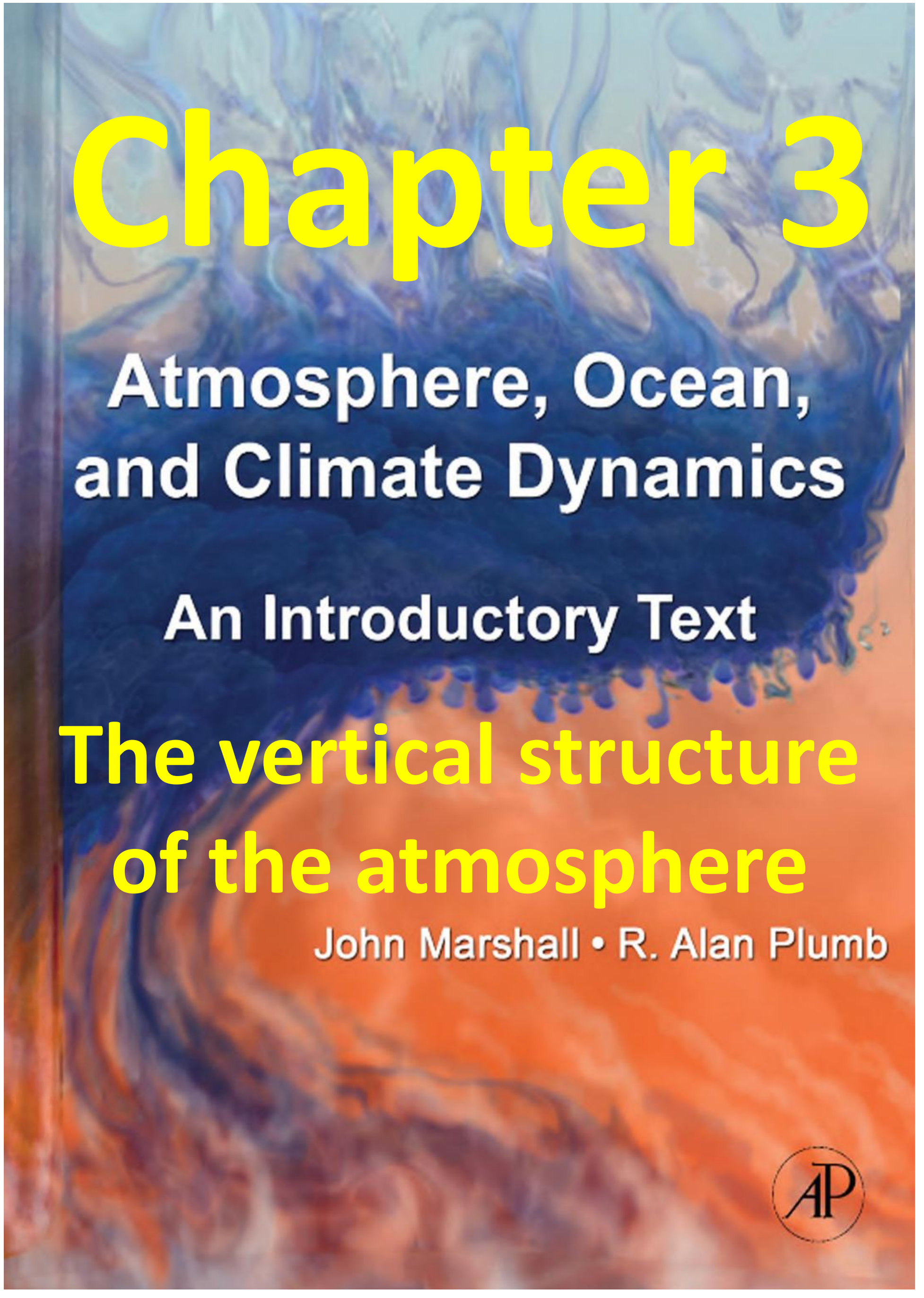
Chapter 3

The vertical structure of the atmosphere



O_2 abs UV

O_3 abs UV

O_3 peak conc

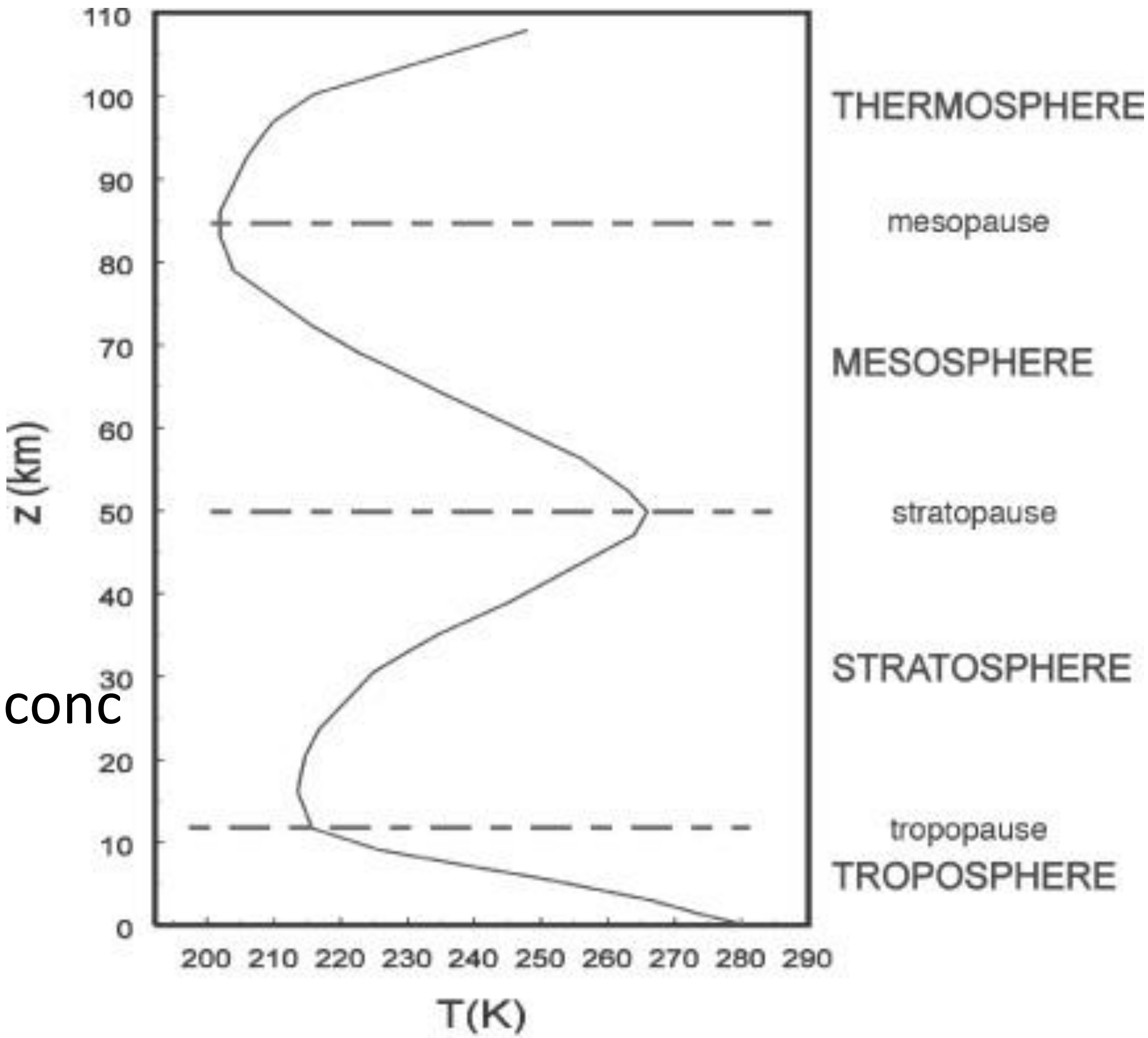


Figure 3.1: Vertical temperature profile for the "US standard atmosphere" at 40° N in December.

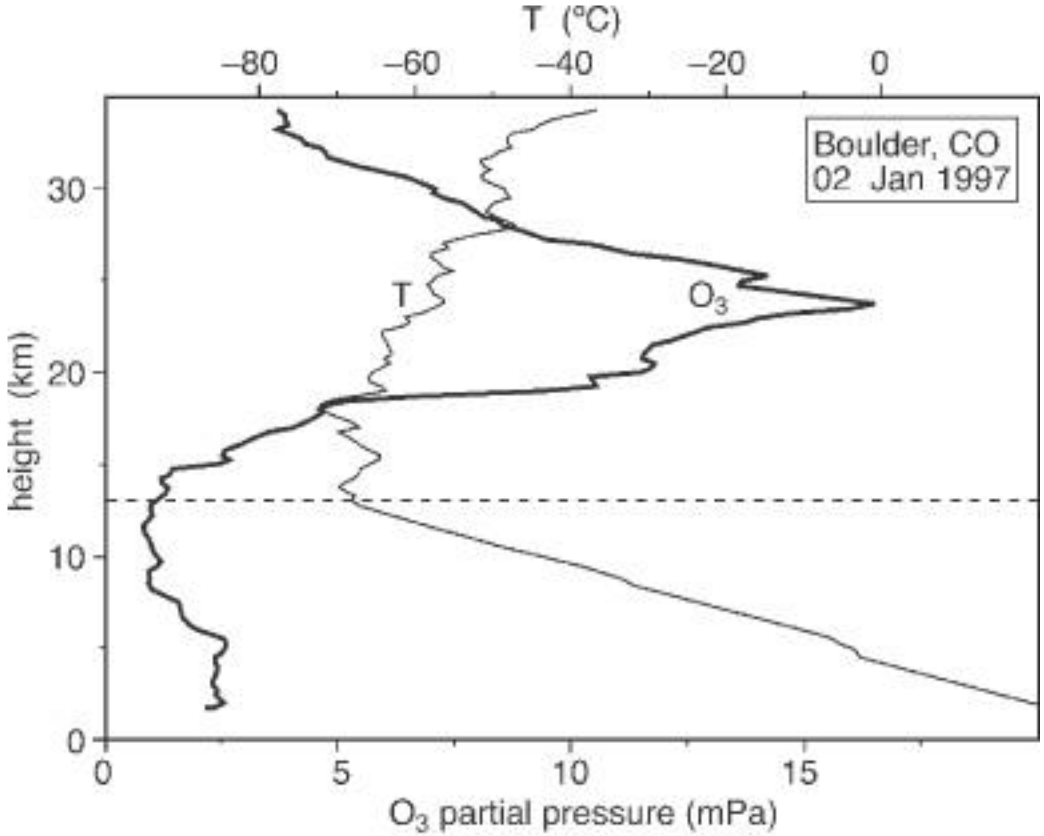


Figure 3.2: A typical winter ozone profile in middle latitudes (Boulder, CO, USA, 2 Jan 1997). The heavy curve shows the profile of ozone partial pressure (mPa), the light curve temperature (°C) plotted against altitude up to about 33 km. The dashed horizontal line shows the approximate position of the tropopause. Balloon data courtesy of NOAA Climate Monitoring and Diagnostics Laboratory.

Spesifikk fuktighet (q): Masse (kg) av vanndamp pr. masse (kg) av luft

$$q = \frac{\rho_v}{\rho}\,, \qquad (4\text{-}23)$$

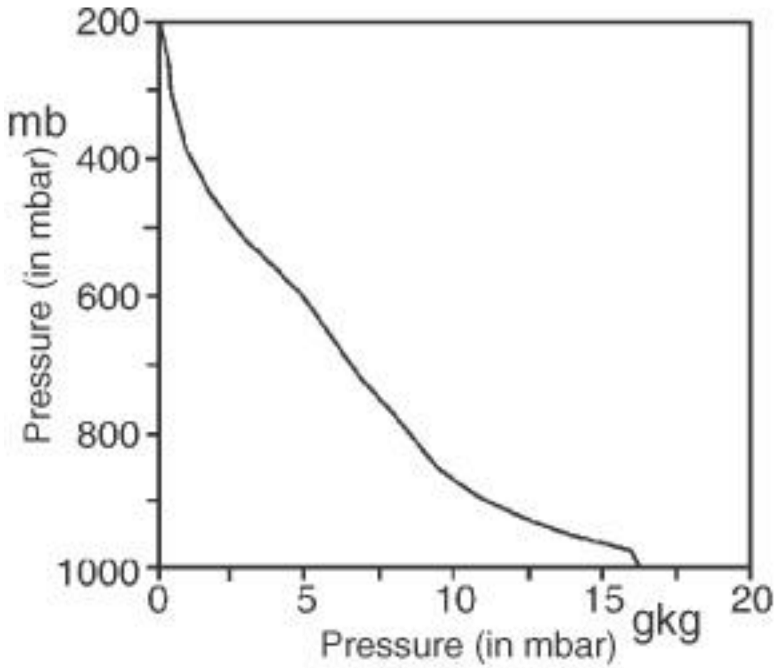


Figure 3.3: The global average vertical distribution of water vapor (in g kg^{-1}) plotted against pressure (in mbar).

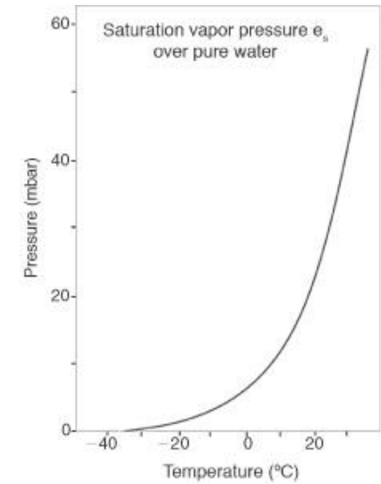


Figure 1.5: Saturation vapor pressure e_s (in mbar) as a function of *T* in °C (solid curve).

(From Wallace & Hobbs, (2006).)

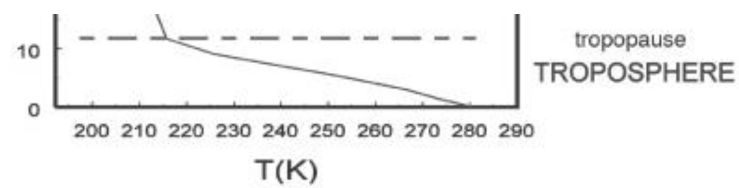
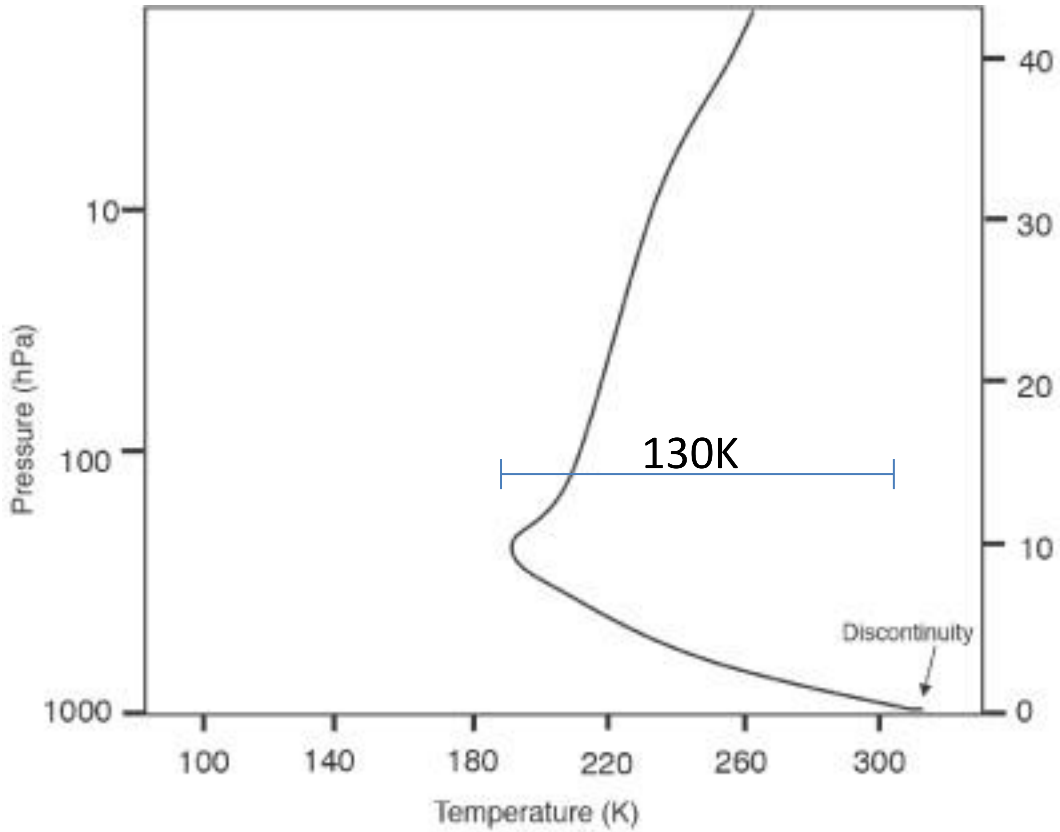


Figure 3.1: Vertical temperature profile for the "US standard atmosphere" at 40° N in December.

Beregnet temperaturprofil ved ren stålingslikevekt

Global årlig middel, realistisk vertikalprofiler av ozon, H_2O og CO_2



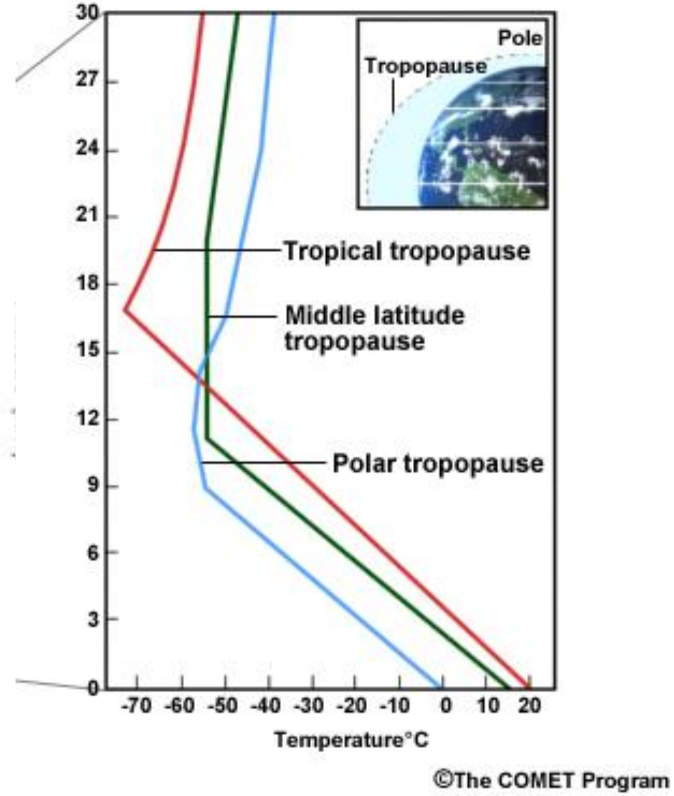


Figure 2.11: The radiative equilibrium profile of the atmosphere obtained by carrying out the calculation schematized in Fig. 2.10. The absorbers are H_2O, O_3, and CO_2. The effects of both terrestrial radiation and solar radiation are included. Note the discontinuity at the surface.

(Modified from Wells (1997).)

65K

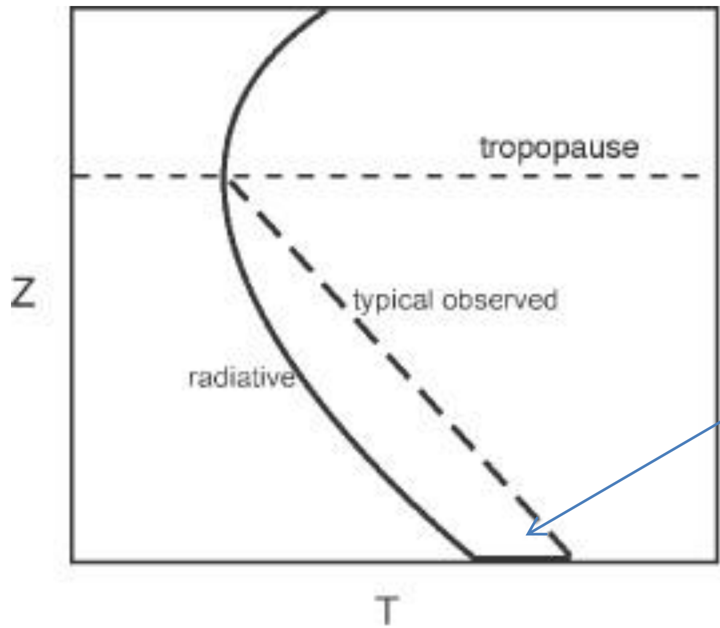


Figure 3.4: A schematized radiative equilibrium profile in the troposphere (cf. Fig. 2.11) (solid) and a schematized observed profile (dashed). Below the tropopause, the troposphere is stirred by convection and weather systems and is not in radiative balance. Above the tropopause, dynamical heat transport is less important, and the observed T is close to the radiative profile.

Diskontinuitet ved overflaten er ikke observert

Hydrostatisk Balanse

- Luften er i ro
- Trykket ved et gitt nivå er bestemt av vekten av atmosfæren over

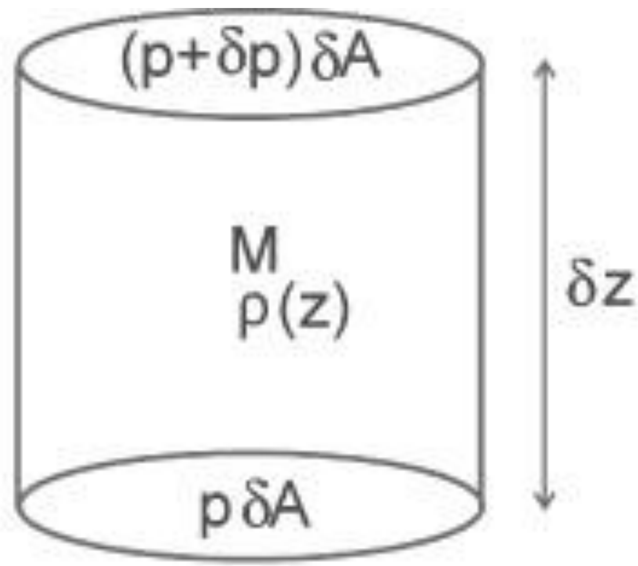
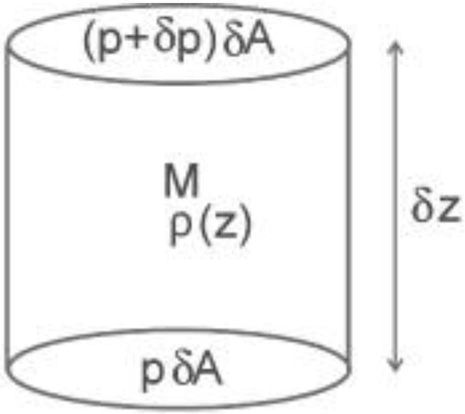


Figure 3.5: A vertical column of air of density ρ, horizontal cross-sectional area δA, height δz, and mass $M = \rho \delta A \delta z$. The pressure on the lower surface is p, the pressure on the upper surface is $p + \delta p$.



Assuming δz to be small,

$$\delta p = \frac{\partial p}{\partial z}\delta z. \qquad (3\text{-}2)$$

the mass of the cylinder is

$$M = \rho\delta A\delta z.$$

Krefter som virker Sylinderen:

i) gravitational force
$F_g = -gM = -g\rho\delta A\delta z$,

ii) pressure force acting at the top face,
$F_T = -\left(p + \delta p\right)\delta A$, and

iii) pressure force acting at the bottom face, $F_B = p\delta A$.

Setting the net force $F_g + F_T + F_B$ to zero gives $\delta p + g\rho\delta z = 0$, and using Eq. 3-2 we

$$\frac{\partial p}{\partial z} + g\rho = 0. \qquad (3\text{-}3)$$

Den hydrostatiske likningen.

God tilnærmelse for mange fenomener i atmosfære og hav.

$$p(z) = g \int_z^{\infty} \rho \, dz. \qquad (3\text{-}4)$$

Trykket i høyden z er gitt ved vekten av atmosfære (+hav) over.

$$\frac{\partial p}{\partial z} + g\rho = 0. \qquad (3\text{-}3)$$

Using the equation of state of air, Eq. 1-1, we may rewrite Eq. 3-3 as

$$\frac{\partial p}{\partial z} = -\frac{gp}{RT}\ . \qquad (3\text{-}5)$$

3.3.1. Isothermal atmosphere

If $T = T_0$, a constant, we have

$$\frac{\partial p}{\partial z} = -\frac{gp}{RT_0} = -\frac{p}{H},$$

where H, the *scale height*, is a constant (neglecting, as noted in Chapter 1, the small dependence of g on z) with the value

$$H = \frac{RT_0}{g}. \tag{3-6}$$

If H is constant, the solution for p is, noting that by definition, $p = p_s$ at the surface $z = 0$,

$$p(z) = p_s \exp\left(-\frac{z}{H}\right). \tag{3-7}$$

Alternatively, by taking the logarithm of both sides we may write z in terms of p thus:

$$z = H \ln\left(\frac{p_s}{p}\right). \tag{3-8}$$

3.3.2. Non-isothermal atmosphere

What happens if T is not constant? In this case we can still define a local scale height

$$H(z) = \frac{RT(z)}{g}, \qquad (3\text{-}9)$$

such that

$$\frac{\partial p}{\partial z} = -\frac{p}{H(z)},$$

where $H(z)$ is the local scale height.

Therefore

$$\frac{1}{p}\frac{\partial p}{\partial z} = \frac{\partial \ln p}{\partial z} = -\frac{1}{H(z)}, \qquad (3\text{-}10)$$

whence

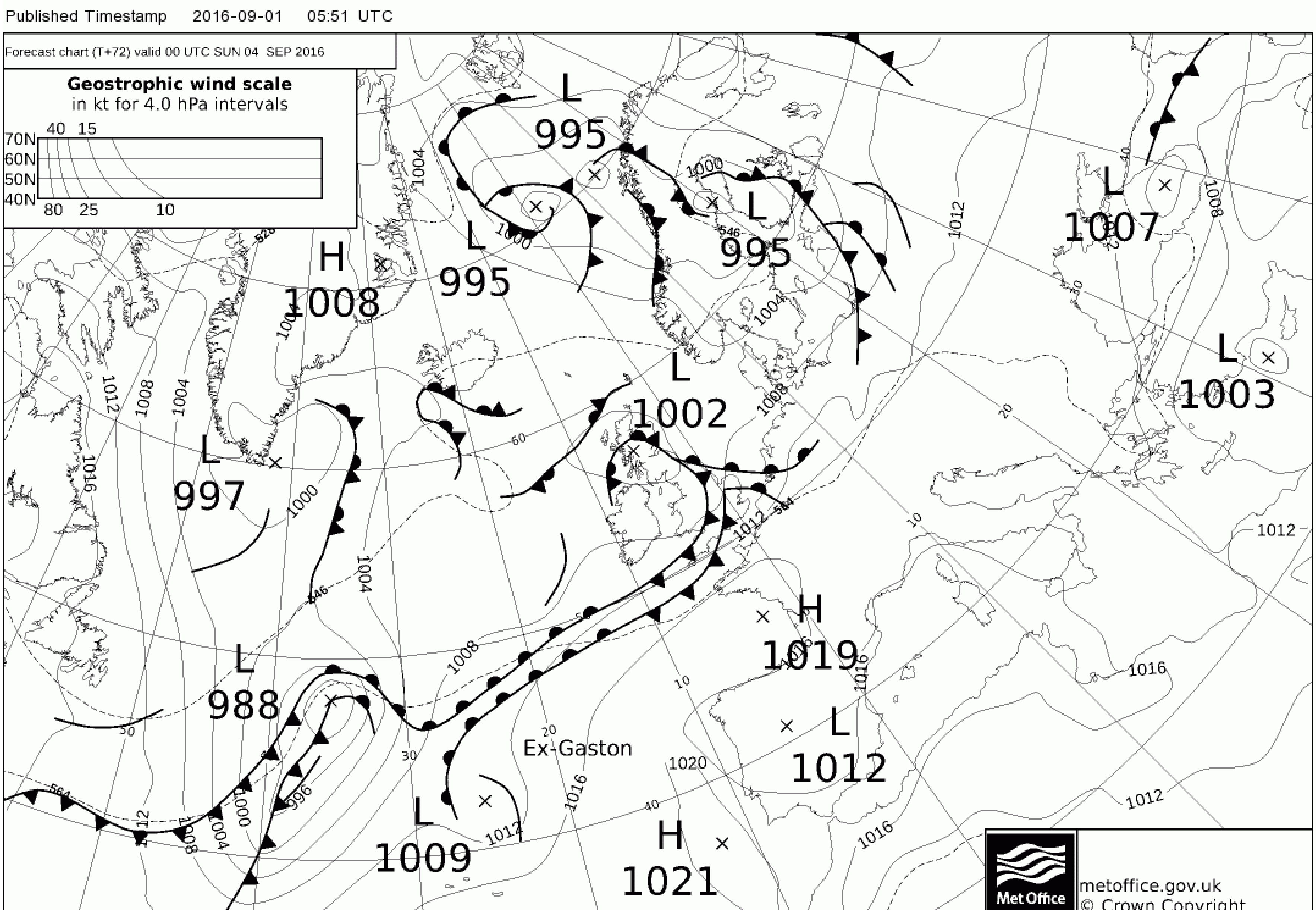
$$\ln p = \int_0^z \frac{dz'}{H(z')} + \text{constant},$$

or

$$p(z) = p_s \exp\left(-\int_0^z \frac{dz'}{H(z')}\right). \qquad (3\text{-}11)$$

---------------------------- Tykkelse 500-1000 hPa gitt i dekameter.

Kald luft ➔ mindre tykkelse



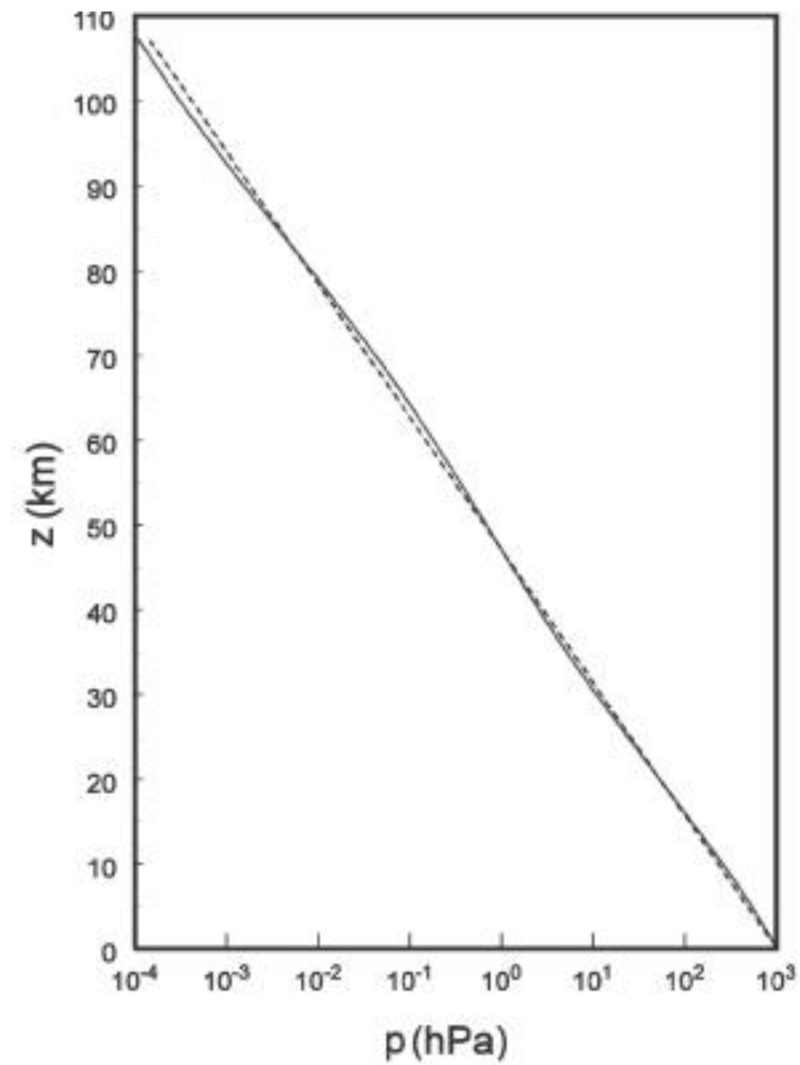


Figure 3.6: Observed profile of pressure (solid) plotted against a theoretical profile (dashed) based on Eq. 3-7 with $H = 6.8$ km.

Ref profile:
T_0=237.08 K
H=6.80 km

3.3.3. Density

For the isothermal case, the density profile follows trivially from Eq. 3-7, by combining it with the gas law Eq. 1.1:

$$\rho(z) = \frac{p_s}{RT_0} \exp\left(-\frac{z}{H}\right) . \tag{3-12}$$

For the nonisothermal atmosphere with temperature $T(z)$, it follows from Eq. 3-11 and the equation of state that

$$\rho(z) = \frac{p_s}{RT(z)} \exp\left(-\int_0^z \frac{dz'}{H(z')}\right) . \tag{3-13}$$