

Chapter 11

Sounding Analysis

Introduction

The term "sounding" is used in meteorology to describe a vertical plot of temperature, dew point temperature, and wind above a specific location. It provides a picture of how these parameters change as you rise into the atmosphere. These vertical data are used to determine atmospheric stability, precipitation type, and a glimpse of cloud location at one point in time.

The primary source for these data is the radiosonde, an instrument package that is launched at numerous locations around the world at 00 UTC and 12 UTC. This coordinated launch provides a three-dimensional picture of the structure of the atmosphere twice a day. The actual launch time is one hour prior to nominal observation time due to the time it takes for the instrument to rise into the stratosphere. On a good day, the radiosonde will reach the 10 mb level.

Another source of sounding data is satellite sounders. Due to their asynoptic nature, these data are usually integrated via data assimilation into computer forecast models. These soundings tend to be smoother than radiosonde profiles.

Soundings are also available from computer forecast models. At most forecaster workstations you can click on a point on a map and get a forecast sounding for that location. This allows you to evaluate future soundings for precipitation type or stability.

Radiosonde Data: A radiosonde is an instrument package that measures temperature, relative humidity and atmospheric pressure (RAOB). These data are transmitted back to the launch site by radio. Temperature, relative humidity and pressure can be used to calculate dew point temperature and the height of a pressure level, and a variety of other variables.

Tracking the instrument by radio-direction finding provides the elevation angle of the instrument as a function of time. These data, combined with an assumed ascent rate, are used to determine wind direction and wind speed. The next generation of

radiosondes will incorporate global positioning system (GPS) data to more accurately locate the instrument.

Sounding Features

Figure 11-1 shows a typical presentation of sounding data on a plot called a skew-T, log p diagram (discussed later in this chapter). Temperature increases toward the right while height increases and pressure decreases upward.

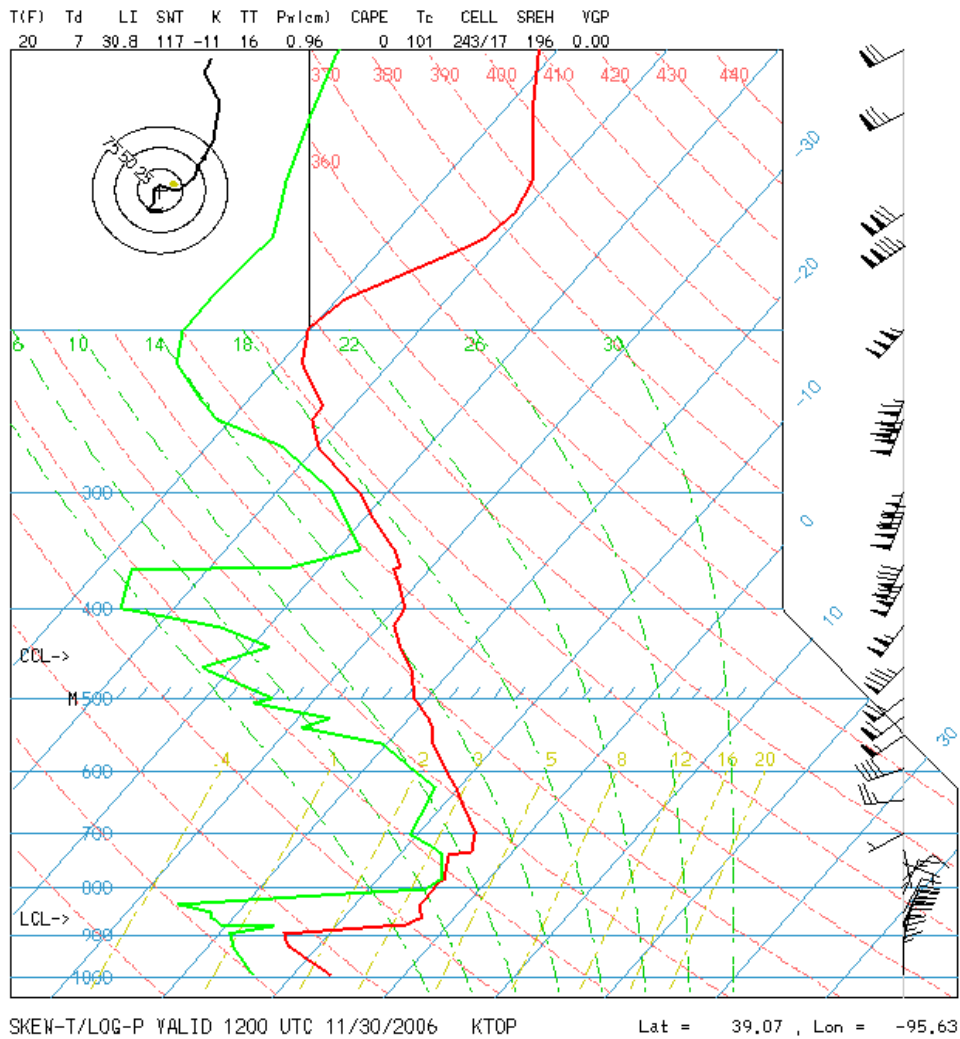


Figure 11-1: Sounding for Topeka at 1200 UTC on 30 November 2006 (source: ucar.rap.edu)

Several features can be seen upon examination of a typical sounding. These features include:

Ambient data: The basic data shown on a sounding (Figure 11-1) are temperature (in red) and dew point temperature (in green). Surface data are found at the bottom of the diagram, in this example around 1000 mb. Examination of the profile helps a forecaster identify several of the features described below.

Winds data are usually plotted on the right in standard station model format. North is assumed to be at the top of the diagram. Changes in wind speed and wind direction with height can be inferred from this plot.

Lapse rate: Lapse rate (Γ or γ) is defined as the slope of the ambient temperature curve. Specifically, $\Gamma = -\partial T/\partial z$, where T is temperature and z is height. The minus sign is used so that a temperature curve decreasing with height is positive. Lapse rate is the negative of the mathematical slope. Lapse rates usually refer to a layer of limited depth.

Several terms are used to describe lapse rate:

- **Normal lapse rate:** This is a positive lapse rate that is associated with the normal decrease in temperature found within the troposphere.
- **Isothermal lapse rate:** This is a lapse rate equal to zero, that is, the temperature does not change with height. The layer from around 830 mb to 790 mb in Figure 11-1 is isothermal.
- **Inversion:** See the next paragraph.
- **Adiabatic lapse rate:** This is the lapse rate associated with an adiabatic process. There are two types of adiabatic lapse rates, dry and moist. These are discussed below.
- **Super-Adiabatic lapse rate:** This is a positive lapse rate with a value greater than the dry adiabatic value.

Inversion: This is a layer with a negative lapse rate. The temperature increases with height, opposite to the normal tropospheric lapse rate. There are several inversions in Figure 11-1 with the most prominent from 900 mb to 885 mb.

Inversions can be caused by fronts (as in Figure 11-1), by subsiding air in a layer, by surface radiation (usually at night), or by turbulence.

Humid layers/clouds: Radiosonde data infrequently measure relative humidity values of 100 percent due to the sensitivity

and response of the instrument. Nevertheless, if the difference between the temperature and dew point temperature (dew point depression) is less than 5°C, it is assumed that clouds are present. In Figure 11-1 there is a cloud layer between 800 mb and 740 mb, some indication of a cloud or moist layer around 615 mb, and another moist layer 350 mb to 270 mb.

Tropopause: The tropopause is the boundary between the troposphere and the stratosphere. It is characterized by an abrupt change in lapse rate. In Figure 11-1 this abrupt change is seen around 200 mb.

Thermodynamic Diagrams

The chart on which sounding data are plotted is called a *thermodynamic diagram*. These diagrams were developed so that operational forecasters and research meteorologist could move air parcels upward and downward to simulate atmospheric motion and evaluate factors such as stability without the need to do any actual thermodynamic calculation. Prior to computer displays, thermodynamic diagrams were plotted by the human hand on a sheet of paper on the order of 2 by 2.5 feet. In the last 20 years, computerized displays of soundings have been developed that reproduce the paper versions and allow the forecaster to manipulate air parcels on a computer display.

The area on thermodynamic diagrams is proportional to energy. The charts typically display isobars, isotherms, and mixing ratio isohumes, and a set of curves that represent changes to due to both dry and moist adiabatic movement. Currently there are two main thermodynamic diagrams used operationally: the skew $T, \log p$ diagram and the tephigram.

The *skew $T, \log p$ diagram* was developed by the U.S. Air Force and is currently used by the U.S. military and the National Weather Service. The x-axis is temperature and the y-axis is $-\ln p$ (negative of the natural log of atmosphere pressure). The x-axis is not orthogonal to the y-axis but is skewed or rotated 45° clockwise to allow better data presentation. This diagram will be discussed in detail later in this chapter.

On a *tephigram* (used by Canadian meteorologists) the coordinates are temperature on the x-axis and $\ln \theta$ (potential temperature) on the y-axis. Figure 11-2 shows an example of a tephigram.

Several diagrams have been developed over the year and have been used both operationally and in research.

- Stüve Diagram: (also known as pseudoadiabatic chart)
 - o x-axis: temperature (T)
 - o y-axis: $-p^{R/c_p}$
 - o where p=pressure, R=gas constant, c_p specific heat at constant pressure
- Aerogram: (previously used by the U.S. Navy)
 - o x-axis: natural log of temperature
 - o y-axis: $-T \ln p$
- Emagram: (orthogonal skew T, log p diagram)
 - o x-axis: temperature
 - o y-axis: $-\ln p$

If you understand how to use one type of diagram, transitioning to the other charts is relatively easy.

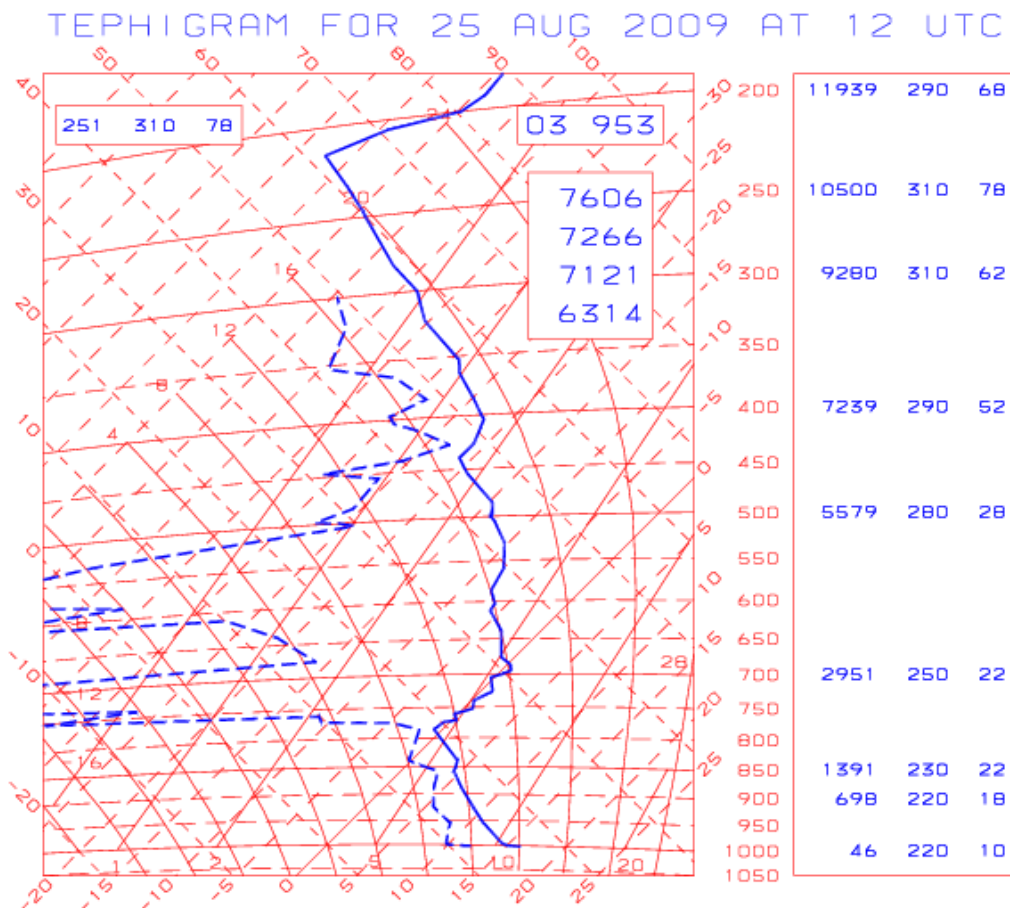


Figure 11-2: Example of a tephigram

Skew T, log p Diagram

As noted above, the skew T, log p diagram uses a coordinate system with temperature along the x-axis and $-\ln p$ along the y-axis, with the x-axis rotated 45° clockwise to improve data presentation. Pressure is plotted as $-\ln p$ so that the vertical scale approximates a linear height scale.

In addition to temperature and pressure lines, other lines on the diagram represent mixing ratio, and the dry and moist adiabatic processes. Figure 11-3 shows a skew T log p.

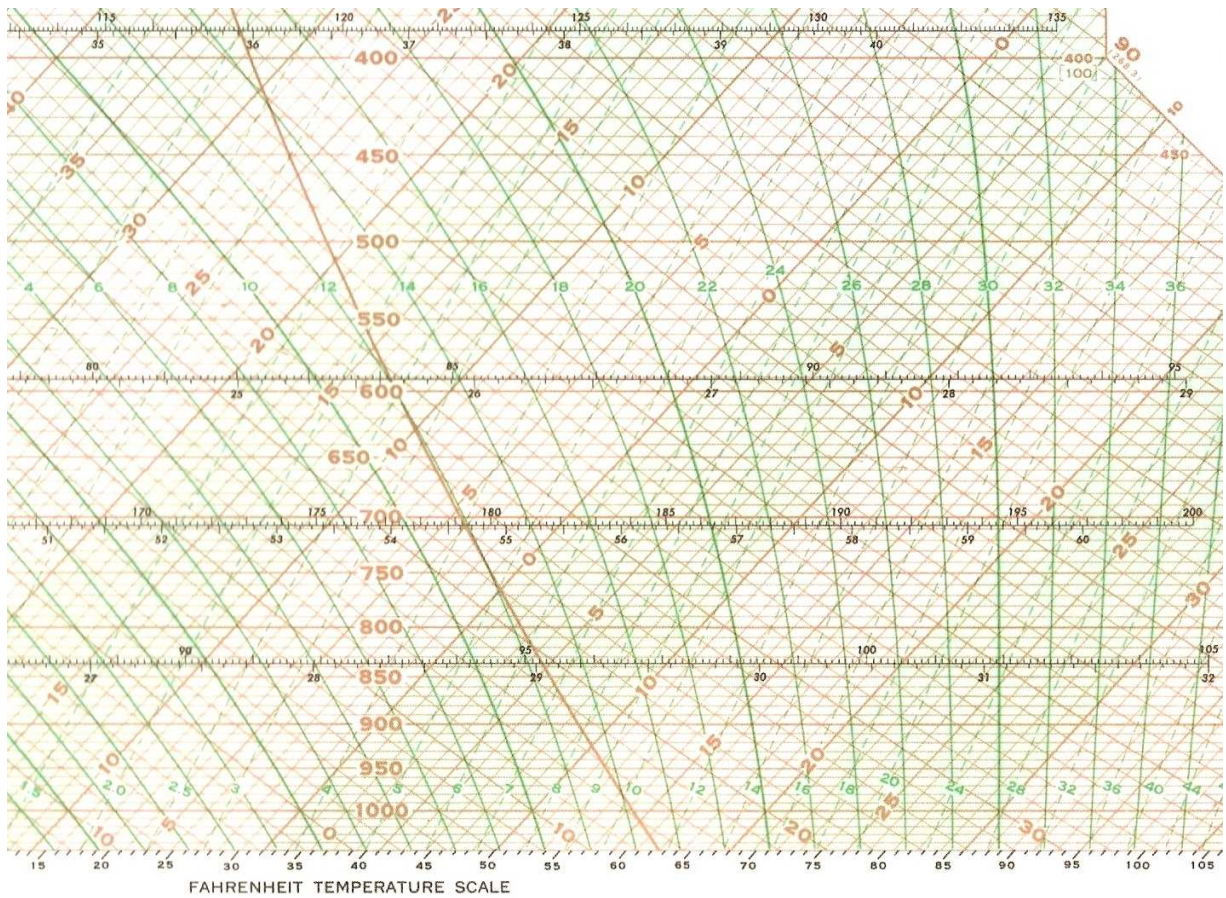


Figure 11-3: Skew T, log p Diagram (U.S.A.F chart)

On Figure 11-3 pressure lines are horizontal brown lines labeled in millibars (mb). Values range from 1050 mb at the bottom of the chart to 400 mb at the top of the figure with lines every 10 millibars. This interval allows pressure to be plotted to the nearest millibar. Temperature lines are brown lines, run from

lower left to upper right, and are labeled in degrees Celsius at one degree intervals. Values over 30°C are found in the lower right; values around -40°C are seen in the upper left of the figure.

The green dashed lines that extend from lower left to upper right are lines of constant saturation mixing ratio labeled in grams per kilogram. *Mixing ratio* is the ratio of the density of the water vapor in an air parcel to the density of the dry air. It is measure of the humidity in the air. Mixing ratio is a dimensionless quantity on the order of 0.010. In order to avoid talking in decimals, the value is multiplied by 1000 and the unit's denominator is expressed in kilograms. Thus, a mixing ratio value of 0.010 is expressed as 10 g/kg.

If you plot a point corresponding to the temperature and pressure of an air parcel on the diagram, you can then read the saturation mixing ratio directly from the chart. For example, if the temperature is 10°C and the pressure is 1000 mb, the saturation mixing ratio is read as 7.7 g/kg.

If you plot a point corresponding to the dew point temperature and pressure of an air parcel on the diagram, you can then read the actual mixing ratio directly from the chart. For example, if the dew point temperature is 0°C and the pressure is 1000 mb, the actual mixing ratio is read as 3.8 g/kg.

If you need a relative humidity (RH) value for an air parcel, you can calculate it as the ratio of the actual mixing ratio to the saturation mixing ratio (times 100 percent). In our example, $RH = (3.8/7.7) * 100\% = 49.4\%$.

The brown lines that run from lower right to upper left are lines of constant potential temperature (θ). They also represent the change in temperature and pressure associated with dry adiabatic movement. As a result, these lines are commonly called *dry adiabats*. These lines are labeled in degrees Celsius. Use of these lines is discussed below.

The solid green lines that turn toward the left as they rise are lines of constant wet-bulb potential temperature (θ_w). They also represent the change in temperature and pressure associated with moist adiabatic lift. As a result, these lines are commonly called *moist or saturation adiabats*. These lines are labeled in degrees Celsius. Use of these lines is discussed below.

Parcel Method

When evaluating vertical movement an imaginary volume of air, called a *parcel*, is used. Thermodynamic properties are assigned to this parcel and allowed to change as the parcel moves. It is usually assumed that the parcel does not exchange mass or energy with the surrounding environment, that is, the parcel undergoes an *adiabatic process*. In an adiabatic process, changes in internal energy (as measured by parcel temperature) are balanced by the work done to or done by the parcel as the pressure changes.

The parcel method is a good first approximation to vertical movement of air. However, it does not take into consideration several factors that impact real parcels:

- Gradients of vertical perturbation pressure.
- Entrainment, or the mixing of the parcel with the surrounding environmental air.
- Compensating subsidence within the surrounding atmosphere.
- Weight of the condensate that is produced when saturated air is lifted.

Discussion of these factors is beyond the scope of this chapter.

The parcel method is used to evaluation stability. Assessment is based on the difference between the temperature of the lifted air parcel and the temperature of the surrounding environment, that is, buoyancy. Stability is discussed later in this chapter.

Adiabatic Lift

During adiabatic lift an air parcel will expand and cool as it rises and will warm and compress as it sinks. It assumes that the pressure of the air parcel adjusts to the environmental pressure as it up and down.

For an unsaturated air parcel the potential temperature (θ) of the parcel remains constant during vertical movement. You can calculate the temperature changes using the formula for potential temperature: $\theta = T \cdot (p_0/p)^{R/c_p}$, where:

- T - temperature (K) at level of interest
- p_0 - 1000 mb

- p - pressure (mb) at level of interest
- R - universal gas constant (287.04 j/kg/K)
- c_p - specific heat at constant pressure (1004 j/kg/K)

The change in temperature during dry adiabatic movement is known as the *dry adiabatic lapse rate* (DALR or Γ_d) and is equal to 9.8°C/km.

In an operational setting performing calculations can be time consuming and tedious. The thermodynamic diagram was designed to simplify this process. If you know the temperature and pressure of an unsaturated parcel, you can estimate the change in its temperature as it rises from one pressure level to another by moving along a dry adiabat. For example, on Figure 11-4, starting with a temperature of 30°C at 1000 mb, move to 900 mb by sliding along the dry adiabat. The temperature at that point is 21°C. This is the temperature that the air parcel will cool to as it moves from 1000 mb to 900 mb.

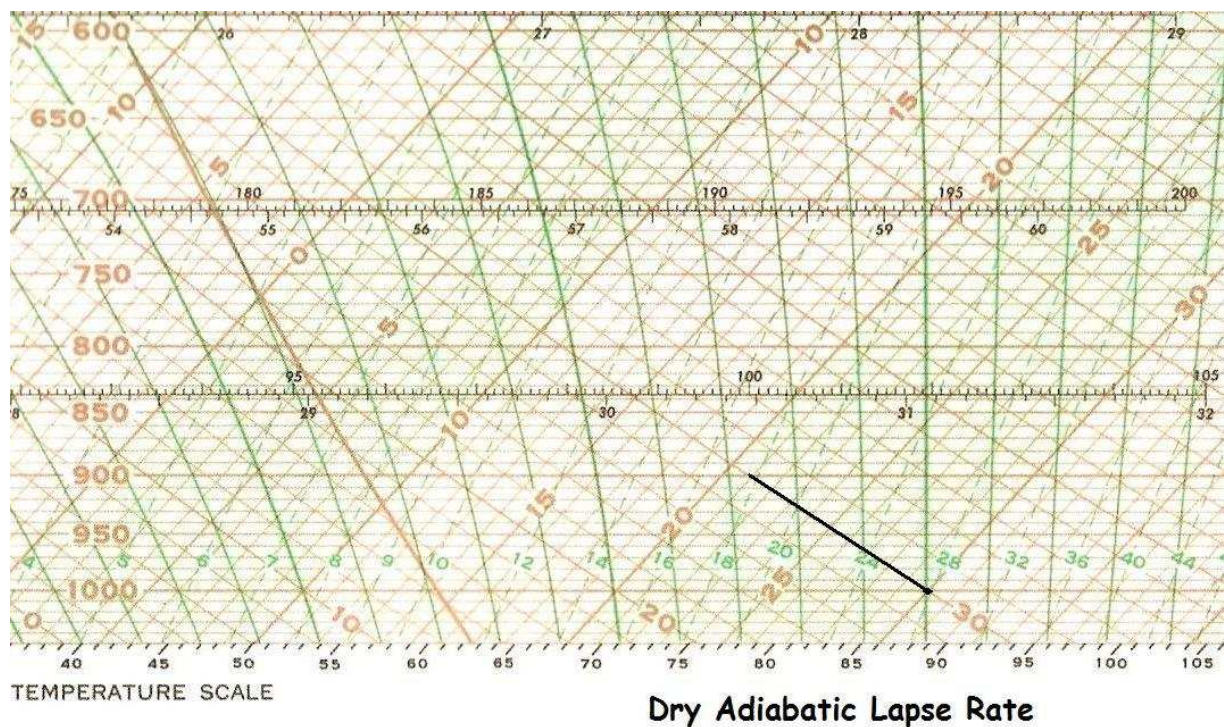


Figure 11-4: Example of Dry Adiabatic Lift

For saturated air as the parcel rises it will cool and expand. However, in order to maintain saturation at 100% RH, excess moisture must be condensed out. The condensation process

releases latent heat that counteracts the cooling. The net result is a cooling rate that is smaller in magnitude than the rate associated with dry adiabatic lift. The change in temperature during moist or saturated adiabatic movement is known as the *moist or saturated adiabatic lapse rate* (MALR/SALR or Γ_s). The MALR/SALR ranges from around 4°C/km at low levels to 9°C/km at high levels.

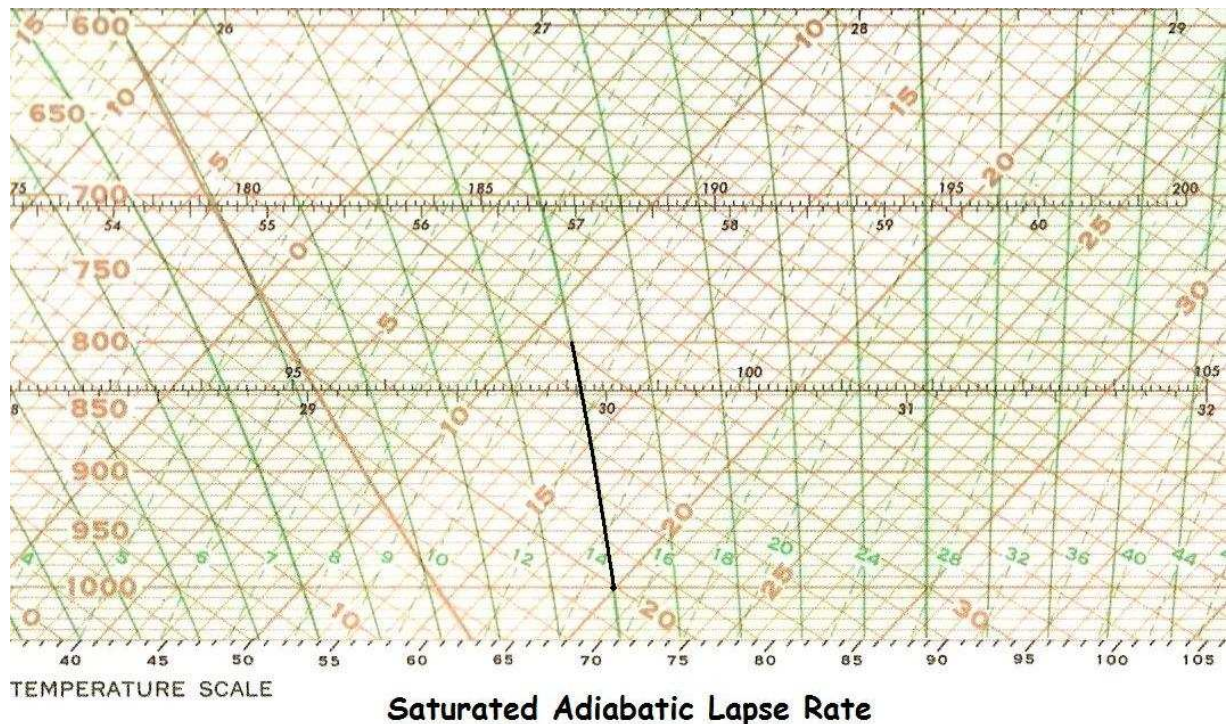


Figure 11-5: Example of Saturated Adiabatic Lift

As an example, if you have a saturated air parcel with a temperature of 20°C at 1000 mb (Figure 11-5) and want to lift it to 800mb, move along the moist adiabat to 800 mb and find that the temperature of the parcel has cooled to 11.4°C.

Graphic Manipulations

Thermodynamic diagrams allow you to easily move air parcel up and down and estimate parcel temperature changes. They also allow the determination of quantities such as lifting condensation level, wet-bulb temperature, and wet-bulb potential temperature.

As part of stability assessment you may have to lift an air parcel from near the Earth's surface upward to some level higher in the troposphere.

- If the parcel is unsaturated, the initial lift will cool the parcel at the DALR without changing the actual moisture content of the parcel. At some point the parcel becomes saturated and the cooling rate changes to the MALR.
- If the parcel is saturated, the parcel will cool at the MALR through the entire lifting process.

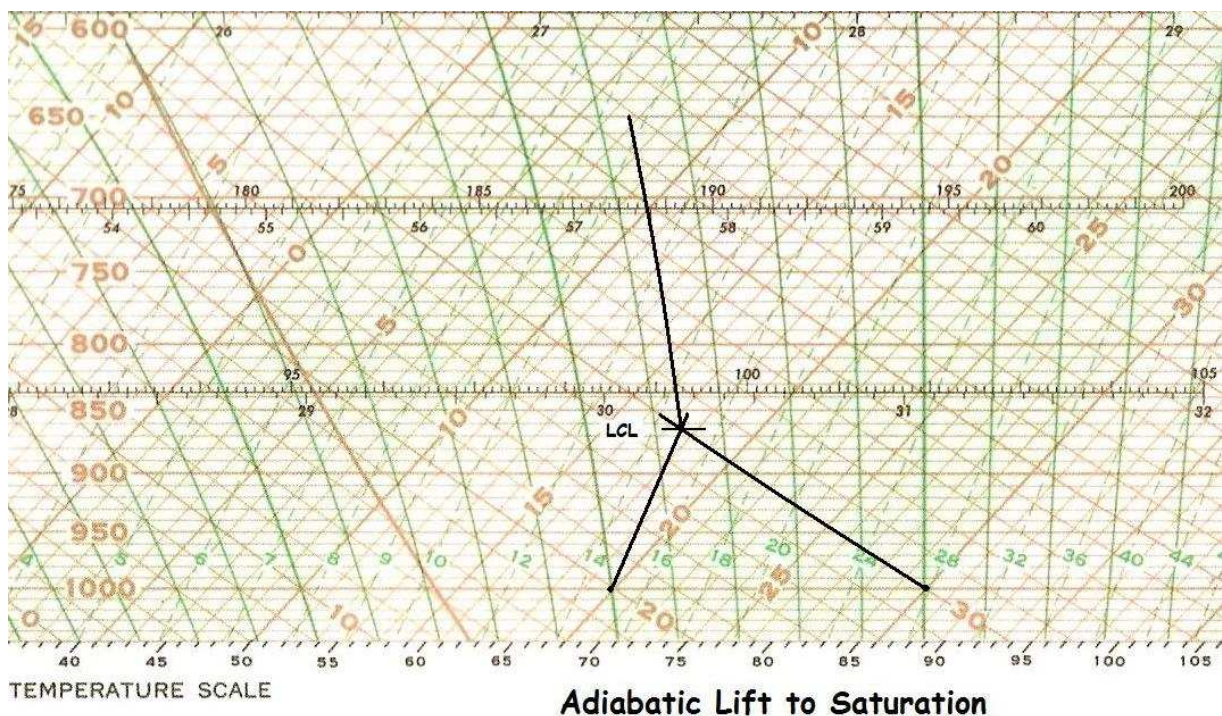


Figure 11-6: Example of Adiabatic Lift

Figure 11-6 illustrates this process. The air parcel starts at 1000 mb with a temperature of 30°C and a dew point temperature of 20°C. The temperature cools along the dry adiabat. Because the moisture content does not change, the dew point temperature changes along a mixing ratio line. At 900 mb the parcel has cooled to 21°C while the dew point temperature has decreased to 18.2°C even though the actual moisture content has not changed.

Around 865 mb the dry adiabat through the temperature crosses the mixing ratio line through the dew point temperature. At this level the lifted parcel is saturated, that is, the temperature

equals the dew point temperature. This level is called the *lifting condensation level* (LCL). This is the point where condensation will be produced by the rising parcel and clouds will likely form. This is also the point where the cooling rate changes to the MALR. If this parcel continues upward to 650 mb, the temperature and dew point cool to 6.7°C.

Figure 11-7 shows what happens if you lift the surface air parcel upward. The parcel is initially unsaturated and cools at the DALR. It becomes saturated at its LCL and continues to cool at the MALR above that point. If the lifted parcel becomes warmer than the environment, the point at which the parcel first becomes positively buoyant is called the *level of free convection* (LFC). This is the level above which the lifted parcel will accelerate upward without forced lift. It is marked on Figure 11-7. Not all soundings have a LFC.

As the parcel continues upward above the LFC, it will eventually become cooler than the environment, usually around the tropopause. This point is the equilibrium level (EL). This point is shown on Figure 11-7. Between the LFC and EL the lifted parcel is warmer than the environment or positively buoyant. The area between the lifted parcel curve and the environment represents the buoyant energy available to accelerate the parcel upward. This area is marked P on Figure 11-7.

Wet-Bulb Temperature: If you lift an air parcel upward to its LCL and then slide downward along a moist adiabat to the original pressure, the temperature at that level is the adiabatic (or pseudo) wet-bulb temperature (T_w) for the parcel. If you continue along the moist adiabat to the 1000 mb level and read the temperature, you have the wet-bulb potential temperature (θ_w). In Figure 11-7, the wet-bulb temperature and the wet-bulb potential temperature are the same.

The adiabatic wet-bulb temperature is slightly different from the isobaric wet-bulb temperature. The latter is determined by cooling a parcel adiabatically to saturation *at constant pressure* while evaporating water into it as all latent heat is absorbed by the parcel. This latter process is similar to determining the wet-bulb temperature with a sling psychrometer.

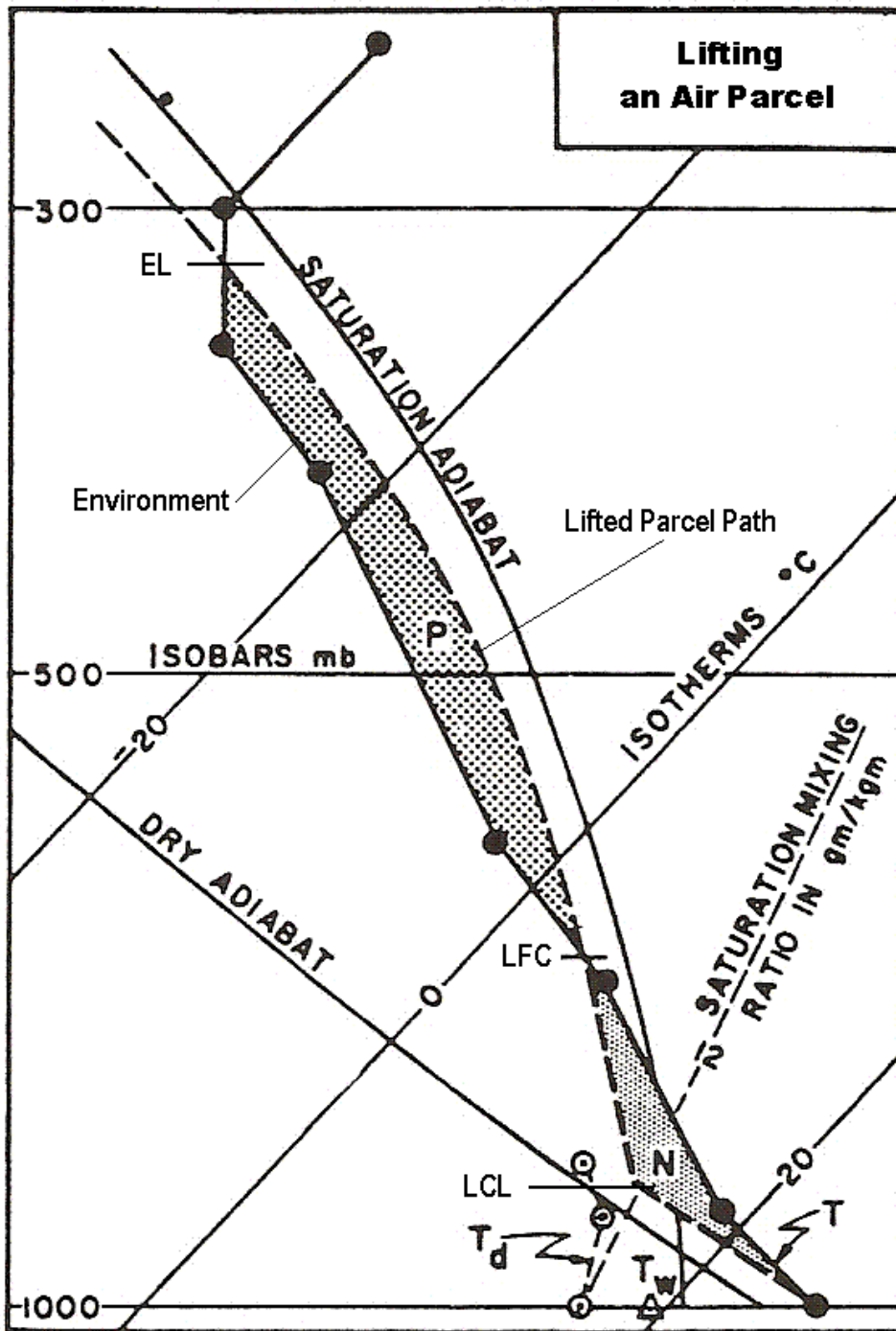


Figure 11-7: Example of Lifting a Parcel
(source: modified USAF diagram)

Another variable that can be determined from a thermodynamic chart is equivalent temperature (T_e). It is found using the following process:

- Dry adiabatic expansion until the parcel becomes saturated
- Moist adiabatic lift until all moisture is precipitated out of the parcel
- Dry adiabatic compression to the original pressure

If the air parcel is brought to 1000 mb after the last step, the resulting temperature is the equivalent potential temperature (θ_e).

Parcel Method Summary

The following summarizes the graphical manipulations of air parcel on a thermodynamic diagram:

For unsaturated air:

- The parcel mixing ratio remains constant.
- Rising parcels cool at the dry adiabatic rate, that is, they follow the DALR upward on the diagram.
- Sinking parcels warm at the dry adiabatic rate, that is, they follow the DALR downward on the diagram.

For saturated air:

- Rising parcels cool at the moist adiabatic rate, that is, they follow moist adiabats upward on the diagram.
- For rising parcels, the mixing ratio decreases as the parcel rises and the parcel remains saturated (temperature equals dew point temperature).
- Sinking parcels warm at the dry adiabatic rate, that is, they follow the dry adiabat downward on the diagram.
- For sinking parcels, the mixing ratio remains constant and the parcel becomes unsaturated.
- If moisture is evaporated into a saturated parcel as it sinks in order to maintain saturation, the parcel will follow a moist adiabat downward on the diagram (assuming the energy is taken from the air for evaporation).

Stability

If you are forecasting thunderstorms, you will be interested in the stability of the vertical column. Assessing this instability is a major part of the forecasting process. In a general physical system *instability* refers to a magnification (or distortion) of the initial state when a small disturbance or perturbation is introduced into the system. Essentially, the unstable system moves away from its initial state.

Figure 11-8 illustrates this for a ball and bowl system. In Part A, a ball is balanced on top of a bowl that is inverted. If the ball is moved a small distance (or perturbed) it will accelerate away from its initial position. This situation is considered **unstable**. In Part B, the ball is at the bottom of a bowl. If the ball is moved a small distance and released, it will return to its original location after a short period of oscillation. This situation is considered **stable**. In Part C, the ball located on a flat surface. If it is moved a short distance, it will remain at its new location. This situation is called **neutral stability**.

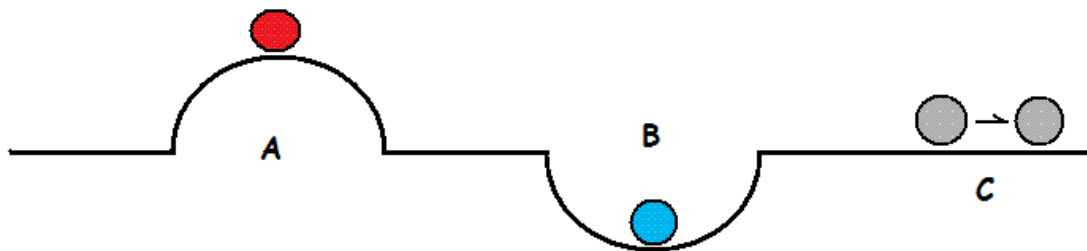


Figure 11-8: Example of Three Types of Instability

In terms of vertical movement in the atmosphere, this basic definition of stability can be interpreted in terms of buoyancy. If a parcel of air is lifted adiabatically to some level in the atmosphere, and:

- The lifted parcel is warmer than the environment, it will accelerate upward; this is considered an *unstable* situation.
- The lifted parcel is cooler than the environment, it will sink back to its original level; this is considered a *stable* situation.
- If the lifted parcel temperature is the same as the environment, it will remain at that level; this is *neutral stability*.

Another approach to evaluating stability is to examine the lapse rate of a layer and compare that lapse rate to the adiabatic lapse rates. If Γ_e is the environmental lapse rate of a layer, the stability of that layer is defined as follows:

- Unstable: $\Gamma_e > \Gamma_d$
- Conditional Instability: $\Gamma_d > \Gamma_e > \Gamma_s$
- Stable: $\Gamma_s > \Gamma_e$

For the unstable case, both saturated and unsaturated parcels lifted adiabatically from the base of the layer will always be warmer than any temperature in the layer. Thus, the layer is unstable in terms of the previous definition.

For the stable case, both saturated and unsaturated parcels lifted adiabatically from the base of the layer will always be cooler than any temperature in the layer. Thus, the layer is stable in terms of the previous definition.

For the conditional instability case, a saturated parcel lifted adiabatically from the base of the layer will be warmer than any temperatures in the layer while an unsaturated parcel lifted adiabatically from the base of the layer will be cooler than any temperatures in the layer. Hence, the stability depends upon the saturation within the layer.

Operationally the best way to evaluate instability for a location is to lift an air parcel upward and assess whether the parcel becomes warmer than the environment. This can be done by manually moving air parcel around on a sounding plot. Another method is to use stability indices.

Indices

A *stability index* is an algorithm designed to evaluate the stability properties of a vertical sounding. Over the years numerous indices have been developed including such indices as the Showalter Index, the Lifted Index, the K Index, and SWEAT. When you run an automated sounding program, you typically get a list of over two dozen stability index values. Developing a new stability index is like constructing a better mouse trap. If you develop a better one, the world will beat a path to your door. What you have to realize is: There is more to forecasting

thunderstorms that instability. You also need a mesoscale lifting mechanism and moisture to feed the "mesoscale beast."

Let's look at several stability indices that have been used operationally:

Showalter Index: This index was one of the first stability indices used operationally. It takes an air parcel from the 850 mb level and lifts it adiabatically to 500 mb. The index is the difference between lifted parcel temperature (T_{lift}) and the environmental temperature (T_{env}) at 500 mb:

$$SI = T_{env} - T_{lift}$$

When SI is negative, that is, the lifted temperature is warmer than the environment, the parcel is unstable. The main short-coming of this index occurs when the low level moisture does not reach the 850 mb level. The index is too stable in these cases.

K Index: The K Index was created by J.J. George for forecasting general thunderstorms. It is a formula that makes use of 850, 700 and 500 mb data to calculate an index value:

$$K = T_{850} - T_{500} + T_{d850} - (T_{700} - T_{d700})$$

In general, K values over 24 indicate showers while values over 30 indicate thunderstorms. K is not a good indicator of severe thunderstorm occurrence.

Total Totals Index: The Total Totals Index (TT) was created by the U.S. Air Force. It is composed of two parts, the Cross Totals index and the Vertical Totals index, based on 850 and 500 mb data.

$$\begin{aligned} CT &= T_{d850} - T_{500} \\ VT &= T_{850} - T_{500} \\ TT &= CT + VT = T_{d850} + T_{850} - 2*T_{500} \end{aligned}$$

VT is essentially a measure of the 850-500 mb layer lapse rate. Values of TT over 44 indicate thunderstorm potential while values over 55 correlate well with severe weather.

Historical Note: These first three indices make use of 850, 700 and 500 mb temperature and dew point temperature information. This reflects how upper data were transmitted in the 1950s and for several decades thereafter. In these pre-computer days

sounding data were processed and coded manually, and standard level data were transmitted earlier than the full sounding. As a result, an "early" look at stability could be had by calculating the above indices based on standard level data only.

Lifter Index: The original lifted index was created by Joe Galway of the Severe Local Storms (SELS) unit. It takes boundary layer air and lifts this parcel upward to 500 mb and compares the lifted parcel temperature to the environment. This is similar to the Showalter Index except the air parcel starts near the Earth's surface. The formula is essentially the same:

$$LI = T_{env} - T_{lift}$$

The biggest challenge when calculating LI is: Where should you start the lift? There are several possibilities:

- **Mixed Layer**: If you assume that the air feeding into a thunderstorm comes from a layer within the boundary layer, start the lift with a mixed layer that represents this inflow layer. Most forecasters use a layer that is 100 mb deep.
- **Afternoon Maximum Temperature**: If you are using a 12 UCT sounding, incorporate afternoon heating into the stability evaluation by using the afternoon maximum temperature as the point for starting the dry adiabatic lift.
- **"Best" Level**: Not all thunderstorms are surface based, that is, the inflow may come from a layer above the Earth's surface. To assess this possibility, some programs lift a series of parcels at 50 mb intervals above the surface (up to 700 mb) and display the most unstable, or "best LI" value for the sounding.

Caution: When looking at automated sounding index values, it is usually not obvious where the program starts the lifting process to determine the LI value.

CAPE: Convective Available Potential Energy (CAPE) incorporates information from the entire sounding into its index values. It represents the buoyant energy (or positive area) between the LFC and the equilibrium level (EL). This is the positive or P area shown in Figure 11-7.

CIN: Convective Inhibition (CIN) represents the negative area below the LFC. When an air parcel is lifted from the surface, it

is initially cooler than the environment. This is the area labeled N in Figure 11-7. The energy corresponding to the area between the lifted parcel path and the environmental temperature curve is the CIN value. This is the energy needed to get the surface air to the LFC to produce convection. It is calculated in the same manner as CAPE.

Index Values: The values of indices discussed above correspond to the potential for thunderstorm occurrence. Please refer to the two "Stability Index Value" pages in the "Useful Weather Information" section of this web page for these correlations.

How to Destabilize a Vertical Column: For any given vertical column of air, you can assess instability by lifting an air parcel upward from the boundary layer. There are three ways that a column can be destabilized:

- *Warm the lowest layer:* If the lowest layer of the atmosphere is warmed, the lifted parcel curve will shift toward warmer temperatures. This will increase the buoyancy and create a more unstable situation. Afternoon heating is an example of this destabilizing process.
- *Add moisture to the lowest layer:* If you add moisture to the lowest layer, the lifted parcel curve will shift toward warmer temperatures and the LCL is lowered. This will increase the buoyancy and create a more unstable situation. The influx or advection of higher mixing ratio values into an area is an example of this process.
- *Cool the mid-levels:* If the temperatures in the middle levels of the atmosphere (700-500 mb) are cooled, the difference between the lifted parcel temperature and the environment increases. This increased buoyancy translates to increased instability.

If you start with a morning (or 12 UTC) sounding, you can apply these ideas by modifying the sounding to incorporate things like afternoon heating and increases in low level moisture. This approach gives you a better idea of what instability will be like during the afternoon compared to just using the morning the morning sounding alone.

Concluding Remark

The best way to learn to evaluate instability on a sounding is to get data from late Spring, plot these data on a skew T, log p

diagram (yes, this is tedious process that has been replaced by computer plotting), and then move air parcels around. Determine a variety of stability indices, and modify the sounding to represent expected short-term changes. Appendix C provides an exercise that allows you to do this.

There are other graphical operations that can be done on a skew T, log p diagram. Please refer to the U.S. Air Force manual listed in the references for a complete description of these processes.

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