Atmospheric Physics

Lecture 2

J. Sahraei

Physics Department,

Razi University

https://sci.razi.ac.ir/~sahraei

Atmospheric models

Unlike laboratory physicists, atmospheric researchers cannot perform controlled experiments on the large-scale atmosphere.

The standard 'scientific method', of observing phenomena, formulating hypotheses, testing them by experiment, then formulating revised hypotheses and so on, cannot be applied directly.

Instead, after an atmospheric phenomenon is discovered, perhaps by sifting through a great deal of data, we develop models, which incorporate representations of those processes that we hypothesise are most important for causing the phenomenon.

Two simple atmospheric models

It is a basic observational fact that the Earth's mean surface temperature is about 288 K

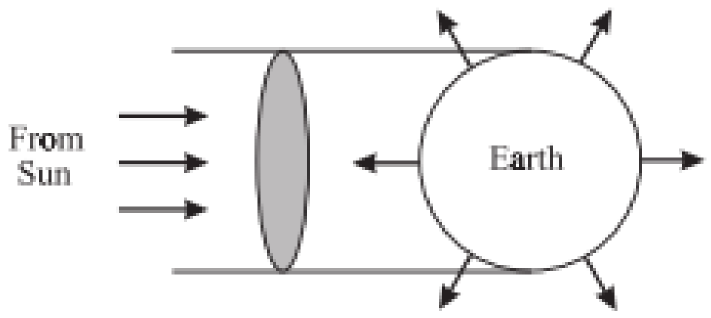
A model with a non-absorbing atmosphere

The solar power per unit area at the Earth's mean distance from the Sun (the total solar irradiance, TSI, formerly called the solar constant) is:

$$F_s = 1370\ \mathrm{Wm^{-2}}$$

in a tube of cross-sectional area πa^2

The total solar energy received per unit time is $F_s \pi a^2$



planetary albedo $A = 0.3$;

that is, 30% of the incoming solar radiation is reflected back to space without being absorbed:

$$0.3 F_s \pi a^2$$

If the Earth is assumed to emit as a black body at a uniform absolute temperature T then, by the Stefan-Boltzmann law, Power emitted per unit area = σT^4

However, power is emitted in all directions from a total surface area

the total power emitted is: $4\pi a^2 \sigma T^4$

$$(1-A)F_s \pi a^2 = 4 \pi a^2 \sigma T^4 \quad \rightarrow \quad T \cong 255^\circ K$$

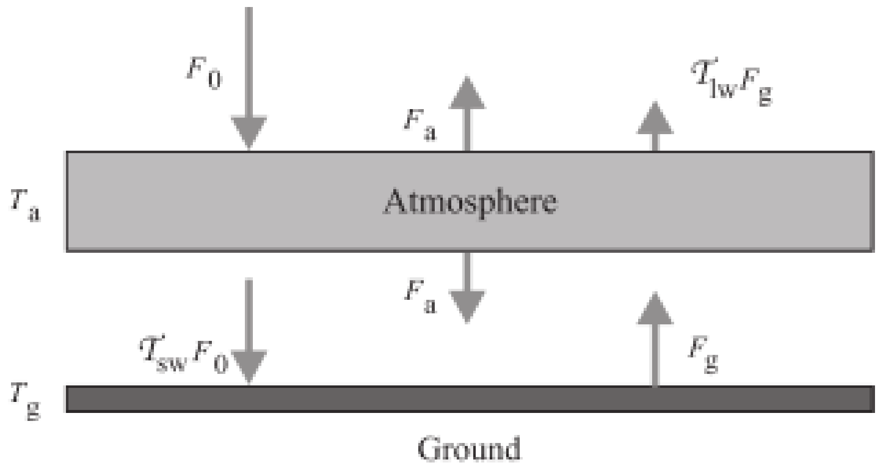
The temperature obtained from this calculation is called the effective emitting temperature of the Earth:

$$T_e \equiv \left(\frac{F_0}{\sigma}\right)^{1/4} \approx 255K$$

Its value is significantly lower than the observed mean surface temperature of about 288 K

A simple model of the greenhouse effect

We now consider the effect of adding a layer of atmosphere, of uniform temperature T_a



The atmosphere is assumed to transmit:

a fraction of any incident solar (short-wave) radiation τ_{sw}

fraction of any incident thermal (infra-red, or long-wave) radiation τ_{lw}

These fractions are called transmittances

We assume that the ground is at temperature T_g

Taking account of albedo effects and the difference between the area of the emitting surface $4\pi a^2$

and the intercepted cross-sectional area of the solar beam πa^2

The mean unreflected incoming solar irradiance at the top of the atmosphere is:

$$(1-A)F_s \pi a^2 = 4\pi a^2 \sigma T^4 \qquad F_0 = \frac{1}{4}(1-A)F_s \approx 240\ Wm^{-2}$$

F_0

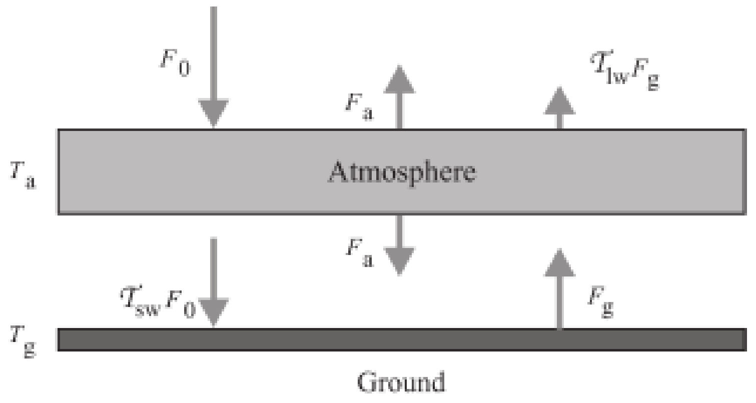
$\tau_{sw} F_0$ is absorbed by the ground

$(1-\tau_{sw}) F_0$ is absorbed by the atmosphere

The ground is assumed to emit as a black body

an upward irradiance $F_g = \sigma T_g^4$

$\tau_{lw} F_g$ reaches the top



The atmosphere is not a black body, but emits irradiances

$$F_a = (1-\tau_{lw}) \sigma T_a^4$$

Kirchhoff's law: the emittance – the ratio of the actual emitted irradiance to the irradiance that would be emitted by a black body at the same temperature – equals the absorptance

$$\text{emittance} = \frac{\textit{the actual emitted irradiane}}{\textit{the irradiance that would be emitted by a black body}} = 1 - \tau_{lw}$$

We now assume that the system is in radiative equilibrium: that is, energy transfer takes place only by the radiative processes described above, and the associated irradiances are in balance everywhere;

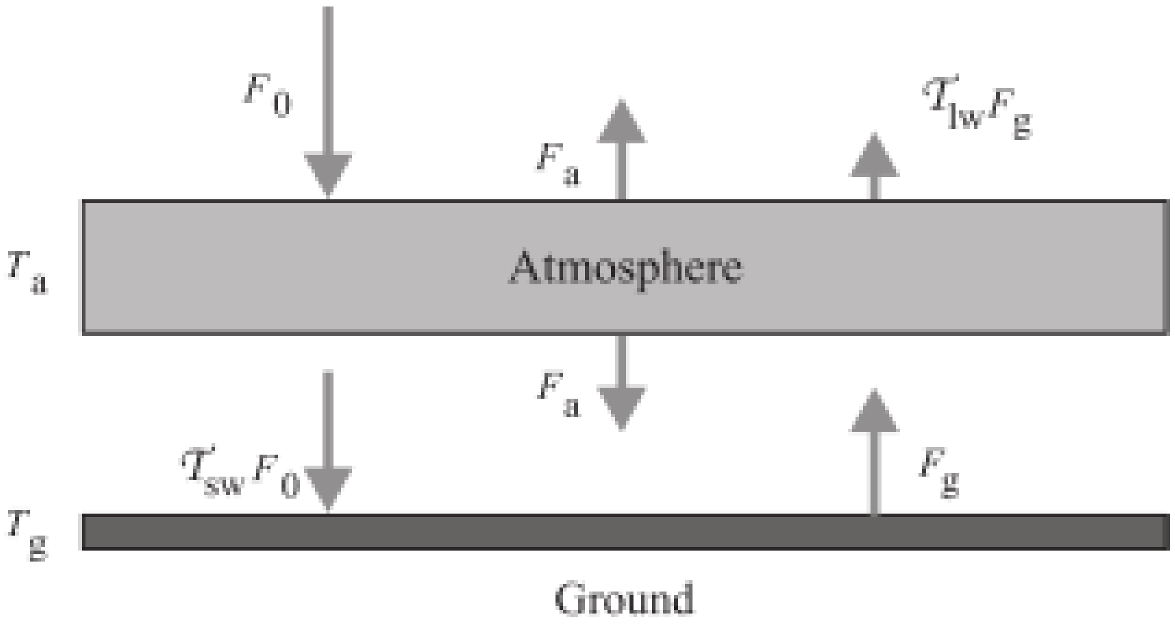
we neglect any energy transfers due to non-radiative processes such as fluid motions.

Equating irradiances, we have

$$F_0 = F_a + \tau_{lw} F_g$$ above the atmosphere, and

$$F_0 = F_a + \tau_{lw} F_g \quad *$$

above the atmosphere



$$F_g = F_a + \tau_{sw} F_0 \quad *'$$

between the atmosphere and the ground

By eliminating F_a from equations we obtain

$$F_g = \sigma T_g^4 = F_0 \frac{1+\tau_{sw}}{1+\tau_{lw}}$$

In the absence of an absorbing atmosphere, we would have

$$\tau_{sw} = \tau_{lw} = 1 \qquad F_g = F_0 \qquad T_g \approx 255K$$

Taking rough values for the Earth's atmosphere to be

$\tau_{sw} = 0.9$ strong transmittance and weak absorption of solar radiation

$\tau_{lw} = 0.2$ weak transmittance and strong absorption of thermal radiation

$$T_g \approx 286K$$

This close agreement is somewhat fortuitous, however, since in reality non-radiative processes also contribute significantly to the energy balance.

We can also find the atmospheric emission from * equations:

$$F_a = (1-\tau_{lw})\sigma T_a^4 = F_0 \frac{1-\tau_{sw}\tau_{lw}}{1+\tau_{lw}}$$

and this gives the temperature of the model atmosphere,

$$T_a \approx 245K$$

One way to quantify the 'greenhouse effect' of an absorbing gas is in terms of the amount $F_g - F_0$ by which it reduces the outgoing irradiance from its surface value: in the case discussed above this reduction is 140 Wm^{-2}

This equals the difference between the amount

$$(1-\tau_{lw})\sigma T_g^4 = 304\ Wm^{-2}$$

of the thermal emission from the 'warm' surface that is absorbed by the 'cool' atmosphere

and the smaller amount

$$F_a = (1-\tau_{lw})\sigma T_a^4 = 164\ Wm^{-2}$$

that the atmosphere re-emits upwards.

Since the atmosphere is in equilibrium, it also equals the difference between the downward emission F_a from the atmosphere and the small proportion

$$(1-\tau_{sw})F_0 = 24\ Wm^{-2}$$

of the solar irradiance that it absorbs.