Clouds and Aerosols

Cloud nomenclature
Cloud climate impact
Hydro-cycle
Cloud ascension
Cloud characteristics

Luke Howard (1772-1864)
“Father of Meteorology”
Cloud classification 1802

Cumulo-stratus
Cirro-stratus
Cumulus
Cirrus
Stratus
(a) **Cirrus.** This is the most elevated of all the forms of clouds; is thin, long-drawn, sometimes looking like carded wool or hair, sometimes in curl-like or fleece-like patches.

(b) **Cumulus.** This form appears in large masses of a hemispherical form above, but flat below, one often piled above another, forming great clouds, common in the summer.

(c) **Stratus.** This form appears in layers or bands extending along the horizon.

(d) **Cirro-cumulus.** This form consists, like the cirrus, of thin, broken, fleece-like clouds, but the parts are more or less rounded and regularly grouped. It is often called mackerel sky.

(a) **Cirro-stratus.** In this form the patches of cirrus coalesce in long strata, between cirrus and stratus.

(b) **Cumulo-stratus.** A form between cumulus and stratus, often assuming at the horizon a black or bluish tint characterized by rolls in layers or patches, being larger and darker than those of cirro-cumulus and smaller than those of strato-cumulus.
Nimbus. This form is characterized by its uniform gray tint and ragged edges; it covers the sky in seasons of continued rain, as in easterly storms, and is the proper rain cloud. The name is sometimes used to denote a raining cumulus, or cumulo-stratus.
The physics of cloud formation

Clouds are one of the most important components for climate, they change the albedo by reflecting more of the incoming solar flux back into the outer atmosphere (gray nasty day) and they also absorb large fractions of the radiated heat by trapping infrared radiation from earth in the troposphere! They change the lapse rate by adding a humidity component to the air.

How do clouds form and grow?

- Evaporation of water
- Ascension of water vapor
- Condensation of vapor
- Precipitation of water

Atmospheric hydro-cycle
Clouds are critical for the earth’s albedo, but also for the earth’s water cycle.

Total volume of water on earth $1.384 \cdot 10^9 \text{ km}^3 = 1.384 \cdot 10^{18} \text{ m}^3$

- 97% is contained in oceans
- 3% is contained in atmosphere or on land
- 2.25% is retained in polar and glacier ice
- 0.715% is freshwater in rivers and lakes
- 0.035% is atmospheric water vapor ($\approx 5 \cdot 10^{18} \text{ m}^3$)

The balance is maintained by the hydrological cycle of evaporation, condensation, and precipitation of water due to heating and cooling mechanisms.
Troposphere mixing

Microscopic molecular mixing only plays a role in diffusion processes within a few centimeters near ground level and at ~ 100 km high levels above the stratosphere at low pressure and density conditions!

In the intermediate troposphere range mixing occurs through motion of air parcels and the conditions can be approximated by two assumptions:
- No heat exchange with external environment - adiabatic behavior
- Pressure balance with external environment – hydrostatic equilibrium
- Minimum on kinetic energy considerations – total energy conservation

\[- \frac{dT}{dz} = \frac{m \cdot g}{C_p} = \Gamma = 9.8 \frac{K}{km}\]  

Dry adiabatic lapse rate

\[V = \frac{R \cdot T}{P}\]  

Temperature decreases linearly with height and pressure decreases exponentially causing a volume expansion.
Adiabatic behavior of gas (cloud) parcels can be expressed in the framework of new specific quantities such as potential temperature $\theta$ and entropy $S$ that for the parcel that remain altitude independent.

\[
\frac{d\theta}{dz} = 0 \quad \frac{dS}{dz} = 0
\]

First law of thermodynamics: \( C_p \cdot dT = V \cdot dP = R \cdot T \cdot \frac{dP}{P} \)

At adiabatic conditions

\[
\frac{dT}{T} = \frac{R}{C_p} \cdot \frac{dP}{P} \quad \frac{R}{C_p} = \kappa = \frac{2}{7}
\]

for a perfect diatomic gas

with \( \frac{dx}{x} = d(\ln x) \)

\[
d(\ln T) - \kappa \cdot d(\ln P) = 0 \quad \text{or} \quad \frac{T}{P^\kappa} = \text{const.} = \frac{\theta}{P_0^\kappa}
\]

\[
\theta = T \cdot \left( \frac{P_0}{P} \right)^\kappa
\]

Poisson’s equation

Potential temperature $\theta$ is defined as the temperature of a volume element of air moving adiabatically ($\Delta Q=0$) at a standard pressure of 1 bar. While temperature $T$ changes under such displacement, the potential temperature $\theta$ is conserved!
\[ \theta = T \cdot \left( \frac{P_0}{P} \right)^\kappa \]

\[ \frac{d\theta}{\theta} = \frac{dT}{T} - \kappa \cdot \frac{dP}{P} = 0 \]

This results shows that the potential temperature does not change under adiabatic displacement, is therefore independent from altitude!

**Graphical example for the concept:**

A volume element of air at \( P = 300 \text{ mbar} \) has \( T = 239 \text{ K} \) and \( \theta = 334 \text{ K} \). If moved adiabatically down to ground level with pressure \( P = P_0 = 1000 \text{ mbar} \), it would conserve \( \theta \) and change the temperature to \( T = \theta = 335 \text{ K} \).

This graphical approach allows to directly derive the temperature in a cloud element from the pressure conditions at different altitudes.
Meteorological chart
pseudo-adiabatic chart

- Pressure in millibars (horizontal lines),
- Temperature in degrees Celsius (vertical lines),
- Dry adiabats $\theta$ (sloping black lines),
- Lines of constant water vapor or mixing ratio* (solid red lines), and
- Moist adiabats (dashed red lines).

Gas element of 500 mbar and -40 °C corresponds to a dry adiabat of 285 K. Moving a dry gas element adiabatically to a pressure of 900 mbar, follow the dry adiabat $\theta \approx 285$ k to $P=900$ mbar to locate the new temperature in the gas parcel $T \approx 4$ °C = 40 °F.

Nowadays computer approach
\[
\frac{d\theta}{dz} = \theta \cdot \left( \frac{1}{T} \cdot \frac{dT}{dz} - \kappa \cdot \frac{1}{P} \cdot \frac{dP}{dz} \right) = 0
\]

potential temperature variation with height = 0!

\[
\frac{d\theta}{dz} = \frac{\theta}{T} \cdot \left( \frac{dT}{dz} + \Gamma_s \right)
\]

\[\Gamma_s = -\kappa \cdot \frac{T}{P} \cdot \frac{dP}{dz}\]

Lapse rate \(\Gamma\) is always a negative number!

\(\Gamma = -9.8 \text{ K/km}\)

Atmosphere stable against convection:

\[\Gamma_s > \Gamma \text{ or } \frac{d\theta}{dz} > 0\]

For increasing potential temperature

Atmosphere unstable against convection:

\[\Gamma_s < \Gamma \text{ or } \frac{d\theta}{dz} < 0\]

For decreasing potential temperature
Potential temperature and wind development

Equivalent potential temperature (color contours) and storm relative wind vectors at 925 hPa (above) and 600 hPa (below).

\[
\frac{\Delta \theta}{\Delta z} = \frac{320K - 330K}{3km - 1km} = -\frac{10K}{2km} = -5 \frac{K}{km} < 0
\]
1st law of thermodynamics in terms of entropy $S$, which is related to the potential temperature.

\[ dQ = C_p \cdot dT - V \cdot dP \]

\[ dS = \frac{dQ}{T} \]

\[ dS = C_p \cdot \frac{dT}{T} - \frac{V}{T} \cdot dP = C_p \cdot \frac{dT}{T} - R \cdot \frac{dP}{P} \quad \frac{V}{T} = \frac{R}{P} \]

Integrating over entropy yields:

\[ S = C_p \cdot \ln T - R \cdot \ln P + const \]

\[ S = C_p \cdot \ln \theta - R \cdot \ln P_0 + const \]

\[ S = C_p \cdot \ln \theta + const \]

\[ \Delta S = C_p \cdot \Delta(\ln \theta) \]

Entropy scales with logarithm of potential temperature
For a rising gas element such as a cloud:

\[
\frac{dS}{dz} = \frac{C_P}{\theta} \cdot \frac{d\theta}{dz}
\]

\[
\frac{dS}{dz} = \frac{C_P}{T} \left( \frac{dT}{dz} + \Gamma_s \right)
\]

Lapse rate is always a negative number!

Adiabatic lapse rate is always a positive number

Same conclusion as before!

Atmosphere stable against convection: \( \Gamma_s > L \) or \( \frac{dS}{dz} > 0 \) For increasing entropy

Atmosphere unstable against convection: \( \Gamma_s < L \) or \( \frac{dS}{dz} < 0 \) For decreasing entropy
Humidity

Humidity is a measure of moisture in air. The specific humidity $q$ is defined as:

$$q = \frac{\rho_{H_2O}}{\rho}$$

$$\rho = \rho_d + \rho_{H_2O}$$

$\rho \equiv$ density of air;
$\rho_d \equiv$ density of dry air;
$\rho_{H_2O} \equiv$ density of water vapor in air
$\rho_{SV} \equiv$ density of water vapor in saturated air

$$q_{SV} = \frac{\rho_{SV}}{\rho}$$

The saturation specific humidity $q_{SV}$ is the specific humidity, at which saturation of air with water vapor occurs (condensation balances evaporation). The vapor pressure $P_{H_2O}$ is the partial pressure exerted by the water molecules in wet air.
If the air is saturated with water vapor, the vapor pressure is called saturated vapor pressure $SVP$. Saturation and therefore $SVP$ depend on the temperature.

The temperature dependence of the $SVP$ of water is described by the Clausius-Clapeyron equation and empirically expressed as:

$$SVP = A \cdot e^{\beta \cdot T}$$

$$A = 6.11 \text{ mbar}$$

$$\beta = 0.067 \degree C^{-1}$$

Alternative empirical formulas exist in abundance in literature and on the web!
Humidity level increases towards tropical regions and accumulates more at lower altitudes.