Condensation in the Atmosphere

- The atmosphere contains a mixture of dry air and a variable amount of water vapor (0-4% or 0-40 g/kg).
- An air parcel is said to be “saturated” when the vapor pressure of the parcel equals the vapor pressure exerted by a plane surface of pure water at the air temperature.
- Another way to state this is that saturation is reached when the flux of water vapor molecules into a plane surface of pure water equals the flux of water molecules escaping the water surface into the air. (What is boiling?)
- If there is no surface of water, vapor content in the air can reach upwards to 500% RH before homogeneous nucleation of water takes place.
- RH never exceeds 100.04% because aerosols always nucleate non-pure water droplets (even with RH as low as 70%), which grow large enough to be a proxy for a plane surface of pure water as $100 + \varepsilon \%$ RH is reached.
Variation of Saturation with Temperature

- Teten’s formula for computing saturation vapor pressure (over plane surface of pure water) is:

\[ e \equiv \text{vapor pressure (hPa)} \]

\[ e_s = 6.1078 \exp \left( \frac{a(T - 273.16)}{T - b} \right) \]

<table>
<thead>
<tr>
<th></th>
<th>water</th>
<th>ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>17.2693882</td>
<td>21.8745584</td>
</tr>
<tr>
<td>b</td>
<td>35.86</td>
<td>7.66</td>
</tr>
</tbody>
</table>

\[ r_{sl} = 0.611 \frac{e_{sl}}{p - e_{sl}}, \quad r_{sl} = 0.611 \frac{e_{sl}}{p - e_{sl}} \]
How does atmosphere form cloud?

\[ e > e_{sl}(T) \]

I. Adiabatic
   A. Increase \( e \)
      a) Evaporation - can only increase RH to 100%! It CANNOT form a cloud by itself!
      \[ \frac{dq_{\text{aq}}}{dt} \propto (e - e_s(T)) \]

   B. Decrease \( T \)
      a) Expansional cooling
      b) Conduction with cold surface in parcel, ie ice hydrometeor
How does atmosphere form cloud?

\[ e > e_{sl}(T) \]

II. Diabatic

A. Increase \( e \)
   a) Evaporation from a water surface - this cannot moisten a parcel to 100% RH because an air parcel must have a temperature only infinitesimally different from the water and therefore can only bring the air in contact with the water to near saturation. Mixing with cooler air above can then bring about saturation, but then saturation was created by mixing, not evaporation.

B. Decrease \( T \)
   a) Radiational Cooling
      i. Tends to occur only at interface between:
         » moist and dry layer
         » between cloud and clear air
   b) Conduction of heat into cold surface (e.g. snow, ice, cold water, cold ground)
      i. Molecular diffusion of heat very inefficient, especially when diffusing a cool layer upward
      ii. Can be enhanced by forced turbulent mixing of thin cold air layer into air above, e.g. winds
   c) Parcel Mixing
      i. Due to curvature of saturation variation with temperature
      ii. Mix two subsaturated parcels to achieve super saturation and the formation of cloud droplets
Cloud formed by breath on cold day
Mixing over a relatively warm lake on a cold day
Fogs

• Fog is a cloud in contact with the ground
• The reasons for fog formation mirror all the ways that saturation can be achieved, i.e.
  – Radiation (radiation fog, ground fog)
  – Conduction (sea, advection fogs)
    • Mixing is still involved
  – Mixing (steam fog, frontal fog, advection fog)
  – Expansional cooling (upslope fog)
Fog Types

- **Advection Fogs**
  - Sea Fog – advection of warm moist air over a cold sea surface leads to mixing of warm moist and conduction cooled air producing saturation and fog
  - Advection of warm moist air over cold land surface leads to mixing of warm moist and conduction cooled air producing saturation and fog (e.g. warm air advection over a snow cover)
  - Land and sea breeze fog
  - Tropical air fog
  - Ice fog
  - Snow fog
Role of Dew

- Cooling of the surface causes moisture of the air in contact with the surface to be deposited as dew.
- This causes a net downward transport of moisture into the ground and the formation of a “dew point inversion”.
- The dew point inversion may inhibit fog formation.
- However, once the sun rises and the surface warms, the dew acts as a reservoir of water to allow fog to persist for several hours.
Role of Droplet Settling

- Small liquid droplets settle very slowly
- Settling depletes liquid water content at top of fog and increases it below
- This weakens radiative divergence at the top
- Hence low CCN contents produce more settling (larger droplets) and lower water contents
Radiation Fog

1. Radiation cools the surface, surface air cools by conduction
2. Radiation divergence across top of moist layer cools the air above, destabilizing air above
3. Static instability at layer top causes turbulence to overturn air, mixing cold air from below, forming saturation
4. Once cloud layer forms, radiational cooling at top of fog layer is greatly enhanced, further increasing overturning and increasing fog water content
Valley Fog

1. Nocturnal radiation cooling along side walls produces sinking motion along sidewalls
2. Dew deposition at the surface creates a dew point inversion at the surface
3. Converging cold and somewhat dry air flows over the valley force upward motion and deepen the inversion
4. About 3 h before fog formation, mountain wind forms, providing continuity for the downslope flow, but restricting upward motion in valley center.
5. Cooling is then capped to low and mid levels of the valley by the strengthening inversion
6. Radiation cooling at the top of the inversion layer leads to the formation of a thin cloud layer
7. The thin cloud layer enhances radiation divergence and deepens to the surface
Marine Fog

• Differs from Radiation fog:
  – Radiation does not rapidly affect surface temperature
  – Less CCN- more drizzle (Giant salt nuclei)
  – Moisture flux up
  – Heat flux down

• Results of model experiments show:
  – Case 1: upward moisture flux, downward heat flux, ie cold water/warm air promotes fog
  – Case 2: upward heat and moisture flux, ie fog if air above is cold and moist
Fog Produced By Marine Stratus Lowering

- Radiational cooling lowers base of stratus cloud:

Fig. 7.10. Schematic representation of fog formation through the stratus-lowering process. [From Pilie et al. (1979).]
Clear California summer afternoon
Strong upper inversion
Avg. rel. humidity 97% in mixed
layer to $z_i$

$z_i =$ inversion height

Fig. 7.11. Typical late afternoon temperature profile measured by Mack et al. (1974). [From Oliver et al. (1978).]
Fig. 7.12. Evolution of fog resulting from the lowering of stratus. Fog shows to form at surface just before downward-propagating stratus reaches the surface et al. (1978).]

Fig. 7.13. Radiative cooling/heating distribution in a stratus-lowering fog. Maximum cooling is concentrated at cloud top. After sunrise, net heating occurs, but deep within the cloud. [From Oliver et al. (1978).]
Fig. 7.16. Schematic representation of the vertical cross section of fog formed as a result of low-level convergence and radiative cooling. [From Pilie et al. (1979).]
Marine Stratocumulus

• Exist over large spans of the eastern Pacific, eastern Atlantic and western Indian Oceans

• These are upwelling regions of cool water so air naturally near saturation in marine PBL

• These are also regions of large scale subsidence aloft
Figure 5.9 Average cloud cover by stratus, stratocumulus, and fog. Data are from standard surface observations and are expressed in percent of sky covered. The years 1952 to 1981 are included in the data set. (a) June, July, August. (b) September, October, November. (Analysis by C. Leovy of data published in the atlas of Warren et al., 1988.)
Fig. 7.17. Temperature, moisture, and wind data from an NCAR Electra sounding at 37.8°N, 125.0°W and between 1522 and 1526 GMT; (a) temperature and dew point; (b) dry static energy, moist static energy, and saturation static energy; (c) wind direction and speed. Dashed lines below 50 m are extrapolations. [From Schubert et al. (1979a).] (Figure continues.)
Fig. 7.17 (continued)
Dynamics of Marine Stratocumulus

• Subsidence Drying Aloft
• Moistening from the cool ocean surface
• Radiation divergence at top of marine PBL
• Entrainment of low theta-e but high theta air from above PBL inversion
Other Factors:

• Drizzle: Weakens radiation divergence
• Shear: enhances entrainment
Fig. 7.31. Diurnal variation of liquid water in stratus cloud formed in a subsidence-capped boundary layer. [From Oliver et al. (1978).]
Fig. 7.32. Distribution of radiative cooling/heating in stratus cloud. Note that net heating occurs deep within the cloud interior. [From Oliver et al. (1978).]
Max turbulence $= 1.21 \text{ m}^2/\text{sec}^2$

Fig. 7.33. Distribution of turbulence in diurnally varying stratus cloud. [From Oliver et al. (1978).]