Review of concepts and methods relating to climate sensitivity

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Questions

What is climate sensitivity useful for?

What determines climate sensitivity?

Is climate sensitivity constant?

What is the role of ocean heat uptake?

How is climate sensitivity evaluated?

What is climate sensitivity useful for?

Climate sensitivity defined in terms of temperature

Define equilibrium climate sensitivity $\Delta T_{2\times}^{\mathrm{eqm}}$ as steady-state global average surface air temperature change ΔT for 2 $\times$ CO_2.
Said to be 1.5–4.5 K in IPCC reports.

Why do we care about ΔT?

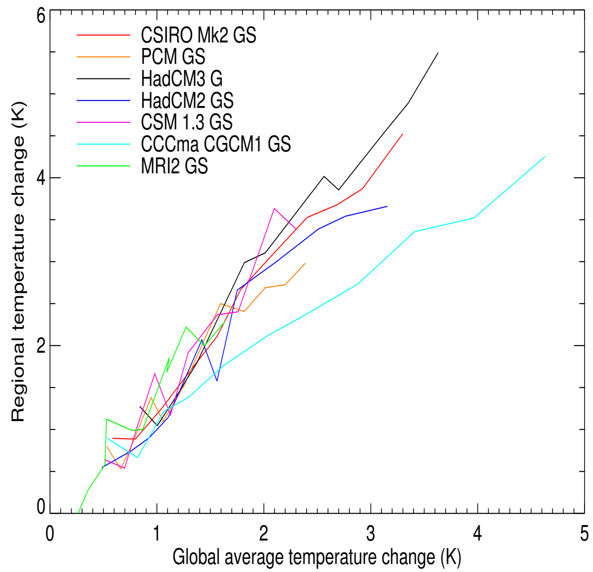
We find that spatial and seasonal patterns of change in all quantities scale quite well with ΔT, which we can regard as measuring the "magnitude" of climate change.

Also ΔT is well measured and has good signal/noise.

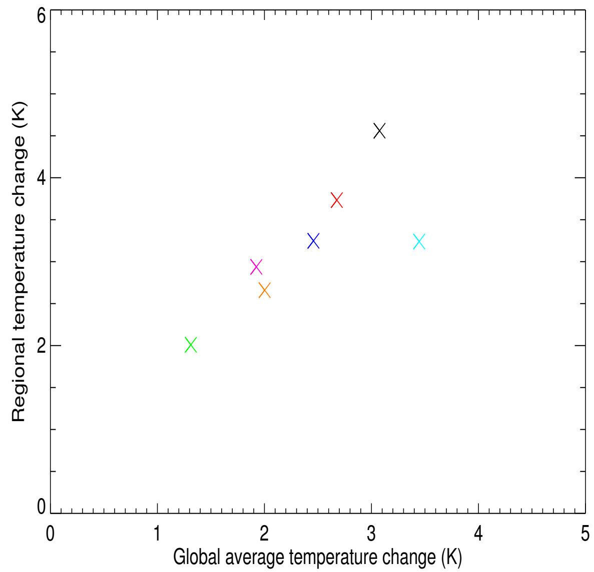
Scaling works within a model, but not across models

For any model, any quantity $V(\text{space}, \text{time}) = \Delta T(\text{time}) \times \text{pattern}(\text{space})$
Maybe this works in the real world too.

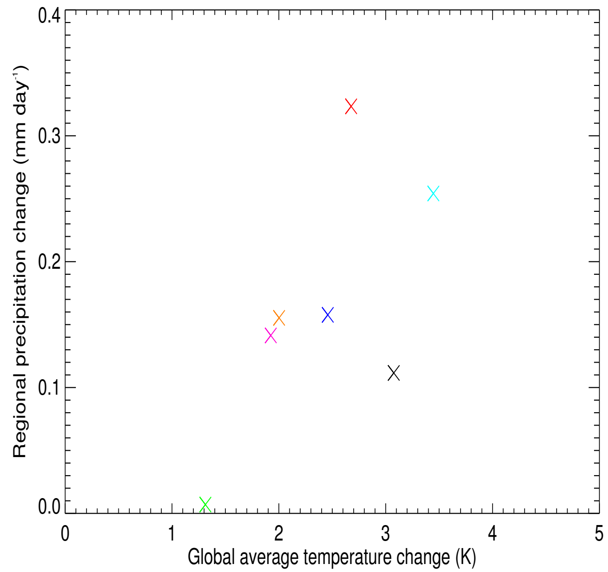
Changes in Northern Europe versus ΔT



Decadal ΔT_L v ΔT



2050–2100 ΔT_L v ΔT



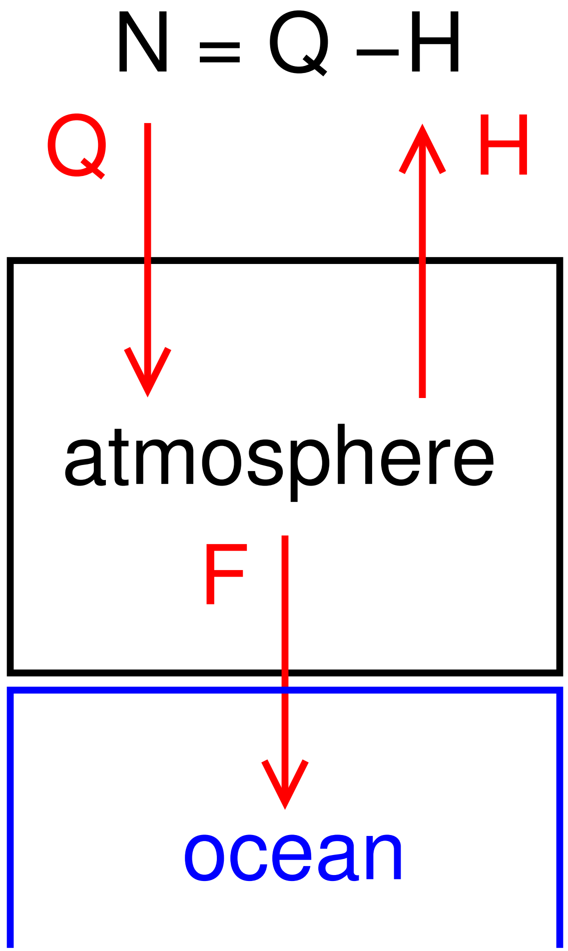
2050–2100 Δppn_L v ΔT

ΔT is not a good predictor of regional changes across models.

Heat budget of the global climate system

Define radiative forcing Q as the change caused by the forcing agent in net downward radiation at the tropopause, everything else being held constant.

Forcing agent *e.g.* CO_2, insolation, land surface change.



N is the net heat flux into the climate system.
$H(\Delta T)$ is the radiative response of the climate system, dependent on ΔT.

Initial steady state $N = Q = H = 0$ and $\Delta T = 0$.

When Q imposed, system responds by adjusting H so as to oppose $Q \Rightarrow H$ must increase with ΔT.

Climate sensitivity characterises H.

While climate is changing, $N \neq 0$ and $F \simeq N$.

In perturbed steady state $N = 0 \Rightarrow Q = H(\Delta T)$.

Radiative forcing and climate response

In any given GCM, for steady state ($Q = H$) with various forcings, we find $\Delta T \propto Q$.
We write $H(\Delta T) = \alpha\,\Delta T$, α constant in W m^{-2} K^{-1}, $\alpha > 0$ for stability.
Hence $Q = \alpha\,\Delta T$ in steady state and $\Delta T_{2\times}^{\mathrm{eqm}} = Q_{2\times}/\alpha$.

If α is really a constant, we can estimate response to different magnitude and nature of forcing agent from Q alone without doing the climate experiment.

FAR uses two formulations

$\Delta T = sQ$ $s = 1/\alpha$ (K W^{-1} m^2) (s written λ in FAR) climate sensitivity parameter How much the climate system changes when Q is imposed	$Q = \alpha\,\Delta T$ α (W m^{-2} K^{-1}) (α written Λ in FAR) climate feedback parameter How strongly the climate system reacts to ΔT

What determines climate sensitivity?

Processes

Climate sensitivity is determined by a sum of processes causing radiative responses of the form $-y_i \, \Delta T$.

$$Q = -\Delta T \sum y_i \Rightarrow \alpha = -\sum y_i$$

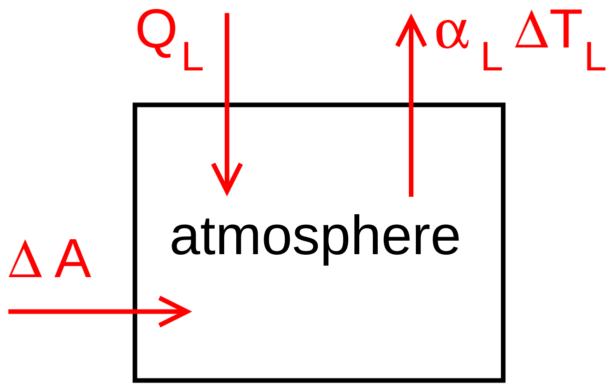
Positive y_i for positive feedback.
The black-body y is large and negative, making $\alpha > 0$ as required.

Response	y	$\Delta T_{2\times}^{\mathrm{eqm}} \pm$
Black-body	-3.3	
Water vapour $+$ lapse rate	1.4$\pm$0.2	1.4
Surface albedo	0.3$\pm$0.1	0.9
Cloud	0.6$\pm$0.3	2.5

Colman (2003)

Regionality

Consider local heat balance, including horizontal atmospheric heat convergence A.



For a steady state,

$$\Delta A + Q_L = \alpha_L \, \Delta T_L = \alpha_L P_L \, \Delta T$$

Then since $\langle \Delta A \rangle = 0$, on the global average,

$$Q = \langle \alpha_L P_L \rangle \, \Delta T$$

Hence $\alpha = \langle \alpha_L P_L \rangle$, where $\alpha_L P_L$ is the local contribution to the global α (Boer and Yu, 2003).

How does regionality really work?

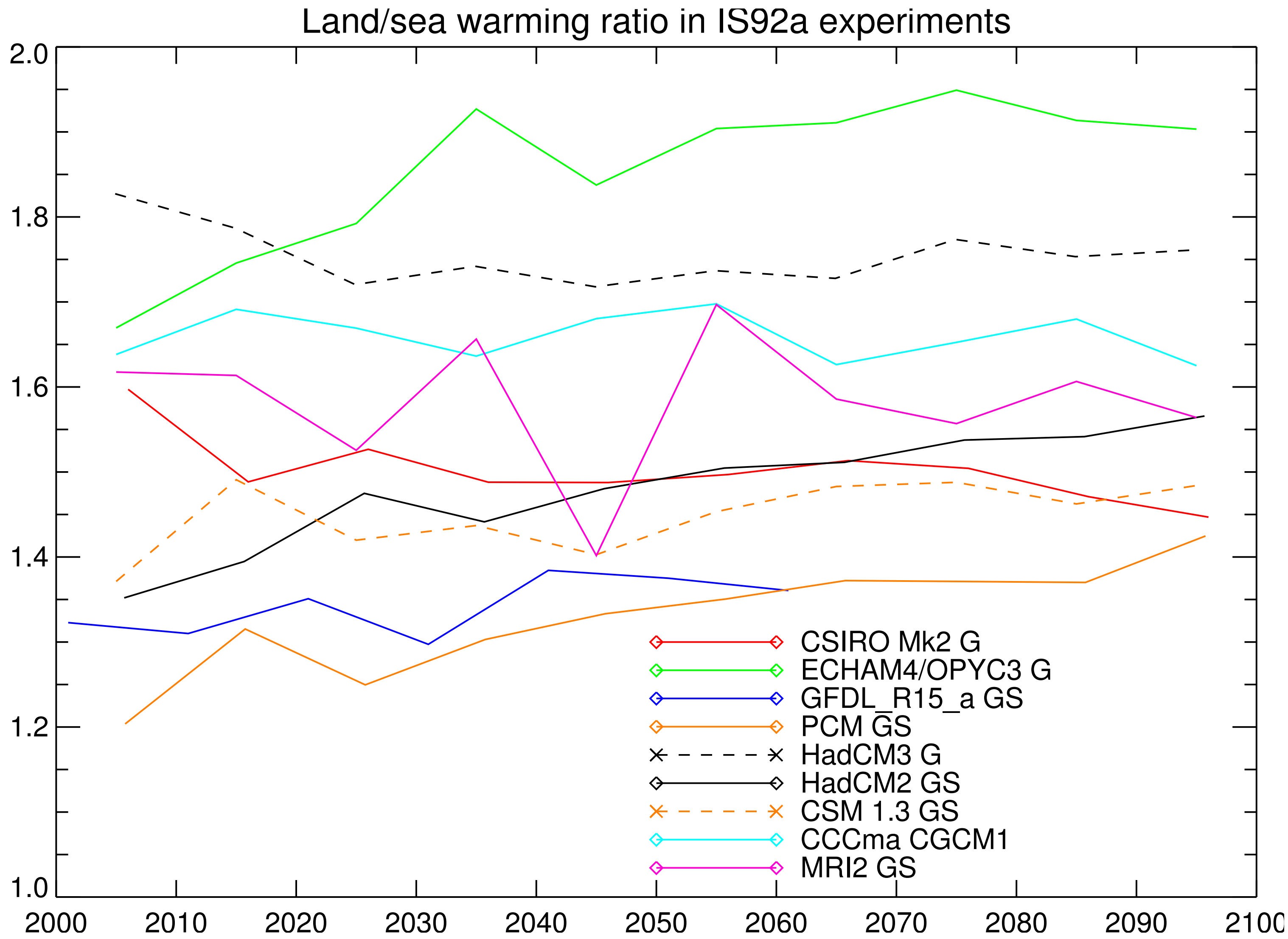
Are all feedbacks really local?

Should we look at surface feedbacks rather than TOA to determine ΔT_L?

Is ΔA a contributory cause of local changes or is it a response to ΔT_L?

What determines how effective ΔA is?

Can we account quantitatively for this, for instance?



Is climate sensitivity constant?

as a function of
- magnitude of forcing
- nature of forcing agent
- climate change

Distinction of forcing and feedback

Forcing is a radiative response to the forcing agent, more quickly than ΔT can change substantially

- Instantaneous forcing
- Induced forcing *e.g.* stratospheric adjustment, indirect and semi-direct aerosol forcings

Feedback is a radiative response to climate change

- Fast feedback scales with ΔT
- Slow feedback has its own timescale, so not $\propto \Delta T$ during climate change

Different “responsivity” for various forcing agents?

Hansen *et al.* (1997) define responsivity

$$R \equiv \frac{\Delta T \text{ for some forcing}}{\Delta T \text{ for } \Delta(\text{solar}) \text{ with same adjusted forcing}}$$

They find $R(\text{CO}_2) = 1.26$ and $R(\text{ozone})$ has a range of values.

Joshi *et al.* (2003) investigated this further.

So is α different for various forcing agents?
Another possibility is that the wrong Q is being used.

Forcing diagnosed by preventing climate change

$$Q = N + \alpha\,\Delta T$$

Add a forcing agent, impose $\Delta T = 0$, so $Q = N$, thus diagnose Q.

Hansen *et al.* (2002) suppress SST changes $\rightarrow$
"fixed SST forcing" Q_{SST}.

Shine *et al.* (2003) suppress T_* changes $\rightarrow$
"adjusted troposphere and stratosphere forcing" Q_{ats}.

$$s_? = \Delta T / Q_? = 1/\alpha_?$$

Forcing	ΔT	Q	Q_{SST}	Q_{ats}	s	s_{SST}	s_{ats}
$2 \times CO_2$	1.9	3.8	4.3	4.3	0.50	0.44	0.44
+2% solar	1.9	4.9	4.9	4.2	0.39	0.39	0.45
aerosol $\omega =$1.0	−1.7	−4.7	−4.6	−4.1	0.36	0.37	0.45
aerosol $\omega =$0.8	2.9	1.6	4.8	6.8	1.8	0.61	0.43
upper tropospheric O_3	0.3	1.0	—	0.68	0.30	—	0.48

Q_{ats} is the sum of instantaneous and induced forcings.

Forcing and feedback from time-dependent climate change

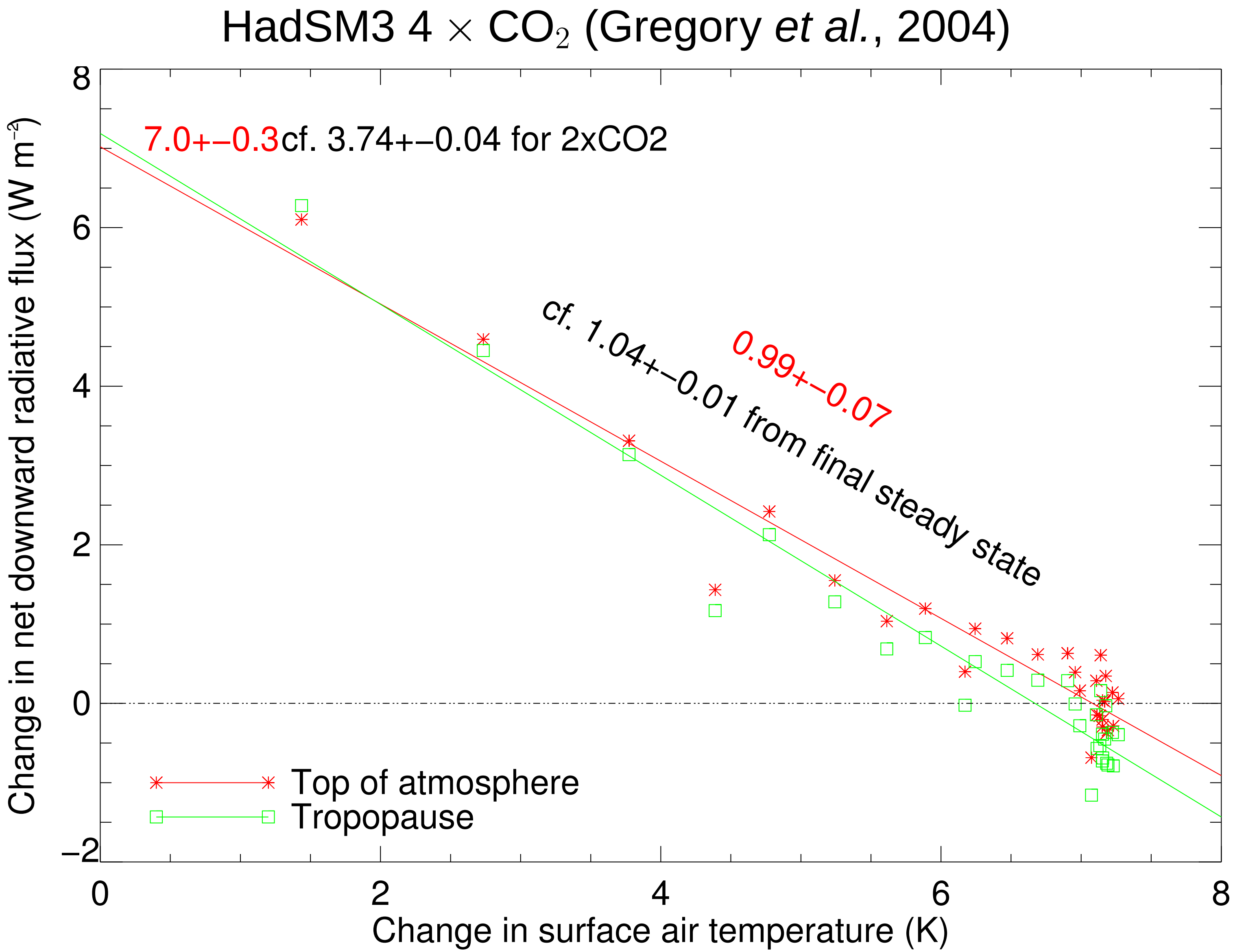
The idea of proportionality of feedbacks to ΔT comes from considering **steady state** for different forcings, $Q = -\sum y_i \, \Delta T$.

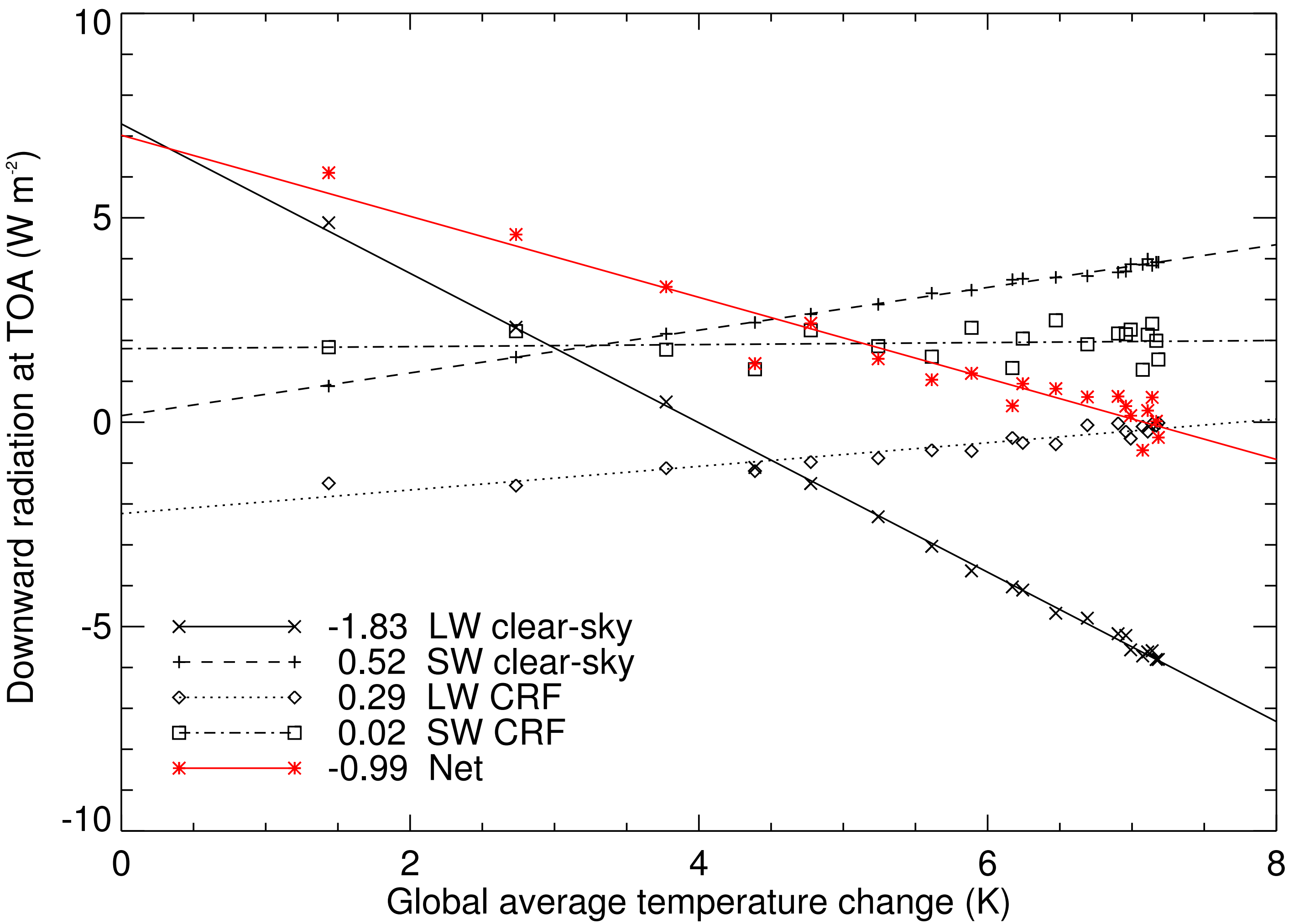
Are climate feedbacks are proportional to ΔT **during time-dependent change**?

For any fixed forcing agent, assume that Q is constant, so $N = Q - \alpha \, \Delta T$ suggests that N versus ΔT will be a straight line, with intercept Q, slope $-\alpha$.

This makes a practical distinction between forcing and feedback.

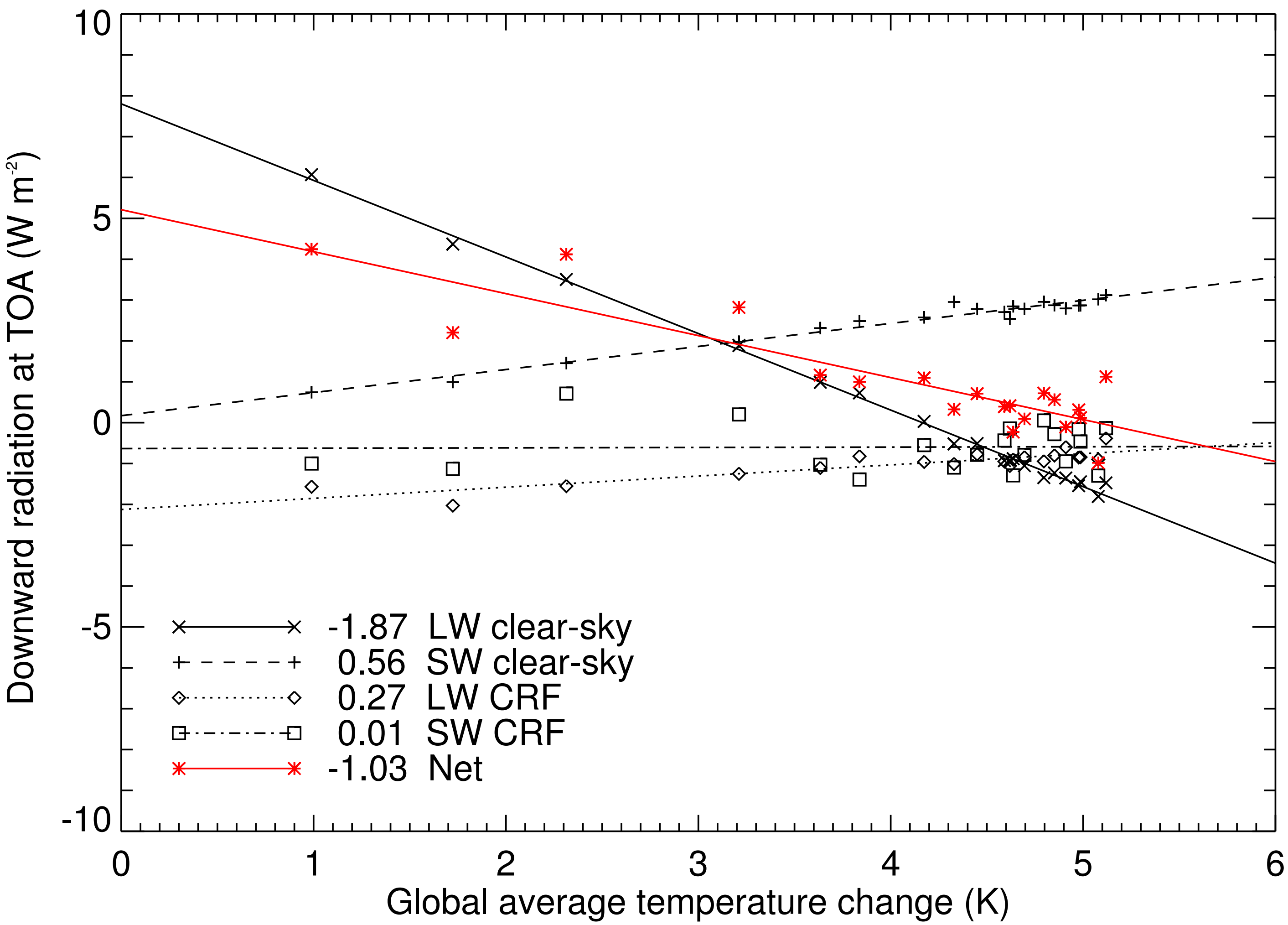
Forcing is the limit of N for $\Delta T \to 0$; this should be consistent with $Q_{\mathrm{ats}} \equiv N(\Delta T = 0)$.


HadSM3 4 × CO2 (Gregory et al., 2004)
7.0+–0.3cf. 3.74+–0.04 for 2xCO2
cf. 1.04+–0.01 from final steady state
0.99+–0.07
Top of atmosphere
Tropopause
Change in net downward radiative flux (W m-2)
Change in surface air temperature (K)
8
6
4
2
0
−2
0
2
4
6
8



Downward radiation at TOA (W m^{-2})
Global average temperature change (K)
10
5
0
-5
-10
0
2
4
6
8
-1.83 LW clear-sky
0.52 SW clear-sky
0.29 LW CRF
0.02 SW CRF
-0.99 Net

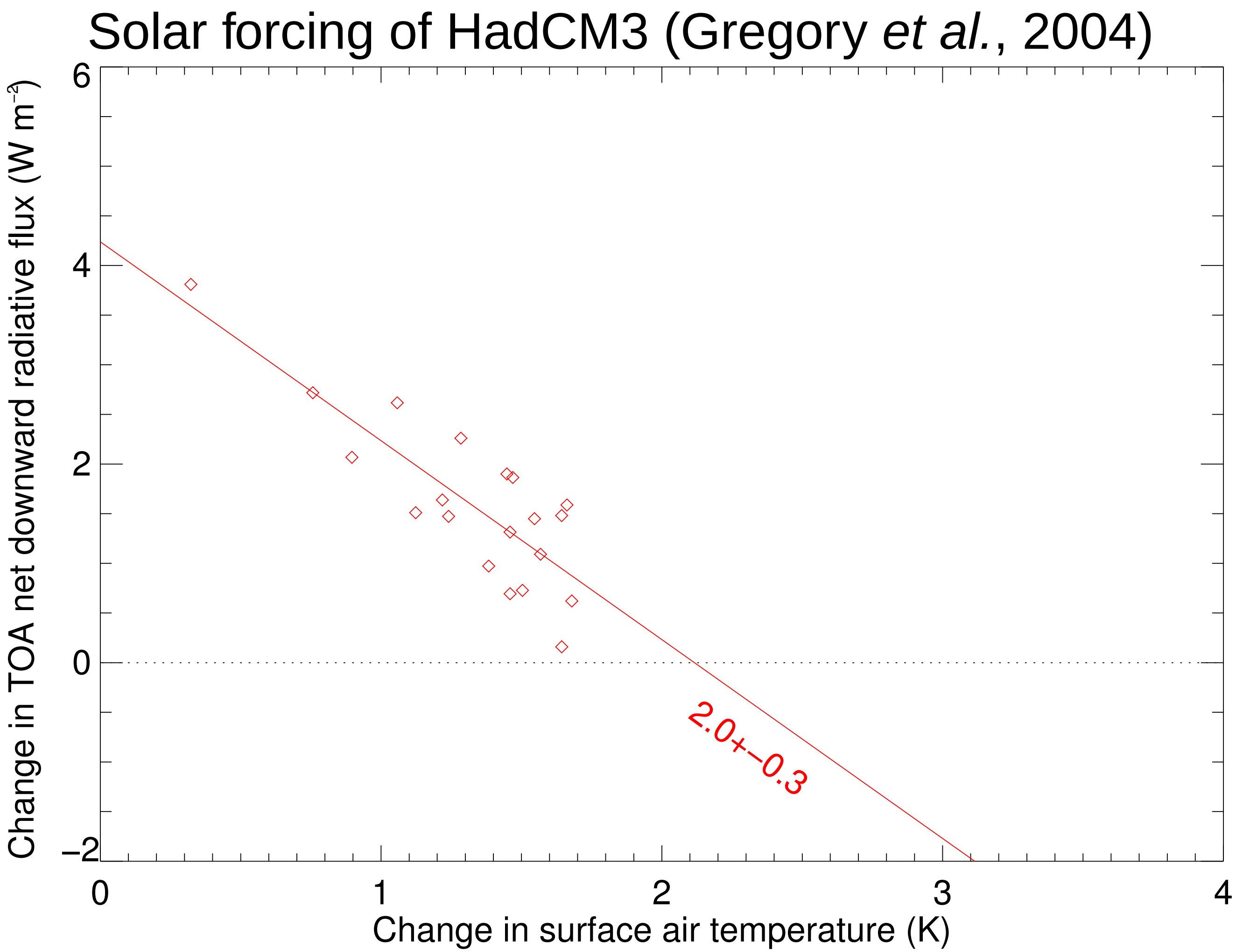
A member of the HadSM3 QUMP ensemble with unusual SW cloud forcing (acknowledgements to Mark Webb).



Could it be that some of the spread in $\Delta T_{2\times}^{\mathrm{eqm}} = Q/\alpha$ is due to spread in Q?

Climate feedback dependent on nature of the forcing

Since $\alpha = \langle \alpha_L P_L \rangle$, α depends on P_L, which could depend on forcing agent.



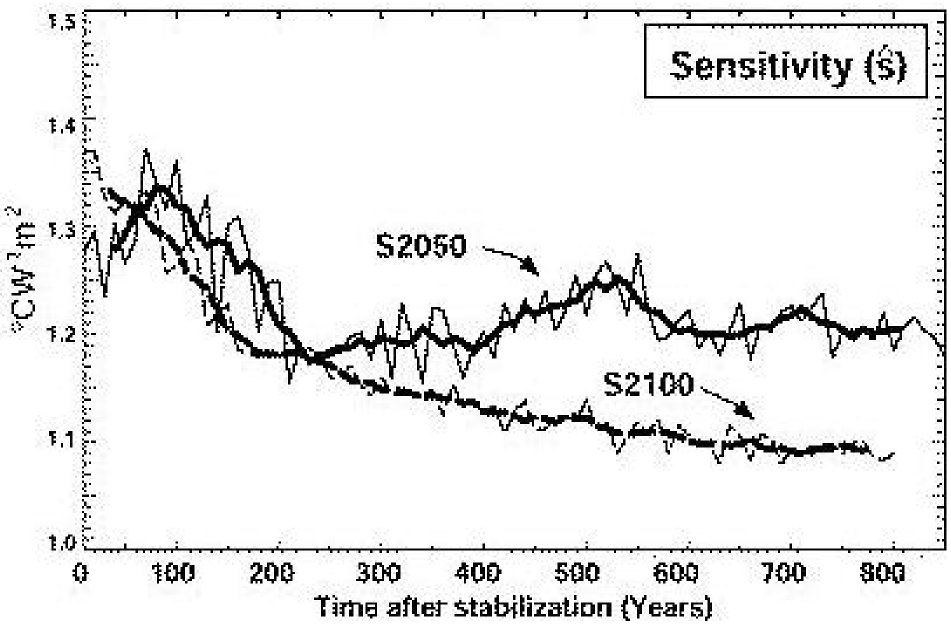
In Cess-type runs, $P_L = 1 \Rightarrow \alpha = \langle \alpha_L \rangle$, probably different from CO_2 runs.

Effective climate sensitivity

To measure variation of α as climate changes, define effective climate sensitivity $\Delta T_{2\times}^{\mathrm{eff}}$ (Murphy, 1995)

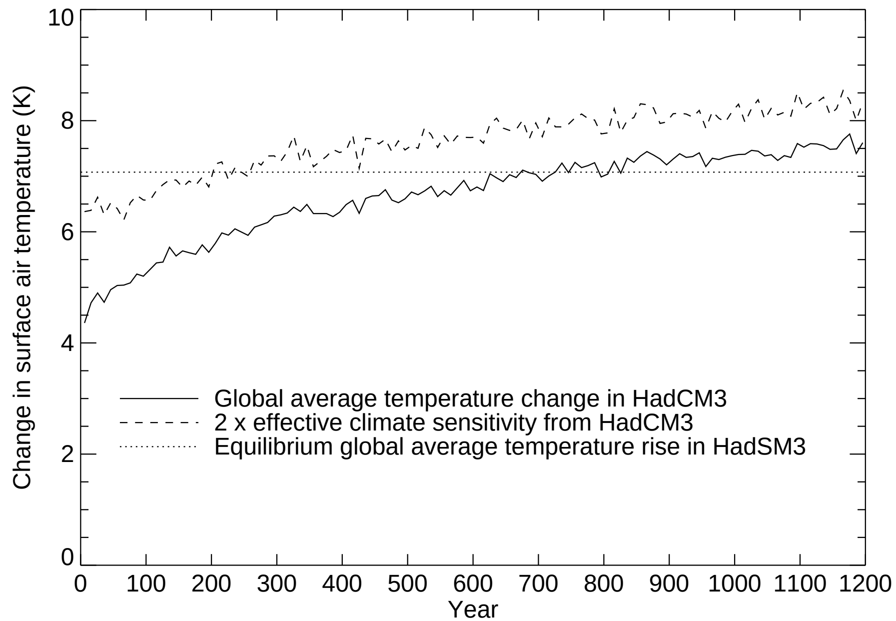
$$N = Q - \alpha_{\mathrm{eff}}\,\Delta T \Rightarrow \alpha_{\mathrm{eff}} = \frac{Q-N}{\Delta T} \text{ and } \Delta T_{2\times}^{\mathrm{eff}} \equiv \frac{Q_{2\times}}{\alpha_{\mathrm{eff}}}$$

CCCma CGCM



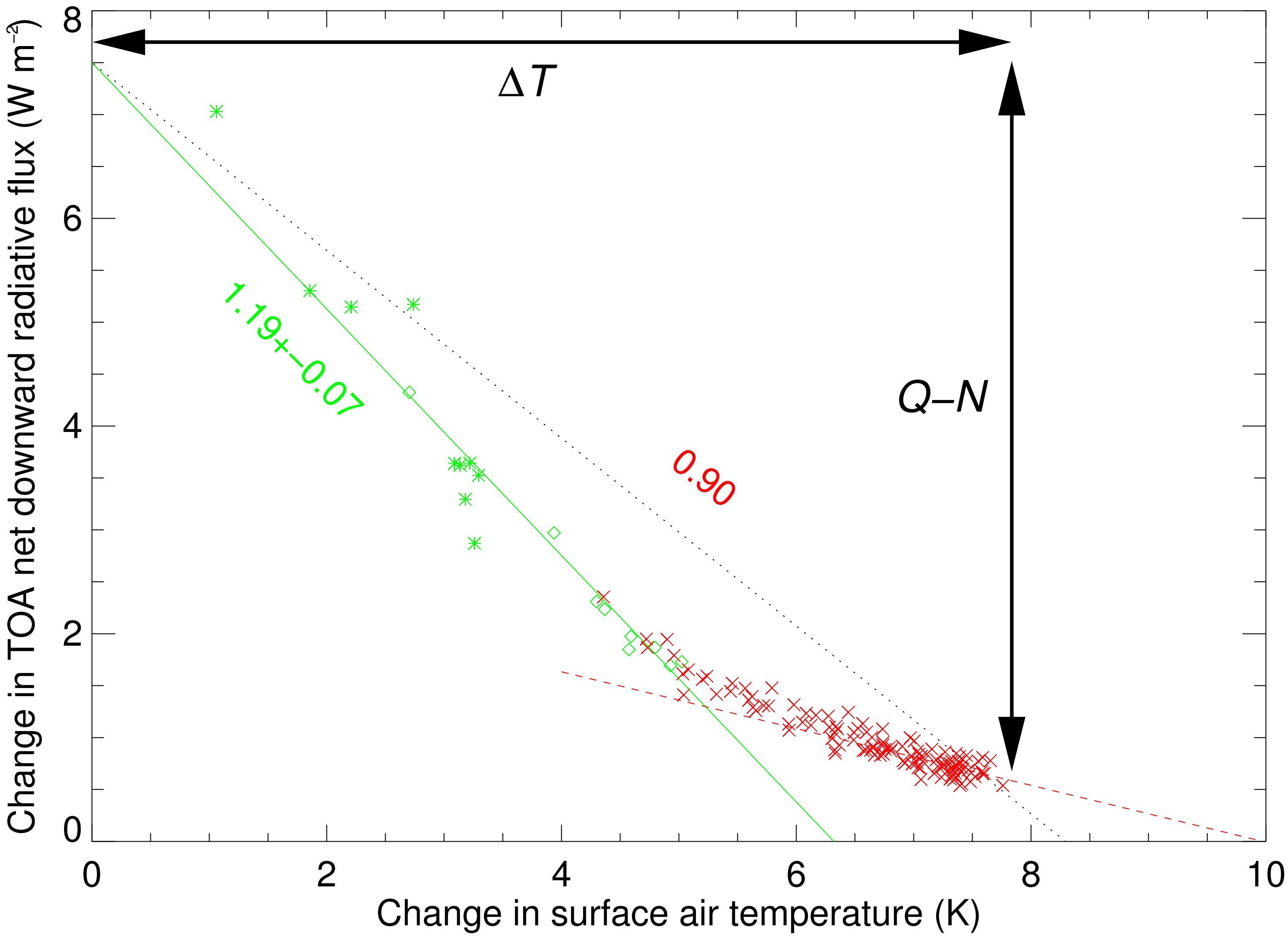
Boer and Yu (2003)

HadCM3



(not shown in) Gregory *et al.* (2004)

HadCM3 4 $\times$ CO_2 (Gregory *et al.*, 2004)



P_L changes as SSTs evolve; hence $\alpha = \langle \alpha_L P_L \rangle$ changes.

Slow feedbacks

Changing SST patterns have two effects:

- Fast feedbacks measured by $\alpha = \langle \alpha_L P_L \rangle$ are modified.
- Further climate change (slow feedback) occurs in response to $\delta\alpha\,\Delta T$.

Ben Booth (PhD thesis) has distinguished these for HadSM3.

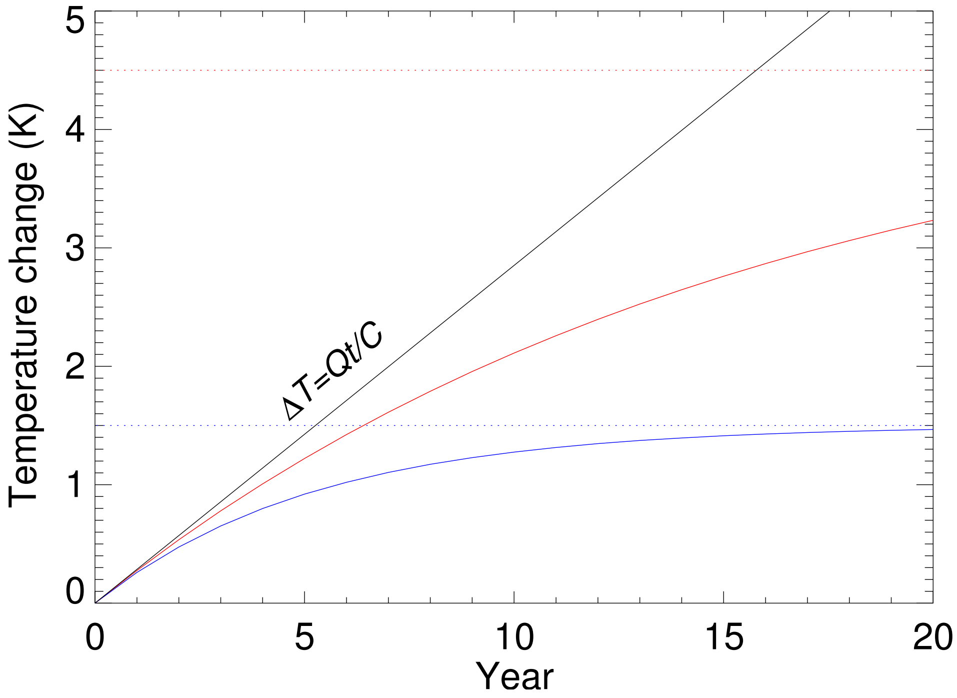
If modelled interactively, other slow (possibly irreversible) feedbacks, could be THC collapse, vegetation change, carbon cycle feedback, slow atmospheric chemistry, ice sheet change.

If imposed non-interactively, these are forcings.

What is the role of ocean heat uptake?

Timescale of climate response

Suppose $F = C\,\mathrm{d}T/\mathrm{d}t$, C a constant heat capacity.



Switching on Q at $t = 0$ gives

$$\Delta T = \Delta T^{\mathrm{eqm}}(1 - \exp(-t/\tau))$$

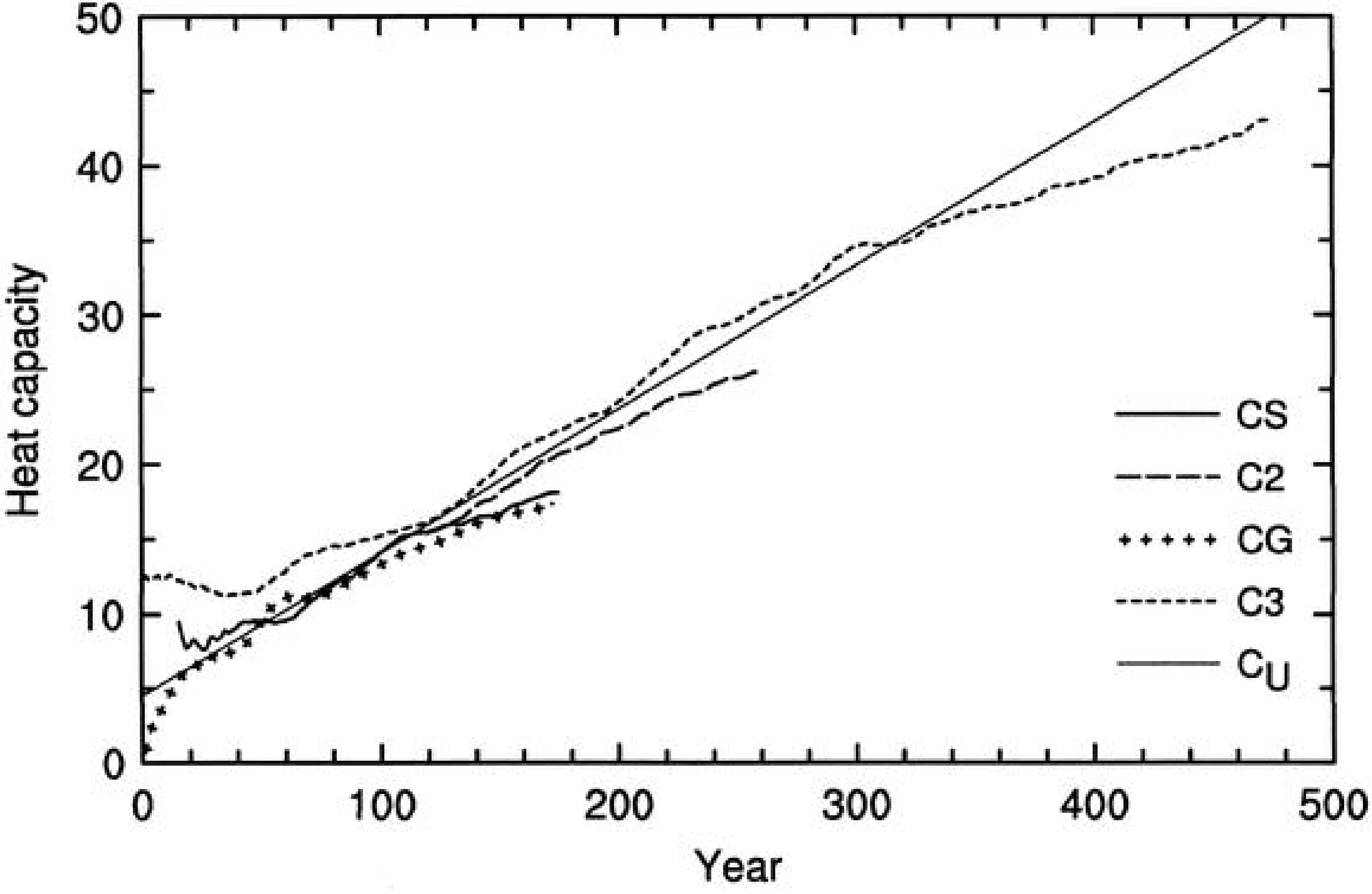
with $\tau = C/\alpha$ and $\Delta T^{\mathrm{eqm}} = Q/\alpha$.

For $t \to 0$, $\mathrm{d}T/\mathrm{d}t \to Q/C$ for all α.

Timescale is longer for larger C and smaller α (larger $\Delta T^{\mathrm{eqm}}_{2\times}$).

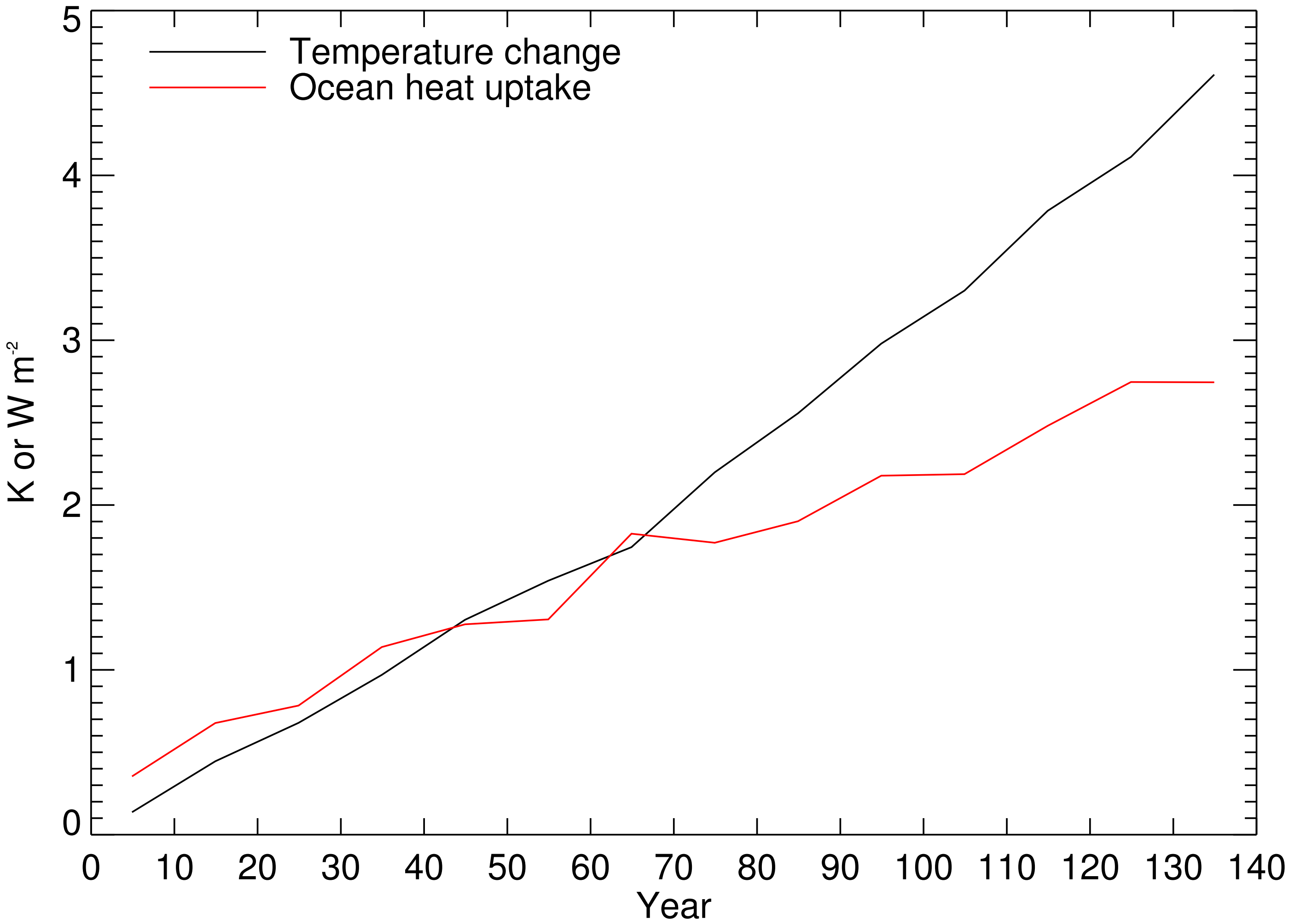
With linearly increasing Q, $\Delta T/\Delta T^{\mathrm{eqm}}$ at any t is smaller for the same conditions.

Effective heat capacity is not constant in an AOGCM



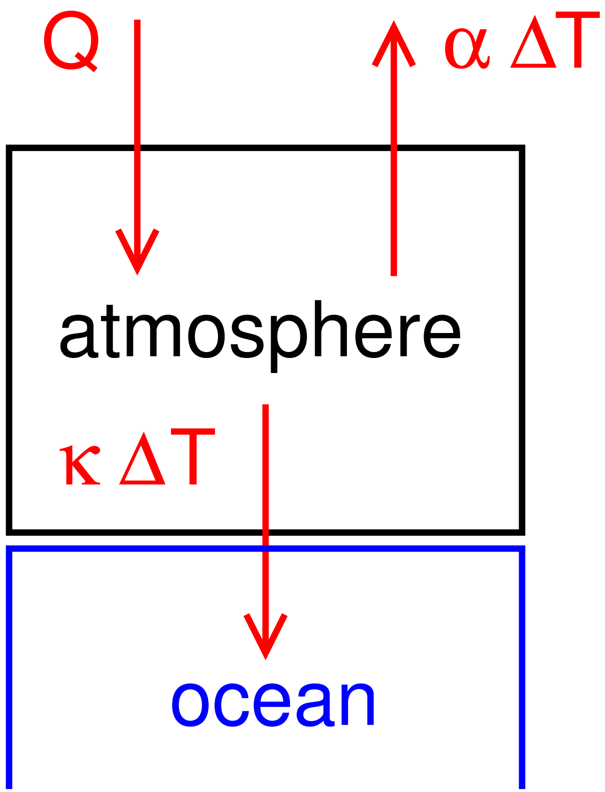
Watterson (2000), heat capacity in 10^8 W m^{-2} K^{-1}.

ΔT and F in HadCM3 1% CO_2 experiment



We could assume $F \propto \Delta T$.

Relative importance of climate sensitivity and heat uptake



Suppose $F = \kappa\Delta T$, κ ($\mathrm{W\,m^{-2}\,K^{-1}}$) constant ocean heat uptake efficiency.

No good for stabilisation, when $N = F \rightarrow 0$.

With this model $Q = F + \alpha\Delta T = \Delta T(\kappa + \alpha)$.

Diagnostic relationship $Q \propto \Delta T$—no timescale!

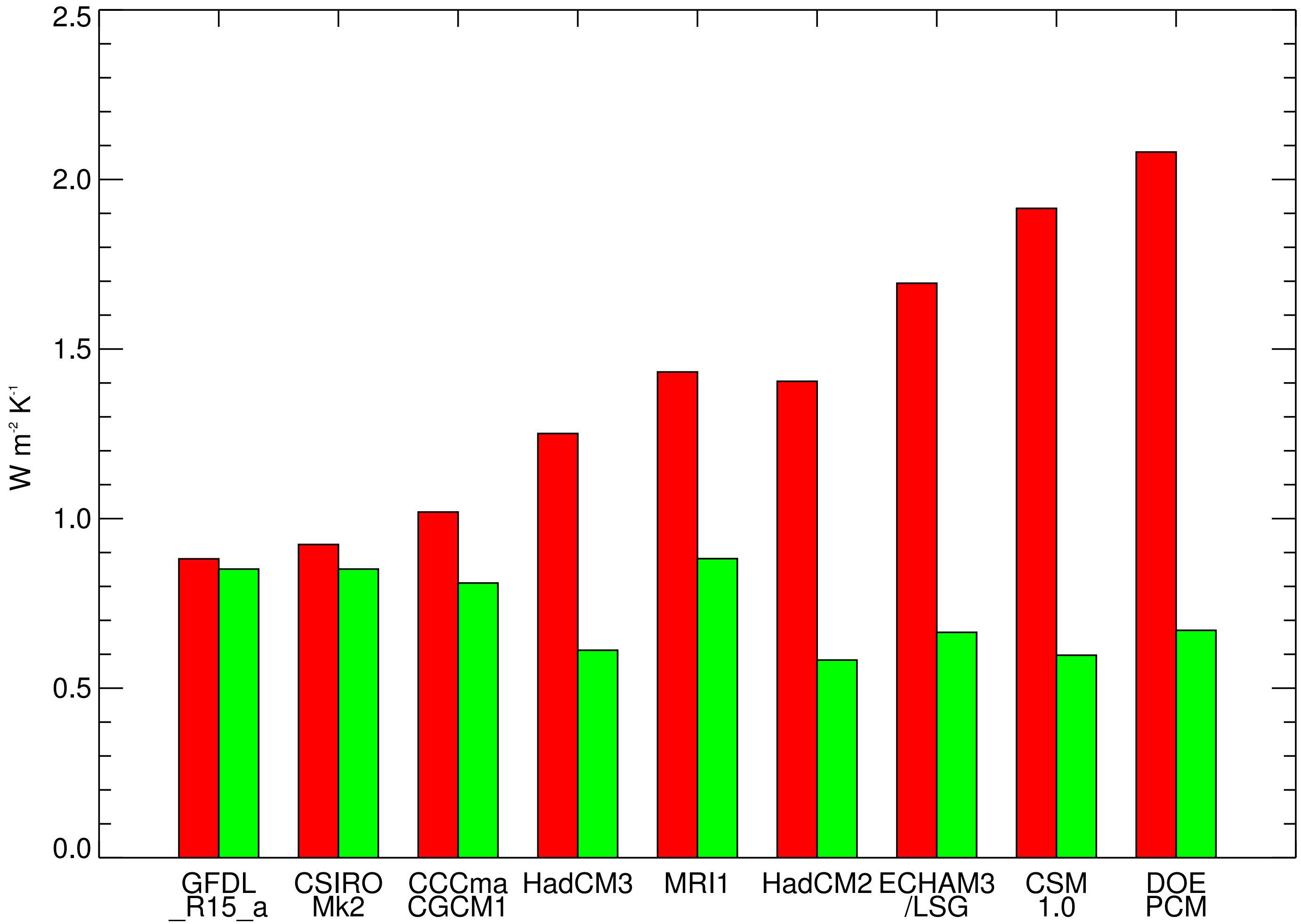
Cf. Allen *et al.* (2000) relation between past and future.

As before, $\Delta T/\Delta T^{\,\mathrm{eqm}} = \alpha/(\kappa + \alpha)$ is smaller for

- smaller α (larger $\Delta T^{\,\mathrm{eqm}}$).
- larger κ (like C).

Climate sensitivity and ocean heat uptake together determine the transient climate response, defined as ΔT at time of $2 \times CO_2$ in a 1% scenario $\Rightarrow$ TCR $= Q_{2\times}/(\kappa + \alpha)$.

Raper *et al.* (2001)
α and κ
from 1% CO_2

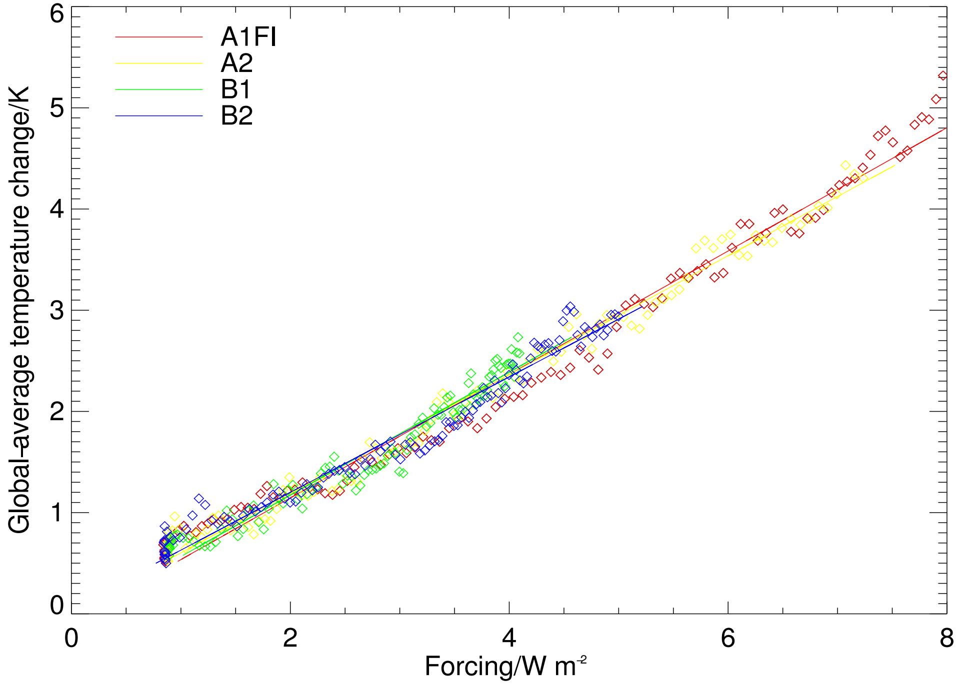


Heat uptake is important. Anticorrelation of κ and α reduces the range of TCR.

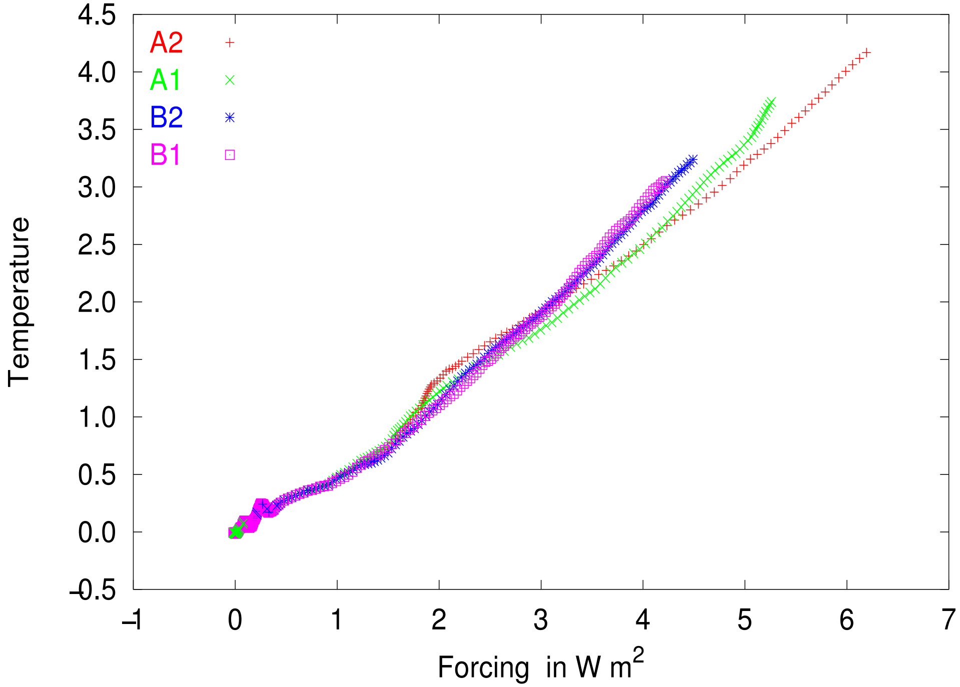
A method for estimating ΔT for SRES scenarios?

$\Delta T \propto Q$ works fairly well, at least for these models; presumably κ is fairly constant.

HadCM3



CSIRO Mk 2



(Ian Watterson, pers. comm.)

How is climate sensitivity evaluated?

A classification of approaches

	Separate processes	Global heat balance
Real-world climate change/variability	**Parametrisation and evaluation of model processes:** Bony Cess Ramaswamy Fu Collins Kinne Allan Williams Hall	$\Delta T_{2\times}^{\mathrm{eqm}}$ **uncertainty from real world:** Jouzel Allen Andronova Knutti Hegerl Joos *Forest&al (2002) Gregory&al (2002)*
Simulated climate change	**Systematic uncertainty in processes to explain spread of model** $\Delta T_{2\times}^{\mathrm{eqm}}$**:** *Hansen&al (1984) W&M (1988) Cess&al (1990, 1996)* Colman Kiehl Meehl Kitoh Sausen Braconnot Otto-Bliesner	$\Delta T_{2\times}^{\mathrm{eqm}}$ **uncertainty from model processes:** *IPCC ARs* Murphy Räisänen *climateprediction.net*

Uncertainty in $\Delta T_{2\times}^{\mathrm{eqm}}$ from model processes

In the models considered

Report	FAR	SAR	TAR
Range of $\Delta T_{2\times}^{\mathrm{eqm}}$ (K)	1.9–5.2	2.1–4.6	2.0–5.1

These are “ensembles of opportunity”.

Met Office QUMP and climateprediction.net are constructing perturbed physics ensembles of models with different formulation.

$$\left.\begin{array}{l}\text{Varying parameters within plausible limits}\\ \text{Weighting by comparison with climatology}\end{array}\right\} \Rightarrow \begin{array}{l}\text{Constraints on } \Delta T_{2\times}^{\mathrm{eqm}}\\ \text{from present climate}\end{array}$$

Systematic application of subjectively chosen constraints.

All models are wrong!

Evaluation of $\Delta T_{2\times}^{\mathrm{eqm}}$ from past climate change

$$N = Q - \alpha\,\Delta T$$

Given Q, ΔT and N, we can evaluate α.

Q is calculated. ΔT is measured or deduced from proxies.

What about N? Three possibilities:

- Consider steady states only, so $N = 0$.
- Obtain measurements of it.
- Use a model to estimate it.

Feedbacks in past climate change (*e.g.* LGM to present, or response to volcanoes) may not be the same as for GHG-forced future climate change

Steady-state palæoclimates

For instance, take present day and LGM as steady states.

The dominant forcings are the result of slow feedbacks.

α might not be the same for these climates or forcings.

	H&M	H++	H&C	B&M
Insolation	0.1			
Ice sheet	-2.9	-2.6	-2.3	-0.9
Vegetation		-0.9	-0.7	-0.7
Snow and sea ice	-0.6			
CO_2	-1.7	-1.6	-1.8	-2.0
Other GHGs		-1.0	-1.0	
Aerosols		-1.0	-0.9	

(Table compiled by H&M)

H&M=Hewitt and Mitchell (1997); H++=Hansen *et al.* (1993); H&C=Hoffert and Covey (1992); B&M=Broccoli and Manabe (1987)

Last 150 years without using a climate model

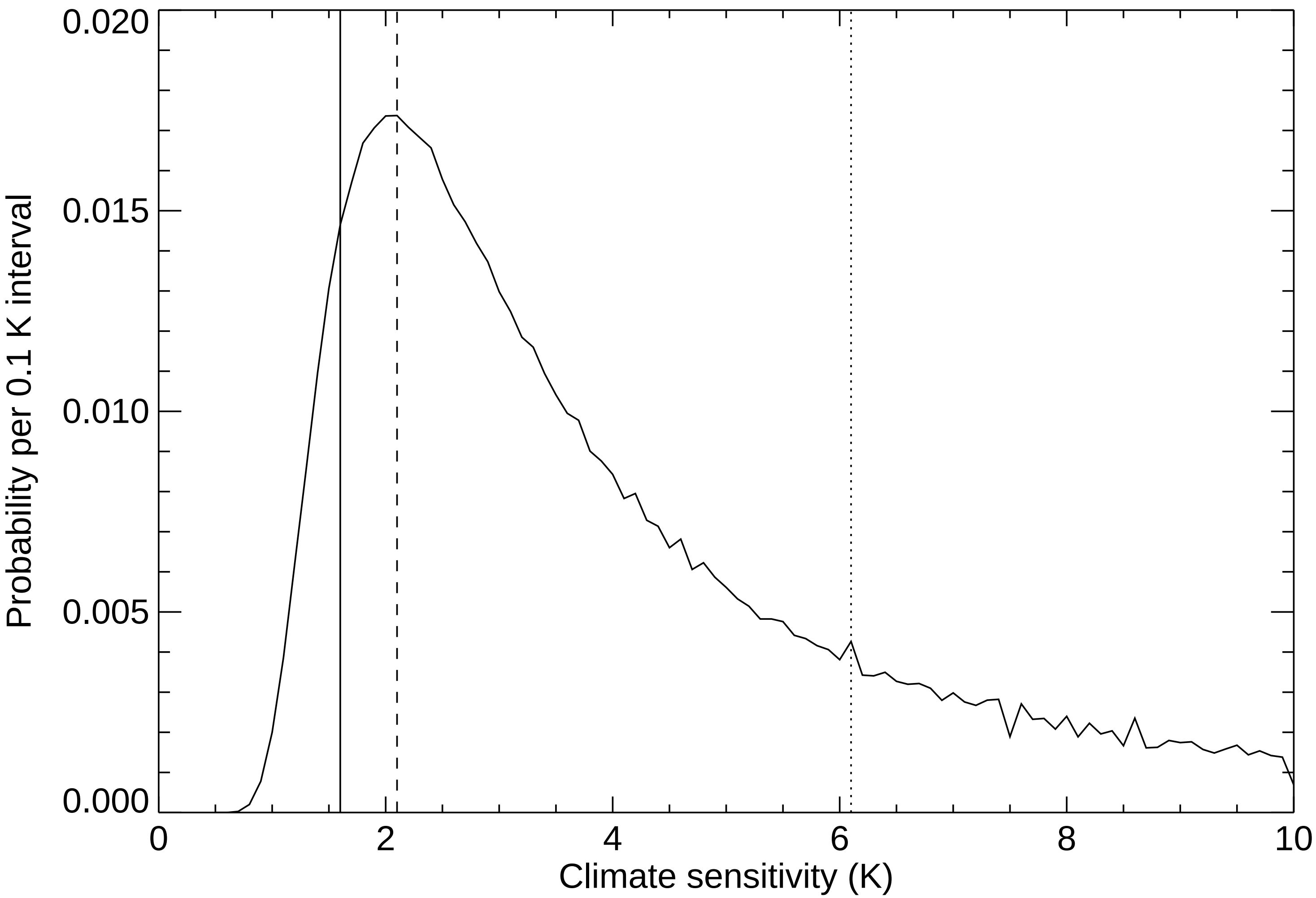
$$N = Q - \alpha\,\Delta T$$

We don't have a control $\Delta T = 0$ for the real world, so consider differences between two unsteady states

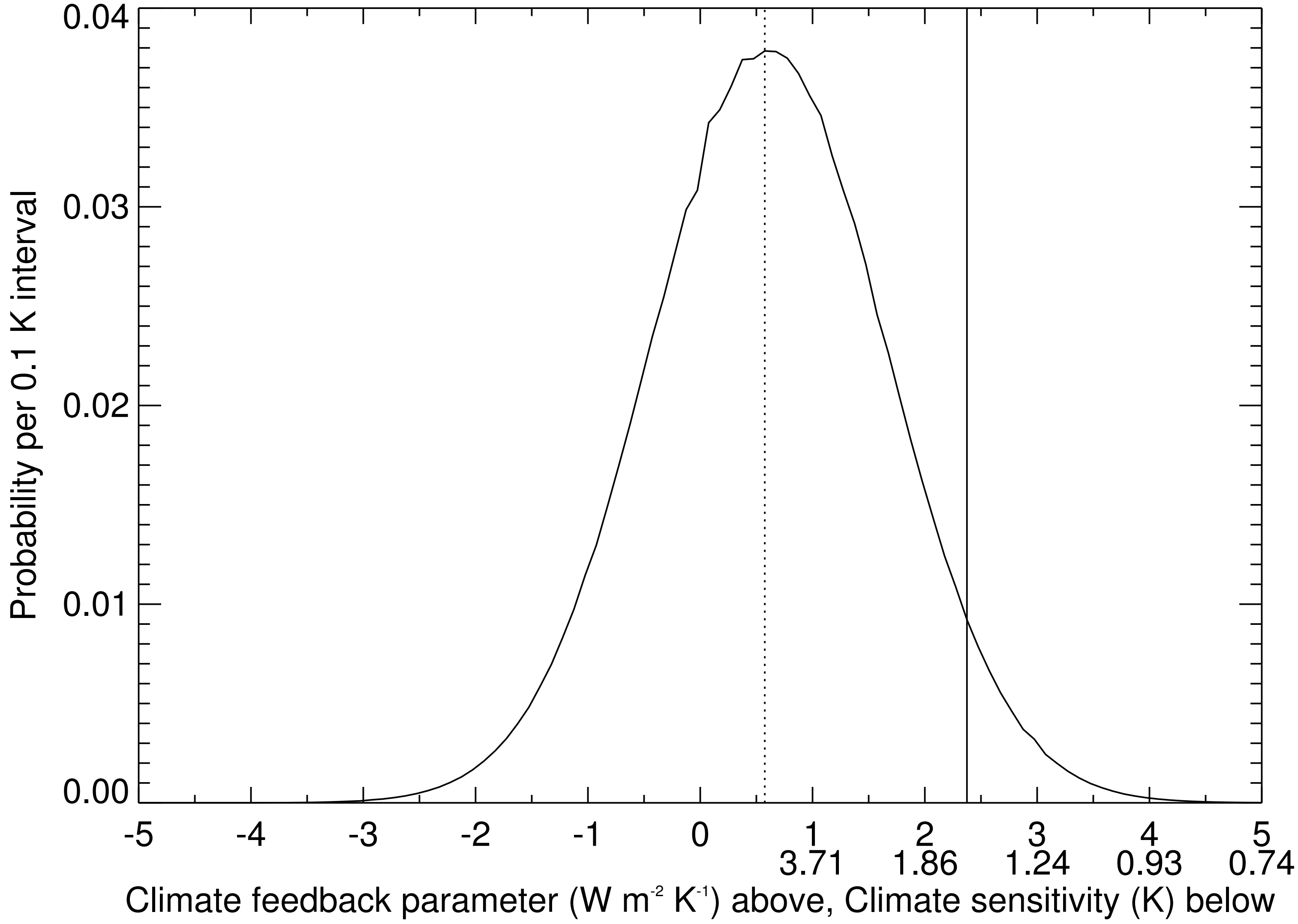
$$\delta N = \delta Q - \alpha\,\delta T \Rightarrow \alpha = \frac{\delta Q - \delta N}{\delta T}.$$

Gregory *et al.* (2002) used 1957–1994 minus 1861–1900.

Bayesian assumption of uniform priors on observables.



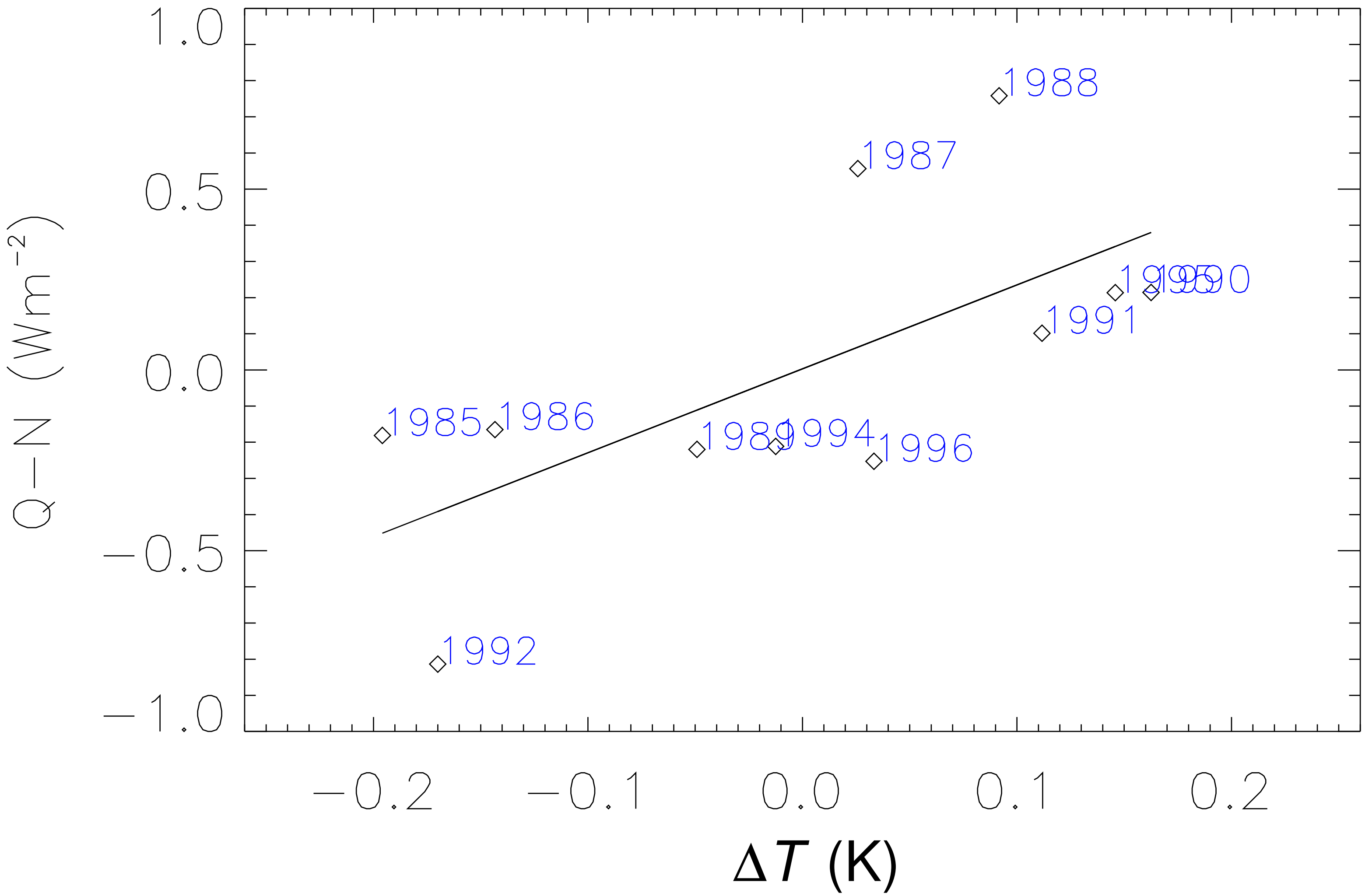
$\Delta T_{2\times}^{\mathrm{eqm}}$ goes to ∞ and beyond!



Probability per 0.1 K interval
0.00
0.01
0.02
0.03
0.04
-5
-4
-3
-2
-1
0
1
2
3
4
5
3.71
1.86
1.24
0.93
0.74
Climate feedback parameter (W m-2 K-1) above, Climate sensitivity (K) below

Recent decades without using a climate model

$Q - N = \alpha\,\Delta T$ Correlation 0.66 $\alpha = 2.32 \pm 1.30\ \mathrm{W\,m^{-2}\,K^{-1}}$

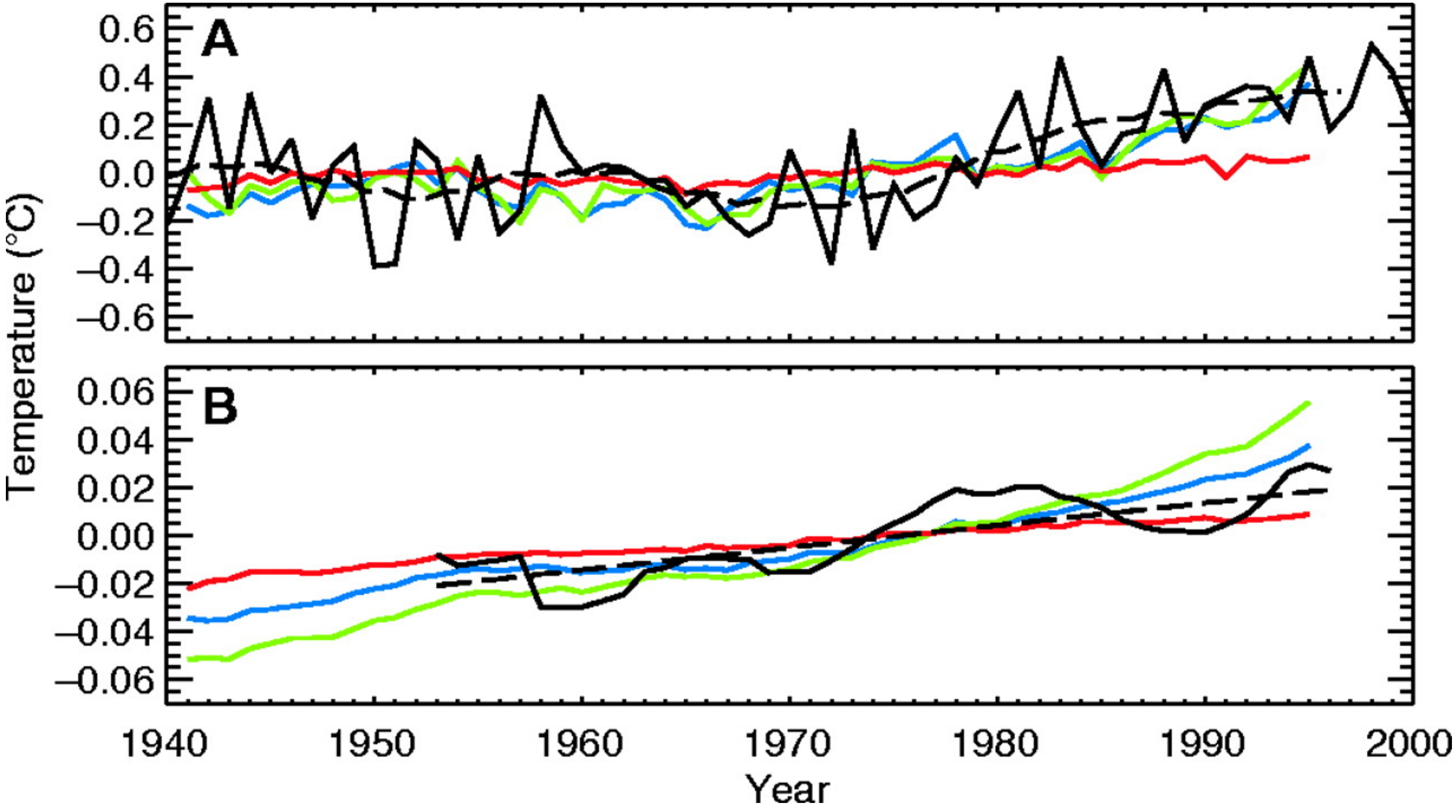


(Piers Forster, pers. comm.)

Recent decades/centuries using a climate model

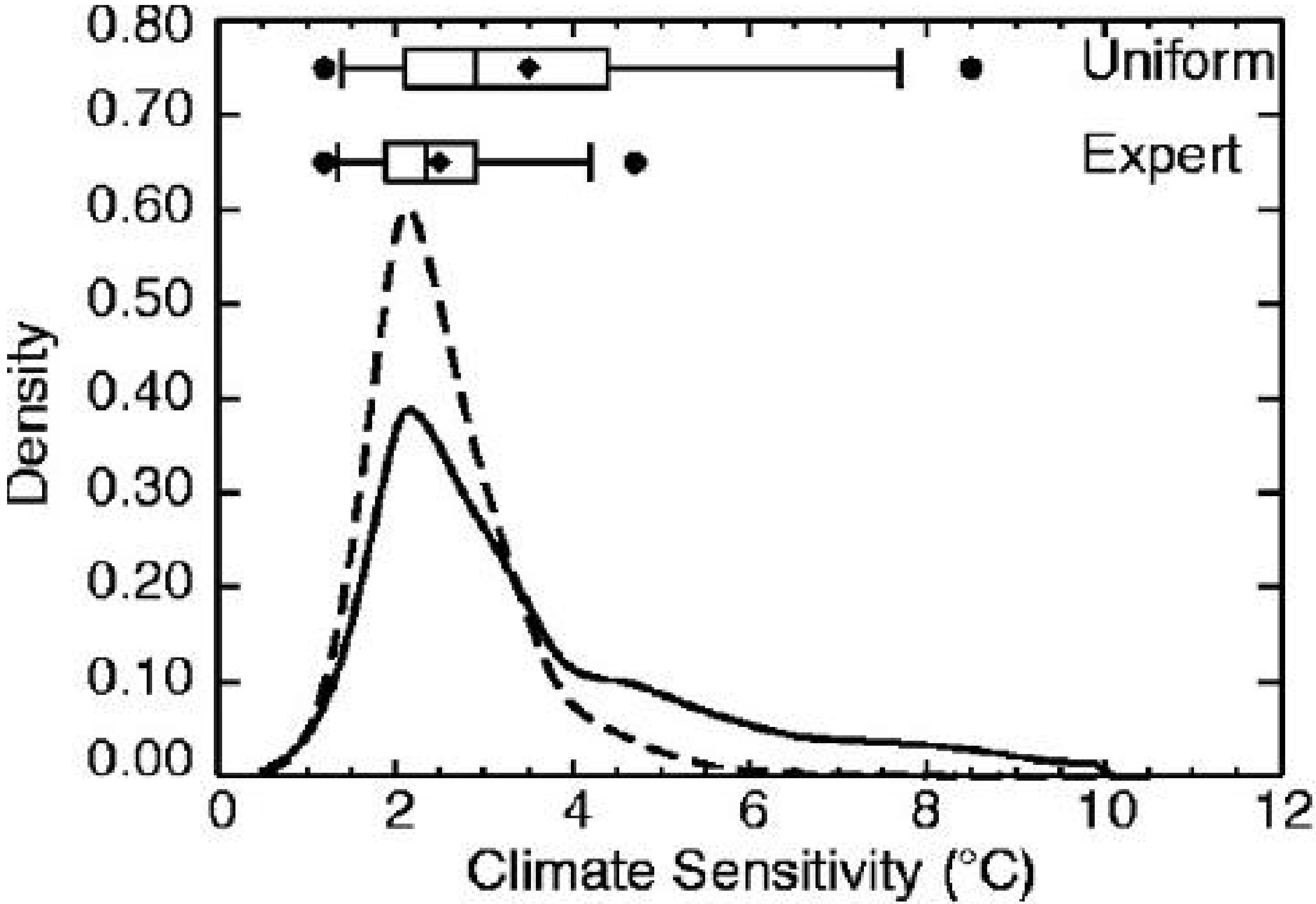
Given Q, choose α etc., use a model for ΔT etc., compare with measurements.

Can use more kinds of data $\Rightarrow$ constrains α more, but depends more on models.



Forest *et al.* (2002)

PDF of α depends on the prior



Forest *et al.* (2002)

Main points

Climate sensitivity is useful for predicting ΔT and all aspects of climate change which scale with it, in a given model or maybe the real world.

We need greater understanding of how the patterns of change are determined.

Climate sensitivity and ocean heat uptake jointly determine global average temperature change ΔT in time-dependent climate change.

Four timescales: instantaneous forcing, induced forcing, fast feedback (scales with ΔT), slow feedback (has its own timescales).

Climate sensitivity measures fast feedbacks. They can depend on the nature of the forcing, but this dependence is reduced by accounting for induced forcings.

Slow feedbacks can modify climate sensitivity. They are forcings if not interactive.

Use of observed variability and change to constrain local and global feedbacks complements evaluation of $\Delta T_{2\times}^{\mathrm{eqm}}$ using GCMs.