Radiative transfer and climate IV

Reading: GPC Ch3

Outline: Role of clouds and aerosols in climate:

- Distribution
- Clouds and SW radiation - dependence on liquid water path, solar zenith angle and droplet size
- Clouds and LW radiation - dependence on liquid water path, cloud top height
- Cloud forcing
- Aerosols - direct and indirect effects

Clouds

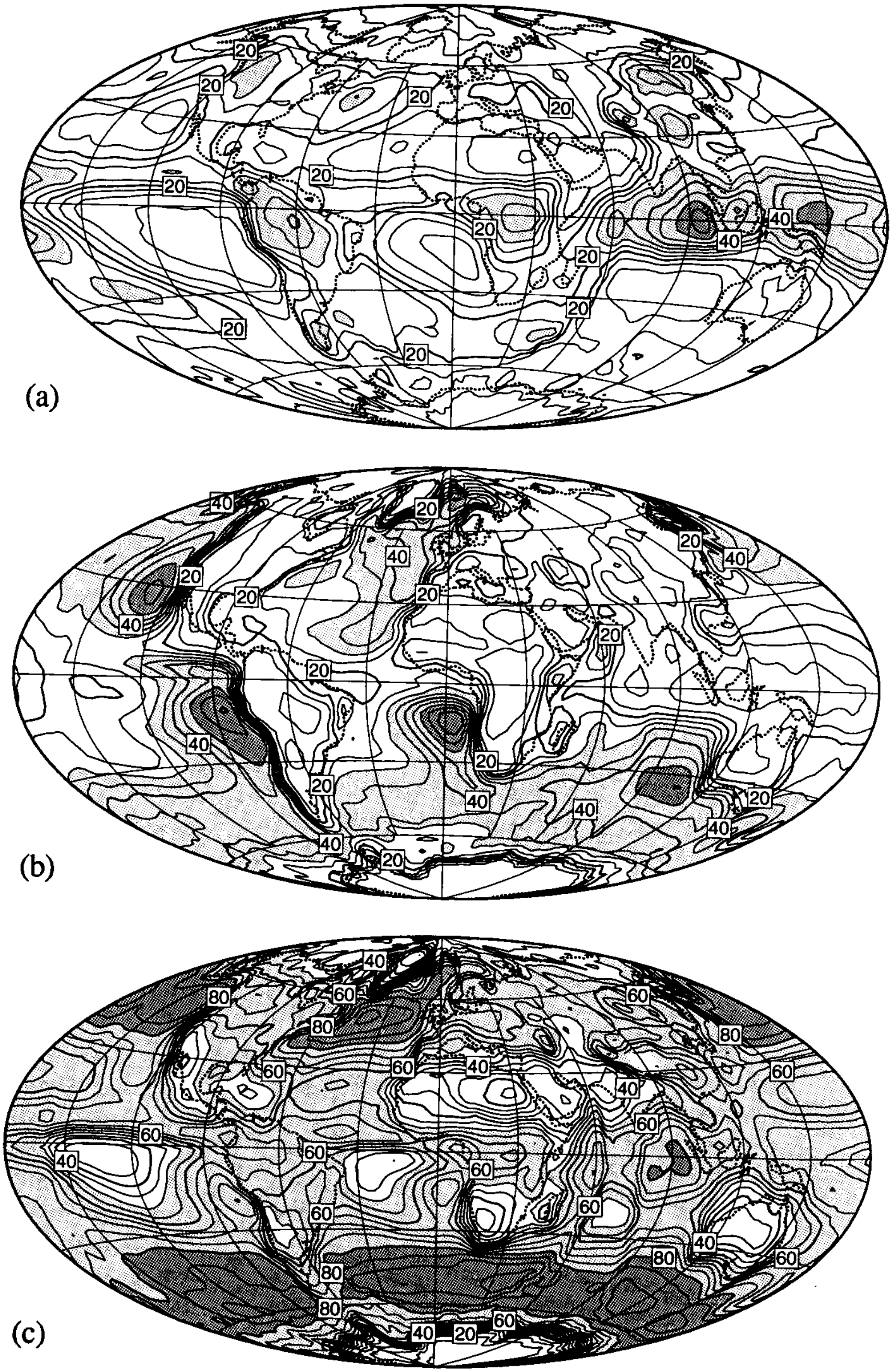
Clouds consist of liquid water droplets (or ice particles) suspended in the atmosphere. They are formed by condensation of water vapor when the temperature falls below the saturation value. Water droplets (and ice particles) have substantial interactions with both SW and LW radiation, whose nature depends on

- the total amount of water
- the size and shape of the droplets (or ice particles)
- the distribution in space of the droplets (or particles).

The problem of describing these interactions is often simplified by assuming that clouds are uniform and infinite in horizontal (plane-parallel assumption).

In the next figure, the annual mean cloud fractional area estimated from satellite data is shown.

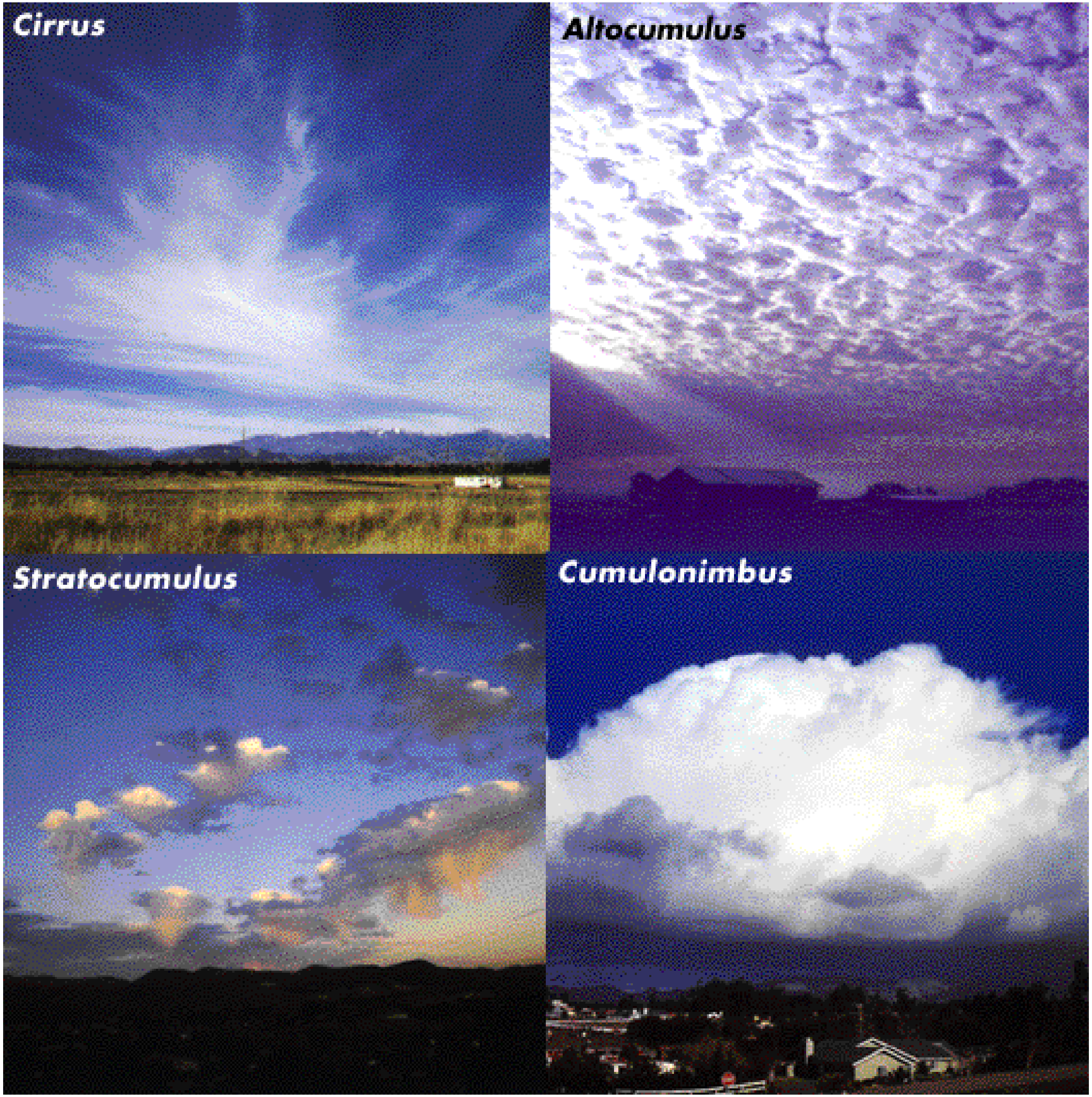
Annual mean cloud fractional area



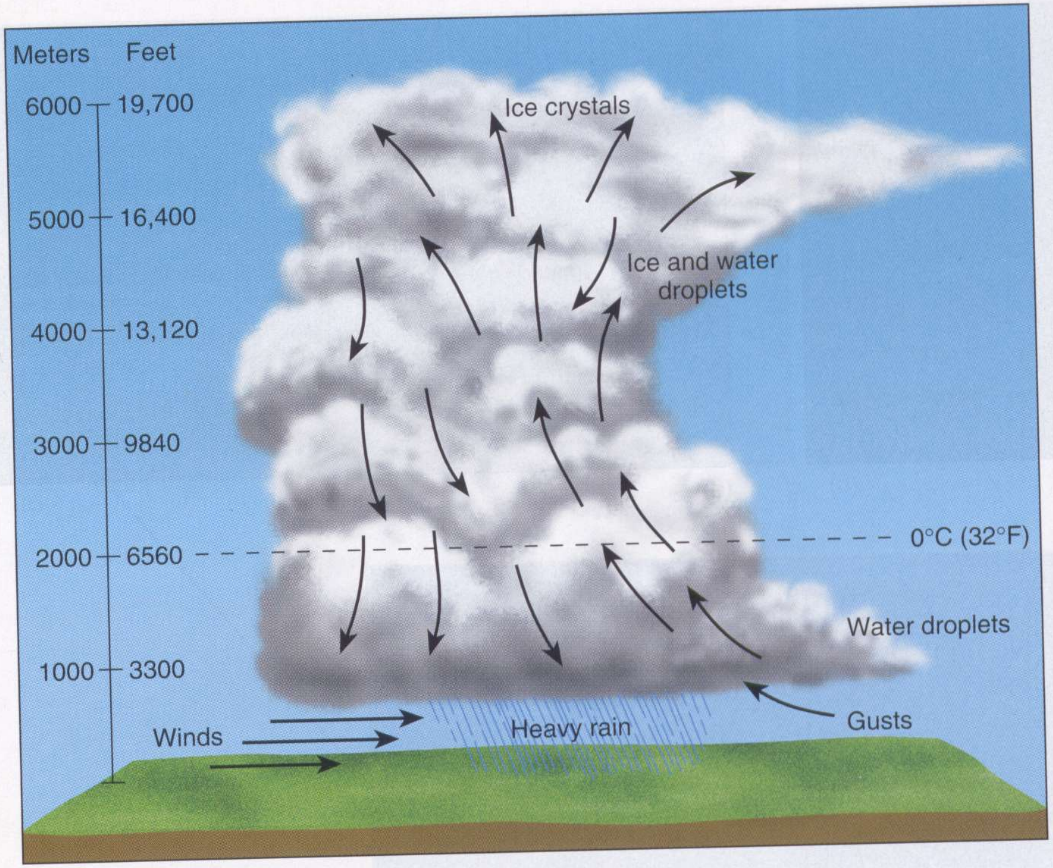
(a) High clouds (p<440mb)

(b) Low clouds p>680mb

(c) All clouds



Source:
NASA

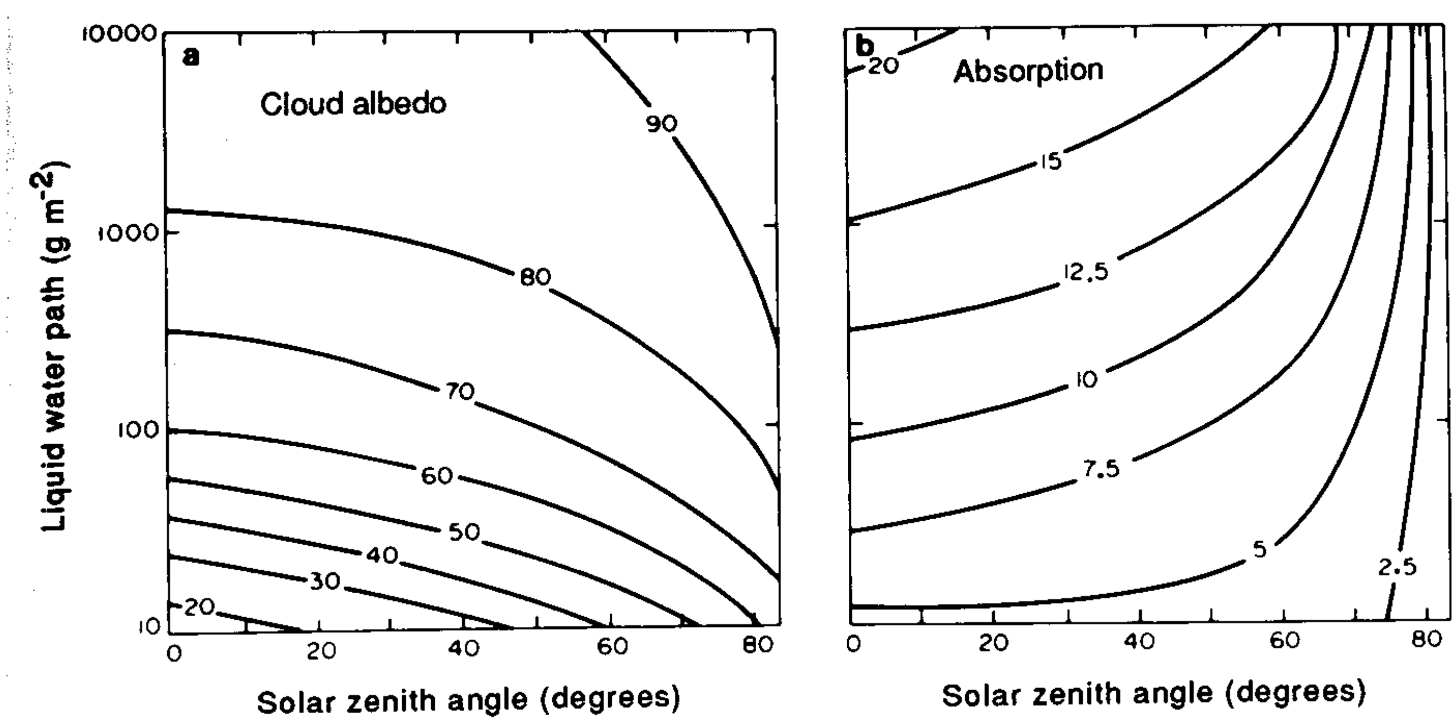

Meters
Feet
6000
19,700
5000
16,400
4000
13,120
3000
9840
2000
6560
1000
3300
Ice crystals
Ice and water
droplets
0°C (32°F)
Water droplets
Gusts
Heavy rain
Winds

(a)

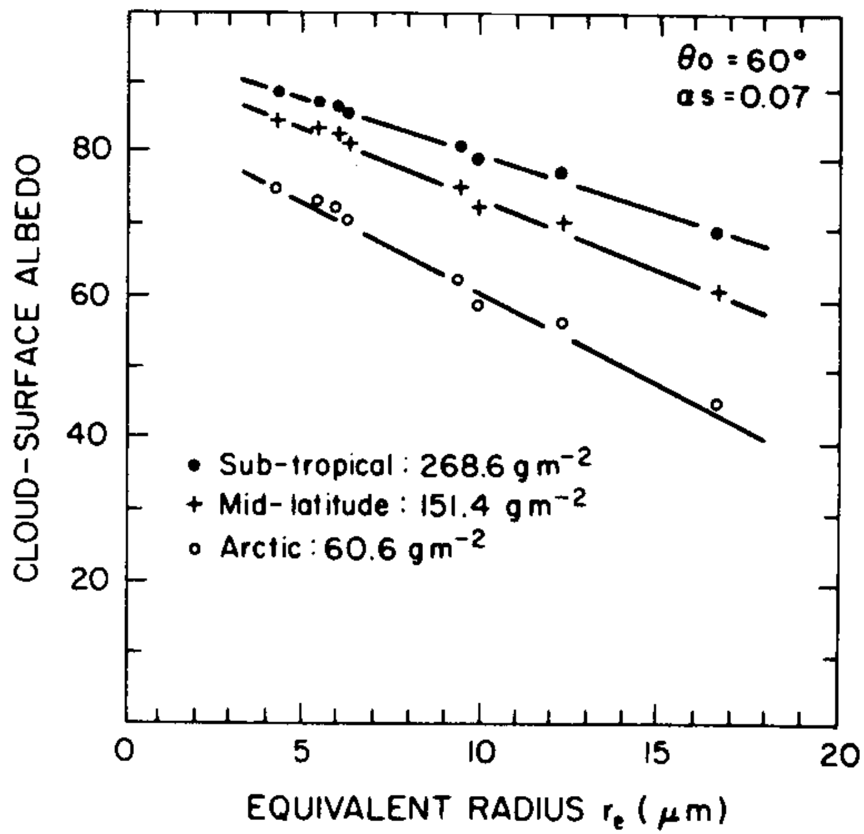
(b)

Clouds and SW albedo and absorption

In the plane-parallel assumption, and assigning the droplet size distribution and the vertical profile of humidity, curves may be derived, giving the cloud albedo and the cloud absorption of solar radiation as functions of solar zenith angle and of "liquid water path", essentially the thickness of the cloud and therefore its mass or total liquid water content in a vertical column of unit area.



Cloud albedo increases with decreasing clouds droplet size (for a given amount of total cloud water): albedo is greater for smaller droplets, that present a larger surface area for the same mass.



Fixed solar zenith angle

Total water content

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

Clouds and terrestrial radiation

Clouds are efficient absorbers/emitters of LW radiation - notice how fast emissivity reaches one.

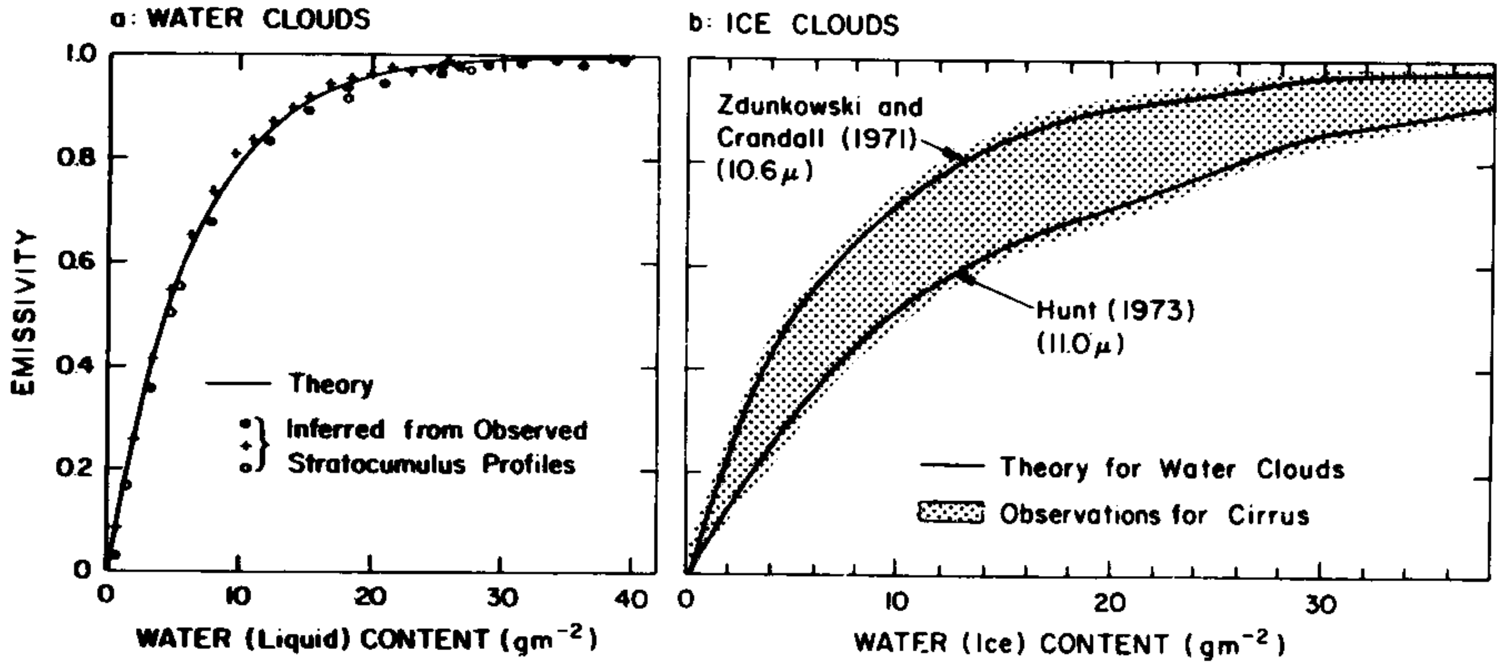
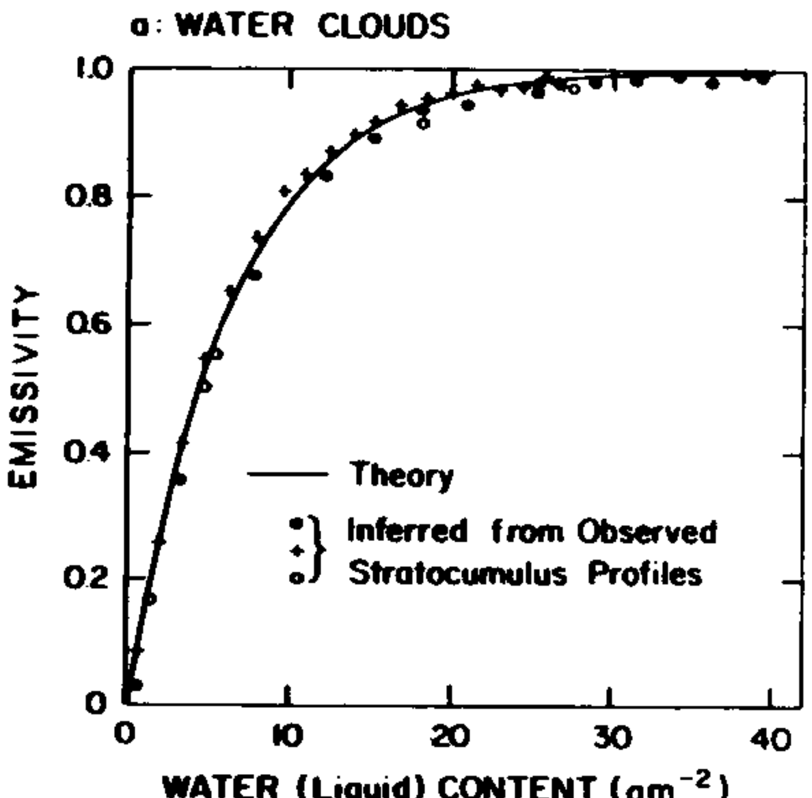
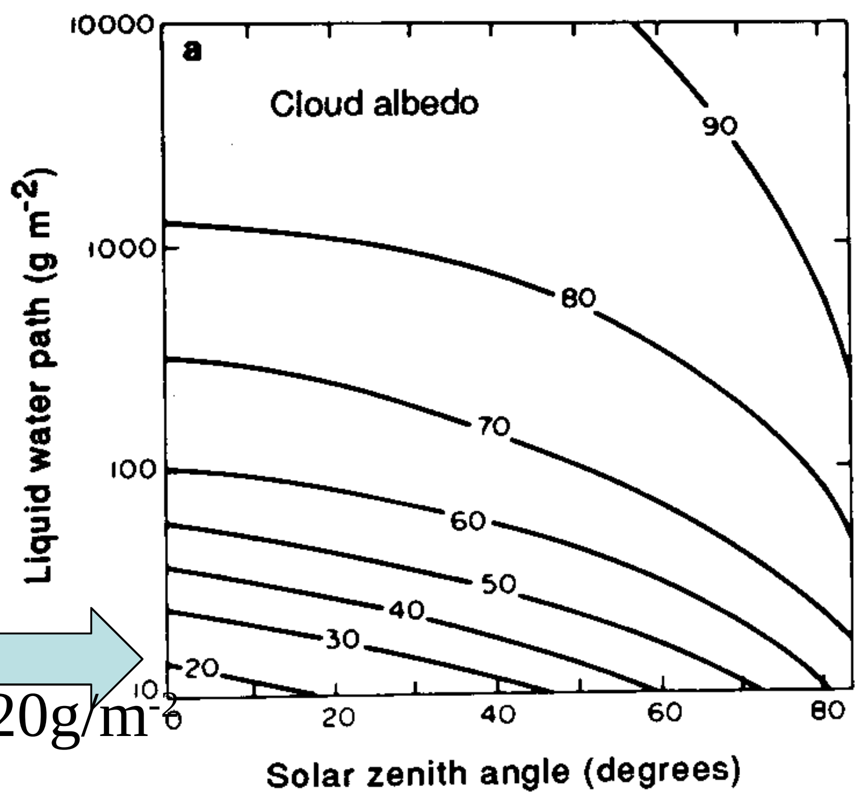


Fig. 3.15 The dependence of the longwave emissivity on (a) liquid water content [from Slingo *t al.* (1982); reprinted with permission from the Royal Meteorological Society] and (b) ice content [from riffith *et al.* (1980); reprinted with permission from the American Meteorological Society].

Notice how fast clouds saturate in the LW emissivity but not in the SW albedo (arrow points to 20g/m^2). It means that even relatively thin clouds are effective blackbodies for LW, whereas a lot more cloud water is needed to make cloud albedo be close to 100%

20 g/m^{-2}





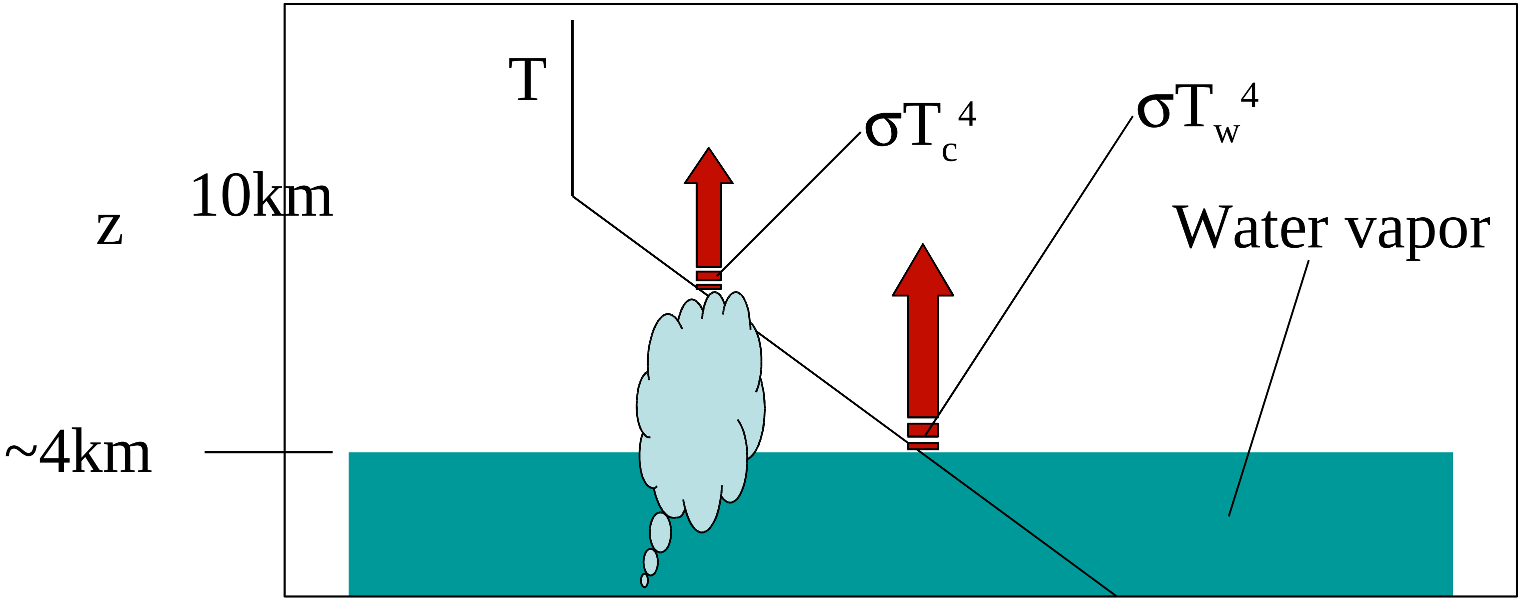
20g/m^{-2}

Outgoing Longwave Radiation (OLR) and clouds

For the case of no clouds, it is the water vapor at the lower levels of the troposphere that is the main contribution to $F^{\Uparrow}$.
So, the temperatures at which water vapor emits are those at ~ 4 km of height, below which the specific humidity is large.
When clouds are introduced, the height z (and hence the T where the effective emission occurs) may change.
If the clouds are low (within the ~ 4 km where the specific humidity is large), then clouds have relatively little effect.
But if clouds are high, then the height at which the effective emission occurs is more or less the height of the cloud top, which can be as high as the tropopause (and hence significantly colder).
In summary, it is usually the first significant concentration of water vapor or clouds looking down from the TOA that dictates the amount of OLR.

OLR (LW radiation at the top of the atmosphere)

•If no cloud, OLR comes from the top of the water vapor layer around ~4km
•If clouds are below ~4km, it has little effect on OLR
•If clouds are above ~4km, OLR comes from the level of the cloud top



Cloud forcing is the cloud contribution to net radiation absorbed by the earth (assume a cloud layer of given properties)

Net SW+LW TOA flux ***clear sky***: $R_{clear} = S_o(1-\alpha_{clear})/4 - \sigma T_w^4$

Net SW+LW TOA flux ***cloudy sky***: $R_{cloudy} = S_o(1-\alpha_{cloudy})/4 - \sigma T_c^4$

Cloud forcing = TOA flux cloudy sky - TOA flux clear sky

$$= S_o(\alpha_{clear} - \alpha_{cloudy})/4 - \sigma (T_c^4 - T_w^4)$$

Cooling since $\alpha_{clear} < \alpha_{cloudy}$

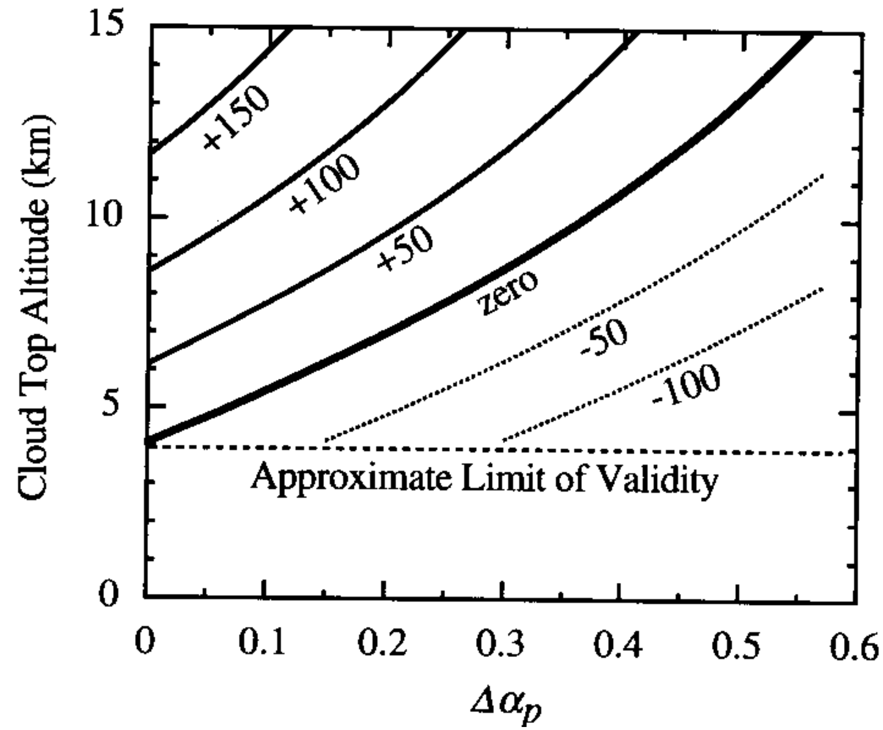
Warming since $T_c^4 < T_w^4$

$\alpha_{clear} \sim 0.15$

$\Delta\alpha_p = \alpha_{cloudy} - \alpha_{clear}$

NOTE: you can define cloud forcing also for just LW or SW by itself (so: LW cloud forcing, SW cloud forcing)

From calculations: change in net radiation at TOA caused by the insertion of a cloud layer into a clear atmosphere→ cloud effect can go in ***either*** direction



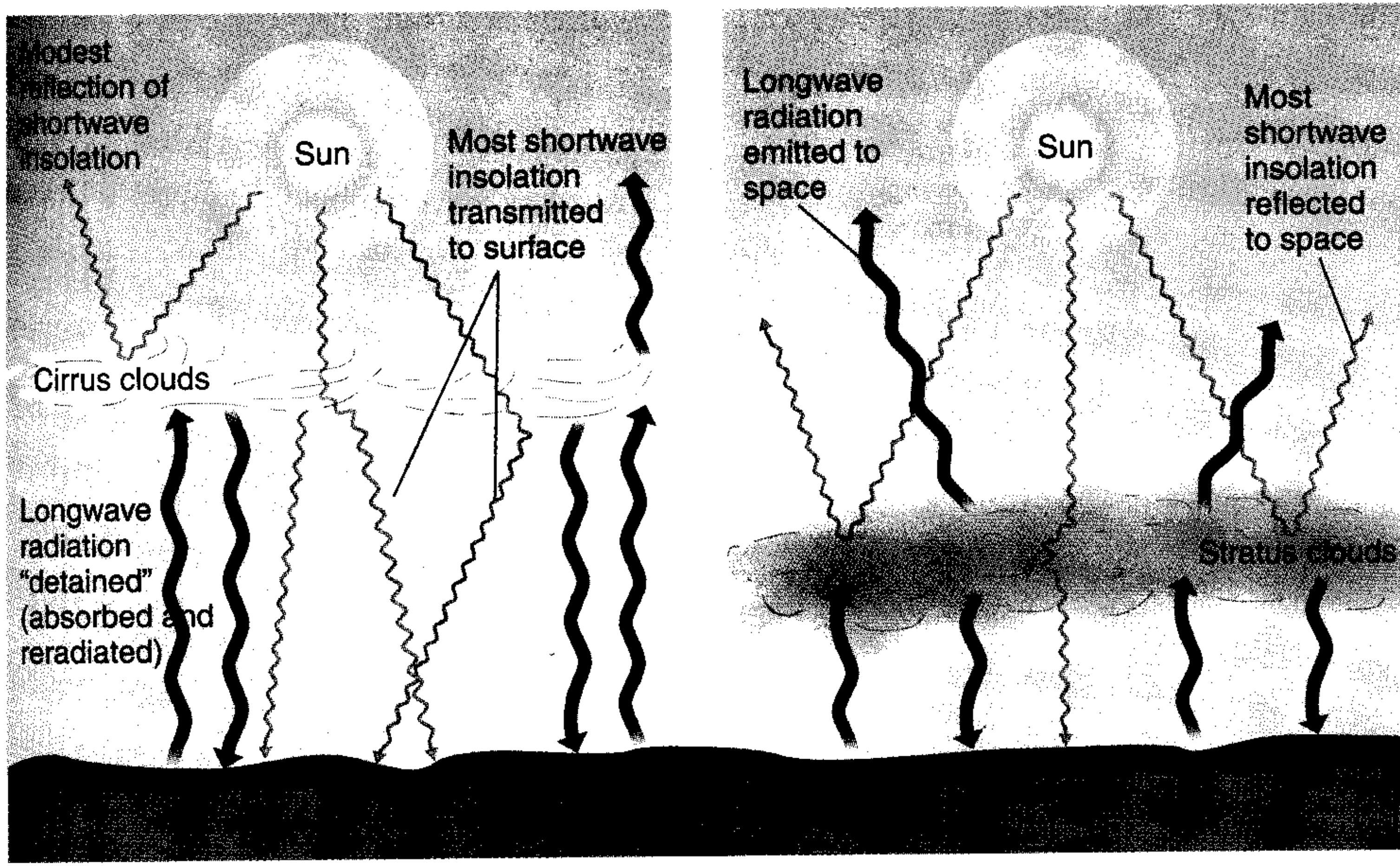
But on the global average, clouds ***cool*** (as estimated from observations): loss of ~ 20 W/m^2

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.] (± 5 W/m^2)



(a) High clouds: net greenhouse forcing and atmospheric warming

(b) Low clouds: net albedo forcing and atmospheric cooling

(source: elemental geosystems)

Radiative-convective equilibrium model

- Low clouds reduce T at the surface and in the troposphere
- High clouds can cause T_s exceed the value obtained without clouds

The albedo assumed for lower clouds are higher, so that the greater reflection of SW radiation from these clouds explains a good part of their strong coling effect (strong albedo forcing). Lower clouds have a weaker effect on escaping LW, however, since their top T are warmer, and this also explains part of the greater cooling effect of low clouds in these calculations (little greenhouse effect).
The result is a net albedo forcing.

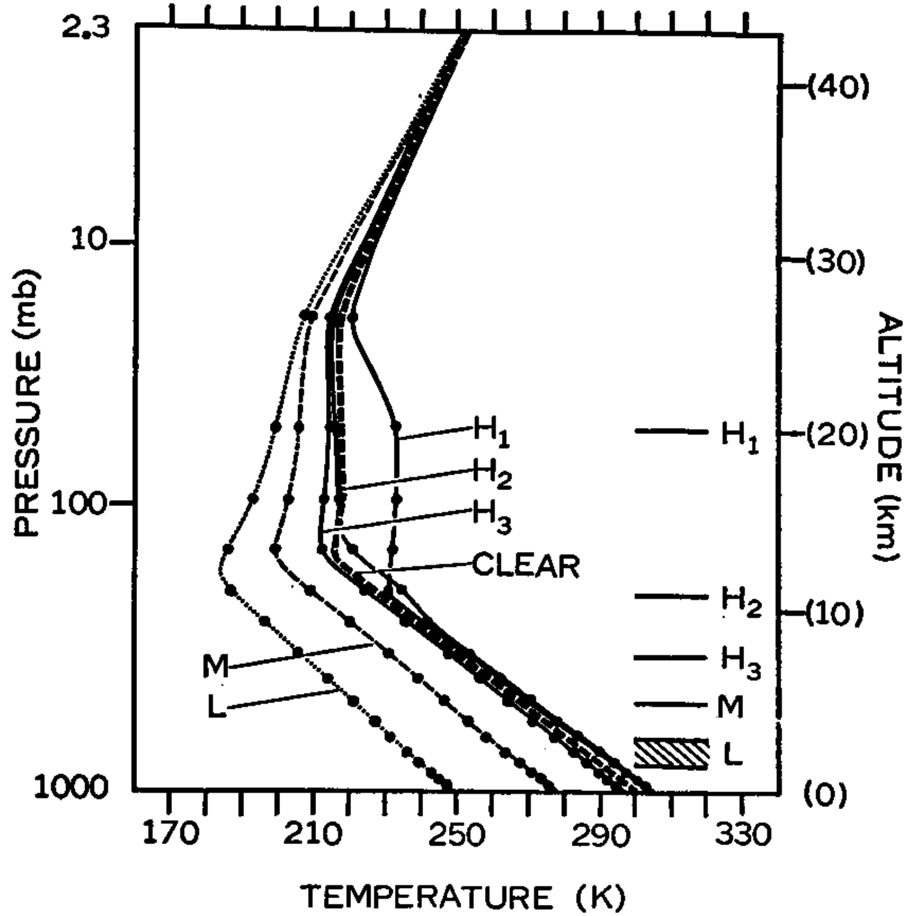
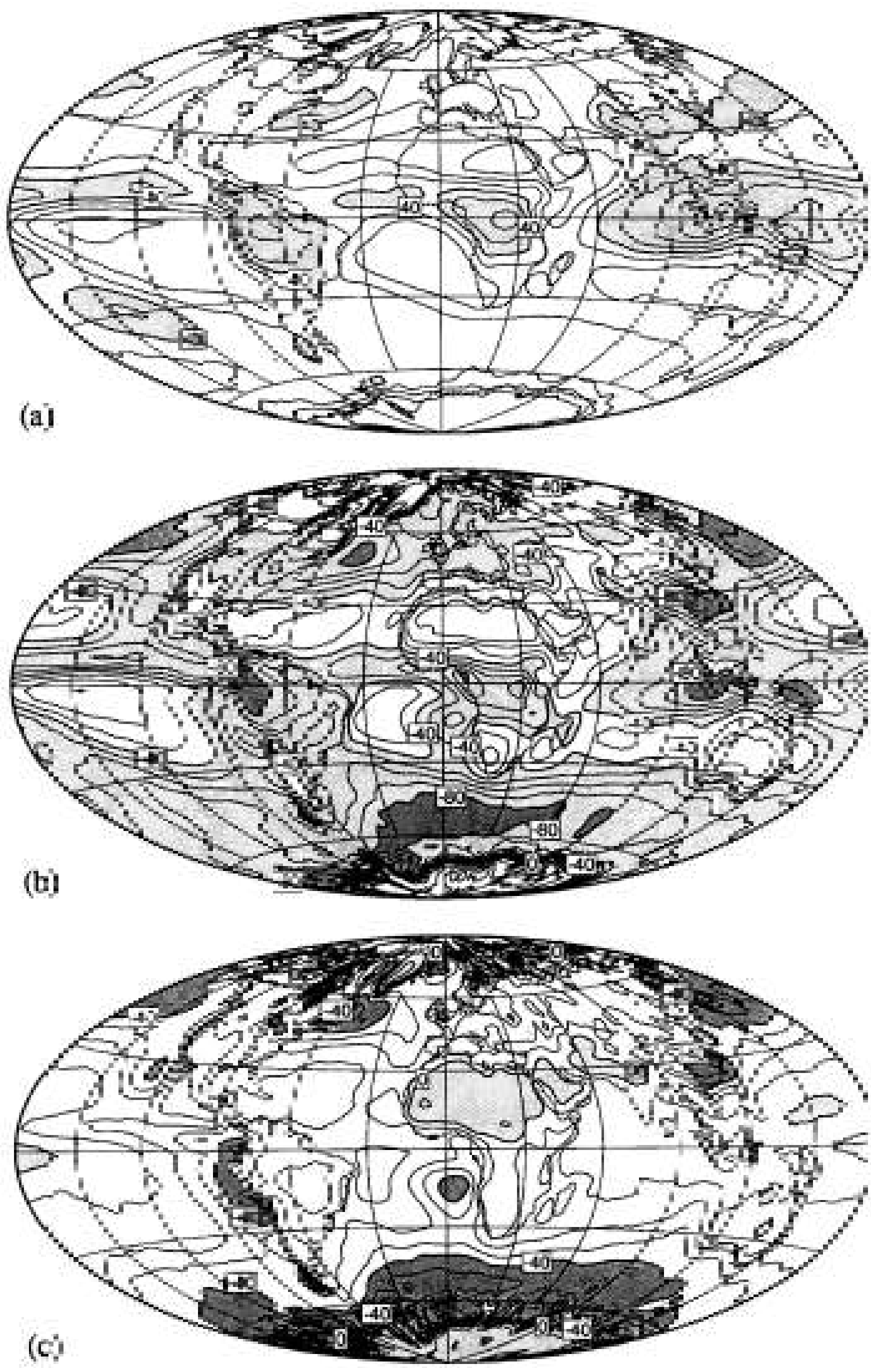


Fig. 3.19 Thermal equilibrium temperature profiles for atmospheres with various cloud distributions. The cloud heights corresponding to each type of cloud are shown on the right (L = low, M = medium, and H = high cloud). The heavy dashed line shows the equilibrium profile for clear skies. [From Manabe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

Cloud radiative effects



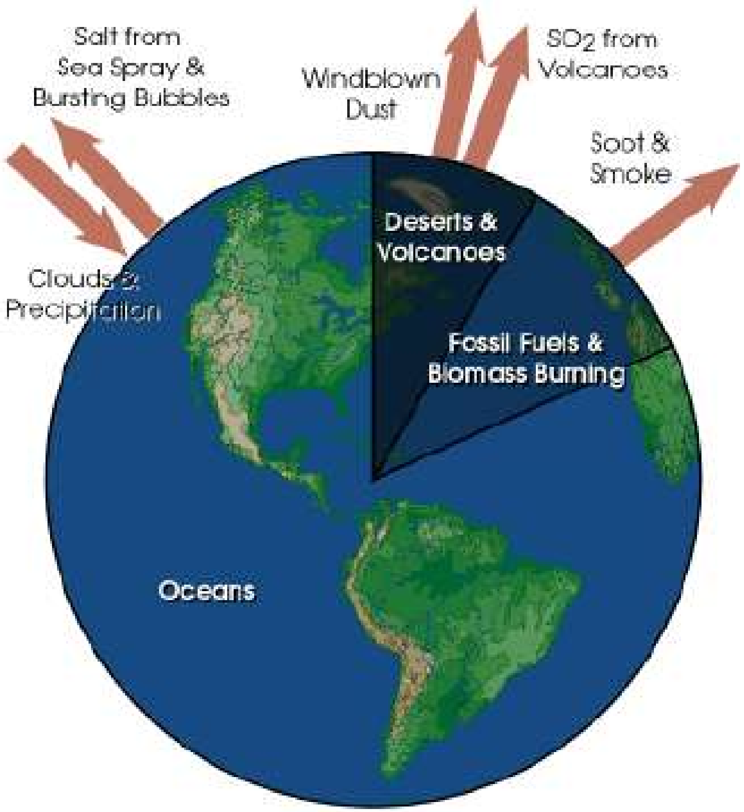
Reduction in OLR

Increase in absorbed solar

Cloud forcing

Aerosol impact on radiative budget

Major sources of aerosol



Aerosols are minute particles (solid or liquid) suspended in the atmosphere.

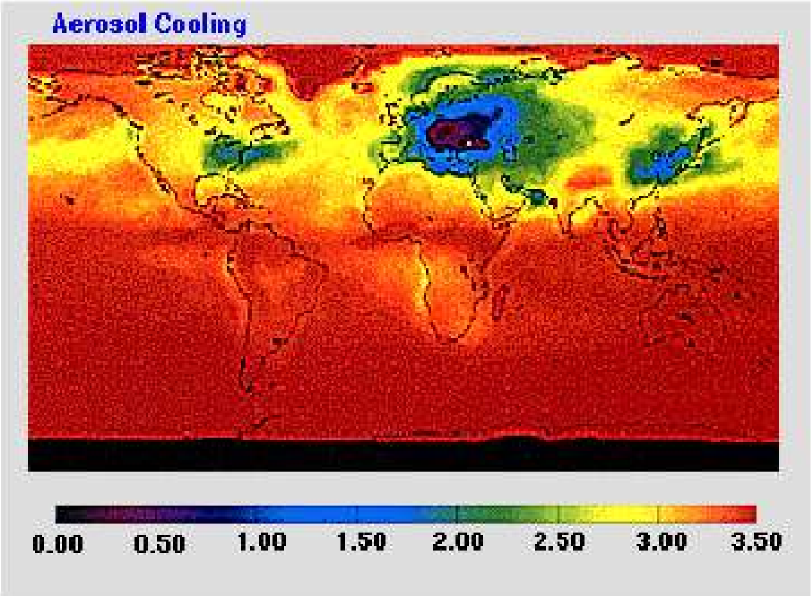
Prominent aerosols: sulfate (salts that contain a charged group of sulfur and oxygen atoms); dust; fog; sea salt, etc…

They affect the climate through changing the radiative properties of the atmosphere.

http://earthobservatory.nasa.gov/Library/Aerosols/

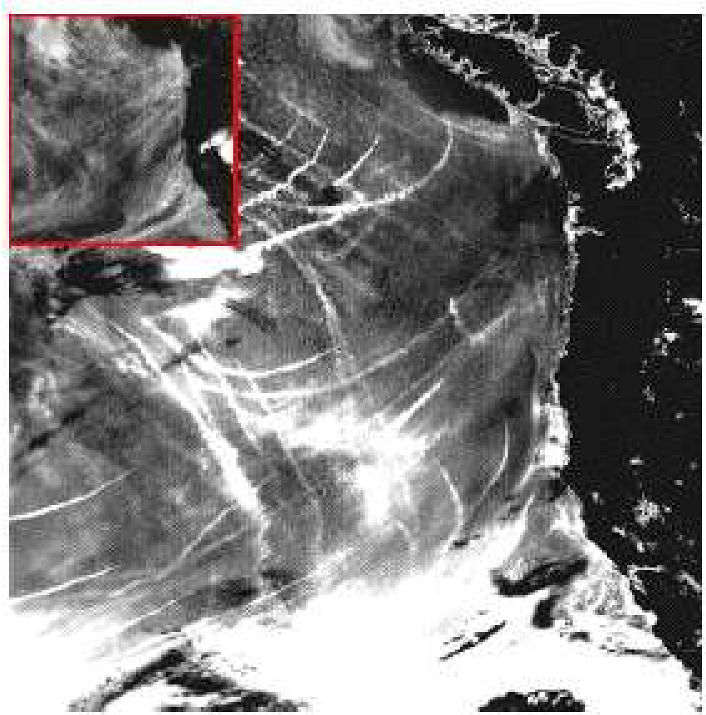
Impact on climate:

Direct impact through reflection / absorption of radiation



(source: NASA)

Indirect effect: aerosols act as cloud condensation nuclei, making cloud droplet size smaller and more numerous→more reflection by clouds, cooling effect



Ship tracks off the west coast of the US

O_3 and aerosols

Aerosols can act as sites for chemical reactions to take place. The most significant are those that lead to the destruction of stratospheric ozone.
During winter in polar regions, aerosols grow to form polar stratospheric clouds of large surface areas where reactions form large amounts of reactive Cl and ultimately lead to the destruction of ozone in the stratosphere.
Similar changes in stratospheric ozone concentrations occur after major olcanic eruptions, with emission of SO_2, HCl and ash.