Pressure decreases exponentially (note logarithmic y-axis) because air is a compressible fluid (i.e. density is variable).

Fig. 1.8 Vertical profiles of pressure in units of hPa, density in units of kg m$^{-3}$, and mean free path (in meters) for the U.S. Standard Atmosphere
Temperature structure more complex than pressure structure

Temperature decreases with altitude in the troposphere because pressure decreases with altitude.

Not ideal gas law:

\[ PV = nRT \]

P decreases, but V increases → how does T change?

T increases with altitude in the stratosphere because of UV absorption by ozone.

Figure by MIT OCW.
Humidity Profiles - Annual

Altitude (km) vs. Specific Humidity (g/kg)

- 10S-10N
- 40-50N
- 70-84N
Potential temperature \( \theta \) is the T of air if it is brought to the ground by the dry adiabatic lapse rate. It can be used to determine the stability of the atmosphere.

Dry Adiabatic Lapse Rate

\[
\Gamma = \frac{g}{c_p}
\]

Earth’s atmosphere:

\[
\Gamma = \frac{1 K}{100 m}
\]

Potential temperature is the T of air if it is brought to the ground by the dry adiabatic lapse rate. It can be used to determine the stability of the atmosphere.
Equilibrium state of the atmosphere and surface in the absence of all energy fluxes other than radiative fluxes.
Problems with Radiative Eq’m

• Too hot in lower troposphere and surface
• Too cold in upper troposphere
• Tropospheric lapse rate too large
  (Stratosphere temperatures pretty good)
Radiative-Convective Eq’m

Account for convection by assuming that convection limits the lapse rate to adiabatic and calculate the new equilibrium (a good assumption for a dry world)
Better, but still too hot at surface, too cold at tropopause
Still missing: moist convection (aka clouds)

Most convection is associated with clouds, which affect energy balance through:

- latent heat release
- changes in planetary albedo
- redistribution of water vapor
Accounting for the wet adiabatic lapse rate, temperature profile is much closer to observed values.

Manabe and Strickler 1964 calculation:
Dashed lines on far left side are reflected sunlight fluxes.

Net longwave radiative cooling is less than half the total cooling.

Latent heat (LH) flux (moist convection) dominant over sensible heat (SH) flux (dry convection)

---

**Fig. 2.4** Radiative and nonradiative energy flow diagram for Earth and its atmosphere. Units are percentages of the global-mean insolation (100 units = 342 W m$^{-2}$).
This curve describes the equilibrium between liquid water and the partial pressure of water vapor.
Classes of typical cloud types.

High clouds 6000 m: Cirrostratus, Cirrocumulus

Middle clouds 4000 m: Altocumulus, Altostratus

Low clouds 2000 m: Nimbostratus, Stratus, Stratocumulus, Cumulus (fair weather), Cumulus

Clouds with vertical development: Cumulonimbus

Earth’s surface
Reflected SW Radiation

Annual Mean Cloud Reflection

- thick frontal
- small cumulus
- deep convective
- small cumulus
- deep convective
- thick frontal
- thick frontal
- small cumulus
- deep convective
- small cumulus
- deep convective
- low-level stratiform

W m^-2

-60 -40 -20 0 20 40 60 80
Large stratocumulus deck off of California and Mexico.

Image courtesy of MODIS.
Stratocumulus deck off the coast of Angola and Namibia.

Image courtesy of MODIS.
Vortex streets occur around the Canary Islands within the stratocumulus deck.

Large scale flow dynamics affects the microphysics!

Image courtesy of MODIS.
Shallow cumulus in Antigua (Caribbean)
Cumulonimbus over Africa – the cirrus shield hides the strong convective clouds
Cumulus clouds with cloud-top temperatures below –5°C are typically mixed-phase, containing both ice crystals and supercooled liquid drops.
Clouds can exist as continental-scale features. Cloud properties are controlled by the interaction between microphysics ($10^{-6}$ m) and dynamics (up to $10^6$ m).
Cloud Drop Formation

Condensation: Conversion of water vapor in a parcel of air to liquid on a CCN after reaching saturation.
It takes 1 MILLION cloud droplets to make 1 rain drop!!

Rain drop

\( D_p = 2000 \, \mu m \)
Collision-Coalescence

Collision: Bumping into each other due to differences in gravitational settling velocities

Coalescence: Merging of smaller droplets with the larger, faster falling collector drops
- Cloud droplet

Rain drop
Heterogeneous deposition
Condensation freezing
(Homogeneous freezing at –40°C)
Contact freezing
Immersion freezing

Ice Nucleation Mechanisms
Relative Humidities of Ice and Liquid Water

<table>
<thead>
<tr>
<th>Temperature</th>
<th>RH wrt H₂O(liq)</th>
<th>RH wrt H₂O(ice)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°C</td>
<td>100%</td>
<td>100%</td>
</tr>
<tr>
<td>-05°C</td>
<td>100%</td>
<td>105%</td>
</tr>
<tr>
<td>-10°C</td>
<td>100%</td>
<td>110%</td>
</tr>
<tr>
<td>-15°C</td>
<td>100%</td>
<td>115%</td>
</tr>
<tr>
<td>-20°C</td>
<td>100%</td>
<td>121%</td>
</tr>
</tbody>
</table>
Ice crystals grow by deposition at the expense of evaporating supercooled liquid drops. These ice crystals then go on to initiate collision/coalescence.
Fig. 6.36  Laboratory demonstration of the growth of an ice crystal at the expense of surrounding supercooled water drops. [Photograph courtesy of Richard L. Pitter.]
Some ice habits

Classification from Magono and Lee, 1966.
<table>
<thead>
<tr>
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</thead>
<tbody>
<tr>
<td>P6b Plate with spatial dendrites</td>
</tr>
<tr>
<td>P6c Stellar crystal with spatial plates</td>
</tr>
<tr>
<td>P6d Stellar crystal with spatial dendrites</td>
</tr>
<tr>
<td>P7a Radiating assemblage of plates</td>
</tr>
<tr>
<td>P7b Radiating assemblage of dendrites</td>
</tr>
<tr>
<td>CP1a Column with plates</td>
</tr>
<tr>
<td>CP1b Column with dendrites</td>
</tr>
<tr>
<td>CP1c Multiple capped column</td>
</tr>
<tr>
<td>CP2a Bullet with plates</td>
</tr>
<tr>
<td>CP2b Bullet with dendrites</td>
</tr>
<tr>
<td>CP3a Stellar crystal with needles</td>
</tr>
<tr>
<td>CP3b Stellar crystal with columns</td>
</tr>
</tbody>
</table>
Zonally-averaged DJF potential temperature:

The atmosphere is, on average, statically stable (with periods of instability – c.f. radiative-convective equilibrium)
Zonally- and annually-averaged mean vertical wind (positive = downwards)
Cloud regimes in thermally direct circulations.
Large stratocumulus deck off of California and Mexico.

Image courtesy of MODIS.
Stratocumulus are forced from the top down – by radiative cooling of the clouds. This makes for air that is unstable (turbulent) below cloud top. This turbulence replenishes the moisture of the cloud (which is slowly lost to the warm air above by entrainment). The moisture is trapped in this thin layer by the warm air above, allowing the clouds to persist. If this lid were not around, then the moisture would be diluted and the cloud less persistent.
Shallow cumulus are forced from the surface – by small differences in temperature.

Plumes of positively buoyant, moist air originate from the surface layer and penetrate into the warmer, subsiding air. These clouds live for short periods (tens of minutes).

The moisture is trapped in this thin layer by the warm air above, allowing the clouds to persist. If this lid were not around, then the moisture would be diluted and the cloud less persistent.
In a conditionally stable atmosphere, there are three important heights:

**LCL** = lifting condensation level  
**LFC** = level of free convection  
**EL** = equilibrium level

Notice the dry adiabat and saturated adiabat. The thick line connecting the closed circles is the temperature sounding (actual temperature profile).

The bottom “negative area” is defined as **CIN** = Convective Inhibition.

The middle “positive area” is defined as **CAPE** = Convective Available Potential Energy.