Climate Dynamics (lecture 2)

Aarnout van Delden, 6 Feb. 2013

http://www.staff.science.uu.nl/~delde102/climatedynamics.htm

Solar irradiance*
Planck's- and Stefan-Boltzman-law*
Bougeur-Lambert-Beer law
Optical depth
Kirchoff's law
Emission temperature*
Albedo*
Energy balance at TOA*
Greenhouse effect*
Radiative equilibrium timescale*

What do you know about these topics?

***Today**

Solar radiation

Sun puts out a nearly constant flux of energy: $L_0 = 3.9 \times 10^{26}$ W

This is the **solar luminosity**

Flux density is luminosity divided by area of the photosphere*:

6.4×10^7 Wm^{-2}

This energy also reaches the Earth **at a mean distance of** 1.5×10^{11} m

Flux density **at this distance** is $S_0 = 1366$ Wm^{-2}

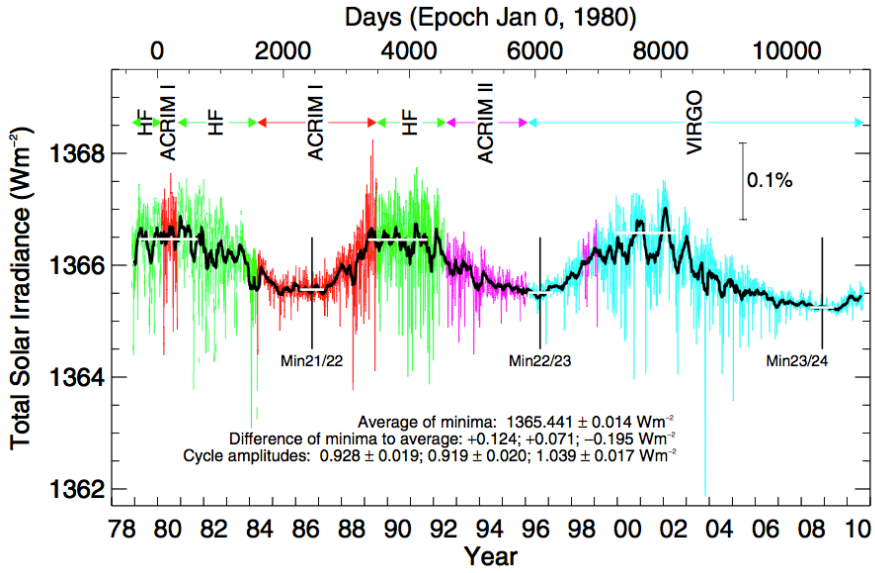
This number is called the **solar "constant" or solar irradiance**

Is actually not constant, though we would like it to be →

* surface area of a sphere is $4\pi r^2$

The “solar constant”:

http://www.pmodwrc.ch/pmod.php?topic=tsi/composite/SolarConstant



The PMOD composite TSI as daily values plotted in different colors for the different originating experiments. The difference between the minima values is also indicated, together with amplitudes of the three cycles.

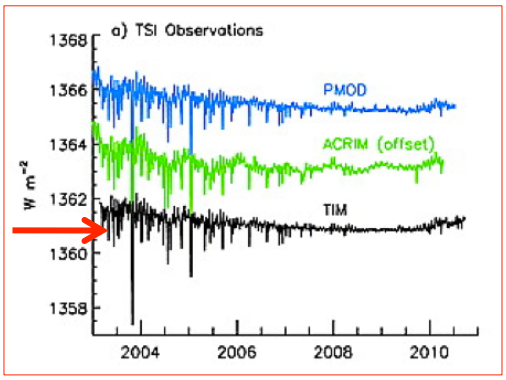
Recent paper: Solar “constant” much lower!?

GEOPHYSICAL RESEARCH LETTERS, VOL. 38, L01706, 7 PP., 2011, doi:10.1029/2010GL045777

A new, lower value of total solar irradiance: Evidence and climate significance

Greg KoppLaboratory for Atmospheric and Space Physics, Boulder, Colorado, USAJudith L. LeanSpace Science Division, Naval Research Laboratory, Washington, D. C., USA

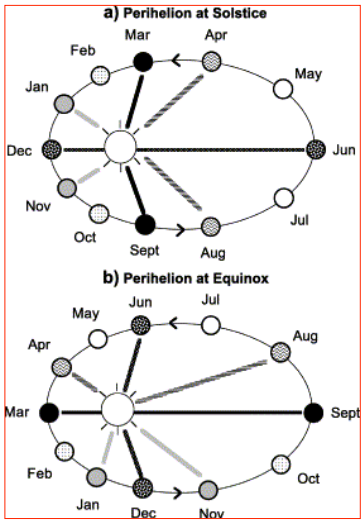
The most accurate value of total solar irradiance during the 2008 solar minimum period is 1360.8 ± 0.5 W m^{-2} according to measurements from the Total Irradiance Monitor (TIM) on NASA's Solar Radiation and Climate Experiment (SORCE) and a series of new radiometric laboratory tests. This value is significantly lower than the canonical value of 1365.4 ± 1.3 W m^{-2} established in the 1990s, which energy balance calculations and climate models currently use. Scattered light is a primary cause of the higher irradiance values measured by the earlier generation of solar radiometers in which the precision aperture defining the measured solar beam is located behind a larger, view-limiting aperture. In the TIM, the opposite order of these apertures precludes this spurious signal by limiting the light entering the instrument. We assess the accuracy and stability of irradiance measurements made since 1978 and the implications of instrument uncertainties and instabilities for climate research in comparison with the new TIM data. TIM's lower solar irradiance value is not a change in the Sun's output, whose variations it detects with stability comparable or superior to prior measurements; instead, its significance is in advancing the capability of monitoring solar irradiance variations on climate-relevant time scales and in improving estimates of Earth energy balance, which the Sun initiates.



Data from the three on-orbit TSI instruments, the ACRIM III (green), VIRGO (blue), and TIM (black), exhibit generally similar fluctuations, but with notable differences that are not easily attributable to solar variations.

Variations in solar irradiance

Precession of the equinoxes



The precession determines the timing of the **perihelion,** i.e. the the date on which the Earth is closest to the Sun. Today this is the case on January 4.

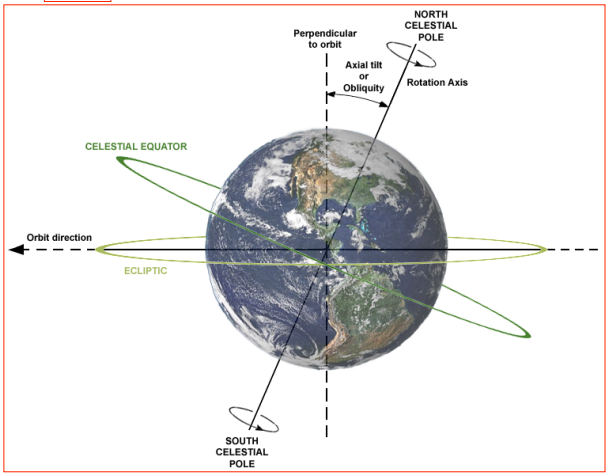
Seasonal cycle Solar irradiance: 90 W m^{-2}!!! *

The time period of the associated precession is about 19,000 to 23,000 years.

***1320-1410 W m^{-2}**

Variations in solar irradiance

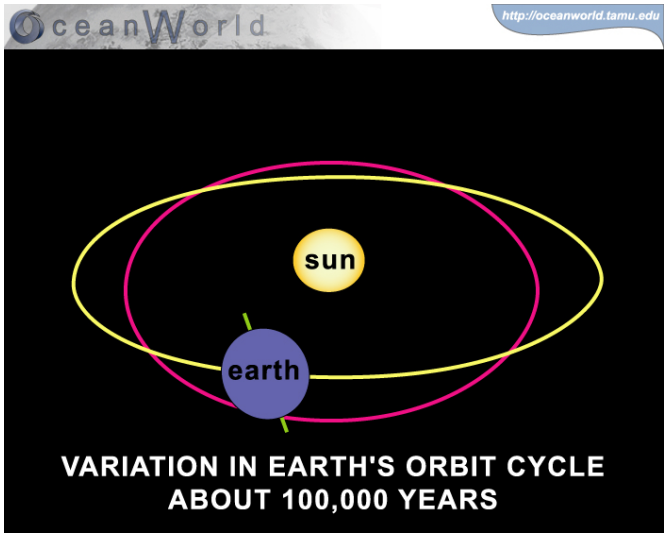
Tilt



Earth shown as viewed from the Sun The axial tilt (or **obliquity**) has varied between 22° and 24.5° with a period of 41000 years. Obliquity determines the amplitude of the **seasonal cycle**.

Variations in solar irradiance

eccentricity



Periodic changes in the "roundness" of Earth's orbit varies in cycles of 100,000 to 400,000 years. This affects how important "timing of the perihelion" is to the seasonal cycle in the intensity of the Solar "constant" (now 6.6%;16.6% 116000 years ago!)

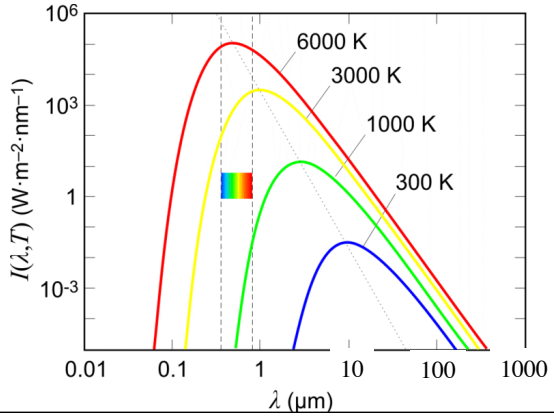
Planck's law

A **black body** with a temperature T emits radiation at wavelength λ with an intensity (energy per unit time per unit area per unit solid angle per unit wavelength)

$$B_\lambda(T) = \frac{2\hbar c^2}{\lambda^5}\frac{1}{\exp(\hbar c/\lambda kT)-1}$$

$\hbar = 6.63\times10^{-34}$ J s
$k = 1.38\times10^{-23}$ J K^{-1}
$c^2 = 3\times10^{8}$ m s^{-1}

Planck's radiation law (dependence of emissivity of **blackbody** surface on the wavelength and temperature). Log-log scale is necessary to display the curves for a large spread in temperature. The vertical dashed lines confine the visible spectral range. The slanted dotted line indicates the Wien's displacement law.



Stefan-Boltzman law

By integration of Planck's law over all wavelengths we obtain

Stefan-Boltzmann law $E_{BB}=\sigma T^4$ $\sigma=5.67\times10^{-8}\text{W m}^{-2}\text{K}^{-4}$

Emission temperature of the sun calculated from

$\sigma T^4=6.4\times10^7\text{Wm}^{-2}$ ← Flux density of energy at the edge of the photosphere

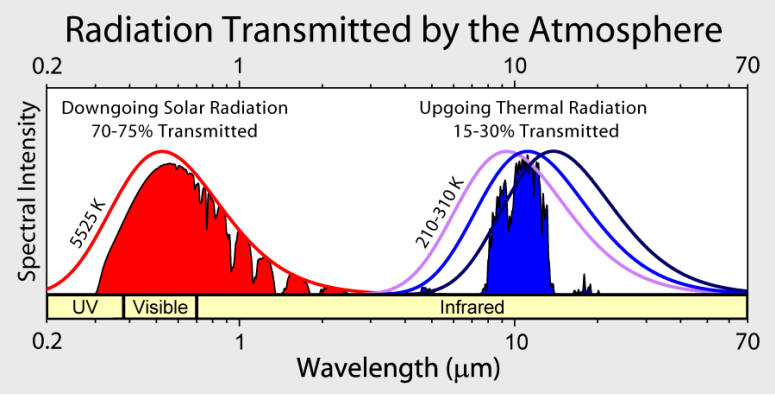
gives $T=5796$ K

This need not be the actual temperature of the sun's surface

Why?

Radiation spectrum

Both the Earth and the Sun emit electromagnetic radiation (e.g. light) that closely follows a black body spectrum, and which can be predicted based solely on their respective temperatures. For the Sun, these emissions peak in the visible region and correspond to a temperature of ~5500 K. Emissions from the Earth vary following variations in temperature across different locations and altitudes, but always peak in the infrared. The bulk of emissions from the Earth radiate from within the colder regions of the atmosphere rather than from the surface directly, and give the Earth an average emission temperature of about 250 K



In *this lecture* we ignore all these spectral details.

We make the following strongly simplifying approximation:

Radiation falls into two spectral categories:

short wavelength (*SW*) solar radiation and

long wavelength (*LW*) terrestrial radiation

Emission temperature of the Earth

Black body temperature of emitted radiation if there is energy balance

Absorbed solar radiation $= S_0(1-\alpha_p)\pi r_p^2 [\mathrm{W}]$

Emitted radiation $= \sigma T_e^4 [\mathrm{W\ m^{-2}}]$

$T_e \equiv$ emission temperature

$\alpha_p \approx 0.3$ is the average planetary albedo of the Earth

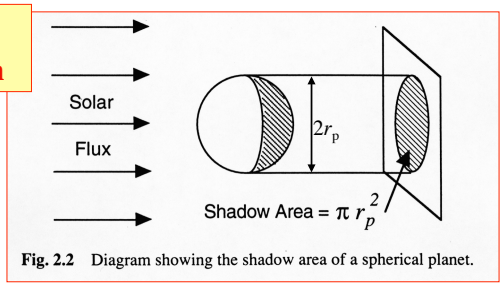


Fig. 2.2 Diagram showing the shadow area of a spherical planet.

Averaged over the whole surface area of the earth, i.e. $4\pi r_p^2$:

Absorbed radiation= emitted radiation → $T_e = \sqrt[4]{\frac{S_0/4(1-\alpha_p)}{\sigma}} = 255\ \mathrm{K} << T_s$

T_s is the temperature of Earth's surface

Why?

Albedo of various surfaces

Soils have higher albedo when they are dry; oceans have a relatively low albedo

It is of course very difficult to get this right in a climate model!

Table 4.2
Albedos for Various Surfaces in Percent

Surface type	Range	Typical value
Water		
Deep water: low wind, low altitude	5–10	7
Deep water: high wind, high altitude	10–20	12
Bare surfaces		
Moist dark soil, high humus	5–15	10
Moist gray soil	10–20	15
Dry soil, desert	20–35	30
Wet sand	20–30	25
Dry light sand	30–40	35
Asphalt pavement	5–10	7
Concrete pavement	15–35	20
Vegetation		
Short green vegetation	10–20	17
Dry vegetation	20–30	25
Coniferous forest	10–15	12
Deciduous forest	15–25	17
Snow and ice		
Forest with surface snowcover	20–35	25
Sea ice, no snowcover	25–40	30
Old, melting snow	35–65	50
Dry, cold snow	60–75	70
Fresh, dry snow	70–90	80

depends on zenit angle and cloudiness

Albedo of the sea surface

clouds scatter radiation

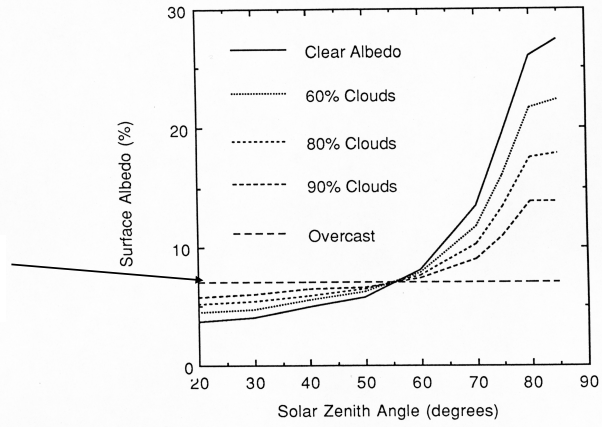


Fig. 4.4 Dependence of the albedo of a water surface on solar zenith angle and cloud cover. [Data from Mirinova (1973).]

Earth's energy balance 1985-1989

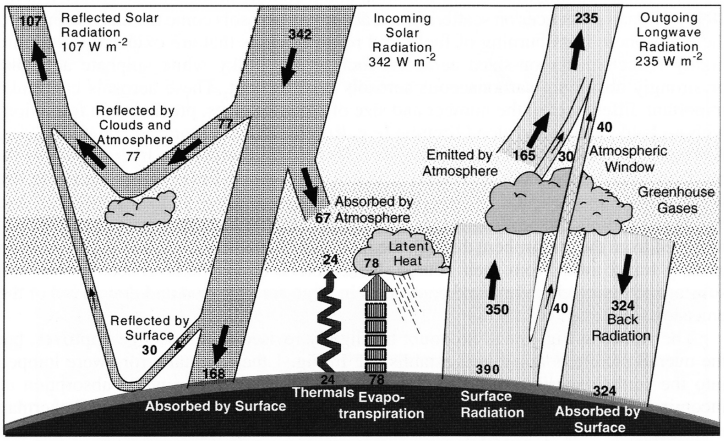


Figure 1. The earth's radiation balance. The net incoming solar radiation of 342 W m^{-2} is partially reflected by clouds and the atmosphere or at the surface, but 49% is absorbed by the surface. Some of that heat is returned to the atmosphere as sensible heating and most as evapotranspiration that is realized as latent heat in precipitation. The rest is radiated as thermal infrared radiation and most of that is absorbed by the atmosphere and re-emitted both upwards and downwards, producing a greenhouse effect, as the radiation lost to space comes from cloud tops and parts of the atmosphere much colder than the surface. From Kiehl and Trenberth (1997).

Figure 2.10

Energy fluxes (2000-2004)

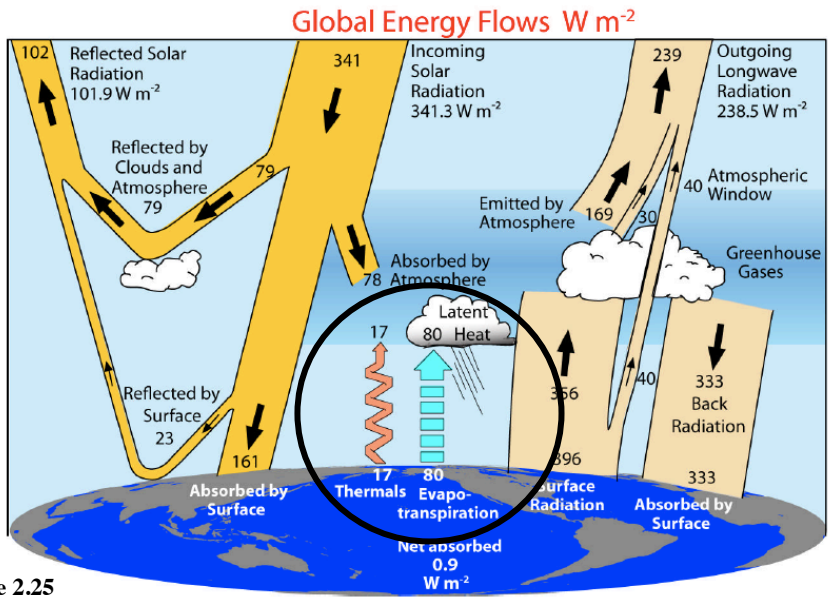


Figure 2.25

Energy balance at the top of the atmosphere (TOA)

JJA

?

DJF

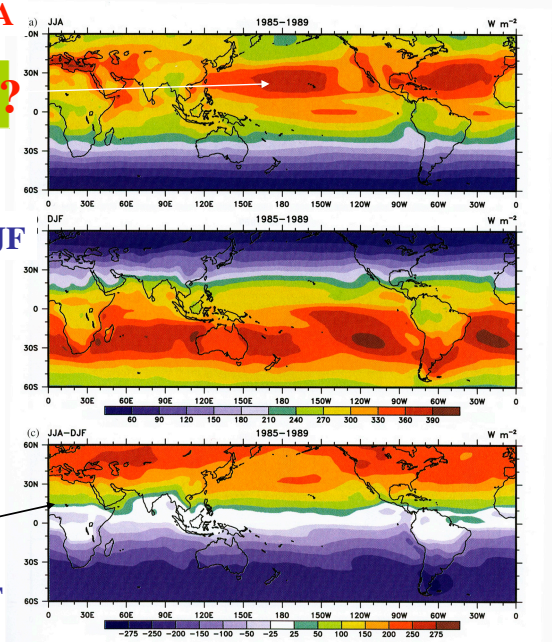


Figure 2. Top of atmosphere (TOA) absorbed solar radiation for the period of the Earth Radiation Budget Experiment, 1985–89 averaged for: (a) June, July and August (JJA), (b) December, January and February (DJF), and (c) their difference JJA–DJF. The contour interval in (a) and (b) is 30 W m^{-2} and in (c) is 25 or 50 W m^{-2}.

Absorbed solar radiation at TOA

Why is the zero-line located north of the equator!?

JJA-DJF

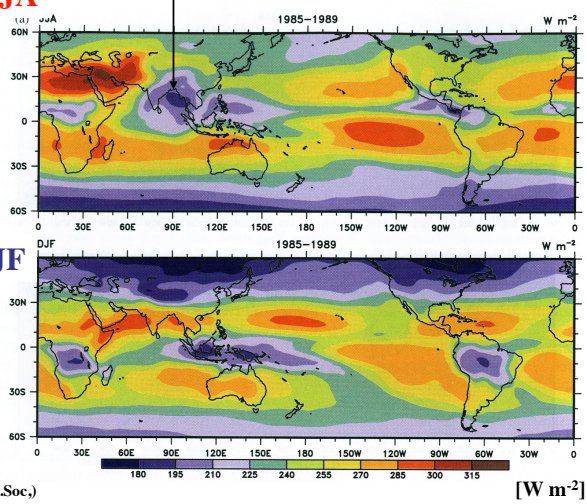
From Trenberth &Stepaniak, 2004 (Q.J.R.Meteorol.Soc,)

[W m^{-2}]

Energy balance at the top of the atmosphere (TOA)

Why minimum?

JJA



DJF

Terrestrial radiation

Also referred to as "Outgoing Longwave Radiation at TOA" (OLR-TOA)

From Trenberth &Stepaniak, 2004 (Q.J.R.Meteorol.Soc,)

[W m^{-2}]

Energy balance at the top of the atmosphere (TOA)

JJA

?

DJF

JJA-DJF

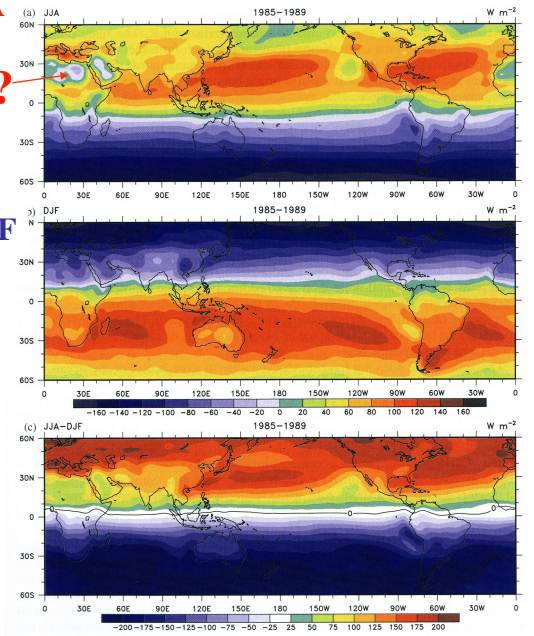
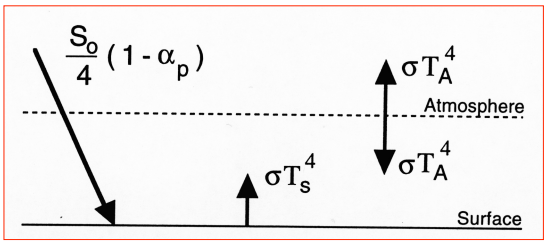


Figure 4. As Fig. 2, but for TOA net radiation. The contour interval in (a) and (b) is 20 W m^{-2} and in (c) is 25 W m^{-2}.

net radiation

From Trenberth &Stepaniak, 2004 (Q.J.R.Meteorol.Soc,)

Greenhouse effect(1)



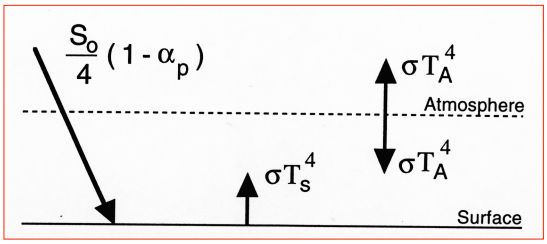
Simplified model of the atmosphere

Assumption: atmosphere is **transparent*** to Solar radiation but **opaque⁺** to terrestrial radiation. Atmosphere and earth are "black bodies"

*transmission coefficient=1

⁺absorbtion coefficient=1

Greenhouse effect(1)



Equilibrium:

System $\quad (S_0/4)(1-\alpha_p)=\sigma T_A{}^4 \equiv \sigma T_e{}^4$

Atmosphere $\quad \sigma T_s{}^4 = 2\sigma T_A{}^4$ } $T_s = \sqrt[4]{2}\,T_e > T_e$

Surface $\quad (S_0/4)(1-\alpha_p)+\sigma T_A{}^4 \equiv \sigma T_s{}^4$

With $T_e = 255$ K, $T_s = 303$ K.

This seems a bit too high. What could be wrong with this model?

Atmosphere absorbs a fraction, ε, of the terrestrial radiation

Greenhouse effect (2)

Emission temperature

$$T_e = \sqrt[4]{\frac{S_0/4(1-\alpha_p)}{\sigma}} = 255 \text{ K}$$

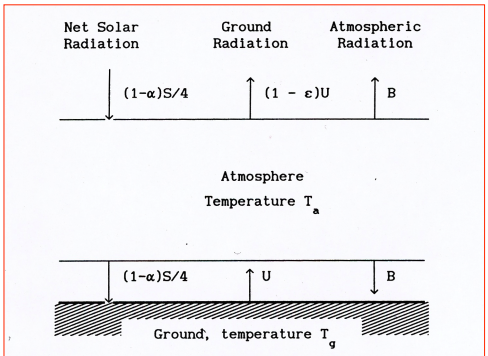


Fig. 3.2. Single slab model of the Earth–atmosphere system.

Atmosphere absorbs a fraction, ε, of the terrestrial radiation

Greenhouse effect (2)

Emission temperature

$$T_e = \sqrt[4]{\frac{S_0/4(1-\alpha_p)}{\sigma}} = 255 \text{ K}$$

Ground (black body) emits energy at rate U of which a fraction, $(1-\varepsilon)$, escapes to space.

$$T_g = \left(\frac{U}{\sigma}\right)^{1/4}, \; T_a = \left(\frac{B}{\varepsilon\sigma}\right)^{1/4}$$

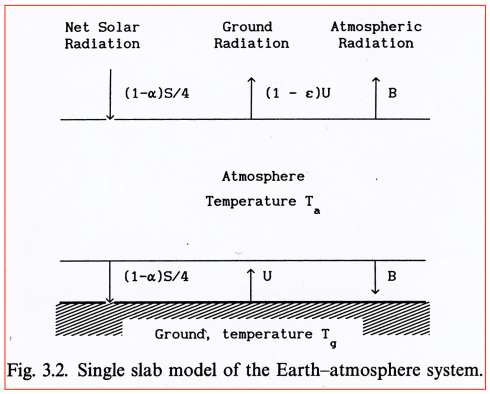


Fig. 3.2. Single slab model of the Earth–atmosphere system.

In a steady state:

$$(1-\alpha)\frac{S_0}{4} + B = U$$

$$2B = \varepsilon U$$

This leads to:

$$T_g = \left\{\frac{2}{2-\varepsilon}\right\}^{1/4} T_e \text{ and } T_a = \left\{\frac{1}{2-\varepsilon}\right\}^{1/4} T_e$$

$$T_g \geq T_e, \; T_a \leq T_e$$

If ε=0.78, T_g=288 K

Section 2.4

How fast is radiative equilibrium reached?

Earth's surface has a negligible heat capacity:

$$(1-\alpha)\frac{S_0}{4} + B = U$$

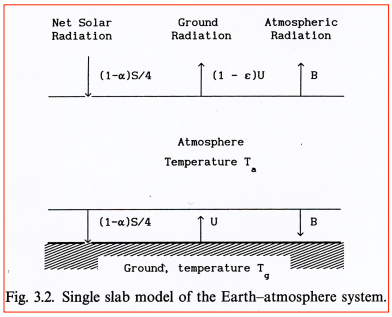


Fig. 3.2. Single slab model of the Earth–atmosphere system.

Let us assume that the atmosphere has a finite heat capacity,

$$C = \frac{c_p p_s}{g} \quad \textbf{[J K}^{-1}\textbf{ m}^{-2}\textbf{]}$$

$$c_p = 1004 \quad \textbf{[J K}^{-1}\textbf{ kg}^{-1}\textbf{]}$$

*p_s/g is the atmospheric mass per unit area; why?

Section 2.4

How fast is radiative equilibrium reached?

Earth's surface has a negligible heat capacity:

$$(1-\alpha)\frac{S_0}{4} + B = U$$

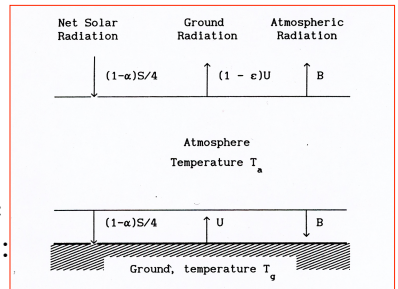


Fig. 3.2. Single slab model of the Earth–atmosphere system.

For an atmospheric column in hydrostatic equilibrium. of unit cross-sectional area *:

$$\frac{c_p p_s}{g}\frac{dT_a}{dt} = \varepsilon U - 2B$$

Assume: $T_a = T_{a0} + \Delta T$

$|\Delta T| << T_{a0}$

Equilibrium value

$$\frac{d(\Delta T)}{dt} \cong -\frac{4(2-\varepsilon)\varepsilon\sigma g T_{a0}^3}{p_s c_p}\Delta T$$

Show this using Taylor's formula+ truncated after the second term

+ See appendix to this lecture

Section 2.4

$$\frac{d(\Delta T)}{dt} = -\frac{4(2-\varepsilon)\varepsilon\sigma g T_{a0}^3}{p_s c_p}\Delta T$$

Solution of this eq. implies exponential decay of ΔT over a time-scale τ_E, i.e.

$$\Delta T \propto \exp(-t/\tau_E)$$

with $$\tau_E = \frac{p_s c_p}{4(2-\varepsilon)\varepsilon\sigma g T_{a0}^3}$$

Which is called the "**radiative equilibrium timescale**".

If τ_E is large relative to the typical time scale temperature changes due to other processes (such as advection and mixing) in the atmosphere, then these processes can maintain the temperature far from radiative equilibrium

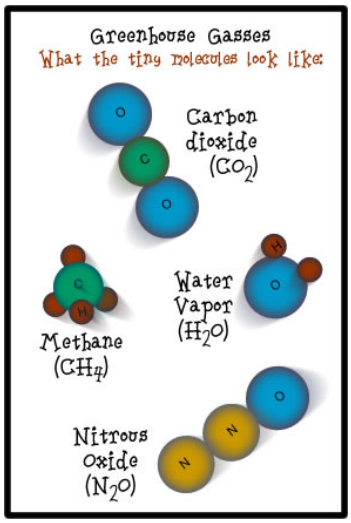
Estimate of rad. eq. time scale

$$\tau_E = \frac{p_s c_p}{4(2-\varepsilon)\varepsilon\sigma g T_{a0}^3}$$

With T=300 K; ε=0.7; g=10 m s^{-2}; p_s=100000 Pa; c_p=1000 J K^{-1} kg^{-1}; $\sigma = 5.67\times10^{-8}$ W m^{-2}K^{-4}

this yields τ_E=20 days

This is rather **slow**, implying that the Earth's atmosphere may **not** be in radiative equilibrium on the daily and seasonal time scale. Daily and seasonal variations in insolation will keep the atmosphere out of radiative equilibrium.



Next lecture

Monday 13/2, 2012, 1100-1245

Oral lecture:
Solution of Schwarzschild equation (box 2.3)
What determines existence and thickness of the stratosphere & troposphere?

Radiative-convective model (RCM) (section 2.7)
Radiative equilibrium (section 2.8)
Radiatively determined state (section 2.9)

After that:
Energy fluxes associated with water cycle (sections 2.10…)

http://www.staff.science.uu.nl/~delde102/climatedynamics.htm

Appendix

Taylor's formula:

$$f(T_0 + \Delta T) = f(T_0) + \Delta T \frac{df}{dT}(T_0) + \text{higher order terms}$$

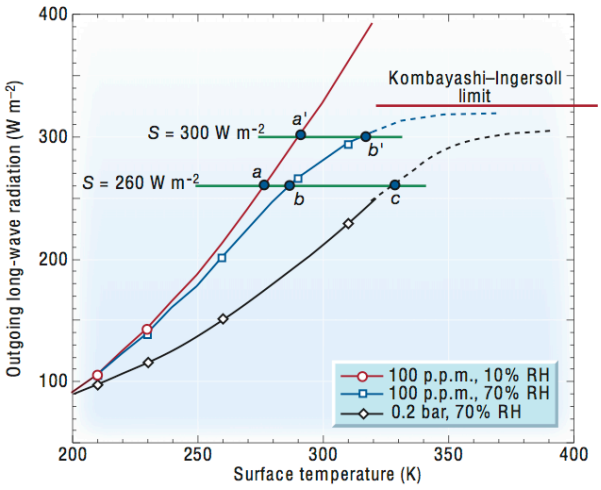
Table 12.1 Physical parameters for the terrestrial planets

	Mercury	Venus	Earth	Mars
Orbital and rotational data				
Mean distance from Sun (km)	5.79×10^7	1.082×10^8	1.496×10^8	2.279×10^8
Eccentricity of orbit	0.2056	0.0068	0.0167	0.0934
Obliquity (°)	0	177	23.45	23.98
Orbital period (Earth days)	87.97	224.701	365.256	686.980
Rotational period (h)	1,407.5	5,832.24	23.9345	24.6229
Solar Day (Earth days)	175.88	117	1	1.0287
Solar Constant (kW m^{-2})	10.4[127]	2.63	1.37	0.593
Planet				
Albedo	0.12	0.76	0.3	0.25
Mass (kg)	3.302×10^{23}	4.870×10^{24}	5.976×10^{24}	6.421×10^{23}
Radius (km)	2,439	6,051.5	6,378 to 6,357	3,398
Surface Gravity (m s^{-2})	3.70	8.60	9.78	3.72
Atmosphere				
Principal Composition	He, Na	CO_2	N_2, O_2	CO_2
Mean molecular weight (amu)	—	43.44	28.98(dry)	43.49
Mean surface temperature (K)	~400	730	288	220
Mean surface pressure (bar)	$<10^{-12}$	92	1	0.007
Mass (kg)	$\sim 10^4$	4.77×10^{20}	5.30×10^{18}	$\sim 10^{16}$

[127] In the case of Mercury this is the mean of a very large range: the insolation varies between 6.3 and 14.5 kW m^{-2} around the orbit.

From Taylor, 2005

Climate sensitivity and the runaway greenhouse effect



Equilibrium outgoing long wave radiation at the top of the atmosphere (OLR-TOA) as a function surface temperature computed with a standard radiative transfer model for (a) an atmosphere with 100 ppm CO_2 and 10% relative humidity, (b) an atmosphere with 100 ppm CO_2 and 70% relative humidity, (c) an atmosphere with 0.2 bar CO_2 and 70% relative humidity. *S* is the absorbed Solar radiation. For further interpretation, see R.T. Pierrhumbert, 2002: The hydrological cycle in deep-time climate problems. **Nature**, **419**, 191-198 (see also the following slides).

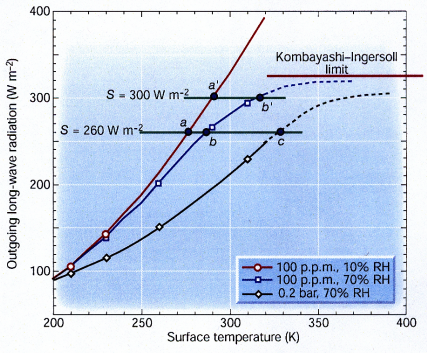
Additional reading

Runaway Greenhouse

Box 1

Climate sensitivity and the dependence of outgoing long-wave radiation on temperature

A planet can lose energy to outer space only through emission of electromagnetic radiation. For the range of planetary temperatures typically considered, this radiation is in the infrared range, and the flux of the outgoing radiation is known as outgoing long-wave radiation (OLR). The temperature of a solar-heated planet settles at the value that allows the OLR to balance the absorbed solar radiation; hence the shape of the dependence of OLR on temperature holds the key to the most fundamental climate phenomena. For an atmosphere-free planet having a uniform surface temperature T, the OLR would satisfy the Stefan–Boltzman law, σT^4. Atmospheres modify the OLR by allowing the planet to radiate at temperatures colder than the planetary surface. The consequent decrease in OLR is a direct consequence of the fact that temperature decreases with height, and thus the strength of the greenhouse effect is sensitive to the lapse rate.



The figure opposite shows OLR as a function of surface temperature, computed under the following assumptions: (1) the vertical temperature profile is given by the moist adiabat, patched to a 200-K isothermal stratosphere, and (2) the specific humidity at each height is a fixed fraction of the saturation value given by the temperature at that height. The latter assumption is known as the 'fixed relative-humidity assumption'. Once these assumptions are made, the OLR can be computed as a function of the surface temperature using any standard radiative-transfer model (the curves here were computed using the model of ref. 25).

Addition of a greenhouse gas to an atmosphere reduces the OLR for fixed T, with the result that temperature must increase in order to bring the radiation budget back into balance. For absorbed solar radiation $S = 260$ W m^{-2}, 100 p.p.m. CO_2 and 10% relative humidity, the temperature is 276 K (point a in the figure). If the relative humidity is increased to 70%, the temperature increases to 288 K (point b). If the greenhouse forcing is further augmented by massively increasing the amount of CO_2 to 0.2 bar, the temperature increases to ~330 K (point c).

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Runaway Greenhouse continued

Because the concentration of water vapour (a greenhouse gas) increases with temperature, it acts to reduce the slope of the OLR curve. This increases the sensitivity of climate to changes in radiative forcing, as is illustrated in the figure by increasing solar absorption to 300 W m^{-2}. With 10% relative humidity, the warming from point a to a' is 14 K, but with 70% relative humidity, the warming from point b to b' is 30 K. This is known as the water vapour feedback. The ultimate water vapour feedback is the runaway greenhouse, manifest in the fact that the flattening of the OLR curve at high temperatures leads it to asymptote to a value known as the Kombayashi–Ingersoll limit. This is the limit to how fast a planet with a moist atmosphere can loose energy by infrared radiation. For a saturated atmosphere, the limit depends primarily on the gravity of the planet, which determines the mass of atmosphere contained above a given pressure level. The precise value one obtains for the limit depends on the treatment of infrared absorption properties of water vapour at high temperature and pressure. In the figure above, we have sketched in the value from ref. 42 (about 320 W m^{-2}), and the reader is referred there for further details.

When the solar forcing exceeds the Kombayashi–Ingersoll limit, the temperature continues to increase until the surface temperature is several thousand degrees, at which point all the ocean has been evaporated into the atmosphere. This is followed by disassociation of water into hydrogen and oxygen, after which the light hydrogen escapes to space making the water loss irreversible. Once there is no precipitating water reaching the ground, it becomes impossible to bind up atmospheric carbon dioxide in carbonate rocks. This scenario, or some version of it, is the prevailing view of how Venus was consigned to a climate so much less habitable than Earth's. The Kombayashi–Ingersoll limit is also relevant to impact heating of planets during their early stages, as it determines how quickly a planet can condense out the transient post-impact steam atmosphere.

The assumption of fixed relative humidity is a convenient starting point for simple models, but as the atmosphere aloft is highly undersaturated, it is dubious to consider humidity to be strictly a function of local temperature. Atmospheric humidity is in reality governed by a bewildering array of dynamical and microphysical processes for which useful idealized models are not yet available. Clouds pose an even greater challenge. The picture of the runaway greenhouse given above was based on clear-sky reasoning, but a saturated atmosphere would be thick with clouds. These clouds would block nearly all solar radiation, but would also be very effective at limiting OLR. The resulting climate would be determined by the delicate balance between a trickle of incoming solar radiation and a trickle of OLR, much as has been proposed to result from CO_2 clouds on early Mars. There have been a few tentative steps towards including clouds in the runaway greenhouse calculation[42], but the cloud physics regime in which the cloud-forming substance is the primary component of the atmosphere is a very unfamiliar one, and offers many exciting prospects for future work.

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