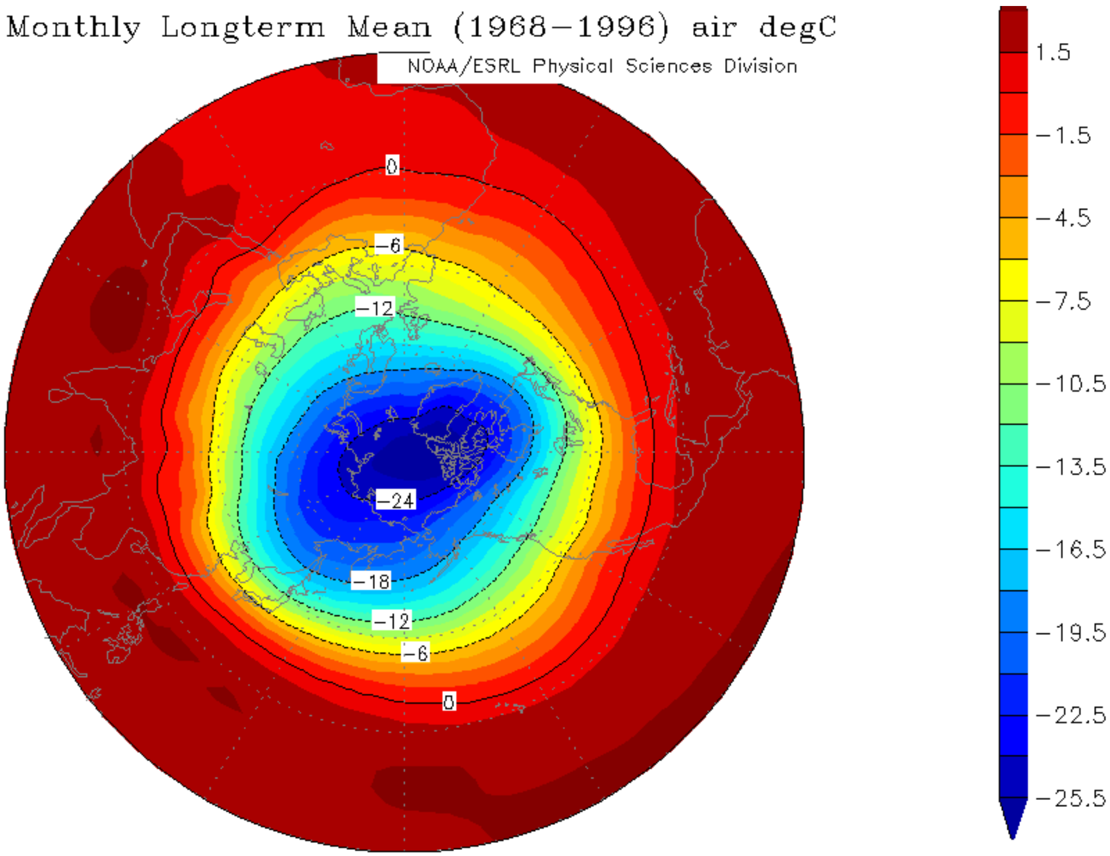
12.842 / 12.301 Past and Present Climate
Fall 2008

What controls the temperature gradient in middle and high latitudes?

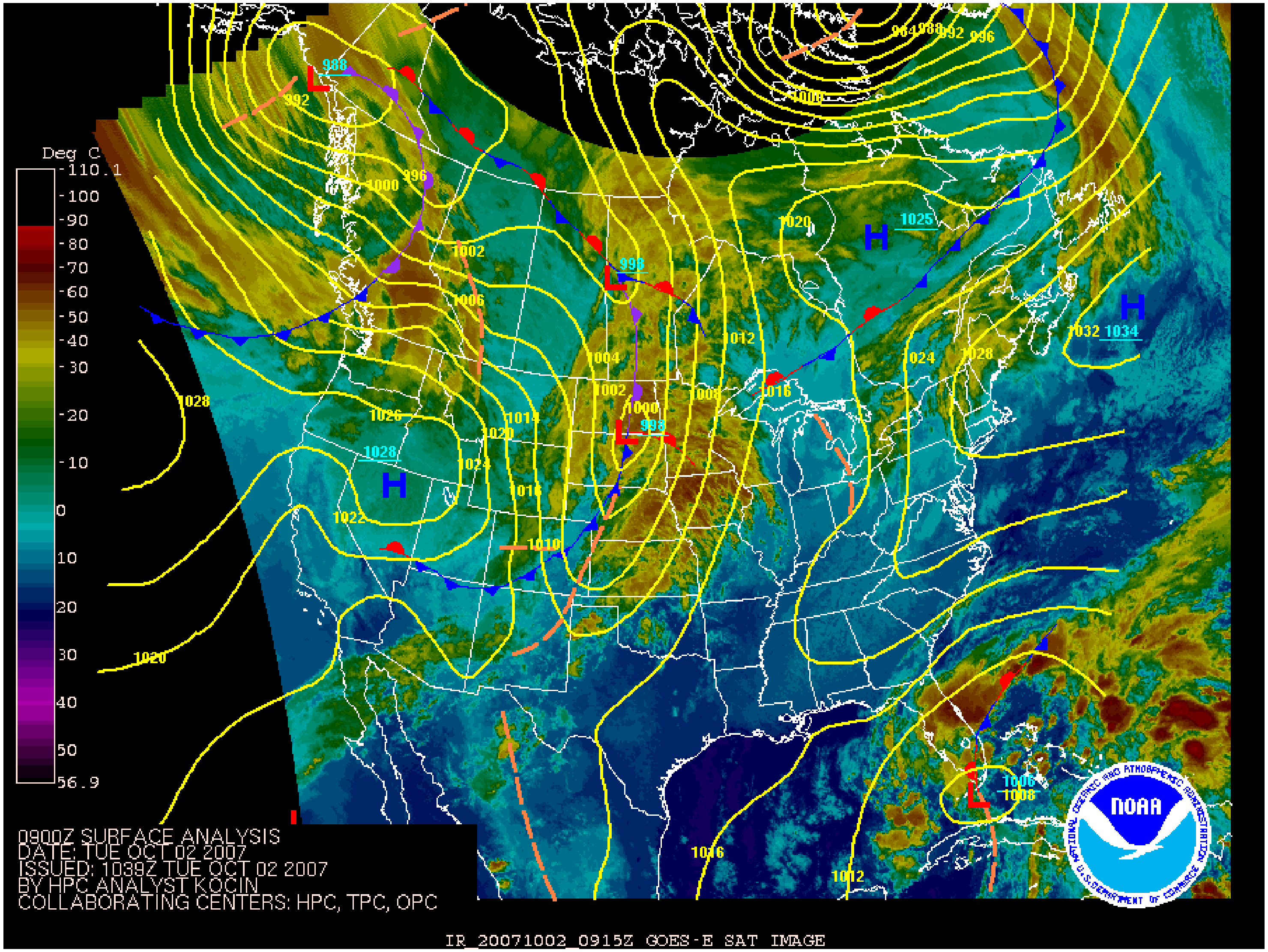


Annual mean temperature, northern hemisphere, about 4 km altitude

Issues

- Temperature gradient controlled by eddies of horizontal dimensions ~ 3000 km
- Familiar highs and lows on weather maps
- Eddy physics not simple
- Concept of criticality does not apply...critical T gradient = 0 (not observed)
- While eddies try to wipe out T gradient, they do not succeed

Example of surface pressure distribution



Concept of Available Potential Energy

- Difference between potential energy integrated over atmosphere and the minimum value that could be obtained by an adiabatic redistribution of mass

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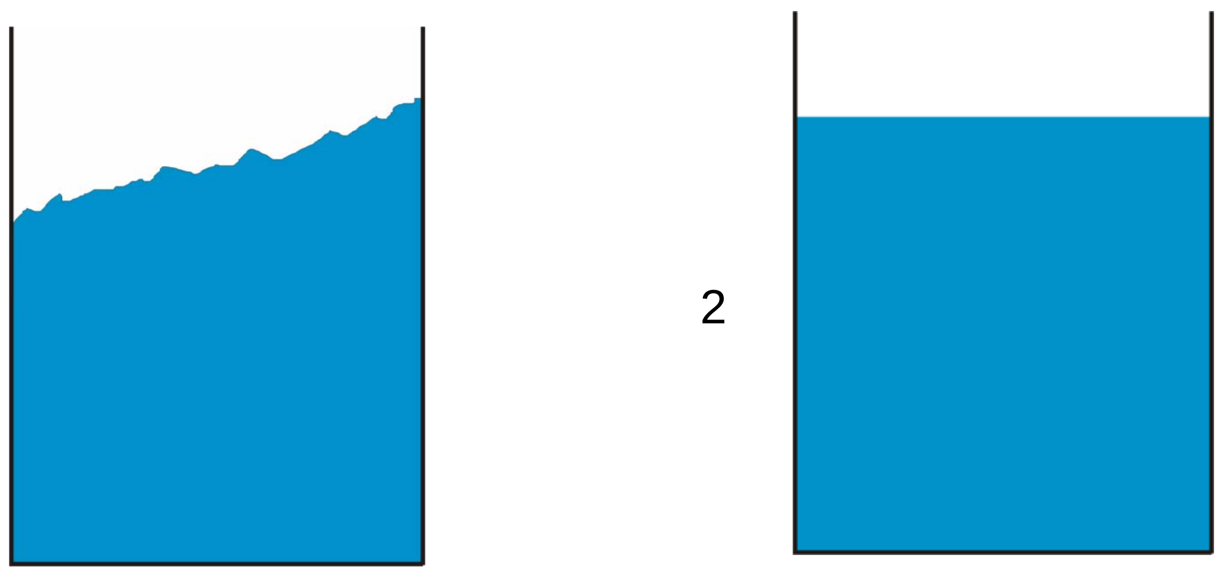

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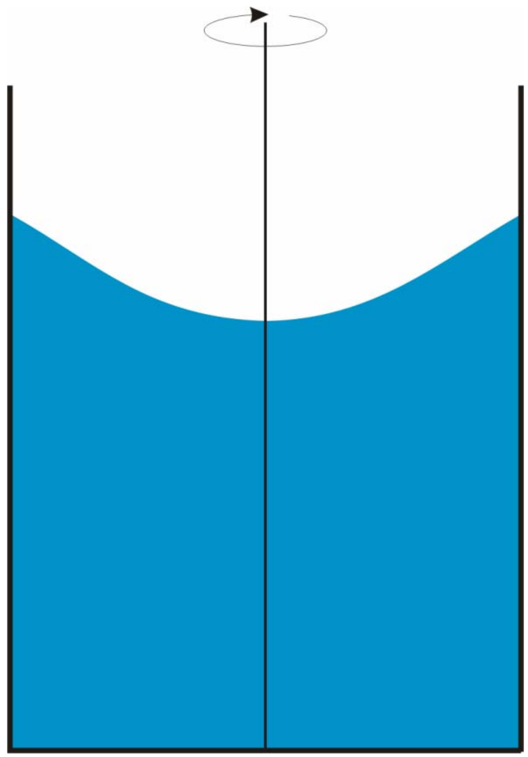
$$APE = PE_1 - PE_2 > 0$$

Complicated by rotation:



Is there a simple principle that governs the middle and high latitude temperature gradient, or do we have to deal with the eddies in all their complexity?

No generally agreed upon answer to this question

One possibility: Atmosphere arranges itself to maximize the rate of entropy production

$$\dot{s} = \int_V \frac{\dot{Q}_{rad}}{T}$$

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Consider two extreme possibilities:

1. Radiative equilibrium in each column:

$$\dot{Q}_{rad} = 0 \quad \rightarrow \quad \dot{s} = 0$$

2. Eddies succeed in wiping out T gradient:

$$\dot{s} = \frac{1}{T} \int_V \dot{Q}_{rad} = 0$$

$\therefore$ Maximum entropy production somewhere between radiative equilibrium and zero T gradient states

Absorption and emission of radiation thermodynamically reversible processes:

$$\dot{s}_{reversible} = \int_V \frac{\dot{Q}_{rad}}{T}$$

But entropy is a state variable, so no net change in long-term average:

$$\dot{s}_{reversible} + \dot{s}_{irreversible} = 0$$

$$\rightarrow \dot{s}_{irreversible} = -\int_V \frac{\dot{Q}_{rad}}{T}$$

$$\simeq -\frac{\dot{Q}_{shortwave}}{T_{surface}} - \frac{\dot{Q}_{longwave}}{T_{emission}}$$

$$But \quad \dot{Q}_{longwave} = -\dot{Q}_{shortwave}$$

$$\rightarrow \dot{s}_{irreversible} = \dot{Q}_{shortwave}\left(\frac{1}{T_{emission}} - \frac{1}{T_{surface}}\right) > 0$$

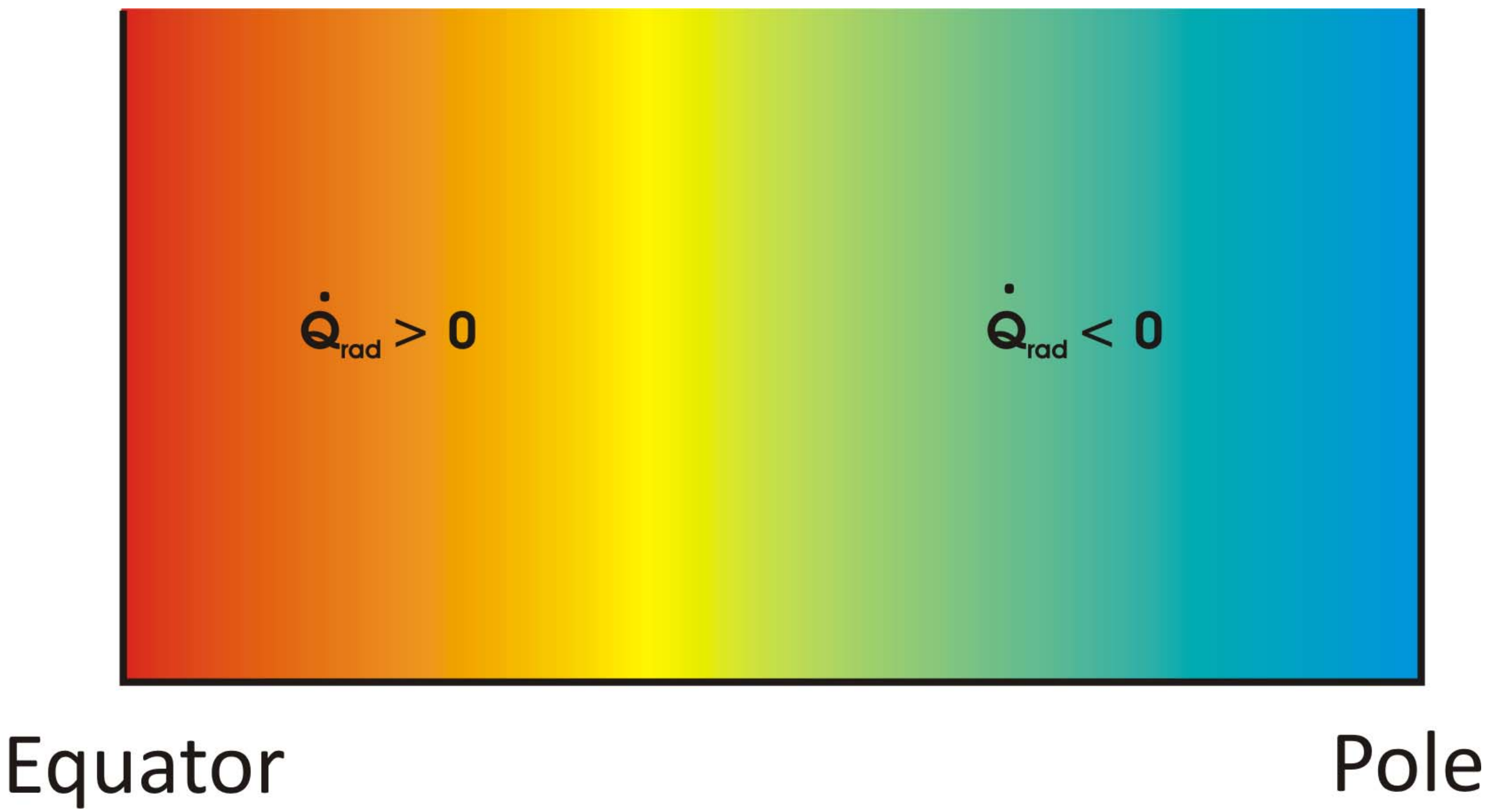
Irreversible entropy production by microphysical processes:

- Mixing of cloudy and clear air
- Fall of rain and snow
- Frictional dissipation of wind energy

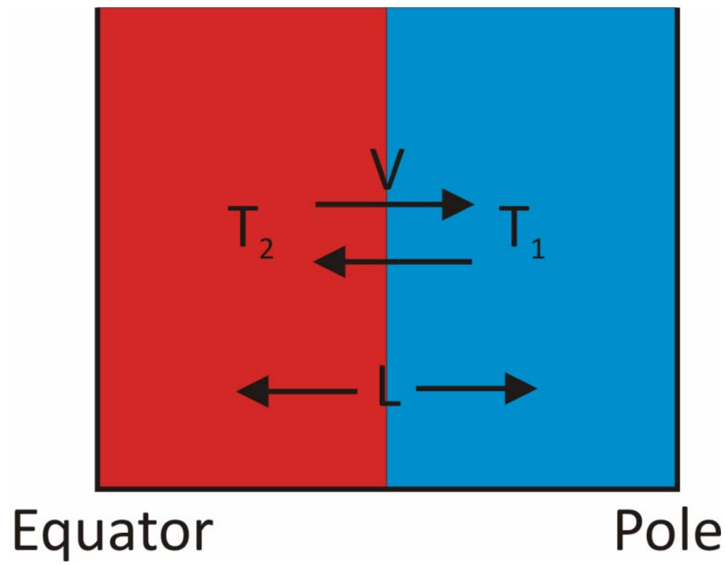
Nevertheless, sum total of all of these constrained by

$$\dot{s}_{irreversible} = \dot{Q}_{shortwave}\left(\frac{1}{T_{emission}} - \frac{1}{T_{surface}}\right)$$

Convective generation of entropy augmented if there are also *horizontal* temperature gradients:



Simple Two-Box Model:



$$\frac{\partial T_2}{\partial t} = 0 = \frac{T_{2_{eq}} - T_2}{\tau_{rad}} - \frac{V}{L}\left(T_2 - T_1\right) \qquad \frac{\partial T_1}{\partial t} = 0 = \frac{T_{1_{eq}} - T_1}{\tau_{rad}} + \frac{V}{L}\left(T_2 - T_1\right)$$

Definitions:

$$\overline{T} \equiv \frac{1}{2}\left(T_{1_{eq}} + T_{2_{eq}}\right),$$

$$\Delta T_{eq} \equiv T_{2_{eq}} - T_{1_{eq}},$$

$$\Delta T \equiv T_2 - T_1$$

$$\chi \equiv \frac{\tau_{rad} V}{L}$$

Solution:

$$T_1 + T_2 = T_{1_{eq}} + T_{2_{eq}},$$

$$\Delta T = \frac{\Delta T_{eq}}{1 + 2\chi}$$

Entropy Production:

$$-\dot{s}_{irreversible} = \frac{T_{2_{eq}} - T_2}{T_2 \tau_{rad}} + \frac{T_{1_{eq}} - T_1}{T_1 \tau_{rad}}$$

$$= \frac{1}{\tau_{rad}} \left(\frac{\Delta T_{eq}}{\overline{T}} \right)^2 \frac{\chi}{(1+2\chi)^2 - \frac{1}{4}\left(\frac{\Delta T_{eq}}{\overline{T}} \right)^2}$$

$$\dot{s}\ maximum\ when \quad \chi = \frac{1}{2}\sqrt{1 - \frac{1}{4}\left(\frac{\Delta T_{eq}}{\overline{T}} \right)^2}$$

$$\Delta T = \Delta T_{eq} \frac{1}{1+\sqrt{1-\frac{1}{4}\left(\frac{\Delta T_{eq}}{\overline{T}}\right)^2}} \cong \frac{1}{2}\Delta T_{eq}$$

Can this be generalized to include more boxes, processes?

Remember that real eddies affect distributions of clouds, water vapor

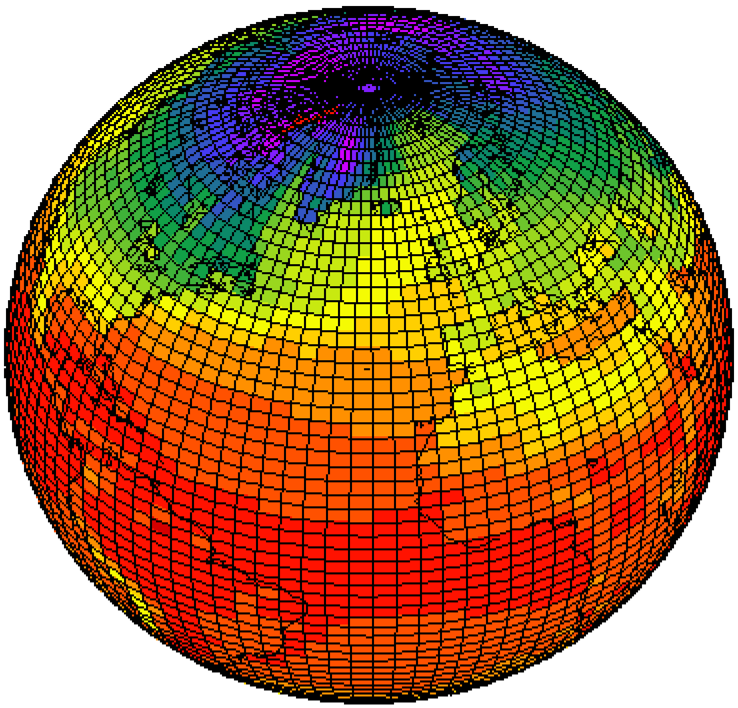
Global Climate Modeling

- **General philosophy**:
  - Simulate large-scale motions of atmosphere, oceans, ice
  - Solve approximations to full radiative transfer equations
  - Parameterize processes too small to resolve
  - Some models also try to simulate biogeochemical processes
  - First GCMs developed in 1960s

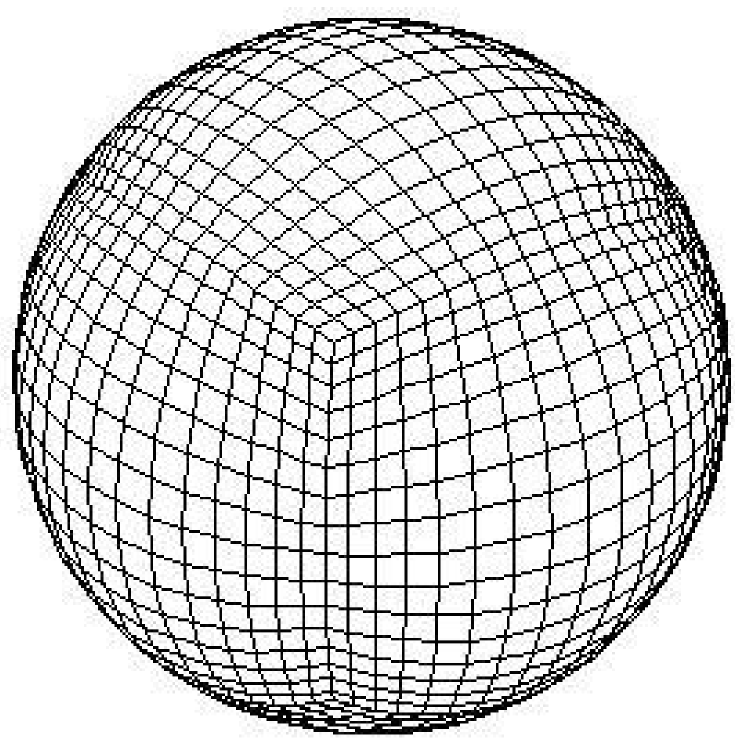
Model Partial Differential Equations

- Conservation of momentum
- Conservation of mass
- Conservation of water
- First law of Thermodynamics
- Equation of state
- Radiative transfer equations

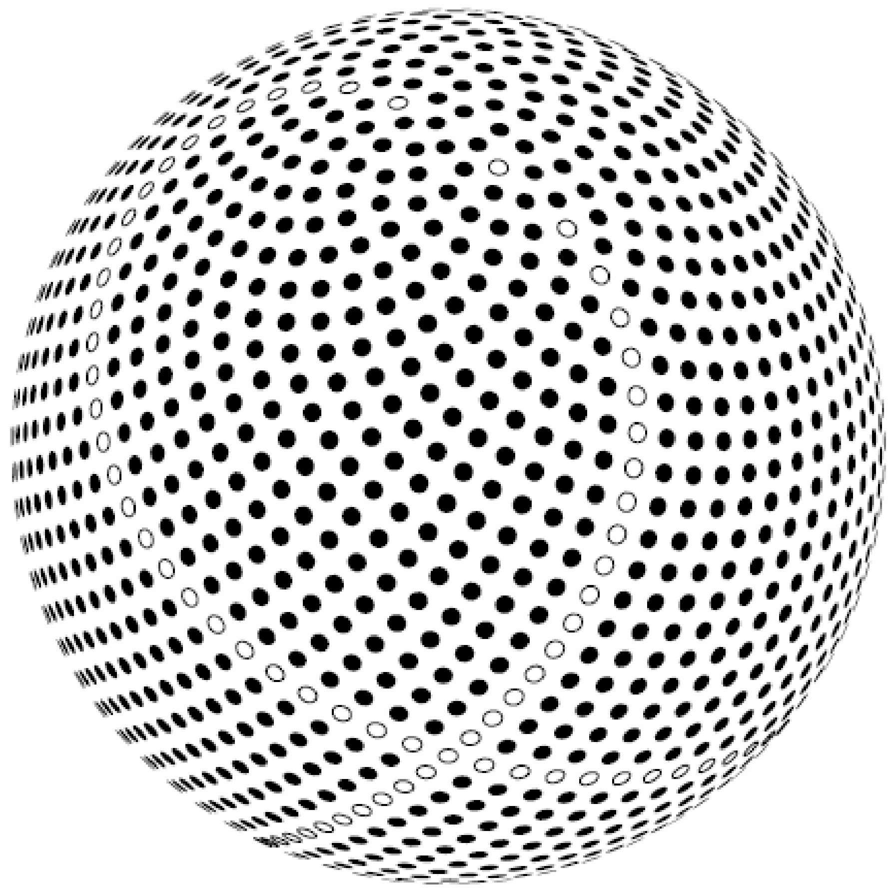
Alternative Grids:



Classical spherical coordinates



Conformal mapping of cube onto sphere



A spherical grid based on the Fibonacci sequence. The grid is highly uniform and isotropic.

Some Fundamental Numerical Constraints

Courant-Friedrichs-Lewy (CFL) condition:

$$\frac{c\Delta t}{\Delta x} < 1,$$

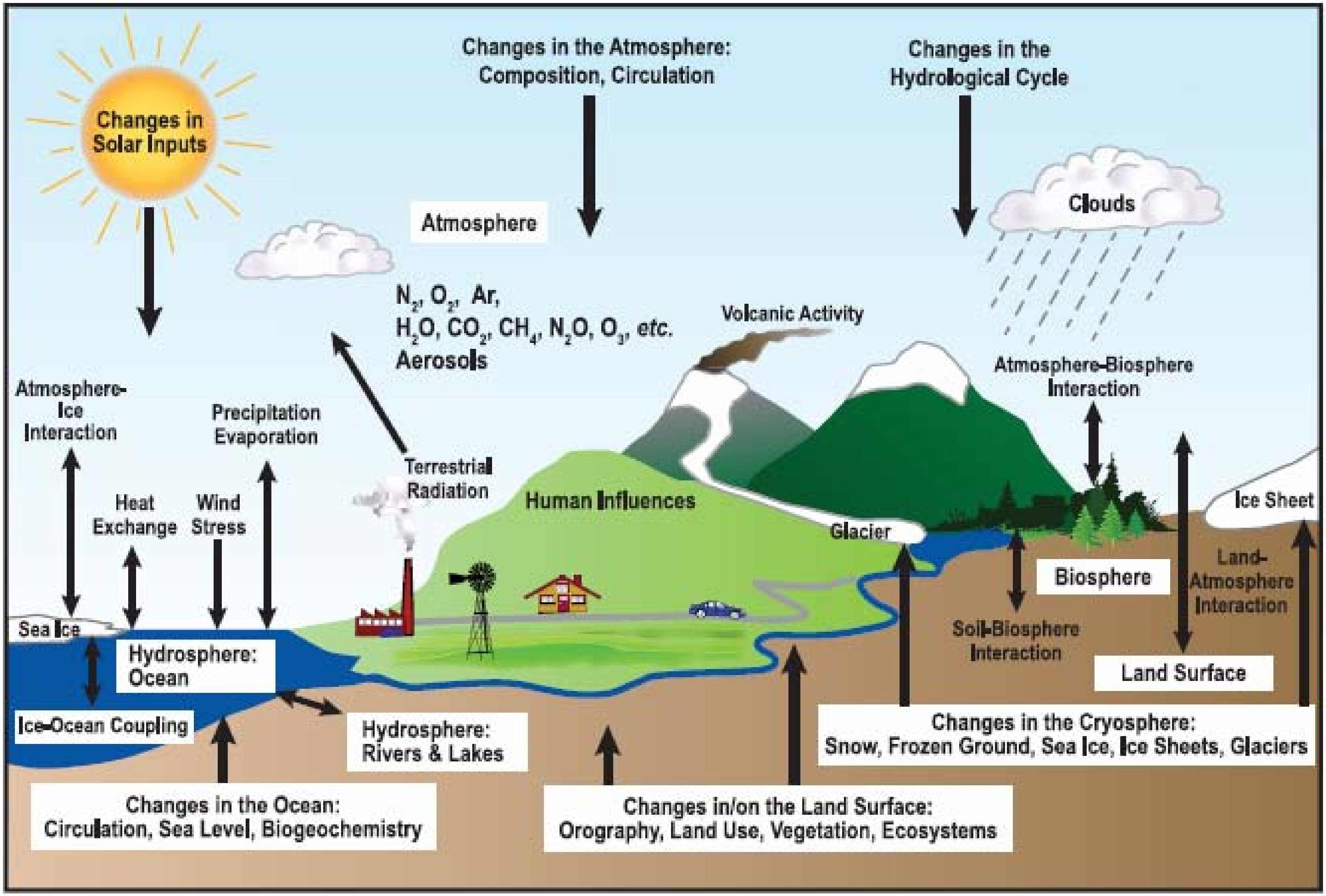
where c is the phase speed of the fastest wave in the system, Δt is the time step used by the model, and Δx is a characteristic spacing between grid points.

Typical size of model: 20 levels, grid points spaced ~120 km apart, 10-15 variables to defines state of atmosphere or ocean at each grid point: ~1,000,000-5,000,000 variables. Typical time step: 20 minutes. Thus 70,000,000 -350,000,000 variables calculated per simulated day.

Unresolved physical processes must be handled parametrically

- Convection
- Thin and/or broken clouds
- Cloud microphysics
- Aerosols and chemistry (e.g. photochemical processes, ozone
- Turbulence, including surface fluxes
- Sea ice
- Land ice
- Land surface processes

Forcings and Feedbacks in Climate Models



Forcings and Feedbacks

Consider the total flux of radiation through the top of the atmosphere:

$$F_{TOA} = F_{solar} - F_{IR}$$

Each term on the right can be regarded as function of the surface temperature, T_s, and many other variables x_i :

$$F_{TOA} = F_{TOA}\left(T_s, x_1, x_2, ...x_N\right)$$

By chain rule,

$$\delta F_{TOA} = 0 = \frac{\partial F_{TOA}}{\partial T_s}\delta T_s + \sum_{i=1}^{N}\frac{\partial F_{TOA}}{\partial x_i}\delta x_i$$

Now let's call the N^{th} process a "forcing", Q:

$$\delta F_{TOA} = 0 = \frac{\partial F_{TOA}}{\partial T_s}\delta T_s + \sum_{i=1}^{N-1}\frac{\partial F_{TOA}}{\partial x_i}\delta x_i + \delta Q$$

$$= \frac{\partial F_{TOA}}{\partial T_s}\delta T_s + \sum_{i=1}^{N-1}\frac{\partial F_{TOA}}{\partial x_i}\frac{\partial x_i}{\partial T_s}\delta T_s + \delta Q$$

Then

$$\frac{\partial T_s}{\partial Q} \equiv \lambda_R = -\frac{1}{\dfrac{\partial F_{TOA}}{\partial T_s} + \displaystyle\sum_{i=1}^{N-1}\frac{\partial F_{TOA}}{\partial x_i}\frac{\partial x_i}{\partial T_s}}$$

$$Let \quad S \equiv \left(-\frac{\partial F_{TOA}}{\partial T_s} \right)^{-1}$$

Climate sensitivity without feedbacks

$$\frac{\partial T_s}{\partial Q} \equiv \lambda_R = \frac{S}{1 - S \sum_{i=1}^{N-1} \frac{\partial F_{TOA}}{\partial x_i} \frac{\partial x_i}{\partial T_s}}$$

Climate sensitivity

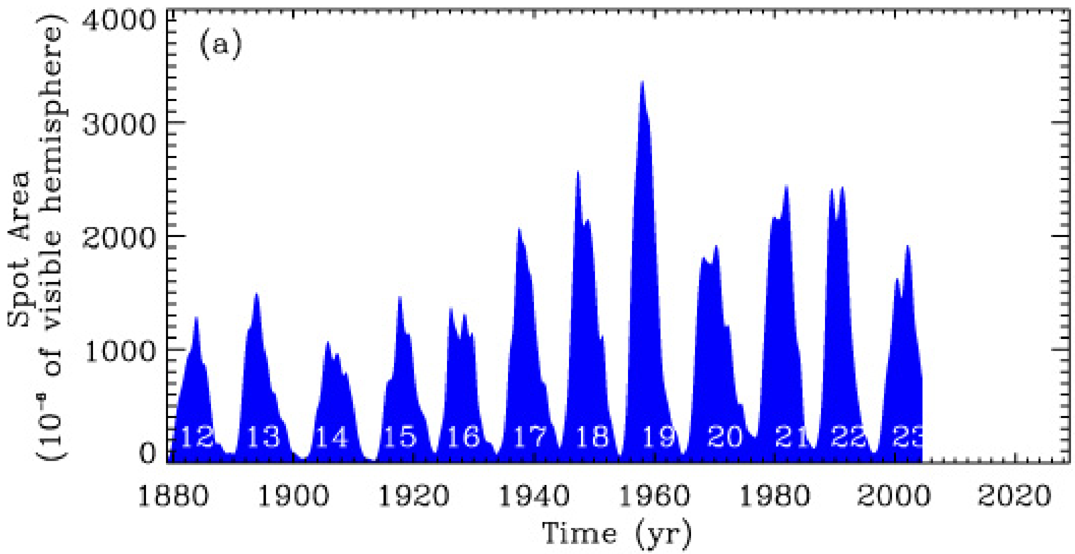
Feedback factors; can be of either sign

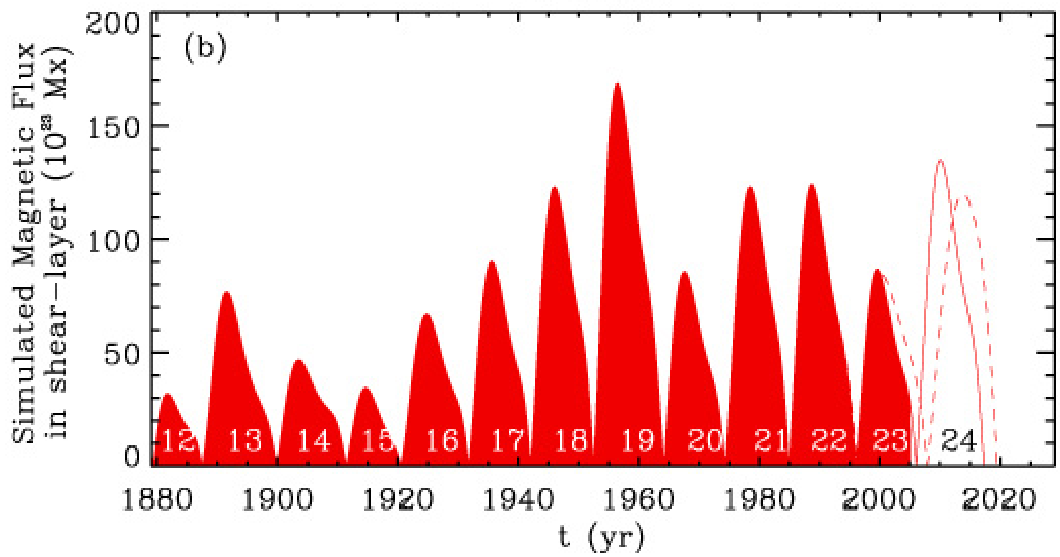
Note that feedback factors do NOT add linearly in their collective effects on climate sensitivity

Examples of Forcing:

- Changing solar constant
- Changing concentrations of non-interactive greenhouse gases
- Volcanic aerosols
- Manmade aerosols
- Land use changes

Solar Sunspot Cycle





Examples of Feedbacks:

- Water vapor
- Ice-albedo
- Clouds
- Surface evaporation
- Biogeochemical feedbacks

Estimates of Climate Sensitivity

$$\frac{\partial T_s}{\partial Q} \equiv \lambda_R = \frac{S}{1 - S\sum_{i=1}^{N-1} \frac{\partial F_{TOA}}{\partial x_i}\frac{\partial x_i}{\partial T_s}}$$

$$S \equiv \left(-\frac{\partial F_{TOA}}{\partial T_s} \right)^{-1}$$

Suppose that $T_s = T_e + constant$ and that shortwave radiation is insensitive to T_s:

$$F_{TOA} = -\sigma T_e^{\,4}, \quad \frac{\partial F_{TOA}}{\partial T_s} = -\frac{\partial}{\partial T_s}\sigma T_e^{\,4} = -4\sigma T_e^{\,3} = -3.8Wm^{-2}K^{-1}$$

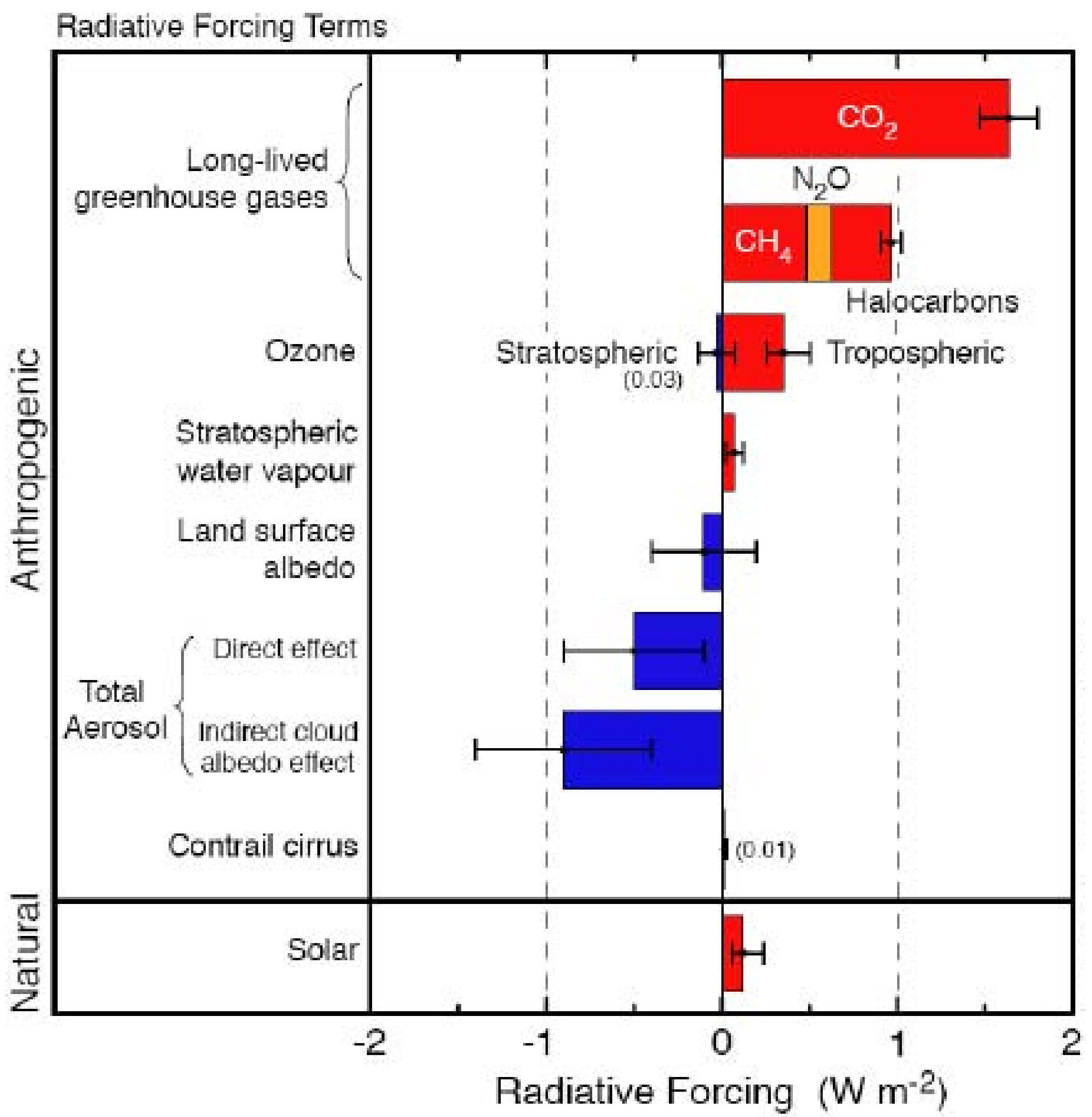
$$S = 0.26K\left(Wm^{-2}\right)^{-1}$$

Examples of Forcing Magnitudes:

- A *1.6%* change in the solar constant, equivalent to $4\ Wm^{-2}$, would produce about 1^oC change in surface temperature

- Doubling CO_2, equivalent to $4\ Wm^{-2}$, would produce about 1^oC change in surface temperature

Contributions to net radiative forcing change, 1750-2004:



Examples of feedback magnitudes:

- Experiments with one-dimensional radiative-convective models suggest that holding the relative humidity fixed,

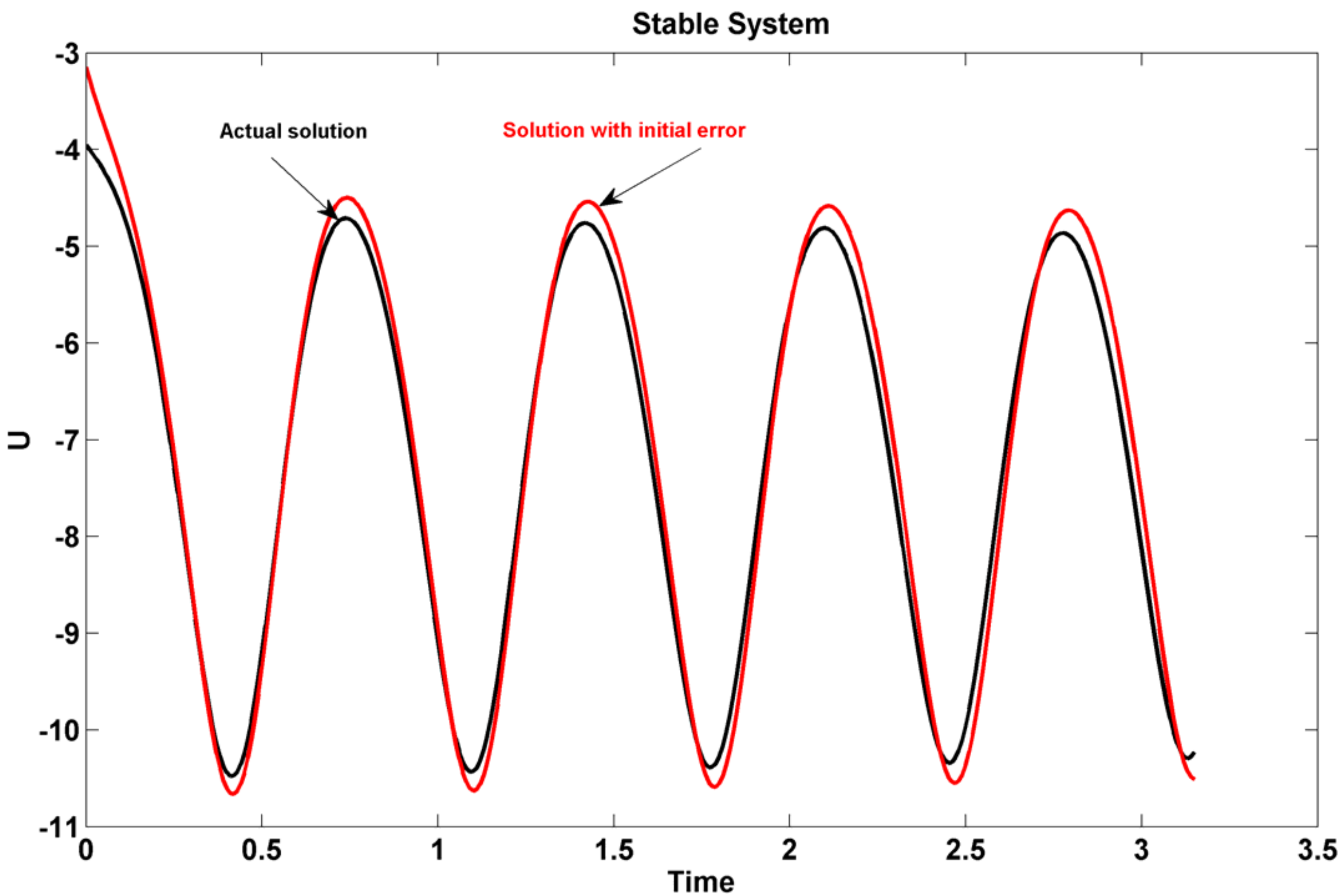
$$\left(\frac{\partial F_{TOA}}{\partial q}\right)\left(\frac{\partial q}{\partial T_s}\right)_{RH} \cong 2\,Wm^{-2}K^{-1},$$

$$S\left(\frac{\partial F_{TOA}}{\partial q}\right)\left(\frac{\partial q}{\partial T_s}\right)_{RH} \cong 0.5$$

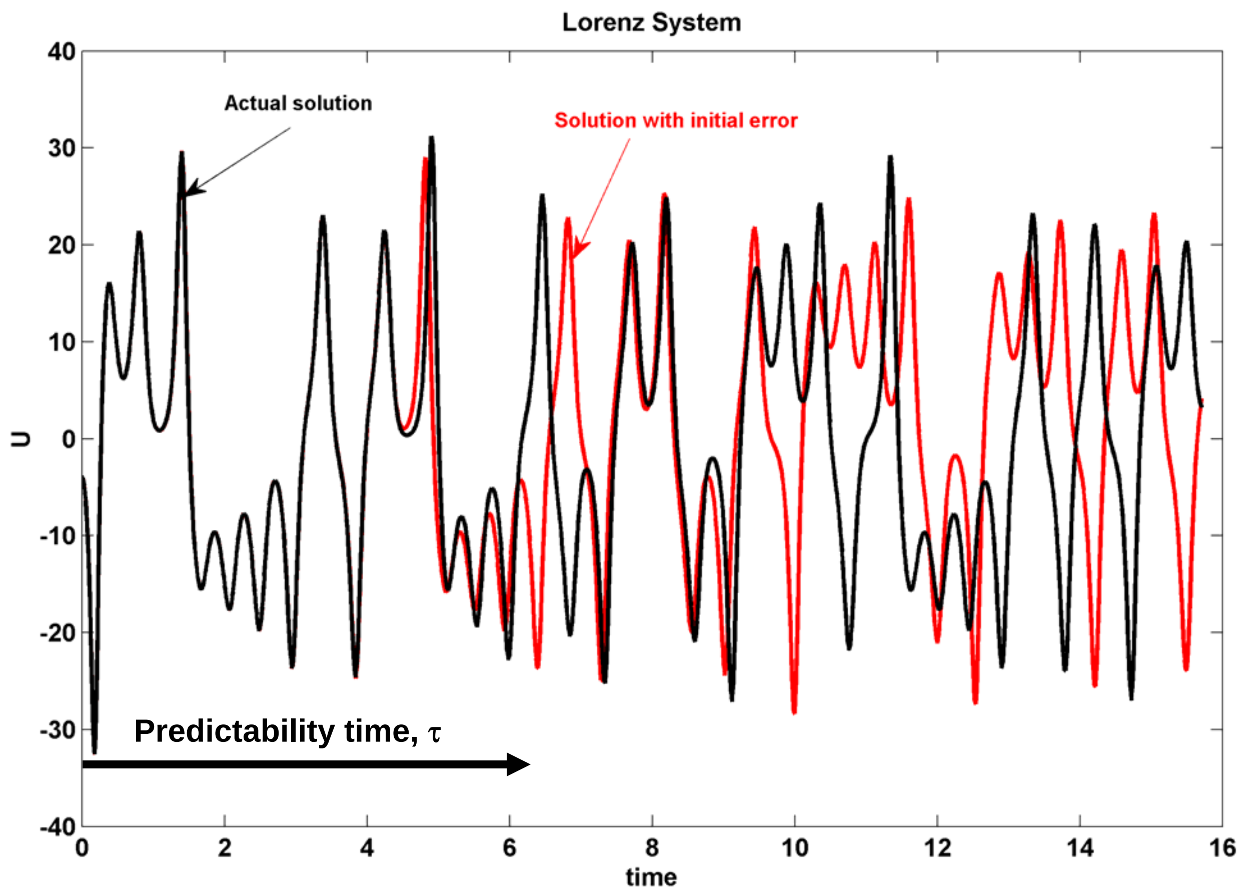
This, by itself, doubles climate sensitivity; with other positive feedbacks, effect on sensitivity is even larger

Free Natural Variability of the Climate System

Deterministic versus chaotic dynamics



Chaotic Dynamics

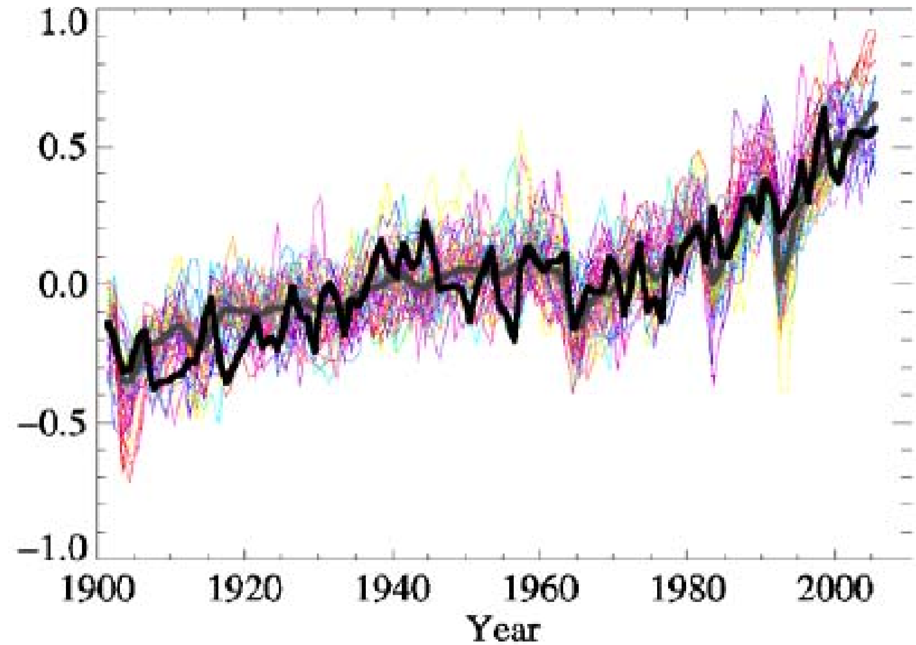


$$Note: \quad \lim_{\varepsilon \to 0}(\tau) = \tau_{pre} \neq 0$$

Climate chaos

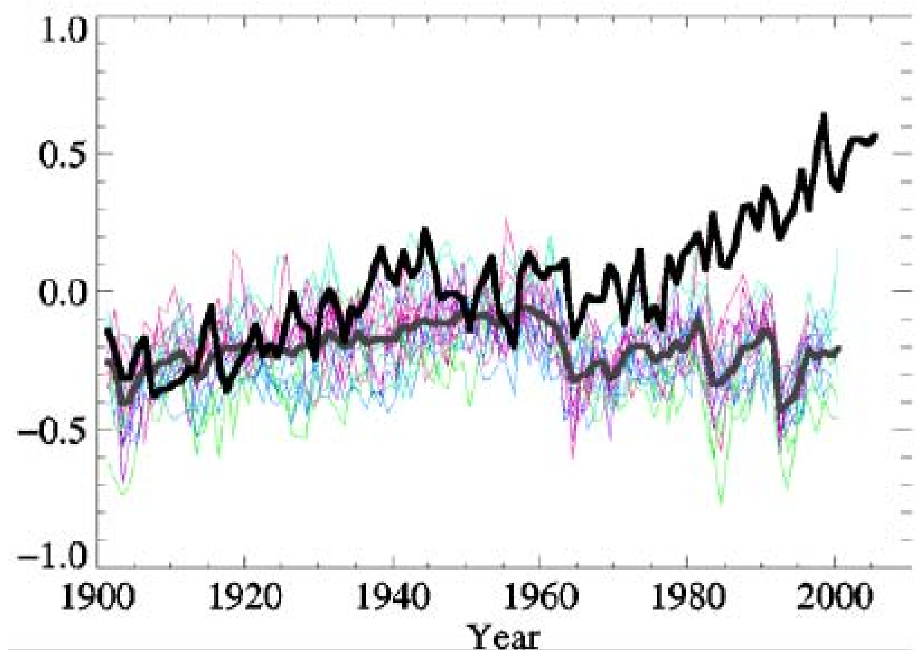
- Atmosphere known to be chaotic on time scales at least as large as several months
- Ocean known to be chaotic on time scales of at least 6 months and perhaps as long as hundreds of years
- Coupled atmosphere-ocean system may be chaotic on time scales as long as several thousand years

Global mean temperature (black) and simulations using many different global models (colors) including all forcings



To quantify natural, chaotic variability, necessary to run large ensembles

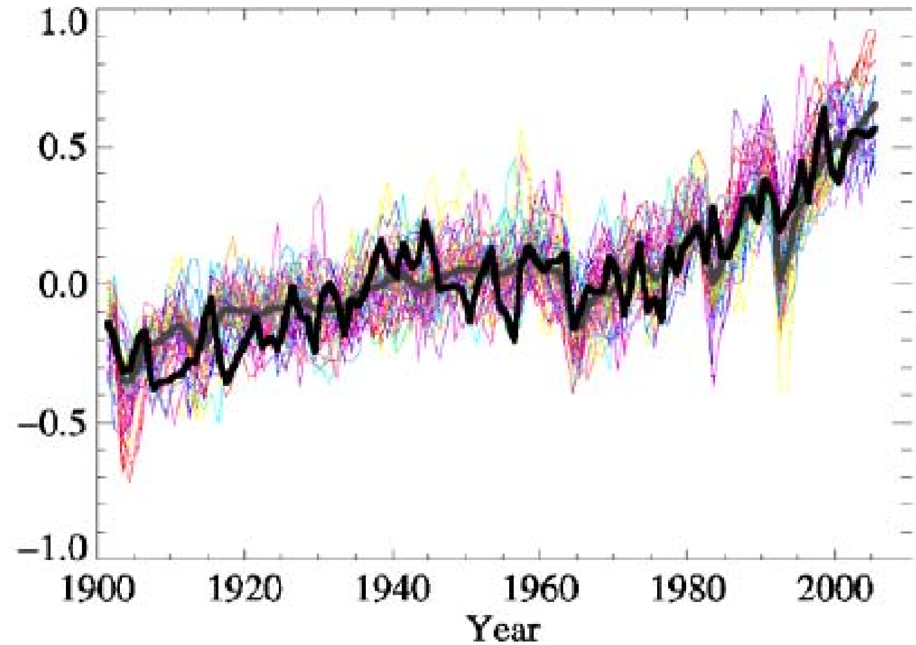
Same as above, but models run with only natural forcings



How Do We Know If We Have It Right?

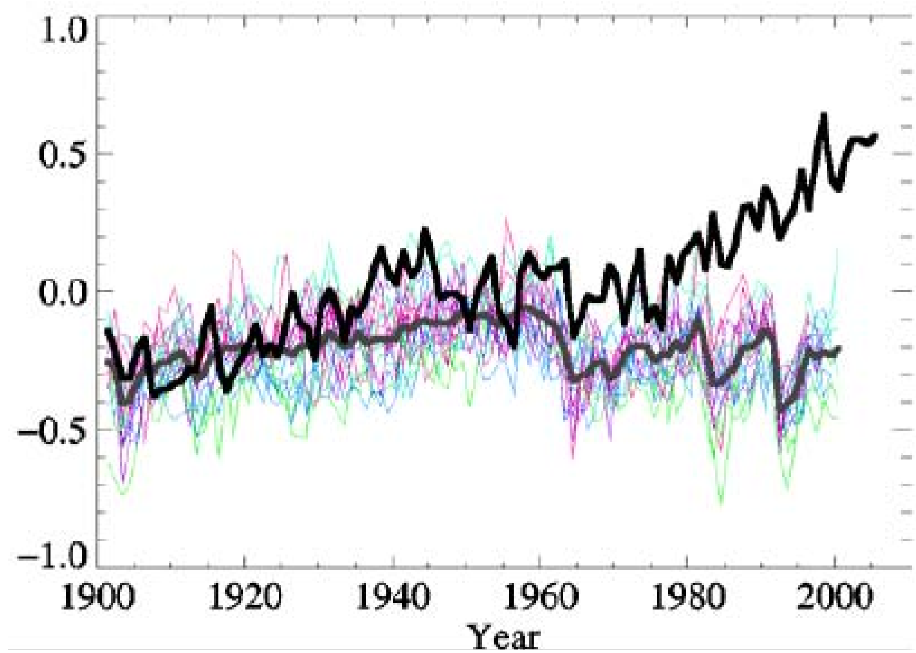
- Very few tests of model as whole: annual and diurnal cycles, 20^{th} century climate, weather forecasts, response to orbital variations
- Fundamentally ill-posed: Far more free parameters than tests
- Alternative: Rigorous, off-line tests of model subcomponents. Arduous, unpopular: Necessary but not sufficient for model robustness: Model as whole may not work even though subcomponents are robust

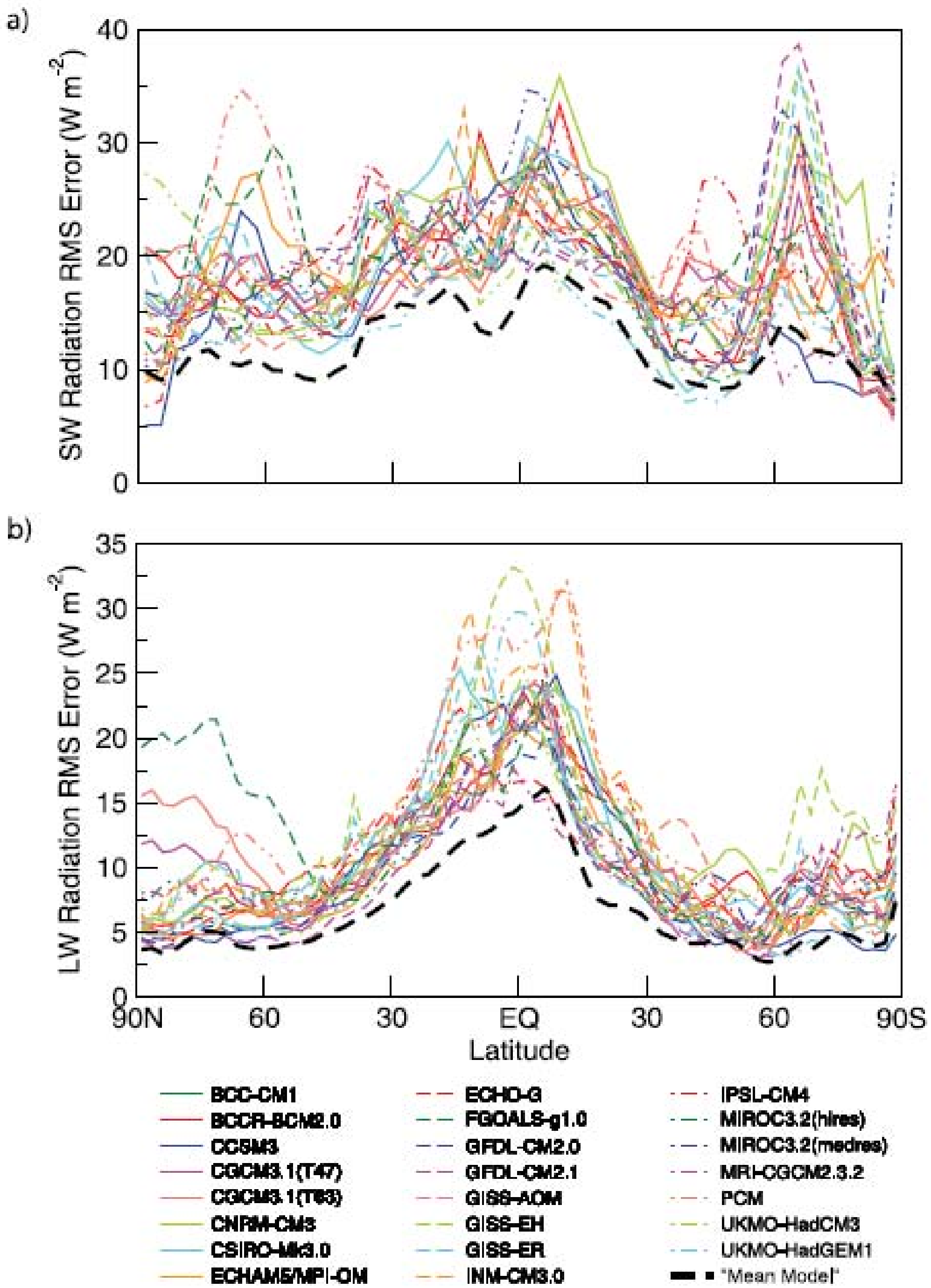
Global mean temperature (black) and simulations using many different global models (colors) including all forcings



To some extent, "success" of 20th century simulations is a result of model curve fitting

Same as above, but models run with only natural forcings





Root-mean-square error in zonally and annually averaged SW radiation (top) and LW radiation (bottom) for individual AR4 models (colors) and for ensemble mean (black dashed)

Observed time mean, zonally averaged ocean temperature (black contours), and model-mean minus observed temperature (colors) for the period 1957-1990