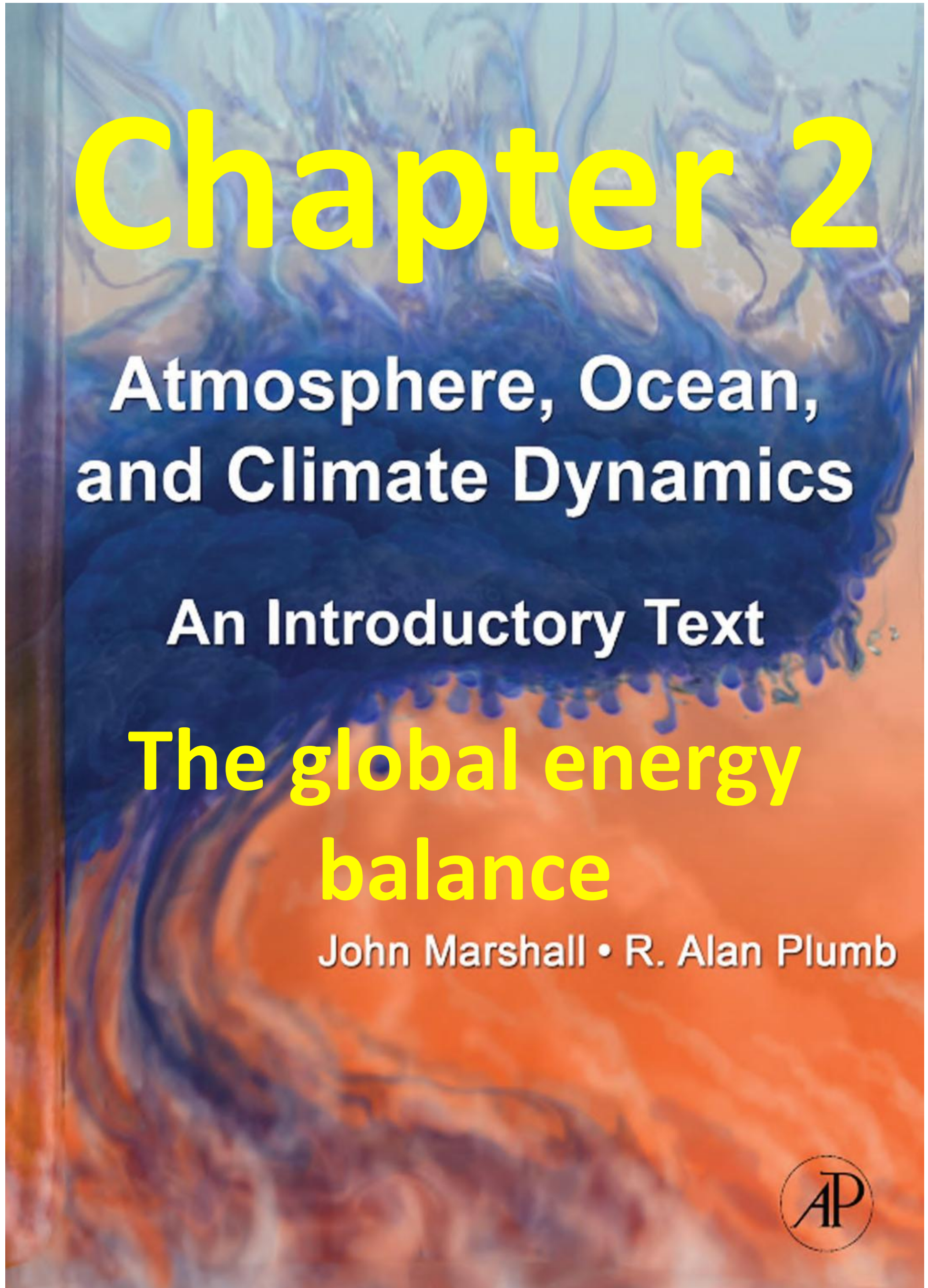
Chapter 2

The global energy balance



Jordas Energibalanse

Verdensrommet er vakuum ➔ Energi kan bare utveksles som stråling
Stråling: Elektromagnetisk stråling

Inn: Solstråling

Ut:
Reflektert solstråling +
Langbølget varmestråling fra Jorda

Stråling: W/m^2

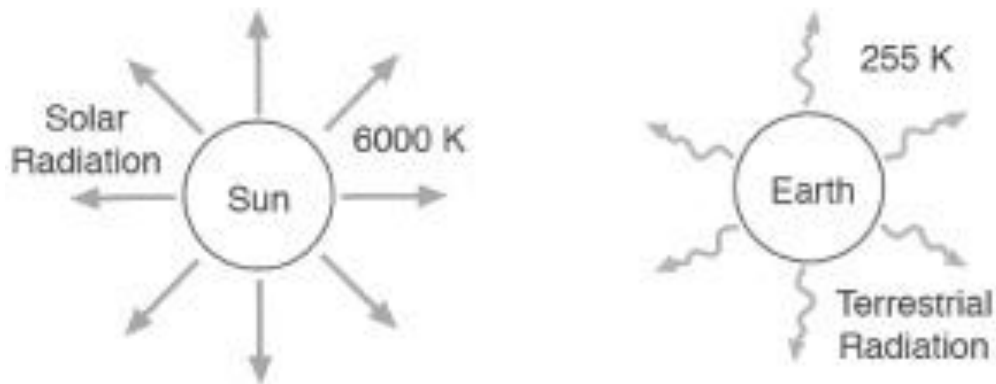


Figure 2.1: The Earth radiates energy at the same rate it is received from the Sun. The Earth's emission temperature is 255 K, and that of the Sun is 6000 K. The outgoing terrestrial radiation peaks in the infrared spectrum; the incoming solar radiation peaks at shorter wavelengths, in the visible spectrum.

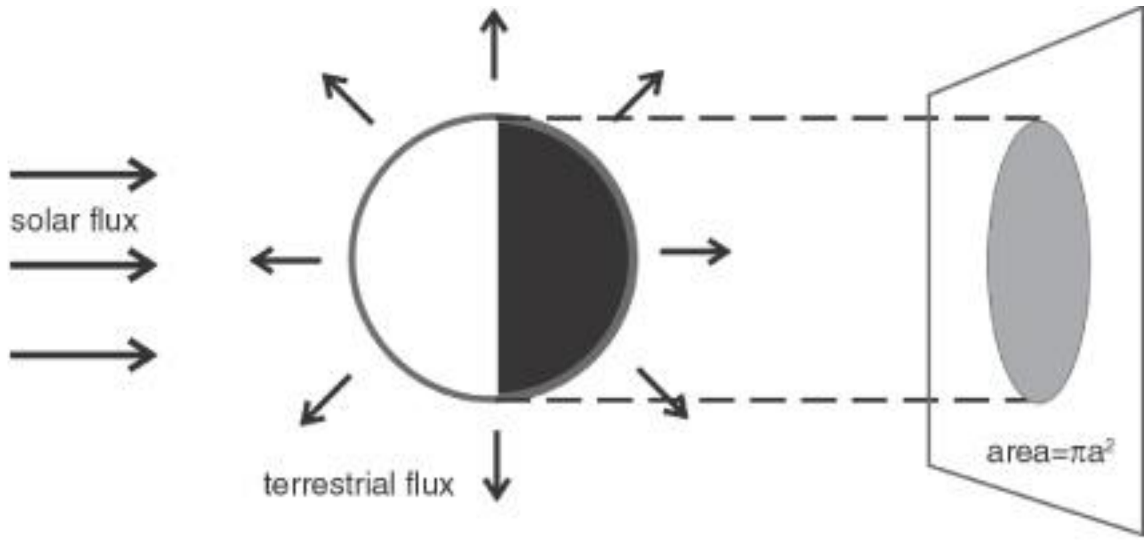


Figure 2.4: The spinning Earth is imagined to intercept solar energy over a disk of radius a and radiate terrestrial energy away isotropically from the sphere.

(Modified from Hartmann, 1994.)

Solarkonstant (S_0): Innkommende solstråling ved toppen av atmosfæren på en flate normalt på solstrålingen

S_0= 1367 W/m^2

Stråling fra et svartlegeme

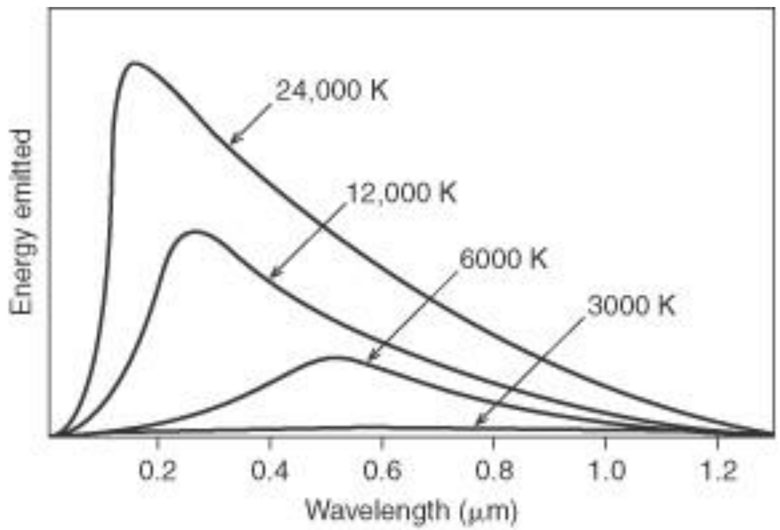


Figure 2.3: The energy emitted at different wavelengths for blackbodies at several temperatures. The function $B_\lambda (T)$, Eq. A-1, is plotted.

$$u_\nu(T) = \frac{8\pi\nu^2}{c^3}\frac{h\nu}{e^{h\nu/k_BT}-1}$$

Et svart legeme (Black Body) absorberer all innkommende stråling.
Det vil også sende ut det maksimalt mulige ut fra sin temperatur

Intensiteten i strålingen som funksjon av bølgelengde er gitt ved Plancks lov (fig 2.3)

Totalt utstrålt energi er gitt ved arealet under kurvene i fig 2.3.

Gitt ved Stefan Boltzmanns lov

$E=\sigma T^4$

σ :Stefans Boltzmanns konstant

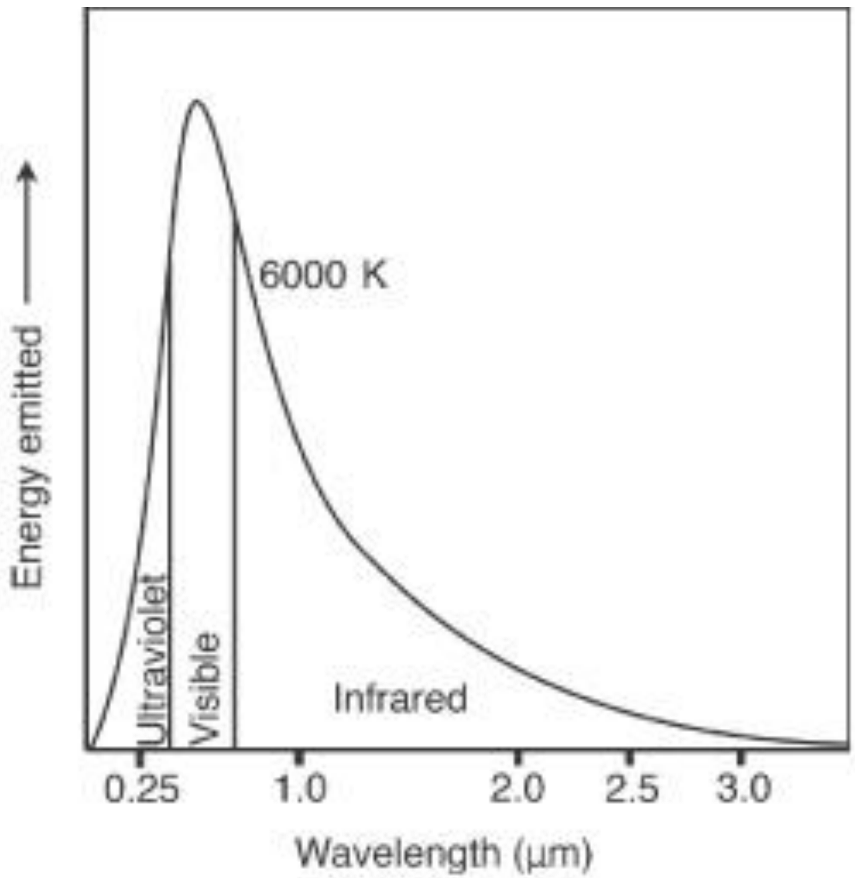


Figure 2.2: The energy emitted from the Sun plotted against wavelength based on a black body curve with $T = T_{Sun}$. Most of the energy is in the visible spectrum, and 95% of the total energy lies between 0.25 and 2.5μm (10^{-6}m).

Planetary emission temperature

solar power incident on the Earth =

$$S_0 \pi a^2 = 1.74 \times 10^{17}\ \text{W},$$

Solar radiation absorbed by the Earth =

$$(1 - \alpha_p) S_0 \pi a^2 = 1.22 \times 10^{17}\ \text{W}. \quad (2\text{-}1)$$

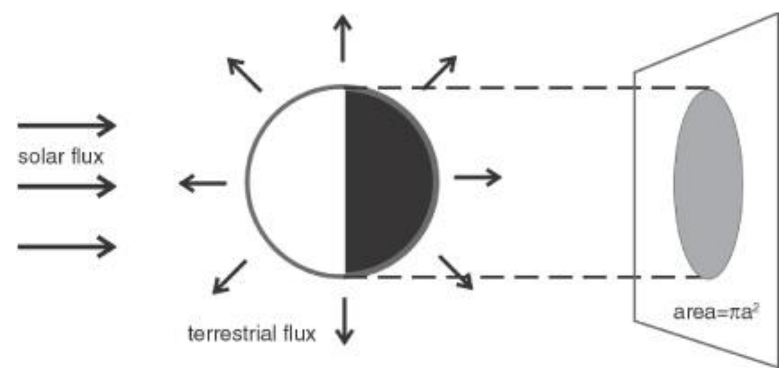


Figure 2.4: The spinning Earth is imagined to intercept solar energy over a disk of radius a and radiate terrestrial energy away isotropically from the sphere.

(Modified from Hartmann, 1994.)

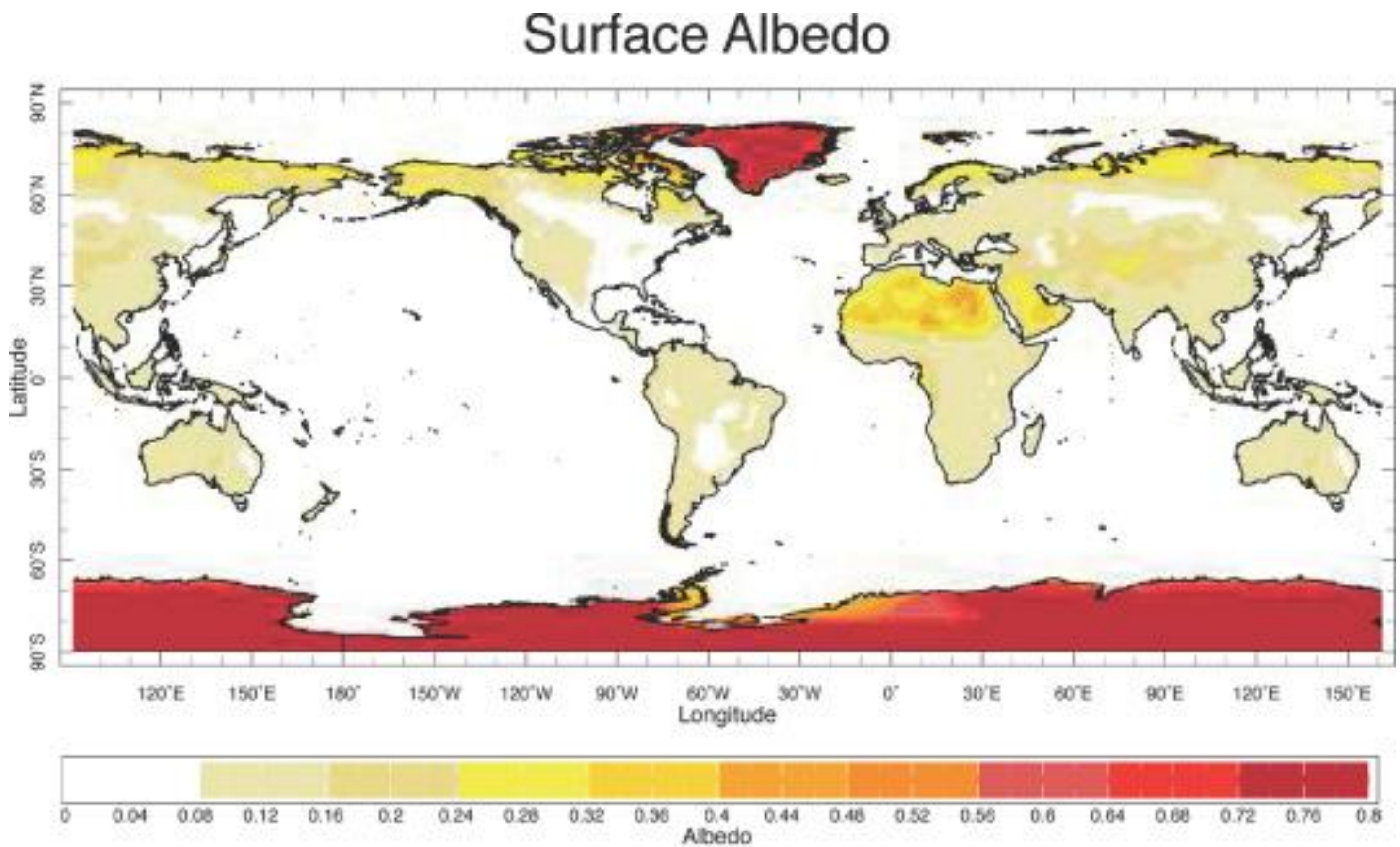
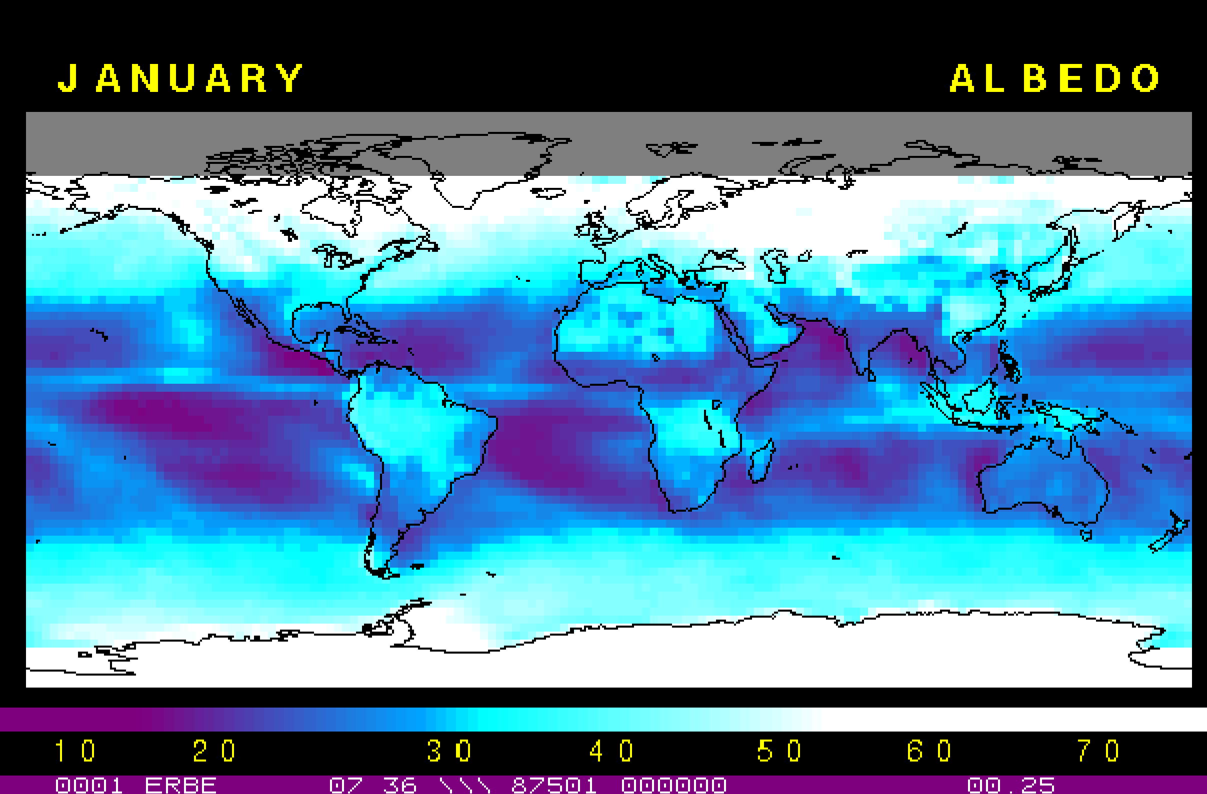
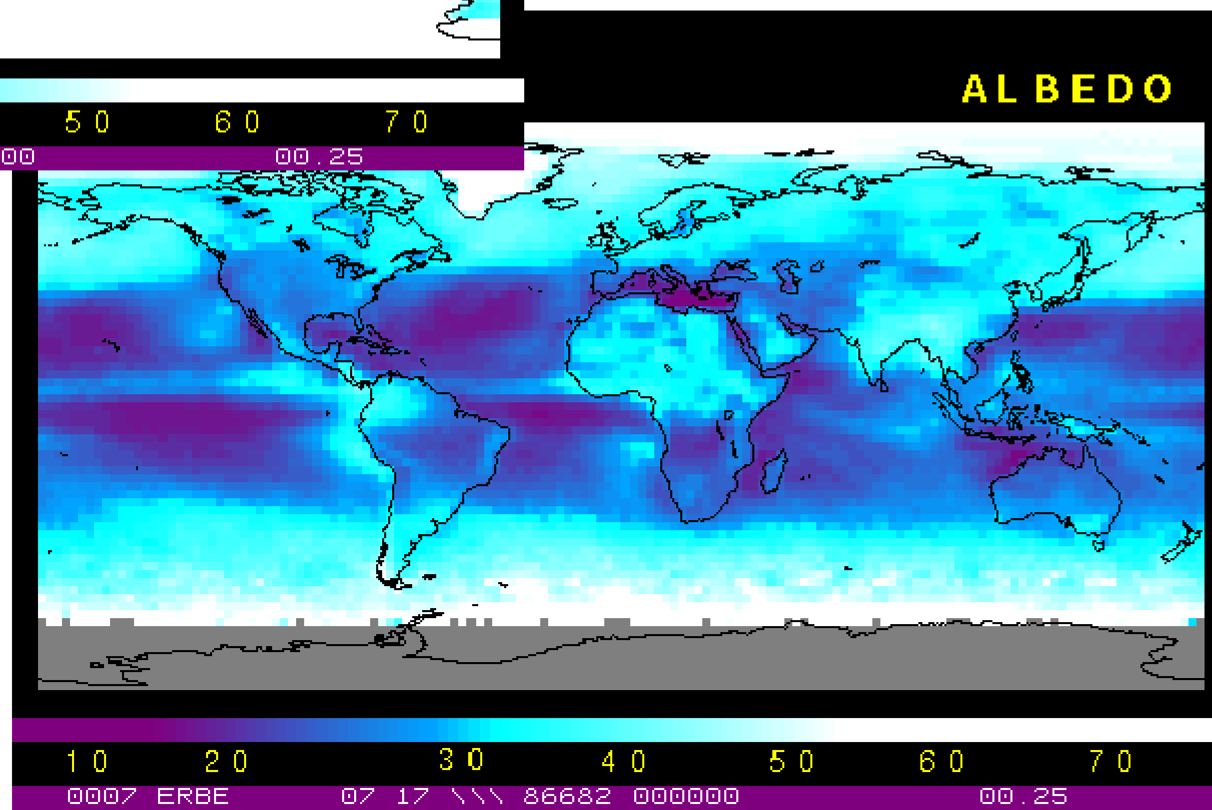


Figure 2.5: The albedo of the Earth's surface. Over the ocean the albedo is small (2–10%). It is larger over the land (typically 35–45% over desert regions) and is particularly high over snow and ice (~80%) (see Table 2.2).





http://cimss.ssec.wisc.edu/wxwise/homerbe.html

Planetær strålingstemperatur

Stefan-Boltzmann law gives:

$$\text{Emitted radiation per unit area} = \sigma T_e^4 \quad (2\text{-}2)$$

where $\sigma = 5.67 \times 10^{-8} \mathrm{W\,m^{-2}\,K^{-4}}$ is the Stefan-Boltzmann constant. So

$$\text{Emitted terrestrial radiation} = 4\pi a^2 \sigma T_e^4. \quad (2\text{-}3)$$

Note that Eq. 2-3 is a **definition** of *emission temperature* T_e. It is the temperature one would infer by looking back at Earth if a blackbody curve was fitted to the measured spectrum of outgoing radiation.

Planetary emission temperature

Absorbert solstråling = Utsendt langbølget stråling

$$(1-\alpha_p)S_0\pi a^2 = 4\pi a^2\sigma T_e^4$$

Equating Eq. 2-1 with Eq. 2-3 gives

$$T_e = \left[\frac{S_0(1-\alpha_p)}{4\sigma}\right]^{1/4} . \qquad (2\text{-}4)$$

T_e = 255 K = -18 °C

- 255K ➔ z≈5 km

NB! Høyden for effektiv utstråling varierer veldig med bølgelengden

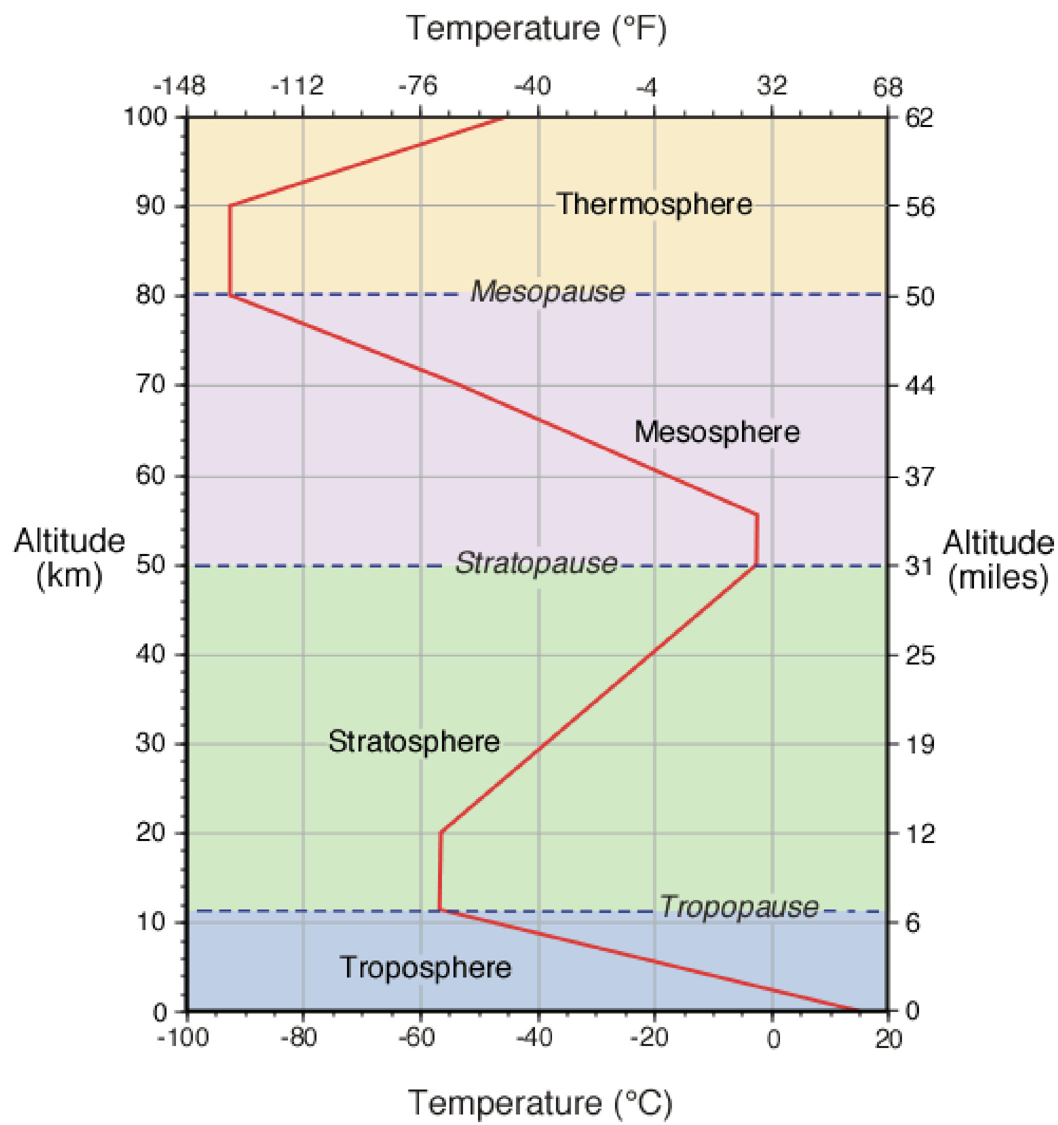


TABLE 2.1. Properties of some of the planets. S_0 is the solar constant at a distance r from the Sun, α_p is the planetary albedo, T_e is the emission temperature computed from Eq. 2 4, T_m is the measured emission temperature, and T_s is the global mean surface temperature. The rotation period, τ, is given in Earth days.

	r 10^9 m	S_0 W m^{-2}	α_p	T_e K	T_m K	T_s K	τ Earth days
Venus	108	2632	0.77	227	230	760	243
Earth	150	1367	0.30	255	250	288	1.00
Mars	228	589	0.24	211	220	230	1.03
Jupiter	780	51	0.51	103	130	134	0.41

TABLE 2.1. Properties of some of the planets. S_0 is the solar constant at a distance r from the Sun, α_p is the planetary albedo, T_e is the emission temperature computed from Eq. 2-4, T_m is the measured emission temperature, and T_s is the global mean surface temperature. The rotation period, τ, is given in Earth days.

TABLE 2.2. Albedos for different surfaces. Note that the albedo of clouds is highly variable and depends on the type and form. See also the horizontal map of albedo shown in Fig. 2.5.

Type of surface	Albedo (%)
Ocean	2–10
Forest	6–18
Cities	14–18
Grass	7–25
Soil	10–20
Grassland	16–20
Desert (sand)	35–45
Ice	20–70
Cloud (thin, thick stratus)	30, 60–70
Snow (old)	40–60
Snow (fresh)	75–95

TABLE 2.2. Albedos for different surfaces. Note that the albedo of clouds is highly variable and depends on the type and form. See also the horizontal map of albedo shown in Fig. 2.5.

The atmospheric absorption spectrum

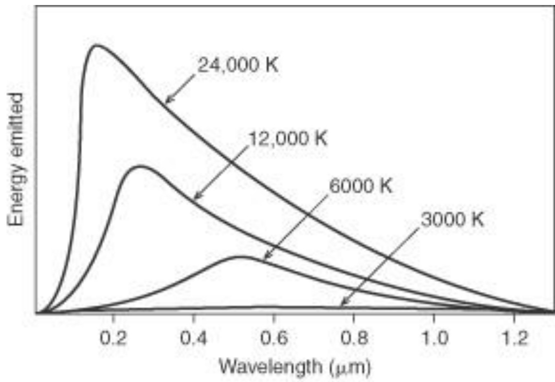
A property of the blackbody radiation curve is that the wavelength of maximum energy emission, λ_m, satisfies

$$\lambda_m T = \text{constant}. \qquad (2\text{-}5)$$

This is known as *Wien's displacement law.*

$\lambda_m = 0.6 \mu m$ for Sola,

T_e er hhv. 6000K og 255K for Sola og Jorda



$$\lambda_m^{Earth} = 0.6\,\mu\text{m} \times \frac{6000}{255} \simeq 14\,\mu\text{m}.$$

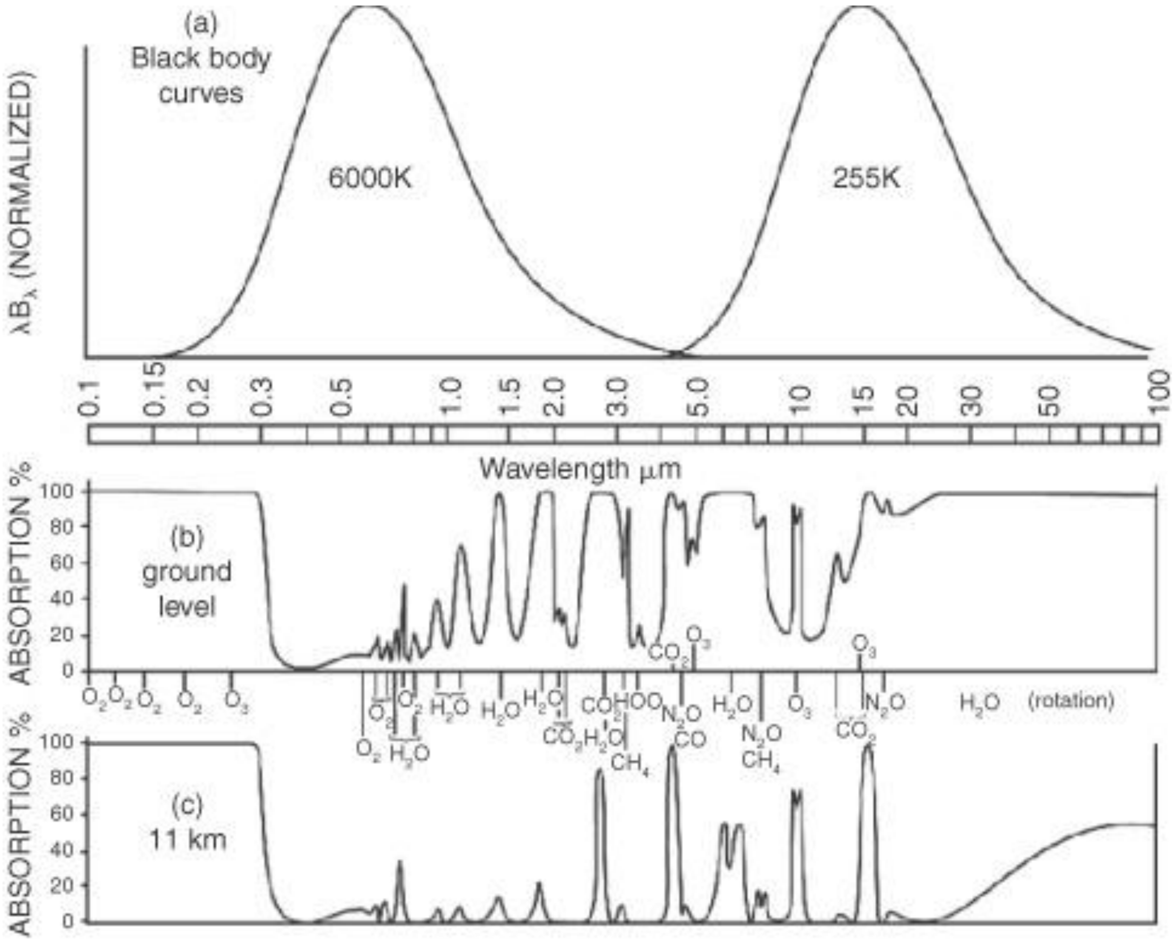
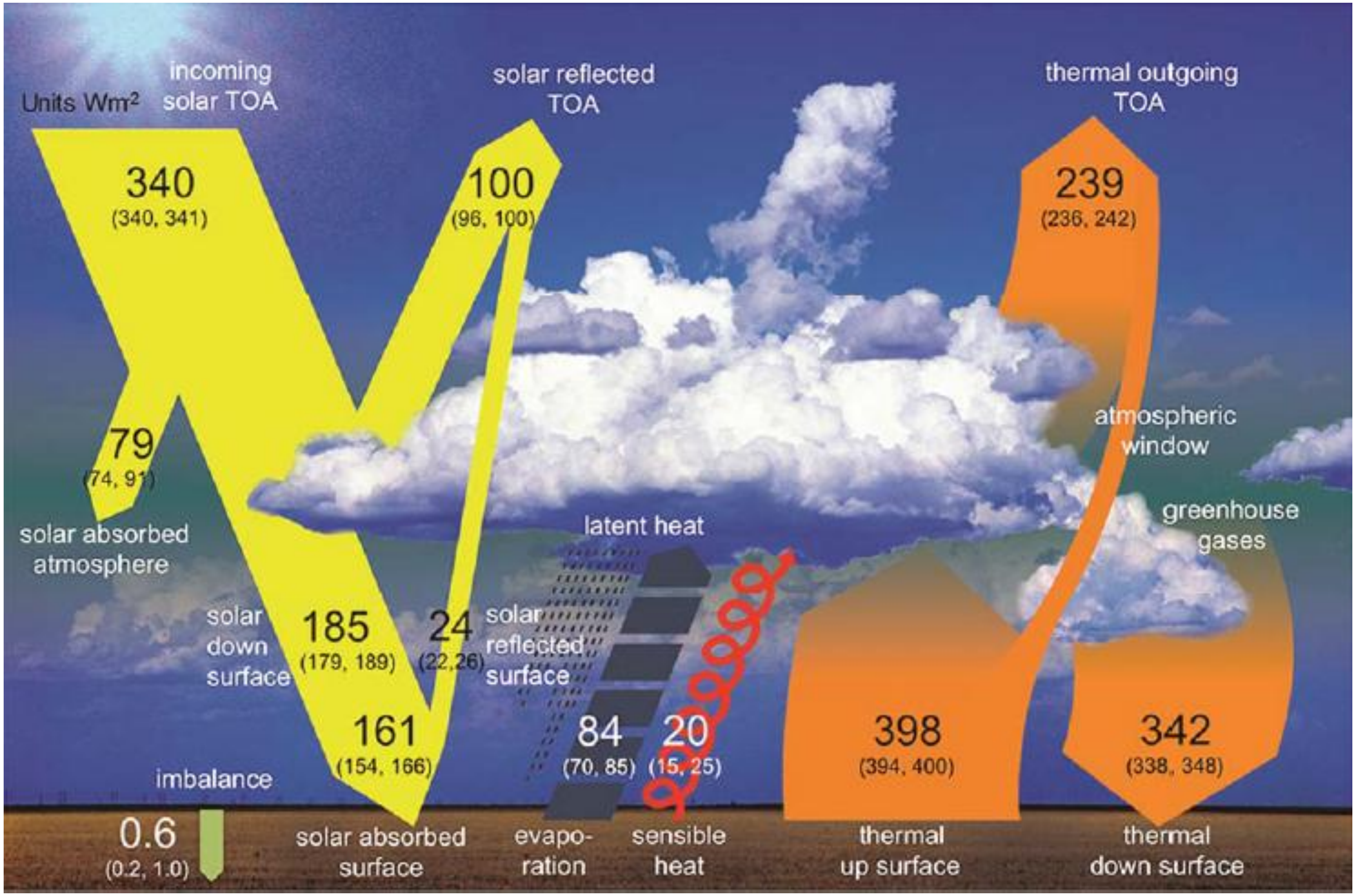


Figure 2.6: (a) The normalized blackbody emission spectra, T-4λ B_λ, for the Sun (T = 6000 K) and Earth (T = 255 K) as a function of ln λ (top), where $B\lambda$ is the blackbody function (see Eq. A-2) and λ is the wavelength (see Appendix A.1.1 for further discussion). (b) The fraction of radiation absorbed while passing from the ground to the top of the atmosphere as a function of wavelength. (c) The fraction of radiation absorbed from the tropopause (typically at a height of 11 km) to the top of the atmosphere as a function of wavelength. The atmospheric molecules contributing the important absorption features at each frequency are also indicated. After Goody and Yung (1989).



Units Wm2
incoming
solar TOA
340
(340, 341)
solar reflected
TOA
100
(96, 100)
thermal outgoing
TOA
239
(236, 242)
79
(74, 91)
solar absorbed
atmosphere
atmospheric
window
greenhouse
gases
latent heat
solar
down
surface
185
(179, 189)
24
(22,26)
solar
reflected
surface
161
(154, 166)
84
(70, 85)
20
(15, 25)
398
(394, 400)
342
(338, 348)
imbalance
0.6
(0.2, 1.0)
solar absorbed
surface
evapo-
ration
sensible
heat
thermal
up surface
thermal
down surface

The greenhouse effect

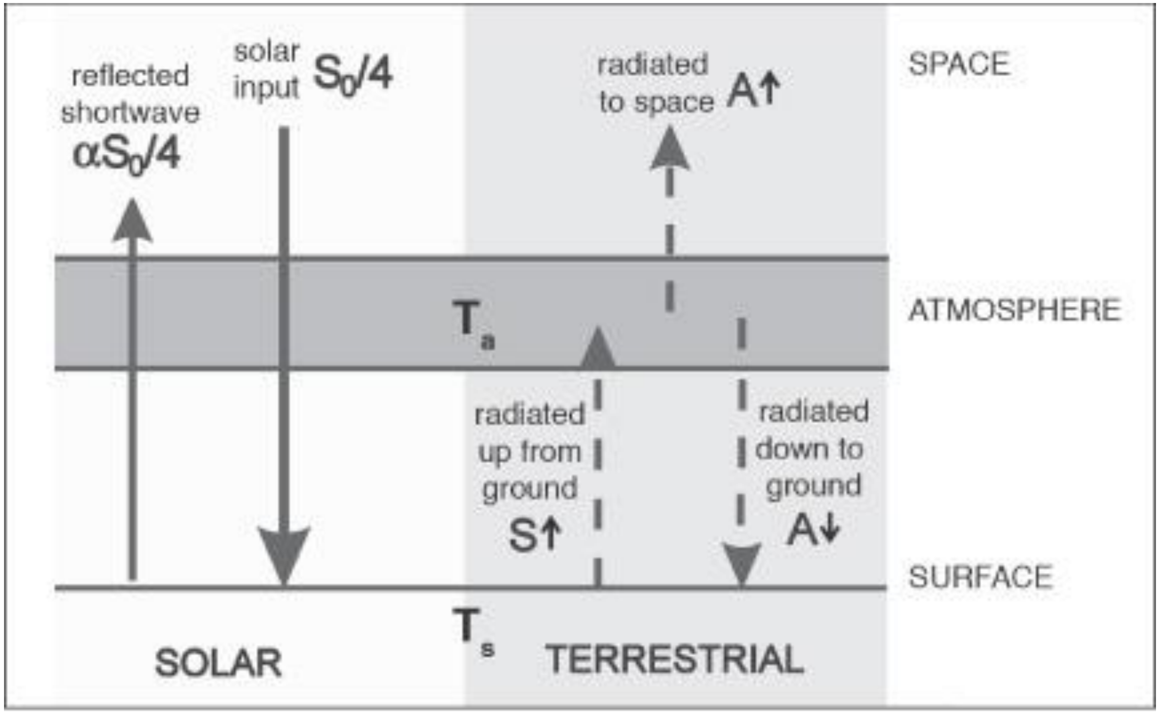


Figure 2.7: The simplest greenhouse model, comprising a surface at temperature T_s, and an atmospheric layer at temperature T_a, subject to incoming solar radiation $S_o/4$. The terrestrial radiation upwelling from the ground is assumed to be completely absorbed by the atmospheric layer.

A simple greenhouse model

Consider Fig. 2.7. Since the atmosphere is thin, let us simplify things by considering a planar geometry, in which the incoming radiation per unit area is equal to the average flux per unit area striking the Earth. This average incoming solar energy *per unit area of the Earth's surface* is

$$\begin{aligned}\text{average solar energy flux}&\\ &= \frac{\text{intercepted incoming radiation}}{\text{Earth's surface area}}\\ &= \frac{S_0 \pi a^2}{4\pi a^2} = \frac{S_0}{4}. \qquad (2\text{-}6)\end{aligned}$$

A simple greenhouse model

Now, since the whole Earth-atmosphere system must be in equilibrium (on average), the net flux into the system must vanish. The average net solar flux per unit area is, from Eq. 2-6, and allowing for reflection, $1/4(1-\alpha_p)\,S_0$, whereas the terrestrial radiation emitted to space per unit area is, using Eq. 2-2:

$$A\uparrow = \sigma T_a^4.$$

Equating them, we find:

$$\sigma T_a^4 = \frac{1}{4}\left(1-\alpha_p\right) S_0 = \sigma T_e^4, \qquad (2\text{-}7)$$

A simple greenhouse model

At the surface, the average incoming shortwave flux is also $1/4\left(1-\alpha_p\right)S_0$, but there is also a downwelling flux emitted by the atmosphere,

$$A\downarrow = \sigma T_a^4 = \sigma T_e^4.$$

A simple greenhouse model

The flux radiating upward from the ground is

$$S \uparrow = \sigma T_s^4,$$

where T_s is the surface temperature. Since, in equilibrium, the net flux at the ground must be zero,

$$S \uparrow = \frac{1}{4}\left(1 - \alpha_p\right) S_0 + A \downarrow,$$

whence

$$\sigma T_s^4 = \frac{1}{4}\left(1 - \alpha_p\right) S_0 + \sigma T_e^4 = 2\sigma T_e^4, \qquad (2\text{-}8)$$

where we have used Eq. 2-7. Therefore

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

$T_s = 2^{1/4} \mathrm{x} 255\ \mathrm{K} = 303\mathrm{K}$

A leaky greenhouse

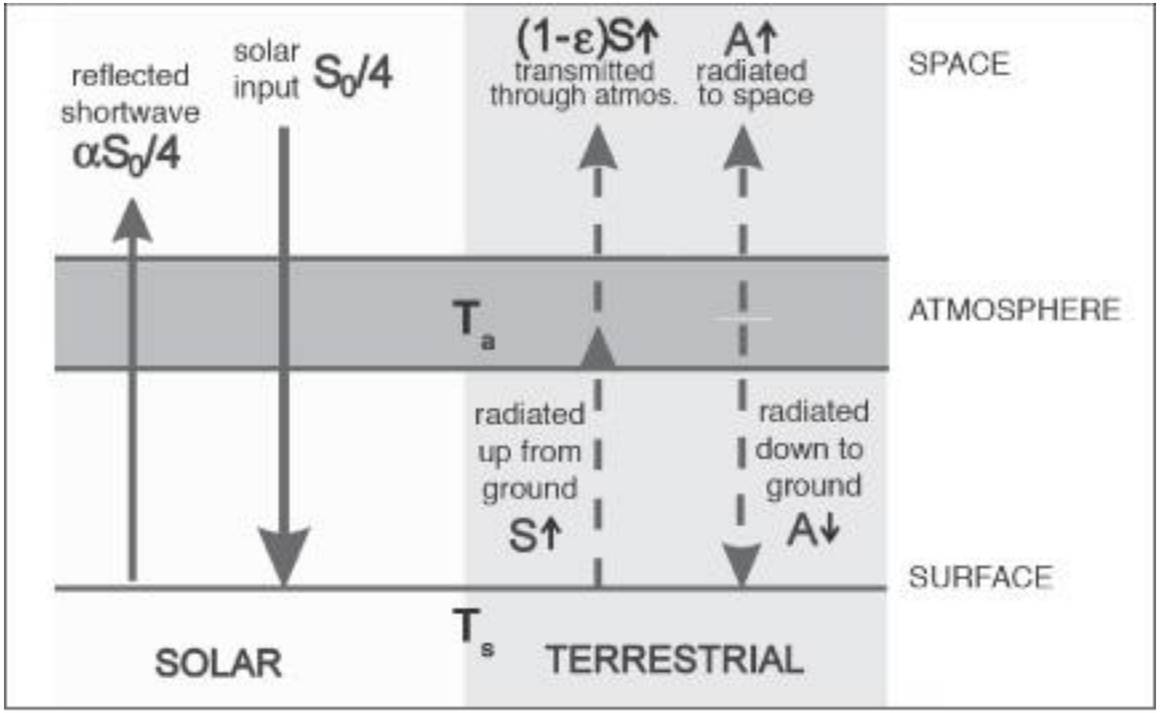


Figure 2.8: A leaky greenhouse. In contrast to Fig. 2.7, the atmosphere now absorbs only a fraction, ε, of the terrestrial radiation upwelling from the ground.

A leaky greenhouse

Consider Fig. 2.8. We suppose the atmosphere has absorptivity ϵ, that is, a fraction ϵ of the IR upwelling from the surface is absorbed within the atmosphere (so the case of Fig. 2.7 corresponds to $\epsilon = 1$). Now again if we insist that in equilibrium the net flux at the top of the atmosphere vanishes, we get

$$\frac{1}{4}(1-\alpha_p)S_0 = A\uparrow + (1-\epsilon)\,S\uparrow. \qquad (2\text{-}10)$$

Zero net flux at the surface gives

$$\frac{1}{4}\left(1-\alpha_p\right)S_0 + A\downarrow = S\uparrow. \qquad (2\text{-}11)$$

Since at equilibrium, $A\uparrow = A\downarrow$, we have

$$S\uparrow = \sigma T_s^4 = \frac{1}{2\,(2-\epsilon)}\left(1-\alpha_p\right)S_0 = \frac{2}{(2-\epsilon)}\sigma T_e^4. \qquad (2\text{-}12)$$

Therefore,

$$T_s = \left(\frac{2}{2-\epsilon}\right)^{1/4} T_e. \qquad (2\text{-}13)$$

A leaky greenhouse

To find the atmospheric temperature, we need to invoke *Kirchhoff's law*,[3] such that the emissivity of the atmosphere is equal to its absorptivity. Thus

$$A \uparrow = A \downarrow = \epsilon \sigma T_a^4. \qquad (2\text{-}14)$$

We can now use Eqs. 2-14, 2-10, 2-11, and 2-12 to find

$$T_a = \left(\frac{1}{2 - \epsilon} \right)^{1/4} T_e = \left(\frac{1}{2} \right)^{1/4} T_s.$$

En Atmosfære med flere lag som absorberer langbølget stråling

Her:
Hvert lag absorberer 100% av langbølget stråling

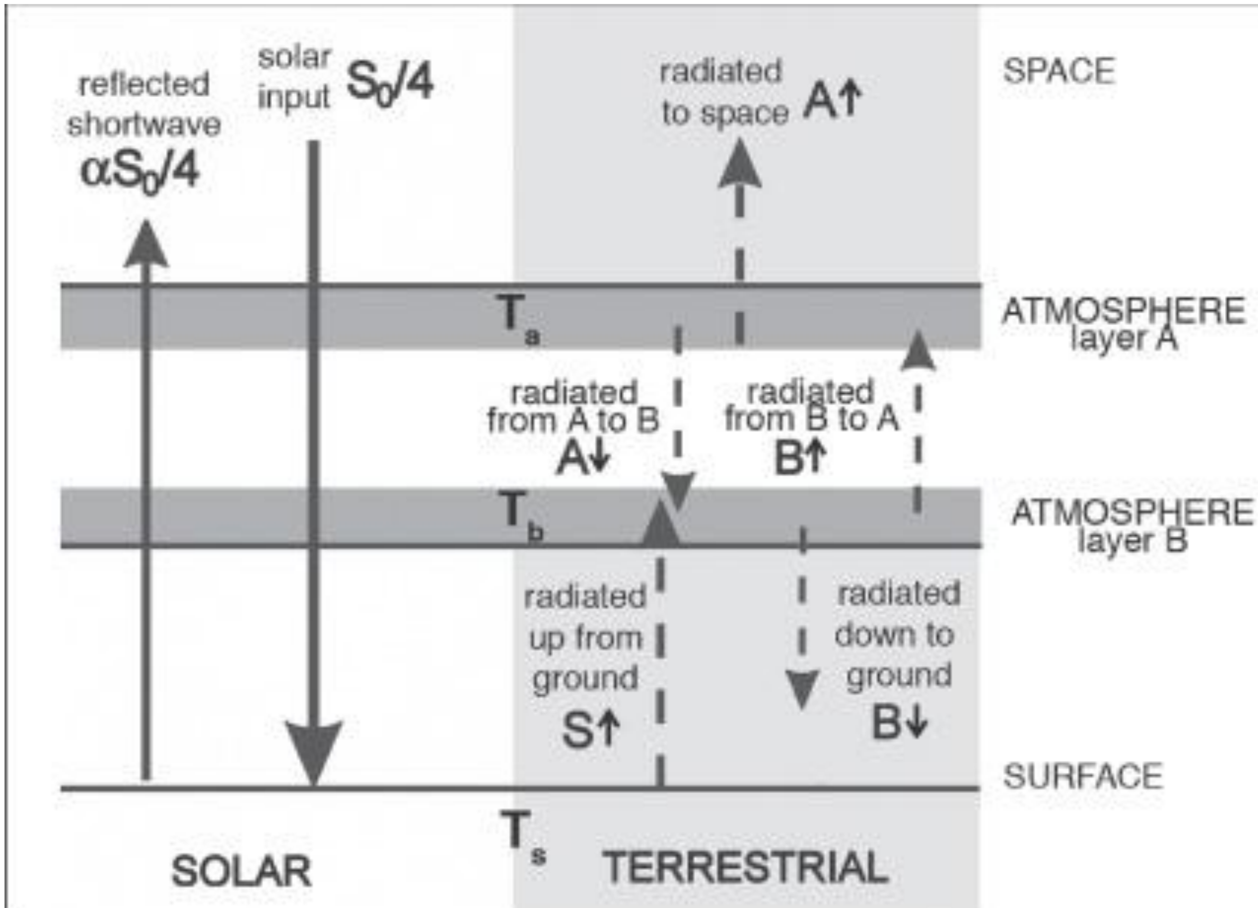


Figure 2.9: An "opaque" greenhouse made up of two layers of atmosphere. Each layer completely absorbs the IR radiation impinging on it.

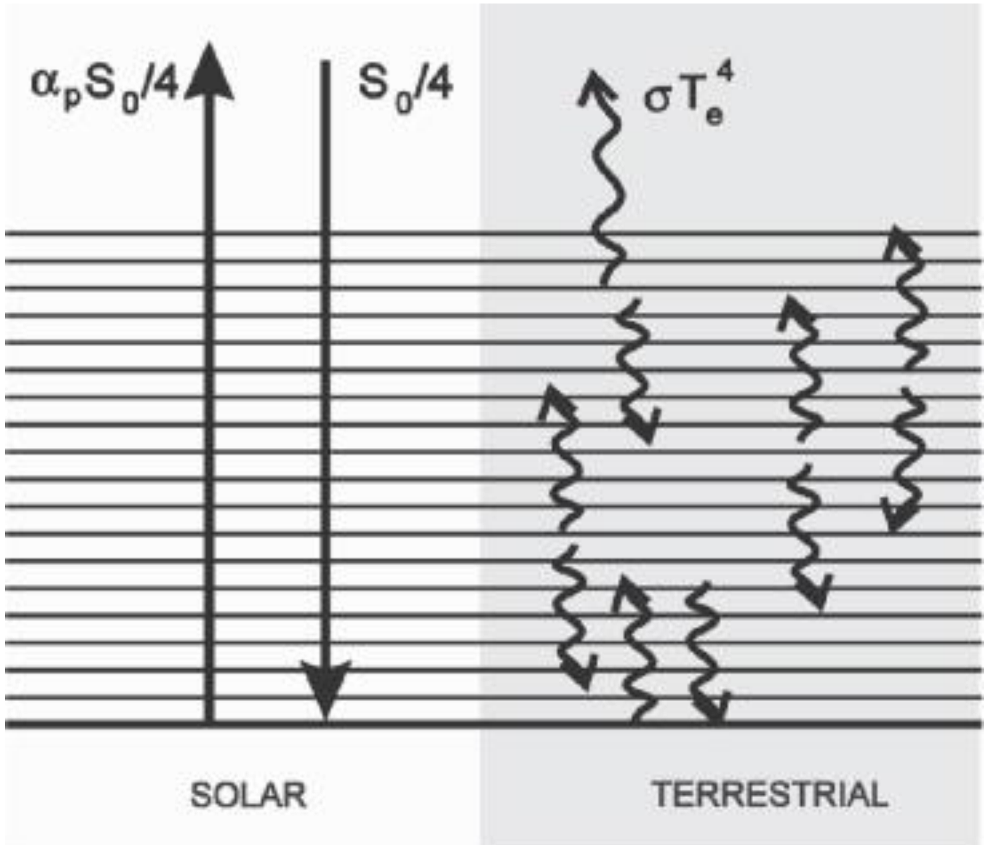
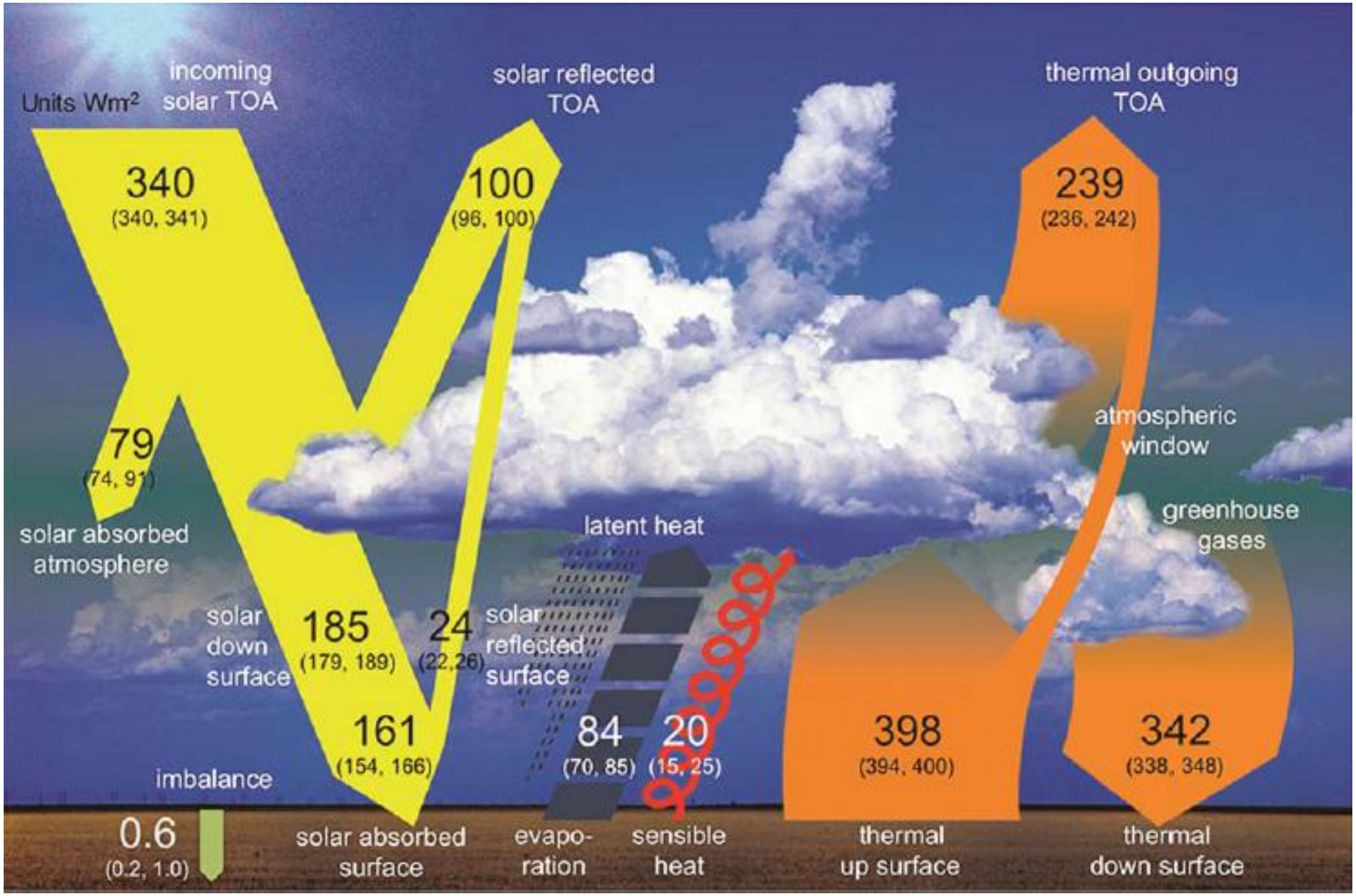


Figure 2.10: Schematic of a radiative transfer model with many layers.



Units Wm²
incoming solar TOA
340
(340, 341)
solar reflected TOA
100
(96, 100)
thermal outgoing TOA
239
(236, 242)
79
(74, 91)
solar absorbed atmosphere
atmospheric window
greenhouse gases
latent heat
solar down surface
185
(179, 189)
24
(22,26)
solar reflected surface
161
(154, 166)
84
(70, 85)
20
(15, 25)
398
(394, 400)
342
(338, 348)
imbalance
0.6
(0.2, 1.0)
solar absorbed surface
evapo-ration
sensible heat
thermal up surface
thermal down surface

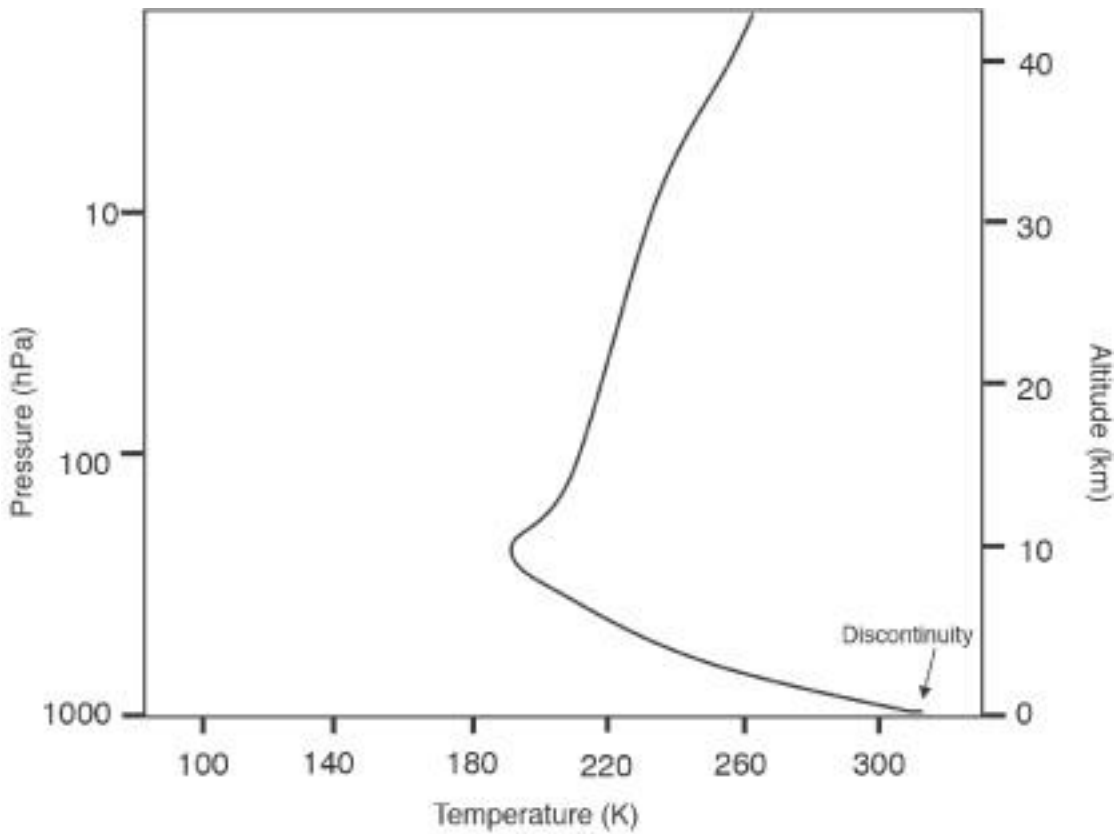
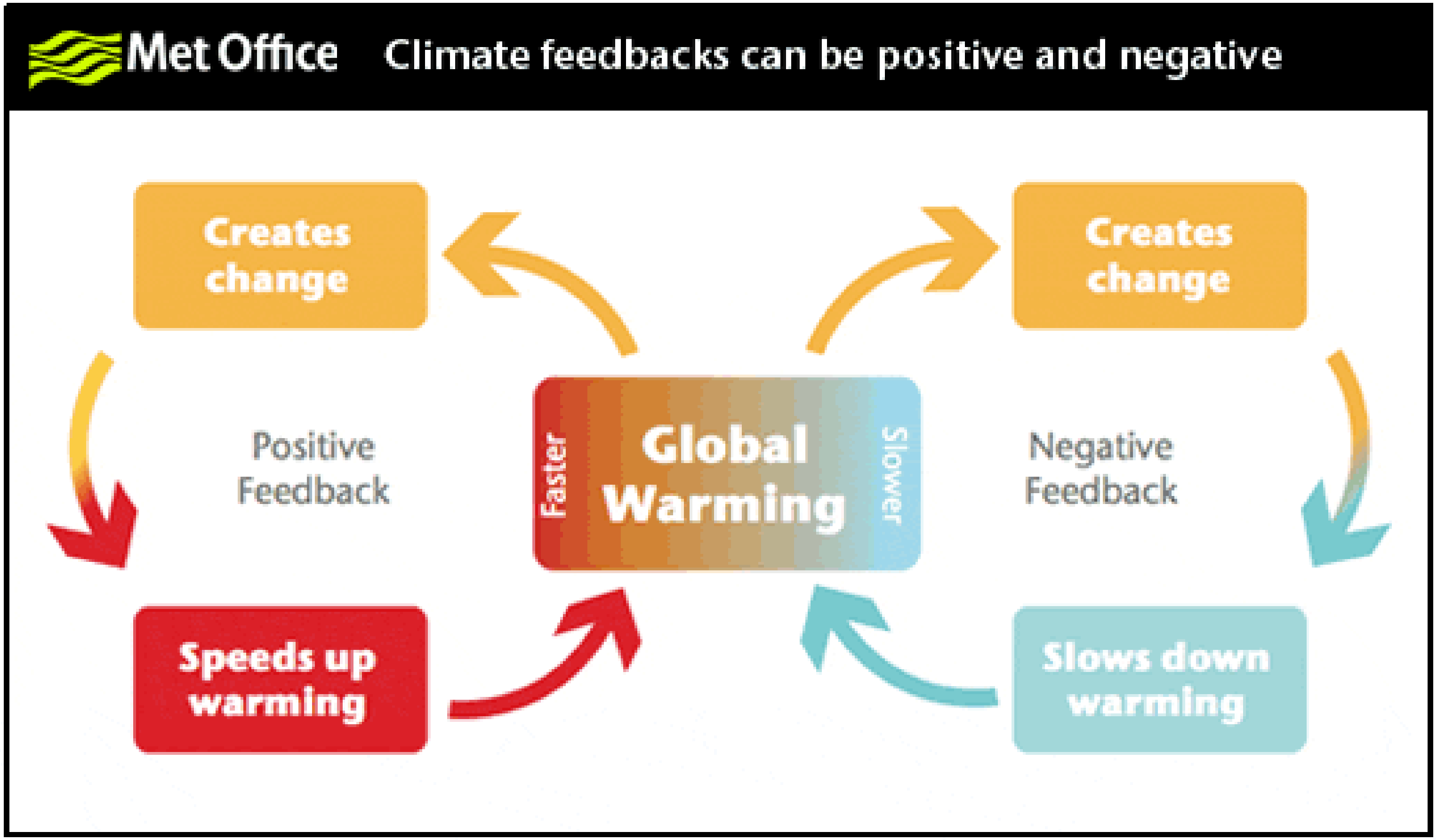


Figure 2.11: The radiative equilibrium profile of the atmosphere obtained by carrying out the calculation schematized in Fig. 2.10. The absorbers are H_2O, O_3, and CO_2. The effects of both terrestrial radiation and solar radiation are included. Note the discontinuity at the surface.

(Modified from Wells (1997).)

Tilbakekoblinger (Climate feedbacks)



Climate feedbacks

The greenhouse models described previously illustrate several important radiative feedbacks that play a central role in regulating the climate of the planet. Following Hartmann (1994) we suppose that a perturbation to the climate system can be represented as an additional energy input dQ (units $\mathrm{W\,m^{-2}}$) and study the resultant change in global-mean surface temperature, dT_s. Thus we define $\partial T_s/\partial Q$ to be a measure of climate sensitivity.

Pådriv og respons

ΔQ : Ytre pådriv (W/m^2)

- Endring i mengden av drivhusgasser
- Endring i innstråling fra Sola (S_0)
- Vulkanutbrudd ➔ Gir partikler i atmosfæren som endrer albedoen

$\frac{d}{dt}\Delta T$: Endring i global temperatur (K/s)

C_m : Jordas varmekapasitet (J/(K·m^2))

$C_m \frac{d}{dt}\Delta T$: Endring i Jordas termiske energi (W/m^2)

Klimafølsomhet og tilbakekoblinger

$$C_m \frac{d}{dt}\Delta T = \Delta Q - \alpha \Delta T$$

Equilibrium: $\Delta Q = \alpha \Delta T$

Climate feedback parameter $\alpha = \Delta Q / \Delta T$

Climate sensitivity parameter $\lambda = \alpha^{-1}$

$\lambda = \Delta T / \Delta Q$

IPCC terminology

Climate sensitivity and feedbacks

$$C_m \frac{d}{dt} \Delta T = \Delta Q - \alpha \Delta T$$

Climate sensitivity parameter $\lambda = \Delta T / \Delta Q$

Climate sensitivity (2 x CO_2) $\Delta Q = 3.7\ Wm^{-2}$

IPCC terminology

Løsning av differensiallikningen over gitt at pådriv og klimafølsomhet ikke varierer med t, og at $\Delta T(t=0) = 0$

$$\Delta T(t) = \frac{\Delta Q}{\alpha}(1 - e^{-\frac{\alpha}{C_m}t})$$

Hva er en rimelig verdi for C_m?

Bidrag fra

- Havet
- Atmosfæren
- Landjorda m/dypere lag i Jorda
- Isbreer

Hva vi tar med kommer an på tidsskalaen vi er interessert i

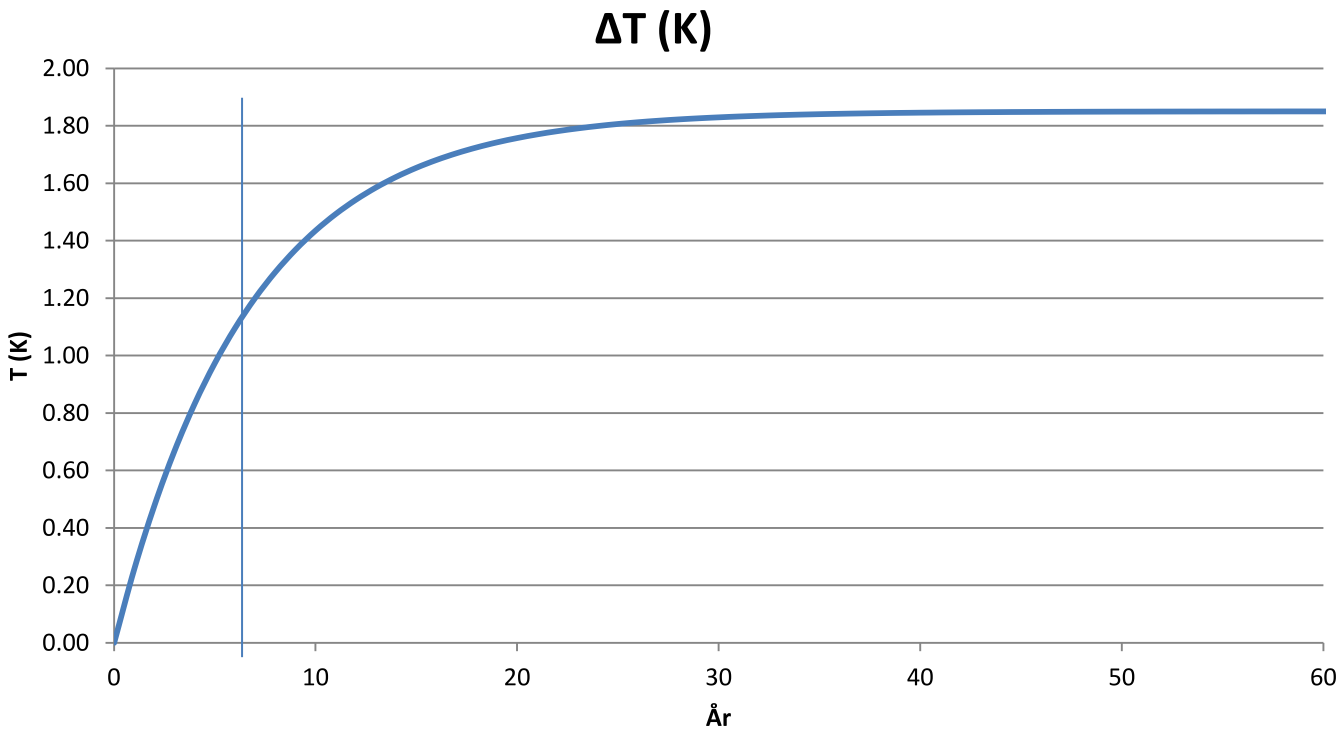
Med blandingslaget i havet (100m, tidsskala dekade)

$C_m = 100 \text{ m} * 10^3 \text{ kg/m}^3 * c_w$ c_w: 4218 J/(K·kg)

➔ $= 4.2 \cdot 10^8$ J/(K·m^2)

Tidsutvikling mot likevekt ved en dobling av CO_2

- $\Delta Q = 3.7\ Wm^{-2}$
- Bruker ca. verdi for α fra IPCC: $\alpha = 2\ Wm^{-2}/K$



Responstid:
$C_m/\alpha = 6.7$ år

Climate feedbacks

The most important negative feedback regulating the temperature of the planet is the dependence of the outgoing long-wave radiation on temperature. If the planet warms up, then it radiates more heat back out to space. Thus using Eq. 2-2 and setting $\delta Q = \delta\left(\sigma T_e^4\right) = 4T_e^3 \delta T_s$, where it has been assumed that T_e and T_s differ by a constant, implies a climate sensitivity associated with blackbody radiation of

$$\frac{\partial T_s}{\partial Q}_{\mathrm{BB}} = \left(4\sigma T_e^3\right)^{-1} = 0.26 \frac{\mathrm{K}}{\mathrm{W\,m^{-2}}}, \qquad (2\text{-}15)$$

Climate feedbacks

Baseline, no-feedback response. At equlibrium

$$\Delta Q = \delta(\sigma T_e^4) = 4\sigma T_e^3 \cdot \delta T_e \stackrel{1}{=} 4\sigma T_e^3 \cdot \delta T_s$$

$$\frac{\partial T_s}{\partial Q}_{BB} = \left(4\sigma T_e^3\right)^{-1} = 0.26 \frac{\mathrm{K}}{\mathrm{W\,m^{-2}}}, \quad (2\text{-}15)$$

Denne størrelsen kalles ofte α_0^{-1}

Definition feedback factor f $\alpha = \alpha_0 (1-f)$

1 Antar at det er en konstant forskjell mellom T_e og T_s .

Climate feedbacks

A powerful positive climate feedback results from the temperature dependence of saturated water vapor pressure, e_s, on T; see Eq. 1-4. If the temperature increases, the amount of water that can be held at saturation increases. Since H_2O is the main greenhouse gas, this further raises surface temperature. From Eq. 1-4 we find that

$$\frac{de_s}{e_e} = \beta dT,$$

and so, given that $\beta = 0.067°C^{-1}$, a 1°C change in temperature leads to a full 7% change in saturated specific humidity. The

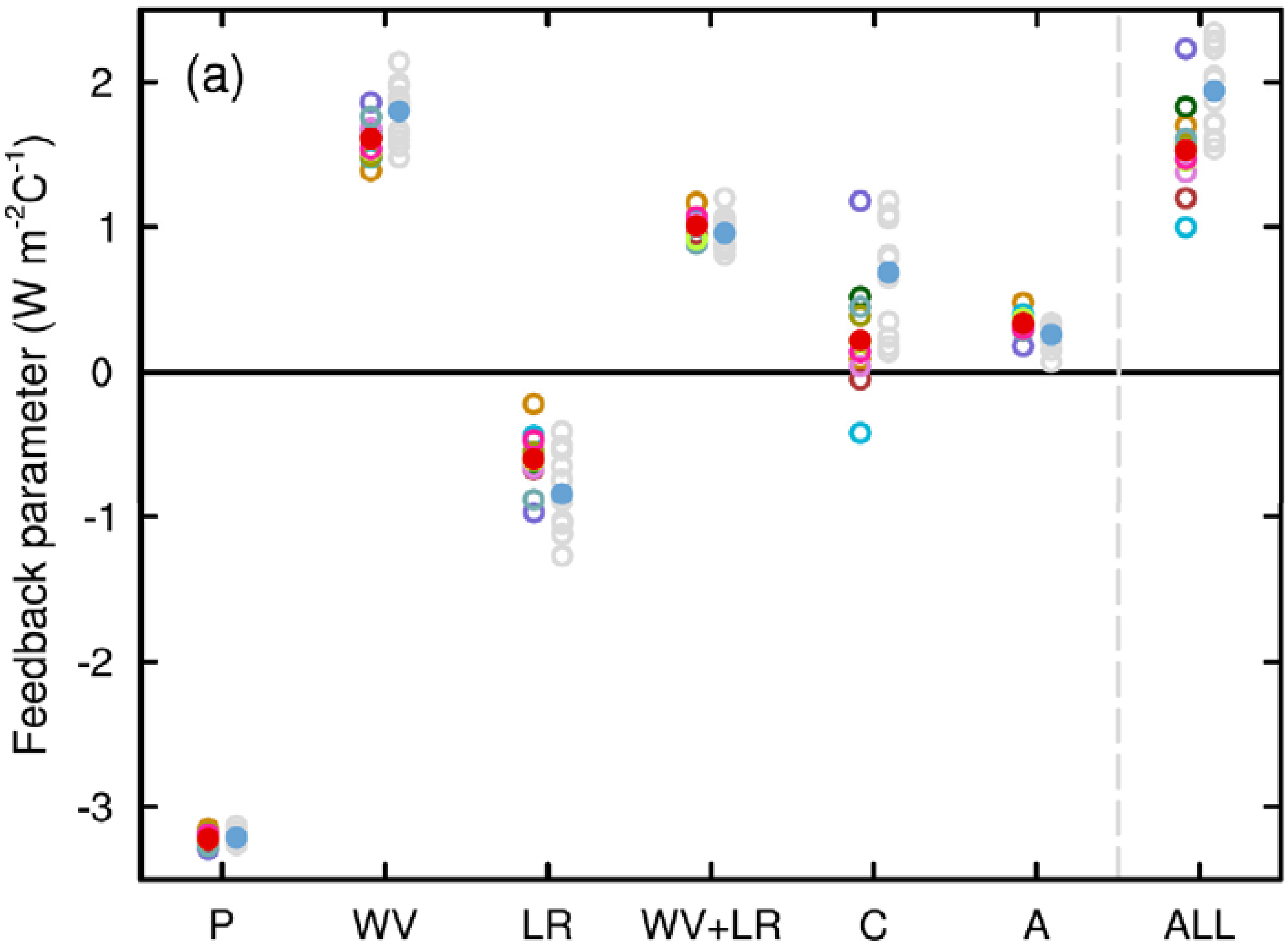
Climate feedbacks

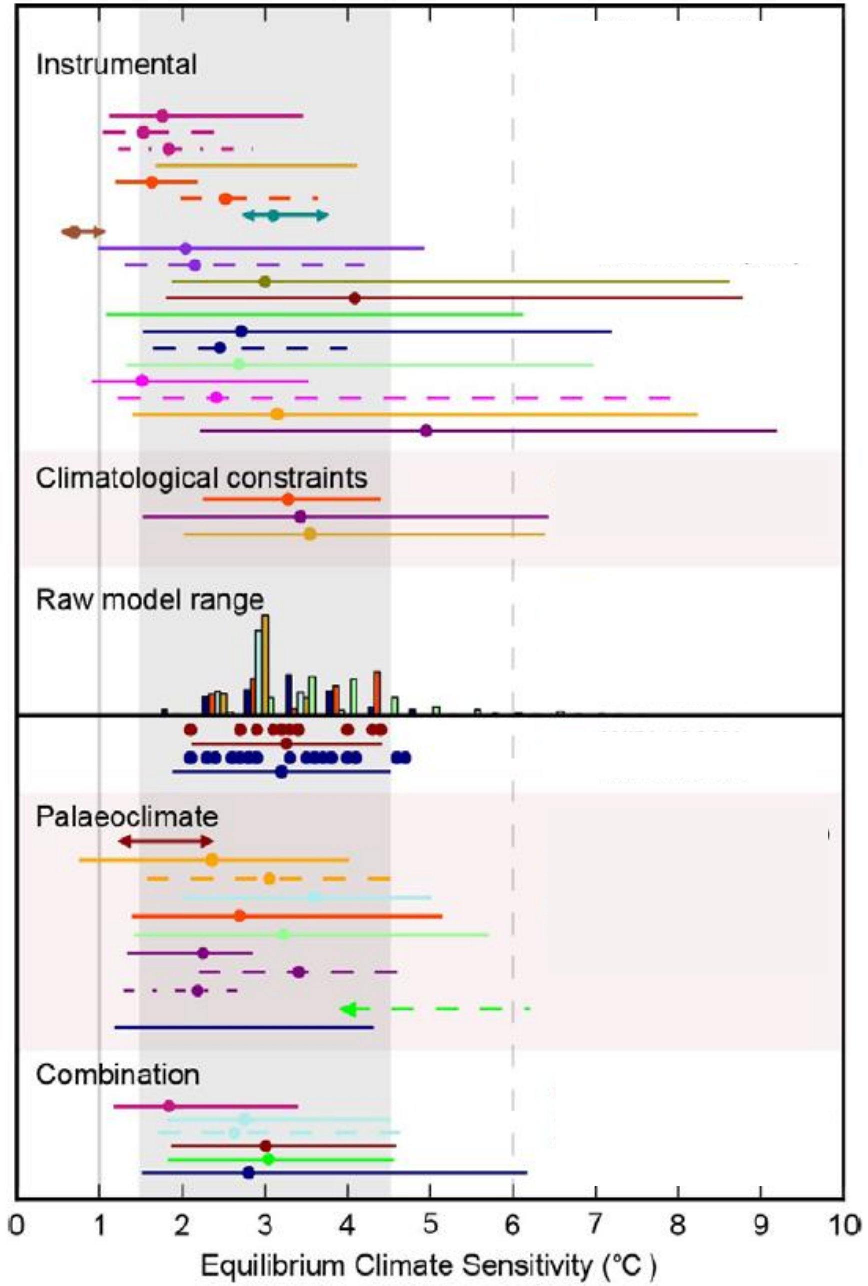
Stefan-Boltzmann and water vapor feedbacks are combined, calculations show that the climate sensitivity is

$$\frac{\partial T_s}{\partial Q}_{\text{BB and H}_2\text{O}} = 0.5 \frac{\text{K}}{\text{W m}^{-2}} ,$$

which is twice that of Eq. 2-15.

Physical feedbacks (Planck, Water Vapour, Lapse Rate, Albedo)





IPCC AR5 Working Group I
Climate Change 2013: The Physical Science Basis

Equilibrium Climate Sensitivity - Klimafølsomhet

Temperatur ved dobling av CO_2 når klimasystemet har nådd en ny likevekt.

Estimater fra perioden med observert temperatur og bruk av klimadrivere.

Estimater fra palaoeklima informasjon om temperatur og bruk av klimadrivere.

Estimater basert på den fysisk forståelsen av klimasystemet fra klimamodellene.

Klimafølsomheten sannsynlig mellom 1.5 og 4.5 °C, ekstremt usannsynlig mindre enn 1 °C og svært usannsynlig større enn 6 °C.

INTERGOVERNMENTAL PANEL ON climate change

WMO UNEP