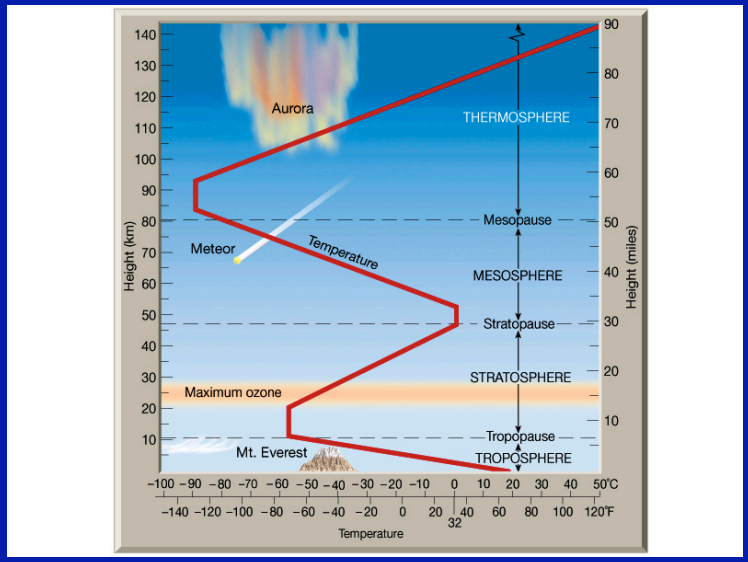
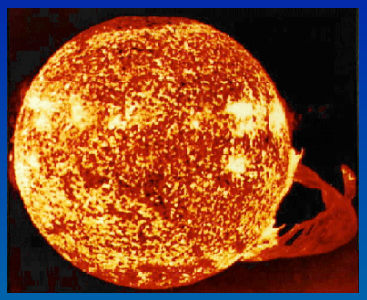
CLIM-111/PHYS-111

Introduction to the Fundamentals of Atmospheric Science

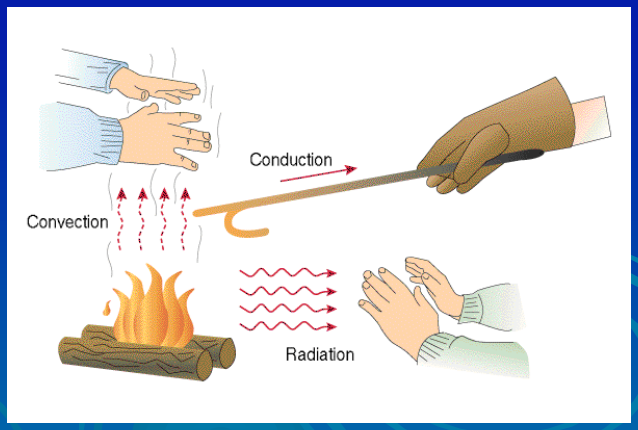
Atmospheric Thermodynamics – Part 2

1. Thermodynamics Review/Tutorial
   - Ideal Gas Law
   - Heat Capacity
   - 1st & 2nd Laws of Thermodynamics
   - Adiabatic Processes
   - Energy Transport
2. Hydrostatic Equilibrium
3. Adiabatic Lapse Rate – DRY
4. Adiabatic Lapse Rate - WET
5. Static Stability
6. SLT and the Atmosphere

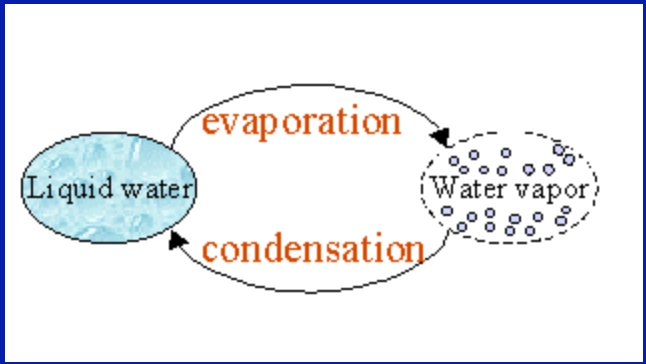




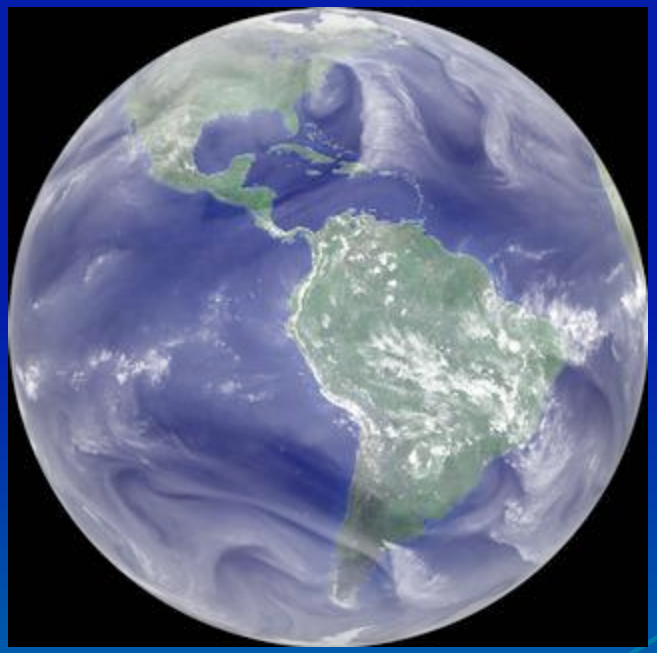




Role of Water in the Atmosphere



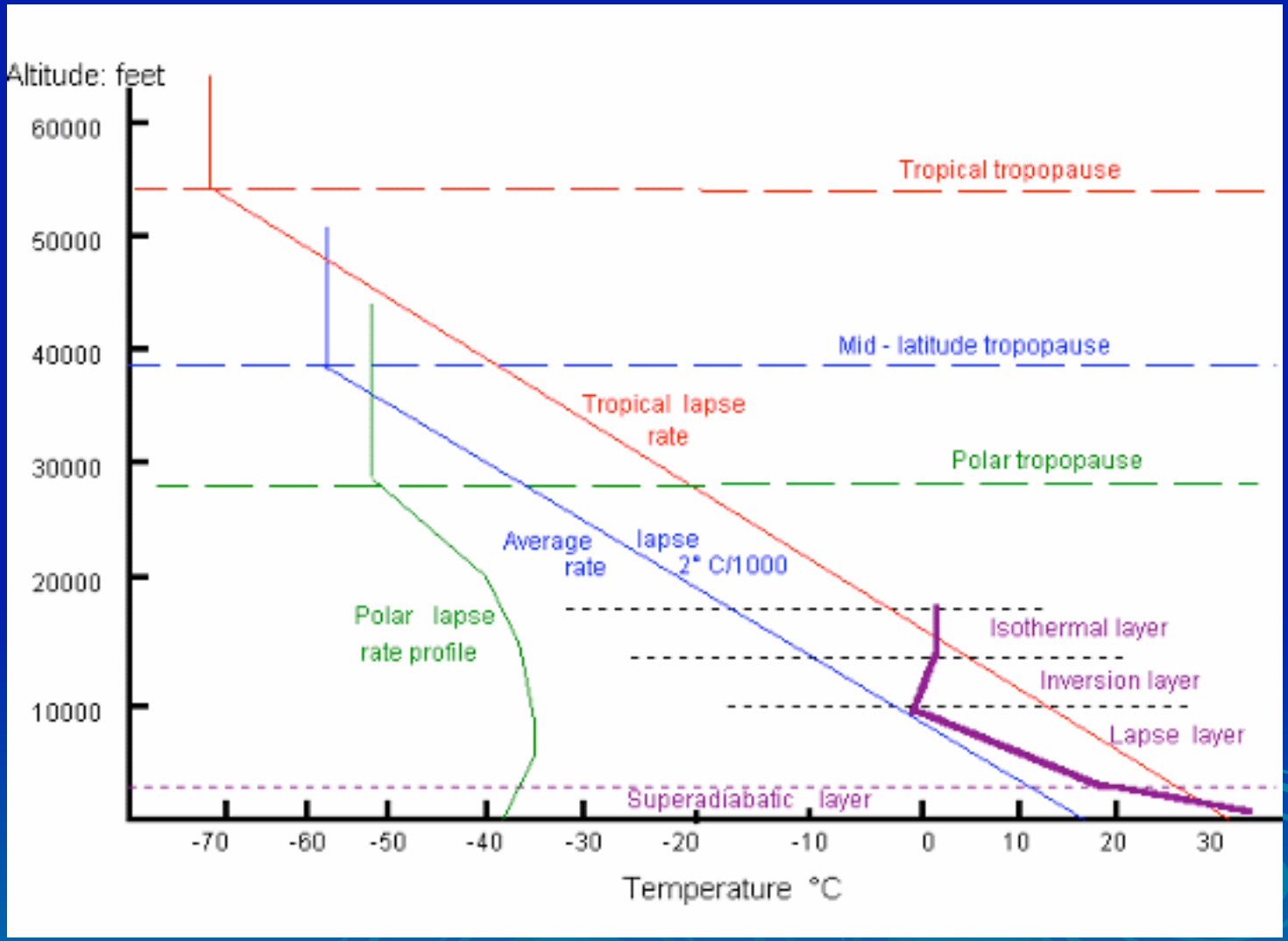






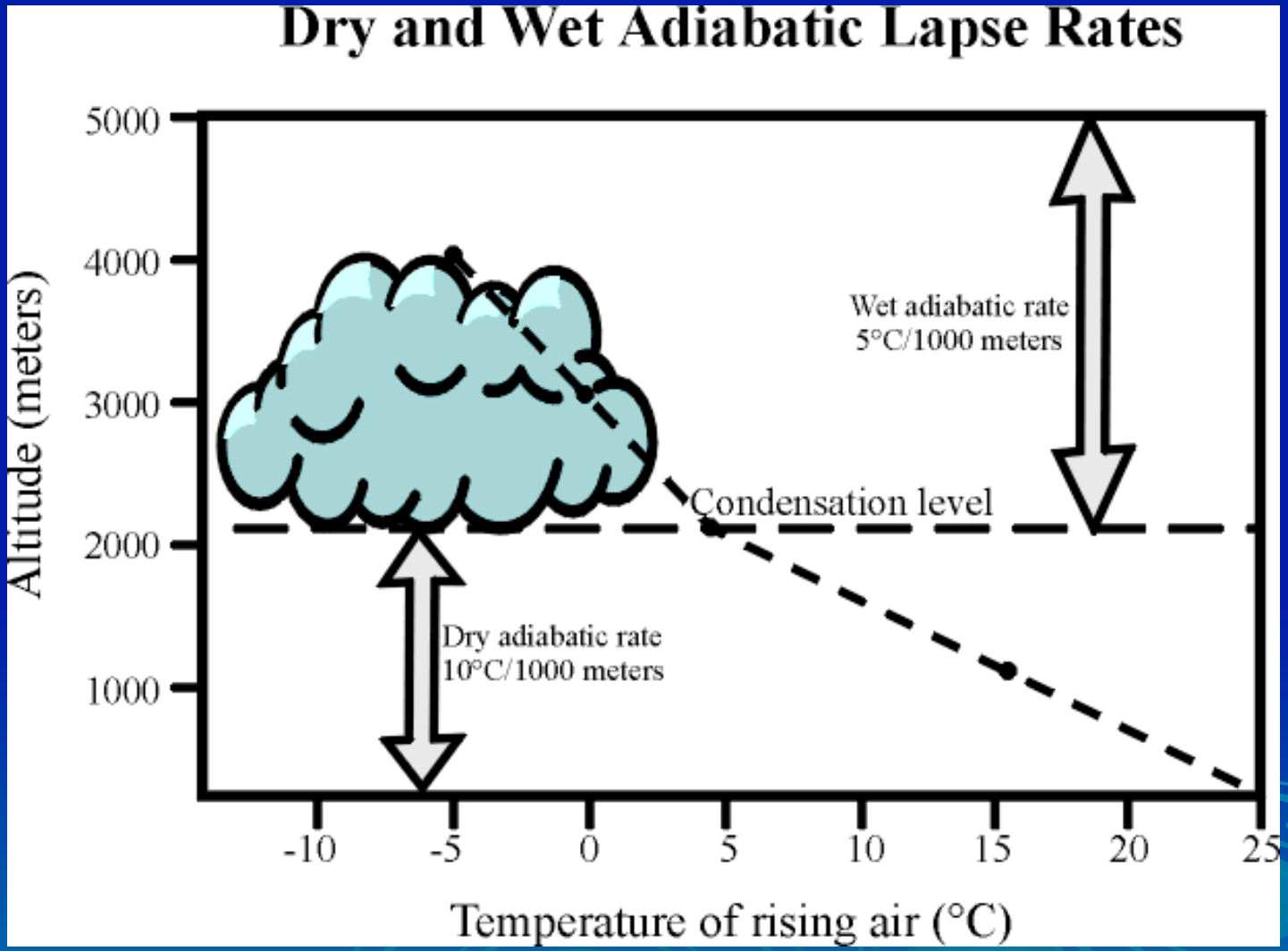


Role of Water Vapor in Atmospheric Thermodynamics of the Troposphere



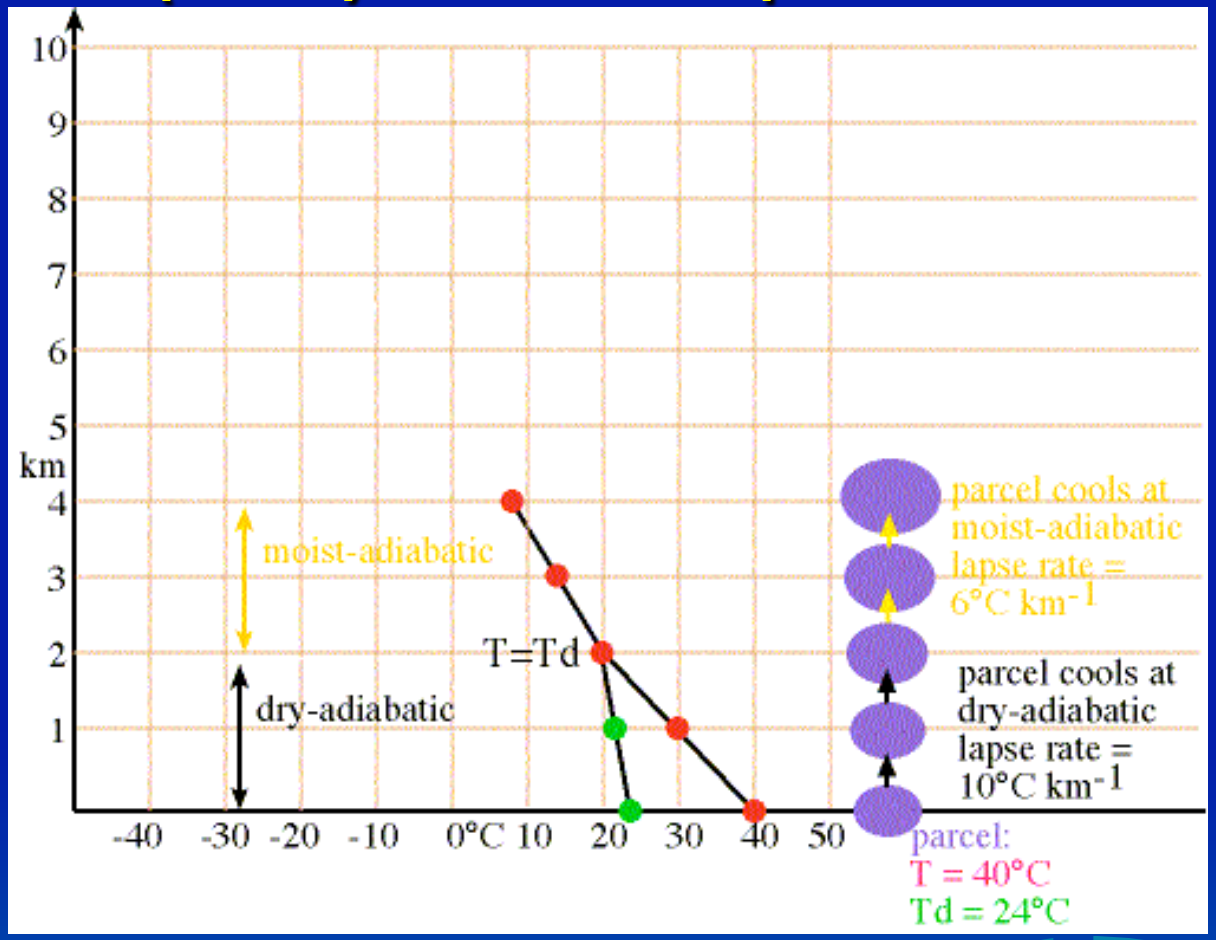
http://www.auf.asn.u/metimages/lapseprofile.gif

4. Adiabatic Lapse Rate - Wet



Water Vapor in the Atmosphere:
The Wet (Moist) Adiabatic Lapse Rate

$\Gamma_d = -g/C_p = DALR$



The Wet Adiabatic Lapse Rate is smaller than the DALR, because the effective heat capacity of a wet atmosphere is larger than that of a dry atmosphere. The phase change of water is an effective heat reservoir.

Saturation Conditions

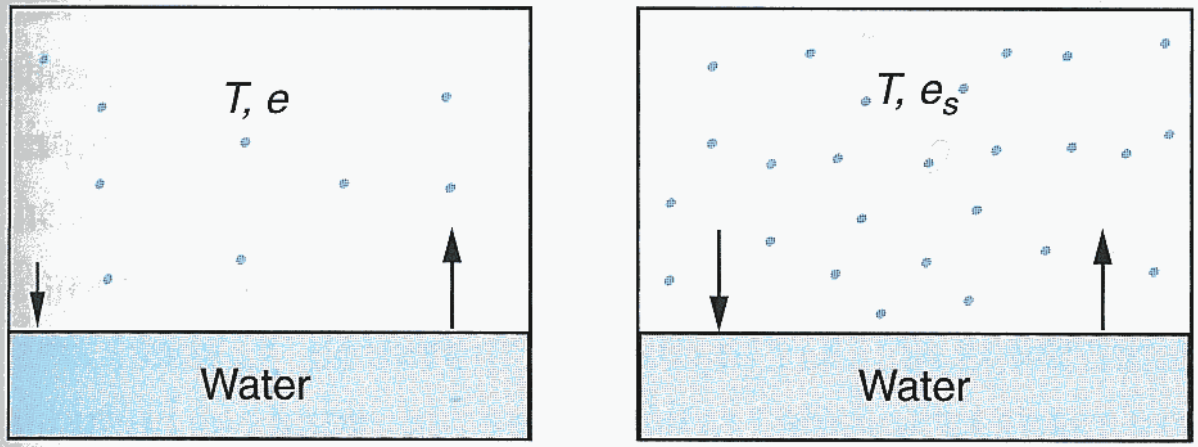


Fig. 3.8 A box (a) unsaturated and (b) saturated with respect to a plane surface of pure water at temperature T. Dots represent water molecules. Lengths of the arrows represent the relative rates of evaporation and condensation. The saturated (i.e., equilibrium) vapor pressure over a plane surface of pure water at temperature T is e_s as indicated in (b).

At saturation, the flux of water molecules into and out of the atmosphere is equal.

Saturation Vapor Pressure of Water Vapor over a Pure Water Surface

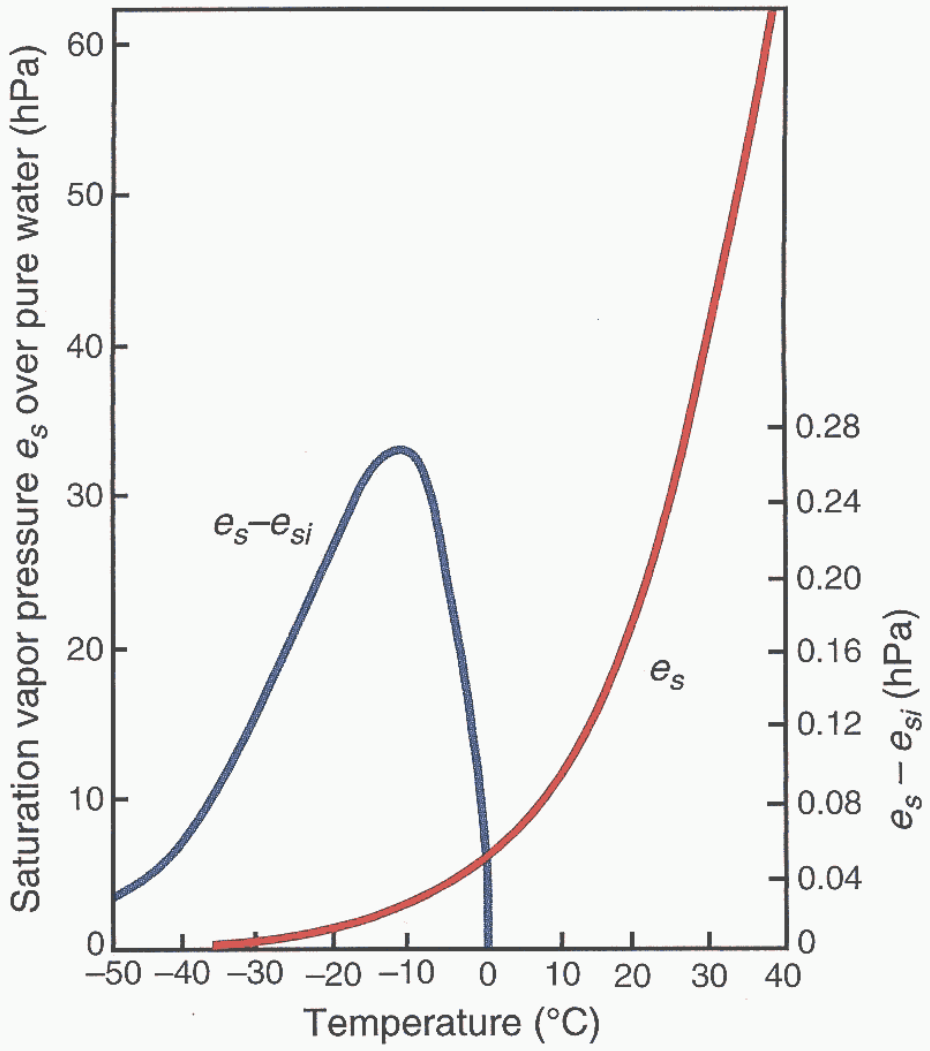


Fig. 3.9 Variations with temperature of the saturation (i.e., equilibrium) vapor pressure e_s over a plane surface of pure water (red line, scale at left) and the difference between e_s and the saturation vapor pressure over a plane surface of ice e_{si} (blue line, scale at right).

Moisture Parameters

The amount of water vapor in the atmosphere may be expressed in a variety of ways, and depending upon the problem under consideration, some ways of quantifying water are more useful than others.

e_s = Saturation Partial Pressure

w = $\frac{m_v}{m_d}$ Mass Mixing Ratios

Where m_v is the mass of water vapor in a given parcel, and m_d is the mass of dry air of the same parcel. This is usually expressed as grams of water per kilogram of dry air. ***w*** typically varies from 1 to 20 g/kg.

$$q = \frac{m_v}{m_v + m_d} = \frac{w}{1 + w}$$

Specific Humidity (typically a few %)

Moisture Parameters for Saturation

e_s = Saturation Partial Pressure

$w_s = \frac{m_{vs}}{m_d}$ Saturation Mixing Ratio

$$w_s = \frac{\rho'_{vs}}{\rho'_d} = \frac{e_s}{(R_v T)} \Big/ \frac{(p - e_s)}{(R_d T)}$$

ρ'_{vs} is the mass density of water required to saturate air at a given T.
p = total pressure

For Earth's Atmosphere:

$$w_s = 0.622 \frac{e_s}{p - e_s} \cong 0.622 \frac{e_s}{p}$$

Relative Humidity

$$RH \equiv 100 \frac{w}{w_s} \cong 100 \frac{e}{e_s}$$

The ***dew point***, T_d, is the temperature to which air must be cooled at constant pressure for it to become saturated with pure water.

Lifting Condensation Level

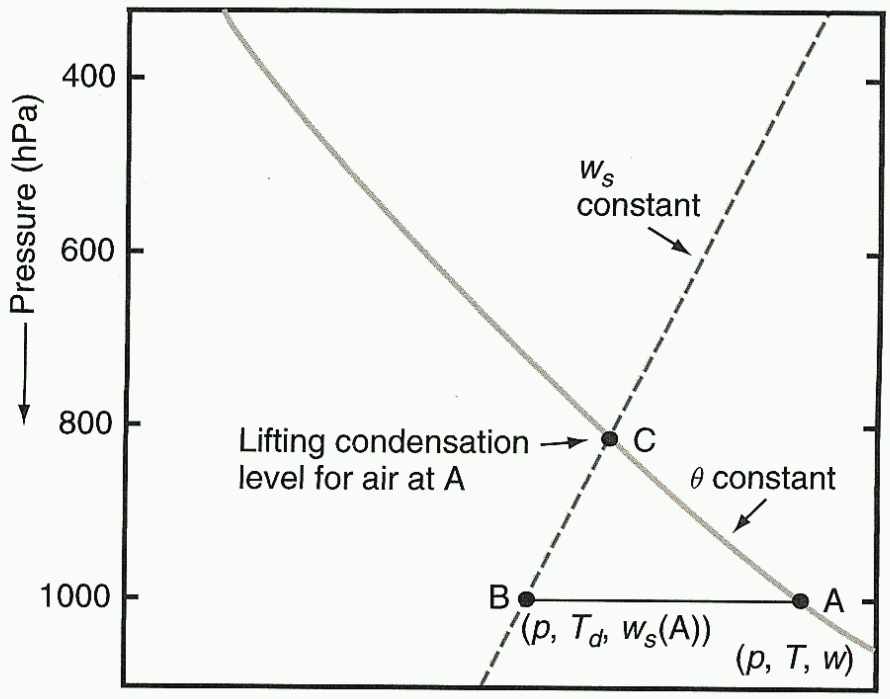


Fig. 3.10 The lifting condensation level of a parcel of air at A, with pressure p, temperature T, and dew point T_d, is at C on the skew $T - \ln p$ chart.

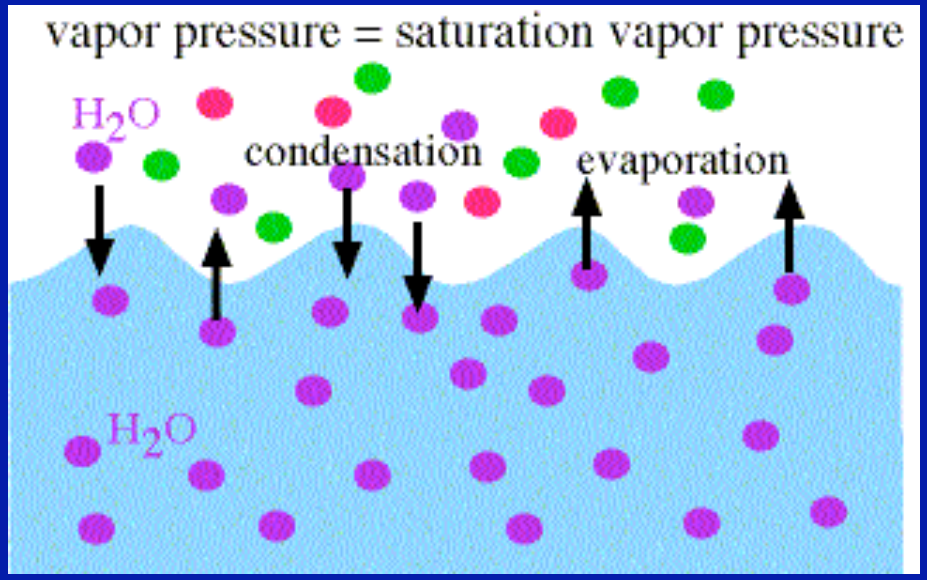
The **Lifting Condensation Level** (LCL) is defined as the level to which an unsaturated (but moist) parcel of air can be lifted adiabatically before it becomes saturated with pure water.

Saturation of Air

Air is **Saturated** if the abundance of water vapor (or any condensable) is at its maximum **Vapor Partial Pressure.**

In saturated air, ***evaporation is balanced by condensation***. If water vapor is added to saturated air, droplets begin to condense and fall out.

Under equilibrium conditions at a fixed temperature, the maximum vapor partial pressure of water is given by its **Saturated Vapor Pressure Curve**.



http://apollo.lsc.vsc.edu/classes/met130/notes/chapter5/graphics/sat_vap_press.free.gif

Relative Humidity is the ratio of the measured partial pressure of vapor to that in saturated air, multiplied by 100.

The relative humidity in clouds is typically about 102-107%, in other words, the clouds are **Supersaturated**.

Saturation Vapor Pressure:

Clausius-Clapeyron Equation of State

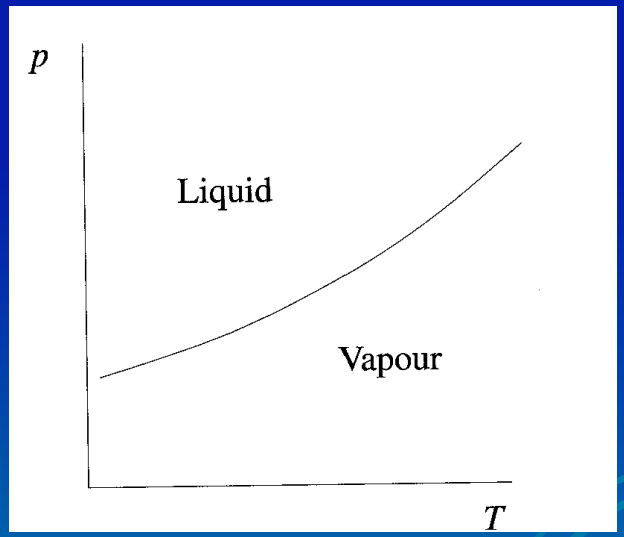
$$P_{sv}(T) = C_L e^{-L_s/RT}$$

$P_{sv}(T)$ = Saturation vapor pressure at temperature T

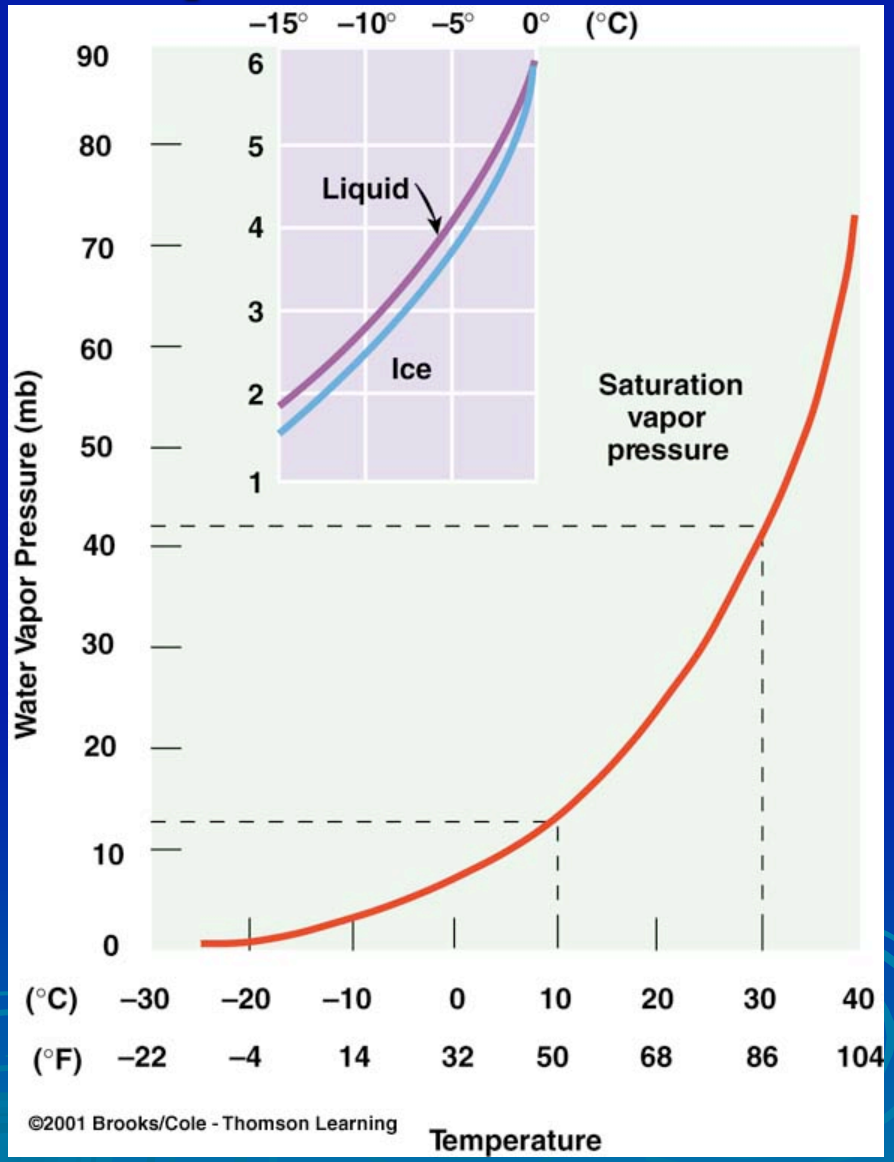
C_L = constant (depends upon condensable)

L_s = Latent Heat of Sublimation

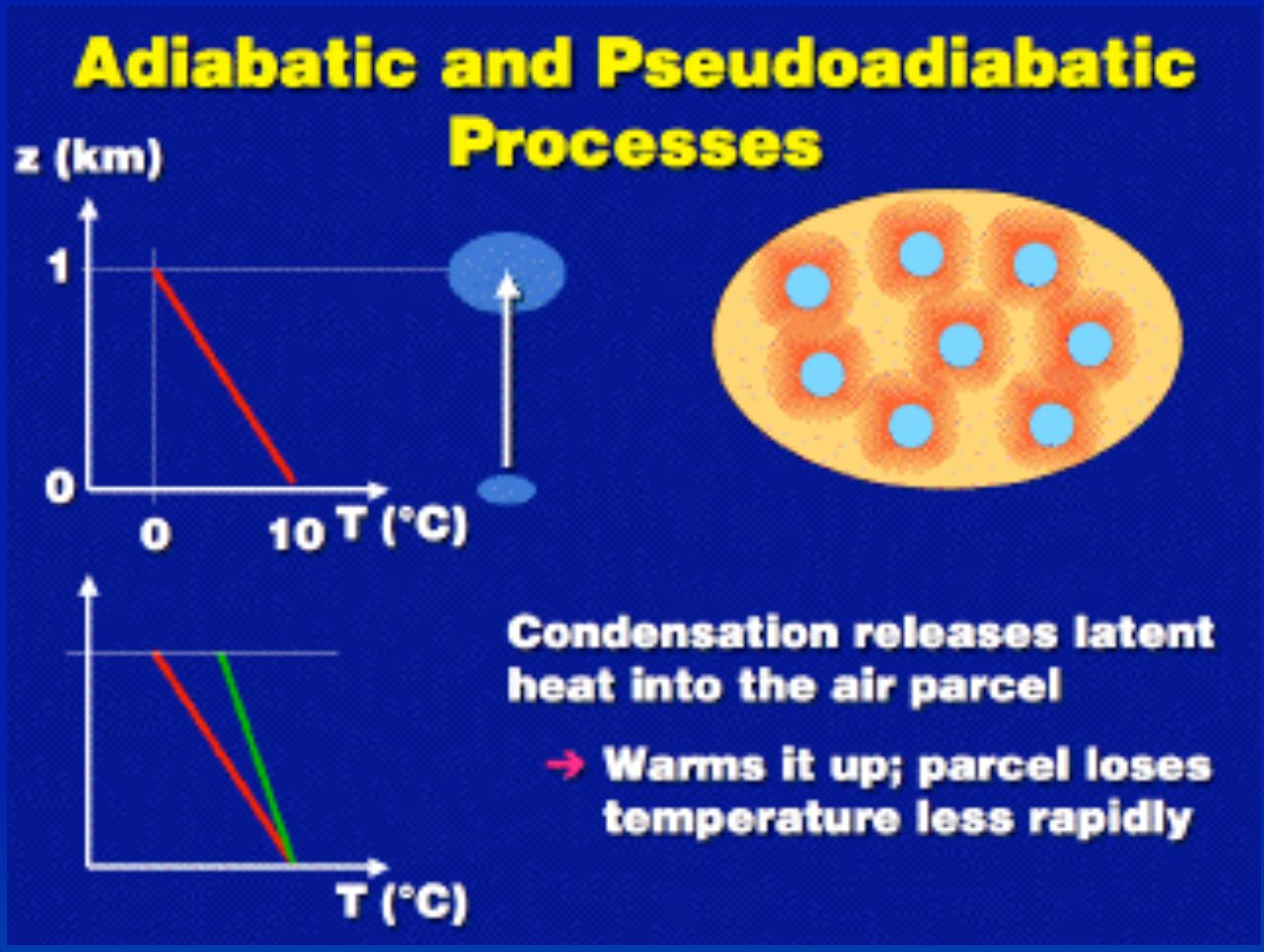
R = Gas constant



Phase Diagram of Water



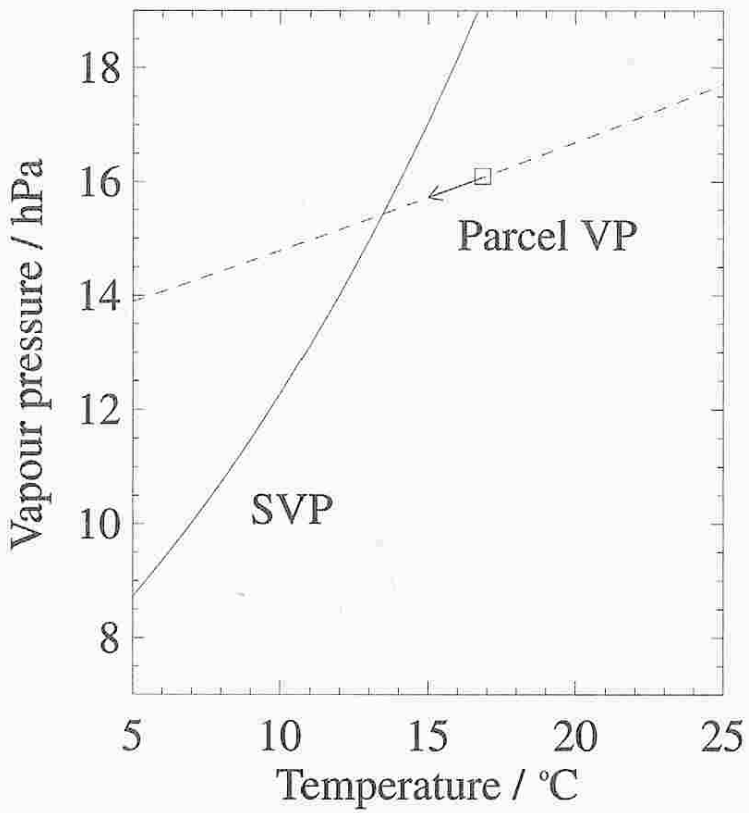
Vertical Motion and Condensation



Upward motion leads to cooling, via the FLT. Cooling increases the relative humidity. When the relative humidity exceeds 100%, then condensation can occur.

Adiabatic Motion of Moist Parcel

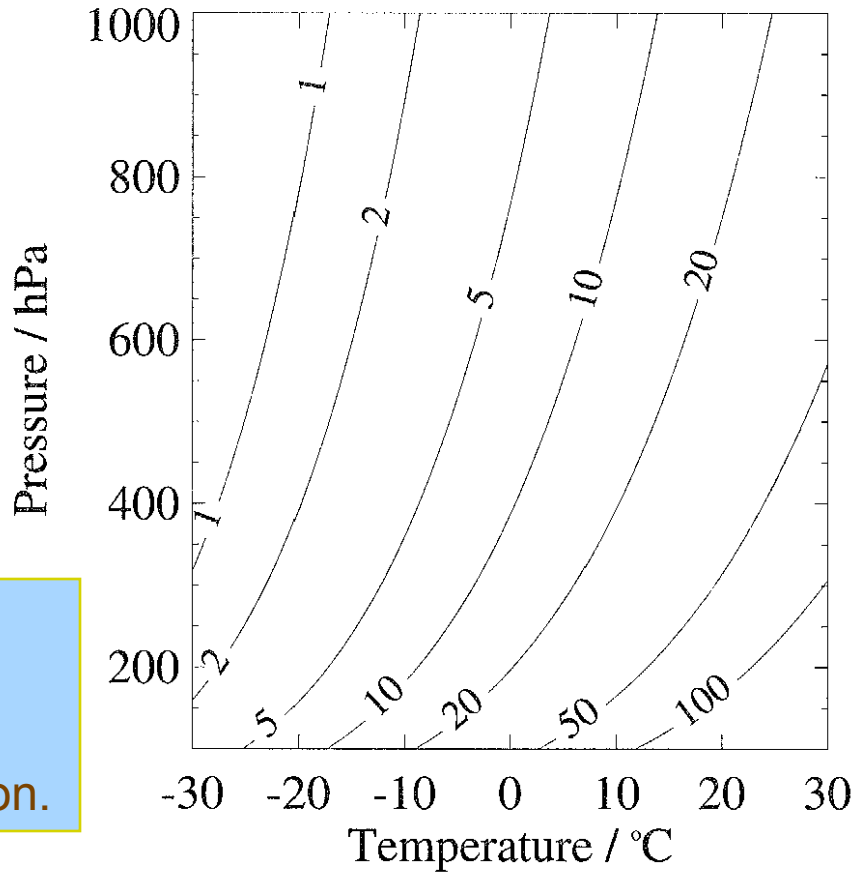
Figure 2.6 Saturation vapour pressure (SVP) curve (solid) and vapour pressure curve for a parcel of air containing water vapour of mass mixing ratio $= 10^{-2} = 10\,\mathrm{g\,kg^{-1}}$ (dashed). The parcel is taken to start at the point marked by a square, at which p_0 = 1000 hPa and T_0 = 290 K ≈ 17 °C. As the parcel rises, its vapour pressure moves down the dashed curve in the direction of the arrow, intersecting the SVP curve at $T \approx 13.3\,^{\circ}\mathrm{C}$.



As a parcel of air moves upwards, it expands and cools. The cooling leads to an increase in the relative humidity. When the vapor pressure exceeds the saturation vapor pressure, then condensation *can* occurs.

Saturation Profile and Temperature

Figure 2.7 Contours of the saturation mixing ratio $\mu_s(T, p)$, in units of $g\,kg^{-1}$.

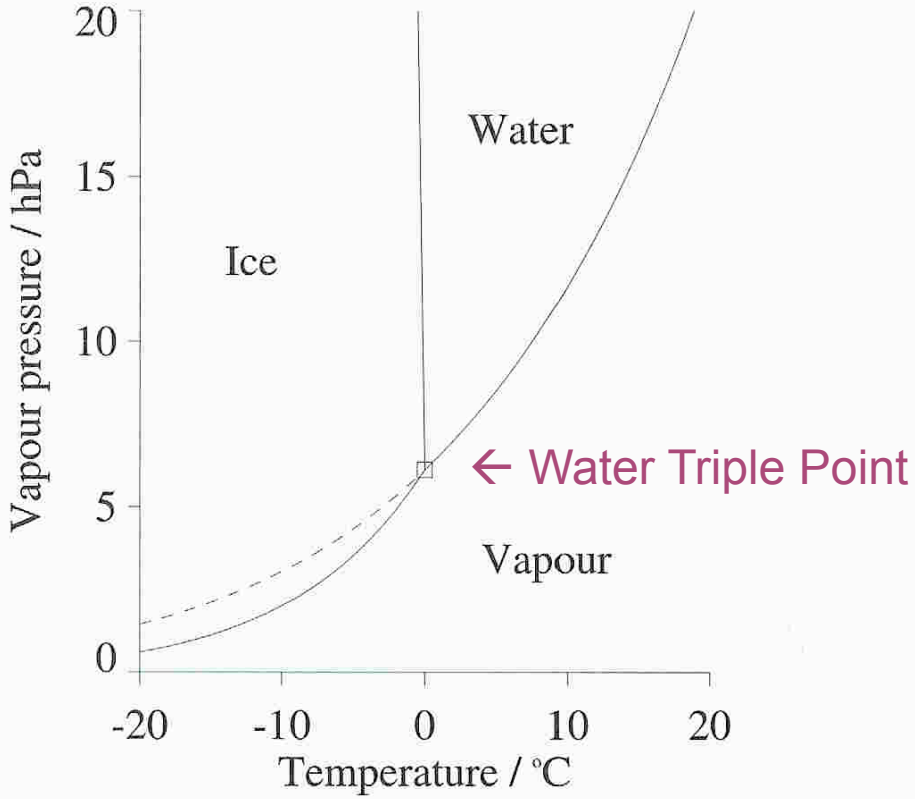


Amounts of water necessary for super-saturation, and thus condensation.

Is it possible to have snow when the atmospheric temperature is below – 30°C?

Water/Ice Transition

Figure 2.8 Schematic diagram showing the phase transitions between ice, liquid water and water vapour. The difference between the ice-vapour and vapour-water curves below 0 °C (where the vapour-water curve is shown dashed) has been exaggerated. Note that the ice-water curve is not quite vertical, but has a large negative slope. The triple point is indicated by the small square.



The saturation vapor pressure of water over ice is higher than that over liquid water. This leads to small, but measurable change is the relative humidity.

Liquid/Ice Transition

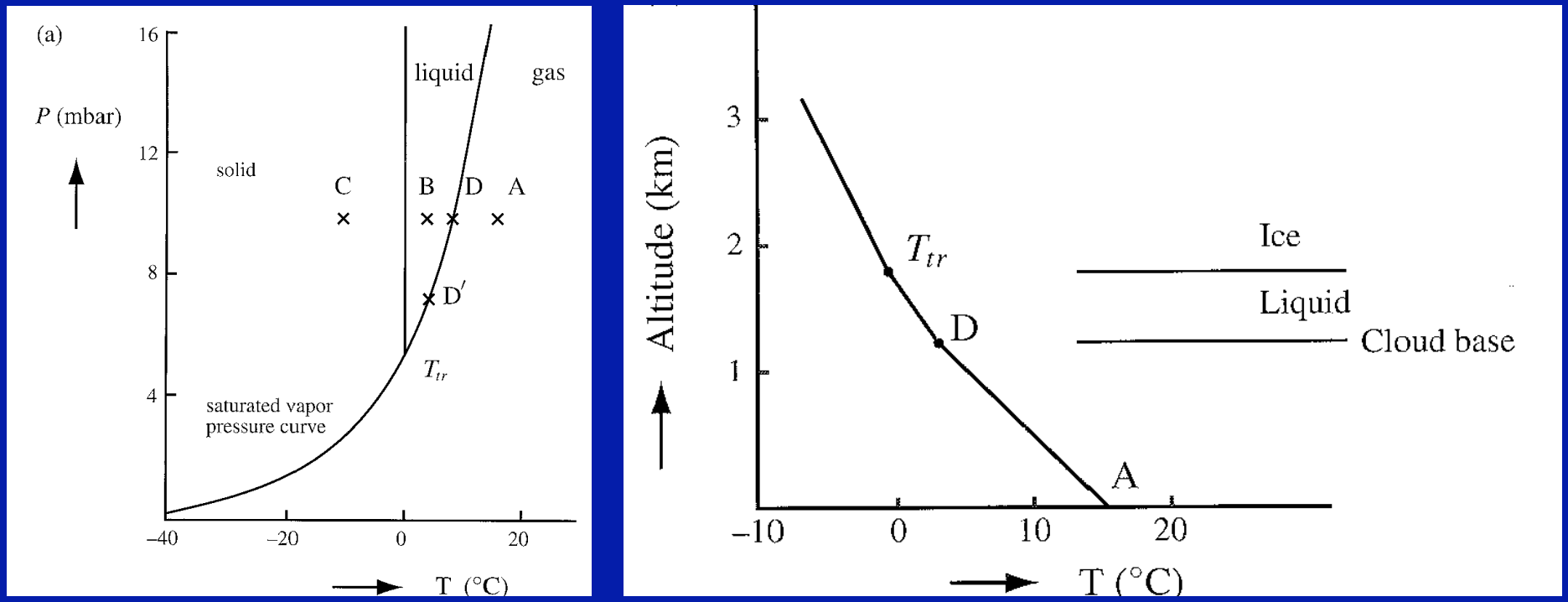


FIGURE 4.15 (a) Saturation vapor pressure curve for water. The y-axis indicates the partial pressure for H_2O vapor at the temperature (in °C) indicated along the x-axis. Water vapor is in the gas state when in point A; the saturated vapor pressure of water in point B is 7.6 mbar (indicated by point D′), and above C there is only 2.6 mbar of H_2O vapor. (b) Ideal sketch of the temperature structure in the Earth's atmosphere. In the lower troposphere, the air follows a dry adiabat. A wet air parcel rising up through the atmosphere starts to condense at point D′, when the water vapor inside the air parcel exceeds the saturated vapor curve (see panel a). The temperature profile in the atmosphere follows the wet adiabat in this region, and steepens even more above T_{tr}, when water freezes into ice crystals.

Wet (Moist) Adiabatic Lapse Rate

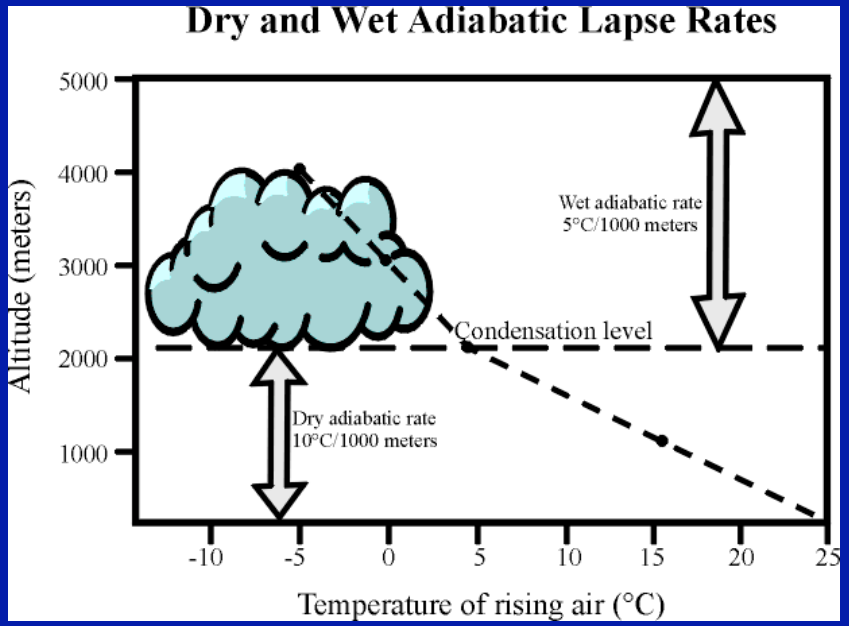
$\Gamma_d = -g/C_p$ = Dry Adiabatic Lapse Rate

In determining the **moist adiabatic lapse** rate, we must modify the First Law of Thermodynamics to include the phase change energy.

Let μ_s = mass of liquid water.

$dQ = C_p dT + gdz$ (FLT for a parcel)

$dQ = -L_s d\mu_s$ (Heat added from water condensation)



Here we assume that the water which condenses drops out of the parcel. Thus this process is strictly irreversible.

Together this implies that the FLT becomes: $C_p dT + gdz + L_s d\mu_s = 0$

Wet Lapse Rate - continued

$C_p dT + g dz + L_s d\mu_s = 0$ (FLT for a saturated parcel)

The mass of water depends upon the degree of saturation:

$\mu_s = \epsilon (e_s/p)$ and by the chain rule $d\mu_s/\mu_s = de_s/e_s - dp/p$

$de_s = (de_s/dT)\, dT$

$(1/e_s)\, de_s/dT = L_s/RT^2$ (Differential form of Clausius-Clapeyron Eqn.)

$dP = -gdz/RT$ (Hydrostatic Law)

This gives us $d\mu_s/\mu_s = L_s dT/RT^2 + gdz/RT$

Using this equation and the FLT form at the top of this page we get:

$(C_p + L_s^2 \mu_s/RT^2)\, dT + g(1+L_s\mu_s/RT)\, dz = 0$

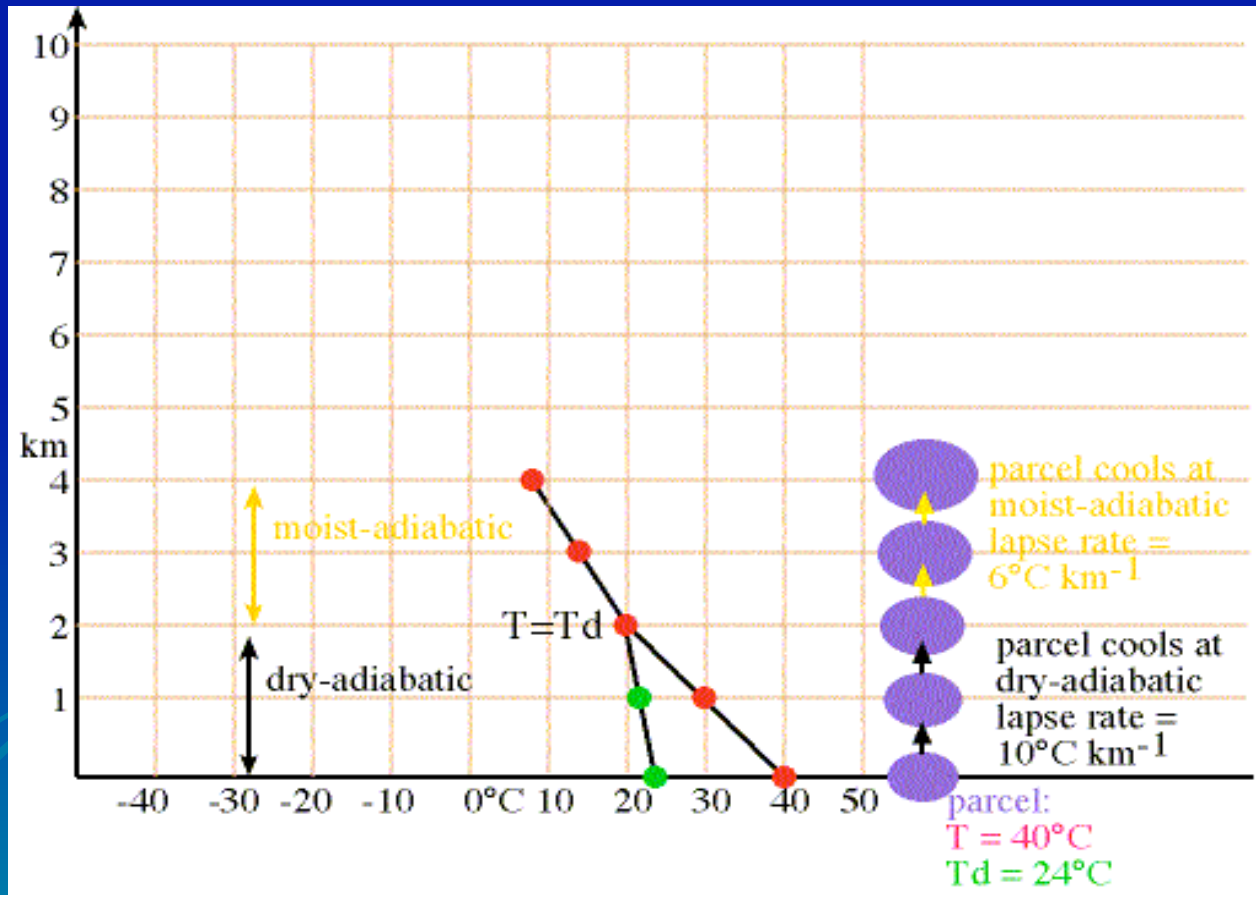
Wet Lapse Rate - Continued

$$\Gamma_w = dT/dz = -(g/C_p)\,((1+L_s\mu_s/RT) / (1 + L_s^2\mu_s/C_pRT^2))$$

Note that when $\mu_s = 0$, this reduces to Γ_d

The factor ((xx)) is always less or equal to1. So $\Gamma_d < \Gamma_w$

Thus, water acts as an agent to increase the effective heat capacity of the atmosphere.



5. Static Stability

Archimedes Principle:
The upward force (buoyancy) is equal to the weight of the displaced air.

The **net force** on a parcel is equal to the difference between weight of the air in the parcel and the weight of the displaced air.

$$F = (\rho - \rho')\, g$$

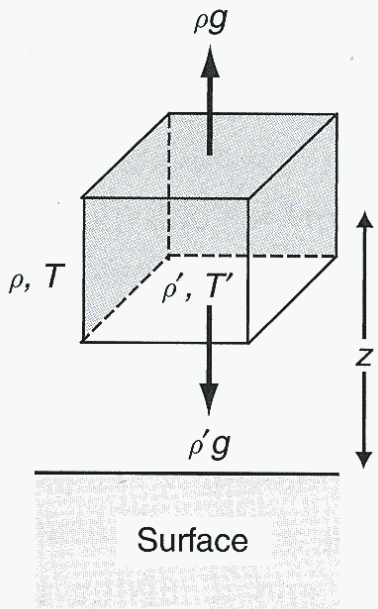
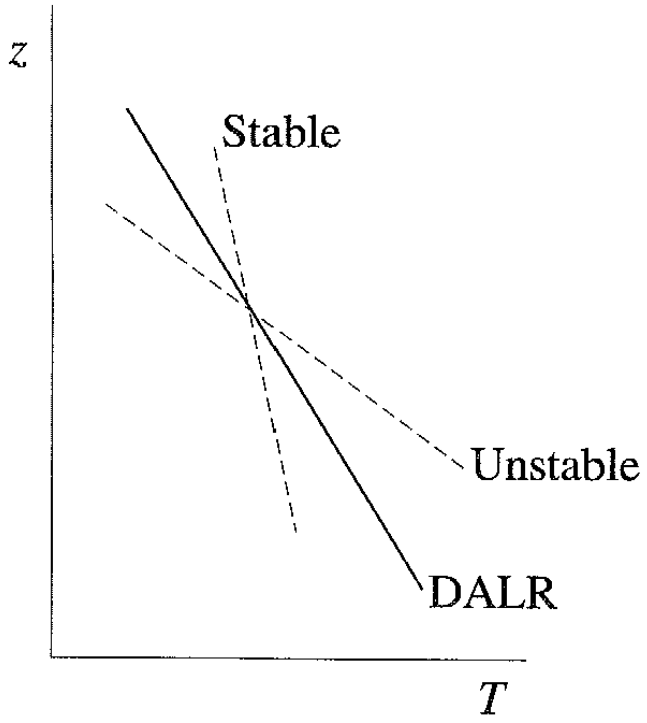


Fig. 3.13 The box represents an air parcel of unit volume with its center of mass at height z above the Earth's surface. The density and temperature of the air parcel are ρ' and T', respectively, and the density and temperature of the ambient air are ρ and T. The vertical forces acting on the air parcel are indicated by the thicker arrows.

Vertical Stability

Figure 2.4 Schematic height profiles of temperature, showing the dry adiabatic lapse rate (DALR) and examples of stable and unstable environmental lapse rates.



$dT/dz = -g/C_p$ = dry adiabatic lapse rate (neutrally stable)

$dT/dz < -g/C_p$ → Unstable

$dT/dz > -g/C_p$ → Stable

Static Stability

Stable

Unstable

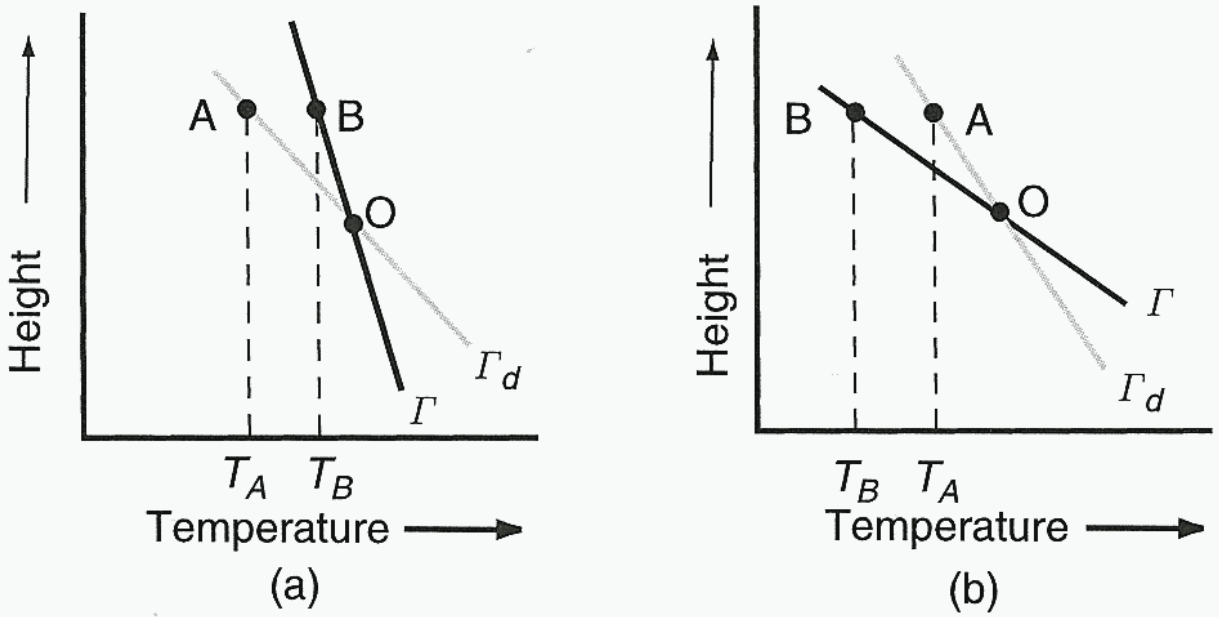


Fig. 3.12 Conditions for (a) positive static stability ($\Gamma < \Gamma_d$) and (b) negative static instability ($\Gamma > \Gamma_d$) for the displacement of unsaturated air parcels.

$$\Gamma_d = -g/C_p$$

Stability and the Effects of Condensation

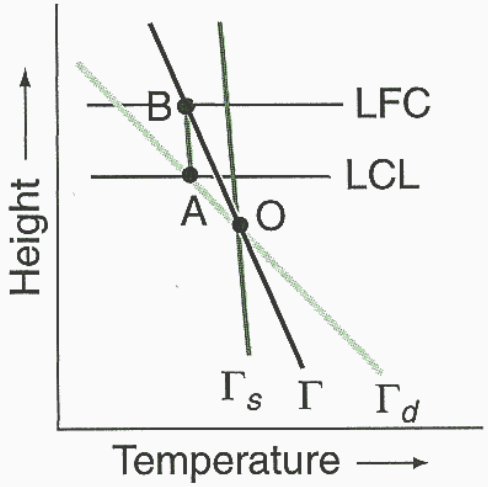


Fig. 3.16 Conditions for conditional instability ($\Gamma_s < \Gamma < \Gamma_d$). Γ_s and Γ_d are the saturated and dry adiabatic lapse rates, and Γ is the lapse rate of temperature of the ambient air. LCL and LFC denote the *lifting condensation level* and the *level of free convection*, respectively.

Thunderstorms seen from the Space Shuttle



Moisture leads to conditional stability in the atmosphere.

Analogs for Stability

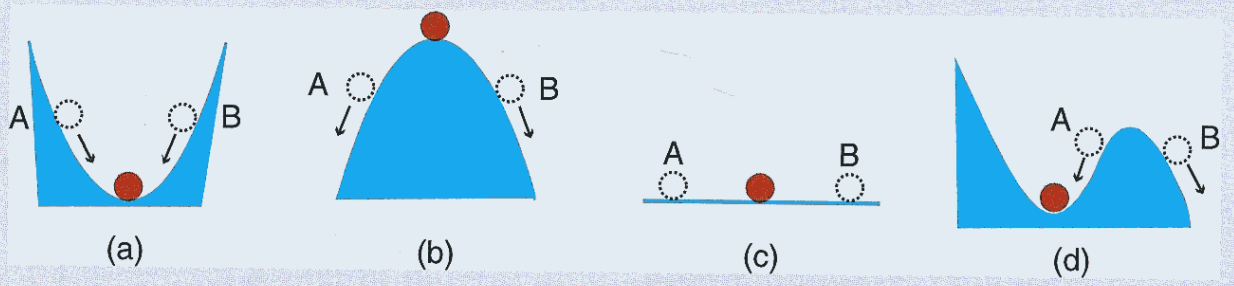


Fig. 3.17 Analogs for (a) stable, (b) unstable, (c) neutral, and (d) conditional instability. The red circle is the original position of the ball, and the white circles are displaced positions. Arrows indicate the direction the ball will move from a displaced position if the force that produced the displacement is removed.

Under **stable** atmospheric conditions, an air parcel that is displaced in the vertical direction will return to its original position.

Neutral stability occurs when the air parcel will remain at it's displaced position without any additional forces acting on it.

For **unstable** conditions, an air parcel that is displaced in the vertical will continue to move in the direction of the displacement.

Conditional instability occurs when a significant displacement of the air parcel must occur before instability can occur.

Regions of Convective Instability

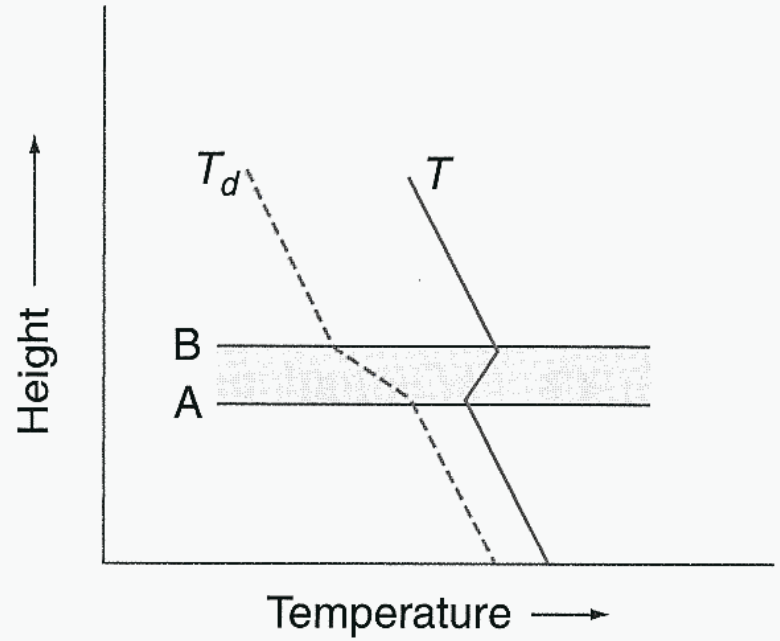


Fig. 3.18 Conditions for convective instability. T and T_d are the temperature and dew point of the air, respectively. The blue-shaded region is a dry inversion layer.

Convective instability may occur in only a small portion of the vertical structure. **Temperature inversions** therefore can inhibit convection.

Atmospheric Oscillations:
Gravity Waves in Stable Air

Consider the force on a parcel of air that has been displaced vertically by a distance z from its equilibrium altitude. Assume that the air is dry and that displacements occur sufficiently slow that we can assume that they are adiabatic. Primed quantities will denote parcel variables.

By Archimedes Principle, the force on the parcel is the buoyancy force minus the gravitational force. The net force is:

$$F = (\rho - \rho')g$$

Acceleration: $$\rho' \frac{d^2 z}{dt^2} = (\rho - \rho')g$$ OR $$\frac{d^2 z}{dt^2} = \frac{F}{\rho'} = \left(\frac{\rho - \rho'}{\rho'}\right)g$$

Substituting from IGL: $$\frac{d^2 z}{dt^2} = \frac{\frac{1}{T} - \frac{1}{T'}}{\frac{1}{T'}} g$$ OR $$\frac{d^2 z}{dt^2} = g\left(\frac{T' - T}{T}\right)$$

Atmospheric Oscillations - continued

If we assume a linear atmospheric temperature profile with rate of change with altitude of Γ, then the temperature profile may be written (z' = displacement)

$$T = T_0 - \Gamma z'$$

The parcel moves adiabatically in the vertical, so its temperature is:

$$T' = T_o - \Gamma_d z'$$

Which gives:

$$T' - T = -(\Gamma_d - \Gamma) z'$$

The equation of motion becomes:

$$\frac{d^2 z'}{dt^2} = -\frac{g}{T}(\Gamma_d - \Gamma) z'$$

Which can be written:

$$\frac{d^2 z'}{dt^2} + N^2 z' = 0$$

Brunt-Väisälä Frequency:

$$N^2 = \frac{g}{T}(\Gamma_d - \Gamma)$$

Atmospheric Oscillations - continued

The equation of motion for the parcel is

$$\frac{d^2 z'}{dt^2} + N^2 z' = 0$$

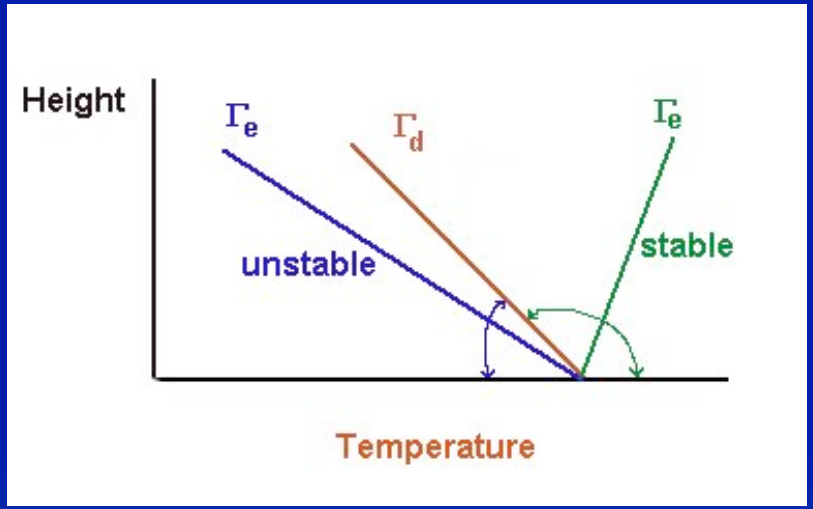
Brunt-Väisälä Frequency:

$$N^2 = \frac{g}{T}\left(\Gamma_d - \Gamma\right)$$

If the air is stably stratified, i.e., $\Gamma_d > \Gamma$, then the parcel will oscillate about its starting position with simple harmonic motion.

These are called buoyancy oscillations. Typical periods are about 15 minutes.

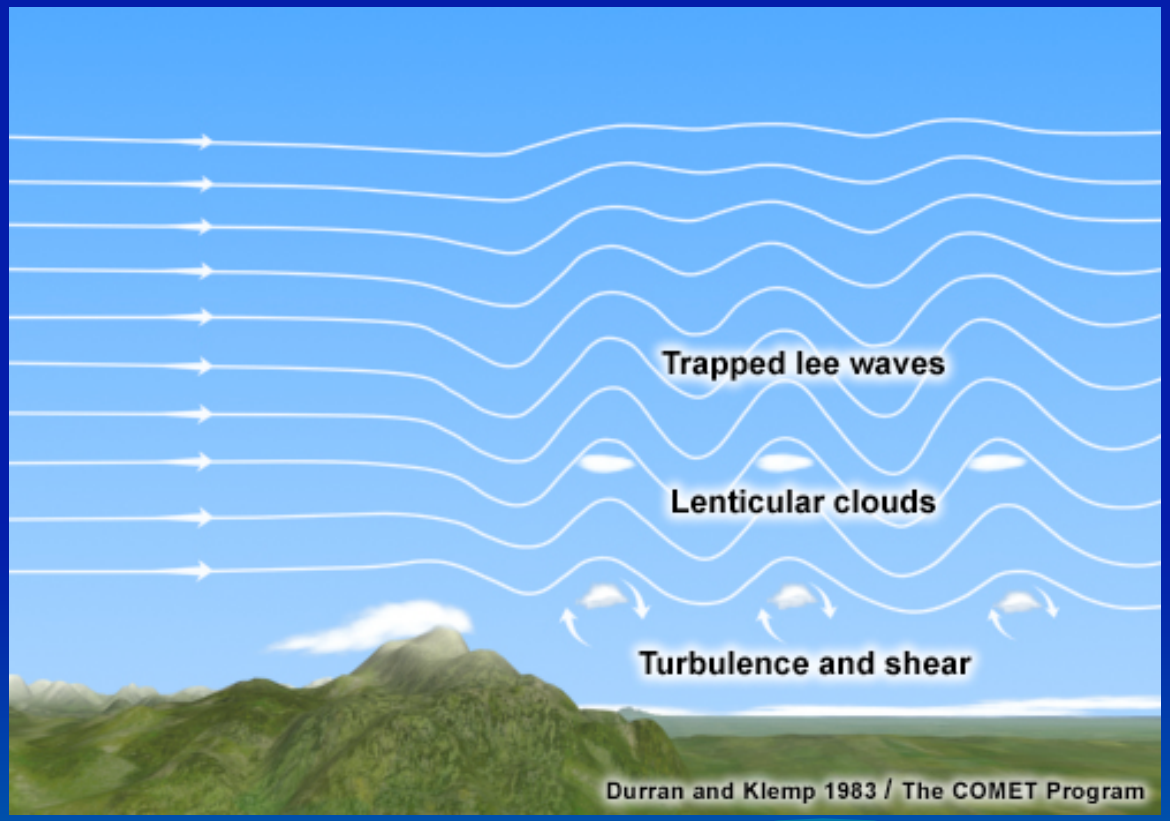
For winds of ~ 20 ms^{-1}, the wavelength is ~10-20 km.



Here $\Gamma_e = \Gamma$ in notes

Mountain (Lee) Waves

Buoyancy Oscillations: $N^2 = \frac{g}{T}(\Gamma_d - \Gamma)$

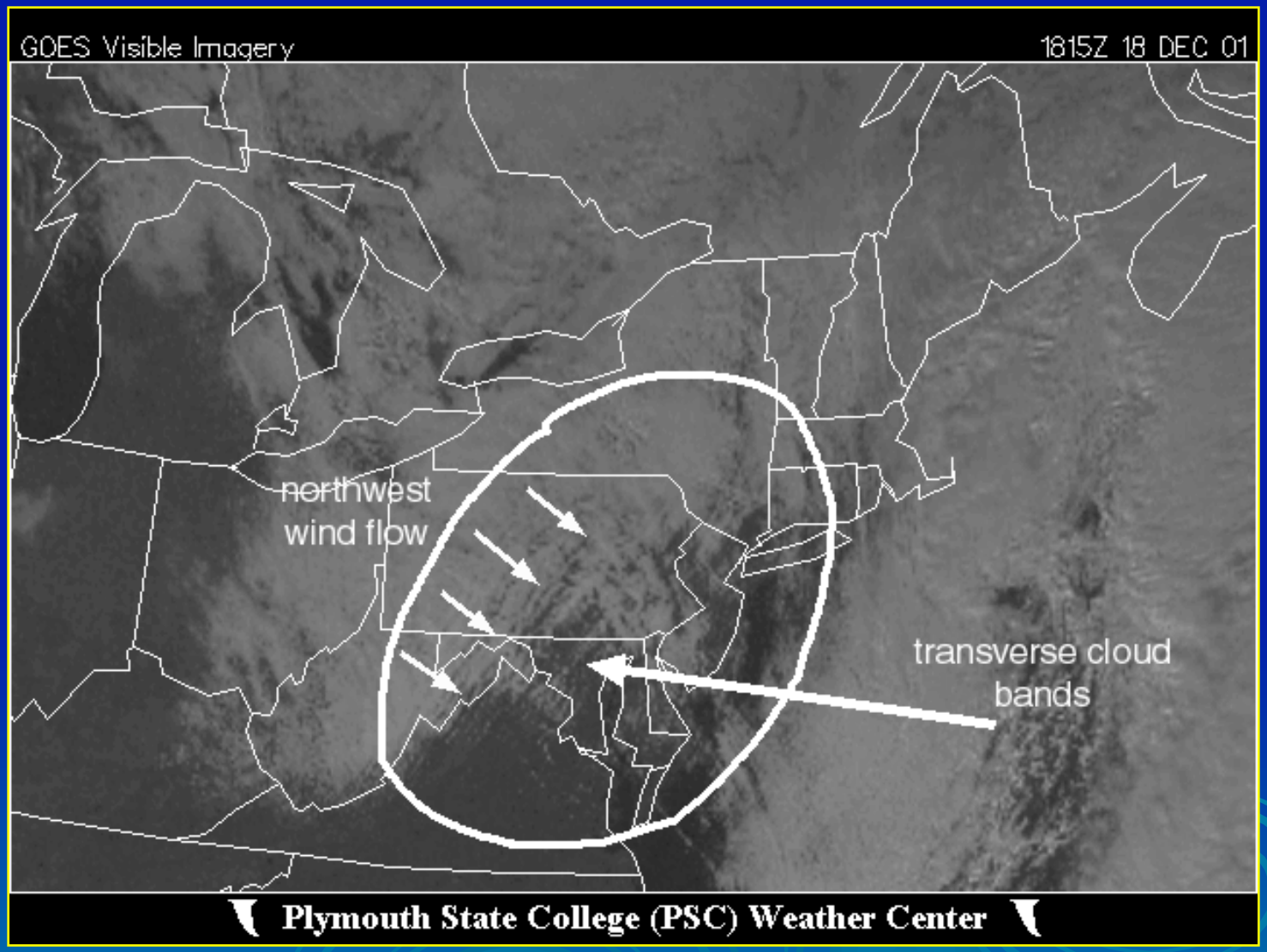


Lee Waves



Observed from ground

Lee Waves Observed from Space



Mountain Winds



Mountain regions display many interesting weather patterns. One example is the valley wind which originates on south-facing slopes (north-facing in the southern hemisphere). When the slopes and the neighboring air are heated the density of the air decreases, and the air ascends towards the top following the surface of the slope. At night the wind direction is reversed, and turns into a down-slope wind. If the valley floor is sloped, the air may move down or up the valley, as a canyon wind. Winds flowing down the leeward sides of mountains can be quite powerful:

Examples are the **Foehn** in the Alps in Europe, the **Chinook** in the Rocky Mountains, and the **Zonda** in the Andes. Examples of other local wind systems are the **Mistral** flowing down the Rhone valley into the Mediterranean Sea, the **Scirocco**, a southerly wind from Sahara blowing into the Mediterranean sea.

Mountain Winds and Climate

Hawaii







6. The Second Law of Thermodynamics

The **Second Law of Thermodynamics** states that it is *impossible* to completely convert heat energy into mechanical energy. Another way to put that is to say that the level of *entropy* (or *tendency toward randomness*) in a closed system is always either *constant* or *increasing*.

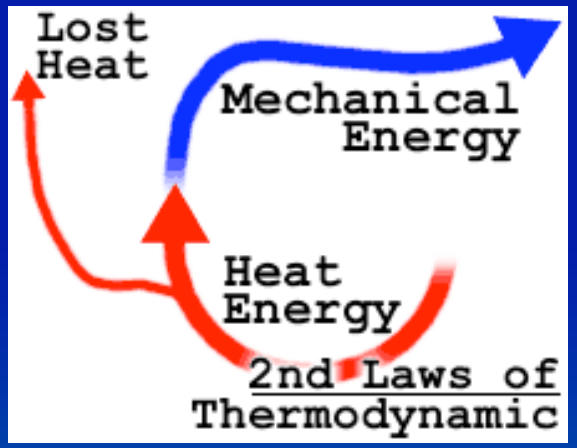
Implications of the Second Law

- It is impossible for any process (engine), working in a cycle, to completely convert surrounding heat to work.
- Dissipation will always occur.
- Entropy will always increase.

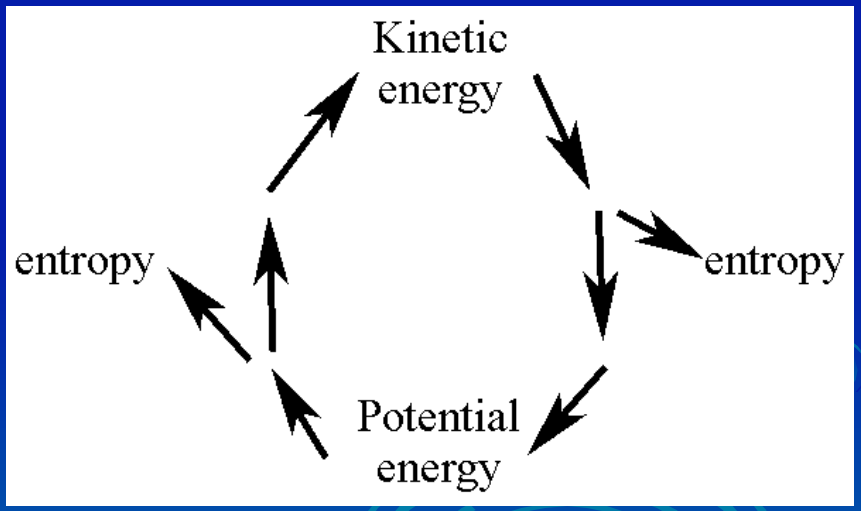
Second Law of Thermodynamics and Atmospheric Processes

The Entropy of an isolated system increases when the system undergoes a spontaneous change.

Entropy is the heat added (or subtracted) to a system divided by its temperature.



$dS = dQ/T$



Second Law of Thermodynamics

The Carnot Cycle

- The First Law of Thermodynamics is a statement about conservation of energy.

- The Second Law of Thermodynamics is concerned with the maximum fraction of a quantity of heat that can be converted into work. There is a theoretical limit to this conversion that was first demonstrated by Nicholas Carnot.

A **cyclic process** is a series of operations by which the state of a substance (called the working substance) changes, but is finally returned to its original state (in all respects).

If the volume changes during the cycle, then work is done ($dW = PdV$).

The net heat that is absorbed by the working substance is equal to the work done in the cycle. If during one cycle a quantity of heat Q_1 is absorbed and a quantity Q_2 is rejected, then the net work done is $Q_1 - Q_2$.

The efficiency is:

$$\eta = \frac{Q_1 - Q_2}{Q_1} = \frac{Work_done_by_engine}{Heat_abosrbed}$$

Carnot's Ideal Heat Engine

$T_1 > T_2$

1. A→B Adiabatic Compression
   Work done on substance
2. B→ C Isothermal Expansion
   Work done on environment
3. C→ D Adiabatic Expansion
   Work done on environment
4. D→ A Isothermal Compression
   Work done on substance

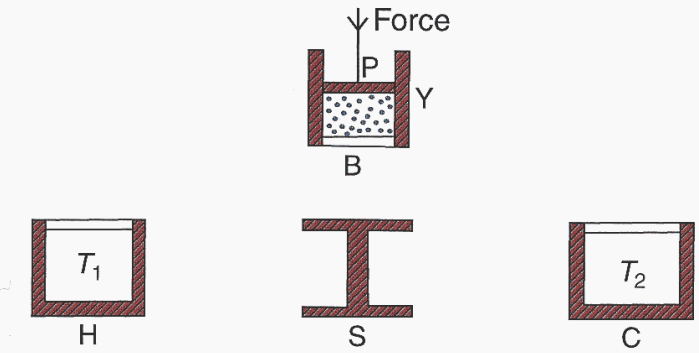


Fig. 3.19 The components of Carnot's ideal heat engine. Red-shaded areas indicate insulating material, and white areas represent thermally conducting material. The working substance is indicated by the blue dots inside the cylinder.

Incremental work done: dW = PdV
So the area enclosed on the P-V diagram is the total Work.

Only by transferring heat from a hot to a cold body can work be done in a cyclic process.

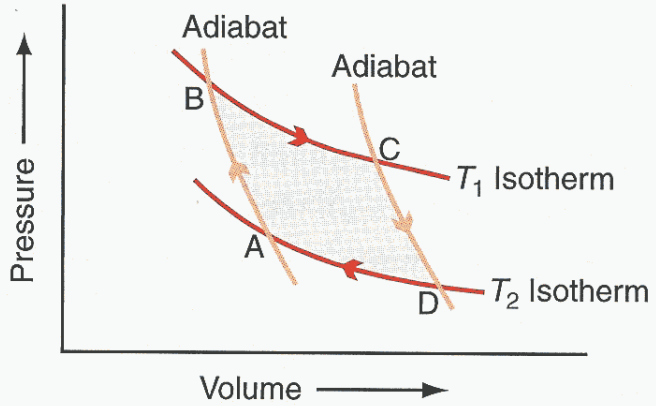


Fig. 3.20 Representations of a Carnot cycle on a *p–V* diagram. Red lines are isotherms, and orange lines are adiabats.

Isotherms and Adiabats

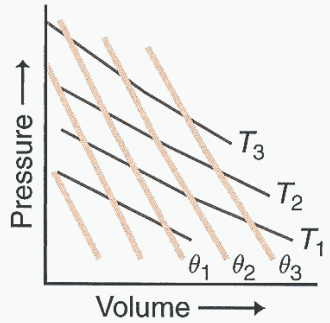


Fig. 3.21 Isotherms (red curves labeled by temperature T) and adiabats (tan curves labeled by potential temperature θ) on a p–V diagram.

P-V diagram

Isothermal Process: T = constant, dT = 0

Adiabatic: dQ = 0

Isentropic: dS = 0

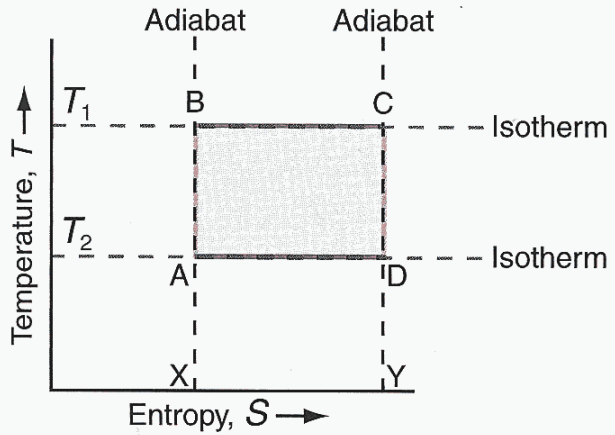


Fig. 3.22 Representation of the Carnot cycle on a temperature (T)–entropy (S) diagram. AB and CD are adiabats, and BC and DA are isotherms.

T-S diagram

Saturation Vapor Pressure: The Clausius-Clapeyron Equation

By application of the ideas of a cyclic process changing water from a liquid to a gas, we can derive the differential form of the Clausius-Clapeyron equation:

$$\frac{1}{e_s}\frac{de_s}{dT} = \frac{L_v}{R_v T^2} = \frac{L_v M_w}{1000 R^* T^2}$$

In its integrated form:

$$e_s(T) = Ce^{-L_v/RT}$$

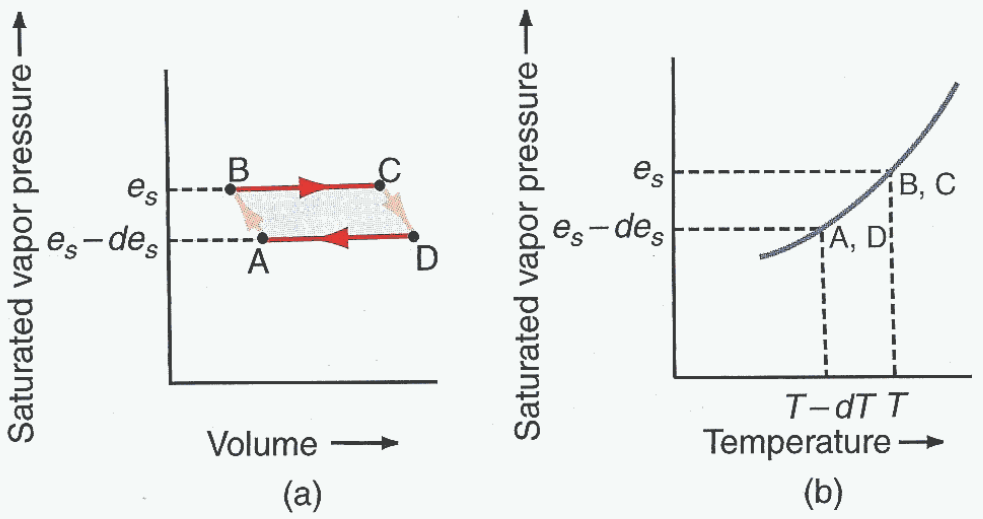


Fig. 3.23 Representation on (a) a saturated vapor pressure versus volume diagram and on (b) a saturated vapor pressure versus temperature diagram of the states of a mixture of a liquid and its saturated vapor taken through a Carnot cycle. Because the saturated vapor pressure is constant if temperature is constant, the isothermal transformations BC and DA are horizontal lines.

Water Vapor and the Carnot Cycle

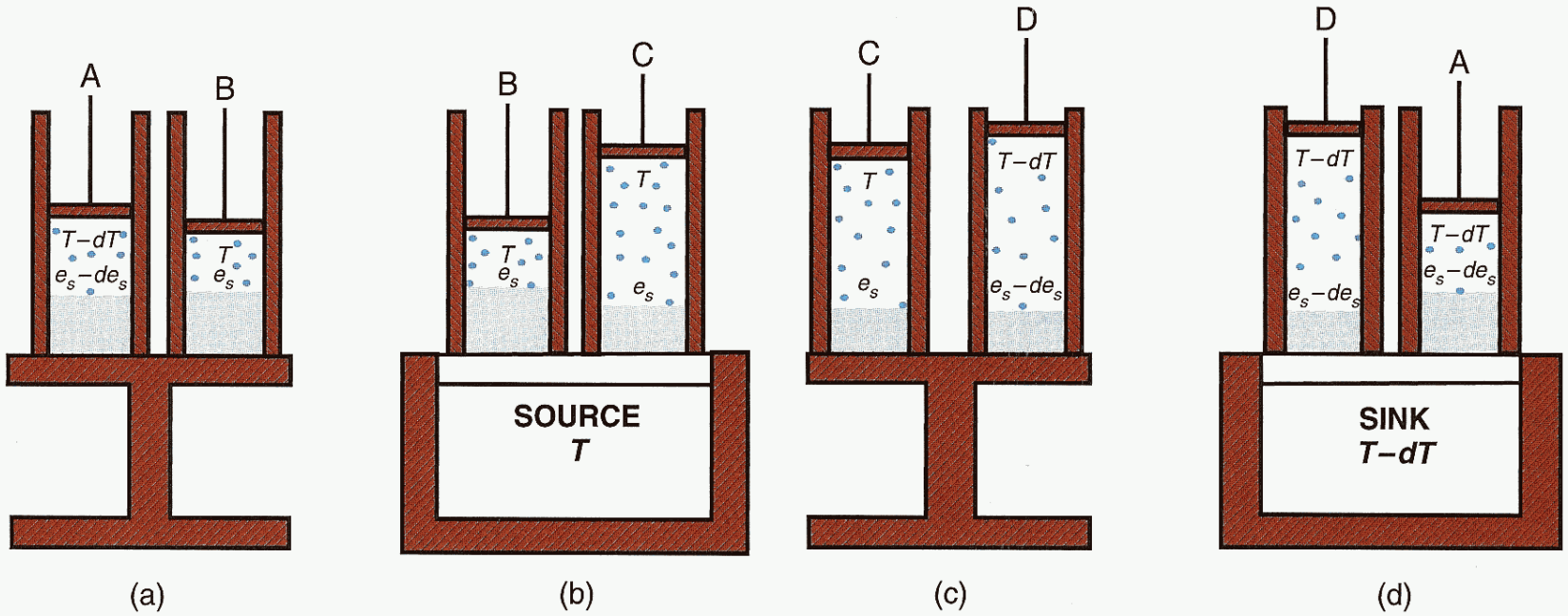


Fig. 3.24 Transformations of a liquid (solid blue) and its saturated vapor (blue dots) in a Carnot cycle. The letters A, B, C, D indicate the states of the mixture shown in Fig. 3.23. Red-shaded areas are thermally insulating materials.

Ambient Pressure and Boiling Point

Water boils at a temperature T_B such that the water vapor pressure at that temperature is equal to the ambient air pressure, i.e.,

$$e_s(T_B) = P_{atmos}$$

The change in boiling point, T_B, as a function of temperature is given by a form of the Clausius-Clapeyon equation:

$$\frac{\Delta T_B}{\Delta p_{atmos}} = \frac{T_B(\alpha_2 - \alpha_1)}{L_v}$$

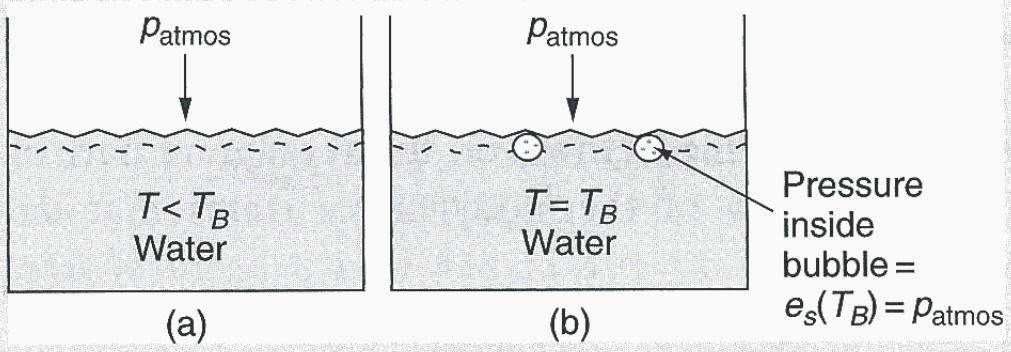


Fig. 3.25 (a) Water below its boiling point (T_B): bubbles cannot form because $e_s(T) < p_{atmos}$. (b) Water at its boiling point: bubbles can form because the pressure inside them, $e_s(T_B)$, is equal to the atmospheric pressure (p_{atmos}) acting on them.

Because $\alpha_2 > \alpha_1$, T_B increases with increasing p_{atmos}. Thus if the ambient atmospheric pressure is less than sea level, the T_B will be lower.

Generalized Statement of the Second Law of Thermodynamics

$$\Delta S \equiv \frac{\Delta Q}{T}$$

If the system is reversible, no dissipation occurs.

- For a reversible transformation there is no change in the entropy of the universe (system + surroundings).
- The entropy of the universe increases as a result of irreversible transformations.

"The Second Law of Thermodynamics cannot be proved. It is believed because it leads to deductions that are in accord with observations and experience."

Questions for Discussion

1. How does one define energy, apart from what it does or is capable of doing?
2. What is Thermodynamics?
3. Why is Thermodynamics relevant to atmospheric science?
4. Why is Thermodynamics a good starting point for discussing atmospheric science?
5. What causes energy transport?
6. Is it possible to perform work with an isothermal system?

Questions for Discussion

7. Why is entropy an important concept in atmospheric physics?
8. Does an atmospheric "parcel" really exist?
9. Is the atmosphere in thermal equilibrium?
10. Is the atmosphere in dynamical equilibrium?
11. What is the difference between steady state and equilibrium?
12. In what ways are the Earth's atmosphere like a heat engine?
13. Why is it impossible to prove the Second Law of Thermodynamics?

Pseudoadiabatic Chart

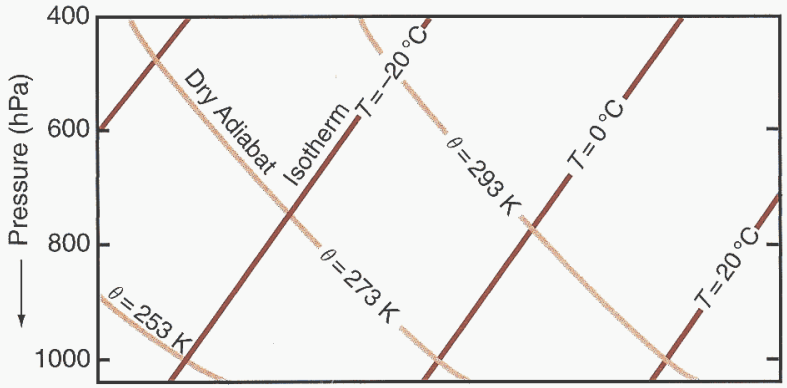


Fig. 3.7 Schematic of a portion of the skew $T - \ln p$ chart. (An accurate reproduction of a larger portion of the chart is available on the book web site that accompanies this book, from which it can be printed and used for solving exercises.)

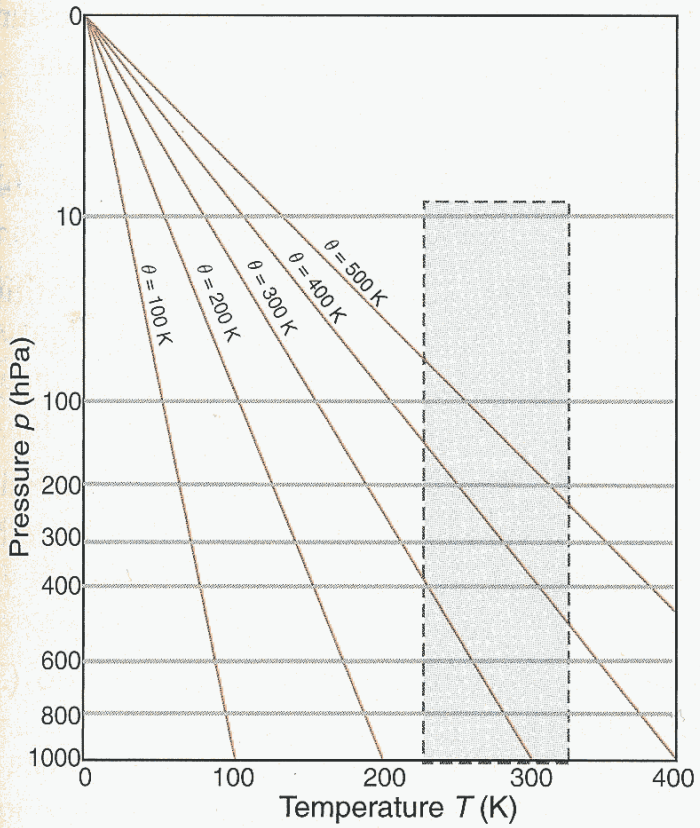


Fig. 3.6 The complete pseudoadiabatic chart. Note that p increases downward and is plotted on a distorted scale (representing $p^{0.286}$). Only the blue-shaded area is generally printed for use in meteorological computations. The sloping lines, each labeled with a value of the potential temperature θ, are dry adiabats. As required by the definition of θ, the actual temperature of the air (given on the abscissa) at 1000 hPa is equal to its potential temperature.

Normand's Rule

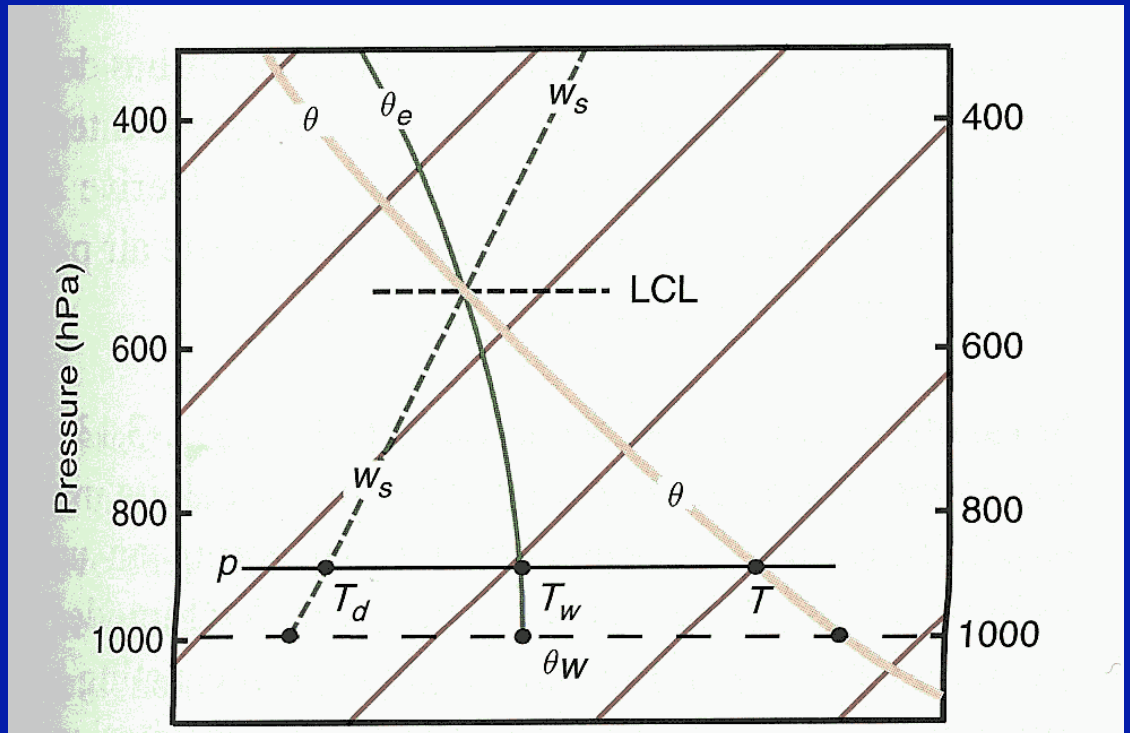


Fig. 3.11 Illustration of Normand's rule on the skew $T - \ln p$ chart. The orange lines are isotherms. The method for determining the wet-bulb temperature (T_w) and the wet-bulb potential temperature (θ_w) of an air parcel with temperature T and dew point T_d at pressure p is illustrated. LCL denotes the lifting condensation level of this air parcel.