

Lab 6

SATURATION AND ATMOSPHERIC STABILITY

Materials Needed

- calculator
- ruler

Introduction

How do clouds form? The concept of saturation, introduced in the previous lab, is important here. The two basic means by which air achieves saturation (relative humidity = 100%) are cooling and increasing water vapor content. In this lab we will examine several common atmospheric processes that will lead to either or both of these phenomena. We will focus particularly on how air changes as it rises from the earth's surface and how these changes relate to atmospheric stability.

Mixing

Saturation occasionally occurs when two air masses mix. Mixing usually involves a change in both water vapor content and temperature. Consider, for example, two unsaturated air samples, A and B:

Air Sample	A	B
Temperature	10°C	38°C
Relative humidity	75%	75%

1. Use Table 6-1 to find the saturation mixing ratio (m_s) of air samples A and B.

$$A: m_s = \text{_____} \text{ g kg}^{-1}$$

$$B: m_s = \text{_____} \text{ g kg}^{-1}$$

2. What are the mixing ratio (m) values of samples A and B?

$$A: m = \text{_____} \text{ g kg}^{-1}$$

$$B: m = \text{_____} \text{ g kg}^{-1}$$

3. Use the temperature and mixing ratio data to plot the two air samples in Figure 6-1.

Assume that equal parts of these two air masses mix together. The resulting temperature is halfway between 10°C and 38°C, and the mixing ratio is halfway between the values that you found for the two samples.

Table 6-1
Saturation Mixing Ratio (g kg^{-1})
at Sea-Level Pressure
as a Function of Dry-Bulb
Temperature ($^{\circ}\text{C}$)

($^{\circ}\text{C}$)	(g/kg)
-40	0.118
-35	0.195
-30	0.318
-25	0.510
-20	0.784
-18	0.931
-16	1.102
-14	1.300
-12	1.529
-10	1.794
-8	2.009
-6	2.450
-4	2.852
-2	3.313
0	3.819
2	4.439
4	5.120
6	5.894
8	6.771
10	7.762
12	8.882
14	10.140
16	11.560
18	13.162
20	14.956
22	16.963
24	19.210
26	21.734
28	24.557
30	27.694
32	31.213
34	35.134
36	39.502
38	44.381
40	49.815

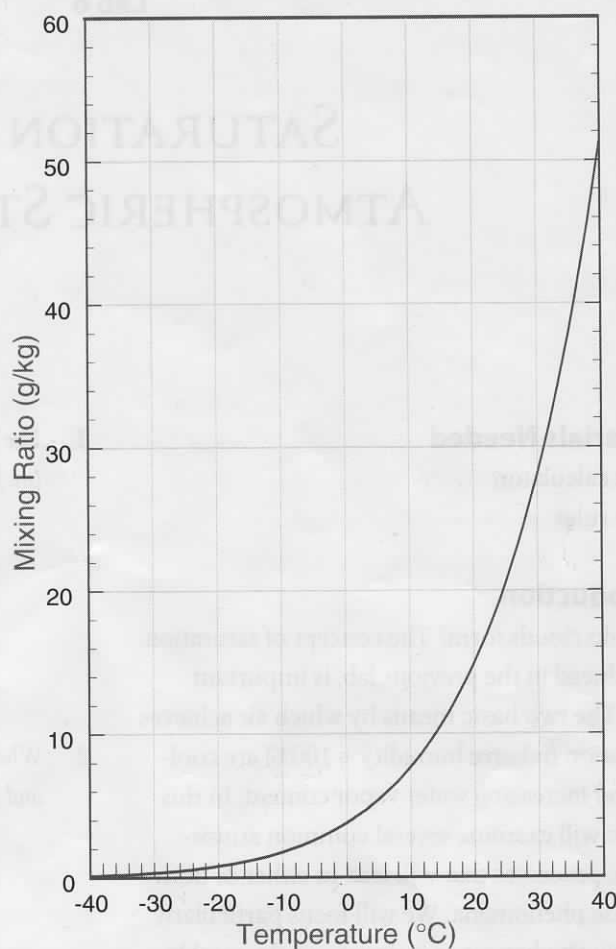


Figure 6-1

4. Plot the mixed air sample in Figure 6-1.

5. What is the new temperature? _____

6. Saturation mixing ratio? _____

7. Mixing ratio? _____

8. Relative humidity? _____

9. The nonlinear relationship between temperature and saturation makes it possible for two unsaturated air samples to mix and form a new saturated air sample. Jet contrails are one example of this as warm exhaust containing moisture mixes with cold air. Some contrails disappear quickly, others linger. What do you think determines the life span of a contrail?

Adiabatic Cooling

Saturation occurs more commonly when air is cooled to its dew-point temperature than when air masses mix. In nature, this cooling often occurs when air rises. Consider a parcel of air that is forced to rise and does not mix with the surrounding environment. As the parcel rises atmospheric pressure around it decreases, allowing the parcel to expand (Figure 6-2). Since the parcel has the same number of molecules but occupies more volume, its average internal energy (i.e., temperature) decreases.

Air can be forced to rise because of:

- Intense surface heating.
- The "collision" or convergence of surface air masses.
- The contact of air masses of unlike temperature along warm and cold fronts.
- A topographic barrier (such as a mountain range).
- Upper-air divergence (a concept we will discuss in a later lab).

As an unsaturated air parcel rises its temperature will decrease at the *dry adiabatic lapse rate* (DALR) of approximately 10°C per kilometer.

10. Calculate the temperature of an unsaturated air parcel at 100-m increments as it is forced to rise from the earth's surface, where its temperature is 35°C .

Height (m)	Temperature
1000 (1 km)	_____
900	_____
800	_____
700	_____
600	_____
500	_____
400	_____
300	_____
200	_____
100	_____
Earth's surface	<u>35°C</u>

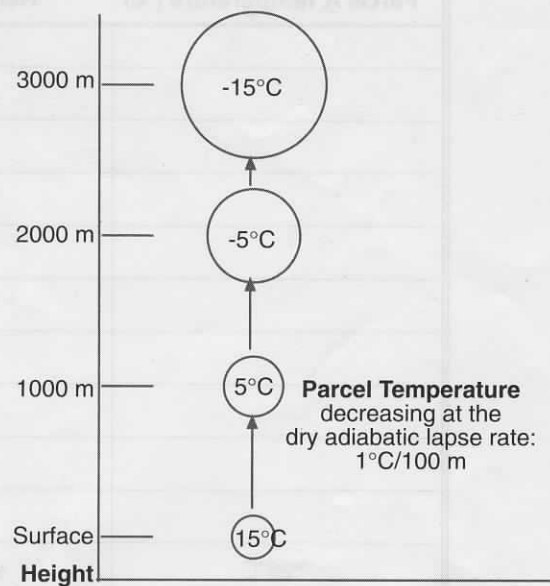


Figure 6-2. Adiabatic cooling of an unsaturated air parcel.

It is possible that rising air parcels will cool to the dew-point temperature. When this happens, a parcel becomes saturated and condensation or clouds form. The height at which this occurs is called the *lifting condensation level* (LCL). If the air parcel rises above the LCL, it cools at a slower rate referred to as the *saturated, moist or wet adiabatic lapse rate* (WALR). The wet adiabatic lapse rate ranges between 5°C and 9°C per kilometer. It is slower than the dry adiabatic lapse rate because latent heat is released within the parcel as water vapor condenses, a warming that partially offsets adiabatic cooling. The WALR varies because the amount of condensation depends both on the amount of water vapor in the parcel and on atmospheric pressure.

Table 6-2

Parcel A Temperature (°C)	Height (km)	Parcel B Temperature (°C)
	5.0	
	4.5	
	4.0	
	3.5	
	3.0	
	2.5	
	2.0	
	1.5	
	1.0	
	0.5	
28°C	surface	10°C

11. Consider air parcel A, which is 28°C at the surface. It is forced to rise to 5 km. The lifting condensation level is 1.5 km, above which the parcel cools at an average wet adiabatic lapse rate of 5°C per kilometer. Fill in the left column of Table 6-2 indicating the parcel's changing temperature.
12. Now consider air parcel B with a surface temperature of 10°C that is forced to rise. It too reaches the LCL at 1.5 km, but the average wet adiabatic lapse rate in this case is 7°C per kilometer. Fill in the right column of the table indicating temperature change in the lower 5 kilometers.
13. Why would the warmer parcel cool at a slower rate between 1.5 and 5 km?

In the previous example, the lifting condensation level was provided for you. We can calculate the LCL from surface temperature and dew-point temperature. The LCL generally occurs at the height where air has cooled to the dew-point temperature. Like temperature, dew-point temperature also decreases with height because of decreasing pressure. As a rule of thumb, dew point drops approximately 2°C per kilometer.

Consider the example shown in Figure 6-3. An air parcel at sea level has the following characteristics: **Temperature = 30°C, Dew-point temperature = 14°C, Pressure = 1010 mb.** If the parcel is forced to rise, its temperature drops at the dry adiabatic lapse rate (10°C per kilometer) until it becomes saturated. The drop in dew-point temperature (2°C per kilometer) can be found by following a *mixing ratio line* associated with dew-point temperature at the surface. The point where the lines intersect is the lifting condensation level.

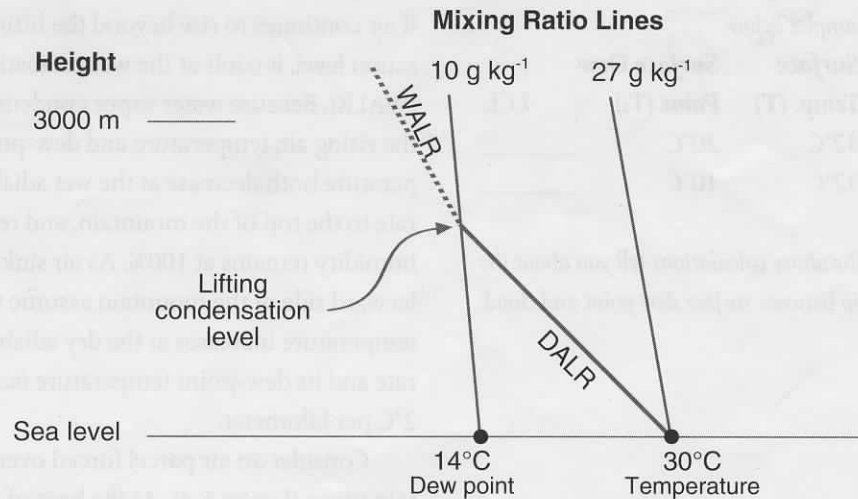


Figure 6-3. Lifting condensation level.

14. Fill in Table 6-3 below to show the temperature and dew-point temperature of the forced air parcel.

If we know the surface temperature and dew point of an air parcel, we can use the following equation to calculate the point of intersection:

$$T_d - 2x = T - 10x$$

The left side of the equation represents the decrease of dew-point temperature with height; the right side represents the temperature decrease with height. At some height x these expressions are equal. Solving algebraically, we find:

$$10x - 2x = T - T_d$$

$$8x = 30 - 14$$

$$8x = 16$$

Now, recalling the units in our equation:

$$x = \frac{16^\circ\text{C}}{8^\circ\text{C km}^{-1}}$$

$$x = 2 \text{ km}$$

Now that you understand the concept, here's a shortcut:

$$\text{LCL (km)} = \frac{T(^{\circ}\text{C}) - T_d(^{\circ}\text{C})}{8^\circ\text{C km}^{-1}}$$

or

$$\text{LCL (meters)} = 125 [T(^{\circ}\text{C}) - T_d(^{\circ}\text{C})]$$

Table 6-3

Height (m)	Temperature ($T^{\circ}\text{C}$)	Dew-Point Temperature ($T_d^{\circ}\text{C}$)
2000		
1500		
1000		
500		
earth's surface	30°C	14°C

15. Calculate the lifting condensation level (LCL) for the two examples below.

	Surface Temp. (T)	Surface Dew Point (T _d)	LCL
Example A	32°C	20°C	_____
Example B	32°C	10°C	_____

16. What do the above calculations tell you about the relationship between surface dew point and cloud height?

Optional Exercise

17. On a day with convective cumulus clouds, measure the outdoor temperature and dew-point temperature and determine the approximate height of the cloud bases.

A Mountain Example

If air continues to rise beyond the lifting condensation level, it cools at the wet adiabatic lapse rate (WALR). Because water vapor condenses out of the rising air, temperature and dew-point temperature both decrease at the wet adiabatic lapse rate to the top of the mountain, and relative humidity remains at 100%. As air sinks on the leeward side of the mountain assume that its temperature increases at the dry adiabatic lapse rate and its dew-point temperature *increases* at 2°C per kilometer.

Consider an air parcel forced over a mountain range (Figure 6-4). At the base of the windward side of the mountain the temperature is 25°C and the dew-point temperature is 13°C.

18. Fill in the spaces for temperature and dew-point temperature at various heights on the windward and leeward sides of the mountain. Assume that the wet adiabatic lapse rate is 5°C per kilometer. (Note that temperature and dew point drop together when a saturated air parcel rises and water vapor condenses.)

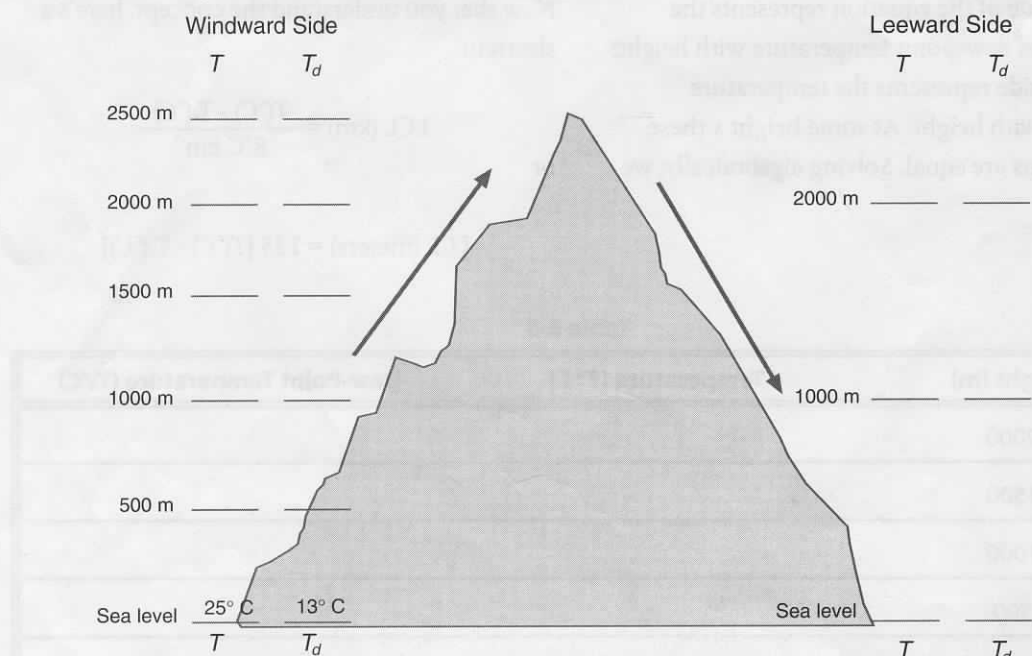


Figure 6-4. Temperature and dew point change as air parcel is forced over a mountain.

19. At what height would cloud bases form?
_____ meters
20. How do the sea-level temperature and dew point on the leeward side compare with the sea-level temperature and dew point on the windward side?
21. Which side of the mountain is more often cloudy and which side is more often clear?

Atmospheric Stability

In our discussion so far we have made two assumptions: (a) that certain processes force air to rise from the earth's surface and (b) that rising air does not mix substantially with the surrounding atmosphere. Once an initial lifting force ceases, the fate of a rising air parcel depends on the state of the atmosphere through which it rises, or *atmospheric stability*. Consider the three examples in Figures 6-5 through 6-7. In each example the environmental temperature is

shown with a solid line, and the parcel temperature follows the dry and wet adiabatic rates of cooling (dashed lines). In each example, the surface temperature is 30°C, surface dew point is 14°C, and the wet adiabatic lapse rate is 5°C/km.

Weather balloons measure the temperature above the earth's surface. These measurements allow us to calculate the rate at which the surrounding environmental temperature changes with height—the *environmental lapse rate* (ELR). The atmosphere is considered *absolutely stable* if an air parcel that is forced aloft cools faster than the surrounding environment (Figure 6-5), that is, when the parcel rate of cooling (the dry or wet adiabatic lapse rate) is faster than the environmental lapse rate. If the force lifting the parcel continued to act, condensation would eventually occur, but clouds would be layered clouds without much vertical development. If the force ceased, the parcel would have a tendency to sink.

In an *absolutely unstable* atmosphere a parcel that is initially displaced will continue to rise because it is warmer and more buoyant than its surroundings at the displaced height. In this case, the parcel rate of cooling is slower than the

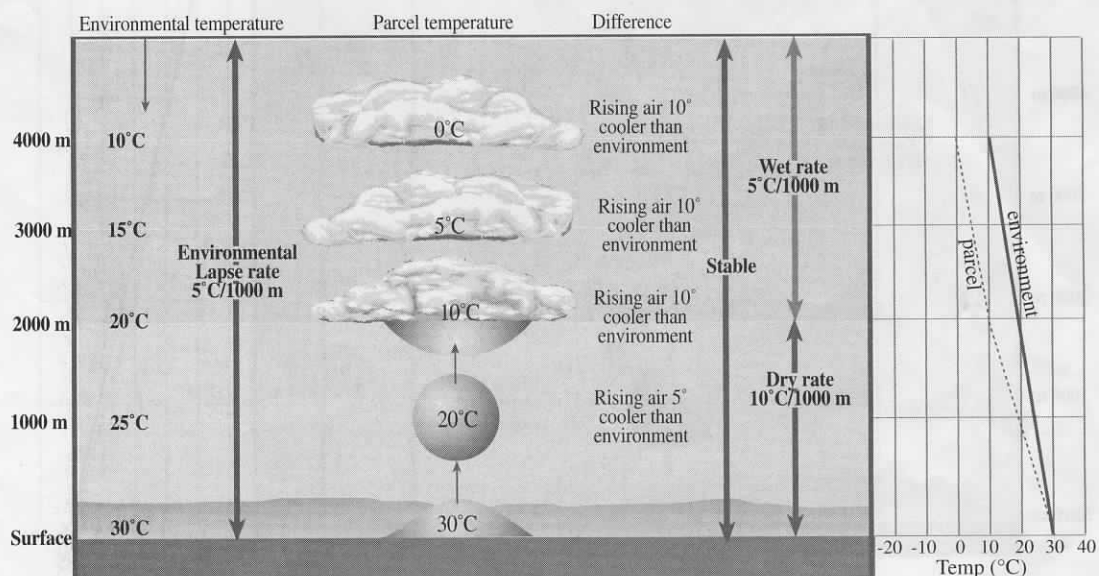


Figure 6-5. Absolutely stable parcel.

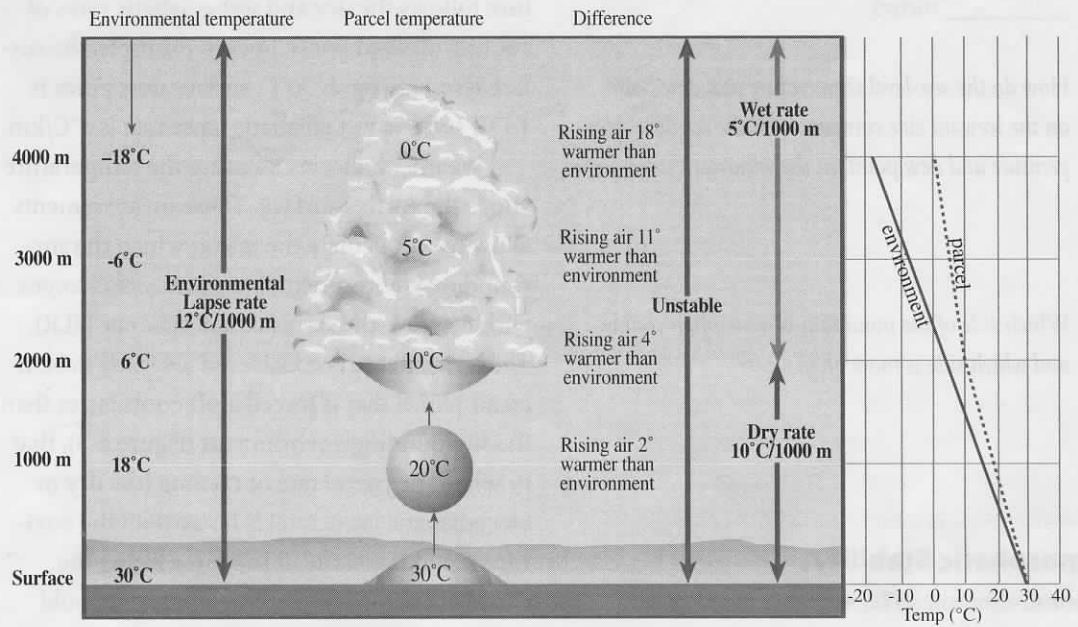


Figure 6-6. Absolutely unstable parcel.

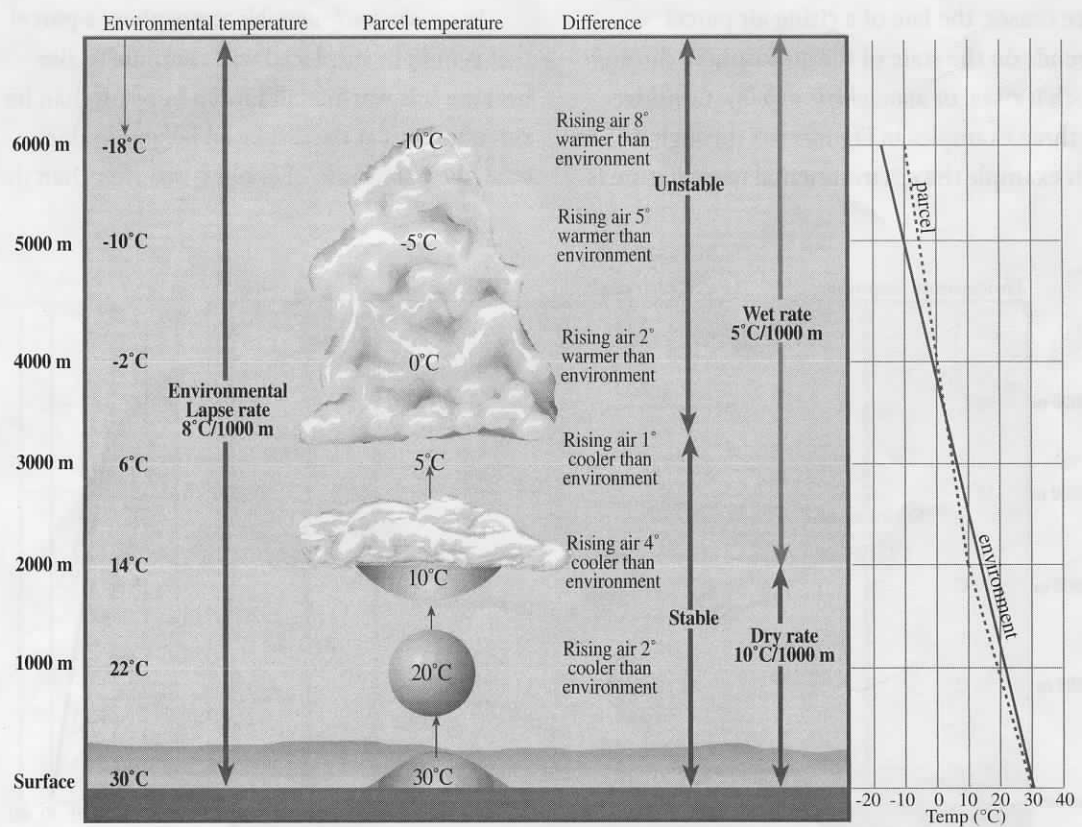


Figure 6-7. Conditionally unstable parcel.

environmental lapse rate. If the air parcel rises to its lifting condensation level, clouds with vertical development will form as the buoyant air rises on its own accord (Figure 6-6). Thunderstorms require such conditions.

A more complex situation is depicted in Figure 6-7, where the lower portion of the atmosphere is stable (the parcel temperature is cooler than the surrounding environment), and higher elevations are unstable (the parcel temperature is warmer than the surrounding environment). In this situation the atmosphere can be described as *conditionally unstable*. It exists, of course, because the wet and dry adiabatic lapse rates are different, and the environmental lapse rate is between the two (DALR > ELR > WALR). Again, notice that the resulting cloud patterns reflect the tendency for vertical development only in unstable layers of the atmosphere (above 3300 meters in this example).

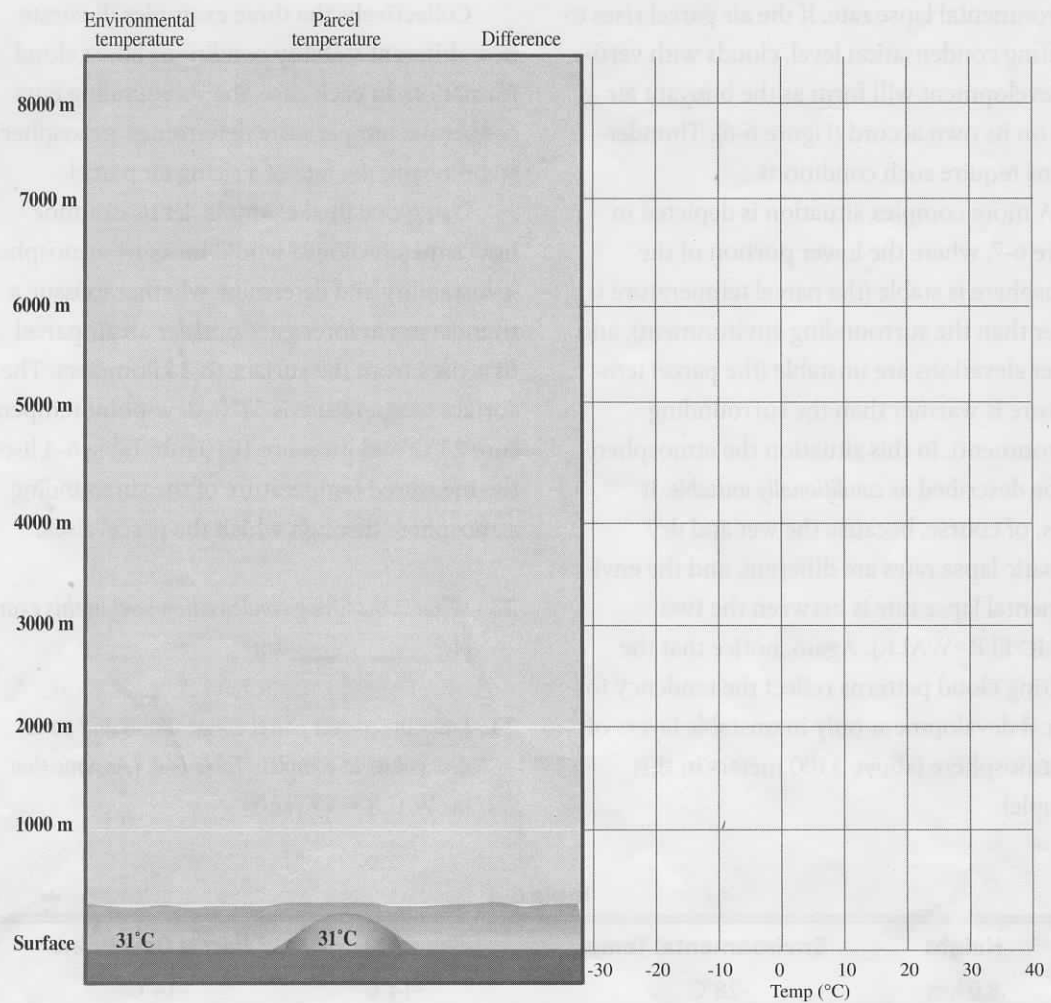
Collectively, the three examples illustrate how different stability conditions affect cloud formation. In each case, the surrounding environmental temperature determines atmospheric stability and the fate of a rising air parcel.

Using one final example, let us examine how a meteorologist would measure atmospheric instability and determine whether to issue a thunderstorm forecast. Consider an air parcel that rises from the surface to 3 kilometers. The surface temperature is 31°C, dew-point temperature 23°C, and pressure 1010 mb. Table 6-4 lists the measured temperature of the surrounding atmosphere through which the parcel rises.

22. What is the lifting condensation level in this example? _____ km
23. Enter the correct parcel temperatures and parcel dew points to complete Table 6-4. (Assume that the WALR = 5°C/km).

Table 6-4

Height	Environmental Temp.	Parcel Temp.	Parcel Dew Point
8.0 km	-28°C	-14°C	-14°C
7.5 km	-25°C	-11.5°C	-11.5°C
7.0 km	-22°C	-9°C	-9°C
6.5 km	-19°C	-6.5°C	-6.5°C
6.0 km	-16°C	-4°C	-4°C
5.5 km	-13°C	-1.5°C	-1.5°C
5.0 km	-10°C	1°C	1°C
4.5 km	-7°C	3.5°C	3.5°C
4.0 km	-3°C	6°C	6°C
3.5 km	1.5°C		
3.0 km	6°C		
2.5 km	10.5°C		
2.0 km	15°C		
1.5 km	19°C		
1.0 km	23°C		
0.5 km	27°C		
surface	31°C	31°C	23°C



24. Using the diagram above and Figures 6-5 through 6-7 as models:
- Plot the parcel and environmental temperatures from Table 6-3.
 - Show the lifting condensation level and draw likely cloud development.
 - Identify any stable or unstable layers.
25. Would you forecast thunderstorms under these conditions?

Review Questions

How is saturation achieved to form most clouds?

What is the difference between a stable and an unstable atmosphere? Which of these two conditions favors vertical cloud development and why?