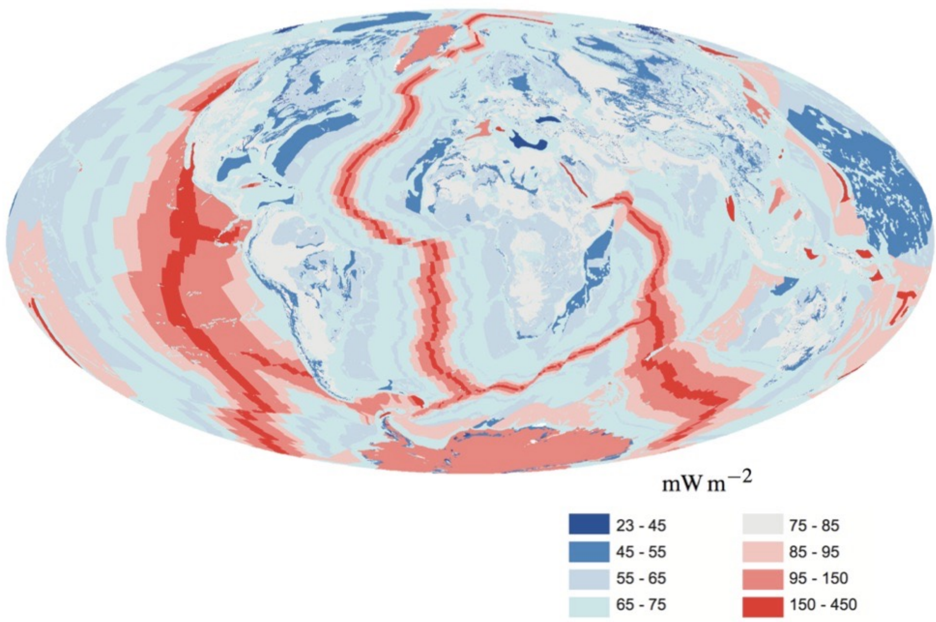
What determines Earth's surface temperature?

Heat flow from Earth's interior is ca. 47 TW



Amount of solar radiation striking the Earth is ca. 173,000 TW
~ 99.97 % of total energy flow to Earth's surface

Energy Fundamentals

1) Energy = Ability to produce change in the state or motion of matter.
   Or the ability to do work, where W = F * D & F = M * A

2) Energy comes in many interchangeable forms: mechanical, electrical, chemical, sound, heat, electromagnetic radiation

3) First law of thermodynamics – Energy in never created or destroyed.

Heat = random, incoherent motion of atoms and molecules.

The more modes of motion, the more heat energy atoms/molecules can transport.

The amount of energy per mode of motion = ½ kT

K = Boltzmann constant = $1.38*10^{-23}$ J/K
T = temperature (K)

Electromagnetic Radiation (EMR)

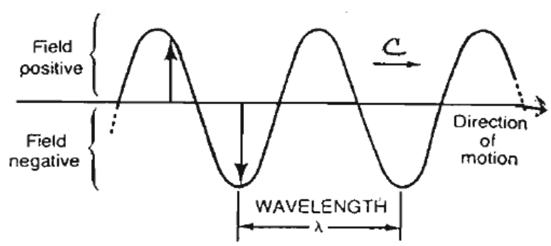
1) Consists of photons – discrete packets of E that can be considered a "ripple" in the electromagnetic field of space.

2) EMR can transport energy through the near vacuum of space.

3) Photons are emitted or absorbed as a result of relatively discrete changes in the electronic energy levels of atoms or the vibrational & rotational energy levels of molecules.

4) To be maintained, these "ripples" (photons) move at the speed of light (3×10^8 m/s)

5) Photons can be described as travelling waves that are characterized by their wavelength.



$$\nu = c/\lambda$$
$$\mathrm{E} = h * (c/\lambda)$$
$$\mathrm{E} = h * \nu$$

ν = frequency (cycles/s or Hz)
c = speed of light = 3×10^8 m/s
λ = wavelength (m/cycle)
E = energy content per photon (Joules)
h = Planck's constant = 6.63×10^{-34} J*s/cycle

Type of radiation	Wavelength range
radio	1–10 m
radar (microwaves)	1–30 cm
infrared	0.71–100 μ
visible	0.40–0.71 μ
ultraviolet	0.10–0.40 μ
X-rays	10^{-5}–10^{-2} μ

Definitions and characteristics of the various wavelength regions of light. The ranges of wavelengths leading to the sensation of a particular color are somewhat arbitrary and vary with individuals. Both frequencies and energies in the table refer to the particular wavelength indicated in column 3 for each wavelength interval. Wavelength magnitudes are those in a vacuum.

C Color	Approximate wavelength range (nm)	Representative wavelength (nm)	Frequency (cycles s^{-1}, or hertz)	Energy (kJ mol^{-1})
Ultraviolet	below 400	254	11.80×10^{14}	471
Violet	400 to 425	410	7.31×10^{14}	292
Blue	425 to 490	460	6.52×10^{14}	260
Green	490 to 560	520	5.77×10^{14}	230
Yellow	560 to 585	570	5.26×10^{14}	210
Orange	585 to 640	620	4.84×10^{14}	193
Red	640 to 740	680	4.41×10^{14}	176
Infrared	above 740	1 400	2.14×10^{14}	85

Blackbody – An object that absorbs all EMR that strikes it and emits the maximum amount of EMR that is possible for each wavelength emitted by an object at a given temperature.

Radiation Laws

The Stephan-Boltzmann Law

For a blackbody:

$$\mathbf{E = \sigma T_e^4}$$

Where: E = energy flux emitted by a blackbody ($J\ s^{-1} m^{-2}$)
σ = Stephan-Boltzmann constant = 5.67 x 10^{-8} ($J\ m^{-2}\ s^{-1}\ K^{-4}$)
T_e= temperature (K)

↳ °C + 273

For a "gray" body:

$$\mathbf{E = \epsilon \sigma T_e^4}$$

Where: ϵ = emissivity

For a blackbody ϵ = 1.
Most natural surfaces have long-wave emissivities between 0.90 and 0.98.

Planck's Law

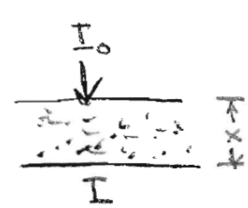
$$\mathbf{E(\lambda) = (2\pi hc^2)/(\lambda^5[e^{(hc/\lambda kTe)}-1)}$$

Where: E (λ) = ($W\ m^{-2}$)/m
c = 3 x 10^8 m/s
h = 6.63 x 10^{-34} J*s
k = 1.38 x 10^{-23} J*K-1
T_e = K = effective radiating temperature
λ = m

Wien's Displacement Law:

$$\lambda_{max} = \mathbf{2897}\ (\mu m\ \ K)/\ T_e(\ \ K)$$

Beer's Law (Beer-Bouger Law):



$$\mathbf{I/I_o = e^{-kx}}$$

Where: I = intensity of transmitted radiation
I_o = incident intensity of radiation
e = base of the natural logarithm
k = extinction coefficient
x = the distance the beam of light travels through the medium

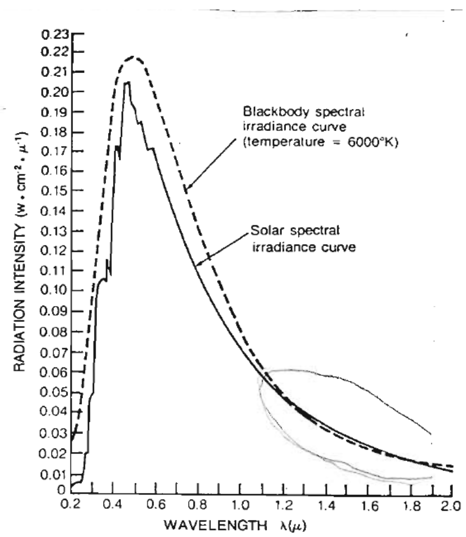
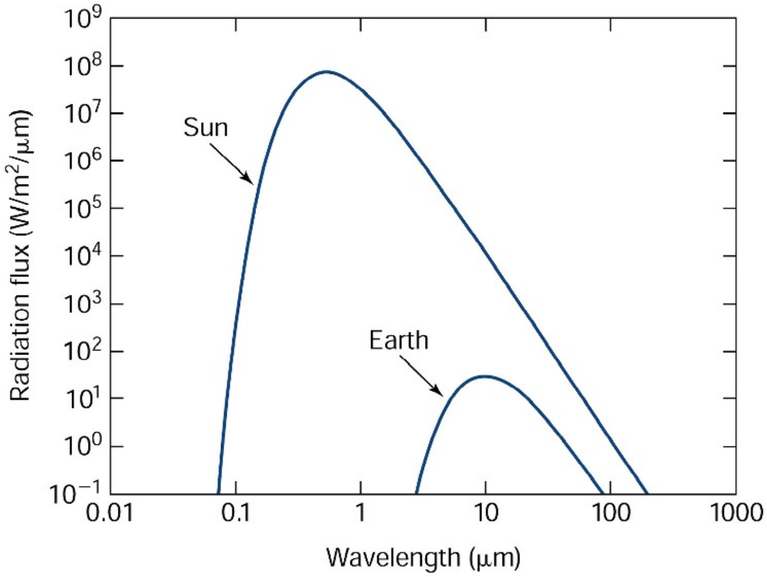


FIGURE 2-11

Solar irradiance spectrum and 6000°K blackbody radiation reduced to mean solar distance. (From Neiburger, Edinger, and Bonner, 1973.) Ehrlich et al. 1977



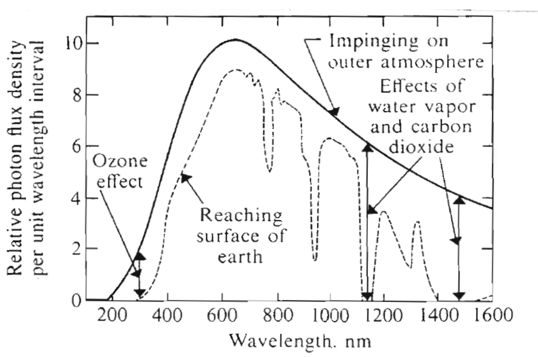


Figure 4.2

Wavelength distributions of the sun's photons incident on the earth's atmosphere and its surface. The curve for the solar irradiation on the atmosphere is an idealized one based on Planck's radiation distribution formula. The spectral distribution and the amount of solar irradiation reaching the earth's surface depend on clouds, other atmospheric conditions, altitude, and the sun's angle in the sky. The pattern indicated by the lower curve is appropriate at sea level on a clear day with the sun overhead (for further details see Bickford and Dunn, 1972; Gates, 1980; Monteith, 1973; and Seliger and McElroy, 1965). Nobel, P.S. 1991

TABLE 2-12

Absorption of Solar Radiation by Atmospheric Gases

Wavelength range (μ)	*Fate of radiation*
UNTRAVIOLET	
Less than 0.12	All absorbed by O_2 and N_2 above 100 km
0.12–0.18	All absorbed by O_2 above 50 km
0.18–0.30	All absorbed by O_3 between 25 and 50 km
0.30–0.34	Part absorbed by O_3
0.34–0.40	Transmitted to Earth almost undiminished
VISIBLE	
0.40–0.71	Transmitted to Earth almost undiminished
INFRARED	
0.71–3	Absorbed by CO_2 and H_2O, mostly below 10 km

Source: Neiburger, Edinger, and Bonner, *Understanding our atmospheric environment.*

Ehrlich et al. 1977

How Light Interacts with the Atmosphere

Light has three fates:

1) transmitted (doesn't interact)
2) absorbed (increased E state of atoms or molecules)
3) scattered/reflected (redirected) without being absorbed

Absorption:

Gases are "choosy" absorbers, condensed matter is less "choosy"

Photons interact with & increase the internal energy state of atoms or molecules through a change in:

e^- excitation
rotational energy
vibrational energy (compression, stretching, & bending)

New energy states are discrete

Limited to certain energy values (frequencies/wavelengths)
Frequency/wavelength specific
$\Delta E_{vib} = E_{photon} = h\nu = hc/\lambda$

https://www.youtube.com/watch?v=aCocQa2Bcuc

To absorb a photon of EMR, the atom/molecule must be able to produce an oscillating electric field with a frequency that matches the frequency of the electric field of the photon. They must be "in tune" with each other.

Molecules can produce oscillating electric fields if their vibrations create an asymmetrical charge distribution (a dipole).

Absorbed EMR energy can be dissipated in several ways including:

Emission of photons of EMR
*Radiationless transitions that produce **heat***

How Earthlight Interacts with Greenhouse Gases

Wavelength/frequency matters

Greenhouse gases are transparent to (not in tune with) solar radiation

BUT they are not transparent to wavelength-specific **portions** of terrestrial radiation.

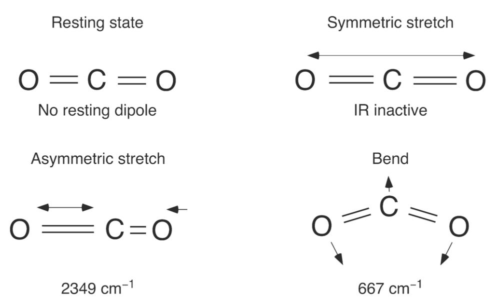


Figure 4-1 Vibrational modes of a CO_2 molecule that interact with infrared light in the atmosphere.

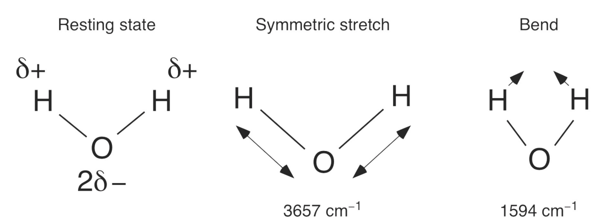
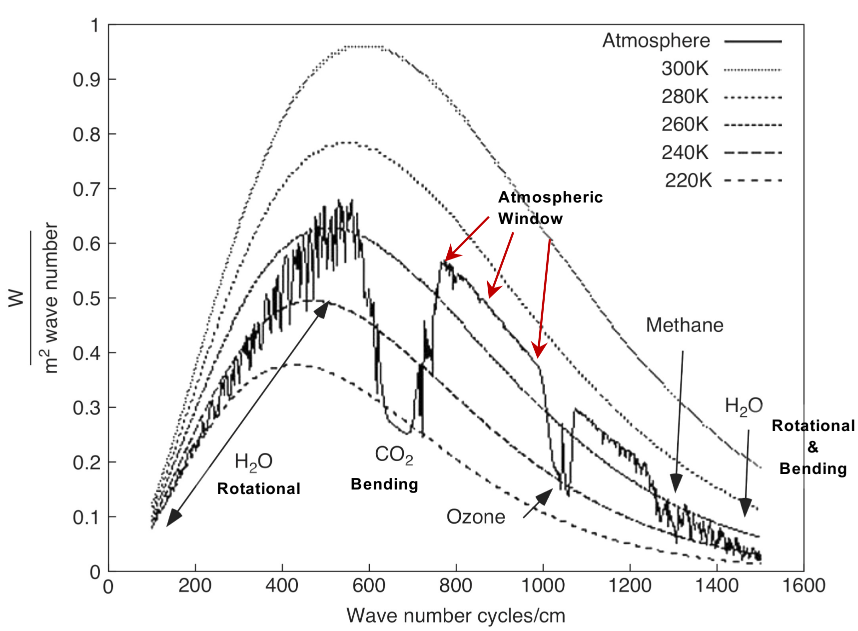
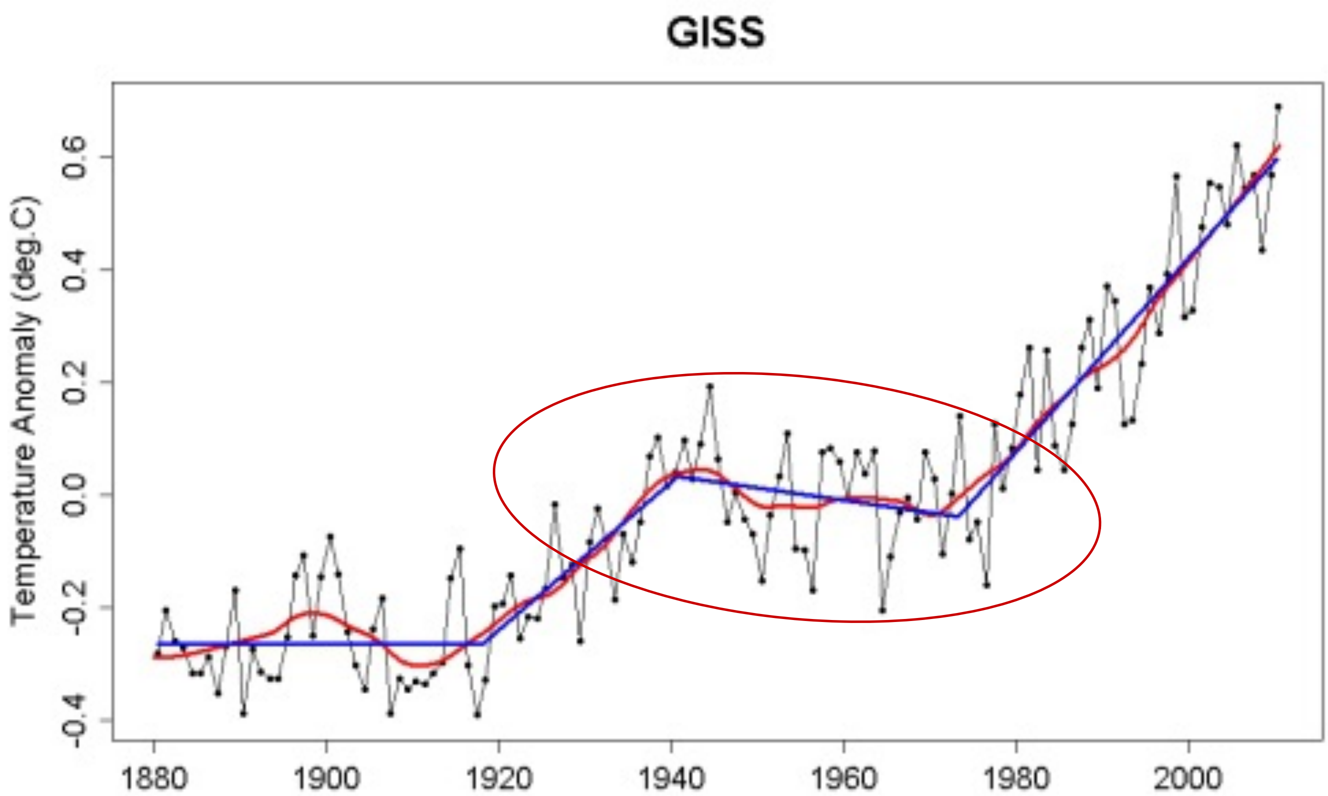


Figure 4-2 Vibrational modes of a water molecule that interact with infrared light in the atmosphere.



Both absorption and scattering of light affect a planet's temperature.



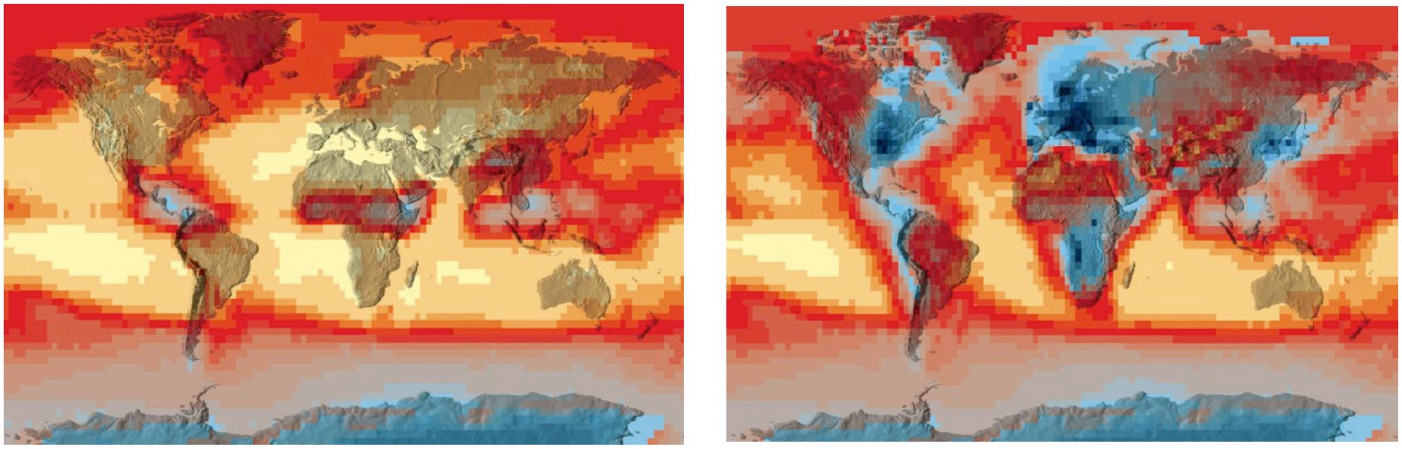
Climate scientists believe that the primary cause of this mid-century cooling was an increase in atmospheric aerosols due to anthropogenic emissions (primarily from the burning of fossil fuels). Aerosols have a complex effect on the climate, because they have both direct and indirect impacts.

Direct Effect

The direct effect of aerosols on climate is the mechanism by which aerosols scatter and absorb shortwave and longwave radiation (a.k.a. "global dimming"), thereby altering the radiative balance of the Earth-atmosphere system. The key parameters for determining the direct aerosol radiative forcing are the aerosol optical properties and distribution in the atmosphere (IPCC 2007).

Indirect Effect

The indirect effect of aerosols on climate is the mechanism by which they modify the microphysical and, therefore, radiative properties, amount, and lifetime of clouds. A key parameter for determining the indirect effect of aerosols on the global surface temperature is the effectiveness of an aerosol particle to act as a cloud condensation nucleus - a function of the aerosol size, chemical composition, mixing state, and ambient environment (IPCC 2007).



CLIMATIC FORCING by human activity is evident in calculations of global heat gain during the Northern summer. Every July greenhouse gases warm the earth by about 2.2 watts per square meter (*left*); the effect is most pronounced over the warm areas of the subtropics. When the cooling by sulfate aerosol is included, however, the forcing drops to about 1.7 watts per square meter (*right*). In fact, the cooling dominates over industrial regions in the Northern Hemisphere.

How Much Light Do Aerosols Reflect Away?

Atmospheric sulfate aerosol scatters light in all directions. About 15 to 20 percent of the light is scattered back into space. The backscattering constitutes the direct effect of atmospheric aerosol on incoming radiation. The light-scattering efficiency of aerosol, represented by the Greek letter alpha (α), is high, even at low humidity: each gram represents an area of about five square meters. Moisture increases the scattering by making the aerosol expand. At the global average relative humidity, the efficiency doubles, to almost 10 square meters per gram. One can use this value to estimate the magnitude of the direct effect of anthropogenic sulfate.

The rate at which light is lost from the solar beam is defined by the scattering coefficient, represented by the Greek letter sigma (σ, expressed in units of per meter). This value is determined by the amount of aerosol mass, M (in grams per cubic meter), multiplied by the light-scattering efficiency: $\sigma = \alpha M$. When both sides of this equation are integrated over altitude, z, a dimensionless quantity called the aerosol optical depth and represented by the Greek letter delta (δ), results:

$$\int_0^\infty \sigma \, dz = \delta = \alpha \int_0^\infty M dz = \alpha B$$

Here B is the world average burden of anthropogenic sulfate aerosol in a column of air, in grams per cubic meter. The optical depth is then used in the Beer Law (which describes the transmission of light through the entire vertical column of the atmosphere). The law yields $I/I_o = e^{-\delta}$, where I is the intensity of transmitted radiation, I_o is the incident intensity outside the atmosphere and e is the base of natural logarithms. In the simplest case, where the optical depth is much less than 1, δ is the fraction of light lost from the solar beam because of scattering. The question, then, is just how large δ is or, more properly, that part of it that results from man-made sulfate.

This global average burden of anthropogenic sulfate aerosol can be estimated by considering the entire atmospheric volume as a box. Because the lifetime of sulfate aerosol is short, the sum of all sulfate sources, Q, and its lifetime in the box, t, along with the area of the earth, determine B:

$$B = \frac{Qt}{\text{area of the earth}}$$

About half the man-made emissions of sulfur dioxide become sulfate aerosol. That implies that currently 35 teragrams (35×10^{12} grams) per year of sulfur in sulfur dioxide is converted chemically to sulfate. Because the molecular weight of sulfate is three times that of the elemental sulfur, Q is about $(3)(35 \times 10^{12})$ or 1.1×10^{14} grams per year. Studies of sulfate in acid rain have shown that sulfates persist in the air for about five days, or 0.014 year. The area of the earth is 5.1×10^{14} square meters. Substituting these values into the equation for B yields about 2.8×10^{-3} gram per square meter for the burden.

This apparently meager amount of material produces a small but significant value for the aerosol optical depth. Using the value of scattering efficiency (α) of five square meters per gram and a factor of two for the increase in scattering coefficient because of relative humidity, the estimated anthropogenic optical depth becomes $\delta \approx 5 \times 2 \times (2.8 \times 10^{-3}) \approx 0.028$. This value means that about 3 percent of the direct solar beam fails to reach the earth's surface because of man-made sulfate. A smaller amount—perhaps (0.15)(3 percent), or about 0.5 percent—is thus lost to space. This scattering operates over the noncloudy parts of the earth. About half the earth is cloudy at any given time, so that globally 0.2 to 0.3 percent is lost.

Charlson and Wigley 1994

How much light do human-produced sulfate aerosols directly reflect to space?

For sulfate aerosols:

$I/I_o = e^{-\delta x}$ Beer's Law with x = 1 because we're considering the entire column of air

Where:

Optical depth = scattering effectiveness of column = $\boldsymbol{\delta = \alpha * \beta}$

α = Empirically know scattering efficiency

β = Average amount of aerosol in column of air

β = Q*t/area ➔ Amount = (Input rate * Mean residence time)/area

Where:

Q = Total input (amount per unit time) => measured

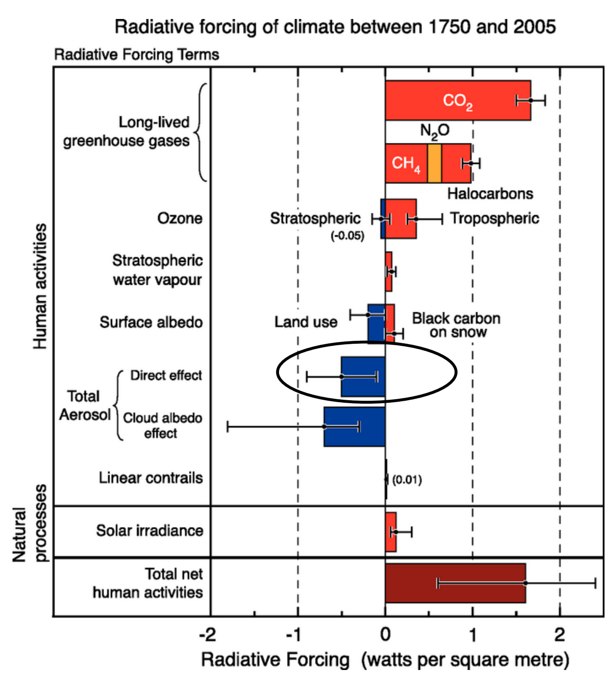
t = Mean residence time (years) => known

Determine δ and plug into Beer's Law to estimate fraction scattered.

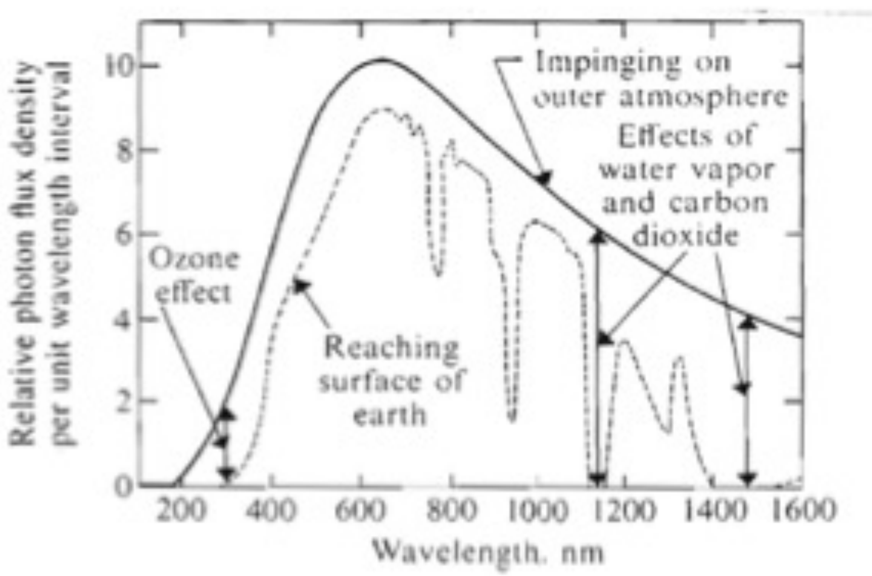
Correct for amount scattered to space (15%) & cloud-free cover (~50%)

When appropriate values are used:

~0.21% of sunlight (340 W/m^2) is directly scattered back to by anthropogenic sulfate aerosols (~0.71 W/m^2).



Both absorption and scattering of light affect both of the quantity and quality of sunlight reaching the surface of our planet in important ways.



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Why is the sky blue & the clouds white?

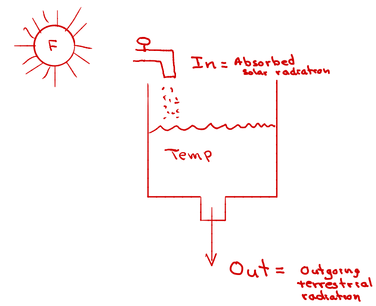
Rayleigh Scattering when scattering material << λ

Scattering efficiency α $1/\lambda^4$

Mie Scattering when scattering material ≥ λ

All λ scattered equally

For a “naked” planet acting as a blackbody

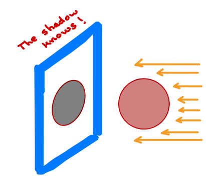


In = Out

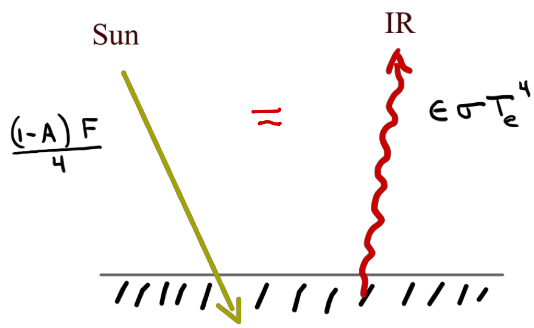
Absorbed solar R = Outgoing terrestrial IR

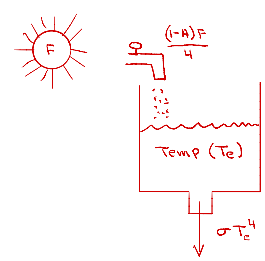
OVERALL RADIATION BALANCE FOR A PLANET

$$\underbrace{4\,\pi\, r^2\, \sigma\, T_e^{\,4}}_{\text{Out}} = \underbrace{\pi\, r^2\, (1\text{-}A)\, (F/R^2)}_{\text{In}}$$



Where: σ = Stefan-Boltzmann constant = 5.67 x 10^{-8} J s^{-1} m^{-2} K^{-4}
r = radius of the planet
A = albedo = ratio of reflected to incident solar radiation
F = solar constant = 1360 J s^{-1} m^{-2} for Earth
R = distance from sun in astronomical units (R = 1 for Earth)





$$T_e = \left[\frac{(1-A)F}{4} \cdot \frac{1}{\sigma}\right]^{1/4}$$

For Earth

$$T_e = \left[\frac{(1-A)F}{4} \cdot \frac{1}{\sigma} \cdot \frac{1}{R^2}\right]^{1/4}$$

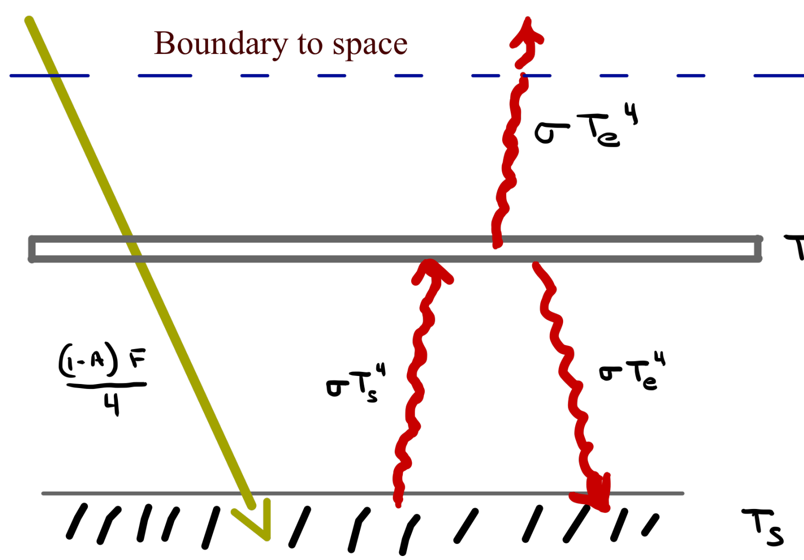
For any planet

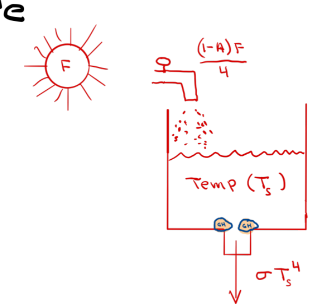
Calculated, measured, and surface temperatures.

	R	A	T(K)	T_m(K)	T_s(K)
Venus	0.72	0.77	227	230	750
Earth	1.00	0.30	255	250	288
Mars	1.52	0.15	216	220	240
Jupiter	5.20	0.58	98	130	134

A partially "clothed" planetary climate model

One-layer Greenhouse Model





For the atmosphere in a one-layer model

In = Out

$\cancel{\sigma} T_s^4 = 2\cancel{\sigma} T_e^4$

$T_s = 2^{1/4}\ T_e$

$= 303\ K$

Too hot !

In general we can determine that for layers (radiation zones) behaving as black bodies

$\boldsymbol{T_s = (n+1)^{1/4}\ T_e}$

where n = # of radiation zones & T_e stays constant

*It has been determined that **Earth's atmosphere has about two radiation zones**.*

What happens if we keep adding layers to the atmosphere?

The layer model is too simple and needs to include other features that cool off surface temperatures.

These include:

Heat transport to atmosphere by convection (atmospheric circulation).

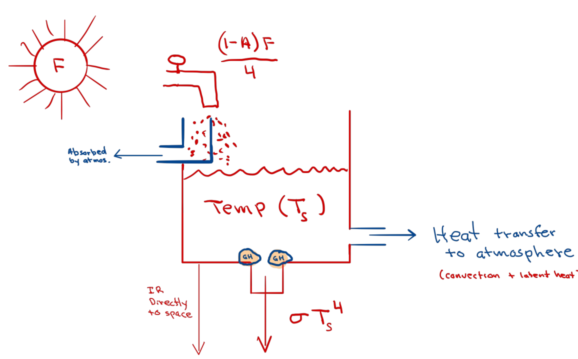
Latent heat transport to atmosphere by evaporated water.

IR emission from the surface that escapes directly to space.

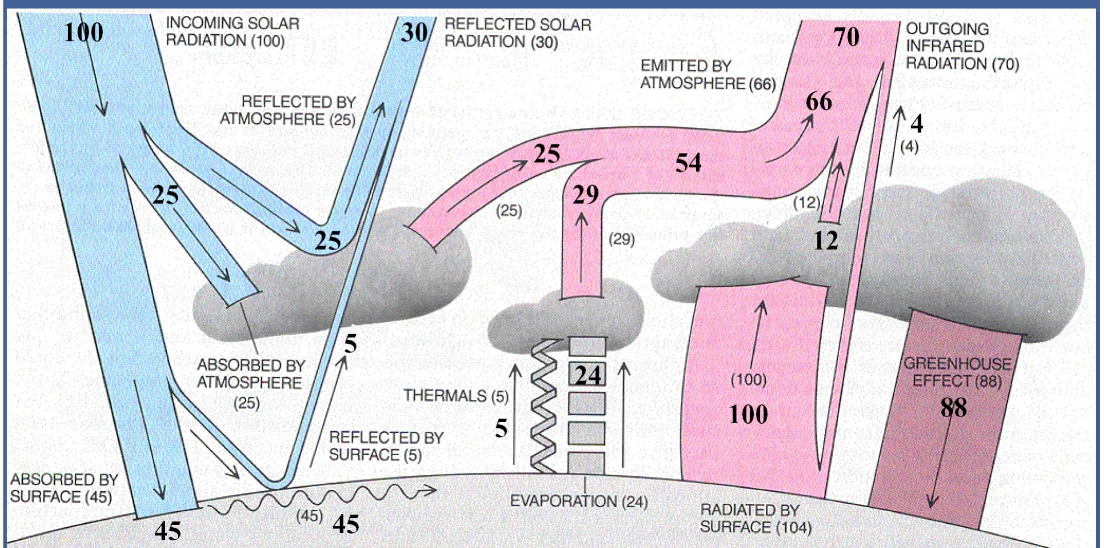
Absorption of solar radiation by materials in the atmosphere.

When these features are included, our models predict

$$T_s = 288.4\ K$$

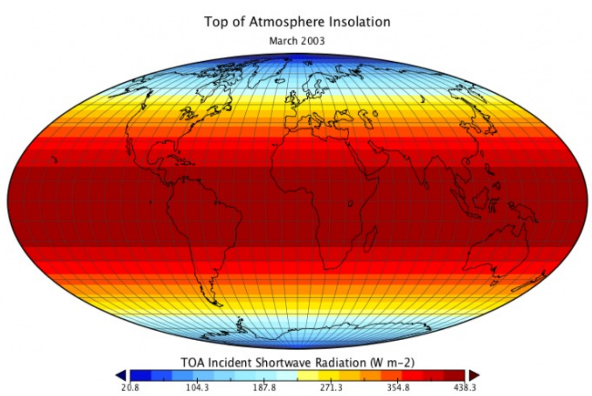


Energy Budget for Earth



Notes:
The total E absorbed by Earth & Atmosphere (70 %) powers processes on Earth.
Units are % of incoming solar radiation entering the top of the atmosphere.

Solar radiation is not evenly distributed in space & the variations are important.



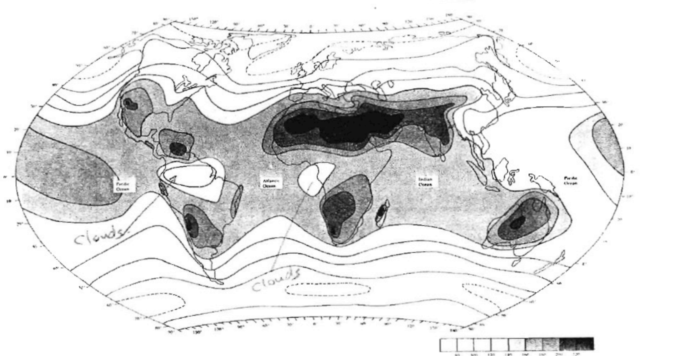
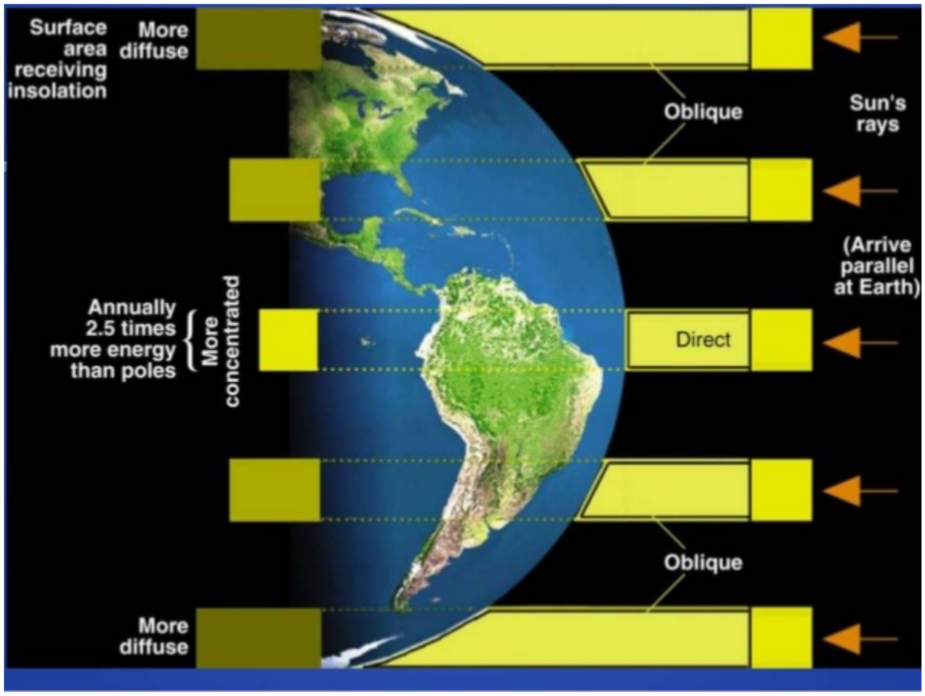
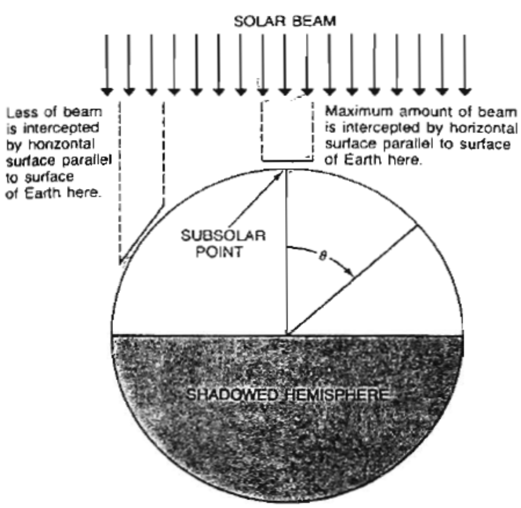
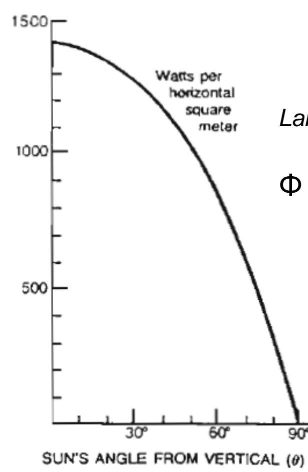


Figure 13-4. World map showing the distribution of solar radiation over the earth's surface in kcal cm^{-1} yr^{-1}. (From Landsberg, H. E., et al. 1966. *World Maps of Climatology.* By permission of Springer-Verlag.) Barbour et al. 1980



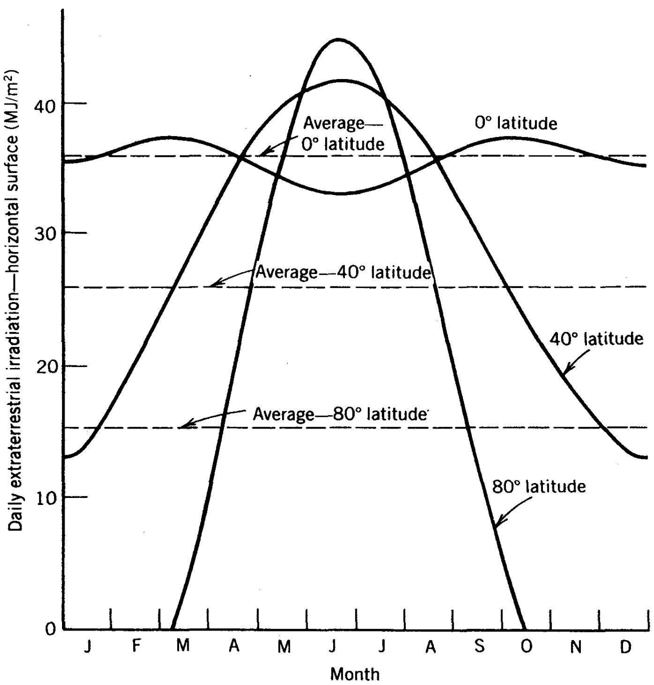




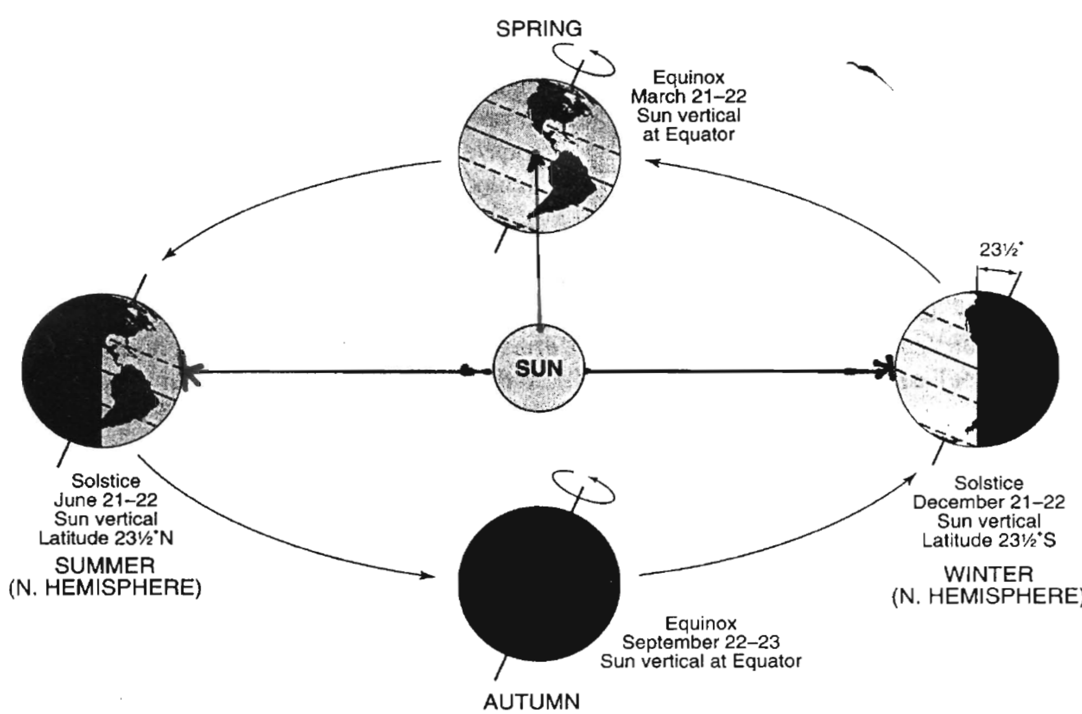
Lambert's Cosine Law

$$\Phi = \Phi_o \cos \theta$$

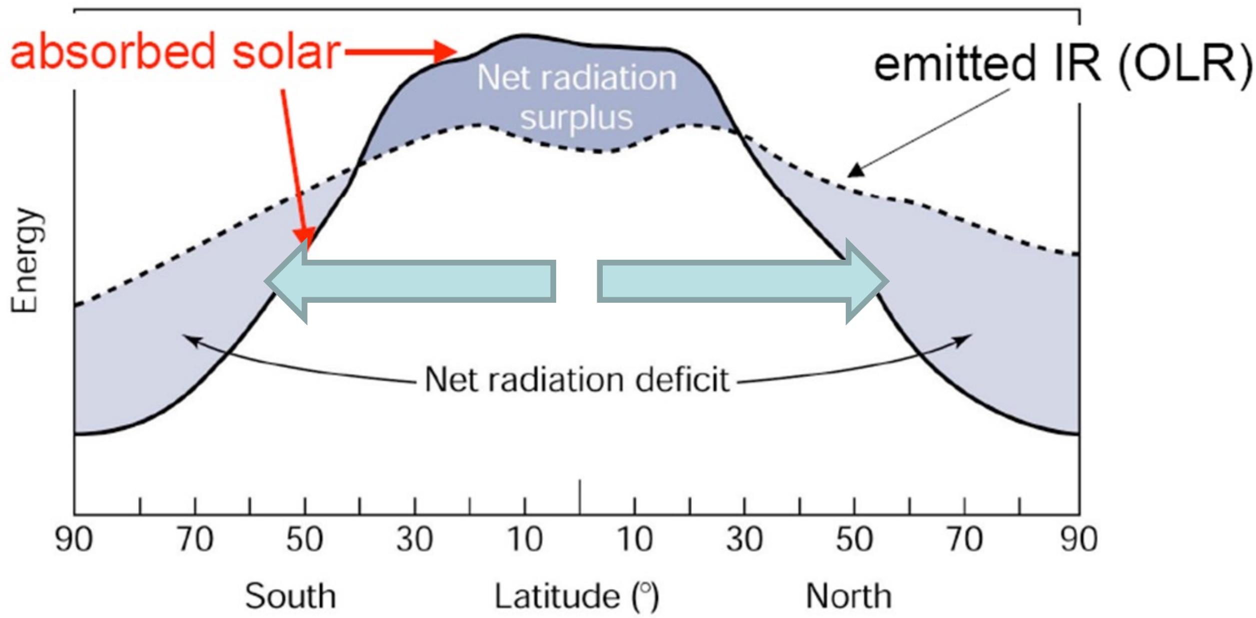
Seasonal variation of solar flux on horizontal surface outside the atmosphere in the N. Hemisphere

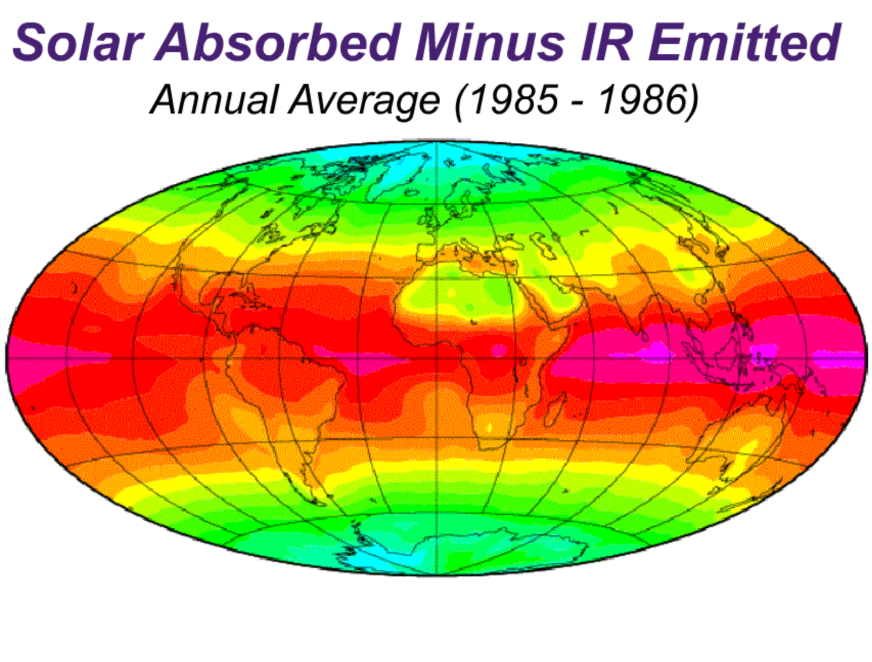


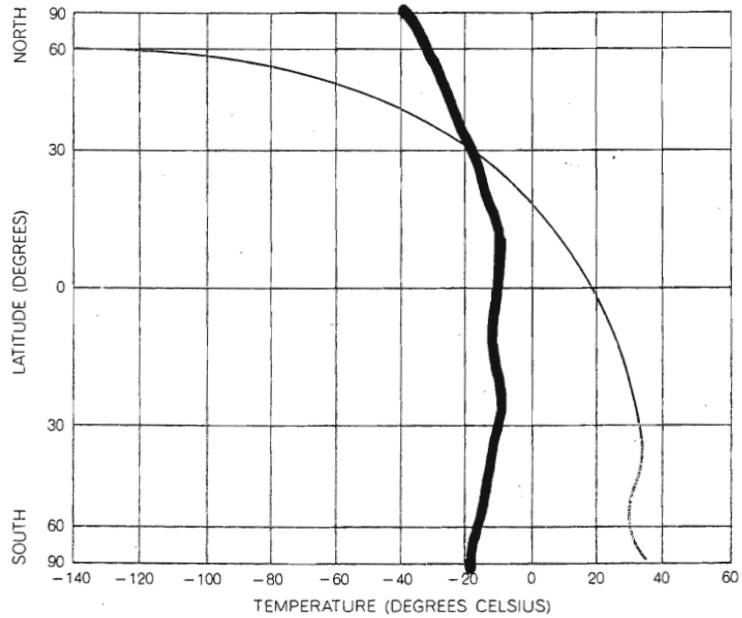
Earth's seasons caused by tilt, not distance.



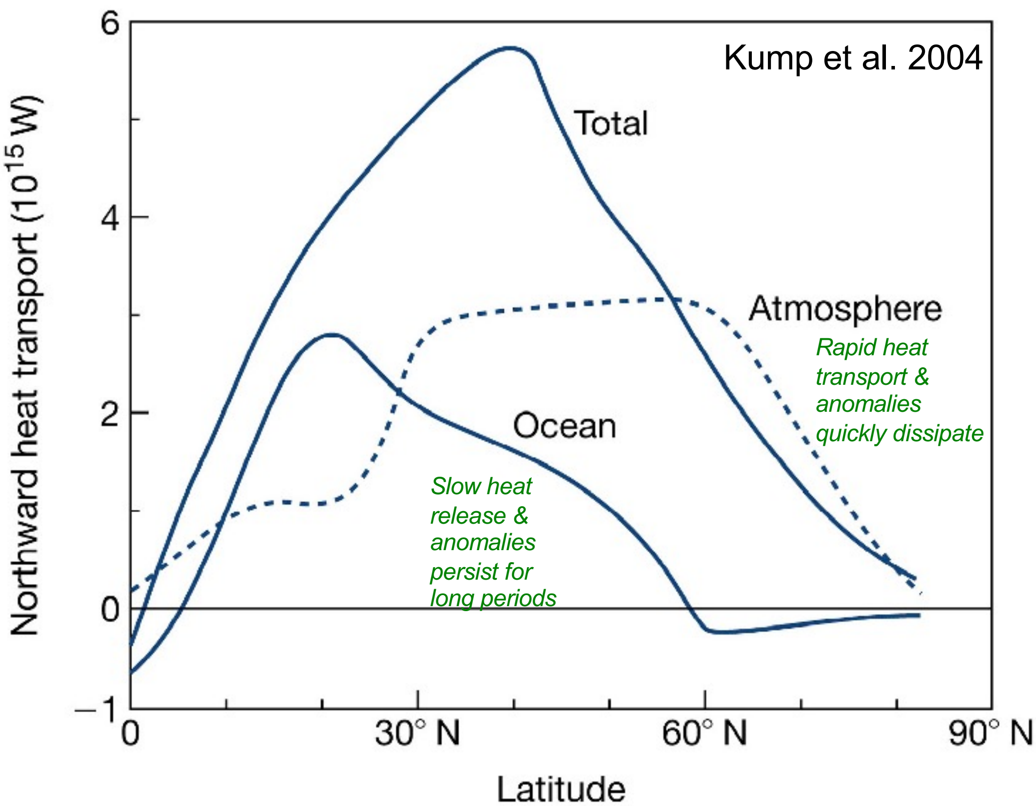
Spatial variations in radiation balance.







IMPORTANCE OF ATMOSPHERIC DYNAMICS in moderating the earth's climate is demonstrated by this graph, which compares the calculated radiative-equilibrium temperature for a "black" earth (*colored curve*) with the observed vertical mean temperature (*black curve*) as a function of latitude during January. At this time no sunshine reaches the earth north of the Arctic Circle; neglecting any lag effects due to the storage of heat, the radiative-equilibrium temperature in the polar cap would go down to absolute zero (−273.2



Climate Forcing

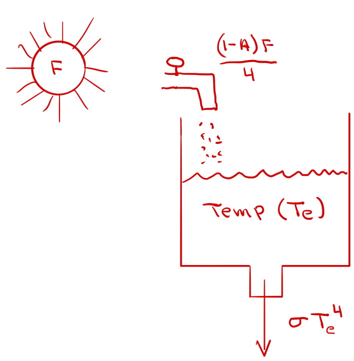
Change imposed on the planetary energy balance that alters the global temperature of the Earth/atmosphere system or the surface temperature.

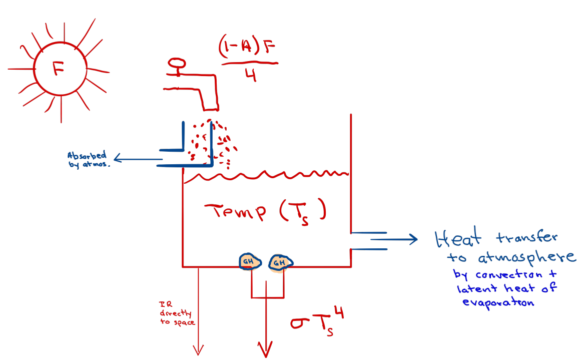
Forcing Factors:

T_e altered only by solar constant & albedo

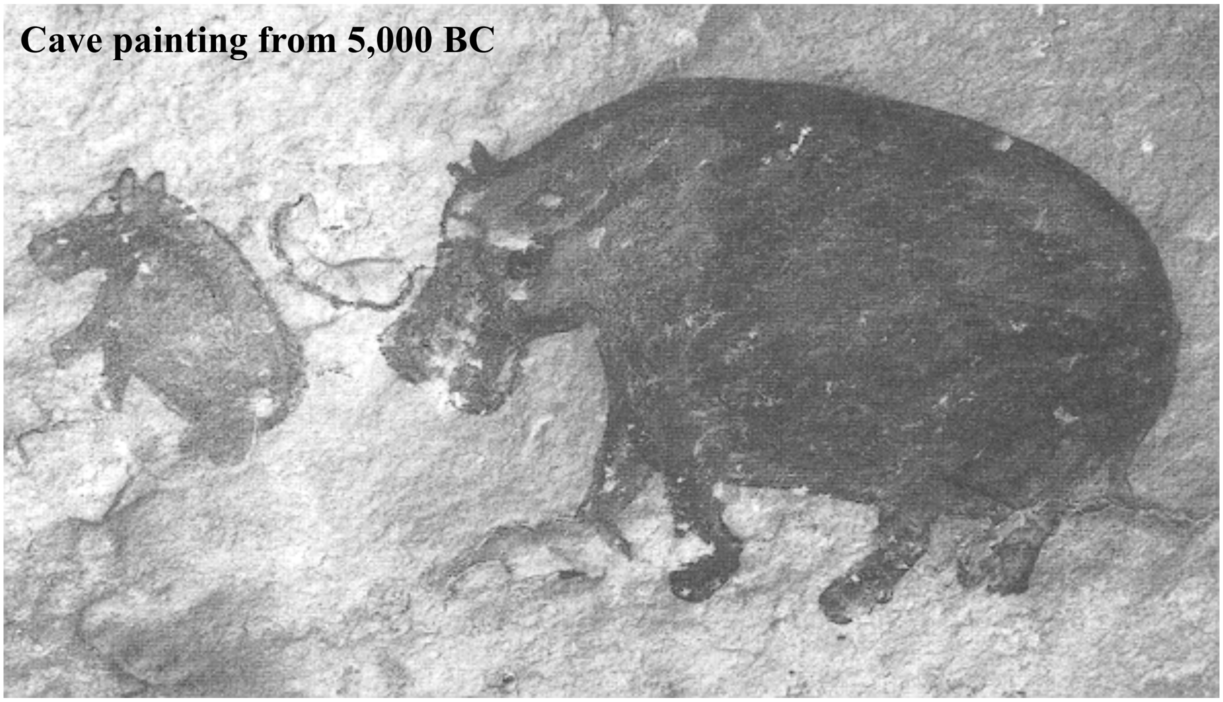
T_s also affected by GH effect; heat transfers; & EMR absorbed by the atmosphere

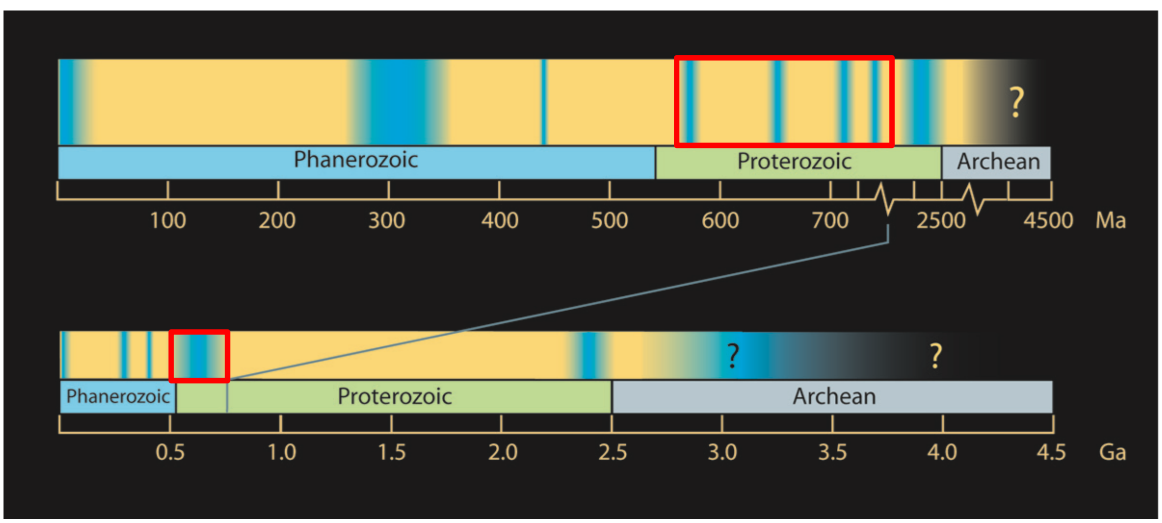
There are both natural & anthropogenic forcings.





Climate Change is Nothing New !



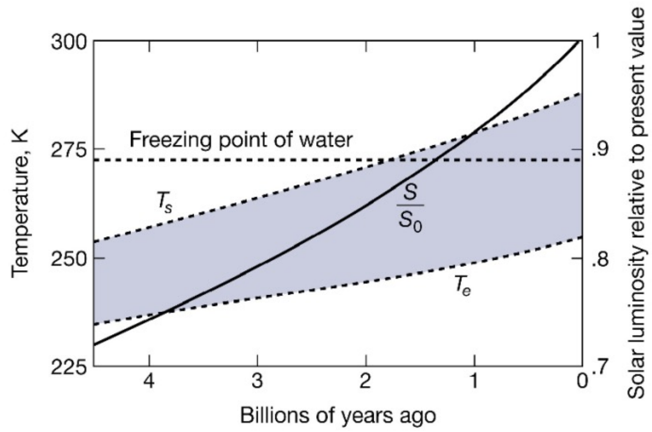


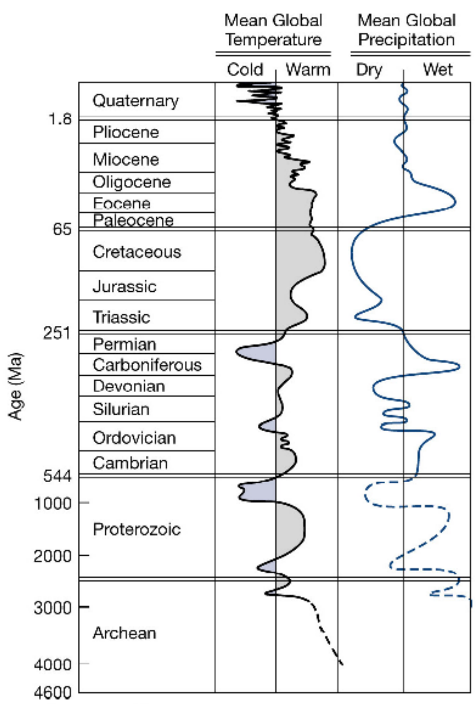
Schematic representation of icehouse-greenhouse intervals through Earth history. Lower bar illustrates all of Earth history, whereas upper bar focuses primarily on the Phanerozoic-Neoproterozoic. Icehouse times are inferred from published records of well-accepted glacial deposits recording the former presence of land-based ice sheets. Thus, greenhouse times are inferred on negative evidence of such deposits. Figure from G.S. Soreghan.

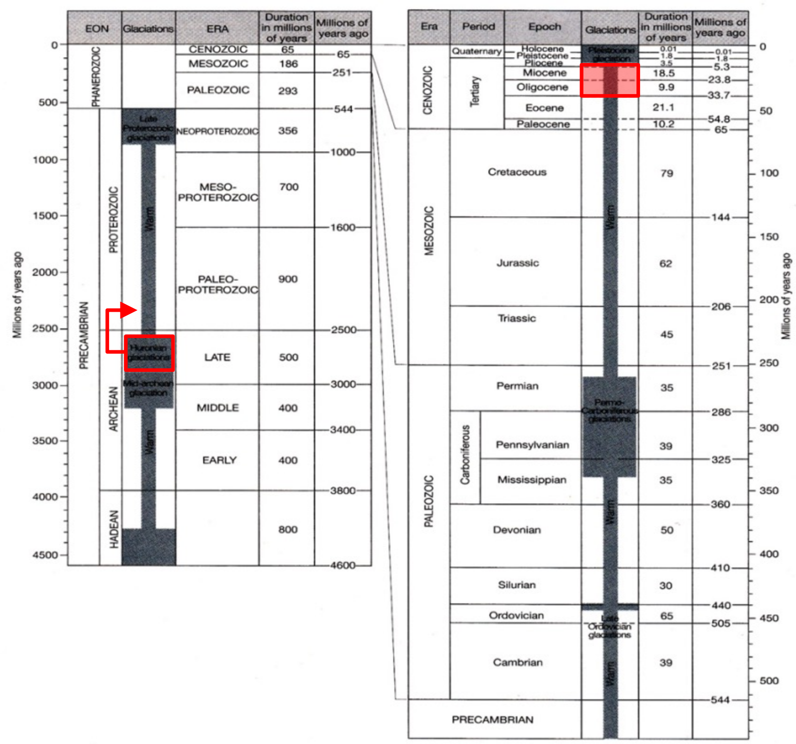
Video lectures on Earth's climate history

Date	Start	Stop	Total
4/23/21	32:24	50:53	18:29
4/26/21	8:22	52:20	44:00
4/28/21	2:28	9:30	7:06

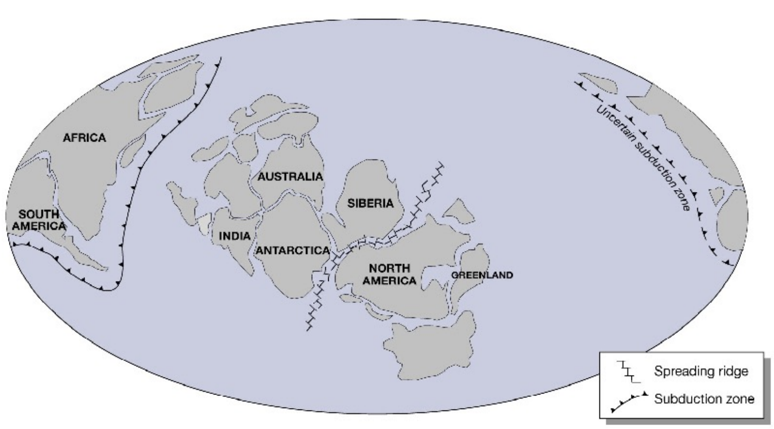
Faint Young Sun Paradox







Snowball Earth



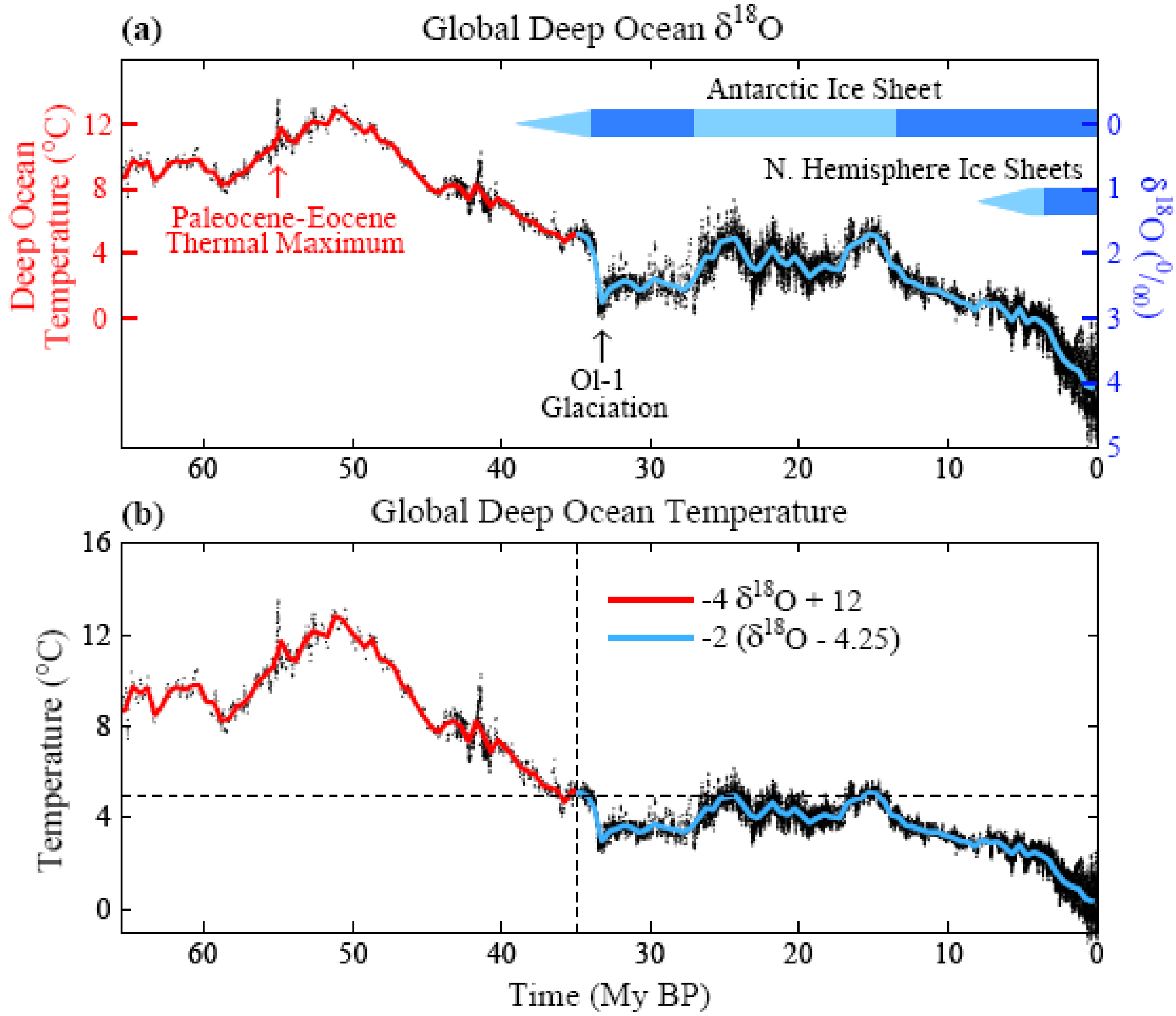
Long-Term Climate Changes

The Faint Young Sun Paradox –

> *When Earth formed, the Sun was ~30% less luminous and surface temperatures should have been below freezing for ~ 2.5 billion years BUT it wasn't.*

Increasing solar luminosity; reduced GH effect

Name	Time	Trigger		Brake
Mid-Archean	~2.9 BYBP	Organic haze from high CH_4/CO_2 ratio	?	$\Downarrow$ silicate weathering & volcanic CO_2 input
Huronian	~2.4 BYBP	Initial $\Uparrow$ O_2 -> $\Downarrow$ CH_4		$\Downarrow$ silicate weathering & volcanic CO_2 input
Late Proterozoic ("snowball Earth") Sun 6% less luminous	~700 MYBP	$\Downarrow$ temperature sensitivity of silicate weathering b/c land concentrated in tropics & snow & ice/albedo feedback		$\Downarrow$ silicate weathering & volcanic CO_2 input & snow & ice/albedo feedback
Late Ordovician	~450 MYBP	Formation of Appalachian Mountains?		$\Downarrow$ silicate weathering & volcanic CO_2 input
Permo-Carboniferous	~300 MYBP	$\Uparrow$ large plants w/lignin causing $\Uparrow$ C storage & $\Downarrow$ GH effect		$\Uparrow$ sea-floor spreading (thus volcanic eruptions) & $\Uparrow$ sea levels ($\Downarrow$ silicate weathering) & stronger circulation (atmos. & ocean)
Pleistocene ("Late Cenozoic")	~80 MYBP	$\Downarrow$ sea-floor spreading; $\Uparrow$ land @ high lats.; formation of Himalayas		Still in it. Humans?



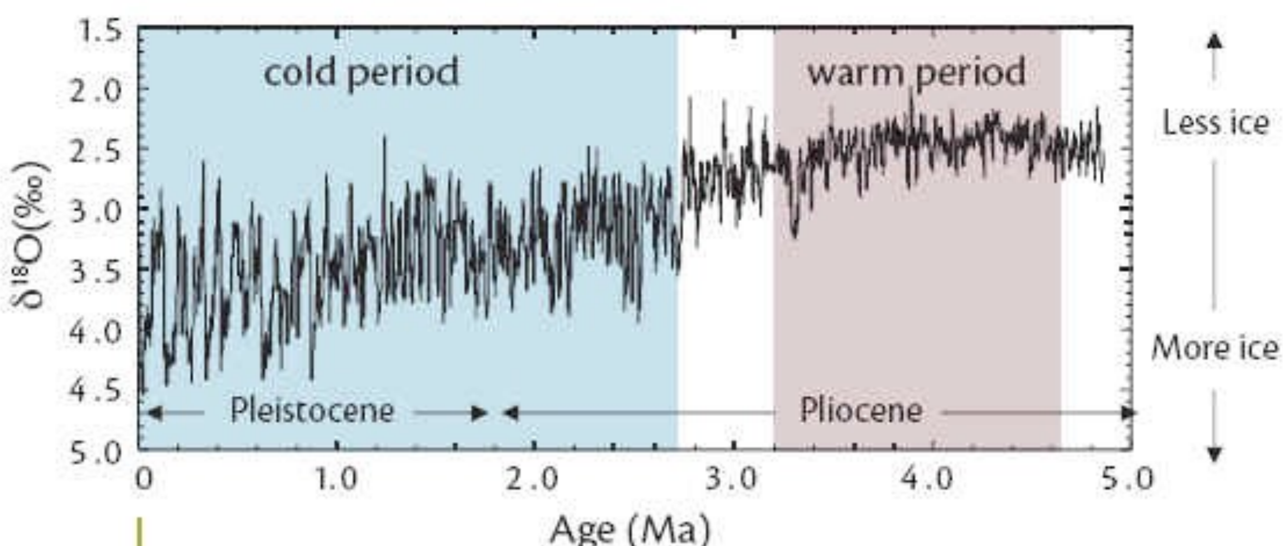
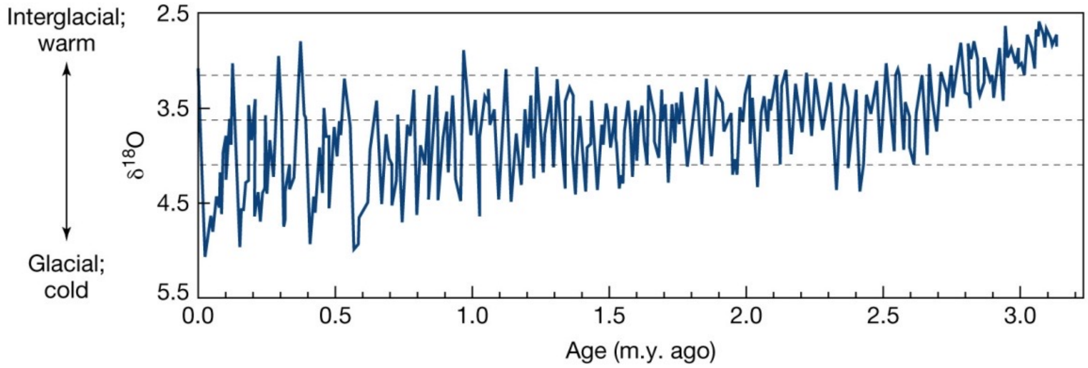
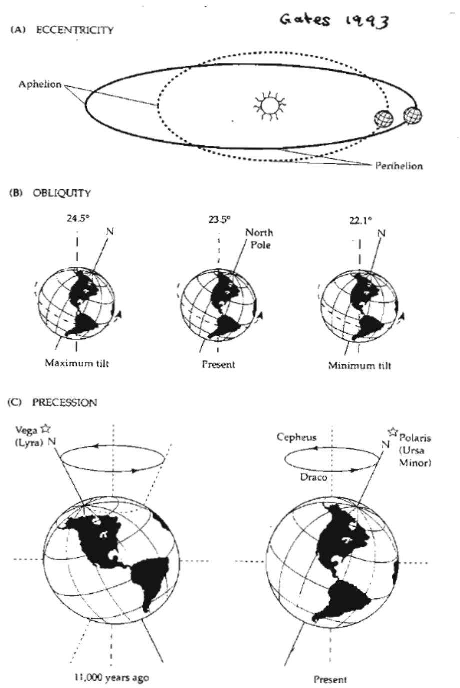
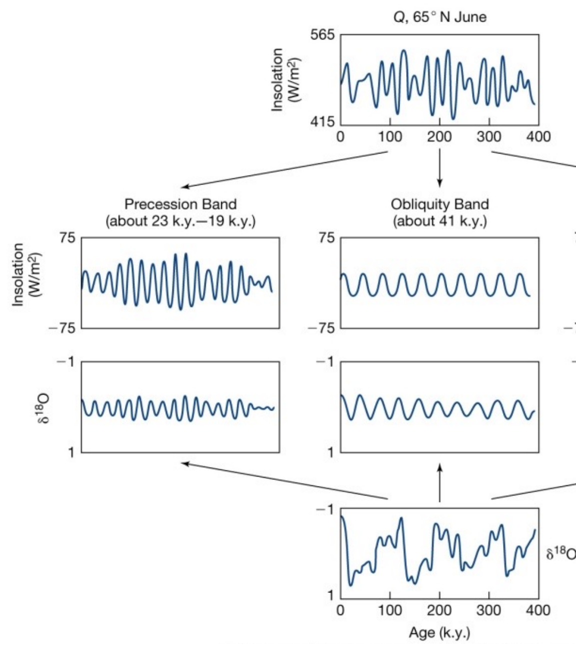


Figure 1. Benthic foraminifera $\delta^{18}O$ record of high-latitude climate (primarily ice volume) from ODP Site 677 (Shackleton et al., 1990) and ODP Site 1085 (Andreasen, 2001). The end of the Pliocene warm period and the onset of Northern Hemisphere glaciation occurred at ~3.0 Ma.

Deep-sea sediment core







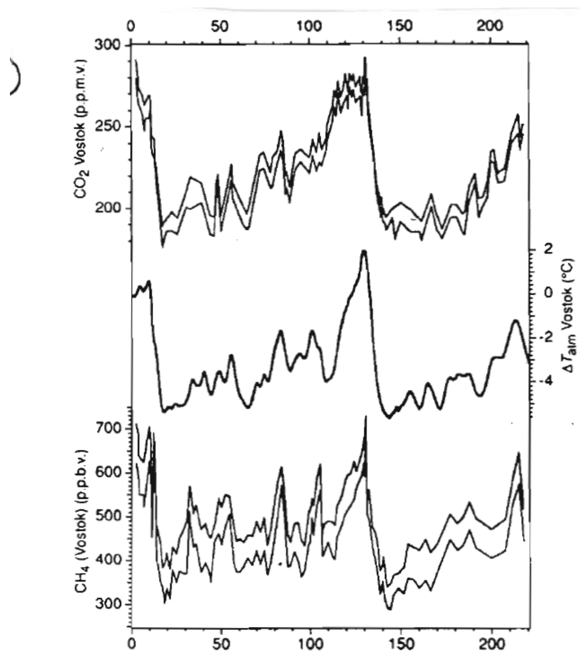


FIG. 5 CO_2, CH_4 and Vostok atmospheric temperature with respect to time (EGT) with the air–ice age difference calculated following Barnola *et al.*[33], taking into account the temperature dependence of the ice density[50]. For CO_2 and CH_4, the envelope corresponds to the measurement accuracy. Jouzel et al.

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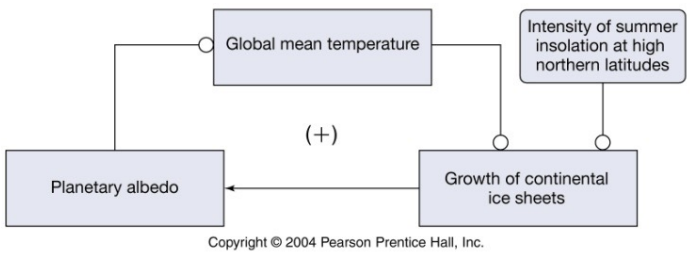
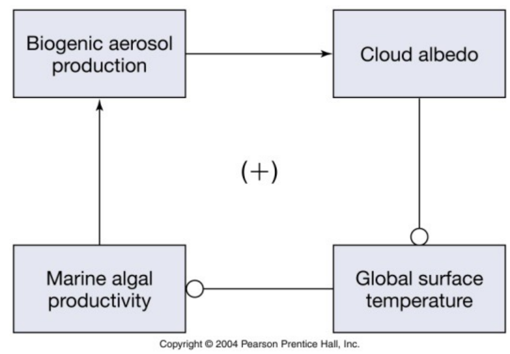
The 100,000 eccentricity signal may have been amplified.

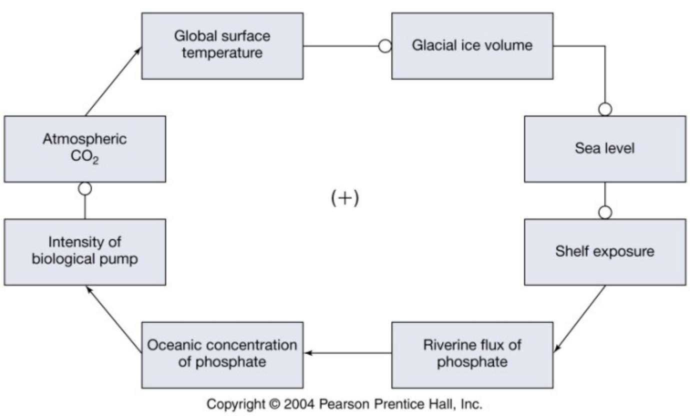
Ice-Albedo Feedback (+)



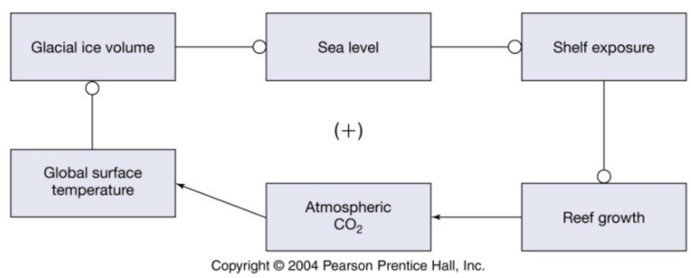
Cloud-Albedo Feedback (+)



Biological-Pump Feedback (+)

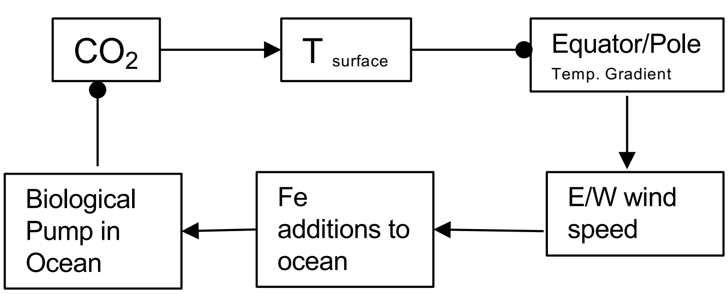


Coral Reef Feedback (+)

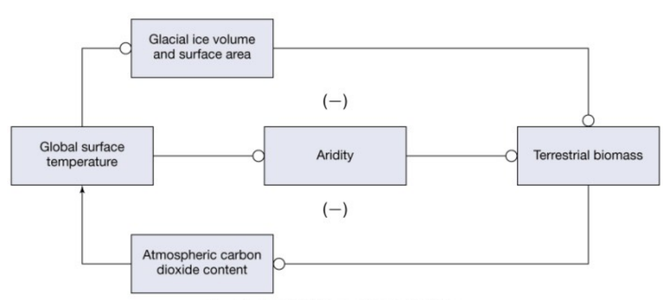


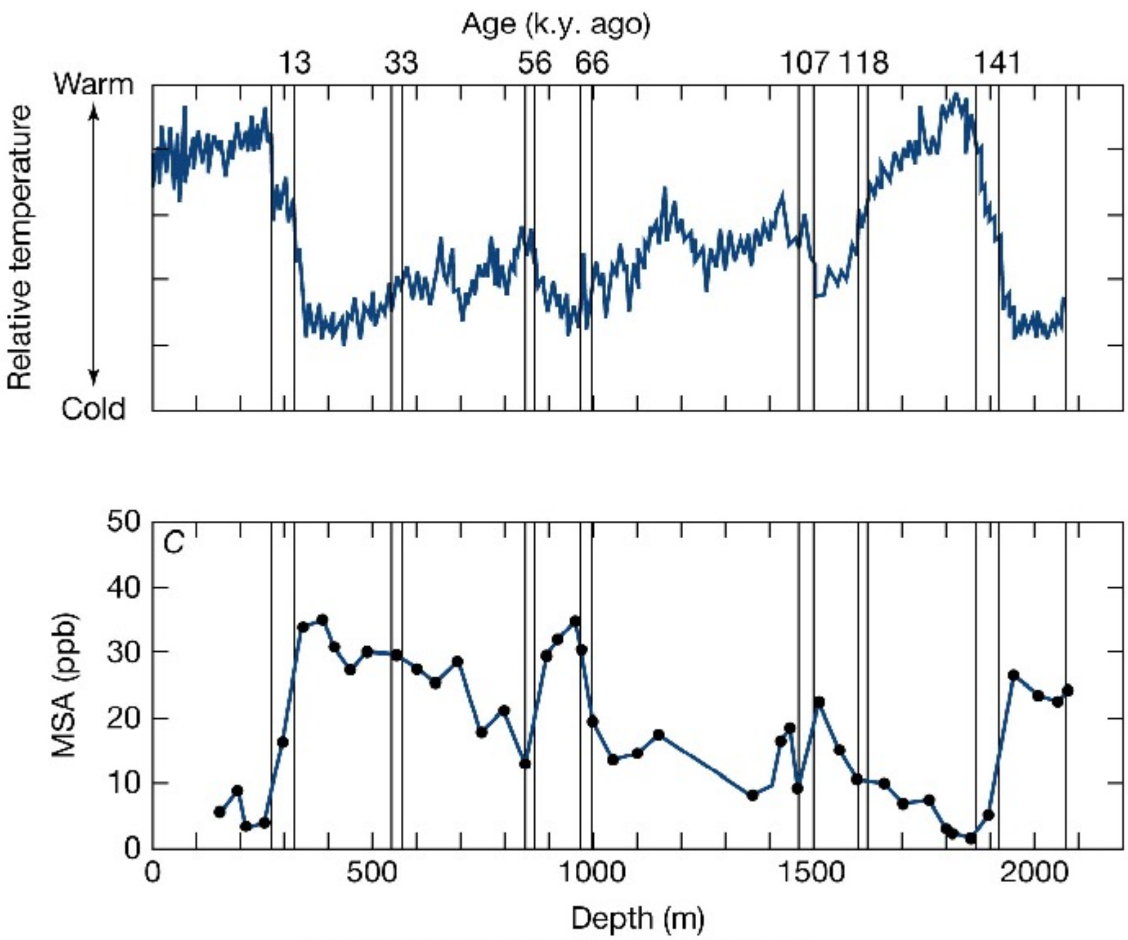
$$Ca_2^{++} + 2HCO_3^- \rightarrow CaCO_3 + CO_2 + H_2O$$

Iron Fertilization Feedback (+)

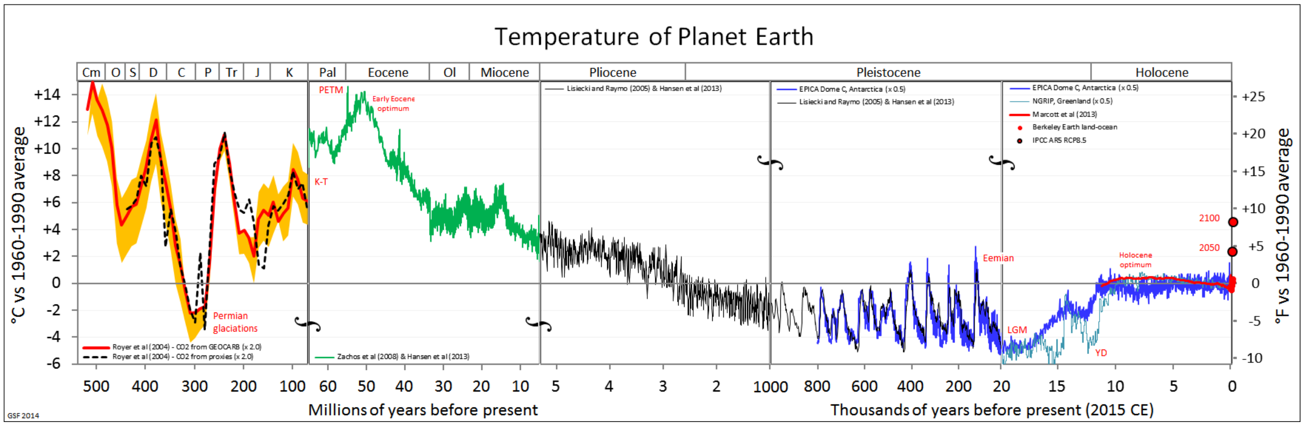


Terrestrial Biomass Feedback (-)





Age (k.y. ago)
13
33
56
66
107
118
141
Warm
Relative temperature
Cold
C
MSA (ppb)
50
40
30
20
10
0
0
500
1000
1500
2000
Depth (m)



Schematic of paleotemperatures for the Phanerozoic. http://en.wikipedia.org/wiki/File: All palaeotemps.png, accessed Oct 20 2018. Compiled from multiple sources (information on website). As the methods used to determine temperature are not identical, the variations should be taken as relative and not absolute. However, the figure represents current thinking about relative temperature changes.

Teachable moments from the history of Earth's climate

Climate optimum of the Pliocene (3.3 – 4.3 MYBP)

Mid-latitudes ~3-4^{o} C warmer than pre-industrial
CO_2 < 450 ppm
Increased annual P

Eemian interglacial (125,000 – 130,000 YBP)

1-2^{o} C warmer than pre-industrial
1-3^{o} C greater warming > 50^{o} N. Lat.
Increased regional P
[CO_2] ~ 300 ppm
Eccentricity driven

Holocene maximum (6,000-5,000 YBP; AKA Hypsithermal)

3-4^{o} C warmer summers in high N. Lats.
1-2^{o} C warmer summers in mid N. Lats.
Increased regional P & lake levels @ high Lats. & sub tropics
[CO_2] ~ 280 ppm
Orbital forcing

Younger Dryas (~12,000 YBP; lasted ~ 1,200 yrs)

6^{o} C cooler in N. Atlantic (took ~100 years)
then warmed 7^{o} C in ~1-3 year at the end
Changes in AMOC – example of climate tipping point

Little Ice Age (~1430 – 1850 AD)

~0.6^{o} C cooler & highly variable climate
Increase in explosive volcanism
Decreased solar activity

The Real Inconvenient Truth Is That …

Global Change is More Than Global Warming

- Changing atmospheric chemistry
- Changes in global nutrient cycles
- Changing atmospheric composition
- Climate change
- Stratospheric ozone depletion
- Land-use change
- Loss of biodiversity