P607 Climate and Energy (Dr. H. Coe)

Syllabus:

The composition of the atmosphere and the atmospheric energy balance;

Radiative balance in the atmosphere; Energy flow in the biosphere, atmosphere and ocean;

climatology of the Earth; circulation of the oceans and the atmosphere;

evidence for natural and anthropogenic climate change;

the pattern of energy consumption now and in the future;

future climate change predictions;

emissions reductions and their impact on future energy consumption;

the useful lifetime of fossil fuels;

the contribution of alternative energy and nuclear resources for the future;

the problem of improved efficiency and reducing waste.

Recommended Texts:

J T Houghton, 'The Physics of Atmospheres', 2nd edition, Cambridge University Press, 1986

R G Barry and R J Chorley, 'Atmosphere, Weather and Climate', Routledge, 1987

J P Peixoto and A H Oort, 'Physics of Climate', American Institute of Physics, 1992

E Boeker and R van Grondelle, 'Environmental Physics', Wiley and Sons, 1999

J T Houghton and others, 'Climate Change 2001', Cambridge University Press, 2001

E Harder, 'Fundamentals of Energy Production', Wiley and Sons, 1982

S S Penner and L Icerman, 'Energy Volume 1', Addison Wesley, 1981

Energy Input to the Earth

The Sun provides almost all of the energy input to the Earth, its oceans and atmosphere, a massive 5 x 10^{24} J/year; in contrast the internal energy of the Earth generates only around 10^{21} J/year.

The Sun is a middle aged, medium sized star with a composition of approximately 75% hydrogen and 25% helium.

The Sun's energy is derived from the fusion of hydrogen into helium nuclei which is then transferred to the surface of the Sun via short wave electromagnetic radiation.

Although the Sun has a radius of 7.0 x 10^{5} km, virtually all the energy received by the Earth is emitted by the outer 500 km, known as the photosphere.

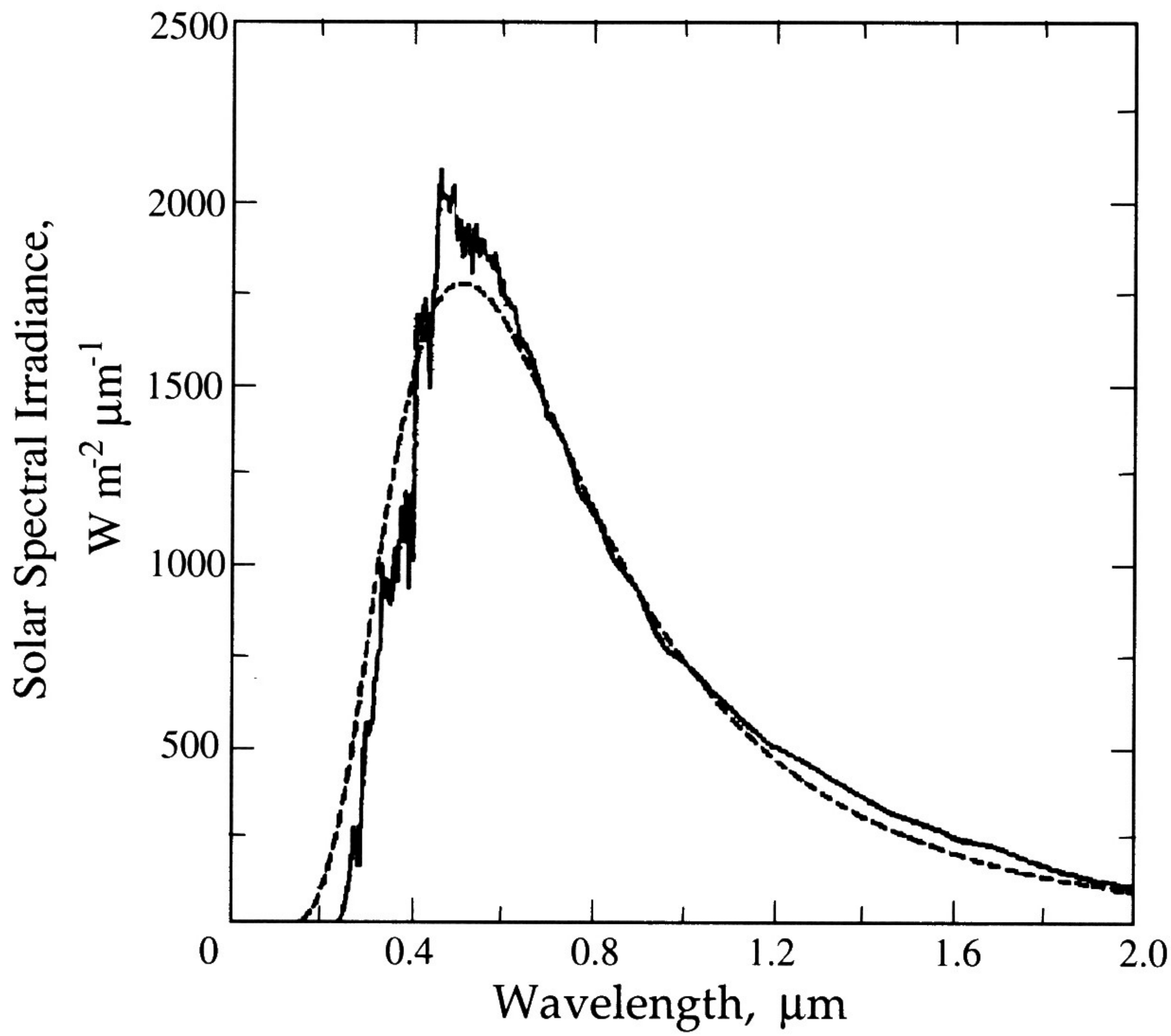
The Solar photosphere emits light across the entire electromagnetic spectrum, from gamma rays to radio waves.

Most of the radiative power incident at the top of the Earth's atmosphere is due to light of wavelength between 200 nm in the ultra-violet to 4 μm in the infra-red.

The peak intensity is at about 490 nm in the green part of the spectrum.

The Sun's photosphere has a temperature of approximately 5800 K and can be thought of as a blackbody emitter of this temperature.

Also shown is the blackbody radiation curve for an emitter at a temperature of 5800 K



The Planck Function

Planck related the emissive power, or intensity, of a blackbody $B(\lambda,T)$ at a given wavelength, λ, to the temperature, T, of the emitter by

$$B(\lambda, T) = \frac{2hc^2}{\left(\lambda^5 \exp^{hc/k\lambda T} - 1\right)}$$

where k is the Boltzmann constant (1.381×10^{-23} J K^{-1}), c is the speed of light in vacuum (2.998×10^{8} m s^{-1}), and h is Planck's constant (6.626×10^{34} J s).

The total flux emitted by a blackbody radiator, F_B, and the total emitted intensity B, can be found by integrating the Planck blackbody function (2.1) over all wavelengths

$$F_B = \pi B = \pi \int_0^\infty B(\lambda, T) d\lambda = \sigma T^4$$

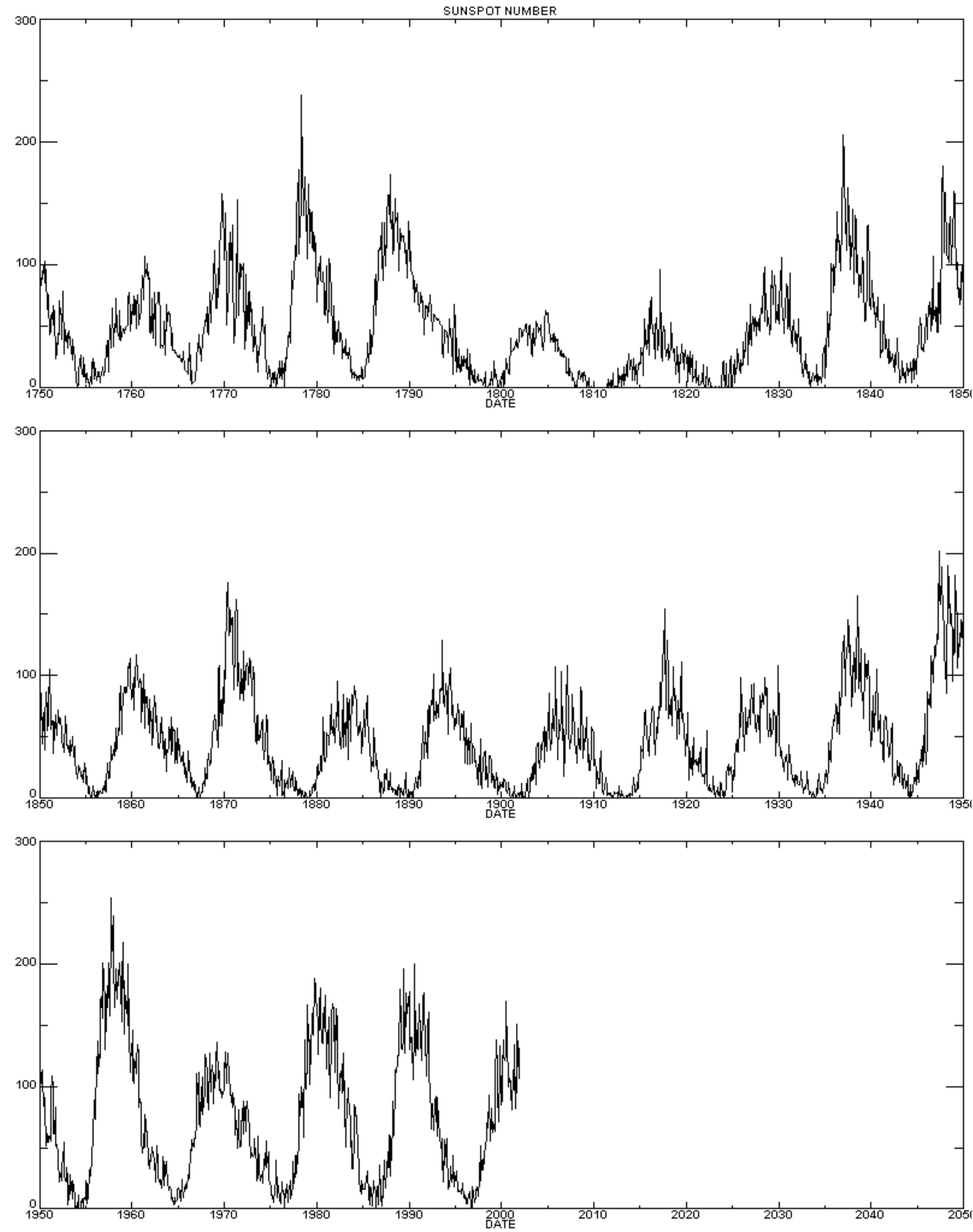
where σ is the Stefan-Boltzmann constant (5.671×10^{-8} W m^{-2} K^{-4}).

Solar Variability

Solar radiation has an average intensity of approximately 1370 W m^{-2} at the distance of the Earth from the Sun, referred to as the solar constant, S. However:

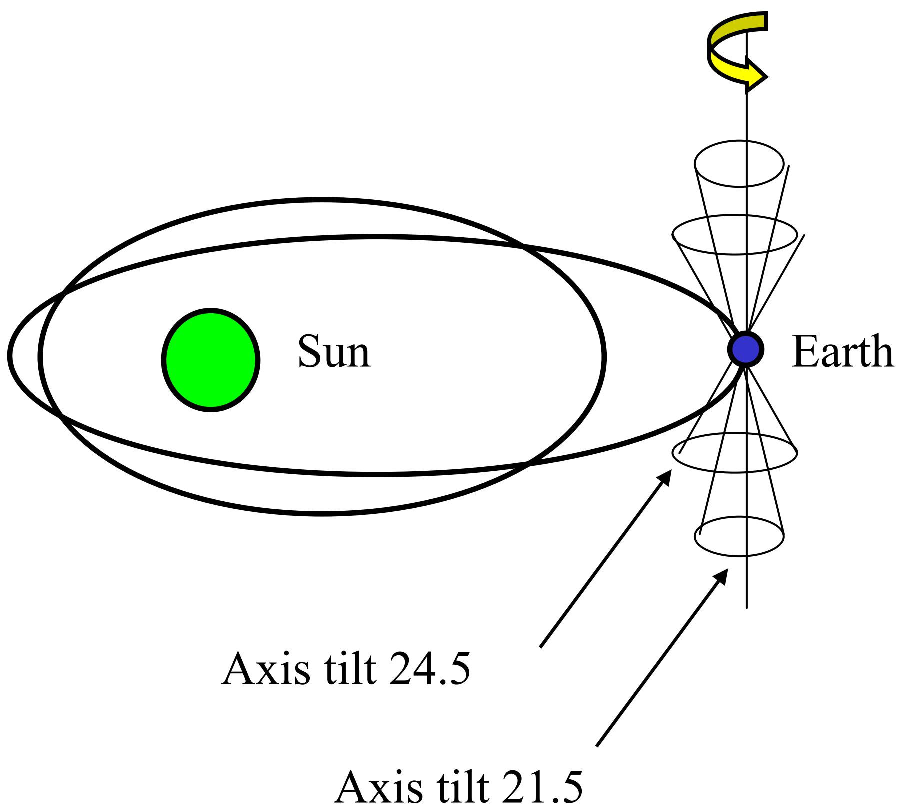
The Sun rotates with a period of 27 days and both active, brighter regions known as *faculae* and less active, darker regions known as *sunspots* face the Earth during each rotation. The output from these different regions of the Sun varies by between 0.1 to 0.3 % of the total flux.

The number of sunspots on the surface of the Sun varies with a cycle of 11 years, causing ~1% variations in radiative flux at the top of the Earth's atmosphere.



Solar Variability 2:

Lower frequency variations in the solar flux, again of the order of 1 to 2 %, have also been inferred from isotopic abundance measurements. These arise from:



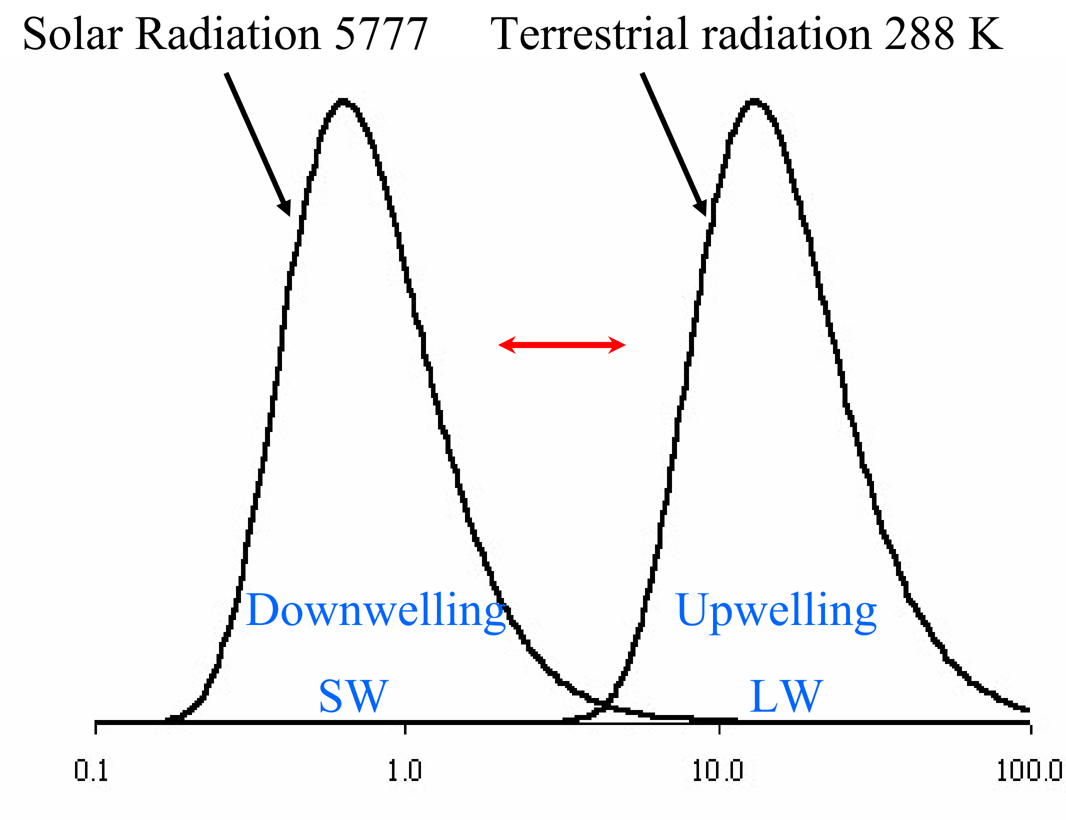
Eccentricity: $\tau \sim$ 100,000 years

Axis tilt: $\tau \sim$ 41,000 years

Precession of axis around normal: $\tau \sim$ 26,000 years

Terrestrial Radiation

The Earth also acts as a blackbody radiator, but as its global mean surface temperature, T_s, is 288 K, most of the irradiance from the Earth is in the infra-red part of the spectrum and peaks at about 10 µm. Compare the two normalised blackbody curves:

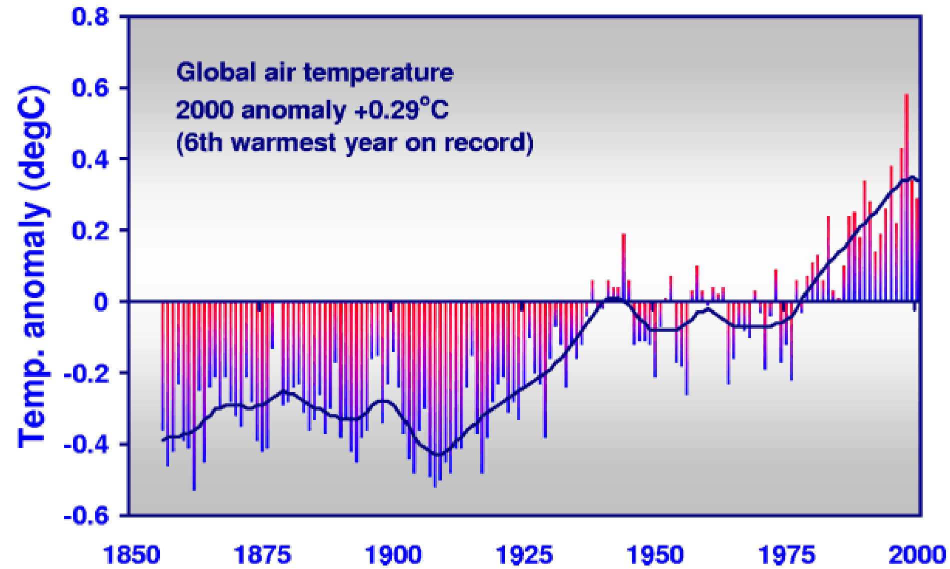


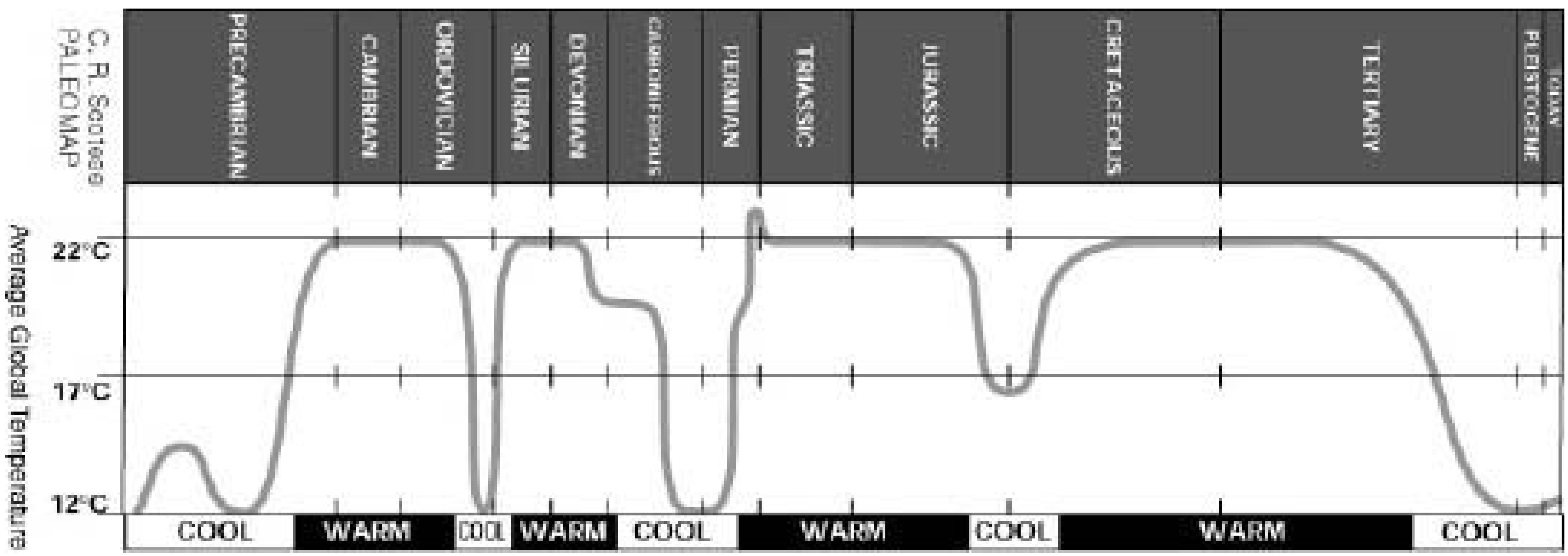
There is very little overlap between the incoming solar radiation at UV and visible wavelengths and the outgoing infra-red radiation from the Earth's surface.

Incoming solar radiation and outgoing terrestrial radiation are separated by a gap at around 4 µm, and are often referred to as shortwave (SW) and longwave (LW) radiation respectively.

Steady State Temperature of the Earth's Atmosphere:

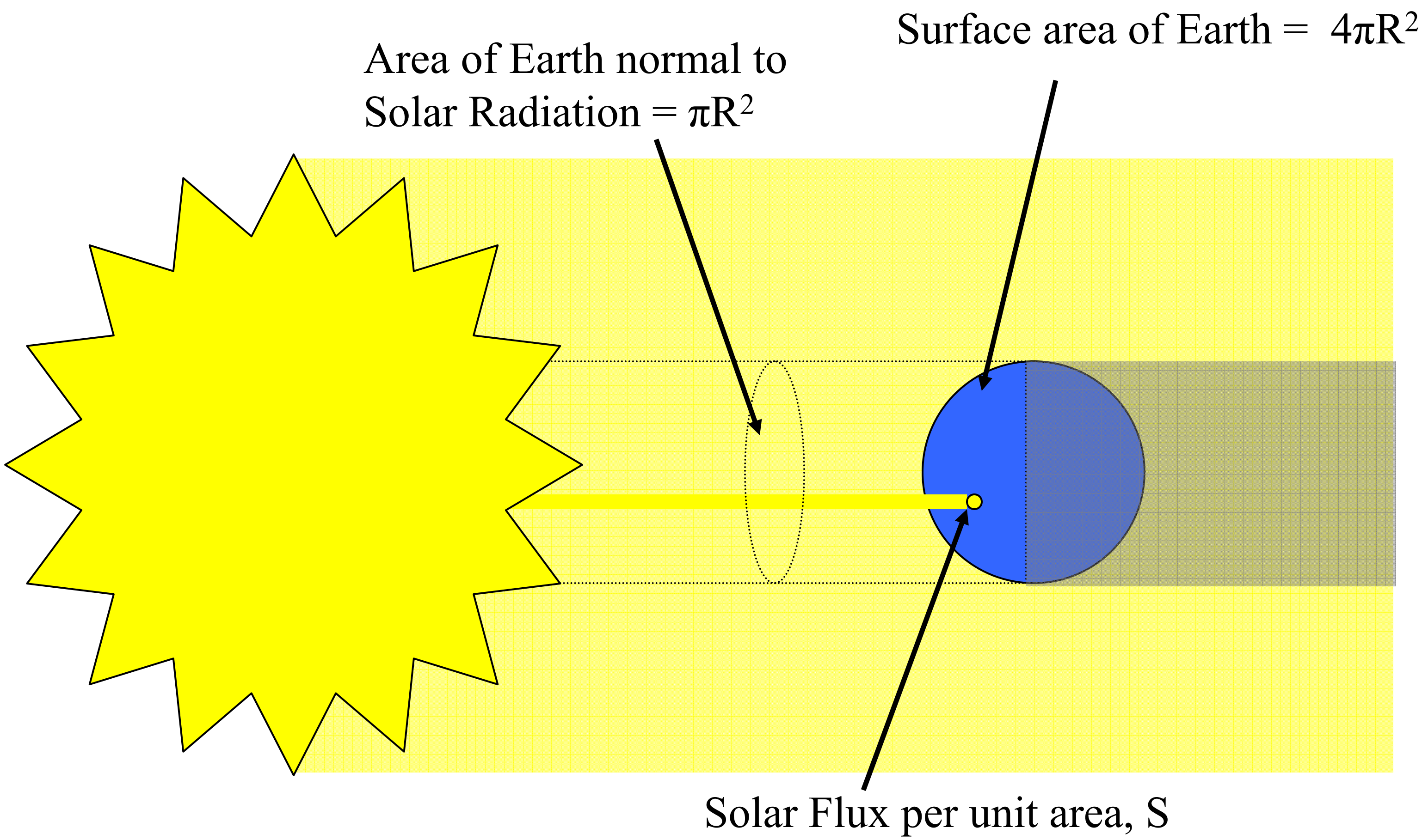
As the mean surface temperature of the Earth changes little from year to year and has varied by less than 5 °C in the last 20000 years and 10 °C over geological time it is clear that the system is in equilibrium and the energy inputs must be balanced by energy losses.





Steady State Temperature of the Earth's Atmosphere 2:

The effective area of the Earth receiving sunlight at any one time is given by πR^2, where R is the radius of the Earth, yet the total area of the Earth is $4\pi R^2$, so the average radiant flux over the Earth is given by S/4.



Steady State Temperature of the Earth's Atmosphere 3:

Not all of the incoming radiation is absorbed by the surface, some is reflected back to space by either the surface, clouds, aerosol particles or scattering from molecules in the atmosphere.

The fractional reflectance is known as the global mean planetary reflectance or albedo, A. The average surface albedo is around 0.15 but the high reflectivity of clouds leads to an overall planetary albedo, A, of 0.3.

Thus the incoming irradiance absorbed by the Earth's surface, F_s, is given by:

$$F_s = (1 - A)\frac{S}{4}$$

and has a value of 240 W m^{-2}.

F_s must be balanced by the outgoing blackbody radiation of the Earth given by σT_e^4, where T_e is the effective blackbody temperature of the Earth-atmosphere system. Equating incoming and outgoing fluxes gives an expression for T_e:

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

an equilibrium temperature of 255 K, compared to 288 K, the average surface temperature of the Earth.

Equilibrium Temperatures of Different Planets:

Planet	**R^***	**Albedo (A)**	**T_e(K)**	**T_m(K)**	**T_s(K)**
Venus	0.72	0.77	227	234	750
Earth	1	0.30	256	250	280
Mars	1.52	0.15	216	220	240
Jupiter	5.2	0.58	98	130	134

R* is the ratio of the planet radius to that of the Earth

T_m is the effective temperature of the planet measured from space. For all planets except Jupiter T_e and T_m are similar, Jupiter's internal energy is a significant heat source for the planet.

There is little difference between T_s, the surface temperature, and T_e for Jupiter and Mars, implying this simple equilibrium is a close approximation. However, for the Earth and in particular Venus, T_s is significantly greater than T_e.

WHY? We need to consider the interaction between trace constituents in the atmosphere and incoming and outgoing radiation.