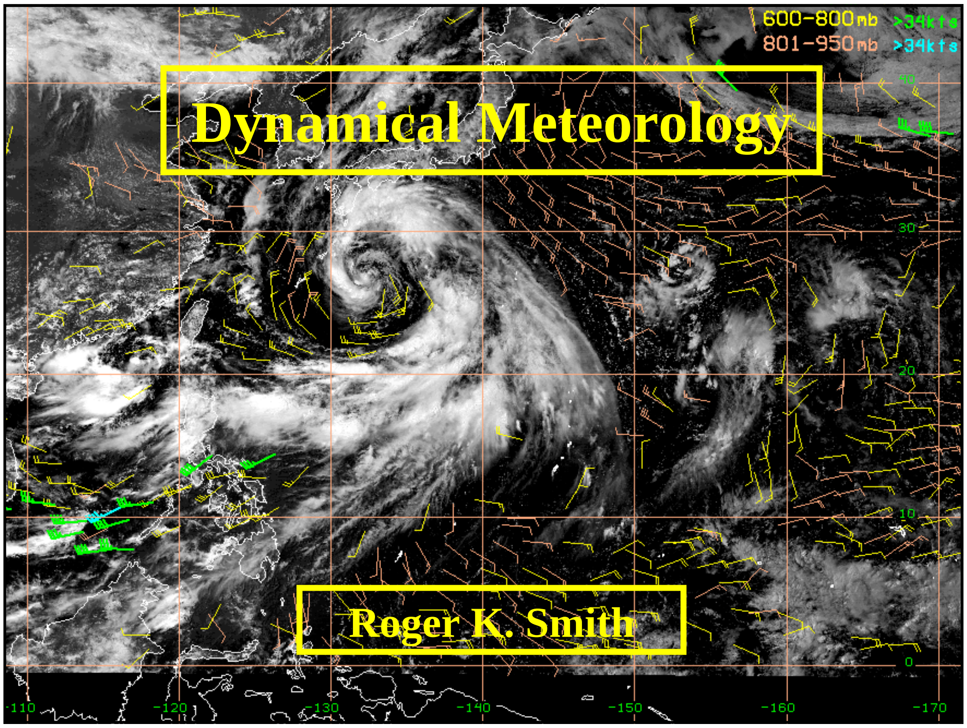

Skript - auf englisch!

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- CHAPTER 12 Gravity currents, bores and flow over obstacles
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Physical sciences

physics
chemistry
biology

Environmental sciences

meteorology
oceanography
geology

Can often isolate phenomena and study under controlled conditions in the laboratory

Cannot control or isolate phenomena

Meteorological and oceanographical analyses tend to be concerned with the assimilation of a body of data rather than with the proof of specific laws.

Dynamical Meteorology

- Dynamical meteorology concerns itself with the theoretical study of **atmospheric motion**.
- It aims to provide an understanding of such motion as well as a rational basis for the prediction of atmospheric events, including short and medium weather prediction and the forecasting of climate.

The Atmosphere

- The atmosphere is an extremely complex system involving motions on a very wide range of space and time scales.
- The dynamic and thermodynamic equations which describe the motions are too general to be easily solved.
- They have solutions representing phenomena that may not be of interest in the study of a particular problem.

Scaling

- We attempt to reduce the complexity of the equations by scaling.
- We try to retain a reasonably accurate description of motion on certain temporal and spatial scales.
- First, we need to identify the essential physical aspects of the motion we hope to study.

Equilibrium and stability

Hydrostatic equation for a fluid at rest

$$\frac{\partial p}{\partial z} = -g\rho \quad \Rightarrow \quad p(z) = \int_z^\infty g\rho(z')dz'$$

p is the pressure
ρ is the density
z is the height
g is the acceleration due to gravity

Pressure as a function of height

The perfect gas equation is $p = \rho RT$

$$\Rightarrow \quad \frac{\partial p}{\partial z} = -\frac{gp}{RT}$$

$$\Rightarrow \quad p(z) = p_s \exp\left[-\int_0^z \frac{dz'}{H(z')}\right]$$

The surface pressure: $p_s = p(0)$. $H(z) = RT(z)/g$

REMEMBER, T is the absolute temperature
for moist air it is the virtual temperature in K.

Potential temperature

The potential temperature θ**, is the temperature a parcel of air would have if brought adiabatically to a pressure of** 1000 mb.

$$\theta = T\left[\frac{1000}{p}\right]^{\kappa}$$

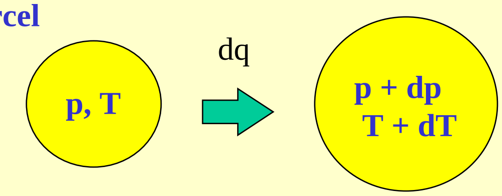
p **in millibars**

T **in K**

$\kappa = R/c_p = 0.286$

Derivation

Add heat dq **to an air parcel**



First law of thermodynamics:

$$dq = c_p dT - \alpha dp$$

$$\alpha = 1/\rho = RT/p$$

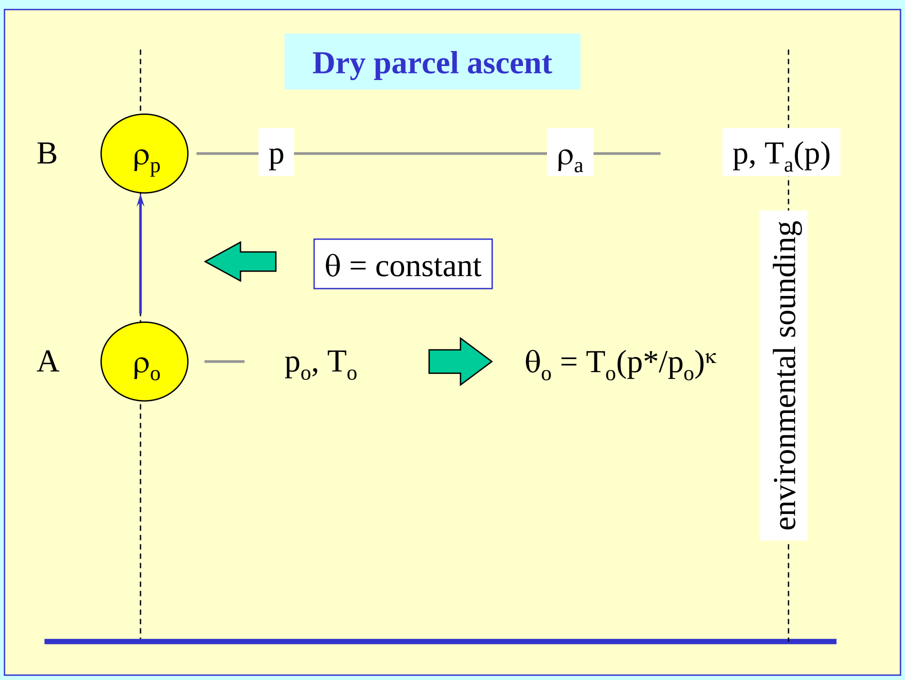
$$\frac{dq}{T} = c_p\left(\frac{dT}{T} - \frac{R}{c_p}\frac{dp}{p}\right) = c_p d(\ln T - \kappa \ln p)$$

Adiabatic process dq = 0 $\Rightarrow$ $\ln T - \kappa \ln p = \text{constant}$

Let $\text{constant} = \ln \theta - \kappa \ln p^*$

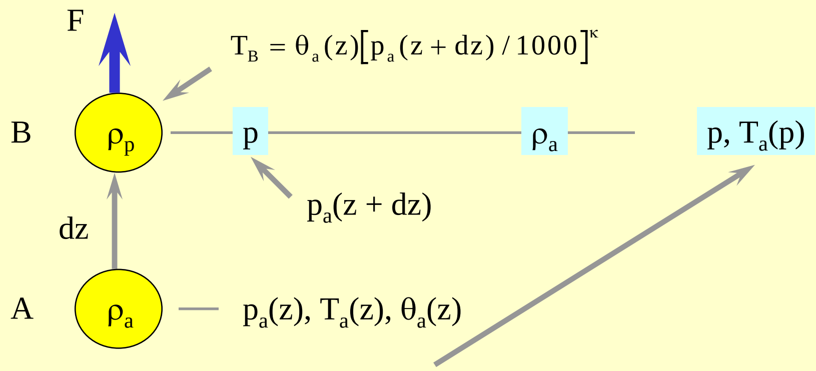
Applications

- A wide range of atmospheric motions are **adiabatic.**
- For such motions the potential temperature is conserved **following air parcels, even when** the parcel of air which experiences a pressure change due to vertical motion.
- The temperature is **not** conserved when the parcel ascends.
- The potential temperature is a fundamental quantity for characterizing the stability of a layer of air.



Parcel buoyancy at level B

$$T_B = \theta_a(z)\left[p_a(z+dz)/1000\right]^{\kappa}$$



$$T_a(z+dz) = \theta_a(z+dz)\left[p_a(z+dz)/1000\right]^{\kappa}$$

Buoyancy force

$$F = \frac{\text{(weight of air displaced - weight of air in parcel)}}{\text{mass of air in parcel}},$$

$$= (g\rho(z+dz)V - g\rho_B V)/\rho_B V,$$

Buoyancy force

$$F = g\left[\frac{\rho_a(z+dz)V - \rho_B V}{\rho_B V}\right] \quad \longleftarrow \quad \rho = p_a(z+dz)/RT$$

$$= g\left[\frac{T_B - T_a(z+dz)}{T_a(z+dz)}\right] \quad \longleftarrow \quad T = \theta[p_a(z+dz)/1000]^{\kappa}$$

$$= g\left[\frac{\theta_a(z) - \theta_a(z+dz)}{\theta_a(z+dz)}\right] \quad \longleftarrow \quad \theta_B = \theta_a(z)$$

$$F \approx -\frac{g}{\theta_a}\frac{d\theta_a}{dz}dz = -N^2 dz$$

N **is the Brunt-Väisälä frequency or buoyancy frequency**

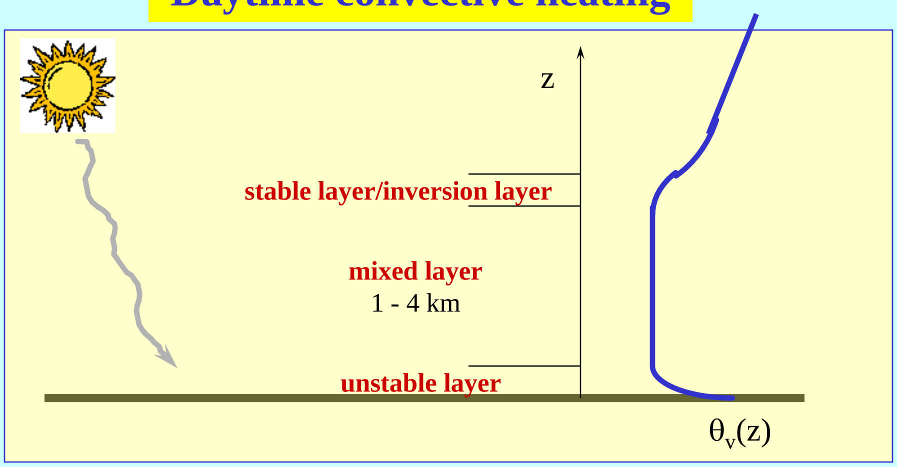
Parcel stability

$$F = -N^2 dz \qquad N^2 = \frac{g}{\theta_a}\frac{d\theta_a}{dz}$$

Three cases:

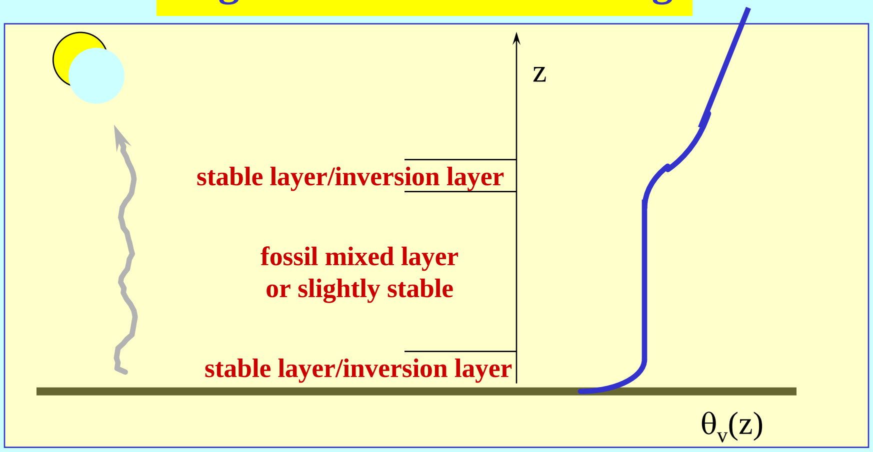
- **Neutral stability** $N^2 = 0 \Rightarrow \theta_a$ **uniform with height.**
- **Stable displacement** $N^2 > 0 \Rightarrow \theta_a$ **increases with height.**
- **Unstable displacement** $N^2 < 0 \Rightarrow \theta_a$ **decreases with height.**
- **These cases apply also to layers of air**
- **Substantial unstable layers are never observed in the atmosphere because even a slight degree of instability results in convective overturning until the layer becomes neutral.**

Daytime convective heating



When the ground is heated by solar radiation, the air layers near the ground are constantly being overturned by convection to give a neutrally stable layer with a uniform (virtual) potential temperature.

Nighttime radiative cooling



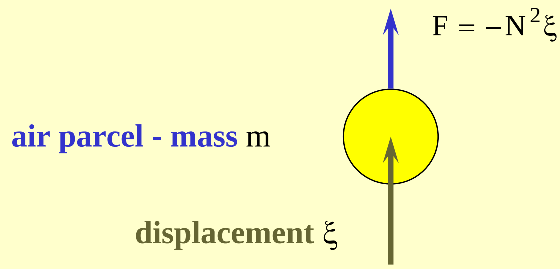
At night, if the wind is not too strong, and especially if there is a clear sky, a strong **radiation inversion** forms in the lowest layers.

An **inversion** is one in which not only θ, but also T increases with height; such a layer is very stable.

The temperature lapse rate

- The **lapse rate** Γ is defined as the rate of decrease of temperature with height, −dT/dz.
- The lapse rate in a **neutrally stable layer** is a constant, equal to about 10 K km, or 1 K per 100m; this is called the **dry adiabatic lapse rate (dalr)**.
- It is also the rate at which a parcel of dry air cools (warms) as it rises (subsides) adiabatically in the atmosphere.
- If a rising air parcel becomes **saturated** at some level, the subsequent rate at which it cools is less than the dry adiabatic lapse rate as condensation leads to latent heat release.

Interpretation of the Brunt-Väisälä frequency



Newton's second law ⇒ $$m\frac{d^2\xi}{dt^2} = -mN^2\xi$$

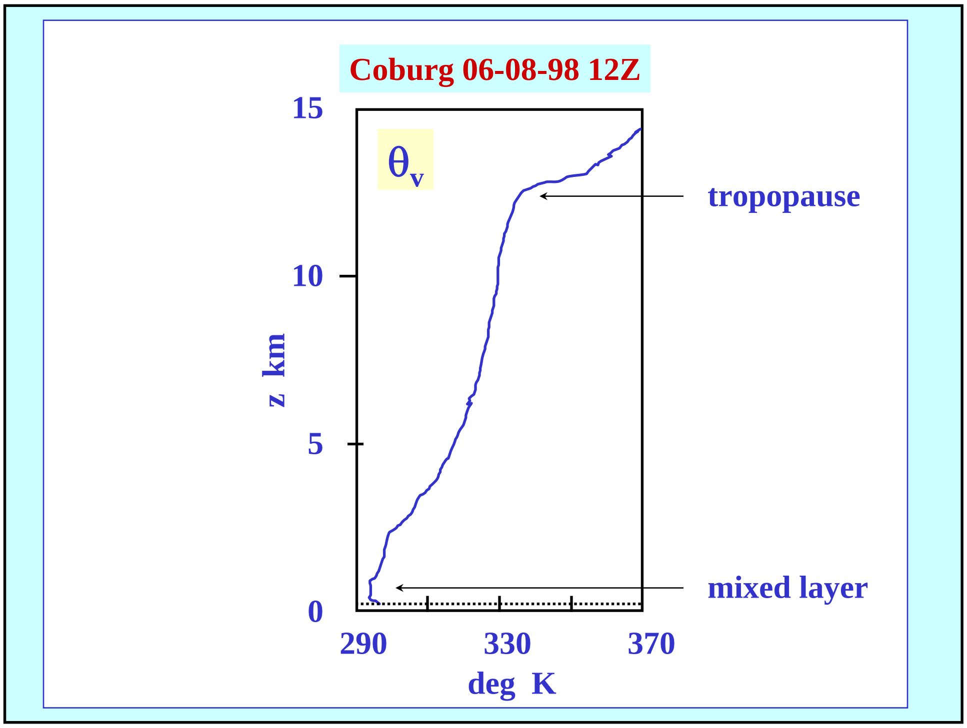
⇒ $$\frac{d^2\xi}{dt^2} + N^2\xi = 0$$

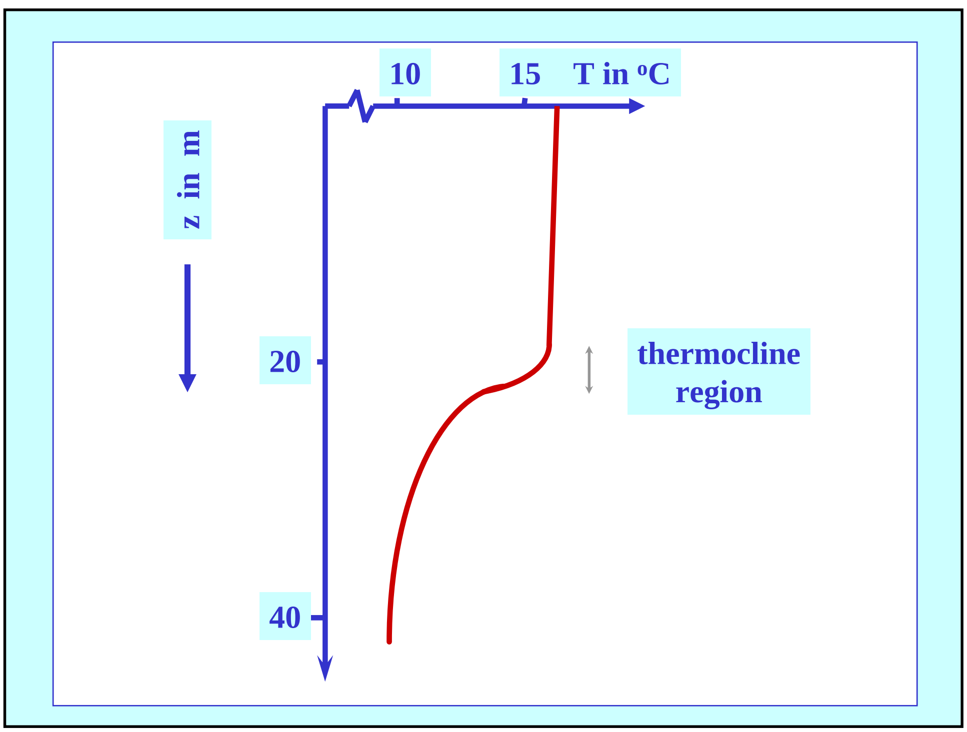
⇒ **motion is simple harmonic (frequency** N **, period** $2\pi/N$ **)**

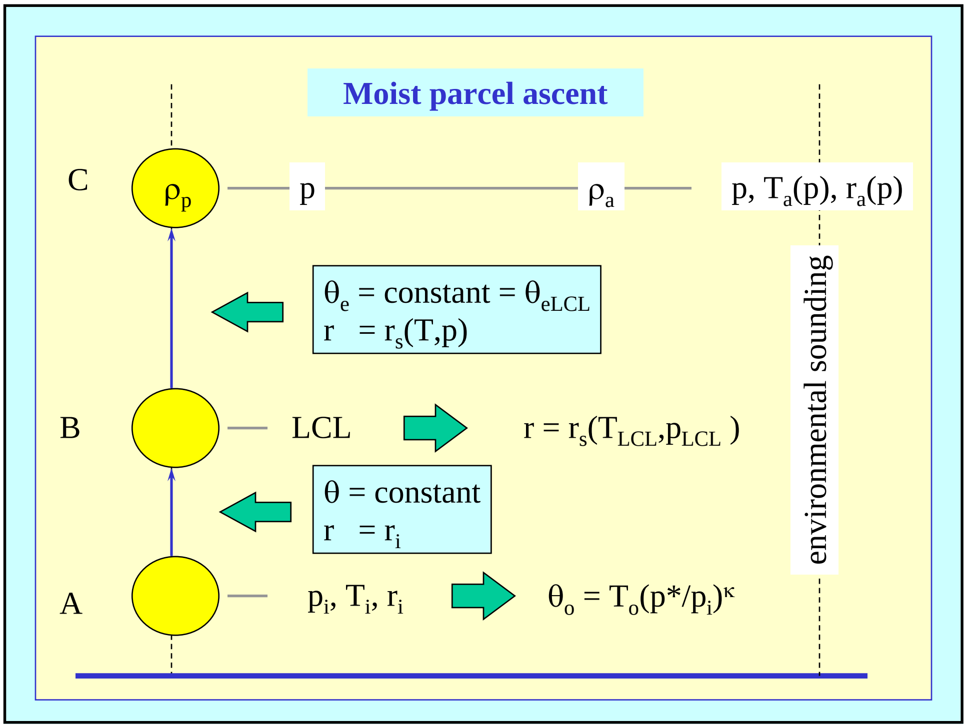
- N **is a key parameter in the theory of gravity waves in the atmosphere.**
- **Since for a fixed displacement, the restoring force increases with** N**, the latter quantity can be used as a measure of the degree of stability in an atmospheric layer.**
- **For an unstable layer,** N **is imaginary and instability is reflected in the existence of an exponentially growing solution to the displacement equation for a parcel.**

$$\frac{d^2\xi}{dt^2} - |N^2|\xi = 0$$

⇒ $$\xi \propto \exp(\pm|N|t)$$


Coburg 06-08-98 12Z
15
10
5
0
z km
θv
tropopause
mixed layer
290
330
370
deg K


10
15
T in °C
z in m
20
40
thermocline region



End of Chapters 1 & 2