Lecture 2: Cloud formation and Physics

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Lectures Outline:

**Cloud fundamentals** - global distribution, types, visualization and link with large scale circulation

**Cloud Formation and Physics** - thermodynamics, cloud formation, instability, life cycle of an individual cloud

**Organization of deep convection at mesoscales** - MCSs, MCCs, Squall lines, Tropical cyclones, Processes, Self-aggregation

**Response of the hydrological cycle to climate change** - mean precip, precip extremes

**Clouds in a changing climate** – climate sensitivity, cloud effect, cloud feedback, FAT
Cloud formation

Courtesy: Octave Tessiot
Atmospheric thermodynamics: instability

Dry convection

T decreases with height.
But p as well.

Density = \( \rho(T,p) \).
How determine stability? The parcel method

Raise parcel adiabatically. Comes back to initial position?
Dry convection

Potential temperature $\theta = T \left( \frac{p_0}{p} \right)^{R/c_p}$ conserved under adiabatic displacements:

First law of thermodynamics: $d(\text{internal energy}) = \Delta Q \ (\text{heat added}) - \Delta W \ (\text{work done by parcel})$

$$c_v \, dT = -p \, d(1/p)$$

Since $p = \rho \, R \, T$,

$$c_v \, dT = -p \, d(R \, T / p) = -R \, dT + R \, T \, dp / p$$

Since $c_v + R = c_p$,

$$c_p \, dT / T = R \, dp / p$$

$$\Rightarrow d \ln T - R / c_p \, d \ln p = d \ln \left( T / p^{R/cp} \right) = 0$$

$$\Rightarrow T / p^{R/cp} = \text{constant}$$

$\Rightarrow \theta = T \left( \frac{p_0}{p} \right)^{R/c_p}$ potential temperature is conserved under adiabatic (reversible) displacement.

Remark 1: ideal gas law: $pV = Nkt \Leftrightarrow p = \rho \, R \, T$, $R = k/m$ where $m=$molecular mass

Remark 2: $c_p = c_v + R > c_v$

Remark 3: We assumed $p_{\text{parcel}} = p_{\text{environment}} \Leftrightarrow$ quasistatic displacement

Remark 4: If we make the hydrostatic approximation, dry static energy $h = c_p \, T + g \, z$ is conserved:

$$c_p \, dT / T = R \, dp / p \Leftrightarrow c_p \, dT = R \, T \, dp / p = -g \, dz \Leftrightarrow c_p \, T + g \, z = \text{constant}$$
Atmospheric thermodynamics: instability

When is an atmosphere unstable to dry convection?
When potential temperature $\theta = T \left( \frac{p_0}{p} \right)^{R/c_p}$ decreases with height!

The parcel method:

Small vertical displacement of a fluid parcel adiabatic ($=> \theta = \text{constant}$).
During movement, pressure of parcel $= \text{pressure of environment}$.

\[
\begin{align*}
\theta_p < \bar{\theta} & \Rightarrow \rho_p > \bar{\rho} & \text{STABLE} \\
\theta_p = \bar{\theta} & \Rightarrow \rho_p = \bar{\rho} & \text{NEUTRAL} \\
\theta_p > \bar{\theta} & \Rightarrow \rho_p < \bar{\rho} & \text{UNSTABLE}
\end{align*}
\]
Atmospheric thermodynamics: instability

Convective adjustment time scales is very fast (minutes for dry convection) compared to destabilizing factors (surface warming, atmospheric radiative cooling…)

=> The observed state is very close to convective neutrality

Dry convective boundary layer over daytime desert <1km

[Renno and Williams, 1995]
Atmospheric thermodynamics: instability

Figure 4  Cartoon of well-mixed, nonprecipitating, stratocumulus layer, overlaid with data from research flight 1 of DYCOMS-II. Plotted are the full range, middle quartile, and mean of $\theta_l$, $q_l$, and $q_l$ from all the data over the target region binned in 30-m intervals. Heights of cloud base and top are indicated, as are mixed layer values and values just above the top of the boundary layer of various thermodynamic quantities. The adiabatic liquid water content is indicated by the dash-dot line.
inversion

Fig. 3.15  Looking down onto widespread haze over southern Africa during the biomass-burning season. The haze is confined below a temperature inversion. Above the inversion, the air is remarkably clean and the visibility is excellent. (Photo: P. V. Hobbs.)
We saw that on a dry adiabat, the potential temperature $\theta = T \left( \frac{p_0}{p} \right)^{R/c_p}$ is constant. If in addition we make the hydrostatic approximation, we can deduce the dry adiabatic lapse rate $\Gamma_d = - \frac{dT}{dz}$.

Recall in that case $c_p T + g z = \text{constant}$

$\Leftrightarrow \frac{dT}{dz} = - \frac{g}{c_p}$

$\Leftrightarrow \Gamma_d = \frac{g}{c_p}$
Atmospheric thermodynamics: instability

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Dry convective boundary layer over daytime desert

<1km

[Renno and Williams, 1995]

But above a thin boundary layer, not true anymore that $\theta = \text{constant}$. Why?…

Most atmospheric convection involves phase change of water
Significant latent heat with phase changes of water = Moist Convection
Atmospheric thermodynamics: instability

Moist variables:

\[ \rho_v = \frac{M_v}{V} \text{ water vapor density} \]
\[ \rho_d = \frac{M_d}{V} \text{ dry air density} \]
\[ \rho = \rho_v + \rho_d \text{ total air density} \]

\[ q_v = \frac{\rho_v}{\rho} \text{ water vapor specific humidity} \]
\[ r = \frac{\rho_v}{\rho_d} \text{ water vapor mixing ratio} \]

\[ e = \text{partial pressure of water vapor} = \rho_v R_v T \text{ (ideal gaz law for water vapor)} \]
\[ p_d = \text{partial pressure of dry air} = \rho_d R_d T \text{ (ideal gaz law for dry air)} \]
\[ p = p_d + e \text{ total pressure (Dalton’s law)} \]

\( T_d \) = dew point temperature : \( T \) at which a parcel must be cooled at constant pressure to reach saturation

\( T_v \) = virtual temperature : \( T \) that dry air would have to have the same density as moist air at same pressure

**Question 1 :** Is moist air lighter or heavier than dry air? In other words is \( T_v \) greater or smaller than \( T \)?

**Question 2 :** Express \( T_v \) as a function of \( T \), \( q_v \) and \( R_d/R_v = \epsilon \)
Atmospheric thermodynamics: instability

**Question 1:** Is moist air lighter or heavier than dry air? In other words is $T_v$ greater or smaller than $T$?

Let’s consider a volume of air $V$ at pressure $p$ and temperature $T$. The ideal gas law implies that $pV=NkT$ where $N$ is the number of molecules in $V$. So regardless of whether the air is moist or not, the number of molecules is the same. In other words, moist air is not formed by adding water molecules to the air, but by replacing dry air molecules with water molecules.

If we compare the molecular masses of $H_2O$, $N_2$ and $O_2$, clearly $H_2O$ is the lightest:

$$m_{H_2O} = (2+16)m_H; m_{O_2} = 2*16m_H; m_{N2} = 2*14m_H.$$  

So moist air is lighter than dry air. Hence $T_v > T$.

In fact the ratio of molecular masses is $m_v/m_d \sim .622 = \varepsilon$

In atmospheric applications we use the ideal gas law with density:

$$pV = NkT \iff p = (N \text{ m / V}) \text{ (k/m)} \ T = \rho \ R \ T$$

**Question 2:** Express $T_v$ as a function of $T$, $q_v$ and $R_d/R_v = \varepsilon$

By definition, $T_v$ satisfies $p= \rho \ R_d \ T_v$ with $p = e + \rho_d = (\rho_v \ R_v + \rho_d \ R_d) \ T$.

So $T_v = \{ \rho_v/\rho \ R_v /R_d + (\rho - \rho_v) /\rho \} \ T = \{ q_v / \varepsilon +1- q_v \} \ T$

$\Rightarrow T_v = \{1 + (1/ \varepsilon - 1) q_v \} \ T \sim (1 + .61 \ q_v) \ T > T$ as expected.
Atmospheric thermodynamics: instability

**Clausius Clapeyron**

\[
\frac{de_s}{dT} = \frac{L_v(T)e_s}{R_v T^2}
\]

- \(e_s\) is saturation vapor pressure,
- \(T\) is a temperature,
- \(L_v\) is the specific latent heat of evaporation,
- \(R_v\) is water vapor gas constant.

\(e_s(T)\)

- \(e_s\) depends only on temperature
- \(e_s\) increases roughly exponentially with \(T\)

**Saturation water vapor amount increases with temperature**

**RH = relative humidity = \(e/e_s\)**

*Remark on « sponge theory »:*

« Warm air can hold more water vapor than cold air » … Has nothing to do with air, similar in other gaz! CC ⇔ Equilibrium between condensation and evaporation
Atmospheric thermodynamics: instability

When is an atmosphere unstable to moist convection?

Equivalent potential temperature $\theta_e = T \frac{p_0}{p} e^{\frac{L_v q_v}{(c_p T)}}$ is approximately conserved under adiabatic displacements:

1st law thermodynamics if air saturated ($q_v=q_s$):

\[
\text{d(Internal energy)} = Q \text{ (latent heat)} - W \text{ (work done by parcel)}
\]

\[
c_v \text{dT} = -L_v dq_s - p \text{d}(1/p)
\]

\[
\Rightarrow d \ln T - \frac{R}{c_p} d \ln p = d \ln \left(\frac{T}{p^{R/cp}}\right) = -L_v \left(\frac{1}{c_p T}\right) dq_s
\]

\[
\Rightarrow T \frac{R/cp}{p} e^{\frac{L_v q_s}{(c_p T)}} \sim \text{constant}
\]

Note: Air saturated => $q_v=q_s$
Air unsaturated => $q_v$ conserved

Hence

\[
\theta_e = T \frac{p_0}{p} e^{\frac{L_v q_v}{(c_p T)}} \text{ equivalent potential temperature is conserved}
\]

Remark: If we make the hydrostatic approximation, MOIST STATIC ENERGY $h= c_p T + g z + L_v q_v$ is conserved:

\[
c_p dT - R T dp / p = -L_v dq_s \Leftrightarrow c_p dT + g d z = -L_v dq_s \Leftrightarrow c_p T + g z + L_v q_v = \text{constant}
\]
Atmospheric thermodynamics: instability

When is an atmosphere unstable to moist convection?

Skew T diagram (isoT slanted), atmospheric T in red

EL equilibrium level

Moist adiabat $\theta_e = \text{cstt}$

Dry adiabat $\theta = \text{cstt}$

LFC level of free convection (= LCL lifted condensation level for simplicity)

CAPE: convective available potential energy
Atmospheric thermodynamics: instability

Moist convection
Parcel = yellow dot

EL equilibrium level

LFC level of free convection

CAPE: convective available potential energy
Atmospheric thermodynamics: instability

If enough atmospheric instability present, cumulus clouds are capable of producing serious storms!!!

Strong updrafts develop in the cumulus cloud => mature, deep cumulonimbus cloud. Associated with heavy rain, lightning and thunder.
Atmospheric thermodynamics: instability

Note that thunderstorms can be:

- single-cell (typically with weak wind shear)

- multi-cell (composed of multiple cells, each being at a different stage in the life cycle of a thunderstorm.

- or supercell, characterized by the presence of a deep, rotating updraft

  Typically occur in a significant vertically-sheared environment
How do those physical considerations explain cloud formation?

=> FOR DEEP CLOUDS:
We saw that for deep clouds, adiabatic ascent from an unstable BL parcel (warm and/or moist) rising through an unstable atmospheric T profile can lead to strong deep convection. Other lifting mechanisms?
Cloud formation: Deep clouds

Other lifting mechanisms:

- orography
- large-scale convergence
- fronts

=> All force ascent, and leads to deep convection if atmosphere above is unstable

Clouds associated with a frontal system

(\textit{blue}: cold front, steep and fast; \textit{red}: warm front, shallower and slower)
Cloud formation: Shallow layer clouds

SHALLOW LAYER CLOUDS

- **Fog and stratus**: in BL cooled from below, by radiation or conduction from cold surface
  => **Stable BL**, reach saturation by cooling

- **Stratus or stratocumulus or shallow cumulus**: in BL heated from below
  => **Unstable BL**, with a stable atmosphere above. Also radiative cooling at the top of the cloud layer destabilizes the layer and contributes to the convection.

When do we have unstable layer capped by stable layer? **Warm air above cold air** "T inversion"

An inversion can develop aloft as a result of air gradually sinking over a wide area and being warmed by adiabatic compression, e.g. associated with subtropical high-pressure areas.
Cloud formation: Shallow layer clouds

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  e.g. - subtropical latitudes west of continents stratus and stratocumulus associated with anticyclones around high pressure
  - middle and high latitudes cold air offshore across the coastlines of cold continents or ice sheets, over warm ocean => stratocumulus
Cloud formation: Shallow layer clouds

MESOSCALE ORGANIZATION OF THOSE CLOUDS: Open and closed cells

Not all is known, but some processes appear to play a role: shear, thermal instabilities of the BL, cloud-top entrainment and precipitation-driven cold pools (aerosols?).

Mesoscale cellularity in marine stratocumulus clouds. This MODIS image (approximately 800 km across) shows the sharp transitions that occur between the closed and open cells.
Cloud formation: Shallow layer clouds

Mesoscale organization of shallow clouds: Open and closed cells shallow convection

Figure 5.24 Global climatology of cellular structure of stratocumulus and small cumulus over oceans. Shaded areas are regions where closed cells predominate. Hatched areas show where open cells are more common. Solid streamlines show locations of warm ocean currents. Dashed streamlines show cold currents. Land masses are blackened. (Adapted from Agee et al., 1973. Reproduced with permission from the American Meteorological Society.)

Houze « Cloud Dynamics »
Cloud formation: Shallow layer clouds

Rayleigh-Benard convection
Gold paint dissolved in acetone. Put it in a shallow dish. Cover it so that the acetone does not evaporate. Let stabilise and then remove the cover. Evaporation of the acetone causes the top layer to cool thus starting convection.
Cloud formation: Shallow layer clouds

SHALLOW LAYER CLOUDS

- **Cirriform clouds:**
  Not much water vapor at those high altitudes => mainly radiation driven. Clouds of (mainly) ice in an **unstable layer between two stable layers**
  SW heating throughout the clouds, while LW cools above and warms below

- Can be detrained from deep convective clouds (most often, consistent with largest cirrus cover in the tropics and in the extratropics where deep convection), or

- Can occur away from generating source when unstable layer aloft

- **Altostratus & altocumulus:** these can be

  - Remnants of other clouds: protruding layers in middle levels due to horizontal wind

  - Altocumulus also sometimes high-based convective clouds => same dynamics as deep convective clouds

  - Altostratus or shallow layer of altocumulus can also resemble a radiatively driven « mixed layer » aloft, leading to a cloud-filled layer radiatively driven at its top (Can lead to rolls in the absence of shear)
Atmospheric thermodynamics: instability

Kelvin Helmholtz Instability: destabilized by shear
Atmospheric thermodynamics: instability

Kelvin Helmholtz Instability: destabilized by shear
Lectures Outline:

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**Cloud Formation and Physics** - thermodynamics, cloud formation, instability, life cycle of an individual cloud

**Organization of deep convection at mesoscales** - MCSs, MCCs, Squall lines, Tropical cyclones, Processes, Self-aggregation

**Response of the hydrological cycle to climate change** - mean precip, precip extremes

**Clouds in a changing climate** – climate sensitivity, cloud effect, cloud feedback, FAT