CSIRO-Mk3.0	3.47	2.21	2.03	1.33
8: ECHAM5/MPI-OM	4.01	3.86	1.22	1.41
9: ECHO-G	3.71	3.01	2.01	1.65
10: FGOALS-g1.0	3.71	1.97	4.57	1.64
11: GFDL-CM2.0	3.50	2.35	1.42	1.47
12: GFDL-CM2.1	3.50	2.28	2.23	1.58
14: GISS-EH	4.06	3.04	2.35	1.21
15: GISS-ER	4.06	2.57	4.42	1.44
16: INM-CM3.0	3.71	2.28	0.79	1.10
17: IPSL-CM4	3.48	3.83	1.94	1.26
18: MIROC3.2(hires)	3.14	5.87	1.18	1.15
19: MIROC3.2(medres)	3.09	3.93	2.29	1.58
20: MRI-CGCM2.3.2	3.47	2.97	1.22	1.45
21: PCM	3.71	1.88	1.57	1.45
22: UKMO-HadCM3	3.81	3.06	1.01	1.65[a]
23: UKMO-HadGEM1	3.78	2.63	1.32	1.20

(Randall et al. – 2007)

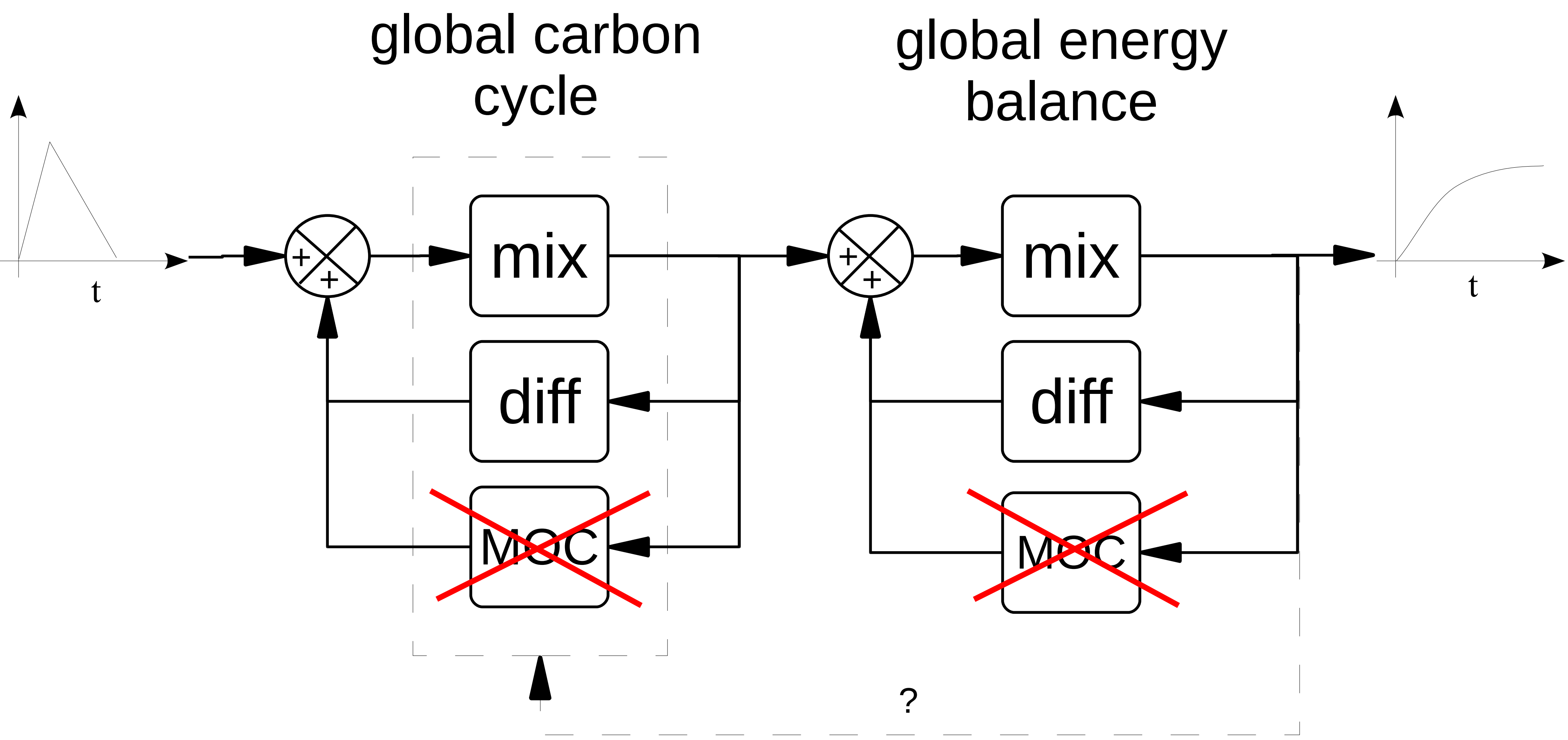
climate system feedbacks





(submitted J. Climate)

very simple climate model



conclusions

- climate models show discernible differences in the timescales of different processes leading to response function type behaviors

- for global temperature approximately 50-60% of the model climate system response is associated with timescales < 100 years. Timescale responses > 100 years are inherently uncertain in these models