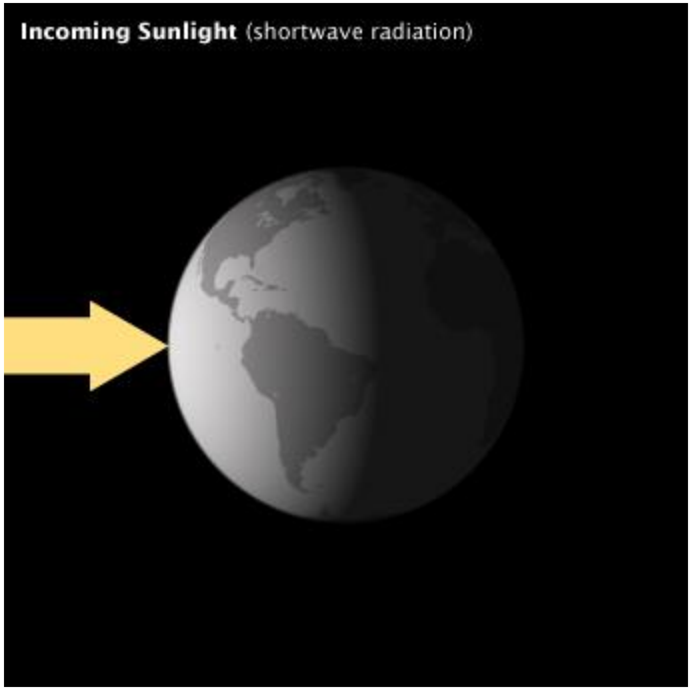
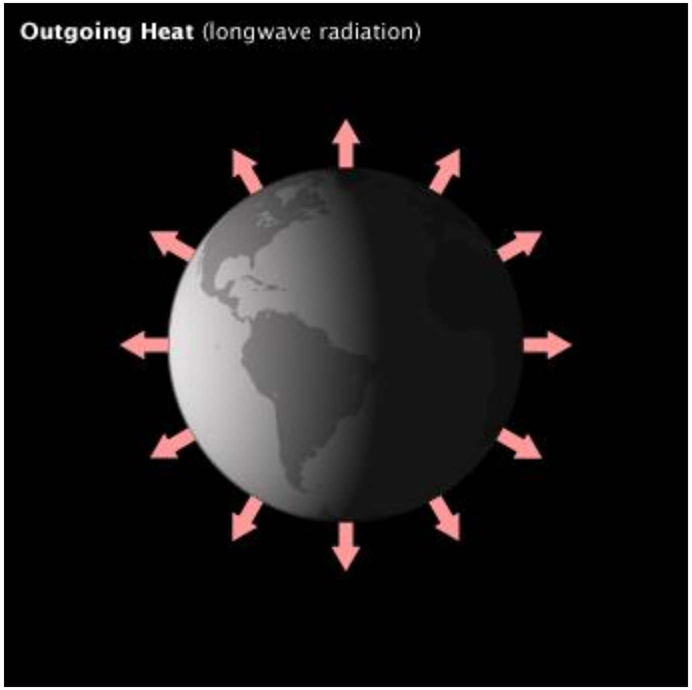
Climate theory – Nili Harnik
Energy Balance continued

From previous class:

Energy balance - what comes in must go out

Earth is heated by short wave (energetic) solar radiation and cooled by emitting less energetic, long wave infra-red (IR) radiation.



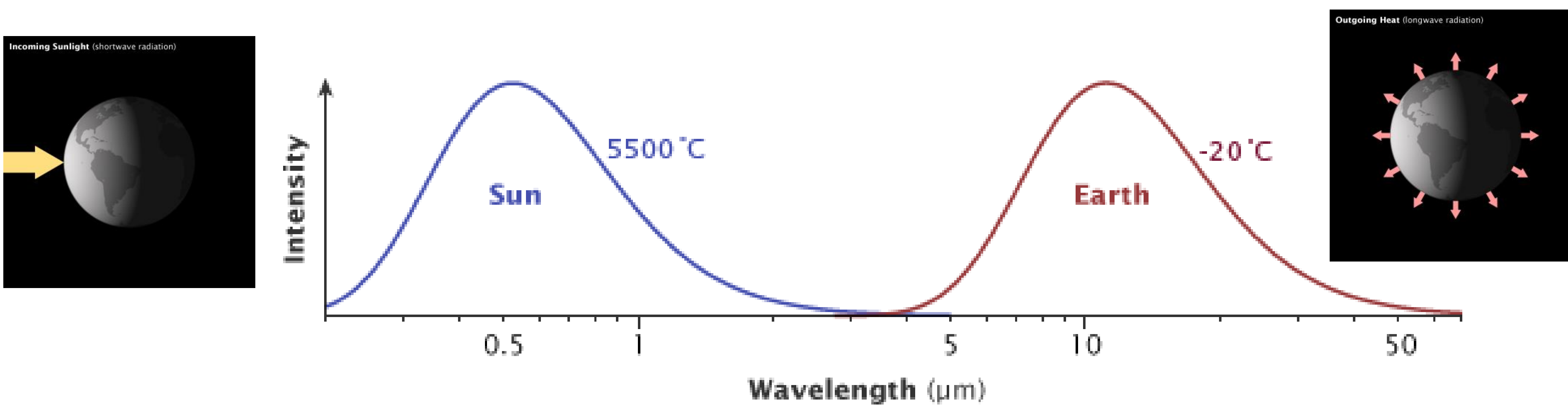


Blackbody radiation

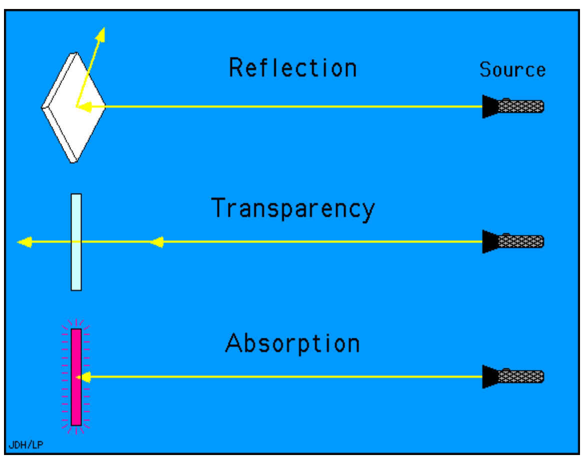
Stefan-Boltzmann Law: $F(T) = \sigma T^4$ warmer bodies emit more radiation

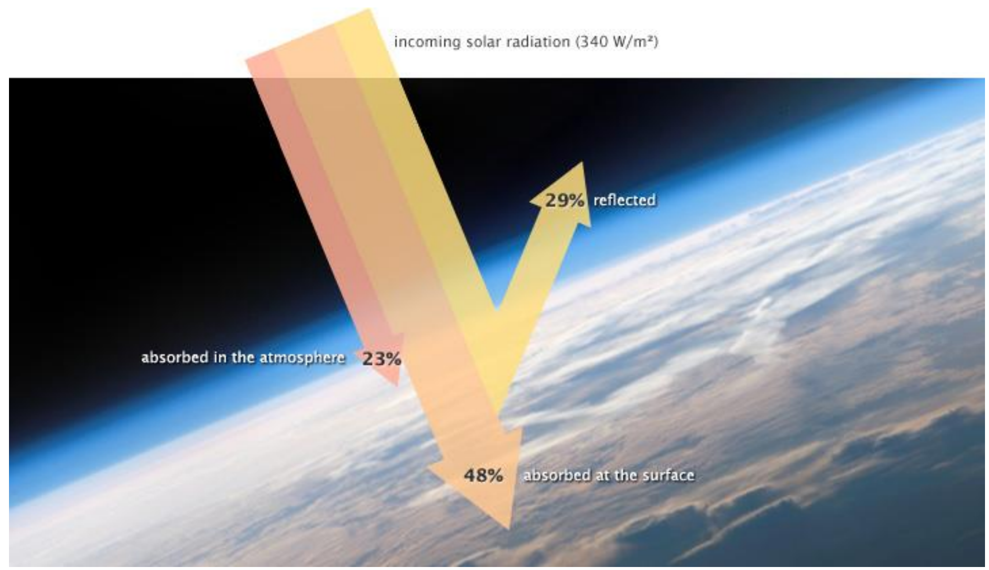
Wein's law: $\lambda max = 2898/T$ warmer bodies emit shorter wavelengths

Earth heats up until the emitted IR radiation equals the incoming solar radiation.



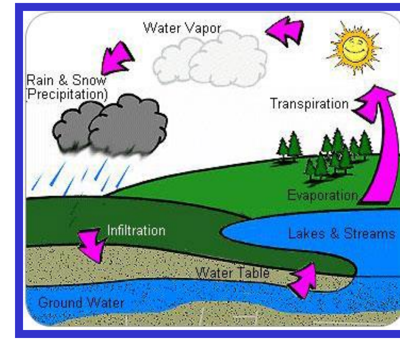
The incoming solar radiation is partly reflected back to space (30%), partly absorbed by the atmosphere, and partly absorbed by the earth (50%).



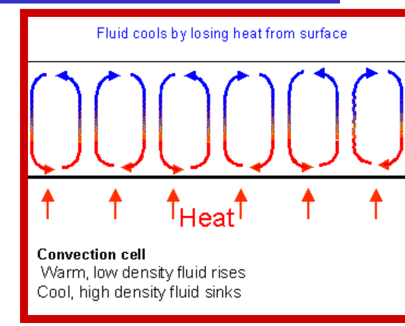


The earth's surface, which absorbs much of the solar radiation, emits heat in 3 ways:

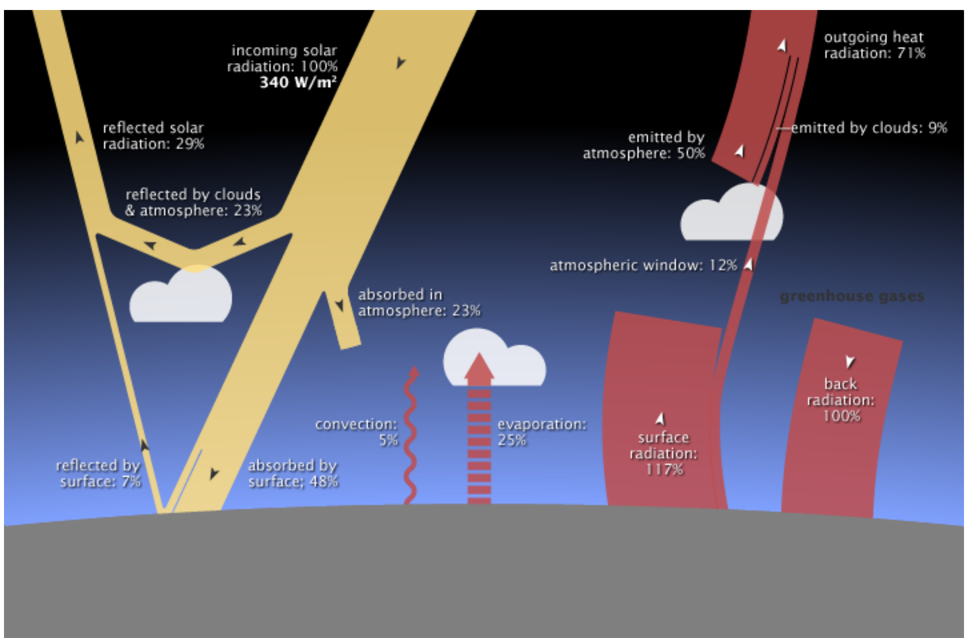
- Latent heat – water cools the surface when it evaporates into the air, and releases heat when it condenses into cloud and rain drops



- Sensible heat – hot buoyant air convection



- IR radiation



on which we will concentrate next…

Measured surface temperate (T_s) is actually 288 K.
The difference is due to the *natural greenhouse effect*.

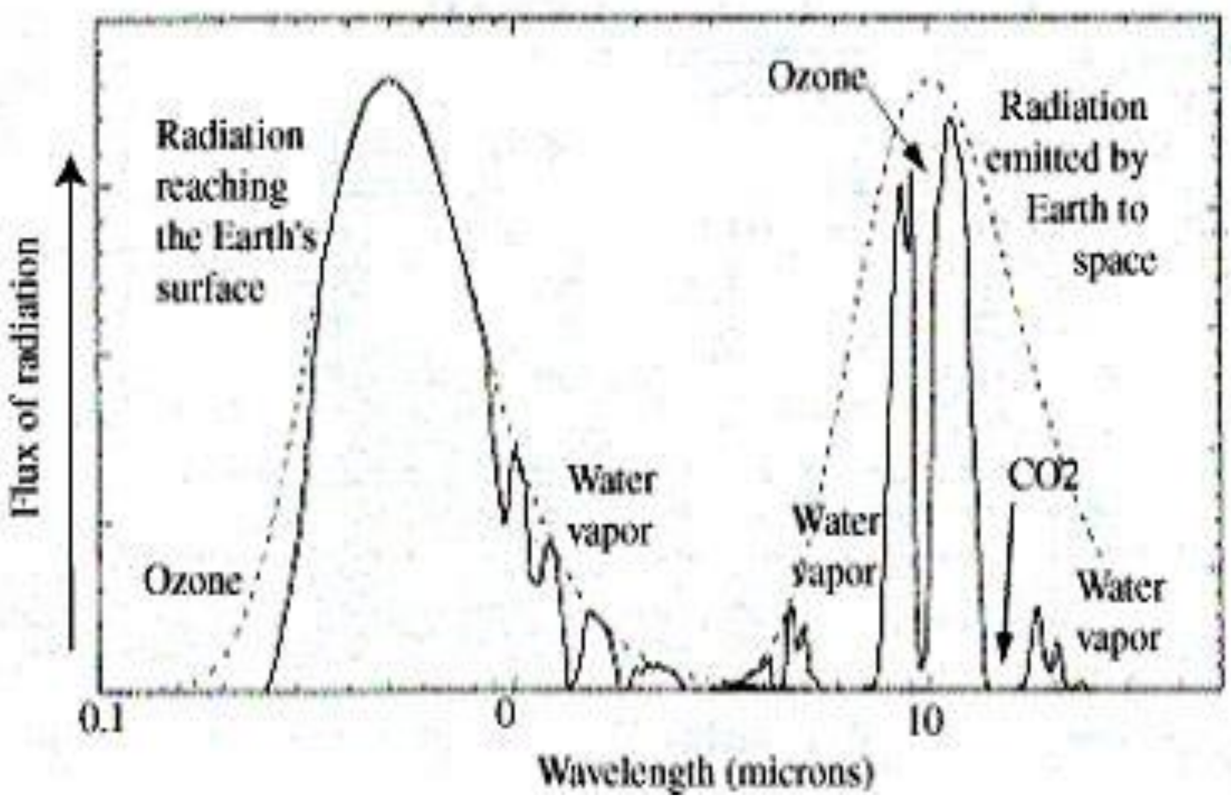
This corresponds to a radiative flux of $F_s = \sigma T_s^4$

The *transmissivity* τ of the atmosphere is the ratio between the flux transmitted out to space and that emitted by the surface:

$$\tau = T_e^4 / T_s^4 = 0.61$$

Absorptivity = $1 - \tau$ = 39%   - the amount of emitted surface radiation absorbed by the atmosphere.

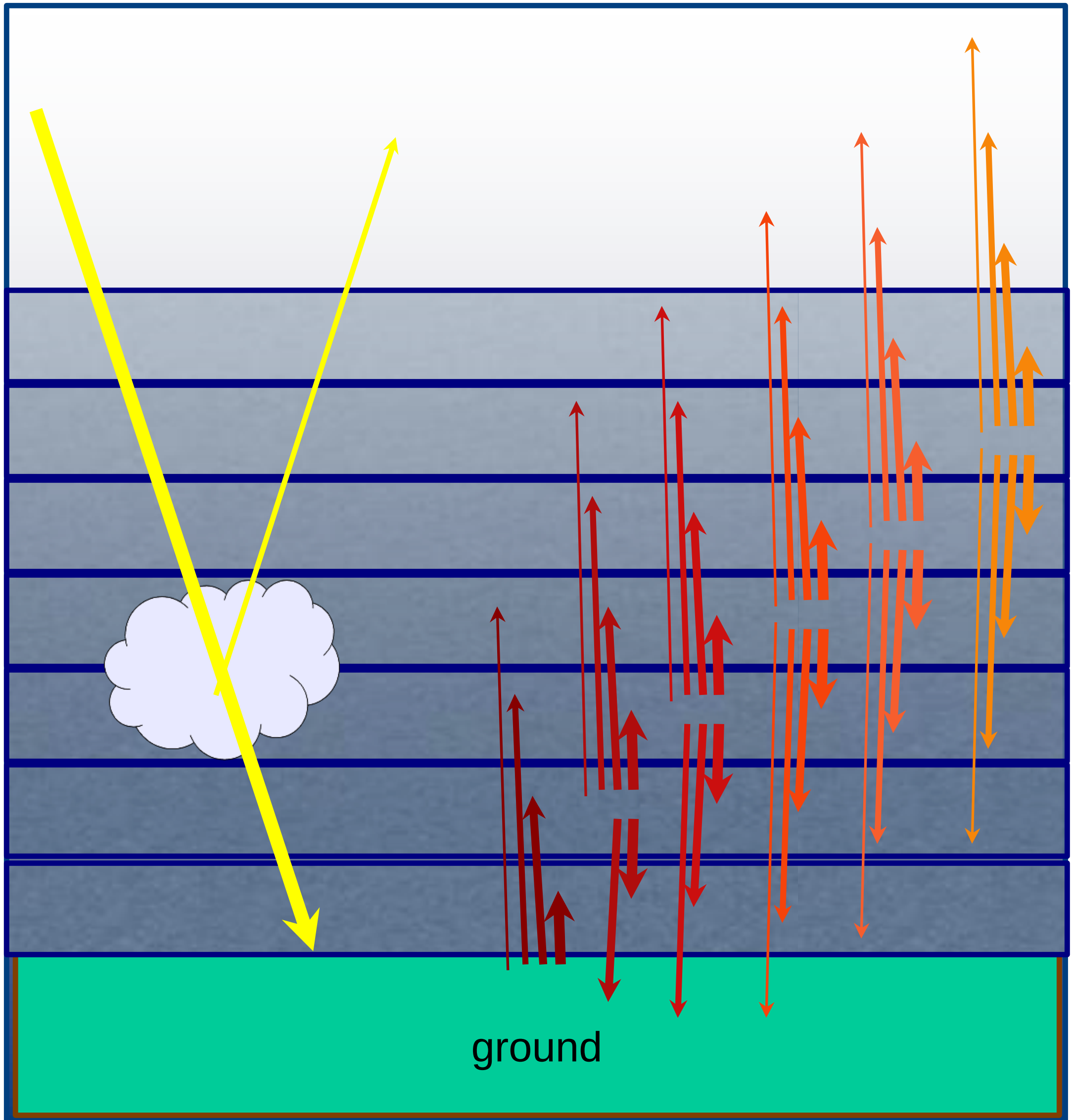
This is done mainly by water vapor, but this picture is too simplistic- lets get a glimpse of the more complex picture



Radiation
reaching
the Earth's
surface
Ozone
Water
vapor
Ozone
Radiation
emitted by
Earth to
space
Water
vapor
CO2
Water
vapor
Flux of radiation
0.1
0
10
Wavelength (microns)

Equations…

The atmosphere is a "continuum of thin layers", with a density which decreases with height

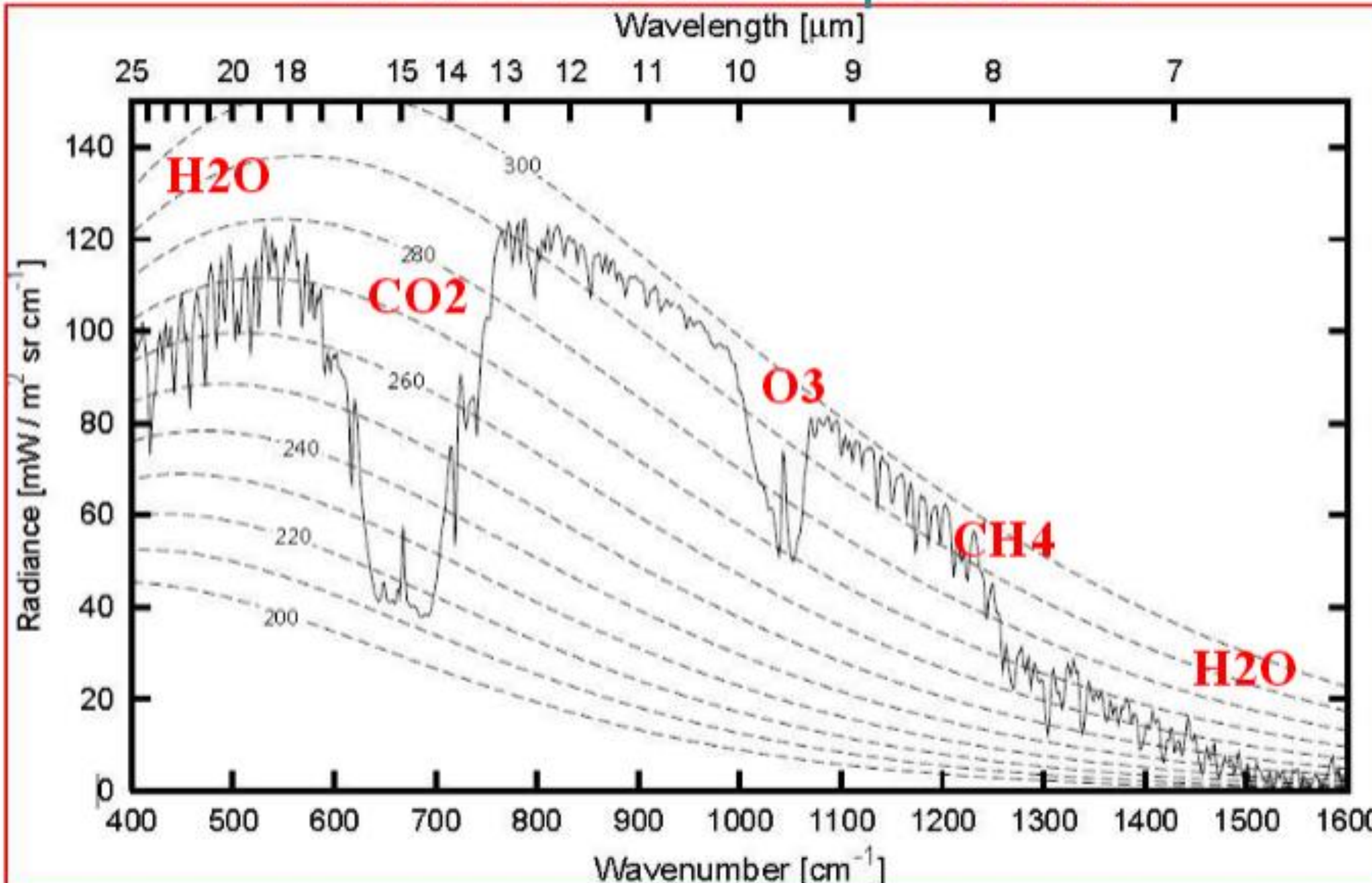


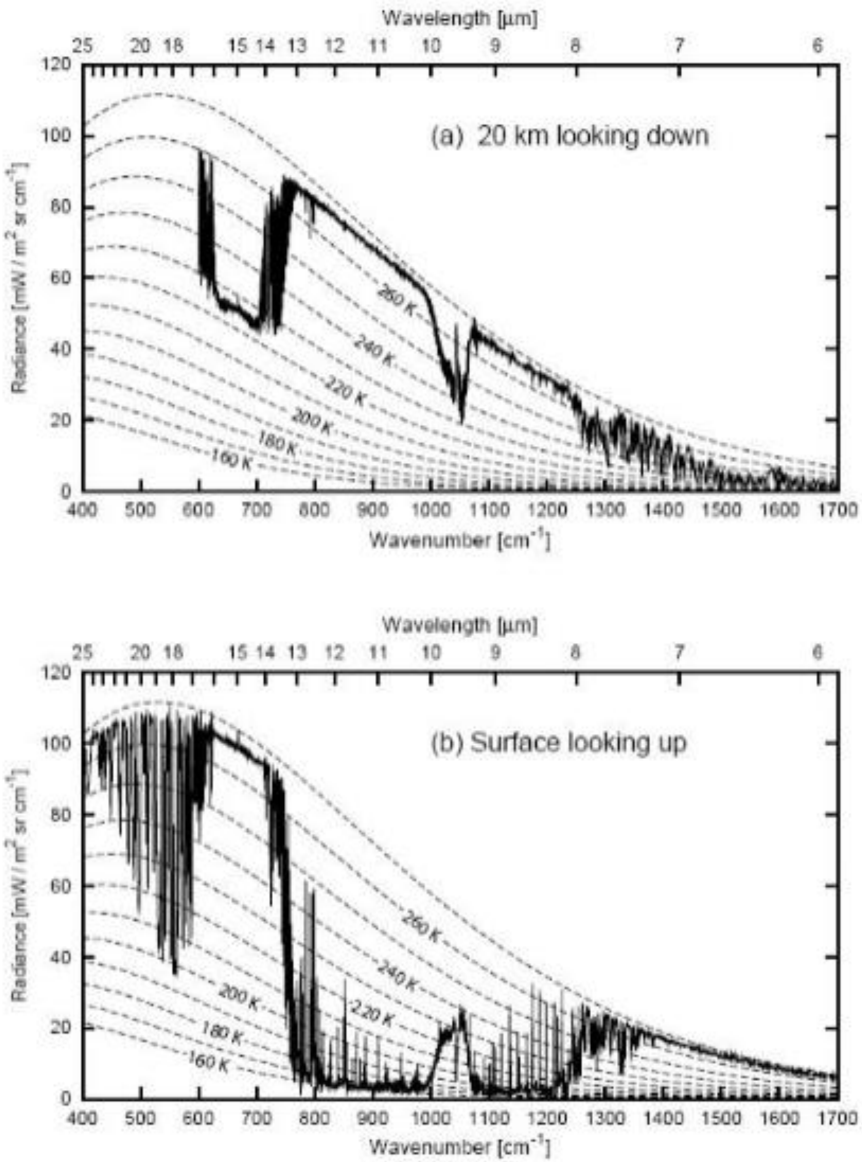
The terrestrial IR radiation gets gradually absorbed as it propagates upwards in the atmosphere.

The IR emitted from each layer, also gets gradually absorbed in the layers above and below.

The full problem consists of summing up the contributions and effects of all layers.

Earth radiation spectrum



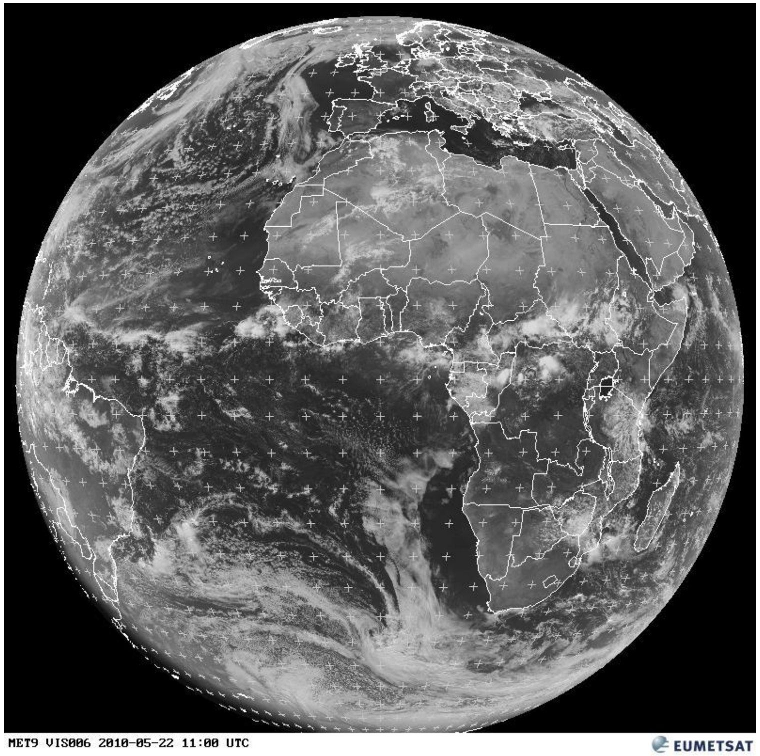


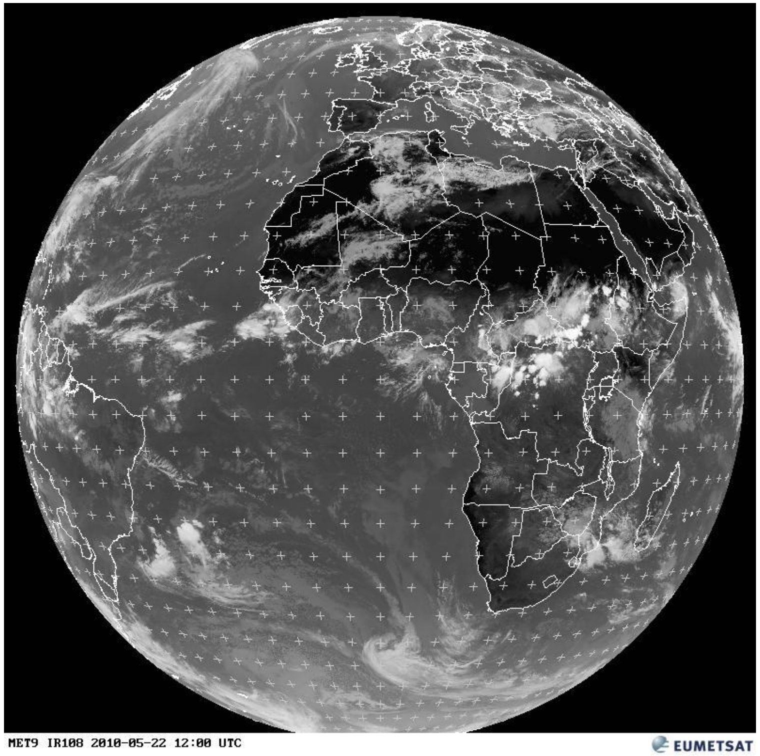
Wavelength [µm]
25 20 18 15 14 13 12 11 10 9 8 7 6
(a) 20 km looking down
Radiance [mW / m2 sr cm-1]
0 20 40 60 80 100 120
260 K
240 K
220 K
200 K
180 K
160 K
400 500 600 700 800 900 1000 1100 1200 1300 1400 1500 1600 1700
Wavenumber [cm-1]
Wavelength [µm]
25 20 18 15 14 13 12 11 10 9 8 7 6
(b) Surface looking up
Radiance [mW / m2 sr cm-1]
0 20 40 60 80 100 120
260 K
240 K
220 K
200 K
180 K
160 K
400 500 600 700 800 900 1000 1100 1200 1300 1400 1500 1600 1700
Wavenumber [cm-1]

For each wave length see different heights at different locations, depending on the cloud and moisture content:

Short wave- e.g. a visible image: We see reflection. Clouds reflect enough to not see surface below. Otherwise see the surface, depending on albedo.

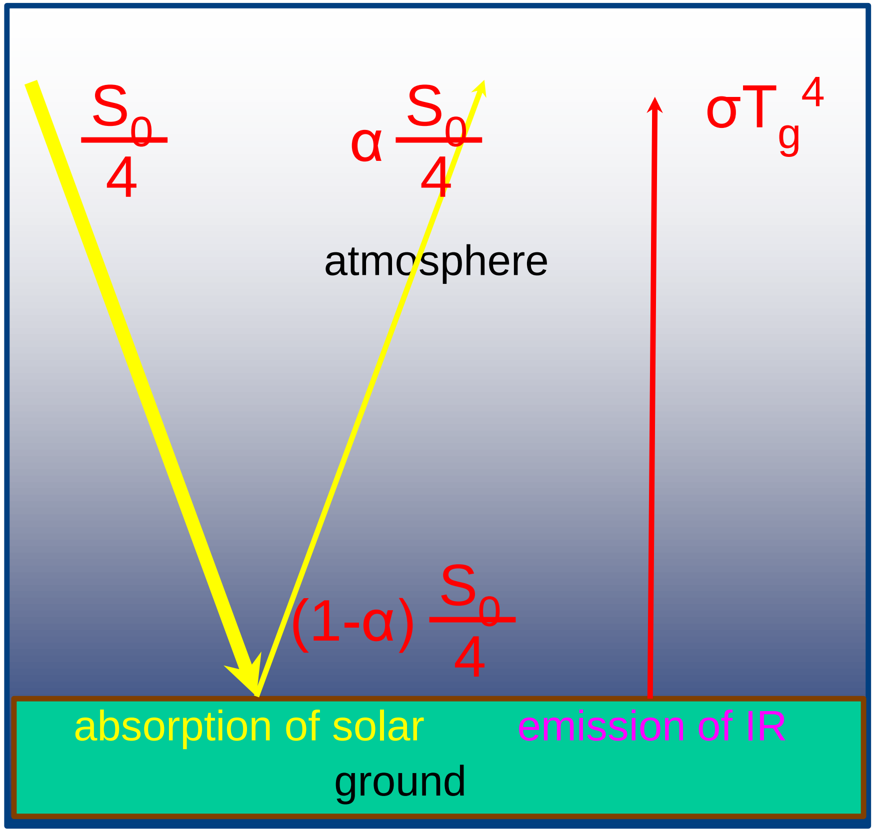
Long wave – e.g. IR image: see outgoing long wave radiation – when have clouds see cloud tops which are cold. In dry regions see ground which is warm (window region).





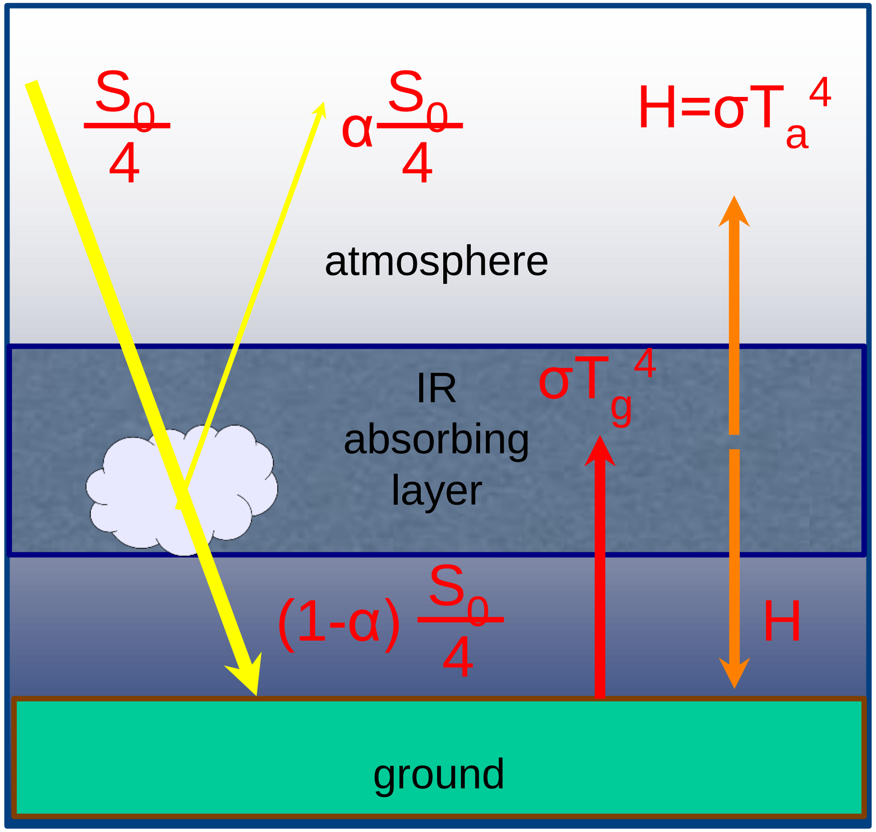
Effect of Atmospheric Absorption: 1 sheet

w/o absorption



$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4}$$

w. absorption



$$\sigma T_a^4 = (1-\alpha)\frac{S_0}{4}$$

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4} + \sigma T_a^4$$

Effect of Atmospheric Absorption: 1 sheet

w/o absorption

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4} = 239\ \mathrm{Wm^{-2}}$$

$$T_g = \sqrt[4]{(1-\alpha)\frac{S_0}{4\sigma}}$$

w. absorption

$$\sigma T_a^4 = (1-\alpha)\frac{S_0}{4}$$

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4} + \sigma T_a^4$$

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{2}$$

$$T_g = \sqrt[4]{(1-\alpha)\frac{S_0}{2\sigma}}$$

The Greenhouse Effect

W/O absorption: $\mathbf{T_g = 255\ K}$ **or -18°C**

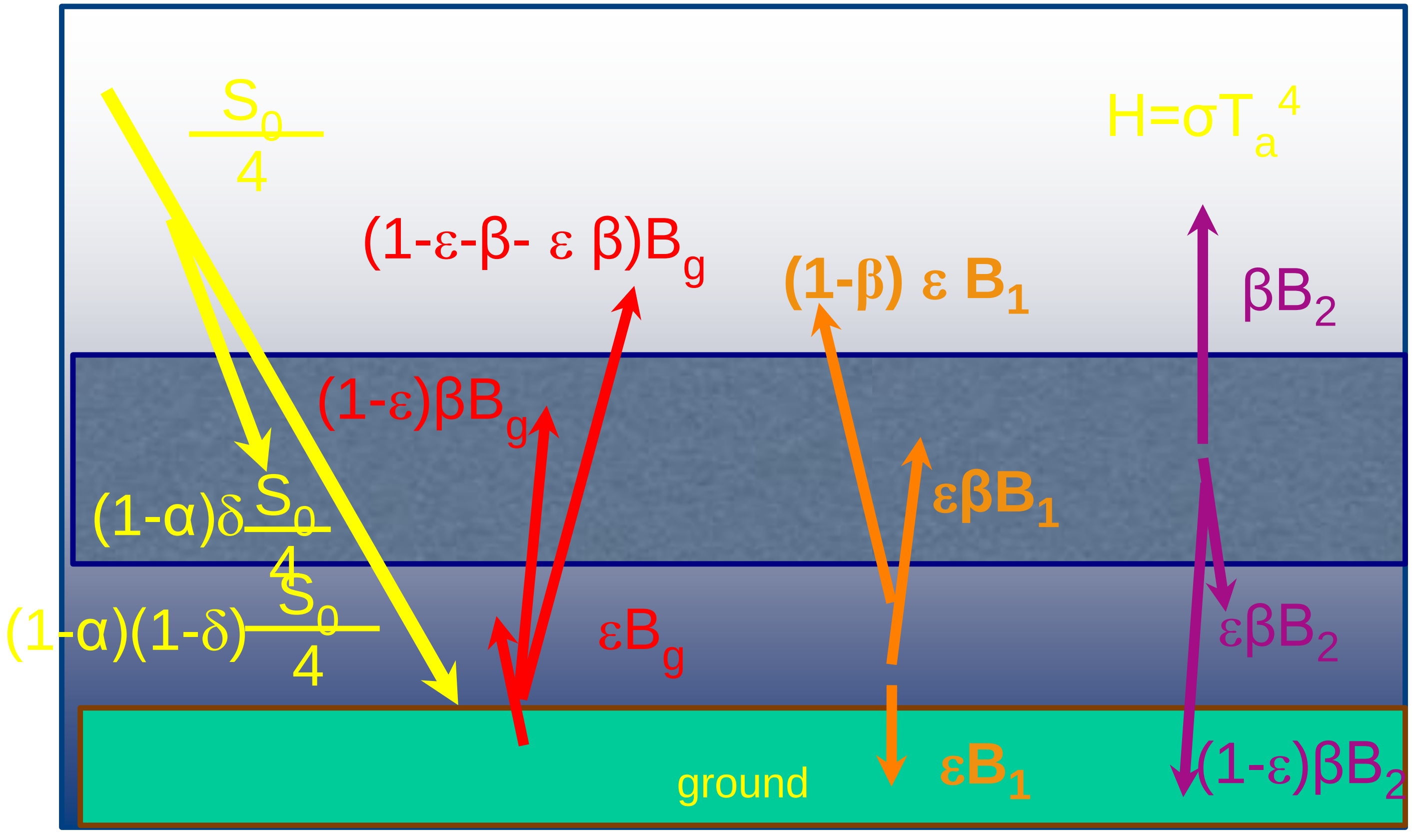
W. absorption $\mathbf{T_g = 303.5\ K}$ **or +30.5°C**

Too warm

2 layers with emissivity

N layers

Surface and surface air temperature



ε Lower level LW emissivity

β upper level LW emissivity.

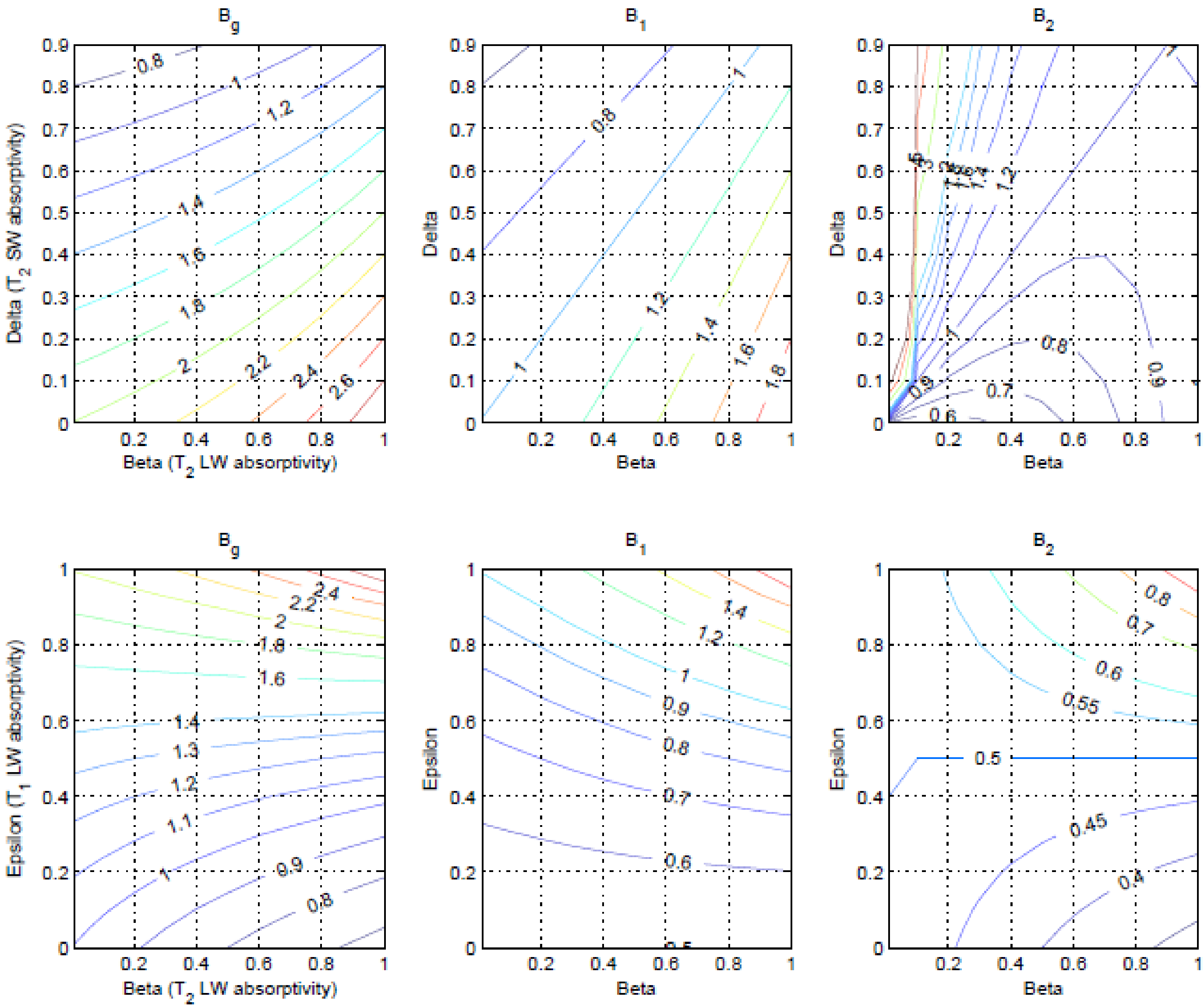
δ Upper level SW absorptivity

Model 1: β<1, δ<1, ε=1

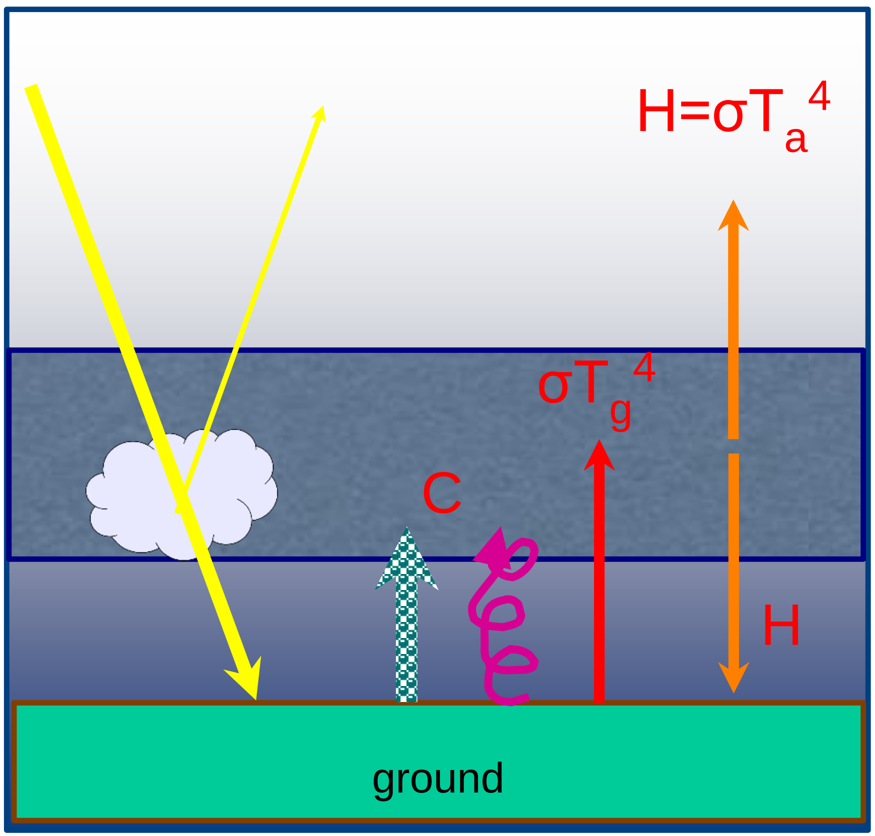
Model 2: β<1, δ=0, ε<1, β<ε

Model 1

Model 2



Since radiation makes the surface hotter than the air just above it, the surface looses heat not only through radiation, but also through latent and sensible heat transports (both upwards and polewards).



A few assumptions:

1) A linear temperature dependence: $C=aT_g+b$

$$\sigma T_g^{\,4} + aT_g + b = H + \sigma T_a^{\,4}$$

2) A flux bluk flux formulation- $C=a(T_a-T_g)$

3) Radiative-convective adjustment: set $(Tg-Ta)/z=\Gamma$ when unstable

This yields colder T_g than with pure radiative surface cooling

Radiative convective equilibrium…
Manabe and Strickler, 1964

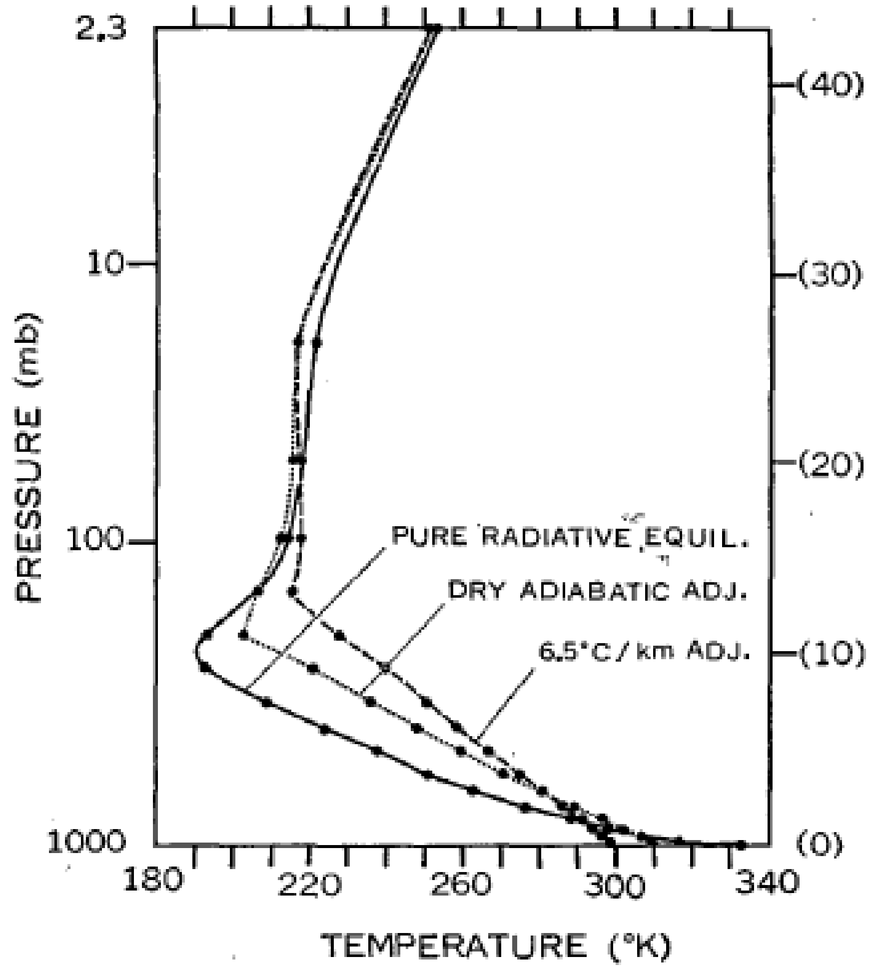


FIG. 4. The dashed, dotted, and solid lines show the thermal equilibrium with a critical lapse rate of 6.5 deg km^{-1}, a dry-adiabatic critical lapse rate (10 deg km^{-1}), and pure radiative equilibrium.

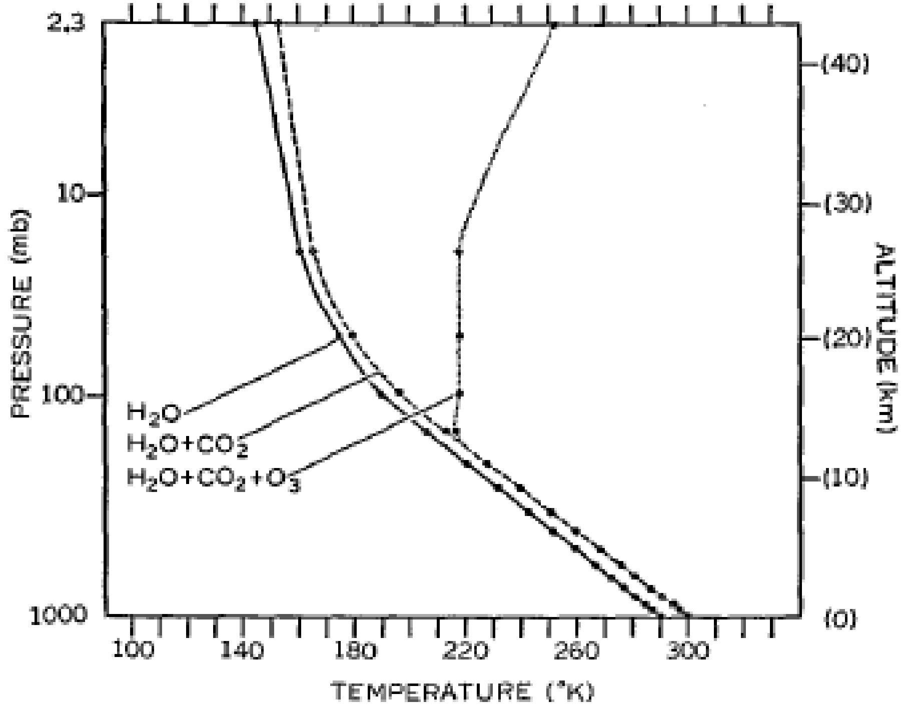
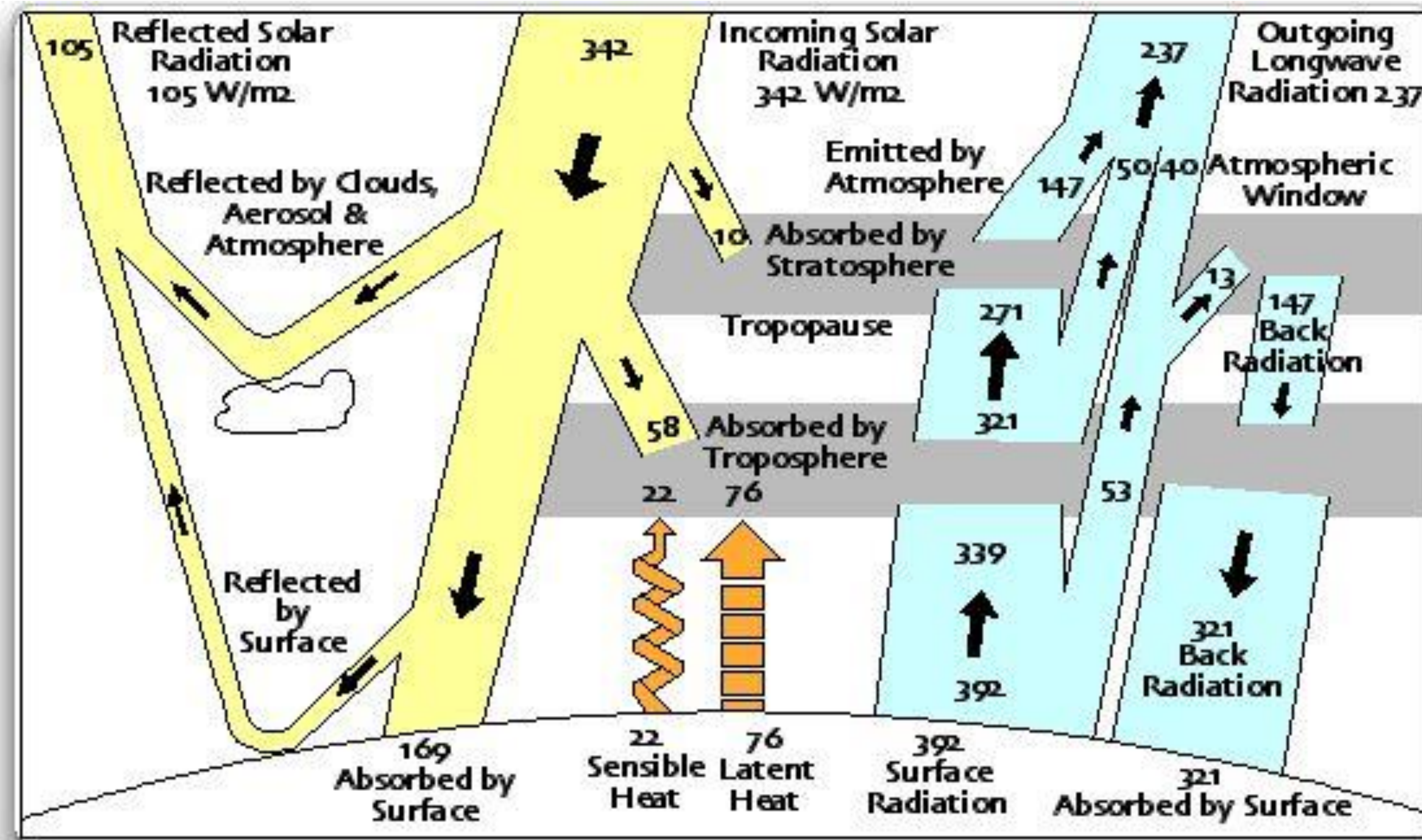
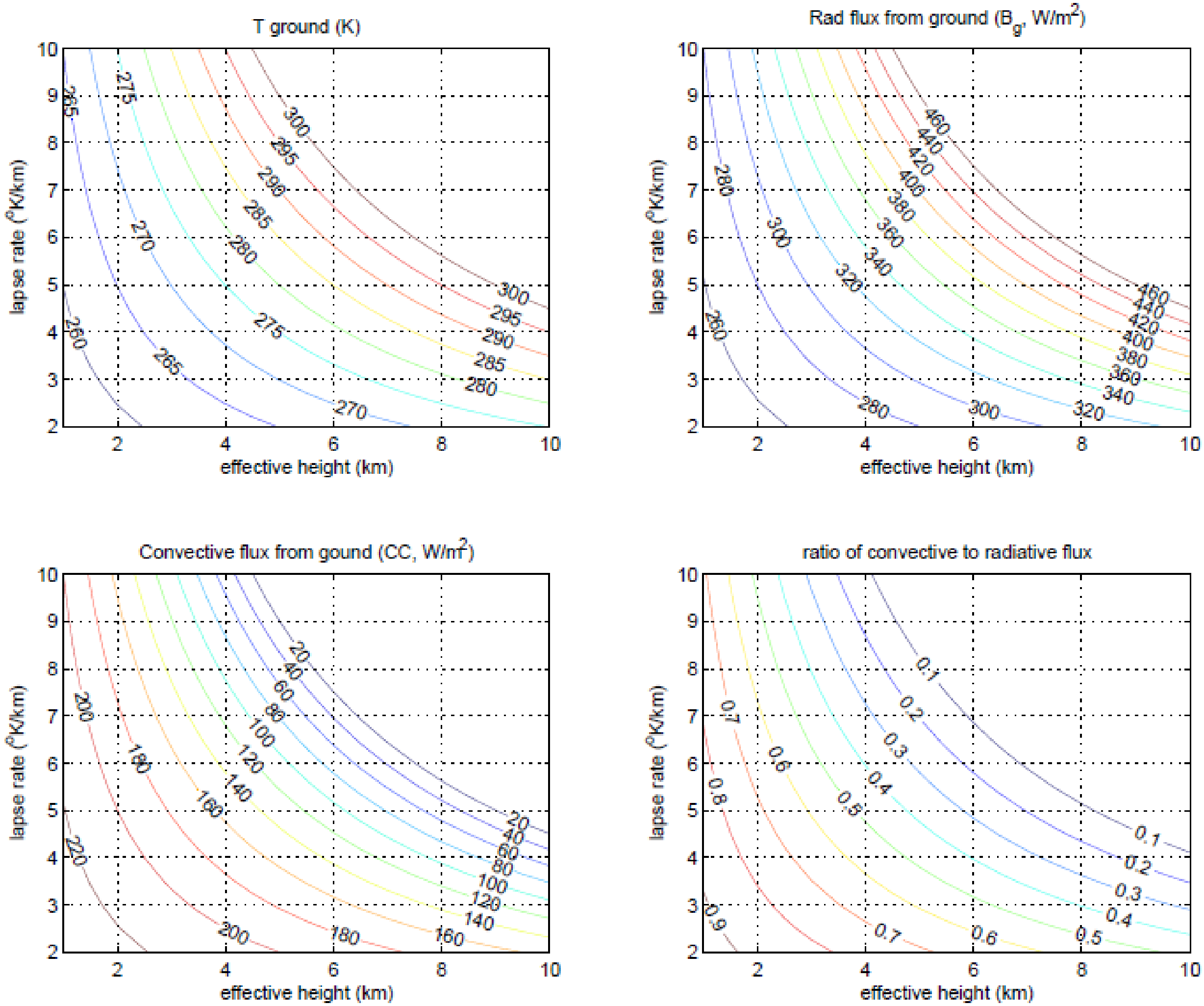


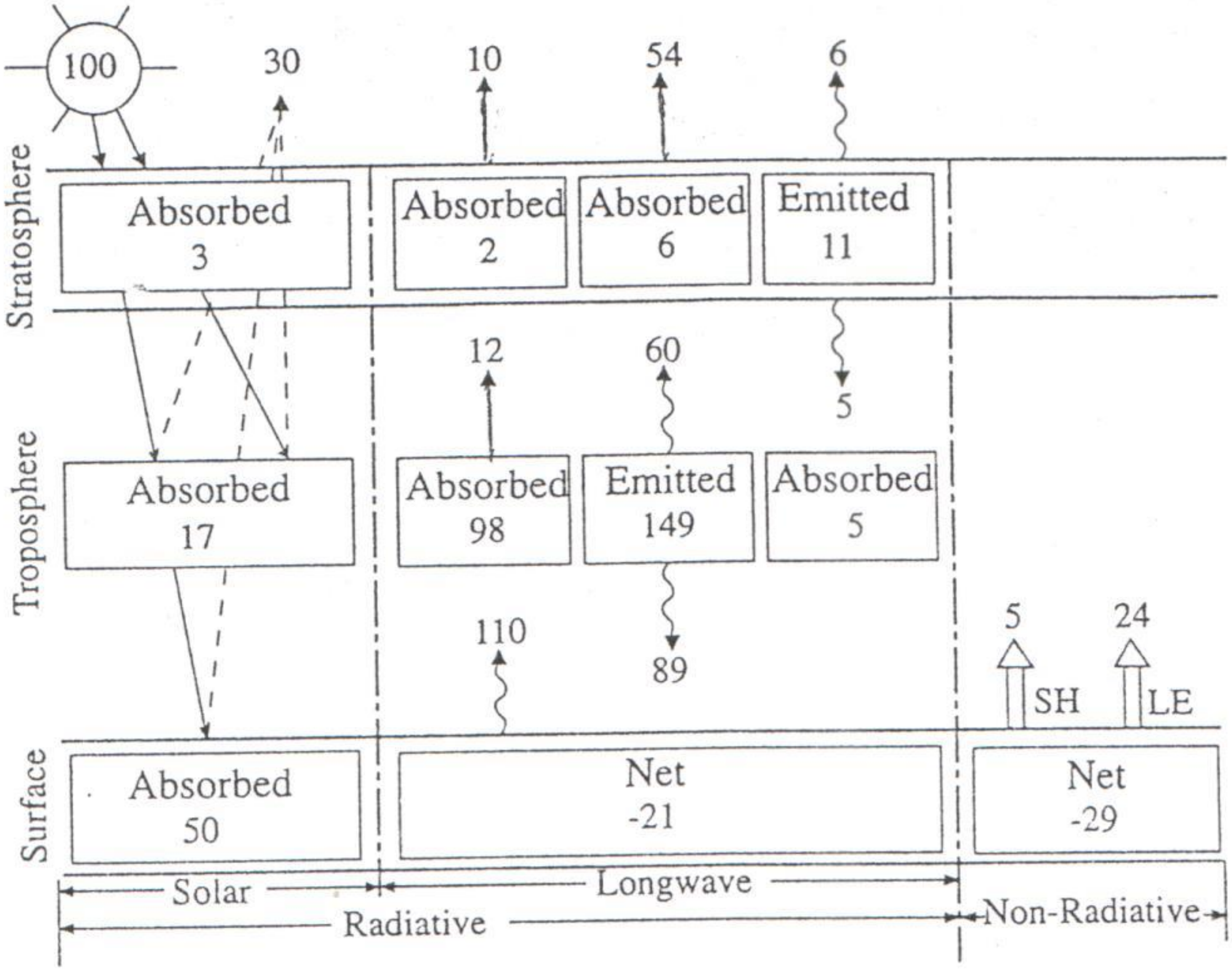
FIG. 6c. Thermal equilibrium of various atmospheres which have a critical lapse rate of 6.5 deg km^{-1}. Vertical distributions of gaseous absorbers at 35N, April, were used. $S_e=2$ ly min^{-1}, $\cos\bar{\zeta}=0.5$, $r=0.5$, no clouds.



Overall, we get a surface temperature of $T_g=(390/\sigma)^{1/4}=288°K=15°C$



Surface temperature decreases, the larger C is. C is largest for lowest emission heights- makes sense since in this model T1 and Tg are fixed, so lapse rate larger, the smaller the emission height.

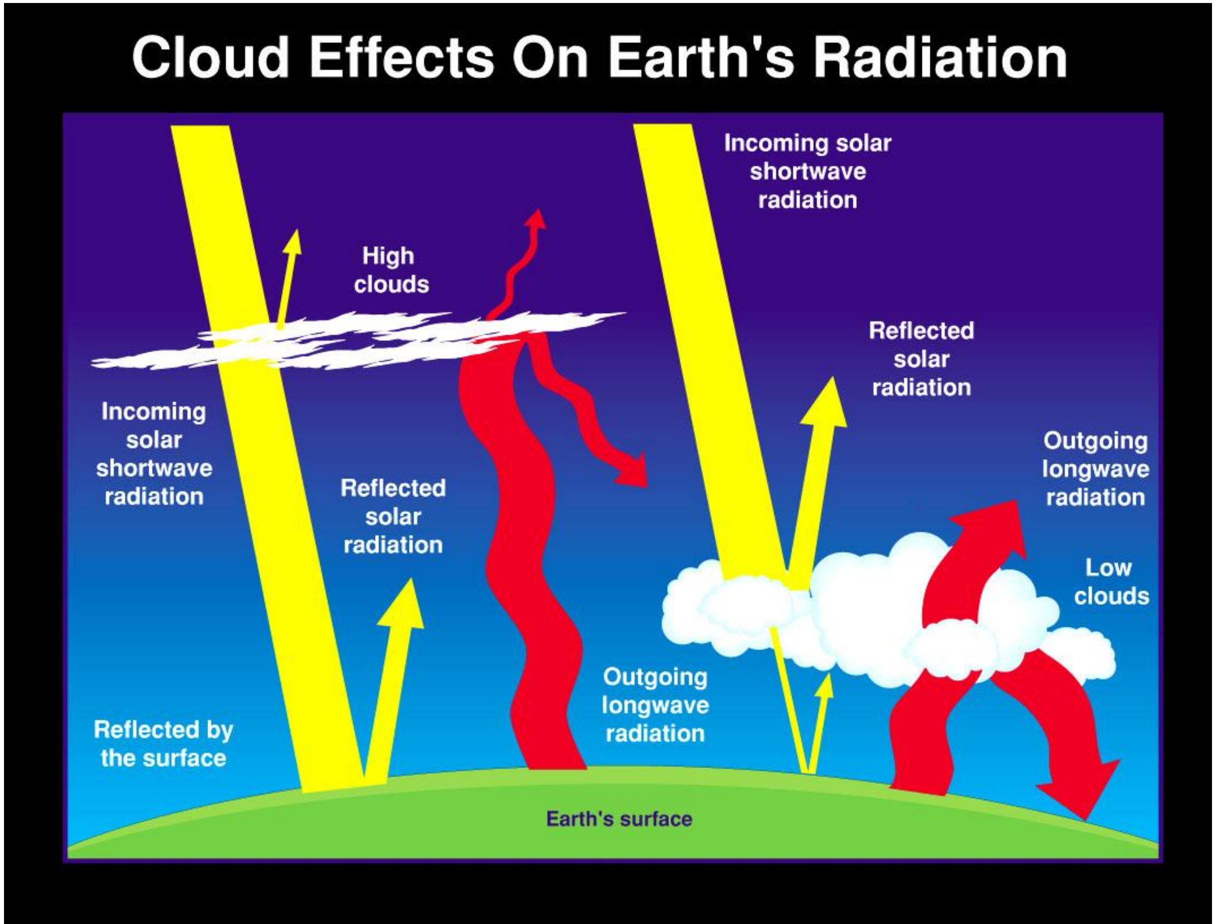


100
30
10
54
6
Stratosphere
Absorbed
3
Absorbed
2
Absorbed
6
Emitted
11
12
60
5
Troposphere
Absorbed
17
Absorbed
98
Emitted
149
Absorbed
5
110
89
5
24
SH
LE
Surface
Absorbed
50
Net
-21
Net
-29
Solar
Longwave
Radiative
Non-Radiative

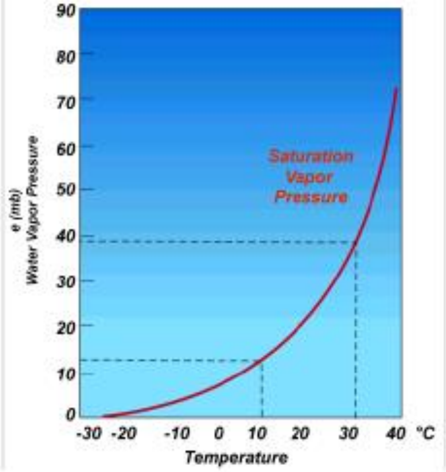
Clouds complicate the picture:

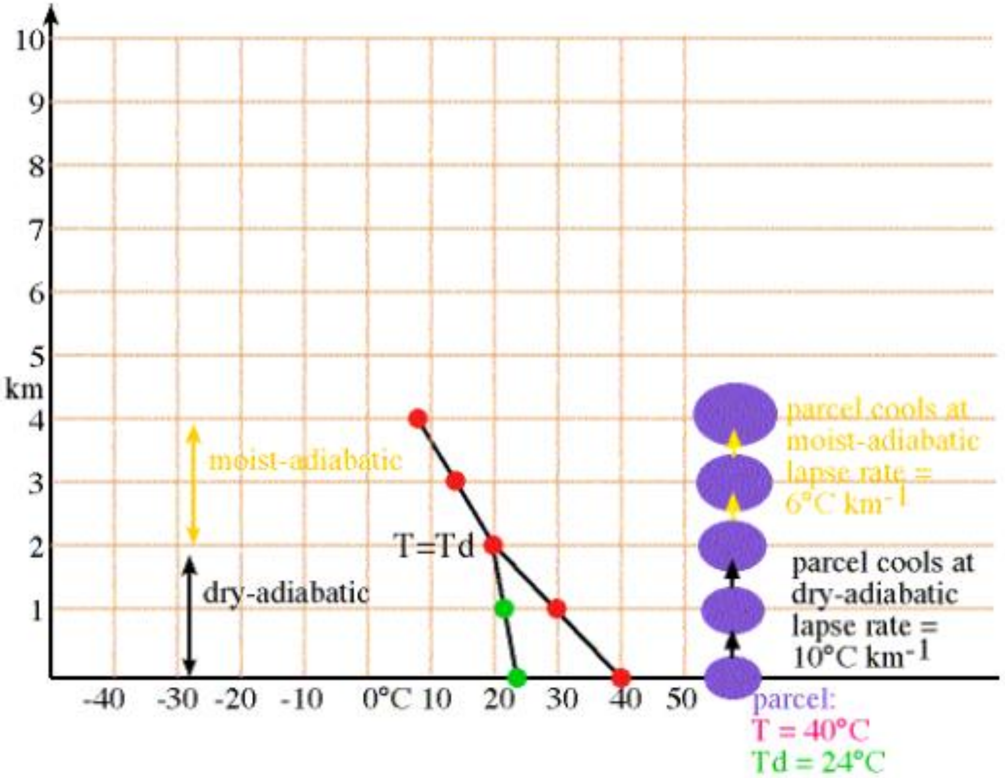
Clouds affect short wave by reflecting it, and long wave by absorbing it. These are opposing effects on surface energy budget.

Different cloud types – high, middle and low. Thick and thin. Each has a different net effect.



$e_s = 6.11 \exp(0.0067T)$ hPa (T in degrees C)

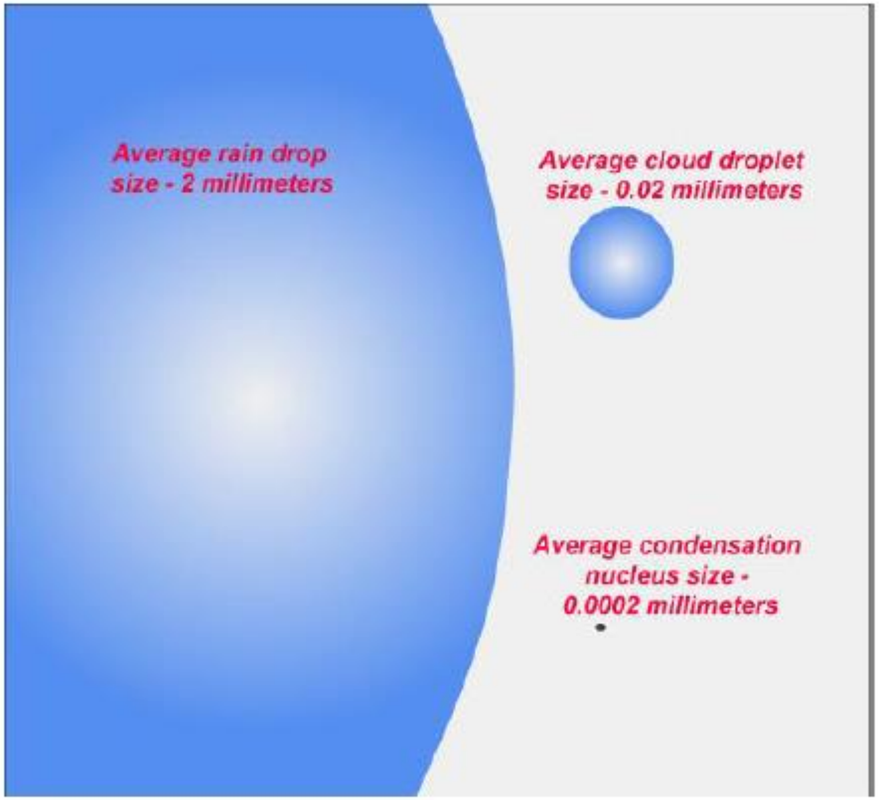




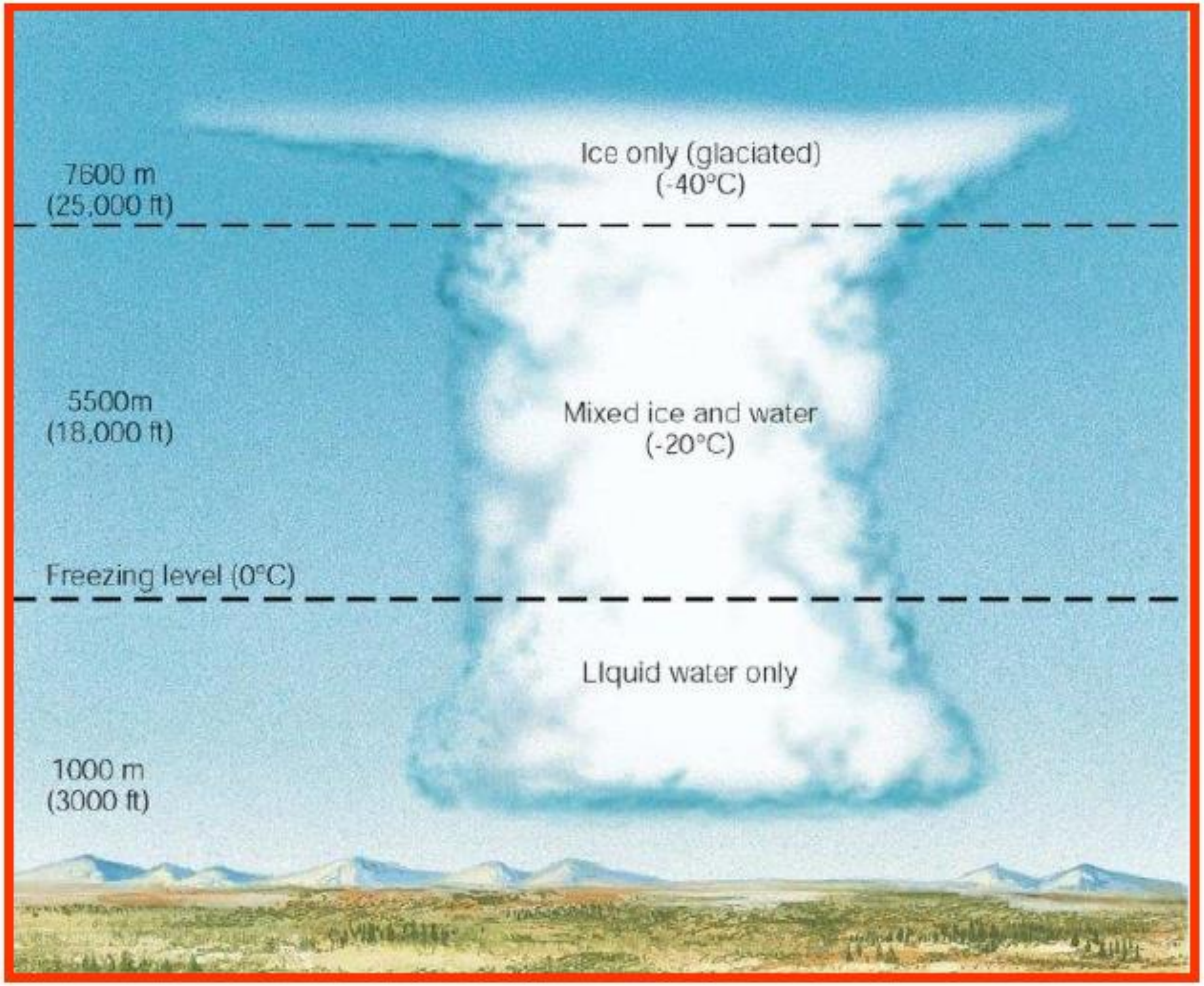
Typical sizes of cloud droplets

- Formation of droplets depends on availability of Cloud Condensation Nuclei (CCN)

- Pure water will condensate at relative humidity of about 1000%
- Ice crystals will form at temperature of -40C.



Water and ice in a cloud



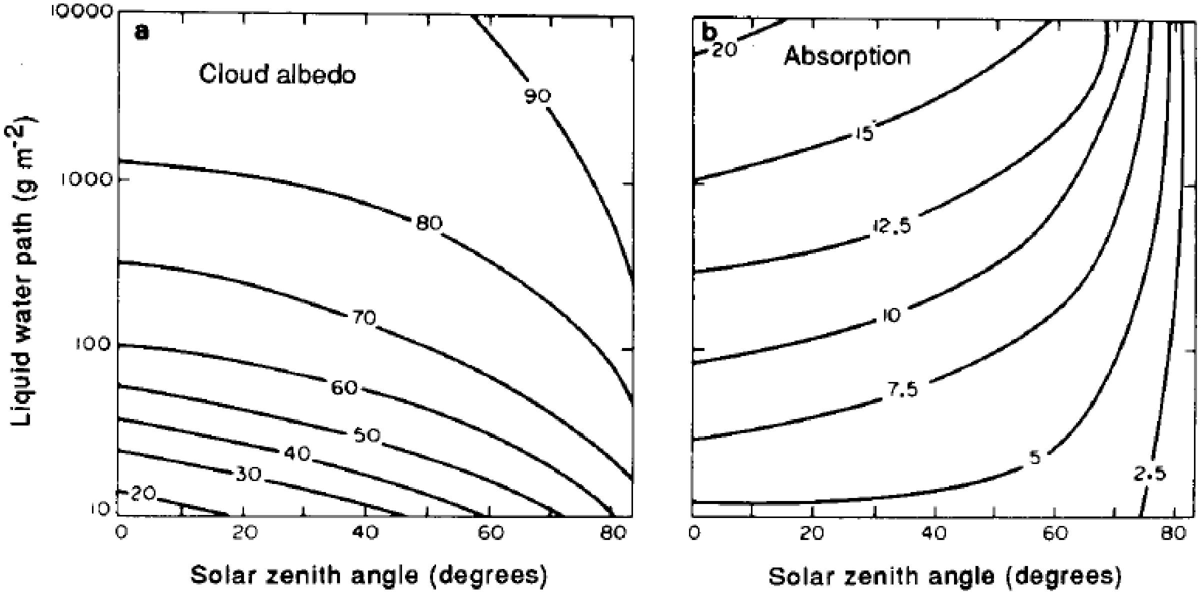


Fig. 3.13 The dependence of (a) cloud albedo and (b) cloud absorption on cloud liquid water path and solar zenith angle. Values are given in percent. [From Stephens (1978). Reprinted with permission from the American Meteorological Society.]

Effects of clouds on solar radiation – for idealized plane-parallel clouds: depends on solar zenith angle, liquid water content, and droplet size distribution.
LWC- effect saturates at high LWC, esp. at high SZA. More small drops reflects more.

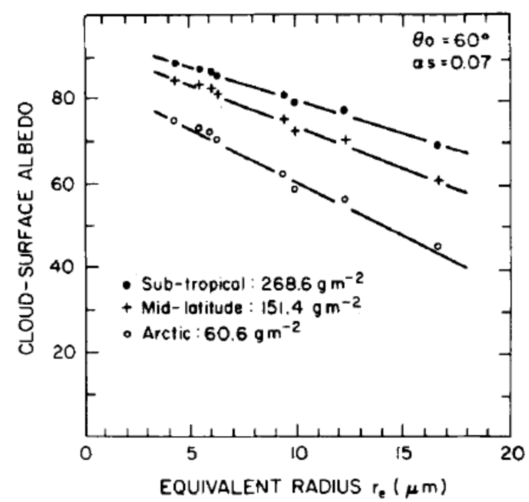


Fig. 3.14 The dependence of planetary albedo on the size of cloud droplets. [From Slingo and Schrecker (1982). Reprinted with permission from the Royal Meteorological Society.]

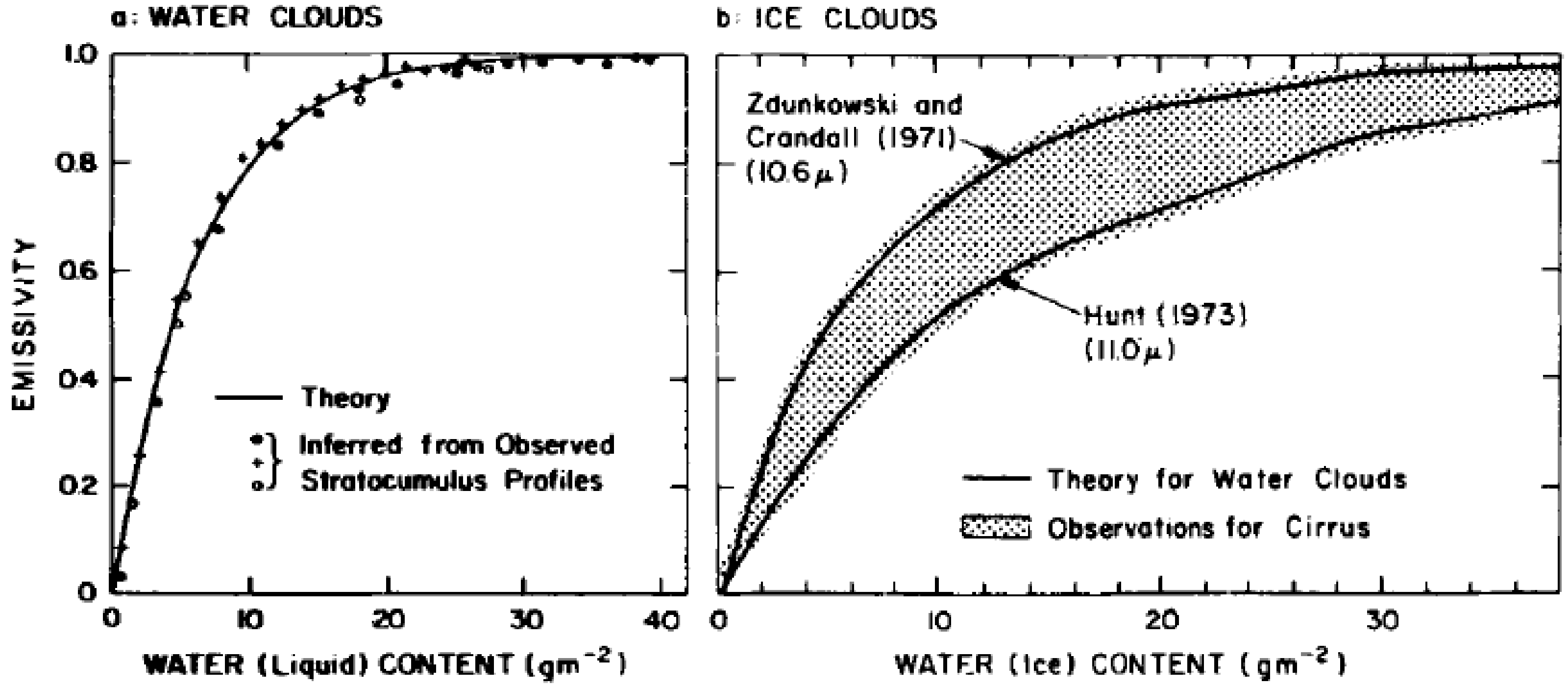


Fig. 3.15 The dependence of the longwave emissivity on (a) liquid water content [from Slingo *et al*. (1982); reprinted with permission from the Royal Meteorological Society] and (b) ice content [from Griffith *et al*. (1980); reprinted with permission from the American Meteorological Society].

Effects of clouds on long wave radiation – depends on liquid water content – for quite small values fully absorbing and can be treated as black body

Saturation with LWC for LW occurs much before it occurs for SW

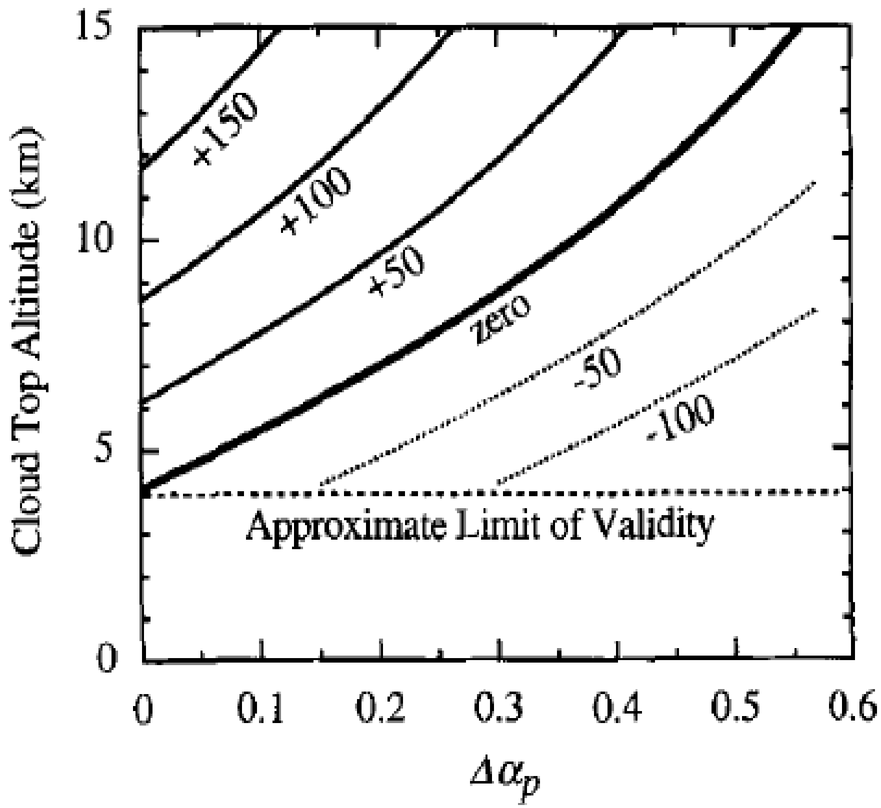


Fig. 3.20 Contours of change in net radiation at the top of the atmosphere caused by the insertion of a cloud into a clear atmosphere, plotted against cloud top altitude and the planetary albedo contrast between cloudy and clear conditions. The net radiation changes are calculated with the approximate model described in Section 3.8 that is invalid for clouds with tops lower than about 4 km. Contours from -100 to $+150$ W m^{-2} are shown at an interval of 50 W m^{-2}. The zero contour where the cloud has no net effect on the radiation budget at the top of the atmosphere is bold and negative contours are dashed.

Table 3.2

Values of Cloud Shortwave Reflectivity and Absorptivity and Fractional Area Coverage Assumed in Manabe and Strickler (1964)

Type	SW reflectivity	SW absorptivity	% of area
High (cirrus)	0.21	0.005	0.228
Medium (cumulus)	0.48	0.020	0.090
Low (stratus)	0.69	0.035	0.313

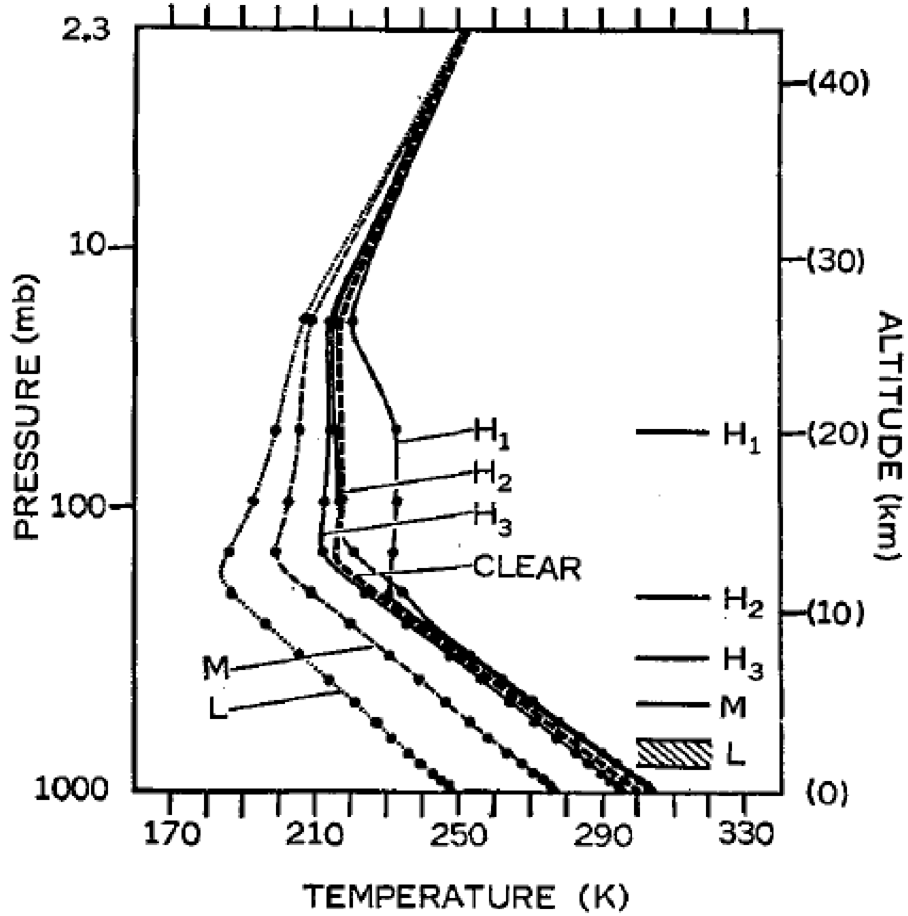


Fig. 3.19 Thermal equilibrium temperature profiles for atmospheres with various cloud distributions. The cloud heights corresponding to each type of cloud are shown on the right (L = low, M = medium, and H = high cloud). The heavy dashed line shows the equilibrium profile for clear skies. [From Manabe and Strickler (1964). Reprinted with permission from the American Meteorological Society.]

Table 3.3

Cloud Radiative Forcing as Estimated from Satellite Measurements

	Average	Cloud-free	Cloud forcing
OLR	234	266	+31
Absorbed solar radiation	239	288	−48
Net radiation	+5	+22	−17
Albedo	30%	15%	+15%

Radiative flux densities are given in W m^{-2} and albedo in percent. [From Harrison *et al.* (1990), © American Geophysical Union.]

Observed cloud fractions

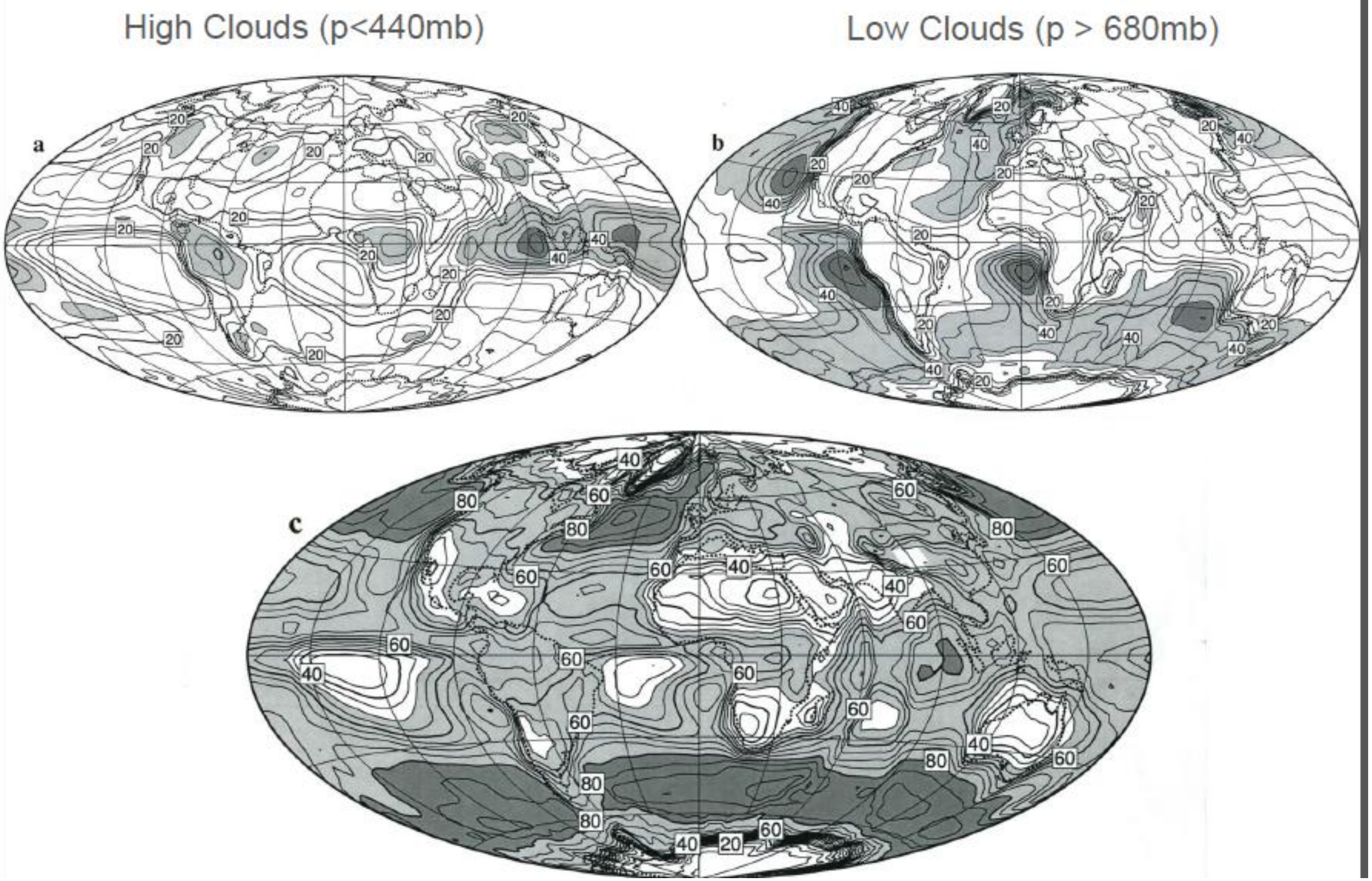


Fig. 3.21 Annual average cloud fractional area coverage in percent estimated from satellite data under the International Satellite Cloud Climatology Project [ISCCP, Rossow and Schiffer (1991)]. (a) Clouds with tops higher than 440 mb, (b) clouds with tops lower than 680 mb, and (c) all clouds. In (a) and (b) the contour interval is 5%, with values greater than 30% lightly shaded, and greater than 50% heavily shaded. In (c) the contour interval is also 5%, but light shading is applied for values greater than 50% and heavy shading for values greater than 80%.

Observed cloud effect on radiation