# ChaserTV Meteorology Tutorials

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## Introduction

This is a compilation of various tutorials that we have written for ChaserTV over the past few months. These tutorials are designed to give an introduction to a wide range of meteorological topics, but are by no means exhaustive. We hope you enjoy these tutorials.

Materials excerpted from METR 4433, Mesoscale Meteorology, School of Meteorology, University of Oklahoma.  Professor Kelvin K. Droegemeier.  Used with permission.

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# Charts and Maps

## Introduction to Hodographs (adapted from material by Kelvin Droegemeier)

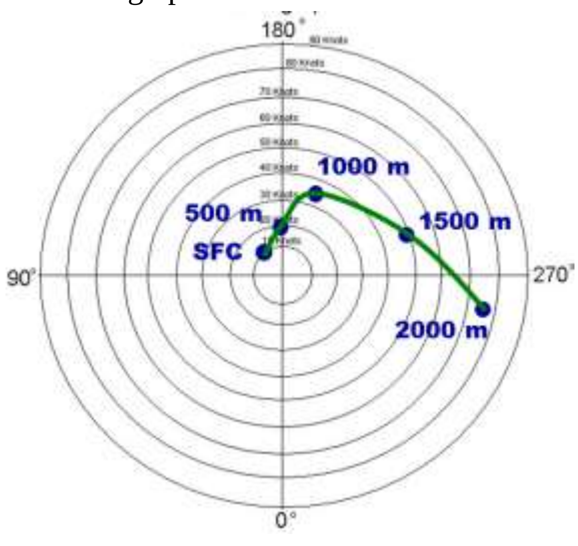
Every day, meteorologists and forecasters are required to analyze and interpret a vast amount of data. How this data is displayed can make analyzing the data much easier depending on what information is being analyzed. Hodographs can be used to quickly assess how the wind is behaving with height, making forecasts somewhat easier. Though a lot of information can be inferred from a quick glance of a hodograph, there is a lot more information that can be extracted than meets the eye. This tutorial explains some of the information that can be extracted from a hodograph.

Hodographs are based on the polar coordinate system. One of the differences that you'll notice between the polar coordinate system that you see in calculus and this system is that on the coordinate system used for hodographs, due south on the hodograph is physically due north. The directions change clockwise from this point; in other words, due west on the hodograph is physically due east, due north on the hodograph is physically due south, and so on.

Suppose we have this wind profile:

| Height | Direction (degrees) | Speed (knots) |
|---|---|---|
| SFC | 160 | 10 |
| 500 m | 180 | 20 |
| 1000 m | 200 | 35 |
| 1500 m | 260 | 50 |
| 2000 m | 280 | 75 |

The hodograph would look like this.



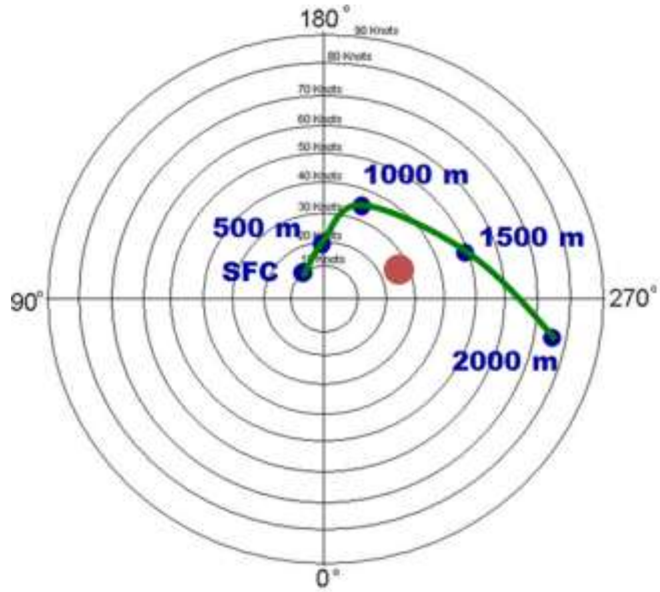
Using this hodograph, we can determine:

- Storm Relative Winds
- Shear Vectors
- Horizontal Vorticity

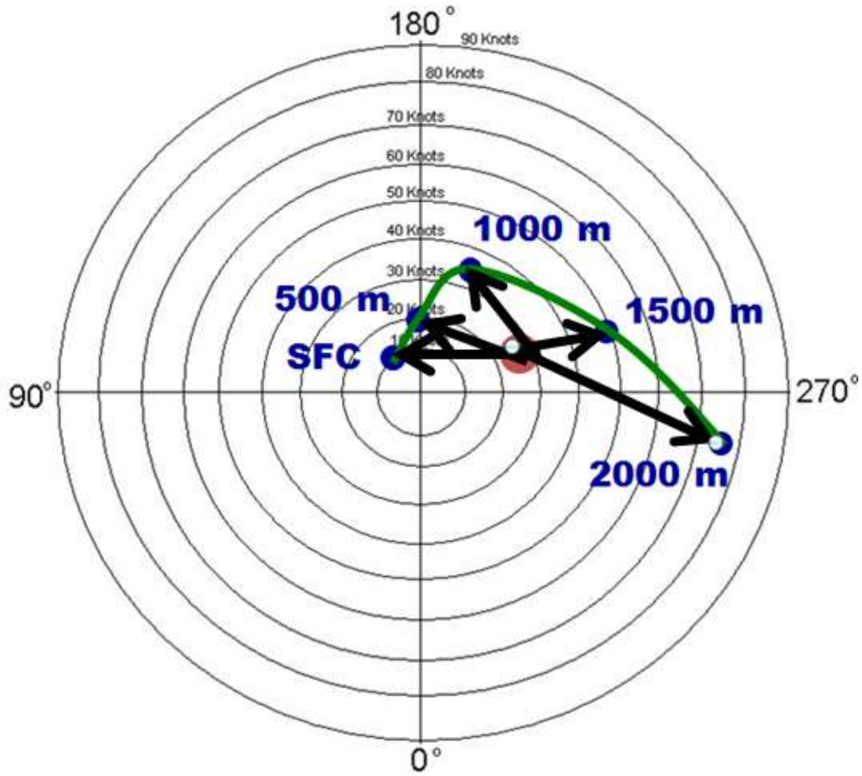
**Storm Relative Winds**

Storm Relative Winds are important in the rotation of updrafts of supercells. When the storm-relative winds are parallel to the horizontal vorticity or perpendicular to the shear vector, strong updraft rotation can result.

The Storm Relative Winds are dependent upon the storm motion, which is denoted by a dot (shown below).




To determine the Storm Relative Winds, draw an arrow from the storm motion back to the hodograph.



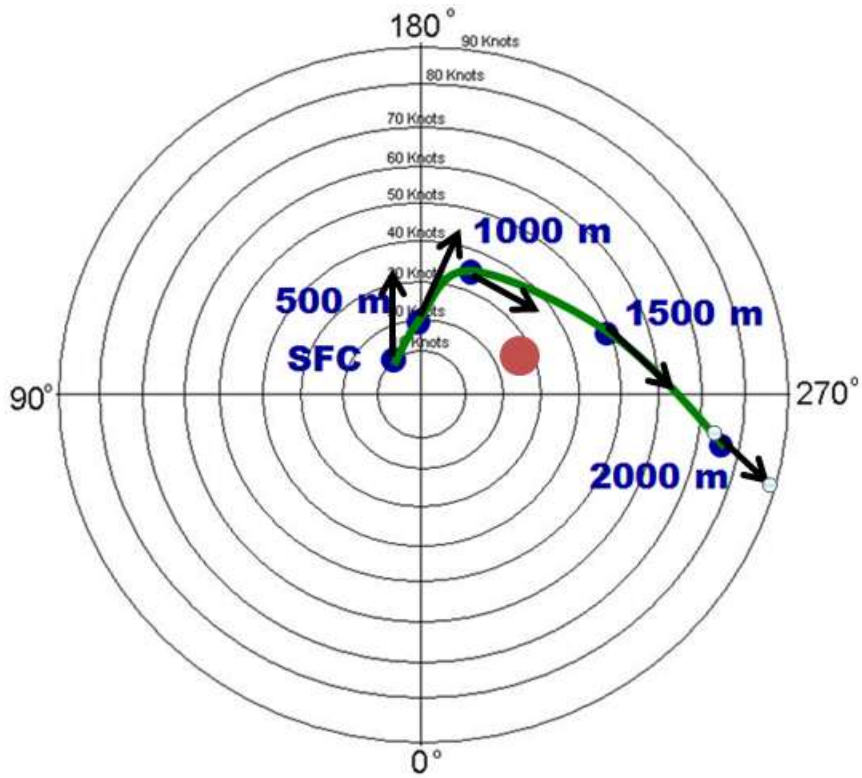

Note: Storm motion data is can be found at the end of the SPC watch text. It would look something like:

```
MEAN STORM MOTION VECTOR 23040
```

This is read as from 230° at 40 knots. Note that this has nothing to do with the hodograph shown above.

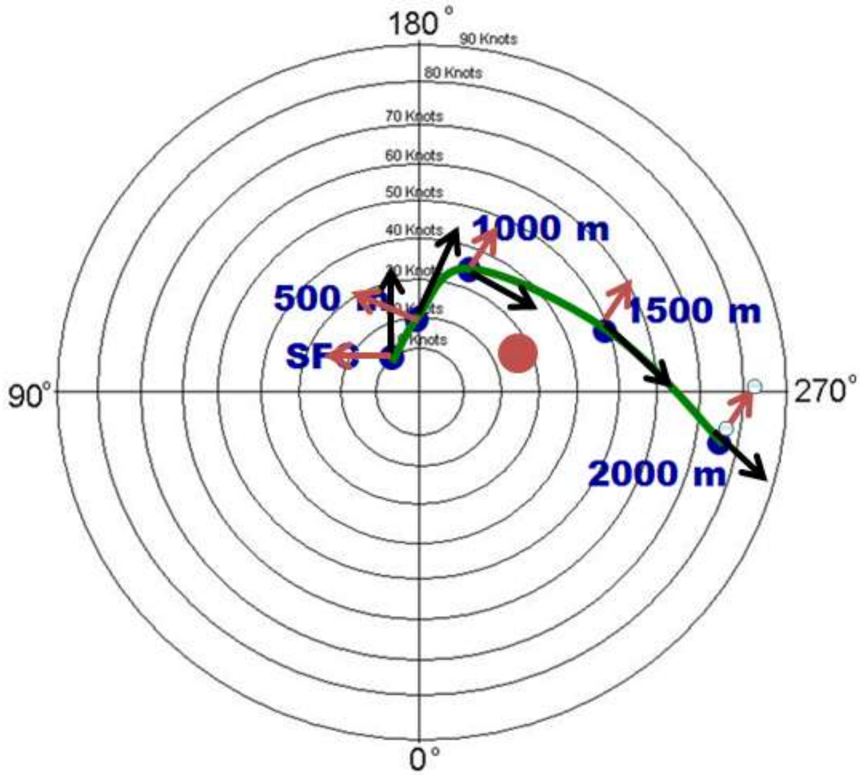
**Shear Vectors**

The shear vector at a given level can be found from the hodograph by drawing a vector tangent to the hodograph at that level. The vector always points toward higher altitude. These are the shear vectors for the given hodograph.




**Horizontal vorticity**

Horizontal vorticity is also important in forecasting and it can be easily determined on a hodograph. The horizontal vorticity vector is found by drawing a vector that is perpendicular and to the left of the shear vector. These vectors are plotted below in red.



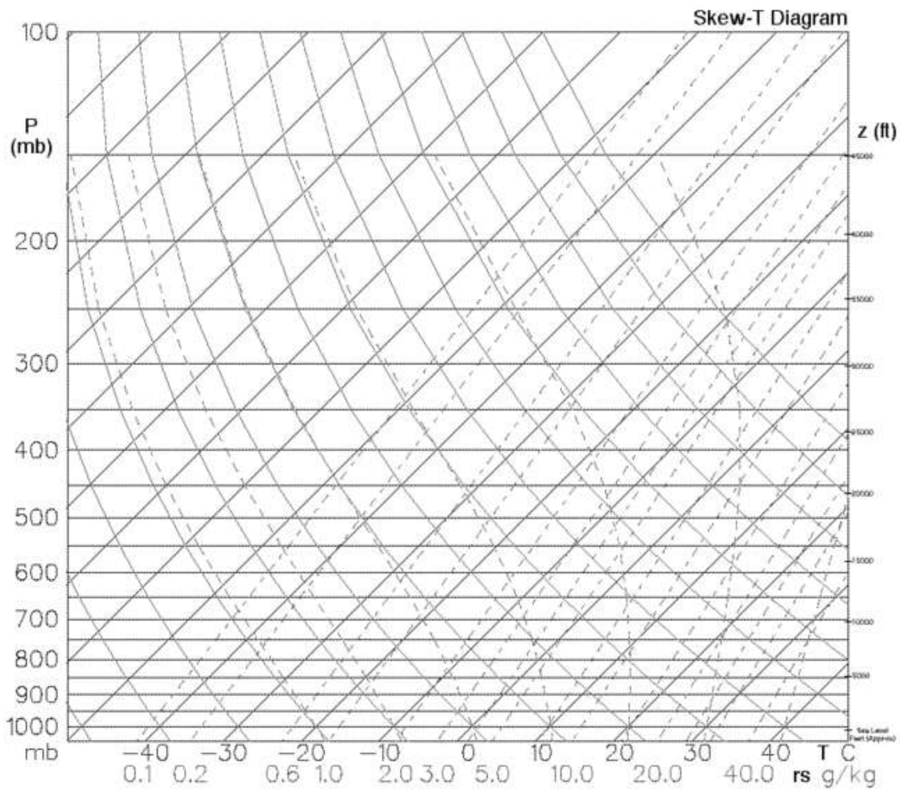

Why do we care about horizontal vorticity? The spin up of horizontal vorticity is related to the horizontal gradient of buoyancy. When this happens, it creates a horizontally oriented vortex. Updrafts near the ground work to tilt this vortex into the vertical, aiding in tornadogenesis.

Notes:
Isolated convection moves approximately with the 0-6 km shear vector, which is found by drawing a vector from the surface measurement to the measurement at 6 km on the hodograph. This vector points toward higher elevation. This does not work for supercell motion; supercell motion can be found with hodographs, but the process is more complicated (called the Bunkers Method).

## Sounding Basics

Upper air data is important in all aspects of forecasting; not just severe weather forecasting. In severe weather forecasting, soundings are important to determine the stability of the atmosphere and how it may change with time. They can also be used to determine what is possible with thunderstorms that develop (hail, damaging winds, tornadoes).

In order to begin to understand how to extract information from a sounding, let's begin by looking at the basics of a sounding. This is a sample Skew-T diagram.




Along the bottom of the diagram, the temperature in degrees Celsius and mixing ratio is plotted. On the abscissa, the pressure is plotted on a logarithmic scale. The isotherms are skewed to the right 45°. There are many reasons for the skewed isotherms. Some of these reasons are:

1. The skewed lines make it easier to assess stability.
2. The ratio of area on the chart to thermodynamic energy is the same on the entire chart.
3. It is easier to plot an entire sounding.

There are many other types of thermodynamic diagrams, but the Skew-T is the most commonly used.

Looking at the diagram, there are many types of lines. To be able to fully use the Skew-T chart, one must understand the purposes of each type of line on the chart. This is a brief description of all of the types of lines on the diagram.

1. Isobars – Horizontally oriented lines spaced logarithmically. The pressure levels are labeled to the left of the diagram.
2. Dry Adiabats – Slightly curved lines that extend from the upper left to the lower right of the diagram. These adiabats are labeled every 10°C.
3. Saturation Adiabats – These are the lines that slope slightly from the lower right to the upper left.
4. Saturation Mixing Ratio Lines – The lines that slope from the lower left to the upper right. These lines are labeled in units of gram per kilogram (i.e. gram of water vapor per kilogram of dry air).

There is a lot of data that can be obtained from a sounding once you understand the basic structure of the diagram. In the subsequent sections, how to obtain various parameters will be explained as well as how to assess the stability of the atmosphere, which is important in severe weather forecasting.

**Determination of unreported parameters**

Often, many of the parameters that we are interested in are not directly reported in the sounding, however, we can determine these parameters ourselves. There are several different parameters that can be obtained from a sounding so only the most common ones will be discussed.

The parameters that will be discussed in this tutorial are:

1. Mixing Ratio
2. Saturation Mixing Ratio
3. Relative Humidity
4. Potential Temperature
5. Wet Bulb Temperature
6. Wet Bulb Potential Temperature
7. Equivalent Temperature
8. Equivalent Potential Temperature
9. Convection Condensation Level
10. Convective Temperature
11. Lifting Condensation Level
12. Level of Free Convection
13. Positive and Negative Areas
14. Equilibrium Level

The determination of these parameters is described below.

1. Mixing Ratio
   In a sample of moist air, the mixing ratio is the ratio of the mass of water vapor to the mass of dry air. It is usually expressed in units of grams of water vapor per kilogram of dry air.

   **Procedure:** The mixing ratio at a given pressure is determined by the saturation mixing ratio line through the dewpoint curve at that pressure.
2. Saturation Mixing Ratio
   The Saturation Mixing Ratio is defined as the mixing ratio a sample of air would have if it was saturated.

   **Procedure:** The Saturation Mixing Ratio at a given pressure is determined by the saturation mixing ratio line through the temperature curve at that pressure.
3. Relative Humidity
   Relative Humidity is the ratio of the amount of water vapor in a given volume of air to the amount the air would hold if it were saturated.

   **Procedure:** Determine the Mixing Ratio and Saturation Mixing Ratio at the level desired. Divide the Mixing Ratio by the Saturation Mixing Ratio and multiply by 100%.
4. Potential Temperature
   Potential Temperature is defined as the temperature a sample of air would have if it were brought down to 1000 mb dry adiabatically.

   **Procedure:** From the temperature at the given pressure, follow the dry adiabat through the temperature down to 1000 mb. The resulting temperature is the potential temperature. (NOTE: Potential Temperature is usually expressed in Kelvin, which is obtained by adding 273 to the Celsius temperature.)
5. Wet Bulb Temperature
   The Wet Bulb Temperature is the temperature a volume of air at constant pressure would have if it was cooled by evaporating water into the air.

   **Procedure:** From the dewpoint curve at the given pressure, draw a line upward along the saturation mixing ratio line. From the temperature curve at the given pressure, draw a line upward along the dry adiabat. From the intersection of these two lines, follow the saturation adiabat back down to the given pressure level. The temperature at this point is the wet bulb temperature.

Note: The Wet Bulb Zero (WBZ) level, where the wet bulb temperature is 0°C, is important in hail forecasting, as well as snow forecasting.

6. Wet Bulb Potential Temperature
   The wet bulb potential temperature is the wet bulb temperature a volume of air would have if it were brought saturation adiabatically back to 1000 mb.

   **Procedure**: Find the wet bulb temperature. The value of the saturation adiabat through this point is the wet bulb potential temperature.

   Another way to do this is to find the wet bulb temperature as explained above. From that point, follow the saturation adiabat back to 1000 mb. The isotherm value at that point is the wet bulb potential temperature.
7. Equivalent Temperature
   The equivalent temperature is the temperature of a volume of air would have if all of the moisture was condensed out of the parcel.

   **Procedure**: From the dewpoint at the specified level, draw a line upward along the saturation adiabat. From the temperature at the same level, draw a line along the dry adiabat until it intersects the first line. From this intersection, follow a saturation adiabat to a level in which the saturation and dry adiabats become parallel. From this level, follow the dry adiabat back to the original level. The isotherm value at this point is the equivalent temperature.
8. Equivalent Potential Temperature
   The Equivalent Potential Temperature is the temperature a volume of air would have if all of the moisture was condensed out of it and then brought down to 1000 mb dry adiabatically.

   **Procedure:** Find the equivalent temperature at the specified level. From that point, follow the dry adiabat down to 1000 mb. The equivalent potential temperature is the resulting temperature.
9. Convection Condensation Level
   The Convection Condensation Level is the height at which a parcel of air, if heated sufficiently from below, will rise adiabatically until it is saturated.

   **Procedure**: Follow the saturation mixing ratio line through the surface dewpoint until the line intersects the temperature curve. This point is the Convection Condensation Level.
10. Convection Temperature
    The Convection Temperature is the surface temperature that must be reached to start the formation of convection clouds by solar heating of the surface air layer.

**Procedure**: From the CCL, follow the dry adiabat down to the surface. The resulting temperature is the Convection Temperature.

11. Lifting Condensation Level

    The Lifting Condensation Level is the height at which a parcel of air becomes saturated when it is raised dry adiabatically.

    **Procedure**: The LCL is located at the intersection of the saturation adiabat through the surface dewpoint and the dry adiabat through the surface temperature.

12. Level of Free Convection

    The Level of Free Convection is the height at which an air parcel that is lifted dry adiabatically until saturated and then saturation adiabatically after would become warmer, and thus less dense, than the surrounding air.

    **Procedure:** The LFC is found by first locating the saturation adiabat through the wet bulb temperature of the parcel. From that point, follow the saturation adiabat until it intersects the temperature profile at a higher level.

13. Negative and Positive Areas

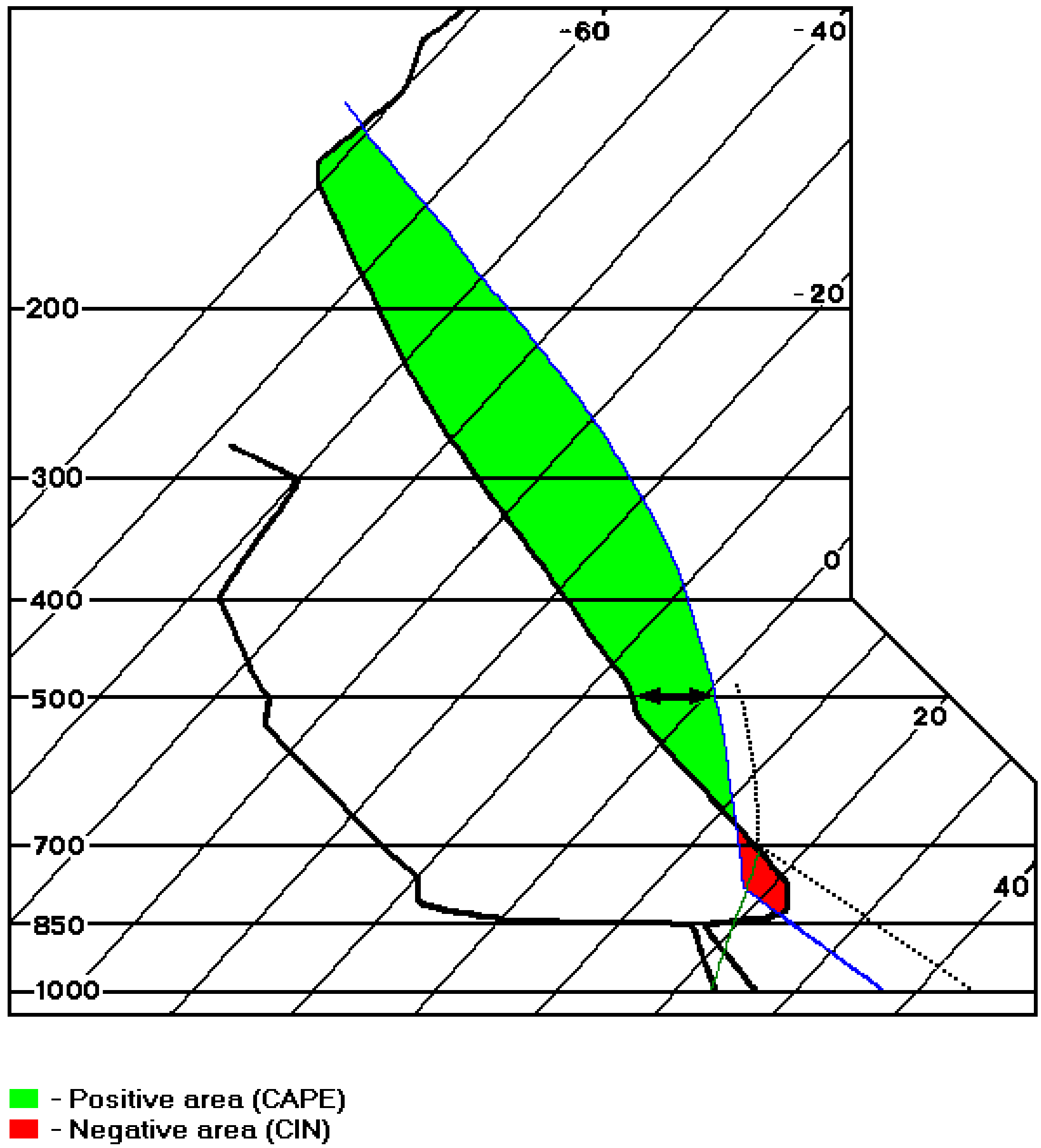
    The Skew-T diagram is set up in such a way that the size of an enclosed area is proportional to the amount of kinetic energy of an air parcel.

    **Negative Area**: When a parcel of air is in a stable environment, energy has to be supplied to the parcel to either move it up or down. The area between the parcel's path along an adiabat and the sounding is proportional to the amount of energy that must be supplied to move it.

    **Positive Area**: When a parcel of air is free to rise because it is warmer than the environment that surrounds it, the area between the adiabat and the sounding is proportional to the amount of energy that the parcel receives from the environment.

    It is important to note that the negative and positive areas are dependent on the parcel and the source of the heating of the parcel.

    Below is a look at what positive and negative areas look like on a sounding.




The green shaded area represents CAPE and the red shaded area represents CIN.

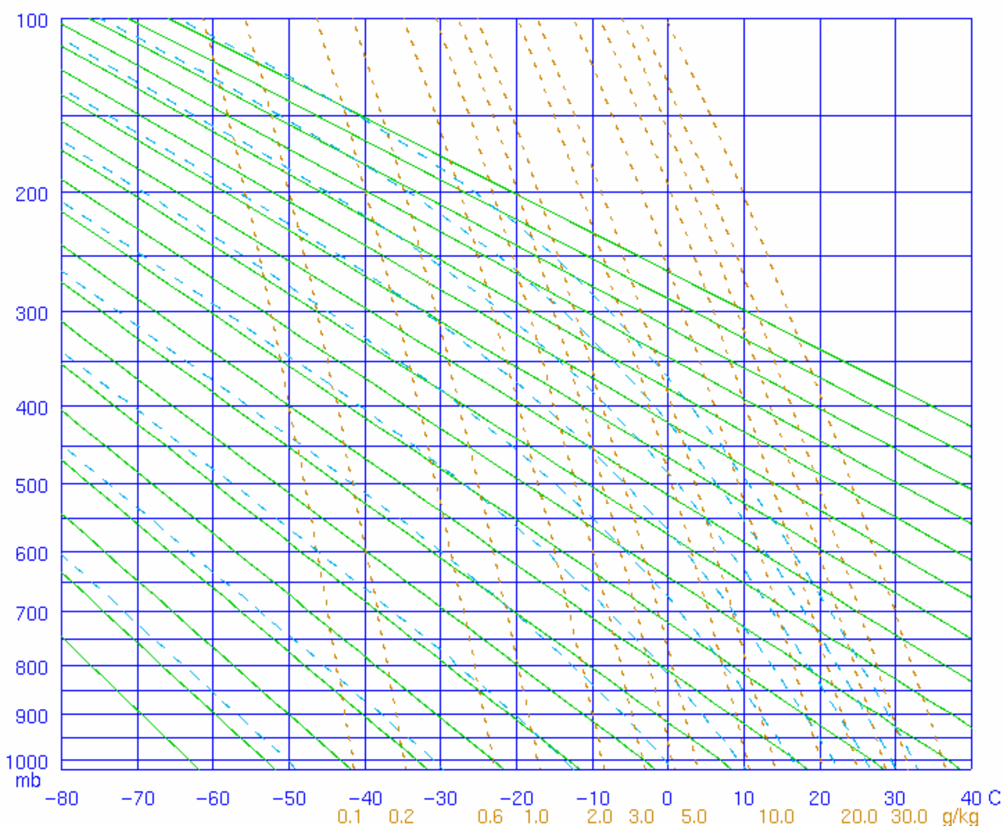
14. Equilibrium Level
The Equilibrium Level is defined as the height where the temperature of a buoyantly rising parcel becomes equal to the environmental temperature.

**Procedure**: Determine the positive area for the parcel. The Equilibrium Level at the top of the positive area where the Temperature and the saturation adiabat through the LFC intersect again.

## Pseudoadiabatic Diagrams

There are a variety of thermodynamic diagrams that are used in meteorology with the Skew-T ln p diagram being the best known. One other diagram is called the pseudoadiabatic diagram. Regardless of the format of the thermodynamic diagram, the basic principle is the same but the layout is different. In this tutorial, we will focus on the pseudoadiabatic diagram and its application.

This is a sample of what a pseudoadiabatic diagram looks like.




In this diagram, the green lines are dry adiabats, the orange lines are saturation adiabats, and the dashed blue lines are isotherms.

In this tutorial, a few sample problems using this chart will be studied to illustrate some practical applications of this diagram.

**Problem:** Consider the following profile.

| Pressure (mb) | Temperature (°C) | Dewpoint (°C) |
|---|---|---|
| 1000 | 35 | 15 |
| 850 | 22 | 9 |
| 700 | 13 | 4 |
| 500 | 4 | -5 |
| 300 | -1 | -9 |

Using this profile, determine the following information:

a. Relative Humidity at 500 mb
b. Lifting Condensation Level
c. Convection Condensation Level
d. Convective Temperature
e. Wet Bulb Temperature at 500 mb
f. Potential Temperature at 500 mb
g. Equivalent Temperature at 500 mb
h. Equivalent Potential Temperature at 500 mb
i. Equilibrium Level

**a. Relative Humidity**
Relative Humidity is defined as the ratio of the mixing ratio to the saturation mixing ratio. The mixing ratio is found by locating the saturation adiabat through the temperature. The saturation mixing ratio is found by locating the saturation adiabat through the dewpoint. In this case, the mixing ratio is approximately 5 g/kg and the saturation mixing ratio is approximately 10 g/kg so the Relative Humidity is (5/10) x 100%, or 50%

**b.Lifting Condensation Level**
The LCL is found by the intersection of the saturation adiabat through the surface dewpoint and the dry adiabat through the surface temperature. In the case of the profile above, the LCL is at approximately 830 mb.

**c.Convection Condensation Level**
Locate the saturation adiabat through the dewpoint at the surface. Follow it upward until it intersects the temperature profile. The level at which it intersects the temperature profile is the CCL. In this case, the CCL is at approximately 580 mb.

**d.Convective Temperature**
From the CCL, follow the dry adiabat through it down to 1000 mb. The resulting temperature is the convective temperature. The convective temperature is approximately 48°C.

**e.Wet Bulb Temperature**

From the dewpoint, draw a line upward along the saturation mixing ratio line. From the temperature, draw a line upward along the dry adiabat. Follow the saturation adiabat down to the original pressure level. The resulting temperature is the wet bulb temperature. In this profile, the wet bulb temperature is approximately -3°C.

**f.Wet Bulb Potential Temperature**

From the wet bulb temperature, follow the saturation adiabat down to 1000 mb and read off the temperature. In this profile, it is approximately 298K (25°C).

**g.Equivalent Temperature**

From the specified level, draw a line from the dewpoint along the saturation mixing ratio line. Also, from the same level, draw a line along the dry adiabat through the temperature until it intersects the first line. From the intersection of these two lines, follow a saturation adiabat upward to where the saturation mixing ratio line and the dry adiabat become parallel to each other. From this point, follow a dry adiabat back to the original level. For this ongoing problem, the Equivalent Temperature is approximately 288K (15°C).

**h.Equivalent Potential Temperature**

From the Equivalent Temperature, follow the dry adiabat through this temperature down to 1000 mb and read off the temperature. In this profile, the Equivalent Potential Temperature is approximately 323K (Some extrapolation was involved since the value was off the chart).

**i.Equilibrium Level**

The Equilibrium Level is located at the top of the positive area (see Skew-T tutorial). In this profile, the Equilibrium Level is at approximately 330 mb.

Many of these parameters require a bit of estimation, but the general idea of each variable is outlined in the above text.

# Severe Weather Basics

## Atmospheric Instability

We know that instability is needed for the development of severe weather, but there is another science related analogy to instability that offers a unique way of looking at things.

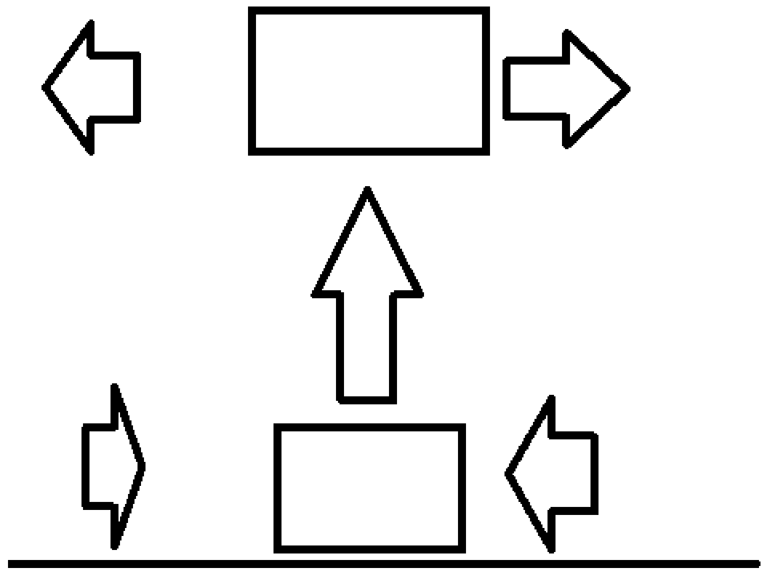
Think of the atmosphere as the human body and weather as the immune system. The atmosphere is always trying to eliminate gradients of any kind (temperature, moisture, etc), but it never gets there. This is what ultimately causes what we see as weather.

In the atmosphere, we have many types of circulations, but here, we're only going to look at two types; primary and secondary circulations. The primary circulation is something that happens in the atmosphere such as air rising. We may not know why the air is rising, and we may not even care why it's happening. The secondary circulation is the process by which the atmosphere tries to eliminate the effects of the primary circulation.

To illustrate this principle, let's use a simple example. Assume air is rising at some point in the atmosphere. The main principle at play here is called Conservation of Mass. The Conservation of Mass is a physical law derived from calculus and physics principles. While we won't go into the details here, the implications are relatively easy to understand.

Essentially, Conservation of Mass says that the volume of air in a parcel is conserved. If air is released from the parcel, the amount of air that is released from the parcel has to be replaced. In the case of the rising air, at the level where the air begins to rise, convergence occurs in order to replace the air at that level. Imagine there's another parcel of air above the level where the convergence is occurring. It also has to maintain a constant mass and in order to maintain this mass, there has to be divergence in order to let the same volume of air out that is coming in.

This is a schematic of the process described above.



In the context of primary and secondary circulations, the primary circulation is the rising air and the secondary circulation is the upper level divergence. This meteorological principle is known as LeChatelier's Principle.

With respect to forecasting, this principle can be used in examining areas that may be favorable for the development of thunderstorms. When looking at an upper air chart, upper level divergence implies convergence somewhere below the level of the chart (not necessarily at the surface). Once these areas are identified, then surface charts should be analyzed to determine the convergence near the surface and soundings to assess the atmospheric stability.

In the Skew-T tutorial, we looked at how to determine the atmospheric stability. These are purely thermodynamic instabilities, but there are other types of instabilities that are only relevant to the mesoscale.

## CAPE and Convective Inhibition

CAPE and Convective Inhibition, also known as CINH, are two important parameters in assessing severe weather potential. This tutorial will briefly explain the utility of both of these parameters.

**CAPE**
CAPE was discussed in the Skew-T basics tutorial in terms of positive area on a Skew-T diagram. As discussed in the Skew-T tutorial, CAPE can be increased in three ways:

1. Warm the surface
2. Increase the low level moisture
3. Cool the midlevels

It is important to note that CAPE is a necessary, but not sufficient condition for thunderstorm development. Once the CAPE is assessed, other aspects, such as ascent, need to be assessed as well. Ways to infer vertical motion will be discussed in a future tutorial. In particular, it is necessary to determine whether or not the vertical motion is sufficient to penetrate the Level of Free Convection.

CAPE can also be used to estimate the potential strength of an updraft. Using parcel theory and some calculus, it can be shown that the updraft speed can be estimated by doubling the CAPE and taking the square root. This will give you the maximum updraft strength in meters per second (1 meter per second is the same as 2.24 miles per hour). It is important to note that this is the maximum possible updraft speed and is rarely, if ever, attained due to various effects such as entrainment and water loading.

## Convective Inhibition

Convective Inhibition (also known as CIN) is also another parameter that is used in forecasting, but more discretion is needed when using CIN in forecasting. CIN is represented by the negative area on a thermodynamic diagram and is important in the initiation of thunderstorms. The physical interpretation of CIN is the amount of work that is needed to lift a parcel of air to the Level of Free Convection.

The calculation of both CAPE and CIN can vary based on a number of factors. These factors include moisture and the initial level of the parcel. In order to account for moisture in these calculations, virtual temperature is used in place of actual temperature. Virtual temperature is a hypothetical temperature that takes the effects of moisture into account. Under normal conditions, the difference between the virtual and actual temperatures is two degrees Kelvin at most.

# Dynamics

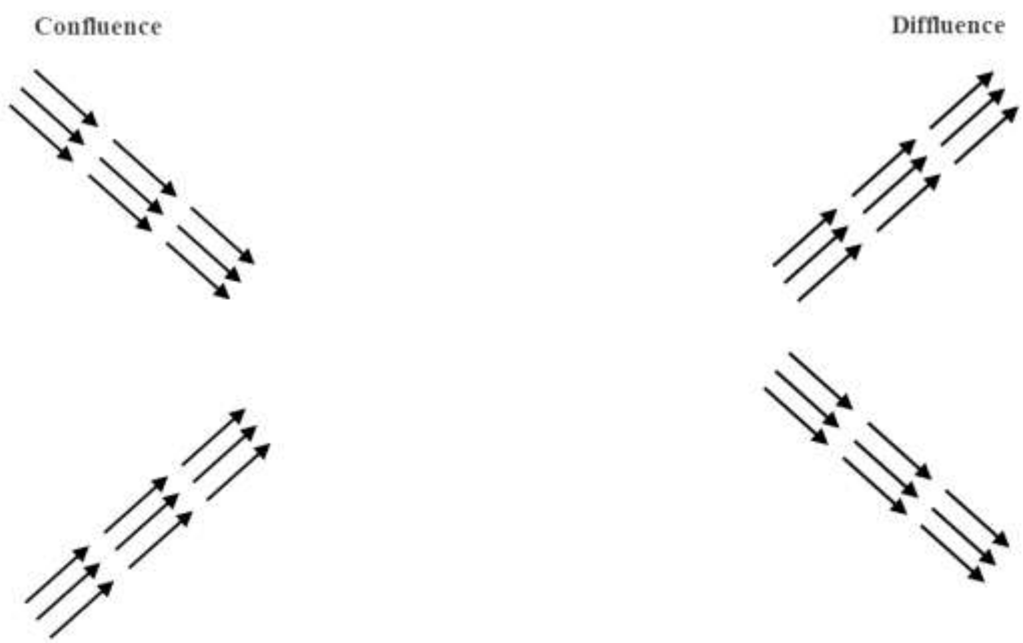
## Convergence and Divergence vs. Confluence and Diffluence

One thing that we look for in analyzing the potential for severe weather is low level convergence and upper level divergence. Convergence and divergence are often mistaken for confluence and diffluence in the process of analyzing synoptic conditions. The difference can mean the difference in whether or not thunderstorms are able to form.

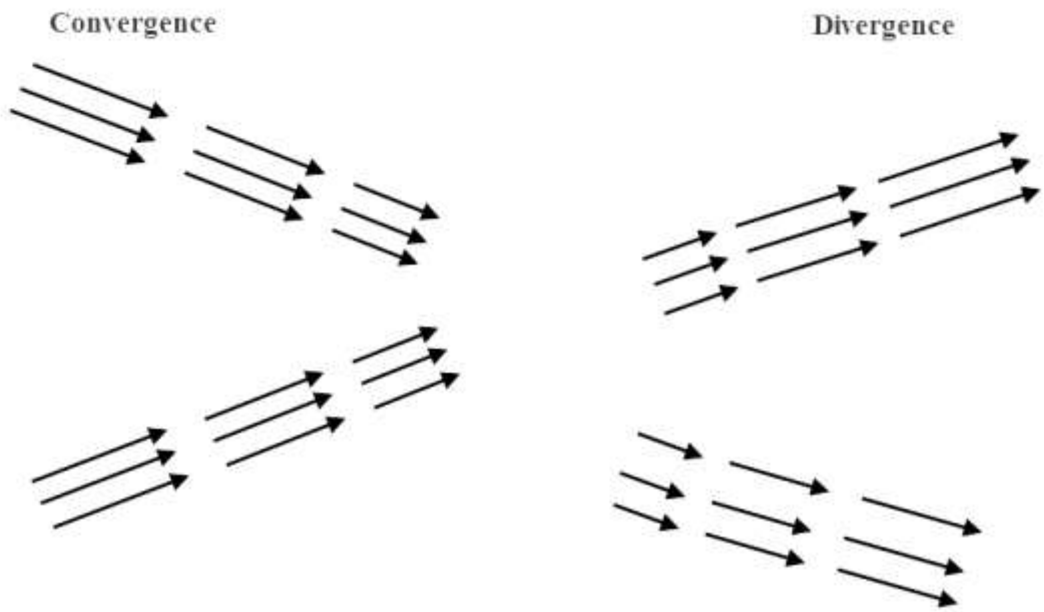
To begin, let's define each of these terms (According to the AMS Glossary of Meteorology).

**Convergence** – The contraction of a vector field.
**Divergence** – The expansion of a vector field.
**Confluence** – The rate at which adjacent flow is converging along an axis oriented normal to the flow at the point in question.
**Diffluence** – The rate at which adjacent flow is diverging along an axis oriented normal to the flow at the point in question.

This is a schematic of confluence and diffluence.



This is a schematic of convergence and divergence.




The lengths of the vectors are correlated to the speed; the longer the vector, the faster the speed. It is important to understand the difference between divergence and diffluence, as well as convergence and confluence. Knowing and discriminating between the two is important in determining if thunderstorm development is possible.

## A primer on vorticity

Vorticity is a very important quantity in meteorology, but it is often misunderstood. In this tutorial, we will explain some of the aspects of vorticity and how it is applied in meteorology.

Simply put, vorticity is the amount of spin in a fluid. If you're familiar with vector calculus, in terms of meteorology and vector calculus, vorticity is defined as the curl of the three dimensional wind field. Because vorticity is the curl of the three dimensional wind field, there is both vertical vorticity and horizontal vorticity components. Vertical vorticity is most applicable to synoptic (large scale) meteorology, whereas horizontal vorticity is more applicable to mesoscale meteorology. Vertical vorticity will be the main focus here, but horizontal vorticity will be briefly touched upon later in this tutorial.

Vorticity is also related to circulation; vorticity is calculated at a point whereas circulation is computed around a closed curve. However, the circulation around a closed curve is the sum of the vorticities around the curve.

Vertical vorticity is made up of two components, relative and earth. Relative vorticity is the vorticity of an object; earth vorticity is due to the spinning of the earth and is dependent upon the latitude. The earth vorticity is proportional to the sine of the latitude of the location. The sum of these two components is called absolute vorticity and this is the variable that you see on vorticity plots generated by numerical models.

## The Vorticity Equation

Probably one of the most crucial things to understand about vorticity is the vorticity equation. This equation won't be derived (it's a lot of calculus), but the terms in the equation will be explained physically.

Since equations cannot be put into a blog, the full equation can be seen at http://www.cimms.ou.edu/~doswell/vorticity/math_4.html.

There are two points to note about this equation:

1. The terms on the right hand side of the equation do NOT create vorticity; they simply spin it up or spin it down.
2. The derivative on the left hand side of the equation is a local derivative, meaning it's referring to how vorticity changes at a particular point with time.

Not all of the terms will be explained, but several of them have very important meteorological implications. The terms that will be explained are the advection, stretching, tilting, and friction terms.

**Advection**
In meteorology, advection refers to the horizontal transport of some quantity by the wind field. One of these quantities that can be advected is vorticity, and this can have important implications on vertical motion (more on this later). If higher values of vorticity are being advected into an area, then positive vorticity advection is occurring; if lower values of vorticity are being advected into an area, then negative vorticity advection is occurring.

Assuming for the moment that the relative vorticity is zero and keeping in mind that Earth's vorticity varies directly with the sine of the latitude from the equator, if we look at an idealized trough in the Northern Hemisphere (winds blowing counterclockwise around it), there is positive vorticity advection to the west of the trough and negative vorticity advection to the east of the trough. Along the trough axis, which is an imaginary line that goes through the lowest point of the trough, there is no vorticity advection.

**Stretching**
If you've taken physics, you've undoubtedly heard of the conservation of angular momentum. You may have also heard it explained in terms of a spinning ice skater; if the ice skater brings their arms in while spinning, they spin faster; if they bring their arms out, they slow down. This is the same principle that occurs in the atmosphere. Assume that there is an area of vorticity with convergence underneath it (we may not know why that convergence is there). The convergence acts to spin up, and stretch, the vorticity. This is one of the mechanisms by which tornadoes are generated by funnel clouds.

**Tilting**
Though this tutorial is mainly discussing the implications of vorticity on the synoptic scale, tilting is best described on the mesoscale. Horizontal vorticity can be generated due to horizontal gradients in air buoyancy. When this horizontal vorticity is generated above an updraft, the vorticity is tilted into the vertical and then it may be stretched into a tornado provided that low level convergence is present.

**Friction**
Friction can also play a role in amplifying vorticity. Suppose that wind is blowing off the ocean onto a beach (or vice versa). The surface of the ocean is smooth whereas the surface of the beach is rough. This change in texture, which also results in a change in friction can help amplify vorticity. This process is especially pronounced in landfalling hurricanes. This effect, coupled with stretching, can cause tornadoes in hurricanes.

**Vorticity Advection and Vertical Motion**

Most storm chasers and forecasters know that there's usually rising motion to the east of a trough, but there are times when this may not be the case. One of the ways to analyze vertical motion is with an area of dynamic meteorology known as Quasi-Geostrophic (QG) Theory. QG theory is derived from a variety of equations, but these equations are manipulated to derive the main QG equations, the omega equation and the height tendency equation. For right now, we will only focus on the omega equation because it is the only one relevant to vertical motion. In the omega equation, there are terms for differential vorticity advection and temperature advection. Before we go further, the word "differential" refers to how some quantity changes with height. In the case of differential vorticity advection, it is referring to the change in vorticity advection with height. In the omega equation, the terms for differential vorticity advection and temperature advection are separated by a negative sign. This means that both terms tend to oppose each other, rather than complement each other, and it's often difficult to tell which term "wins out". There are ways around this (such as Q-Vectors and the Potential Vorticity form of the Height Tendency Equation), but that will be left for a future tutorial.

On the left side of the QG omega equation is the Laplacian of omega. Omega is the variable used for vertical motion. The Laplacian is an operator you usually see in Partial Differential

Equations, but in QG theory, the application is very easy (though not exact). We assume that the atmospheric disturbances are sinusoidal when applying QG theory. Because of this assumption, the sign of the difference of the differential vorticity advection and temperature advection switches. For instance, if this difference is negative, then omega is positive and sinking motion is implied; if this difference is positive, then omega is negative and rising motion is implied.

## Potential Vorticity

When we looked at vorticity in the vorticity tutorial, we implicitly assumed that we were working on isobaric surfaces. In this tutorial, we will give a brief overview of potential vorticity, which is essentially vorticity on an isentropic surface.

Almost every explanation of potential vorticity is rather technical and filled with mathematical equations. For the purposes of this tutorial, the physical implications of potential vorticity will be emphasized over the mathematics.

We begin by defining a parcel of air that is confined between two isentropic surfaces, separated by a finite pressure interval. While the parcel of air is in motion, the mass of the parcel must be conserved. One way to look at what potential vorticity is physically is the ratio of the absolute vorticity of a column of air to the depth of the column.

Recall that the absolute vorticity is the sum of the relative vorticity and the earth's vorticity. Assuming adiabatic flow (which is one of the basic assumptions of isentropic analysis), the potential vorticity of a parcel of air must be conserved as it moves. If, for example, the parcel begins to move northward for instance, the earth's vorticity has to increase, which means that either the relative vorticity has to decrease or the depth of the parcel has to increase.

Potential vorticity can be illustrated with a physical example. Assume that there's a column of air that is moving over a mountain. Let's look at three cases.

1. The column of air is moving northward.
2. The column of air is moving along a line of latitude.
3. The column of air is moving southward.

With all of the scenarios, let's start off by assuming the relative vorticity is 0. This means that the only component of vorticity that is at play (initially) is Earth's vorticity. In all of these situations, it is also assumed that potential vorticity is conserved.

**Scenario #1**: As the column of air moves northward over a mountain range, the earth's vorticity increases, but the depth of the column of air decreases. Because the earth's vorticity increases, the relative vorticity becomes negative to compensate for the decrease in columnar depth and increase in relative vorticity. Once the column passes the mountain, the depth of the column increases and earth's vorticity continues to increase so the relative vorticity must decrease.

**Scenario #2**: In the case where the column of air moves along a line of latitude, the earth's vorticity does not change. As the column moves over the mountain, the depth decreases. In order to compensate for this, the relative vorticity becomes negative (since we initially assumed that the relative vorticity is zero). Once the column passes the mountain, the depth of the column increases and the relative vorticity begins to increase.

**Scenario #3**: When the column begins to move southward over a mountain, the earth's vorticity and the depth of the column decreases. This causes the relative vorticity to increase. After the column moves over the mountain, the depth of the column begins to increase and earth's vorticity continues to decrease. Relative vorticity also continues to increase to compensate for the decreases in columnar depth and earth's vorticity.

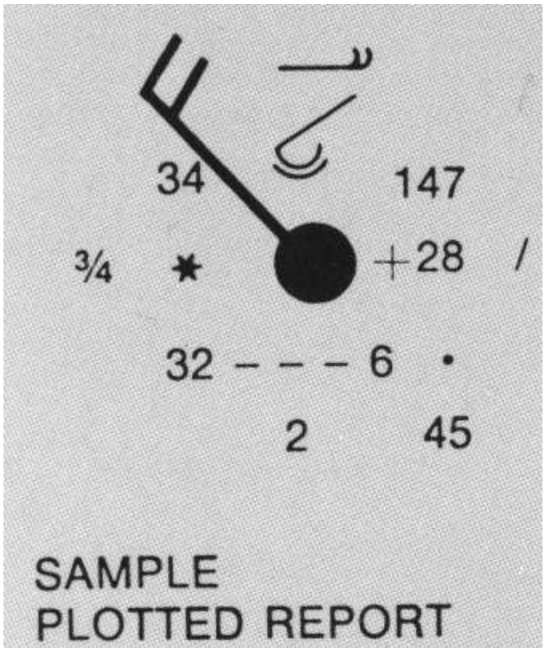
## Upper air maps

Knowing what's going on above the surface is important in forecasting. In order to do this, upper air maps are constructed from radiosonde data that are collected at 00Z and 12Z. There are several levels that are routinely examined for various reasons. The intent of this tutorial is to summarize the significance of each level and what sort of data can be gained from these charts, with specific emphasis on the 500 mb chart.

### Surface Chart

For completeness, we will begin by looking at the surface chart. The surface chart has many key features that are important to keep in mind. The surface chart will have the most data because there are more surface stations than there are upper air stations. In addition, the temporal resolution is better than that of upper air charts (One hour versus 12 hours).

On a weather map, the data is organized into what is known as a station model and a station model looks something like this:




Most station plots do not have as much data as the one pictured above. The number in the upper left is the temperature in degrees Fahrenheit; the number in the lower left is the dewpoint in degrees Fahrenheit. In the upper right is the last three digits of the pressure in millibars. The digit(s) in front of this number is either a 9 or a 10, whichever brings the number closest to 1000 and the decimal goes between the second and third digit. In the figure above, for example, the pressure is 1014.7 mb. The barb with the flags indicates the wind speed. The direction of the barb indicates the direction that the wind is from. The flags indicate the speed of the wind in knots. A long barb indicates 10 knots and a short barb 5 knots. So, for instance, if there are two long barbs and a short barb on the staff, the wind speed is 25 knots. The other data shown is not always displayed so it will not be discussed here.

Unlike upper air charts, surface data is not at a constant pressure or elevation. The contours plotted on the surface chart are isobars, lines of constant pressure. At the surface, the frictional force is important and causes convergence into low pressure regions. Isobars that are kinked usually indicate frontal boundaries.

The surface chart is important for identifying fronts, troughs, ridges, and other boundaries. Once fronts are identified, then convergence can be assessed for assessing the potential for thunderstorm development.

Fronts, especially cold fronts, are associated with deformation zones. Deformation is an important process in frontogenesis (the intensification of a front). Deformation zones can be identified by horizontal wind shifts. The winds will be northerly behind the cold front and southerly ahead of the cold front.

Warm fronts are somewhat more difficult to identify. As with cold fronts, wind shifts are useful in identifying the position of the warm front, but dewpoint depressions (the difference between the temperature and the dewpoint) are more useful in identifying the position of the warm front. The main two features of the warm front are an area of low dewpoint depressions surrounded by higher dewpoint depressions and winds turning counterclockwise as you go toward the north.

One other boundary that is of importance, especially in the Southern Plains, is the dry line. The dry line separates cool, dry air from warm, moist air. On a surface chart, the dry line can be identified by a sharp change in dewpoint, as well as winds changing from westerly behind the dry line to easterly ahead of the dry line. Thunderstorm development usually occurs ahead of the dry line if the conditions are right. Convergence ahead of the dry line, as well as stability should be assessed in order to determine thunderstorm potential.

**Upper Air Charts**

The upper level charts are labeled in terms of pressure and not geometric height. The reason for this is that it is easier to apply certain meteorological principles in pressure coordinates than in height coordinates. Any given pressure level, the level will be closer to the ground in colder environments and higher up in warmer environments.

Each of the levels has a certain way of denoting the height of the pressure surface. Before beginning to discuss the significant features of each level, let’s discuss the procedures for encoding height for each level. In the station model for all charts, excluding the surface, the height is identified by the number in the upper right of the station model. The height is always in meters.

850 mb: Put a 1 in front of the height figure. (Example: 508 corresponds to a height of 1,508 m)
700 mb: The thousands figure is either a 2 or a 3, depending on which brings the number to 3,000. (Example: 958 corresponds to 2,958 m; 014 corresponds to 3,014 m)
500 mb: Height is expressed in decameters (Example: 564 corresponds to 5,640 m)

A few other notes:
1. On an upper level chart, the number in the lower left is the dewpoint depression. Both the temperature and the dewpoint depression are in degrees Celsius.

2. The same convention for wind speeds on surface charts also applies to upper air charts.

3. Temperature advection can also be assessed on upper air charts by looking at the angle between the wind and the isotherms. If the wind is perpendicular to the isotherms, then the temperature advection is maximized; if the wind is parallel to the isotherms, there is no temperature advection. Temperature advection is also a factor in vertical motion.

4. Differential advection refers to how a specific quantity changes with height. For example, Differential Temperature Advection refers to how temperature advection changes with height.

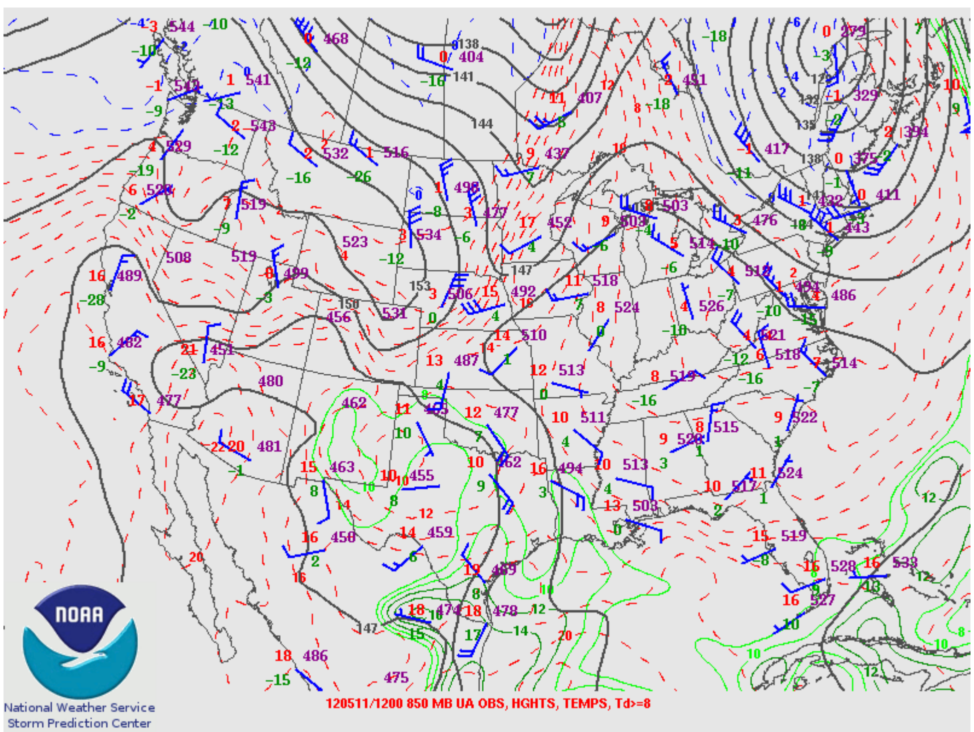
5. In general, warm air advection will increase heights while cold air advection will decrease heights.

**850 mb level**

The 850 mb level is typically the first level that is plotted above the surface (some sites and other sources provide a 925 mb level, but the same ideas apply to both levels). The 850 mb level is usually around 5,000 feet above ground level (except for in the mountains).

The 850 mb level is used primarily to assess warm and cold air advection, as well as moisture advection. It can also be used to help in identifying frontal zones that may not be evident on a surface map.

This is an example of an 850 mb analysis.



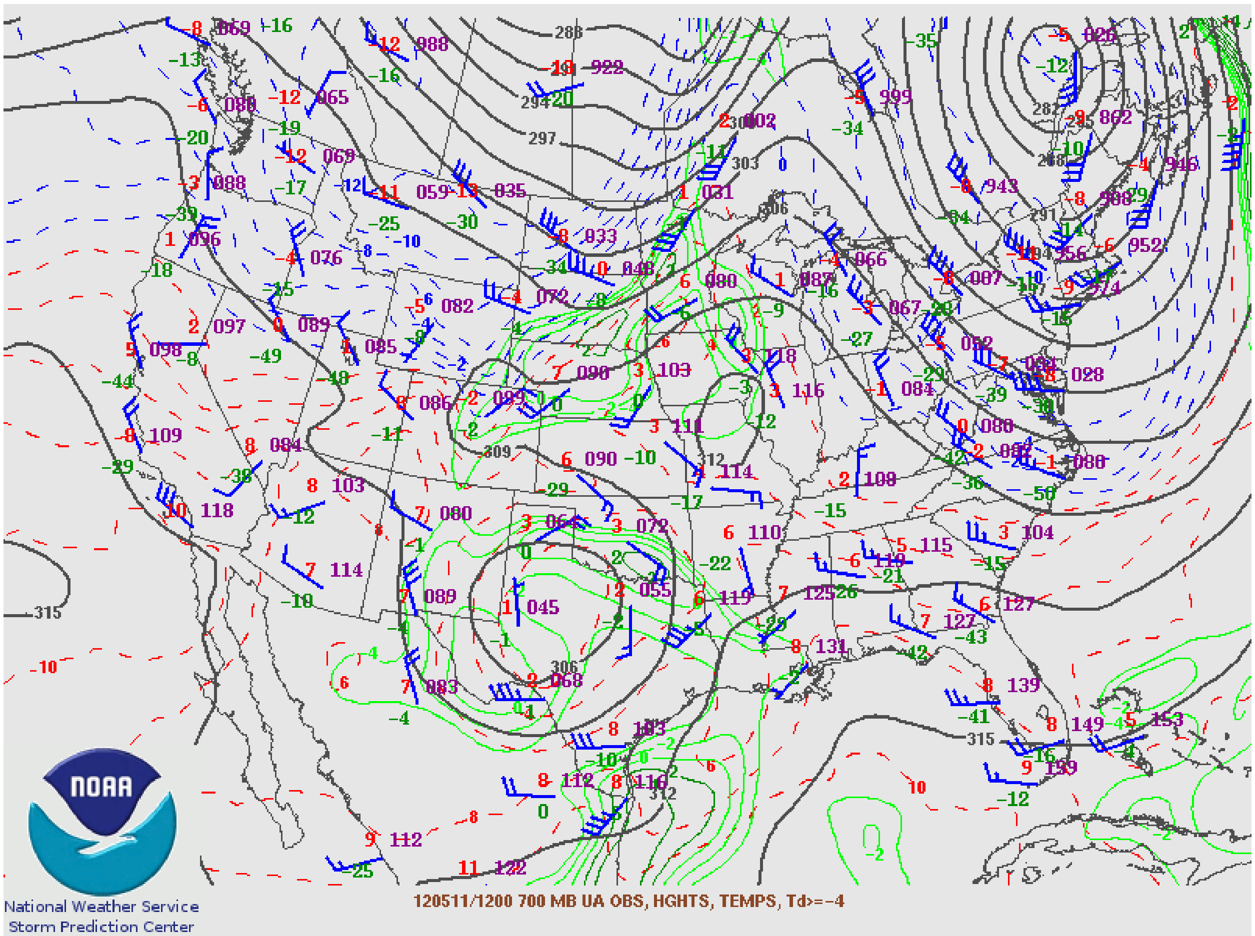

The red dashed lines are isotherms (lines of constant temperature). If you look across Nebraska, for example, you see tightly packed isotherms with winds blowing across them. In this case, the temperature advection is strong whereas if you look over Missouri, the temperature advection is weak because the isotherms aren’t tightly packed and the winds are light. Areas of warm air advection indicate rising motion, whereas areas of cold air advection indicate sinking motion. This level is also useful in identifying low level jets, if there are any. Temperature advection is maximized when the wind is perpendicular to the isotherms.

**700 mb chart**

This level can be used to determine whether or not convection will be able to develop. Studies have shown that if temperatures at 700 mb are warmer than 12-15°C*, then convection will likely not be able to develop unless this level is cooled by some process or mechanism. In these cases, it may be useful to use a sounding for the area that you are forecasting to determine whether warm or cold air advection is occurring.

* – There is some debate on the actual numerical value of this.

This is an example of a 700 mb chart.




Many of the features that would be examined at the 850 mb level can also be examined at 700 mb. Low dewpoint depressions at this level can indicate the potential for midlevel clouds. In the map above, you can see an upper low over West Texas. Because the low is open and the winds to the west of the low are blowing down the mountain, this could be a sign that the low is intensifying.

**500 mb level**

The 500 mb level is probably the most important level to analyze in the atmosphere. One of the reasons for this is that the 500 mb level is near the level of nondivergence. Because of this, we assume that the effect of divergence on vorticity is negligible.

At the 500 mb level and above, we assume that the wind is quasi-geostrophic. Geostrophic wind is a theoretical wind in which the Pressure Gradient and Coriolis forces balance each other exactly. It's a pretty good approximation in the upper atmosphere, but it is not exact. The geostrophic wind is non-divergent. Because of this property, if the wind in nature was purely geostrophic, there would be no rising motion and thus no clouds, precipitation, thunderstorms, etc. At levels below 500 mb, the geostrophic balance is disrupted by friction due to land and mountains. Typically, at the 500 mb level, a theory of dynamic meteorology called quasi-geostrophic theory is described.

Note: The geostrophic wind can be found on an upper air chart keeping in mind two points:

1. The geostrophic wind is parallel to the height contours.
2. With your back to the geostrophic wind, lower pressure is on your left.

Quasi-Geostrophic Theory builds on the approximation of geostrophic balance. It essentially takes the basic model of geostrophic flow and adds a divergence term in order to allow for vertical motion. Quasi-Geostrophic Theory does not completely describe the processes involved with frontogenesis and is best used for large scale weather systems.

A large part of applying Quasi-Geostrophic theory is being able to identify areas of Positive (Anticyclonic) Vorticity Advection and Negative (Cyclonic) Vorticity Advection. Positive Vorticity Advection occurs when larger values of vorticity are brought into a region whereas Negative Vorticity Advection occurs when lower values of vorticity are brought into a region. For example, Positive Vorticity Advection will occur to the east of an upper level ridge whereas Negative Vorticity Advection will occur to the east of a trough. Remember: Along a trough (or ridge) axis, there is no vorticity or temperature advection.

**Movement of pressure systems**

The movement of pressure systems can be described using Quasi-Geostrophic Theory. While the motivation behind Quasi-Geostrophic Theory uses some fairly advanced mathematics, the application of it is fairly straightforward.

These are the rules of thumb regarding the movement of upper level cyclones and anticyclones.

Cyclones:

1. Move from areas of anticyclonic vorticity advection toward areas of cyclonic vorticity advection.
2. Move from areas of cold air advection toward areas of warm air advection.
3. Move from areas of diabatic heating toward areas of diabatic cooling.

Anticyclones:

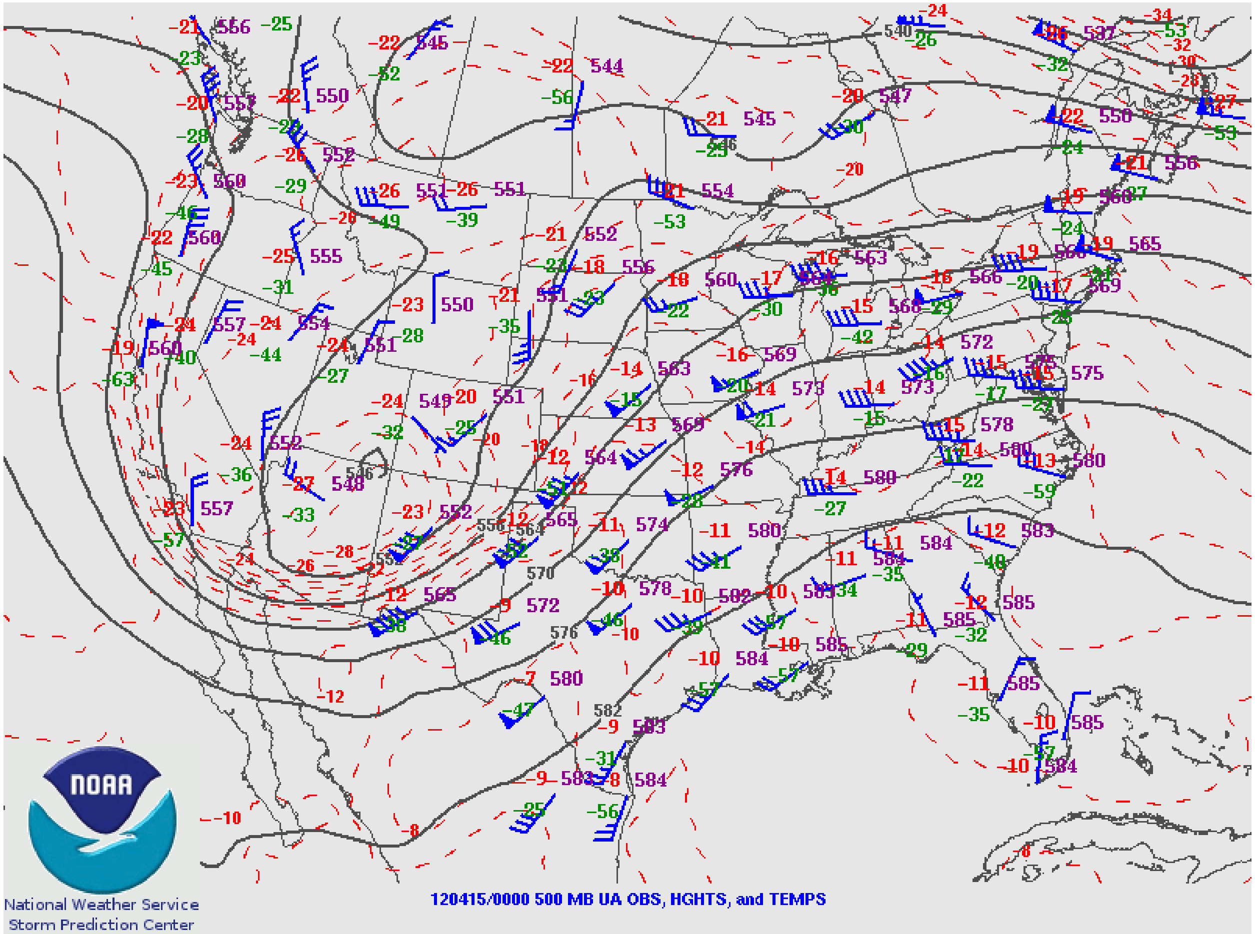
1. Move from areas of cyclonic vorticity advection toward areas of anticyclonic vorticity advection.
2. Move from areas of warm air advection toward areas of cold air advection.
3. Move from areas of diabatic cooling toward areas of diabatic heating.

**The effect of terrain**

Terrain can also play an important role in the movement of upper level systems. As air blows down the side of a mountain. As the air descends, it warms and also compresses, causing height falls. Often, upper level systems will intensify (or deepen). In general (though not always true), a trough will often move with higher elevation to the right.

Real world example:

Quasi-Geostrophic Theory is perhaps best illustrated with an example. Below is a look at the 500 mb chart from the evening of April 14, 2012.



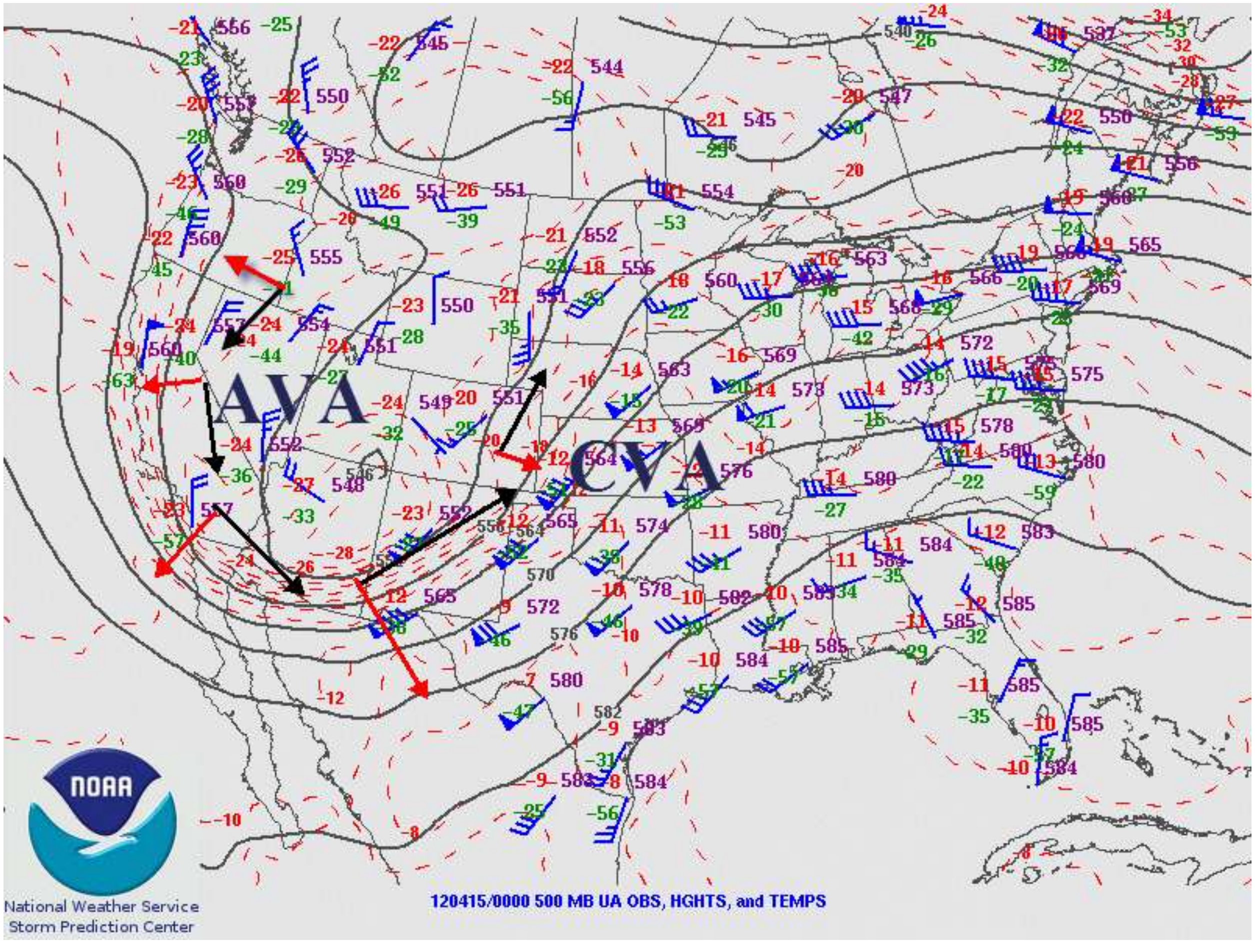

As you can see, there's a fairly significant trough over the Rockies. One of the first things many meteorologists will do is identify areas of CVA and AVA. With respect to the trough, you can see winds from the north to the west of the trough; this is an area of AVA. To the east of the trough, winds are southwesterly; this is an area of CVA. Utilizing the rules that were outlined above, the trough will likely move toward the east, where the CVA is located.

Often, you will hear people say that there's rising motion to the east of a trough. This is not always the case. According to the Quasi-Geostrophic Omega Equation, vertical motion is predominately determined by differential vorticity advection (how vorticity advection changes with height) and temperature advection. These two terms are separated by a negative sign. Because of this, differential vorticity advection and temperature advection tend to oppose each other. If this is the case, how do we determine the proper sign for omega? We can use a formulation of the omega equation called the Q-Vector Formulation. The application of this formulation resolves the issue of ambiguity when it comes to determining the sign of omega.

We won't discuss why this works, but the main emphasis here is in application. To get a completely accurate representation of the Q-Vector field, one would need to use a computer software package, but we can approximate the Q-Vector field using two simple rules.

1. The length of the Q-Vector is proportional to the temperature gradient.
2. The Q-Vector is perpendicular to and to the right of the geostrophic wind.

In the case of the map shown above, the Q-Vectors would looks like this:




In this map, the black vectors represent the geostrophic wind and the red vectors are the Q-Vectors.

The Q-Vectors can be drawn anywhere on the map where there are height contours and can be used to assess vertical motion, as well as frontogenesis and frontolysis, the intensification and decay of the front, respectively.

Rules for vertical motion:

1. If the Q-Vectors converge, then rising motion is occurring.
2. If the Q-Vectors diverge, then sinking motion is occurring.

Rules for Frontogenesis/Frontolysis

1. If the Q-Vectors point toward warmer air, then frontogenesis is occurring.
2. If the Q-Vectors point toward colder air, then frontolysis is occurring.

**Height Tendency**

The elevation of a pressure surface above ground level is called the height. In analyzing upper level systems, both height rises and height falls are important. Ridges will move from areas of height falls to areas of height rises; troughs will move from areas of height rises to areas of height falls.

According to the Quasi-Geostrophic Height Tendency Equation, height tendency is mainly dependent upon vorticity advection and differential temperature advection. In terms of vorticity advection, Negative Vorticity Advection is consistent with height rises; Positive Vorticity Advection is consistent with height falls. From the point of view of differential temperature advection, temperature advection increasing with height implies height falls whereas temperature advection decreasing with height implies height rises.

As with the omega equation, the two main terms of the height tendency equation are separated by a negative and thus they work in opposition of each other. Like the omega equation, the height tendency equation has an equivalent form that has only one term, eliminating the opposition problem. This is called the potential vorticity form of the Height Tendency Equation. With this form, the only forcing term is the advection of potential vorticity by the geostrophic wind. Positive Potential Vorticity Advection is consistent with height falls whereas Negative Potential Vorticity Advection is consistent with height rises.

Note: A plot of Differential Potential Vorticity can be found at
http://www.spc.noaa.gov/exper/mesoanalysis/new/viewsector.php?sector=19.

**Positively and negatively tilted troughs**

The tilt of the trough can provide important clues to whether the trough is intensifying or weakening.

If a trough is tilted toward the east, the trough is called positively tilted. A positive tilt implies that the system is giving energy back to the atmosphere and thus weakening. A negatively tilted trough is tilted toward the west and implies that the trough is extracting energy from the atmosphere and thus is intensifying.

**Digging and lifting troughs**

Troughs can either dig or lift as they progress. Digging refers to a southward movement of a trough whereas lifting refers to a northward movement of a trough.

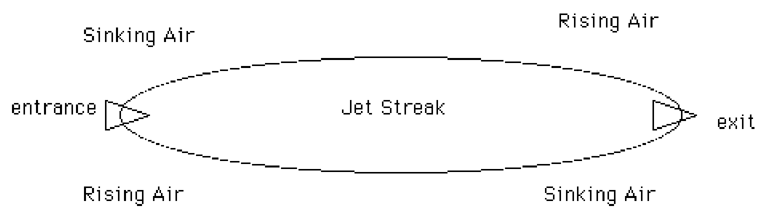
In a digging trough, the winds to the west of the trough are stronger than the winds east of the trough. Because of this, the heights to the west fall faster than the heights to the east of the trough are rising. This causes the trough to dig further to the south.

In the case of a lifting trough, the winds to the east of the trough are stronger than the winds to the west of the trough. This causes the heights to the east of the trough to fall faster than the heights to the west of the trough are rising, causing the trough to lift.

**250 mb level**

At the 250 mb level, we're not concerned with height contours, but rather contours of constant wind speed, or isotachs. This level is near the jet stream level and it is usually easy to identify the jet stream. Jet streaks, the region of the jet stream with the greatest winds are especially of interest at this level.

This is a schematic of a jet streak.




(http://ww2010.atmos.uiuc.edu/(Gh)/guides/mtr/cyc/upa/vrt.rxml)

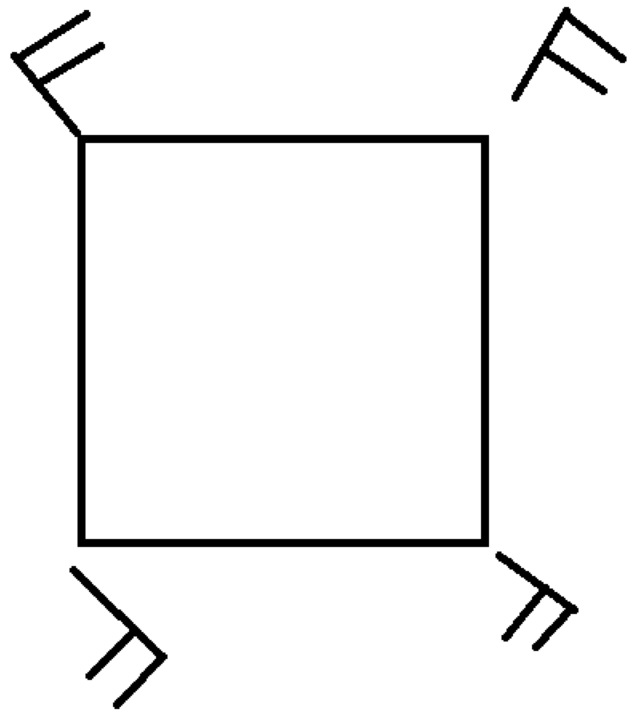
Air rises in the left front and right rear quadrants of the jet streak while air sinks in the left rear and right front quadrants. Because of the rising air, Mass Conservation says that there has to be convergence below the left front right rear quadrants of the jet streak. Typically, the regions below the left front and the right rear quadrants are especially favorable for severe weather if other conditions are met.

**A primer on frontogenesis**

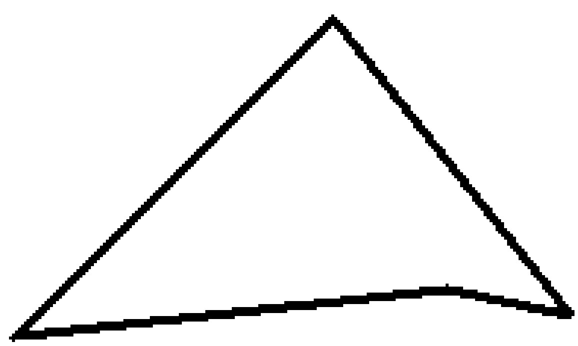
Fronts are important to meteorology because of the vast range of weather conditions that they are responsible for. In this tutorial, we will examine fronts and the processes related to them. According to the Glossary of Meteorology, a front is defined as the interface of the transition zone between two different air masses of different density.

Though the temperature gradient is discontinuous across a front, temperature is not. As a frontal zone becomes narrower, the boundary tends to act like an outflow boundary that is generated by a thunderstorm. The frontal boundary is characterized by a strong horizontal temperature gradient and can tilt with height. When this zone of strong temperature gradient tilts toward the warm air, the stability is low; if the temperature gradient tilts toward cold air, the stability is high. The strength of the cold front is determined by the horizontal gradient of potential temperature.

One of the most important mechanisms is called deformation. Deformation is the change in shape of a fluid mass. To illustrate this, let's look at a diagram.



In this diagram, the square is a parcel of air. For the sake of argument, let’s assume that there is a station at each vertex of the parcel. Given the winds at the four stations, the parcel, as a result of the deformation, may look something like this.



**Frontogenesis**

The formation of a front is called frontogenesis and the decay of a front is called frontolysis. These processes are described mathematically through the frontogenetical function. The frontogenetical function is comprised of three forcing terms. The terms that are in the function are the confluence/diffluence term, the tilting term, and the diabatic term. In the frontogenetical function, these three terms are related to the time rate of change of the horizontal potential temperature gradient. Each of these terms will be explained below.

The first term represents the effects of confluence and diffluence on the potential temperature gradient; confluence acts to increase the gradient, whereas diffluence acts to decrease the gradient. This term also contributes to deformation.

The second term, the tilting term, represents the tilting of the vertical temperature gradient onto the horizontal. If the atmosphere is stable, then the temperature gradient is increased by rising motion on the cold side and sinking motion on the warm side.

The last term represents the quasi-horizontal gradient of diabatic heating. Heating on the warm side of a front without heating on the cold side is a frontogenetical process whereas cooling on the clear side with cooling on the cold side is a frontolytical process.

**Identifying fronts**

When forecasting, it’s often important to be able to identify fronts on a chart when the fronts aren’t explicitly shown. Below is a summary of what to look for.

**Cold fronts**

When identifying the location of a cold front, the first thing you want to look for is a strong temperature gradient with the cold side to the north and the warm side to the south of the front. The winds will also typically veer across the front with winds northerly behind the front and winds southerly ahead of the front.

**Warm fronts**

Warm fronts are bit more difficult to identify. The first thing to keep in mind is that the warm air will move northward with the warm front. In identifying the warm front, winds will turn counterclockwise toward the north. In addition, there will be a minimum of dewpoint depressions concentrated in an area (in other words, an area of low dewpoint depressions surrounded by higher dewpoint depressions).

**Frontogenesis and Q-Vectors**

In the upper air tutorial, we discussed Q-Vectors and how they can be used to diagnose vertical motion. They can also be used to diagnose frontogenesis and frontolysis. Here are some rules to remember when assessing frontogenesis/frontolysis.

1. If the Q-Vector is pointing toward warmer air, frontogenesis is occurring.
2. If the Q-Vector is pointing toward colder air, frontolysis is occurring.

**Mesoscale Boundaries**

Up until now, we focused on synoptic boundaries. Mesoscale boundaries are also important in the development of thunderstorms. These boundaries will be summarized below.

**Drylines**

The dryline is a boundary that separates dry air and moist air. Drylines are best known in the South Central Plains during the spring.

The formation of a dryline begins with the formation of a lee trough (for instance, in the lee of the Rockies). The lee trough causes downslope flow. As the air descends, the air warms and dries. This causes a pronounced boundary between a hot, dry air mass and a warm, moist airmass.

Once the dryline becomes pronounced, the movement depends on the environment. Assuming relatively calm conditions, the drier air aloft mixes down toward the surface and this is what causes the apparent move of the dryline. After dark, this vertical mixing ceases, which causes the dryline to retreat back toward the west.

If there is a strong burst of momentum aloft, such as in a jet streak, this momentum is mixed down to the surface and the portion of the dryline under this jet streak bulges out further than the rest of the dryline. This is caused a dryline bulge. Severe weather usually occurs to the east of a dryline bulge provided that the environment is favorable for the development of severe weather.

When looking for the location of the dryline on a map, we look for two things; a sharp gradient in dewpoint and winds veering across the dryline (winds westerly behind the dryline and easterly ahead).

# Convection

## Cellular convection

Up until now, we have dealt with synoptic scale meteorology. Beginning with this tutorial, we will begin to discuss some principles from mesoscale meteorology.

Before we begin discussing mesoscale meteorology, let’s begin by distinguishing the difference between synoptic scale and mesoscale meteorology.

**Synoptic Scale**- Phenomena that is on the order of thousands of kilometers in length and lasts on the order of days.

**Mesoscale**- Phenomena that is on the order of hundreds of kilometers and lasts on the order of hours.

For the purposes of this tutorial, we will discuss all cellular convection with the exception of supercells; supercells will be reserved for a separate tutorial.

**The role of wind shear**

There are many variables that are involved in the evolution of thunderstorms, including CAPE, wind shear, and moisture. Wind shear can play a role in the longevity and organization of storms.

One of the ways that shear is assessed is the 0-6km shear, which is the vector difference of the wind between the wind at 6 km above ground level and the surface. CAPE and shear can be combined into a parameter called the Bulk Richardson Number. The Bulk Richardson Number is defined as the ratio of the CAPE to the vector difference between the 0-6km mean wind and the 0-500m mean wind.

For low values of BRN, the inflow tends to balance the outflow, which leads to longer lived storms. For large values of BRN, the outflow overwhelms the inflow and therefore the cells are shorter lived. Wind shear enhances the longevity of thunderstorms in two ways. First, as the environmental shear increases, the interference between the precipitation and the outflow increases. Secondly, the vertical wind shear aids in the development of vertical pressure gradients. This shear plays a significant role in the lifting ahead of a gust front and is responsible for triggering new cells.

**Single cell thunderstorms**

A single cell thunderstorm is defined as a thunderstorm that has a single updraft and does not trigger any subsequent convection. Single thunderstorms usually occur in the late afternoon when the peak heating is usually observed. These storms are usually relatively short lived and usually die out with the loss of daytime heating.

The development of a single cell thunderstorm begins with an initial updraft. Eventually, precipitation particles grow big enough to fall through the updraft. As the precipitation falls, it reduces the buoyancy in the updraft. The evaporation of the precipitation induces a downdraft, which spreads out when it reaches the ground. This cuts off the inflow from the main updraft, which prevents additional cells from developing.

**Multicell thunderstorms**

Multicell thunderstorms that continually generate new cells along the gust front provided that the Level of Free Convection can be penetrated. Individual cells typically only last on the order of an hour, but cell regeneration can occur for hours, causing hail and damaging winds. Also, new cells can be forming while the older cells are weakening and dying out.

The environments in which multicell thunderstorms develop are characterized by moderate amounts of shear, but the amount of CAPE can vary greatly.  The strength of the initial updraft can be estimated by the height of the first echo on radar. With multicells, it is important to distinguish between propagation and advection; propagation refers to the total motion of the individual cells whereas advection refers to the motion of individual cells.

The development of subsequent cells usually occurs downshear of the outflow since the environmental shear enhances lifting along the gust front and the horizontal vorticity that is generated baroclinically opposes the horizontal vorticity created by the horizontal wind shear. Boundaries, such as fronts or outflow boundaries, and terrain can also initiate new cell development and also backbuilding.

## Supercell Thunderstorms

In the last tutorial, we looked at single- and multicell thunderstorms. One other classification, and perhaps the most prolific, is the supercell thunderstorm. Because of the complexity of supercell thunderstorms, a separate tutorial is devoted to the topic.

Let's begin by defining a supercell. According to the AMS Glossary of Meteorology, a supercell thunderstorm is defined as "An often dangerous storm that consists primarily of a quasi-steady rotating updraft, which persists for a period of time much longer than it takes an air parcel to rise from the base of an updraft to its summit (often much longer than 10-20 minutes)".

Though supercells are usually defined by one dominant updraft, there can be other updrafts on the right rear flank. These updrafts are typically weaker than the main dominant updraft. The core of the heaviest precipitation is closer to the main updraft where vertical velocities can exceed 50 meters per second (112 miles per hour). This updraft and the associated mesocyclone can be seen on radar imagery with a couplet of inbound and outbound velocities.

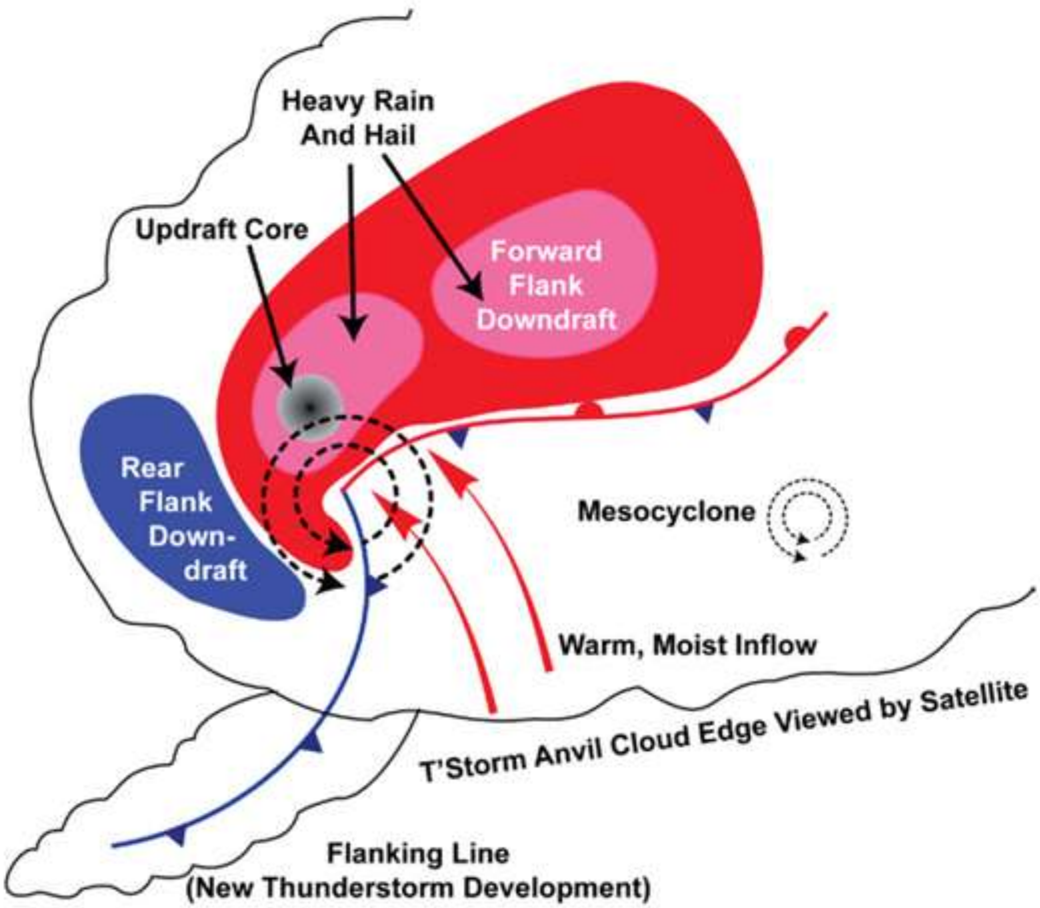
The main updraft can be identified by the location of the reflectivity minimum on a radar image; this area is called a Bounded Weak Echo Region. Within this region, it is assumed that the updraft is too strong to support the formation of precipitation. The hook echo descends from this region and can be seen on low elevation reflectivity images. It is believed that the advection of hydrometeors from the main echo core plays a role in the formation of the hook echo, but the formation process can be complicated by precipitation processes.

There are two downdraft regions in supercells. The first is in the area of the hook echo region in the rear of the storm, known as the rear flank downdraft. It is currently believed that the rear flank downdraft is formed when the winds in the mid to upper levels of the atmosphere act upon the back side of the updraft. This leads to evaporative cooling and the air becomes negatively buoyant, which forms the downdraft. The other downdraft, located on the forward flank of the storm is appropriately called the Forward Flank Downdraft. Both the Rear Flank and the Forward Flank Downdrafts act in unison to produce a gust front.

**The Supercell Spectrum**

The distribution of precipitation across supercells is not uniform. The three classes of supercells are classic, Low Precipitation (LP), and High Precipitation (HP). Supercells that have precipitation patterns that are similar to those originally observed in supercells are called Classic Supercells. In Low Precipitation supercells, most of the precipitation is evaporated before it reaches the ground. High Precipitation supercells typically have the heaviest rain within the hook echo region and on the back side of the storm. Tornadoes in High Precipitation supercells are not as common as in the other two types; the reason for this is believed to be because the downdrafts are undercut by the updrafts with outflow. Despite this, large hail and damaging winds are typical with HP supercells. The classification of supercells is highly subjective and is largely dependent on visual and radar observations.

This is a schematic of an idealized supercell.




(Photo from [http://insidetheforecast.fox19.com/2011/06/structure-of-supercell-thunderstorms.html](http://insidetheforecast.fox19.com/2011/06/structure-of-supercell-thunderstorms.html))

**Storm Relative Helicity**

Helicity is defined as “the measure of the degree to which the direction of fluid motion is aligned with the vorticity of the fluid” (Markowski and Richardson 2010). The idea of helicity is attributed to fluid dynamics, however, it is applied in meteorology because it is thought that helicity may be related to the longevity of supercells.

The only component of helicity that is relevant to storms is the storm relative helicity because the updraft of the storm is related to the tilting of vorticity. Storm Relative Helicity can be evaluated from a hodograph and the tip of a storm motion vector (NOTE: The storm motion vector is usually noted toward the end of the text of a watch). Storm Relative Helicity is equal to twice the area that is swept out by the storm motion vector and is bounded by the hodograph.

In addition to the 0-6 km shear, the 0-3 km helicity is also used in the forecasting of supercells. Storm Relative Helicity values within the lowest 3 km of the atmosphere of 150 m^2/s^2 indicate the potential for supercell development. It is important to note, however, that helicity values should not be considered a magic bullet for the potential for supercell development.

**Development of midlevel rotation**

The theory behind the development of midlevel rotation is highly mathematical and so the full details won’t be mentioned here, but a summary will be provide below.

The development of midlevel rotation begins as a vorticity couplet that is formed by the tilting of horizontal vorticity into the vertical. This vorticity is then advected by the storm relative winds into the updraft. The rotation that develops in supercells causes pressure gradients, which in turn affect updraft motion.

**Motion of supercells**

In the discussion of unicellular thunderstorms, it was said that the thunderstorm motion approximately follows the 0-6 km shear vector. In supercells, there are other dynamics and mechanisms that contribute to the propagation of the storm.

There are many methods that have been proposed to predict the motion of a supercell. Many of these methods determine the motion based on a certain fraction of the mean wind velocity and some deviation from the mean wind. The main problem with these methods is that they are Galilean invariant. This means that the relationship between storm motion and the hodograph is dependent upon the mean wind.

One popular method to predict the motion of a supercell is called the Bunkers (or Internal Dynamics) Method. To use this method, a hodograph is required. Here’s how it works.

Remember from the hodograph tutorial, that the mean shear vector always points toward higher altitude. The Bunkers method works by applying the following technique.

1. Plot the 0-500 m and 0-6 km mean shear vectors.
2. Draw the mean shear vector between the vectors in step 1.
3. The right mover of the supercell will move 7.5 meters per second perpendicular and to the right of the shear vector described in step 2.

This method is least accurate for High Precipitation supercells. It is thought that the reason for the inaccuracy relates to the cold pool of the storm.

## Mesoscale Convective Systems

Up until now, we've focused on isolated convection. In this tutorial, we're going to shift toward linear convection, also known as Mesoscale Convective Systems. According to the AMS Glossary of Meteorology, a Mesoscale Convective System (MCS) is defined as "a cloud system that occurs in connection with an ensemble of thunderstorm and produces a contiguous precipitation area on the order of 100 km or more in horizontal scale in at least one direction".

In the early stages of evolution, MCSs tend to begin as isolated cells. The outflows from these cells tend to merge, creating a large cold pool. Along this cold pool, new cells can be generated; this is referred to as upscale growth and it occurs most rapidly in environments with deep layer shear. MCSs are characterized into two types. A Type 1 MCS develops as a result of strong forcing over a front whereas a Type 2 MCS are driven by their own cold pools. It is Type 2 MCSs that are most responsible for the development of severe weather. In a sense, MCSs can be regarded as multicellular because new cells are being continually triggered by the gust front. Because of this, the entire system has a lifetime much greater than that of an individual cell.

Shear, and the orientation thereof, plays an important role in the evolution of an MCS. When the shear is both deep and strong, the line may be made up of supercells. The role of shear is maximized when the shear is perpendicular to the line. When the shear is strong and perpendicular to the boundary that initiated the convection, cells begin to split and merge with each other. This causes the line to grow rapidly. When the shear is oriented approximately 45° to the initiating boundary, the left moving cells move roughly parallel to the initiating boundary and the right moving cells move approximately perpendicular to the boundary.

Below is a sample of what a squall line might look like on radar.


(Droegemeier)

The leading edge of the line has the most intense precipitation and is called the convective region. Behind the convective region, the intensity of the precipitation is decreasing and the intensity decreases all the way to the back of the line. Behind the convective region is the transition zone and the furthest west region has the lightest precipitation and is called the stratiform region.

**Squall line maintenance**

When discussing the maintenance of squall lines, a theory that is quite possibly one of the most debated theories of meteorology, RKW Theory, is used. This theory essentially quantifies the effects of the balance between the horizontal vorticity produced by the buoyancy gradient across the gust front and the horizontal vorticity produced by the environmental low level wind shear.

One of the underlying assumptions of RKW Theory is that the intensity and the longevity of the squall line are related to the tilt of the updraft. Updrafts that tend to have a significant tilt in the horizontal tend to have their buoyancy reduced due to entrainment as opposed to updrafts that are more upright. This leads to an overall decrease in intensity of the system.

RKW Theory quantifies the relationship between the strength of the cold pool circulation and the low level wind shear. When the cold pool circulation and the low level wind shear balance each other exactly, this is called the optimal condition.

When the cold pool circulation is overwhelmed by the environmental shear, the updraft tilts downshear (with respect to the environmental shear). When the cold pool circulation overwhelms the environmental shear, the updraft tilts upshear. For the purposes of RKW Theory, the shear is measured over the lowest 2.5 km of the atmosphere and is measured perpendicular to the squall line.

RKW Theory, despite its application, has been the subject of some controversy. The main criticism is that many environments are more weakly sheared than is required for RKW's optimal condition. Some possible reasons for the discrepancy have been proposed. One of these proposed reasons is that the optimal condition may not need to be achieved in order for severe weather to occur. This, however, raises questions from a forecasting point of view. Another reason is that the occurrence of straight line winds may not be an accurate indicator of an erect updraft. Studies have shown that those lines with a significant upshear tilt often produce damaging winds. This upshear tilt plays an important role in the development of bow echoes.

### Rear inflow jets and bow echoes

Bow echoes may form when the rear inflow jet is especially strong. When this happens, the momentum from the rear inflow jet forces the portion of the squall line in front of the rear inflow jet forward and thus the bow shape forms. The vortices at the ends of the bow echo can play an important role in the intensification of the rear inflow jet. Damaging winds are usually found at the apex of the bow.

Environments in which bow echoes form tend to have CAPE that is much greater than those of squall lines. In the most extreme cases, CAPE values can be as high as 5000 J/kg and shear can be as high as 25 meters per second in the lowest two kilometers of the atmosphere.

Bow echoes can develop from HP supercells largely due to the strong environmental shear. Whether or not this happens is dependent upon various effects such as the initiation mechanism or the interactions between cells. Despite the strong shear and high CAPE required for the development of bow echoes, changes in the low level relative humidity can greatly affect the development of bow echoes.

## Downbursts

One of the effects of severe thunderstorms is damaging winds, which can be caused by downbursts. These downbursts can be characterized as microbursts or macrobursts and these categorizations are based upon the diameter of the downdraft. If the downdraft's diameter is 4 km or less, it is a microburst; if the downdraft's diameter is greater than 4 km, it is a macroburst.

While damage from damaging winds can be mistaken for damage from a tornado, they both have two distinct damage patterns. In this tutorial, the main topics to be investigated will be how downbursts form, how they are forecast, and how to tell the difference between downbursts and tornadoes in damage patterns.

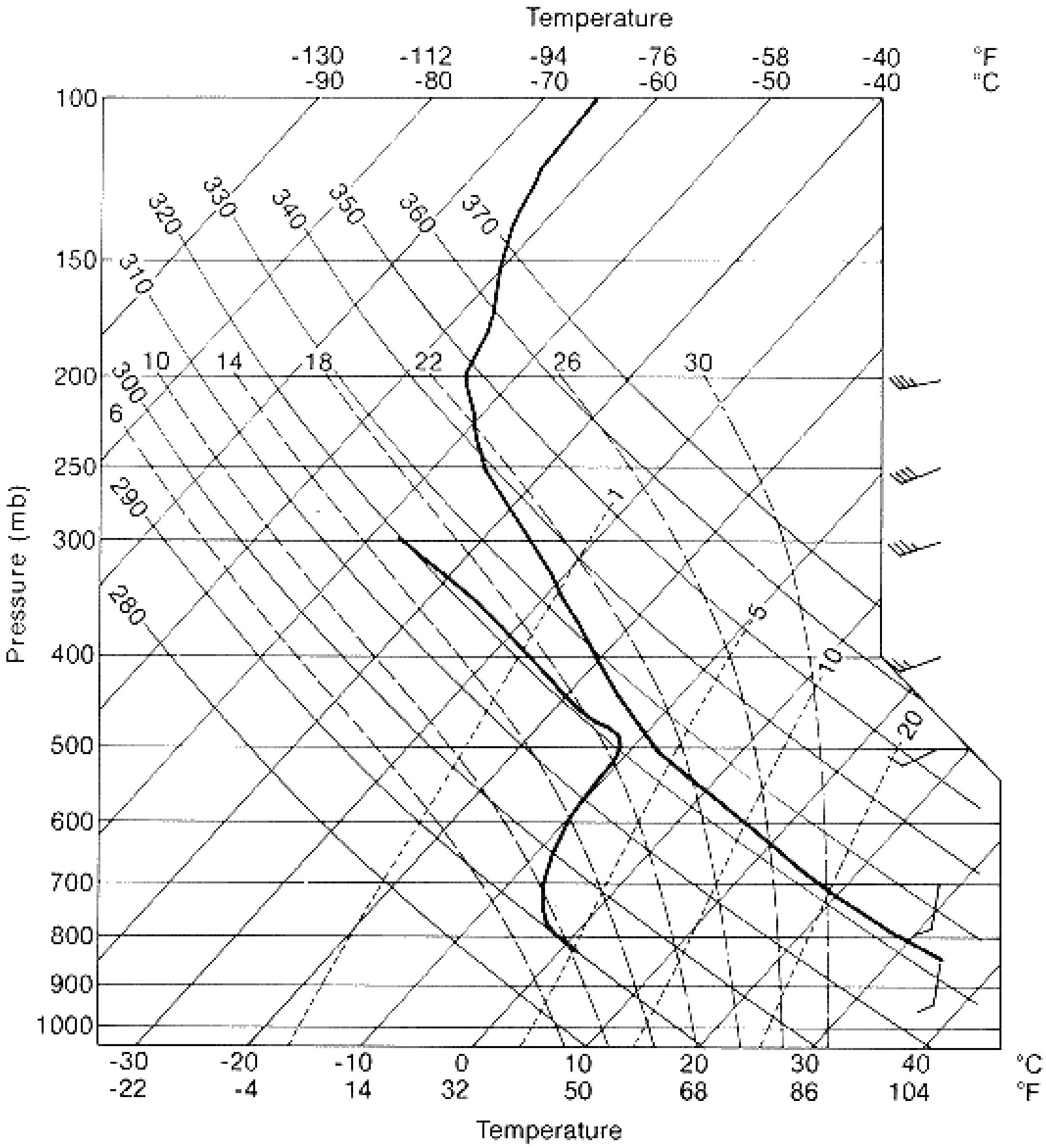
**How do they form?**

A downburst begins with dry air under a thunderstorm cloud. As rain begins to fall from the cloud, the rain evaporates and cools the air around it. As the air cools, it becomes denser as well as negatively buoyant. As the air becomes denser, it begins to descend. Upon reaching the ground, the air spreads out and this is what causes the damaging winds at the surface.

**Forecasting downbursts**

The forecasting of downbursts is largely a thermodynamic problem and soundings play a big role in forecasting them. When forecasting them, it's important to look at the depth of the dry layer in the lowest portion of the atmosphere. Some literature suggests an ideal depth of three kilometers. The typical sounding associated with downbursts is known as an "Inverted V" sounding. Inverted V soundings are also characterized by steep lapse rates.

This is a look at a typical inverted V sounding.




(http://www.wdtb.noaa.gov/workshop/psdp/dmb/sec2_dmb.htm)

Because soundings are usually released only twice a day, it is important to consider how the profile is modified by various processes between soundings. Some of the possible effects are daytime heating, moisture advection or drying, and precipitable water. As daytime heating warms the lower levels of the atmosphere, the dewpoint depression increases, which indicates drying. Another consideration is winds. If winds in the lower atmosphere are advecting drier air, then the dewpoint depression increases as well. Precipitable water is another consideration. While there is no known threshold in precipitable water, if high precipitable water values, say over an inch, are present with an inverted V sounding, this could be a sign that a strong downdraft could form and the potential for damaging winds should be considered.

**Damage from downbursts vs. damage from tornadoes**

One of the problems meteorologists face is determining whether damage was determined by downbursts or tornadoes, however, the damage patterns provide distinct clues to discriminate the event type.

In a damaging wind event, the damage pattern will be highly divergent with debris scattered away from a central point. Tornadoes, on the other hand, have a highly convergent pattern and the damage is pointed toward a central point. If the damage is determined to be from a tornado, then a survey is performed to rate the tornado and then a rating is assigned.

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