Climate Sensitivity and Ocean Heat Uptake

Probabilistic Estimates of Global Warming

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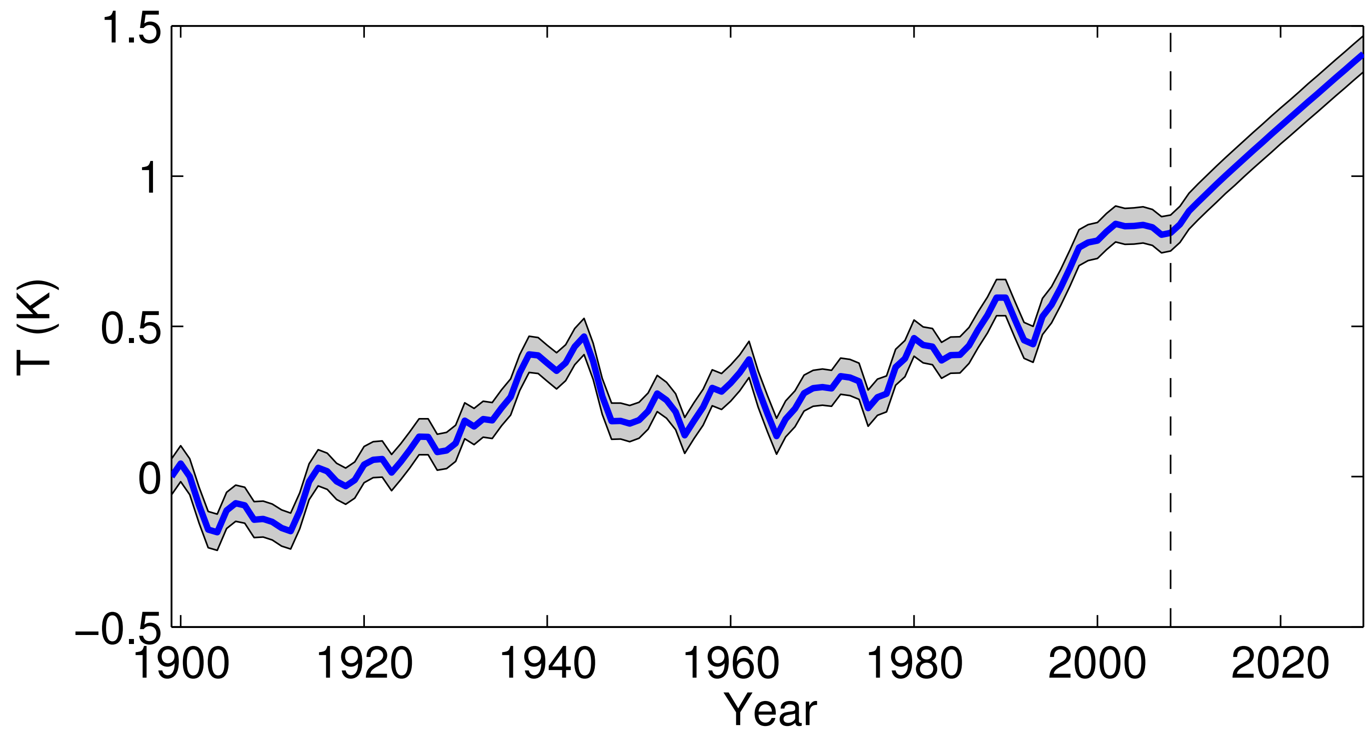
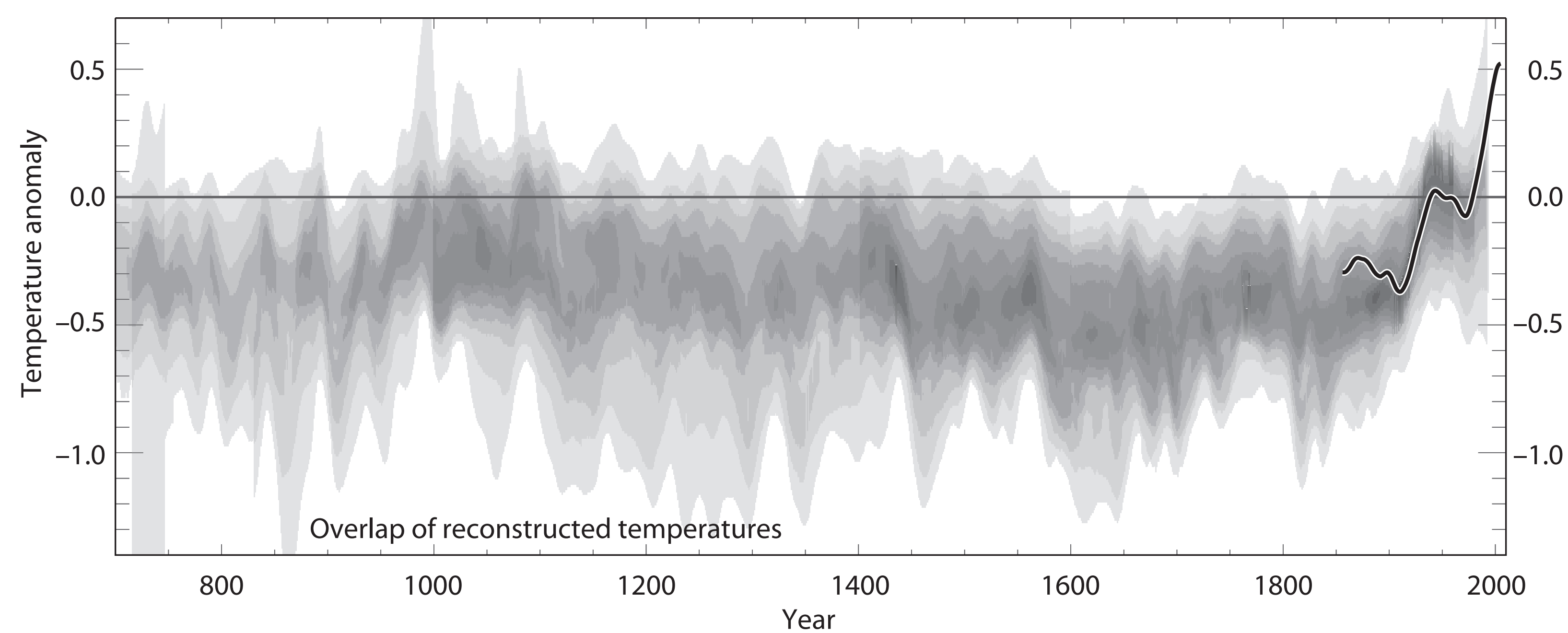
Motivation (If one is needed)

Increase of temperature

Over last century
(NASA/HadCRUT)



Over last millennium
(Mann et al)



What warming can we expect over the next 50, 100, 1000 years?

Q. How do we know global warming is not just natural variability?

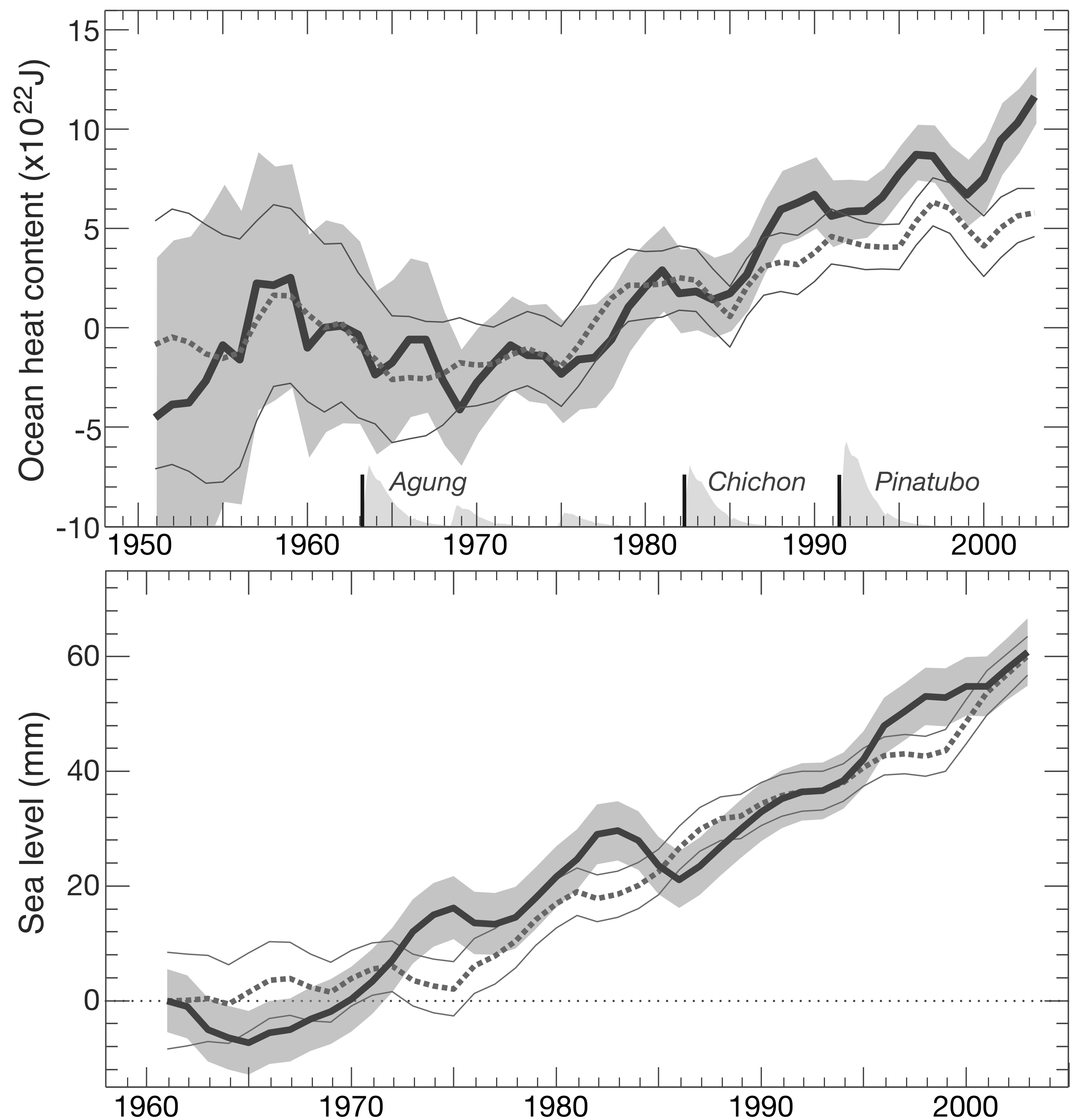
A. The Ocean

1. No known natural mechanism that is consistent with the observations.

2. In particular, record of ocean heat content. Ocean is not giving up heat to the atmosphere.

   Rather, the ocean is warming *because* it is taking up heat *from* the atmosphere.

Heat uptake in top 100 m (mixed layer) and in top 700 m (main thermocline).



Adapted from Domingues et al (2009).

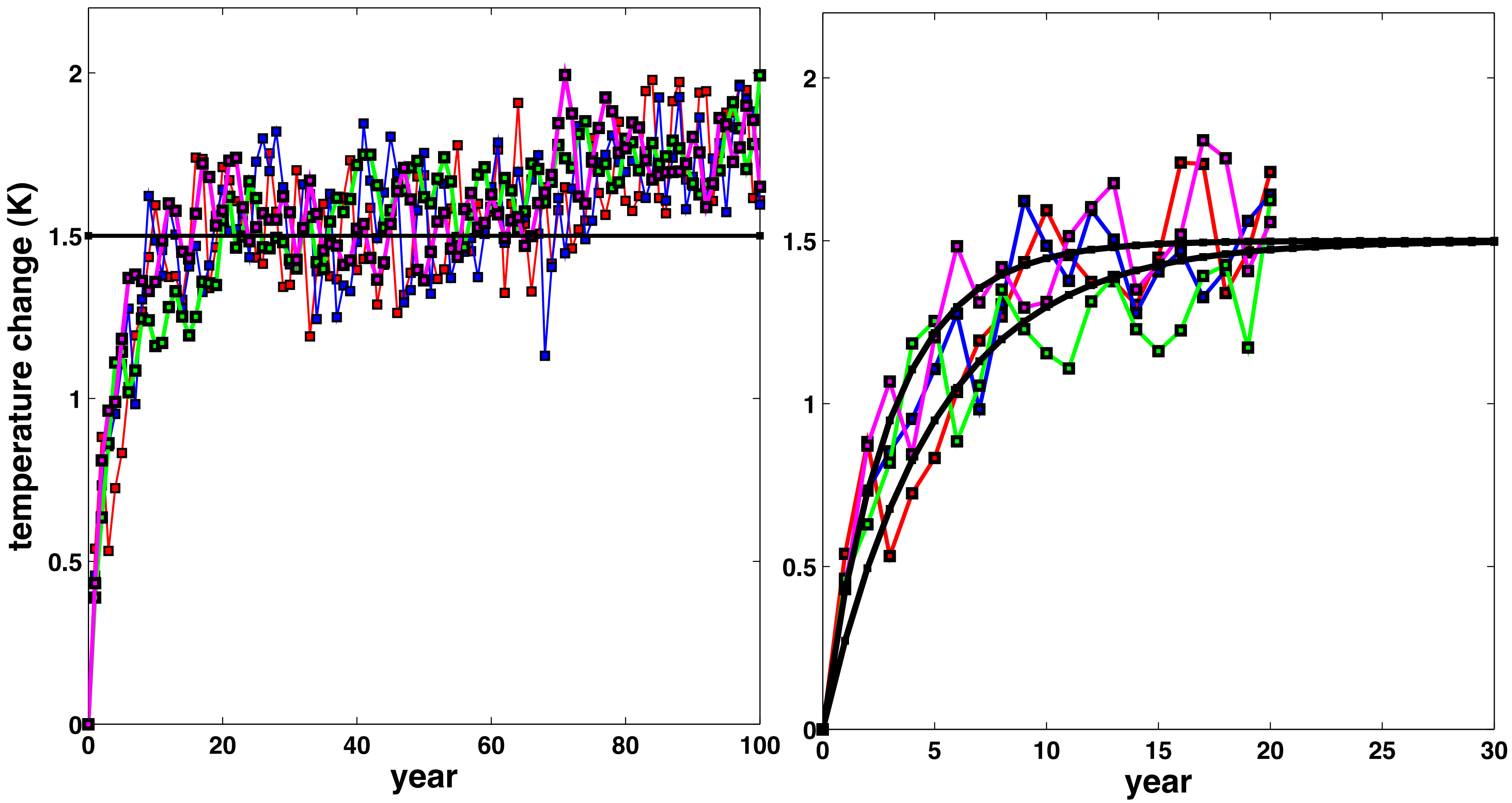
Climate Sensitivity

Suppose we double CO_2 levels in the atmosphere suddenly. (i) What is the amplitude of the response? (ii) What are the timescales of response? (iii) What is the role of ocean in this?

Specifically, suppose we suddenly double the CO_2 levels:

1. How long does it take for the surface temperature to increase to, say, 60% of its final value? Is it:
   (a) 10 years ?
   (b) 100 years ?
   (c) 1000 years ?
2. Suppose we suddenly scrub the atmosphere of all the excess CO_2. How long does it take for the surface value to fall back to near-preindustrial levels?
3. Can we explain the above?
4. Can we predict what the rise in surface temperature will be for a doubling of CO_2, given the observed rate of increase so far? ('climate sensitivity')
   - After 100 years
   - After 1000 years

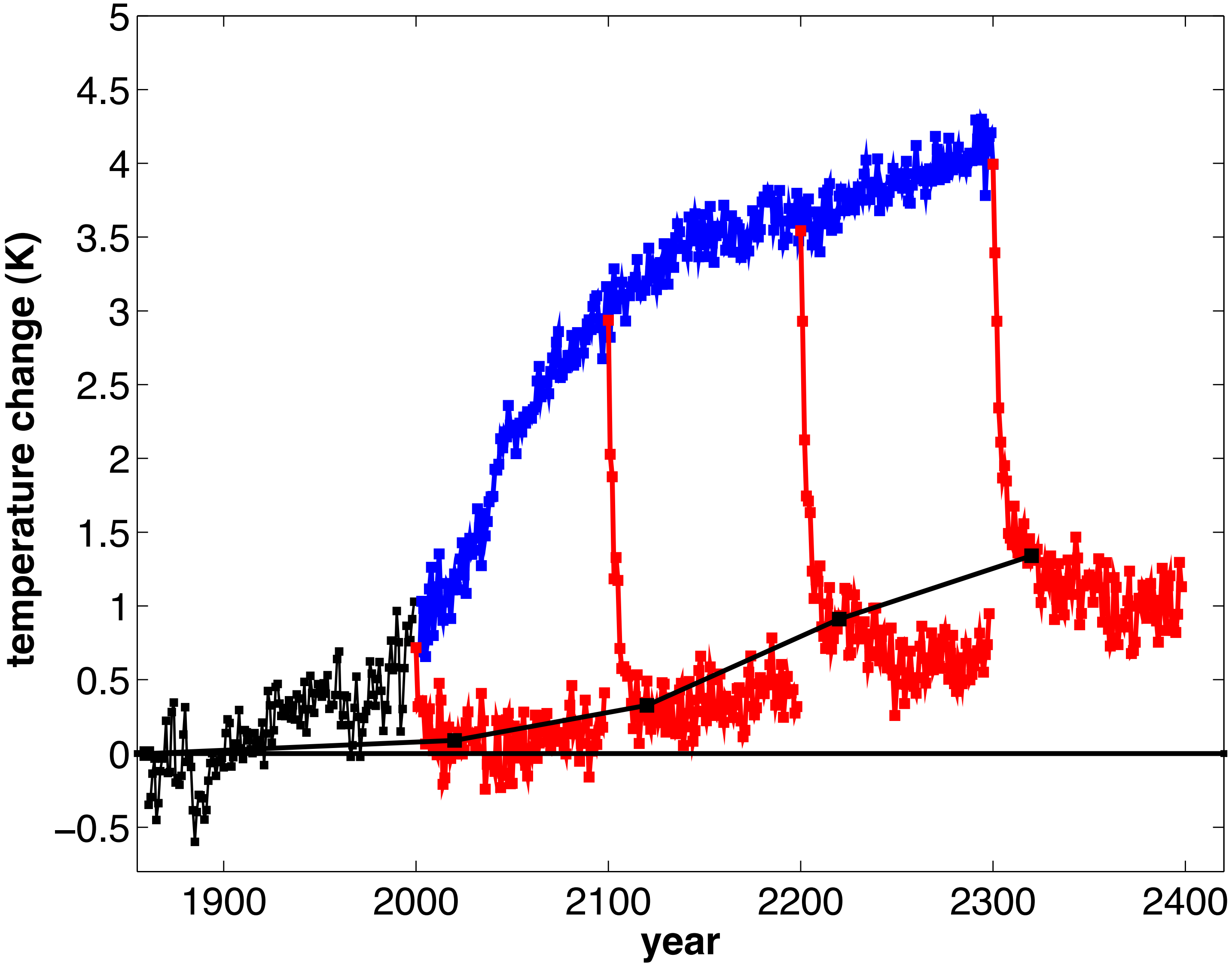
Response of a GCM to instantaneous doubling



- An ensemble of four integrations with GFDL climate model.
- Rapid increase to a quasi-equilibrium in about 10 years, much slower growth after that.

But we know that the ocean will take centuries to equilibrate, from simple calorific considerations. The MOC also takes hundreds of years to fully ventilate the deep ocean.

Suggests that the ocean has a fast response as well as a (or many) slow responses. The fast response is determined by the time taken by just the upper ocean (the mixed layer) to equilibrate.



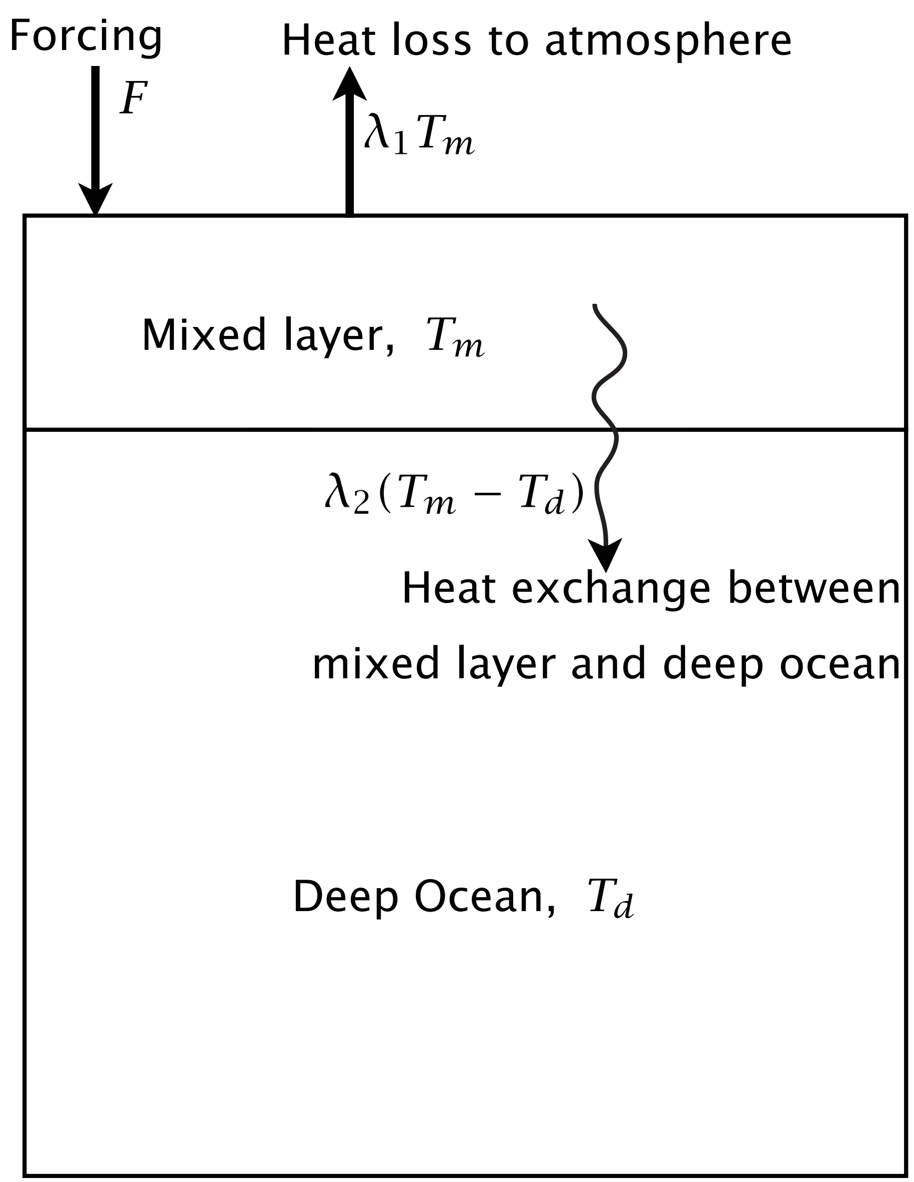
Extract CO_2 from model atmosphere instantaneously.

Model quickly reverts to near pre-global-warming temperatures.

A fast component and slow component (a 'recalcitrant' component) to climate system.

Slow times are determined by the ocean.

A two-box model of the climate system



$$C_m \frac{\mathrm{d}T_m}{\mathrm{d}t} = F - \lambda_1 T_m - \lambda_2 (T_m - T_d),$$
$$C_d \frac{\mathrm{d}T_d}{\mathrm{d}t} = \lambda_2 (T_m - T_d).$$

On decadal timescales $T_d = 0$ so that:

$$C_m \frac{\mathrm{d}T_m}{\mathrm{d}t} = F - \lambda T_m.$$

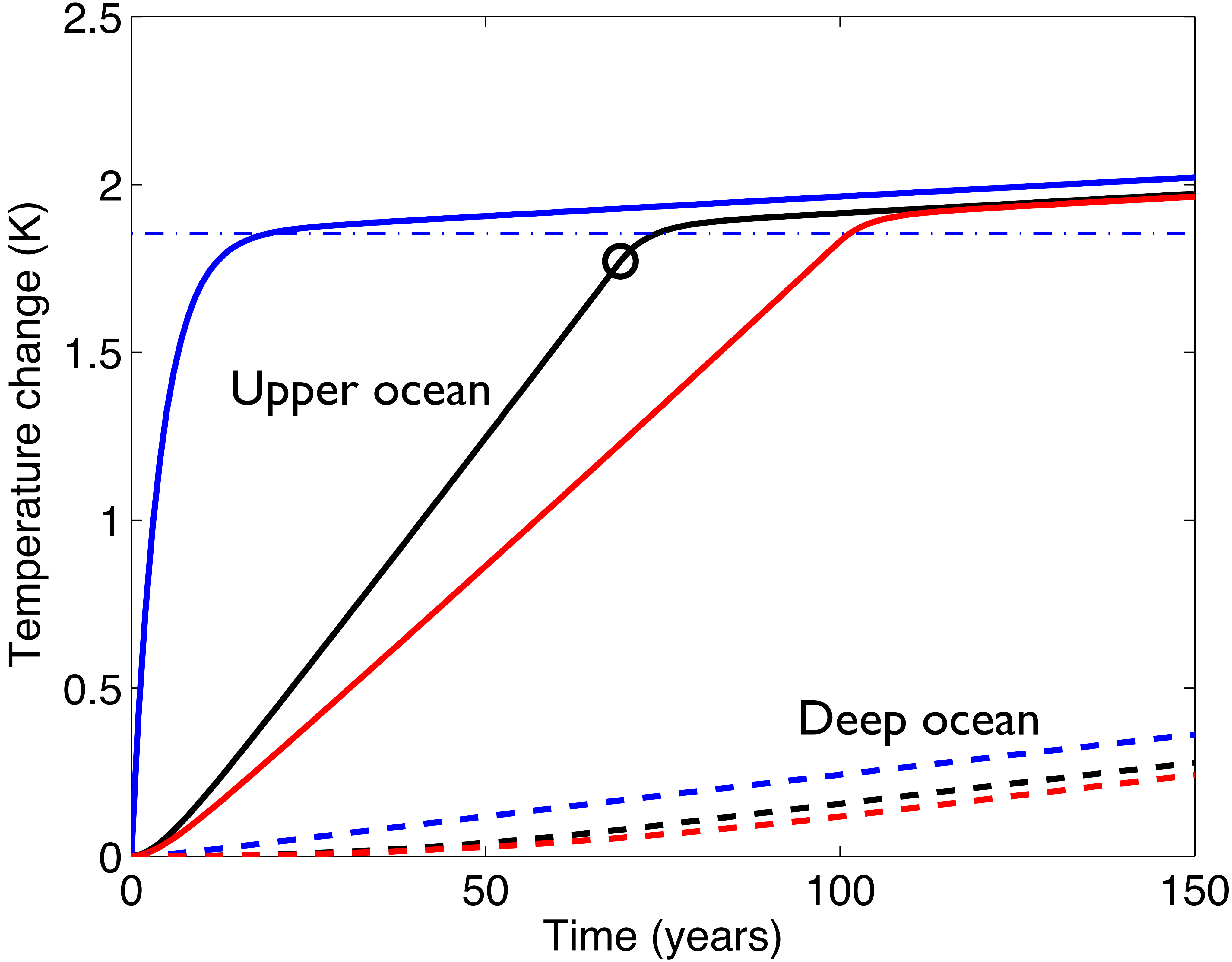
where $\lambda = \lambda_1 + \lambda_2$.
N.B. For instantaneous forcing change, solution is an exponential:

$$T_m = \frac{F}{\lambda}\left(1 - \mathrm{e}^{-t\lambda/C_m}\right).$$

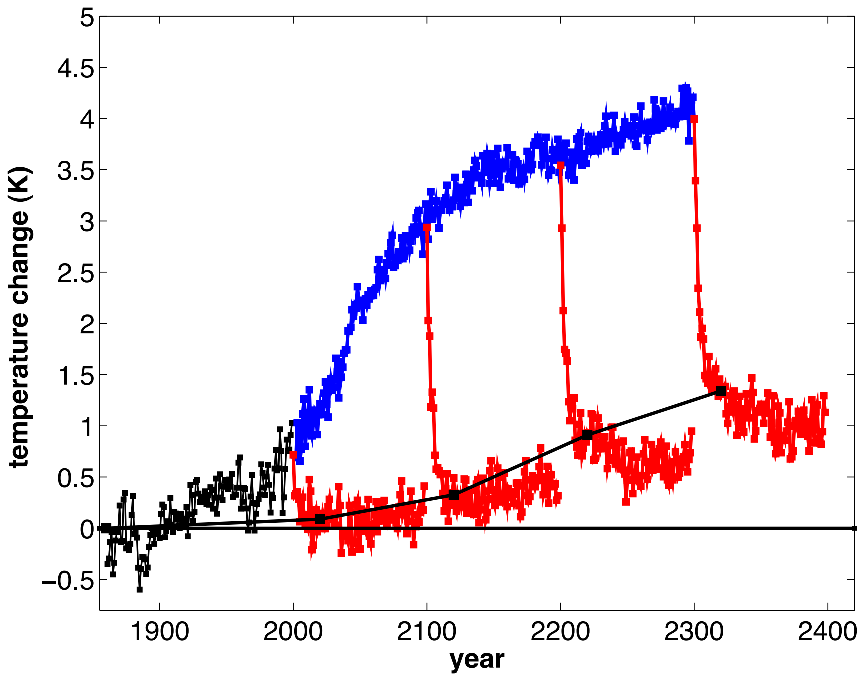
In quasi-equilibrium, $T_m = F/\lambda$. In true equilibrium, $T_m = F/\lambda_1$.

If we can determine F and λ from observations, we have an estimate of the climate sensitivity.

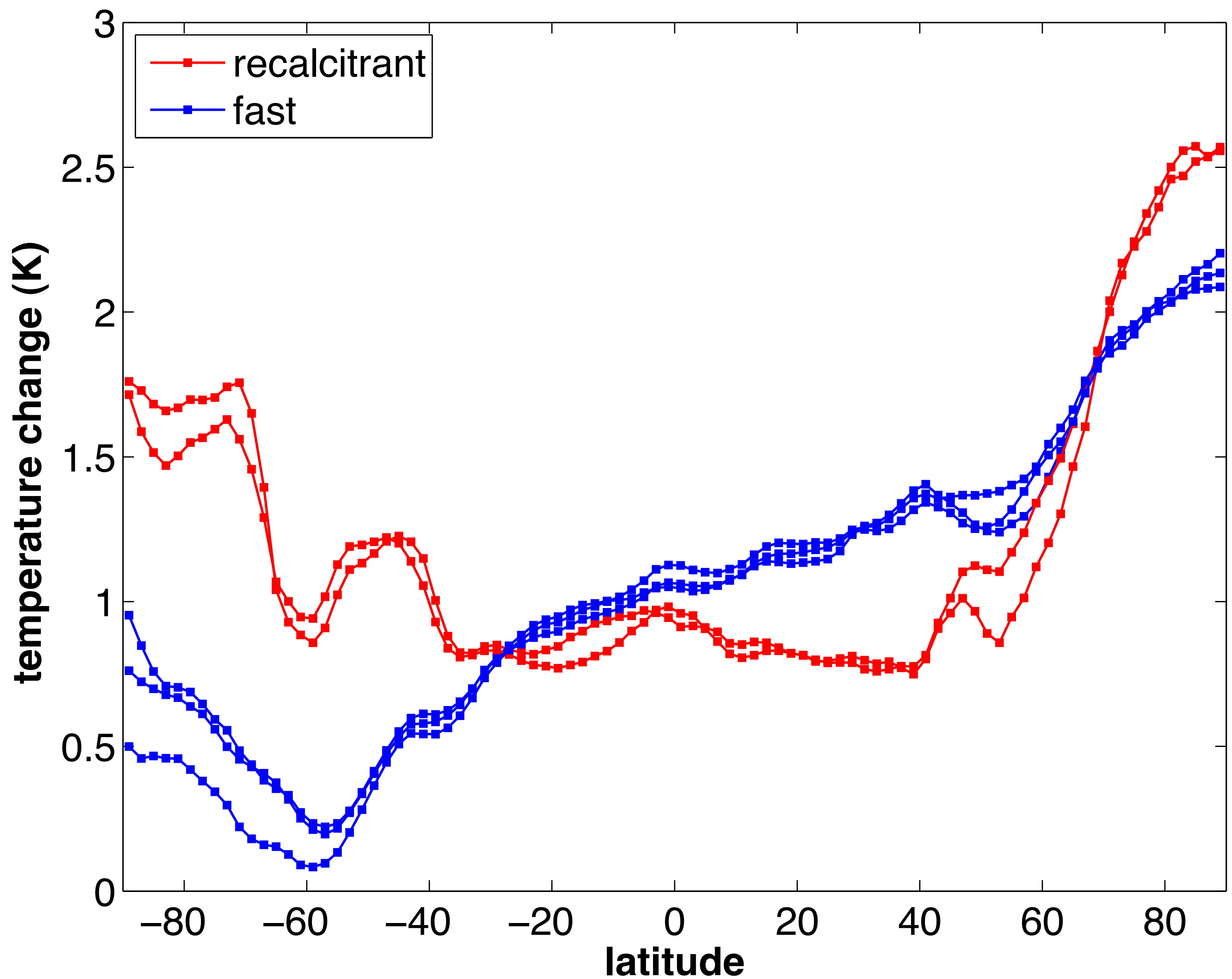
Response in a two-box model



Extracting the fast and slow components of climate change



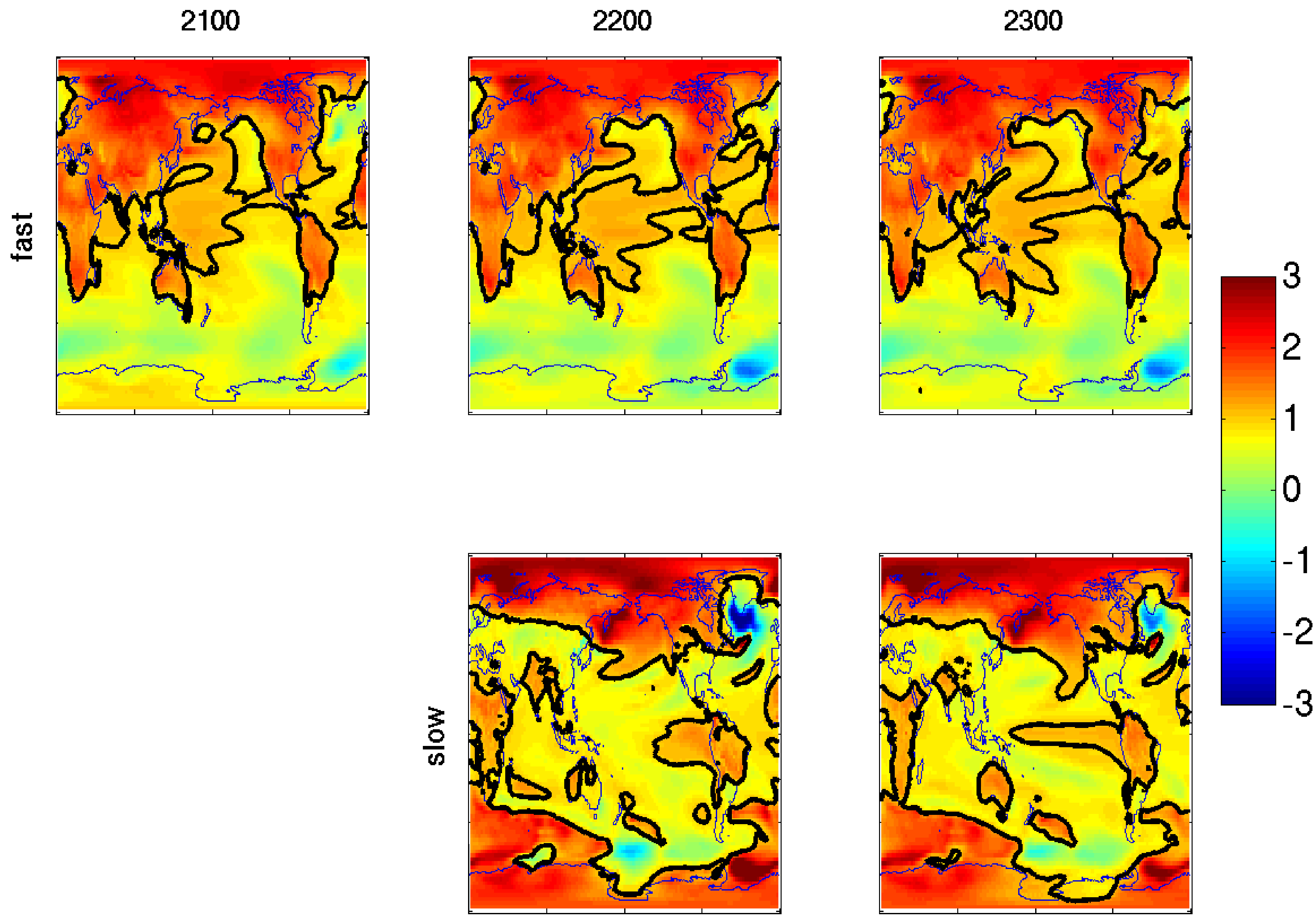
1. Slow or 'recalcitrant' component is the state after the CO_2 scrubbing. It is a surface manifestation of deep ocean warming.
2. Fast component is the difference between total warming and slow warming.



- Polar amplification in slow component in north and south.
- Polar amplification only in north in fast component.

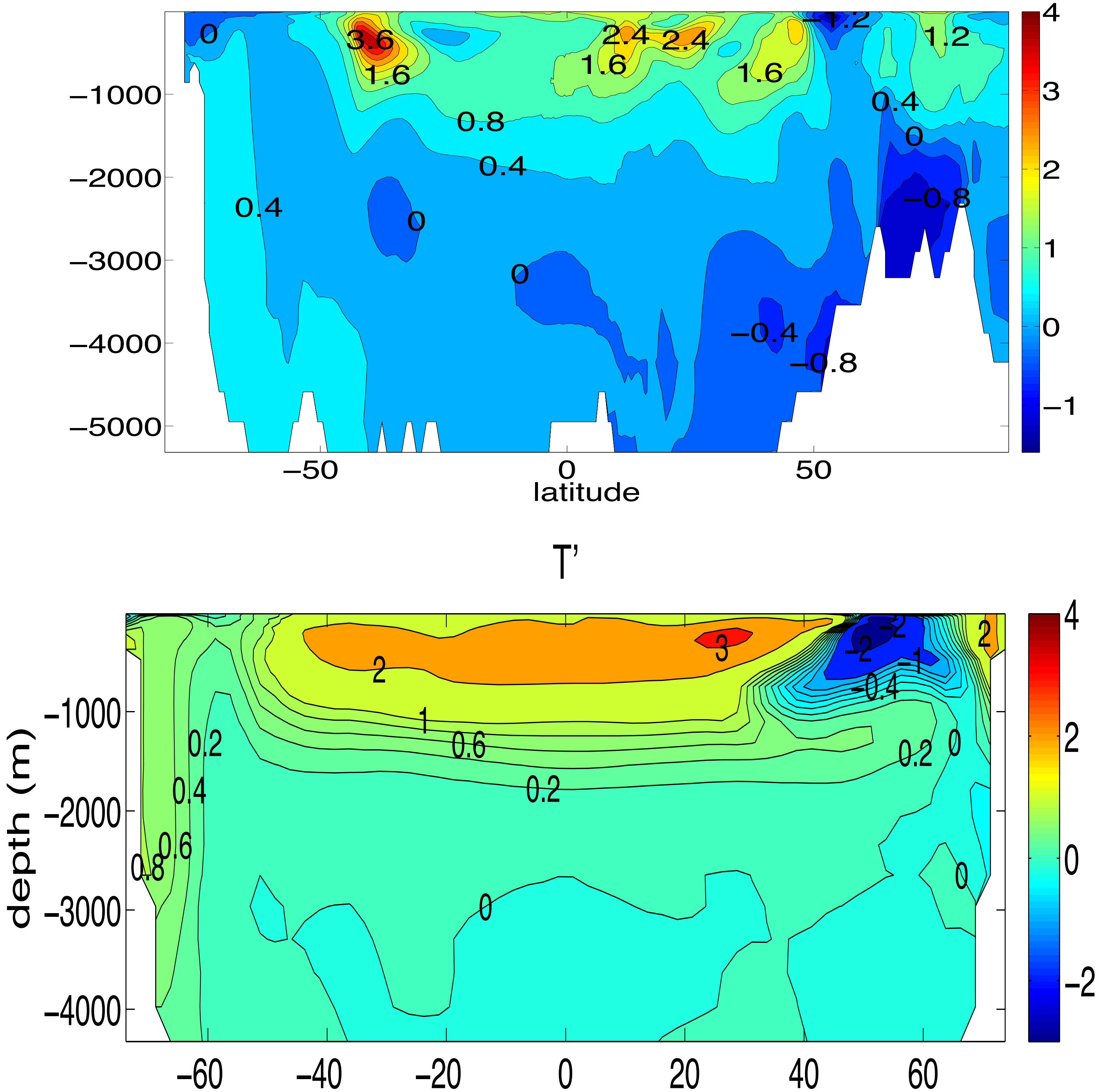
Spatial variation of components

Fast and slow response of the warming (surface air temperature), normalized to unity.



Note difference in Southern Ocean and North Atlantic.

Heat uptake in a GCM



Southern Ocean has a higher heat capacity than elsewhere, so in the short term (decades) the temperature increase in inhibited.

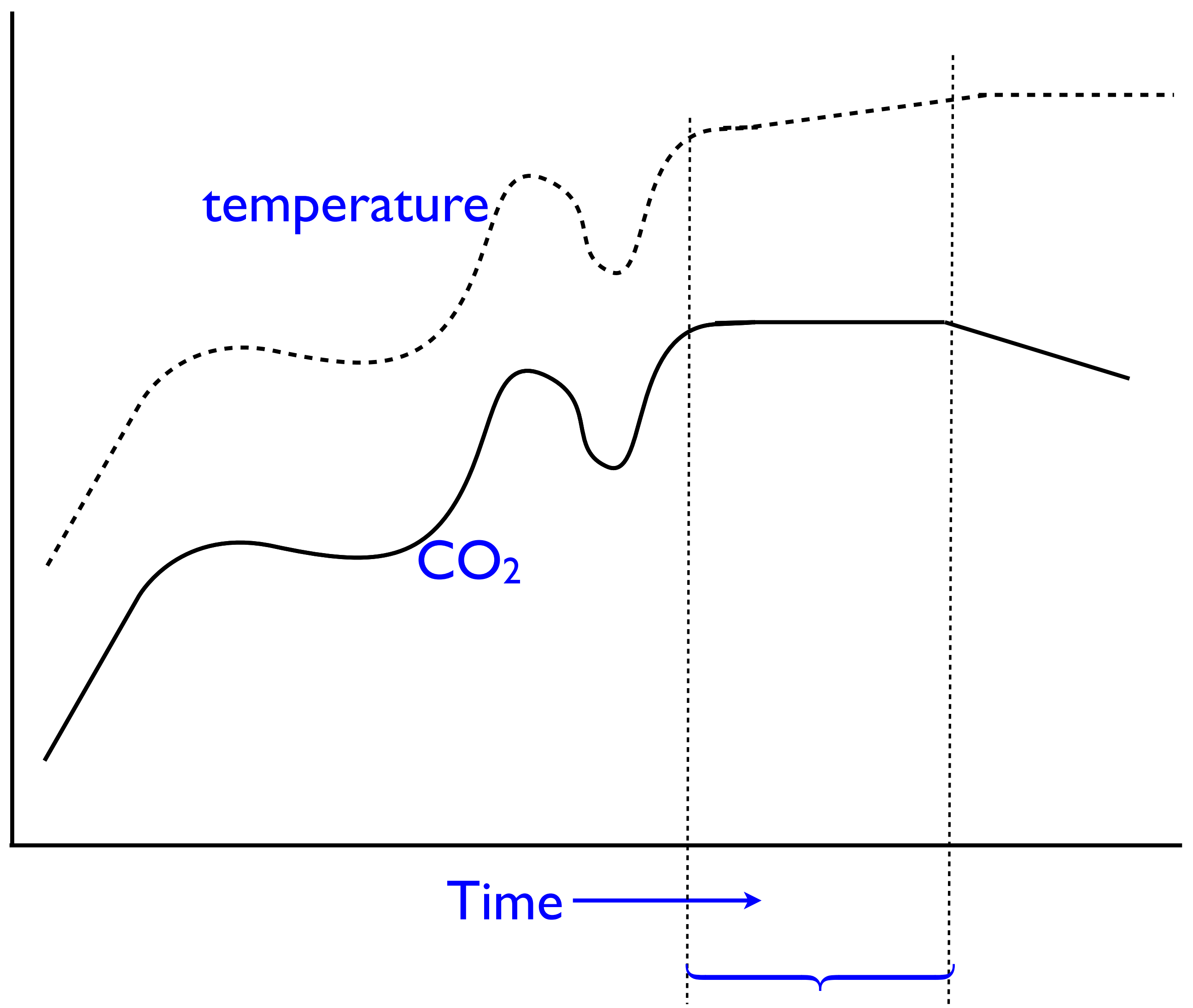
On the long term the polar amplification applies in both hemispheres.

Lessons from the GCM

1. There are distinct fast and slow components to global warming. Probably there are many slow components, but they are separated from the fast component.
2. Slow component is primarily associated with ocean. (Also land ice and other 'very slow' components. Another day.)
3. The warming we are likely to see over the coming century is the 'fast' warming. The historical record of the warming over the past century also reflects the fast response.
4. We may be able to estimate the future warming by extrapolation from the past, for a given forcing, assuming there are no unpredictable nonlinear responses.

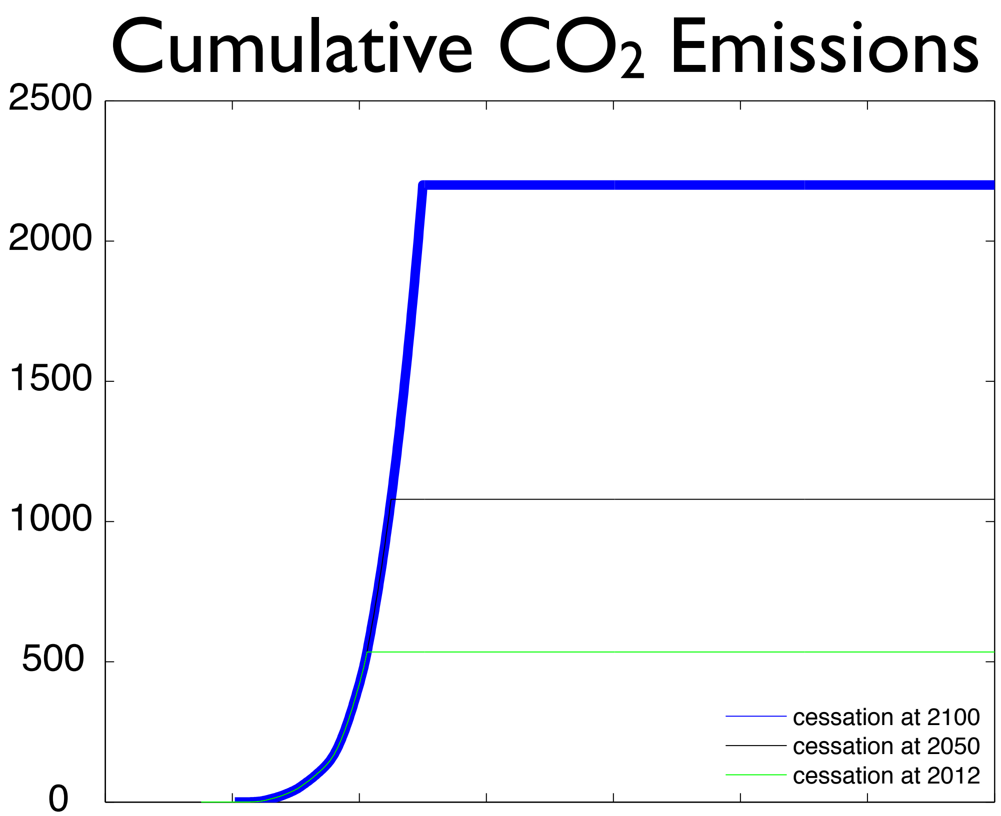
Carbon dioxide and temperature

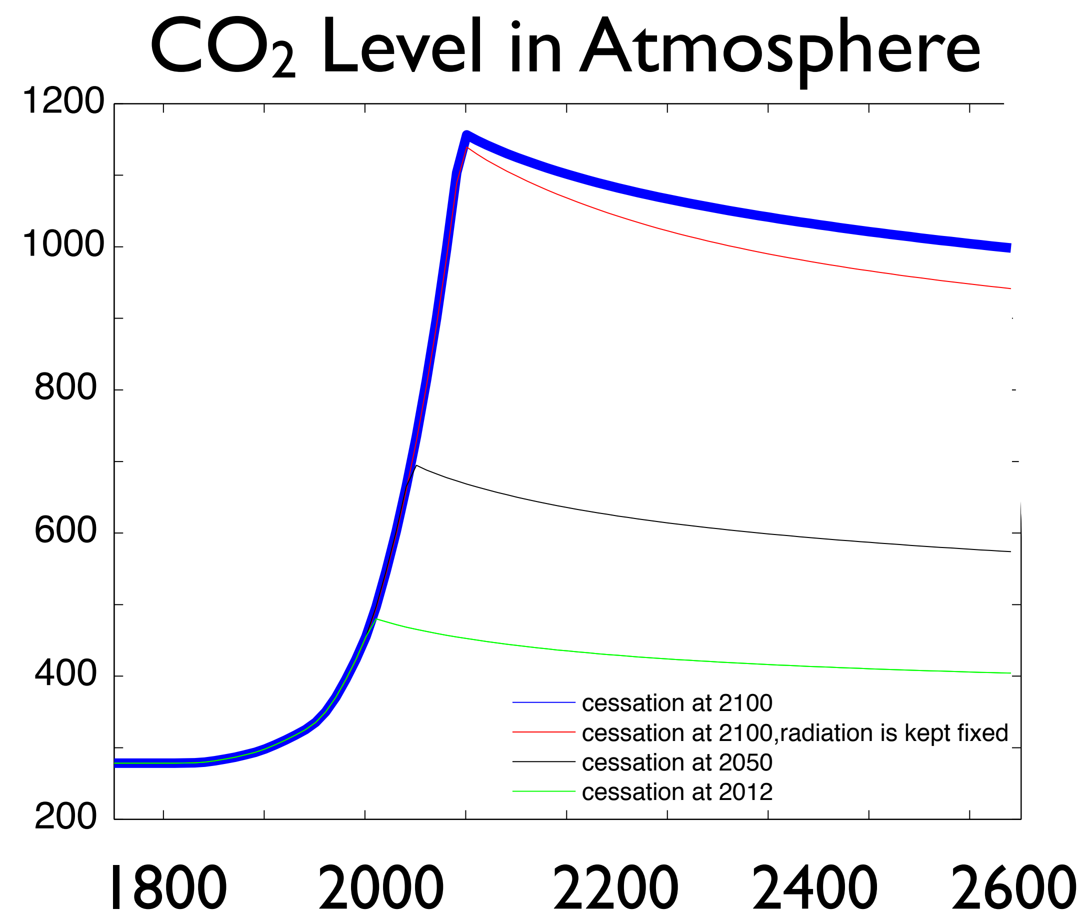
Illustrative scenario

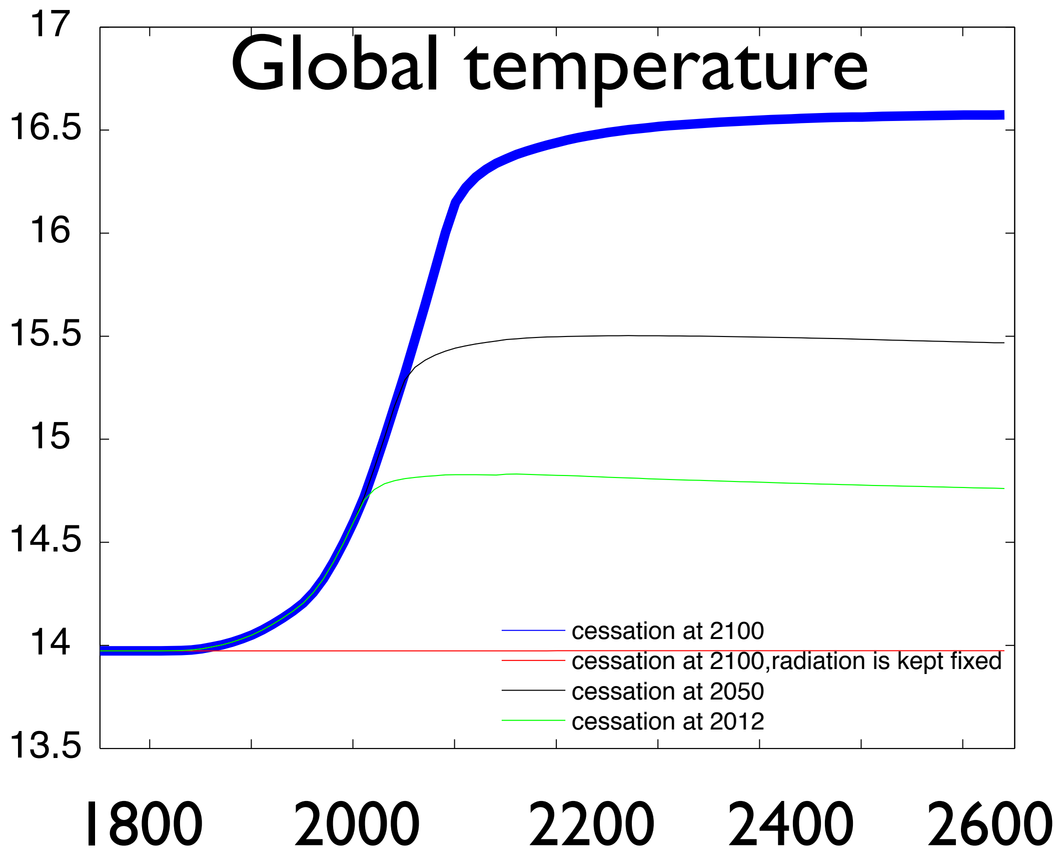


CO_2 levels fixed, temperature rises (‘committed warming’), from ‘transient’ value to ‘equilibrium’ value.

CO2 and Temperature, an Example using a GCM w/carbon cycle





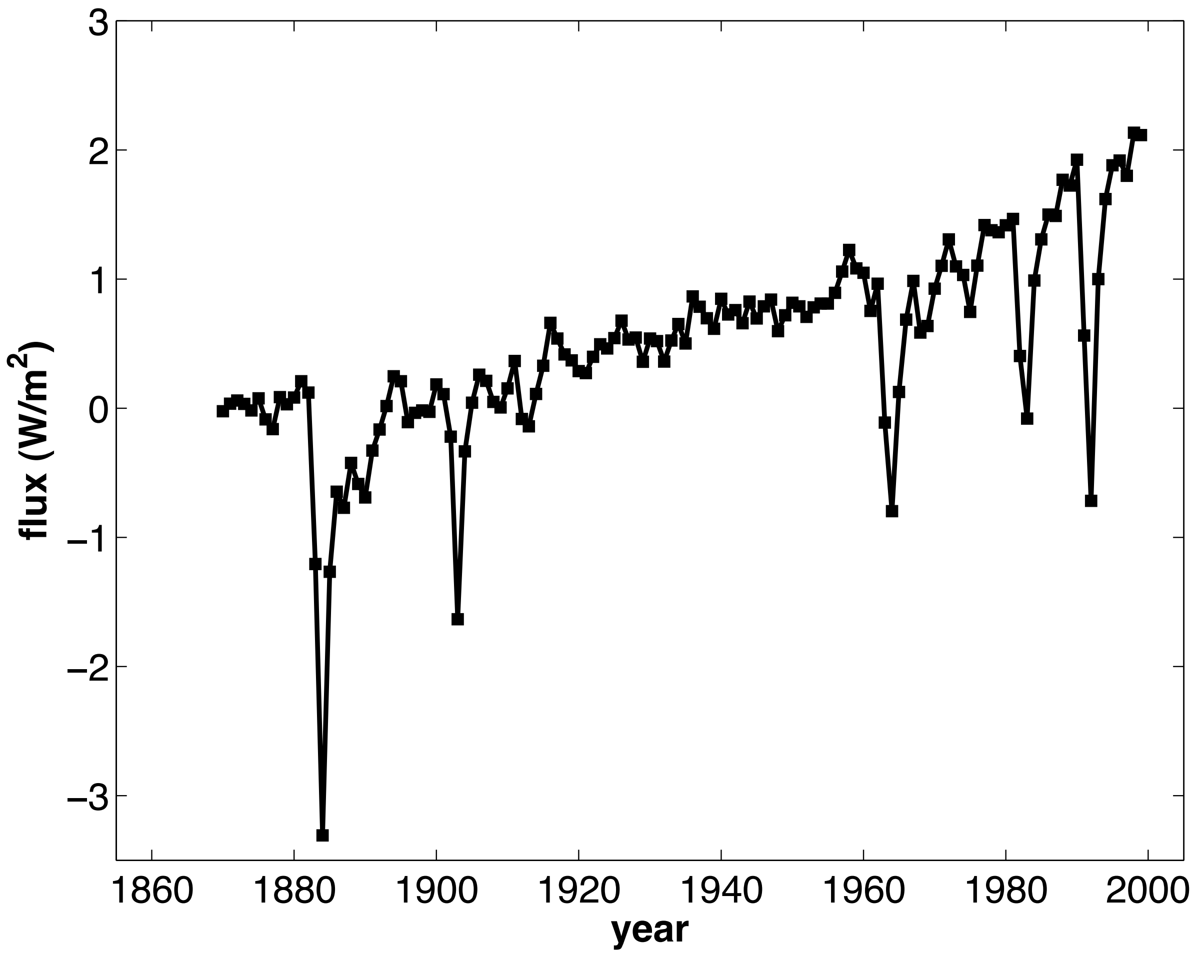


Temperature stays constant when emissions cease.

There is really no 'committed warming'...

Forcing and response

What forcing produces the response in 20 century?



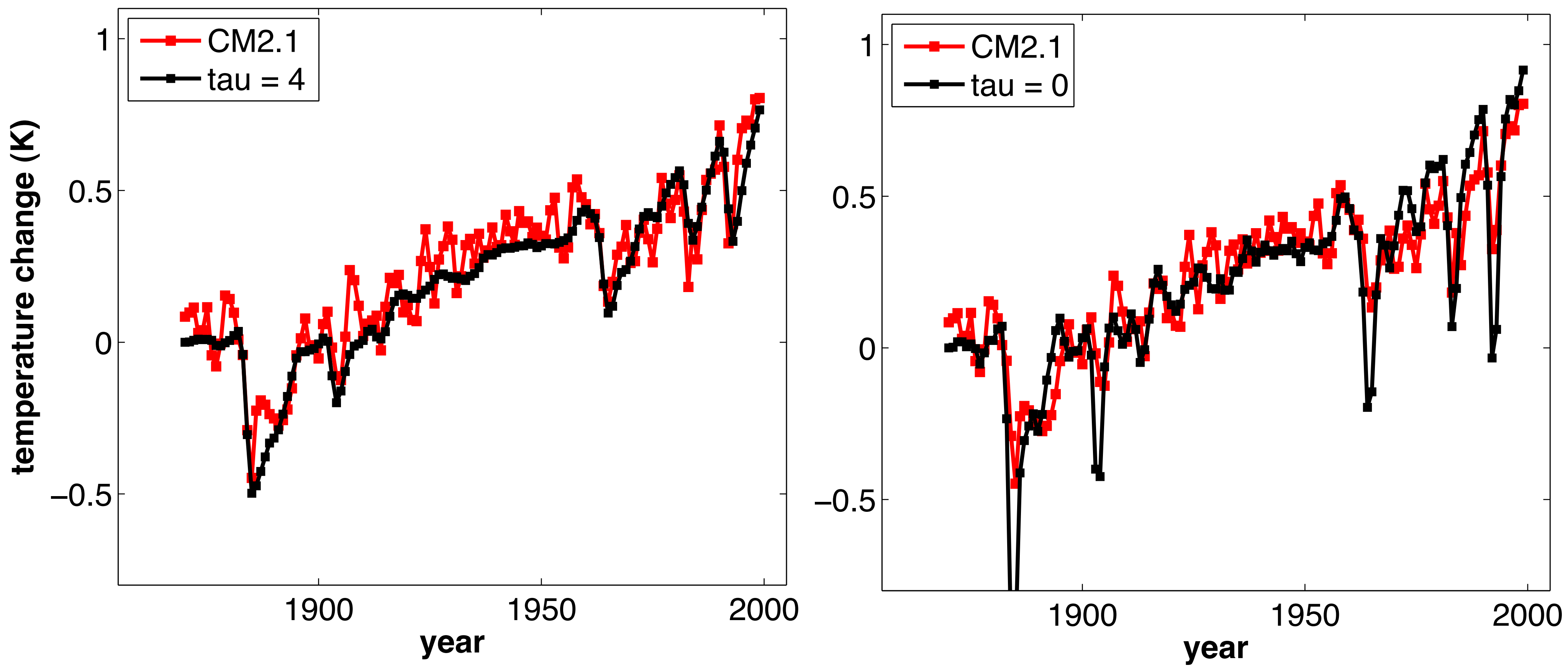
Volcanoes: Krakatoa (1883), Pelee & Santa Maria (1902), Agung (1963), El Chichon (1982), Pinatubo (1991).

- The radiative forcing over the 20th century, as computed by the GFDL GCM. Mainly caused by an increase in CO_2, plus aerosols, plus volcanoes.
- Compute this by increasing the greenhouse gases (and other forcings) while keeping surface temperatures fixed (following Hansen).
  - Fixing temperatures while increasing greenhouse gas levels gives the *radiative* effect of the greenhouse gases, without any 'feedbacks' from increased water vapor etc.

Fit the 20th Century Response to a Simple Model

$$\boxed{C\frac{\mathrm{d}T}{\mathrm{d}t} = F - \lambda T}$$

Model temperatures (red) and fit to a simple one-box model (black).



$\frac{C}{\lambda} = 4$ years

$\frac{C}{\lambda} = 0$ years

Observational estimates of climate sensitivity.

Procedure

1. Estimate the radiative forcing F over the last century.
2. Use the semi-empirical model:

$$C\frac{\mathrm{d}T}{\mathrm{d}t} = F - \lambda T + \dot{W}.$$

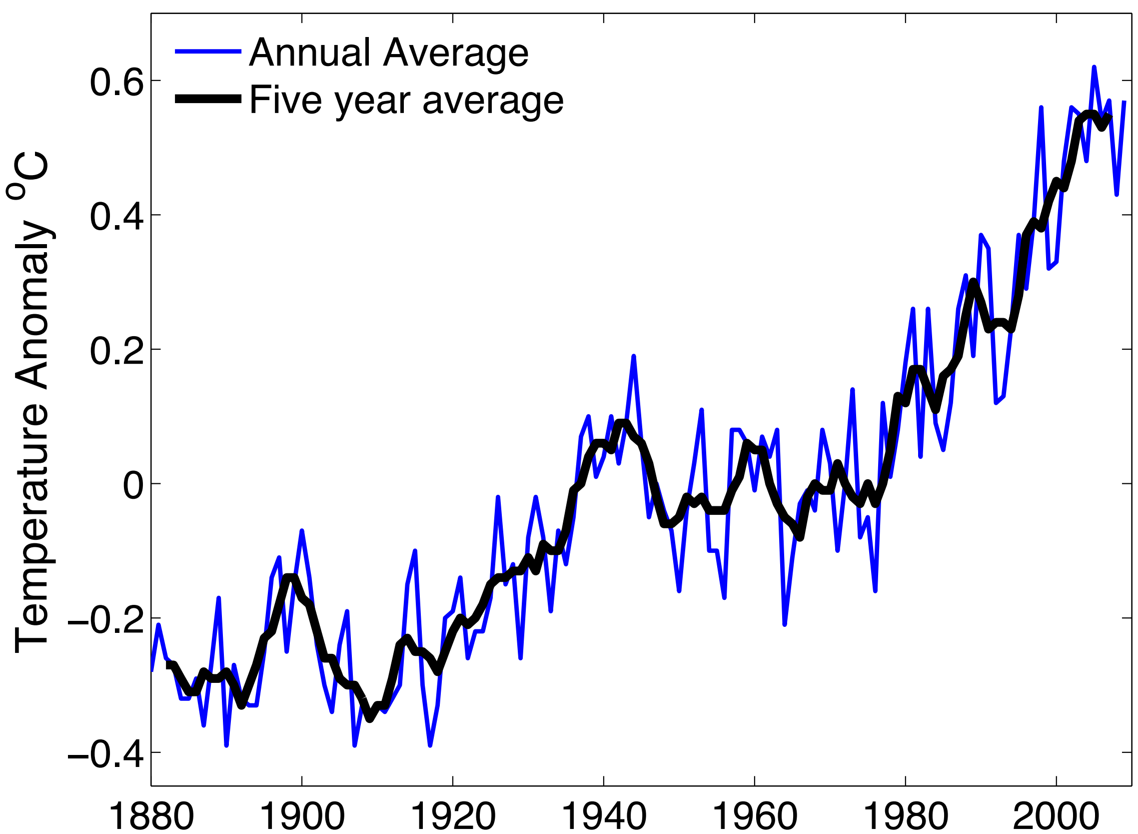
   where $\dot{W}$ represents natural variability. Fit the timeseries of F and T to obtain λ.
3. Use a Kalman filter to obtain *probabilistic estimates* of climate sensitivity.
   - Given probabilities of F, T, and $\dot{W}$, and given some prior estimates, Kalman filter essentially updates a PDF (Gaussian) of the model temperature and λ ('Bayes plus best least-squares fitting.').
   - Same as data-assimilation, except that the parameter λ (and perhaps F) is regarded as an unknown as well as temperature itself.
   - Obtain a PDF of climate sensitivity, λ (and of forcing, F).
   - For a commonly-used number,

$$T_{tr} = \frac{F_{2\mathrm{CO}_2}}{\lambda}$$

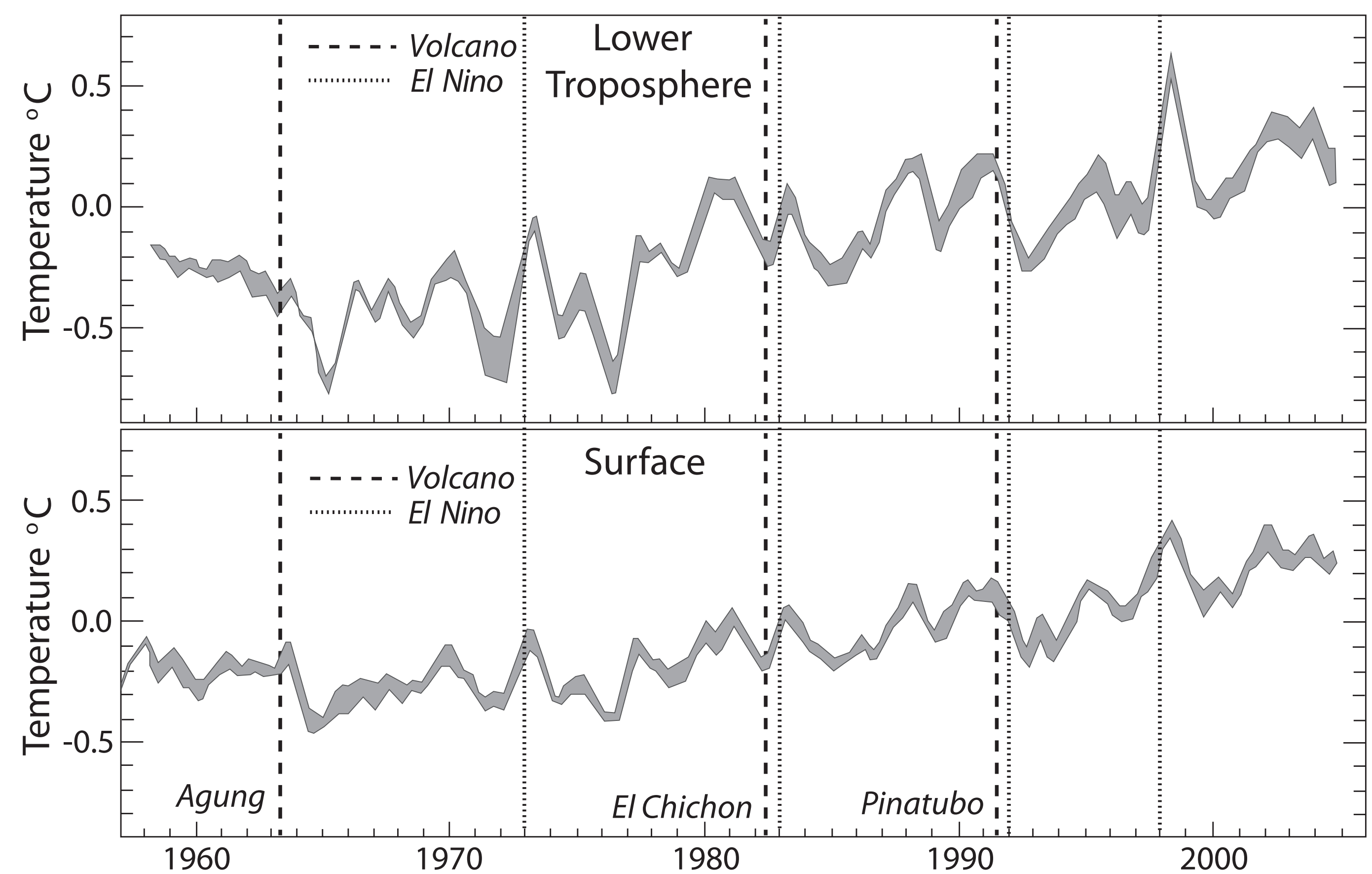
     where $F_{2\mathrm{CO}_2} \approx 3.5\,\mathrm{W\,m^{-2}}$.

20th Century Global Warming

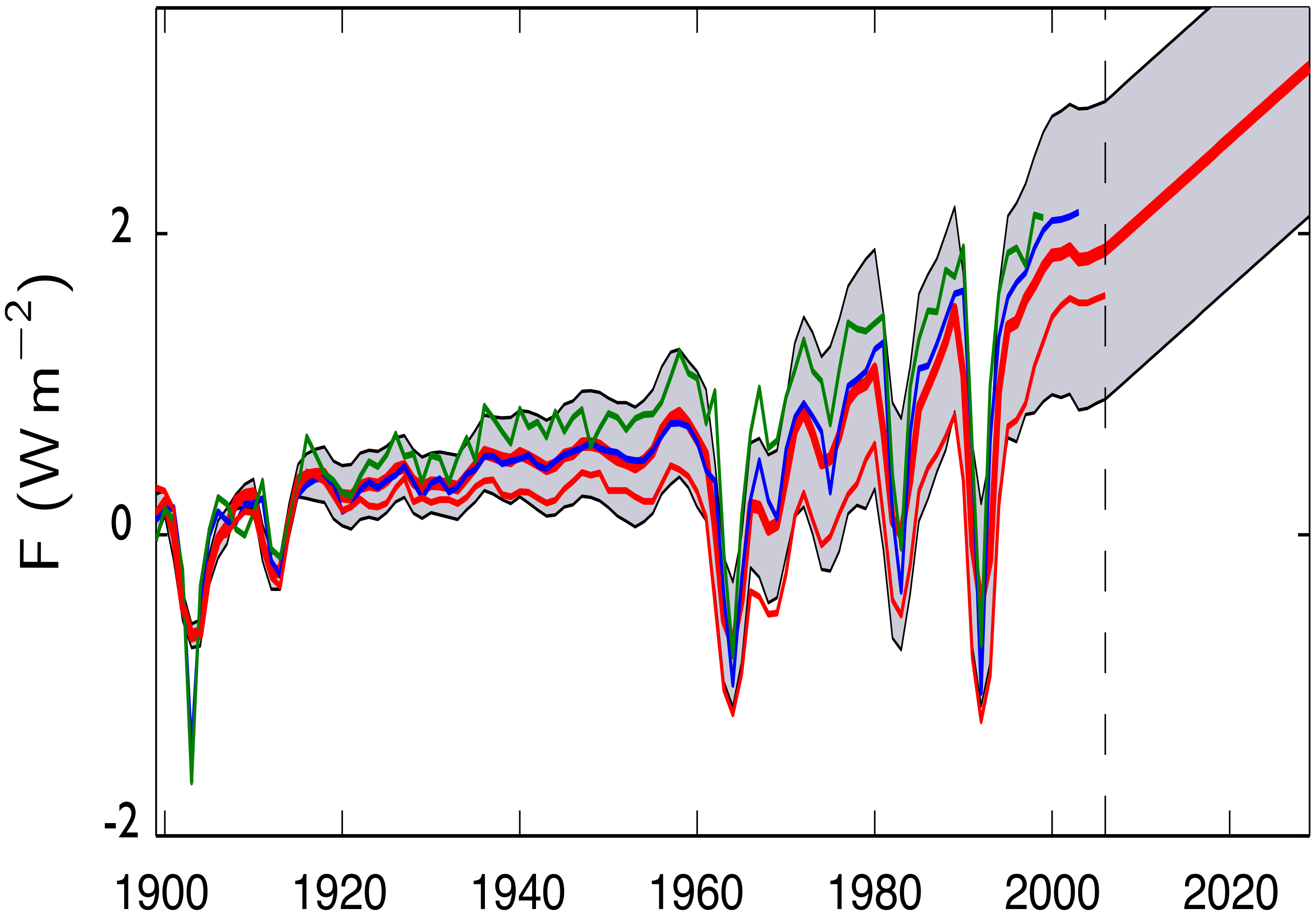
Global average surface temperature, year-by-year and 5-year running mean:



Last 50 years, including satellite measurements



What Forces Climate Change



Three different calculations of the total forcing:

(i) GISS calculation (original);

(ii) GFDL (local)

(iii) Forster and Gregory (2008) (most comprehensive).

Differences largely in aerosol formulation.

Uncertainty of Forcing

GHG gases

Approximately

$$F_{CO_2} \approx 5.5 \log\left[CO_2/CO_2(\mathrm{ref})\right]$$

So CO_2 doubling is about 3.5 W/m^2.

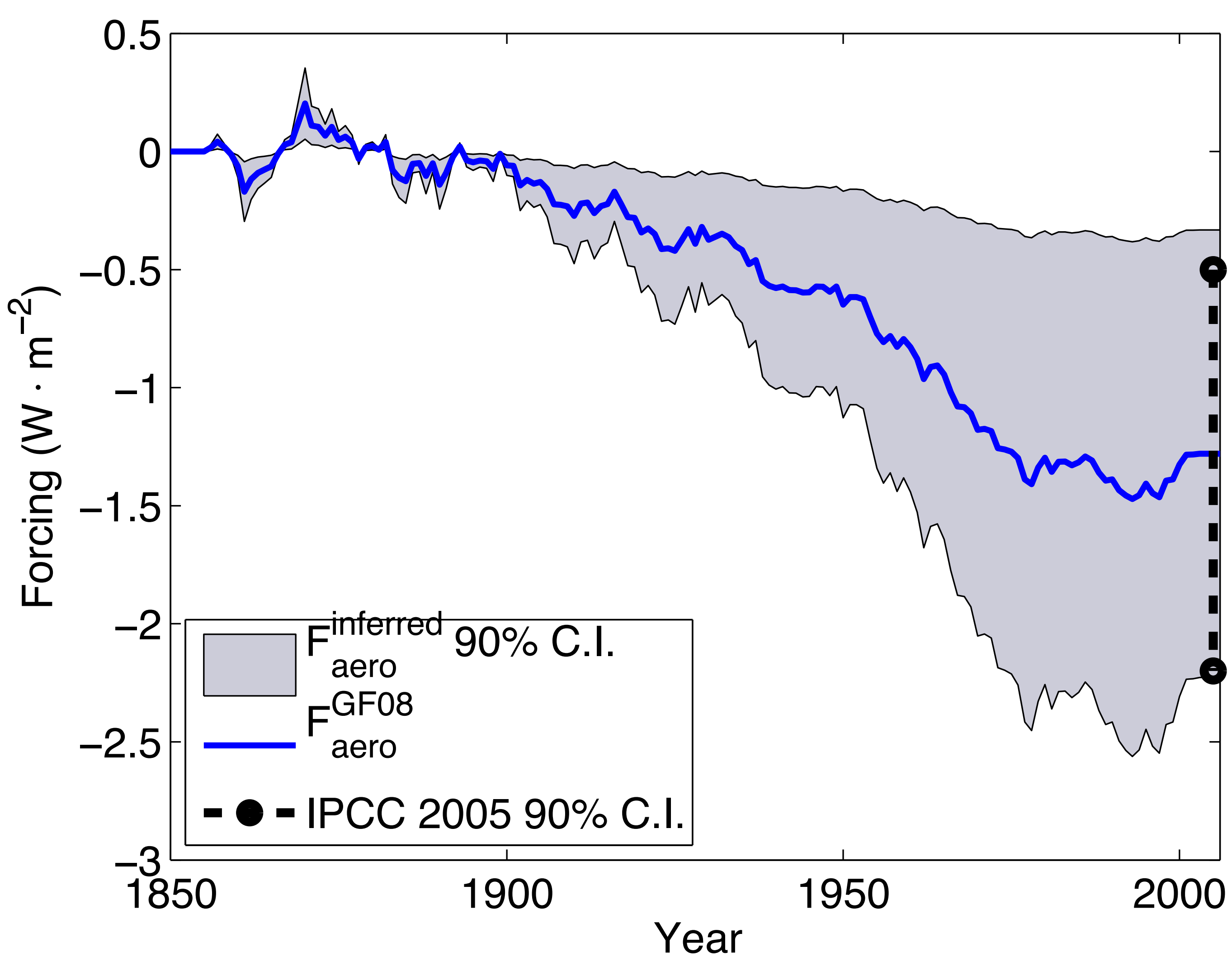
Aerosols

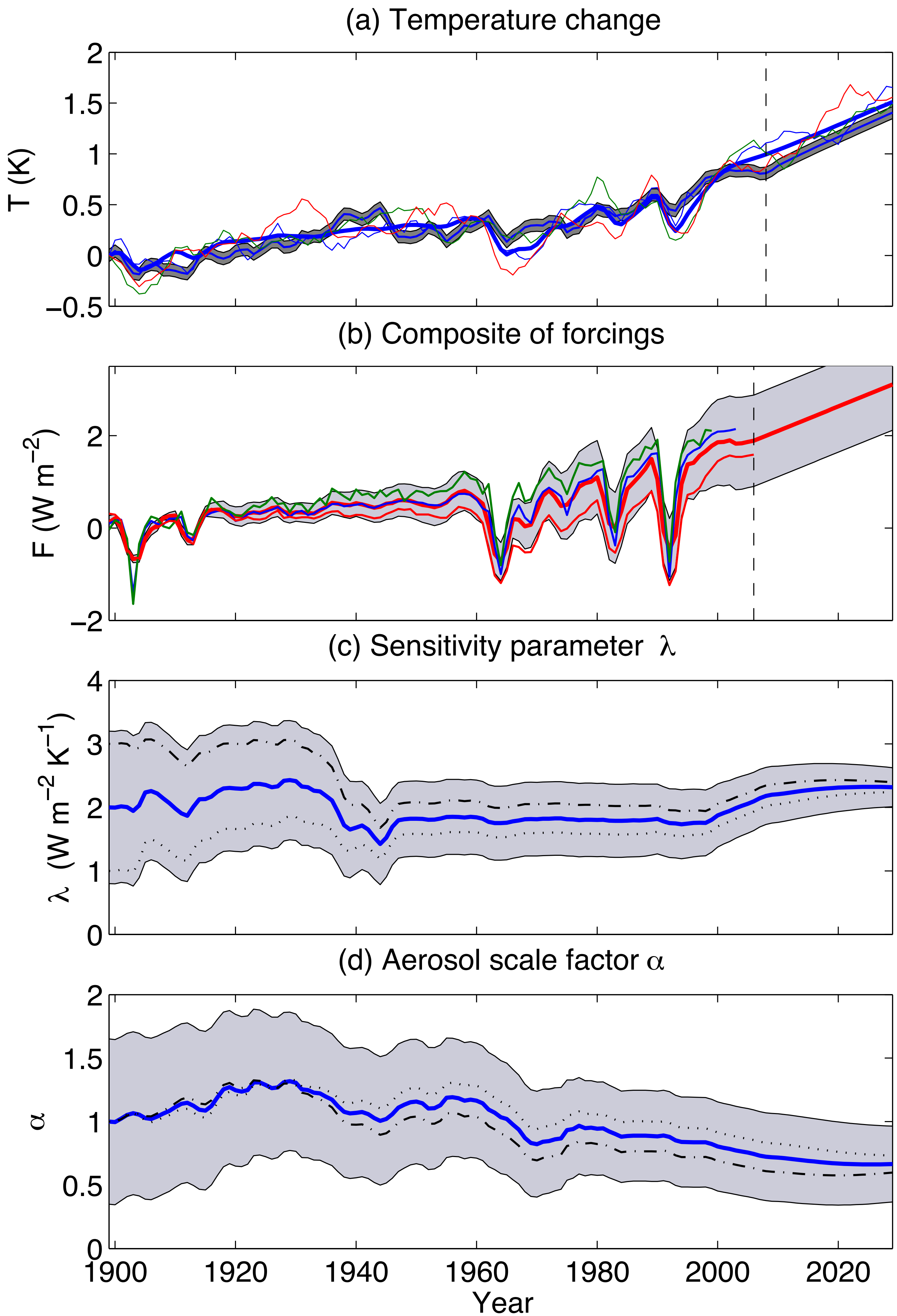
Difference in the forcings is mainly from aerosols. We let

$$F(t) = F_{GHG} + \alpha F_{\mathrm{aer}}(t)$$

where $F_{\mathrm{aer}}(t)$. Aerosol uncertainty grows with time. We assume uncertainty proportional to aerosol levels.

Aerosol forcing and uncertainty:





Temperature record. Observed plus simple model realizations.

Radiative forcings

Climate sensitivity parameter, λ

Aerosol scale factor

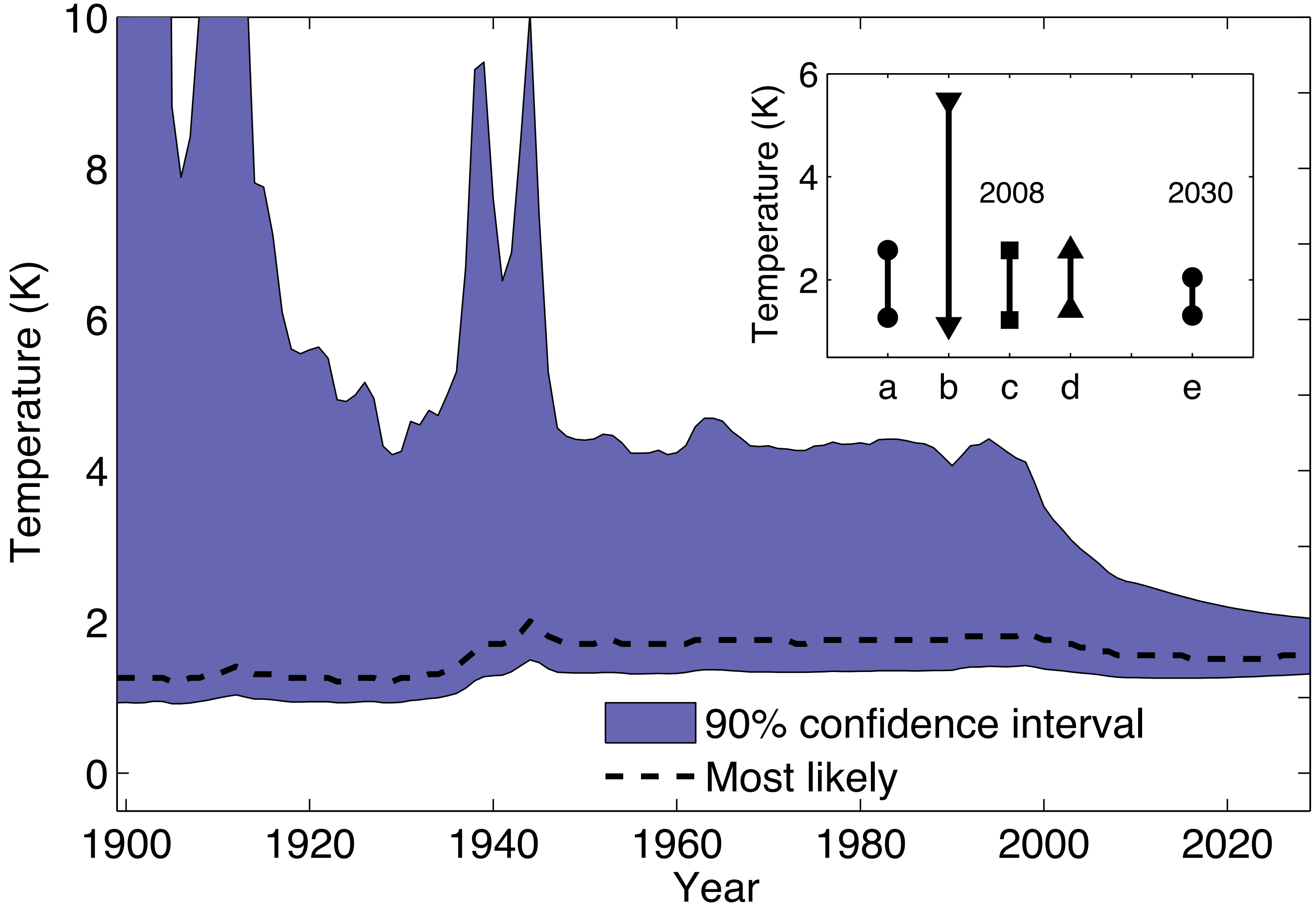
TCS probability evolution

The model produces a PDF for λ, where $C\partial T/\partial t = F - \lambda T + \dot{W}$.
The transient climate sensitivity (TCS) is given by

$$\text{TCS} = T_s = \frac{F_{2CO_2}}{\lambda}$$

and from PDF(λ) we compute PDF(T_s).

PDF of TCS:



Legend:

(a) Most probable

(b) high natural variability

(c) high forcing uncertainty

(d) low forcing uncertainty

(e) In the year 2030 (note time axis).

TCS probability evolution, modified

The model produces a PDF for γ, where

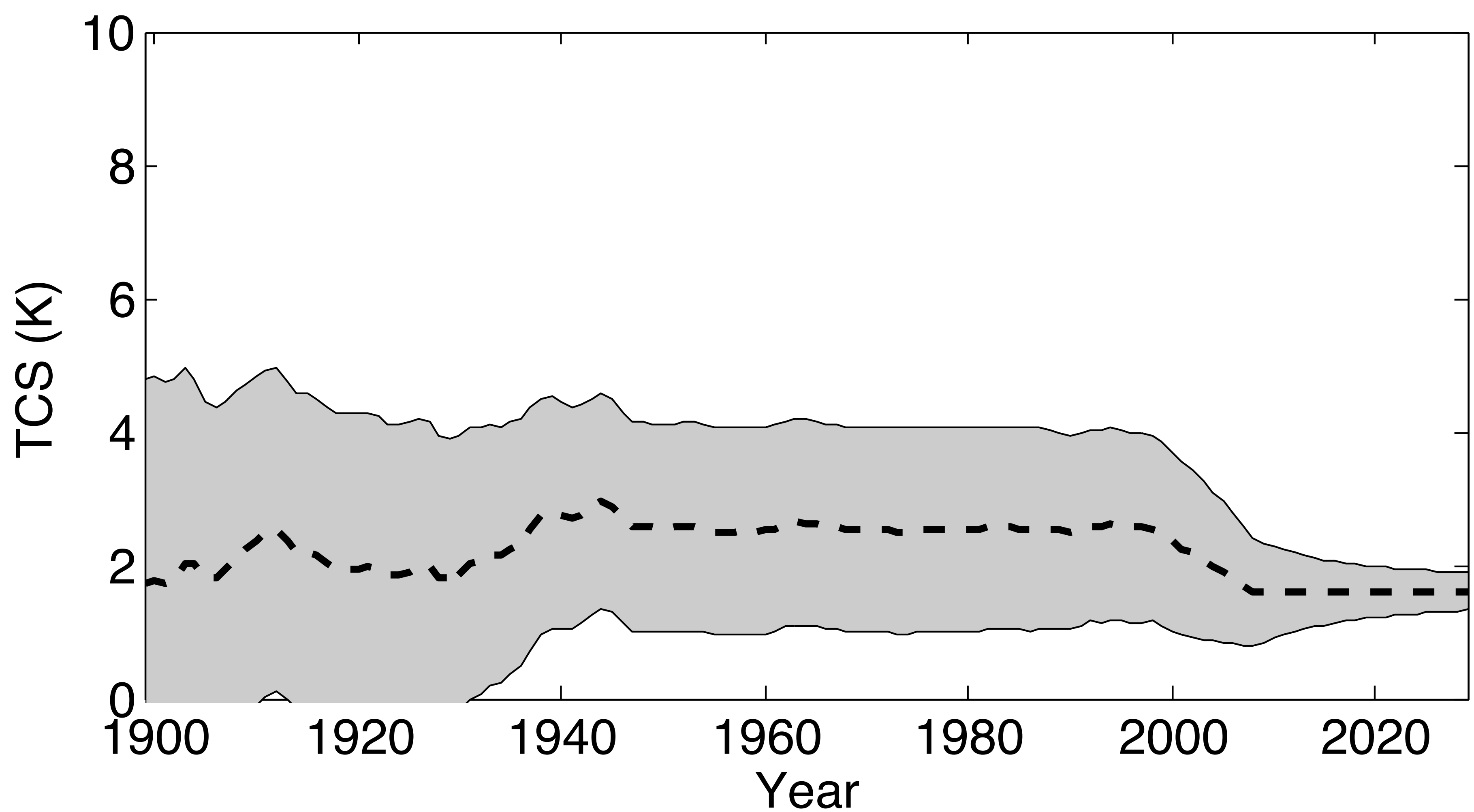
$$C\frac{\partial T}{\partial t} = F - \frac{T}{\gamma} + \dot{W}.$$

The transient climate sensitivity (TCS) is given by

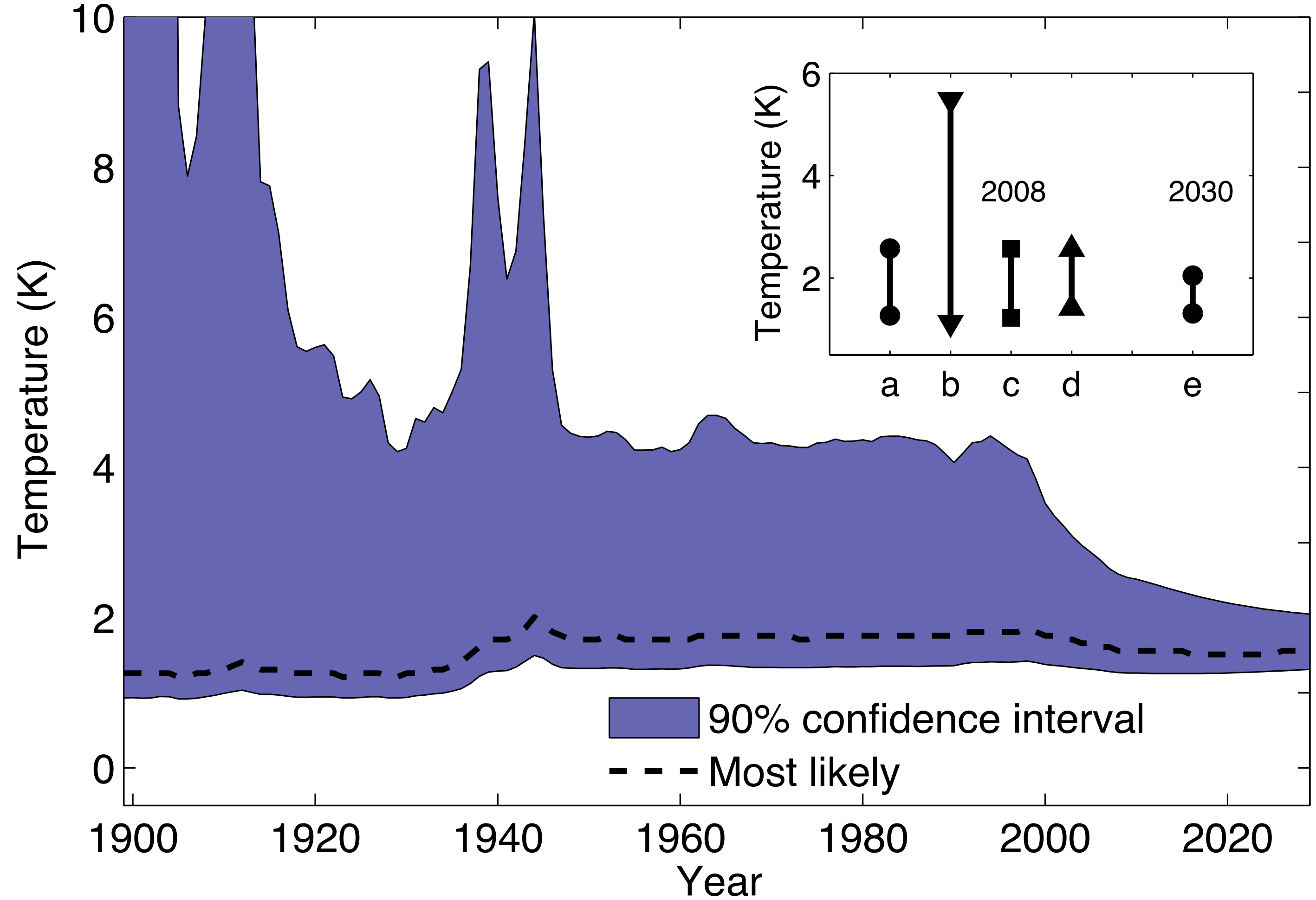
$$\text{TCS} = T_{tr} = F_{2\text{CO}_2} \times \gamma \qquad \text{and} \qquad \text{PDF}(T_{tr}) = F_{2\text{CO}_2} \times \text{PDF}(\gamma).$$

PDF of TCS:

(Now Gaussian)



Climate Sensitivity (Two methodologies)

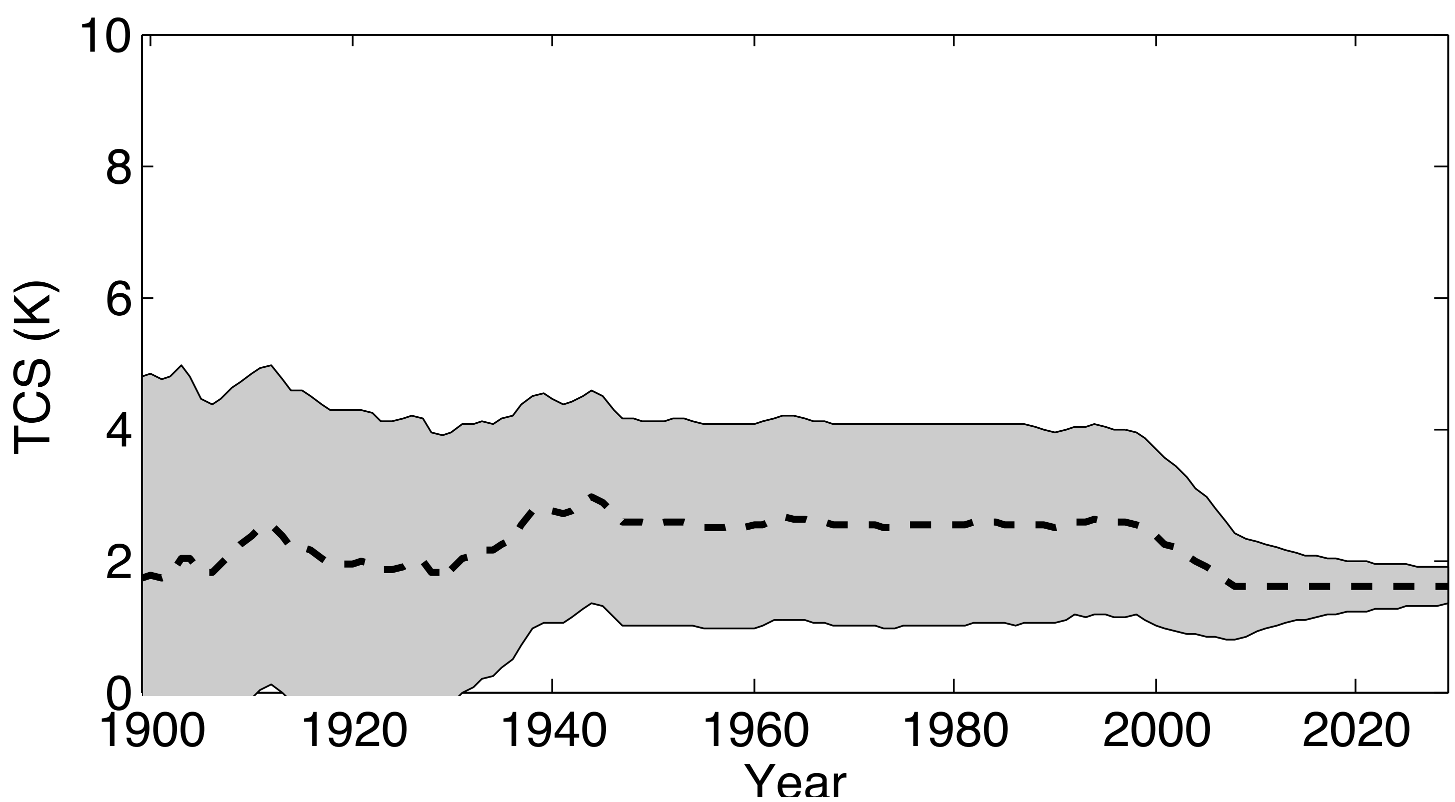


Summary of Method:

1. Obtain the radiative forcing, with error estimates

2. Fit the simple model to the data using a Kalman filter to give probabilistic estimates of climate sensitivity.

3. Results shown use two different techniques (different assumptions about what is Gaussian).

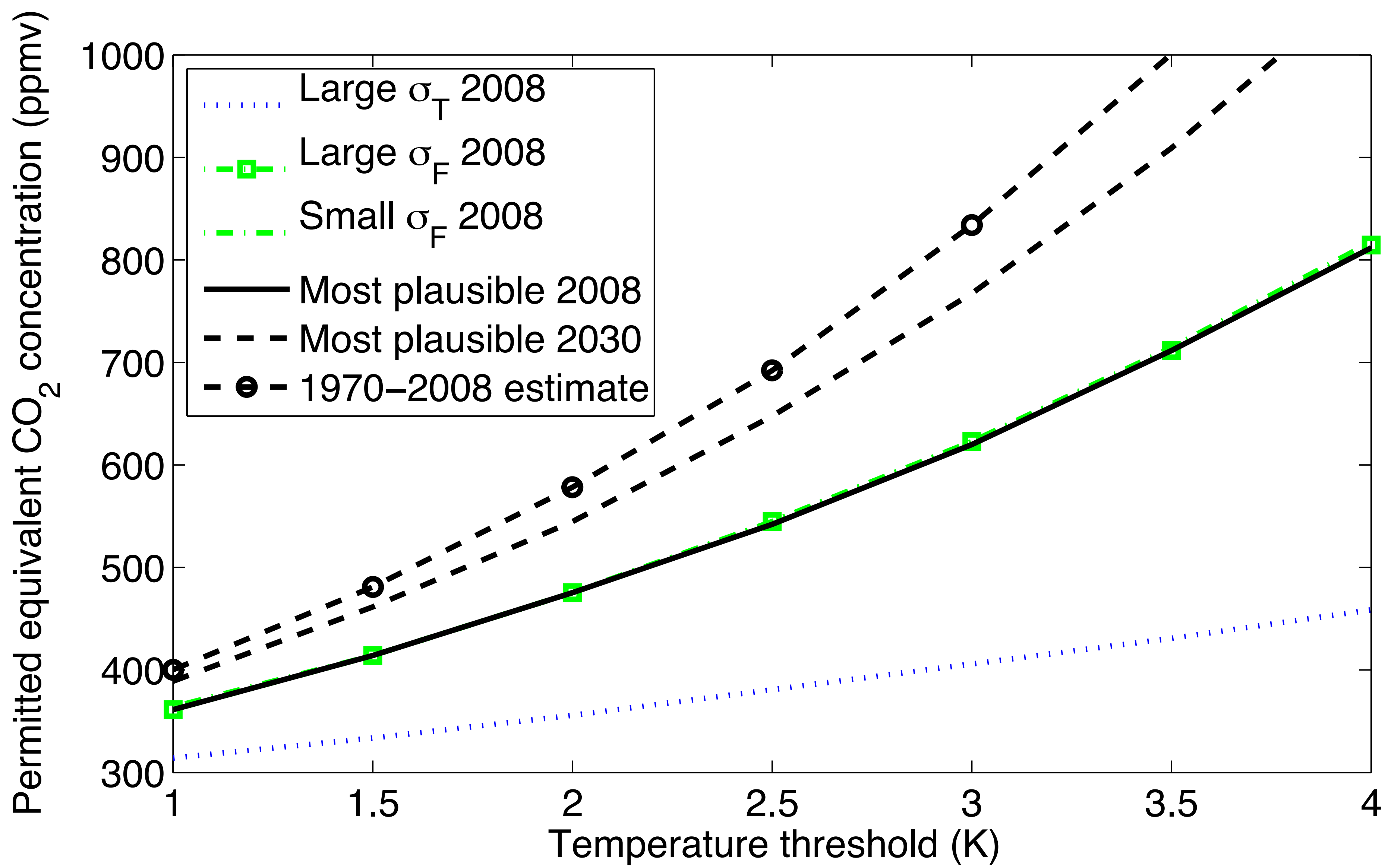


Dashed line is most probable value.

Shading indicates uncertainty.

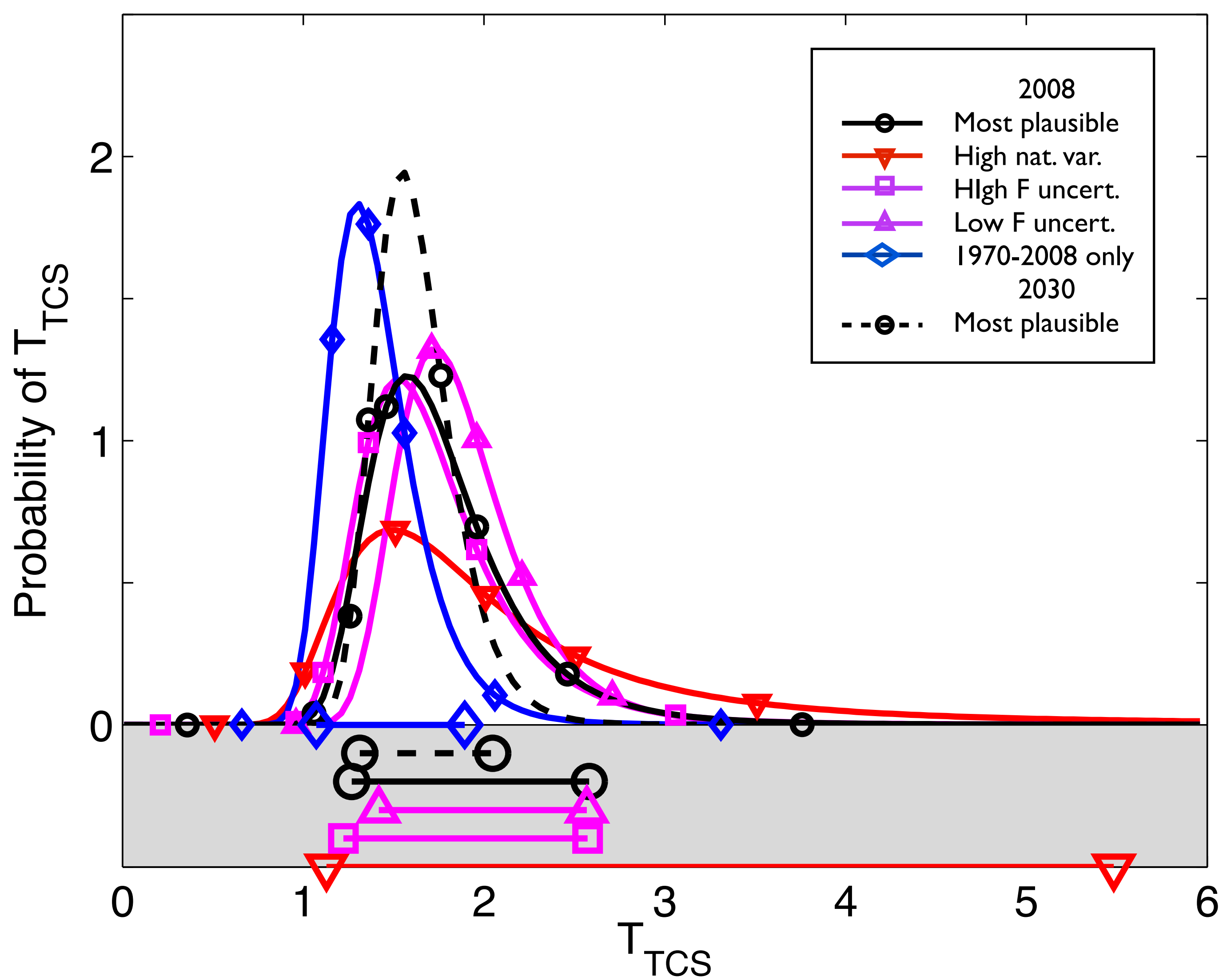
Forcing thresholds

Maximum forcing required to keep TCS below a given threshold with 95% confidence, under various scenarios.



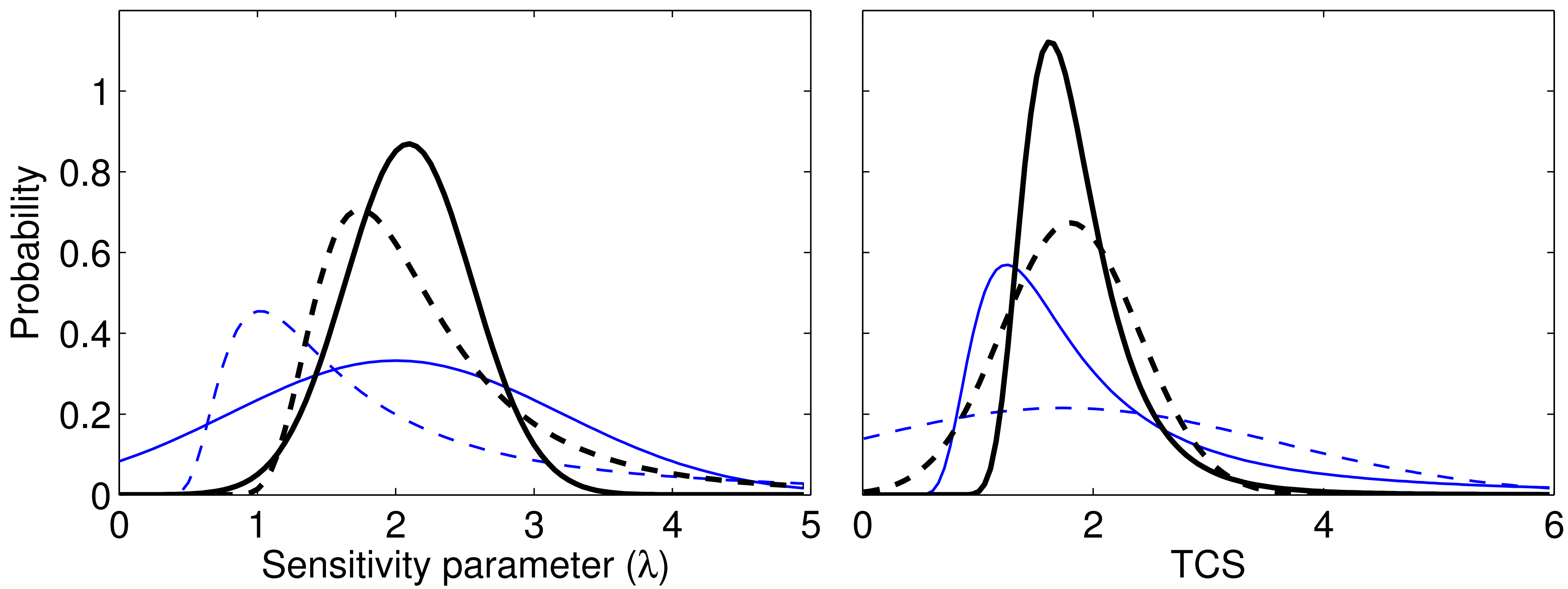
A Range of Ranges of TCS

Range of PDFs of Transient Climate Sensitivity from multiple experiments with various assumptions: different mixed-layer depths, different forcings, different strengths of natural variability etc.



TCS with two Methodologies

TCS parameter λ (1/K) TCS itself (K)



Prior distributions (1900): blue.

Posterior distributions (2008): black.

Gaussian λ: solid (—).

Gausssian $\gamma = 1/\lambda$: dashed (- - -)

Conclusions

- There are, usefully, slow (100–1000 years) and fast (10 years or less) oceanic responses, as determined by the mixed layer and the deep ocean respectively.
  - Polar amplification is suppressed in the Southern Hemisphere by ocean heat uptake on fast timescales.

- Transient climate sensitivity (fast response to a specified greenhouse gas increase) is the quantity of interest for most of the 21st century. Equilibrium value is higher — but may not be relevant.

- Transient climate sensitivity estimated as between 1.3 K and 2.6 K, with a most likely value of about 1.6 K for CO_2 doubling ($CO_2 = 580$ ppm).
  - Much uncertainty stems from knowing the aerosol effects on radiative forcing in the past.

- In order to be almost certain that temperature increase will be less than 2 K in medium term, need to limit CO_2 level to less than about 470 ppm.

- There are some methodological and scientific limitations, but they will not affect the results substantially (I think!).

The Ocean and Global Warming

The good news:

- The ocean will slow down the development of global warming.
  - Without the ocean, the global temperature responds almost immediately and completely to greenhouse gas increases. If the equilibrium climate sensitivity (ultimate response to a doubling of CO_2) is 4° C, we achieve that temperature just a few years after the greenhouse gas increase.
  - With the ocean, it will take several centuries to reach that level, and we will probably *never* reach it. Doubling of CO_2 gives us < 2° C warming on decadal timescales.

The bad news:

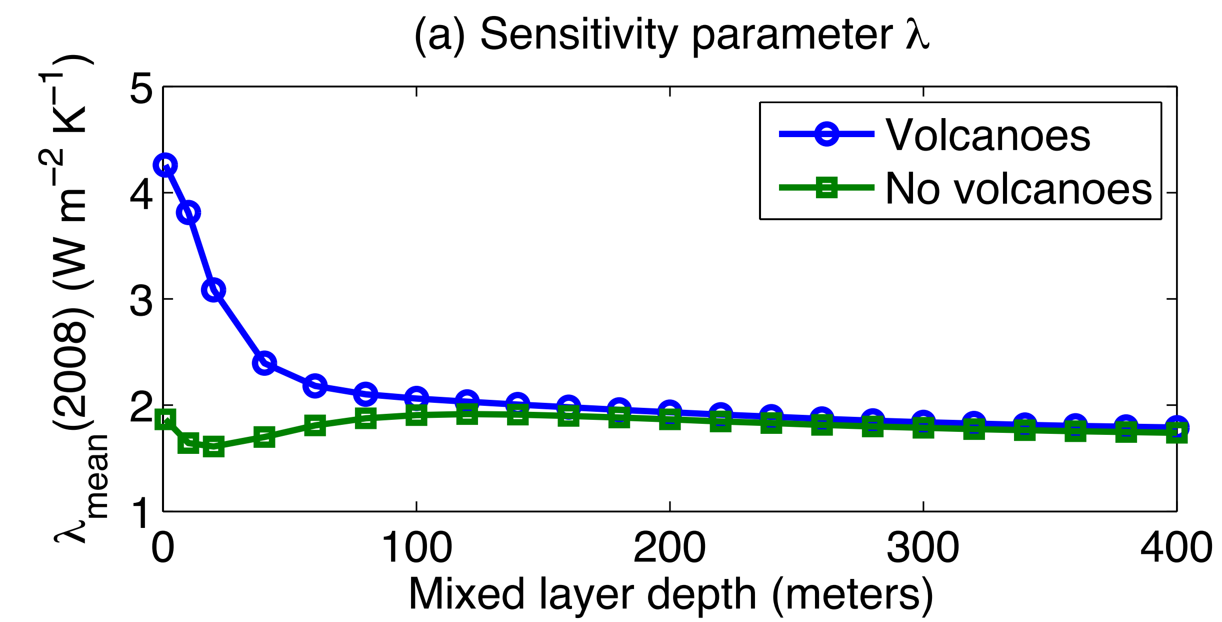
- The level of global warming is maintained even after we have stopped emitting greenhouse gases.
  - Because oceanic CO_2 uptake is slow, it will take centuries (and centuries) for the level of CO_2 to fall once emissions cease. So the temperature will stay high for a very long time!
  - Burning of all our oil and coal reserves could well increase CO_2 levels by a factor of 4 or 6.
  - The sustained warming could cause Greenland ice sheets to melt and sea-level to rise by six metres over the next few hundred to a thousand years.

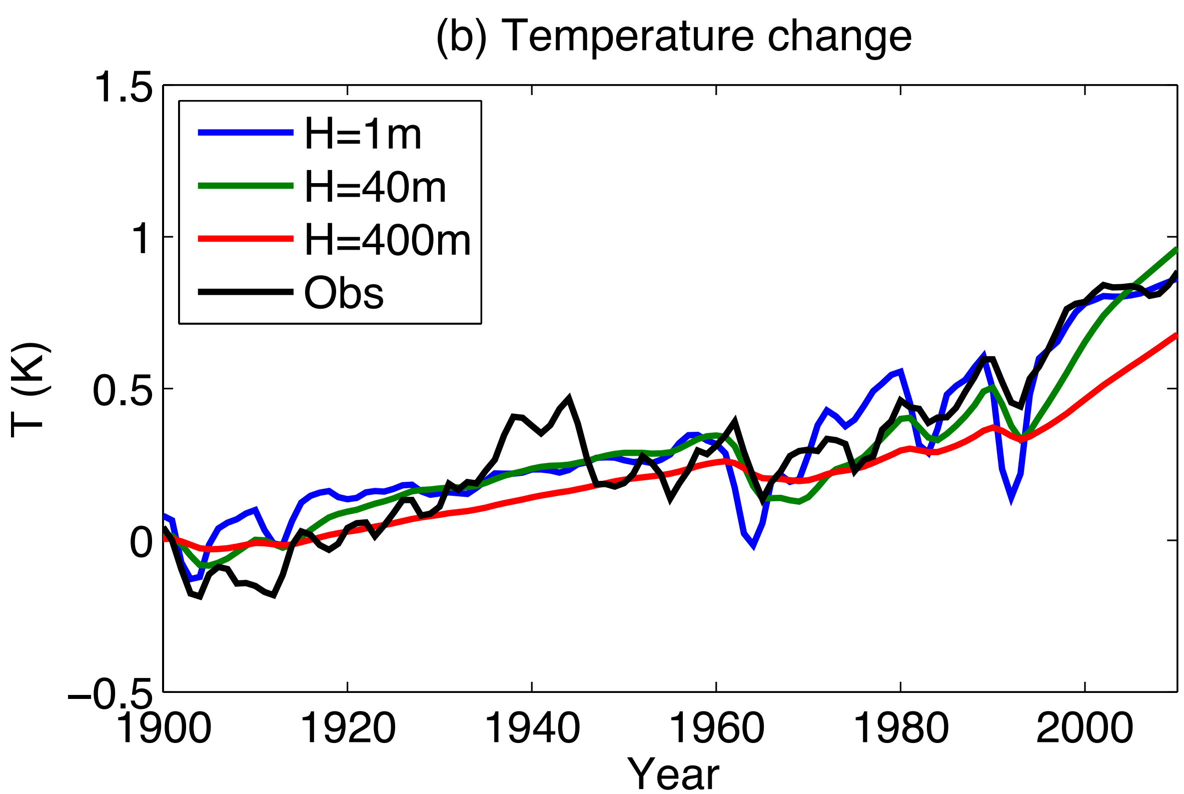
Limitations

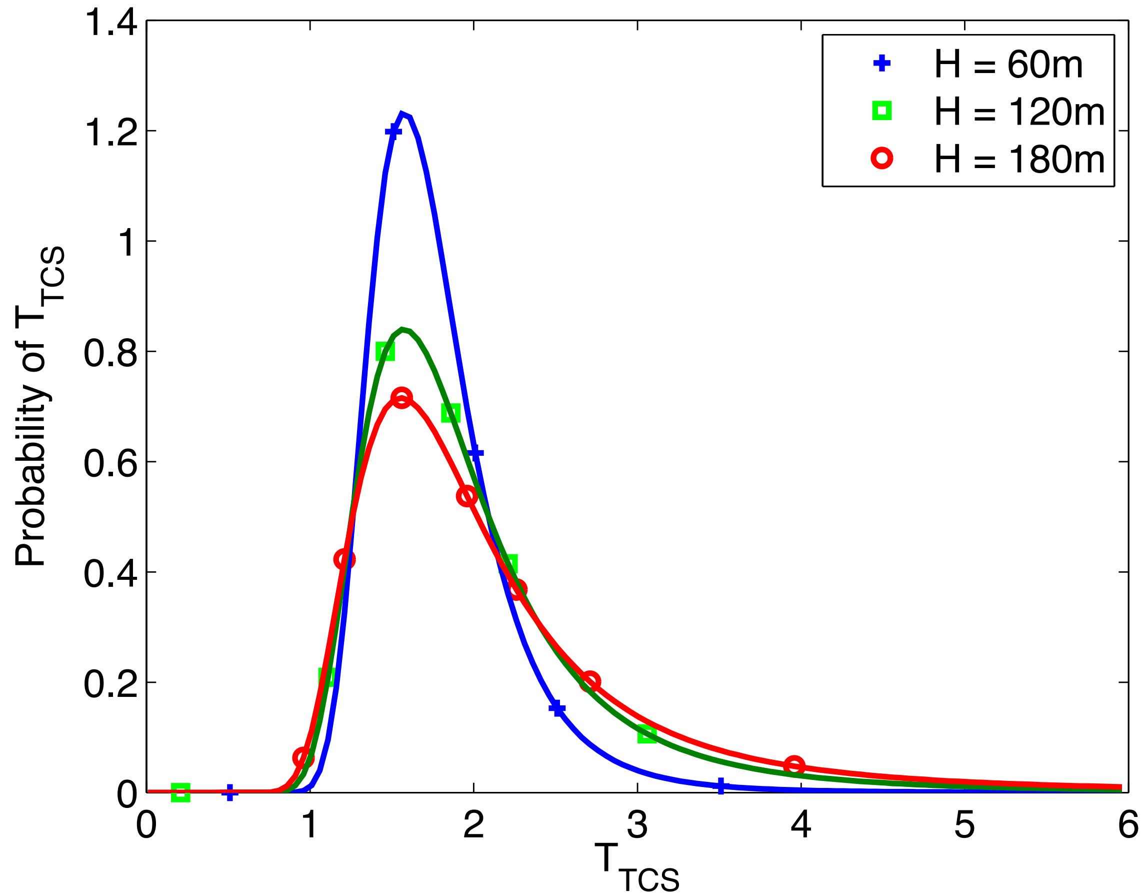
1. Forcing uncertainty is larger than ideal, because of aerosols.
    - Construct a model that treats Northern and Southern Hemispheres separately. In Southern Hemisphere aerosol effects are smaller.
    - Or actually try to predict spatial variations using a more complex energy-balance model.
    - But: more arbitrariness, and more parameters to estimate.

2. The PDFs are assumed Gaussian. Thus, λ is Gaussian, and PDF for T_{TCS} has a long tail, which arguably is artifactual.
    - Use γ directly, so no long tail.
    - Try to predict the shape of the PDF, using particle filters instead of Kalman filters.
        - Unlikely that the data exists to make this viable.

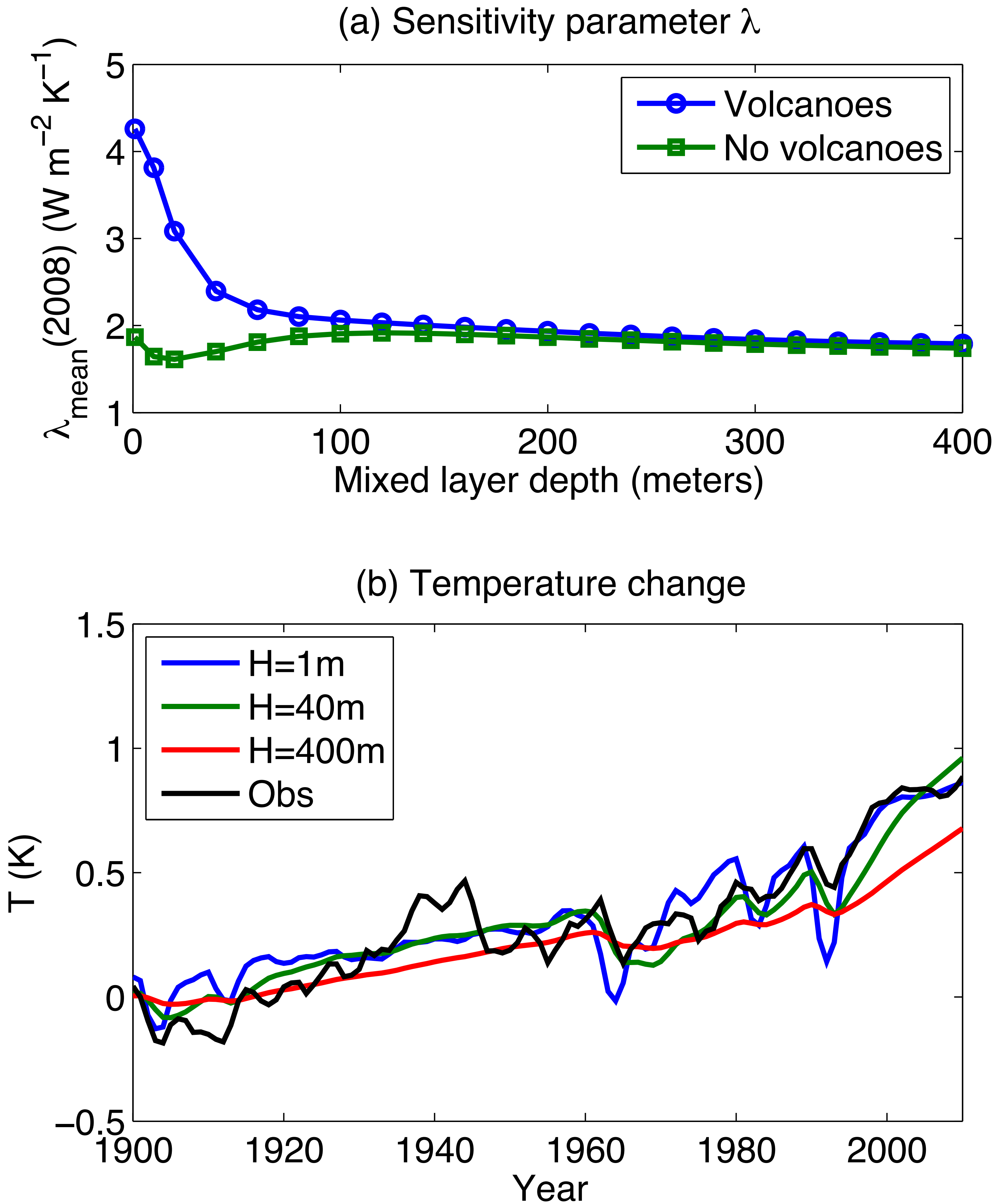
Effect of Choosing Different Mixed Layer Depths



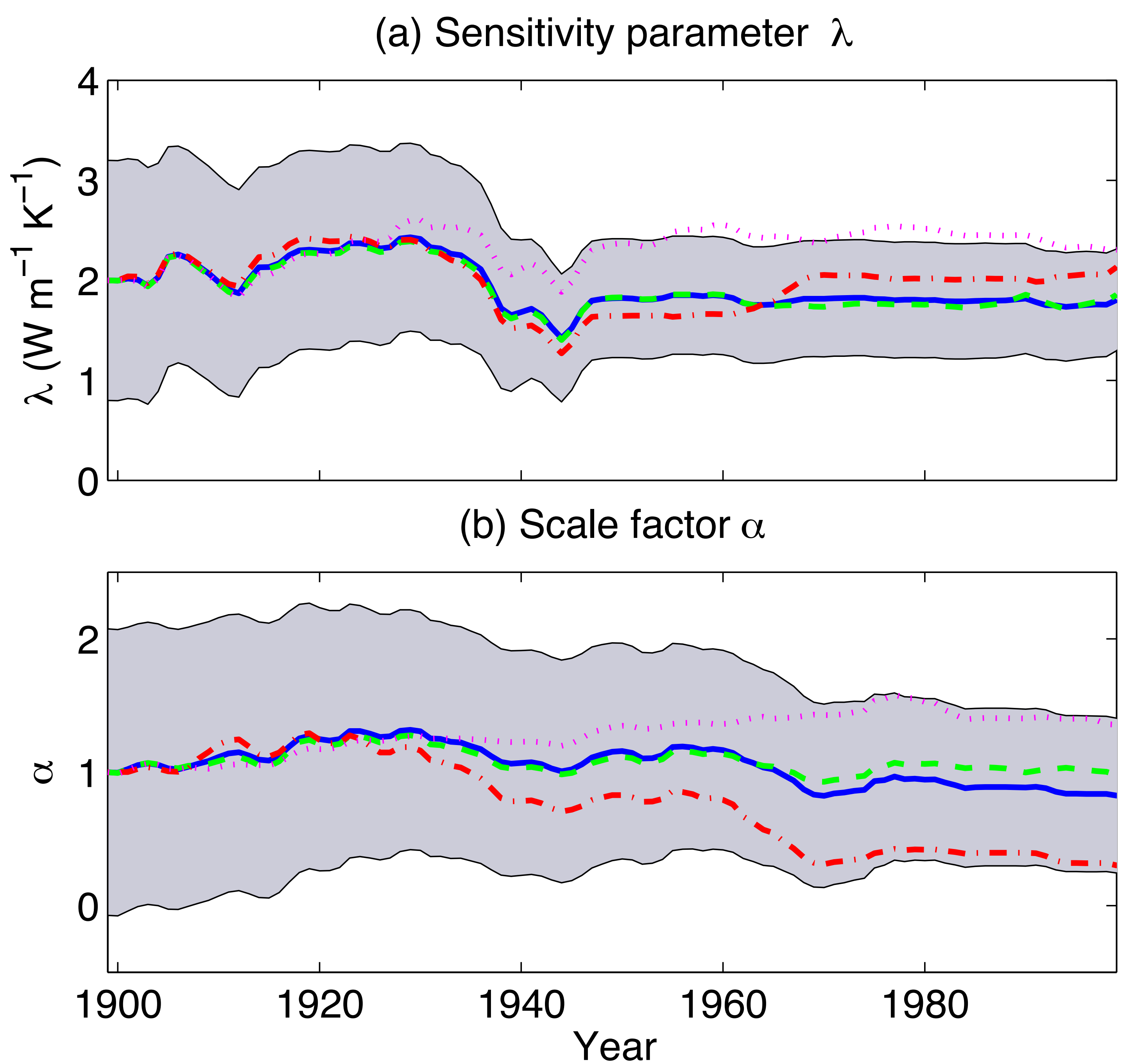




Effects of volcanoes

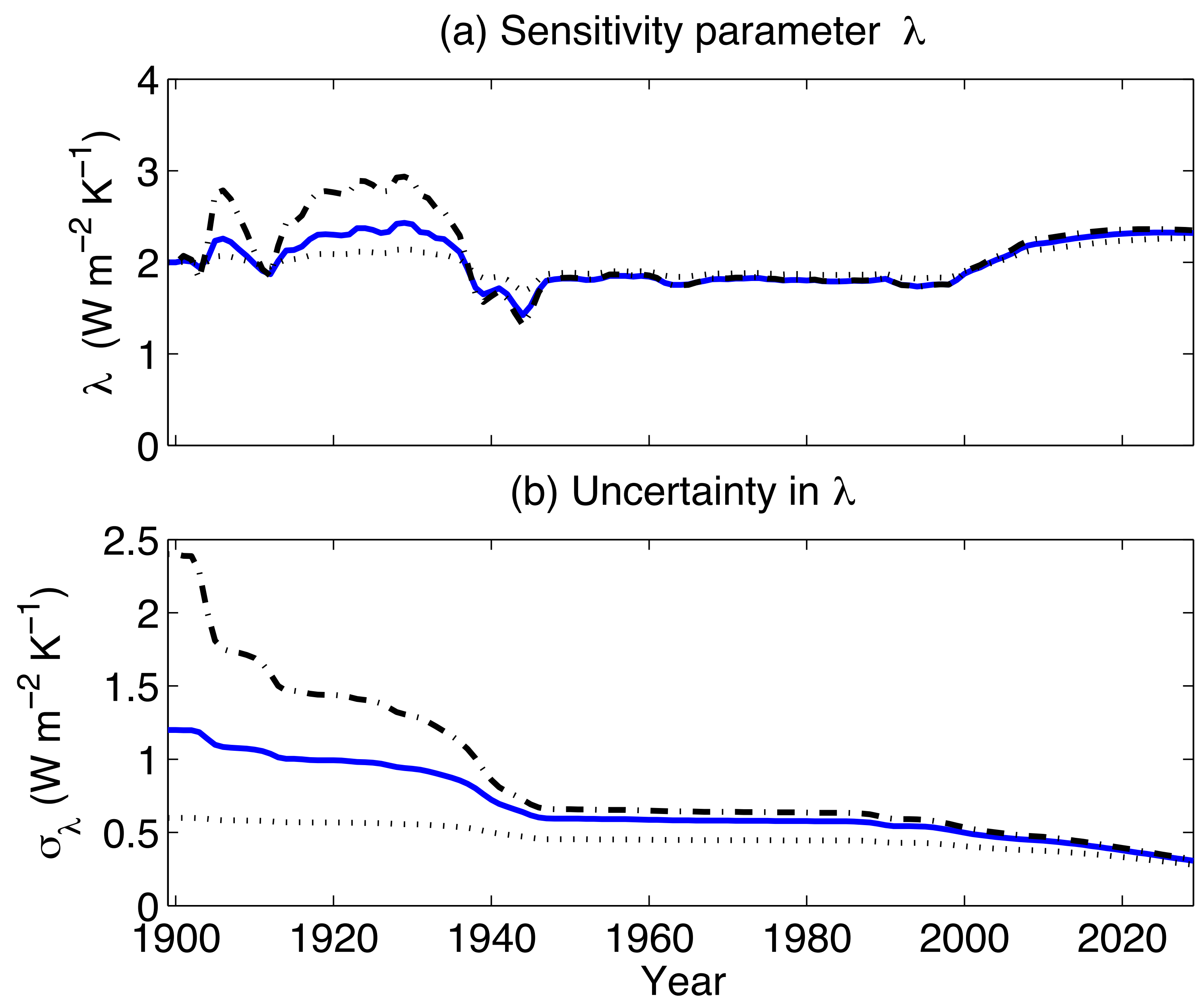


Using different forcings

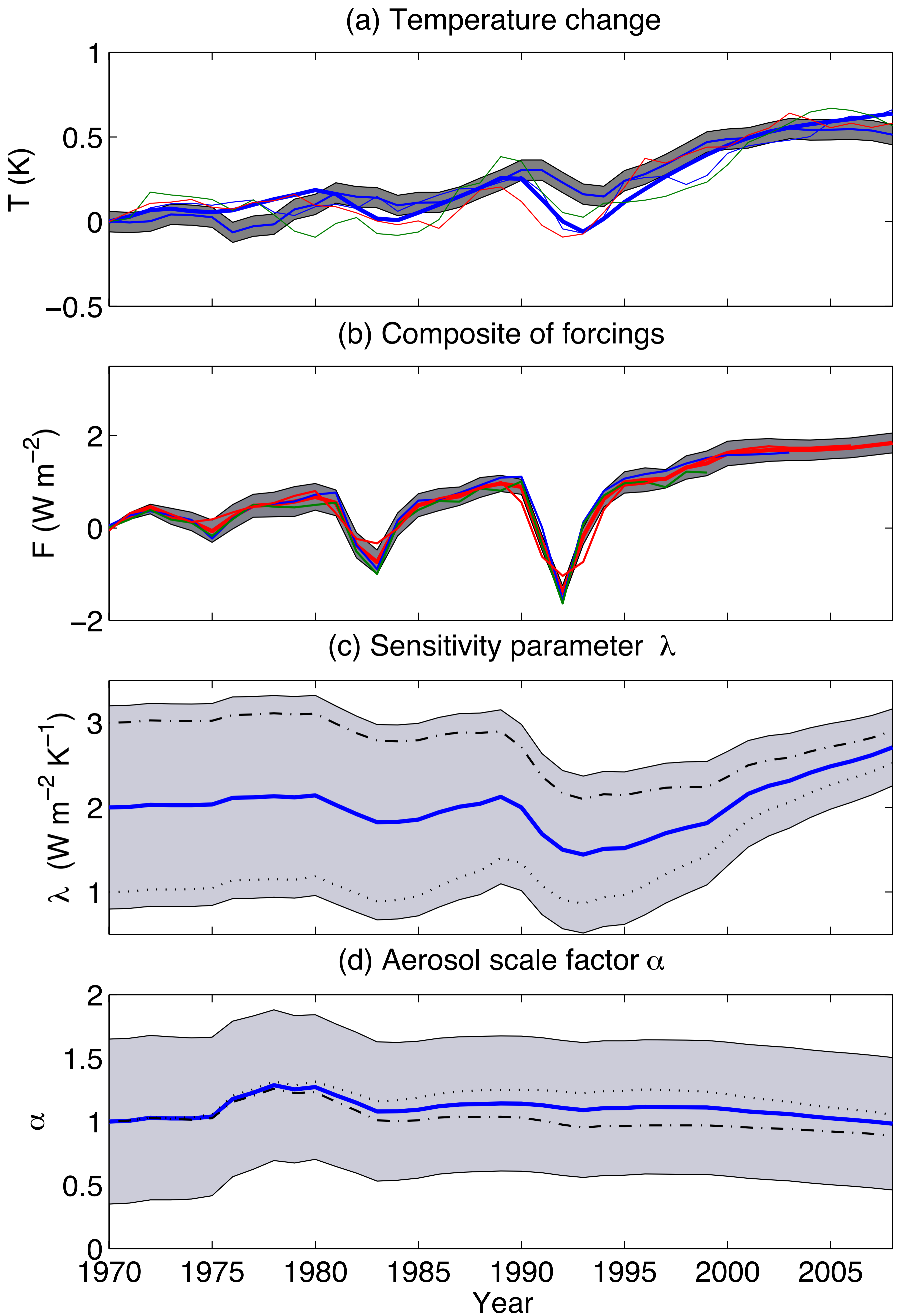


Effect of using forcing from GISS (green-dash), Gregory-Forster (red dash-dot), GFDL (magenta dot) or their mean (blue solid)

Effect of the prior uncertainty



Three different prior assumptions on uncertainty of λ.



Data from 1970 only

Temperature record. Observed plus simple model realizations.

Radiative forcings

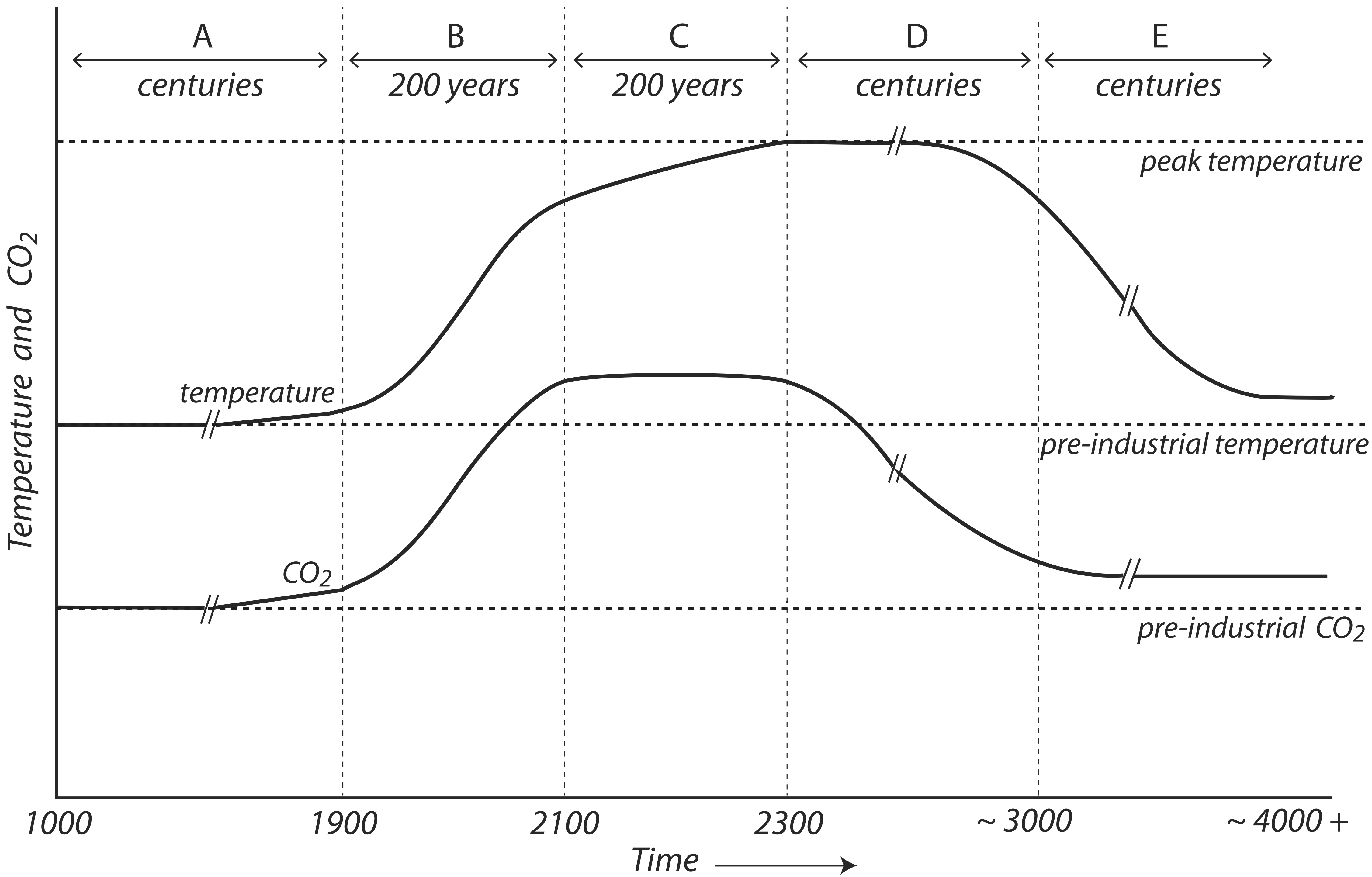
Climate sensitivity parameter, λ

(Less uncertainty from the forcing, but more from natural variability.)

Aerosol scale factor

Carbon dioxide and temperature

Any plausible scenario looks like this, modulo changes of the dates



Equilibrium climate sensitivity

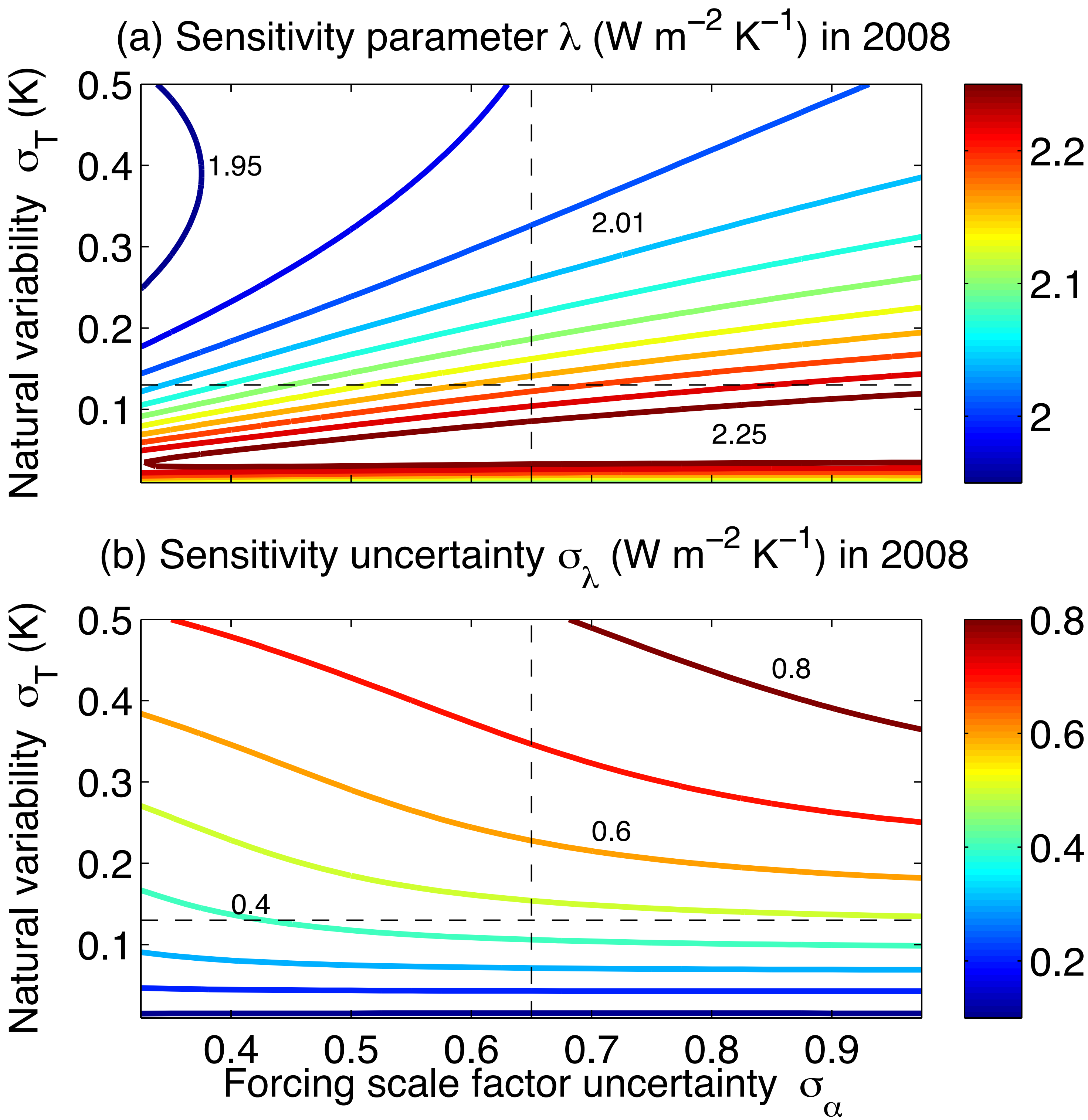
Equilibrium climate sensitivity is a harder problem, because:

1. It is likely to be much larger than the transient.
2. Therefore, more feedbacks at work. Ice-albedo feedback, melting of glaciers, changes in clouds.
3. The transient problem is, in a sense, extrapolation from 20th century record.
   - Unlikely that same approach works for long term.
4. So one point of view is:
   Global warming will be a smaller problem than is sometimes thought in the short and medium term (i.e., decades). However, it proceeds relentlessly and may be a worse problem on longer — centuries — timescales as ocean equilibrates.
5. Estimates of equilibrium climate sensitivity have large error bars.

Effects of natural variability and forcing uncertainty

in 2008

$$C\frac{\mathrm{d}T}{\mathrm{d}t} = F - \lambda T + \gamma \dot{W}$$



Effects of natural variability and forcing uncertainty

in 2030

$$C\frac{\mathrm{d}T}{\mathrm{d}t} = F - \lambda T + \gamma \dot{W}$$