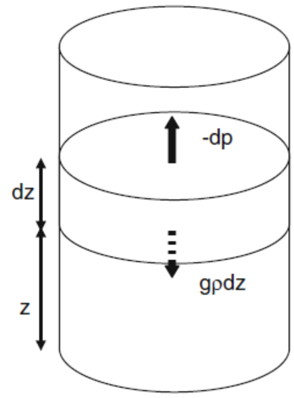
Lecture 1

The Static Equilibrium in the Atmosphere

1.1 The Results of Hydrostatic Balance

Newton's law requires that the upward force acting on thin a layer of air from the decrease of pressure with height is generally closely balanced by the downward force due gravity (as in the figure). The hydrostatic equation is then:



$$\frac{\partial p}{\partial z} = -g\rho$$

Typically, deviations from the hydrostatic balance occur locally, e.g. in updrafts and downdrafts or when the air hits a small obstacle. In contrast to the hydrostatic balance, then an air particle undergoes acceleration:

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{dp}{dz} - g$$

We shall discuss two typical models that approximate the atmosphere.

A. The Homogeneous Atmosphere

In this atmosphere, the density is considered constant anywhere.

$\rho = \rho_o = constant$ (where ρ_o is the air density at the surface?)

From the hydrostatic equation $\frac{dp}{dz} = -g\rho_o$

$$\int_{p_o}^{p} dp = -g\rho_o \int_0^z dz$$

$$p - p_o = -g\,\rho_o\,z$$

$$p = p_o - g\rho_o z \qquad (1.1)$$

Where p_o is pressure at $z = 0$

The homogeneous atmosphere has a finite height H,

when p=0 (at the top of the atmosphere), z=H

$$0 = p_o - \rho_o g H$$

$$\therefore H = \frac{p_o}{\rho_o g}$$

And from the hydrostatic equation and the equation of state ($p_o = \rho_o \, R \, T_o$) we get:

$$H = \frac{\rho_o \, R \, T_o}{\rho_o \, g}$$

At $T_o = 283^o K$, $R = 287$, $g = 9.8 \, ms^{-2} \Rightarrow H \approx 8000 \, m$

(Homework: solve for T_o = 293, 300, 310, 320, 330 ^{o}K)

We may define a temperature in the homogeneous atmosphere from gas equation:

$$p = \rho_o R \, T$$

$$T = \frac{p}{\rho_o R} \qquad (1.2)$$

Put eqn. (11.1) in eqn. (11.2)

$$T = \frac{p_o - \rho_o \, g \, z}{\rho_o \, R} \quad \Rightarrow \quad T = \frac{p_o}{\rho_o R} - \frac{\rho_o \, g \, z}{\rho_o R}$$

$$\boxed{T = T_o - \frac{g}{R} z}$$

This equation shows that T decreases linearly with height in a homogeneous atmosphere.

Question: From the atmospheric model (homogeneous atmosphere) show that the lapse rate $\gamma = \frac{dT}{dZ} = -3.4 \;\; ^o K/100m$

B. The Isothermal Atmosphere

In this model we have $T = T_o = const.$ (where T_o is the temperature at the surface)

From the hydrostatic Equation we get:

$$dp = -\rho g \, dz$$

Recall that $\rho = \frac{p}{RT_o}$

$$dp = \frac{p}{RT_o} g \, dz$$

$$\int_{p_o}^{p} \frac{dp}{p} = -\frac{g}{RT_o} \int_0^z dz$$

$$\ln\frac{p}{p_o} = -\frac{g}{RT_o}z$$

Taking exponential to both sides

$$\frac{p}{p_o} = e^{-\frac{g}{RT_o}z}$$

This equation shows that the isothermal atmosphere is of infinite extent because $p \to 0$ when $z \to \infty$

$$\boxed{p = p_o e^{-\frac{g}{RT_o}Z}}$$

The scale height for an isothermal atmosphere is often defined as the height at which the pressure has decreased to e^{-1} of the surface pressure.

$$z = H_s$$

$$p = p_o\, e^{-\frac{g}{RT_o}H_s}$$

$$p = p_o\, e^{-1}$$

$$p_o e^{-\frac{g}{R\,T_o}H_s} = p_o\, e^{-1}$$

$$-\frac{g}{RT_o}H_s = -1$$

$$\therefore H_s = \frac{R\,T_o}{g} = 8000\,m$$

Or, that the scale height is equal to the height of the homogeneous atmosphere having the same surface temperature as the isothermal atmosphere.

The density in the isothermal atmosphere can be calculated from gas equation

$p_o = \rho_o\, R\, T_o \quad , p = \rho\, R\, T_o$:

$$p = p_o e^{-\frac{g}{RT_o}Z}$$

$$\rho\, R\, T_o = \rho_o\, R\, T_o\, e^{-\frac{g}{RT_o}z}$$

$$\therefore \rho = \rho_o\, e^{-\frac{g}{RT_o}z}$$