PRESENTS

WEATHER FORECASTING:
SEVERE
THUNDERSTORMS
# TABLE OF CONTENTS

Copyright and Disclaimer  3
Preface  4
  About this book  5
Chapter One: Introduction to Thunderstorms  6
  1.1 Definition of a thunderstorms  6
  1.2 Lifting Mechanisms  7
  1.3 Regional Influences  10
Chapter Two: Thunderstorm Structure  13
  2.1 The Lower Levels: Surface to 850 MB  14
  2.2 The Mid Levels: 850 MB to 500 MB  30
  2.3 The Upper Levels: 500 MB to 250 MB  33
Chapter Three: Thunderstorm Features  35
  3.1 Thunder and Lightning  36
  3.2 Rain  39
  3.3 Hail  45
  3.4 Wind Gusts and Damage  47
  3.5 Tornados  51
Chapter Four: Forecasting Tools  53
  4.1 Surface Observations and Soundings  53
  4.2 Upper Level Observations  55
  4.3 Skew-T Log And Soundings  57
  4.4 Radar  74
  4.5 Satellite  78
Chapter Five: Forecasting Philosophy  82
  5.1 SPC Discussions  83
  5.2 Model Guidance  85
  5.3 Observations  86
  5.4 Making Your Forecast  88
Copyright and Disclaimer

Copyright 2010 by Steven DiMartino. All rights reserved.

No part of this publication may be reproduced or transmitted in any form or by any means, mechanically or electronic, including photocopying and recording, or by any information storage and retrieval system, without permission in writing from the Author.
PREFACE

About the Author:
Steven DiMartino is a consulting meteorologist with over a decade of experience forecasting for the Philadelphia and New York City metropolitan area and other locations throughout the United States and the world. Steven is a graduate of the State University of New York at Oswego with a bachelors of science in meteorology. Steven has forecasted for clients ranging from some of the most important energy providers in the world to individuals simply fascinated by the latest weather events.

Steven would like to thank SUNY Oswego for the excellent education and experience he received while at the college. Also special thanks go out to the teachers and staff at Applegate Elementary, Eisenhower Middle School, and Freehold Boro High School for the support and encouragement over the years.

Steven would also like to thank his parents, Jack and Grace, who had to deal with countless hours of The Weather Channel in the 1980’s and 1990’s, enthusiastic ranting through winter storms, and the support and love that only two great parents can provide. This first eBook is for you!
About this book:
Consider this book as an introduction to the study of mesoscale dynamics and severe weather. In this book I will first go over the basic definitions of what makes a thunderstorm, how a thunderstorm develops, and explains how various features of a thunderstorm form. Then we’ll go more in depth on the structure of thunderstorms at each level from the surface to the upper levels, covering features like mesoscale boundaries to jet streaks at 250 MB. Next, I’ll cover the various forecasting tools we use for forecasting for severe weather, giving you the pros and cons to each feature along with how to properly use the data that is provided from soundings. I’ll also go over the various uses of satellite and radar data and discuss some of the clues that can make you look like a genius in front of your friends! Finally, I’ll discuss forecasting philosophy including how to use the SPC forecasts, the various models available and how to use them, and how to put a forecast together. Also I’ll walk you through a live coverage event and show you what I’m looking for when NOW-casting a severe weather event!
1.1 Definition of a thunderstorm
One of the most striking and destructive forces in nature is a thunderstorm. A thunderstorm develops due to the interaction of three key ingredients; moisture, instability, and a mechanism for lifting. Thunderstorms are capable of producing destructive wind gusts that can level a house, hail that can destroy crops and damage property, produce lightning that can spark raging fires, heavy rains that can create a wall of water from flash flooding, and/or produce a tornado that can have the capability to wipe a town off the map.

Thunderstorms develop due to a clash of air masses in the atmosphere. This clash of air masses can develop at a mesoscale (small scale) or in the synoptic (large scale) throughout the United States. In most cases, large scale thunderstorm development occurs due to a Polar air mass interacting with a Tropical air mass. Depending on the frontal system and associated air masses, thunderstorms develop different types of characteristics.
1.2 Lifting Mechanisms
With warm fronts, thunderstorms tend to develop into isolated cells to the south of the warm front boundary. The thunderstorms develop due to the warm, moist air rising (lifting) over a cooler air mass. The isolated cells initiate usually 5 to 15 miles south of the warm front boundary as these locations exhibit a combination of lifting, instability, and strong vertical wind shear. Because of the structure of a warm front, the wind direction from the surface to 500 MB veers with height, which creates a favorable environment for mesoscale low pressure systems to develop and tornadoes. This type of environment also tends to support thunderstorms capable of very large hail, heavy downpours, and very strong down burst wind gusts due to the cool, dry air that is present at 700 to 500 MB. I will go into the process of hail production, down draft winds, and heavy rainfall rates in later chapters. Severe thunderstorms that are isolated in nature are called super cell thunderstorms, and have the capability to become extremely intense.

With cold fronts, the atmospheric lifting properties and environment is different. Isolated thunderstorms and super cell thunderstorms are still able to develop between the warm front and the cold front. This area is usually called the “warm sector” of the low pressure system. However, the primary lifting feature with the cold front is found with a lifting of the warm air over a cold air mass. In this case, once the cold front exits, the threat for
thunderstorms comes to an end. While cold front can produce super cell thunderstorms, the thunderstorms with a cold front usually end up becoming a linear line of thunderstorms along or ahead of the cold front boundary. This type of structure of thunderstorms can lead what is called a Derecho. A Derecho, which I will also discuss in detail in later chapters, can produce damaging wind gusts in a much larger area than a downburst wind gust. Derechos, weak and strong, can easily stand out on the radar as a line of thunderstorms bow out. The bowing out occurs due to mid level winds working to the surface and enhancing the speed in which the line of thunderstorms are moving.

Thunderstorms can also develop along mesoscale lifting boundaries without any influence of synoptic scale features. For example, when a strong sea breeze front moves inland and interacts with the hot, humid, and unstable air mass inland; thunderstorms can form along the boundary. This boundary can be seen in observations via temperature, dew point, and wind direction data but also on the radar. These thunderstorms associated with a sea breeze front can develop rapidly and collapse just as fast depending on the strength of the mesoscale circulation. Another feature that could cause mesoscale lifting is an outflow boundary from another thunderstorm. The out flow boundary develops due to rain-cooled air moving away from the thunderstorm, usually south and southwest from the initial thunderstorms, into the
surrounding warm, humid, unstable air mass. These outflow boundaries are best observed on the radar. Outflow boundaries do not always lead to thunderstorms, but should still be monitored as they can produce strong wind gusts called gusts fronts. The best way to look at an outflow boundary is to consider the feature like a mesoscale cold front.
1.3 Regional Influences
Another factor in thunderstorm development is the region where thunderstorms develop. There is a significant difference between thunderstorms in the Plains, the Southeast, and in the Northeast and northern Mid Atlantic.

One significant difference is the type of geography that these thunderstorms develop in. For a brief overview, let’s look at the Plains, the Southeast, and northern Mid Atlantic.

The Plains for example are flat with a river valley to the east and the Rocky Mountains to the west. This type of topography leads to a variety of mesoscale influences like gravity waves off the Rockies for example. The flat nature of the Plains also allows for mesoscale low pressure systems to rapidly intensify as wind friction is limited in this environment. This is one of the reasons why tornadoes are able to become so powerful. The air masses that interact in the Plains are unique to the world. Only in the Plains do air masses interact that feature hot and dry air (Desert Southwest), cool and dry air (Canadian Plains), and hot and humid air (Gulf of Mexico). Thunderstorms in this location can develop due to a dry line moving east or a powerful cold front dropping south and east from the north. Due to the clash of these air masses, the atmosphere can become extremely unstable with impressive vertical wind shear. With no inhibiting frictional factors, these thunderstorms can develop powerful
mesoscale systems that lead to the infamous tornado producing thunderstorms in this region.

The Southeast and Gulf Coast on the other hand is under the direct influence of the Gulf of Mexico, which feeds the atmosphere with a significant amount of moisture. The influence of the Gulf of Mexico tends to support air masses that are saturated with moisture, which leads to thunderstorms capable of producing very heavy
downpours. These thunderstorms are usually more scattered in nature and tend to develop due to mesoscale influences in the summer. For example, a weak mid level disturbance, which may have initially developed due to the convection over the Plains, can ignite an area of thunderstorms rather easily in this location. Another factor is when temperatures reach what is called a convective temperature, which means that temperatures reach a point where a parcel, say water vapor and some dust, can rise in that atmosphere at free will, which leads to condensation, cumulus cloud development, and eventually thunderstorms.

Finally, in the northeast and northern Mid Atlantic, the Atlantic ocean has a significant influence on thunderstorm development along with the Appalachian Mountains. The lack of flat ground due to the rolling hills in the region limits the potential for thunderstorms to develop strong mesoscale low pressure systems and can even weaken thunderstorms due to subsidence to the east of the Appalachian Mountains. The Atlantic Ocean meanwhile can also create a more stable air mass at the surface, which can limit instability over the region as well. The thunderstorms in this part of the country usually develop due to the interaction of strong upper level disturbances, synoptic scale cold fronts, or mesoscale features such as sea breeze fronts. Of the threat locations discussed above, severe thunderstorm development is the most difficult over the Northeast and northern Mid Atlantic.
The structure of a thunderstorm is one of the most interesting and at times most complex features in meteorology. Thunderstorm development is a process that can be easily disrupted by the smallest of disturbances at any scale, and severe thunderstorms tend to be difficult to develop in the northern Mid Atlantic due to a variety of inhibiting factors that I touched on in the previous chapter.

In this chapter I will discuss the important features at each level of the atmosphere that is needed for thunderstorms to develop, their role in the maturation process of a thunderstorm, and the inhibiting factors as well. The levels of the atmosphere will be organized in the following manner:

Lower Levels: Surface to 850 MB
Mid Levels: 850 MB to 500 MB
Upper Levels: 500 MB to 250 MB
2.1 The Lower Levels: Surface to 850 MB

The lower levels of the atmosphere have a significant influence on the development and type of thunderstorms. There are three key factors that influence thunderstorm development; temperatures, moisture content of the atmosphere, and low level wind shear. These three factors influence almost every measurement and index used to determine the threat of severe thunderstorms. If a forecaster has a strong understanding of these factors, then the forecasting of severe thunderstorms is much easier.
Temperatures 2.1.1

The thermal structure of the lower levels of the atmosphere is extremely important in the development of thunderstorms. The thermal structure of the atmosphere dictates the stability of the air mass, which is extremely important factor in the support of thunderstorm development.

Thunderstorms are convective systems. Convection is a heat transfer process in which a parcel or piece of a gas or liquid, in this case water vapor, can move up or down within the whole mass of the gas or liquid (atmosphere) due to the parcel being warmer. A parcel in the atmosphere usually is water vapor, which is a key factor for thunderstorm develop. The best example of this is to watch boiling water. When you boil water you notice that the water bubbles as the water gets hotter. That bubbling is a portion of the water, evaporated into a gas, rising within the water to the surface of the water. This rising motion is convective lifting. Basically every time you make tea, coffee, or one large pot of pasta; you are using the same process that makes thunderstorms!

There are three states of stability in the atmosphere. The first state is unstable stability, neutral stability, and stable stability.
Unstable stability is when a parcel in the environment is able to rise even after lifting exerted on the parcel stops.

In other words, the temperatures the surface are warmer than the temperatures at 850 MB.

A neutral stability is when a parcel remains at rest, neither rises or nor falls after being lifted. This is usually seen in a thermal profile as temperatures remaining the same from the surface to 850 MB.

Finally, a stable air mass is when a parcel sinks after being lifted. In other words, temperatures are colder at the surface than at 850 MB.

A great way to determine stability of an air mass is to look at a smoke plume. When smoke comes out of the stack,
the fire that produced the smoke is the initial lifting mechanism. The opening of the smoke stack is your ground. Now, when the smoke rises, the smoke will do one of three things. If the smoke continues to rise higher into the atmosphere, that means the atmosphere is unstable. If the smoke rises for a short distance and then remains at a level until the smoke disperses, that means stability is neutral. Finally, if the smoke rises for a short distance and then begins to sink, that means the air is stable. Naturally, the parcel in the smoke stack is going to be warmer than an actual parcel in the atmosphere, however the difference between a natural parcel and the one from the smoke stack drops off significantly as the parcels cools, so the observation is still a good gauge of the nature of the atmosphere.

If you don’t have a smoke stack to look at to determine the stability of the atmosphere, there are two other ways to measure stability, which is provided by the National Weather Service via the Severe Prediction Center.

One measurement is the vertical lapse rate. The vertical lapse rate is the measurement of the rate of cooling from the surface to, in this case, 850 MB. The vertical lapse rate is calculated in degrees Celsius per Kilometer. The more positive the lapse rate, the more buoyant the atmosphere is. So, if the lapse rate is 7 degrees Celsius per kilometer that means the temperature at the surface is seven degrees warmer than the temperatures one
kilometer above the ground. Conversely, if the lapse rate was negative 7 degrees Celsius per kilometer, that means the temperature at the surface is seven degrees colder at the surface than the temperatures one kilometer above the ground.

The other measurement index is called the lifting index. The Lifting Index is a thermodynamic parameter that compares the theoretical 500 MB parcel temperature (temperature of a parcel from the surface lifted to 500 MB) to the actual temperature at 500 MB. A lifting index value has three states; positive, negative, and zero. A lifting index that is positive represents a stable atmosphere. A lifting index that is negative is an unstable atmosphere. Finally, a lifting index of zero represents a neutral stability. A lifting index of negative 1 to 3 is generally unstable, negative 4 to 6 very unstable, and over negative seven extremely unstable. While the lifting index value does not tell a forecaster if a thunderstorm will form, the index gives the forecaster a good idea of the unstable nature of the atmosphere and the amount of lifting needed for thunderstorms to develop.

**Moisture 2.1.2**

Moisture in the lower levels is a key factor in the growth and development of any convective property. When lifting is initiated, the key “parcel” discussed in the previous section is moisture. When moisture from the lower levels,
be that at the surface, 900 MB, or at 850 MB is lifted into higher levels; that moisture saturated parcel cools and condenses which forms clouds. This process is a form of latent heat release, which can create a serious problem with model guidance.

The process of condensation is a form of Latent Heat Release in which water vapor transitions to liquid on a parcel. The parcel is usually dust in the atmosphere. The process of this change in state creates heat due to the change in molecular state. Thus, this causes a release of heat in the atmosphere, which further increases the thermal gradient between the parcel and the surrounding atmosphere. The warmer air, which is also saturated with water vapor, continues to rise in the surrounding cooler atmosphere leading to thunderstorms that can grown to enormous heights! The process continues until the parcel reaches a stable level of stability in the atmosphere.

The amount of moisture in the atmosphere is a key factor in determining the type of thunderstorm that develops. When the atmosphere is saturated at the lower levels, the ability for the thunderstorm to produce significantly high rainfall rates increases. Further, the more moisture in the atmosphere, the more potential energy is available in the atmosphere. An air mass that is both warm and moisture saturated is capable of supporting the development of a severe thunderstorm because there is plenty of moisture for that thunderstorm to feed on.
However, in cases where the atmosphere is lacking moisture, usually below 50% in humidity, the thunderstorms lack the necessary potential energy to mature. The lack of a ready supply of buoyant parcels in the atmosphere limits condensation and thus cloud development. Naturally, lack of clouds equals lack of moisture.

So when looking for potential thunderstorm development, the atmosphere not only has to be unstable, but also have plenty of moisture to support condensation and the latent heat process.

There are several ways to measure water vapor in the atmosphere. There is the most basic, which is the dew point. The dew point is the temperature at which a parcel of air must be cooled, at a constant barometric pressure for the condensation process to begin. The best way to measure the dew point is to use a wet-bulb temperature. A wet bulb thermometer is a basic thermometer where the bulb or sensor is kept wet. The temperature of that thermometer is your dew point.

From your dew point, you can also get your Relative Humidity, which is what is seen on most television reports when discussing heat and humidity. The Relative Humidity Equation is the ratio of the partial pressure of water vapor in the mixture to the saturated vapor pressure
of water at a prescribe temperature. The equation is a bit tricky at first, however once you get the components to the equation, the calculation is easy.

\[ \text{RH} = \left( \frac{e}{\text{es}} \right) \times 100 \]

where

\[ e = \text{ew} - \text{Psta}(T - Tw)0.00066[1+(0.00115Tw)] \]

is the actual water vapor pressure

\[ \text{es} = 6.112\exp\left(\frac{17.67T}{T+243.5}\right) \]

is the saturated water vapor pressure at the dry bulb temperature

\[ \text{ew} = 6.112\exp\left(\frac{17.67Tw}{Tw+2.43.5}\right) \]

is the saturated water vapor pressure at the wet bulb temperature

T is the temperature and Tw is the wet bulb temperature

Psta = the station pressure

Winds 2.1.3

The speed and direction of wind throughout the atmosphere is an important factor in all meteorological forecasting problems at all levels and sizes of the atmosphere. In this section I will focus on low level wind considerations from the surface to 850 MB.

Low Level Jet Streaks 2.1.3.1

A feature that is at times overlooked is the presence of low level jet streams. Often on the news, you’ll hear your local meteorologist discussing upper level winds and jet
streams. Well, jet streams are not just at the upper level, but they can be found in the mid and lower levels as well!

Low level jet streaks from the surface to 850 MB can and often do play a key role in thunderstorm development. Of course due to additional friction at the surface, the low level jet streaks at the lowest portions of the boundary layer are very weak and disorganized unless focused by mesoscale and synoptic scale features like a cold front or sea breeze front. However, jet streaks do develop at 950 MB to 850 MB which have the ability to enhance low level wind shear, transport additional heat and moisture, and produce an environment for mesoscale lifting.

Low level jet streaks normally are enhanced in the evening and overnight when the thermal gradient between boundary layer and the higher levels of the atmosphere are greatest. Still, low level jet streams can enhance a severe threat no matter the time of day.

The Low Level Jet Stream is no stronger than 60 KT and typically averages 15 to 30 KT under normal conditions. When the Low Level Jet Streak is developing, an increase in moisture and temperatures usually follows due to the strong warm air and moisture advection. Also, the low level jet stream does not always lead to lifting in the typical front-right quadrant as air is usually pretty quick to fill the void of air being lifted. However, I do find that the most widespread convection tends to develop to the north of
this quadrant due to enhanced speed shear creating a rotation in the vertical atmospheric regime.

**Wind Shear 2.1.3.2**

Wind shear is the difference in wind speed and wind direction on a vertical and/or horizontal plain. There are two different types of wind shear. Vertical wind shear is the difference between height of wind speed and/or direction in the atmosphere. Horizontal wind shear is the difference within the same height of wind speed and/or direction. Wind shear plays a very important role in the development of thunderstorms by creating mesoscale and micro-scale circulations in the atmosphere.

To lay out the complexity of wind shear in thunderstorms would take far more research and complex description of equations that I think would be appropriate for this e-book. However, for further study I would strongly suggest visiting the [American Meteorological Society](https://www.ametsoc.org) for some excellent books including *Severe Convective Storms* edited by Charles A. Doswell III.

Basically, what is needed to understand for forecasting severe thunderstorms is that wind shear is needed to produce up draft and down draft circulations in a thunderstorm. The more wind shear that is present, the longer the thunderstorm can live. Also, the stronger the
directional wind shear in the vertical plain, the more twisting of the updraft occurs, and thus a tornado forms.

There are several ways to measure the wind shear in the atmosphere. Understanding the tools used by the Severe Prediction Center will better help you use these tools for your own forecast.

Effective Bulk Shear is the measurement of the bulk vector difference from the effective inflow base upward to 50% of the equilibrium level height for the most unstable parcel in the lowest 300 MB. This parameter accounts for the storm depth (effective inflow base to Elevated Level) and is designed to identify both surface-based and elevated super cell environments. Super cells are more likely in effective bulk shear vectors over 25 KT. ([http://w1.spc.woc.noaa.gov/exper/mesoanalysis/help/help_eshr.html](http://w1.spc.woc.noaa.gov/exper/mesoanalysis/help/help_eshr.html)). Basically, this tool covers all the lower levels of the atmosphere for potential vertical shear. In some cases, the winds from the surface to 950 MB are in the same direction or speed, however higher in the atmosphere, at say 850 MB the wind shear rapidly increases leading to elevated severe thunderstorms.

Bulk shear from the Boundary Layer to 6 KM above the ground demonstrates the change in wind direction with height. Supercell thunderstorms are usually associated with shear values of 35 KT or greater. ([http://www.spc.nssl.noaa.gov/exper/mesoanalysis/new/#](http://www.spc.nssl.noaa.gov/exper/mesoanalysis/new/#)) This
tool is perfect for examining the inflow into thunderstorms, which keeps a thunderstorm circulation in balance. In other words, the inflow feeds the thunderstorm with warm, moist air to replace the exiting rain-cooled air. Without this feature of shear, the thunderstorm would die immediately after the first burst of precipitation. This tool is not good for using in elevated thunderstorms as this process can still happen above 6 KM in the atmosphere.

Bulk shear up to 8 KM basically is the same as the 6 KM, however observations have found that Bulk Shear above 50 KT supports longer lasting supercell thunderstorms. (Bunkers, M.J., J.S. Johnson, L.J. Czepycha, J.M. Grzywacz, B.A. Klimowski and M.R. Hjelmfelt, 2006: An observational examination of long-lived supercells. Part II: environmental conditions and forecasting. Wea. Forecasting, 21, 689-714.)

\[ H = \int \vec{V}_h \cdot \vec{\zeta}_h \, dZ = \int \vec{V}_h \cdot \nabla \times \vec{V}_h \, dZ \]

\[
\begin{align*}
Z &= \text{Altitude} \\
V_h &= \text{Horizontal velocity} \\
\zeta_h &= \text{Horizontal vorticity}
\end{align*}
\]

The Bulk Shear from the surface up to 1 KM basically focuses in on the lowest levels of the atmosphere. The stronger the shear at this level, the more likely thunderstorms will be able to feed on low level moisture and to lift a rotating column of air, which could lead to tornadoes. I like to use this tool to determine the threat for tornadoes in thunderstorms and to see if these thunderstorms can feed on the low level moisture. If the shear is above 15 KT, then the potential for severe thunderstorms of the super cell variety is very high.
The Bulk Richardson Number shear value is the combination of the difference in direction and speed of low level winds and the density of the atmosphere. Values over 35 m$^2$/s$^2$ usually will support the development of supercell thunderstorms. I like to use Bulk Richardson Number shear values to get a feel of the buoyancy of the atmosphere in combination with the lifting provided by the wind shear. If the atmosphere is already supportive of upward motion, then the additional wind shear will lead to thunderstorms rapidly developing. I tend to watch this measurement in situations where the atmosphere is very unstable in the morning soundings.

A byproduct of wind shear in the atmosphere is helicity. Helicity is the transfer of vorticity from the environment to an air parcel in a convective environment. The following is the equation of Helicity:

When $H$ equals zero, there is no change in wind direction with height, thus a tornado will not form. When $H$ equals a positive number, the winds are turning clockwise and counter-clockwise with negative numbers. The Helicity is a measure of energy of units of m$^2$/s$^2$. There are several ways that the SPC measures Helicity.

Effective Storm Relative Helicity is a calculation that takes into account the presence of stable air at some lower level of the atmosphere. In most cases, this is a situation
where a thunderstorm has already moved through a location producing rain cooled air yet the overall air mass, above the immediate boundary layer is still unstable. This measurement basically determines how strong the inflow of a thunderstorm truly is and therefore which thunderstorm is likely to strengthen. I use this calculation when there are multiple supercell thunderstorms in an area.

Effective Storm Relative Helicity at 3 KM and 1 KM measure the same parameters but at a specific range of up to 3 KM and 1 KM in the atmosphere. Analyzing these levels can help determine which thunderstorms are more likely to produce tornadoes and which thunderstorms will not. In cases where there are numerous thunderstorms in the area, determining which thunderstorm has the edge in helicity can save a lot of time in warning locations of a dangerous situation.

To get a better picture of the energy involved with helicity, the Energy-Helicity Index (EHI) is suggested. The EHI is the measurement of storm rotation if the CAPE is large and the Storm Relative Helicity is large. EHI greater than 2 has been linked to significant tornadoes in super cell thunderstorms. The EHI can be measured up to 1 KM and 3 KM.

Finally, another excellent parameter to study is the Vorticity Generation Parameter. This is a measurement of
tilting and stretching of the atmosphere at the lower levels by a thunderstorm updraft. The stronger the updraft motion of a thunderstorm, the stronger the Vorticity Generation Parameter. Parameters over 0.2 m/s$^2$ usually indicates thunderstorms are capable of producing tornadoes.

**Mesoscale Boundaries 2.1.4**

Lifting to initiate thunderstorms can occur at many different levels. In this section I want to focus on the mesoscale boundary levels. There are two main boundaries I want to focus on, outflow boundaries and sea breeze fronts.

Outflow boundaries are the result of rain-cooled air from a thunderstorm expanding out and away from the initial thunderstorm. Basically, the outflow boundaries are mesoscale cold fronts. These mesoscale cold front help to lift the air mass away from the thunderstorm provided that the air mass is sufficiently unstable to begin with. The outflow boundary usually ignites thunderstorms to the southwest of the initial storm, in a process sometimes referred to as back building.

Sea breeze fronts develop due to the mesoscale process that is unique to the immediate coast. In a typical sea breeze, initially the land rapidly warms which leads to air rising. The cooler air over the ocean races inland to fill the
void of the rising air. Meanwhile, the warm air that initially rose into the atmosphere has to sink back down to the surface. This creates a mesoscale circulation. The edge of the cool air moving in from the ocean acts like a cold front. Once again, if the air mass is unstable initially then thunderstorms will form along this boundary.
2.2 The Mid Levels: 850 MB to 500 MB

While lower levels provide the fuel and initial lifting for thunderstorms to develop, the area between 850 MB and 500 MB is where the mechanisms of a thunderstorm’s features like lightning, hail, damaging wind gusts, and other features.

2.2.1 Temperatures and Moisture

The temperature regime at the Mid Levels is a key factor in the development of any severe weather threat. For a thunderstorm to grow, the air above the lower levels must be progressively colder to support the continuation of rising motion of the parcel. If the air at this level becomes inverted or warmer with height, then the parcel will no longer rise and the cumulonimbus cloud can no longer grow. As such, the greater the mid level lapse rate, the more buoyant the parcel will be and thus the greater threat for a thunderstorm to mature.

Naturally, another important feature at this level is the amount of moisture present. Even if the temperatures are neutrally stable or even stable, if significantly drier air is present at this level, the atmosphere can become rapidly unstable via evaporational cooling. This process would
lead to a rapid drop in temperatures, which would have several effects on the atmosphere.

The evaporational cooling of the atmosphere at the mid levels leads to another risk, which is a down draft wind gust. A down draft wind gust develops when rain-cooled air sinks to the surface to that air being more dense than the surrounding atmosphere. This type of descending motion can also create a momentum transfer of moving air or strong winds from a jet streak to the surface, which can lead to damaging wind gusts.

Additionally, the colder and drier the air becomes at this level, the higher the threat for hail production due to the enhancement of updraft winds. As the parcels at the lower levels become increasing buoyant, the rate of updraft motion also increases leading to a higher potential for hail production along with very heavy downpours and potential tornadoes.

### 2.2.2 Wind Shear

As with the lower levels, there are two key forms of wind shear. Both vertical directional and speed shear play an important role in the development of severe thunderstorm development.

Vertical directional wind shear, typically turning from southeast to northwest with height, creates a cyclonic
circulation in a vertical column. This circulation creates mesoscale cyclonic circulation or mesoscale low pressure systems within the thunderstorms, which further enhance the updraft circulation in a thunderstorm. Depending on the strength of the vertical wind shear, the mesoscale low can become strong enough to produce a wall cloud and eventually a tornado.

Speed shear on the other hand provides a different threat. Speed shear can create mesoscale and microscale low pressure systems within thunderstorms as well due to the air moving slower at higher and low levels. One of the most impressive influences though is the ability for strong jet streaks at this level to transport additional momentum down to the surface. The jet streaks at the mid levels are capable of producing strong wind gusts call Derechos, which is basically a line of strong wind gusts. On the radar, these wind gusts are seen as a bow structure within an individual thunderstorm or in a line of thunderstorms. A derecho can produce wind gusts over 55 mph covering a long distance, leading to wide spread wind damage.
2.3 Upper Levels: 500 MB to 250 MB

Finally, we reach the top of the thunderstorm structure. The key components here deal with the structure of the jet stream and the influences of divergence on convective systems.

2.3.1 Temperatures and Moisture

Temperatures and dew points at this level are still important. Naturally, temperatures at this level of the atmosphere are much colder than at the surface. However, the presence of high level moisture does have some influence. For one, a high level of moisture content at this level of the atmosphere would lead to a higher threat for heavy downpours due to the significant amount of moisture available in the atmosphere. However, the higher the moisture content also leads to a lower potential for evaporational cooling, and thus less potential for the atmosphere to become unstable and the development of down draft wind circulations.

2.3.2 Jet Stream Influences

In the previous sections, we’ve looked at the influences of low level and mid level winds on a mesoscale system. The influences of upper level jet streams are based in the synoptic and mesoscale regimes.
First I’ll touch on the synoptic. The structure of the jet stream in a synoptic scale trough produces two kinds of motion in the atmosphere, divergence and convergence. When upper level winds convergence over an area, subsidence or sinking occurs. This limits convective and non-convective precipitation. When the right front exit region or left back exit regions of a jet streak are over a location, divergence occurs which allows air at this level to rise. When strong divergence is present at the upper levels, air can rise to the optimal levels, thus leading to a completely developed thunderstorms.

On the mesoscale level, a jet streak from 500 MB to 250 MB is needed for the parcel of air which has risen to this level to escape out of the column of air, or else that air will sink back down to the surface. Sinking air kills thunderstorms, and the upper level environment must allow for the air that has risen and cooled to exit the convective system thus completing the convective cycle. The exiting of the cooler air from the thunderstorm at this level can be seen in what is called an anvil cloud.

Anvil clouds develop due to air racing out and away from the surface based location of the thunderstorm. The anvil cloud is usually made of ice crystals due to temperatures being so cold. The more expansive the anvil cloud, the strong the thunderstorm will get.
CHAPTER THREE
THUNDERSTORM FEATURES

Thunderstorms have a large variety of features that can impact a location from deadly lightning to the rapid currents of flash flooding to the devastation of tornadoes. Thunderstorms have the capability to inflict a large amount of damage in a short period of time. In this chapter, I will discuss each feature, how they develop, and what impacts they can have on a location.
3.1 Thunder and Lightning:

Lightning and the resulting thunder are key elements that are part of a thunderstorm. In an almost instinctual nature, everyone of us knows when a thunderstorm is on the way when we hear thunder and see lightning. Thunder and lightning is the world wide calling card for thunderstorms, but what causes these iconic features?

3.1.1 Lightning:

There is still great debate on exactly what causes lightning in a thunderstorm. What meteorologist do know is that lightning develops between negatively and positively charged particles in a thunderstorm cloud, called a cumulonimbus. What we don’t know is exactly how these charges are separated to produce the flashes of lightning and lightning strikes from thunderstorms.

There are a variety of theories including that the idea that ice crystals at the high levels of thunderstorms help separate the charged particles with the negative charges towards the higher portion of the thunderstorm and the positive particles towards the base of the thunderstorm.

Personally, my theory on lightning development falls on two levels. First, I look for an atmospheric sounding with a very cold mid and upper level environment. I have found through experience, that the more unstable the mid levels,
the more likely the upper portions of the cumulonimbus will become exceeding cold. Why is this important?

Well, the reason for this is that when the atmosphere is colder, the atoms within air mass at the molecular level are closer together and moving slower. This is called static electricity. The static electricity process leads to the potential for electrons within the thunderstorm to exchange between atoms, leading to an electrical charge. The exchange of electrons creates a positive charged state at higher portions of the cumulonimbus and a negatively charge state at the base of the cumulonimbus.

So if a cold air mass is in place in the soundings, we can expect plenty of lightning, but will that lightning reach the ground? For the most part, the difference between cloud to cloud lightning and ground to cloud lightning is dependent on the amount of positive and negative charges in the atmospheric column.

When I am looking at a sounding, I look for an environment with a large value of particles available for lifting, in other words, plenty of low level moisture. As the water vapor rises in the atmosphere, the atmospheric gas carries a growing charge. The more water vapor available, the better potential for an electrical charge provided that there is colder air located in the upper levels.
Note, that in tropical systems thunderstorms are the primary mechanism to produce latent heat release to fuel the warm core low. However, because the air mass is relatively warm from the surface to 250 MB, the ingredients for lightning are lacking due to no cold air mass aloft.

Lightning can heat up to 54,000 degrees Fahrenheit, which leads us to thunder!

### 3.1.2 Thunder:

Thunder develops due to the lightning rapidly heating the air. First remember, that sound is a vibration so what is vibrating to cause thunder? The air!

As lightning rapidly warms the atmosphere to extreme levels, the atoms within the column of the atmosphere effected moves rapidly. This rapidly warming air with the rapidly moving atoms, expands into the cooler air surrounding the lightning. This causes the atoms in the hot air to slam into the atoms in the significantly cooler air, which creates a vibration. That vibration is thunder!

So when you hear thunder you are actually hearing the movement of atoms in the air mass!
3.2 Rain

The primary producer of rainfall in the summer is from thunderstorms. The process in which thunderstorms produce rainfall includes several different factors which will be discussed in this section.

3.2.1 Water Vapor: The Fuel

I’m about to make a rather obvious statement. Rain cannot be produced without sufficient water vapor in the atmosphere. However, that does not mean that water vapor has to be concentrated at the surface. In some cases the water vapor is concentrated at the mid levels while the lower levels are relatively dry. In these situations, the thunderstorms are what is called elevated as the base is usually higher in the atmosphere. Water vapor can be considered the fuel source for thunderstorms. This is an important statement as the process that transitions water vapor into water droplets, called condensation, causes the release of latent heat energy, which further builds a thunderstorm. Think of water vapor like the gasoline for your car. You need to have gas in your tank before the combustion energy can take that fuel and produces heat.

One of the key factors in water vapor content is the type of air mass that is in place at the surface and mid levels. In terms of a tropical air mass, there is usually a large
percentage of water vapor per volume of air. The more water vapor in the lower and mid levels of the atmosphere, the more parcels can be lifted. The more parcels that can be lifted, the more water droplets can form and potentially collide and coalescence.

There are several processes in which a water droplet can form. I will briefly go over these process here. However to gain a full knowledge of cloud development, I would suggest studying cloud physics.

First and foremost, cloud condensation nuclei (CCN) must be present in the atmosphere. CCN are very small particles in the atmosphere in either liquid or solid form that water droplets condense onto. Water vapor can not transition into liquid form without a nuclei. This is why after a shower water vapor in the air, when cooled, forms on your mirror. The mirror provides a surface to condense on. CCN can include dust, black carbon, sea salt, pollution from factories and cars, volcanic activity, and other small particles. Without these particles, clouds could not form and rain would never fall.

One process that occurs in thunderstorms is called Coalescence. This is a process in which two or more droplets or particles in the air mass collide and form into one large droplet. This process happens in up drafts of thunderstorms until the droplet is too heavy and falls back to the earth. This process is also import in hail production.
The other process is the Bergeron-Findeisen process in which in cold clouds, ice crystals grow. This process, usually is found at the highest levels of severe thunderstorms were the air mass is cold enough to support this process. The equilibrium vapor pressure over water is greater than the saturation vapor pressure over ice, at the same temperature. At this state, the vapor pressure must be balanced out, which forces the water vapor to condense onto the ice crystal, forcing the ice crystal to grow until the crystal is too heavy and falls to the base of the cloud and to the surface.

In thunderstorms, the primary process for the development of rainfall is the Coalescence process.

3.2.2: Lifting Mechanism

Of course of the Coalescence process to begin, the parcels in the atmosphere must be lifted. There are several processes that can develop in the summer that can create strong enough lifting to produce heavy rain in thunderstorms.

At the synoptic level, the primary lifting process is from large scale troughs and frontal systems, which are the surface reflection of these upper level troughs. As discussed previously, the type of frontal boundary is key in
determining the type of thunderstorms that develop. However, all lifting process accomplish the same goal, to create upward motion in the atmosphere.

Aside from frontal boundaries, there is also the influence of upper level disturbances. The key with upper level disturbances is that these feature typically further enhance instability in an atmospheric column. Combined with mid level forcing, the parcel in the column of air is able to rise further and faster in height due to the introduction of a colder, drier environment. These types of thunderstorms usually are found to be elevated thunderstorms initially, however outflow boundaries from the initial thunderstorms support the development of other thunderstorms, which leads us to the mesoscale.

The process to produce thunderstorms and thus rainfall at the mesoscale level can be initiated by a variety of different ways. As discussed above, out flow boundaries from other thunderstorms can support lifting to initiate the coalescence process. Another feature is a sea breeze front, which acts like a low level cold front. These process also produce upward motion, which enhances up draft motion in the atmosphere thus supporting the process of coalescence.
3.2.3 Mid and Upper Level Winds:

The orientation and strength of mid and upper level winds are key in determining the impact rainfall can have from thunderstorms on a location in a variety of ways.

In cases where upper levels winds are parallel to a surface cold front, the cold front tends to move very slowly. As a result, the line of thunderstorms tends to move east slowly, which individual thunderstorm cells continue to move over the same location. This is one form of thunderstorms training over a location, in which thunderstorms can form and develop over a sustained area of weakness.

Sometimes, the mid and upper level winds are very weak while the atmosphere is already favorable for thunderstorm development. With the help of some lifting, usually in the mesoscale, thunderstorms can develop but move very slowly, leading to hours of heavy rainfall. Strong mid and upper level winds can also aid in the addition of moisture and lifting into the atmosphere. As discussed previously, a low level jet streak has the ability to add water vapor and lifting into a location thus enhancing the Coalescence.
3.2.4 Flash Flooding:

Flash flooding localized and regionally can and does occur from thunderstorms. In some cases, the simple nature of the atmosphere, tropical with a high level of water vapor per atmospheric volume at a specific temperature, can be enough fuel to produce extremely impressive rainfall rates. In other cases, the process of thunderstorms developing and redeveloping over an area of weakness or a mesoscale boundary can lead to rainfall amounts exceeding monthly averages. While in others, the addition of an upper level disturbance enhancing the up draft in the atmosphere can support flash flooding. Of course, the forecasting of the potential for flash flooding must include not only the period of time in which the rain falls, but also if previous weather events enhance this potential.
3.3 Hail:

The formation and size of hail is influenced by several important properties; depth of upper level cold air, available low/mid level moisture, and strength of updraft. Hail is formed via the process of Coalescence, however unlike in the process of rain droplet formation, temperatures at the mid and upper levels must be below freezing in order for the coalesced water droplet to freeze.

When forecasting for hail, the first feature one must look for is the presence of a cold, dry air mass from 700 MB on up. Why is this important? Hail can not form in a warm environment because the water droplets must freeze. As such, the formation of hail is usually found in areas where a clash of air masses is taking place. This is also why hail is not usually found in “warm core” low pressure systems like hurricanes because the atmosphere is too warm at mid and upper levels.

The next feature to look for in hail development is the presence of vertical wind shear. The stronger the wind shear, the stronger the upward motion in the atmosphere. A strong updraft is important because the upward motion of the atmosphere in an updraft is what forces a hail stone into higher portions of the atmosphere.
The influence of an up draper in hail formation works like this. Initially, a strong up draft forces a water droplet higher into the atmosphere where the air is below freezing. This water droplet freezes and begins to fall back to the surface. However, as the droplet falls back into the mid and lower levels, the up draft picks up the now semi-frozen droplet and forces that droplet back into the upper portions of the cumulonimbus. While doing so, the droplet can grow in size via water vapor condensing onto the semi-frozen droplet or by other water droplets colliding and coalescing with the semi-frozen water droplet. The now larger water droplet is forced into the higher portions of the atmosphere, freezes and then begins to fall again. This process continues until the now frozen hail is too heavy for the up draft to lift. The hail then falls to the ground as a hail stone.

The stronger the updraft, the larger a hail stone can become providing other atmospheric properties are supportive of formation and growth. The updraft can become as strong as 100 mph in the strongest thunderstorms (National Center for Atmospheric Research (2008). "Hail". University Corporation for Atmospheric Research. http://www.ncar.ucar.edu/research/meteorology/storms/hail.php. Retrieved 2009-07-18.).

3.4 Wind Gusts and Damage:

Strong wind gusts in a thunderstorm can create a significant amount of damage with wind gusts exceeding hurricane strength in the strongest thunderstorms. Winds gusts in thunderstorms are typically developed via the process called Downburst.

3.4.1 Downbursts:

A down burst develops when through the process of evaporational cooling, a portion of the atmosphere becomes cooler and thus heavier than the surrounding environment. Due to the cooler air becoming more dense, this air sinks to the surface. This movement of air is the wind gust, which moves to the surface and then expands outward. The strength of the down draft can be influenced by a variety of processes and environments like upper level jet streaks, the nature of the air mass at the mid/upper levels, and momentum transfer potential.

3.4.1.1 Mid Level Air Mass Qualities:

Much like when forecasting for hail, one key factor to examine is the atmospheric characteristics of the air mass from 700 MB to 250 MB. First aspect to determine is the nature of the air mass’s temperature and dew point at these levels. Assuming of course that the lower levels are unstable and high in humidity, the colder and drier the air
mass at these levels, the more likely that a down draft will develop. The reasoning behind this statement is that the colder and drier the air mass is in relation to the lower levels, the more extreme is the evaporational cooling. The colder the air mass after evaporational cooling, the faster this air mass sinks to the surface due to the rain-cooled air being significantly more dense than the surrounding warmer environment.

**3.4.1.2 Upper and Mid Level Jet Streaks:**

The position and influence of upper level jet streaks can also significantly enhance the influence of down draft wind gusts. Jet streaks at the mid and upper levels have the ability to advect more cold, dry air into the mesoscale processes of a thunderstorm. This addition of cold, dry air can enhance the unstable nature of the atmospheric column and create a higher potential for evaporational cooling. At the same time, the momentum of the air in the jet streak becomes entrained in the down draft column, which can cause a thunderstorm line to “bow” to the east. This bowing develops because of portion of thunderstorm line is moving faster than the rest of the thunderstorms within the line. At times, if the down draft becomes strong enough, the momentum or speed of movement of the air at 700 to 250 MB is transported to the surface, which leads to the development of very strong wind gusts.
The mid and upper level jet streaks can develop in the synoptic scale via influence from the Polar jet stream or at a mesoscale level via the development of a mesoscale low within a thunderstorm line.

Downbursts are very strong down draft winds, capable of reaching speed of 150 mph, that hit the surface and then expand outward at all directions. A downburst with no precipitation is called a dry downburst. In this case the precipitation within the column of sinking air evaporates because the sinking air is very cold and dry. A downburst with precipitation is called a wet downburst. A downburst the size of 2.5 miles or less are called microbursts, while downbursts larger than 2.5 miles are called macrobursts.

3.4.2 Derecho:

A derecho is an exceptionally large downburst that can exceed 200 miles wide and 1000 miles in length. The duration of a derecho can exceed 12 hours in extreme cases. A derecho develops due to the interaction of upper level jet streaks with the down draft motion of individual thunderstorms. As a line of thunderstorms matures and the down draft becomes stronger, the jet streak at the upper levels enhances the momentum transfer and drives the storms east in a bow fashion. The difference between an average downburst and a derecho is the length of time and area of coverage. Due to the wind shear that
derechos create, weak tornados can develop in the thunderstorm complexes.

There are three different types of derechos. A serial derecho has multiple bow embedded in a squall line usually around 250 miles long. Serial derechos normally are more likely to spawn tornados. A progressive derecho is a small line of thunderstorms that can travel hundreds of miles along a stationary boundary at the synoptic or mesoscale level. Progressive derechos normally do not produce tornados. A hybrid derecho has both characteristics of serial derechos (associated with a deep low pressure system and progressive frontal boundary) but are smaller in size like progressive derechos.

3.4.3 Straight-line winds:

Once a downburst hits the ground, the air expands outward beyond the thunderstorm. This expansion of the cool air is called straight-line winds. Typically, straight line winds are very strong wind that are found in the Plains ahead of thunderstorms, but can develop in the northern Mid Atlantic. The straight line winds are normally stronger in the Plains due to a lack of geographic friction compared to locations in the northern Mid Atlantic. Straight-line wind gusts are typically associated with gust fronts and outflow boundaries.
3.5 Tornadoes:

Tornadoes are one of the most destructive weather disasters a person could face. The damage associated with tornadoes can transport a car into another county in a matter of seconds, clear a house clean to the foundation, and erase the existence of an entire town. Tornadoes are normally strongest in the Plains where a lack of surface friction allows tornadoes to move over large areas of distance and grow to over 250 miles in circumference.

Fig. 2.4-1 The Fujita tornado scale (F scale) pegged to damage-causing windspeeds. The extent of damage expressed by the damage scale (f scale) varies with both windspeed and the strength of structures.
Tornadoes in the northern Mid Atlantic are typically weak due to the lack of open level landmass, which leads to high levels of surface friction, which destabilizes the development of rotating columns of air.

A tornado forms due to two processes. The first is the presence of vertical wind shear in which the direction of the wind is veering with height and speed increases. For example, winds at the surface from the southeast at 15 mph while winds at 500 MB are from the northwest at 70 mph. The wind shear at the lower levels of the atmosphere creates a horizontal column of rotating air. The next process is a strong up draft motion with the thunderstorm. The up draft takes the horizontal rotation column and twists the air in a vertical orientation. This twisting and pulling causes the column of air to stretch, which causes the rotating air to exellerate.

When Helicity (measurement of directional and speed shear) and CAPE (measure of Convective Available Potential Energy) are both large, the potential for a significant tornado is greatest as both properties described above are present.

3.5.1 Enhanced Fujita Scale:

After a tornado is observed, the National Weather Service sends out a team of meteorologist to determine the
strength of the tornado. The following scale is used for that determination.

CHAPTER FOUR
FORECASTING TOOLS

In this chapter, I will break down the tools and philosophy I use to create a forecast for severe weather. While no forecast is always going to be completely accurate at every location, if you use these tips more times than not you will end up with a verified severe weather forecast.

4.1: Surface Observations and Soundings:

For thunderstorms to develop, the atmosphere has to be in a state of instability. So step one to forecasting for severe weather is to examine the surface observations.

I first make sure that extensive low level cloud cover is not an issue. Low level cloud cover, especially fog, is a clear sign of a stable air mass over the region. Now, this does not mean severe weather can not develop later in the day, but getting an understanding of what inhibiting factors the air mass must overcome can provide insight in the amount of energy the atmosphere must use to support severe weather. Usually, if morning fog does not start to break up
after 11 AM, the threat for severe weather begins to decline significantly.

Equally as important is the relationship between the dry bulb air temperature and the dew point. If an air mass is relatively dry, thunderstorms will not form in most cases. Low level moisture is the fuel that feeds a thunderstorm and a lack of fuel means a thunderstorm is going to have a hard time remaining sustained.

Next I look for low level boundaries. These boundaries could be a developing sea breeze front, an old out flow boundary from thunderstorms in the morning, and other types of mesoscale boundaries. These boundaries, aside from the actual cold/warm fronts can be initiation points for thunderstorm development.

Another important tool is the observed sounding. For the Northern Mid Atlantic, the closest sounding for the region is found at OKX, which is from the National Weather Service in Upton, New York. The problem with this sounding is that the location is on Long Island and does not provide an accurate representation of the air mass over the Philadelphia metropolitan area or even the immediate New York City metropolitan area. You see, the soundings at OKX or Upton, New York are influence by the marine air mass of the Atlantic, so the sounding is typically more stable than the rest of the region. Still, my suggestion to you is to take an average of the sounding in
Washington, D.C. and the sounding from Upton, New York to get an accurate feel of the air mass in place.

4.2: Upper Level Observations:

While surface observations are updated every hour, upper level observations are typically updated only twice a day. Once at 12Z which is around 6 AM and again at 00Z which is around 6 PM. In rare cases the National Weather Service office may perform another sounding to determine the nature of the atmosphere ahead of a significant severe threat.

There are several key features I look for in severe weather environments at the mid and upper levels. One feature is the position of the jet streaks at various levels of the atmosphere. The orientation and strength of the 850 MB jet streak for example, can give me an idea of the moisture advection potential into the forecast area, mid and low level wind shear via the strong jet streak, and enhanced lifting from the low level jet streak. Meanwhile, at 500 MB and 250 MB, we can learn the potential of enhanced upper level divergence which allows a thunderstorm to grow to optimal heights. Further, jet streaks can also introduce dry air in the upper levels from 500 MB on up that can support the development of down burst winds, large hail, and frequent lightning.
Besides jet streaks, the position of heights in the atmosphere can also detail the strength of upper level disturbances. These disturbance can enhance lifting of certain locations while produce subsidence over other locations. The position of the enhanced rising and sinking motion within the atmosphere would lead to the development of severe thunderstorms in one area and only a few scattered showers in other areas. Forecasting for severe weather hinges on the ability to pick out these disturbances as early as possible. I prefer using the water vapor satellite picture to find these disturbances, which can be observed very easily. A strong disturbance typically produces enhanced sinking air to the north and northwest of the axis of the short wave trough. The stronger the sinking air (usually seen as black and red), the strong the disturbance. Analyzing these disturbances can help determine if your location will be impacted by severe weather or remain dry with no impacts at all.
4.3: SKEW-T LOG AND SOUNDINGS:

Understanding and using the Skew-T Log soundings is extremely important in forecasting severe weather. The Skew-T log gives a forecaster a unique view of the atmosphere with a three dimensional analysis of the direction and magnitude of winds, the state of the thermal profile of the atmosphere, and location of potential mechanisms for lifting.

Not all sounds are the same of course and different attributes of a sounding leads to different convective features. For example, if the atmosphere is saturated into the mid levels of the atmosphere, there is a very strong likelihood that a heavy rain event will materialize. If in this same sounding there is a pocket of dry air, a strong down draft can develop as the moisture at the lower levels is forced to evaporate in that level. Then you have to look at the winds at that level.

Let’s say there is strong 500 MB jet streak where there also happens to be dry air. As that air cools via evaporational cooling and sinks to the surface, some of the air momentum (wind) is transported to the surface. This is a sounding suggesting down draft winds are possible.
If up draft and down draft winds are developing, then you have to examine the direction of the winds from the surface to the mid levels. Are the winds unidirectional (same direction) or changing with height? If the wind direction is change with height with an up draft and down draft circulation, then this sounding would suggest a high probability of a tornado. If the winds are all in one direction, then down burst and straight line winds are more likely. Note however that in cases where the speed shear is exceptional (where winds are very fast at one level and much slower at the surface), brief, weak tornadoes are possible.

Before going forward, let’s go over the details on a Skew-T chart.

4.3.1: Parts of a Skew-T Chart:

The Skew-T chart is made up of several important features that helps us look at the atmosphere in a three dimensional way.

Isobars are lines of equal pressure. They are the horizontal lines on the skew-T chart, which are labeled on the left side. In most charts, the pressure is given in 50 to 100 MB increments and usually range from 100 MB down to the surface.
Isotherms are lines of equal temperature and run from southwest to northeast on the chart. These lines are solid. The lines are typically in 10 degrees Celsius increments and are defined at the bottom of the chart.

The Saturation Mixing Ratio lines represent the mass of water vapor divided by the mass of dry air at grams per kilograms. These lines are from southwest to northeast as well but are dashed. These lines are also labeled at the bottom.

Wind barbs are on the right hand side of the chart. The wind speed and direction is defined at 50 MB increments in Knots. The barbs are orientated via the direction of the wind. The wind speed is defined by the following:

Half barb = 5 KT, Full barb = 10 KT, Triangle = 50 KT

The Dry Adiabatic Lapse Rate is the rate of cooling of a rising unsaturated parcel of air. These lines represent cooling at 10 degrees Celsius per kilometer and are oriented from southeast to northwest. The lines are solid on the chart.

The Moist Adiabatic Lapse Rate is the rate of cooling of a rising saturated parcel of air. These lines are oriented from south to northwest and rapidly increases with height. In other words, the rate of cooling is dependent on the amount of moisture in the air. This is why dry air at the
mid and upper levels of a sounding is so important. The drier the mid and upper layers of the atmosphere, the faster the parcel being lifted can rise in the atmosphere.

The sounding itself is the actual temperature and dew point reading of the atmosphere from the surface to 100 MB.

Finally, the Parcel Lapse Rate itself is a parcel at the surface that is raised along the Dry Adiabatic Lapse Rate until that parcel reaches condensation (read below on Lifting Condensation Level) and then is raised via the Moist Adiabatic Lapse Rate. This lapse rate is how all indices including Lifting Index and CAPE are calculated.
Example:

Skew-T Diagram

- Dry Adiabatic Lapse Rate
- Saturation Mixing Ratio
- Isotherms
- Moist Adiabatic Lapse Rate

Pressure

P (mb)

Temperatures

-30 -20 -10 0 10 20 30 40

Saturated Mixing Ratio

T g/kg

Sea Level (Approx)
4.3.2: Lifting Condensation Level:

The Lifting Condensation Level or LCL is the pressure level a parcel reaches saturation via lifting the parcel to a higher level of the atmosphere. For example, when water vapor at the surface is lifted to the mid or upper levels, there is a point in which the parcel cools to the point where water vapor condenses into a liquid form. That level is the Lifting Condensation Level.

You can find the LCL on a Skew T chart by take the temperature and dew point at the surface and lift the temperature along the dry adiabatic line and the dew point along the mixing ratio line. Where those lines meet is the Lifting Condensation Level!

The LCL can be used to determine the level at which the cloud base will form. This is important in severe thunderstorm develop in order to determine how much lifting is needed in the atmosphere to initiate thunderstorm development. The lower the LCL, the less amount of lifting is needed for thunderstorm initiation. Further, the lower the LCL is in the atmosphere, the higher the potential for tornadoes to develop in supercell thunderstorms because the potential for a large CAPE (we’ll discuss CAPE in the following sections) value, which means more buoyant energy is present in the atmosphere.
HOW TO FIND THE LIFTING CONDENSATION LEVEL
4.3.3: CAPE (Convection Available Potential Energy)

The Convection Available Potential Energy or CAPE is the area on a Skew-T chart sounding in which the parcel being lifted is warmer than the surrounding atmosphere. The CAPE is calculated first by lifting a parcel to the Lifting Condensation Level. From that point, the parcel is lifted along the Moist Adiabatic Lapse Rate. The area between the actual sounding temperature and the path in which the parcel is rising along the Moist Adiabatic Lapse Rate. The larger the area, the more impressive the CAPE. The CAPE is calculated in units of Joules per kilogram which is energy per unit mass.

The larger the value of CAPE, the faster a parcel will rise. The faster a parcel rises, the faster a thunderstorm can build in the atmosphere. Typically a CAPE of 1,000 J/kg is where I like to start in examining the potential for thunderstorms. A CAPE above 1,500 J/kg is considered a large CAPE, capable of supporting damaging hail, strong down drafts, and frequent lightning. A CAPE value over 2,500 J/kg is considered an extreme CAPE value. CAPE values this high are usually associated with thunderstorms that produce large hail, frequent lightning, very heavy downpours (depending on the nature of the low level air mass in terms of moisture content), and tornadoes.
Remember, the CAPE is an indication of how warm the parcel being lifted compared to the surrounding atmosphere. The colder and drier the air mass is aloft, the larger the CAPE value thus the higher potential for large hail, strong up draft and down draft circulations, and the more enhanced the charge is in the atmosphere.

4.3.4: Level of Free Convection:

The Level of Free Convection or LFC is basically the level at which CAPE begins in the troposphere. It is the point at which the temperature of the parcel being lifted is equal to the temperature of the surrounding environment. At this point, the parcel is able to rise independent of lifting because the parcel is warmer than the surrounding environment.

The LFC is very easy to find on a Skew T Chart by taking a saturated parcel or a parcel at the Lifting Condensation Level and raising that parcel Moist Adiabatically to the point in which the parcel is at an equal temperature as the environment. The closer the LFC is to the LCL, the larger the CAPE value and thus the more unstable the atmosphere.
4.3.5: Equilibrium Level:

The Equilibrium Level is found at the top of the CAPE and is the level in the atmosphere at which the parcel rising in the atmosphere is equal to the surrounding environment. Basically this is where the thunderstorm stops growing because the parcel (usually water vapor) can no longer rise on its own or is no longer buoyant.

Let’s look at this via an Infrared Satellite picture. When we look at the satellite picture and see oranges, reds, and even grey colors we know that the cloud tops are very cold. The cloud tops are very cold because the parcel at the top of the cloud has reached the evaporation point. So when we see cloud tops getting colder, what we are really seeing is the Equilibrium Level expanding higher into the atmosphere!

The higher the Equilibrium Level, the higher the thunderstorm can build into the atmosphere. If the thunderstorm is able to build high up into the atmosphere, this is an indication of a very strong updraft and also supports the potential for large hail, down burst wind gusts, heavy rainfall, frequent lightning, and tornadoes.
4.3.6: Convective Inhibition:

Convective Inhibition or CINH is the area on a sounding between the surface and the Level of Free Convection or LFC. The larger this area is, the more lifting is needed in order for thunderstorms to develop. This is basically what meteorologist mean when they talk about breaking the cap on the atmosphere as basically the parcel is not able to rise independently as the parcel is cooler than the surrounding environment.

The CINH can be reduced in four ways. The first is simply day time heating. In the morning, the atmosphere is typically capped because the air mass above the surface is warmer than the parcel at the surface. Thus the parcel can not rise, but only sink. As long as the temperatures at the higher levels of the atmosphere do not rise, typical day time heating warms the surface and makes the atmosphere unstable.

The other examples usually produces elevated thunderstorms. The parcel can be lifted synoptically via an upper or mid level disturbance or a cold front. The parcel can be lifted via low level convergence via a mesoscale feature like an outflow boundary or a sea breeze front. Or a parcel can be lifted via warm air advection, which basically is a warm front.
A CINH value below 50 J/kg is typically very weak and can be overcome rather easily. A CINH value between 50 and 200 J/kg typically needs some additional forcing beyond day time heating like a strong mid or upper level disturbance. Finally a CINH over 200 J/kg is typically a very strong cap and limits the development of thunderstorms.
4.3.7: Severe Weather Indices:

There are a variety of indices that meteorologists use to determine the potential for severe weather. The following are a brief overview of these indices and how to calculate them.

4.3.7.1: Lifted Index:

The Lifted Index or LI is used to determine the instability of the atmosphere in the troposphere. The LI is calculated by taking the temperature of the environment at 500 MB minus the temperature of the parcel being lifted at 500 MB. If the LI is positive, this means the atmosphere is stable. If the LI is negative, then the atmosphere is unstable. The more negative the LI the more unstable the atmosphere is at the mid levels. I tend to be very careful with using the LI as the sole measure of the instability of the atmosphere as the parcel has to be able to break the CINH first. The lifting index is in degrees Celsius.

A lifting index of 0 to -4 Celsius indicates marginal instability. A lifting index of -4 to -7 Celsius indicates large instability. A lifting index of -8 or greater indicates an extreme instability in the atmosphere.
4.3.7.2: Showalter Index:

The Showalter Index or SI is best used when in combination with the Lifting Index. The Showalter Index is the measure of change in temperature from 850 MB to 500 MB. The SI is calculated by taking the environmental temperature at 500 MB and subtract that temperature from the parcel temperature at 500 MB raised from 850 MB. In other words, instead of starting with the temperature at the surface, you are starting at observed temperature at 850 MB. Just like at the surface, first you must find the LCL of the 850 MB temperature and then raise the parcel moist adiabatically to 500 MB. The drier the atmosphere at 850 MB, the more stable the atmosphere. This is basically a measure of the stability at the mid levels.

If the LI is negative and the SI is positive, then a cap exist in the mid levels of the atmosphere and must be overcome.

If the LI is positive and the SI is negative, this means the cap is at the lower levels of the atmosphere.

If the LI and SI are negative then the atmosphere is significantly unstable.

If the LI and SI are both positive than the atmosphere is significantly stable and no thunderstorms will develop.
The SI index is best used in set ups where a warm front is moving through the region as the index gives us a better idea of the degree of instability moving into the mid levels of the atmosphere and thus the potential for elevated thunderstorms.

4.3.7.3: Total Totals Index:

The Total Totals Index or TT is calculated by combining the difference between temperatures at 850 and 500 MB and the dew points at 850 MB and temperatures at 500 MB.

TT= \( (T_{850\ MB} - T_{500\ MB}) + (T_{d\ 850\ MB} - T_{500\ MB}) \)

The TT index is not my favorite index to use. There are a lot of problems with using this index as the sole determination for severe thunderstorm potential, but can be used as a supportive tool to collaborate with other indices. For one, the TT can not pick up caping layers and does a poor job with handling moisture in the atmosphere. Below 850 MB. Further, the index can not be used in rising elevations nor has a stable value from season to season.
If you must use the TT index the following are operational significant levels:

TT of less than 44 suggests no convection

TT of 44 to 50 has a likely potential for thunderstorms

TT of 51 to 52 has an isolated potential for severe thunderstorms

TT of 53 to 56 has a widely scattered potential for severe thunderstorms

TT of over 56 has a scattered potential for severe thunderstorms

4.3.7.4: Helicity:

Helicity is the measure of vertical wind shear in the atmosphere. The higher the helicity, when there is a large CAPE, the higher the potential for tornadoes.

The Helicity is calculated by deriving the increase in wind speed from the surface to 3 KM and the change in direction of the wind speed from the surface to 3 KM. I’ll spare you the calculus, but honestly it is one of the easiest calculation you’ll do in meteorology.
Helicity Values:

Helicity values of 150 to 300 suggest a possible supercell.

Helicity values between 300 to 400 suggest supercell development is favorable.

Helicity values over 400 suggest high tornadic possibilities.
4.4: Radar:

When ever severe weather develops, you can count on meteorologist showing radar scan after radar scan, but what are you really looking at?

Today, the radar method used is called Doppler radar. The Doppler effect, named after Austrian physicist Christian Doppler, is the change in frequency of a wave for an observer moving relative to the source of the wave. The type of Doppler effect weather radars use is called Pulse-Doppler which not only detects the target’s bearing, range, and altitude but also measures the radial velocity.

On the radar, precipitation is measured in dBz or decibels of Z. Z stands for reflectivity which is $1 \text{ mm}^6/\text{m}^3$. The dBz, via observations as the radar system was developed has the following scale:

- dBz of 40 or more equals heavy rain
- dBz of 24 to 39 equals moderate rain
- dBz of 8 to 23 equals light rain
- dBz below 8 is drizzle or no precipitation

The National Weather Service of the United States uses NEXRAD, which stands for Next Generation Radar. Is it me or does NOAA try to put everything into acronyms?
Anyway, NEXRAD is basically a radar that used the Doppler effect to obtain weather information. The radar emits a Doppler wave frequency. That wave hits an object, like rain, snow, or a bird and that wave returns back to the source. The bird issue is why a meteorologist should keep an eye on other observations and satellite pictures before trusting the radar completely.

There are a variety of features you can use on a radar, which are all useful depending on the situation.

Base Reflectivity is the display of reflectivity or the amount of energy sent back to the radar at the lowest tilt of angle from the horizon, which is 0.5 degrees. Base Reflectivity is measured in dBz.

Composite Reflectivity is the display of reflectivity at any angle at all ranges of the radar. Composite Reflectivity helps define intensity trends, storm structures, and the maximum intensity of thunderstorms, especially when compared to Base Reflectivity.

When forecasting, it is important to use both radar features to get a true feel on the development of the thunderstorm. For example, the composite reflectivity pin points the growth of stratiform rainfall behind a thunderstorm and also the development of the thunderstorms at higher elevations. However, the base reflectivity is by far better in picking up outflow boundaries,
sea breeze fronts, and developing thunderstorm cells. When forecasting in an environment when thunderstorms are expected to rapidly develop once the convective temperature is reached, the base reflectivity radar data must be used!

The great advantage in using the Doppler Effect is the ability to use the data to define velocity, storm motion, and precipitation rates.

The Base Velocity feature represents the wind field around the radar. Green colors indicate winds moving towards the radar and red represents winds moving away. So why is this useful? Well, if a tornado begins to develop the greens and reds will be right next to each other within the thunderstorms. If winds are moving towards and away from the radar within a thunderstorm than a circulation is developing. The radar can pick up the winds moving away and toward the radar base up to 143 miles away.

Storm Relative Motion feature takes the base velocity idea and takes one step further. The display shows wind relative to the specific storm’s motion. This process allows the meteorologist to clearly observe the development of rotation within a thunderstorm, without trying to separate the storm motion from the rotation within the thunderstorm. Just like the Base Velocity, green colors indicate winds moving towards the radar and red colors indicate winds moving away from the radar.
The Doppler Radar also can help determine precipitation amounts out to 143 miles. The one hour radar precipitation data is an estimate of rainfall over a 60 minute period. The rainfall estimates are taken via correlating dBz returns to the intensity of the rainfall rate. The Storm Total Precipitation data is the rainfall rate back to the last full hour of no precipitation. This feature can be used to highlight flood prone areas. I should note that features like hail, can cause significant errors in rainfall amount estimates due to the hail having a very high reflectivity.
4.5: Satellite:

The National Oceanic and Atmospheric Administration or NOAA uses the Geostationary Operational Environmental Satellites or GOES to provide meteorological data to the general public via the National Weather Service. There are three forms I will discuss here that can be used when studying severe weather potential, which are visible, infrared, and water vapor.

4.5.1: Visible Satellite:

The visible satellite picture is basically an image of the earth in visible sun light. When the sun sets, the satellite is basically useless. Despite that obvious fact, the visible satellite is a very important tool that can be used to pick up developing thunderstorms and mesoscale boundaries. For example, when looking for an area of developing super cell thunderstorms or even a line of thunderstorms, the best way to look for development is on the visible satellite picture. If there is an area of scattered cumulus, rapidly growing in the afternoon hours, this area is likely where thunderstorms will begin to develop. Conversely, the visible satellite picture can also show a meteorologist where strong subsidence is occurring via very clear skies developing ahead or behind an area of thunderstorms.
The visible satellite picture can also be used to detect low level boundaries like sea breeze fronts or mesoscale outflow boundaries.

The visible satellite picture can also be used to detect the thickness and heights of clouds. Clouds that are thicker with water vapor and thus reflect more light show up as bright white.

4.5.2: Infrared Satellite:

The infrared satellite picture uses heat or the lack of heat of a given object to detect radiant energy. The warmer(colder) the cloud, the less(more) energy is given off. The infrared has several different advantages over the visible satellite picture, one of which happens to be that the satellite can be used at any time of the day or night.

The infrared satellite picture shows the temperature of cloud tops. The colder the cloud top, the higher the cloud is building into the atmosphere and thus how much lifting is associated with the thunderstorm. Typically, infrared satellite clouds with temperatures under -40 degrees Celsius end up being severe thunderstorms. When cloud tops become this cold or colder, the cumulus cloud has built high into the atmosphere, which would indicate a strong updraft is associated with the thunderstorm. Cloud cloud tops would also indicate the existence of a dry, cold layer of air in the upper level which would suggest the
potential for a strong down draft component as well. These features of course would lead to large hail, heavy downpours, frequent lightning strikes, strong down burst winds, and potentially tornadoes.

The draw back with the infrared satellite picture is that the mesoscale features can become masked. While the infrared is an excellent tool to pick up areas of lifting as the lifting is intensifying, the satellite lacks the ability to pick up on developing thunderstorms or intensifying mesoscale boundaries until the lifting is strong enough to produce a thunderstorms, which by then is too late.

4.5.3: Water Vapor Satellite:

The water vapor satellite picture detects water vapor in the top third of the troposphere (where weather happens). The water vapor satellite picture is an excellent tool for picking up areas of mesoscale lifting and sinking in the atmosphere. Specifically, this tool is excellent for picking up the influences of mid level disturbances that typically creates a lot of volatility in a forecast. When a strong disturbance interacts with a line of thunderstorms, strong sinking air can unexpectedly inhibit thunderstorm development in an area where the rest of the thunderstorm line is developing. The water vapor satellite picture shows the forecaster (exactly where mid level disturbances are in the atmosphere due to the increase in moisture in one area (water vapor is rising from the
surface to the upper levels) and the decrease of moisture or dry air in another area (cold, dry air from the upper levels is sinking). We can determine how strong a disturbance is via the sharp gradient between these two contrasting areas. The sharper the gradient, the more intense the disturbance and thus the higher potential for severe thunderstorms. At the same time, by being able to track the mid level disturbance in real time, we can also quickly update a forecast of what locations will get severe weather and what locations will miss out.
5.0: Forecasting Philosophy:

When developing a forecast for severe weather, my best advise is to expect the unexpected. Forecasting for thunderstorms, let alone severe thunderstorms, is an extremely difficult task. You as the forecast have to take into account not only the nature of the atmosphere in your location, but also in areas surrounding your location. The forecaster must take into account of mesoscale features, mid level dynamics, upper level disturbances, developing subsidence boundaries, and how fast clouds will lift and break up before noon. That’s just for starters!

The key to understand here, is that if you are looking to be 100% in forecasting severe weather, you are in the wrong field. That’s just being honest. You’ll never catch everything, but you can pick up the changes in the atmosphere as they happen if you know what you are observing and how what you are observing will impact your forecast.

So in this section I am going to take you threw a forecast and what I look for. You’ll develop your own steps as you
get used to your own forecasting techniques. Some work, some don’t but you’ll never know unless you try.

5.1: SPC Discussions

A man is not an island, and the same can be said for a meteorologist. The first place I check when I’m thinking that severe thunderstorms will be a threat is look at the Severe Prediction Center. Are they perfect? No, of course not, but what you do have is a source of information from meteorologist, some of the best in the country, breaking down every region of the country for severe threats.

The SPC discussions include a Convective Outlook Discussion that is broken down in typically three 24 hour periods, sometimes 12 hour periods in the morning initially and Mesoscale Discussions that detail immediate severe threats that are developing.

The Convective Outlooks are excellent for getting a general picture of how the atmosphere is evolving throughout the country. This outlook provides an excellent overview of the country with detailed analysis without having to spend hours breaking down each severe threat.

Why is a severe threat in Ohio important to New Jersey? Well, by learning the nature and evolution of the severe
event to the west of your location, you can begin to establish the severe threat for your location. Was there a great deal of wind shear? Was there a low level jet streak or a strong upper level disturbance? Did the severe weather develop due to mesoscale or synoptic considerations? All of these answers can help you in including or ruling out severe weather characteristics in your forecast.

The other discussion is the Mesoscale Discussion, which provides in depth analysis of a specific severe weather threat. Again, paying attention to not only your forecast area but locations around your forecast area comes into play. Let's say for example there is a Mesoscale Discussion for Baltimore, where a strong disturbance is starting to intensify. Well, if a strong disturbance is starting to intensify to the south of the Philadelphia metropolitan area, this means that sinking air or subsidence is more likely over the Philadelphia metropolitan area and New Jersey. Thus, the severe threat that might have been possible due to other factors is now overwhelmed by the sinking air from the disturbance. So, while the Mesoscale Discussion was not issued for your location, you now have verified information on the evolution of the atmosphere!
5.2: Model Guidance

Model guidance, especially in the mesoscale level, has come a long way over the past 5 to 10 years, however no model is never 100% accurate. When using model guidance, be sure to use the appropriate models for the forecasting situation.

Global models tend to not have the resolution high enough to handle severe thunderstorms due to convective feedback errors. This is why the models tend to have circular “bulls eyes” in the QPF forecasts. These models include the GFS, ECMWF, CMC, UKMET, and Ensemble guidance. However, global models are good to use in forecasting for several synoptic features like mid level jet streaks, upper level troughs, and the temperatures of the atmosphere. Global models can show a forecaster where cold pockets at 500 MB are likely to develop, which can help in finding areas of potential severe weather several days ahead of the event.

Mesoscale models include the NAM-WRF, SREF, and RUC guidance. These models have a very high resolution and are very good at picking up mid level disturbances, low level boundaries, and other small scale disturbances. The problem with these models is the limited time period these models are useful. I would strongly suggest not using the RUC past 9 hours. The NAM-WRF and SREF guidance is reliable up to 60 hours in my opinion.
5.3: Observations

Never take a model for granted! Before using models you must first examine your observations. By examining the observations at all levels of the atmosphere, you can pick up what the models may be missing or perhaps are under/over forecasting in the short range period.

There are several features you can find via observations, some of which can have a significant impact on your forecast. The tools you can use range from radars, satellite pictures, surface observations, and soundings.

By examining the local observations I first want to get a look at the latest soundings in and around the Northern Mid Atlantic. Where is the most unstable air mass located? Is this air mass moving towards the forecast area? Atmospheric soundings are usually available at 00Z (7 PM) and 12Z (7 AM), so in between those time periods the surface observations are equally important.

With the surface observations, I want to look for areas of wind convergence, moisture convergence, cloud cover, temperature trends, and dew points. By examining these trends I can find low level boundary areas, the rate at which the atmosphere is destabilizing, and potential inhibiting factors like dry air.
A website that is a major tool in examining observations is the SPC Mesoscale Analysis Page. This page provides a variety of information for the latest analysis and observations from the surface to 200 MB, including temperatures, dew points, wind speed, wind direction, pressure falls, height falls, and other features. In addition, you can use this page to track the development of wind shear, various thermodynamics like the low and mid level lapse rates, and even severe weather composite indices like the Bulk-Richardson numbers!

I would strongly suggest using this source of information at all times when forecasting for severe weather.
5.4: Making Your Forecast

Now that you have your observations, radar, and satellite data combined with the latest model guidance, you can now prepare your forecast.

When making a forecast for severe weather, I make sure I pinpoint the areas with the best combinations to support severe weather. By starting at this point, I can branch out and set up the parameters for what type of severe weather to expect.

Experience from previous severe weather events helps a great deal here. Be sure to write down and record every severe weather event that you have observed and forecasted for (this includes all weather forecasting problems), as you may have experienced a similar set up and thus can afford mistakes or errors from the past.

The forecast should have the basic details for the public; the location of expected severe weather, the type of severe weather that will develop and severity of threat, and the time of the threat. By using the information you gather, you should be able to break down each of these details. The sounding tells you how cold the upper levels are and also the whether winds are unidirectional or changing with height. These observations will produce specific severe weather characteristics. The surface observations and observations from the SPC mesoscale
page gives you the areas that are most likely to experience severe weather via the latest trends. Finally, the models, especially the mesoscale models, can be used to given a time period of expected impact on the region. Now, you have your first basic forecast!

Your job isn’t done after the forecast is out. Forecasting for severe thunderstorms or thunderstorms in general will provide one big factor, the unexpected! When dealing with convective systems, any feature from a 800 MB disturbance to a sea breeze front can have a drastic impact on your forecast. This is why you must stay on top of every severe event and constantly monitor the threat. Radar and satellite data are excellent tools for doing this, but I would also suggest keeping an eye on the SPC Mesoscale Analysis page as well. By using these tools you can quickly and correctly alter your forecast without having to wait for the forecast to go wrong to make updates. Remember, this type of forecasting goes beyond just verifying, the forecast or change of the forecast may save someone’s life. Just ask anyone who has ever been through a tornado how much a few seconds or minutes of warning could mean between living or being seriously hurt or worse.
SUGGESTED LINKS

PENN STATE E-WALL

GOES EAST SATELLITE

SEVERE PREDICTION CENTER

SEVERE PREDICTION CENTER TOOLS

NATIONAL WEATHER SERVICE RADAR

WEATHER TAP

PLYMOUTH STATE WEATHER

RAP REAL TIME WEATHER DATA

90
SOURCES

American Meteorological Society Glossary

Mesoscale Meteorology and Forecasting
    edited by Peter S. Ray

Severe Prediction Center