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Topic: Convective Storm Structure and Evolution

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Welcome to the RAC Convective Storm Structure and Evolution lesson on Hodograph Essentials for Convective Storms. I’m Greg Schoor, WDTD Instructor. We will use this lesson to go over basics of hodographs and how to analyze them to get a better picture of the environment. This lesson is actually a prerequisite for the remainder of this topic, because it provides the foundation for the forecasting and understanding of different convective modes and the contribution of atmospheric shear. So, let’s get started!

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Learning Objectives

Given a list:

1) Identify the primary benefit of visualizing vertical winds on a hodograph versus a SKEW-T.
2) Identify the primary reason that hodographs are plotted on a polar diagram.
3) Define each of these types of shear parameters, vertical “bulk”, total, and mean shear.
4) Identify the difference between ground-relative and storm-relative reference frame.
5) Identify the relevance of clockwise curved versus counter-clockwise curved hodograph layers.
6) Identify the difference between streamwise and crosswise vorticity.
7) Identify the two primary applications of using storm-relative helicity (SRH).

Here are the learning objectives. There are quite a few of them but this is a foundation lesson, so – a lot of important elements to it. Please take a moment to read through them – and refer back to them if need be.

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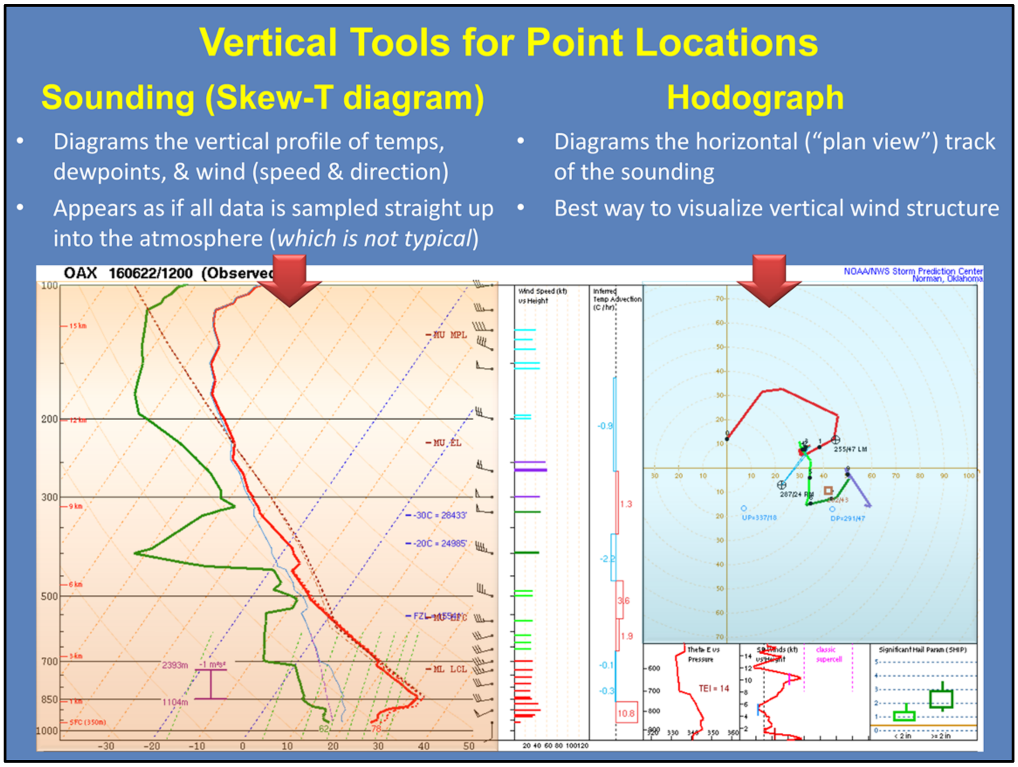

In terms of measuring the atmosphere, one radiosonde can serve two purposes – and really, both are necessary for proper analysis and forecasting of the atmosphere surrounding a point location. First, the sounding – on the left – also known as the Skew-T diagram, depicts the vertical temperature, dewpoint, and wind profile. But the way it is plotted on this chart, makes it seem like this information is being taken straight and directly up from the point where the weather balloon was released. In a controlled environment, with absolutely no vertical shear, that may be true, but realistically, it is not. But what we will focus on is the hodograph – on the right – which shows us in a horizontal, or "plan view" sense, where the radiosonde has travelled, start to finish. This gives the best visualization of the vertical wind structure and goes a long way to helping predict potential convective modes.

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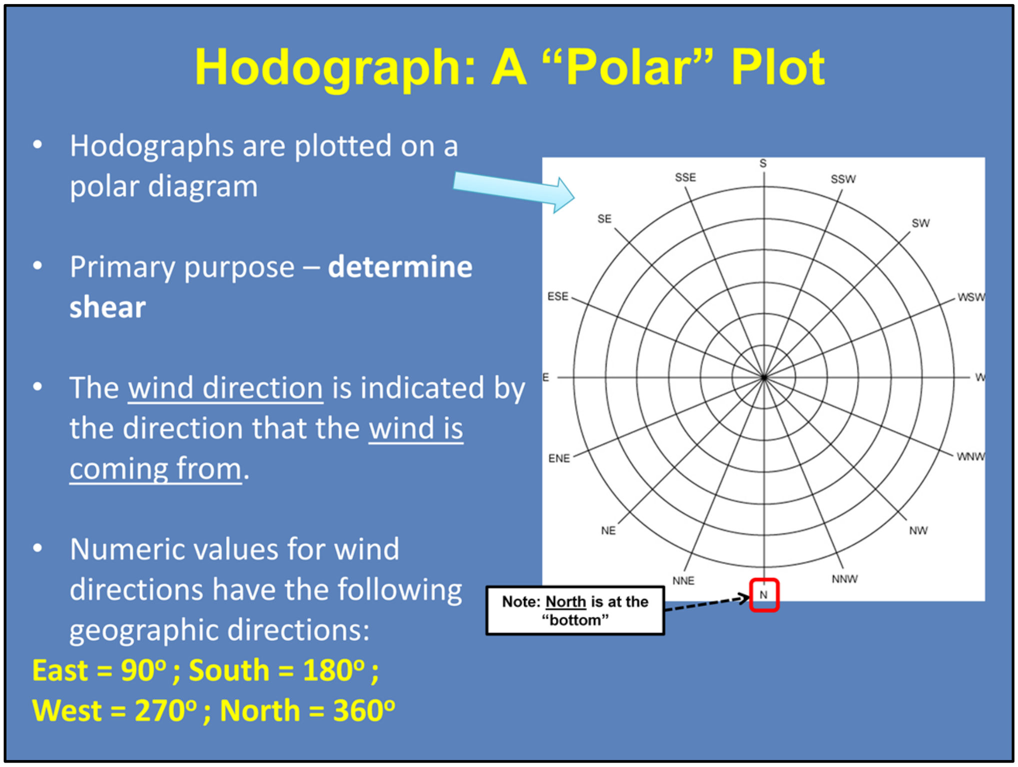
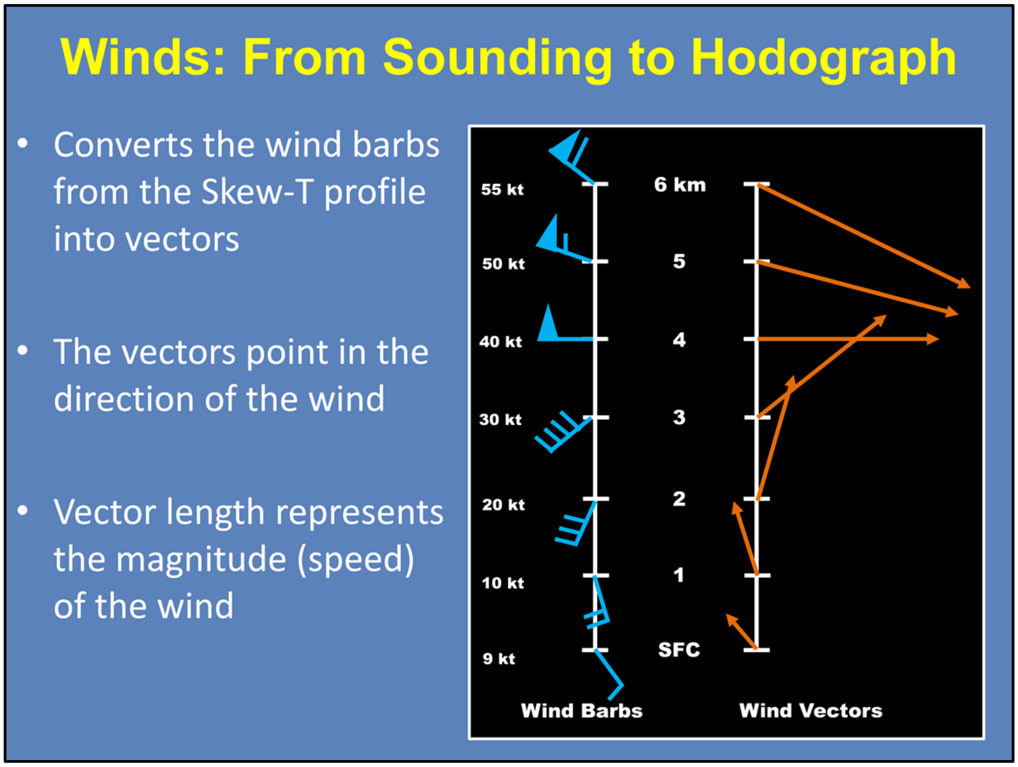


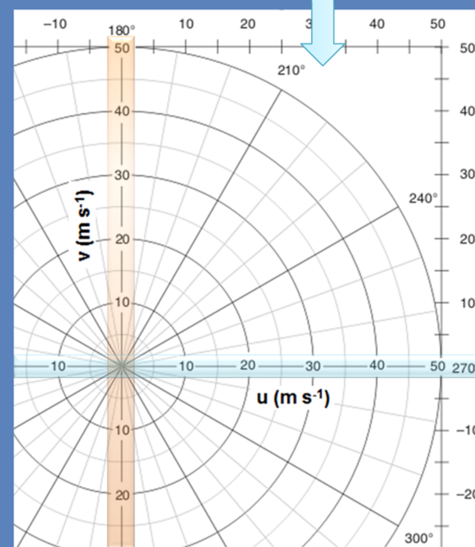
With Skew-T diagrams, three different elements have to be plotted on the same chart, but with hodographs, we are only concerned with wind. Plotting wind on a polar diagram allows us to determine various types of shear. Recall that shear is the change in wind speed and direction between 2 levels or with height, so we're talking about the gradient with the term, "shear". Also recall that wind direction is indicated by the direction that the wind is coming from, so if it's a "south wind", that means it is coming from the south. The hodograph's polar coordinates appear upside-down from what you're probably used to, with north being on the bottom, south on the top – for the Northern Hemisphere.



Next, we have to go back to the Skew-T to grab the wind information before it can be plotted onto a polar diagram. Wind barbs, in blue, look like they would on a Skew-T diagram, showing the magnitude – or speed – with either the tick marks or the flag symbols. The direction at each level goes from the tip of the barb to where the stem of the symbol meets the vertical line. For a hodograph, these values are converted into vectors, orange arrows on the right, in this example. Same information as the wind barbs, but these vectors now point in the direction the wind is going, instead of where it's coming from and they get longer with higher magnitudes. But what are the magnitudes? What does a 20 knot wind barb look like in vector form on a hodograph?

Hodograph: Magnitude Labels

- In the Northern Hemisphere, this quadrant is the most typical



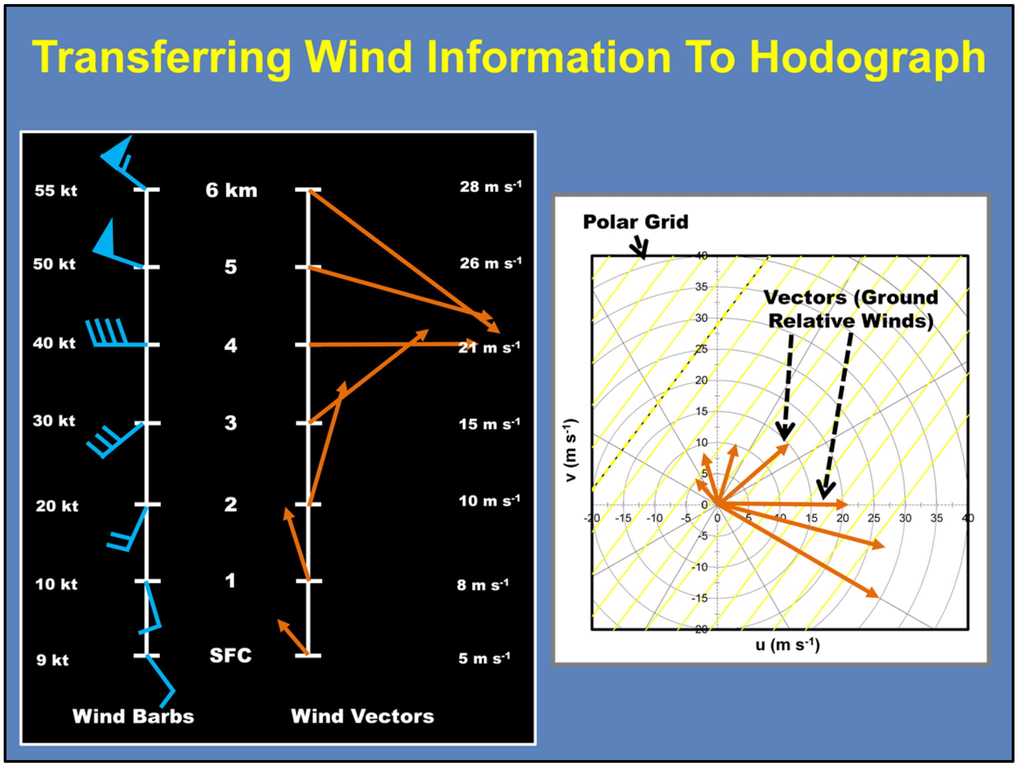
When plotting wind vectors…

- The wind vector is in knots (kts)
- The X & Y Axis are in meters per second ($m\ s^{-1}$)

- The rough conversion is: $m\ s^{-1} = ½(kts)$

Examples:
10 kts = ~5 $m\ s^{-1}$ ($5.144\bar{4}$ $m\ s^{-1}$)
20 kts = ~10 $m\ s^{-1}$ (10.2889 $m\ s^{-1}$)
50 kts = ~25 $m\ s^{-1}$ ($25.722\bar{2}$ $m\ s^{-1}$)

A couple of slides ago, we introduced the polar diagram. So what about the X and Y axis and why are the units from 0 to 50? On a sounding, the wind information is in knots but on a hodograph, we convert that to meters per second. Fortunately, the conversion from knots to meters per second is just dividing the knots by 2…roughly. It's not an exact conversion but very close as you can see from the examples, all the way from 10 knots to 50 knots, the conversion to meters per second is just about half for each one.



Now that we have converted wind barbs into vectors, we'll transfer them onto a polar diagram. So, we basically just pluck them at each level from the diagram on the left, taking the vector's origin and placing it on the center point of the hodograph, at 0, 0 – as you can see from this animation. Each vector points in the horizontal direction the wind is headed at that level. Next, we are going to dive into the concept of shear.

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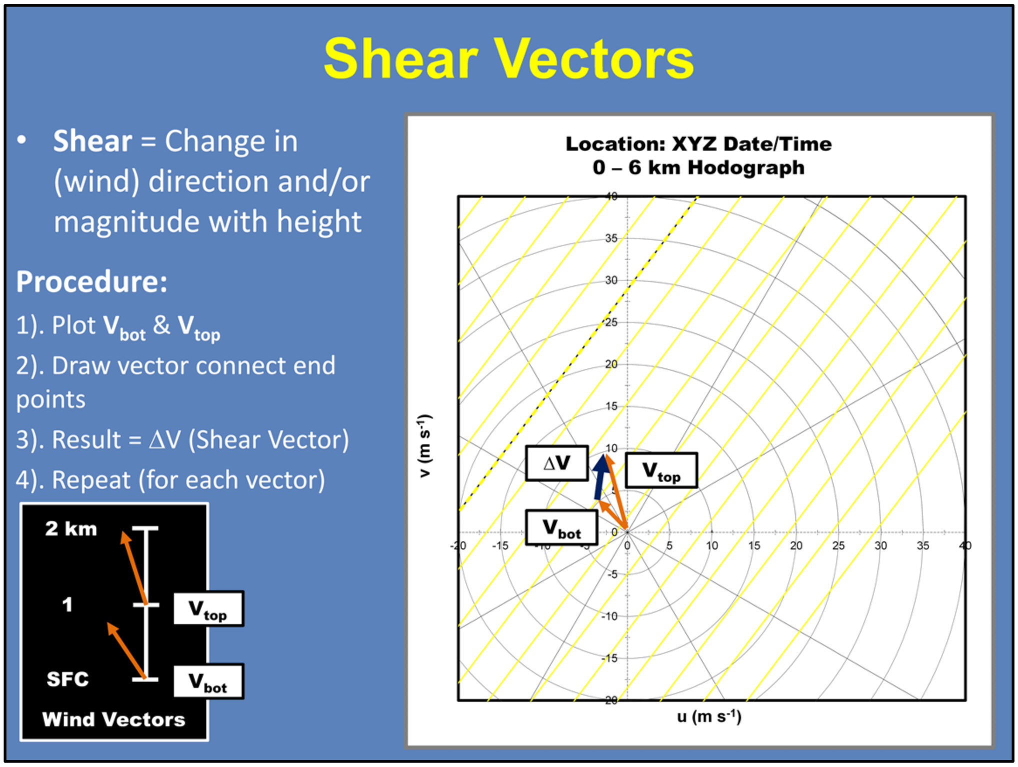

The foundation of the hodograph is this concept of shear…so what is it? Shear – and if we're talking about wind, it's wind shear – is the change in direction and/or speed, with height. From here on out, we're talking about the difference in the wind at two levels, every step of the way. Obviously, hodographs, like Skew-T's, are automatically generated but it is important to know how they are generated – so you can see where the various severe convective parameters come from. So, we'll start at the bottom, with the lowest two wind vectors from the earlier example. Plot them on the diagram and then place a connecting vector at the tips of the wind vectors, pointing the direction from bottom to top of the layer. The result is the Shear Vector, if it was just a line without an arrow at the end, we'd call it a Shear Segment – but a vector has direction. Then, you do this step for the remainder of your wind vectors through the column.

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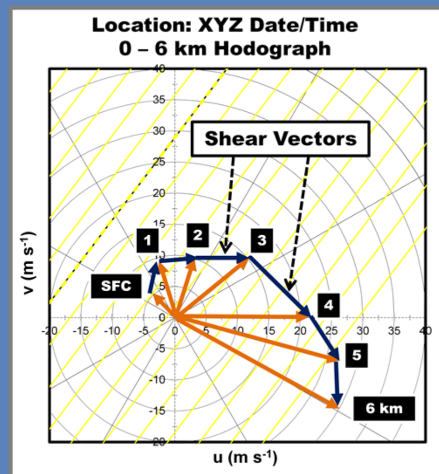
Tracing the Hodograph

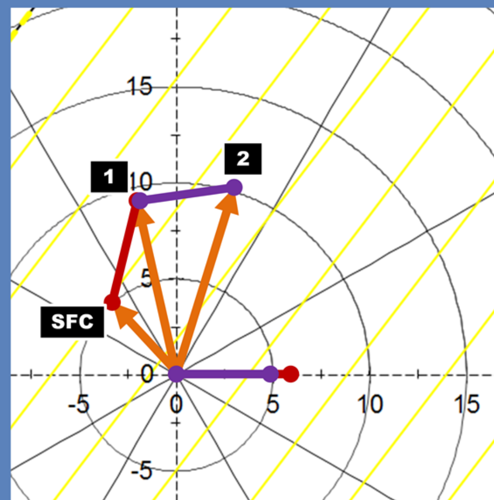
- Once the vectors (ground-relative winds) are plotted – the "**Shear Vectors**" connect the end points
- When complete, this trace is the "hodograph"
  - Then the shear vectors disappear
  - Also, notice that the "hodograph" (blue line) doesn't start at the origin (unless it's calm at the surface)



Now we have taken all of the wind vectors, plotted them, and generated Shear Vectors for each level. Remember, with this example, we had a vector at each 1 kilometer up, from the surface to 6 km. When all of the Shear Vectors are plotted through the column, that trace is referred to *as* the hodograph, the trace of all your wind shear values from the surface to the top of the sounding. Once this is plotted, then you can ignore the shear vectors. Also, you may be wondering why the hodograph line, which is the result of the blue arrows, doesn't start at the origin, or 0,0 on the chart. That would only happen if you have no wind at the surface and therefore your lowest point is a zero vector. Otherwise, you will see the line starting at some point away from the origin, depending on the strength and direction of the wind at that lowest elevation.

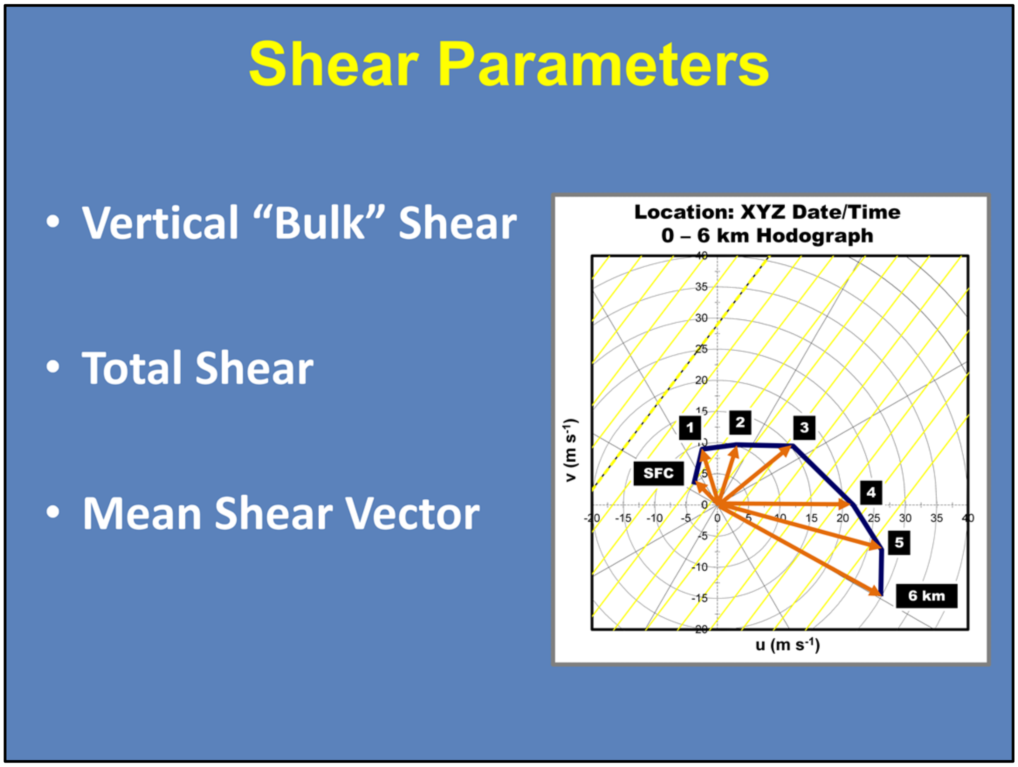
Shear Segments

- **Shear Segment = Magnitude (length) between two levels**
  - Take first segment down to (0, 0) then determine x-axis length (estimate)
  - Same all additional segments
  - Then, add them all end-to-end for the "total" segment length



Segment	Length
SFC → 1	6 m s^{-1}
1 → 2	5
Total (SFC → 2)	11

The final important component of plotting the hodograph is determining the magnitude – or the length – of each shear segment. How do we do that? In this coordinate system, we can certainly complicate these calculations very quickly. However, we can actually just pluck the segments and transfer them down to either the X or Y axis and then estimate the length on that axis. In this example, we'll use the X-axis, starting with the surface to 1 km segment, which is the red line. Plopping that onto the X-axis, we come up with 6 m/s for the first segment and then 5 m/s for the 1 km to 2 km layer. Now, to add them up…we just add them up, end to end, which is 11 m/s. This is the total shear amount for the surface to 2 km layer. So now, let's build on that.



Once the hodograph is drawn, the real work starts because we need to make sense out of this trace. What does this shape mean in terms of the environmental wind and this time? Does this profile help or hinder convection? If we do have convection, what will become of it, what mode will it be? We can begin to answer these and other questions through variations of this concept of shear – which again, is what the hodograph is plotting, the differences in wind speed and direction with height. So, we're going to explore these parameters, Vertical or Bulk Shear, Total Shear, and the Mean Shear Vector listed in the next few slides.

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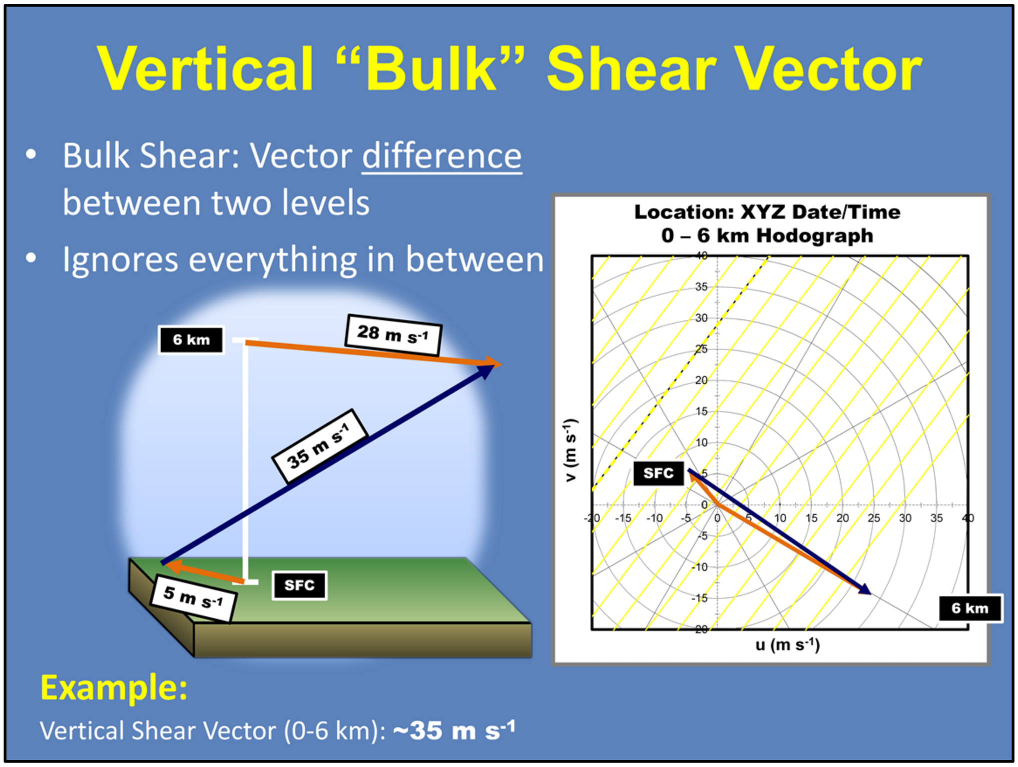

First, we can determine the Vertical or Bulk Shear Vector which is just the vector difference between two given horizontal levels. In our example, we subtract the vectors from the surface to 6 km. This can be done for any two levels, 0 to 1, 3 to 6, and on and on. This is vector subtraction though and it is not as easy and just subtracting 28 minus 5 to get 35 m/s. So, where did 35 m/s come from then? We will find out how to determine magnitudes in the coming slides.

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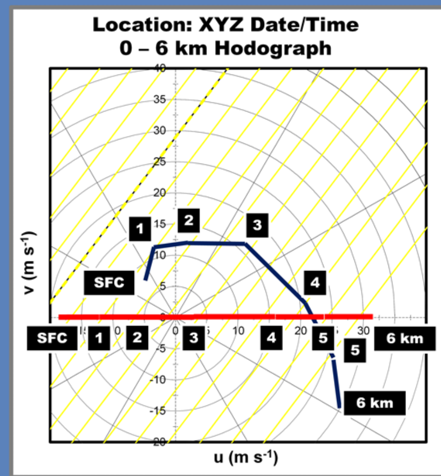
Total Shear

- **Total Shear**
  - "Net Length" of the hodograph
  - How is it derived?
  - Place the segments on the X-axis & add them up (the magnitudes)



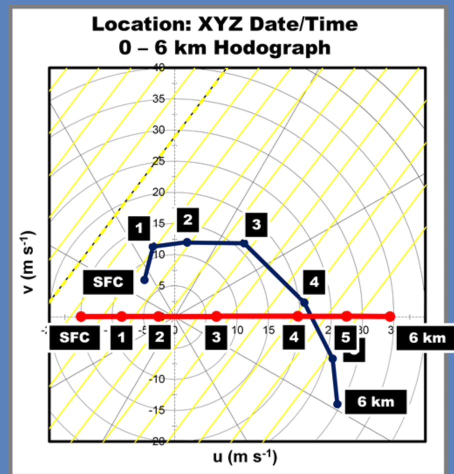
Remember how we came up with shear segment lengths? Now we're going to build on that and figure out the Total Shear for this entire plot. If you have a hodograph with sounding data that goes all the way to the troposphere, you will probably have data well past 6 km, maybe even between 10 and 15 km. Generally speaking though, most of the convective processes in terms of shear are found below 6 km, so that's where we will focus on. But if you do truncate levels higher than 6 km, then you have to denote this as the Total Shear from 0 to 6 km, for instance.

Since we've turned these vectors into segments, which is just magnitude, independent of direction, we can add them up end-to-end on the X-axis and get the total shear for that layer. On the next slide, we'll put them together.

Total Shear

- **Total Shear**
  - Determine the length of all segments
  - Put them end-to-end on the x (or y) axis

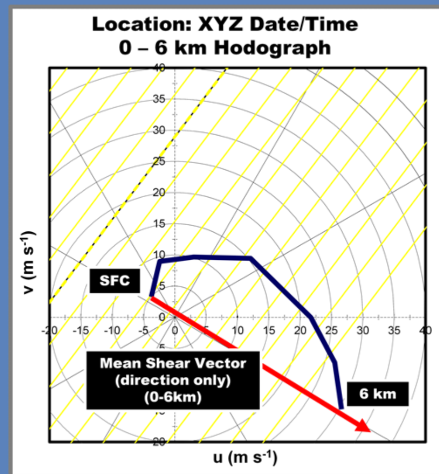
Segment	Length
SFC → 1	6 m s^{-1}
1 → 2	5
2 → 3	10
3 → 4	14
4 → 5	8
5 → 6	7
Total (SFC → 6)	50 m s^{-1}



Going through the same exercise as we did a few slides ago to determine the shear segment lengths, I've plotted the remainder of the magnitudes – or lengths – of each of the segments from 0 to 6 km from the same example. You can see the lengths there in the table. Then, we plop them all onto the x-axis, end-to-end, and normally, I would have started the end-to-end process at (0, 0) but as you can see with the total length of 50 m/s, the right side wouldn't have fit, but as you can see, on the X-axis it starts at -15 and goes to +35, which the absolute value is 50, so it works either way. So then, this is our Total Shear from 0 to 6 km for this example, a whopping 50 m/s.

Mean Shear Vector (Direction)

- **Mean Shear Vector can help determine supercell motion**
  - Total Shear is easier to deduce (just add up the magnitudes through a specified layer depth)
  - Mean Shear is a bit more labor intensive
  - (easy) First, determine ***JUST*** the Mean Shear vector **DIRECTION** by drawing a vector from beginning point to the end point (of the defined layer)



Example:

[red vector] Points in the direction ESE but is "West-northwesterly" (~300°)

Building on total shear, we delve into this idea of the Mean Shear Vector, which can help determine supercell motion – and again, this 0 – 6 km layer is a typical one to use, so we'll continue with this example. But the mean shear vector, hand-calculated, is not nearly as easy as finding the total shear – which we just took the shear segments end-to-end and added them up. Luckily, the direction of the Mean Shear Vector is pretty easy – just draw a line from the point where the SFC segment starts and connect it to the end of the 6 km segment and that direction this is your Mean Shear Vector for that layer. In this case, it's pointing toward around 300 degrees, which is West-Northwesterly, which means it is coming out of the West-Northwest but it is pointing in the direction of East-Southeast. Now, we have the task of determining its magnitude.

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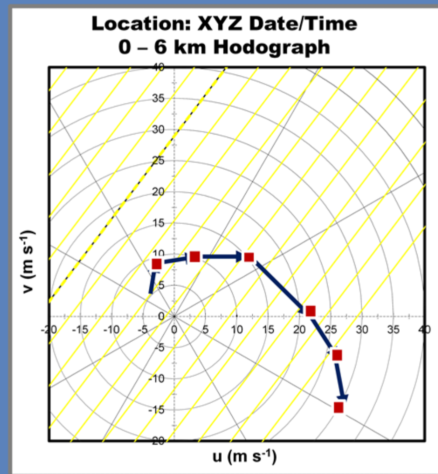
Mean Shear Vector (Magnitude): Part 1

- Convert segment points into the (x, y) coordinate system
- Treat each segment individually
- Act as if each segment's **starting** point is (0, 0) then find the (x, y) of the **end** point

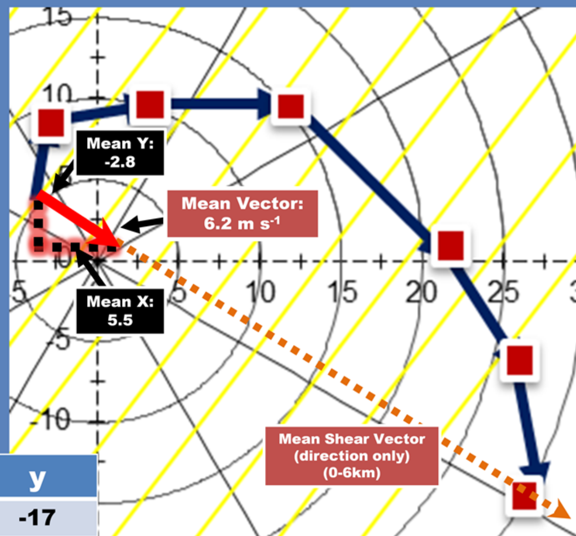


Segment	x	y
SFC → 1	+1	+6
1 → 2	+6	+1
2 → 3	+10	0
3 → 4	+10	-10
4 → 5	+5	-8
5 → 6	+1	-6
Total	+33 m s^{-1}	-17 m s^{-1}

Now, on to a bit more labor-intensive but critical step of finding Mean Shear magnitude. For each segment, we need to get an X and Y coordinate for the END point of that segment. How do you do that? By treating the START point as (0, 0) for each segment. Then, find the x-axis extent of that segment… for the first segment from 0 to 1 km, it only goes forward 1 in the positive x-direction, but it goes 6 in the positive y-direction. The next segment is 6 in the positive x-direction, then only up 1 in the positive y-direction… and so on until we reach 6 km. You see most of the y-directions being negative for these segments because we're going the opposite way on the chart from positive y. Add them all up for X and Y and you have 33 m/s total shear in the x-direction and -17 in the y-direction.

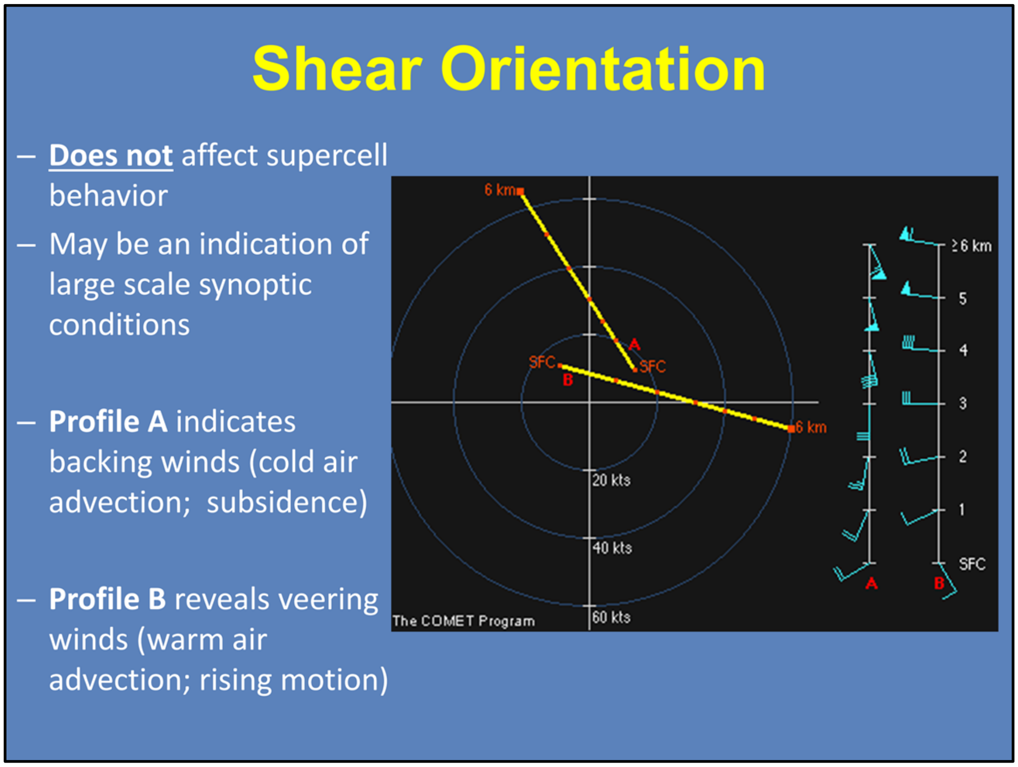
Mean Shear Vector (Magnitude): Part 2

- Divide the X and Y sides by 6 km
- Result is the **Mean Shear** (x, y)
- Plot this (x, y) back on the hodograph, starting at the SFC (0 km) point
- **Mean Shear: (5.5, -2.8)**
- **Vector: 6.2 m s^{-1} (ESE)**



Segment	x	y
Total (0 → 6 km)	+33	-17
(divide each by 6 km)		
Mean / Avg.	+5.5 m s^{-1}	-2.8 m s^{-1}

Almost there, but first we have to divide both the total length from X and Y and by 6 km and get the mean of each one. Then, we plot those back on the graph, starting with first hodograph point, down 2.8 on the y-axis and then over 5.5 on the x-axis, which equates to a 6.2 m/s magnitude line – the red line – that points to the ESE, which was the first thing we determined back a couple of slides ago.

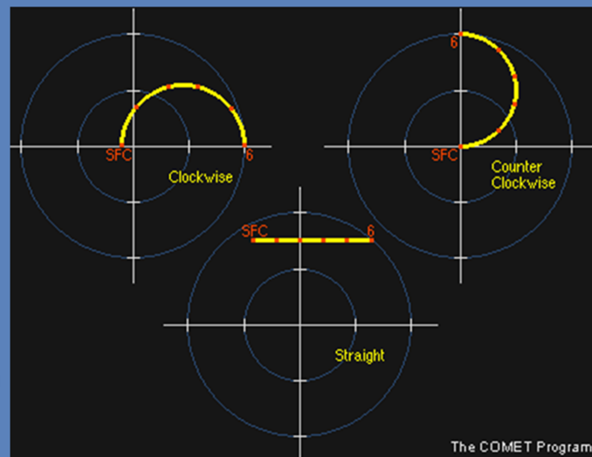


Let’s go over some functions of hodograph shapes and different types of plots, which can tell us a lot about the environment. First, the Shear Orientation – what does mean? We’ll take these two different wind profiles and just plot the direction or mean orientation of each profile. And so the orientation is more of a large-scale or synoptic behavior model as opposed to showing the behavior of individual storms or supercells. Profile A, which shows a backing orientation which is an indication of cold air advection – and therefore subsidence. Meanwhile, Profile B shows a veering winds, showing warm air advection in that layer and possible rising motion.

Relative Curvature

- **Does** impact storm behavior

- Concerned with:
  - Relatively curvature
  - When it does curve
  - The level through which it curves
  - Whether it curves clockwise or counterclockwise with height



- These variations all have implications for storm structure.

- Match up (from previous slide) the **Counter-Clockwise hodograph with Profile A** and **Clockwise with Profile B**

What does make a difference for individual storm behavior, especially supercellular ones, is the curvature of the shear on a hodograph. We are concerned about various aspects of curvature, including… the relative amount of curvature, i.e. how strong or intense, when it occurs – morning, noon, or night – where it occurs in the profile, and whether it is a counterclockwise or clockwise curvature. All of these aspects can be crucial in determining the type of environment you have for potential convection. In terms of what they mean, look at the wind profiles from the previous slide and match up the effects of the Counter-Clockwise hodograph with Profile A and Clockwise with Profile B.

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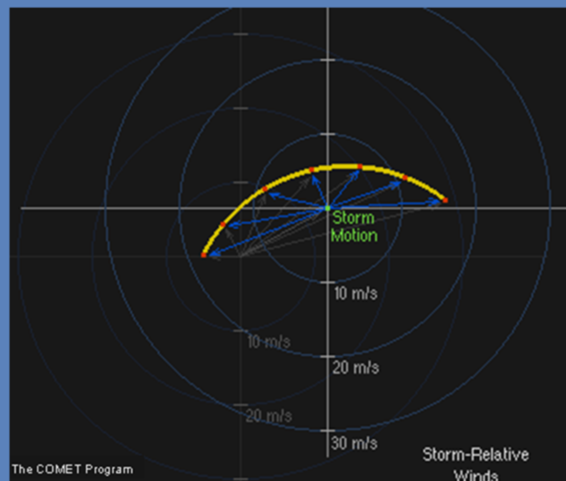
Storm-Relative Flow

- An **observer** sees the winds in this hodograph marked by the **red vectors**
- The **storm** sees the winds in this hodograph marked by the **blue vectors**



After going through the Radar Products lessons, you should be familiar with the concept of "Storm-Relative", which has to do with the relative point of view difference between and observer at a fixed point and the storm itself. Similar to the SRM or Storm-Relative Map product, on the hodograph, the environmental winds are plotted from a fixed point, but sometimes we want to place that point on the storm itself and see how the environment affects the storm. This brings us to Storm-Relative Flow, starting off with the ground-relative location, which is the location of the observer and how the winds change from their perspective. Then, we transport the X and Y axis over to this Storm Motion point, to see how this environment is affecting the storm itself and thus, the storm-relative flow. But what is Storm Motion?

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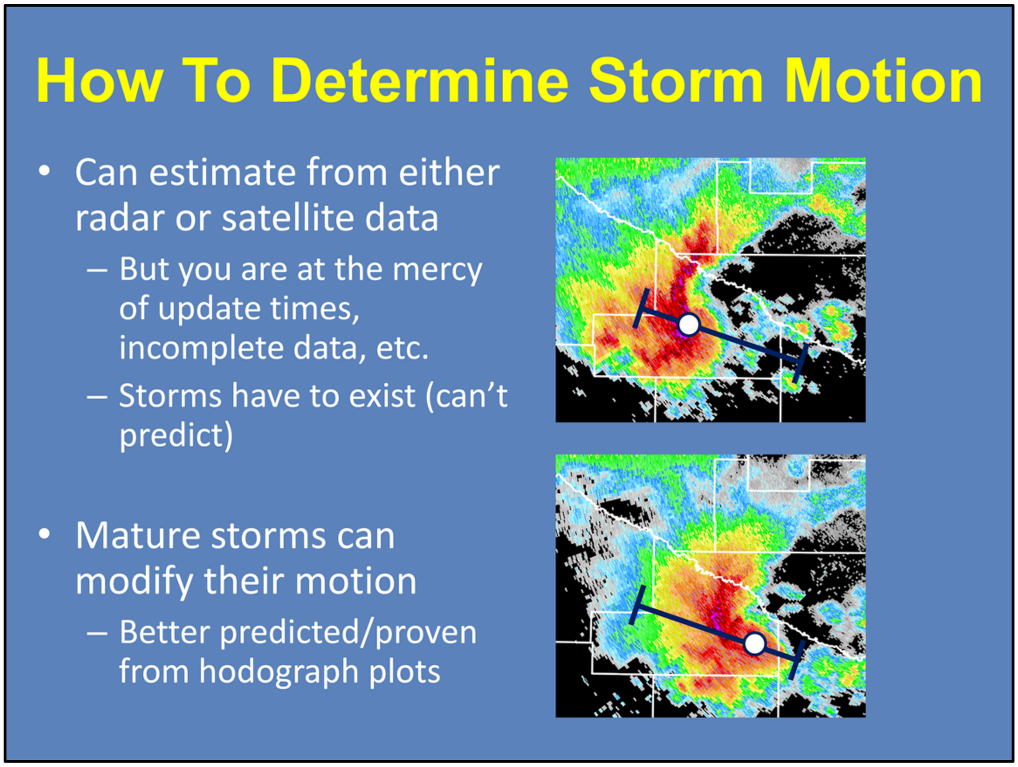

Next, we want to determine the storm motion, which will play into the relative effects that an environment can have on individual convective cells in a specific atmosphere. One can estimate storm motion, the direction and forward speed, based on either radar or satellite data. But, you have to have a storm to track, this won't do much good if you're trying to predict storm motion hours or even minutes in advance if there's nothing there. You are also at the mercy of how fast or slow the data updates and potential data quality issues. So, how do we do this on a hodograph?

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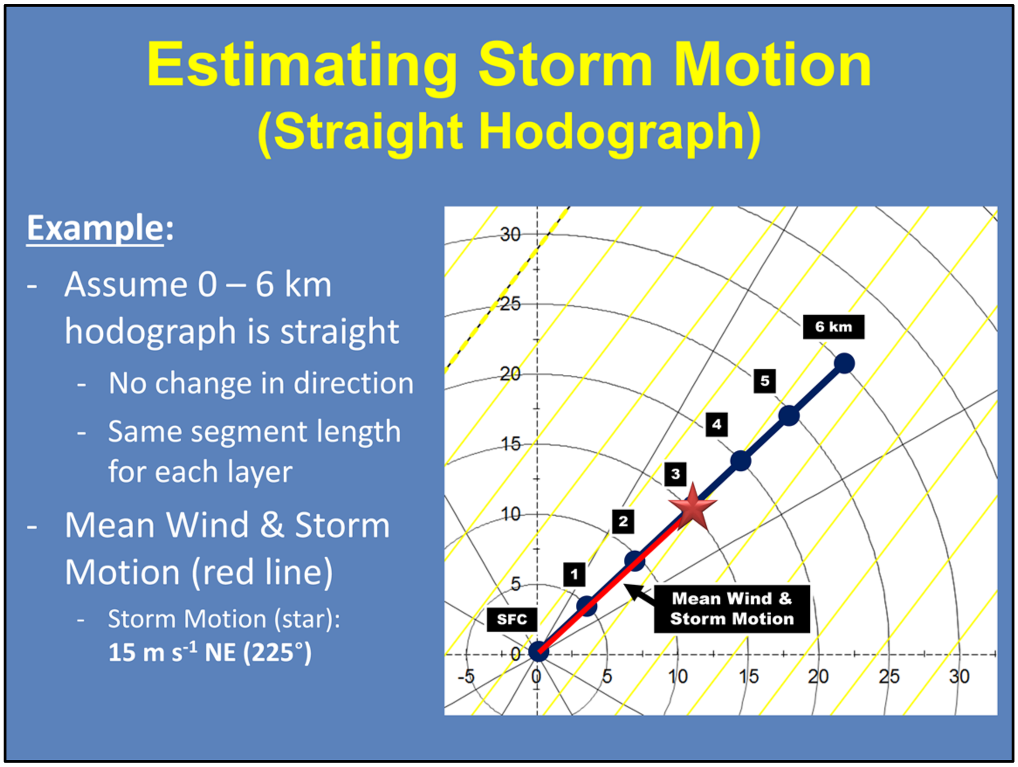

How do we find Storm Motion on a hodograph? Let's start out with a very basic example with a straight hodograph, meaning, the winds throughout the column, 0 to 6 km in this case, are in the exact same direction, top to bottom. In this case, we have winds out of the SW that are also the same magnitude throughout the column, making it easy to average out both the mean wind speed and direction for this profile, which is 15 m/s and the vector would point toward the NE, or at 225 degrees, which is a Southwesterly wind. But in this case, we're not talking about vectors, we want a point which will be our new reference point, acting as the storm itself. So then, this red star is going to be our Storm Motion and you see where it is placed on the chart.

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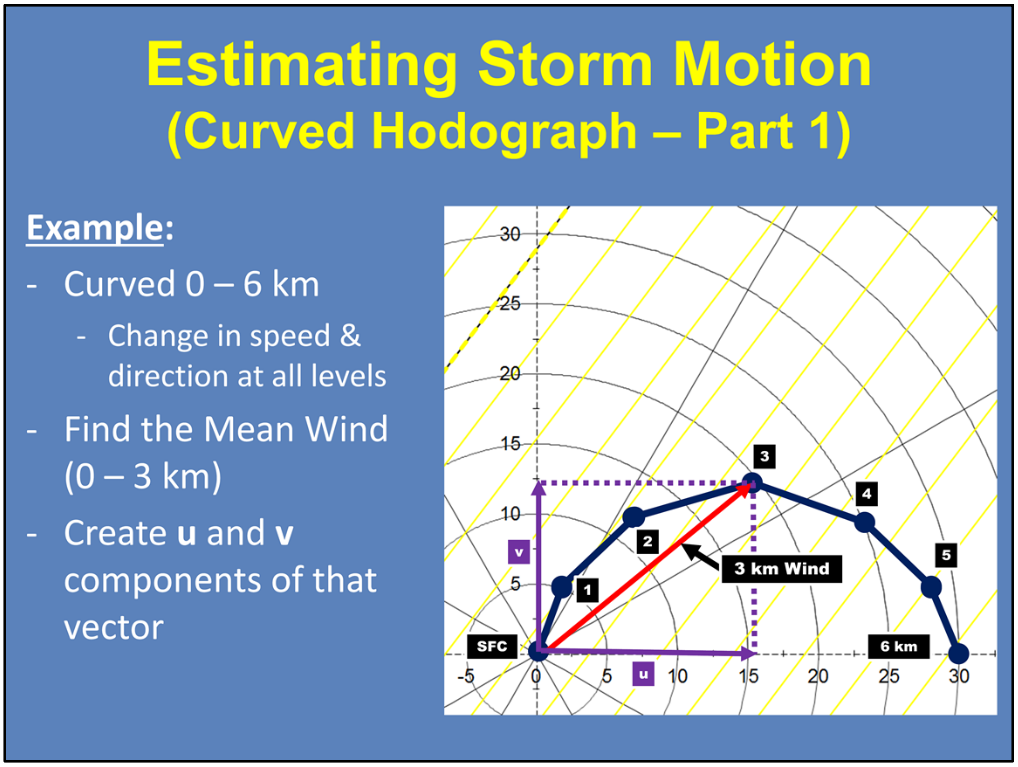

Unfortunately for manual calculations, reality is usually much more complicated, so we'll add a little curve to the profile. You can see how the winds change in speed and direction as we go from the surface to 6 km. Recall a few slides back we determined the mean wind by making a vector go from the surface to the elevation of interest. In this case, we can just make a vector go from 0 to 3 km, with this red arrow. Then, we can break down the vector into its u and v components and plot them as well. To calculate the mean wind between 0 and 3 km, we'll average out u and v, like we did a few slides ago, then add these means together to get the total mean wind vector.

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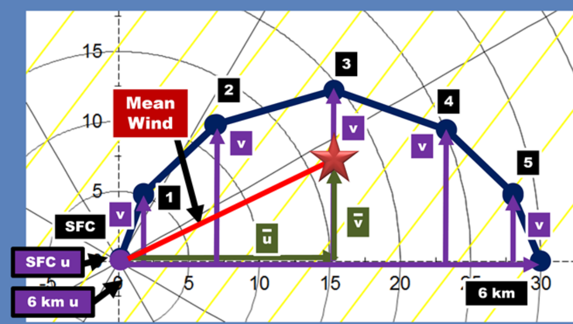
Estimating Storm Motion
(Curved Hodograph – Part 2)

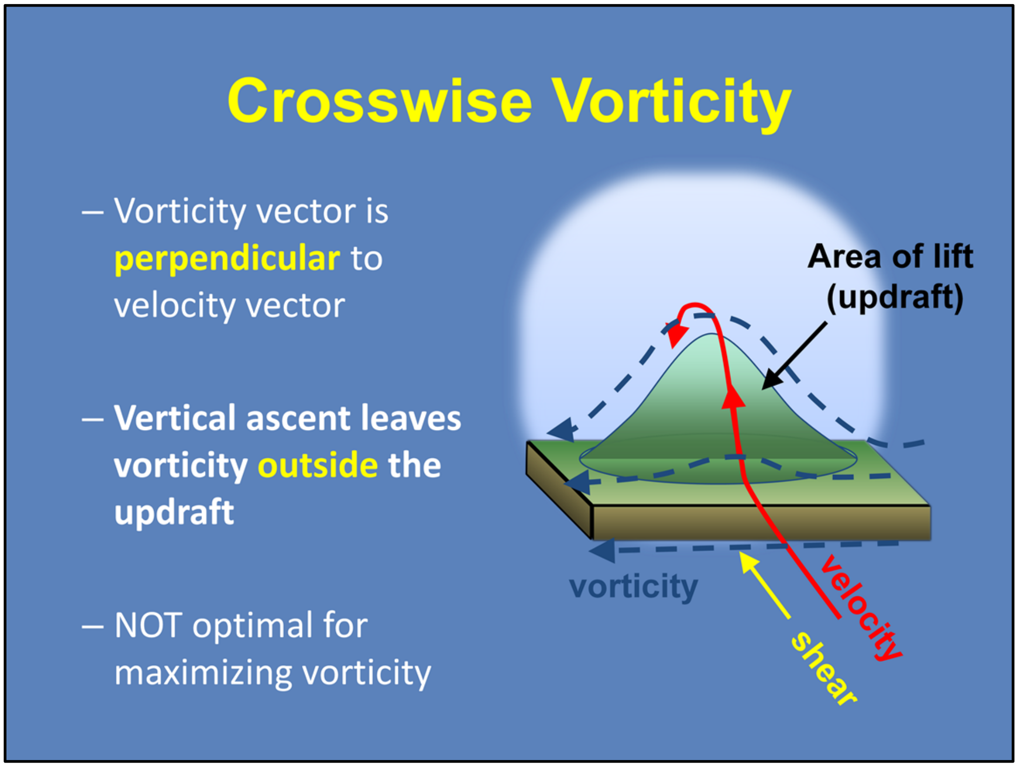
Example cont'd:

- First, estimate mean u for 0 & 6 km winds
  - (if you assume segments are relatively equal)
  - u (0 km) = 0 m s^{-1}
  - u (6 km) = 30 m s^{-1}
  - $\bar{u}$ (0 – 6 km) = 15 m s^{-1}
- Next, find the mean of all the v components
  - $\bar{v}$ (0 – 6 km) = 7 m s^{-1}
- Add u + v components (red line)
- Then, plot the Storm Motion (red star)



$\bar{u}$ = mean u
$\bar{v}$ = mean v

Continuing with this example, first we'll take the surface and 6 km wind and average them, to find the mean u. This is easy if all the segments are roughly equal, like in this example. It would be a bit more labor intensive if you had very different segment lengths. Then, do the same with the v components, which is about 7 m/s. Once the mean u and v are found, you connect them with the vector that is the mean wind, the red line. Then, the point with the star is the Storm Motion. Now…from here it gets more complicated with variations that can occur, even with this example, where this whole profile that, for instance, can be re-oriented onto a different axis – say, we re-orient this entire profile and all the vectors 45 degrees to the left – which would look a bit more realistic. Also, you can have multiple curves and make this profile look more like an S-shape which would be more difficult to compute but again, is not uncommon in the real world.



Now, we're going to add in this idea of vorticity, which – for hodographs – is broken down into two main types, Crosswise and Streamwise. Recall that vorticity is tendency for something to spin and a vorticity vector basically gives us the information about how that object will spin, so then it's up to the environment to effectively or ineffectively utilize the vorticity that is present. Crosswise, which is the example on the right, is the most absolutely inefficient method to utilize vorticity. In other words, if you need a certain amount of vorticity to go into an updraft and cause it to spin – you won't get it with crosswise vorticity because the shear and the mean flow are oriented perpendicular to the vorticity and passing by each other with no real interaction. If you have an updraft, the vertical ascent is not able to utilize the vorticity because of this skewed orientation.

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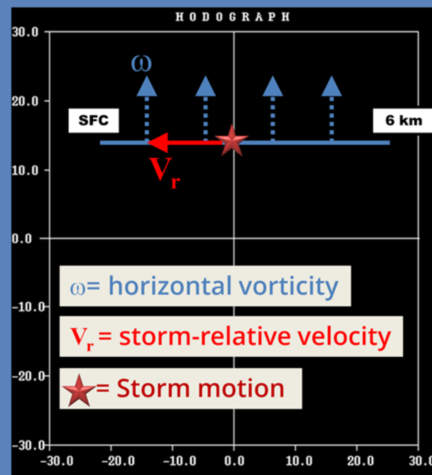
Crosswise Vorticity Plotted

- Example of a storm experiencing **crosswise** vorticity
  - Updraft not correlated (same direction) with vertical velocity



Here is an idealized straight hodograph, from 0 to 6 km, with the Storm Motion plotted with the red star. The storm-relative velocity on this chart, the red V-sub R vector, represents the mean wind. You can see that this mean wind vector and the storm motion are essentially on top of the hodograph plot, which – also notice – is straight. The horizontal vorticity vectors are perpendicular to the hodograph which means that they are basically not affecting each other. They would need to be in the same direction, in order to maximize the potential of vorticity in rising motion. If you're trying to make sense of this in a conceptual format, head back to the previous slide and check out the diagram, to see how the vorticity lines just continue on their path, unaffected by the ambient wind, which is the point of crosswise vorticity.

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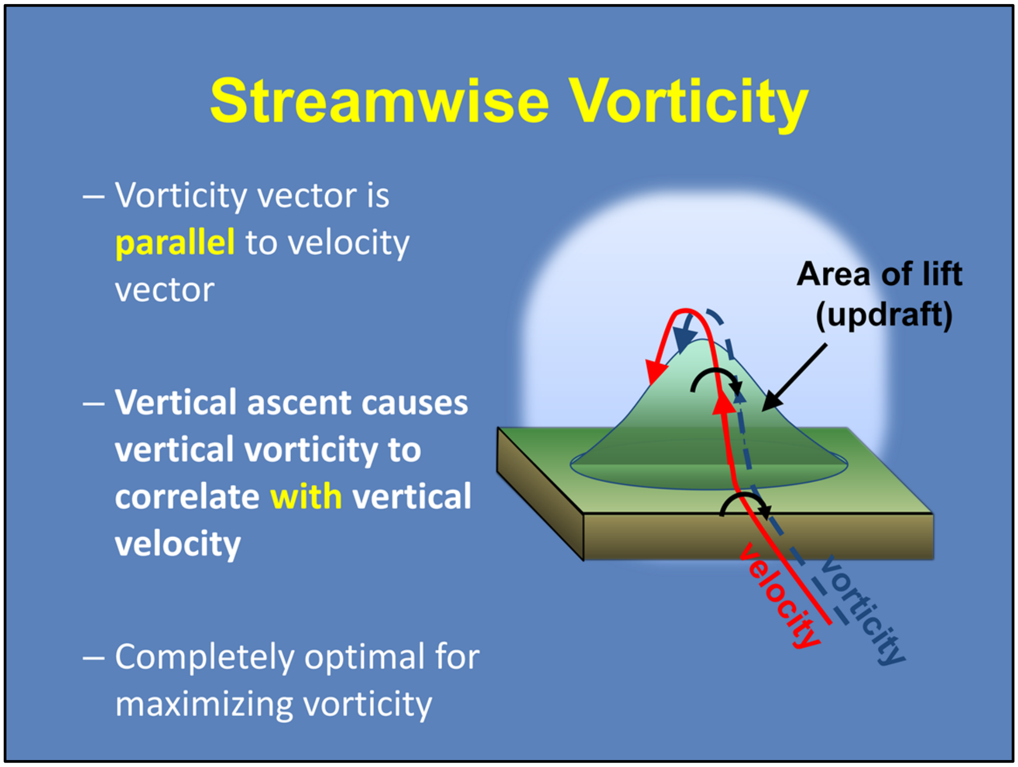

In direct contrast, there is streamwise vorticity, which means that the vorticity vectors are completely parallel to the velocity vector. Again, think of the velocity as the ambient wind and if you're trying to achieve a rotating updraft, you will efficiently utilize the vorticity into the updraft when both vectors are parallel, working in sync – in other words. Vorticity is your "spin" and when the velocity can efficiently tilt that spin into upward-moving air, then your potential for supercellular storms increases.

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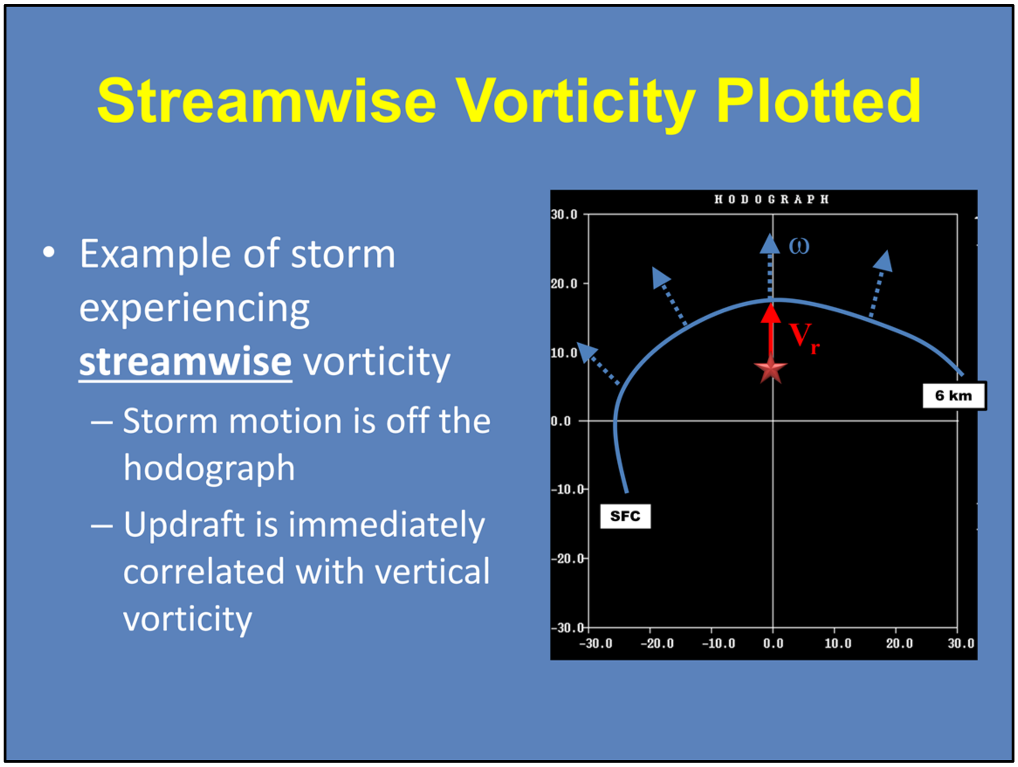

This is an example of what streamwise vorticity looks like on a hodograph. You have probably heard that for supercells and especially for possible tornadic potential, you would look for a “curved” hodograph, even if you aren’t completely sure why that is. In this example, the storm motion is off of the hodograph, which allows the storm to utilize vorticity at all angles throughout the layer of interest. When you determine the mean wind, which again, is basically the red arrow, you see how it is in the same direction as the blue horizontal velocity vector. Again, if you need a more conceptual reference for this, go back to the previous slide and check out the diagram to see how both of these elements need to be parallel in order to maximize the available vorticity.

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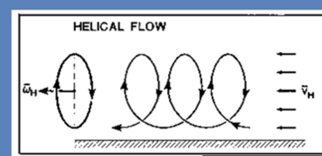
Helicity

- **Helicity:** The property of a moving fluid to evolve into a helical flow
  - Corresponds to the transfer of vorticity from the environment to an air parcel (through convection)
- Proportional to the strength of the flow, the amount of vertical wind shear, and the amount of turning in the flow (i.e. vorticity)
- **Helical flow**: flow in the pattern of a corkscrew

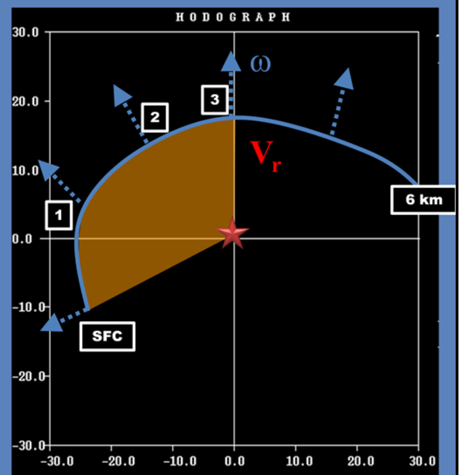


Doswell 1991

Coming down the home stretch, we need to add in this last concept of helicity which builds into some of the most useful parameters that can be derived from hodographs. Helicity, in meteorology, refers to the property of a moving fluid to evolve into a helical flow and helical flow is basically in the pattern of a corkscrew. The transfer of vorticity from the environment to an air parcel occurs through the convective process and the amount of available helicity is a reflection how strong this convective process is.

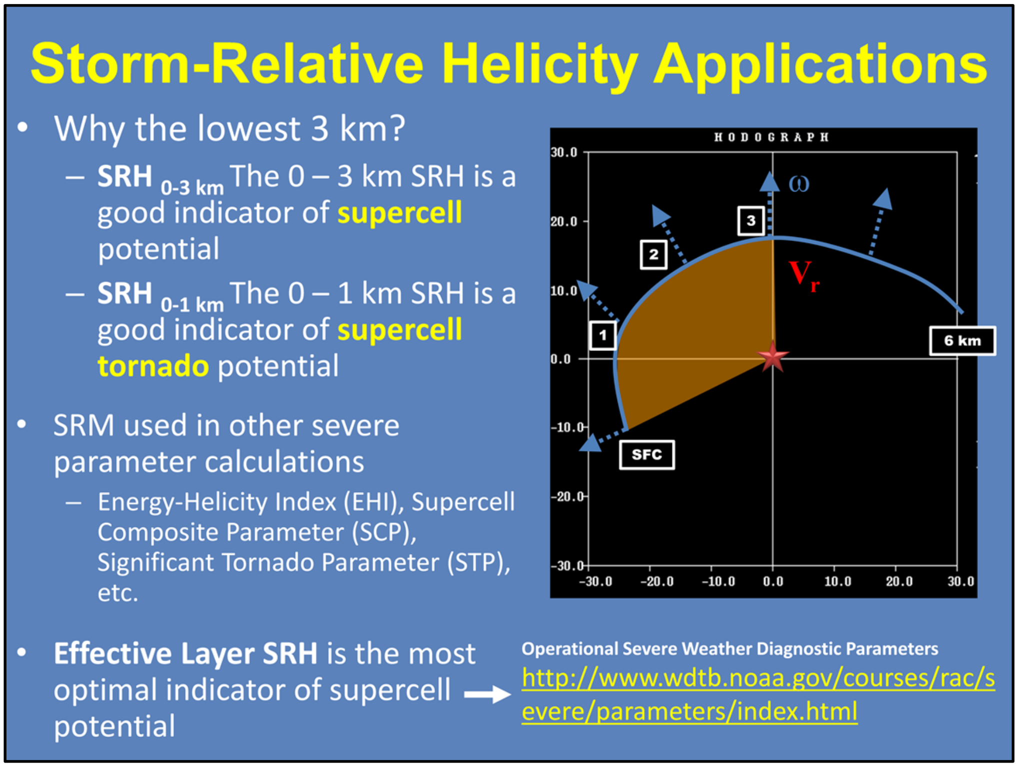
Storm-Relative Helicity (SRH)

- **SRH** Definition: The potential for cyclonic updraft rotation in right-moving supercells
  - Integrated over a vertical layer
    - 0-1 km, 0-3 km
  - SRH is twice the area swept out between the hodograph and the Storm Motion between two levels *(0 – 3 km in the example)*



So why did we cover all these terms with helicity and vorticity? Why do they matter? Now, we can wrap together a number of these concepts into this term, Storm-Relative Helicity or SRH. Again, like the concept of Storm-Relative Velocity, as you learned about in the products lessons, we want to know how much relative streamwise vorticity is or could be ingested into a storm's updraft. Just like on a hodograph, when we experiment with the lifting parcel levels to get different values of CAPE for different situations, we use these concepts and calculations on a hodograph to find out a storm's potential for developing a mesocyclone (i.e. rotating updraft) which is the defining characteristic of a supercell thunderstorm.

SRH is integrated over a defined layer and is defined as basically the area under the curve between the hodograph and the Storm Motion, or the red star. We'll find out on the next slide why we use the 0 to 1 and the 0 to 3 km layers for this calculation.



Researchers using this parameter have found that the 0 to 3 km layer SRH is a good indicator of supercell storm potential, while the 0 to 1 km layer which is a more compact area near the surface, which is a good indicator of supercell tornado potential. The 0 to 3 km layer is more of a mean approximation of the storm's inflow and how the base portion of the storm is affected by the environmental shear, while the 0 to 1 km layer really takes into account the smaller-scale effects that normally determine whether a supercell tornado will form or not. And the story doesn't end there with SRM. It is used a number of other severe parameter calculations, each of them using this concept of storm-scale helicity usage in different ways. And then, Effective Layer shear is likely the most optimal of all parameters for indicating supercell potential. More on this can be found in the interactive module "Operational Severe Weather Diagnostic Parameters," just follow the link on the bottom right.

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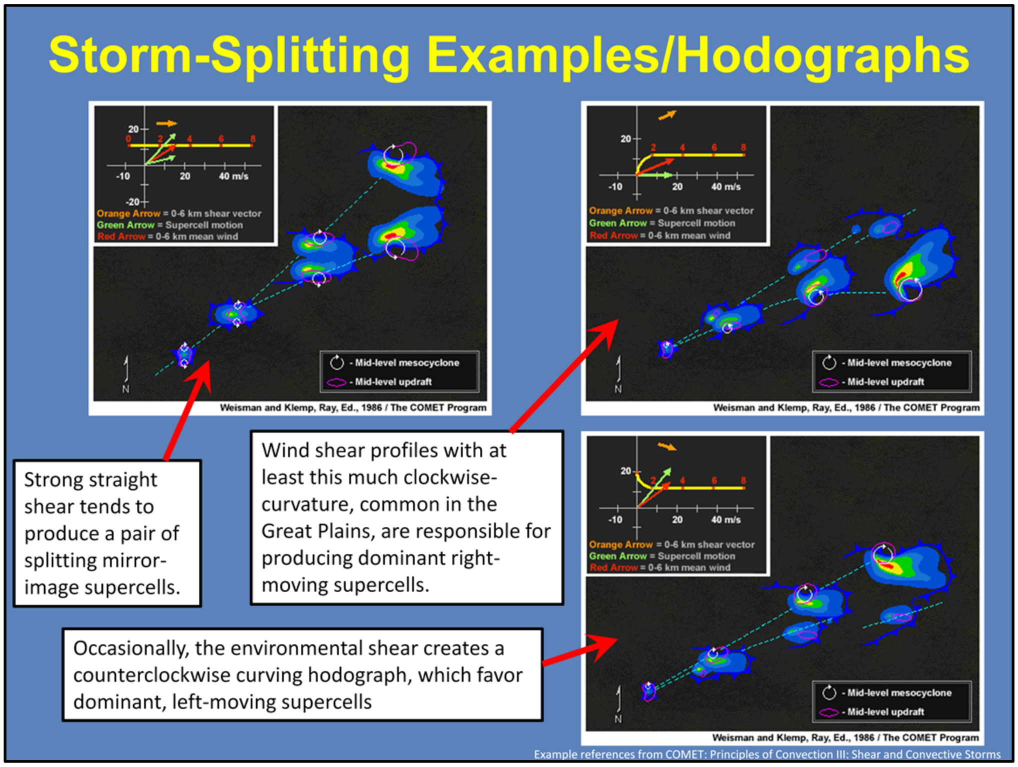

Lastly, one more application that groups together multiple concepts we learned about, here are 3 different idealized examples of a specific type of hodograph profile and what affect it has on storm-splitting. Look at the hodograph at the top-left of each image and see how each type of profile affects the production and maintenance of supercells. The terms left and right-moving refer to a storm's movement either to the left or right of a theoretical line in between two cells. These concepts will be covered in greater detail in some of your subsequent lessons in this Convective Storms Topic. But, at least you are getting a start on the basics and theory behind some of the more complex features and cases.

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Takeaways (part 1)

- **Hodograph is a plot of the vertical wind structure**
- **Shear**
  - Wind difference between two layers
  - Important for determining characteristics of convective storms
- **Certain specified layers of the atmosphere…**
  - are important for specific severe parameters
- **Bulk, Total, and Mean Shear**
  - Foundational for other severe parameters from the hodograph plot

Now, let's summarize what we covered. This lesson showed how hodographs are plots of the vertical atmospheric wind structure, using the wind information from a sounding. The hodograph is all about shear – which is the change or difference between two layers, any two layers – both speed and direction. From this information, we can determine what the possible characteristics, or you could even think of it as the side-effects of a storm being in a certain type of environment. We saw how specific layers of the atmosphere are interrogated for certain types of features, or effects, that could have implications on convective cells. And in order to know these things, we look at several different base parameters derived from the hodograph, which are the foundations for more specific calculations.

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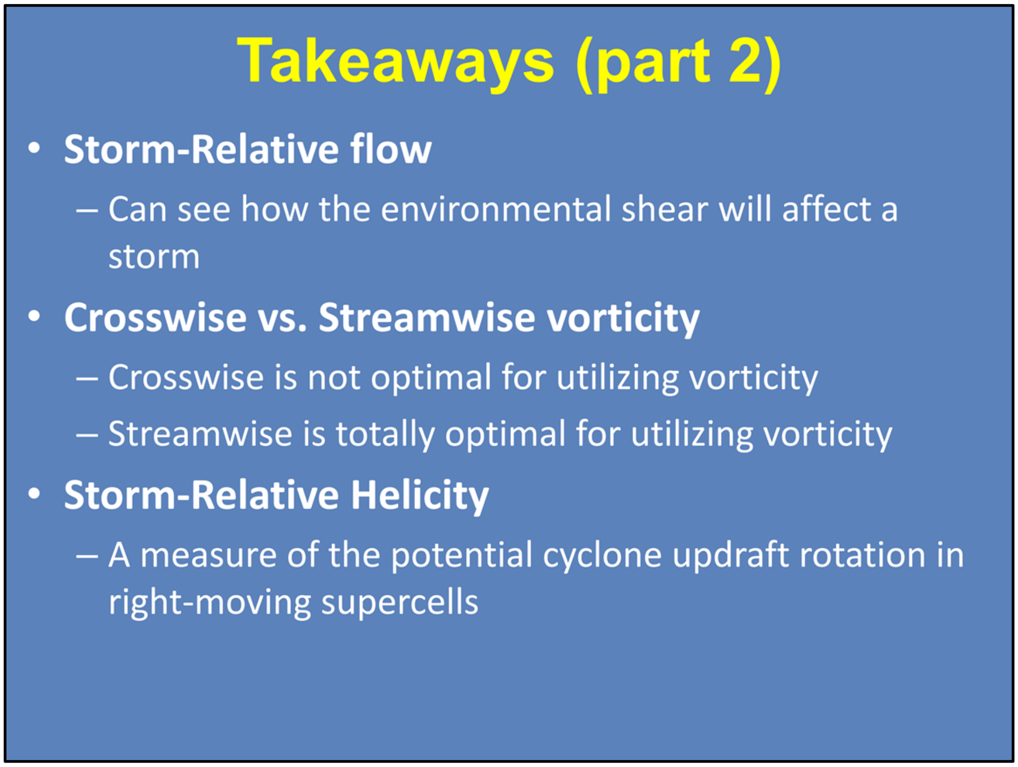

Then, we looked at this idea of storm-relative flow and how we need to see things from an individual storm's perspective. And then, getting into the heart of a storm's characteristics is how the two types of vorticity, crosswise and streamwise, will affect a thunderstorm in-particular, the updraft. And finally, the calculation for how vorticity and helicity measure-up for a thunderstorm with the Storm-Relative Helicity parameter.

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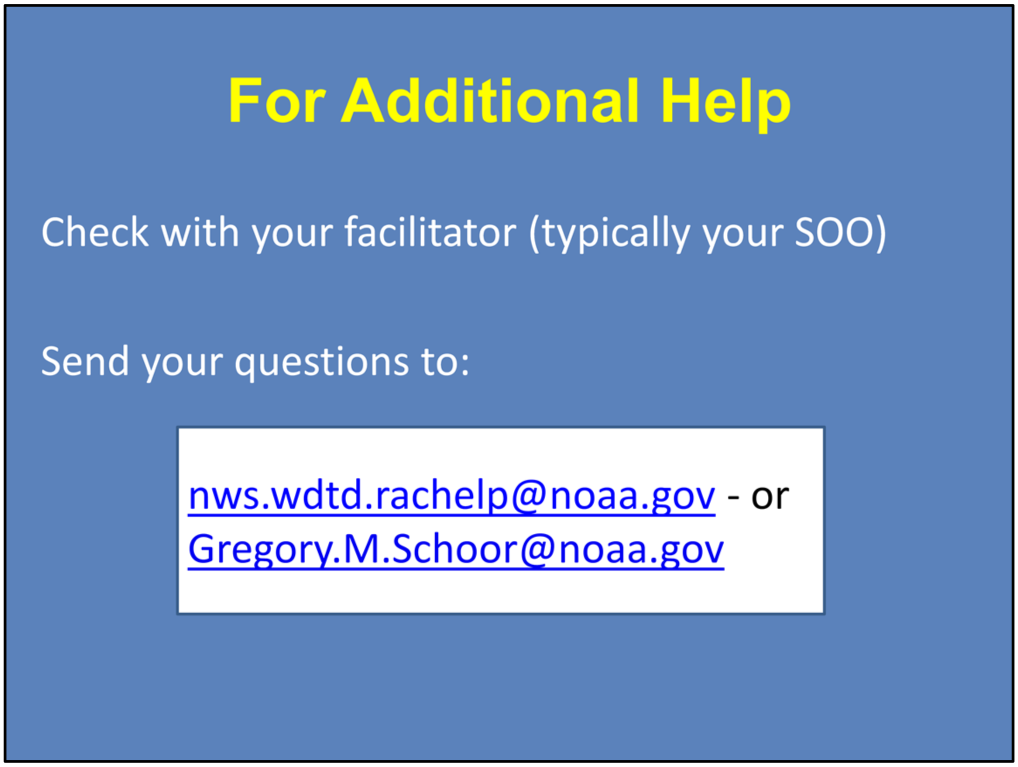

For additional help, check with your facilitator (typically your SOO) or send your questions to the either of the e-mail address listed – and thanks for your time!

Welcome to this RAC Convective Storms lesson on the Fundamental relationships between shear and buoyancy on convective storm structure and type. I'm Justin Gibbs of the Warning Decision Training Division.

Learning Objectives

1. Identify the types and characteristics of shear most commonly used in operations
2. Identify the primary operational uses for different measurements of shear
3. Identify applications and limitations of CAPE/parcel theory to thunderstorm forecasting
4. Identify the effects of the depth/direction of shear on storm organization and evolution

Here are the learning objectives for this lesson.

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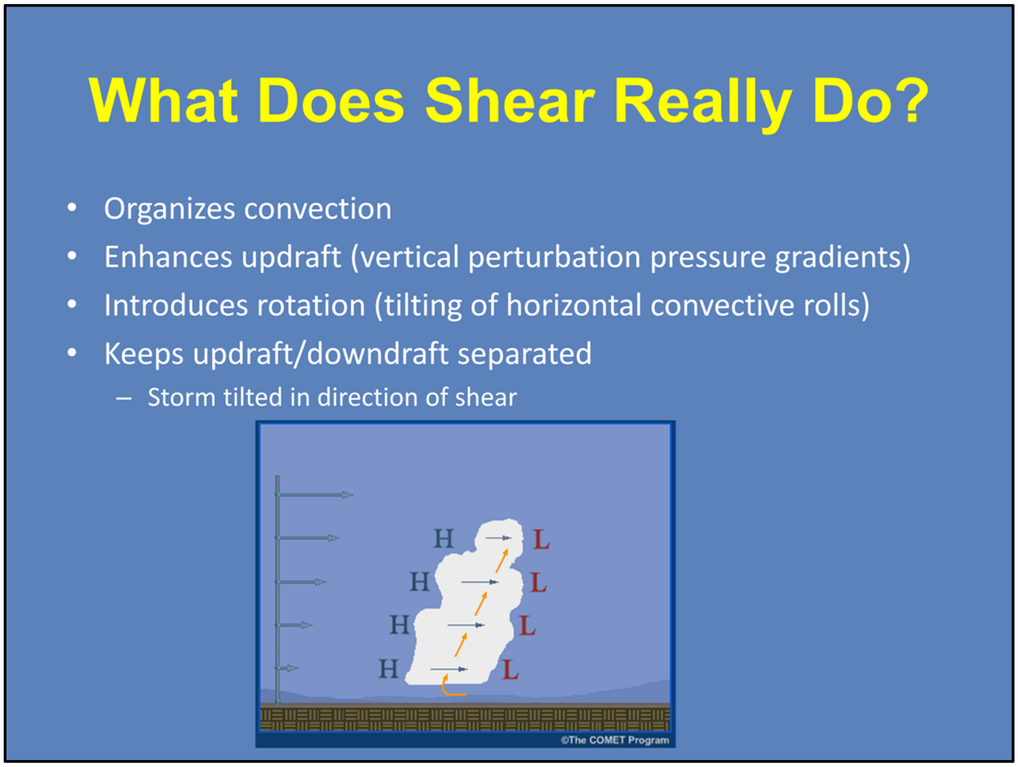


So we have been over this before, but its easy to take it for granted and both over simply and over complicate what shear really does

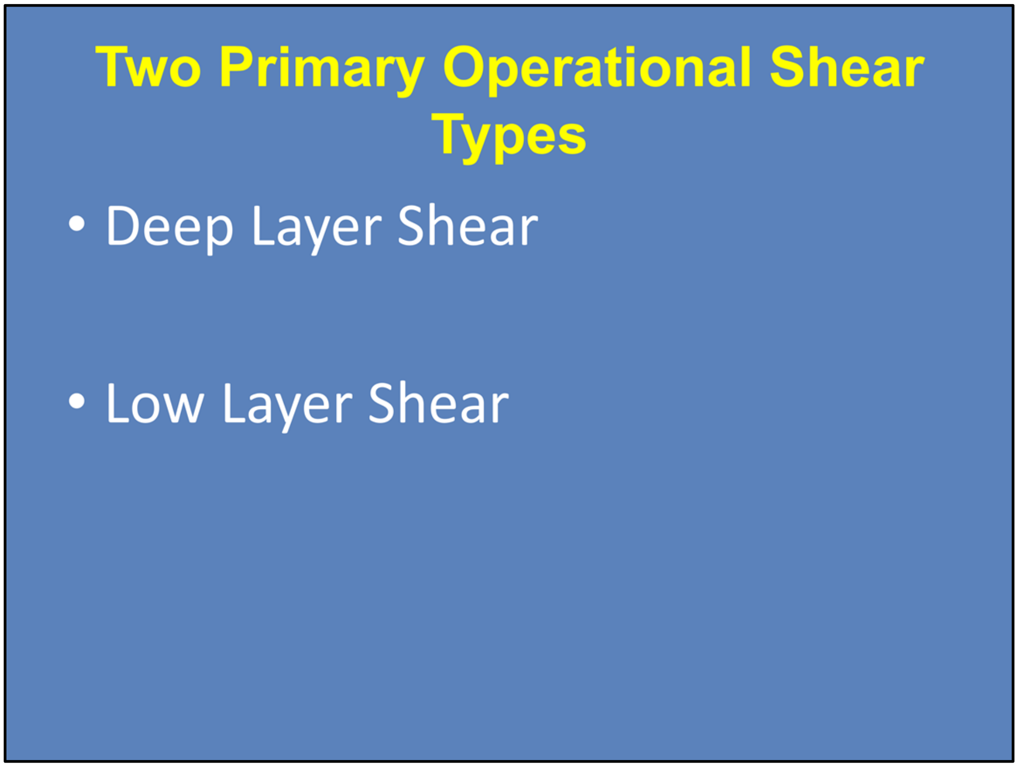
For all intents and purposes it organizes convection

It serves to enhance the updraft through vertical perturbation pressure gradients, decreasing pressure above the parcels that serve to enhance lift.

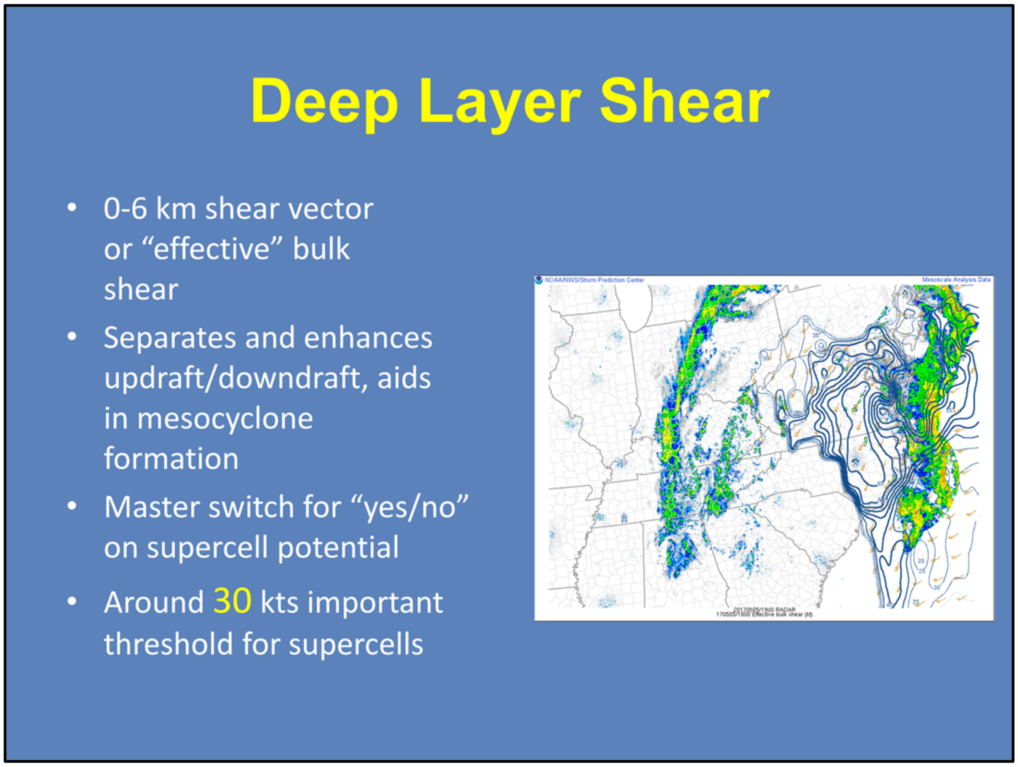
It introduces rotation through the tilting of horizontal convective rolls

It also keeps the updraft and downdraft separated by tilting the storm in the direction of the shear. Without shear the updraft would collapse on itself.

Bluestein, 2013



In severe weather operations there are two primary types of shear, deep layer shear, and low layer shear.



Deep layer shear, is usually measured by the 0-6km bulk shear vector, or as effective bulk shear.

Deep layer shear, separates the updraft and downdraft, keeping the downdraft from overlapping the updraft and killing off storm inflow. It also enhances the updraft through the development of vertical pressure gradients and is a primary driver in mesoscylone formation.

Basically, from a severe weather standpoint deep layer shear is the on/off switch for supercell potential.

30 to 35 kts of deep layer shear is a good approximate threshold for that switch, below it, supercells will be much less likely. Above that value and you might have the shear necessary for increased storm organization and supercell development.

Thompson, et al. 2007
Doswell and Evans 2003
Thompson, et al. 2007
SPC Mesoanalysis

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Low Layer Shear

- 0-1km bulk shear
- Effective/0-1 km Storm Relative Helicity
- Respectable tornadic/non-tornadic supercell discriminator
- 150 m^2/s^2 a practical threshold for increased tornado concern (except for local boundaries!)
- Large values perpendicular to storms organized in groups or arcs motion can increase damaging wind potential

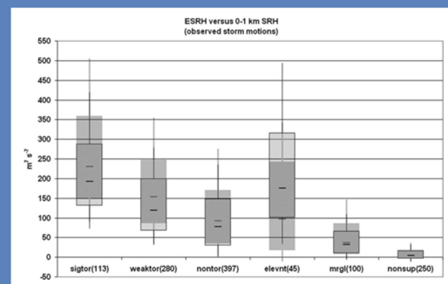

FIG. 8. Box and whiskers overlay plots of effective SRH ($m^2\ s^{-2}$) based on the CAPE $\geq$ 100 J kg^{-1} and CIN ≥ -250 J kg^{-1} parcel constraints (solid gray box and thick gray whiskers), and 0–1-km SRH (dotted black box and thin black whiskers). Plotting conventions are the same as in Fig. 7.

Low layer shear in terms of severe weather operations is usually described as 0-1km bulk shear, or 0-1 km slash effective storm relative helicity.

It is a primary discriminator of tornadic versus non tornadic supercells. Its not perfect but if you have strong deep layer, and low layer shear in a supercell environment you better start paying attention.

150 meters squared per second squared is a good first line threshold, but keep in mind strong local boundaries, like a stalled gust front or differential heating boundary, can locally increase available helicity to a supercell even if helicity is fairly weak on the mesoscale.

When you look at the 0-1km shear vector large values perpendicular to storms organized in large groups or arcs will increase damaging wind potential.

Thompson et al 2007

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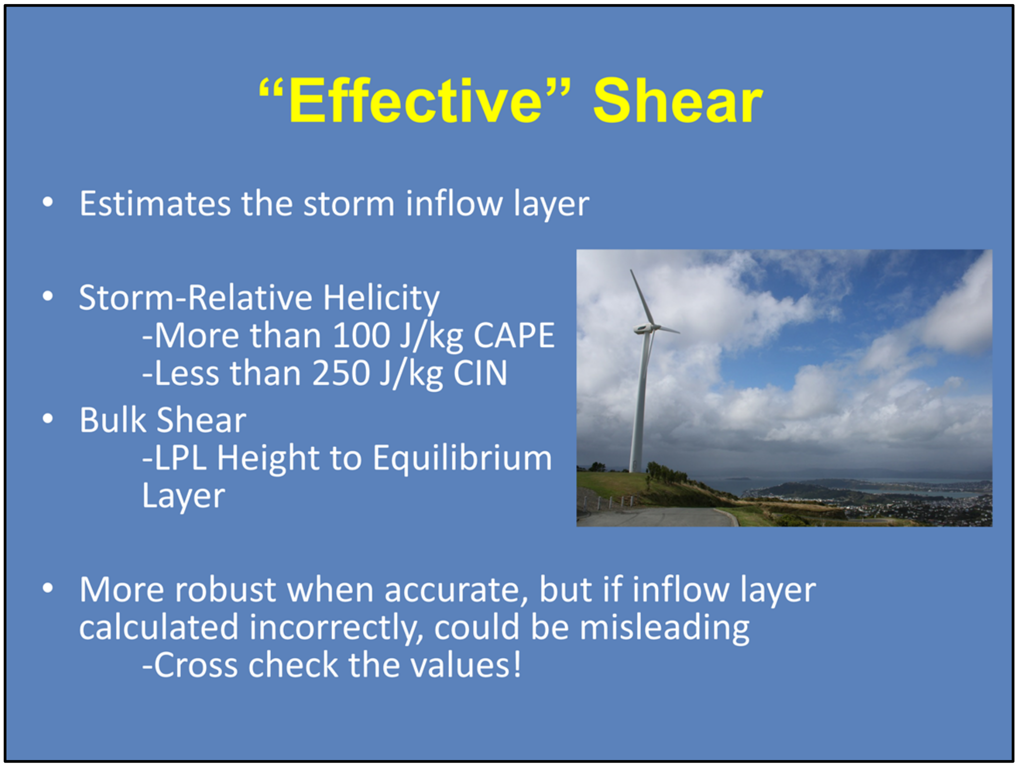

Traditionally Deep layer shear was measured in the 0-6 km layer, and low level shear was measured in the 0-1 or 0-3 km layer. Now we attempt to estimate the actual storm inflow layer, using model derived values.

For example to determine effective storm relative helicity tests parcels for more than 100 j/kg of CAPE and less than 250 j/kg of CIN

Bulk effective shear measures the Lifted parcel level height to the equilibrium layer.

Now in many cases effective values and the traditional layers will be very close or identical, but in general the effective parcels will be more robust as discriminators when accurate. But if the thermodynamic data is wrong going into the calculation they could be misleading. As a result it makes sense to look at for example, effective and 0-1km storm relative helicity to check for differences and what those might mean meteorologically.

Thompson et al 2007 & 2004
SPC MesoAnalysis

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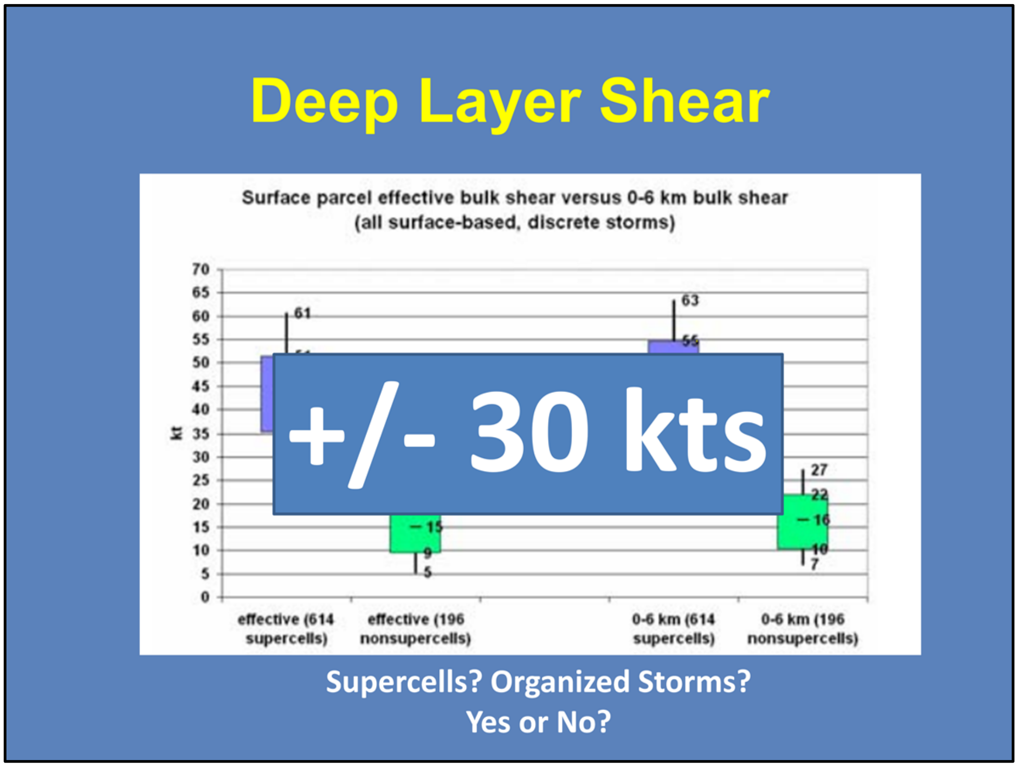

So on a day you are expecting storms, deep layer shear answers the questions, supercells? Organized storms, or maybe slow moving pulse storms?

And the switch flip at and above AROUND 30 knots.
Thompson et al. 2004

The amount of 0-1 km shear in the environment seems to differentiate between favorable and less favorable tornado environments.

Blowing this up a little you can see the significant tornado cases lead the pack, with values of 200-300-400 meters squared per second squared with lower values for weak tornado and lower values still for non tornadic supercells. Although be aware there is some overlap there, and how well you have modeled or sampled the winds, and in the event of effective SRH, the themodynamics of an event will matter greatly in the usefulness of this . If its straight off a sounding or VWP, versus say MesoAnalysis that might not have initialized correctly in that hour based on limitations in the RAP model.

Beyond that very small physical differences can exert a big influence. Vertical shear in the lowest km tends to be about 5-10 kt stronger for the significant tornadic supercells. So the margin for error is pretty small.

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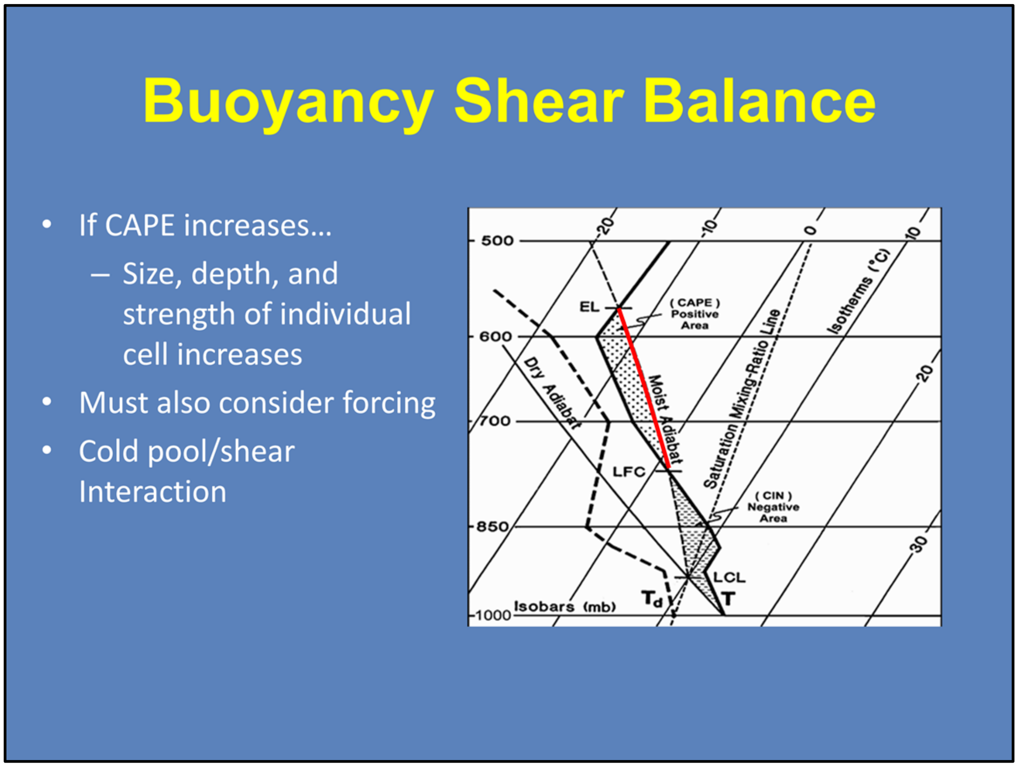

The amount of available CAPE also naturally plays a role in convective potential with large cape tending increasing the size, depth and strength of individual cells.

But its not just more cape equals more storms you also have to factor forcing are thunderstorms developing on a prefrontal trough, along an outflow boundary, a cold front, the seabreeze?

And the way the storm will interact with any available shear and developing cold pool. A seabreeze storm developing in a weak shear environment will probably be single celled, short lived and propagate slowly. A series of cells developing just along and ahead of a cold front with strong large scale ascent from an approaching trough in a modestly unstable and highly sheared environment might be ripe to become a well organized MCS with high damaging wind potential.

Holton and Hakim, 2013
Bluestein, 2013

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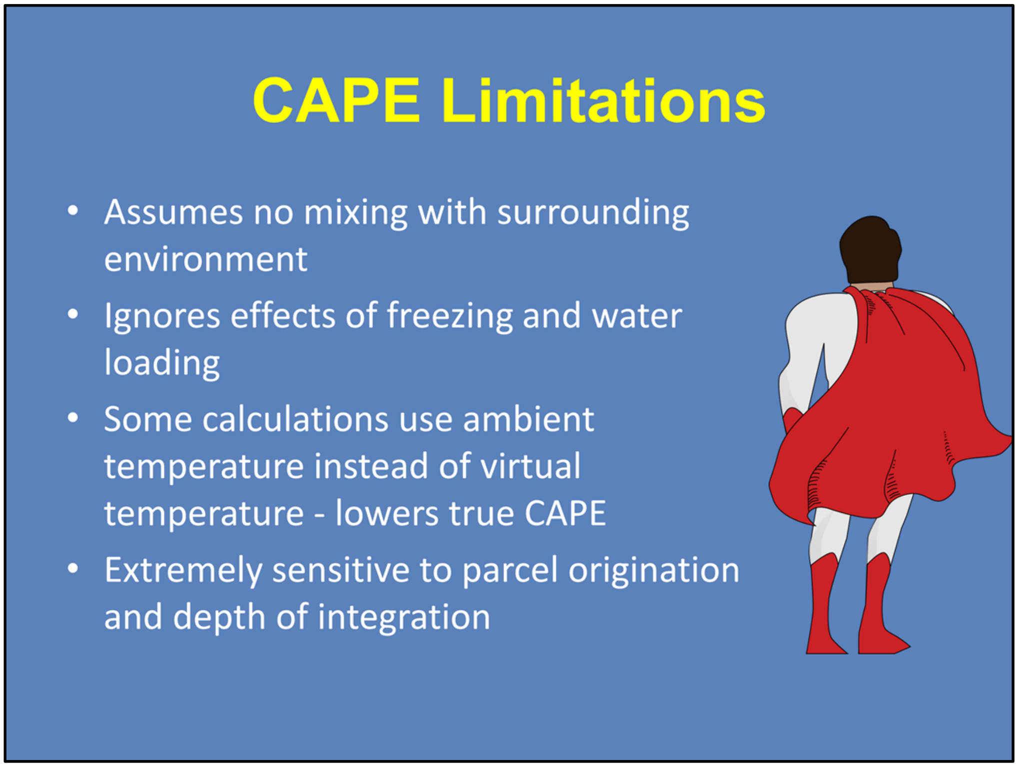


We all covered this in undergrad, and probably derived it in unspeakable ways, but lets pause for a second to talk about the limitations of CAPE calculations.

First, it assumes no mixing with the surrounding environment, which isn't actually how it works, we get by with it, but its still a limitation.

It ignores the effects of freezing and water loading, which also happens.

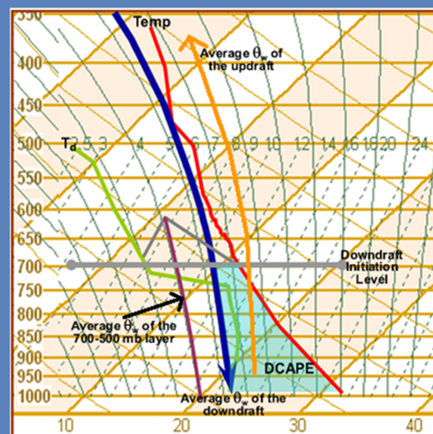
You would want the virtual temperature calculated CAPE for true energy available in an idealized environment

And it is extremely sensitive to where you start your parcel from. Physically, the 100mb mixed layer CAPE which attemps to calculate an integrated mixed layer, is going to be the most physically realistic and should be a front line source operationally.

Holton and Hakim, 2013
Craven, Jewell, and Brooks 2002
Doswell and Rasmussen 1994

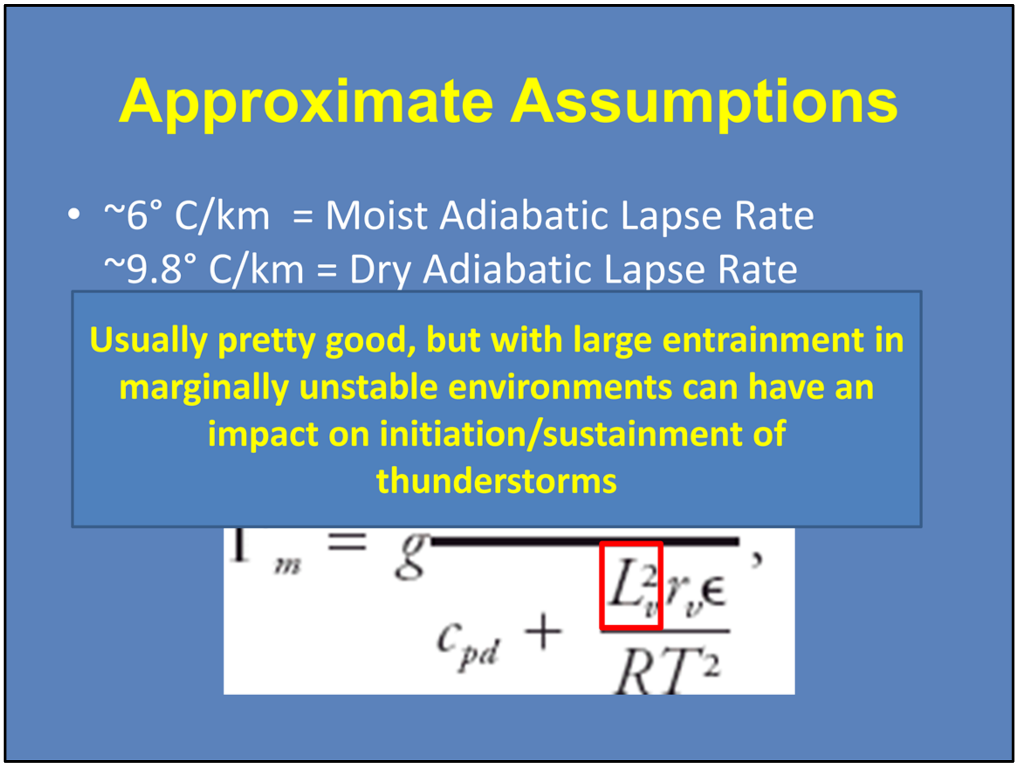
Role of Dry Air

- Above surface dry air is entrained into both convective updrafts and downdrafts
- Decreasing positive buoyancy by increasing the parcel lapse rate



So CAPE calculations, the little line on your screen showing the theoretical parcel path do not factor in outside entrainment of dry air in the calculations, but in reality of course both updrafts and downdrafts entrain less than saturated air. This causes a decreasing positive buoyancy on updrafts by increasing the effective parcel lapse rate.

Holton, 2013
De Rooy et al. 2013



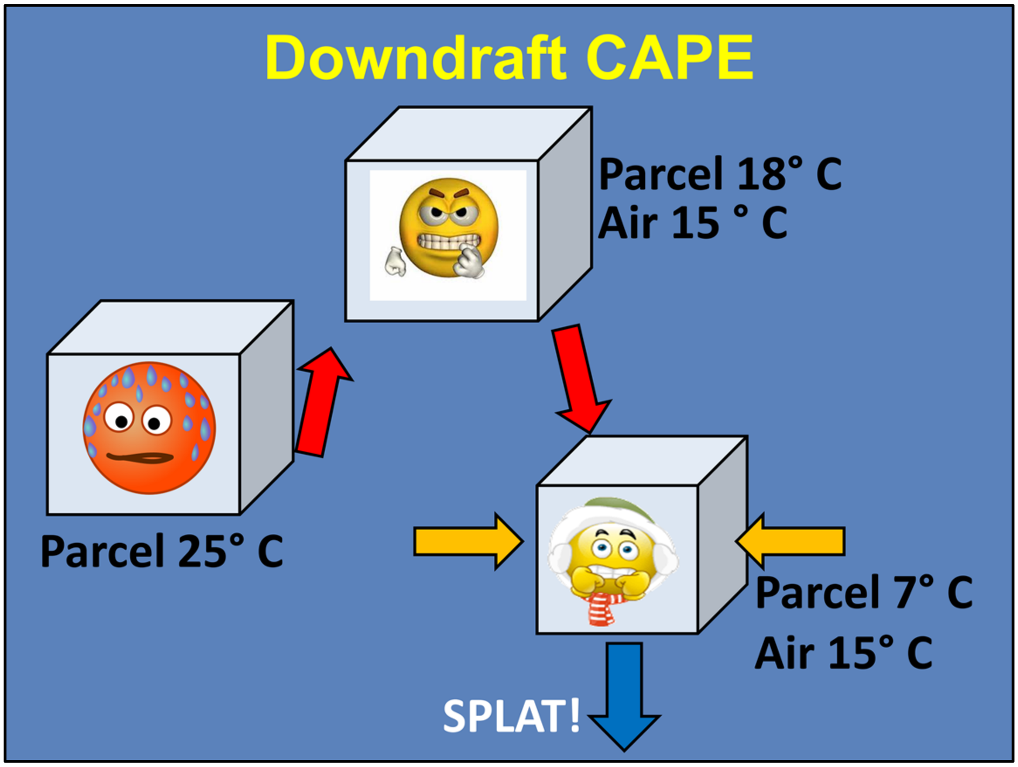
In school we approximate the Moist Adiabatic Lapse Rate as around 6 degrees C/km and the dry adiabatic lapse rate is 9.8 to 10 C/km. On our skew-Ts we draw our parcel line flat along the dry adiabat until we hit the LCL then follow the moist adiabats.

The problem is the actual lapse rate is a little more complicated

With the mixing ratio of the parcel factored in, drier air causes less condensation, less latent heat release makes the parcel cool more quickly increasing its effective lapse rate.

Now usually these assumptions are pretty good, but with large entrainment of dry air just off the surface, like the 850 to 700 mb layer, it can have an impact on initiation and sustainment of storms, and their intensity, especially if that layer is only marginally unstable with weak lapse rates. Just one of those things we want you thinking about when you are out there operating.

Holton 2013
De Rooy et al 2013
Gibbs and Butts 2015



Downdraft CAPE has similar advantages and limitations. When a parcel get started
And gets a nudge to the point
It becomes warmer than its environment,
Its determined to become a thunderstorm in most cases, especially from this theoretical standpoint.
When that parcel descends, especially if it descends below the cloud layer
Dry air entrains into that parcel from all sides and a saturated parcel undergoes a lot of evaporation.
This obviously makes the parcel colder and when the parcel is colder than the air surrounding it
It will accelerate towards the ground, which can lead to pretty big consequences on the ground if its strong enough.

Holton and Hakim 2013
Bluestein 2013

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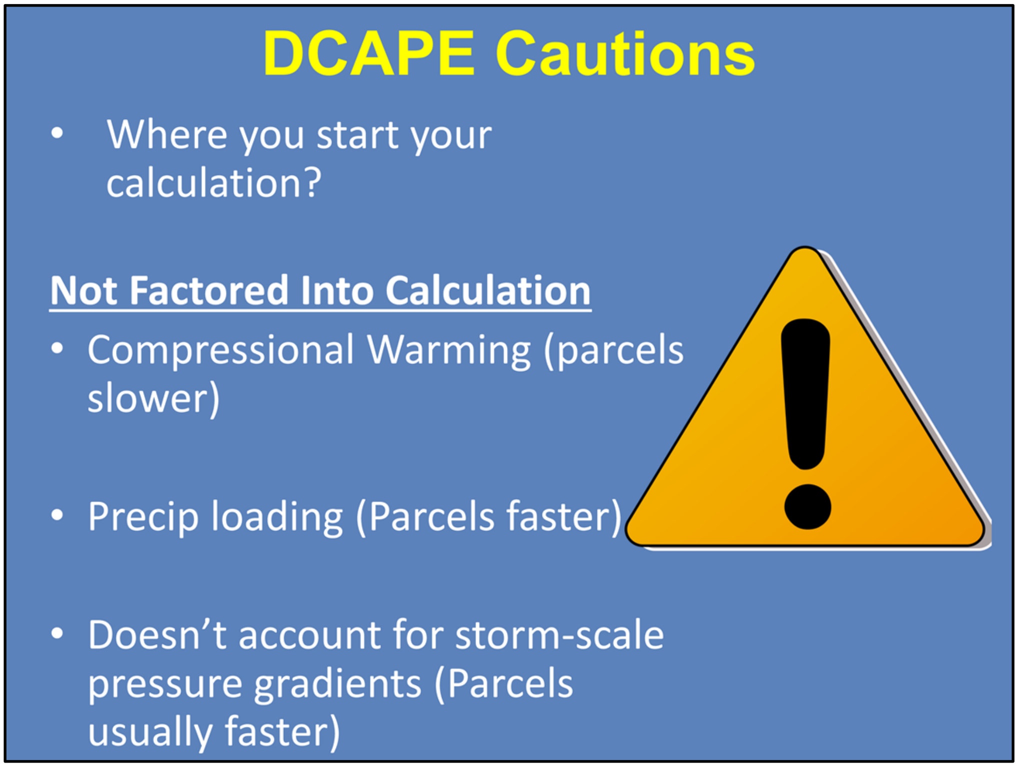

There are a few cautions to apply to properly interpret DCAPE. First, note that we started the integration at a specific level (700 mb) and called it the downdraft initiation level. In actuality, there is less certainty as to where the downdraft initiates than an updraft. Most downdrafts initiate over a layer rather than a level.

Secondly, compressional warming, the heating of the air as it sinks, usually outpaces the cooling by evaporation, so the parcel never really follows the theoretical curve, but warms more quickly.

As a third caution, DCAPE does not account for precipitation loading. A high reflectivity core interacting with gravity could cause more downward acceleration than from DCAPE process alone.

Finally, DCAPE does not account vertical pressure gradients from strong mesocyclones or divergence such as develops in well organized MCS systems with a rear inflow jet, or in a supercell with a rear flank downdraft.

Holton 2013

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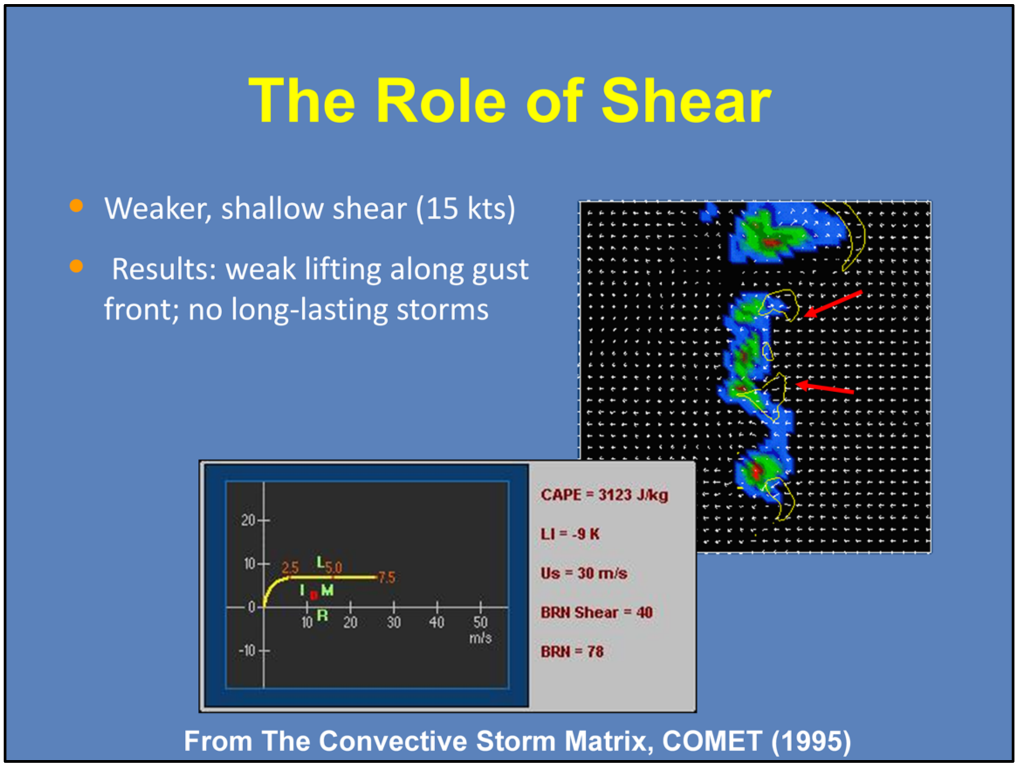

Vertical wind shear is crucial to the organization of convective systems. But just like CAPE, it's HOW the shear is distributed that is important in analyzing potential convective storm type. Large CAPE, but relatively weak shear, as is in this case would result in weaker pulse storms, with no cells lasting too long.

Bluestein 2013

Doswell and Evans 2003

Evans and Doswell 2001

Corfidi 2003

Thompson et al 2012

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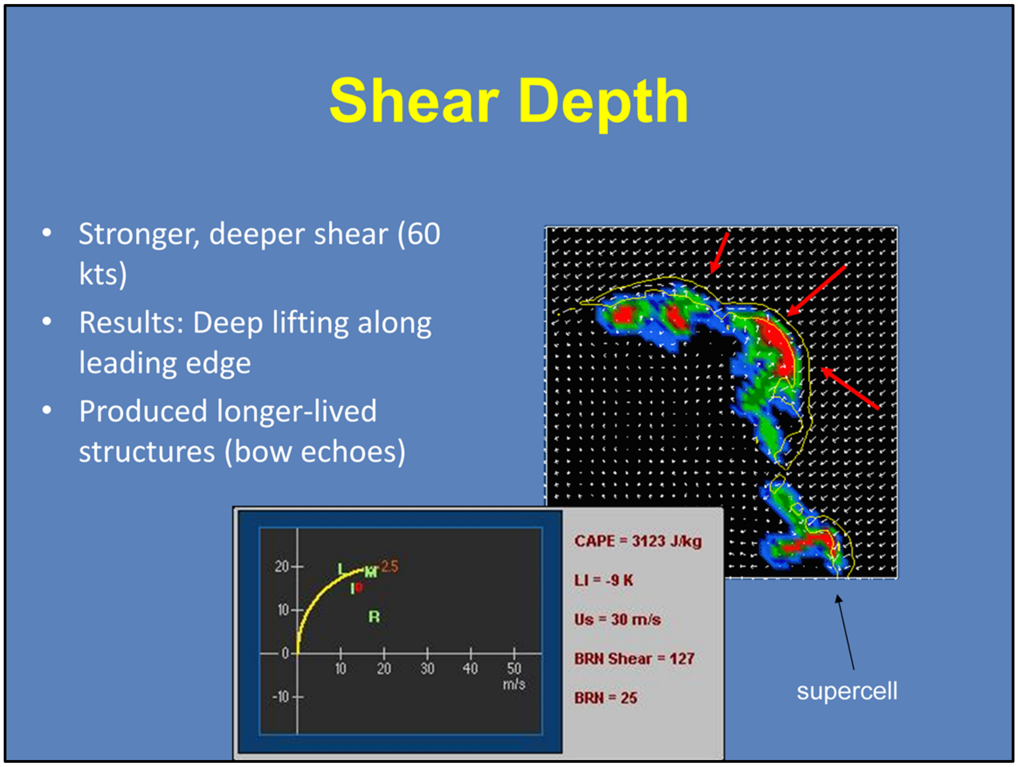


On the other hand, where strong, deeper shear environments exist in this case about 60 kts of mid to deep layer shear, this leads to deeper lifting along the leading edge of convection, and longer-lived organized squall lines including bow echoes and occasional supercells. Environmental instability and system relative flow must also be considered when predicting eventual storm type.

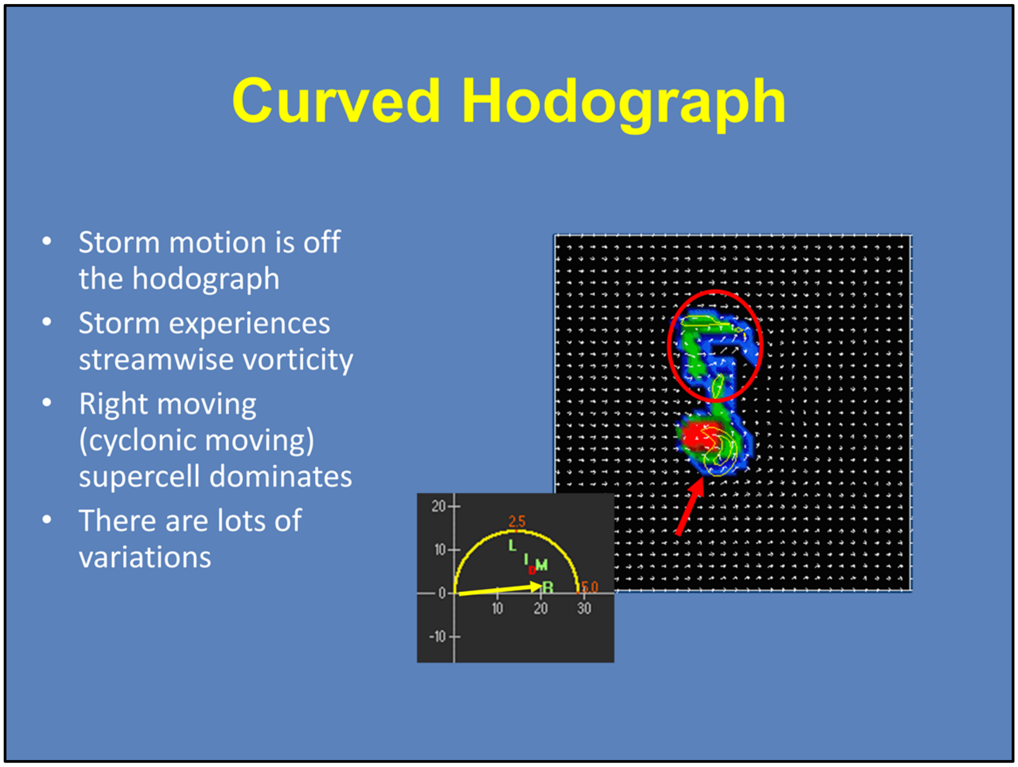
Bluestein 2013

Doswell and Evans 2003

Evans and Doswell 2001

Corfidi 2003

Thompson et al 2012



Clockwise (counterclockwise) turning hodographs favor the right-moving (left-moving) supercell and weakens the left-moving (right-moving) member. Remember, there are lots of variations.

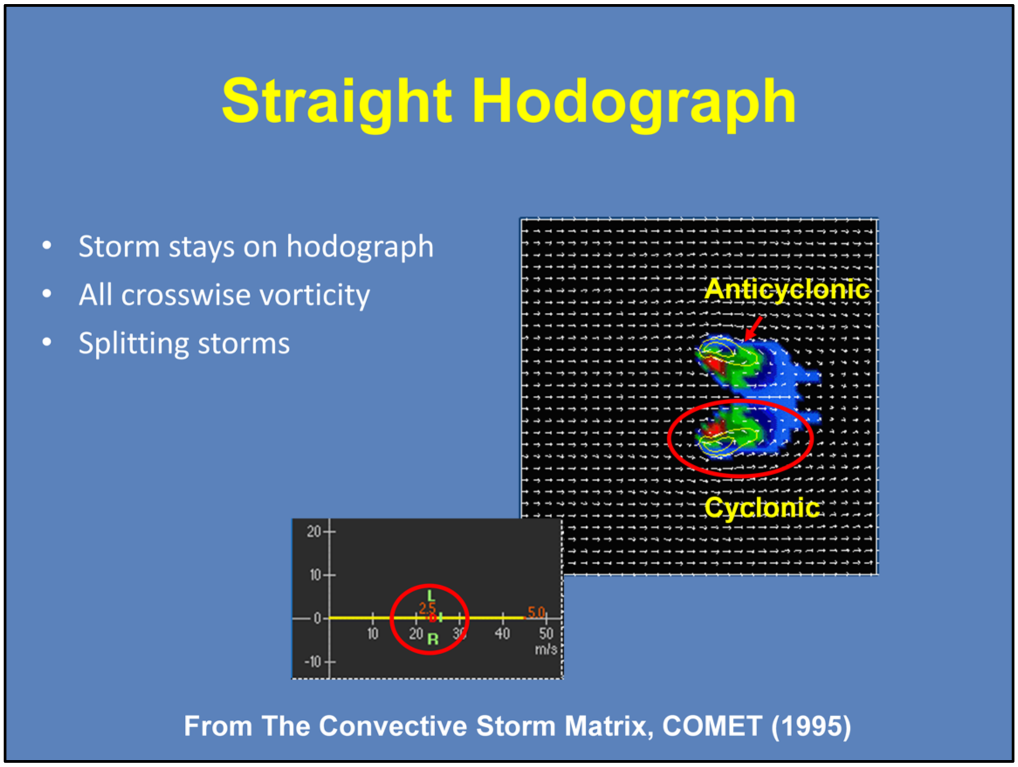
Bluestein 2013

Doswell and Evans 2003

Evans and Doswell 2001

Corfidi 2003

Thompson et al 2012



Both straight and curved hodographs produce equally strong supercells given enough shear. But, straight hodographs allow both the right (cyclonic) and left (anticyclonic) moving supercells to be equally strong. Note the mirror image cyclonic and anticyclonic supercells in this simulation from an environment characterized by unidirectional shear.

Bluestein 2013

Doswell and Evans 2003

Evans and Doswell 2001

Corfidi 2003

Thompson et al 2012

Summary

- Deep Layer Shear (Storm organization)
- Low Layer Shear (tornado/wind gust potential)

- Buoyancy
  -Limitations of CAPE/DCAPE (parcel entrainment)
  -Storm Size (but also depends on other factors)

- Shape of the hodograph
  - Curved (right movers/streamwise vorticity)
  - Straight (splitting storms, crosswise vorticity)

So in summary we briefly discussed the role of shear in thunderstorms with deep layer shear the 0-6km layer usually, dictating overall storm organization, while low layer shear gives you a discriminator for tornado and damaging wind gust potential once storms, and especially supercells exist.

The role of buoyancy in governing storm size and updraft intensity but also a reminder those calculations are just perfect enough to be problematic sometimes.

And the shape of the hodograph, with curved hodographs usually producing right moving storms, streamwise vorticity, those are your big tornado days. With straight hodographs producing more crosswise vorticity and splitting storms, those are the days you could get big left moving hailers, but you probably would have a little lower tornado threat.

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For Additional Help

1. Check with your facilitator (typically your SOO)
2. Send your questions to:
   - nws.wdtd.rachelp@noaa.gov

For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.


Radar & Applications Course
(RAC)
Convective Storm Structure and Evolution
Ordinary Cell Convection
Warning Decision Training Division (WDTD)

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Learning Objectives

1. Identify the characteristics of convection associated with ordinary cells.
2. Identify how to anticipate the motion of ordinary cells.

These are the learning objectives for this lesson. The test at the conclusion of this lesson includes questions based on these objectives.

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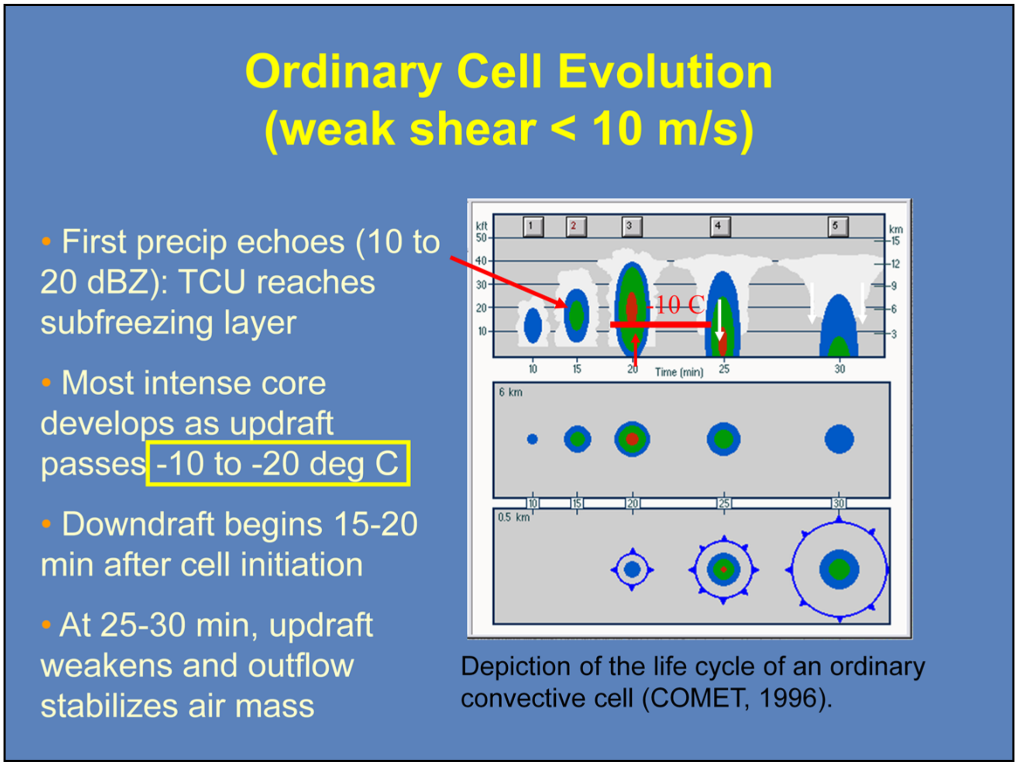

In this three panel graphic we see the cross section and plan views of an ordinary cell. The initial towering cumulus causes sharp gradients in the refractive index of the atmosphere along the cloud edges. These gradients scatter just enough of the incident WSR-88D energy back to result in -10 to 0 dBZ echoes just above the boundary layer. The first real precipitation echoes (10-20 dBZ) develop as the towering cumulus rises into the subfreezing layer. The most intense core develops as the updraft passes through the -10 to -20 deg C layer. The onset of downdraft is likely to occur as the precip core exceeds 45-50 dBZ. A downdraft usually begins between 15 and 20 minutes after cell initiation. The base of the descending precipitation core and the downdraft are typically, but not always, coincident. Therefore, when the core has reached the ground the downdraft begins to spread out into a cold pool. At this time, the updraft remains strong around a preferred side of the descending core. At 25-30 minutes after initiation, the updraft begins to weaken as the outflow stabilizes the low-level environment at its roots. Without a continuous feed of unstable low-level air in a weakly sheared environment, the updraft dies in the lowest several km above ground leaving an anvil behind.

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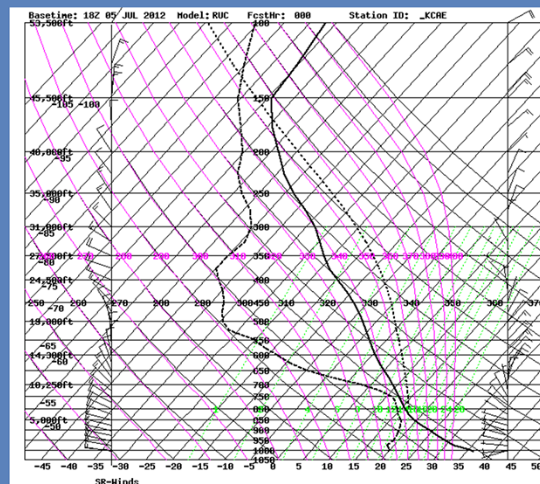
How Do Ordinary Cells Move?

- Cells move with the mean wind
- Use cloud bearing mean wind layer (could be lower or higher than 0-6 km layer)



0-8 km mean wind = 335°/4 kts

Single cell storms in the absence of significant shear move with the flow at any level (which is not surprising since the flow at any one level is nearly the same as any other level). Based on Byers and Braham (1949) in the Thunderstorm Project, they found that the best estimate for steering layer flow was the mean wind in 0-6 km AGL layer. A common mean wind calculation is weighted by density giving more influence to the low-level flow. If the average uses a deeper layer (ex. 0-12km), then weighting the average may provide more accurate results. Also, low- (high-) topped thunderstorm motion may be influenced better by a shallower (deeper) mean wind. So, use the cloud-bearing mean wind layer, which could be lower or higher than 0-6 km layer. In this example, we can see from the sounding analysis output, the 0-8 km layer best represents the mean cloud-bearing layer is a mean wind vector from 335 deg at 4 kts, which would suggest isolated storms that develop would move very slowly SSE, which is exactly what they did.

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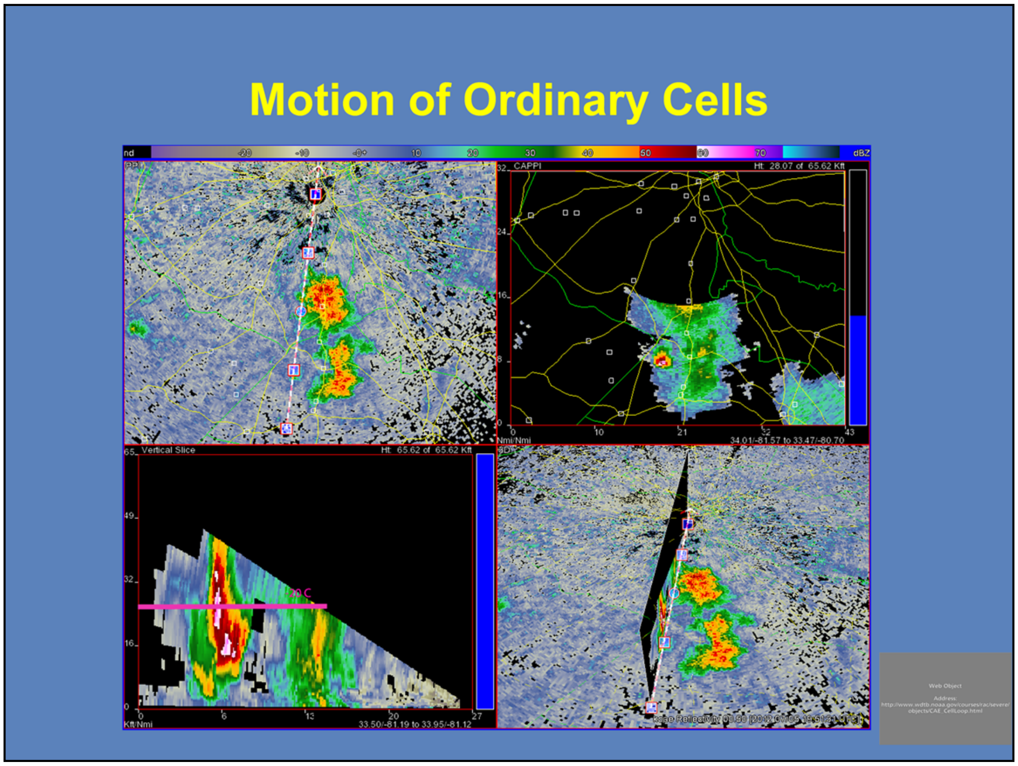

To illustrate an example of ordinary cells moving in weak shear, I am going to show an example from July 5, 2012 in Columbia, SC. There will be a flash animation (popping up in a new window) which illustrates a general southward drift of storms which developed south of the radar. In this animation, note the outflow boundary spreading outward from 20z to 21z. Pulse storms in this example tended to maximize their overall vertical extent after the leading edge of the outflow boundary and associated cold pool had passed. Outflow boundaries such as these typically spread out equally in all directions from a collapsing storm formed in an environment of weak ambient shear. The depth and orientation of the convergence in the boundary, plus the ambient air profile, were all factors in determining when and where storms would initiate. The still image shown on this slide is a screen capture at 1951z which shows a cross-section of a storm developing a 60 dBZ core above -20 deg C air. This pulse storm's updraft developed rapidly, then died off as the downdraft commenced within 45 minutes, but not before producing pea-size hail at 20z and a brief severe wind gust that downed trees at 2030z. A low-level divergence signature indicative of the small downburst associated with the downdraft stage of the storm is seen around 2034Z in the Velocity data.

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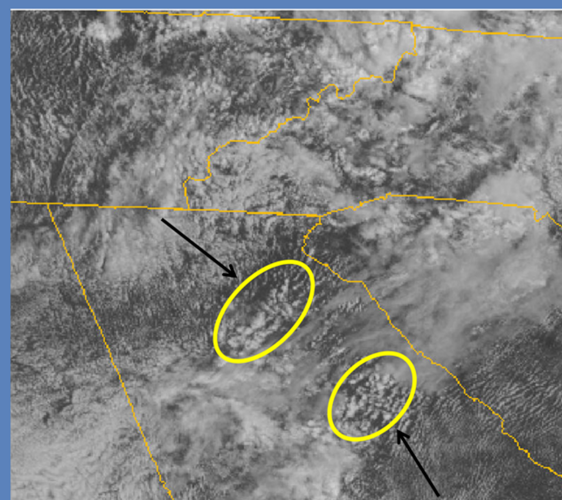
Updrafts in Weakly Sheared Storms

- Most ordinary cell updrafts only reach 50% of W_{max}
- W max limited due to water loading or dry air entrainment
- Look for wider/secondary updrafts
- Look for TCUs near mesoscale ascent



GOES 1 km Visible Image from 5 Sept 2012

The intensity of updrafts in ordinary cells is limited by their small size and speed. Most ordinary cell updrafts only reach 50% of their maximum updraft velocity due to water loading and/or dry air entrainment. A weaker updraft acceleration increases the chance that precipitation loading will diminish the strength of the updraft before it has a chance to reach the high theoretical speeds. Given the same CAPE, not all updrafts will be the same. Narrow updrafts are likely to be affected by lateral dry air entrainment so wider updrafts will allow the core to be protected. Use visible satellite imagery or radar representation of the midlevel core to estimate updrafts that will be less prone to entrainment. Secondary updrafts developing near a previous storm may grow in a more moist midlevel environment than what the models or RAOBs indicate.

A large area of towering cumulus growing in a region of mesoscale ascent (such as due to convergence from a sea breeze boundary or even differential heating) may provide clues that the environment is moister than expected. In this GOES-14 1 km visible image, you can see an updraft growing due to deeper moisture and stronger mesoscale ascent.

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Summary

- Motion of ordinary storms
  - Follow the mean (cloud-bearing) wind
- Ordinary cell updrafts are only 50% of V_{max}
  - Precip loading
  - Dry air entrainment
- Pre-existing boundaries / storms will support better growth

In summary, the motion of ordinary storms in the absence of even moderate shear, is dictated by the mean wind. Be cautious of using just 0-6 km layer alone to estimate the motion as the actual cloud bearing layer may vary based on the environment that the storms develop within. The actual updraft velocity of a storm is about 50% of its maximum updraft velocity due to water loading and/or dry air entrainment. A weaker updraft acceleration increases the chance that precipitation loading will diminish the strength of the updraft before it has a chance to reach the high theoretical speeds. To help compensate for these factors, look for updrafts developing near pre-existing boundaries or around previous storms as these will provide greater moisture and vertical growth.

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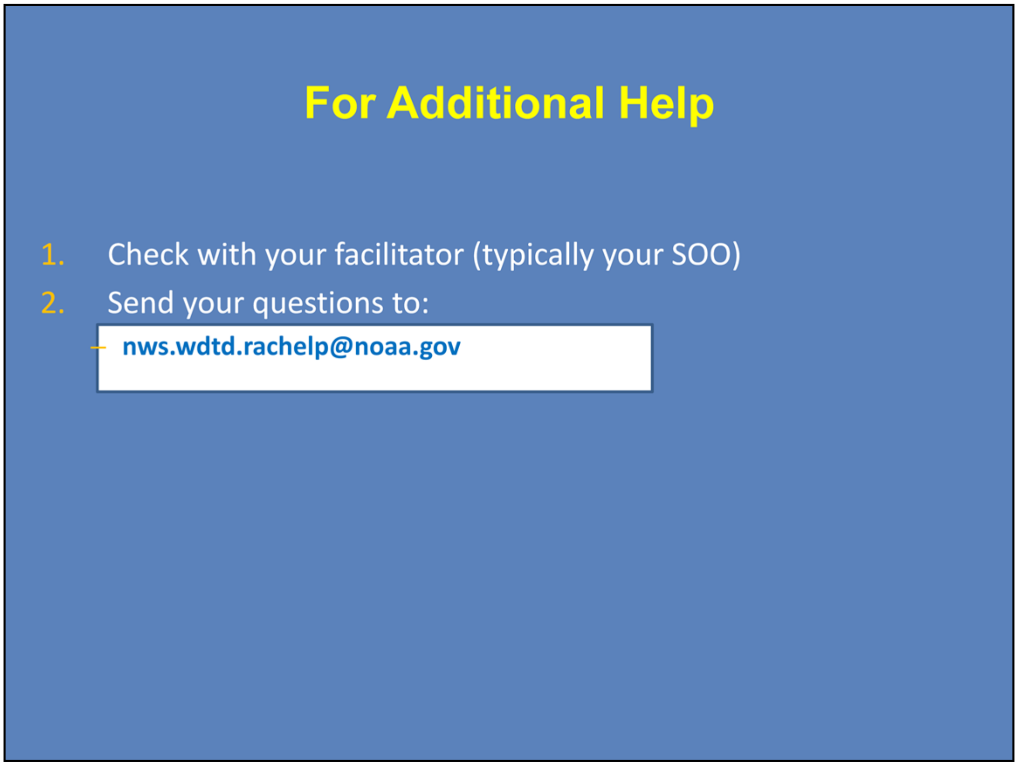

For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.

Welcome to the RAC Convective Storm Structure and Evolution lesson on anticipating lightning

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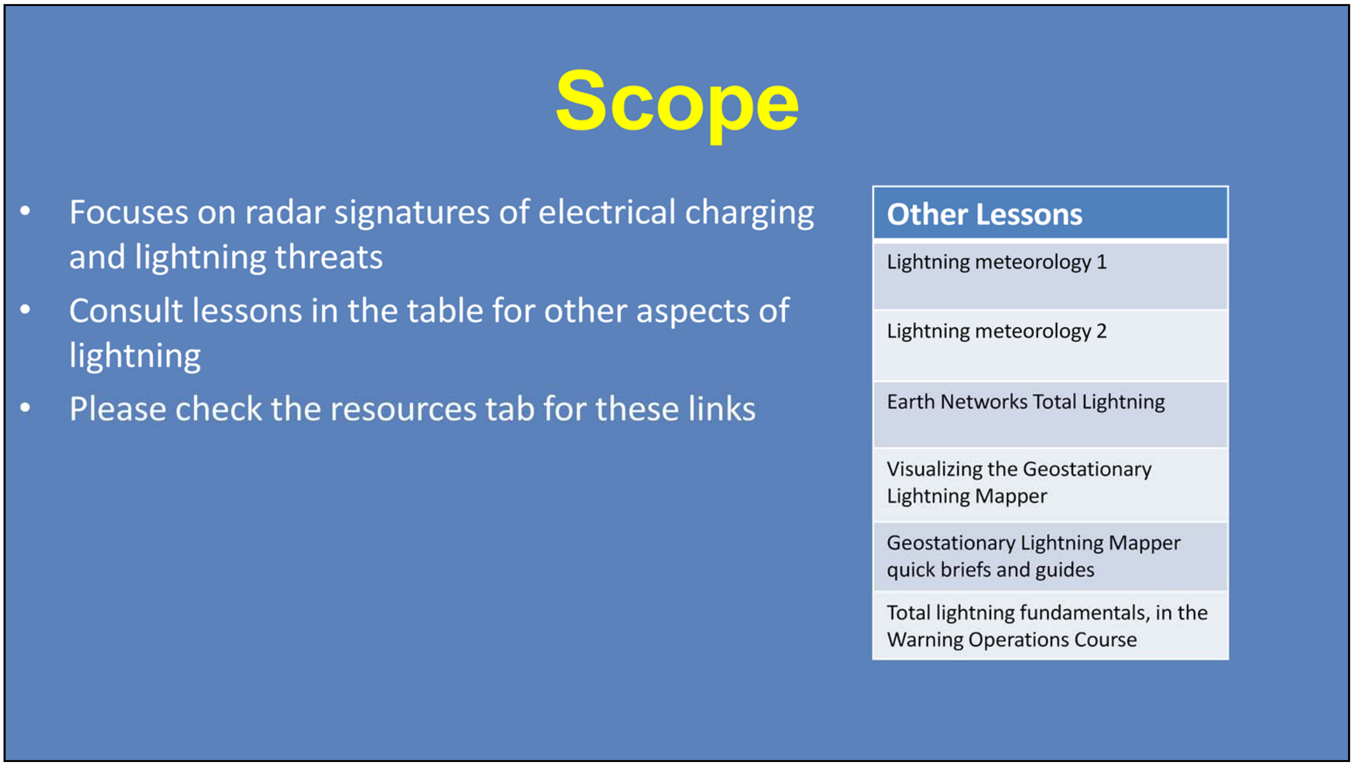
Objectives

Given this lesson,

1. Recall the approximate reflectivity threshold within a growing cumulus, before electrical charging becomes sufficient to generate the first cloud-to-ground lightning discharge.
2. Recall the most important temperature level in which to track the reflectivity critical to anticipating first cloud-to-ground lightning discharges
3. Identify the typical lead time between onset of significant charging signatures, and first cloud-to-ground lightning discharge
4. Identify the typical lead time between the relevant radar-based storm signatures and frequent cloud-to-ground lightning
5. Restate the applications of MRMS products in anticipating first cloud-to-ground lightning in a specified convective cell and compare with single radar.
6. Identify the cautions when applying radar and/or MRMS to anticipating first lightning.

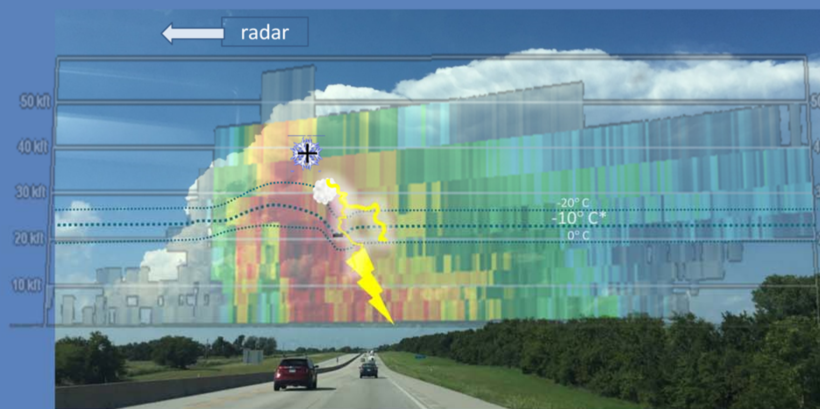
Here are the objectives. Please take a moment to review them.



The scope of this lesson focuses on identifying the radar signatures associated with electrical charging and lightning threats. There are other lessons that spend more time on the foundations of understanding the electrical behavior of thunderstorms, and the usage of lightning detection networks to detect lightning and anticipate severe weather. Check the references tab in the upper right of this player for links to these lessons.

Background

1. Given a moist updraft rising > 5ms^{-1} through the environmental melting layer.
2. Graupel forms from riming crystals or frozen drops. Note reflectivity exceeds 40 dBZ at the -10° C level.
   – We use temps adjacent to the storm
3. New ice crystals collide with graupel causing charge separation – noninductive charging
   – Ice crystals transport positive charge upward
   – Graupel falls with net negative charge
4. Intracloud discharge likely before first CG strike.
   – But not all the time.



MacGorman and Rust 1998
Photograph courtesy of James LaDue, NWS

First I do want to provide a little background on the specific charging mechanisms, and evolution to first lightning discharge, that we want to infer from radar signatures. So I transport you to a storm visible from a highway with a relatively young updraft, in this case, exposed to westerly shear. We're viewing the storm facing north. Let's assume the visual updraft you're seeing has at least a 5 m/s speed. Five m/s is a commonly used threshold that is associated with electrical charging sufficient to discharge lightning. The temperature contours show that much of the updraft is above the melting level.

Graupel forms from riming ice crystals and/or frozen drops within a few minutes of the updraft passing above the melting level, especially above the -10° C level. For convention we'll go ahead and use the -10° C level adjacent to the thunderstorm noting that within the updraft core the temperature may be several degrees warmer.

New ice crystals are always forming and they collide with the graupel. I depict one collision where charge separation occurs. This is called noninductive charging. Under typical conditions, ice crystals transport positive charge upward while graupel falls with the recently acquired negative charge.

This process multiplied sufficiently, leads to build up of a negative charge layer closer to the melting level and a positive charge layer toward the anvil.

Most of the time, an intracloud lightning discharge occurs between the elevated charge layers followed by a cloud-to-ground discharge under the negative charge layer. But sometimes the cloud-to-ground discharge ay be first.

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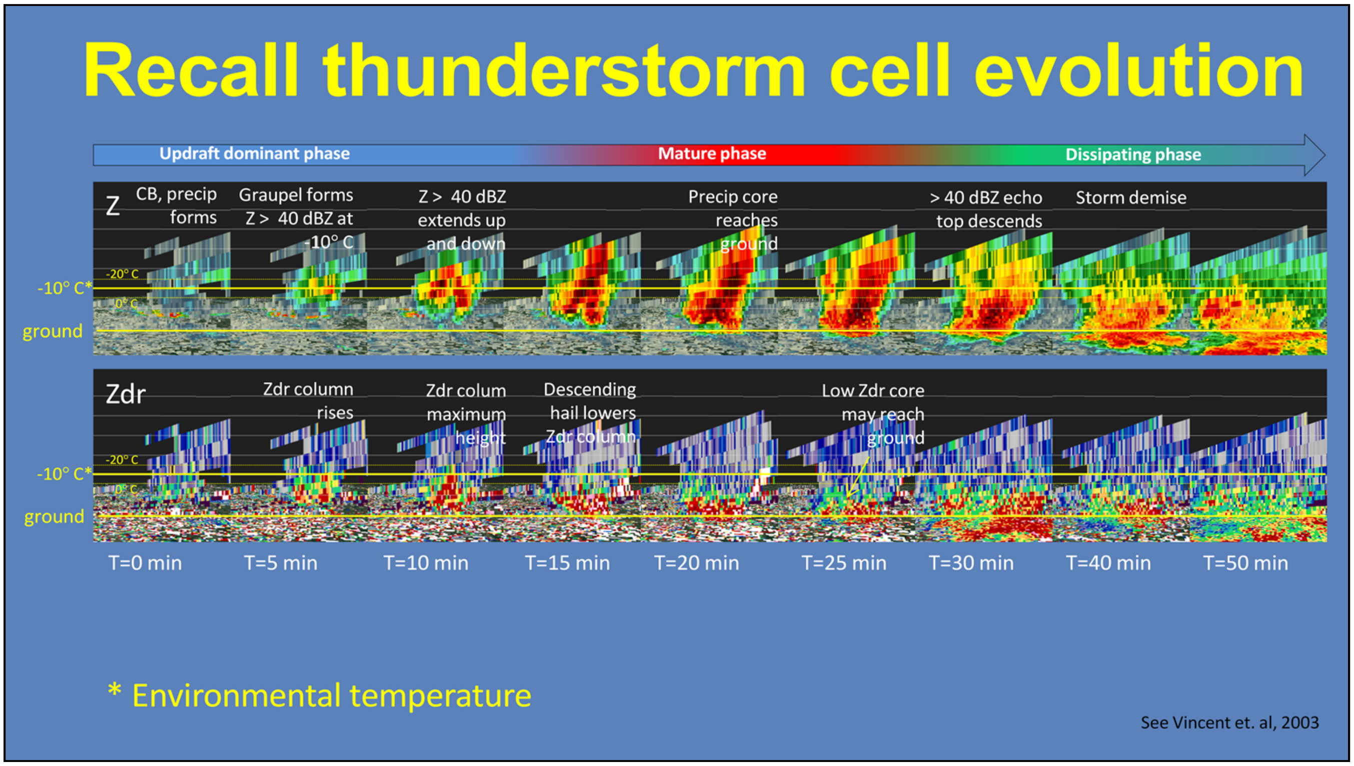


How does this evolution appear on radar? I take a typical low-shear but high CAPE pulse thunderstorm and stretch a cross-section through the storm's evolution. Reflectivity appears on top and Zdr appears in the bottom. The first elevated echo begins to develop around the -10° C level but its not strong, only 20 dBZ. Some of the echo higher up is associated with an anvil of an adjacent storm.

Since the lower echo rapidly formed, it's likely from developing snow flakes and rain drops within a strong updraft. This is time zero.

At T=5 min, the reflectivity core dramatically increases to 50 dBZ near -10° C. The Zdr shows high values at this level strongly suggesting much of this reflectivity is made up of liquid water. Remember that the temperature levels are from the environment and thus the temperature in the updraft may be closer to melting. We can see that the sudden drop in Zdr at higher levels with still high reflectivity suggests liquid drops are freezing and forming cores to graupel. Other snow flakes are probably experiencing strong riming and also turning into graupel. The stage is set for new snow flakes to collide with graupel and the charging process is beginning.

At T=10 min the 40+ dBZ reflectivity core is rising upward in the updraft and descending with a developing downdraft at low levels. The Zdr column reaches its highest level within the updraft pulse but note it's displaced a little from the descending core.

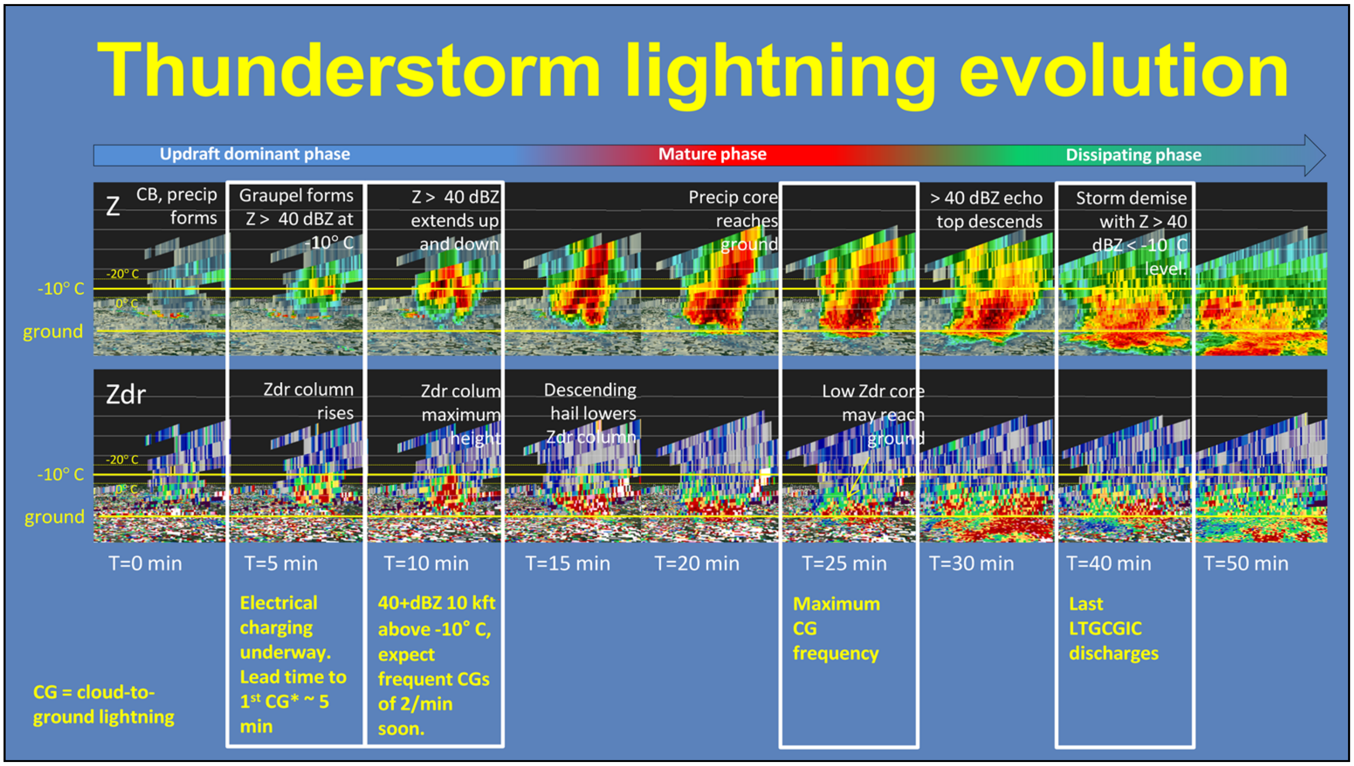
At T=15 min the 40+ dBZ reflectivity core continues to extend up and down. But note that the lowering of the Zdr column reflects that hydrometeors are experiencing significant freezing above the melting level.

At T=20 min, the 40+ dBZ reflectivity core reaches the ground with a likely downburst and outflow. It's also reaching the highest level as the storm is fully mature.

The mature phase of the storm may last another several minutes with simultanous updraft in upper levels and downdraft below the melting level. Notice that the low Zdr core has almost reached the lowest scan. Some of that graupel has turned to hail and is possibly hitting the ground.

After T=25 minutes, the storm is dissipating and the reflectivity core is raining out.

Next we'll take a look at where first lightning is most likely in this storm.



There are four key phases of lightning behavior in this storm's lifetime.

The first significant electrical charging begins at only T=5 min when the 40+ dBZ echo forms at and above -10° C environmental temperature level. The first cloud-to-ground lightning discharge occurred by T=10 min, giving us 5 minutes of lead time.

At T=10 min, when the 40+dBZ reflectivity extends 10 kft above the -10° C level, expect charging to become sufficient to generate more frequent cloud-to-ground lightning. In this case, frequent means 10 strikes every five minutes or more.

By T=25 min it is likely that this storm is producing the maximum cloud-to-ground lightning frequency given the volume of graupel and hail falling toward ground and bringing significant negative charges downward.

The last lightning discharge is likely in the storm dissipation stage once the graupel and hail have fallen out into low levels, time = 40 min.

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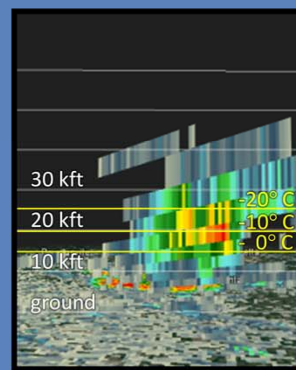
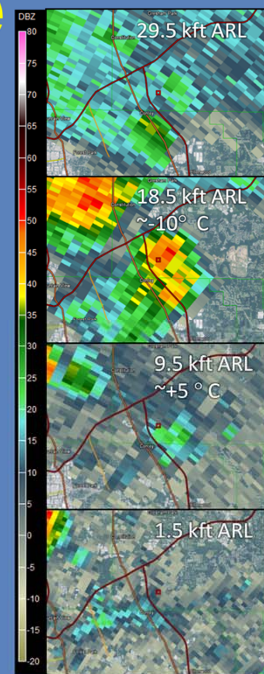
Before 1st CG strike

- Onset of Z ≥ 40 dBZ @-10° C level
- Updraft dominant phase
- Lead time ~ 5 min





So as a forecaster providing DSS, watch for the onset of 40+ dBZ echo at the environmental -10° C level in a new convective updraft. In planview, look at the scan crossing closest to -10° C. Note the lack of echo below and above, especially before this scan indicating a young storm. This is the time to advise your users of lightning danger. You may have only five minutes lead time.

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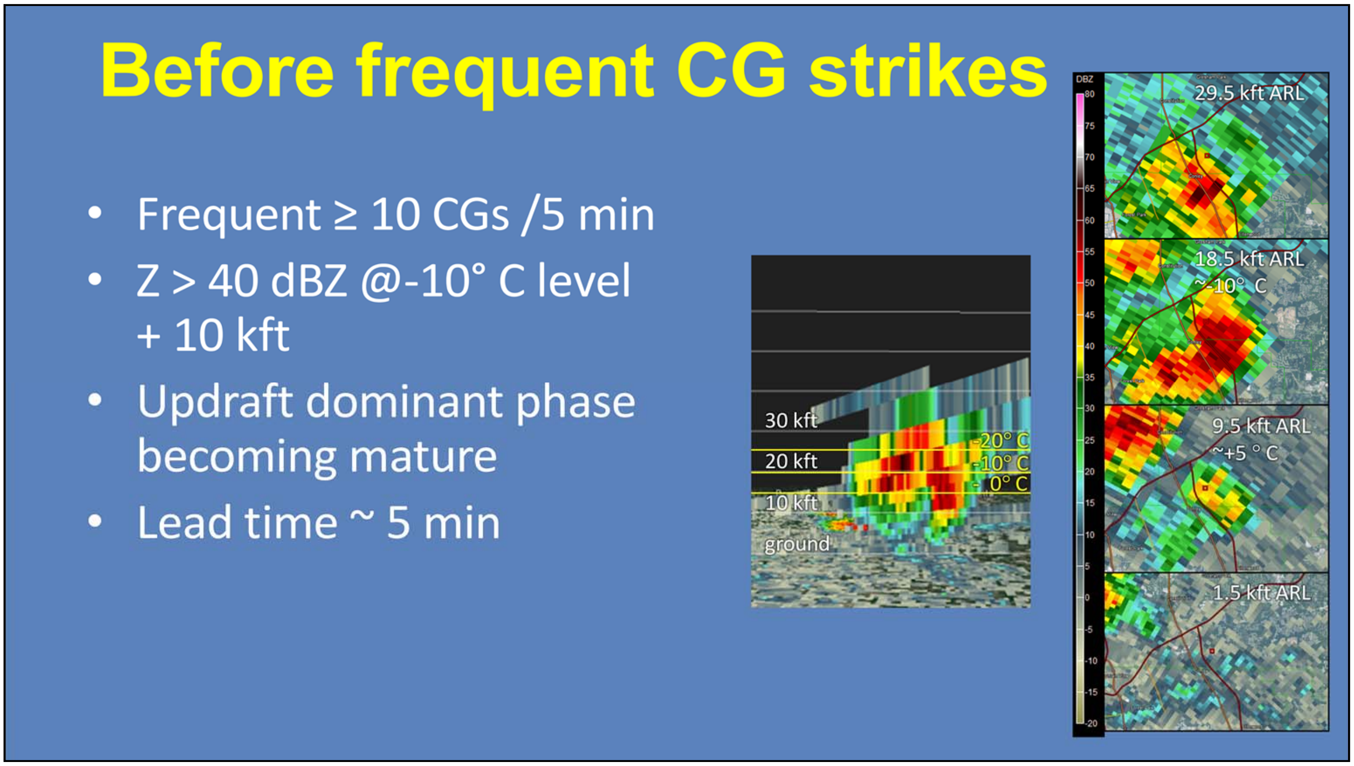

Frequent cloud-to-ground lightning is imminent after the 40+ dBZ echo extends to at least 10 kft above the environmental -10° C level. In planview, notice the 50 dBZ reflectivity at 29.5 kft ARL, or about 11 kft above the -10° C level in this updraft dominant phase of the storm. You'll have about five minutes lead time before getting multiple CG strikes.

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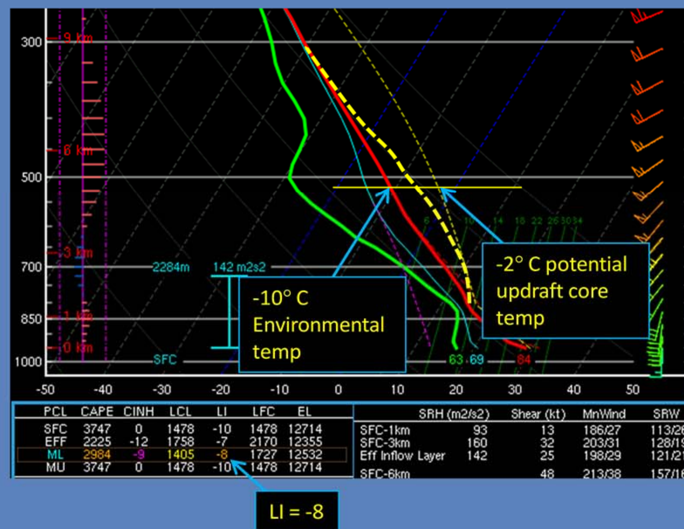
Caution: the -10° C layer

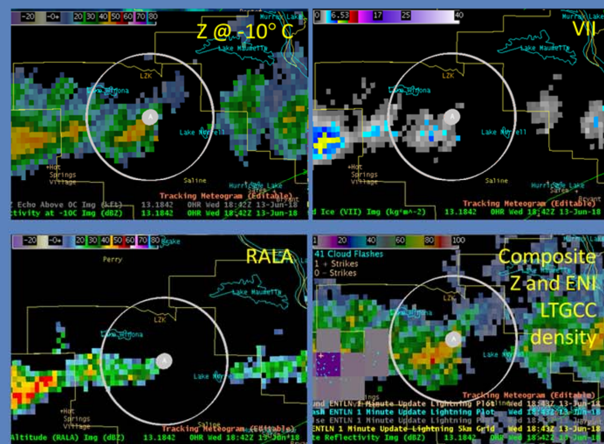
- Updraft core temperatures are warmer than the environmental -10° C layer.
- Precip core may be too warm for charge separation if the Lifted Index is strongly negative (< -4° C).
- Consider viewing precip core at higher altitudes.
- Account for higher lead time and false alarm potential to first LTGCG.
- Fortunately, initiating cells rarely have such relatively warm updraft cores.



I mentioned previously that we're using the environmental temperature level. Yes the updraft core temperature is warmer, especially in highly buoyant atmospheres. So that may mean that the reflectivity at -10° C may not consist of enough graupel/ice crystal mixture to result in charge separation. You may want to look at a higher scan. Or more likely, given this highly buoyant atmosphere, and the understanding that this is a growing updraft, go ahead and anticipate first strike even though there may be a slightly higher false alarm potential. You'll at least get better lead time. And you must consider that even though the sounding may indicate a very high updraft temperature, this is a new cell may experience some entrainment and is likely to have a lower temperature and thus plenty of graupel and ice crystals.

Using Multi Radar/Multi Sensor MRMS

- Fast way to compare Reflectivity at -10° C level with Reflectivity at Lowest Altitude (RALA).
- Other products:
  - Vertically Integrated Ice (VII)
- Note new >40 dBZ core aloft at -10° C with lower reflectivity in RALA.
- VII shows > 5 kg/m^2
- No lightning yet



Instead of single radar, you can use MRMS, especially the reflectivity at -10° C product. This is an easy way to detect new convective cells without having to scan through multiple elevation angles from multiple single radars. You can also use vertically integrated ice (VII).

In this case note the new >40 dBZ core in the reflectivity at -10° C whereas the RALA indicates lower reflectivity. The VII increased to more than 5 kg/m squared. Note there's no lightning discharge detected by the lightning network…yet.

Remember to compare the elevated reflectivity and VII to the RALA to make sure you're detecting a new elevated core.

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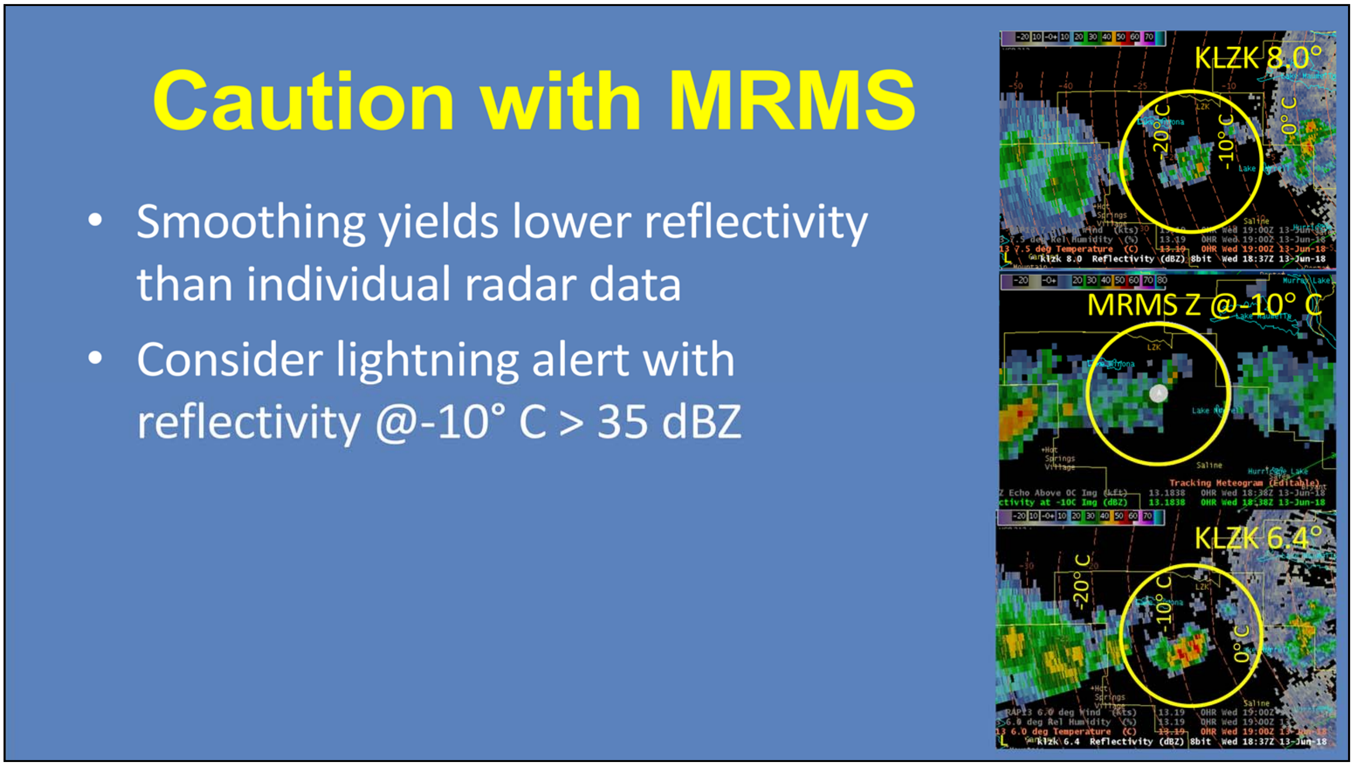

There are a couple disclaimers with using MRMS to anticipate first lightning. MRMS smooths the reflectivity from individual radars and so the peaks of small cores may be rounded somewhat. Consider a lightning alert when the reflectivity at -10° C exceeds 35 dBZ.

Looking at the same storm as the last slide, the MRMS reflectivity at -10° C is sandwiched between the KLZK 6.4 and 8 deg scans, both of which show higher reflectivity than MRMS. Still, with a 35 dBZ threshold, the MRMS still implies significant electrical charging.

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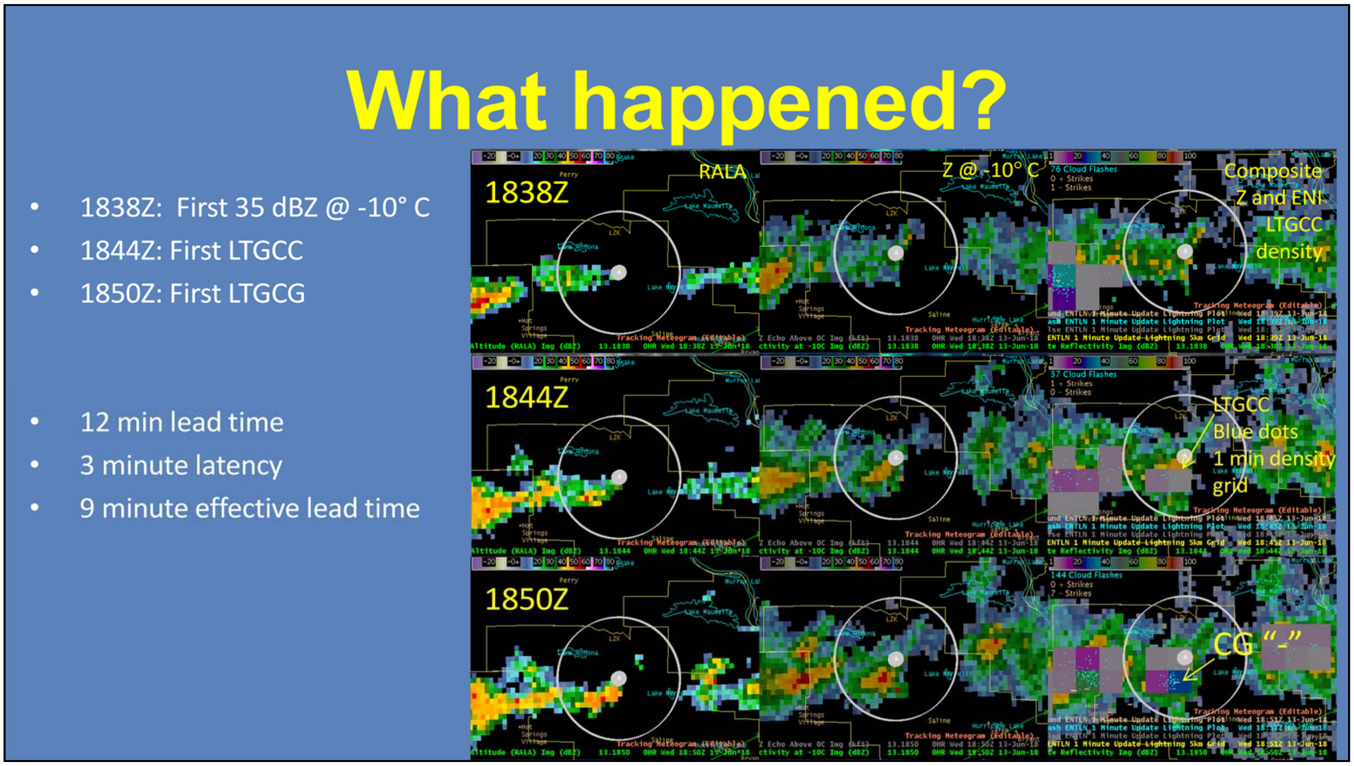

So what happened with this storm? At 1838Z the first 35+ dBZ appeared at -10° C whereas RALA showed lower reflectivity.

Six minutes laster at 1844 Z, the midlevel echo grows in strength and the first intracloud lightning discharges occur as shown by the blue dots and initial lightning flash density plot.

Six minutes after we get the first cloud-to-ground discharge yielding a 12 minute lead time from detecting this core in MRMS -10° C reflectivity product. There's a typical 3 minute latency in the MRMS products and so effective lead time is 9 minutes in this event.

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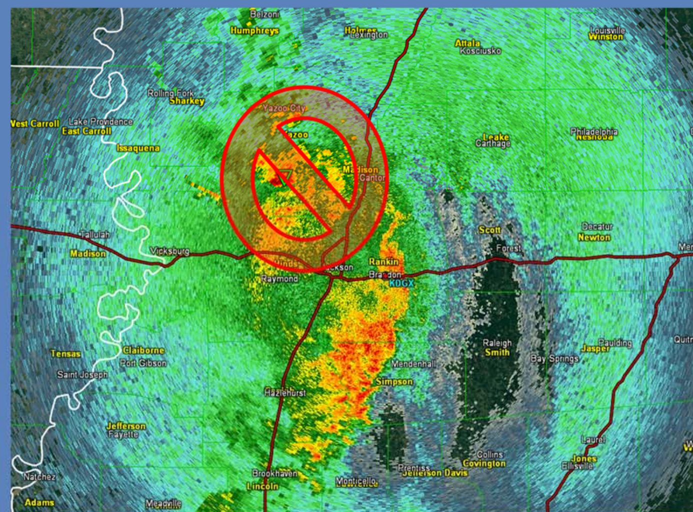
Constraints

- Use only with new cells
- Does not apply to pre-existing anvil regions



With either single radar or MRMS, you can only use this technique on new convective cells. Top-heavy precipitation profiles exist elsewhere such as in extensive anvil regions. However, lightning activity is likely already ongoing and so this technique wouldn't apply.

Summary

- Identify likely graupel-ice crystal collisions to predict first cloud-to-ground lightning in a new convective cell.
  - Identify onset of significant reflectivity at the environmental -10° C level in a new updraft.
    - ≥ 40 dBZ in single radar
    - ≥ 35 dBZ in MRMS
  - Reinforced by Zdr column
  - Rapid growth of MRMS VII
  - Higher reflectivity aloft than at low level
- Lead time to first CG ~5 – 10 min.
  - Lesser lead times with stronger updraft pulses
- For onset of frequent cloud-to-ground lightning (CGs)
  - ≥ 10 CGS/5 min
  - Identify onset in single radar of ≥ 40 dBZ reflectivity reaching ≥ 10 kft above the environmental -10° C level in a new updraft.
- Cautions
  - Applies to new convection only.
  - Not applicable to established anvil reflectivity > 40 dBZ

In summary, the goal of anticipating lightning is met by identify the onset of graupel-ice crystal collisions in new convective cells.

You do this by identifying the onset of significant reflectivity at the environmental -10° C level in a new updraft. Significant reflectivity is usually at least 40 dBZ from single radar and 35 dBZ in MRMS. You should see a Zdr column associated with the increasing reflectivity of young cells. You may also look at MRMS VII to see how quickly it's growing. Consider that you should be viewing higher reflectivity aloft than at low levels to identify a new cell.

The lead time to first cloud-to-ground lightning is roughly 5 – 10 min, more likely on the shorter end of this range with vigorous updraft pulses.

For onset of frequent cloud-to-ground lightning, or at least 10 CGs per 5 min period, look for at least 40 dBZ echo top to rise at least 10 kft above the environmental -10° C level.

As for caution, please apply this technique to new convection only. This wouldn't be applicable to substantial anvil reflectivity even though the reflectivity is higher aloft than at low levels.

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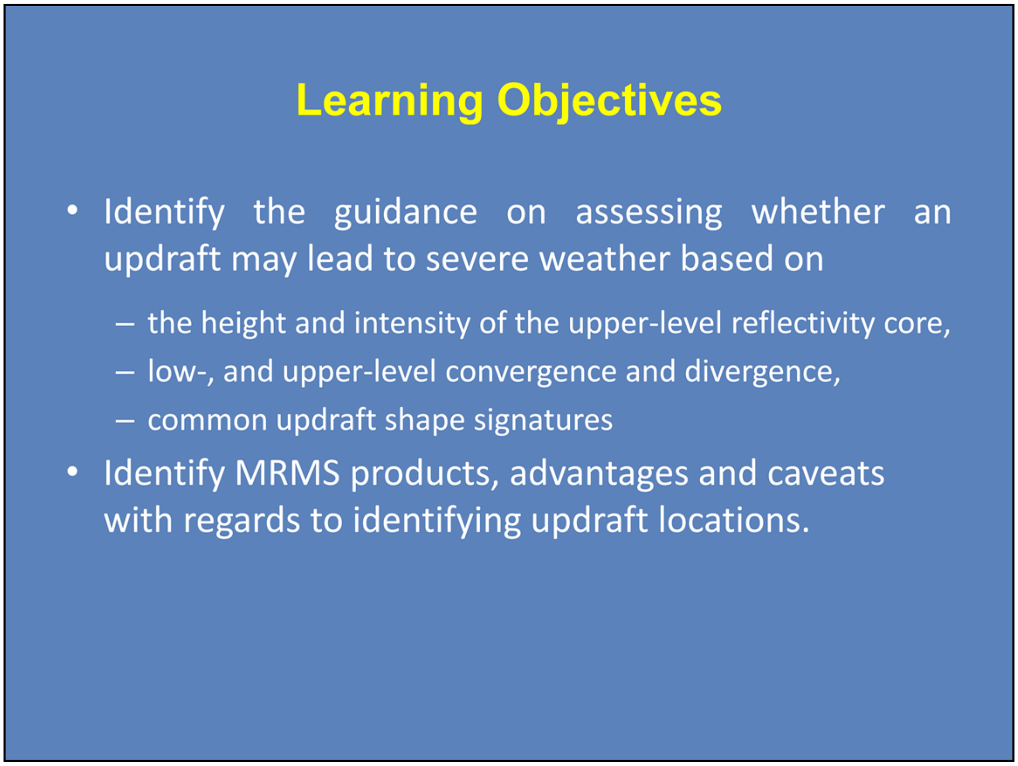
Thank you for your time. For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.

Welcome to the Convective Storm Structure and Evolution lesson on assessing updraft strength and location.



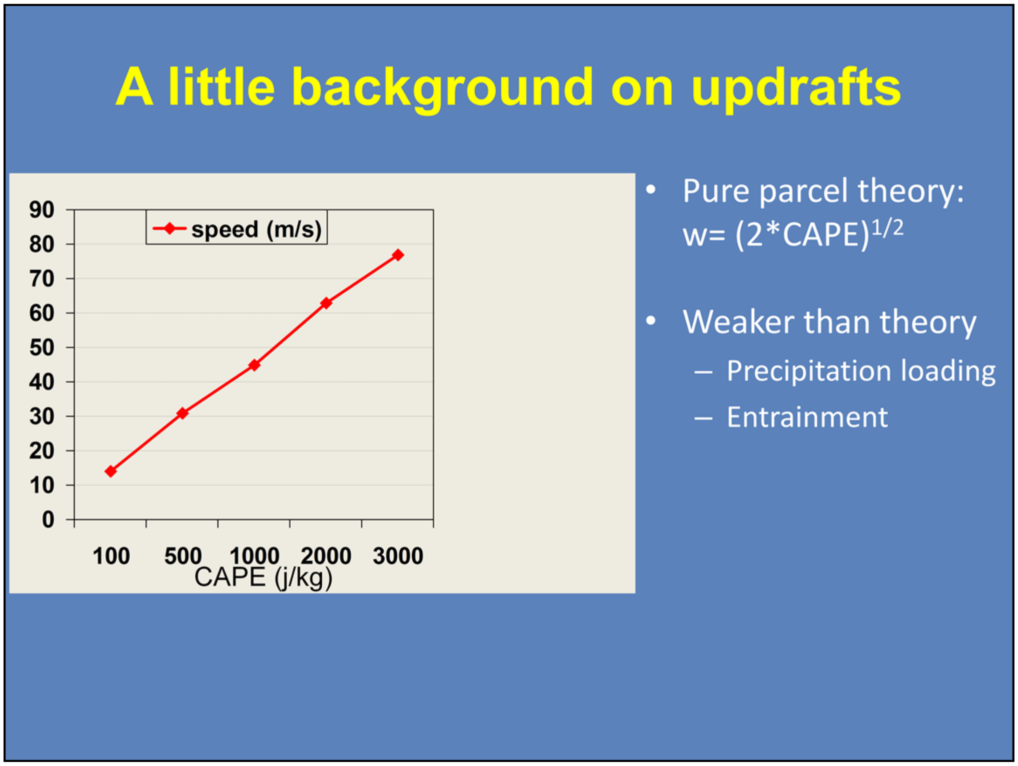
The objective of this lesson is to provide you guidance on assessing whether an updraft may lead to severe weather based on:

the height and intensity of the upper-level reflectivity core,

low-, and upper-level convergence and divergence,

and common updraft shape signatures.

Identify MRMS products, advantages and caveats with regards to identifying updraft locations.



The estimation of the maximum updraft strength (Wmax) (based on the CAPE) does not take into account precipitation loading or dry air entrainment. Therefore, most ordinary cell updrafts reach only about 50% of Wmax due to these effects. That's the blue part of this domain where the red line is the pure parcel theory updraft speed.

Weaker updrafts can result from precipitation loading. A storm with 3000 J/kg of CAPE over 18 km of depth will have a weaker updraft acceleration than one with the same CAPE over 12 km. **A weaker updraft acceleration increases the chance that precipitation loading will diminish the strength of the updraft before it has a chance to reach the high theoretical speeds**. Stronger updraft accelerations advect cloud condensation nuclei upward so quickly that significant hydrometeor growth does not occur. Perhaps CAPE density can inform you about the updraft acceleration potential in a storm.

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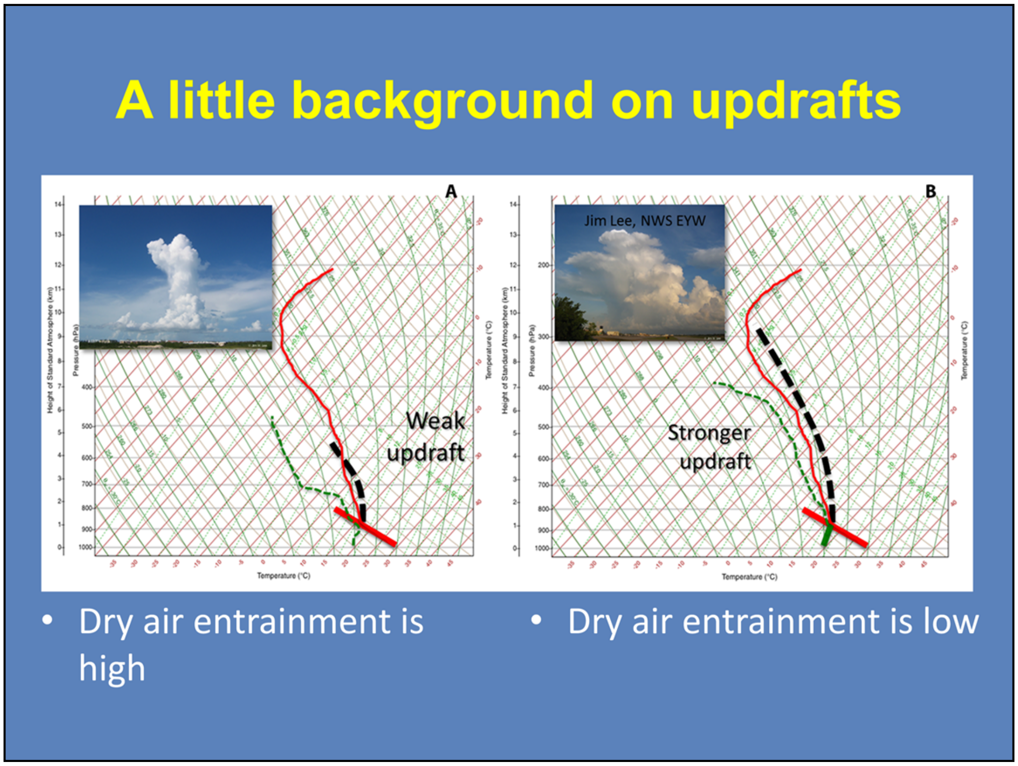
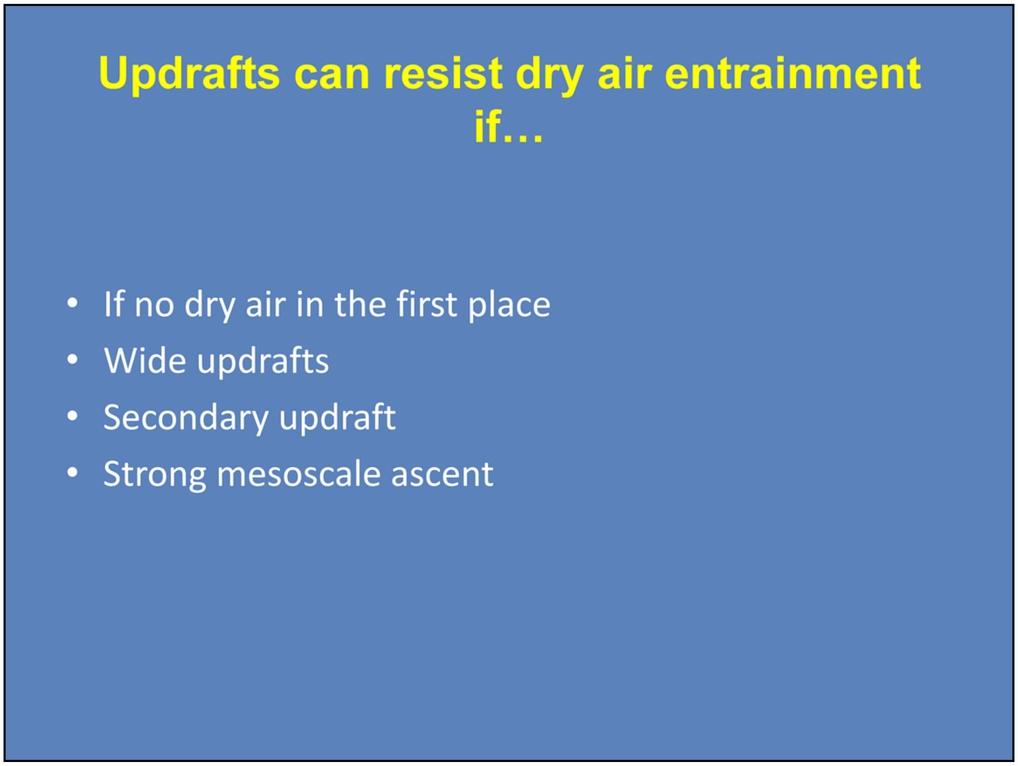


However, given the same CAPE and CAPE density, not all updrafts will be the same. Some storms remain weak regardless of the environmental CAPE. **Narrow updrafts are likely to entrain dry air to the core limiting updraft strength**. Also, significant midlevel dry air can increase the entrainment efficiency reducing the strength of an updraft even given large values of CAPE.

The impact of midlevel dry air is graphically represented by the more severe loss in parcel theta-E despite the same CAPE (or MLCAPE).



Given the effects of entrainment, look for these factors when considering the storm with the greatest updraft potential.

• **Look for the presence of dry air in a sounding** that could mix with the updraft air diminishing its buoyancy.

• **The widest updrafts allow the updraft core to be protected**. Satellite imagery of the width of the cumulus, or radar imagery of the midlevel precipitation core width, are two ways to estimate which storm will have the least entrainment potential. Large Bounded Weak Echo Regions (BWER)s can be used to infer updraft size. Wide updrafts may also manifest themselves as areas of low spectrum width.

• **Secondary updrafts developing near a previous storm may grow in a more moist mid level environment than what the models or**

RAOBs indicate.

• **A large area of towering cumulus growing in a region of mesoscale ascent (e.g., a boundary) provides a clue that the environment will be more moist than analysis show**.

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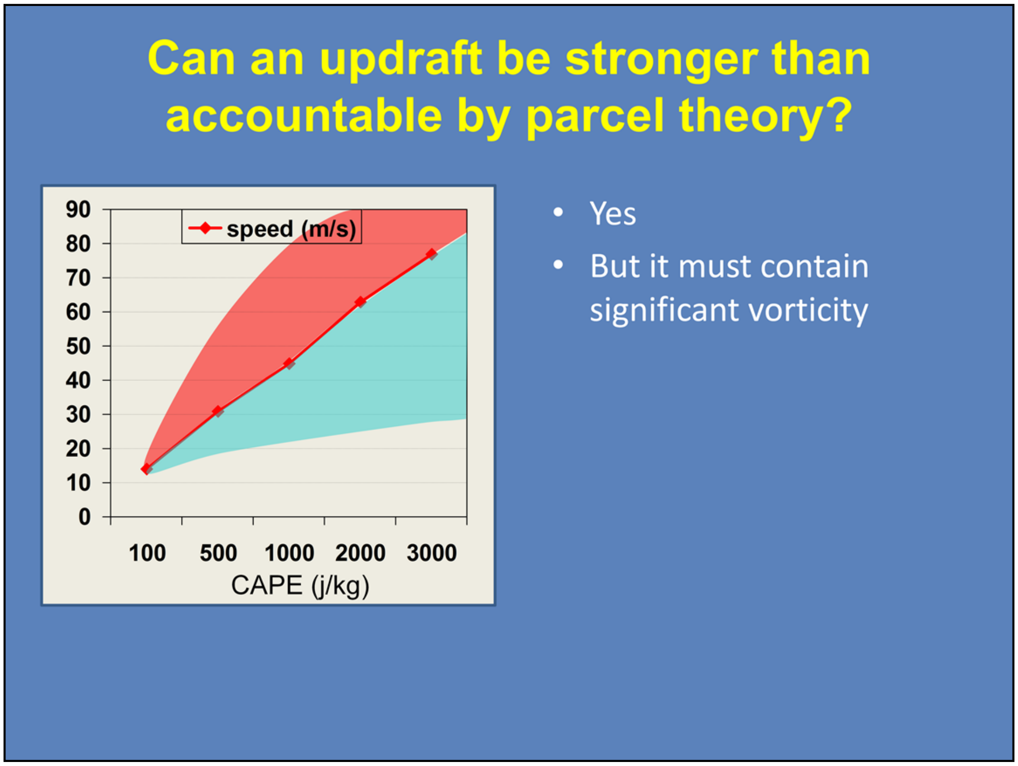
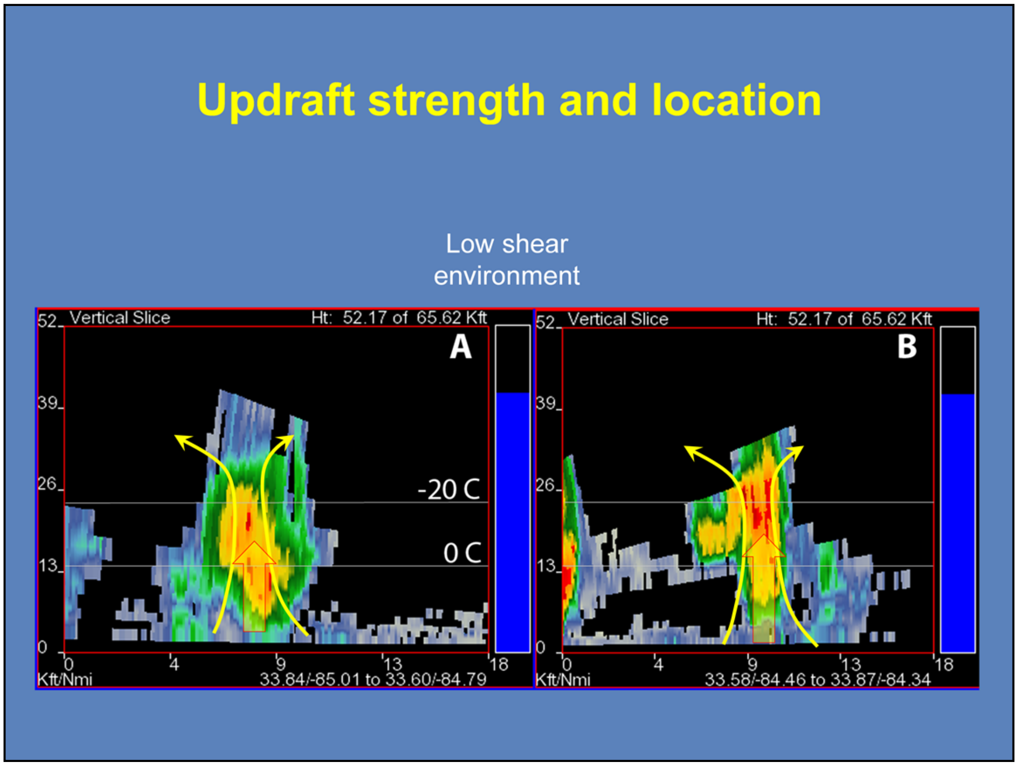


A very intense updraft can form in a relatively low updraft buoyancy environment if it is well correlated with significant vertical vorticity in mid-levels. As will be discussed in later lessons, a significant mid-level mesocyclone is occupied by a dynamic pressure perturbation pressure minimum that can significantly boost updraft strength. Some estimates based on numerical model studies suggest more than 50% of the updraft strength can be attributable to dynamic pressure forcing.



The most common technique for inferring an updraft's location is to observe the location of its upper-level reflectivity core as it reaches the maximum height in its lifecycle**.** Hydrometeor growth is maximized as the most intense part of the updraft passes through the -12° C to -20° C layer. Therefore, the highest reflectivity core in a layer centered just a bit higher should reveal the location of the strongest updraft.

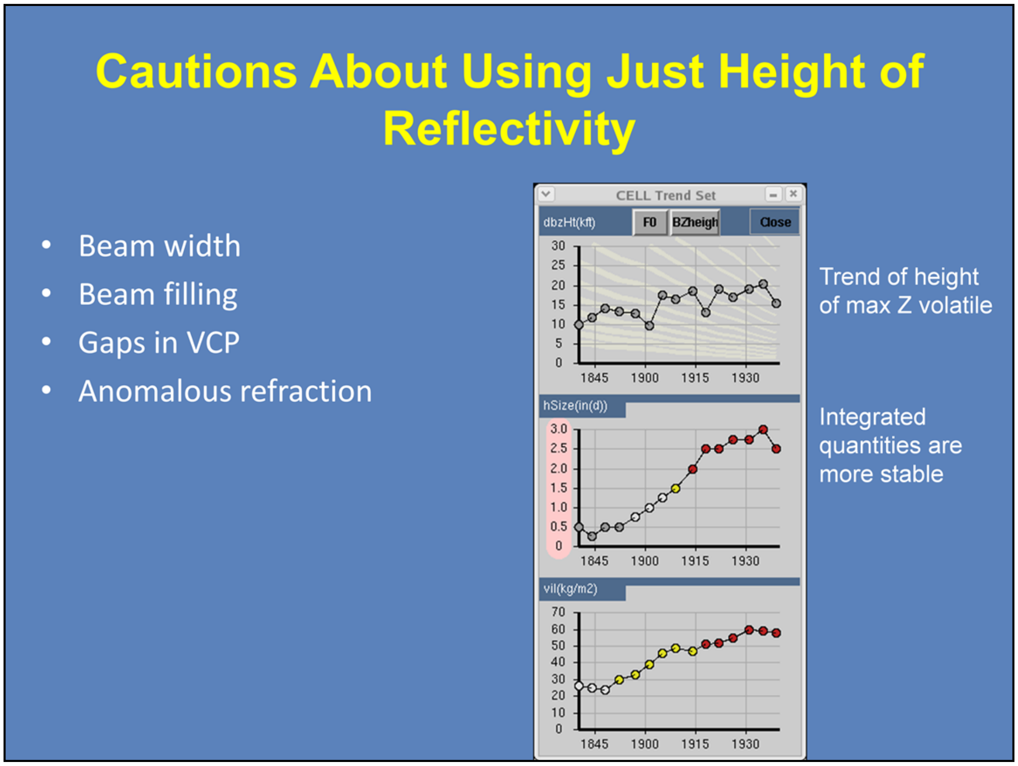
This figure is a good example of a reflectivity signature of 40-50 dBZ surpassing the -20 ° C. Both storms were sampled by KFFC at the same stage in their development in the weakly sheared environment. However, note that the maximum reflectivity in the storm in B was at a higher altitude. It went on to produce a severe downburst in Atlanta while no severe reports were received from the storm in A.

So these examples point to what has been found before. **The intensity and altitude of the elevated core both increase as an updraft's intensity Increases**

As updraft intensity increases, the likelihood for intense downdrafts and large hail also increases given the same environment.

In fact, Cerniglia and Snyder (2002) noted that as the 55 dBZ reflectivity reaches higher altitudes, the False Alarm Ratio (FAR) decreases for some types of severe reports (wind or hail). Another study (Gerard, 1998) examined 64 storms from either the Jackson, MS CWA or the Cleveland, OH CWA, and found that those storms with 65 dBZ above the 0° C level were severe 96% of the time. At the time of these studies, the severe hail criterion was 0.75" in diameter.

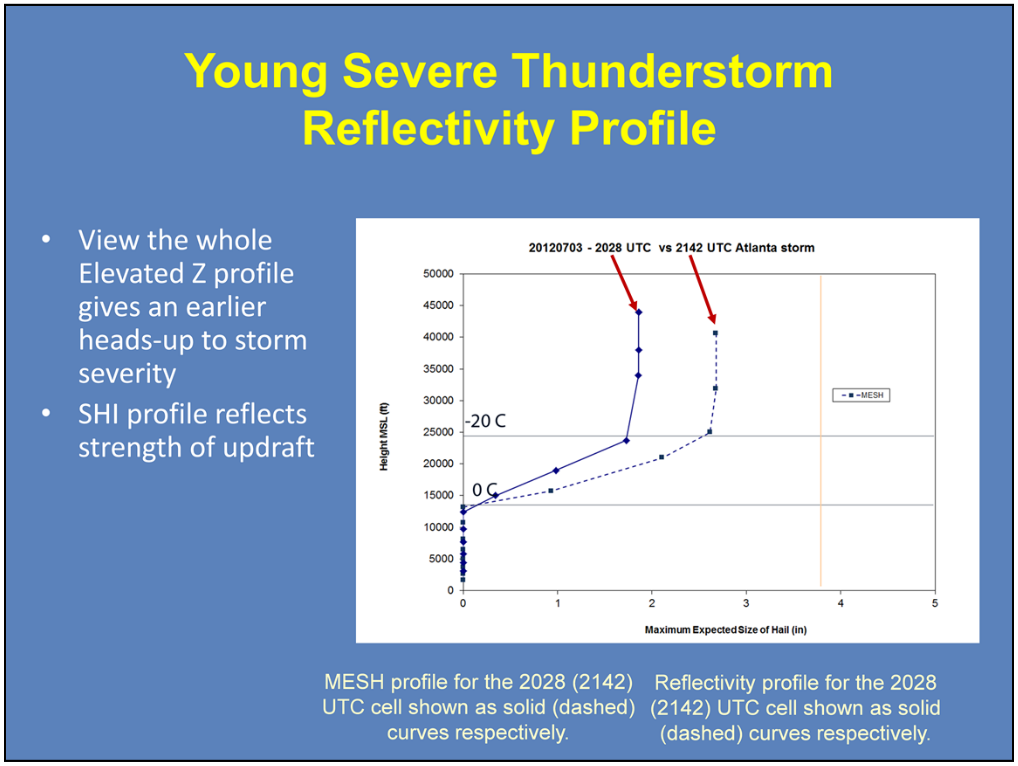
The next slide is a video overview of the storm shown in the right panel.



Though with any single height level radar observation, single height reflectivity guidance suffers from a rapid decrease in accuracy with increasing range owing to beam width, beam filling, gaps in VCP, and refractive index. Please be careful when you consider these radar-calculated heights in this lesson or in scientific literature. You will likely have more success comparing heights for storms at the similar ranges.

One way to reduce the impact of radar sampling limitations on estimating updraft intensity is to vertically integrate the vertical reflectivity profile of a storm from the freezing level to the top. Vertically integrated reflectivity is more resistant to changes in sampling from one volume scan to the next. Recall from the products topic that the Hail Detection Algorithm actually accomplishes such a task in its calculations (see Witt et al. 1998). The Severe Hail Index (SHI) is essentially a vertical integration of strong reflectivities above the freezing level.

This trend set shows how volatile the height of the maximum reflectivity in the storm can be relative to the VIL and maximum expected hail size (both integrated quantities). Take note that VIL and VIL density vertically integrates reflectivity through the whole storm depth, and so part of that reflectivity may be occupied by downdraft.



Contrast the reflectivity profile in this figure between two ordinary cells in their updraft-dominant phases. The dashed reflectivity profile of the cell that initiated at 2142 UTC showed higher values above the freezing level and lower values closer to the ground compared to the earlier storm at 2028 UTC.

The highest reflectivity for both storms was located above the freezing level. However the 2142 UTC cell had stronger reflectivities in the 0 to - 20° C layer. In fact, they helped to boost the Maximum Expected Size of Hail (MESH) much higher than the 2028 UTC cell. While these values were greatly overestimated for a weakly sheared and short-lived cell, the differences in the MESH helped to highlight the much stronger updraft in the 2142 UTC cell.

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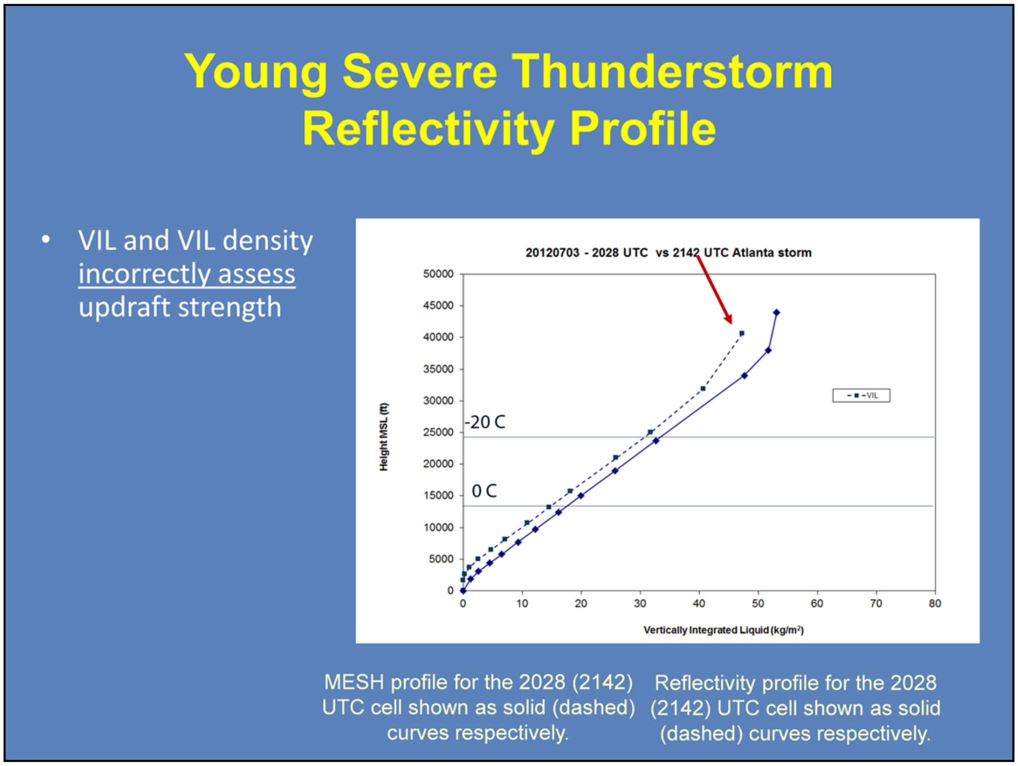
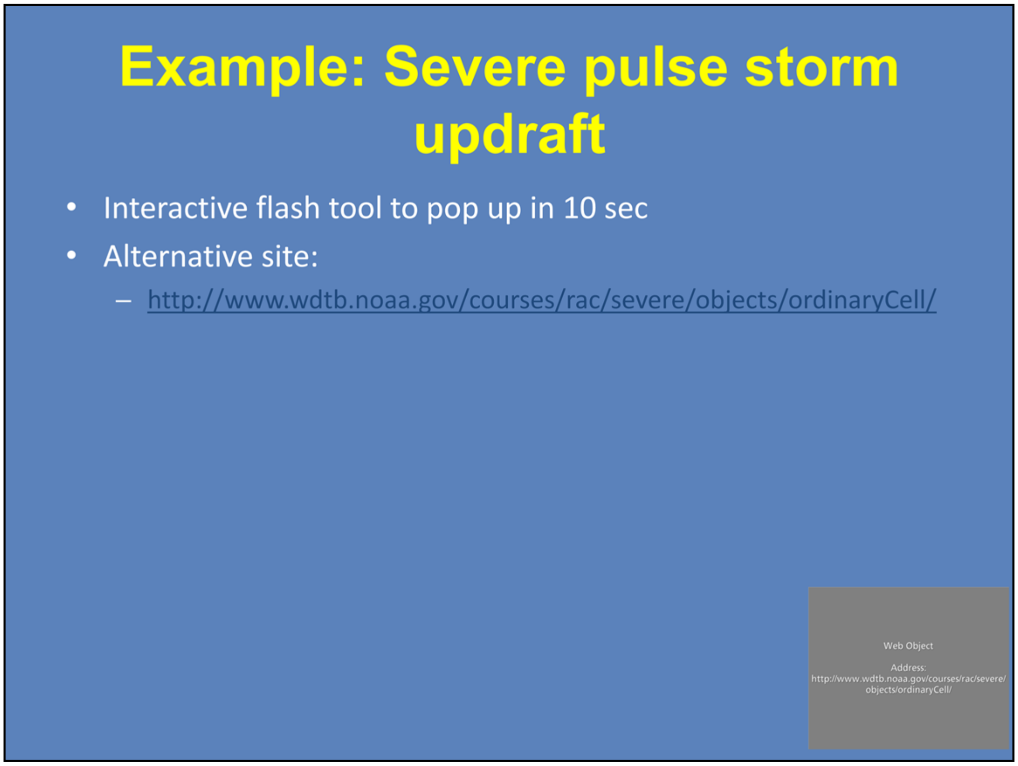


The VIL failed to show a significant difference in value between these two cells in their updraft dominant phase. In fact, the cell with the weaker updraft had a slightly higher value of 51 kg/m^2. This was because the 2028 UTC scan of the cell showed more reflectivity at low levels. VIL density also showed higher values for the 2028 UTC cell. VIL is a great tool for showing the cells with the deepest and most intense reflectivities but it is not so great to evaluate which storm may have the strongest updraft.



In 10 seconds an interactive flash loop of the Atlanta, GA severe downburst producing pulse storm will appear. This is the same storm you saw in the last example but the loop starts at 2128 UTC.

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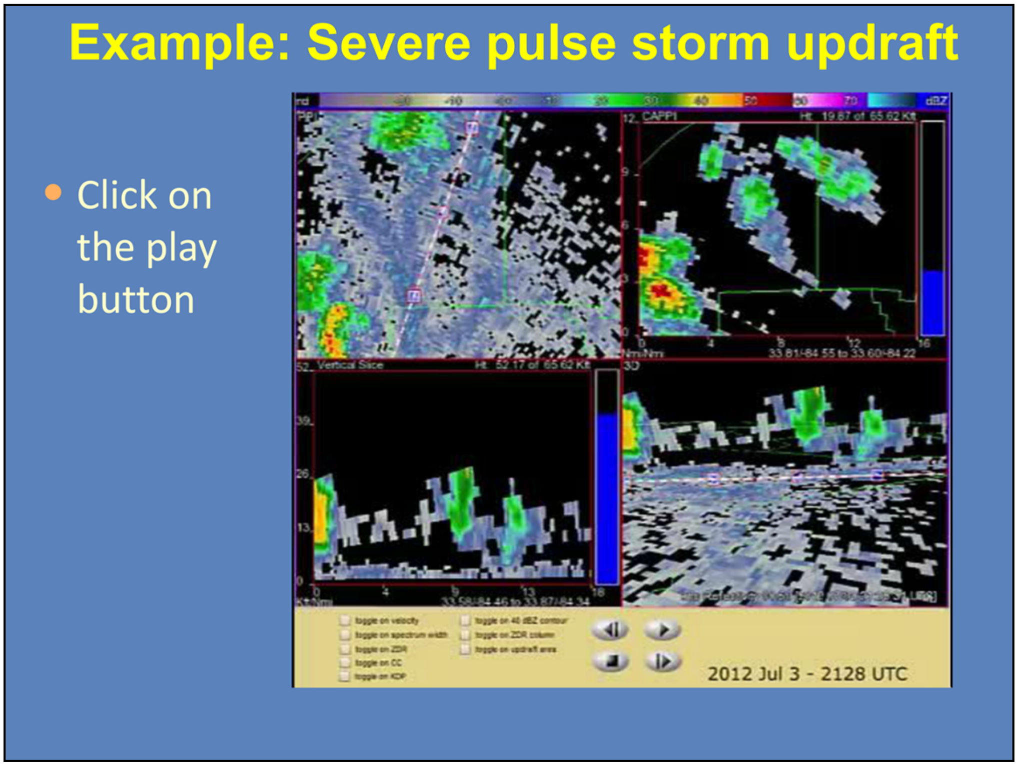

I have an FSI loop of the Atlanta, GA storm where in the upper left panel lies the lowest elevation PPI, and the upper right shows the CAPPI at relevant altitudes. For reflectivity that's at about 20 kft or some just below -20 deg C. In the lower left is a cross-section and the lower right is the 3-D-like display with a cross section in the vertical and PPI in the horizontal.

Starting at 2128 UTC, the first sign of the significant updraft shows up as an area of 20-30 dBZ echoes starting around 13 kft (3 deg C) to above 26 kft (<-20 deg C). The CAPPI shows it nicely too. The lowest PPI shows two boundaries that just collided.

Clicking on velocity shows rather unimpressive convergence at the collision point but otherwise some faint hint of cloud top divergence in the cross-section.

If we were there we'd see an impressive towering cumulus cloud transitioning to a CB. The cell is all updraft.

Going to 2133 UTC, the reflectivity goes from around 30 dBZ to > 50 dBZ in just 5 minutes! The base of the core begins to descend to 13 kft and the top is brought upward to 35 kft. The PPI shows nothing yet but see how the CAPPI snags the core at near 20 kft. That's why the CAPPI is so useful.

Velocity is again weak in the PPI and even in the CAPPI. I'm looking for signs of a downdraft to form. I don't quite see one yet as there is no elevated convergence. But look at the storm top divergence! We know the anvil is spreading out fast. And since there's no shear, the updraft is occupying the whole core. But things are about to change.

Five minutes later at 2138 UTC, the main body of the reflectivity core still hasn't reached the surface but the echoe has exceeded 65 dBZ up to 26 kft. It takes one intense updraft to develop such a strong echo. For a weakly sheared case, this updraft is powerful. No doubt there's significant hail and the subsequent downdraft will likely be severe.

Velocity now begins to show an area of inbound on the far edge of the precip core centered around 14 kft, where the CAPPI is located. That's the beginning of the downdraft and now the updraft area is likely pushed to the south and up above the mid-altitude radial convergence, all the way to storm top.

Going to 2142 -2147 UTC, the downdraft rushes to the ground but the updraft continues its ascent and diverges more quickly at anvil level. Notice the storm top divergence increase even more. But down below the -20 C level, the updraft is likely pushed further to the south or may even be dissipating.

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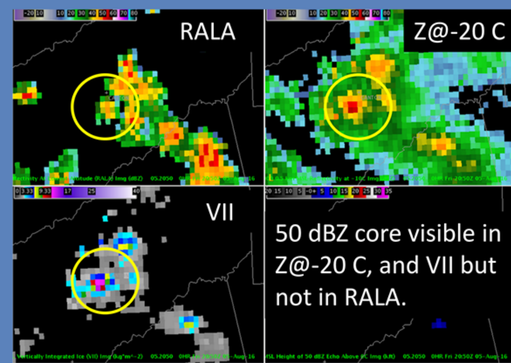
Identifying new cells with Multi-Radar/Multi-Sensor (MRMS)

- These MRMS products depict developing cores aloft collocated with new updrafts.
  - Reflectivity at -20 C (Z@-20 C)*
  - Reflectivity at -10 C (Z@-10 C)*
  - Vertically Integrated Ice (VII)*
  - 30 dBZ Echo Top*
  - 18 dBZ Echo Top*
  - Height of 50 dBZ Echo above 0 C
- Compare these products with RALA to identify onset of updrafts with new deep, moist convection.



* Products with the most lead time to first severe weather report

Updrafts of new cells can be identified using the Multi-Radar/Multi-Sensor (MRMS) suite most readily using the isothermal reflectivity products, (e.g., Reflectivity at -20° and -10° C), Vertically Integrated Ice, the echo top products for 18 and 30 dBZ, and the height of the 50 dBZ echo top above 0° C. However waiting to identify a new updraft using the 50 dBZ echo top above 0° C may reduce your lead time to first lightning or severe weather report.

In the 4-panel display to the right, a new updraft created an elevated precipitation core intense enough to appear as a 50 dBZ echo in the Reflectivity at -20° C and 5 kg/m2 in VII before significant reflectivity appeared in the RALA product. Looping the MRMS helps a forecaster to quickly identify new updrafts using this display. Thus use these products with the RALA to identify new updrafts associated with deep, moist convection.

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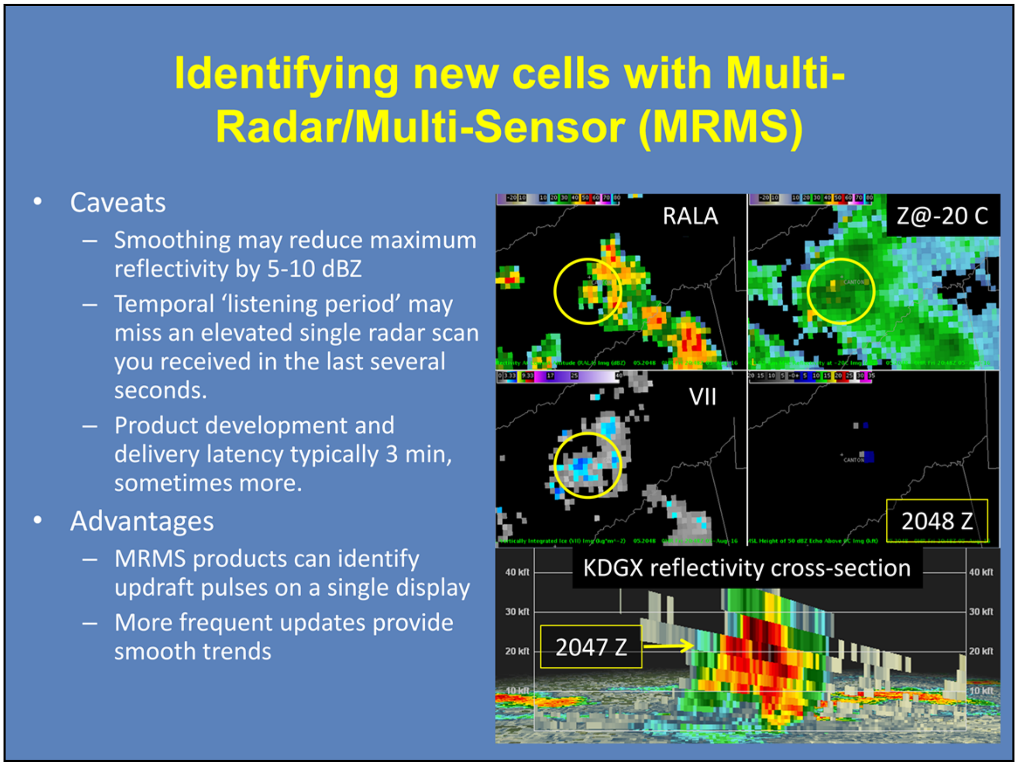

There are some caveats to using MRMS in identifying new cells. The reflectivity analysis smooths the highest peaks by sometimes up to 10 dBZ thus requiring a mental adjustments to thresholds you have set. Also, a new scan from a single radar may show you development of an elevated precipitation core but this new data may have missed the ‘listening period’ of the MRMS with the same time label as the single radar elevation scan. I illustrate this point here with a reflectivity cross-section from KDGX showing an elevated core where reflectivity exceeds 50 dBZ for the 2047 Z scan that cuts along the -20° C level. This data appeared in the 2050 Z MRMS reflectivity at -20° C but not at 2048 Z. Add a typical 3 minute latency, and it may be up to six minutes before you see the onset of this core. Thus there is one more reason to reduce your reflectivity threshold for identifying new updrafts.

Given these caveats, when you need a single display for identifying new thunderstorm updrafts, MRMS is the go-to-product suite. And the frequent updating means you can establish a trend more quickly, even if you have to reduce your thresholds.

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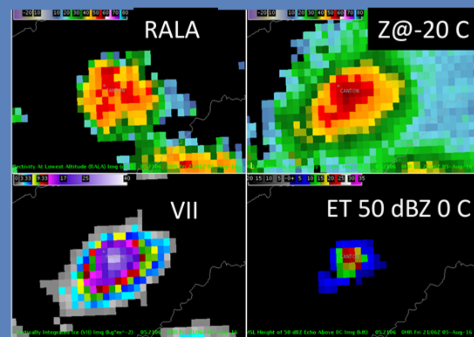
Multi-Radar/Multi-Sensor (MRMS) signatures example: mature phase

- These MRMS products depict developing cores aloft collocated with mature updrafts.
  - Reflectivity at -20 C (Z@-20 C)*
  - ~~Reflectivity at -10 C (Z@-10 C)*~~
  - Vertically Integrated Ice (VII)*
  - 50 dBZ Echo Top*
  - 30 dBZ Echo Top*
  - 18 dBZ Echo Top*
  - Height of 50 dBZ Echo above 0 C*
  - Height of 50 dBZ Echo above -20 C*



For mature thunderstorms, the split updraft/downdraft pattern means that the importance of the higher level products increases in locating updrafts in MRMS products. Now the reflectivity at -10° C may be occupied by downdraft so scratch this product. Add the echo top of the 50 dBZ echo above 0° and -20° C.

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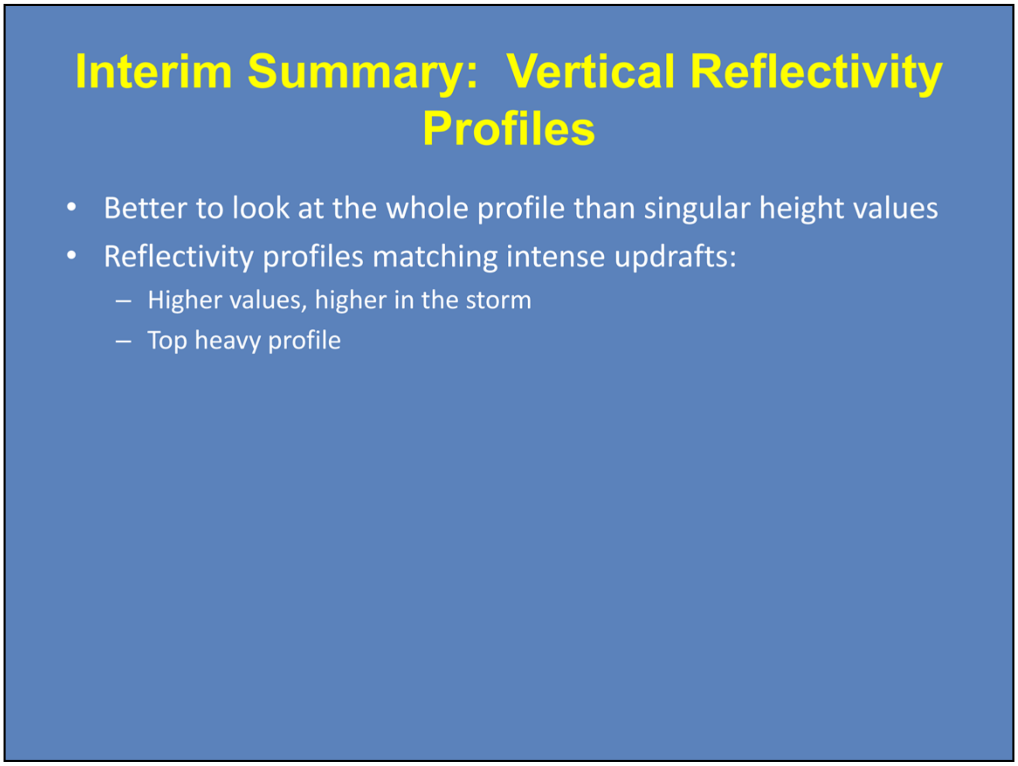

To summarize, the radar beam height uncertainties are just too great to allow you to hinge your warning decision on a single height threshold of reflectivity (e.g., height of the 55 dBZ echo). Instead, determine the shape and intensity of the reflectivity profile as it extends into the subfreezing layer in a storm as it's intensifying. Storms with a the most top heavy reflectivity profile (e.g., highest reflectivity at the highest levels) at this early stage are more likely to be severe than those with bottom heavy profiles. Additionally, storms with a top heavy reflectivity profile signifies the production of a core with large hail aloft before it descends to the surface.

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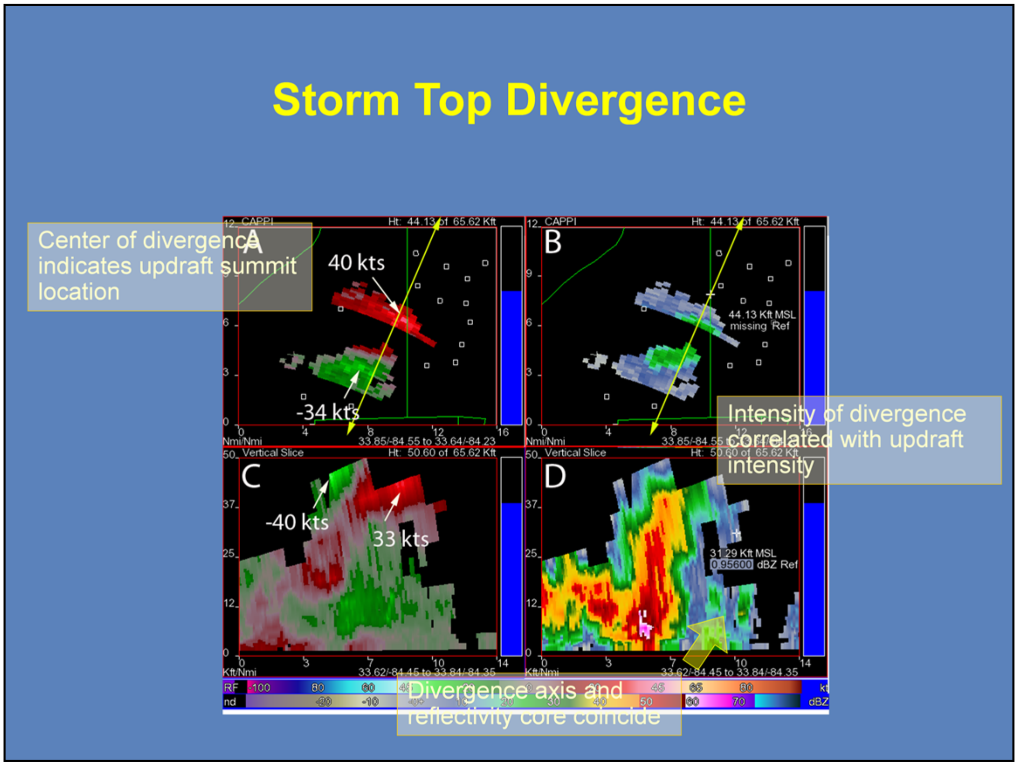
Velocity Signatures for Updraft Intensity Estimation

- All storms exhibit low-level convergence and upper-level divergence
- Can the WSR-88D detect this?
  - Many, but not all, times, yes

All convective storms exhibit some amount of low-level convergence as air enters the updraft, and upper-level divergence near the equilibrium level as air diverges from the updraft. For severe storms, the updraft intensity is likely to be higher and so is the intensity of the convergence and divergence signatures. The ability of the WSR-88D to detect these signatures depends on how well the total convergence and divergence velocity patterns are reflected in the radial velocity-only patterns. In many cases, the WSR-88D is able to adequately sample convergence and divergence affording you another tool to evaluate updraft intensity.



Radar base velocity or storm-relative velocity data show a divergent flow pattern at the storm summit once the equilibrium level has been reached and an anvil begins to form. **The center of the divergence indicates the updraft summit location. The intensity of the divergence is positively correlated to the intensity of the updraft** (Witt and Nelson, 1990). The maximum inbounds and outbounds can be quite strong, exceeding 50 kts in both directions in the stronger storms. Note that the divergence axis and the reflectivity core roughly coincide. This storm top divergence (ΔV) was about 80 kts as determined from sampling the maximum and minimum velocities found around the overshooting top sampled by lifting and dropping the CAPPI and moving the vertical cross section east and west in FSI. What you see here is about 74 kts labeled.

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Storm Top Divergence Sampling Issues

- Shallow Divergence not sampled
  - Minimize with VCP 11, 211, 12, 212
  - Choose radar within 80 nm range
- Asymmetric storm top outflow results in differing radial divergence
- Individual cell divergence lost in multicells
- Don’t use very high elevation angles
  - Interior range limitation
- Choose higher velocity increment

There are sampling issues that may inhibit you from detecting the full divergence that a storm is generating:

Shallow divergence may be missed by wide beams or VCP gaps. Beamwidths and/or gaps should be less than 2 km. To reduce its impact make sure you are using one of the 14 elevation angle VCPs (11, 211, 12, 212). Make sure you are sampling the storm inside of about 80 nm (120 km).

The storm top divergence, or outflow pattern may not be symmetric about the updraft summit. Thus radial divergence may change according to viewing angle. Try using an alternate radar for a better angle.

Individual storm top divergence signatures within large multicells may be difficult to detect.

Be careful about interpreting storm top divergence at very high elevation angles. Each individual beam may not sample both the minimum and maximum velocities. Try using the FSI CAPPI or a more distant radar.

Some anvils produce greater than 123 kts velocity. You will need to change the velocity increment to 2kts (1m/s) to adequately measure anvil divergence in these cases.

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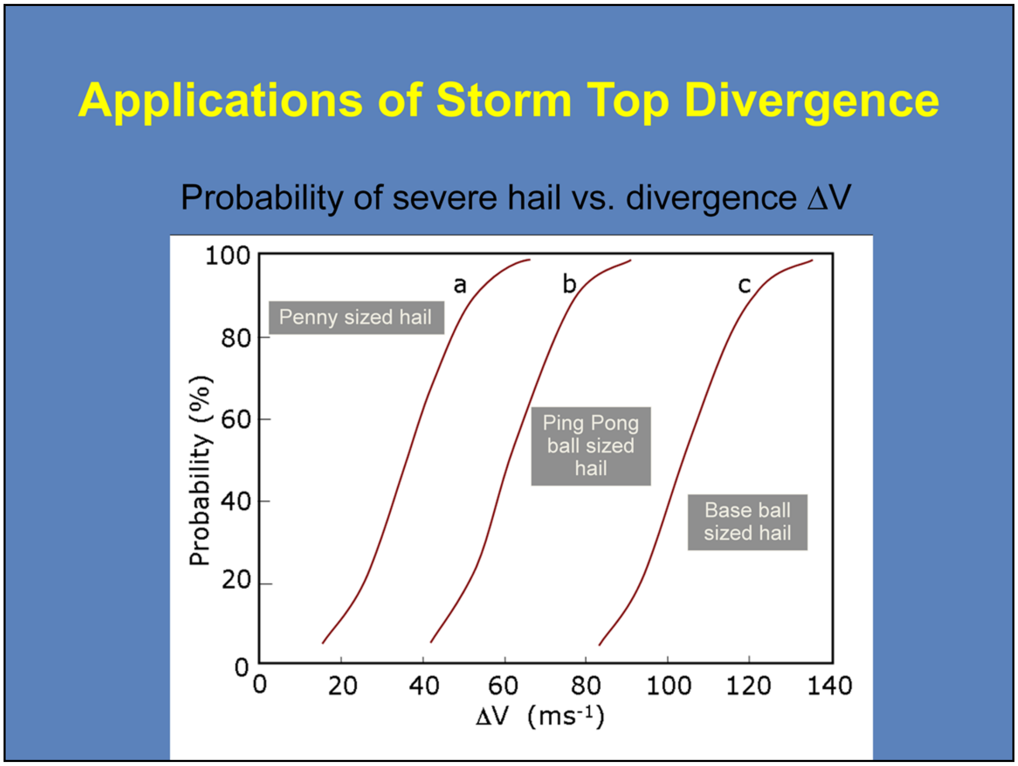

Updraft intensity estimations have been most closely tied to estimating hail size. Witt and Nelson, 1990 derived a useful correlation between maximum storm top divergence and probabilities of maximum hail size shown in this Figure . As a matter of caution, this graphic does not take into account the diversity of environments that you may face when estimating hail size.

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Low-level Convergence

- Weakest with first storm
- Stronger with cold pool lifting for following storms
- Colliding gust fronts create strongest lift

In addition to storm top divergence, you can attempt to detect the air converging into an updraft. During the initial stage of the first surface-based cell of the day, there may be a weak radial convergence feature within the lowest two kilometers of the ground as air flows in to feed the updraft. **Maximum radial velocities typically are very small for the first nonsevere cells of the day.** After cold pools develop, new surface-based updrafts receive much stronger initial baroclinic forcing along their edges leading to stronger updrafts.

Colliding cold pools, or cold pool interactions with other boundaries can generate vertical velocities exceeding 25 kts (12 m/s) within a few km above the surface. That's enough vertical velocity to generate graupel and split electric charges if the temperatures were cold enough.

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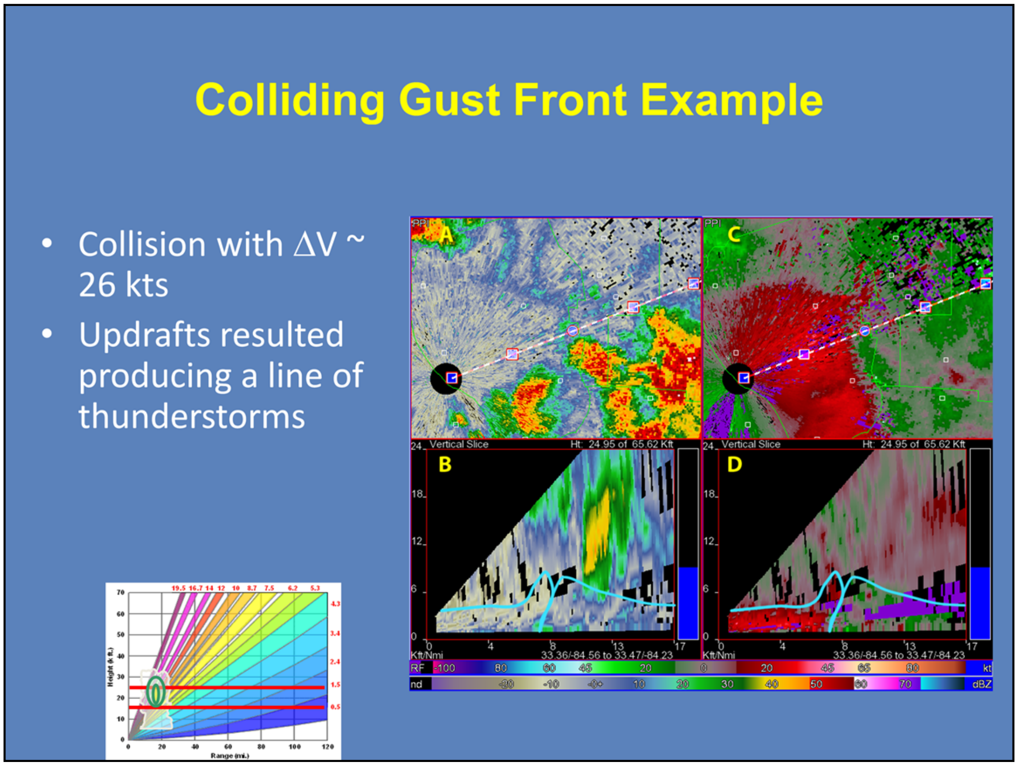

An example of colliding gust fronts initiating strong, but low-shear convection, is shown in this loop from just southeast of Atlanta GA. These gust fronts resulted in a rapid initiation of a line of broken ordinary cells, one of which we have seen before in this lesson.. The gust fronts collision resulted in approximately a ΔV ~ 20 kts over a 1 km distance.

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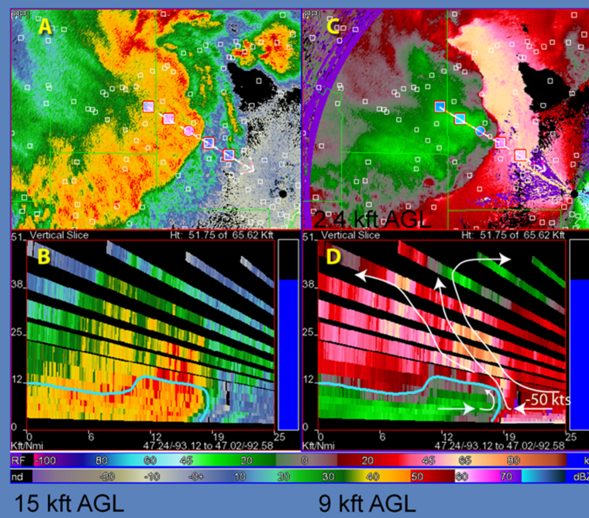
Depth of Convergence is Important

- Deep convergence zones lead to strong updrafts
- Convergence was > 9 kft for the severe bow echo on right
- 30 m/s updraft @ 3km AGL



For stronger storms, stronger low-level convergence is much more likely to develop as cold pool boundaries tend to be stronger and deeper. **It is not just the magnitude of the convergence but also the depth that is important to the strength of the updraft**, especially in the lower half of a storm. In the most severe multicell events, a cold pool leading edge may be up to 5 km (15kft) deep with more than 50 kts of velocity difference, leading to updraft strengths exceeding 70kts.

An example of a relatively deep convergence zone is highlighted in this image of a bow echo approaching Duluth. The gust front maintained strong convergence through a depth of at least 10 kft (3 km) with a DV ~ 20 kts across a boundary of 1 km wide (e.g., convergence > 0.01 s-1). By continuity, an updraft exceeding 30 m/s would occur at the top of the gust front's vertical interface.

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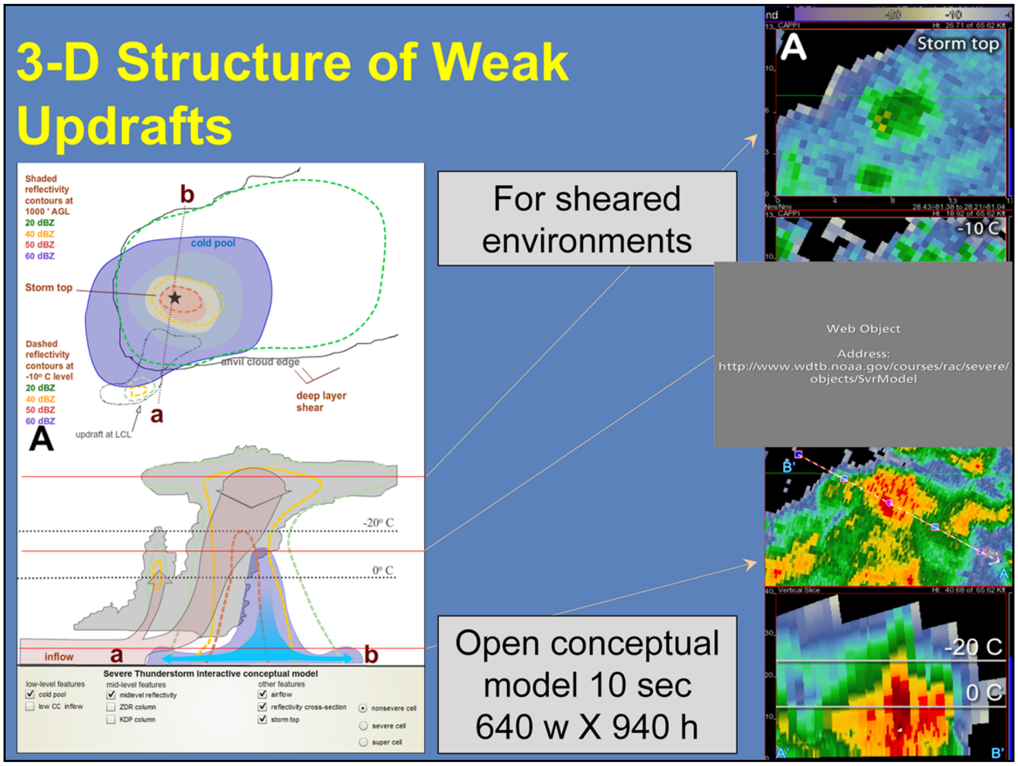

Before analyzing specific severe weather threats (e.g., hail and wind), it is important to take advantage of the volumetric radar data available to analyze the three-dimensional nature of storms. Lemon (1980), derived a methodology for volumetric discrimination between non-severe and severe convection. The technique focuses on the three-dimensional distribution of reflectivity through low-, mid- and upper-level indicators. Since updraft strength is an important controlling factor in severe weather, analyzing the 3-D shape of the reflectivity core is important. The following conceptual model is intended to represent convection as it evolves from nonsevere to severe modes in significant vertical shear environments. About a minute in, an external browser window will appear with an interactive version of this model. Follow along. This slide will stop at the end. The conceptual drawing on the left shows a horizontal planview on top and a vertical cross section on the bottom of a relatively weak updraft storm in a sheared environment. This storm can represent the onset of a severe storm to be, or be a mature storm that fails to utilize the instability and shear to its full potential. On the right, you are seeing a real example of a nonsevere storm as visualized in FSI.   The cross section is displayed exactly as the conceptual drawing shows on the left with point A (B) facing toward (away) from the low-level inflow ingesting into the updraft. The shaded reflectivity shield in the conceptual drawing corresponds the low-level reflectivity shield and matches roughly to the scan second from the bottom on the right. The dashed reflectivity contours in the planview part of the schematic corresponds to the reflectivity map third from the bottom on the right and is roughly where the -20° C level should be located. The storm top reflectivity echo lies at the top of the figure on the right.

A weak updraft in a sheared environment slopes downwind and is typically unable to suspend any precipitation. Convergence at low-levels and the corresponding storm top divergence is relatively weak. Severe weather possibilities are relatively low with this kind of structure. An example of nonsevere convection in a sheared environment shows the least reflectivity overhang as the echo top lies mostly on top of the low-level reflectivity core. What overhang exists is mostly an artifact caused by the storm motion as the radar antenna ascended in elevation scans. Another example appears from KS on a high risk day in 2012 April 14 which shows similar structure.

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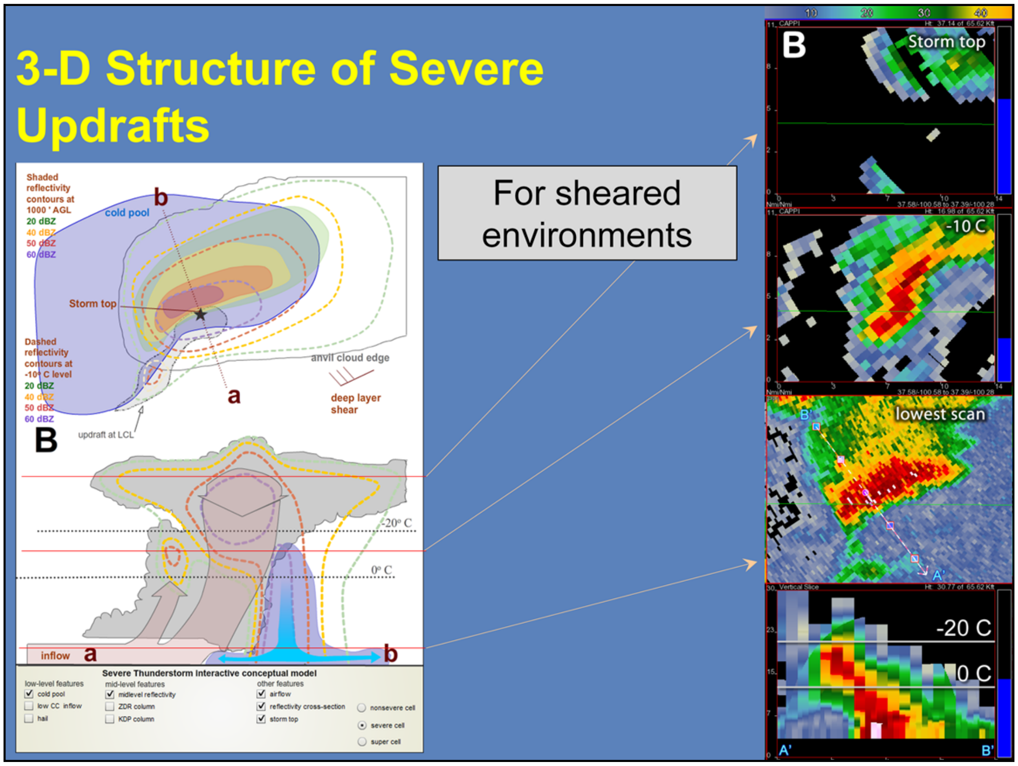

As updraft strength increases, it becomes more erect and is able to suspend a heavy core of precipitation resulting in the Weak Echo Region (WER) (Fig. b). In this **Figure** , the reflectivity core shows obvious overhang above the WER in a direction facing the low-level storm-relative inflow and where the low-level core boundary exhibits concavity and a tight gradient. The echo top extends over the low-level reflectivity gradient next to the WER. The storm is more capable of producing severe weather. Large hail is quite likely when much of the echo overhang is above the freezing level.

Another example of a cell on 2012 April 14 in KS shows a similar structure of an echo overhanging the low-level WER on the side facing the low-level storm-relative inflow. The only difference is that the storm top significantly displaced downshear or into the plane of the cross section and so the echo appears somewhat shallow in the cross-section.

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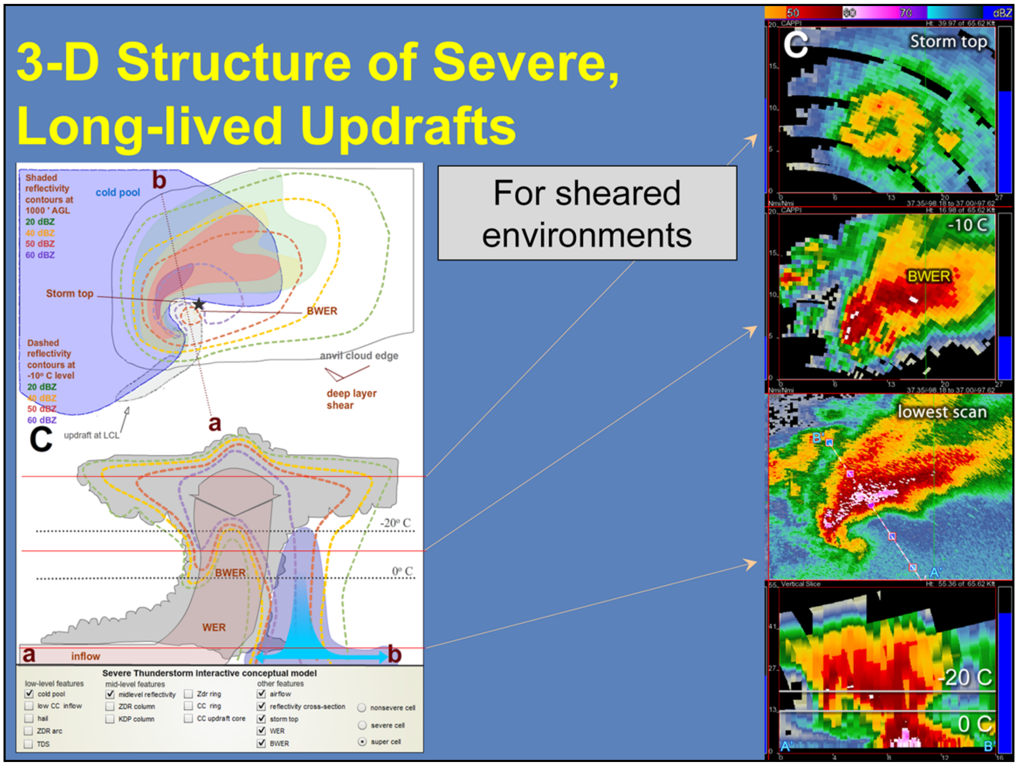

The most intense updrafts are erect and may exhibit a BWER . In this conceptual model, the BWER becomes more pronounced as an upward extension of the WER. The updraft in this storm is most likely stronger than in the non-supercellular severe storm model. The low-level reflectivity gradient facing the WER exhibits more curved concavity while the echo top may extend directly over the low-level WER and BWER. This reflectivity configuration is associated with the strongest updrafts, and the most severe weather reports.

The base of a severe updraft is typically located under the WER and BWER and next to strong reflectivity gradients and inflow notches. The WER is typically much larger than the saturated updraft; much of it exists because of intense reflectivities resident in the overlying anvil and the intense storm summit. You may use the low-level velocity to look for areas of strong convergence. The updraft base is most often rooted in the convergence.

The second example shows a more tornadic supercell where the BWER is a little more difficult to find at -10° C. The tornadogenesis may have helped weaken the updraft at the typical level of the BWER as the low-level rotation intensified and helped forced a rear flank downdraft.

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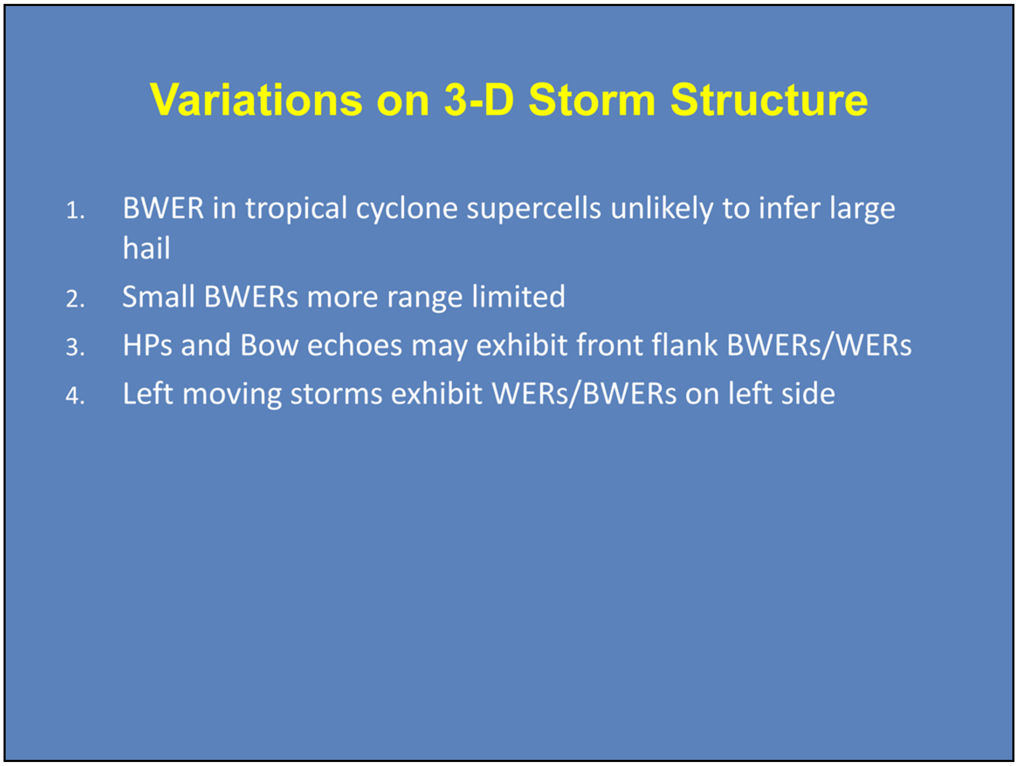

The updraft signatures discussed previously may result in different severe weather types depending on the storm environment. Here are a few examples:

A BWER in a nearly saturated, warm sounding (e.g., a tropical cyclone) is unlikely to infer the presence of large hail because the environment is too warm at the top of the BWER.

An environment with a low equilibrium level supporting mini supercells may indicate to the forecaster that BWERs may be too small to detect beyond a close range.

Many HP supercells and bow echoes may show WERs, and BWERs ahead of the main core with respect to the deep layer shear vector. In other words, these storms have front flank updrafts.

Straight or cyclonically curved hodograph environments favor left-moving storms with updraft signatures to the left front flank of the main core when you face its direction of motion.

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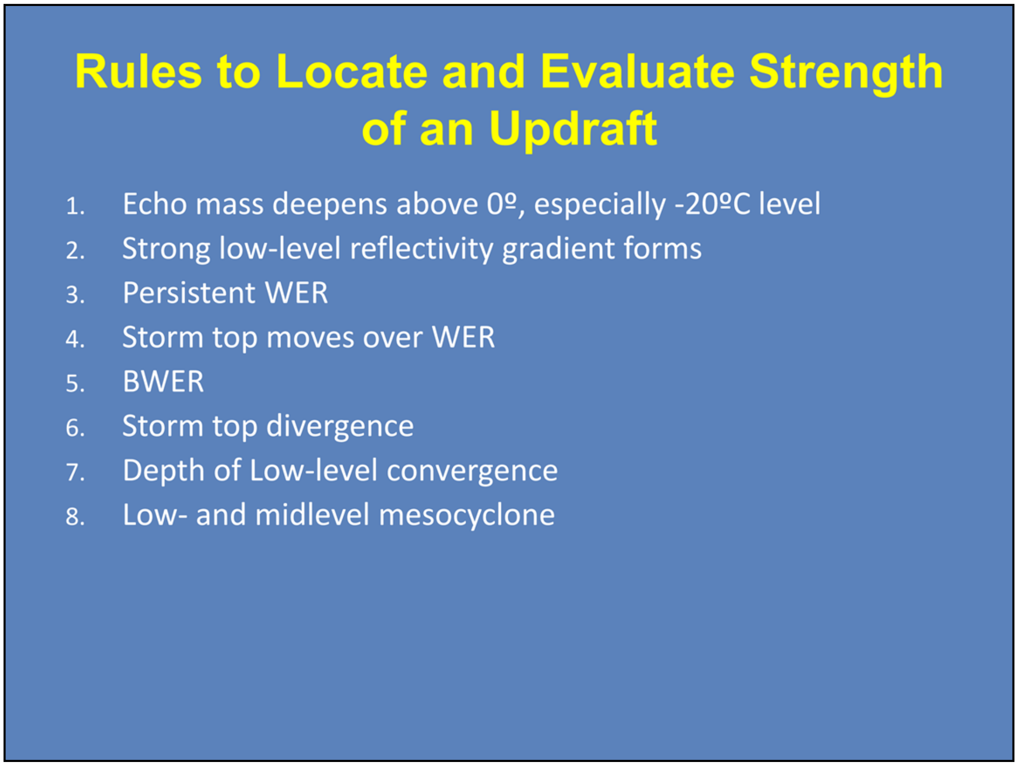
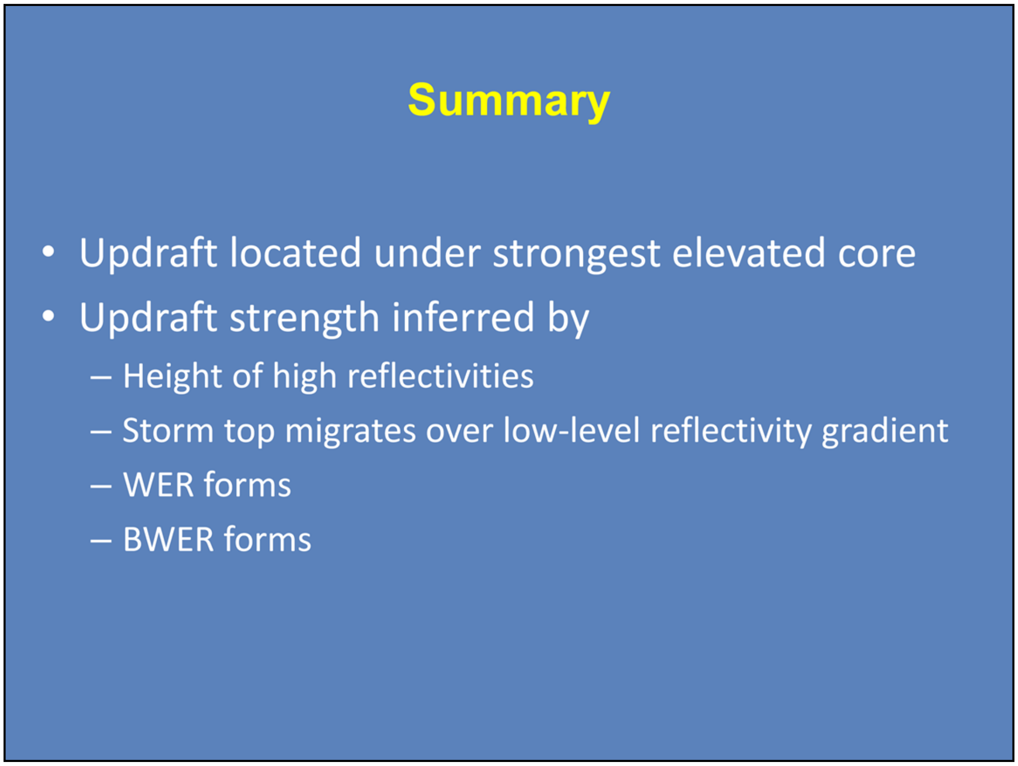


Remember the following rules for locating and evaluating an intense updraft in a sheared environment:

1. Echo mass deepens above the freezing level, especially above the -20° C level.
2. Strong low-level reflectivity gradient develops.
3. A persistent strong echo overhang extends over the low-level, concave reflectivity gradient forming a WER.
4. The storm top moves over the lower level WER.
5. A BWER may form in the stronger updrafts as an upward extension of the WER.
6. Strong storm top divergence becomes strong.
7. Low-level convergence intensifies and becomes deep.
8. Low- to midlevel mesocyclone forms (not all of the mesocyclone is updraft).



For ordinary cell convection, the most generic proxy for locating an updraft is to locate the strongest elevated reflectivity core where it corresponds to temperatures less than -15° C. The higher you locate the core, the more likely it is dominated by updraft. The base of the updraft may begin with its roots in the boundary layer in the young phase of a cell. But over time, the base of the updraft becomes more elevated as intense core forces downdraft to form. As the storm nears its demise, the updraft may only be confined to the anvil layer.

The relative updraft strength can determine the maximum height achieved by the high reflectivities forming in the precipitation core. Generally the higher they form, the more likely there may be severe winds and/or hail.

Other reflectivity-based signatures come into play once the storm becomes more persistent. The more severe storm updrafts including severe multicells and individual supercells, exhibit a tendency for the high level echo core to migrate over a low level weak echo region in the vicinity of a tight reflectivity gradient. Sometimes the tight low-level reflectivity gradient becomes concave as low-level storm-relative inflow increases in response to the intensifying updraft. The size and extent of the Weak Echo Region (WER) increases with increasing severity of the updraft. BWERs form in the strongest updrafts.

Welcome to the lesson on assessing updraft strength and location with polarimetric radar data. This lesson should last approximately 30 minutes.

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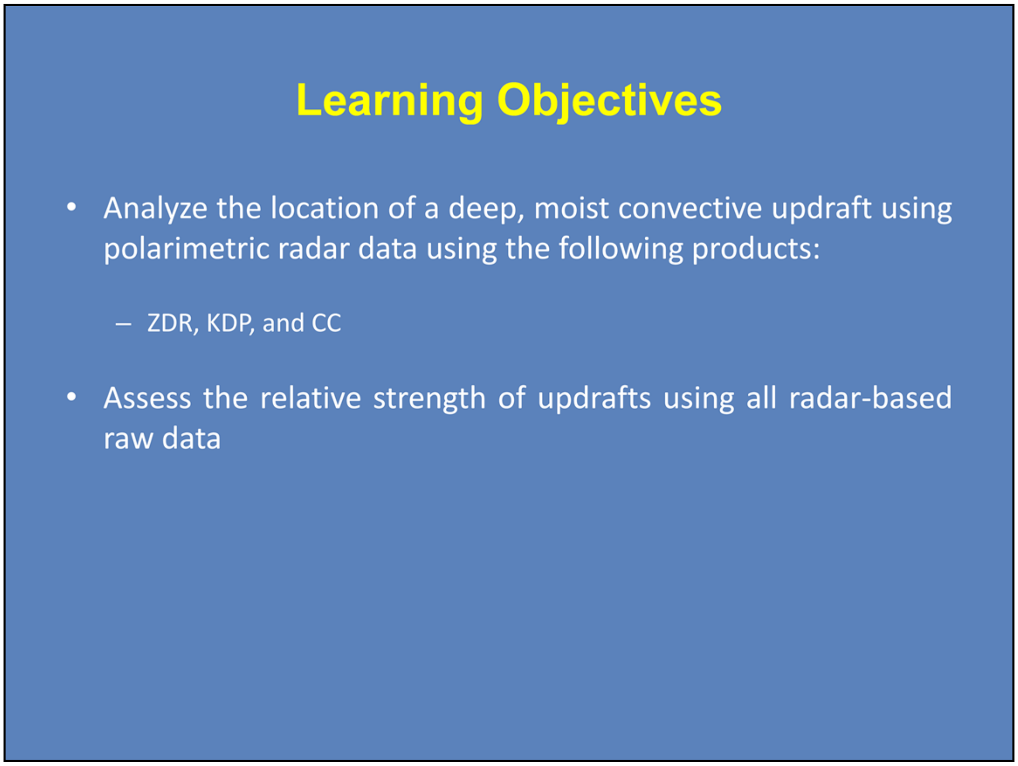

Analyze the location of a deep, moist convective updraft using polarimetric radar data using the following products: Differential Reflectivity (ZDR), Specific Differential Phase (KDP), and Correlation Coefficient (CC).

Assess the relative strength of updrafts using all radar-based raw data.

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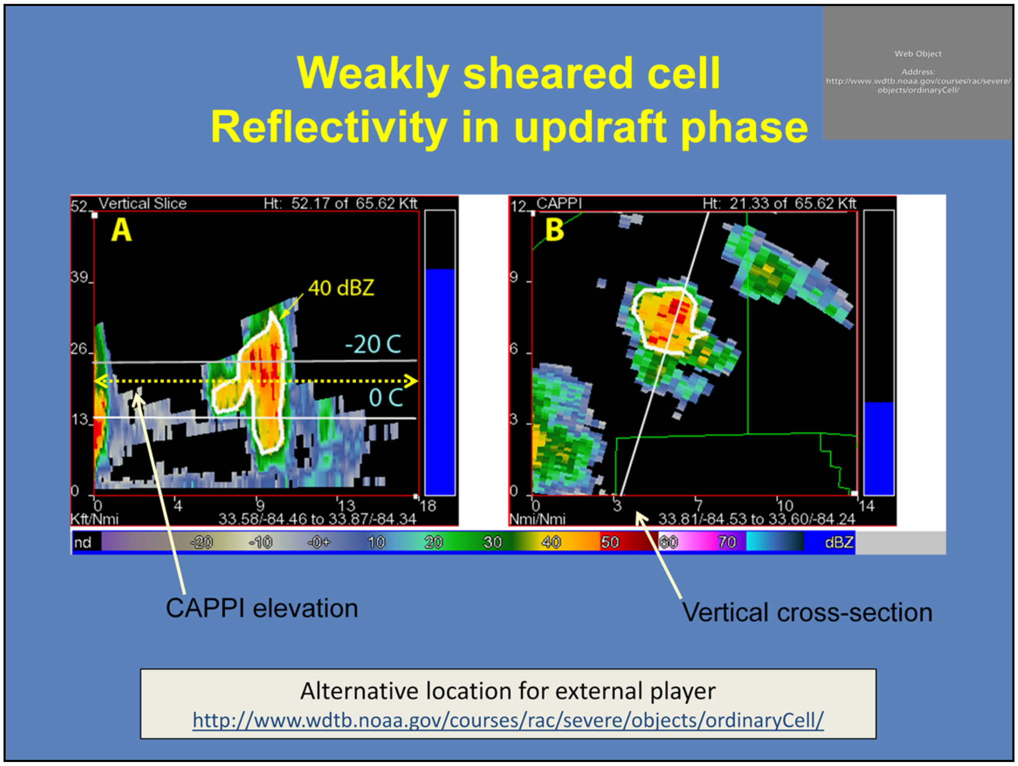
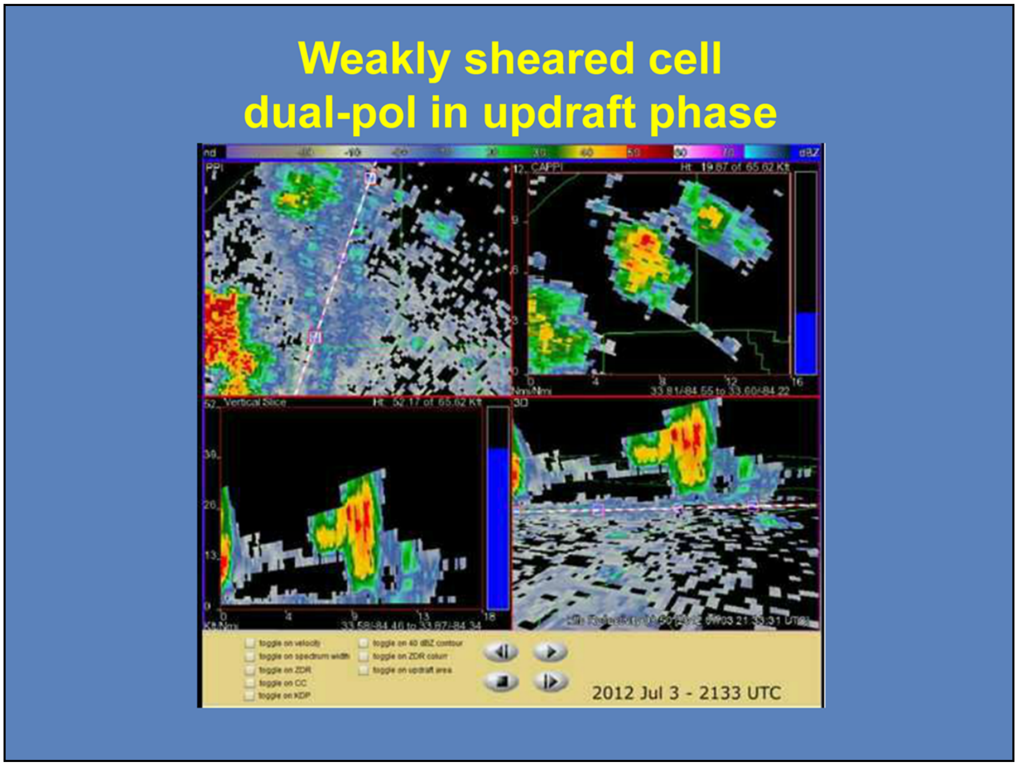


In a few seconds an interactive flash display will appear that you can use to follow along with my discussion on dual-pol updraft signatures of an ordinary cell. It is the same display you saw in an accompanying lesson on convective updraft identification, however there may be slight differences in the CAPPI elevation and vertical cross-section placement.

Capturing updraft signatures in weakly sheared ordinary cells depends on the timing of the radar's volume scans relative to the life cycle stage of the cell in question. The dual-pol signatures are transitory as the updraft bubble ascends toward its equilibrium level.

Let's start with viewing how these signatures **Updraft-dominant Phase** appear as the updraft pulse begins to produce precipitation. The case that we'll use is the 3 July 2012 Atlanta pulse severe thunderstorm case starting at 2133 UTC. This storm is quite typical in its appearance in the dual-pol products.

Keep the interactive display next to the articulate for the video walk-through of what's happening to the dual-pol products at this time in the next slide.



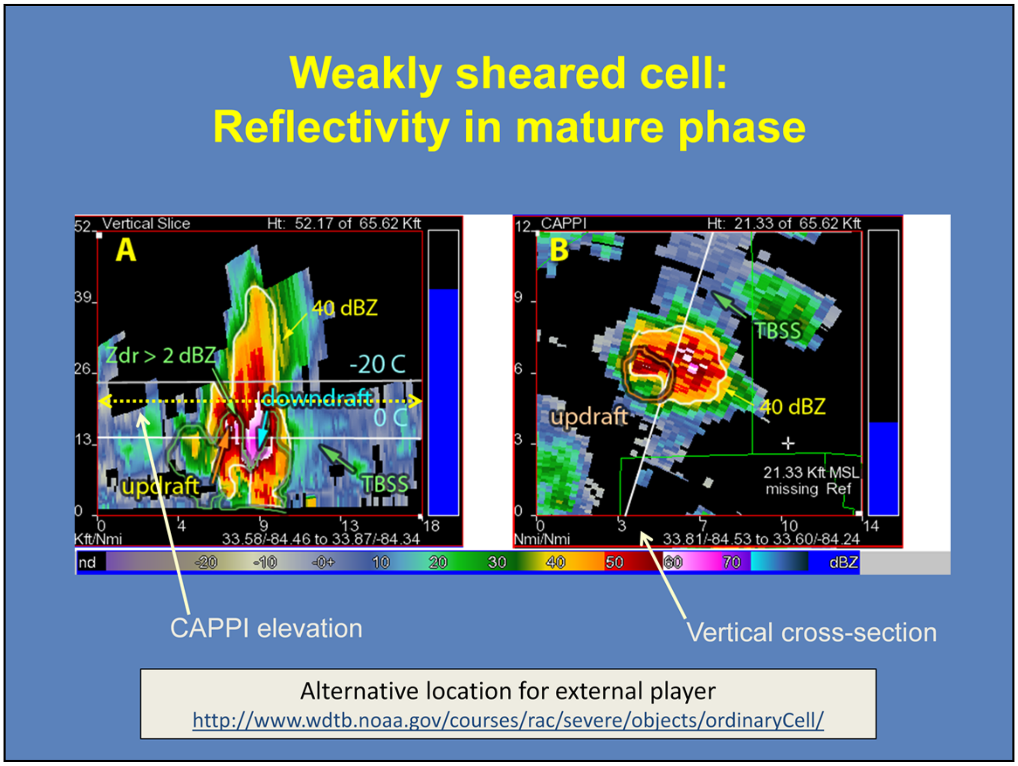
Click on the play button to start the video. You may follow along with the interactive viewer.

As the cell begins to produce precipitation, almost immediately, most of it is located above the freezing level (Figure 7-27A - Note from here on these are figure references in the student guide). Note that the ZDR shows large values in excess of 2 dB in the bottom portion of the reflectivity envelope that exceeds 40 dBZ. This is where large drops most likely dominate. Especially notable is the upward extension of likely large liquid drops above the freezing level (Figure 7-27C). The only way to get liquid drops above the environmental freezing level is through updraft. This is the area where we identify a ZDR column.

The vertical cross section shows a peak altitude of the ZDR column reaching about 18 kft above the radar (ARL), but the CAPPI (Figure 7-27D) shows the ZDR column extends up to its level at 21 kft ARL marked by a small area exceeding 2 dB. The vertical cross section missed the highest extent of the ZDR column. This is the area of strongest updraft within the broader updraft that is developing the echo. Note that the 2 dB value selected here is on the high side of a 1-2 dB threshold to consider the bounds of a ZDR column.

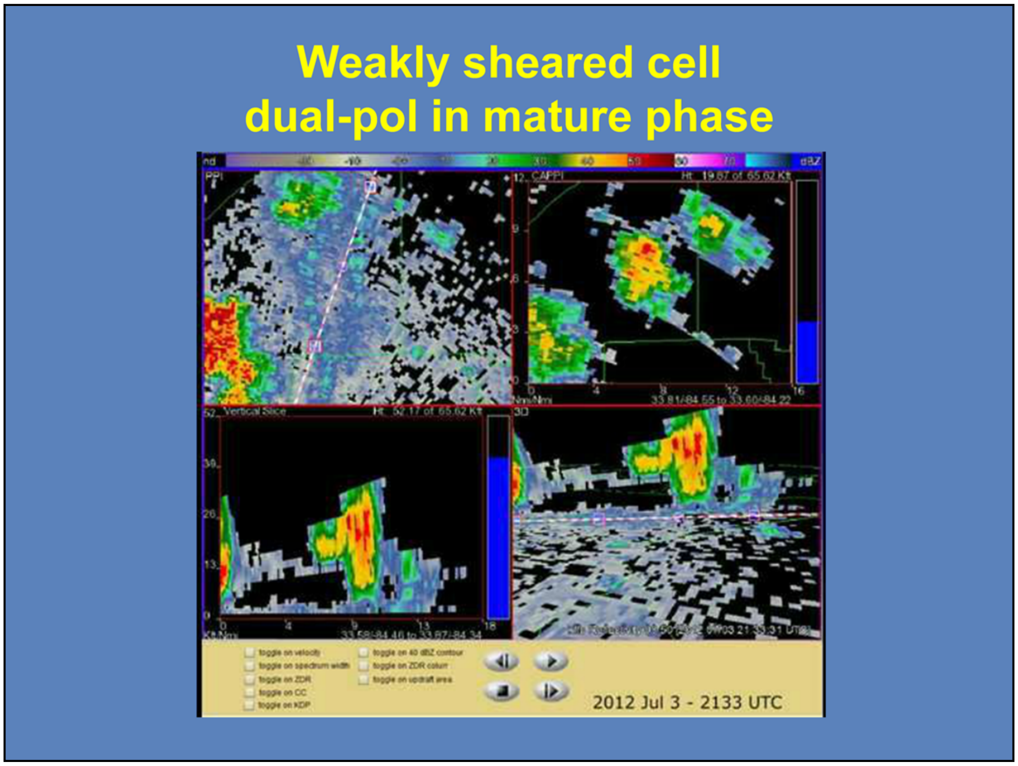
Where ZDR stands out with a large column exceeding 2 dBZ, the KDP remains small with only a small area exceeding 1deg/km near the freezing level (Figure 7-27E). This is because there is likely not much integrated water volume in the updraft. Only widely scattered large drops are lifting above the freezing level in the updraft.

The updraft within the ZDR column top exhibits slightly reduced CC values (Figure 7-27G and Figure 7-27H) that could be the result of a few ice particles forming amongst the large liquid drops. These values remain between 0.9 and 0.95.



Picking the location of an updraft through the depth of a mature ordinary cell is a bit more challenging. Now that the downdraft has begun, not all of the reflectivity core is occupied by updraft. Conventionally, we teach that the strong reflectivity core (> 40 dBZ) above the -20 deg C level is most likely occupied by updraft while that below this level is more than likely downdraft. Now with dual-pol data, we have the ability to better discriminate the location of the updraft.

Let's explore with another video tour in the next page. Meanwhile, keep your interactive graphic handy to follow along. When the next page loads, click the play button.



This video walks you through the dual-pol signatures of a weakly sheared cell in its mature phase. Hit the play button when you're ready.

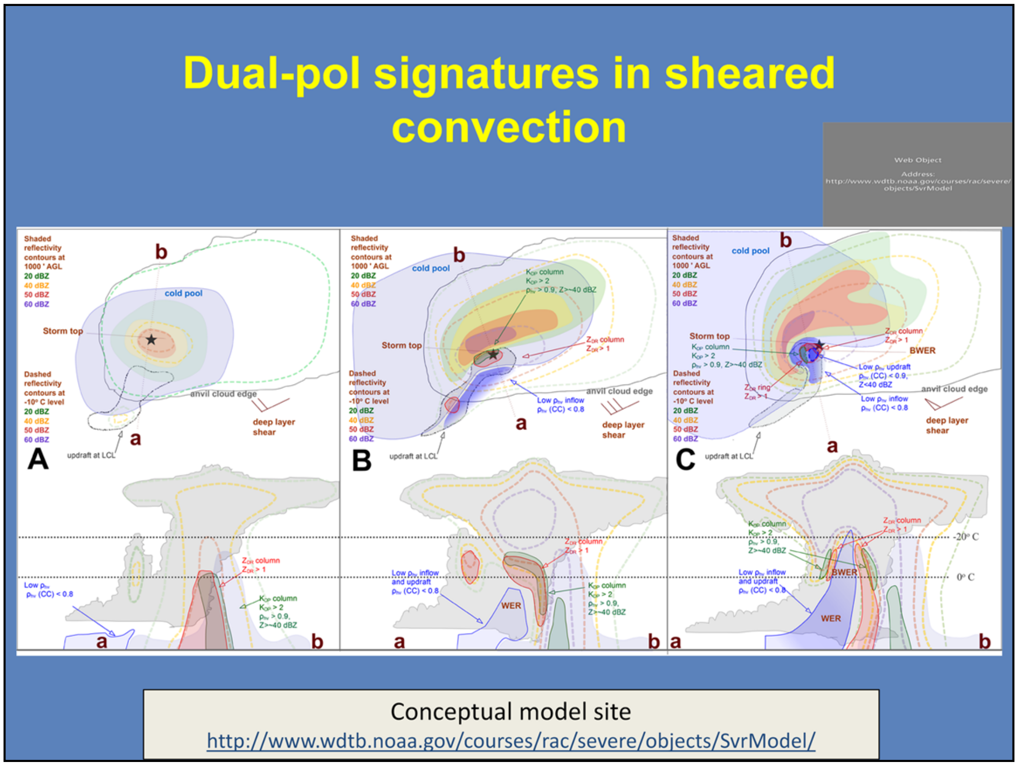
The onset of reflectivities > 55 dBZ above freezing level shown in Figure 7-28A is a strong signal that graupel and hail have formed 10 minutes after initiation (Figure 7-27). The ZDR images in Figure 7-28C and Figure 7-28D shows the depressed values < 1 DB that helps support the idea that the precipitation was dominated by ice. The downward plunge of the low ZDR precipitation core is quite likely associated with downdraft. At this time, the ZDR column has bifurcated somewhat with the primary updraft likely on the southwest flank of the precipitation core. The ZDR column there still reaches up to the 21 kft as the CAPPI in Figure 7-28D highlights.

Note that very high ZDR values exist down-radial from the precipitation core. This is a Three-Body Scatter Spike (TBSS) and not a ZDR column. Always suspect high ZDR values down-radial of a ZDR column when the reflectivity is low.

KDP indicates high concentrations of liquid water along the path of the radar beam. In Figure 7-28E the KDP is high (~2-3 deg/km) in the axis of low ZDR

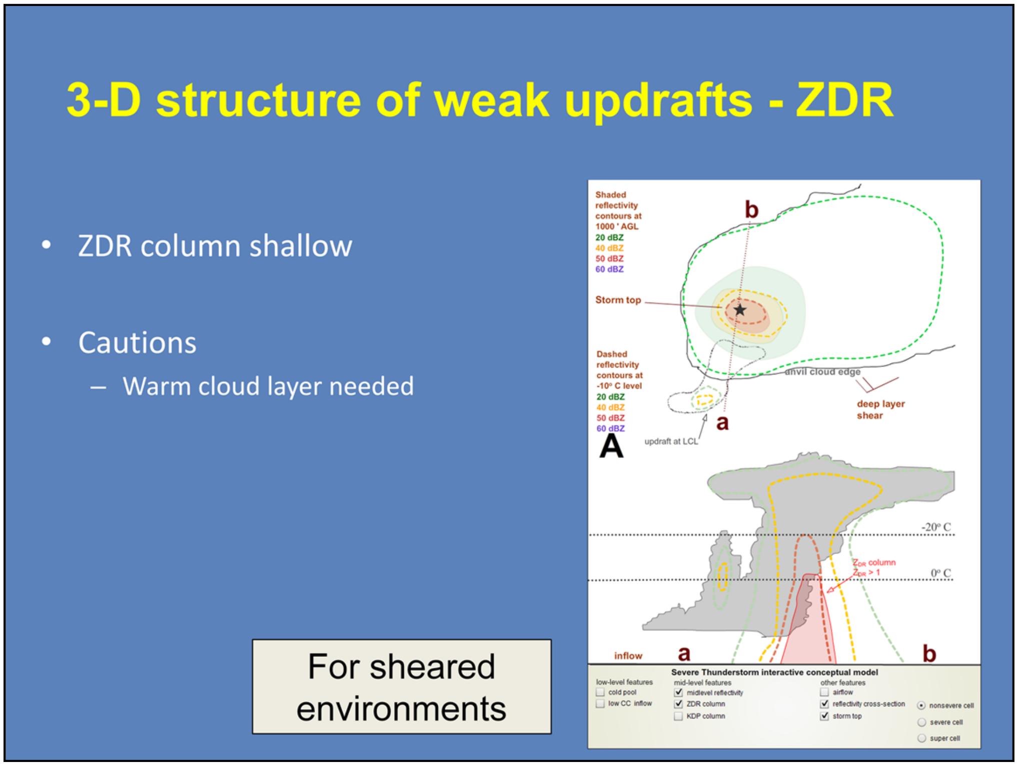
Values from straddling both sides of the freezing level. These high values mean there is a large amount of liquid water amidst the hail and graupel. How did so much liquid water wind up above the freezing level within what we believe is now downdraft? Two possibilities exist, the air may be downward moving but the temperature is still warm compared to the environment to melt hail or at least force drop shedding off of existing hail stones. Or perhaps the liquid water hasn't frozen from when they formed within the updraft in the previous 10 minutes. Most likely, the downdraft has just commenced, and the air is likely still warmer than the environment. Thus, using the KDP column can give some clues that liquid water exists, but there is considerable question as to whether or not the air is still ascending. Note that the KDP is much lower closer to the 21 kft level and provides little information as to the location of the updraft (Figure 7-28F).

The CC is perhaps least associated with updraft. In Figure 7-28G, the ZDR column indicates somewhat depressed values (CC = 0.95-0.97) extending above the freezing level, perhaps on two areas straddling the heavy precipitation core and downdraft. This may indicate some mixture of rain drop sizes in these regions.



Instead of a short-lived pulse-like updraft, updrafts in sheared environments will exhibit a more steady state behavior. Indeed, there will always be multicellular behavior with updraft pulsing, but now at any one time in a storm's lifetime, updraft exists, either in a new cell or a mature one. You can use similar methods with dual-pol data to detect updraft location as with ordinary pulse cells. And there are some new signatures that appear in the strongest storms in a sheared environment. Now, let's add a dual-pol component to the reflectivity-based conceptual model of a non-severe, severe, and supercell thunderstorm in Figure 7-29.

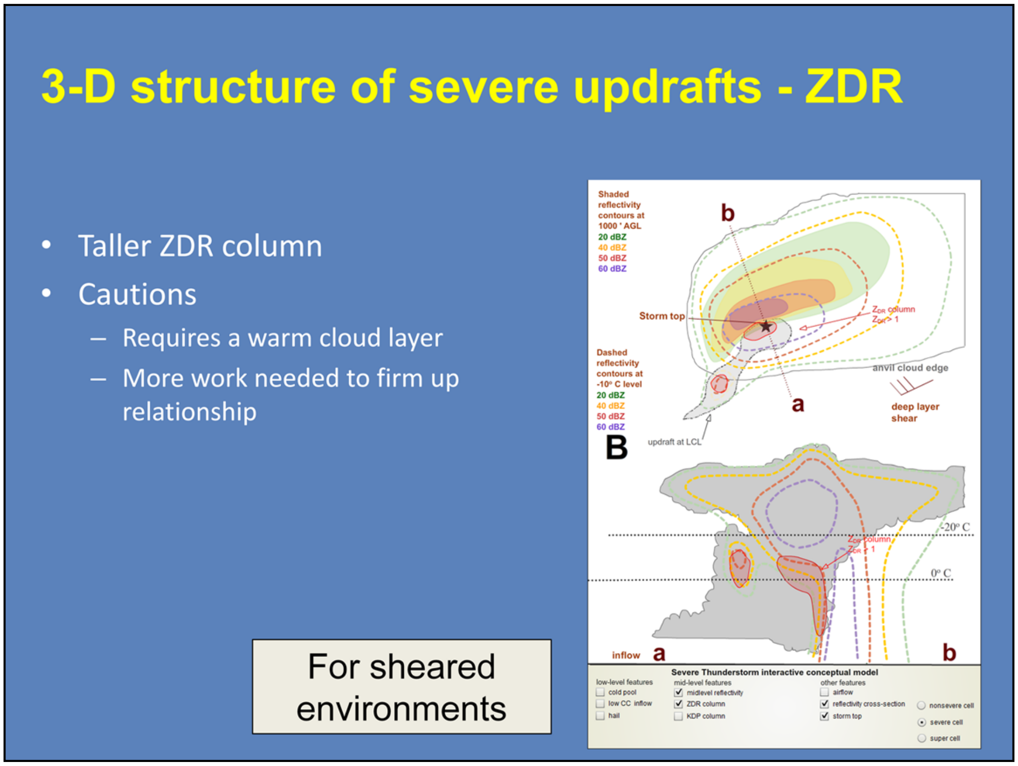
The convective storm severity conceptual model will appear in a separate browser window. Use it for the up and coming discussion.



In the external conceptual model diagram, make sure you have the nonsevere cell radio button selected. Then select the following checkboxes: midlevel reflectivity, ZDR column, reflectivity cross-section, and storm top.

The ZDR column may appear for a longer duration than in an ordinary cell, yet the shape of the ZDR column may change. In an updraft that is weak, the ZDR column may only extend a few degrees above the freezing level (Figure 7-29A). It may also not exhibit any kind of overhang, just like the high reflectivity envelope.

Remember that in order to get a ZDR column there must be a warm layer.



In the external conceptul model diagram, make sure you have the severe cell radio button selected. Then select the following checkboxes: midlevel reflectivity, ZDR column, reflectivity cross-section, and storm top.

As the updraft intensifies, the ZDR column expands upward and over the WER along the bottom of the strong reflectivity overhang (Figure 7-29B). The updraft and ZDR column are collocated.

Remember that in order to get a ZDR column there must be a warm layer. More work is needed to firm up guidance regarding the ZDR column and it's relationship to updraft severity.

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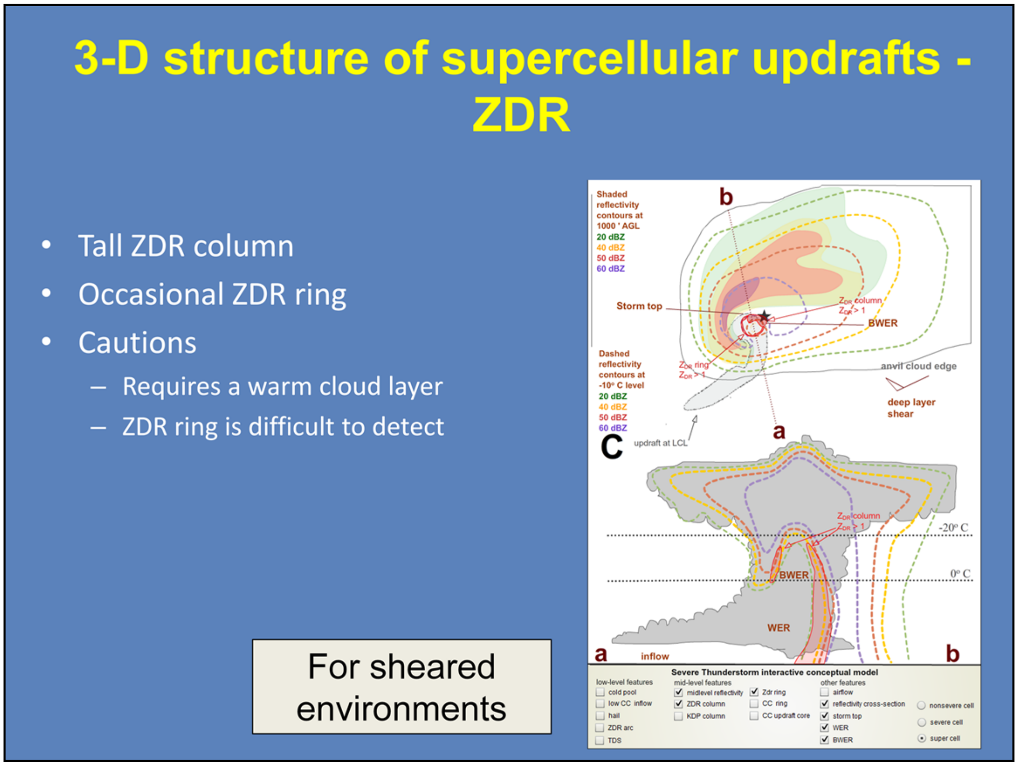
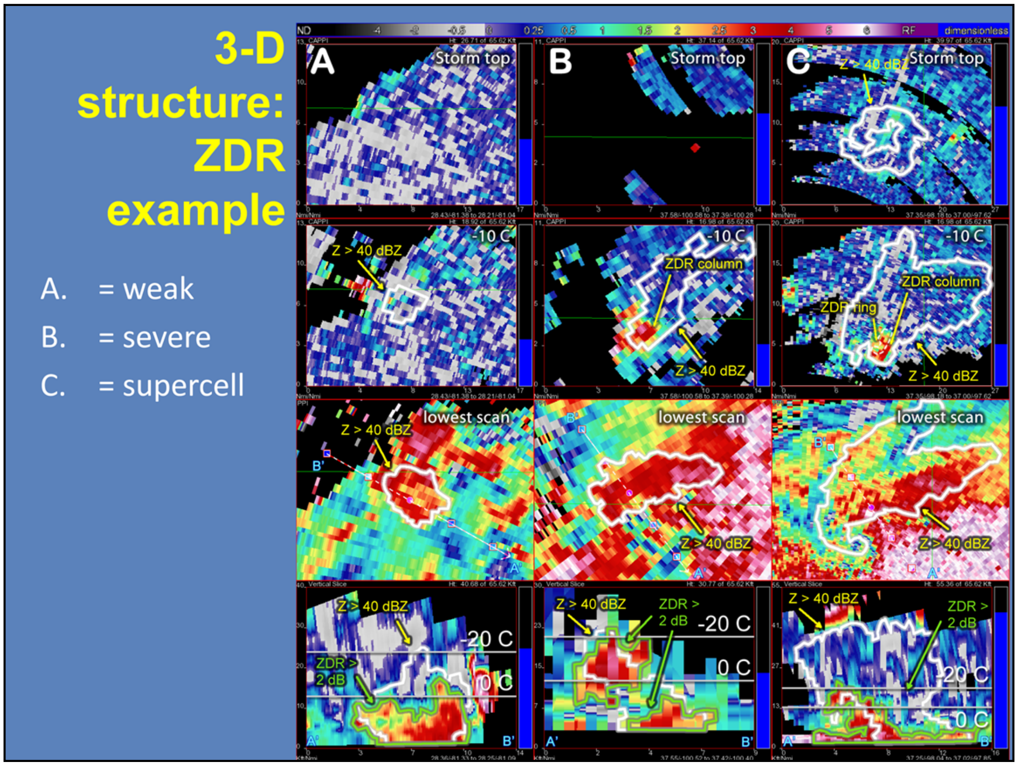


Now the thunderstorm is strongly rotating and produces an updraft so strong that even large liquid rain drops don't have time to form by the time the air flows above the freezing level. Instead, what happens is that large drops reside along the periphery of the updraft into the sub-freezing air. This is why the ZDR column may lie on the periphery of the BWER (Figure 7-29C). The strong circulation may actually transport the large rain drops around the exterior of the updraft resulting in a ZDR ring. The ZDR ring, like the BWER itself, is often small and ephemeral meaning that poor radar sampling may prevent you from detecting many true events.

Remember that in order to get a ZDR column there must be a warm layer. More work is needed to firm up guidance regarding the ZDR column and it's relationship to updraft severity.



An example of how the conceptual model plays out with real storms appears in Figure 7-30.

In the non-severe storm case (Figure 7-30A), the ZDR column extends up close to the -10 deg C level on the western end of the updraft. At no place is there an overhang in the ZDR column.

The severe storm case in this column B (Figure 7-30B) shows a much taller ZDR column that extends above the -20 deg C level. The vertical cross-section reveals a substantial ZDR column overhang along the reflectivity extending over the low-level inflow and WER.

Finally, in the supercell example in Figure 7-30C, the ZDR column at -10 deg C is forced along the sides of the BWER. There is even indication of a ZDR ring around the lowest reflectivity portion of the BWER. This ZDR column extends upward to the -20 deg C level in the vertical cross-section and then descends down to the low-levels in the forward flank reflectivity gradient.

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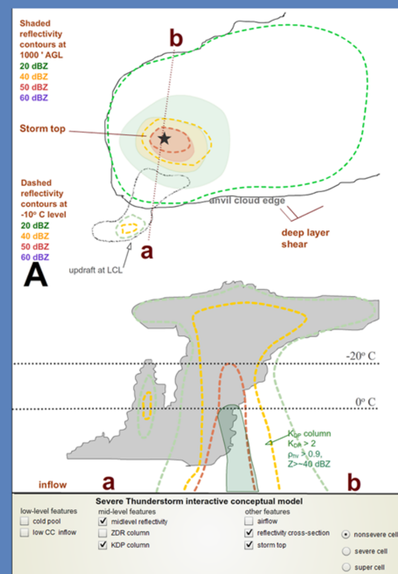
3-D structure of weak updrafts - KDP

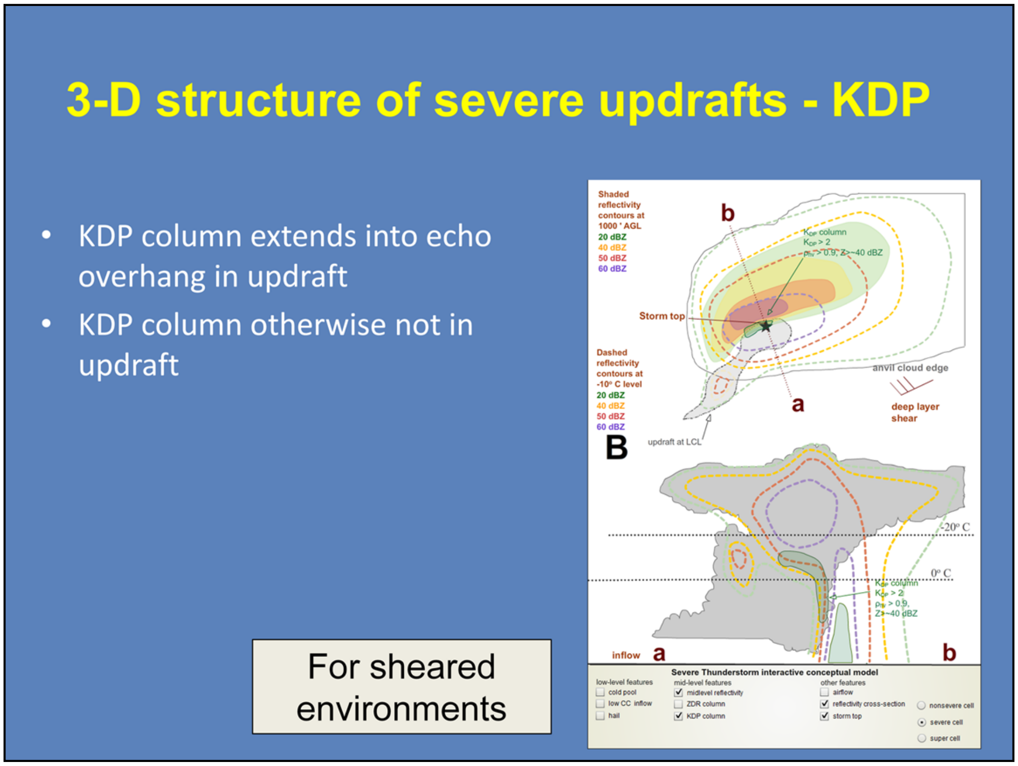
- KDP column primarily in warm layer
- Mostly occupies precipitation core
- Mainly outside updraft



For sheared environments

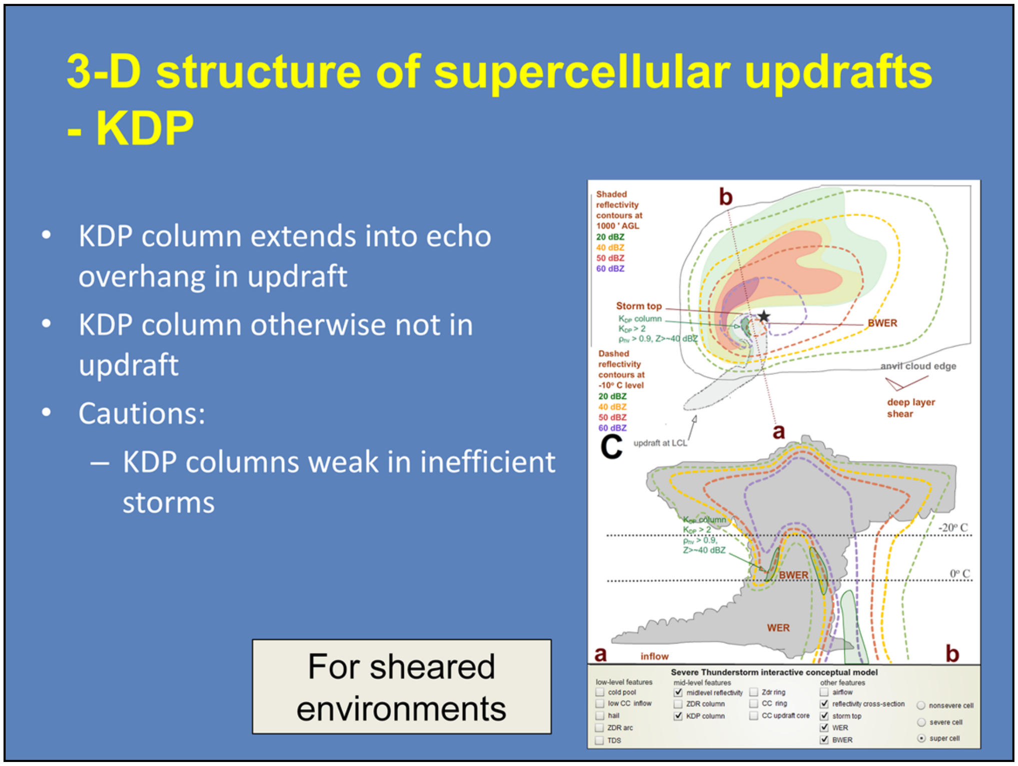
KDP columns do appear somewhat similar to ZDR columns, but there are differences that reflect the focus on integrated water content that KDP measures as opposed to the shape of large rain drops. In the non-severe storm conceptual model the KDP column extends up to just above the freezing level and it occupies more of the heavy precipitation where large volumes of rain and wet hail descend to the ground (Figure 7-30A).

Remember that in order to get a ZDR column there must be a warm layer.



In the severe thunderstorm case the KDP column extends upward and over the WER but at a slightly higher height than the ZDR column (Figure 7-30B).

In order for high KDP values to show, there needs to be larger quantities of liquid water. This area is likely to see that within the updraft, whereas the ZDR column may only be occupied by widely scattered large drops. This overhang of KDP often connects with the column of KDP extending to ground within the core of the storm. This cascade of high KDP falls outside of the main updraft. The higher extent of the KDP also can reflect the stronger updraft than with updraft in Figure 7-31A.



In the supercell, the KDP column also extends well above the freezing level. However, the highest values may lie on the upshear side of the BWER rather than the downshear location of the ZDR column. The KDP column extends into the core and descends to the ground in the heavy precipitation shield, often further into the core than the region of high ZDRs. Only in the highest parts of the KDP column would there be an association with the updraft. However much of the KDP, even at this altitude, may be further away from the updraft core than the ZDR column. Part of the reason why is that this region may be where the updraft is even weaker allowing more precipitation to fall out or be recycled. Yet, at an environmental temperature of -10 deg C, the presence of this much liquid water indicates that the updraft is still warming the area.

Sometimes, severe thunderstorms, especially supercells, may not exhibit much of a KDP column. This happens when severe storms produce a small amount of very large hail but not much volume of liquid or ice. Reflectivities may be high but KDPs stay low.

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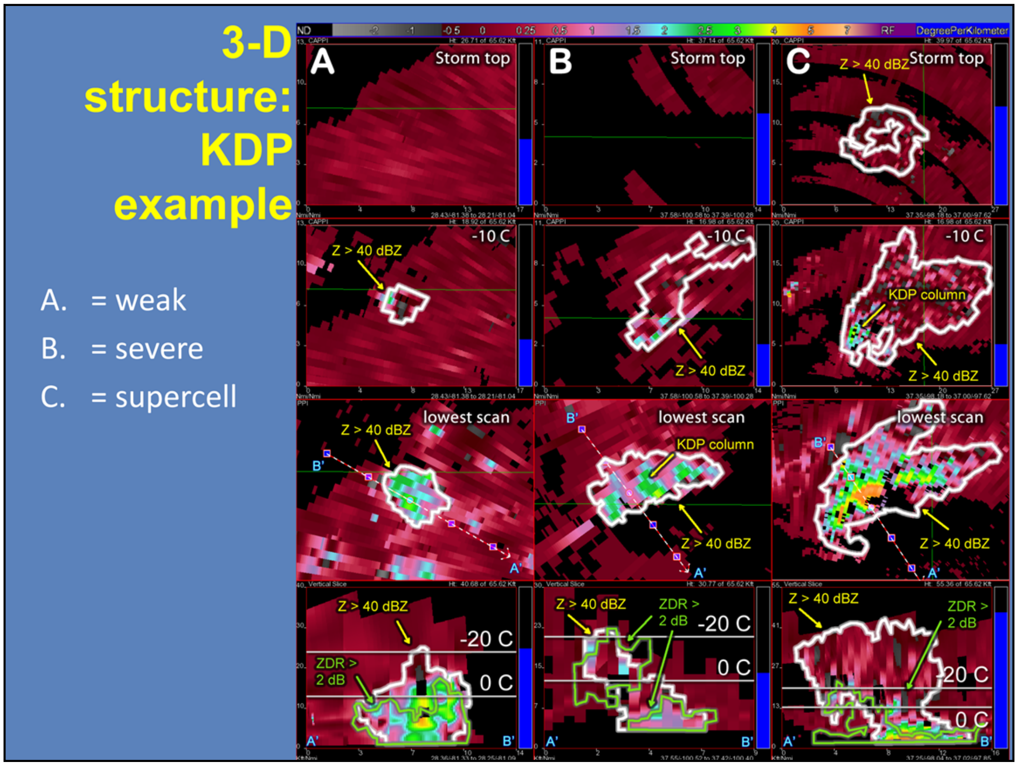
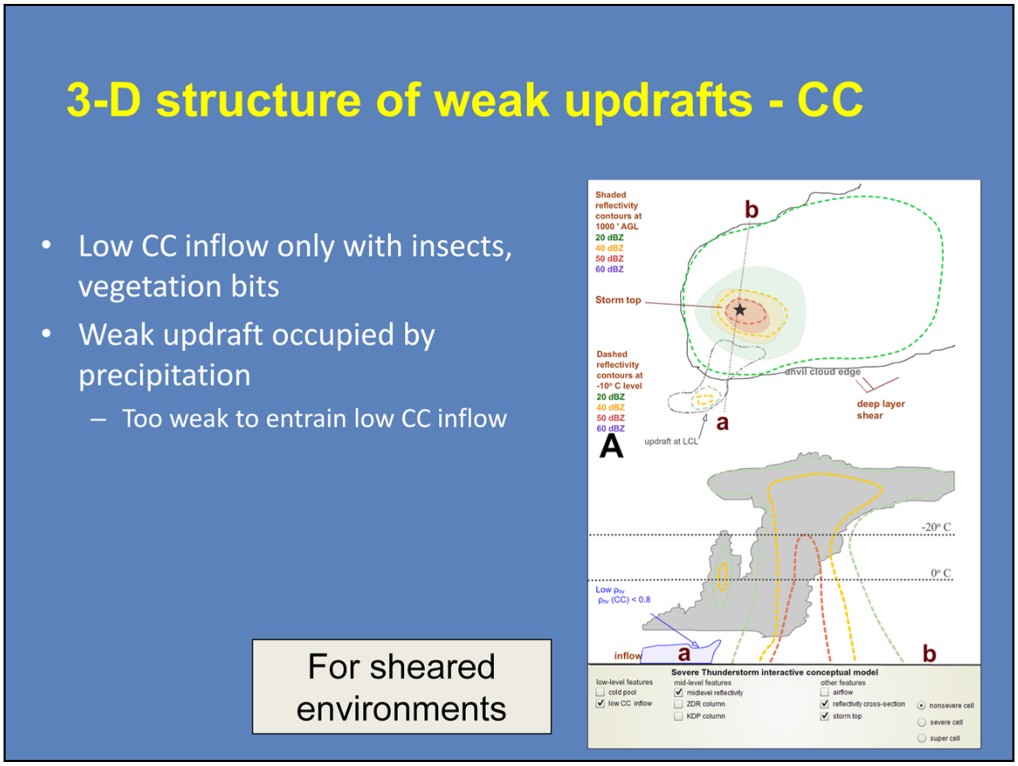


The same set of storms is now shown in KDP where the conceptual model is confirmed except for one case (Figure 7-32). Talking about where it is confirmed, note that the non-severe thunderstorm exhibits a substantial KDP column that rises almost to -10 deg C. Going to the severe and supercell thunderstorm cases, the KDP fields become more diminished aloft. Only down low do the KDP fields rise. Both of these storms appear to be dominated by drier hailstones. Nevertheless, at -10 deg C, there are regions of high KDP values, especially upshear (west) of the BWER in Figure 7-32B.



Correlation coefficient (CC) is perhaps one surprising tool to identify updrafts. Sometimes, thunderstorm updrafts show low values of CC (CC < 0.8) where reflectivity is very low. These low CC areas manifest themselves as an upward extension from the low CC clear echoes often found within a boundary layer occupied by flying insects or lofted light vegetative debris. Thus, the appearance of the low CC inflow and updraft is very dependent on the presence of these scatterers. It is also dependent on at least part of the updraft being free of any precipitation. Very small amounts of precipitation rapidly increase the CC and mask the detection of the non-meteorological scatterers.

In Figure 7-33, let's assume that the air is filled with insects or light vegetation and that the clear air boundary has a typical CC of less than 0.8. A non-severe thunderstorm typically sports an updraft too weak to result in an upward extension of the low CC air (Figure 7-33A). Most likely, the weak updraft is unable to produce a precipitation-free WER and the low CC signal is masked.

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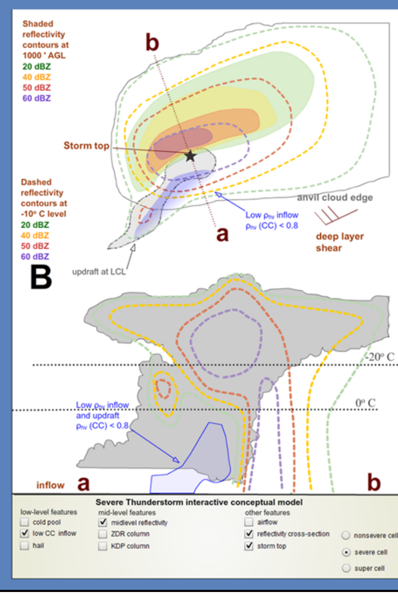
3-D structure of severe updrafts - CC

- Low CC inflow only with insects, vegetation bits
- Strong updraft entrains low CC inflow up into WER
  - Any rain obscures the signature

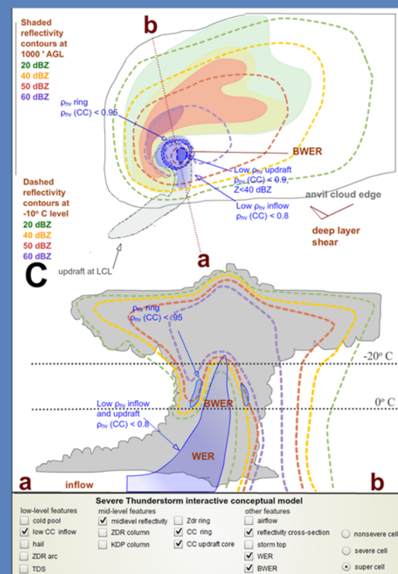


For sheared environments

As the thunderstorm updraft intensifies, the likelihood of a precipitation-free updraft increases, allowing for the opportunity for the radar to detect the low CC echoes from the non-meteorological scatterers (Figure 7-33B) like insects and vegetation bits. Then the updraft can entrain this low CC inflow into the WER. Any rain in the updraft obscures this signal.

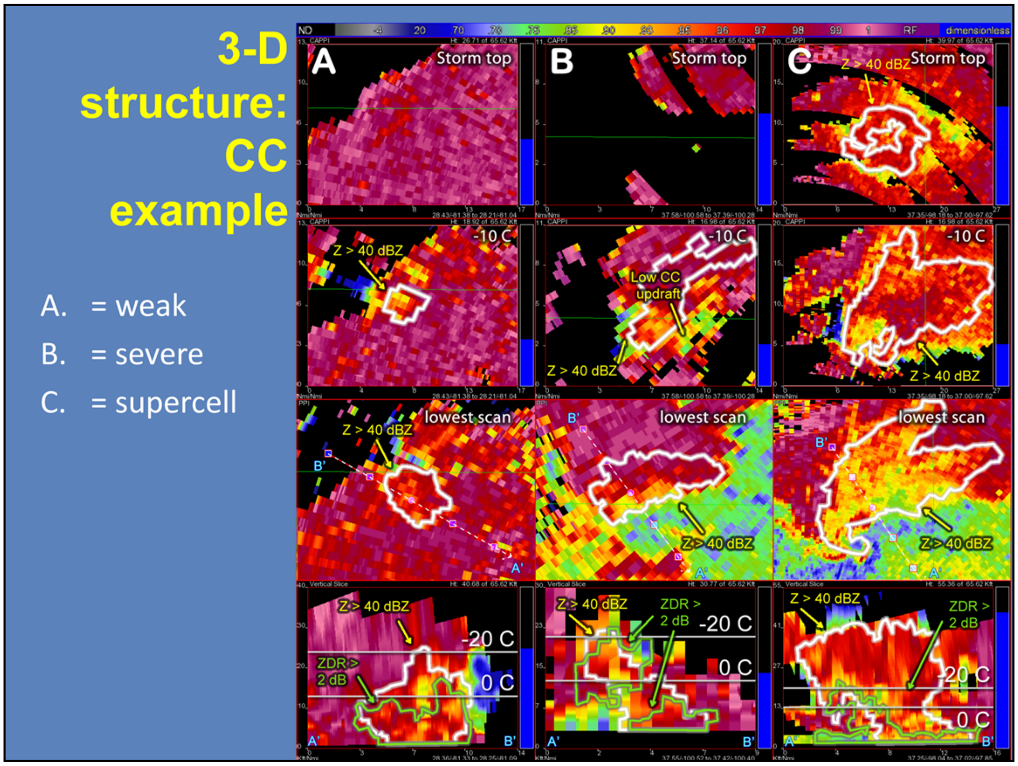
3-D structure of supercellular updrafts - CC

- Strong updraft entrains low CC inflow (if available) up into WER
  - And even the BWER
- Low CC ring around exterior of BWER



For sheared environments

Strong supercells exhibiting a pronounced BWER present a challenge sometimes in that within the center of the BWER may show the weak CC signal from insects and/or debris as you can see from Figure 7-33C. But at the edges of the BWER, a low CC ring may appear just above the freezing level as precipitation encounters the melting between the environment and the updraft.



Finally, we have the appearance of CC with the same three storms shown in Figure 7-34. The non-severe case in Figure 7-34A did not have a low CC boundary layer owing to extensive light stratiform precipitation. As a side note, there was a TBSS confined to the subfreezing air. This was perhaps due to hail production after an updraft pulse. No severe hail was reported, however.

In Figure 7-34B there was a low CC boundary layer with values falling below 0.7 at times. Some of that low CC air was actually being entrained upward into the updraft within the WER high enough to be detected at the -10 deg C level adjacent to the precipitation core. In the supercell case in Figure 7-34C, there is substantially low CC in the low-level inflow, but the radar cannot detect the non-meteorological scatterers up to the -10 deg C due to light precipitation filling in the WER.

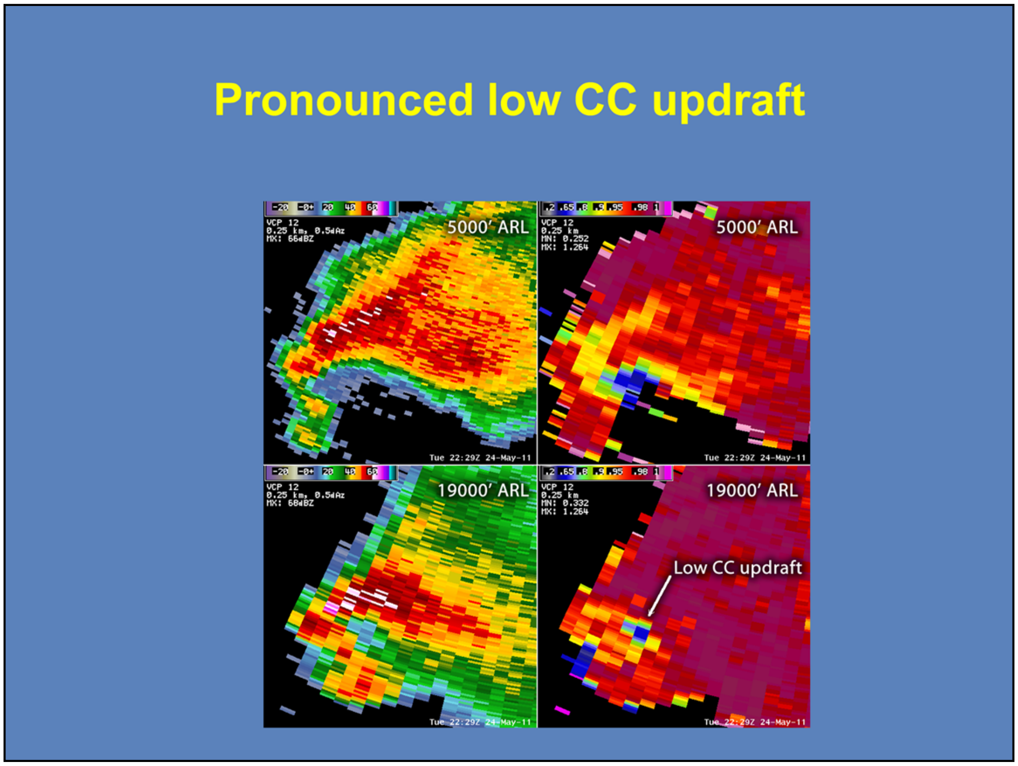


Figure 7-35 shows an example of a supercell with a prominent low CC updraft where the low values coincide well with the inflow notch at low levels extending into the BWER in the subfreezing air.

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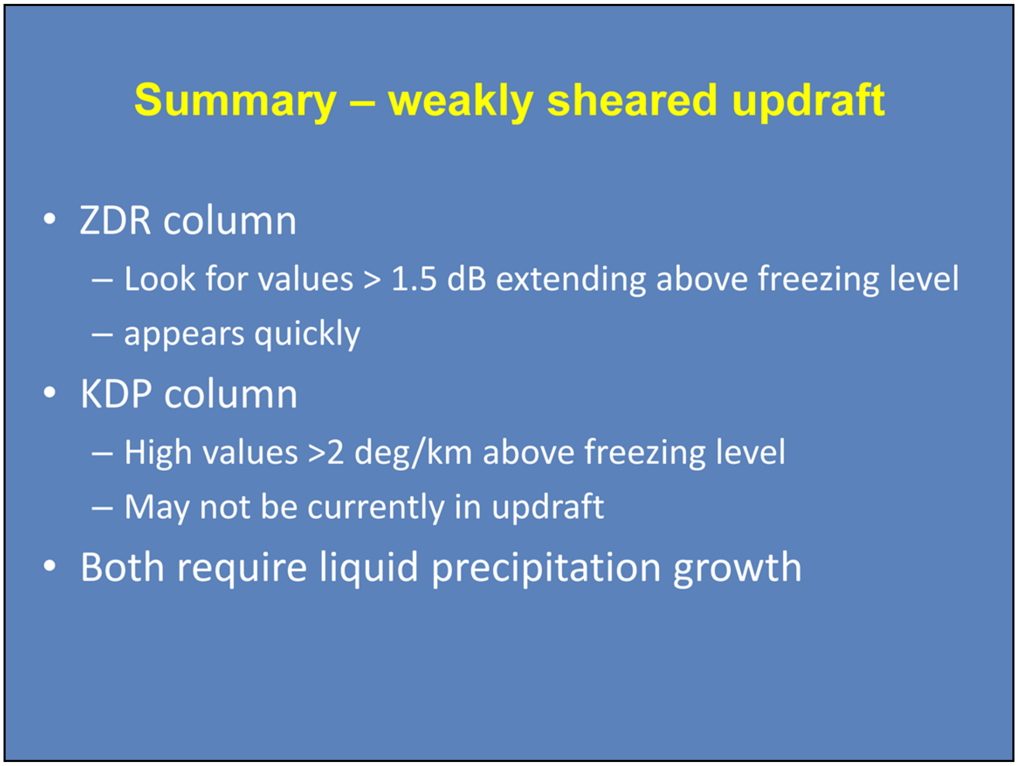

ZDR values in excess of 1.5 dB extending from low-levels up to and above the freezing level. The ZDR column will appear as soon as the precipitation core develops and is the best dual-pol updraft signature.

A KDP column appears after the ZDR column as the liquid water content increases in the updraft above the freezing level. However, by the time this happens, part of the column may be occupied by downdraft.

For both ZDR and KDP, there needs to be a warm cloud layer to enable liquid precipitation growth within the updraft.

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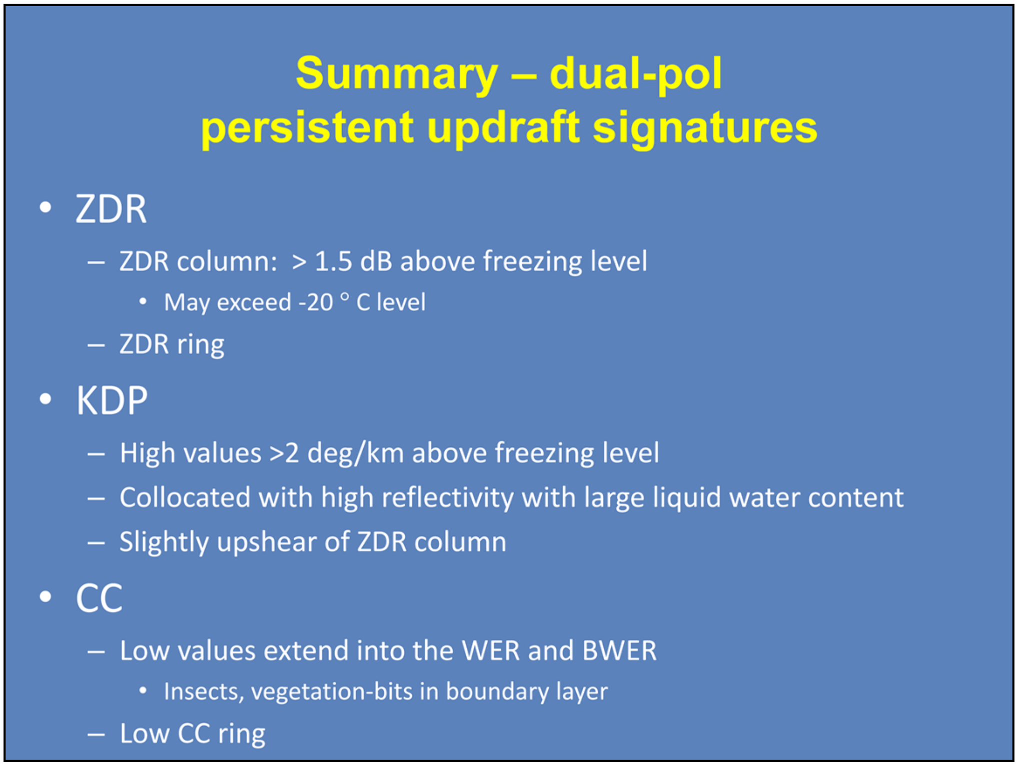

ZDR

- Look for ZDR columns to extend into higher altitudes for the stronger updrafts in a given environment. Some ZDR columns may extend to -20 deg C.
- Strongly rotating updrafts may produce a high ZDR ring surrounding a BWER.

KDP

- The KDP column neighbors the ZDR column and will produce highest values where reflectivity is high.
- In supercells, the KDP column is often displaced slightly upshear of the ZDR column above the freezing level.
- KDP columns may not appear in some severe storm updrafts if they produce mostly large dry hail.

CC

- Low CC columns may appear in severe thunderstorm updrafts if the following conditions are met:
- Presence of low CC non-meteorological scatterers in the boundary layer (e.g., insects and/or light vegetation debris)
- Relatively weak reflectivities in the updraft.
- A low CC ring may appear in the periphery of the updraft just above the freezing level as a result of mixed phase precipitation.

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If you have passed the quiz, then you have successfully completed this lesson. If you have any questions, please contact us using any of the e-mail addresses listed on the bottom of the slide.

Welcome to Single Cell Downburst Detection, a module in the Convective Storm Structure and Evolution topic in the Radar and Applications Course.

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Learning Objectives

1. Identify the environmental and storm signatures favorable for dry and wet microbursts.
2. Identify favorable environmental and radar signatures of a high wind threat from supercells.

These are the learning objectives for this lesson. The end of lesson test includes some questions based on these objectives.

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What is a Downdraft?

- Pulse downdraft spectrum
  - ➢Microbursts < 4 km < Downbursts

- Dry (<.01 inches, cell reflectivities < 35 dBZ)
- Wet (>.01 inches, cell reflectivities > 35 dBZ)

Wet downburst near Norman, OK on June 14, 2011

The downdrafts discussed in this section are typically on the same scale as the individual ordinary cell updraft. Downbursts and microbursts are outflows of an ordinary cell downdraft. The only difference is that downbursts are considered outflows larger than 4 km in diameter while microbursts refer to outflows less than 4 km in diameter.

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What Drives Downbursts?

1. Lateral dry air entrainment (lower Theta-E in mid layers)
2. Subcloud cooling (forcing for dry microbursts)
3. Sublimation (LCL < 0 deg)
4. Precip loading (Z > 45 dBZ)
5. Rear-Flank Downdraft (RFD) forcing

These are the primary factors that drive downburst processes: 1) Lateral dry air entrainment, which is measured by lower equivalent potential (theta-E) temperature in mid layers, 2) Subcloud cooling (which is the forcing for dry microbursts), 3) Sublimation, which occurs when the LCL is below freezing, and 4) Precipitation loading (which occur when lapse rates drop below 8 deg K/km). This factor (process) is observed with microbursts with a descending precipitation core of > 45 dBZ. Lastly, the fifth factor is Rear Flank Downdraft (non-hydrostatic) forcing in supercells. All of these factors will be covered in subsequent sections of this lesson.

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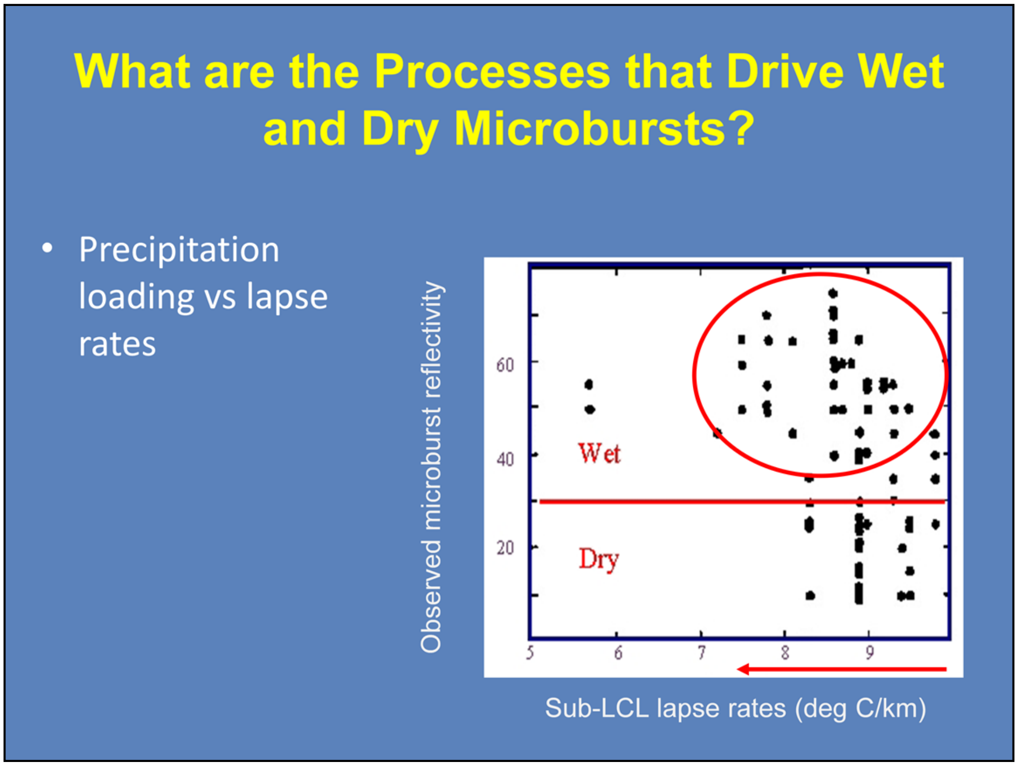
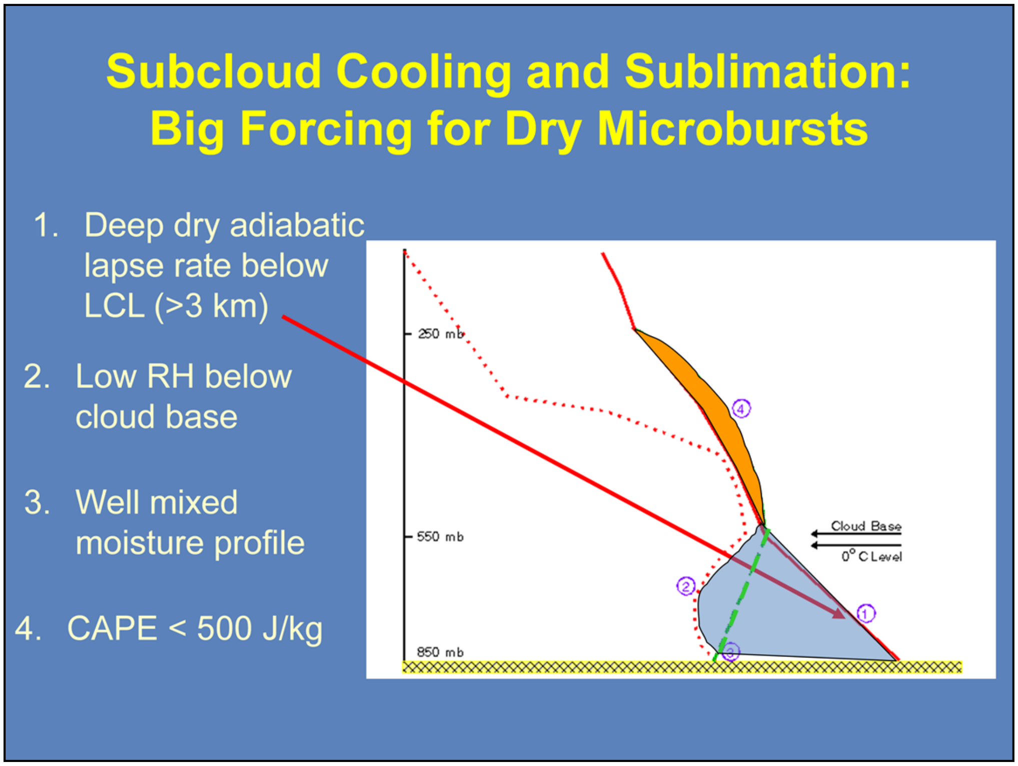


Taking a look at two of the primary processes that drive wet and dry microbursts, the primary factor is negative buoyancy and what forces it. Two forces at play are precipitation loading and lapse rates. They are both related. Let’s look at this figure from Srivastava. As lapse rates decrease, downdrafts have an increasingly difficult time of maintaining their descent based on negative buoyancy effects alone. Heavy precipitation (especially those with reflectivity cores greater then 40 dBZ), can force the descent of a downdraft even if it loses negative buoyancy.



Dry microbursts are forced by evaporating precipitation below the LCL. These events are most common in the arid or semi-arid regions where LCLs are at least 3 km AGL. There are four basic characteristics of dry microbursts (kind of like clues in the environmental data: 1) A deep, dry adiabatic lapse rate below LCL, 2) Low relative humidity below the cloud base, 3) A well-mixed moisture profile (you can see the constant mixing ratio line from the surface to the LCL, and 4) Weak CAPE usually 500 j/kg or less with moist midlevels above the LFC. The typical dry microburst sounding is termed an “inverted V”.

In these type of environments, the updrafts are relatively weak so precip loading and/or lateral entrainment are not major factors in contributing to downdraft strength. If the LCL is below freezing, the precipitation that does form initially cascades down as snowflakes which maximizes the surface area exposed to dry air.

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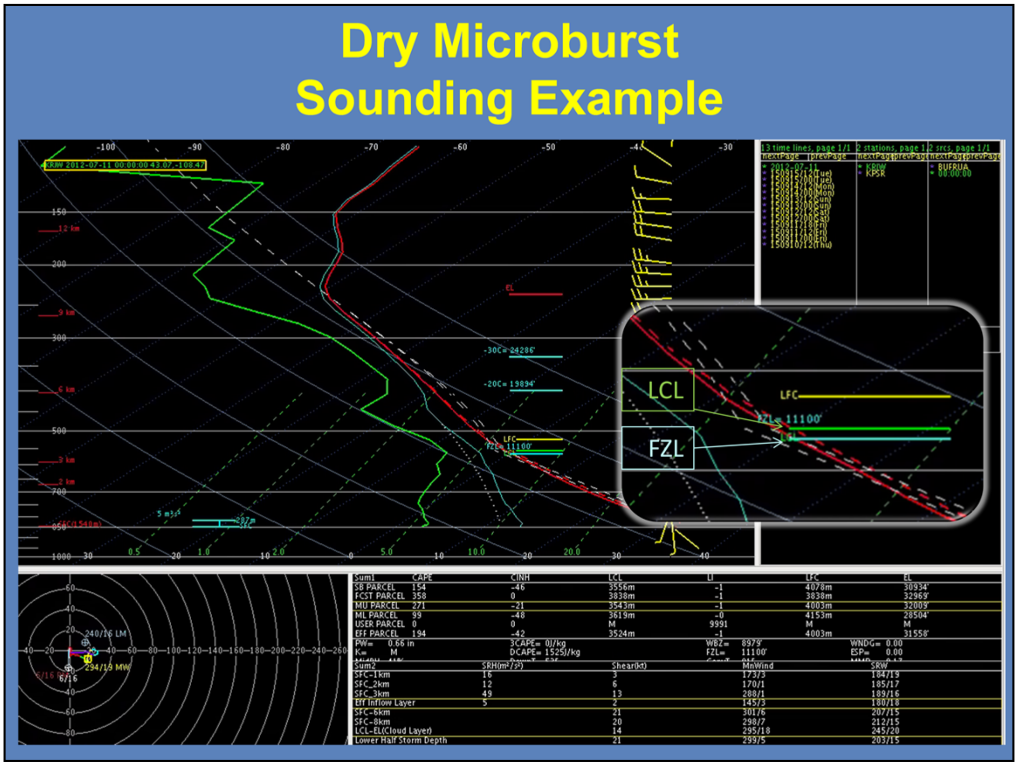

We are going to show you an example of a typical dry microburst event. Here's the plotted RAP analysis sounding taken at Riverton, WY (KRIW) on 00 UTC July 11, 2012.

Note the LCL height is below freezing and there lies a deep, dry adiabatic layer extending to the ground with very little CAPE.

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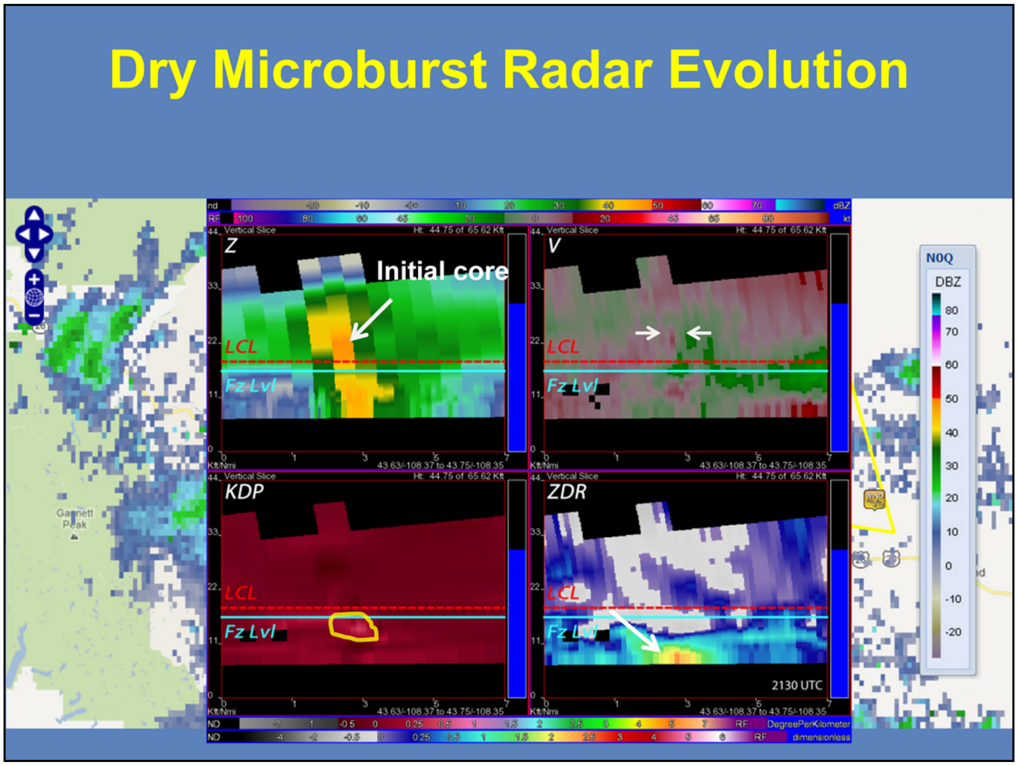

Next is the storm scale evolution of a microburst as seen by a multi-product cross section product from KRIW. There were two storms initially just northeast of the radar. The second one shown here is the one that became severe. You can see the warning polygon out of WFO Riverton. The microburst-producing storm initially produced an elevated core structure, as shown in the static 4-panel cross section at 2130 UTC (this is figure 7-40 in the Student Guide). Note the development of a 30-35 dBZ core 15-24 kft above the surface. This core of frozen precipitation eventually descended and sublimated just below the LCL. Radial convergence was weak (25-30 kts) as is typical for most dry microbursts. The ZDR column was apparent extending down to the surface and simultaneously weakening slightly prior to microburst impact. KDP showed larger values in the core indicative of a some intense rainfall. The loop which will pop up in a separate window when you advance to the next slide shows the core descending through the melting layer. Monitoring the descent of the reflectivity core helps provide some lead time of a dry microburst.

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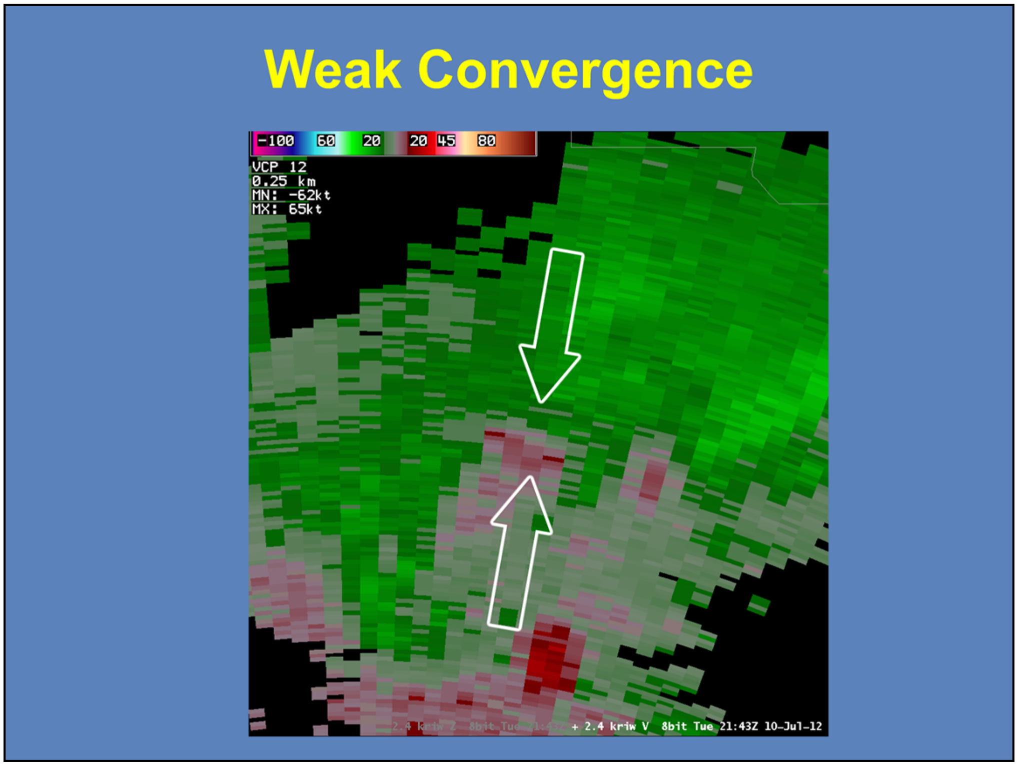
Flash Animation of a Dry Microburst

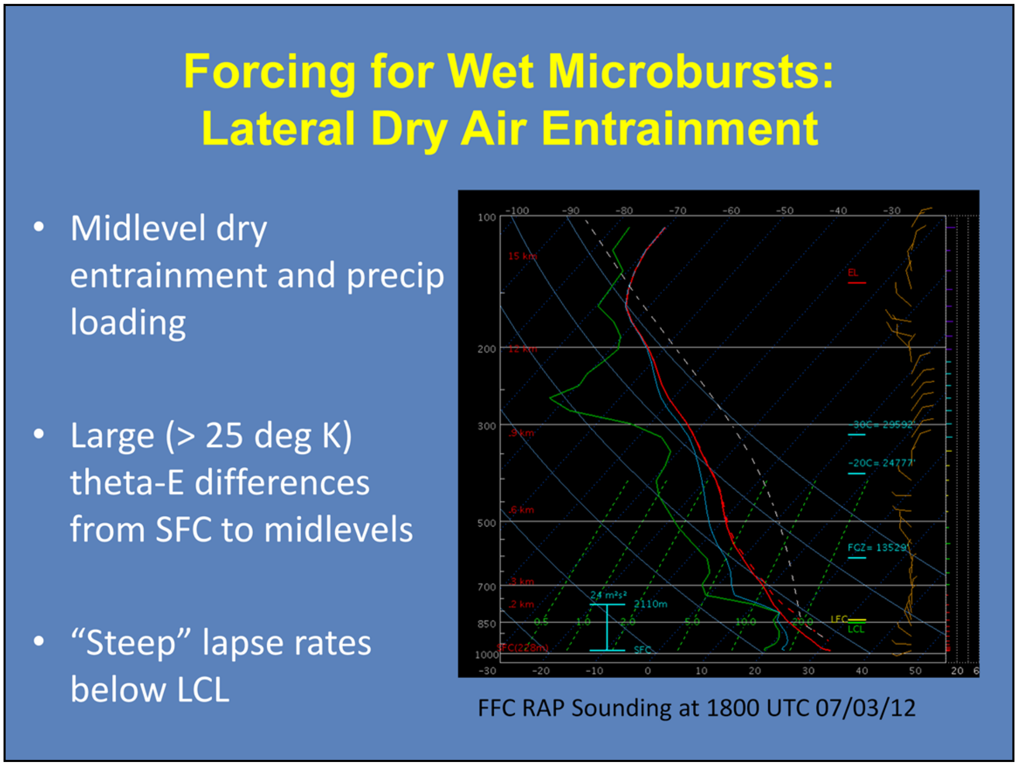
Web Object

Address:
http://www.wdtb.noaa.gov/courses/rac/severe/objects/RIW_DryMicro_FSI.html

This is a flash animation of an evolving dry microburst from KRIW.



This is a velocity image of the dry microburst producing storm showing weak convergence near and just below cloud base (~ 20, 000 ft) indicating the downdraft has initiated. Radial convergence is weak (25-30 kts), as is typical for most dry microbursts, and more often than not, will be not be definable due to radar sampling limitations or flow tangential to the beam.



Wet microbursts are forced by midlevel entrainment and precip loading. Large (> 25 deg K) theta-E differences from the surface to midlevels are observed on days with wet microbursts. A "steep" lapse rate below the LCL is also important to allow the strength of the downdraft to reach the surface with strong outflow. However, the height of the LCL is not as important as with dry microburst cases. In this sounding from Atlanta from July of 2012, there was a SBCAPE of 2583 J/kg, no CINH, 4 kts of shear from 0-6 km, and substantial mid level dry air to support wet microburst generation. For reference, the avg. RH in midlevels of 63%. We are going to examine this case in more detail in a bit

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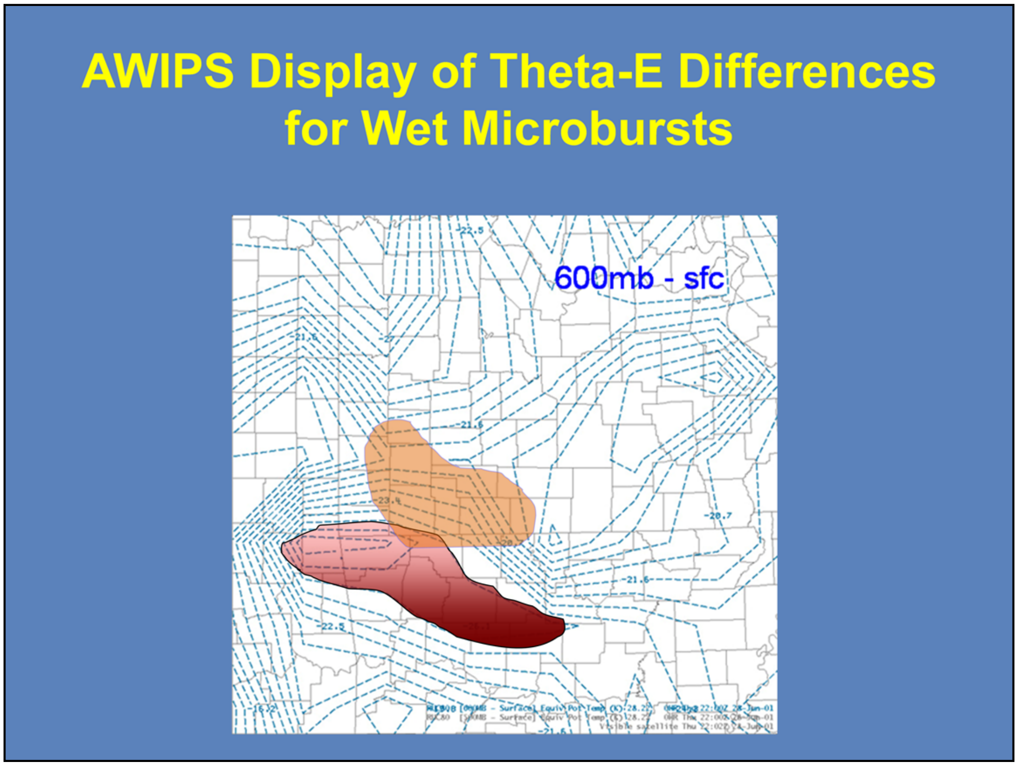
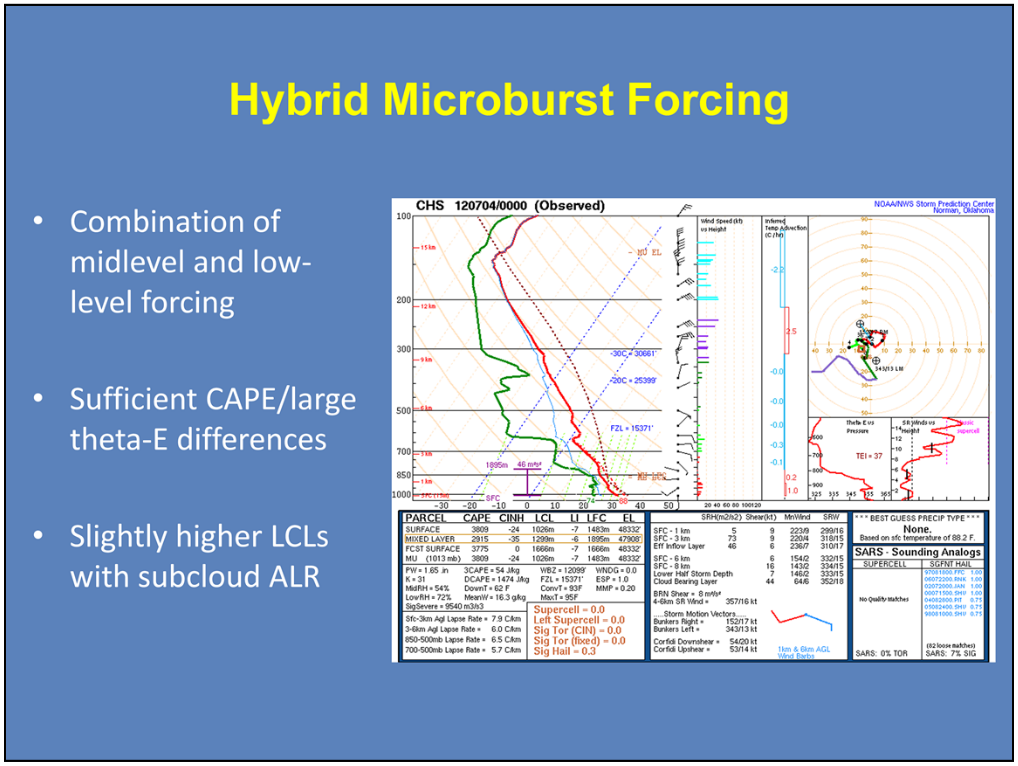
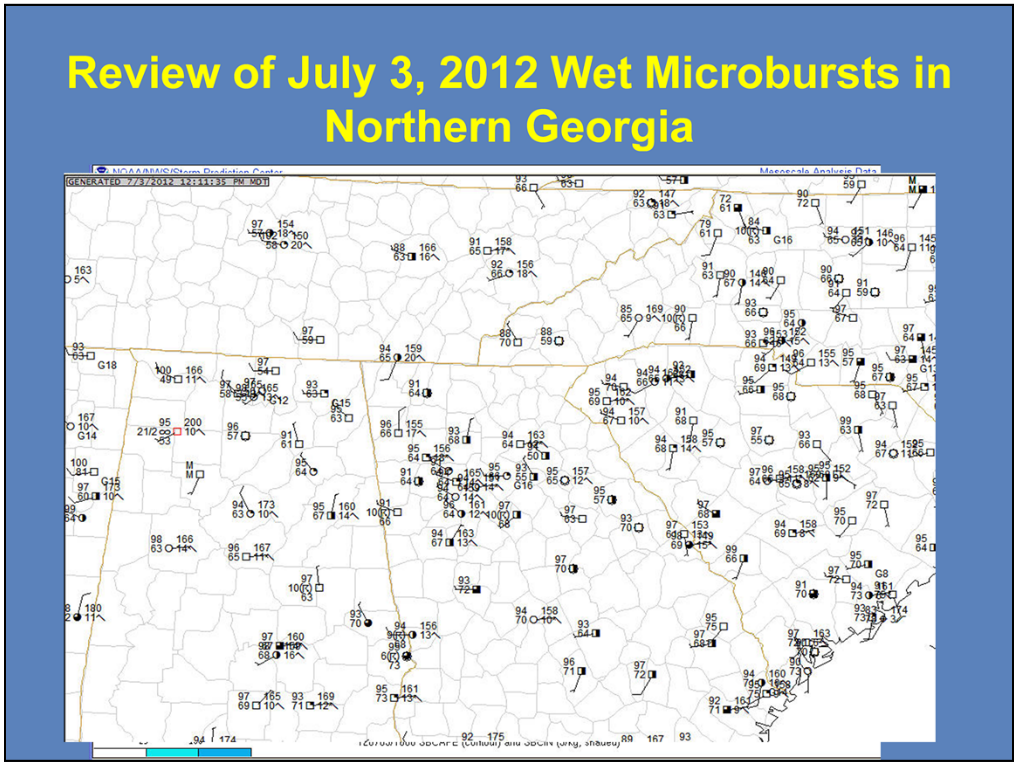


The AWIPS volume browser allows you to subtract theta-E from two layers. In this figure, we see the isolines computed as theta-E difference between 600mb and the surface for a wet microburst event in Southwest Missouri. Values exceeding 25 deg K are shaded in red. The areas where significant microbursts occurred are shaded in blue. As is often the case, the risk area is not always confined to the maxima of Theta-E difference but throughout the gradient regions.



Hybrid microburst environments display characteristics of both dry and wet microbursts. Environments that support hybrid microbursts usually have sufficient CAPE, large theta-E differences from surface to mid-levels, and slightly higher than normal LCLs with adiabatic subcloud lapse rates. In a sense, hybrid microburst environments represent the middle of the microburst sounding spectrum. It could be said that most microburst soundings are hybrids.

This observed SPC NSHARP sounding from Charleston, SC was taken on the same day (July 3, 2012) that several microburst producing storms occurred all across GA, SC, NC, and VA. Note the theta-E vs Pressure plot right below the hodograph. There was an observed difference of 37 deg C from around 600 mb to the surface. This is a great environment for hybrid and wet microbursts.



A surface plot at 2011 UTC showed a large area of temperatures in the mid-90s and dew point temperatures in the mid-60s to around 70F in portions of central and northern Georgia. There was very weak surface convergence but with moderate instability increasing throughout the afternoon of 3 July 2012, forcing was accentuated by numerous outflow boundaries generated by thunderstorms across northern Georgia. You are going to see these outflows in motion when you advance to the next slide.

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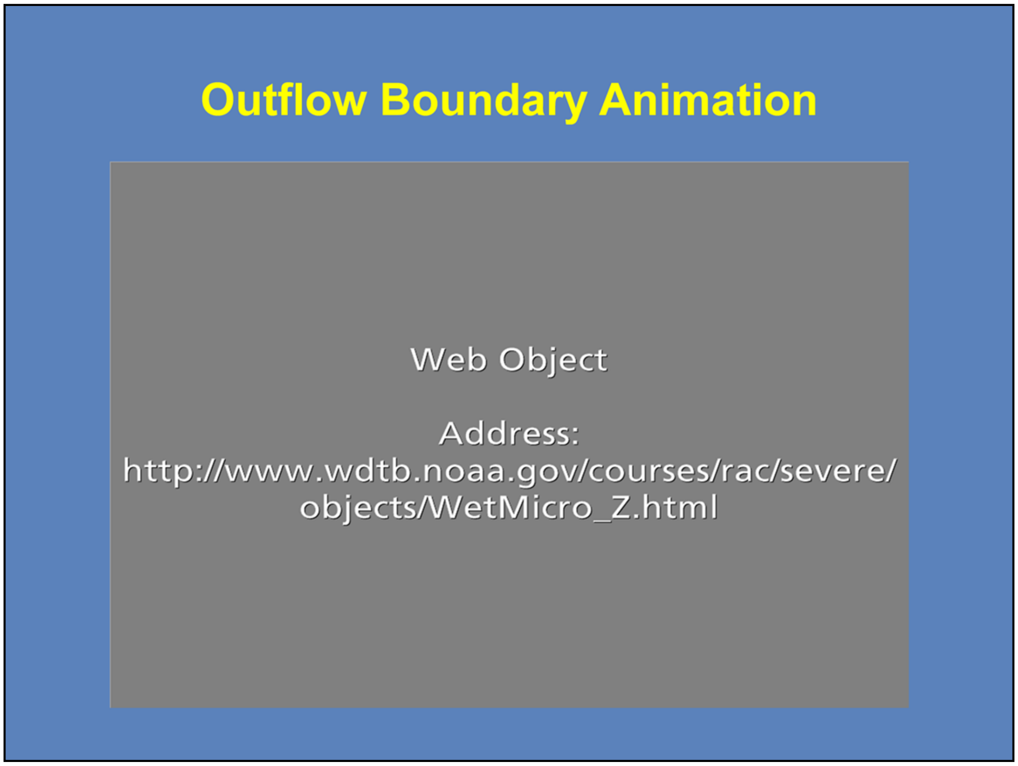

The window that has just popped up displays an animation of 0.5 deg Reflectivity from 2042Z to 2238z from KFFC (Peachtree City radar). As this was a weak shear environment, a number of thunderstorm outflow boundaries with large boundary motions were apparent, many of moved toward the radar. By 2105 UTC, which coincides with Figure 7-47 in the Student Guide, there was increasing thunderstorm development along and behind the outflow boundaries, especially to the east of KFFC. Some of the biggest storm intersections occurred along colliding outflow boundaries east of the radar from 2115z to 2205z. Next we will show a FSI cross section at 2142z through a storm that developed NE of the radar and produced severe winds.

The loop now playing shows a series of FSI cross-sections of Z, V, ZDR, and KDP from 2134 to 2210 UTC through a storm of interest northeast of the radar. Radar signatures show some of the signs of an impending microburst. First, you can see a large elevated core of high reflectivity (> 60dBZ) from 13-26 kft which had a TBSS. Then, the core begins to descend as the velocity shows some mid-level radial convergence as air flows into the top of the downdraft. About this same time (2142z) , there is a core of values of ZDR near zero indicating some hail in the downdraft column. If you toggle over to the KDP product, you can see high values of KDP (3-5 deg/km) in the downdraft core, indicating a high concentration of liquid water and potential hail accompanied the microburst. It is speculated that the large increase of KDP values prior to microburst may be due to the addition of hail meltwater to the rain already falling within the downdraft.

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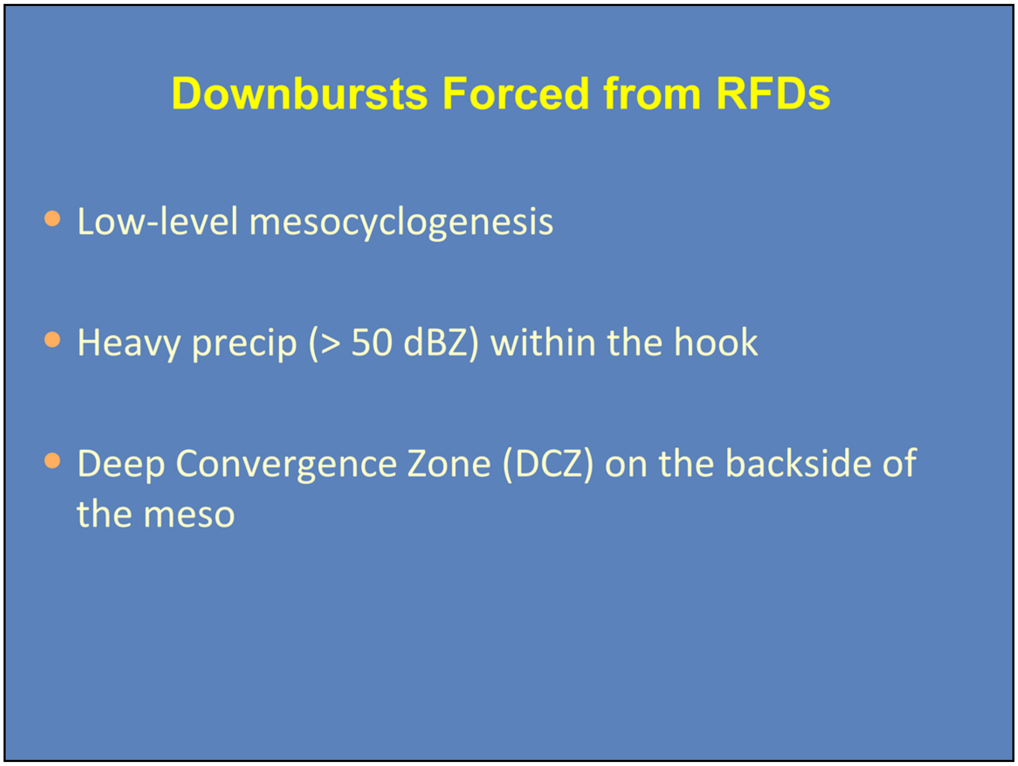
What are the Storm Precursors of a Wet/Hybrid Microburst?

- Stronger than normal initial updraft
  - Long lead time, relatively high FAR
- Descent of strong core/Mid-altitude radial convergence increase
  - Short lead time (< 10 min), moderate FAR
- Storm collapse
  - Short lead time (< 10 min), but high FAR!
- Decrease in ZDR (if hail is also occurring)

So, what are the signatures of an impending wet or hybrid microburst? Well, after the underlying threat assessment of the environment which can give clues to an increased possibility to an event, you should detect a stronger than normal initial updraft pulse. Now, in terms of verification of this signature, the stronger than normal initial updraft may provide a good lead time but also result in lots of false alarms, as the process is dependent on the dry air entrained to maintain or accelerate the downdraft. Secondly, as we saw in the Peachtree City microburst example, the decent of the strong core is a likely location for the base of the downdraft. This is often associated with the development of mid-altitude radial convergence as air flows into the downdraft source. However, this signature is often ill-defined and may yield a short lead time (on the order of 5 – 10 min) . And, it has a moderate False Alarm Rate (FAR).

Also, most severe pulse storms collapse as they are about to produce a downburst. This process will be evidenced by a simultaneous decrease in cell-based VIL and max reflectivity with time. But, not all weakening storms end in a downburst. And, since you have to wait till the end of volume scan for a time trend, the lead time is short. In addition, the lack of any signals of a storm collapse does not necessarily mean you won't get a downburst. Thus, very high FAR on this precursor. Finally, another microburst detection signal may be a simultaneous decrease in ZDR if hail is occurring in the microburst. But note that not all microbursts will have hail and again not all weakening storms end in a downburst.



Now the same environmental parameters favoring hybrid or wet microbursts help to encourage a severe RFD**.** However, since we are talking about forcing from non-hydrostatic pressure deficits, they are not necessary if a strong, low-level mesocyclone is developing. Some indications in the supercell include heavy precipitation (> 50 dBZ) within the hook as this favors precipitation loading and evaporation cooling. Next, the indication of a deep convergence zone through 18-20 kft on the backside of the meso is a signal that strong damaging winds will occur usually just the right of the primary meso. Advance to the next slide to see an example of a northern OK HP supercell that produced a severe RFD.

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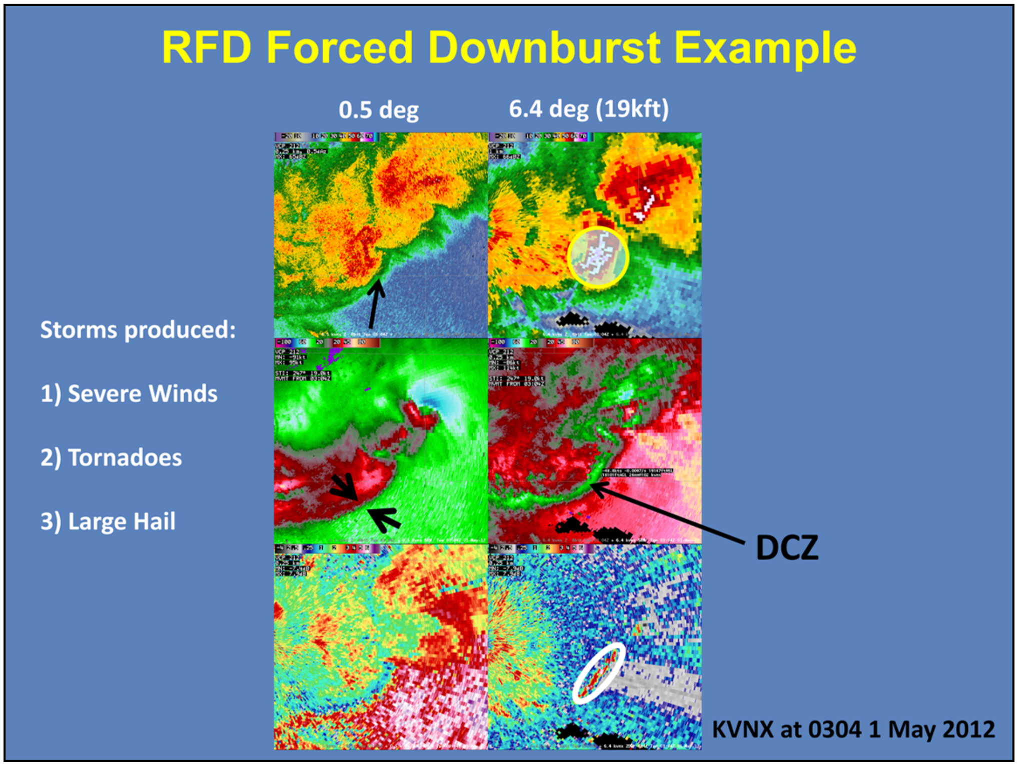

Here is an example of two supercell storms from the evening of April 30, 2012, in northern OK that illustrates most of the previously mentioned signatures of RFD forced downburst winds. With the storm closest to the radar, note a fat reflectivity hook at the 0.5 deg slice with huge inflow notch, a very high elevated Z core with a large 70 dBZ core at 6.4 deg slice (~ 19 kft MSL), elongated convergence zone with 70-80 kts of velocity difference extending up to 21 kft. The ZDR products (lowest two panels) display SW-NE oriented line of values > 1 dB on the 6.4 deg slice which indicates the updraft column extending above -20 deg C. Thus, the storm likely contains hail as well. It was just mentioned that the most damaging RFD winds likely occur with HP supercells, just to the right of the primary mesocyclone track. In this case, damaging winds and tornadoes occurred with both of the two distinct mesocyclones.

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Types of Downbursts	Dry Microbursts	Wet Microbursts	Hybrid Microbursts	Supercells
Forcing	Evaporative Cooling • Sub-Cloud (Low reflectivity) • Sublimation	Evaporative Cooling • Lateral dry air entrainment • Precip Loading • Boundary lifting	Same as Wet	Add Non-Hydrostatic Vertical Pressure Gradients
Environment	- Minimal CAPE - Very Deep Dry ALR - LCL Height Below Freezing - Weak Boundary Layer winds - Weak 0-6 km shear	- Steep midlevel Lapse rates - Moderate CAPE - Large Theta-E differences (> 25 to 30K) from surface to midlevel min) - Low cloud bases Dry sub-cloud ALR	- Sufficient CAPE (> 500 J/Kg) - Steep midlevel Lapse rates - Large Theta-E Differences - High LCLs - Deep, dry sub-cloud ALR	Supercell environment • 0-6 km shear > 15 m/s • Low LCL • Large CAPE • Steep sub-cloud ALR
Pre-cursor Radar Signatures	- Intense core above Freezing Level - Descent of max reflectivity core - Lowering ZDR (~0 DB); Midlevel radial convergence	- Descent of max reflectivity core - Mid-level radial convergence - Decrease in ZDR High KDP in descent	- Descent of max reflectivity core - Mid-level radial convergence - Decrease in ZDR	- Intensifying Mesocyclone with a RFD - Heavy precip (>50 dBZ) in the hook; DCZ

First, recognize clues in the environment. There will be parameters in proximity soundings suggesting an enhanced threat. Look for precursor signals based on the type of forcing. For dry microbursts, look for LCLs below 0 deg C, and an inverted V sounding. For wet microbursts, look for sources of dry air aloft and a subcloud ALR. In radar, you may detect a large, reflectivity core aloft with the initial pulse, then as the downdraft descends, a collapsing core with mid level convergence above the downdraft. For RFD forced events, watch for intensifying low-level mesos, large reflectivity in the hook, and a DCZ. These are some of the main detection signatures.

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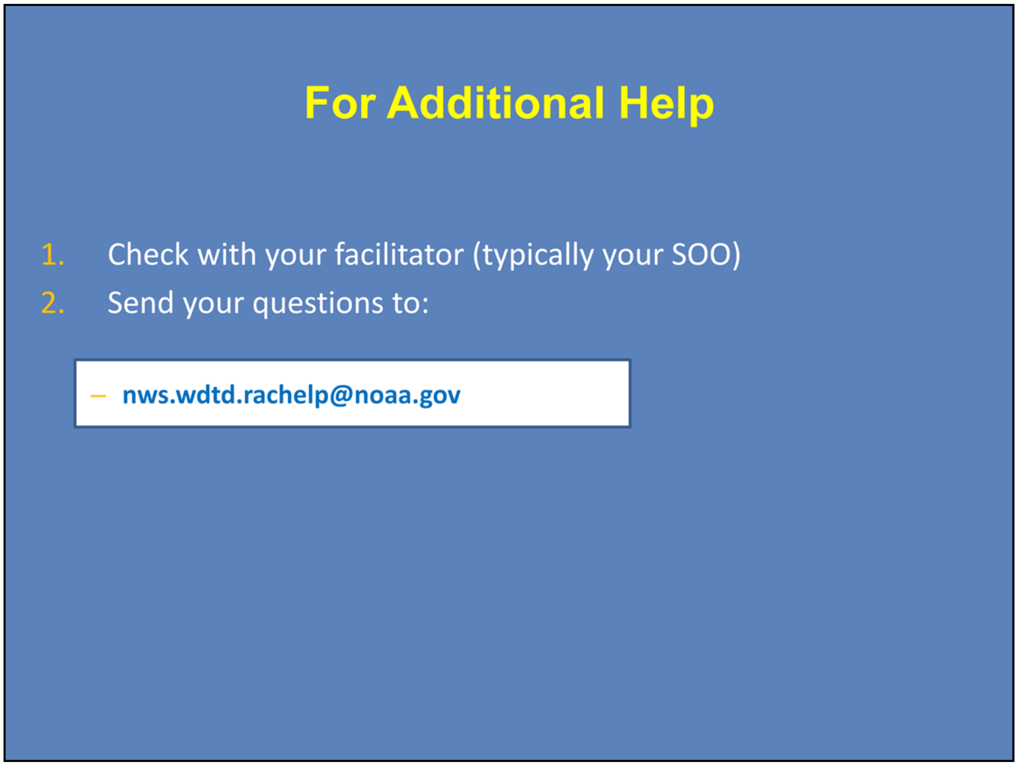

For additional help, check with your facilitator or send your questions to the listserv e-mail address here.

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Severe Hail

1. Introduction

Notes:

Welcome to the RAC Convective Storm Structure and Evolution lesson on Severe Hail.

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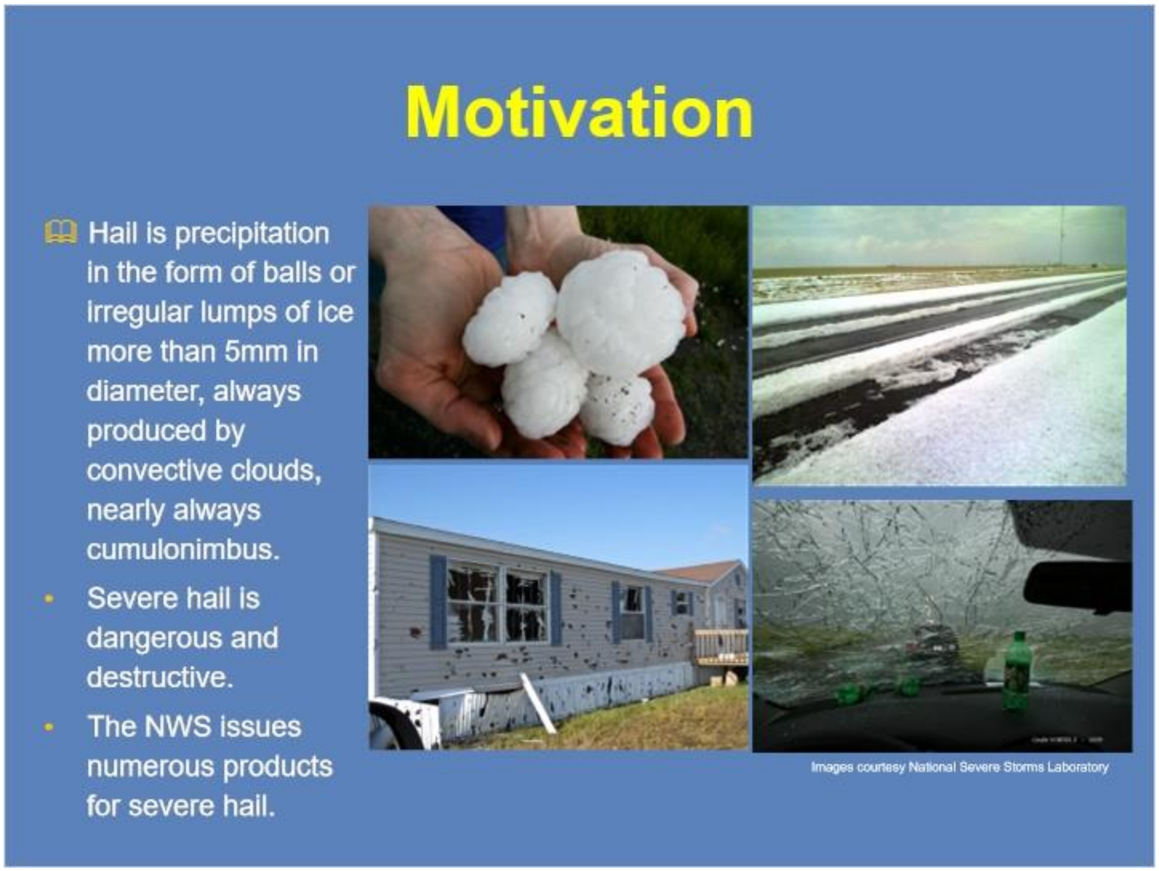
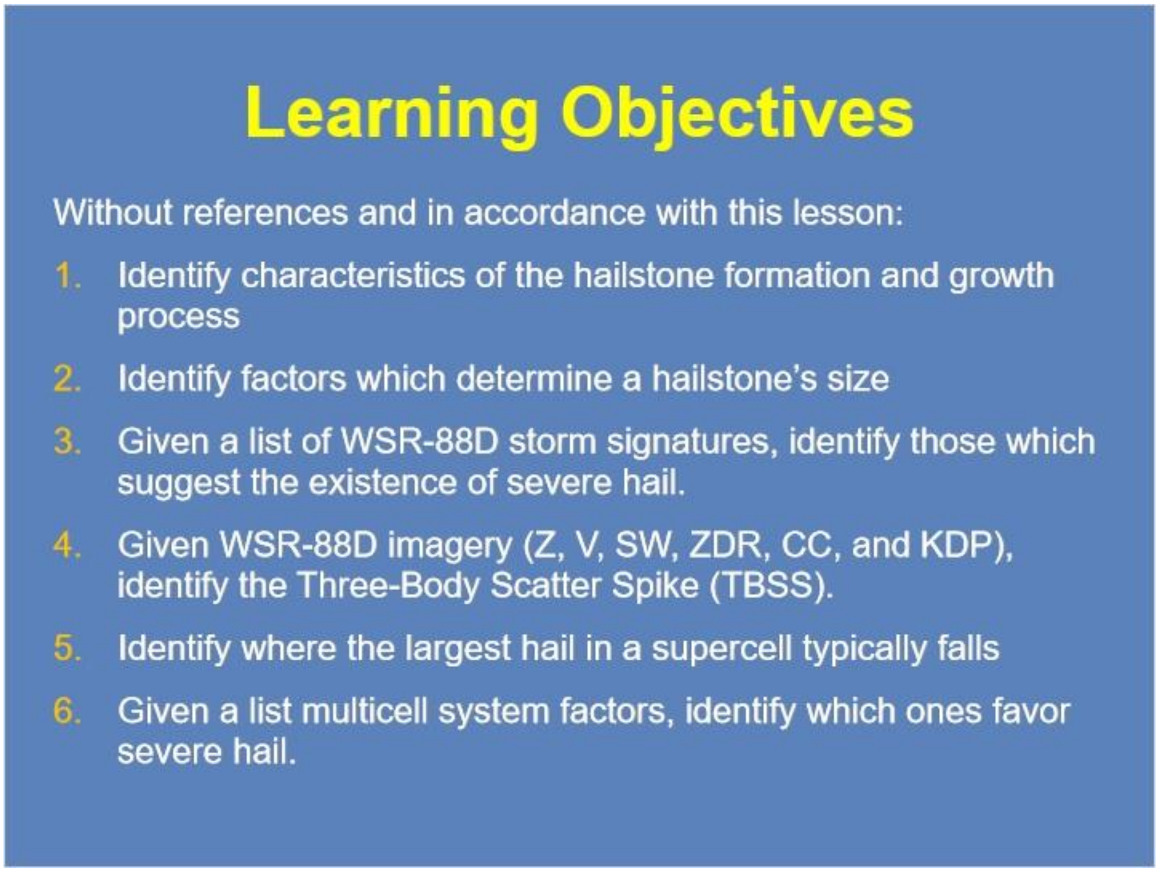


Notes:

Hail is precipitation in the form of balls or irregular lumps of ice more than 5mm in diameter, always produced by convective clouds, nearly always cumulonimbus. Hail can be deadly to people and livestock, and can damage property including homes, businesses, automobiles, and aircraft.

The National Weather Service issues numerous public and aviation products for severe hail. That’s why it’s important for you to learn about it.



Notes:

Here are the learning objectives. Please take a moment to read them.

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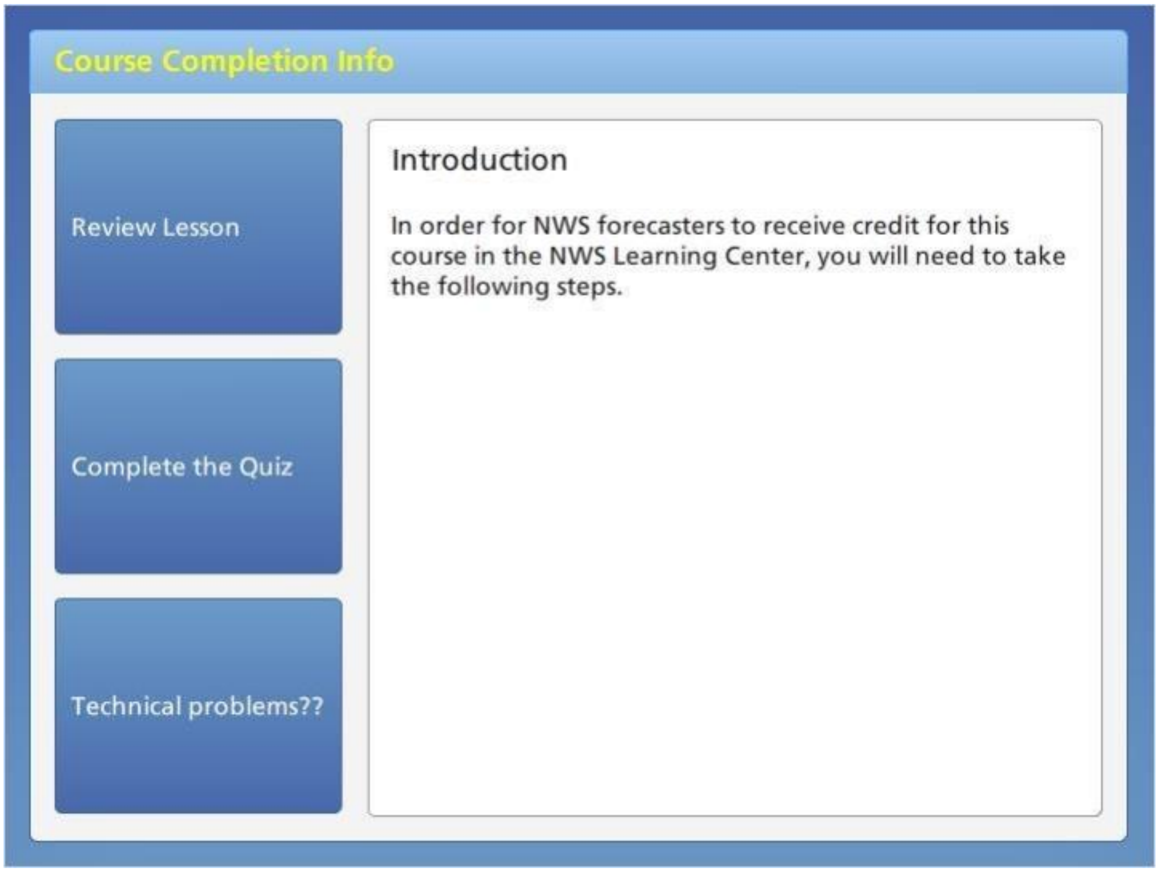
Course Completion Info
Review Lesson
Complete the Quiz
Technical problems??
Introduction
In order for NWS forecasters to receive credit for this course in the NWS Learning Center, you will need to take the following steps.

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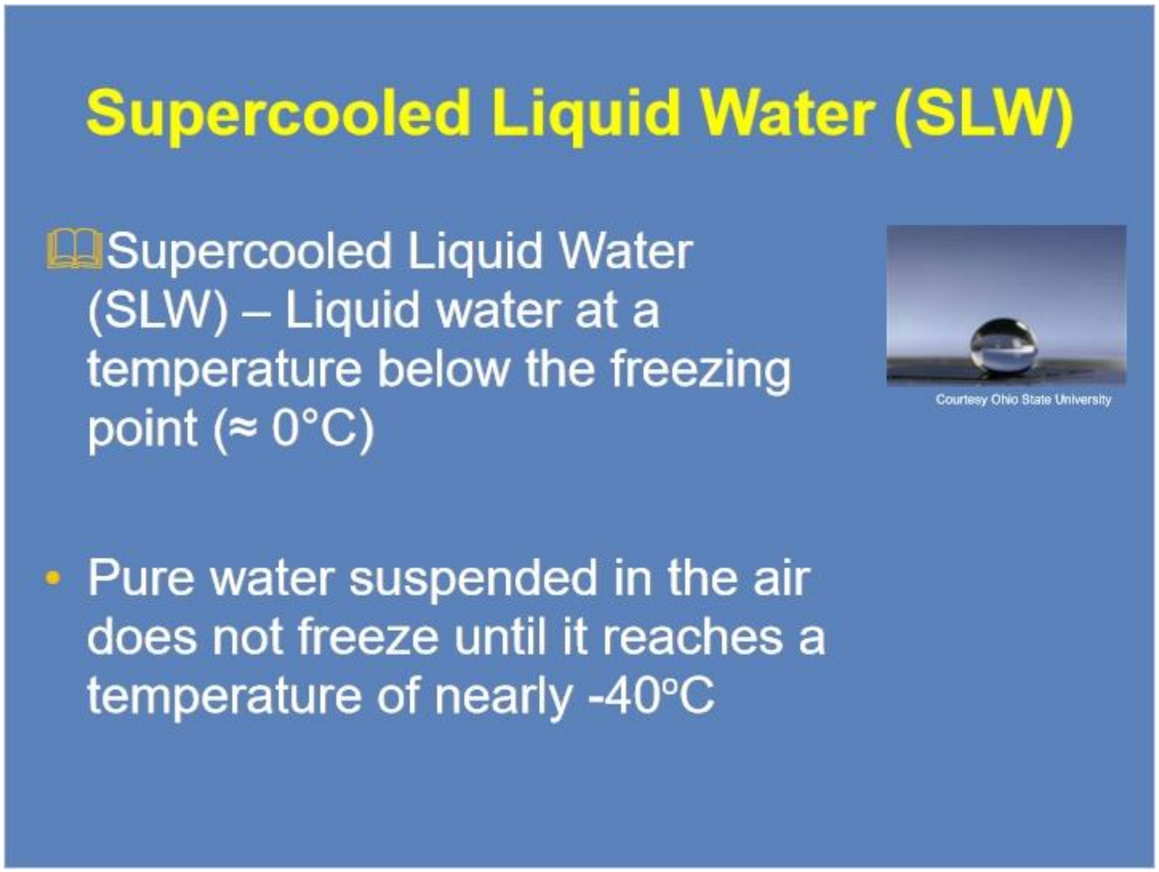
2. Formation and Growth



Notes:

Let's begin by discussing the formation source of hail which is Supercooled liquid water (SLW). This is liquid water at a temperature below the freezing point (~0°C). It can exist at temperatures below zero because freezing is a complex process. Pure water suspended in the air does not freeze until it reaches a temperature of nearly -40°C.

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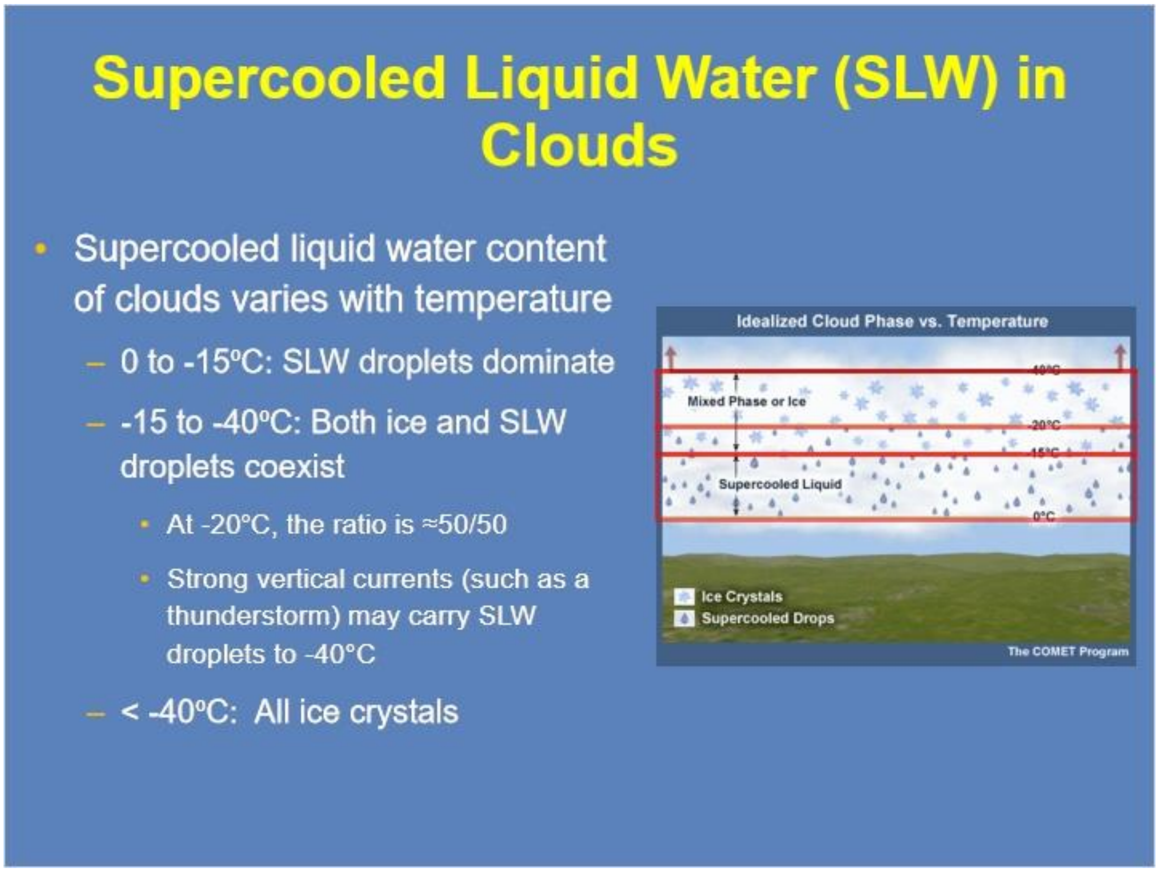

Notes:

The supercooled liquid water (SLW) content of clouds varies with temperature. Between 0^{o} to -15^{o}C, supercooled liquid water droplets dominate. Between -15^{o} to -40^{o}C, both ice and supercooled liquid water droplets coexist. At -20^{o}C, the ratio is about 50/50. However, strong vertical currents such as a thunderstorm may carry SLW droplets to -40°C. At temperatures colder than -40^{o}C, clouds consist entirely of ice crystals.

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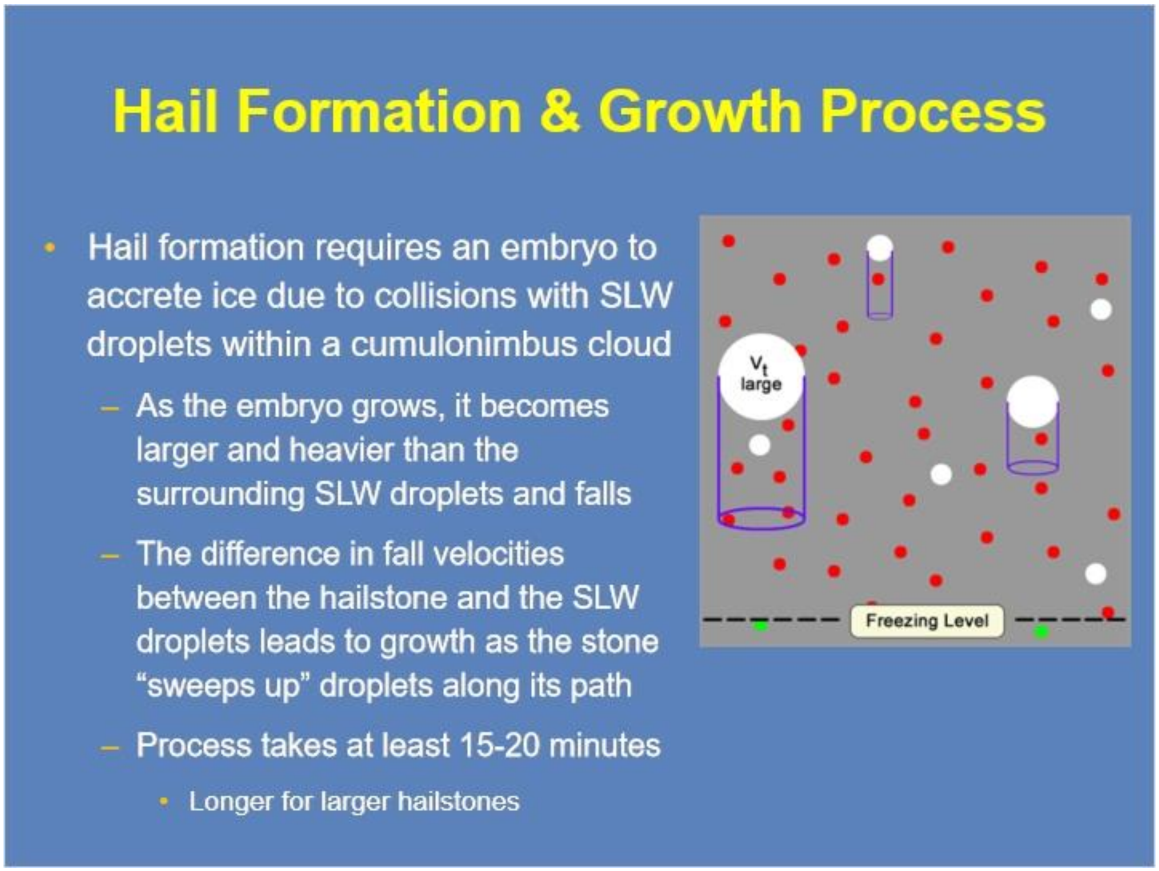
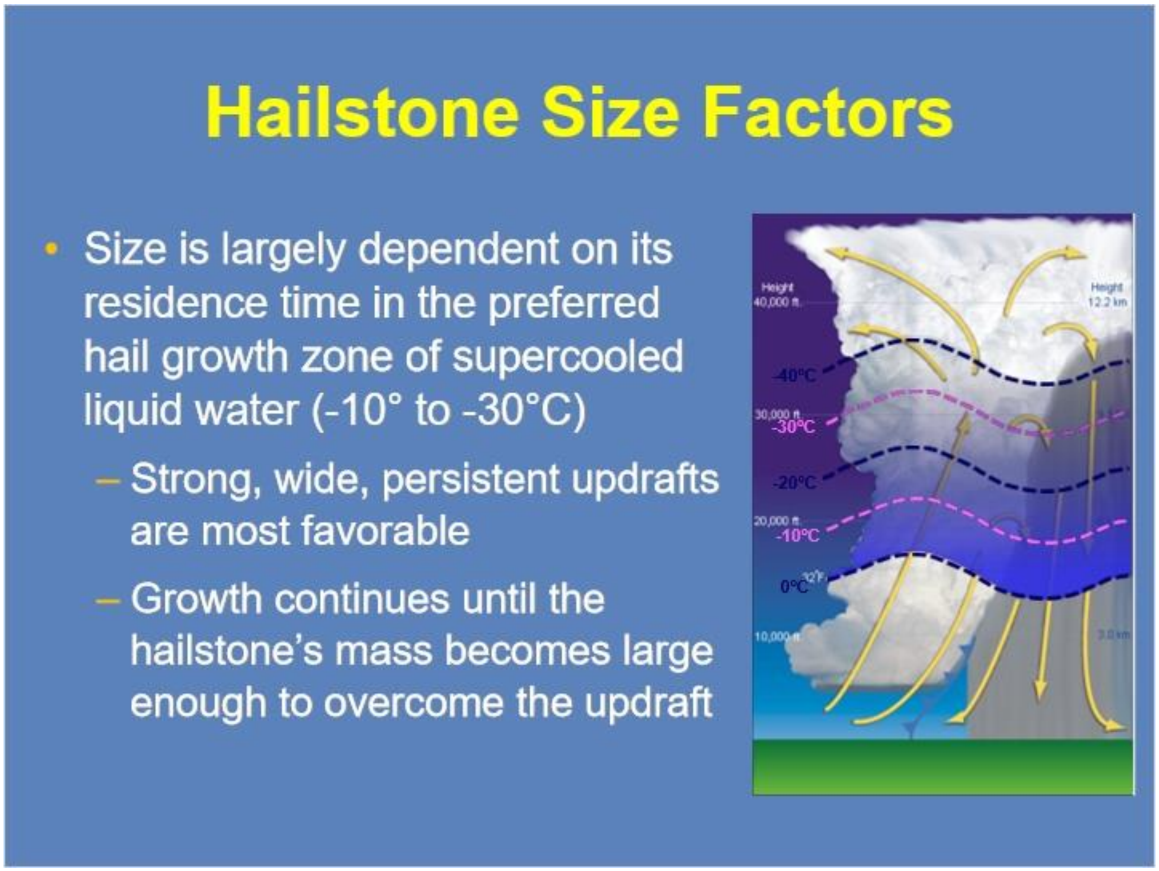


Notes:

Hail formation requires an embryo such as an ice crystal, frozen raindrop, dust or some other nuclei to accrete ice due to collisions with supercooled liquid water (SLW) droplets within a cumulonimbus cloud. As the embryo grows, it becomes larger and heavier than the surrounding supercooled liquid water droplets and falls (relative to the supercooled droplets). The difference in fall velocities between the hailstone and the SLW droplets leads to growth as the stone "sweeps up" droplets along its path. Process takes at least 15-20 minutes; longer for larger hailstones.



Notes:

A hailstone's size is largely dependent on its residence time within the preferred hail growth zone of supercooled liquid water between -10º to -30ºC. Strong, wide, persistent updrafts are most favorable. The growth rate is maximized near -13ºC and rapidly diminishes at temperatures approaching -30ºC as supercooled water droplets become rare at these colder temperatures. Growth continues until the hailstone's mass becomes large enough to overcome the attendant updraft.

Hail Size Descriptions

Description	Diameter	Updraft Speed
BB	< 0.25"	< 24 mph
Pea	0.25 – 0.375	24 mph
Small marble	0.5"	35 mph
Dime	0.7"	38 mph
Penny	0.75"	40 mph
Nickel	0.88"	46 mph
Quarter	*(Severe)* 1"	49 mph
Half Dollar	1.25"	54 mph
Walnut / Ping-Pong Ball	1.5"	60 mph
Golf Ball	1.75"	64 mph

Description	Diameter	Updraft Speed
Lime	*(Significant)* 2"	69 mph
Tennis Ball	2.5"	77 mph
Baseball	2.75"	81 mph
Large Apple	3"	84 mph
Softball	*(Giant)* 4"	98 mph
Grapefruit	4.5"	103 mph
CD / DVD	4.75 – 5"	105 mph

Modified from NWSI 10-1605, Table A3. Hail Conversion Table

Notes:

Here is a table which equates hail description with its diameter and approximate updraft speed. The National Weather Service criteria for severe hail is one-inch diameter (or about quarter-size), which is based on research that indicates this is the threshold at which significant damage occurs.

Storm spotters are taught to report the largest hailstone they observe. That's what WFOs log into their Local Storm Reports and is subsequently published in the National Climatic Data Center's official Storm Data publication.

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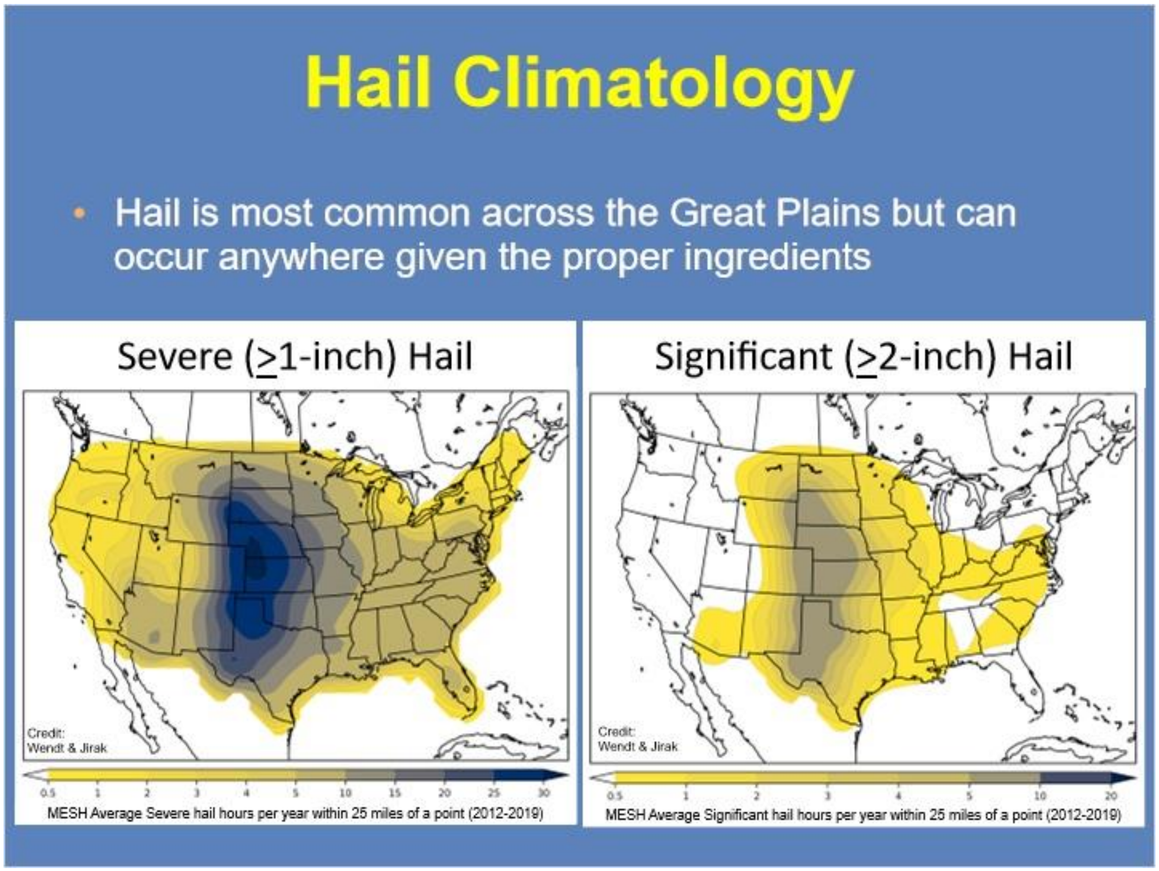

Notes:

Hail is most common across the Great Plains, but can occur anywhere given the proper ingredients. Please take a moment to view these two hail climatology graphics.

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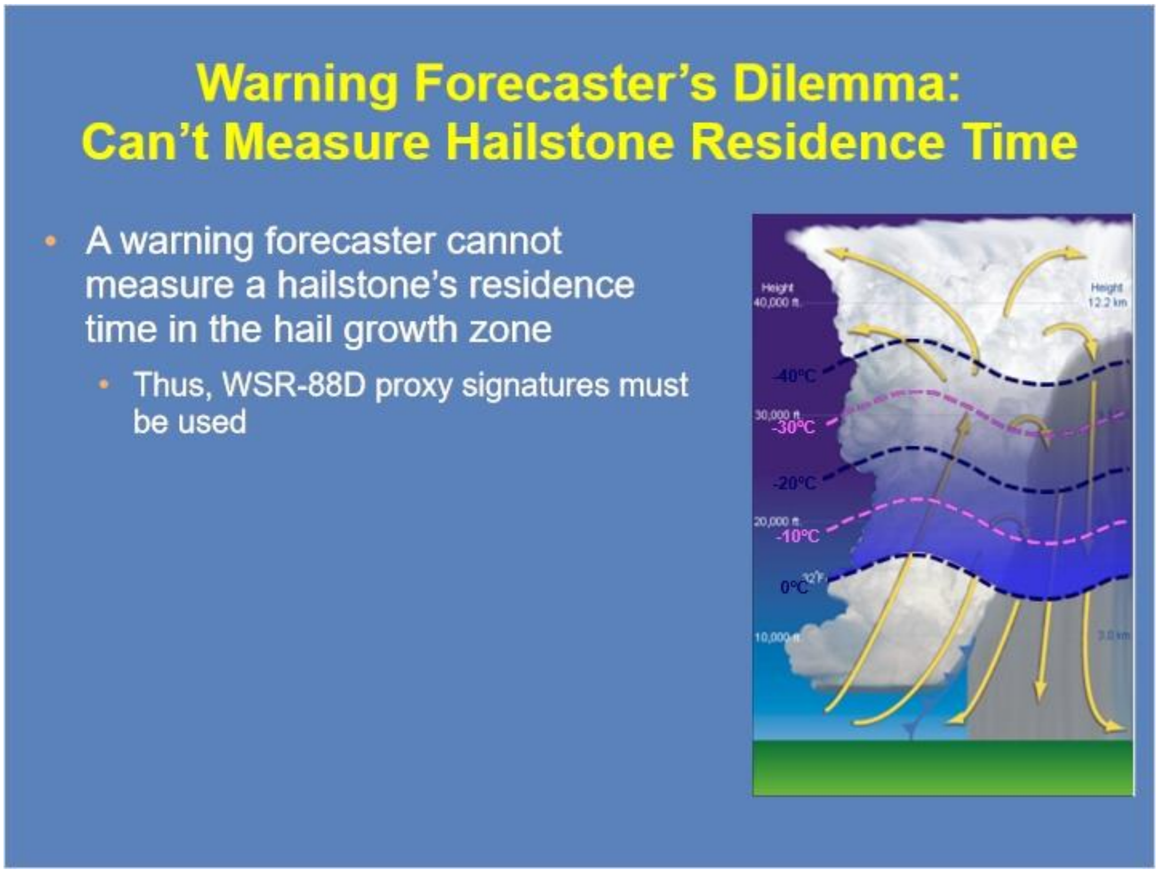
3. WSR-88D Detection



Notes:

A warning forecaster cannot measure a hailstone's residence time in the hail growth zone. It's difficult to estimate updraft strength, let alone perform a real-time trajectory analysis. Thus, WSR-88D proxy signatures must be used to detect severe hail.

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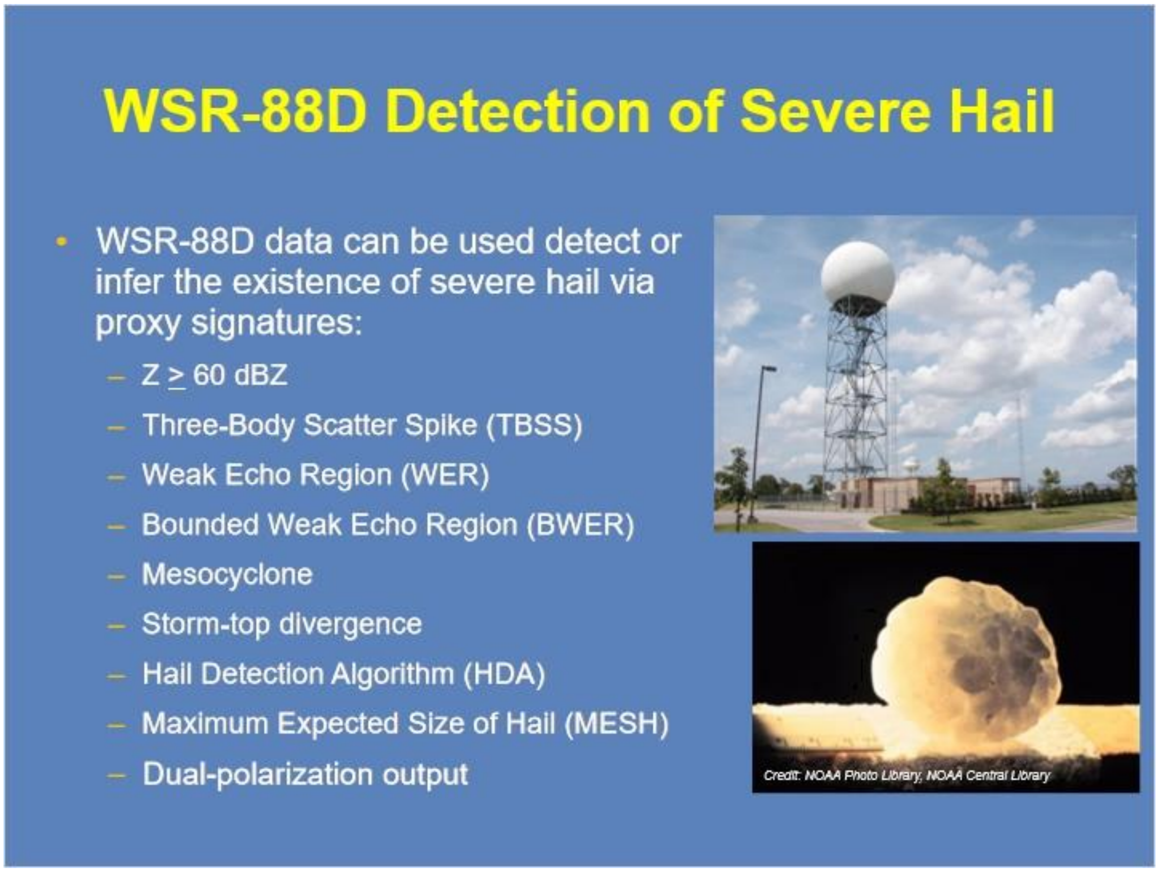
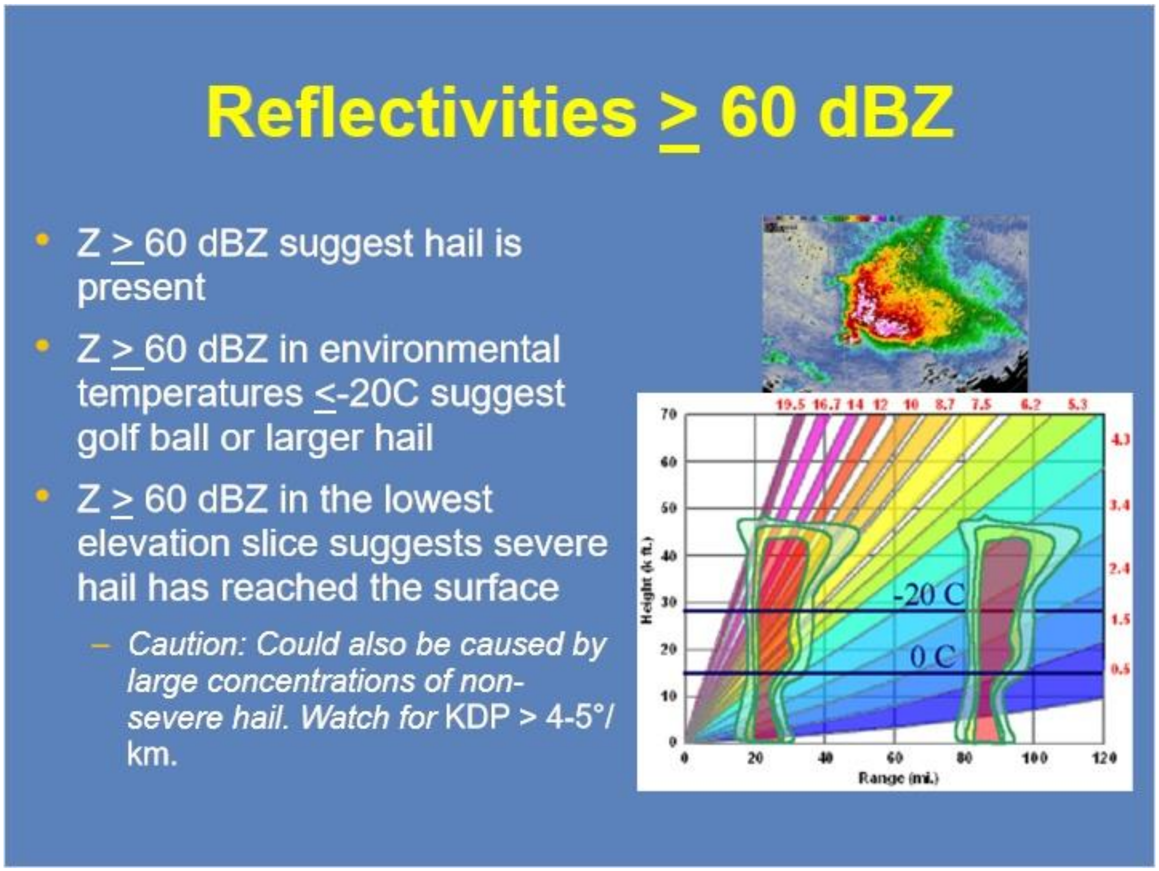


Notes:

WSR-88D data can be used to detect or infer the existence of severe (≥ 1-inch) hail via the proxy signatures listed here. Please take a moment to view them before we discuss them in detail.

Note: Dual-polarization radar hail analysis is covered in another lesson.



Notes:

Reflectivities greater than or equal to 60 dBZ suggest hail is present. This is because pure water is almost impossible when reflectivity is greater than 55 dBZ. Thus, you can be confident that hail is present.

Reflectivities greater than or equal to 60 dBZ in environmental temperatures less than or equal to -20°C suggest hail greater than or equal to golf ball size (1.75”). Hail this large is unlikely to melt before it reaches the surface regardless of the melting level height. Increase reflectivities in this layer and the odds of severe hail increase dramatically. Reflectivity greater than or equal to 65 dBZ through the entire preferred hail growth zone (-10°C to -30°C) suggests the potential for giant (≥4”) hail (Blair et al., 2011). Thus, it’s important to know the height of the -20°C level.

Reflectivities greater than or equal to 60 dBZ in the lowest elevation slice strongly suggest the presence of hail that has reached the surface. Beware that is could also be caused by large concentrations of sub-severe melting hail so watch out for KDP values greater than 4-5°/km. Also, beware that giant, very dry hail in low quantities have been observed with reflectivities between 35 and 50 dBZ. When this occurs, the hail is usually falling in the strong reflectivity gradient immediately adjacent to an inflow notch and updraft with a supercell storm.

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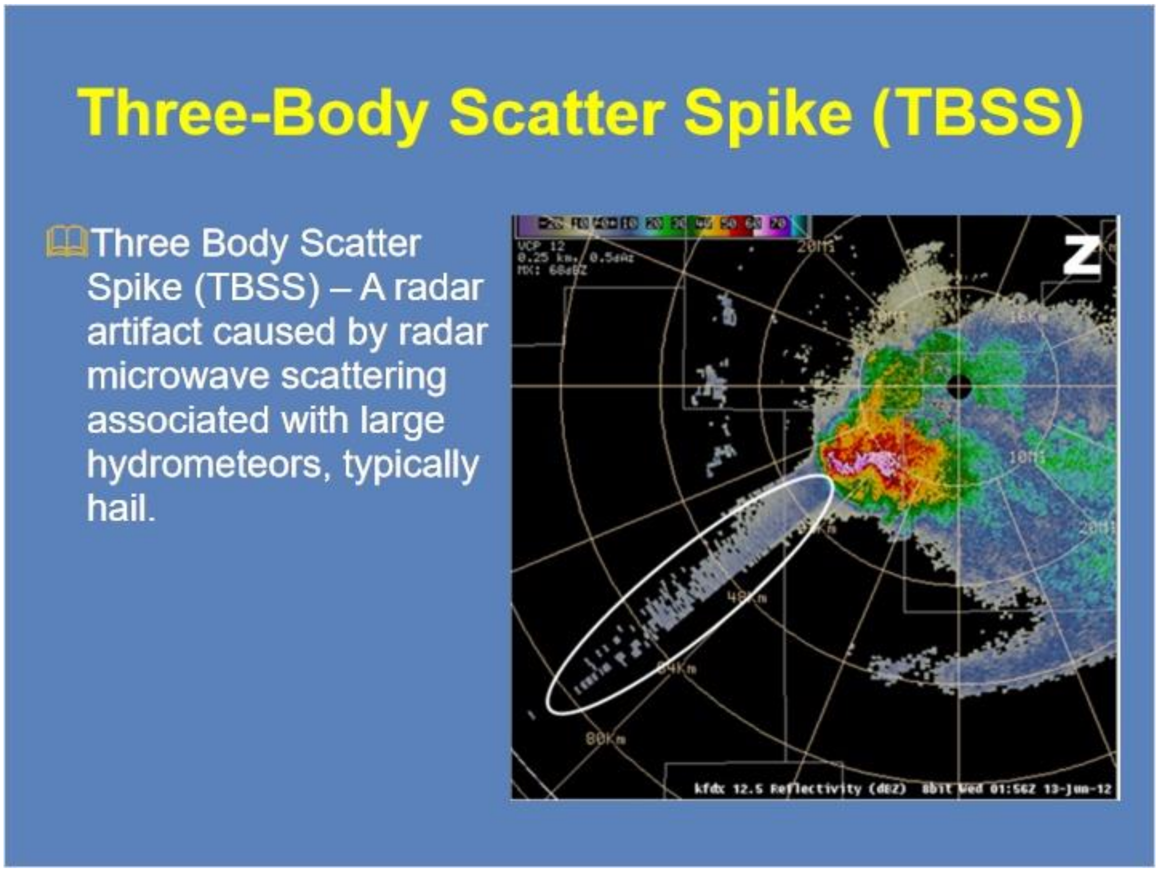

Notes:

A Three-Body Scatter Spike (TBSS) (also known as a “hail spike”) is a radar artifact caused by radar microwave scattering associated with large hydrometeors, typically hail. The TBSS is characterized by extremely high ZDR and very low CC located (radially) just behind high reflectivity cores in hail-bearing storms.

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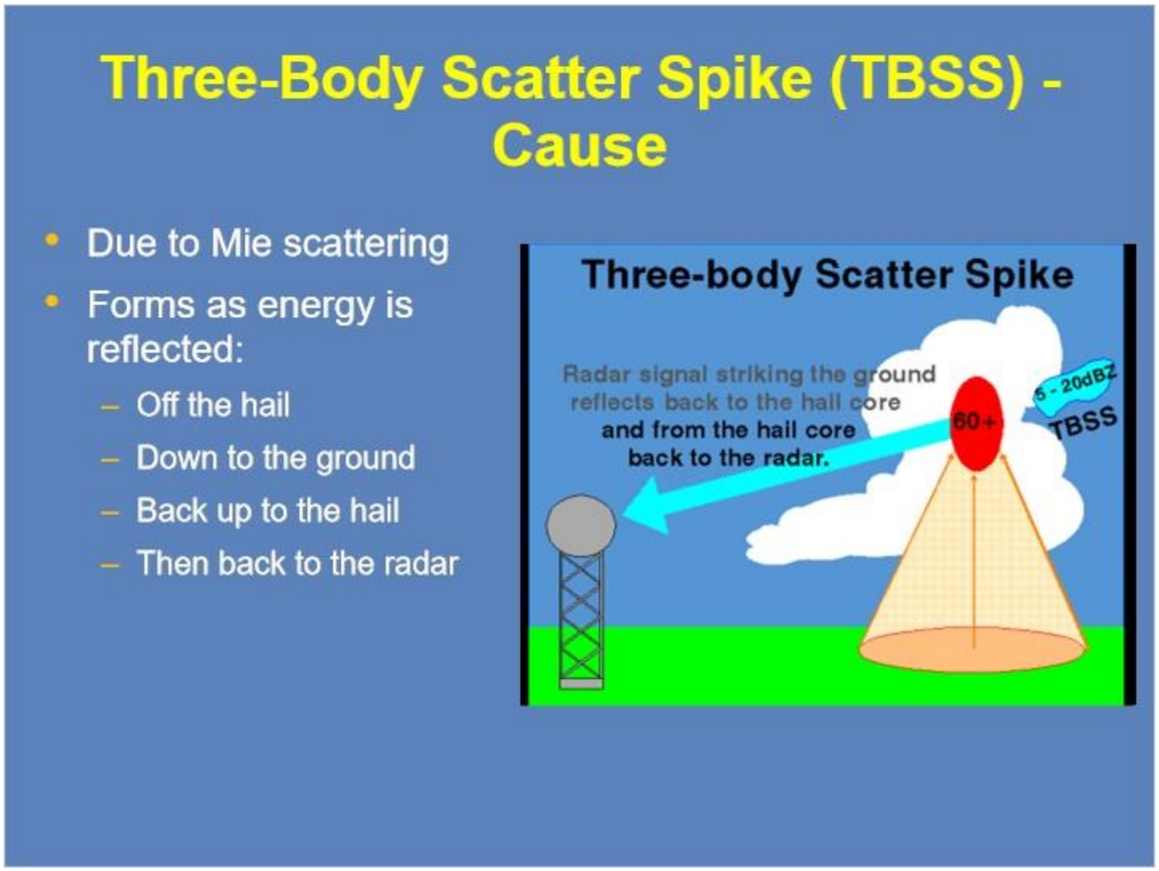
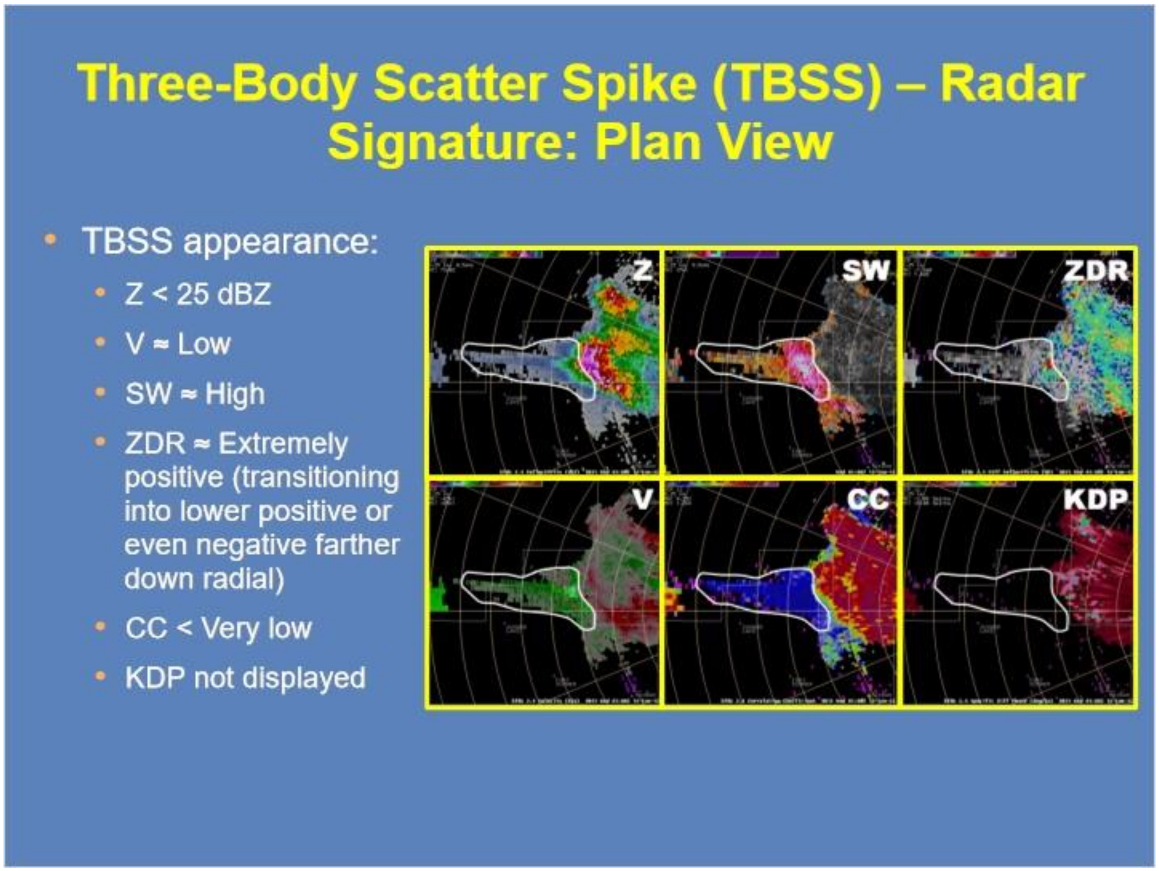


Notes:

The TBSS is an artifact of the electromagnetic radar beam being subject to “Mie scattering” instead of the usual “Rayleigh scattering” process.

A TBSS forms as incident energy from the radar is reflected off the hail, down to the ground, then back up to the hail and back to the radar. Because of the delay in reception of the pulses, the radar circuitry displays the TBSS as downrange from the hail core.



Notes:

The TBSS signature produces low reflectivities (generally less than 25 dBZ), low radial velocities (V), and high spectrum widths (SW). In ZDR, the TBSS appears as an area of extremely positive values just down-radial of the hail core (transitioning into lower positive or even negative values farther down-radial). In CC, the TBSS shows up very clearly as a spike of extremely low values (generally < 0.5), on the down-range side of the hail core. CC is especially useful in cases when the TBSS reflectivity signature is short or masked by precipitation echoes. The spike is not seen in KDP because of the 0.90 CC threshold filter discussed in previous lessons.

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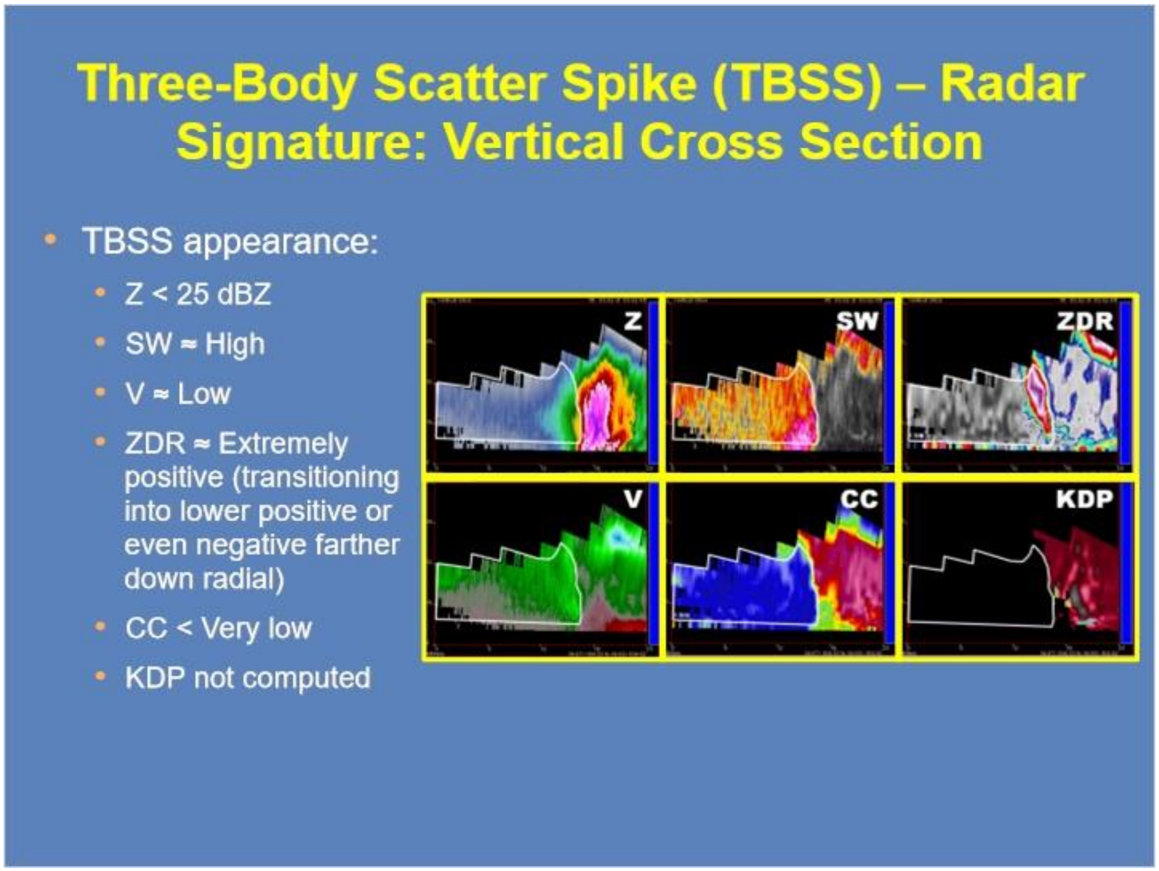

Notes:

Taking a look at cross sections of the same storm, the TBSS is seen down-radial of the high reflectivity (Z) core. High spectrum widths (SW) and low radial velocities (V) exist down-radial from the hail core. In ZDR, the area of extremely high, positive values located immediately behind the hail core is usually somewhat wedge-shaped, while farther down-radial, there is a transition to negative values of ZDR. In CC, the TBSS is marked by very low values. The spike is not seen in KDP because of the 0.90 CC threshold.

The strength and length of the TBSS is related to the intensity and vertical extent of the reflectivity core. Therefore, a TBSS should be easier to detect with a more intense and elevated reflectivity core. Also, the larger the highly reflective core area, the more extensive the TBSS.

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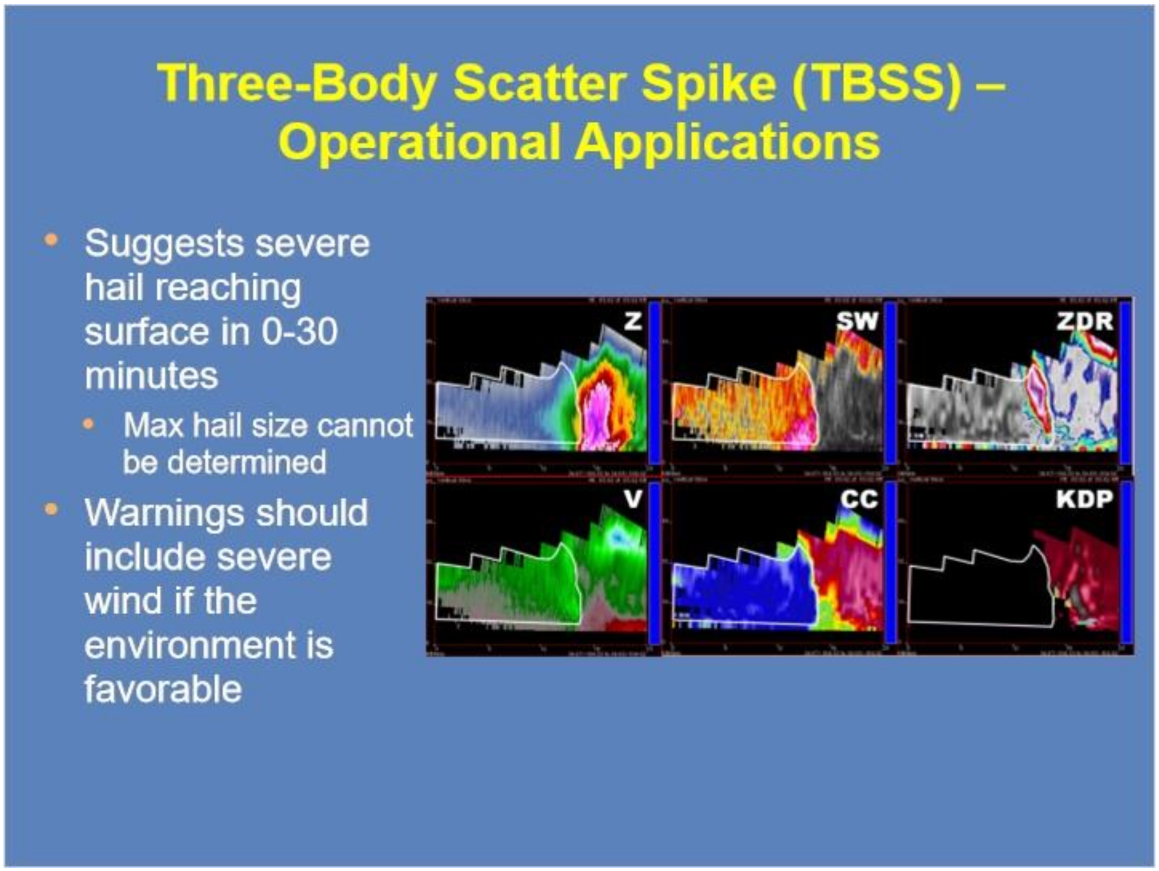

Notes:

For S-band (10 cm) radar such as the WSR-88D, the presence of a TBSS generally suggests the thunderstorm possesses severe hail which is either reaching the surface or will do so within the next 10-30 minutes. It is likely that the hailstones responsible for the three-body scatter spike (TBSS) range from around 0.8 to 2 inches in diameter. However, the largest hailstones probably do **not** contribute to the three-body scatter spike in many cases, making it difficult to gauge maximum hail size. Additionally, because storms with a TBSS often produce damaging surface winds, resulting warnings should contain mention of severe wind if the near storm environment is favorable.

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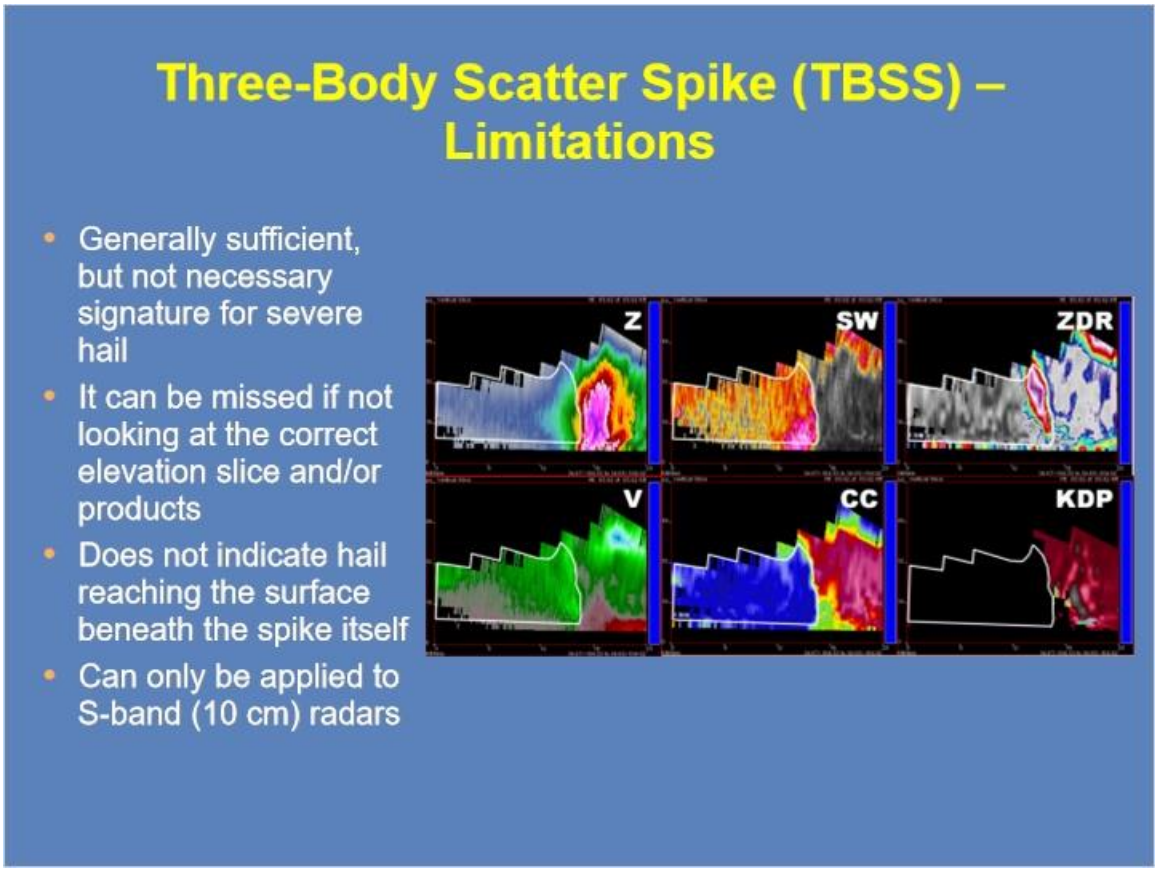
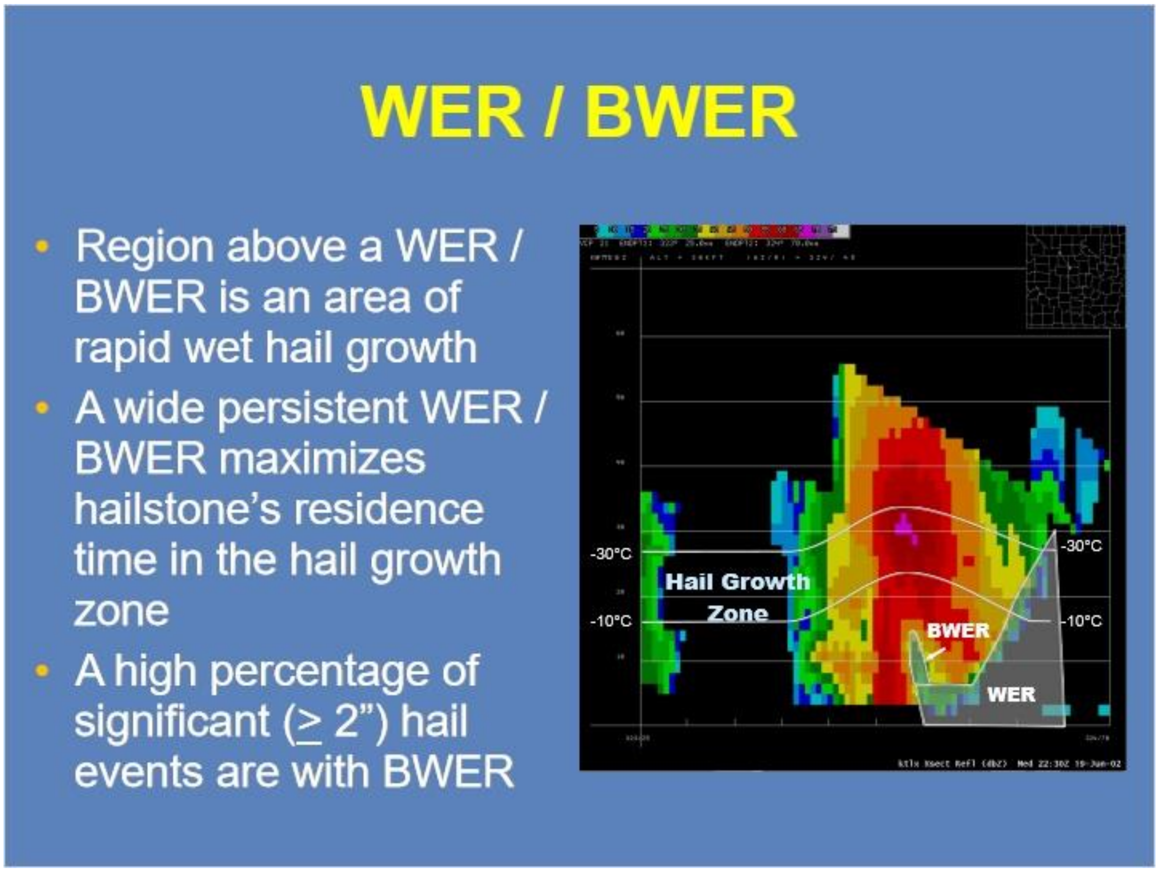


Notes:

Be aware of the limitations of using the TBSS signature. It is a generally sufficient, but not necessary signature for severe hail identification. It can be missed if not looking at the correct elevation slice and/or products. This signature is an artifact of Mie scattering and must not be construed as hail actually reaching the surface *beneath the echo spike itself*. The TBSS signature can only be applied on S-band (10 cm) radars, such as the WSR-88D. On C-band (5 cm) radars, the TBSS can be related to large raindrops rather than hail.



Notes:

Studies have shown that the high reflectivity region above the top of the Weak Echo Region (WER) and Bounded Weak Echo Region (BWER) is an area where rapid wet hail growth is occurring in the core of an intense updraft. The existence of a WER/BWER suggests that these hailstones are subject to a massive influx of supercooled cloud water and growth, especially if in the favored hail growth zone (-10ºC to -30ºC). A wide, persistent WER/BWER helps to maximize a hailstone's residence time in the favored hail growth zone before it cascades into the core. A high percentage of significant (2-inch diameter and larger) hail events are associated with BWERs. Remember that a bona fide WER/BWER must be topped by intense reflectivities in order for it to be associated with updraft.

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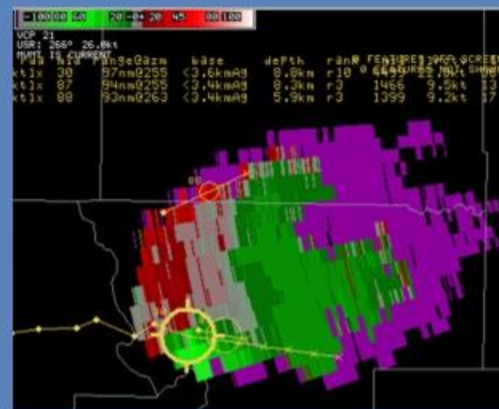
Mesoscyclone

- A strong, persistent mesocyclone is a strong indicator of severe hail
- Larger mesos are more favorable
  - Provides a larger hail growth zone
- A very high percentage of significant (>2”) hail is produced by supercells
- Strong, persistent mesoanticyclones (left-moving supercells) are often prolific producers of severe hail



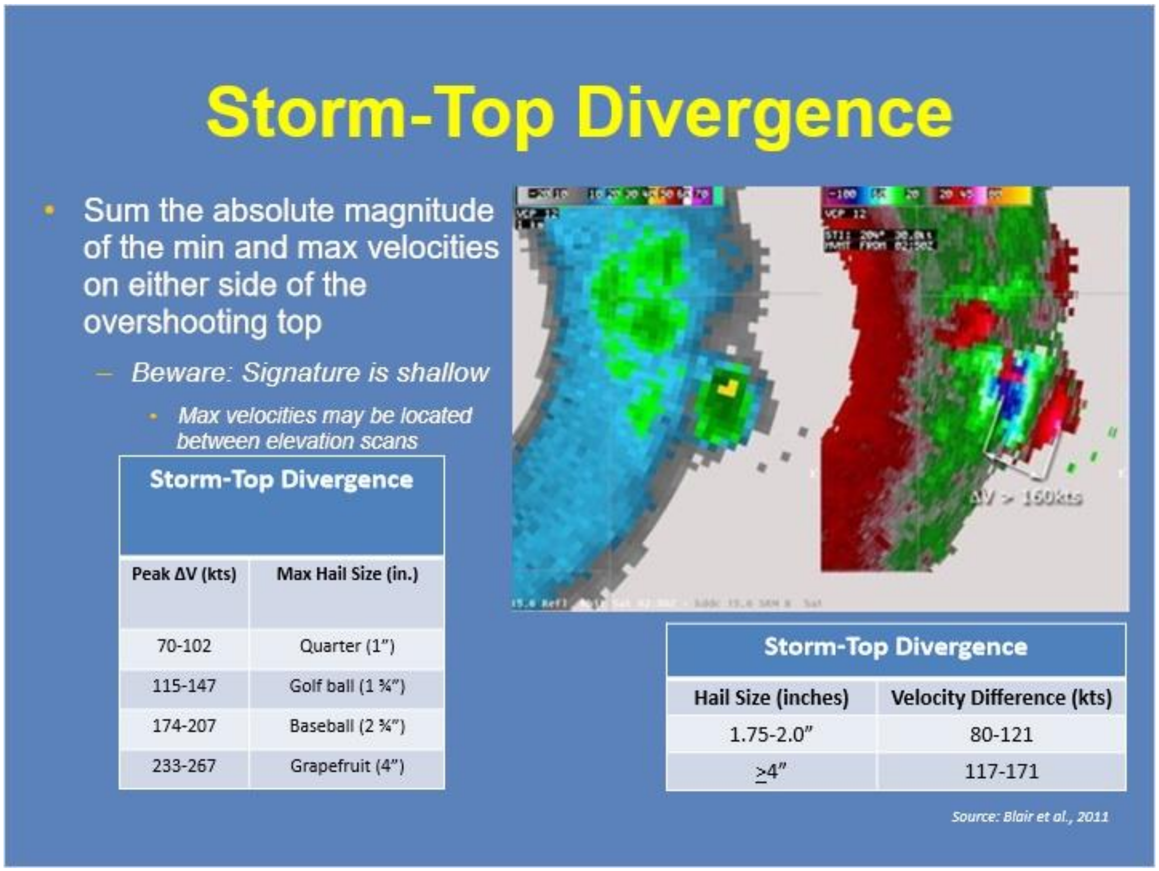
Mesocyclone	
Hail Size (in.)	Peak Rotational Velocity (kt)
1.75” to 2.00”	27-41
>4”	39-56

Notes:

Severe hail can be inferred from velocity signatures as well.

A strong, persistent mesocyclone, the defining characteristic of a supercell, is a strong indicator of severe hail. Dynamic pressure drops, especially in a strong mid-level mesocyclone, accelerate the updraft in the hail growth zone. It is hypothesized that mesocyclones produce favorable trajectories that lead to enhanced hail growth. Larger diameter mesocyclones are more favorable for severe hail because they provide a larger hail growth zone which also increases residence times and hail growth potential. Mesocyclone strength is especially important for the growth of large hail. A high percentage of significant (>2”) and giant (>4”) hail is produced by supercells.

Strong, persistent mesoanticyclones (left-moving supercells) are often prolific producers of severe hail. This may be because they move faster and cover more area than right-movers.



Notes:

Storm-top divergence can be used to assess a storm's maximum potential hail size. An example of this storm-top divergence technique can be seen here. First, use the proper elevation slice to sample the storm's overshooting top. Sum the absolute magnitude of the minimum and maximum velocities found on either side of the overshooting top and you'll get a representative sample of storm top velocity difference, in this case > 160 kts. If you cannot sample the overshooting top as shown here, then the technique may fail. There are other factors that may cause this technique to fail (e.g., poor data quality, mini-supercells, etc.), so use with caution.

Two tables relating maximum storm top divergence velocity difference to hail size are shown here.

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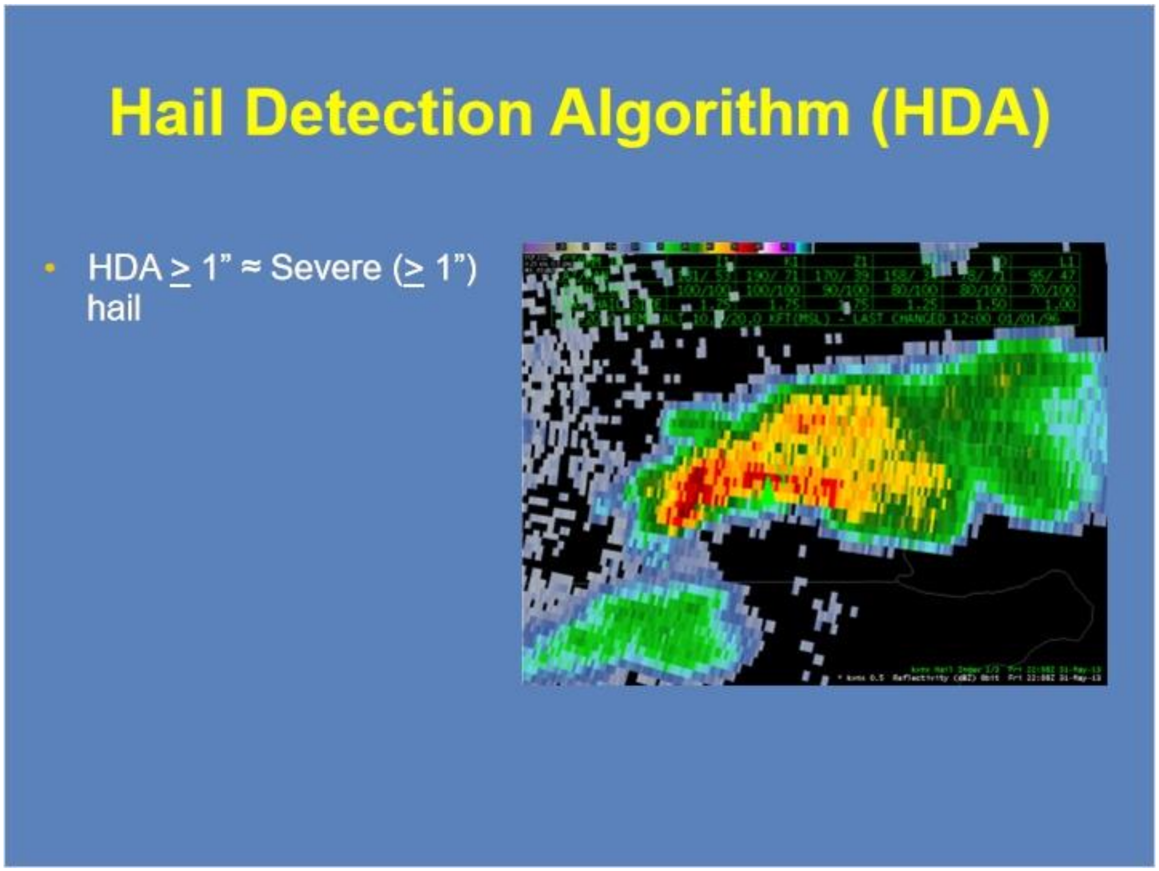

Notes:

The Hail Detection Algorithm (HDA) can be used to infer the existence of severe hail. Values greater than or equal to 1-inch suggest the existence of severe hail.

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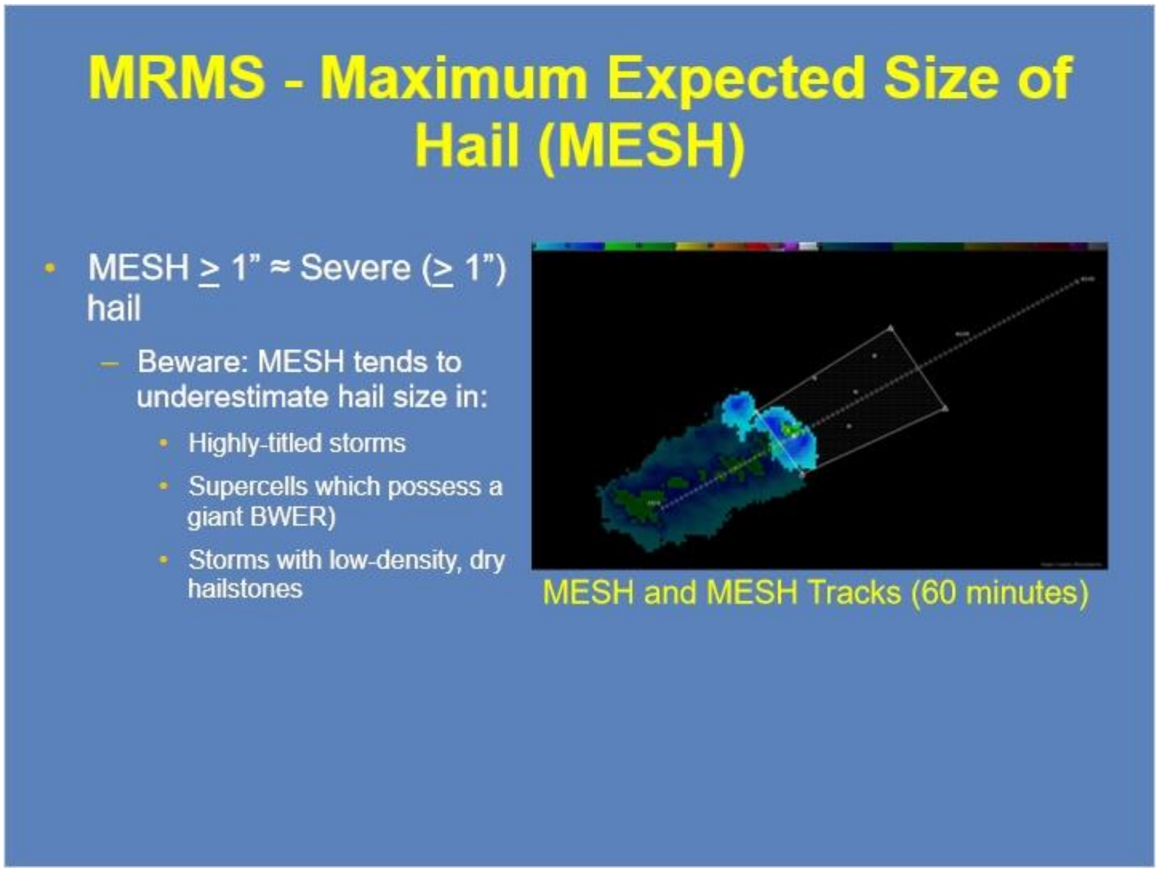
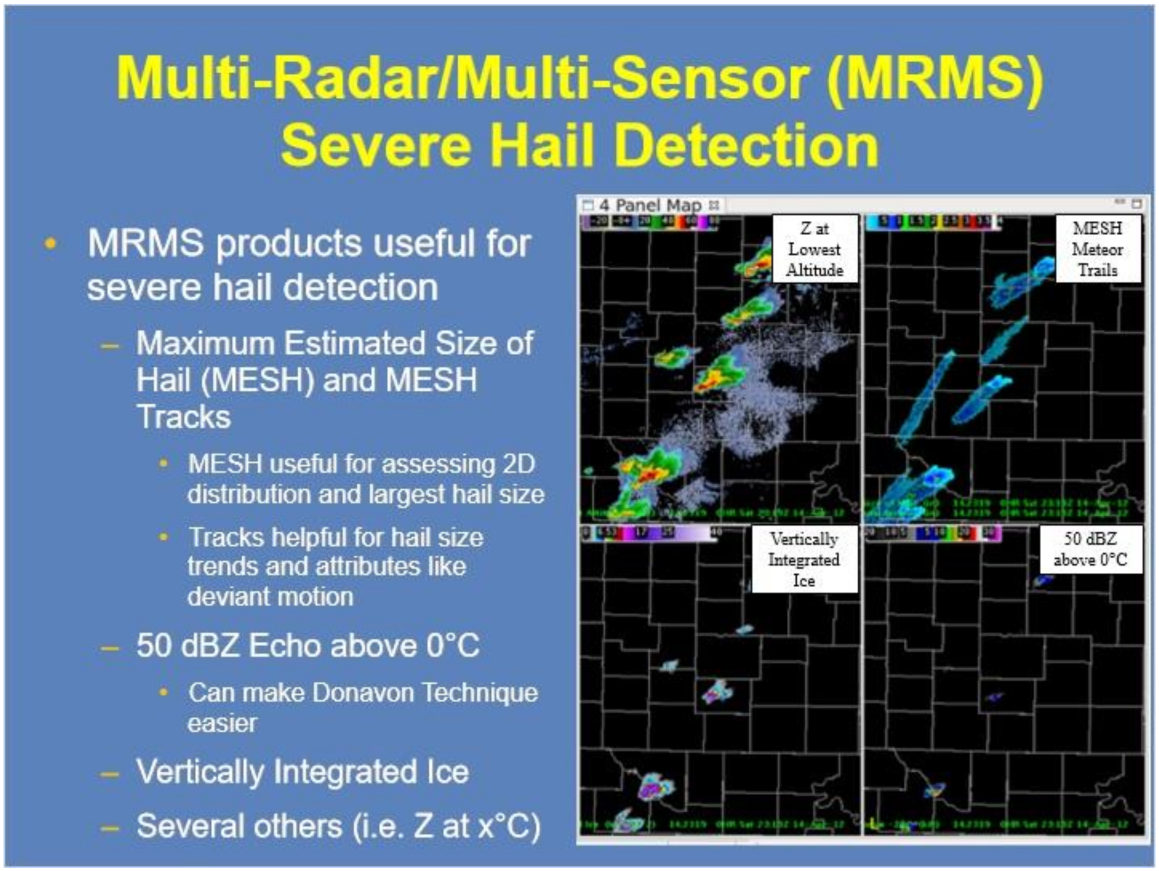


Notes:

The Multi-Radar/Multi-Sensor (MRMS) Maximum Expected Size of Hail (MESH) product (and its accompanying MESH Tracks products) can be used to infer the existence of severe hail. Values greater than or equal to 1-inch suggest the existence of severe hail.

Beware, MESH tends to underestimate hail size in: Highly-tilted storms, supercells which possess a giant Bounded Weak Echo Region (BWER), and storms with low-density, dry hailstones.



Notes:

Multi-Radar/Multi-Sensor (MRMS) products are useful for severe hail detection. Perhaps the most useful is Maximum Estimated Size of Hail (MESH), and its associated MESH Tracks products. MESH has been shown to be very useful for assessing both the 2D distribution of hail and largest hailstone size associated with a storm. MESH Tracks products are useful for assessing both storm intensity trends and deviations in storm motion. Forecasters have discovered the benefits of image combining the MESH on top of the MESH Tracks products to create a MESH “Meteor Trails” display which is useful for the orientation of warning polygons.

Another group of MRMS products that can be used to detect severe hail are the reflectivity thickness products. In particular, the 50 dBZ Echo above 0°C product can help you to streamline the Donavon Technique.

MRMS Vertically Integrated Ice (VII) can be used to assess storm severity including hail potential. It can also be used to assess changes in updraft intensity. Sudden increases (decreases) in VII often occur when an updraft is intensifying (weakening). It can be used to identify the region of a storm where new cell growth is occurring which is particularly useful for warning polygon orientation.

Several other MRMS products can be useful for severe hail detection, including the Reflectivity at x°C products. Make sure you keep a single-site radar display visible when using MRMS hail products during warning operations.

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Notes:

Giant (≥4-inch diameter) hail is a relatively rare phenomenon, accounting for less than 1% of all hail reports in the United States.

Blair et al. (2011) examined several radar signatures to assess their utility in discriminating storms most favorable for giant hail. It was found that virtually all giant hail-producing storms were supercells with a well-organized structure. They were characterized by median values of mesocyclone rotational velocitiy (Vr) of 47 kts (24 m s^{-1}), storm-top divergence value of 140 kt (72 m s^{-1}), 50-dBZ echo top of 43,000 ft (13,100 m), and 60-dBZ echo top of 34,800 ft. (10,600 m). Median reflectivies ≥65 dBZ were present through the entire vertical depth of the preferred hail growth zone (-10° to -30°C), suggesting that high reflectivity should reside throughout that layer with giant-hail producing supercells.

They also noted that: Vertically Integrated Liquid (VIL)-based products, maximum reflectivity within the storm, and reflectivity within the preferred hail-growth zone showed little to no skill in discriminating between giant and smaller hail sizes.

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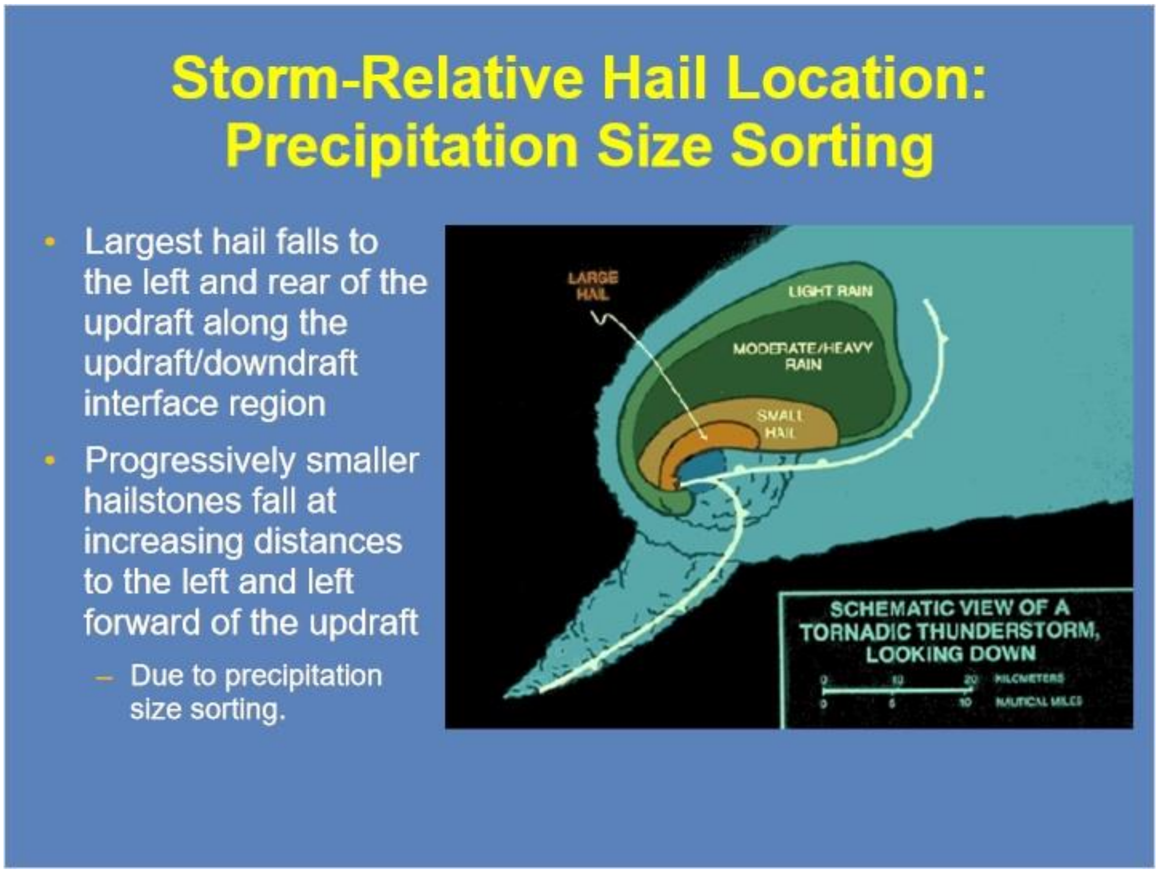
Note: Anecdotal evidence suggests there are at least some giant hailstones produced by elevated, non-supercell thunderstorms.



Notes:

Remember to always keep the radar products in context with the storm features, especially the updraft location as denoted by the Weak Echo Region (WER) and Bounded Weak Echo Region (BWER). For a cyclonic (right-moving) supercell in the northern hemisphere, the largest hailstones typically fall to the left and rear of the supercell's updraft (relative to its movement), along the updraft/downdraft interface region denoted by the strong low-level reflectivity gradient, inflow notch, and hook echo. Progressively smaller hailstones fall at increasing distances to the left and left forward of the updraft due to precipitation size sorting.

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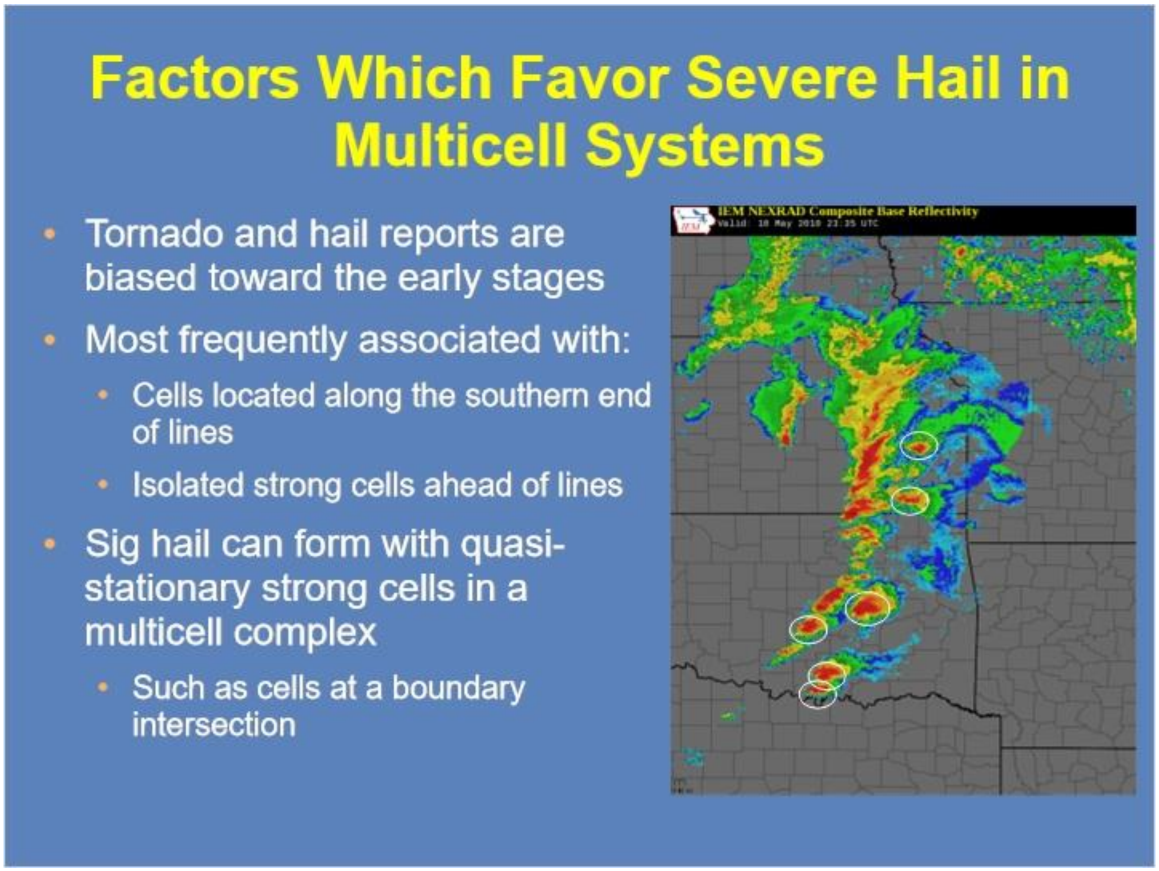
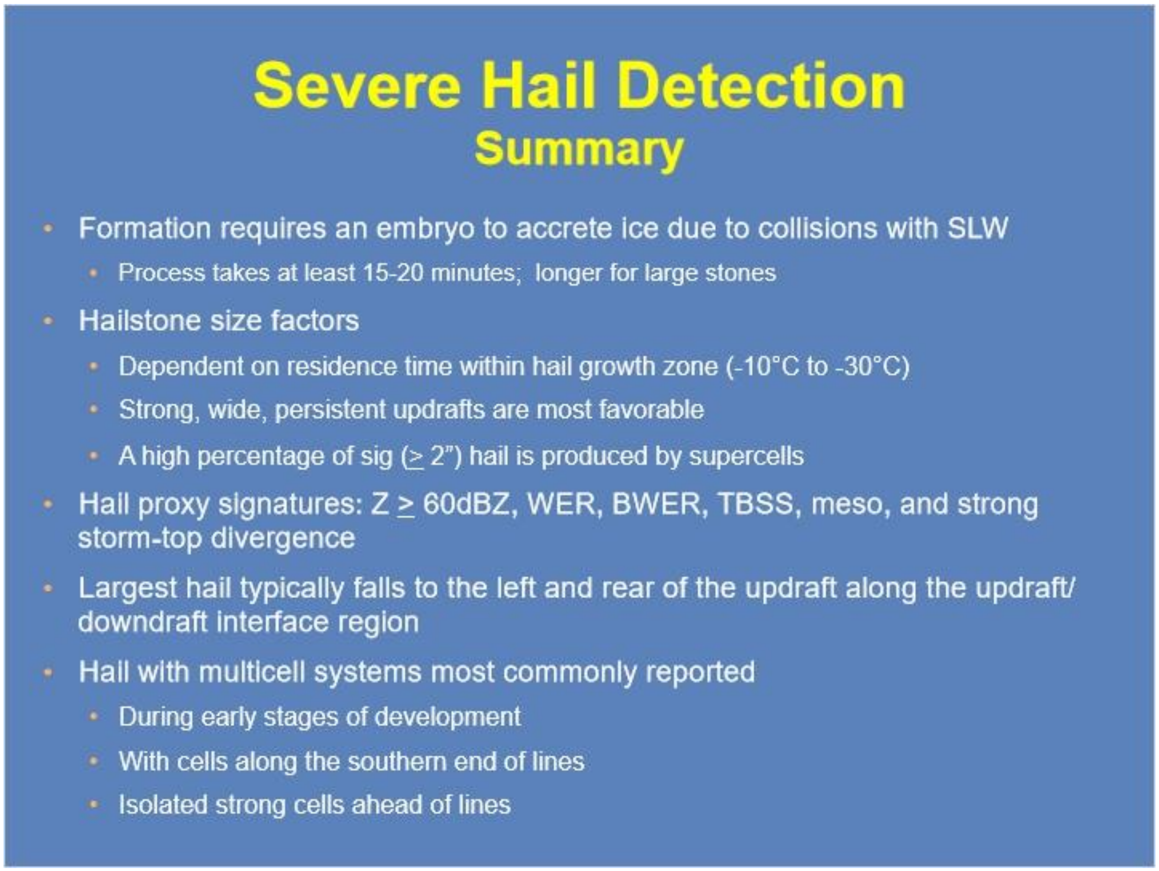


Notes:

Houze et al. (1990) documented severe weather locations for various mesoscale precipitation systems and found that tornado and hail reports were biased toward the early stages of multicell system development. They were most frequently associated with 1) cells located along the southern end of squall lines and 2) isolated strong cells ahead of the squall lines. This contrasts with high wind reports which are sometimes reported with isolated cells but are more numerous along well-developed convective lines. As multicell systems intensify, the effects of the cold pool and resulting increasing rear-to-front flow in the system tend to force an upright updraft along the leading edge. Any significant hail fall will likely occur in this region, not in the downdraft region or wake of the multicell system, which becomes dominated by cooler, saturated air.

Significant hail can occasionally form with quasi-stationary strong cells in a multicell complex, such as cells which form in the vicinity of a surface boundary where strong low-level convergence is focused near the updraft region of the complex.



Notes:

Hail formation requires an embryo to accrete ice due to collisions with supercooled liquid water (SLW) droplets within a cumulonimbus cloud. The process takes at least 15-20 minutes; longer for larger hailstones.

A hailstone’s size is dependent on its residence time within the preferred hail growth zone of supercooled water between -10ºC to -30ºC. Strong, wide, persistent updrafts are most favorable. A very high percentage of significant (≥2-inch) and virtually all giant (≥4-inch) hail events are produced by supercells.

WSR-88D signatures can be used to detect or infer the existence of severe hail including: Reflectivities ≥60 dBZ, Weak Echo Region (WER), Bounded Weak Echo Region (BWER), Three-Body Scatter Spike (TBSS), Mesocyclone, and strong storm top divergence. Hail detection is more robust with the inclusion of dual-pol data the inclusion of which can be used to detect hail type. *Note: Dual-polarization radar hail analysis is covered in another lesson.*

The largest hailstones typically fall to the left and rear of a supercell updraft, along the updraft/downdraft interface region denoted by the strong low-level reflectivity gradient, inflow notch, and sometimes hook echo.

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Hail reports with multicell systems are biased toward the early stages of development and are most frequently associated with cells located along the southern end of squall lines and isolated strong cells ahead of the squall lines.

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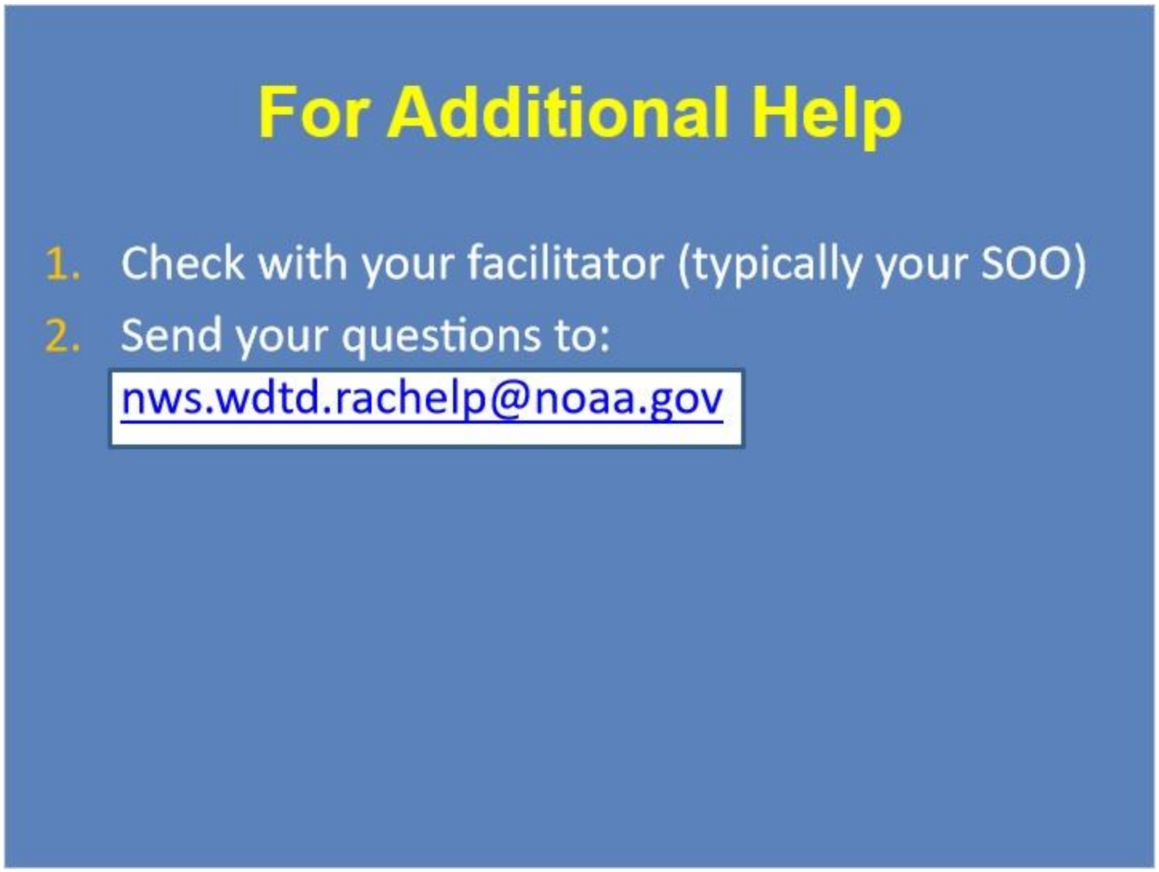

Notes:

For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.

1. Introduction

Notes:

Welcome to the RAC Convective Storm Structure and Evolution lesson on Dual-Pol Hail Analysis.

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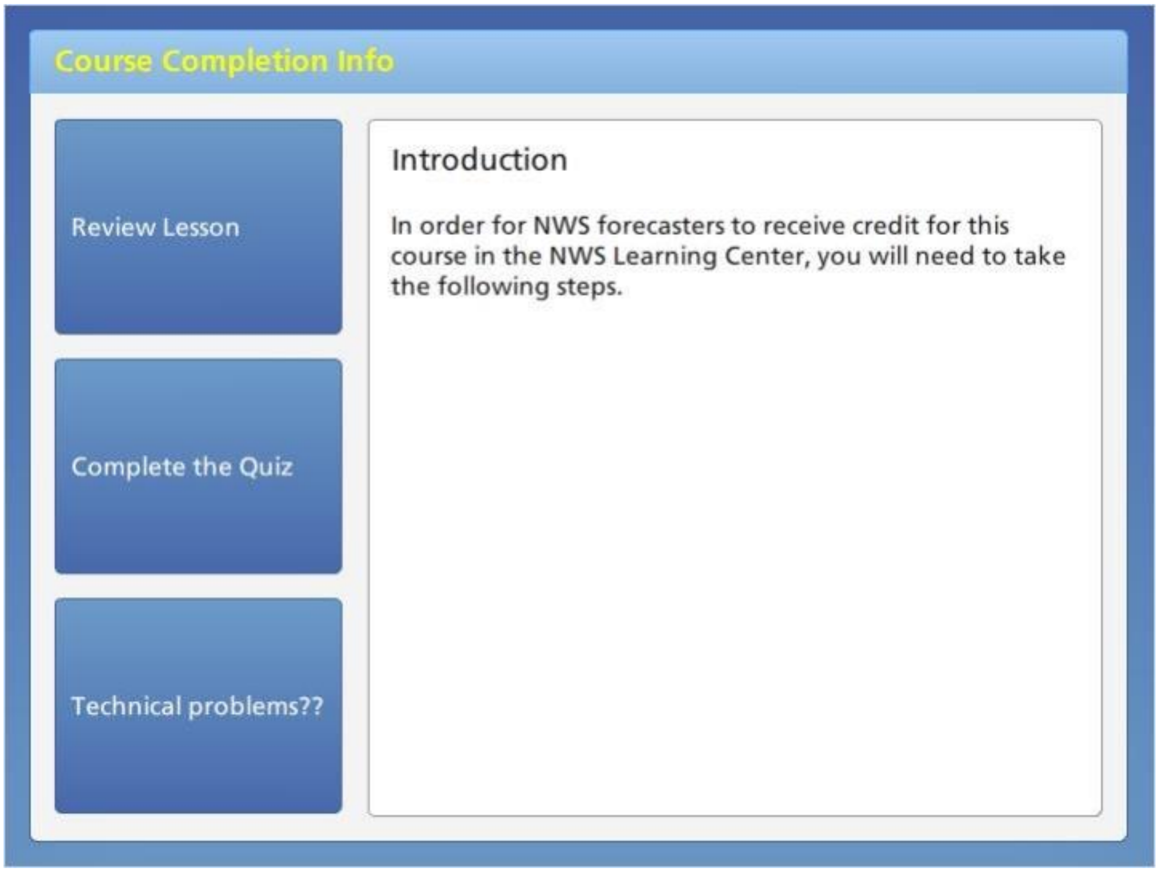
Course Completion Info
Review Lesson
Complete the Quiz
Technical problems??
Introduction
In order for NWS forecasters to receive credit for this course in the NWS Learning Center, you will need to take the following steps.

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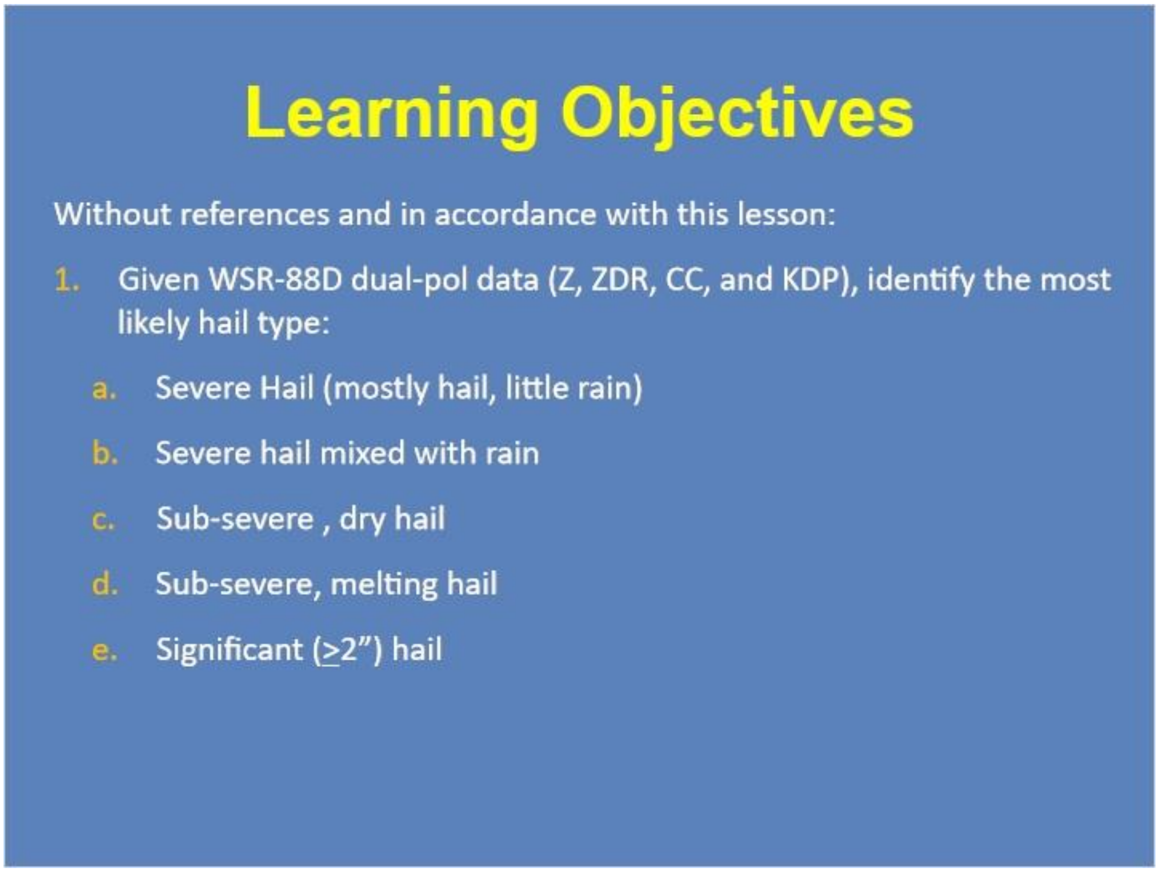

Notes:

Here are the learning objectives. Please take a moment to read them.

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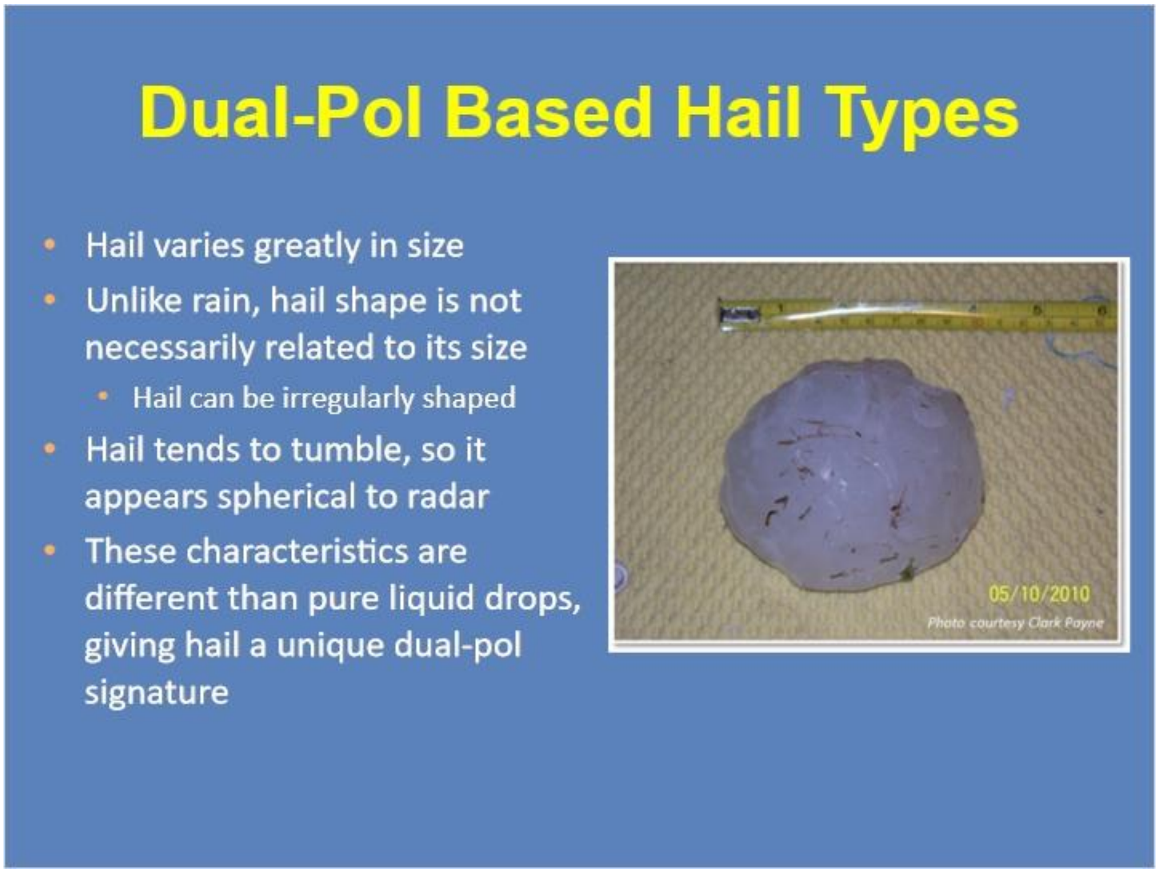
2. Dual-Pol Analysis



Notes:

Dual-pol data is very useful for hail detection.

Hail varies greatly in size, from as little as a quarter of an inch up to 8 inches in diameter. Unlike rain, the shape of hail is not necessarily related to its size. Hail can be irregularly shaped, with some hailstones having large protuberances, and in some cases be elliptical with one particular dimension much larger than the other. Hail also has the tendency to tumble, so it tends to appear effectively spherical to the radar. These characteristics are different from pure liquid drops, giving hail a unique signature in dual-pol data.

Let's discuss the hail detection capability of dual-pol products individually and then show how these products can be used in combination with reflectivity (Z) to detect certain hail event types.

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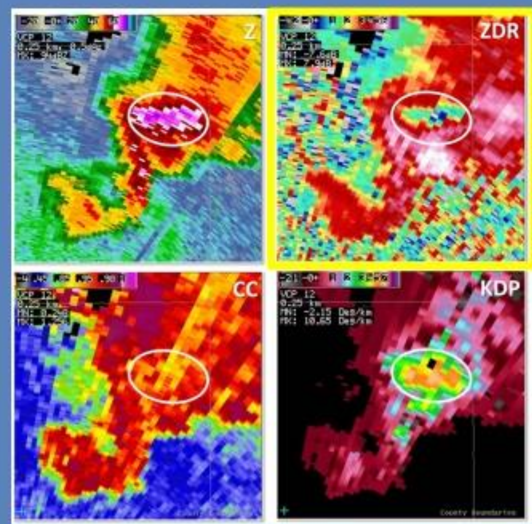
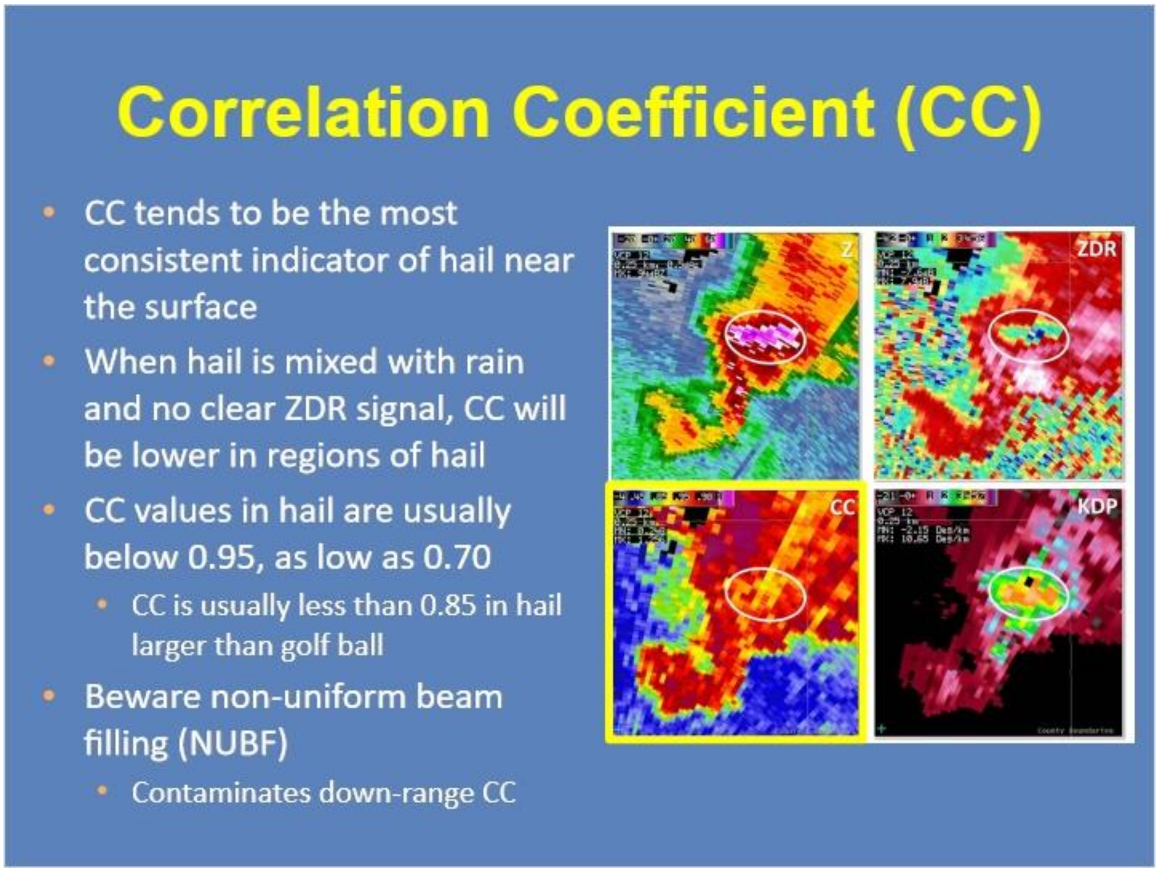
Differential Reflectivity (ZDR)

- ZDR will usually be fairly low (-0.5 to 1.5 dB) due to tumbling motion of hail fall
- ZDR reduction to near 0 dB coincident with high Z is a guarantee of hail
- ZDR can be quite variable
  - Melting hail or hail mixed with rain may result in very little ZDR reduction



Notes:

Differential Reflectivity (ZDR) will usually be fairly low, between -0.5 dB and 1.5 dB, due to the tumbling motion of the hail as it falls. A reduction in ZDR to near 0 dB coincident with high reflectivity is a guaranteed detection of hail. ZDR can be quite variable though, and in cases where the hail is melting or mixed with rain there may be very little reduction in ZDR.



Notes:

Correlation Coefficient (CC) tends to be the most consistent indicator of hail near the surface. In cases when hail is mixed with rain and there is not a clear signal in ZDR, CC will be locally lower in the regions containing hail. Values of CC in hail are usually below 0.95 and can be as low as about 0.70. For hail larger than roughly golf balls (> 1.75-inches), CC is normally less than 0.85.

It is important to note that the CC values down range from the hail region in this figure are characterized by radial streaks of depressed CC values. This is known as non-uniform beam filling (NUBF) and is one of the limitations of the Correlation Coefficient product. It occurs when at least some of the radar pulse volumes are characterized by significant gradients of PhiDP across these pulse volumes. When this occurs, these down-range CC (and other dual-pol base data) values are “contaminated,” or compromised. This means that the CC values and all other dual-pol values along the affected radials cannot be trusted or used.

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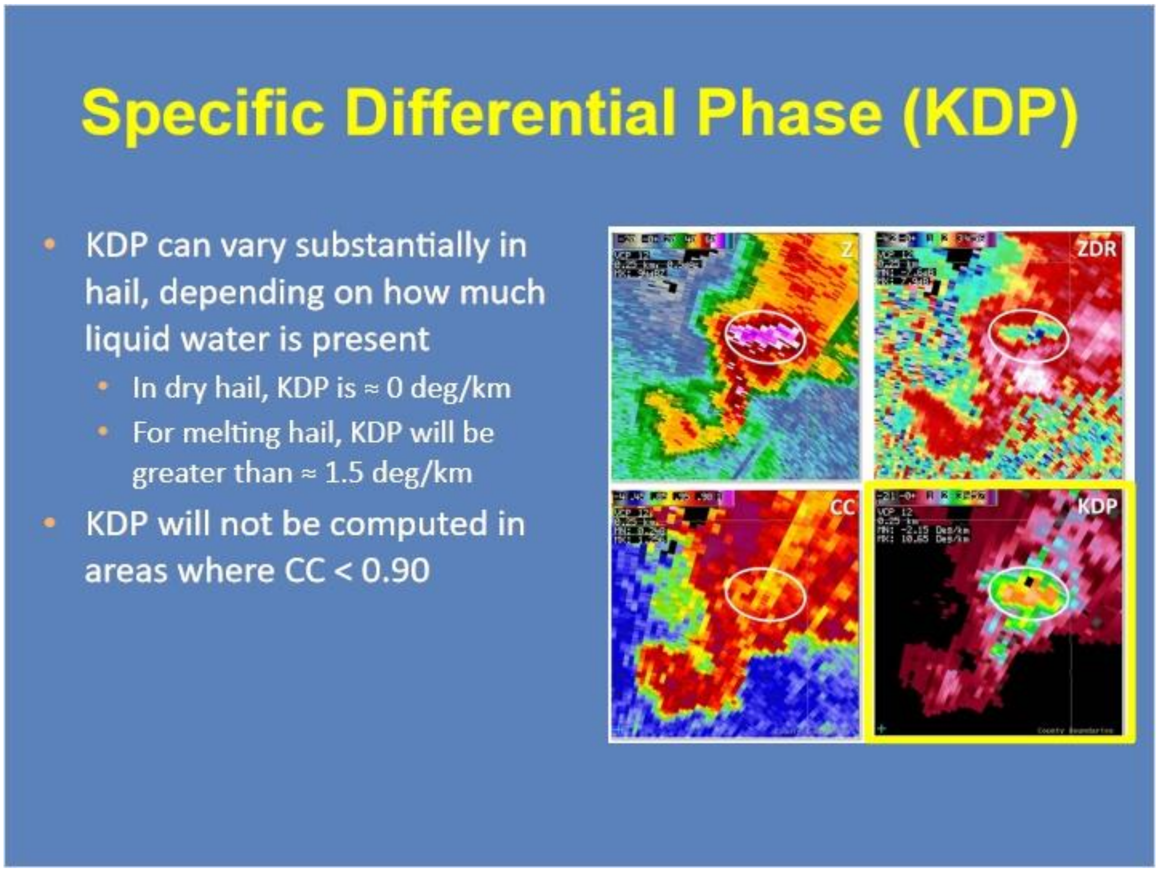
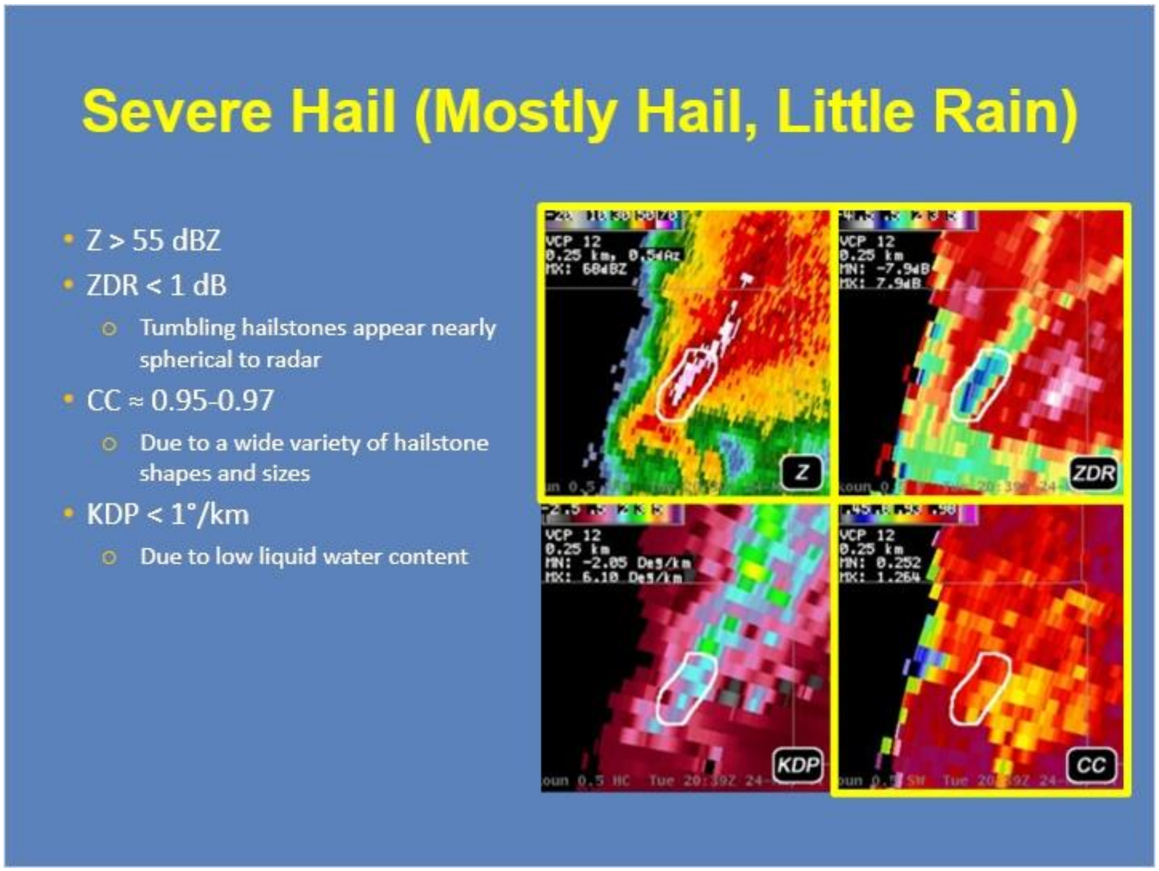


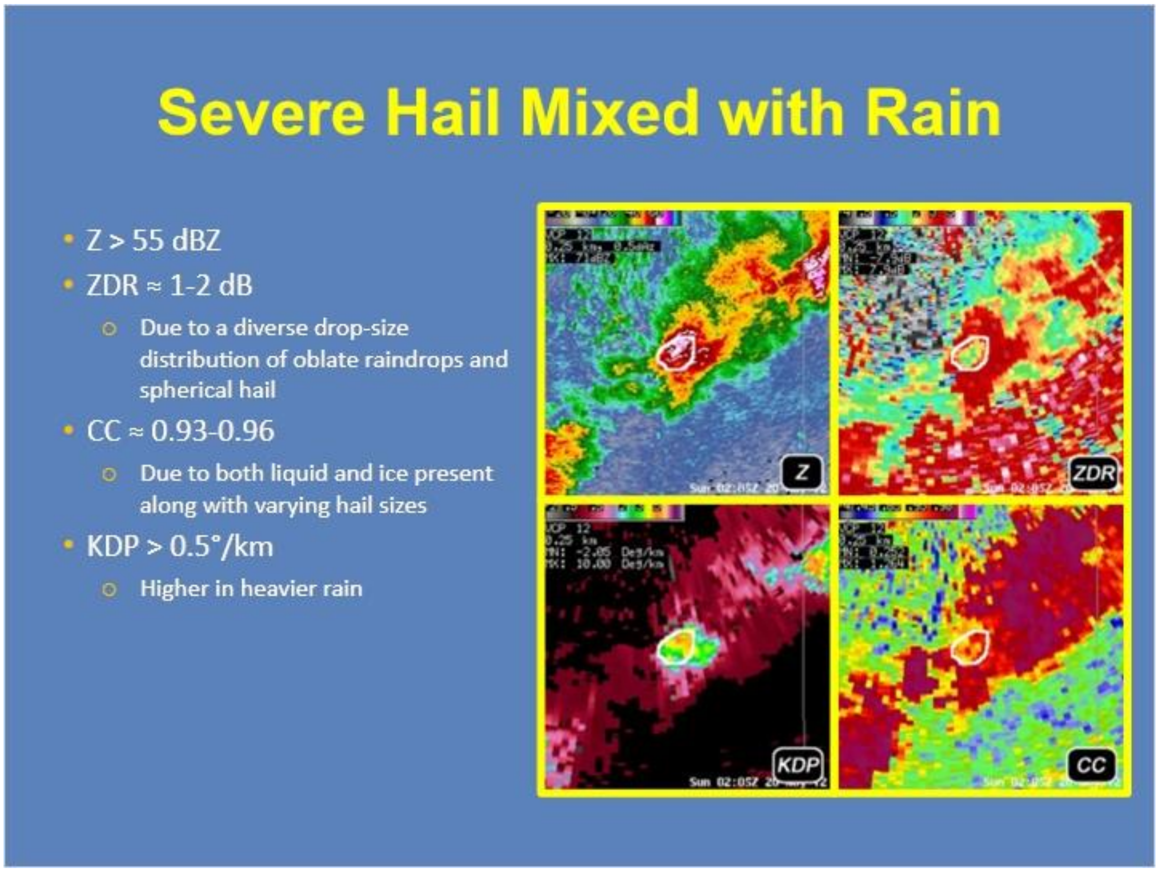
Notes:

Specific Differential Phase (KDP) can vary substantially in hail, depending upon how much liquid water is present with the hail. In dry hail without much rain, KDP is near 0. For melting hail, KDP will be greater than about 1.5 deg/km. This assumes that KDP is computed in areas containing hail, which is not the case when CC is < 0.90 (notice the black range gates within circle).



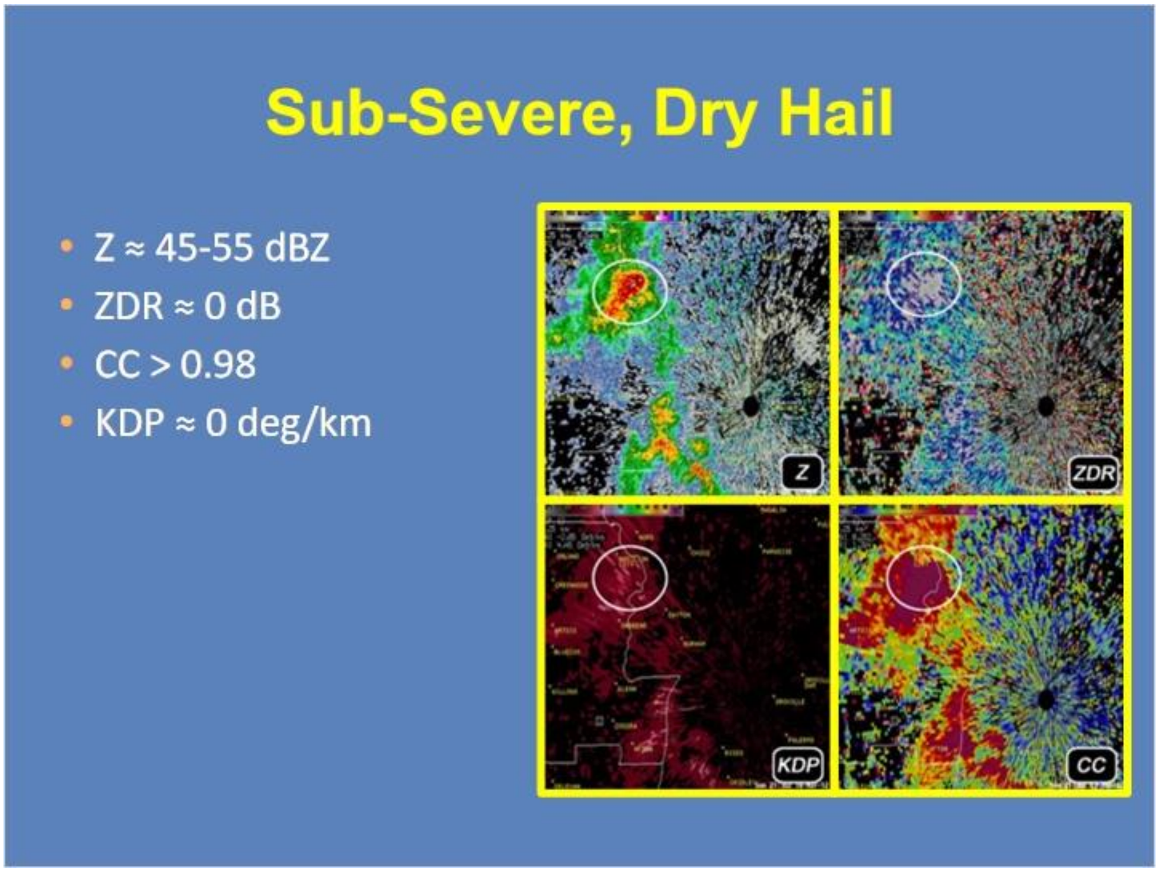
Notes:

Many research papers refer to the “classic, severe hail signature.” Severe hail, by definition, has a diameter of at least 1 inch (although some research papers were written when severe hail was threshold was ¾-inch diameter). Reflectivity values for severe hail are usually larger than pure rain events (Z > 55 dBZ). Since hail often tumbles as it falls, hailstones appear nearly spherical to the radar (ZDR typically < 1 dB). A wide variety of hail shapes and sizes result in CC tending toward the 0.95-0.97 range. Finally, KDP values are lower than pure rain (KDP typically < 1 deg/km).



Notes:

When rain is mixed with the “classic, severe hail signature,” the dual-pol variables behave a little differently. Reflectivity values will still be very high (Z > 55 dBZ) because of the size dependence of Z. ZDR values will be more positive (ZDR ~ 1-2 dB) as the diverse drop-size distribution of oblate rain and spherical hail both contribute significant power returns. Likewise, CC will be slightly lower than the “classic, severe hail signature” (CC ~ 0.93-0.96) because there is now both liquid and ice present along with varying sizes of hail. KDP will increase (KDP > 0.5 deg/km) because it's not dependent upon drop size like Z and ZDR. KDP only depends upon drop shape and number concentration. As a result, tumbling hail doesn't contribute significantly to KDP, but oblate rain drops will. The heavier (and more concentrated) the rain, the higher the KDP values will be.



Notes:

Sub-severe (< 1-inch), dry hail also has some unique dual-pol characteristics. Reflectivity will still be high, but not as high as other hail cores (Z ~ 45-55 dBZ). Small hail will tend to be smooth on the surface and appear spherical on radar (ZDR ~ 0 dB). CC should be near uniform (CC > 0.98) since the hail stones are similar in shape and size and little liquid water content is present. Likewise, KDP will be low (KDP ~ 0 deg/km).

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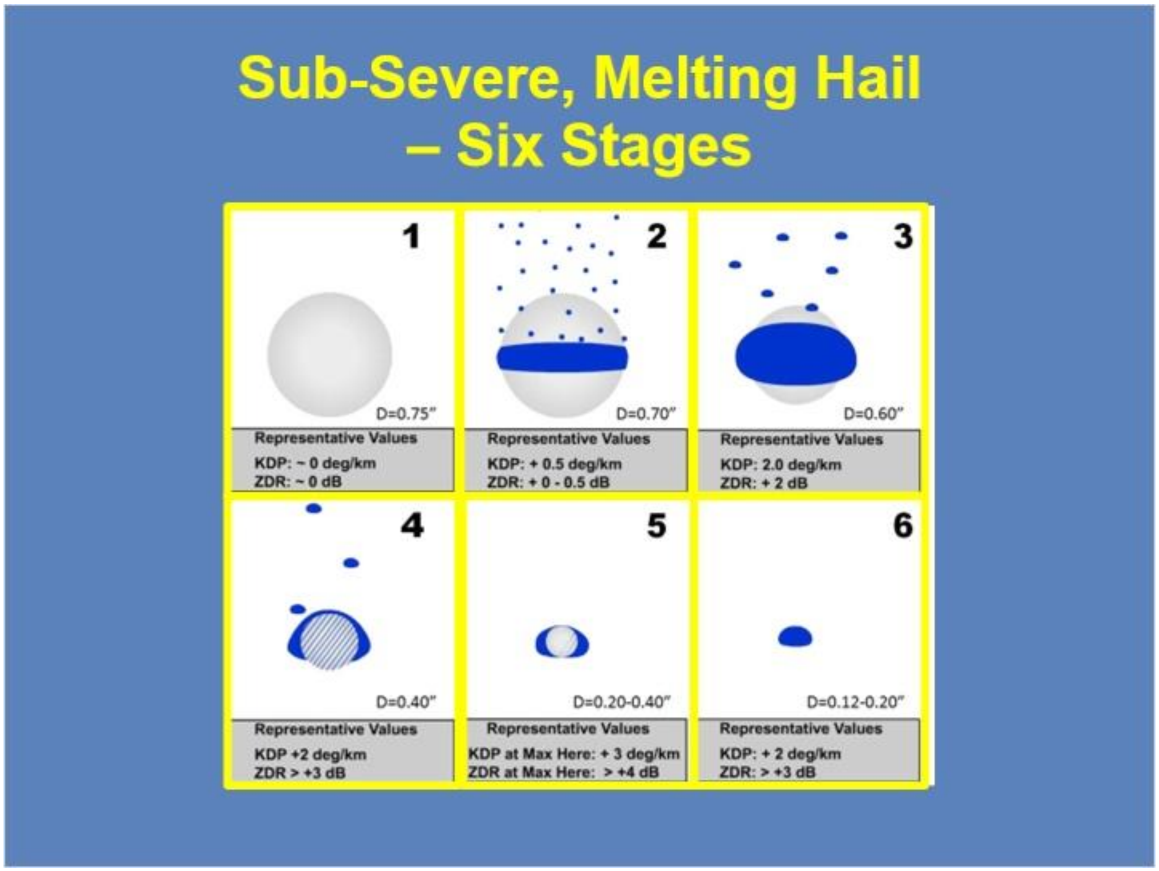

Notes:

Dual-polarization provides the capability to detect hail that has significantly melted, to the point that it's not likely to be severe. This signature is important because hail signatures can be found in just about any convective storm near and just above the freezing level. Therefore, observing a hail signature doesn't mean the hail will reach the surface. So let's discuss how hail melts and what it should look like on dual-pol products. A melting hailstone goes through six stages.

Stage 1: Begin with a solid ice sphere (D=0.75") falling outside of the updraft below the 0^oC level. Melting begins on the surface of the hailstone.

Stage 2: As the surface melts, the meltwater is advected into a torus (blue band around the equator of the hailstone) due to drag as it falls. Continuous shedding of small drops (~1 mm) occurs from the torus of water. Shed drops fall much slower. Hail diameter now 0.70."

Stage 3: Hail continues to melt and the torus moves upstream as the size of the ice particle decreases. Intermittent shedding of large drops (~3 mm) occurs from the unstable torus. Hail diameter now 0.60."

Stage 4: The torus loses its distinction, and a water cap forms around the top (lee side) of the ice core. Intermittent shedding of a few large drops (~3 mm). Ice core is now 0.40" diameter.

Stage 5: Meltwater forms a stable raindrop shape around the ice core. There is no drop shedding any longer. Horizontal axis diameter of ice and water coating ~0.2" to 0.4" (~5-9 mm).

Stage 6: Eccentric melting of ice core occurs until ice is completely melted. All that is left is a large, cold rain drop ~0.12" to 0.20" (3-5 mm) in diameter.

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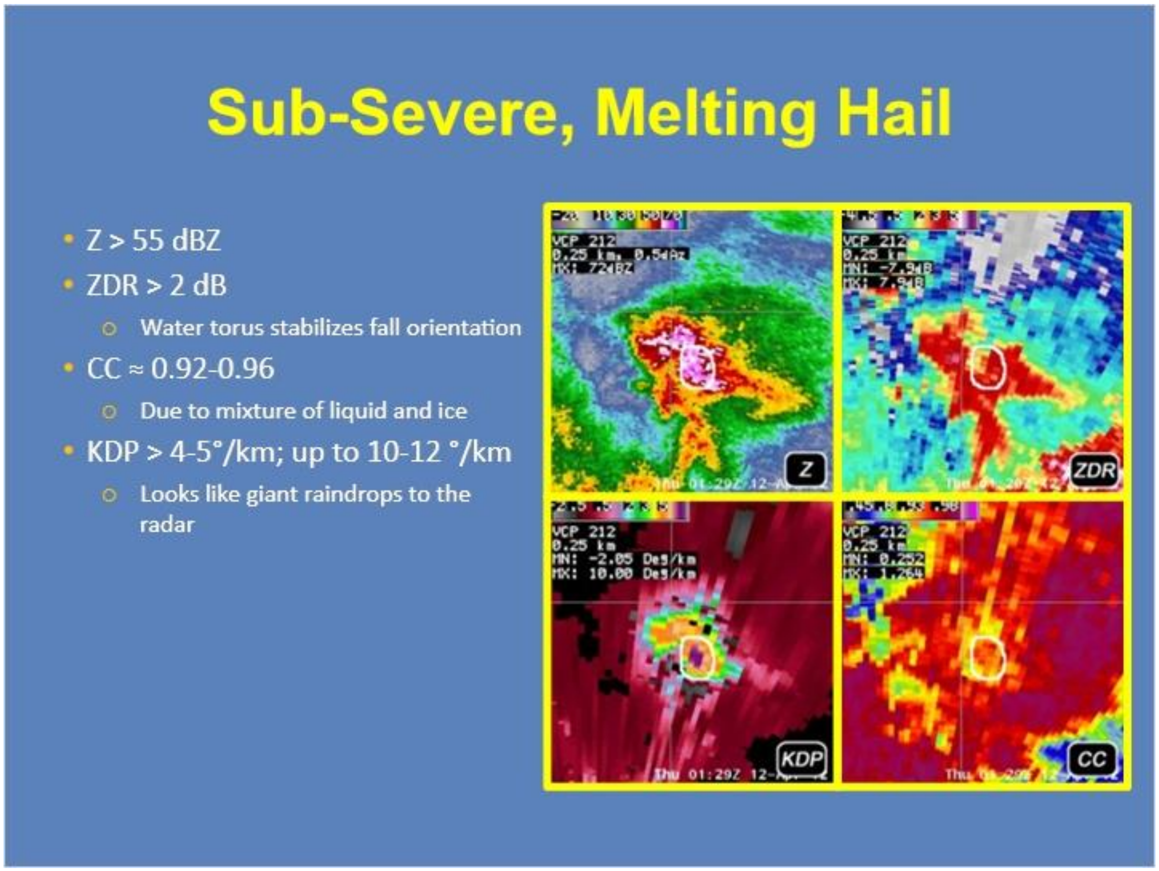

Notes:

Now that you have seen how smaller hail melts, let's discuss how it appears to dual-pol radar. As you have seen, when sub-severe hail melts, it develops a water torus on the surface around its center. This water torus tends to stabilize its fall orientation, making it look like a giant rain drop to the radar. As a result, reflectivity values tend to remain high (Z > 55 dBZ). Likewise, ZDR increases (ZDR > 2 dB; possibly as high as 6 dB). CC decreases to around 0.92-0.96 due to the mixture of ice and liquid.

KDP is very revealing in this case. KDP in small, melting hail can become extremely large (KDP up to 10 deg/km!). Why is that? When there is a high concentration of these “giant raindrops” (that is sub-severe, melting hailstones with a water torus), then KDP values can become extremely large. In pure rain situations, KDP values will rarely go above 4-5 deg/km. When KDP is larger than those values, you can confidently assume that there is some small melting hail present.

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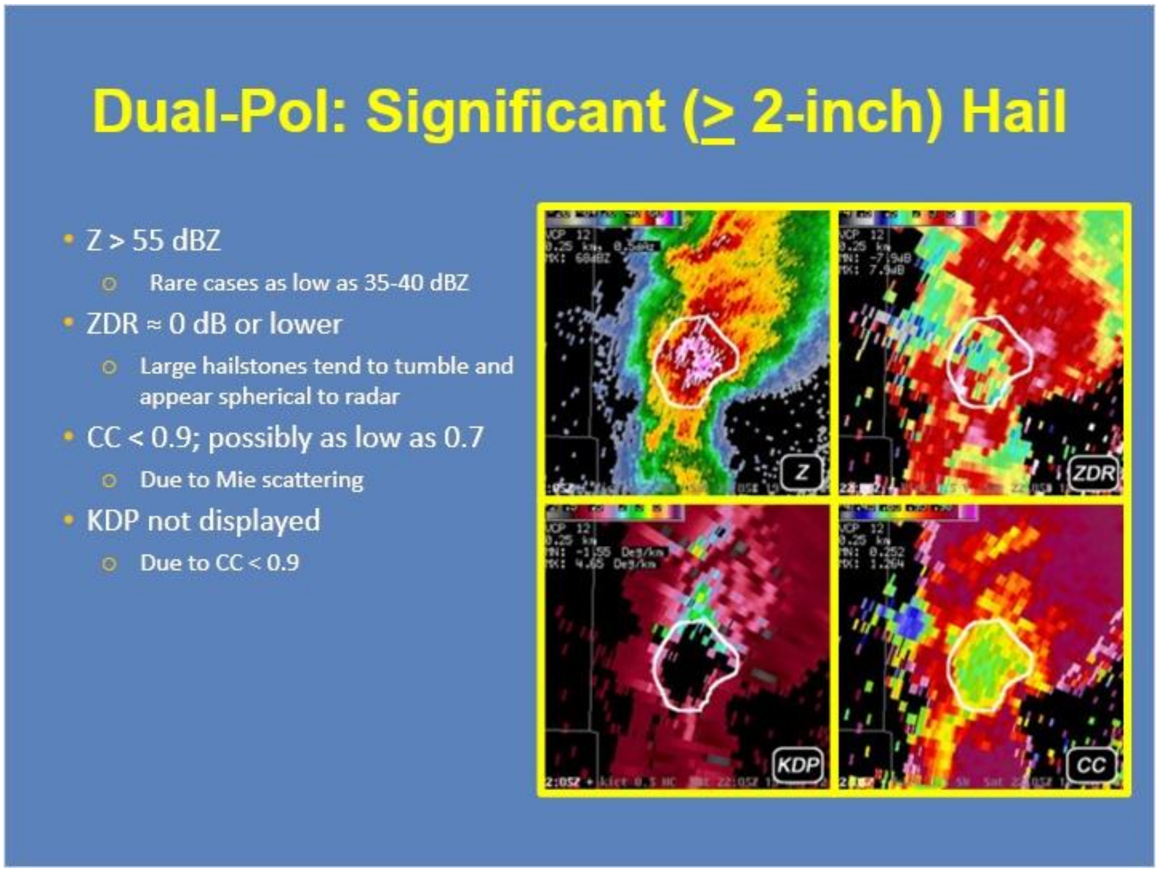

Notes:

When significant (2-inch diameter and larger) hail is present, the signature in the dual-pol products can be very pronounced. When hail gets to be larger than golf balls, Mie scattering effects begin to alter the way the dual-pol variables appear, and this signature is unique. Reflectivity (Z) will remain high (Z > 55 dBZ) except for rare cases when only a few, large hailstones fall in the Z gradient near the updraft/downdraft interface region (Z as low as 35-40 dBZ). Differential Reflectivity (ZDR) will still be near zero or even be mostly negative (ZDR ~ 0 dB or lower).

Correlation Coefficient (CC) is the most revealing product in this case. Mie scattering will cause CC values to drop significantly lower (CC < 0.9; possibly as low as 0.7!). Dropouts in the Specific Differential Phase (KDP) data will appear since gates where CC < 0.9 are filtered from the product. Therefore, if you see high reflectivity (or moderate reflectivity near a supercell updraft) and CC < 0.9, you can confidently say significant hail is present.

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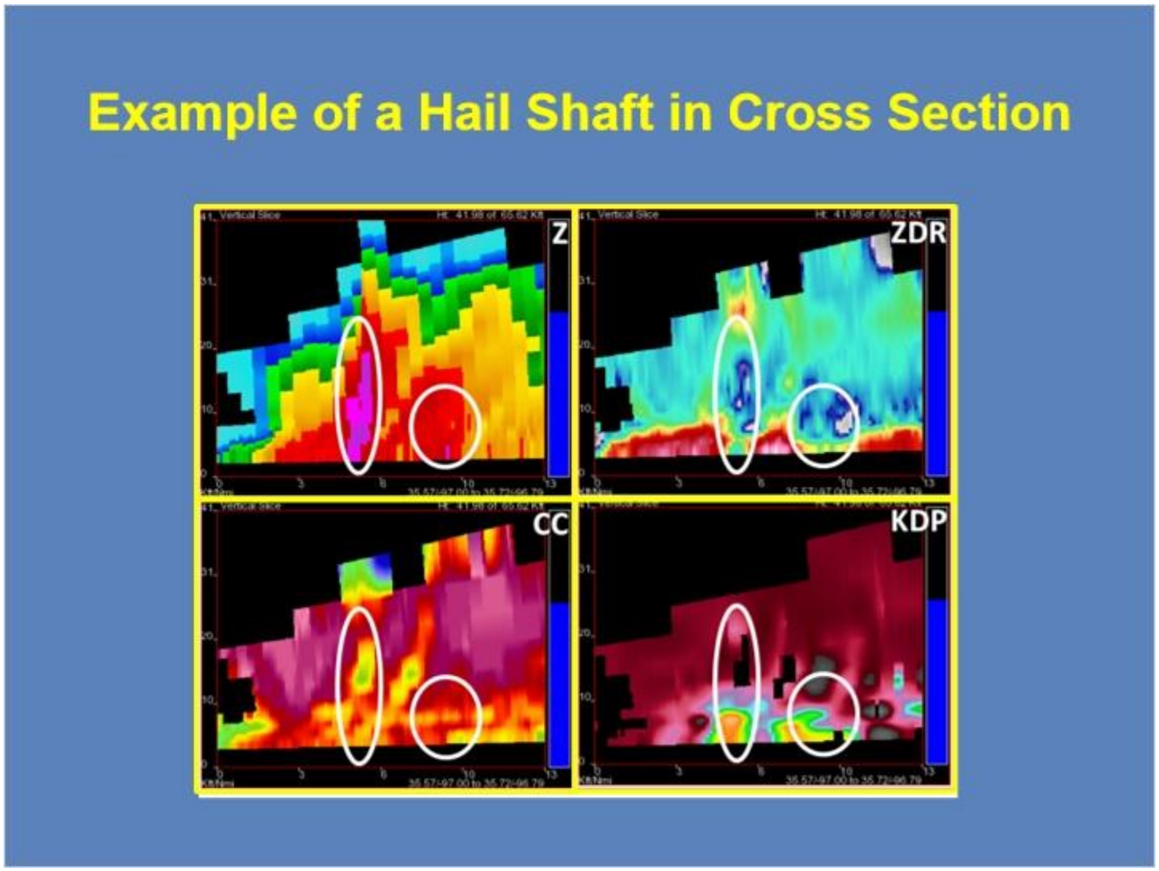

Notes:

The AWIPS Four-Dimensional Stormcell Investigator (FSI) tool can be used for hail detection with dual-pol products. In the FSI cross section seen here, there are actually two hail shafts, marked by the white circles. As expected, both hail shafts have high reflectivity, although the one on the left, associated with a newer updraft, is deeper with higher values.

Taking a look at ZDR, both hail shafts are associated with fairly low positive to slightly negative values just above the melting layer, with the lower values extending higher aloft for the (younger) hail shaft on the left. Note also the depression in the transition to higher, positive values in each hail shaft. The signal in CC is not as easy to pick out as in the reflectivity or ZDR in this case, but notice the values of CC are generally low in both hail shafts. KDP is generally very high in both hail shafts reflecting increased liquid water content associated with melting hail and/or a rain/hail mixture.

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Notes:

For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.

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Welcome to the Convective Storm Structure and Evolution's lesson on supercell dynamics and motion. This lesson is about 30 min long.

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Learning Objectives

- Identify the typical environment, storm structure, and evolution of supercells.
- Identify the effects of shear on storm propagation.
- Identify the technique to anticipate the motion of supercells.

These are the learning objectives that you need to learn in this lesson. Identify the typical environment, storm structure, and evolution of supercells.

Identify the effects of shear on storm propagation. Identify the technique to anticipate the motion of supercells.

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Reasons for Storm Intensification in Significant Vertical Shear

- As deep vertical shear exceeds 20 m/s
  - Increased updraft/downdraft separation
  - Precipitation removal from updraft
  - Lower boundary-relative storm motion
  - Stronger storm-relative low-level inflow
  - Increased nonhydrostatic vertical pressure gradient
    - Due to updraft vorticity
    - Due to interaction of shear and updraft exterior

As updrafts encounter an increasingly sheared environment (e.g., 0-6 km shear >20 m/s), they become enhanced by: increased updraft/downdraft separation, precipitation removal from updraft, lower boundary-relative storm motion, stronger storm-relative low-level inflow, increased nonhydrostatic upward directed pressure forcing due to updraft vorticity, increased nonhydrostatic upward directed pressure forcing due to shear interacting with the updraft boundary.

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What Describes a Supercell?

- Updrafts last longer than individual air parcel transit time
- Updraft well correlated with vertical vorticity

If an updraft begins to persist for longer than an individual air parcel takes to traverse it, and it is well correlated with significant vorticity, the updraft is then called a **supercell**. Sometimes the effects of nonhydrostatic pressure forcing on updraft strength can exceed that of buoyancy.

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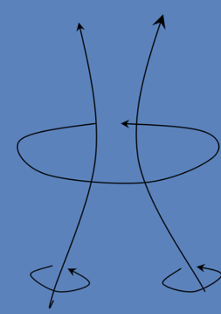
Definition of Updraft Rotation and a Vortex

- Vortex is a local maximum collection of vortex lines
- Rotating Updraft is a local concentration of vortex lines with curvature vorticity



Before we proceed with the following discussion, there are some definitions of vortices, updraft rotation and updraft vorticity that you should know so as to avoid any misconceptions. When we discuss the term ‘vortex’, we refer to a local concentration of vortex lines. The vortex may or may not be rotational. In other words, a vortex could be a locally intense region of shear vorticity. Likewise a vortex could result from a concentration of vortex lines eminating from curvature vorticity or rotation. The term rotating updraft is a vortex with curvature vorticity. However air parcels within a rotating updraft may not complete a closed circuit as observations have shown. Supercell updraft air trajectories often show anticyclonic curvature even though the vertical vorticity is positive.

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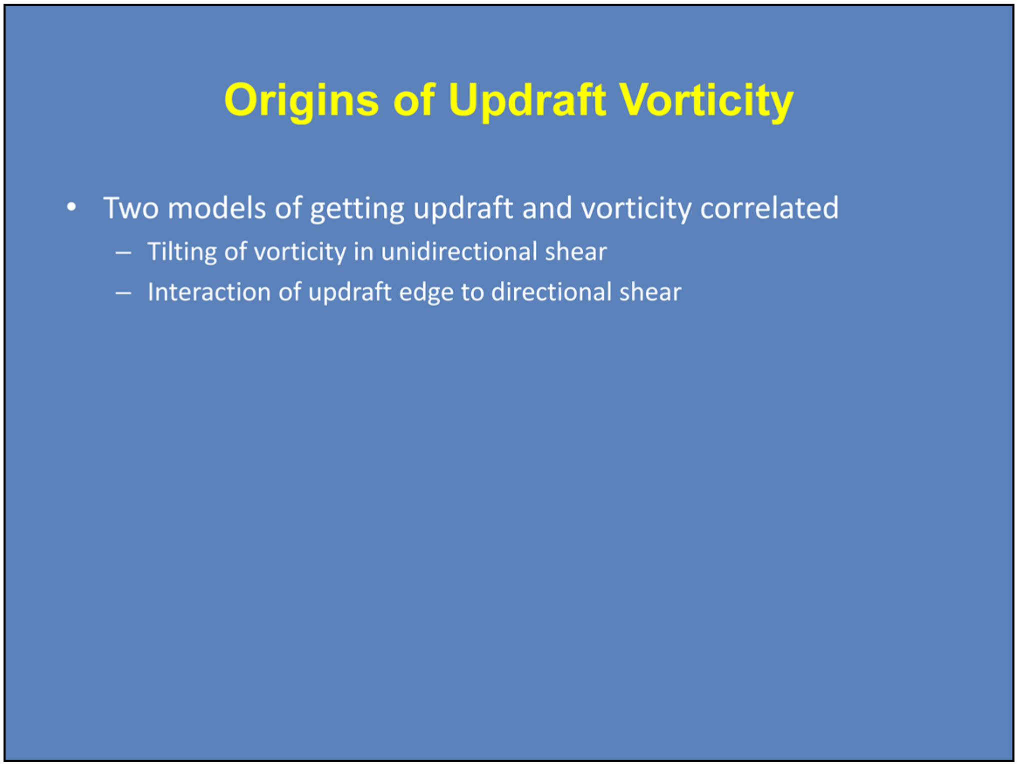

The origins of updraft vorticity and storm motion deviant to the steering layer wind can both be explained by how the updraft is influenced by vertical wind shear. There can either be unidirectional or directional vertical shear in supercell environments. Fundamental origins of updraft vorticity and propagation are shared by both straight and curved sheared environments. However, there are important differences in the origins of updraft vorticity and propagation between unidirectional and directional vertical shear. These differences will be covered in this section.

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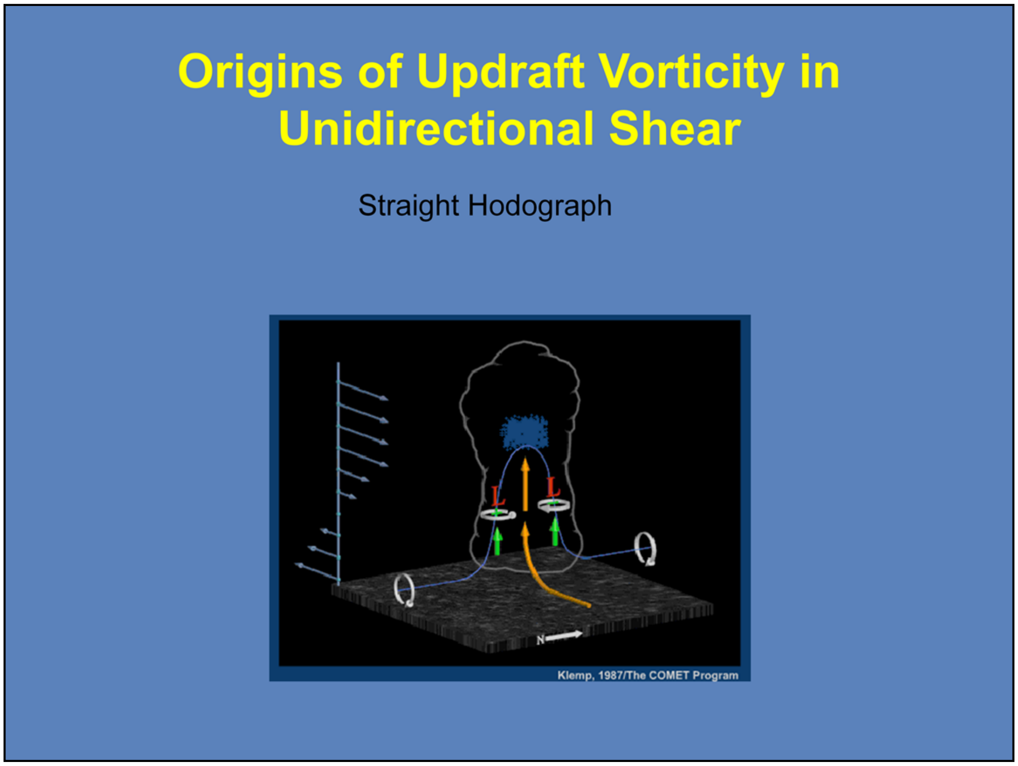

We can visualize vertical shear as a continuous series of vortex lines oriented horizontally. A good analogy is a sheet of rolling logs. **As an updraft extends into a sheared environment, horizontal vorticity tilting acts to create two vertical vortices.** The strength of these vortices depends on the strength of the shear and the intensity of the updraft. Facing toward the direction of the shear, from left to right in Figure , on the right (left) side of the updraft lies a cyclonic (anticyclonic) vortex. Initially in this figure, the vortices lie along the periphery of the updraft, and thus contain no updraft within them. In other words, the updraft and vorticity are not correlated.

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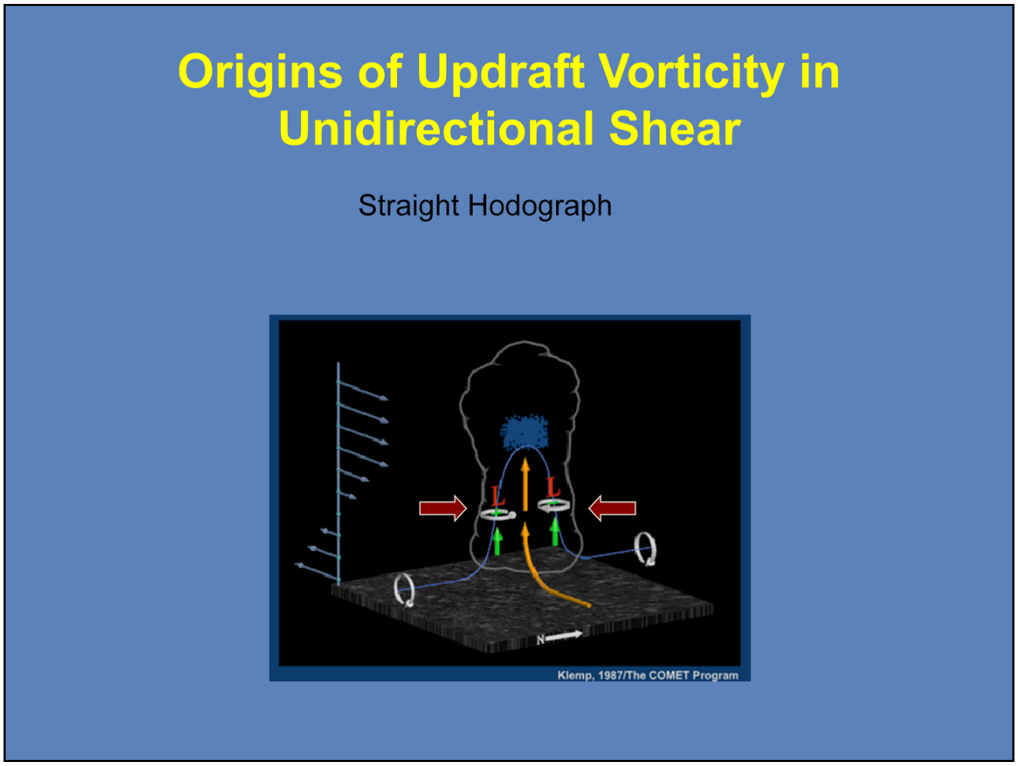

Both counter rotating vortices create a dynamic low. The stronger the vortex, the lower the pressure in its center. **Since tilting of the originally horizontal vorticity is most pronounced where the updraft is strongest (at midlevels), the vertical vortices are most intense there.** With the dynamic pressure at its lowest aloft, an enhanced upward directed pressure gradient force promotes the development of new updraft within their centers of rotation. The effect is a widening of the updraft and increasing correlation between updraft and vorticity on both flanks. Updraft strength is also augmented through this process.

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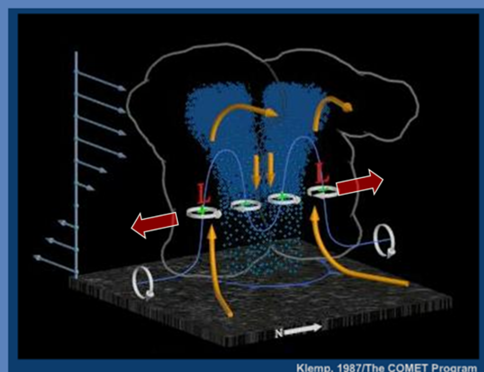
Origins of Updraft Vorticity in Unidirectional Shear

Straight Hodograph



The greatest tilting of horizontal vorticity occurs right and left of the shear vector. **This means that the development of rotation and new updrafts also occur to the right and left of the shear vector.** Precipitation developing in the middle of the widening updraft acts to develop a downdraft which, in turn, helps to split the widening updraft into two parts. **The cyclonically (anticyclonically) rotating member moves to the right (left) of the shear vector.** Since both the cyclonic and anticyclonic updrafts experience similar upward dynamic pressure forcing, they are equally strong supercells in a straight hodograph environment. Once the supercell is deviating off the hodograph, it experiences streamwise vorticity, and storm-relative helicity in its inflow layer. Tilting of the streamwise vorticity into the updraft immediately produces vertical vorticity well correlated with updraft.

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Origins of Rotation in Directionally Varying Shear

- Directional shear implies curved hodograph
- Curved hodograph suggests streamwise vorticity

The processes that develop rotation in the unidirectional hodograph, also apply to curved hodographs. However, a curved hodograph implies that streamwise vorticity and helicity are available for the updraft to directly ingest upon its initial growth. Instead of the rolling log analogy to describe the vorticity in the environment, here the analogy is the thrown spinning football. This analogy represents the available streamwise vorticity that merely needs to be tilted into the vertical by the updraft in order for rotation and updraft to be well correlated. Therefore the evolution from ordinary cell to supercell is much quicker.

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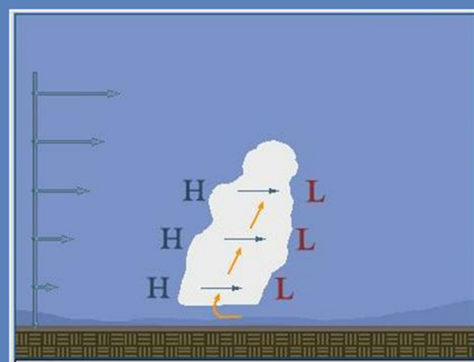
Interaction of Shear with Updraft Exterior



While the same processes that promote deviant motion in unidirectional hodographs will work in curved hodographs, the interaction of the changing shear vector with height will result in additional nonhydrostatic vertical pressure gradient forcing that promotes growth on only one flank of an updraft. This additional process is related to the same processes that force an updraft to tilt in the presence of vertical shear. On the upshear side of an updraft, high dynamic pressure forms as a result of partial flow blockage, while low pressure forms on the other side forcing the updraft to tilt. While this illustration deals with unidirection shear, we will next discuss how directional shear extends this concept to explain the origin updraft deviant motion and preference for the cyclonic member of a supercell to intensify.

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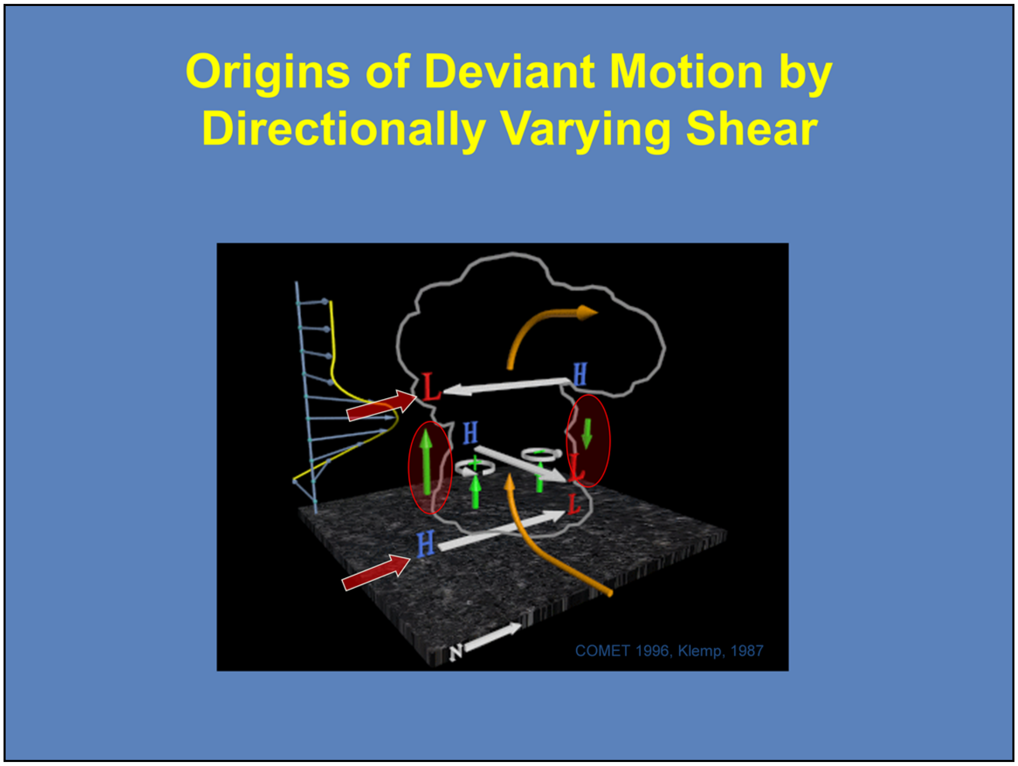

When the shear direction changes with height so do the locations of the dynamic pressure maxima and minima. **We know that a dynamic high (low) forms on the up (down) shear side of an updraft.** In the example shown in this figure , the relative high is on the south side of the updraft at low levels. At higher levels, the shear vector pointing south would produce a relative low on the south side of the updraft. The result is an upward directed pressure gradient force that causes new updraft development and therefore, storm propagation to the right of its original motion. Meanwhile, the left side of the updraft would experience a downward directed dynamic pressure gradient force weakening, or even destroying, the side of the updraft containing the anticyclonic member of the rotational couplet. This is why a left-moving storm, given the hodograph in , would be suppressed.

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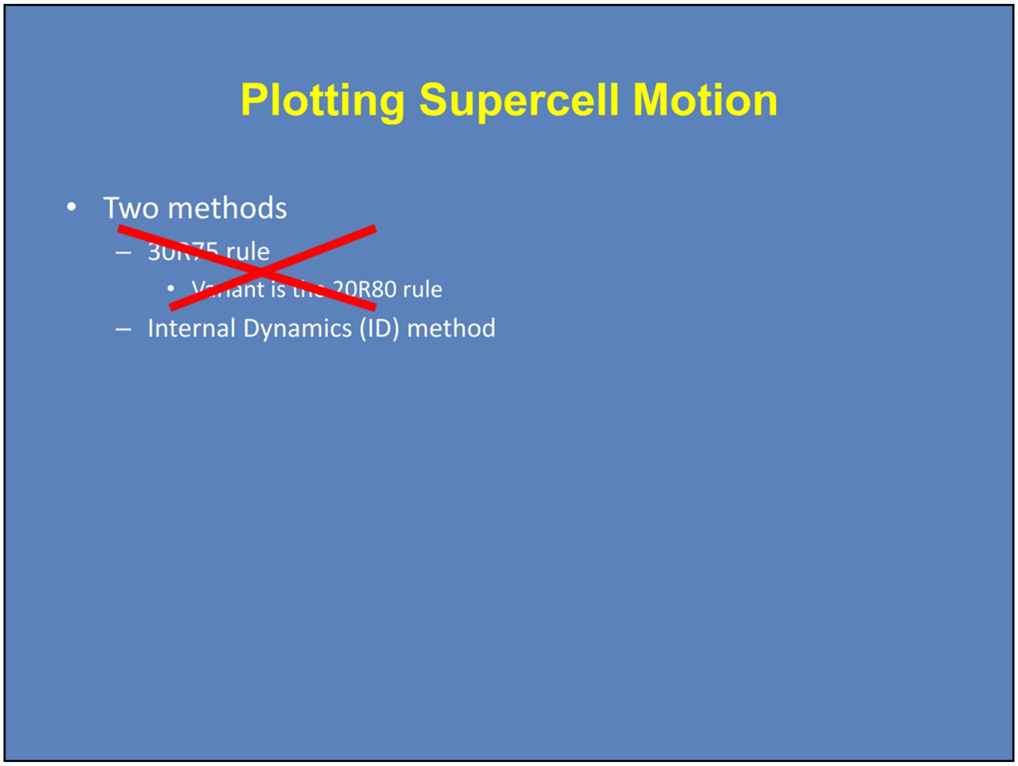
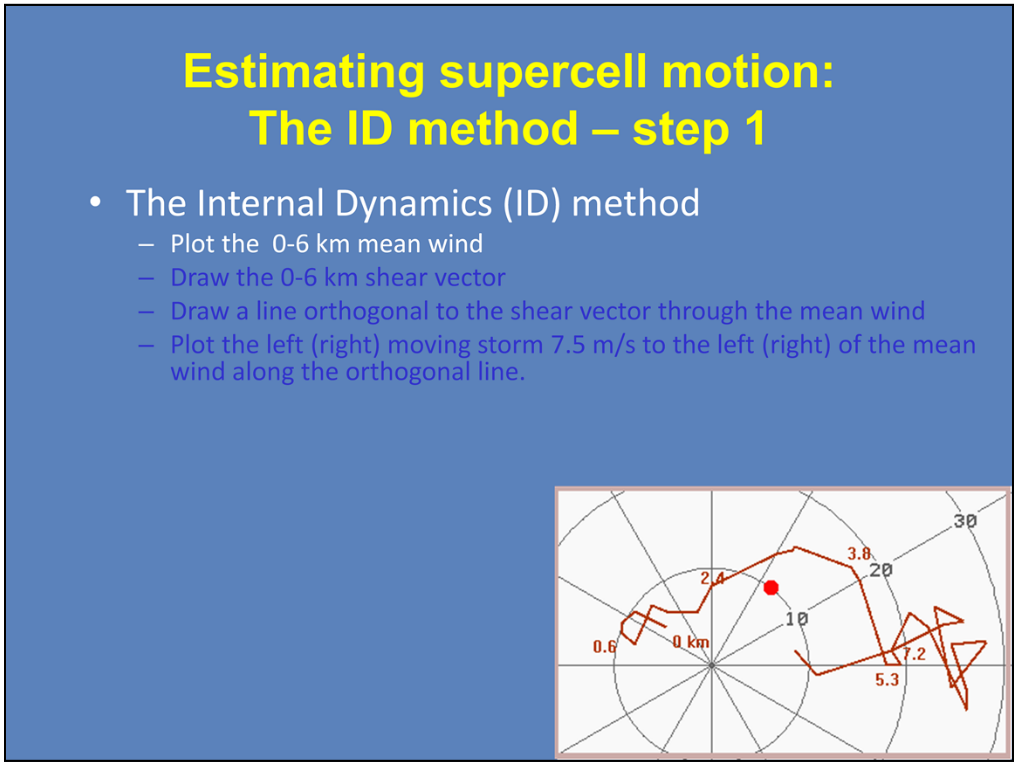


There are two methods for estimating supercell motion for which to be aware. The “legacy Supercell method” and the ID method, which is the preferred technique, are presented next.

In the past, forecasters often based supercell motion on the 30R75 (Maddox, 1976) or 20R85 (Davies and Johns, 1993) rules. **The 30R75 rule estimates the cyclonically rotating supercell motion by adding 30° to the right of the 0-6 km steering layer flow direction and 75% of the speed**. The 20R85 rule was an adjustment for those supercells embedded in very strong flow. Unfortunately, these estimations are non-physically based and only apply in the Northern Hemisphere with the typical counterclockwise turning hodographs. The AWIPS skew-T program still uses this technique to estimate SRH.



Bunkers et al. (2000) developed a better method called the **ID method (Internal Dynamics)**, which uses the mechanisms by which updraft and shear interact to cause deviant motion. **This method can be used to calculate storm motion for both the cyclonically and anticyclonically rotating supercells resulting from a storm split**. The ID method is Galilean invariant allowing for its use in atypical hodographs (i.e., westerly shear with northerly mean winds). To estimate supercell motion using the ID method, the following steps work well: 1) Plot the 0-6 km non-pressure-weighted mean wind. An example in this Figure shows the mean wind as a red dot.

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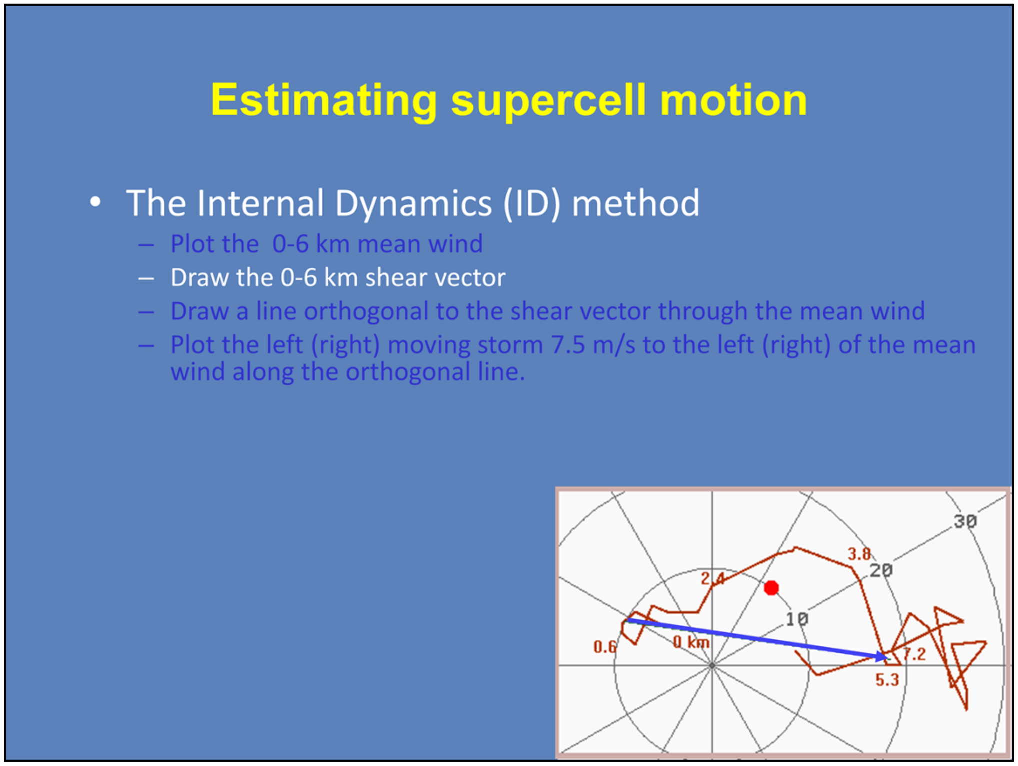
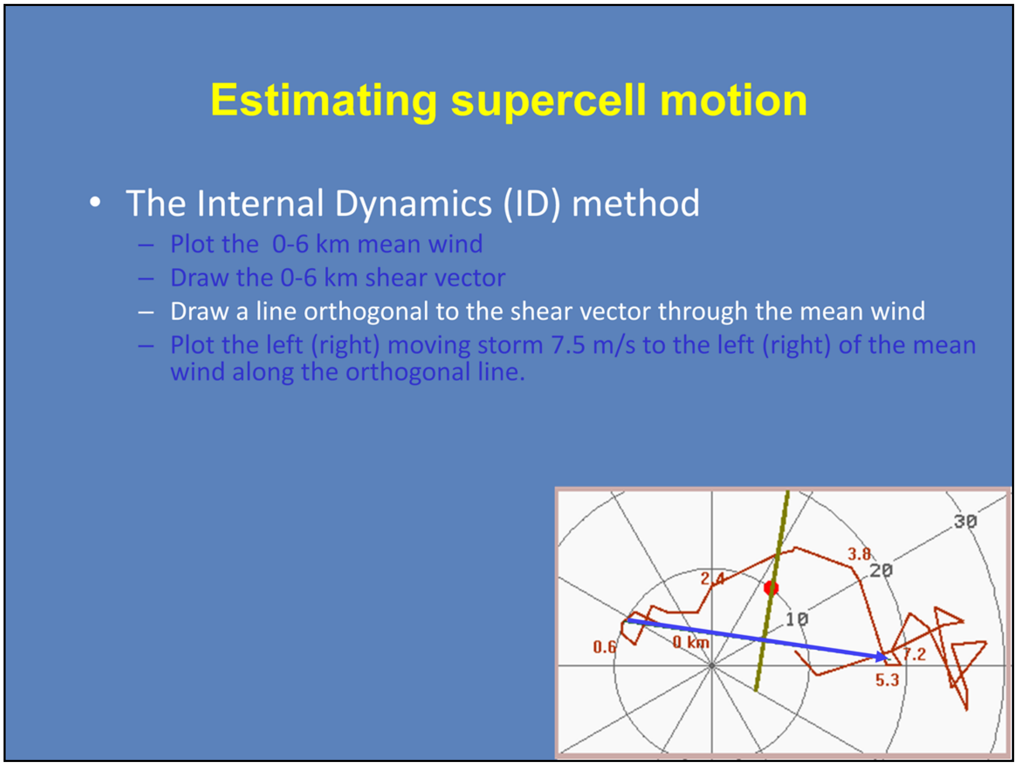
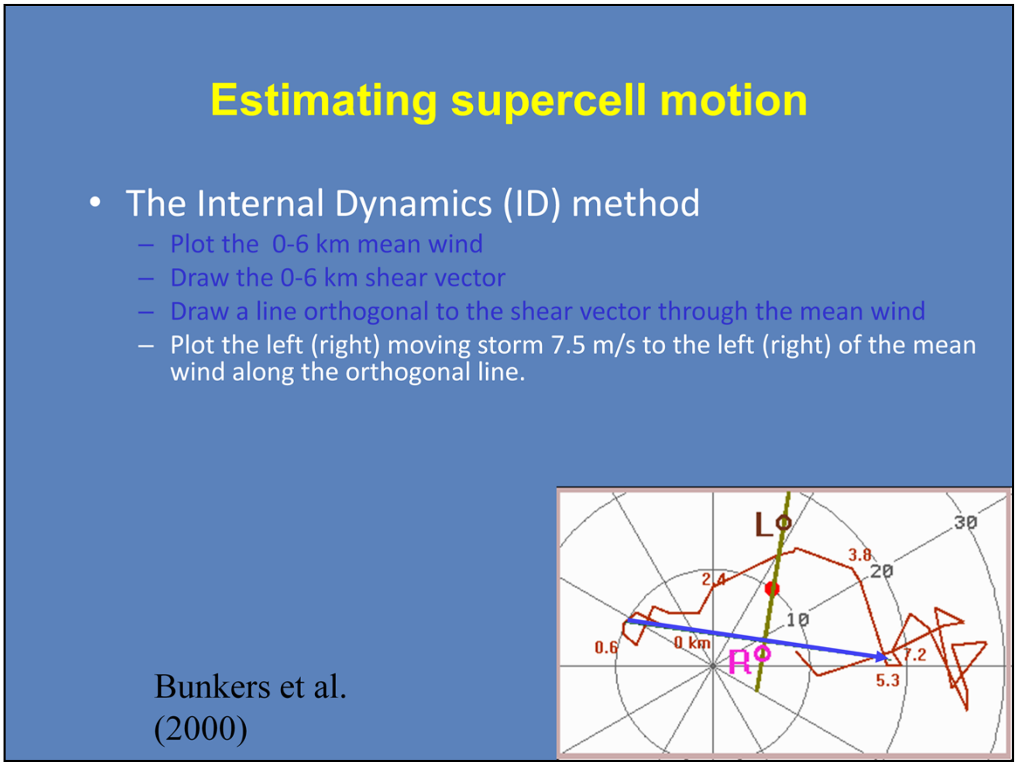


Draw the shear vector from the mean wind in the lowest 0.5 km to the mean wind from 5.5-6 km.



Draw a line orthogonal to the shear while passing through the mean wind. Note that the shear vector can be placed anywhere on the hodograph as long as it retains the same direction and magnitude.



The right- (left-) moving supercell is drawn 7.5 m/s to the right (left) of the shear vector where shear vector intersects the shear-orthogonal line at the 0-6 km mean wind. Note that the storm motion remains on the shear-orthogonal line.

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Limitations in ID Method

- ID Method contains uncertainties
- What is the “best” Deviant motion?
- How is it Modulated?
- Propagation Effects due to Boundaries

ID Method contains uncertainties. For example, what is the “best” Deviant motion? 7.5 m/s was chosen on a representative sample size. How is motion modulated? What about propagation effects due to boundaries?

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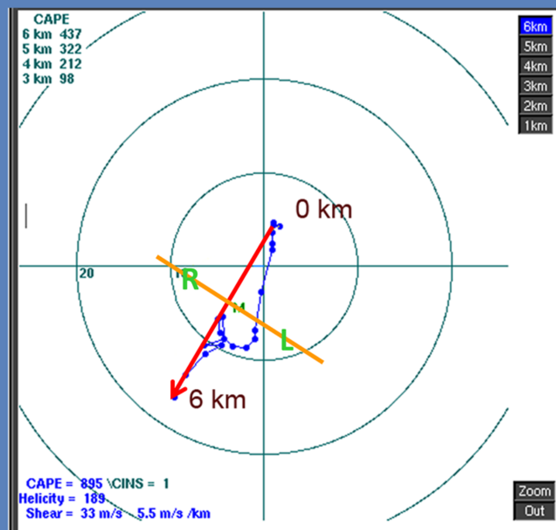
Motion Explained by Hodograph



This is the author’s explanation of one of the previous quiz items.

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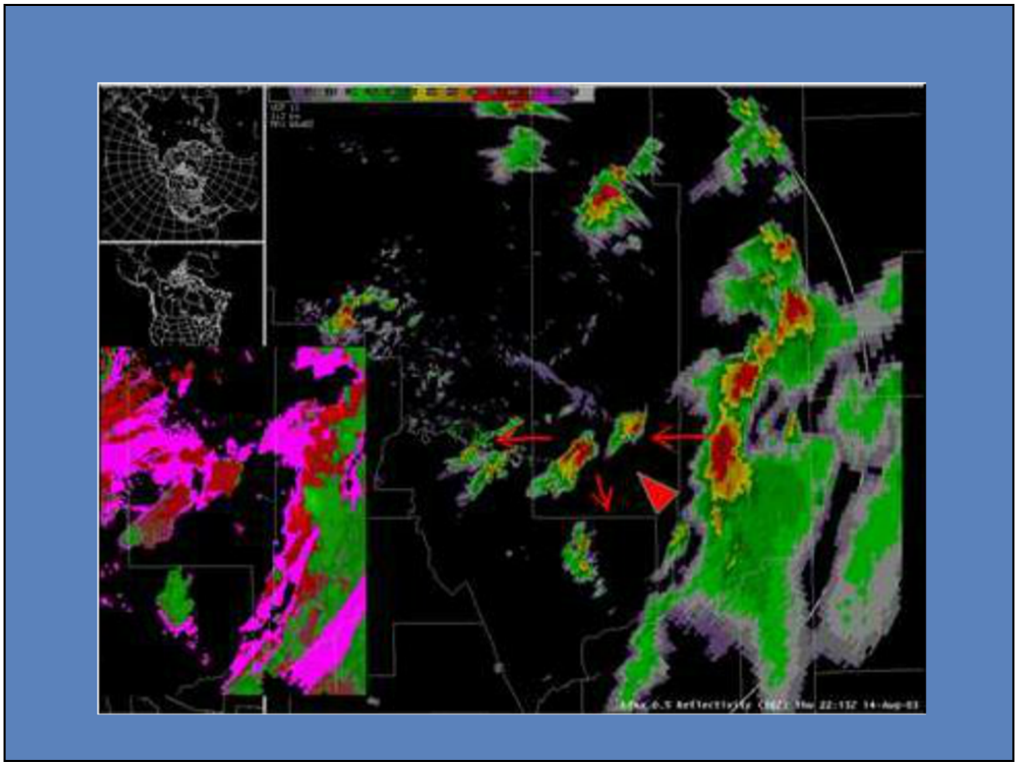

This is a radar loop (0.5 deg Z and 0.5 deg SRM in the lower left) from KFGZ of the resulting storm motions from the hodograph just analyzed.

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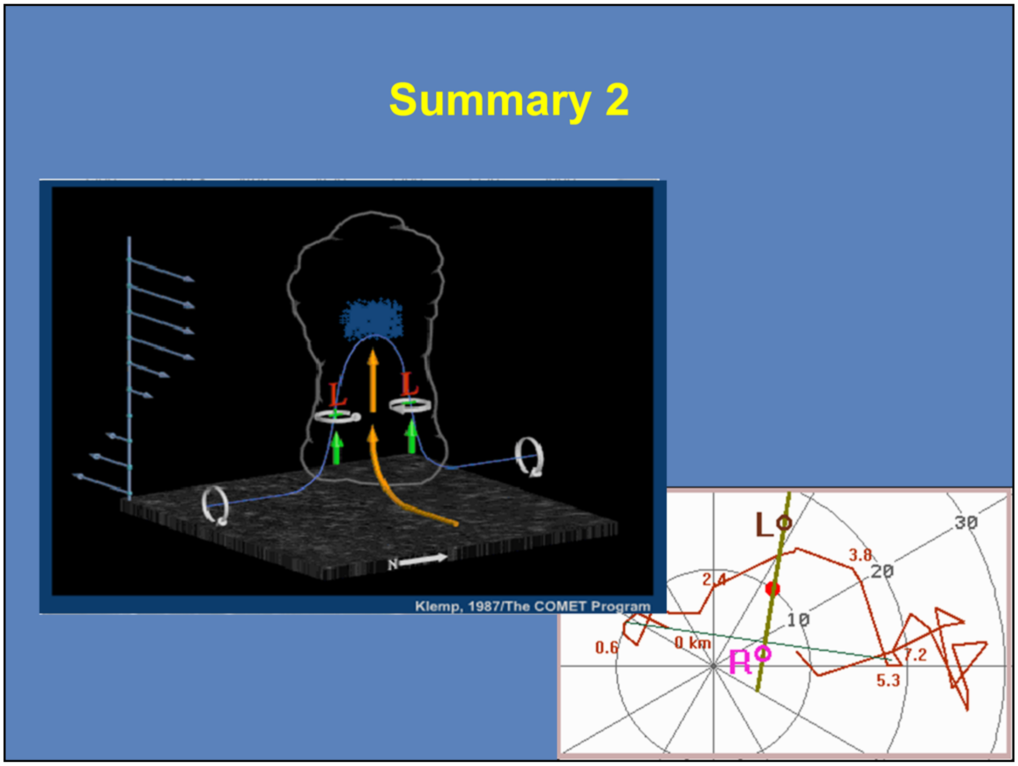
Summary 1

- Strength of wind shear (> 20 m/s through 6 km) defines supercell longevity
  - Increased updraft/downdraft separation
  - Precipitation removal from updraft
  - Lower boundary-relative storm motion
  - Stronger storm-relative low-level inflow
  - Increased nonhydrostatic vertical pressure gradient
    - Due to updraft vorticity
    - Due to interaction of shear and updraft exterior
- Persistent updraft vorticity defines a supercell

This is the summary slide part 1.



Summary slide part 2.

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Welcome to the Convective Storm Structure and Evolution's lesson on Supercell Morphology: Radar Reflectivity Signatures.

Lesson Objective

- Identify radar reflectivity characteristics of supercells

The persistence and strength of a supercell thunderstorm updraft yields a distinctive appearance to its precipitation distribution. This lesson describes the common radar-based reflectivity characteristics associated with supercells.

The objective is: Identify radar reflectivity characteristics of supercells.

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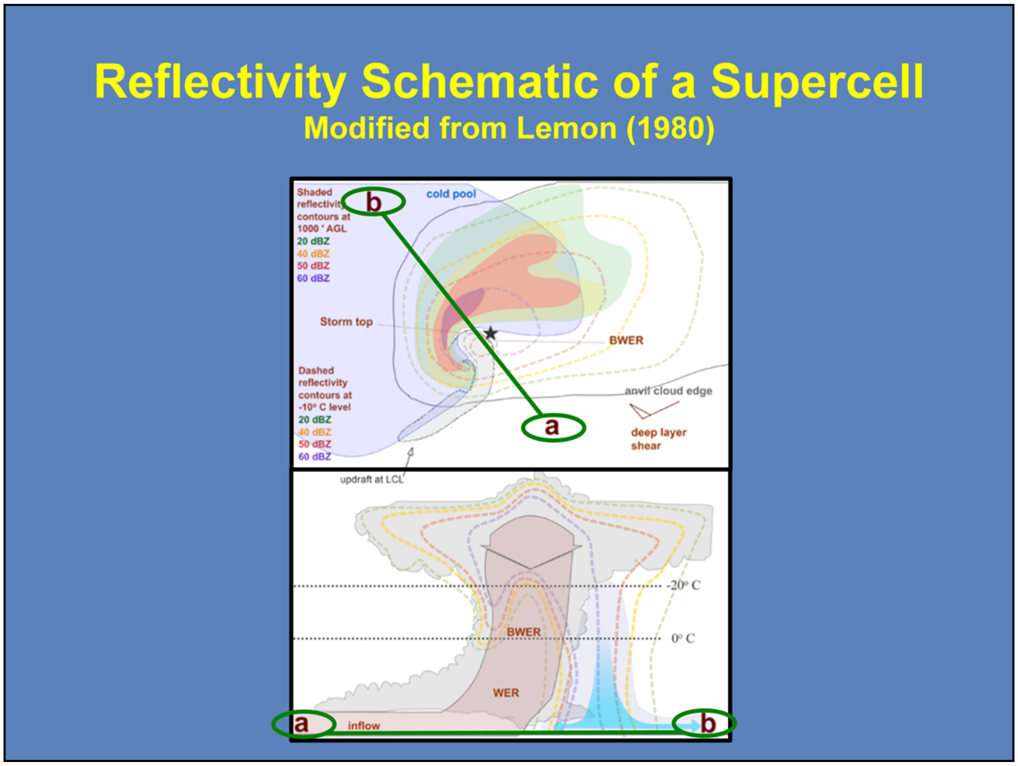

In 1980, Lemon identified radar reflectivity characteristics associated with supercells. The schematic shown here is a conceptual model of the reflectivity structure of a cyclonic, right-moving supercell in the northern hemisphere, modified from his paper. The horizontal plane is shown at the top, while a vertical cross-section shown at the bottom. The letters a and b denote the endpoints of the vertical cross section. Let's discuss those radar reflectivity characteristics using this schematic.

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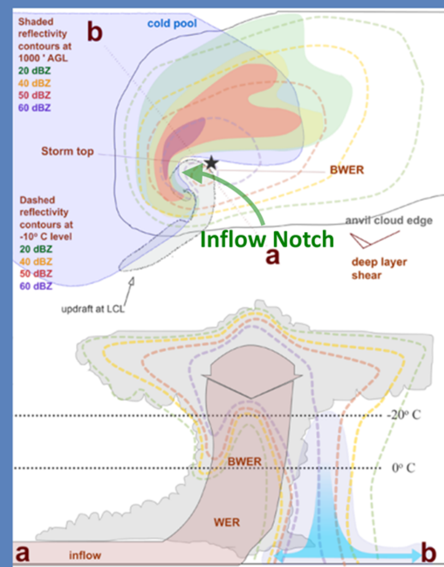
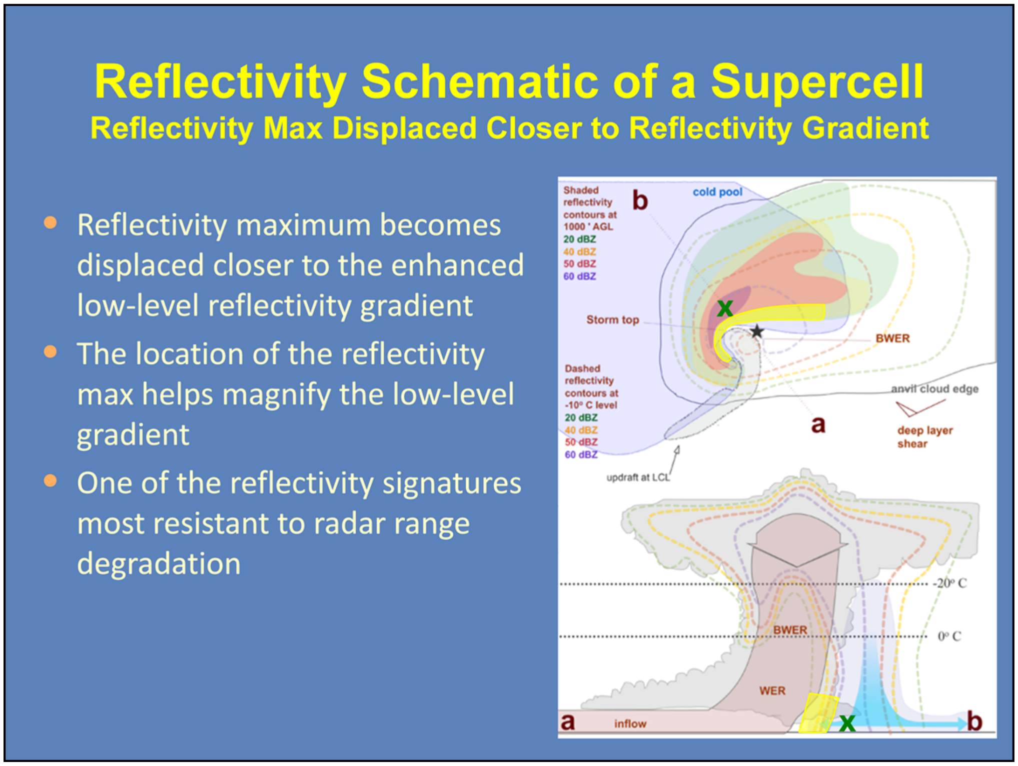
Reflectivity Schematic of a Supercell

Inflow Notch

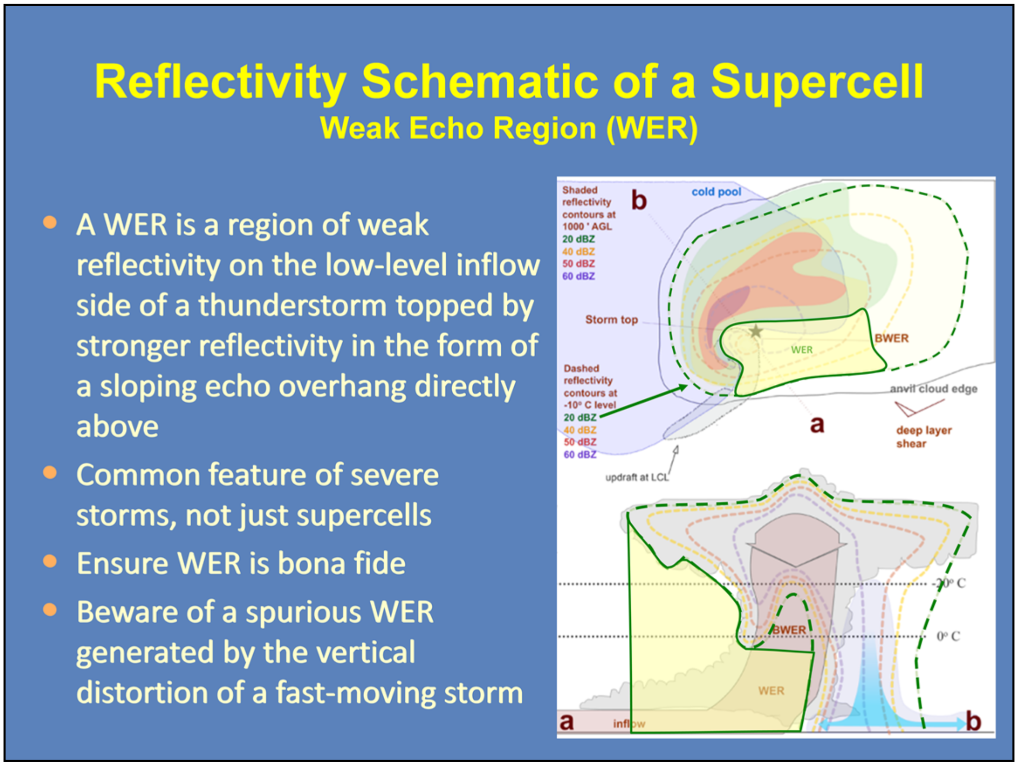
- The inflow notch is a low-level, concave, enhanced reflectivity gradient open to the low-level inflow side of the cell
- Indicates a very strong updraft with enhanced low-level inflow
- One of the reflectivity signatures most resistant to radar range degradation



The most common radar reflectivity characteristic of supercells is the **"Inflow Notch"** which is a low-level, concave, enhanced reflectivity gradient open to the low-level inflow side of the cell. This signature indicates the presence of a very strong updraft with associated enhanced low-level inflow. If the storm is close to the radar, a surface trailing gust front may be seen wrapping into the region of the notch. The inflow notch is one of the reflectivity signatures most resistant to radar range degradation.



Another characteristic that indicates a storm has transitioned into a supercell is **the reflectivity maximum becomes displaced closer to the enhanced low-level reflectivity gradient.** The location of the reflectivity maximum helps magnify the low-level gradient. The enhanced reflectivity gradients is one of the features most resistant to radar range degradation.



A **Weak Echo Region (WER)** is a region of weak reflectivity on the low-altitude inflow side of a thunderstorm topped by stronger reflectivity in the form of a sloping echo overhang directly above. The WER is produced by strong updraft and associated strong storm-summit divergence that carries large amounts of precipitation particles in all directions creating a high reflectivity echo-canopy (slopping echo overhang) over the low-level inflow of a strong or intense convective storm. The slopping nature of the overhang is created when precipitation begins to fall from the far edges of the overhang (visually, the edge of the thick anvil) and descends through the storm relative environmental winds finally reaching the ground in the strong low-level reflectivity echo. A WER is a common feature of severe storms in vertically sheared environments, not just with supercells. Note that the key ingredient that distinguishes storms is the strongly sheared environment. Therefore, features such as the WER, are not found with storms in a weakly sheared environment, such as the pulse storm.

Care must be taken to ensure that a WER is on the updraft and inflow flank of the storm. A bona fide WER should be persistent (~ 10 minutes) and capped by high reflectivities (>45 dBZ) with the base of the slopping overhang beginning as high as the -20 to -30 degrees Celsius environmental temperature. False WERs not capped by strong reflectivity imply weak updraft, such as with an overspreading anvil layer. In addition, the WER should be found above the low-level inflow notch and strong reflectivity gradient.

Because a radar's volume coverage pattern (VCP) samples a storm from bottom to top, beware of a spurious WER oriented in the direction of storm motion generated by the vertical distortion of a fast-moving storm. For example, a storm moving at a speed of 60 kts can have its upper-level scans displaced up to 5 miles in the direction of storm motion.

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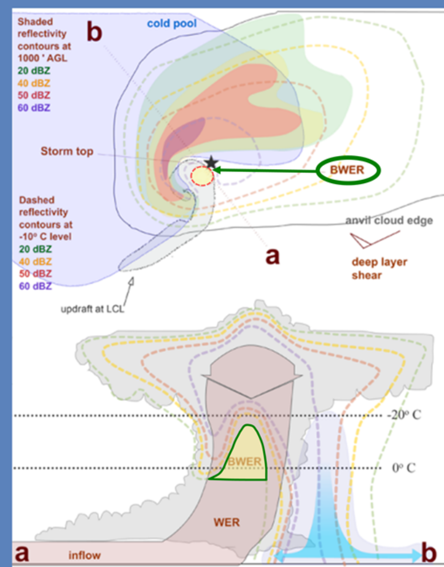
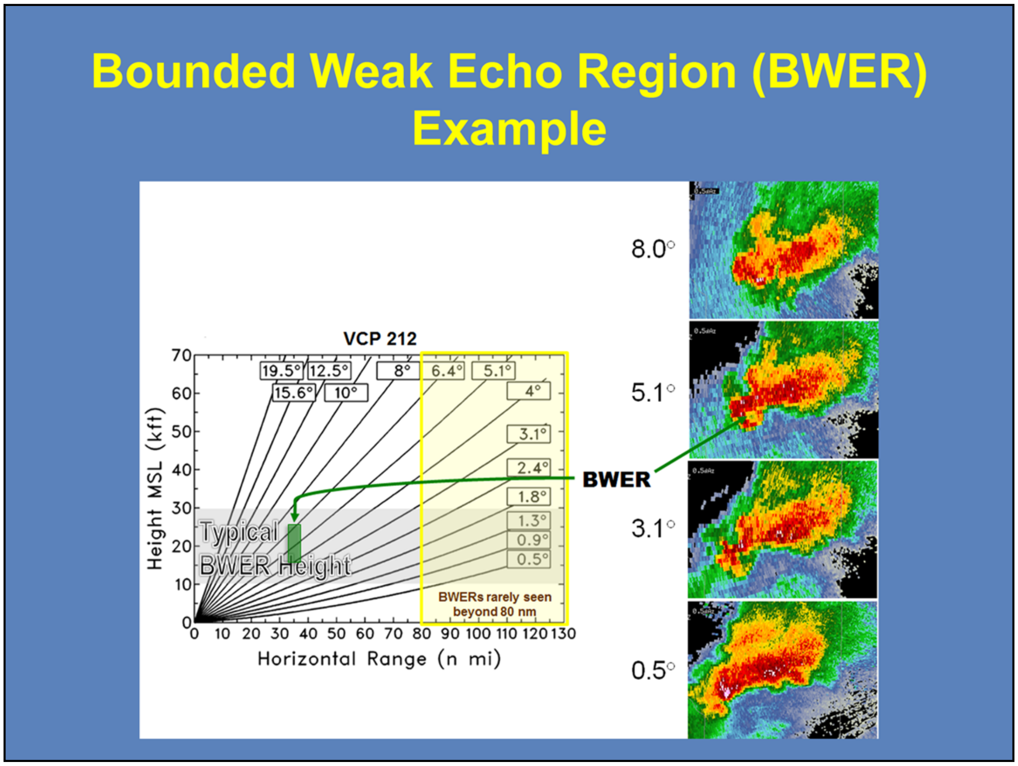
Reflectivity Schematic of a Supercell

Bounded Weak Echo Region (BWER)

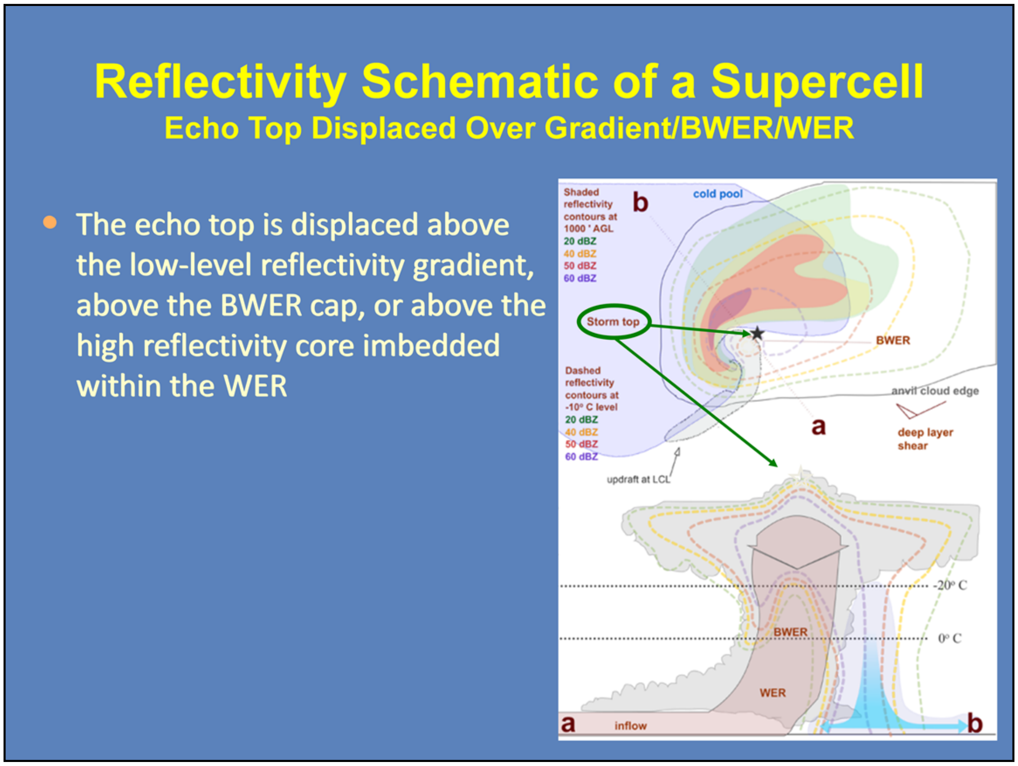
- A BWER is a conically-shaped, nearly vertical channel of weak radar echo, encompassed and capped by strong echo
- Typically found 3-10 km (10-33 kft) AGL and are a few km (1-4 nm) in diameter
  - Observed up to 5-6 nm wide and extending to storm summit
- Small features rarely detected beyond 80 nm
- Almost always associated with very large hail and a supercell



A **Bounded Weak Echo Region (BWER)** (also known as a "vault") is a conically-shaped, nearly vertical channel of weak radar echo, encompassed and capped by strong echo. The cap is composed of large concentrations of supercooled liquid water and rapidly growing hail. The BWER is the core of an intense updraft that carries newly formed hydrometeors to high levels before they can grow to radar-detectable sizes. BWERs are typically found imbedded in the slopping echo overhang and aloft above the apex of the low-level inflow notch. They are typically found 3-10 km (10,000 - 33,000 ft.) AGL and are a few kilometers (1-4 nm) in horizontal diameter. However, on rare occasions, they have been observed up to 5-6 nm wide and extending to storm summit. BWERs are small features rarely detected beyond 80 nm due to radar resolution limitations. The presence of a BWER is almost always associated with very large hail and is associated with a supercell. Note that the BWER is not associated with updraft rotation.



Here is an example of a BWER observed with a supercell 35 nm to the west-southwest of the radar. It is located between 15,000 to 26,000 feet above ground level which certainly falls within the typical BWER range of 10,000 to 33,000 feet above ground level. Remember that BWERs are rarely detected beyond 80 nm due to radar resolution limitations.



Another supercell reflectivity characteristic identified by Lemon is that the **echo top is displaced above the low-level reflectivity gradient, above the BWER cap, or above the high reflectivity core imbedded within the WER.**

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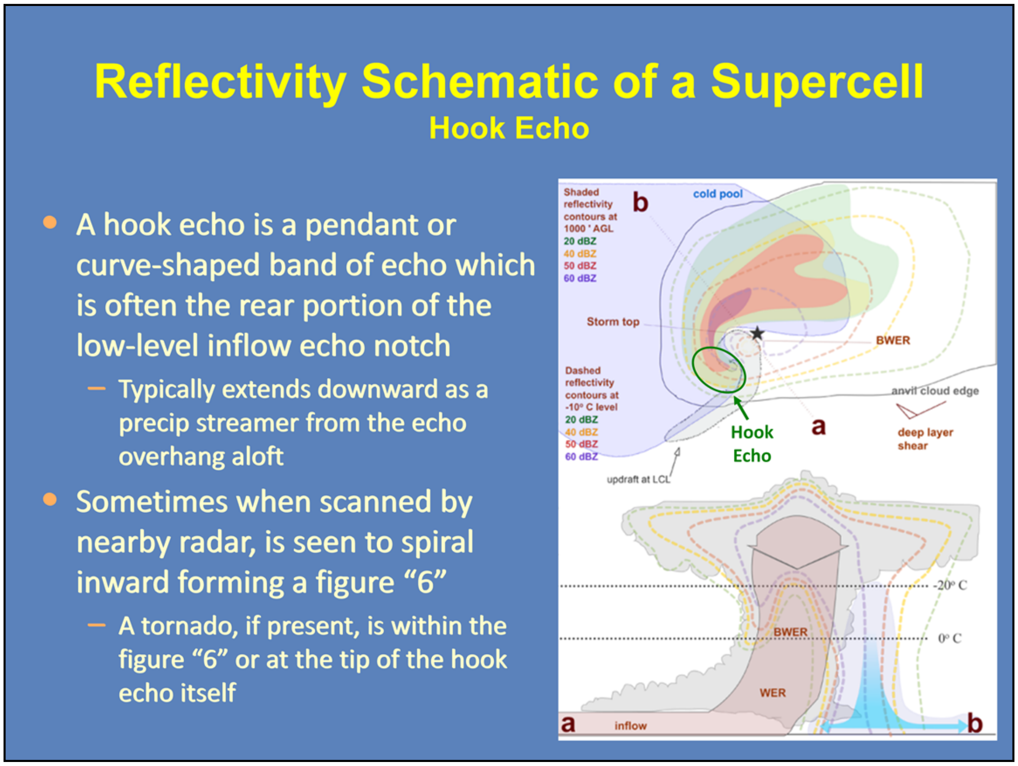
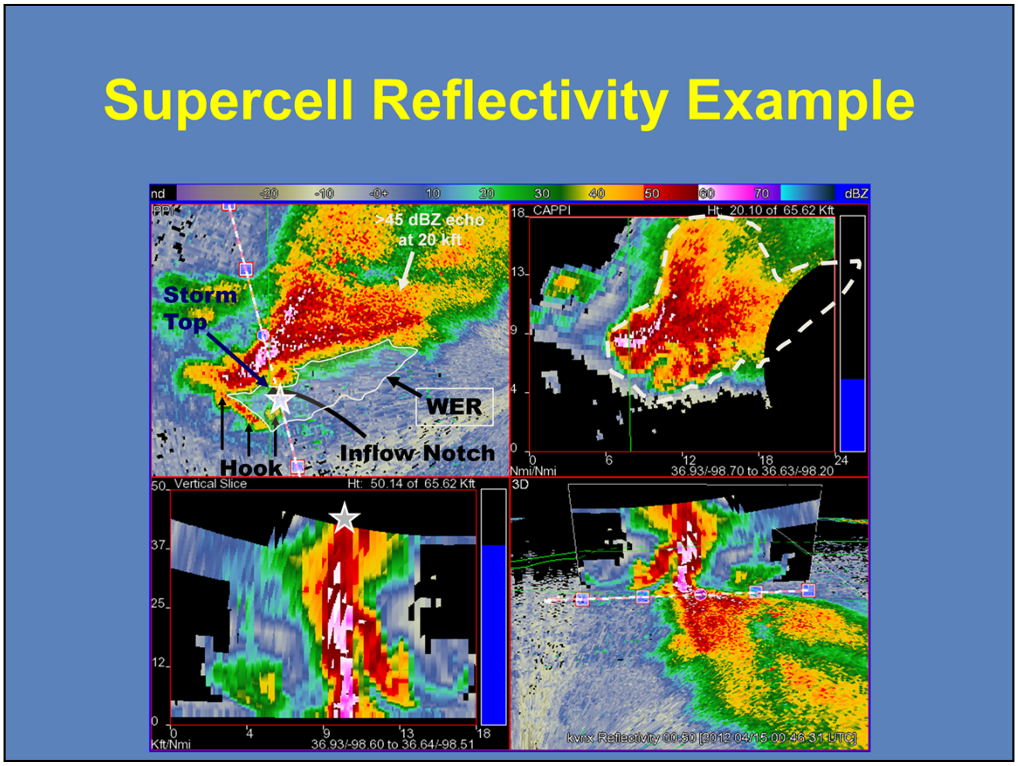


The final supercell reflectivity characteristic identified by Lemon is a **"hook echo"** which is a pendant or curve-shaped band of echo which is often the rear portion of the low-level inflow echo notch. It typically extends downward as a precipitation streamer from the echo overhang aloft. It is often a portion of echo bounding the BEWER on the rear. It may also be precipitation carried downward rapidly by the RFD or associated with the storm mesocyclone. Sometimes when scanned by nearby radar, is seen to spiral inward forming a sharply defined figure "6." A tornado, if present, is within the figure "6" or at the tip of the hook echo itself.



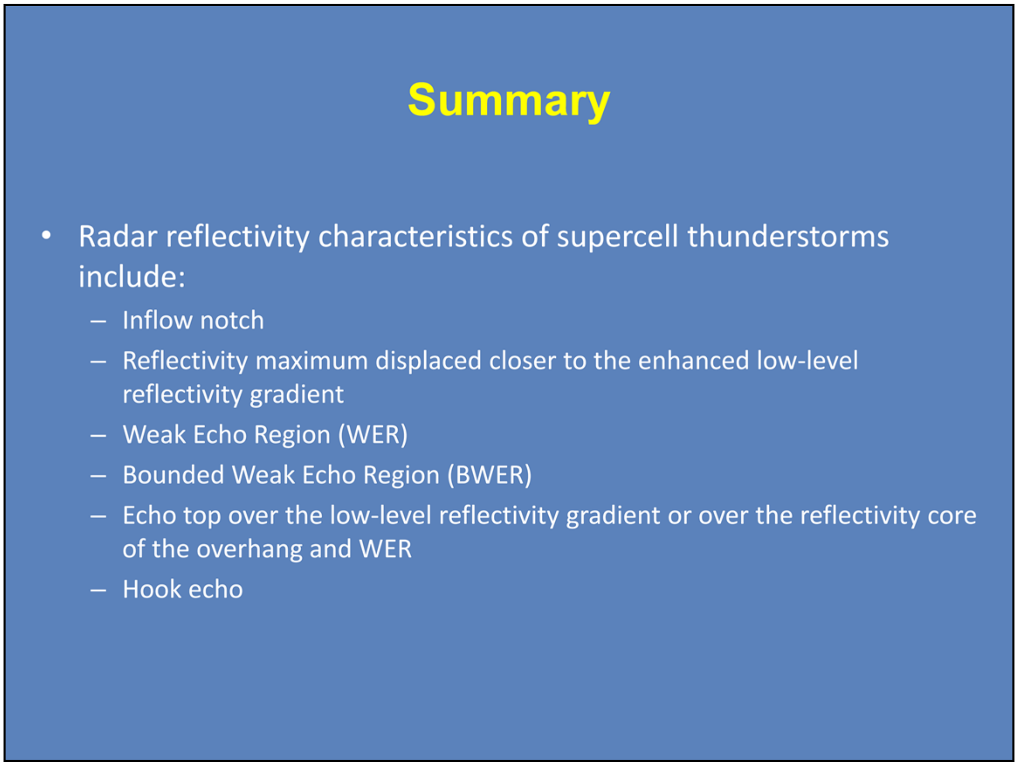
Here's a Base Reflectivity FSI example of the Cherokee, Oklahoma tornadic supercell of April 15, 2012 at 0046 UTC sampled from the Vance AFB WSR-88. It exhibits many of the characteristics we've discussed.

- An **"Inflow Notch"** was located on the low-level inflow side of the cell. This signature indicated the presence of a very strong updraft with associated enhanced low-level inflow.
- **The reflectivity maximum (in pink) displaced close to the enhanced low-level reflectivity gradient.**
- A **Weak Echo Region (WER)** was apparent as a region of weak reflectivity on the low-altitude inflow side of the storm topped by a sloping high-reflectivity echo-canopy directly above.
- The Cherokee supercell did not exhibit a well-defined, vertically-oriented, conically-shaped **Bounded Weak Echo Region (BWER)**, perhaps due to the rapid, cyclic tornado-genesis which occurred with this storm.
- However, the **echo top was displaced above the low-level reflectivity gradient, above the high reflectivity core imbedded within the WER.**
- And finally, a **"hook echo"** was evident as pendant shaped band of echo, spiraling inward to form a sharply defined figure "6." A tornado was located within the circular portion of the figure "6" on this volume scan.
- Bear in mind that both the reflectivity schematic and this example depict a cyclonic, right-moving supercell in the northern hemisphere. An anticyclonic, left-moving supercell in the northern hemisphere would appear as a mirror image of this.

Beware of Relying on Just One Signature or Volume Scan

- All weather radars have spatial and temporal limitations which can hinder your analysis
  - Resolution may be insufficient to resolve small features like BWERs
  - Radar beam may overshoot lower-level features
  - Features may occur between volume scans
- Also, beware of the collapse phase of some supercells when all these features disappear and the storm produces a tornado!

Beware of relying on any one signature or volume scan in isolation when trying to identify a supercell. All weather radars have spatial and temporal limitations which can hinder your analysis of storm structure. Radar resolution may be insufficient to resolve smaller features at longer ranges such as BWERs or even hook echoes. The radar beam may overshoot lower-level features such as some hook echoes and WERs. Features may occur between volume scans. Plus, this lesson doesn't discuss deviant motion from the mean wind which is perhaps the most easily identifiable and reliable supercell characteristic. Finally, beware of the "collapse phase" of some supercells when all of these distinctive features disappear and the storm produces a tornado!



Radar reflectivity characteristics of supercell thunderstorms include:

- A strong reflectivity gradient bounding a concavity or “inflow notch”

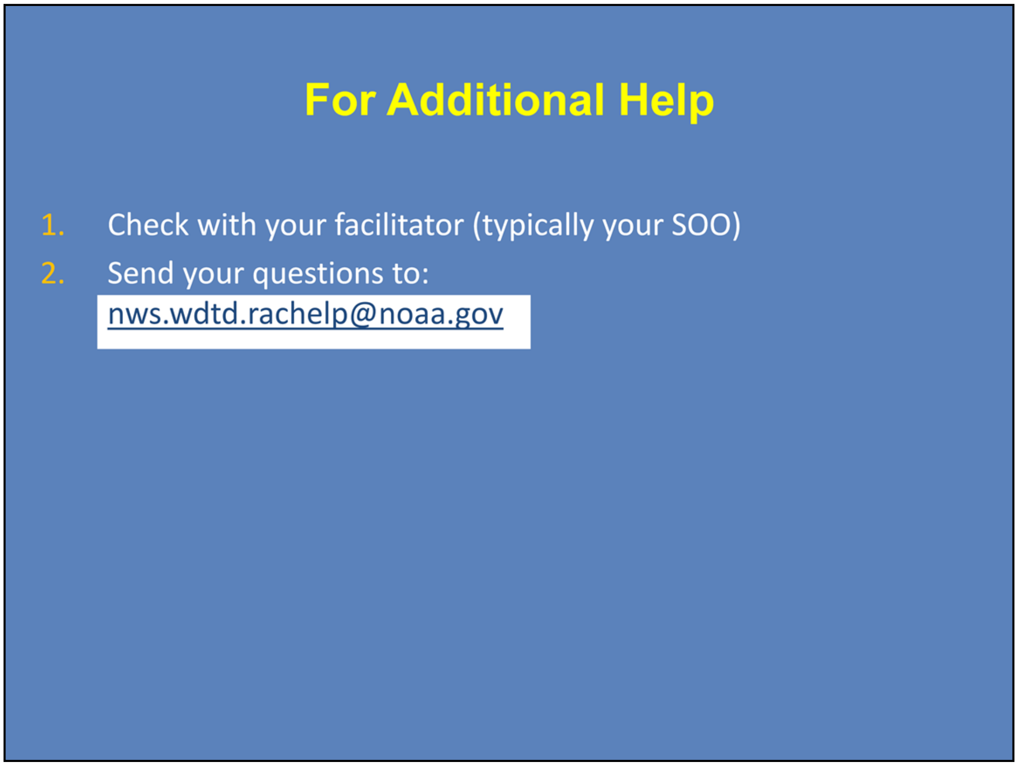
- Reflectivity maximum displaced closer to the enhanced low-level reflectivity gradient

- Weak Echo Region (WER)

- Bounded Weak Echo Region (BWER)

- Echo top over the low-level reflectivity gradient or over the reflectivity core of the overhang and WER

- Hook echo



For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.

Welcome to the Convective Storm Structure and Evolution's lesson on supercell velocity signatures. This lesson is approximately 20 minutes long.

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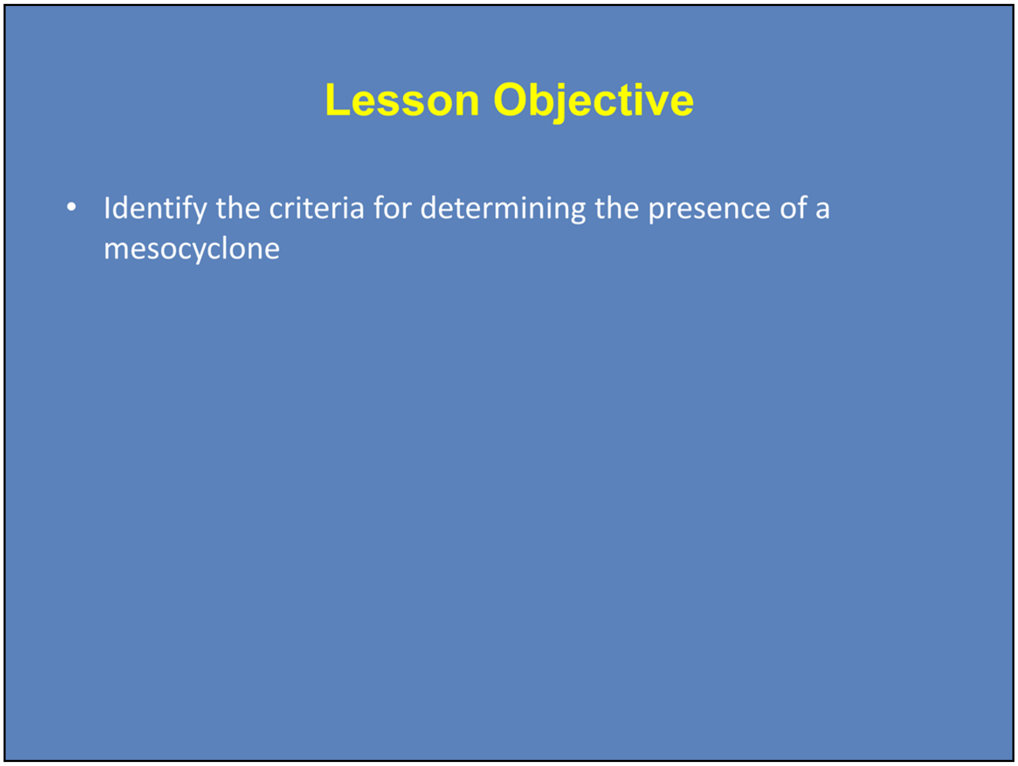

In many cases, a warning forecaster may have difficulty in distinguishing an ordinary from a supercell based solely on the reflectivity pattern. This is one of the major reasons that we have the WSR-88D network; to identify velocity patterns that compliment reflectivity by playing a critical role in identifying a supercell. This lesson describes the structure and morphology of supercell velociy signatures; focusing on aspects of the mesocyclone structure.

Identify the criteria for determining the presence of a mesocyclone

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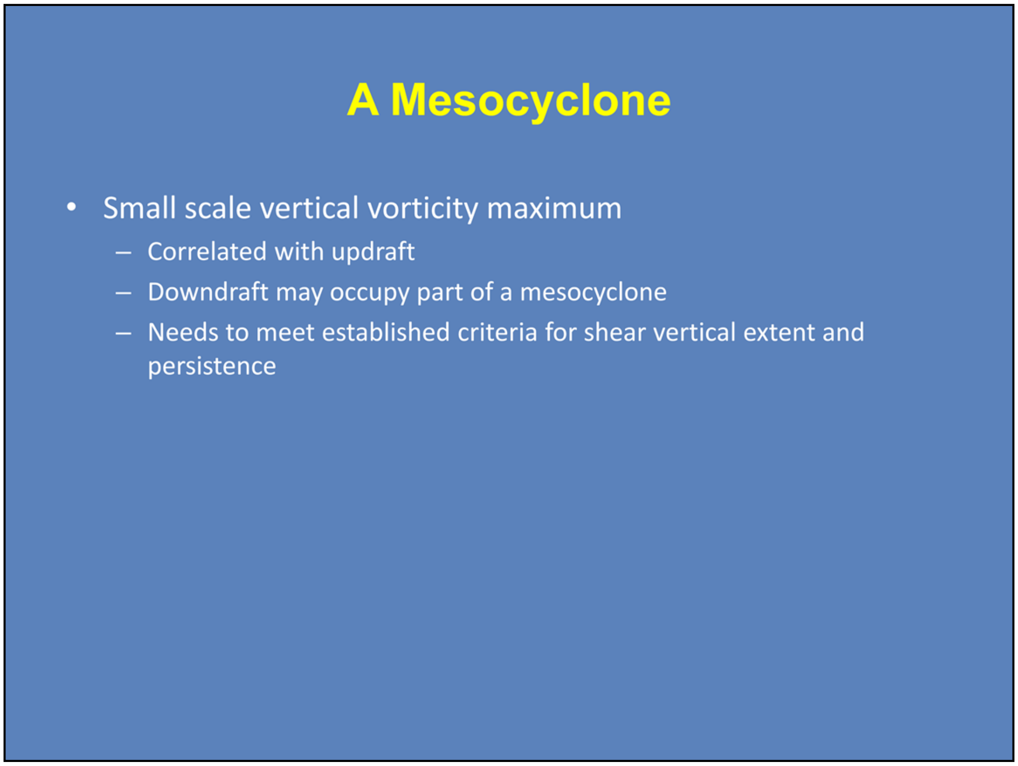
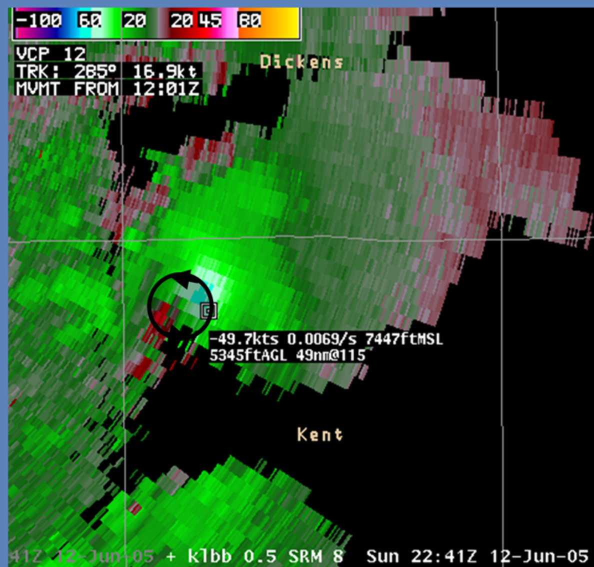
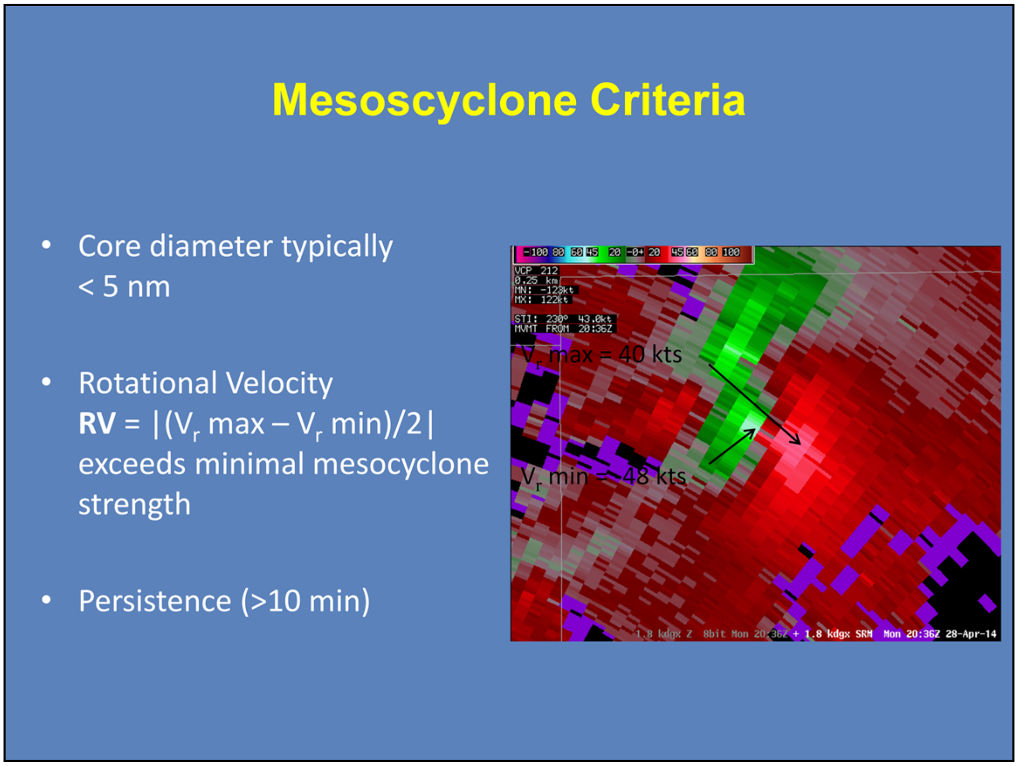


A localized region of vertical vorticity partially, or fully embedded within an updraft of a DMC (termed a mesocyclone) is one of the defining characteristics of a supercell. **By definition, a mesocyclone is a small-scale vertical vorticity maximum closely associated with the updraft, and downdraft, of a convective storm that meets or exceeds established criteria for shear, vertical extent and persistence**. Each of these criteria will be discussed.

Basic Mesocyclone Structure:
2-D velocity profile



Mesocyclone velocity structure is similar to that of a Rankine Combined Vortex. The core of the mesocyclone rotates as a solid body with the tangential velocity proportional to radius. Beyond this core, the velocity decreases exponentially with increasing radius from the mesocyclone center. Since only the radial velocity component is detectable from Doppler radar, only the radial components of the velocity can be detected. Therefore, the mesocyclone appears as a range adjacent couplet of inbound and outbound velocity.



To establish the validity of a mesocyclone, we use a set of criteria for shear persistence and vertical depth. A circulation feature is labeled a mesocyclone when:

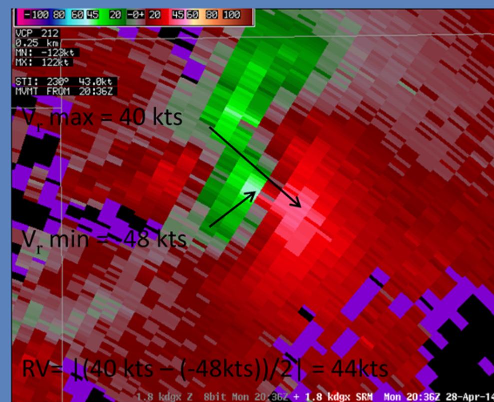
The core diameter (distance between the maximum and minimum velocities) is < 5 nm, **and**

The rotational velocity (RV = |(Vr max – Vr min|) / 2) equals or exceeds minimal mesocyclone strength. **Vr** min (Vr max) is the minimum (maximum) radial velocity in the circulation.

The feature persists for at least 10 minutes.

How to Measure RV

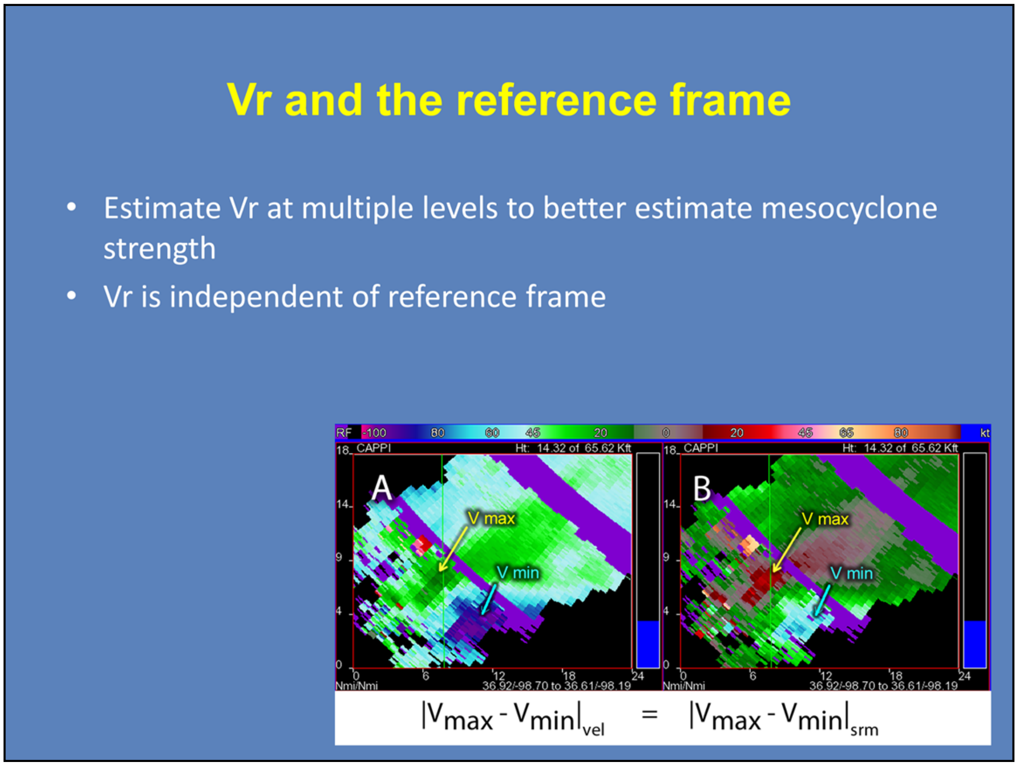
- Rotational Velocity
  RV = $|(V_r \max - V_r \min) / 2|$
- Can use the Vr shear tool in AWIPS
  - Typical mesocyclone values are 10^{-2} s^{-1}
  - Vr shear is very sensitive to the length of the baseline
  - Pick the endpoints to overlay the middle of the gates containing Vr max and Vr min.



The inputs into calculating RV should represent the maximum and minimum velocities as illustrated by the inset in this figure. Note that the minimum and maximum velocities that contribute to the calculation of rotational velocity should be measured using representative peak values of the data levels in the velocity or SRM products.

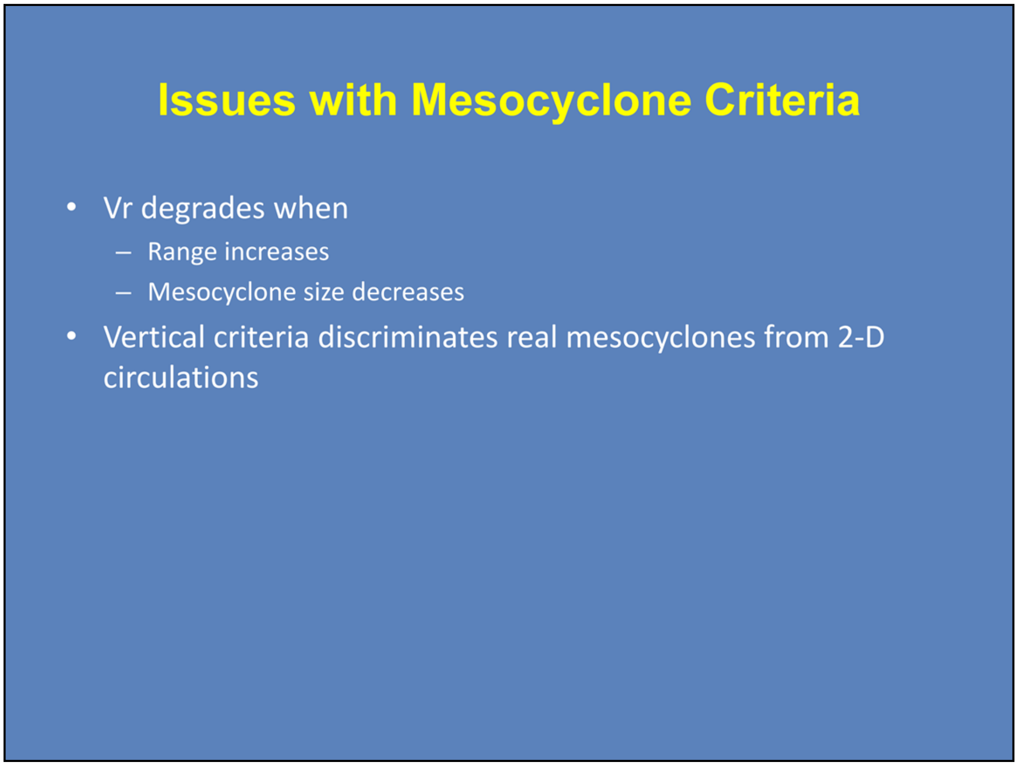
In this example, the maximum Vr is 40 kts for a good representative value, and the minimum Vr is -48 kts. Taking the difference, dividing by two, and taking the absolute magnitude reveals a RV of 44 kts.

Vr shear is calculated by dividing RV by the distance between Vmin and Vmax. It can be easily calculated using AWIPS using the Vr shear tool. Values are on the order of 10**-2 s**-1 for mesocyclones. However, Vr shear can change by orders of magnitude just by changing the baseline distance without any actual increase in mesocyclone intensity. **Therefore, V**r shear should be calculated with great caution and consistency through successive volume scans. Pick the endpoints to overlay the middle of the gates containing Vr max and Vr min. Be aware of that you will need to adapt your baseline as the actual mesocyclone diameter changes. Estimating mesocyclone strength from RV alone is just as valid as that from Vr shear.



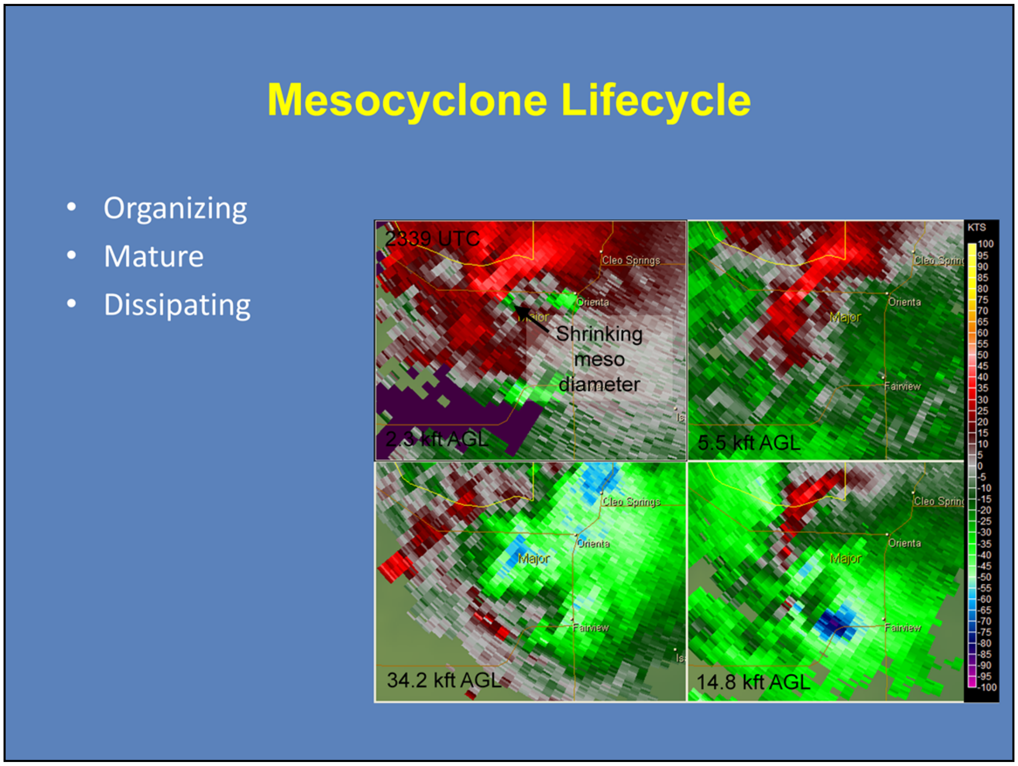
Note that estimating mesocyclone strength is more representative when assessing Vr from multiple levels rather than one alone.

A mesocyclone need not have in- and outbound velocities. The velocity difference, rotational velocity and shear across a mesocyclone are identical no matter the motion of the reference frame. Using the example in Figure 7-92, a forecaster may sample different velocity maxima and minima between the velocity and the SRM product for the mesocyclone moving toward the radar to the northeast. However, the Vr is identical.



Establishing a minimal rotational velocity threshold requires knowledge of the distance of the feature, and the size of the supercell. **As radar sampling resolution degrades either by distance or by circulation size, the warning forecaster must reduce the minimal rotational velocity that discriminates mesocyclones from weaker circulations**.

The vertical criteria are required because of the number of shallow circulations uncorrelated with deeper vertical velocity features. Deep, vertically correlated circulations are most likely associated with updrafts and downdrafts because of vertical vortex stretching and advection of vorticity.



Mesocyclones typically undergo a life span where there is an organizing stage, mature stage and dissipating stage. The typical organizing mesocyclone begins at the level of maximum tilting or in the mid-levels of an updraft. The mesocyclone then begins to build downward and upward. The mid-level mesocyclone is dominated mostly by updraft. If the radar is close enough to the circulation, a convergent signature may be detected in association with the mesocyclone in the lowest slices.

An idealized mature mesocyclone has significant low-level convergence (panel 'G' in Figure 7-93), nearly pure rotation at mid-levels (panels 'C' and 'E' Figure 7-93), divergent rotation at upper-levels (panel 'A' in Figure 7-93).

In the decaying phase of a mesocyclone, the convergent rotational signature in the low-levels gradually transitions to that of divergent rotation as outflow begins to dominate. Mesocyclone depth decreases as does the maximum rotational velocity. As the mesocyclone weakens, it also broadens and becomes diffuse. **If the mesocyclone is tornadic and undergoes a dissipating stage, the tornado could persist for a period of time after all evidence of the parent mesocyclone has dissipated.**

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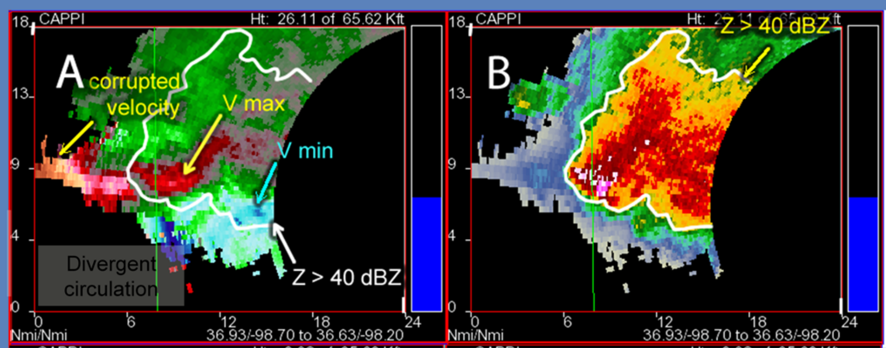
Mature Mesocyclone



The example in Figure 93 shows a little more complexity than the ideal model. This is because the vorticity from the occluding low-level mesocyclone has been advected upward by the updraft within the larger mid-level mesocyclone producing an interior couplet of peak velocities.

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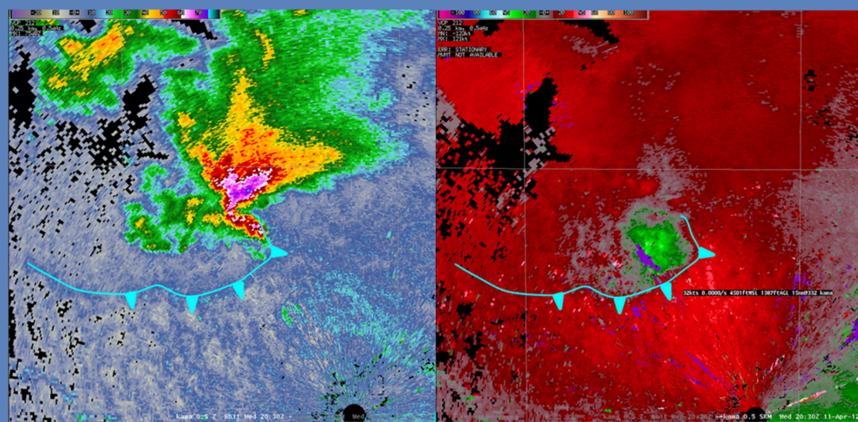
Mature Mesocyclone
Rear Flank Downdraft



The lower half of a mature mesocyclone is occupied by the rear flank downdraft, usually on its trailing side. The rear flank downdraft can be marked by the presence of strong localized convergence between the inbound to outbound velocities (Figure 7-94). Do not confuse the gust front with the RFD itself. The RFD is often associated with the hook, or pendant echo, and is a divergent outflow that creates the gust front. But this RFD divergence is often difficult to identify in contrast to itsassociated gust front.

Additionally, the convergence along the RFD gust front should not be mistaken for the transition from in- to outbound velocities in a symmetric vortex.

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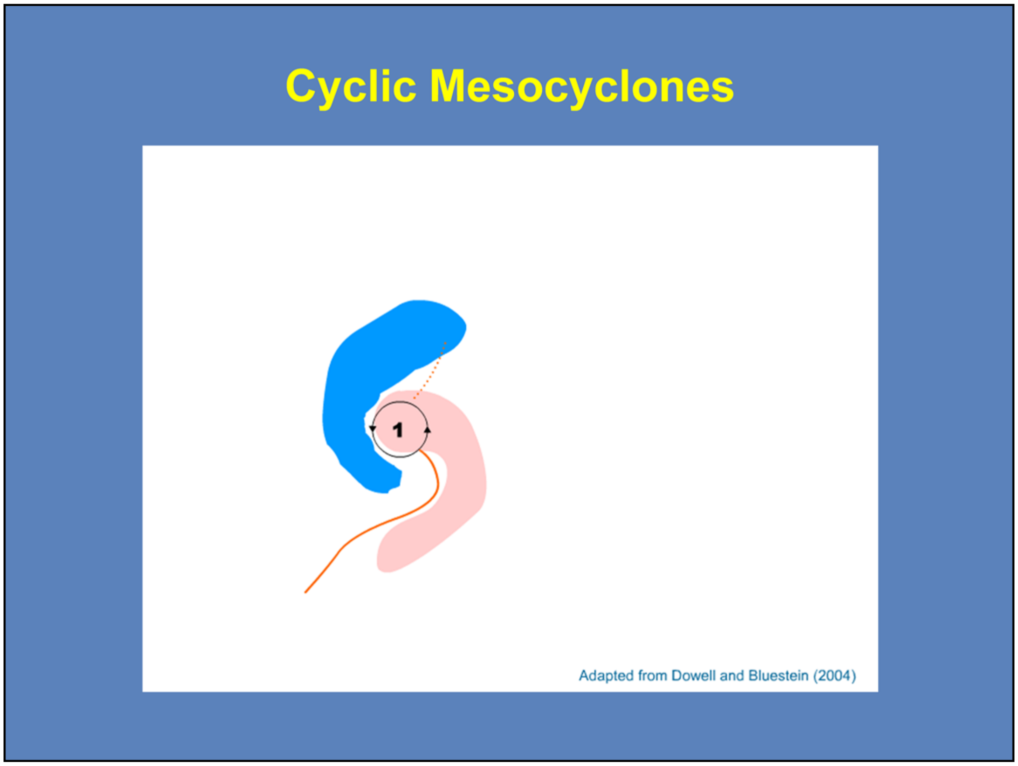
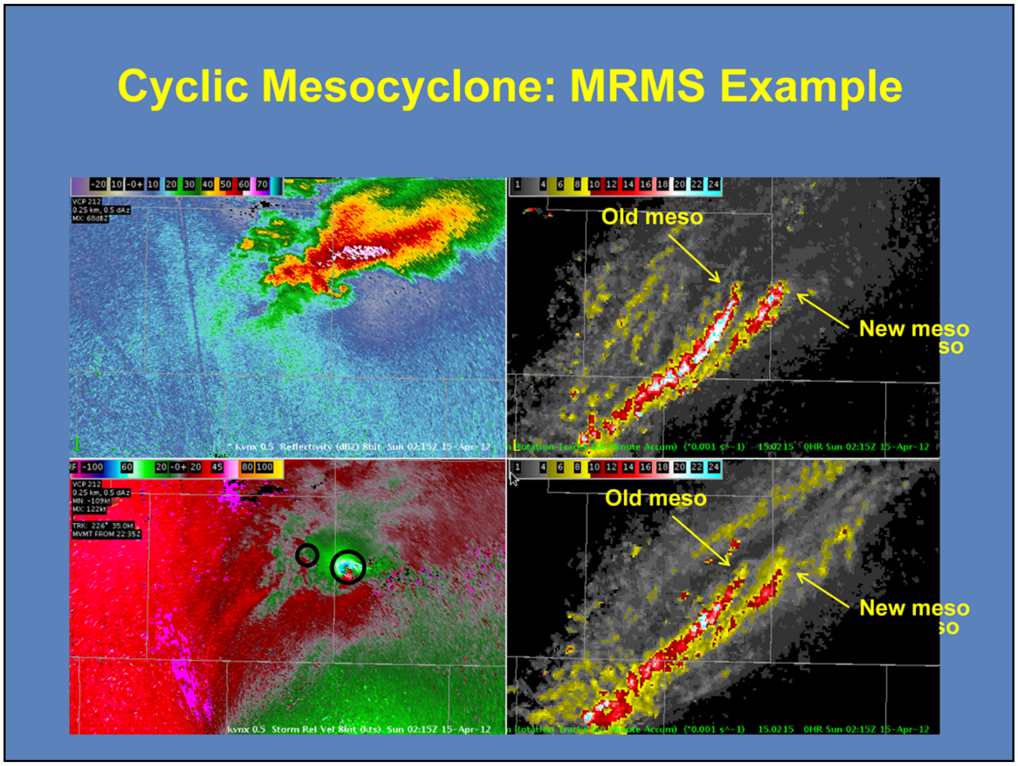


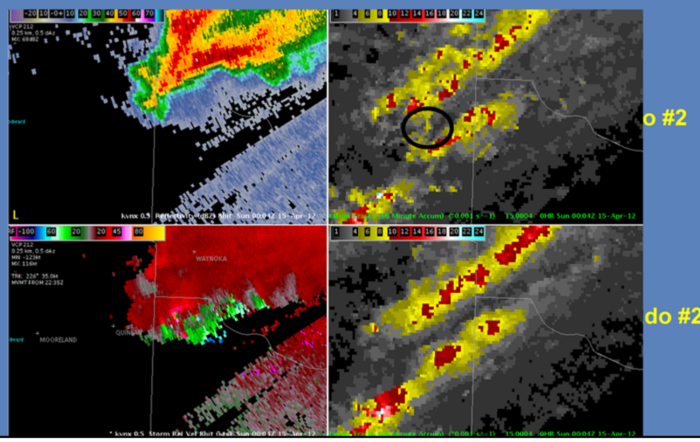
A supercell may produce more than one mesocyclone during its lifetime. In the flash movie coming up, you will see the first mesocyclone mostly occupied by updraft (denoted by the orange filled area) with downdraft on its backside (denoted by the blue area). The first mesocyclone typically takes the longest time to mature as the supercell remains outflow deficient. **Successive mesocyclones mature much more rapidly as they have the advantage of stronger lifting and vortex tilting from a stronger gust front** (denoted by the light brown border) . The life spans of successive mesocyclones may or may not be longer than the first one. The first mesocyclone extends to low-levels as the RFD reaches the ground. When the RFD matures, the outflow wraps cyclonically around the center of circulation, eventually closing it off from the inflow. **If the RFD is thermodynamically unstable, the primary mesocyclone can continue for an extended time**. However, the leading edge of the gust front associated with the RFD can quickly produce successive updrafts and mesocyclones owing to increased convergence and vertical low-level vorticity. In turn, the successive mesocyclones become wrapped by local RFD enhancement, and the process continues for possibly several hours. A family of tornadoes is produced in this fashion from one supercell.



Here's an example of a storm with a cyclic mesocyclone. We'll examine it using Multi-Radar/Multi-Sensor (MRMS) Rotation Tracks. At this time, a tornado is ongoing, as seen in the 0.5 degree Z panel. On the Low-Level Rotation Tracks product, this meso is exhibiting very high azimuthal shear levels near the surface. Strong rotation is also evident at mid-levels. At the next timestep, the old mesocyclone is still very strong and likely still producing a tornado. But now another Rotation Track is beginning to form south of the current one, indicating the growth of a new meso. Nine minutes later, the old meso has drastically weakened and arced to its left. It is probably not tornadic at this point. The new meso, however, as explained on the previous slide, has quickly strengthened into a large, tight circulation that could be tornadic. Finally, just five minutes later, the first meso has almost completely dissipated and the new one has taken over and is most likely producing a tornado.

Cyclic Mesocyclone: MRMS Limitations

- Can be difficult to view cycles when:
  - Training meoscyclones cause overlapping Rotation Tracks
  - Azimuthal shear kernel size obscures small vortices



MRMS Rotation Tracks are not able to help identify cyclic mesocyclones in every situation. It can be difficult to view cycles when training meoscyclones cause overlapping Rotation Tracks and when the MRMS azimuthal shear kernel size obscures small vortices. This is the same supercell we examined on the previous slide, but at this point in time it cycles through 2 tornadoes without obvious signatures. A brief tornado was reported at this time, but the velocity couplet is being partially obscured by beam blockage and MRMS Rotation Tracks don't show strong azimuthal shear. Five minutes later, the couplet is still obscured and the Rotation Tracks have belatedly noted some strong rotation. This delay could have been due to the relatively fast speed of the storm. Five minutes after that, another brief tornado was reported. Again, the vortex is so small the azimuthal shear kernel doesn't register it. The velocity couplet is visible but still partially obscured and another possible mesocyclone appears to its west. By our last timestep, the original couplet still isn't sampled well and there still appears to be another to its west. Also note that there's been some uncertainty of the northern edge of this storm's Rotation Track because of its close proximity to a previous one.

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Summary

- **Shear Criteria**
  - Shear varies as to a threshold.
  - Vorticity ~ 10^{-2} s^{-1}
- **Size**
  - typically less than 5 nm in diameter
- **Vertical continuity**
  - Mesocyclones should extend through at least two elevation slices
- **Persistence**
  - Mesocyclones should typically last at least 10 minutes

The criteria for determining a mesocyclone is:

Shear Criteria

Shear varies as to a threshold. Some mesocyclones may be poorly resolved and yet still carry considerable severe weather risk.

Mesocyclone vorticity lies on the order of 10–2 s-1

Size

Mesocyclones are typically less than 5 nm in diameter

Vertical continuity

Mesocyclones should extend through at least two elevation slices

Persistence

Mesocyclones should typically last at least 10 minutes


Radar & Applications Course
(RAC)
Convective Storm Structure and Evolution
Supercell Morphology: Dual-Pol WSR-88D Signatures
Warning Decision Training Division (WDTD)

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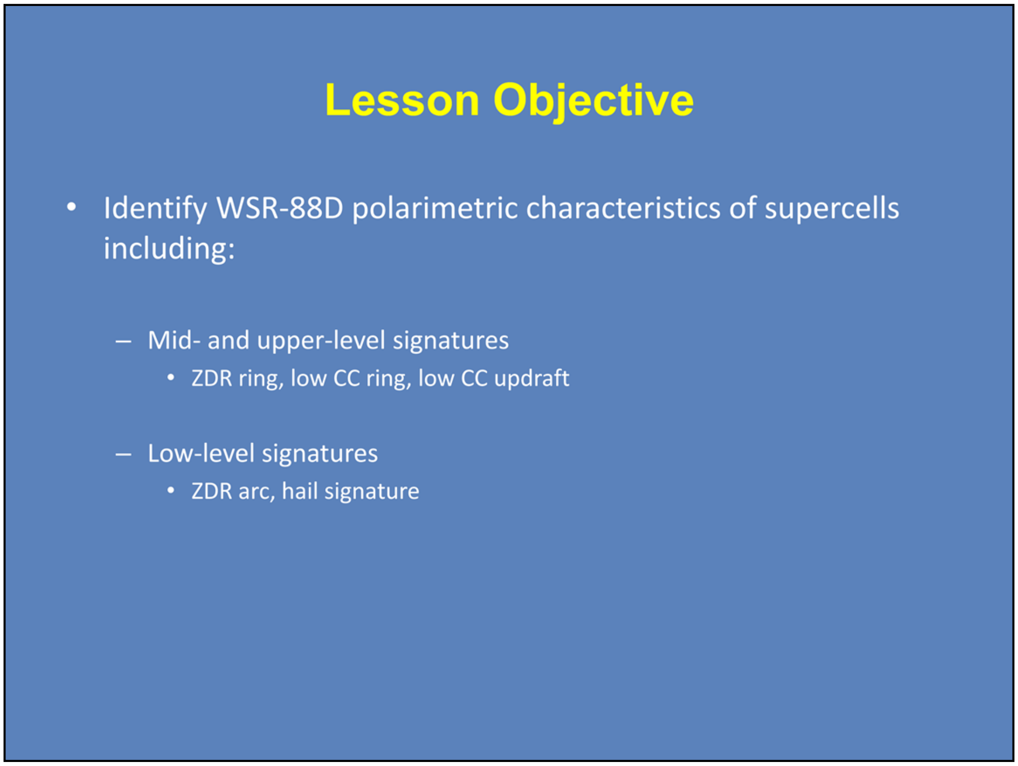

As the supercell produces distinctive patterns in reflectivity, the WSR-88D polarimetric variables like ZDR, CC and KDP, also produce their own distinctive signatures.

The objective is: Identify WSR-88D polarimetric characteristics of supercells.

Mid- and upper-level signatures

ZDR ring, low CC ring, low CC updraft

Low-level signatures

ZDR arc, hail signature

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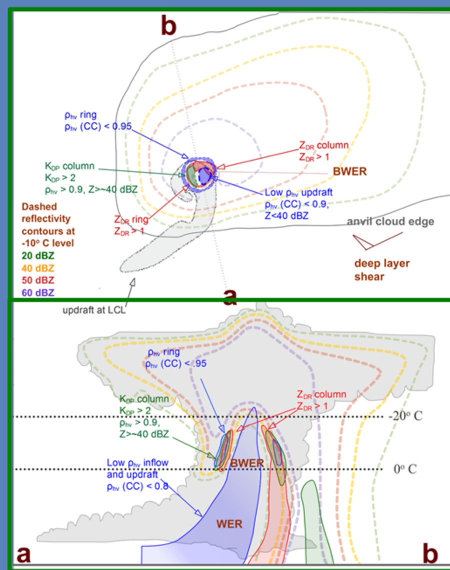
Mid-level polarimetric conceptual model of a Supercell

Conceptual model website
http://www.wdtb.noaa.gov/courses/rac/severe/objects/SvrModel/

Keep for later use



Web Object

Address:
http://www.wdtb.noaa.gov/courses/rac/severe/objects/SvrModel/

With dual-pol WSR-88D we now have the opportunity to observe important micro-physical structures that are unique to supercells. Detecting these structures is critical to successful warning decision making. This lesson overlaps a little with the lesson on dual-pol updraft signatures, and then it expands to describe other common radar-based dual-pol characteristics associated with supercells.

This conceptual model shows the dual-pol signatures that commonly accompany supercells at mid-levels within or close to the updraft. The flash version of this model will appear shortly and we'll frequently use it as we go on through this lesson. Since we've covered some common signatures with other lessons, we'll focus on the ones that most likely appear with supercells.

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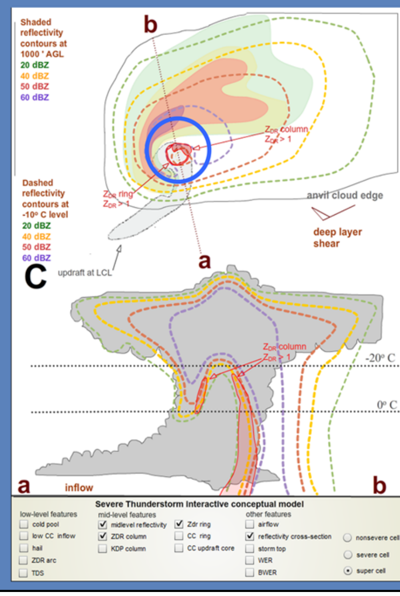
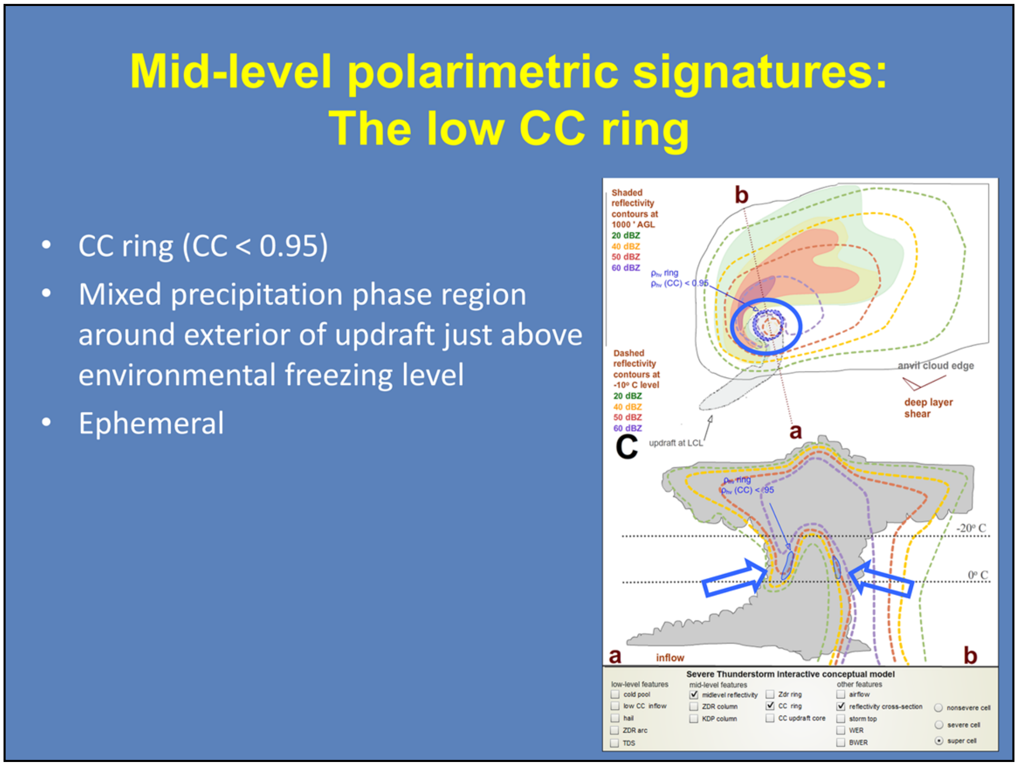
Mid-level polarimetric signatures: The ZDR ring

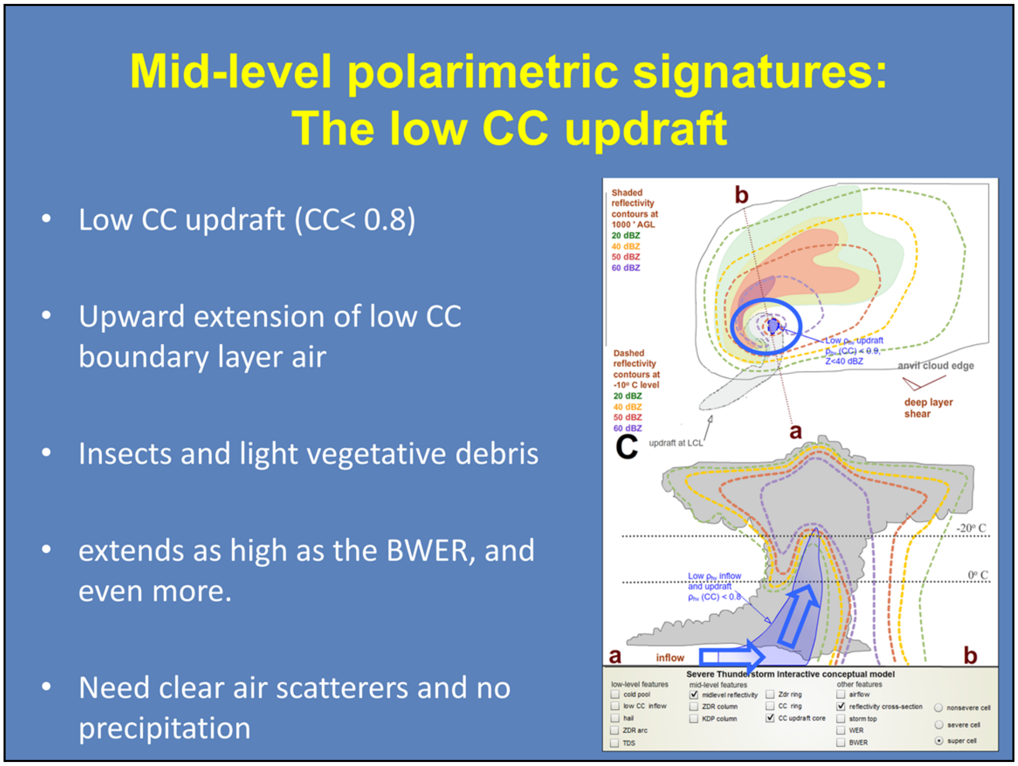
- Large liquid drops wrap around the BWER at mid-levels to form the ZDR ring.



The ZDR ring often surrounds a BWER between the environmental freezing and -20 deg C levels. The reason that a ring forms has been attributed to the strong rotation within the updraft. Large water droplets rising within the updraft's outer edges advect around the center of the mesocyclone resulting in a ring-like structure. Not all supercells exhibit a ZDR ring and others exhibit a partial ring.



Above the environmental freezing level and along the outermost perimeter of the updraft just outside the ZDR ring lies a ring of reduced CC values from 0.9 to 0.95. This ring forms as frozen particles from the main core interact with a raised melting layer of the updraft resulting in a region of mixed phased precipitation. Here, graupel, abundant liquid water, and growing hail are likely present. Because the perimeter of updraft where this occurs is very narrow, and the updraft edge changes quickly, this ring may not always be apparent on radar. This is especially true as the supercell undergoes mesocyclone occlusion processes when the updraft is partially disrupted.



Sometimes, a core of very low CC air can be seen within the BWER at mid-levels.

This core appears to be an upward extension of the low CC boundary-layer inflow ahead of the storm. This signature is dependent on the low-level inflow exhibiting low CC.

This low CC inflow may be associated with non-meteorological scatterers, such as light vegetative debris or insects. In some stronger storms, this signature can even be observed at altitudes above the BWER. If the inflow is relatively clear of insects and light vegetation debris, then this signature may not appear. A low CC updraft column might also fail to appear if precipitation from this storm or an adjacent storm is entrained by the updraft. If so, then the precipitation signal will dominate.

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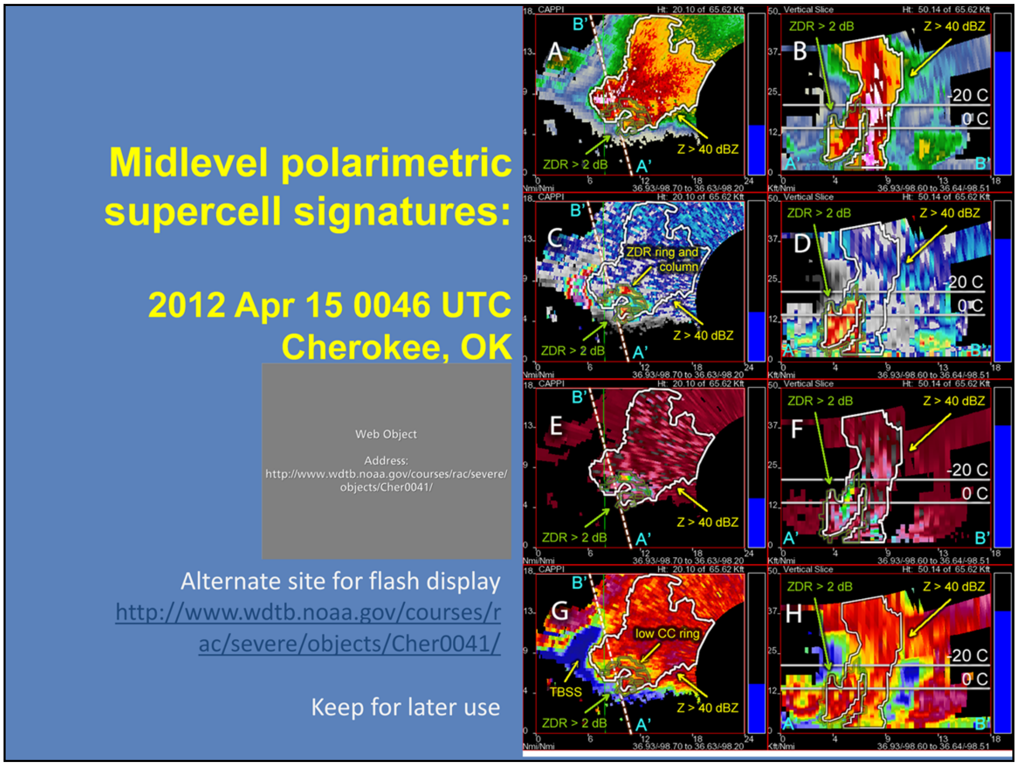
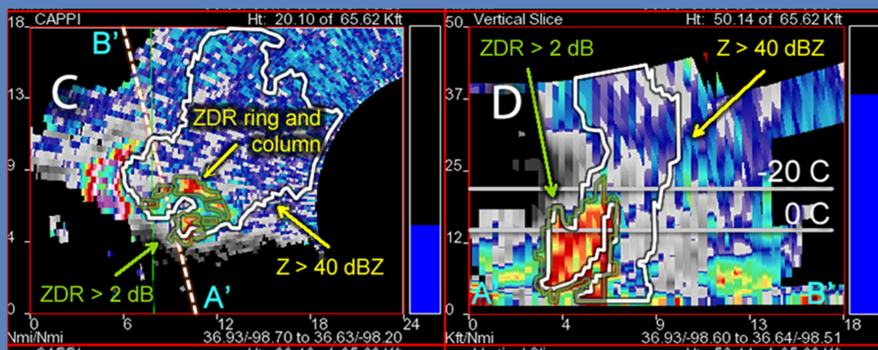


The example supercell thunderstorm, a cyclic tornado producer, was about to produce another tornado. Let’s begin with a discussion of the features at the mid-levels and in the vertical cross-section.

An interactive flash display of this storm will appear with many CAPPI levels for this, and the previous time period. Keep it handy for the rest of the lesson.

Midlevel: ZDR



Note that the vertical cross-section shows what appears to be a substantial BWER. In actuality, it was still a WER since there was an open end pointing into the cross section (Figure 7-97A, B). The reflectivity CAPPI was set at the -15 deg C level (20 kft ARL), high enough to isolate the ZDR column and nearly a ring (Figure 7-97C). Since the WER didn't really close off into a BWER, the ZDR ring also had an open end.

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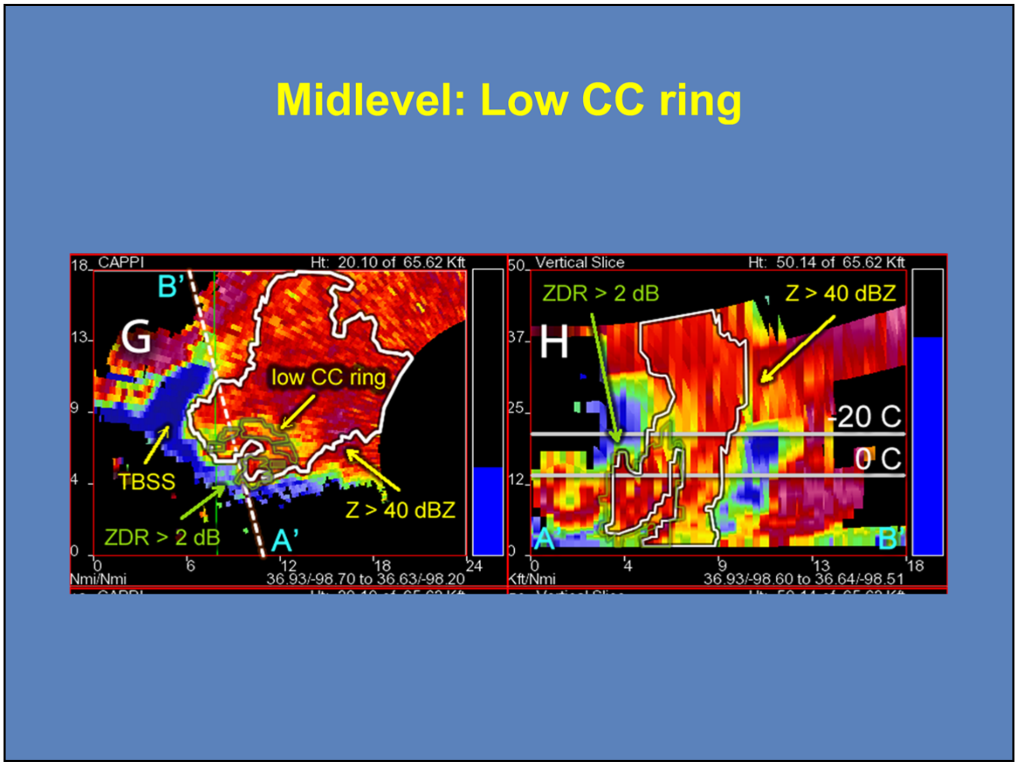


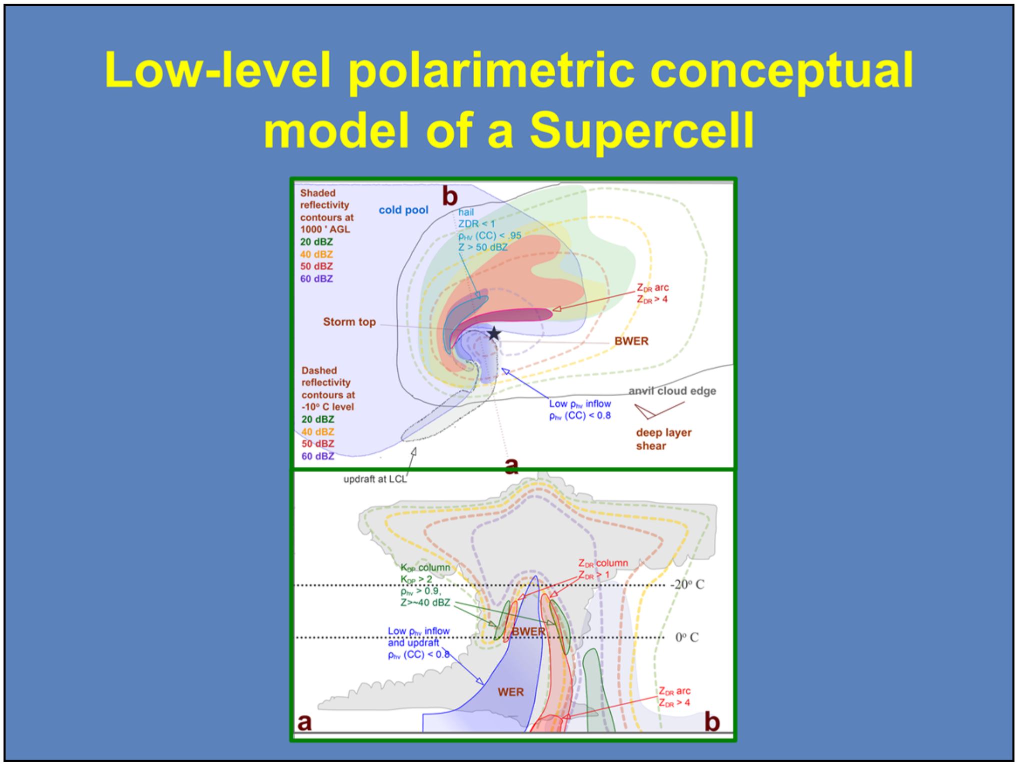
Likewise, the low CC region or arc in Figure 7-97G could also have been a ring had the high reflectivities closed off a BWER at -15 deg C. However, the same mechanism applies, and so, for convenience, we identify the low CC arc with the same name as the conceptual low CC ring.

One dual-pol signature that is notably absent is the low CC updraft signature. While the storm inflow CC was notably low (near the A' in Figure 7-97H), there was a substantial collapse of the high reflectivity and high ZDR echo overhang that resulted in some light precipitation entraining into the updraft, weakening the low CC signature within its core.

In Figure 7-97G, there is a field of very low CCs west (down-radial) of the precipitation core. This feature could be identified as the low CC updraft air since it appears to be in a notch surrounded by high CC echoes, but this is a Three-Body Scatter Spike (TBSS). Be skeptical of any adjacent low CC echo masquerading as a low CC updraft signature if it is:

1. Down-radial of intense reflectivities (> 60 dBZ) and, therefore, could be a TBSS, and

2. Contains a low signal-to-noise ratio (high spectrum width) on the edge of a reflectivity area.



At low-levels, the updraft signatures are replaced with others that are commonly associated with supercells. Low-levels are defined as typically from the ground to 3 km AGL for most surface-based supercells.

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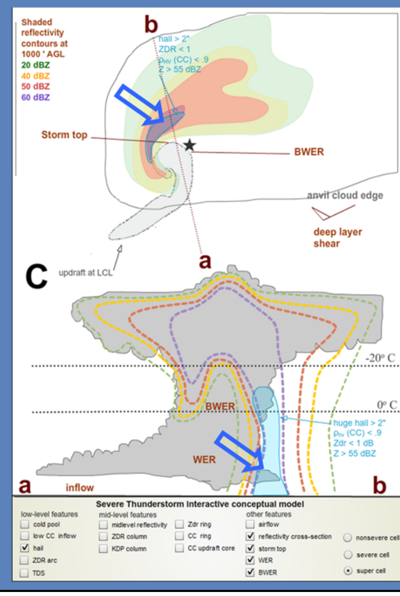
Low-level polarimetric signatures: significantly large hail

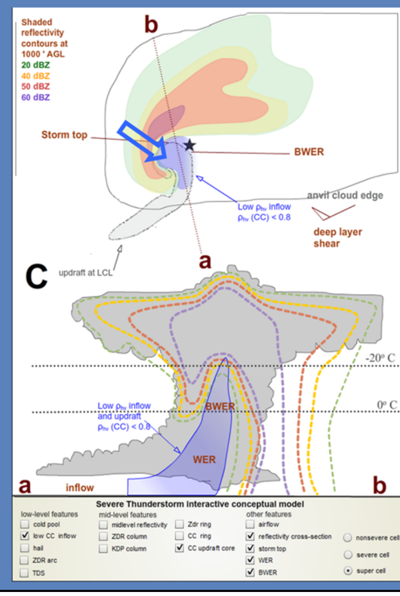
- Significant large hail (diameter ≥ 2”)
  - ZDR < 1 dB
  - CC < 0.9
  - Z > 55 dBZ
    - (preferably > 60 dBZ)



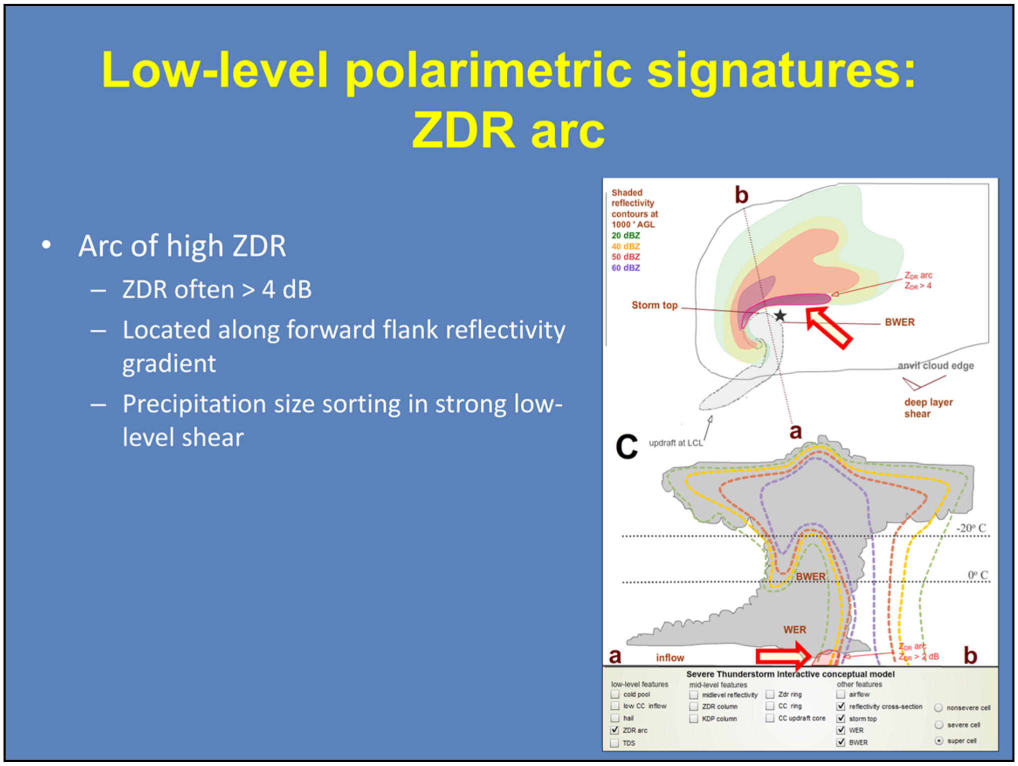
Supercells produce the vast majority of very large hail (diameter > 2”). Thus, the dual-pol signatures associated with this size hail commonly appear within the low-level cores of thunderstorms. These include CC roughly less than 0.9, ZDR less than 1 dB and preferentially near zero, and reflectivity greater than 60 dBZ (Figure 7-98). There is a severe hail identification lesson that explores the dual-pol hail signatures in greater detail.

Low-level polarimetric signatures: Low CC inflow

- Inflow accelerates into supercell updraft base
  - In addition to insects
  - May loft light vegetative debris



A little more about the low CC low-level inflow needs to be said. The lower extension of the low CC updraft core starts in the inflow layer. Usually this low CC inflow is an extension of a low CC precipitation-free boundary layer that is full of insects. However supercells have a tendency to accelerate air into the base of the updraft. If the inflow gets strong enough, then light vegetative debris also gets lofted and the CC may actually decrease in close proximity to the supercell. That's why the low CC inflow area in Figure 7-98 is shaded darker as the air flows into the base of the updraft.



One of the most intriguing signatures is the ZDR arc. This is a region of high ZDR precipitation echoes that lie along the sharp low-level reflectivity gradient facing the storm-relative inflow. Some of these hydrometeors are from the sloping echo overhang and others are from the edge of the precipitation cascade region. Recent research has theorized that the ZDR arc originates as the precipitation falling from aloft, is sorted by the vertical wind shear present in the environment, and enhanced along the forward flank outflow.

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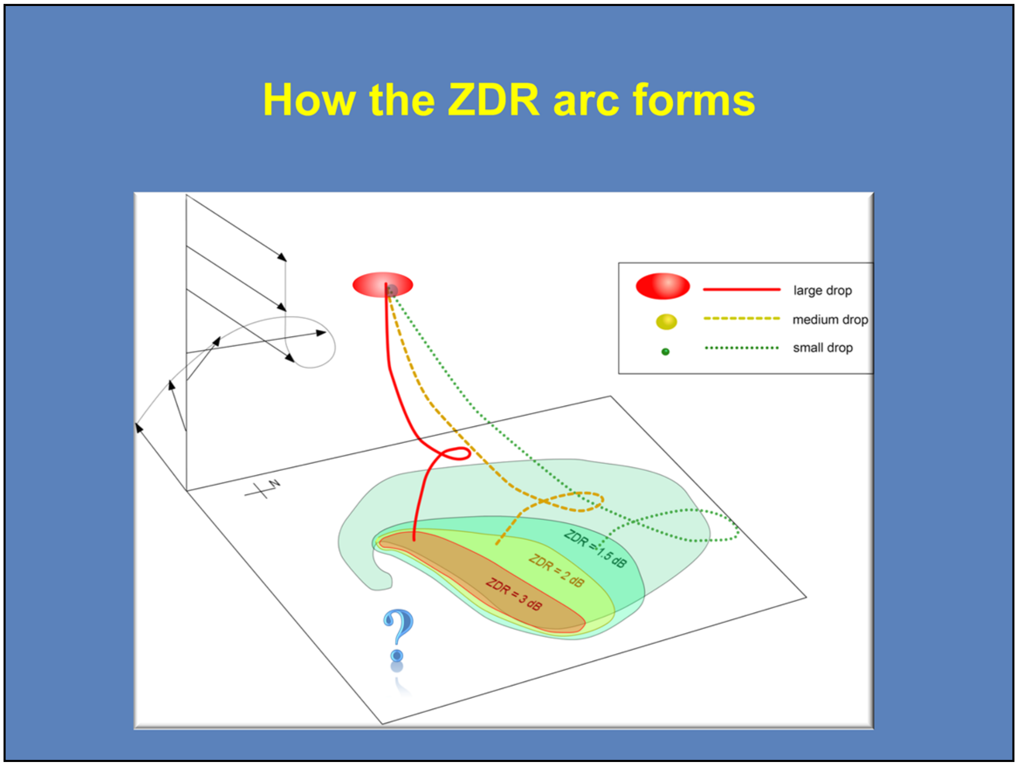
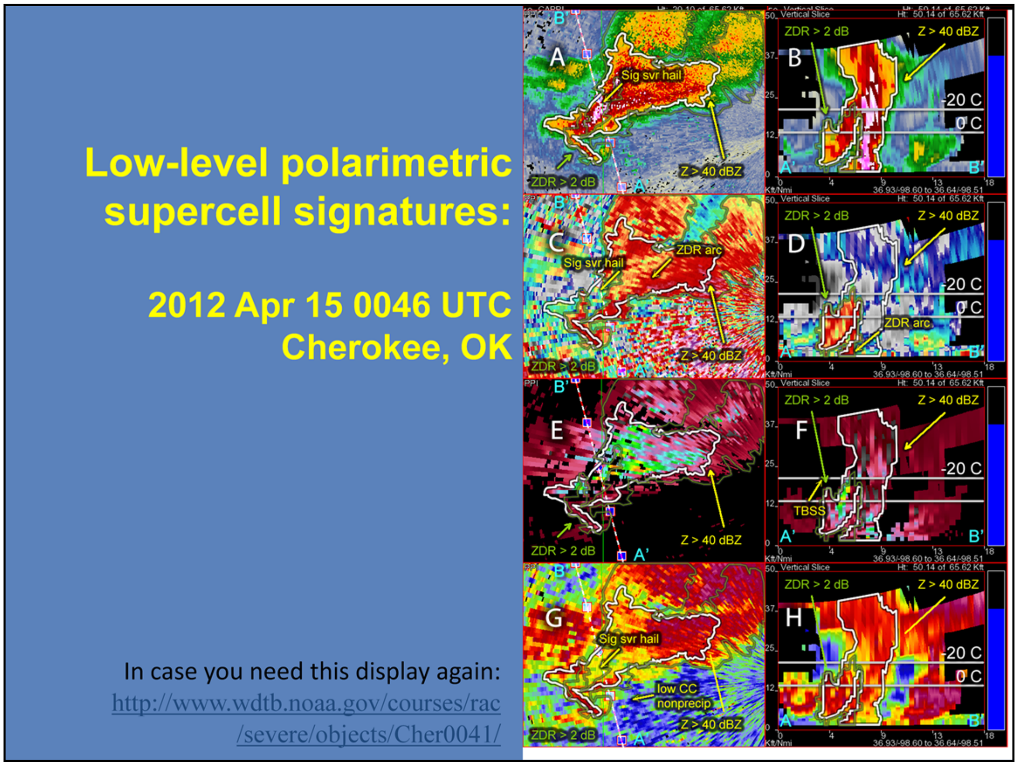


Imagine the wind profile with height changing in magnitude and direction as the vectors in Figure 7- 99 show. Then release a large, medium, and small size drops from the same position above the edge of the forward flank precipitation curtain. The larger droplets would respond less rapidly to the changing winds as they descend and therefore would fall closest to the precipitation-free low-level inflow. The smallest drops would respond most quickly to the changing winds and be carried away into the main core unless they evaporate first.

Because the size sorting continues to the ground, this feature is shallow, often below 6 kft above the ground. The fact is that strong vertical wind shear is required to produce the ZDR arc; therefore, this feature appears most commonly in supercells. In fact, Kumjian and Ryzhkov (2009) have suggested that the magnitude of the ZDR arc increases as the low-level storm-relative helicity increases.

It is enticing to think that monitoring the strength of the ZDR arc would give forecasters an assessment on the strength of the storm-relative helicity feeding the storm updraft, and that it could be used as a tornado precursor signature. However, there is no solid evidence yet and more research is needed on the ability of this signature to help in anticipating tornadogenesis.

In the meantime, the best way to detect this feature is to choose a radar close enough so that the lowest 6 kft can be sampled. Then choose either the lowest scan reflectivity image or drop the CAPPI into the lowest, clutter-free elevation possible.



Let’s look at an example of the low-level dual-pol signatures in supercells by going back to the Cherokee, OK storm shown in the previous example.

An interactive flash display of FSI output will appear in a separate browser window. Use the right/left step buttons to go up in elevation in the CAPPI. As you continue to step to the right, you will go from the 0041 UTC scan to the 0046 UTC scan whereby continuing to step to the right will increase the CAPPI elevation.

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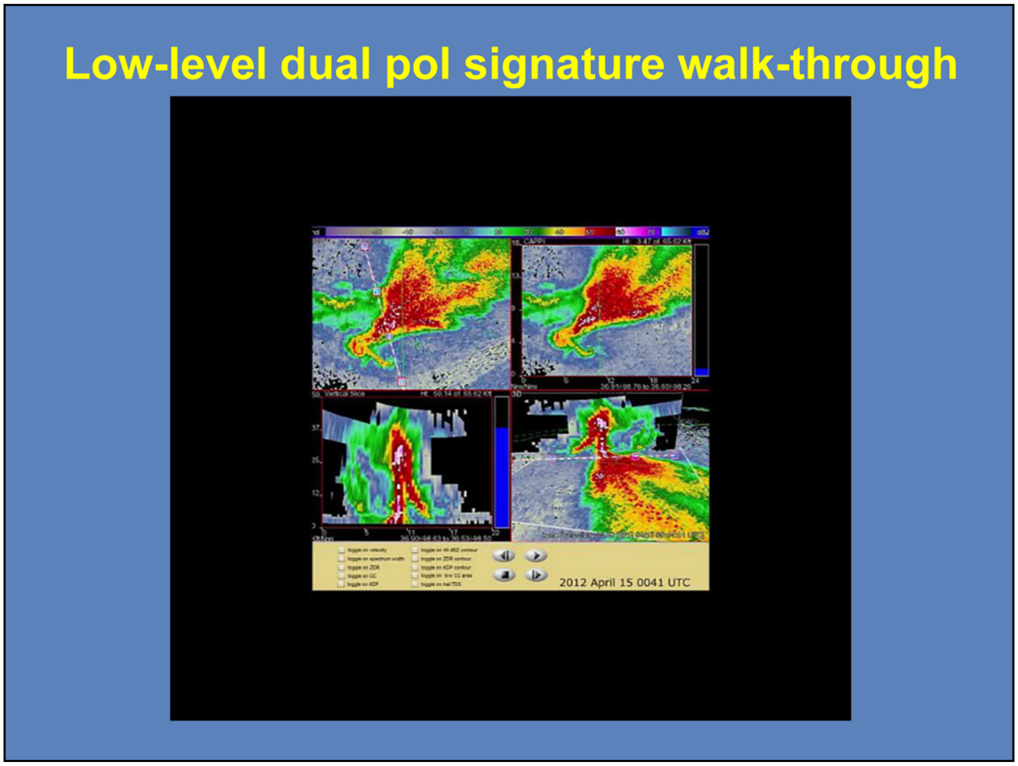

Use the interactive flash display to follow along this video tour.

Okay in this video you may want to have your own display available and have adjacent to this.

Stop at any time

We have reflectivity in 4 panels. Again, it's the same display with PPI in upper left, CAPPI in upper right, then the cross-section lined up in the PPI.

We have a classic supercell with all the accoutrements of what you would expect in reflectivity including concave reflectivity hook e cho. Very high reflectivities exist just northwest of the concave gradient where values exceed 60 dBZ. Then in the cross-section you can see above the concave gradient we have a BWER.

So the highest reflectivities exceeding 60 dBZ indicate large hail. Toggle to Zdr and we see low values in these areas.

Going to CC and you can see values fall below 0.9.

Go to Kdp where cc < 0.9 and they drop out.

But I think you see enough of low Zdrs, right here, low CCs, and high reflectivities indicate very large hail close to the ground.

Let's look at ahead of the storm in CC and you see low values. This is indicative of insects, chaff, dust in the strong inflow.

Take a look at velocity and indeed this is the area of Low CCs overlapping with strong inflow, which supports the idea of more dust and insects in the inflow.

Now going to Zdr, toggling on the 40 dBZ contour in white, and we'll take a look at this area of high values exceeding 4-5 dB here.

So this area represents the Zdr arc. If we look at the cross-section and we see those values and we'd expect to see a Zdr arc. This is the area we'd ex pect to see the Zdr arc due to precipitation size sorting.

The velocity here in the Zdr arc is where you can see the sorting as the storm accelerates the flow and the storm-relative flow helps in precipitation sorting.

It's not just the low-level storm inflow but the environmental storm-relative flow too . You'd expect the storm-relative flow start from the southeast and then going up to the north at mid altitudes and then to the northeast at high altitudes.

All that indicates that small droplets are pushed to the north and large droplets fall quickly along the southern edge of the reflectivity forward flank.

You do note the high values of Zdr in the arc wraps around in the interior side of the hook.

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Summary
dual-pol signatures in supercells

- Mid-level signatures
  - ZDR ring
  - CC ring
  - Low CC updraft
- Low-level signatures
  - Significant severe hail
  - Low CC inflow
  - ZDR arc

We break down the dual-polarization signatures associated with supercells into two regions; the mid-levels between the freezing and -20 deg C level, and the low-levels focusing in the lowest 6 kft above ground.

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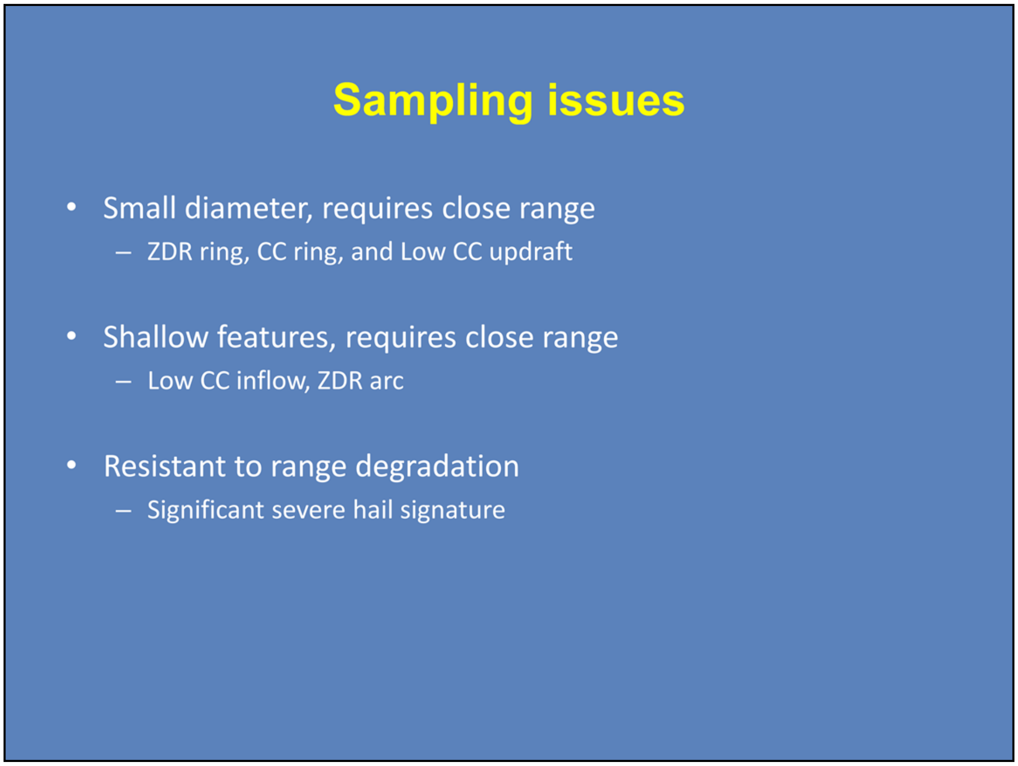

Sampling some of these signatures is sometimes problematic given their small size or shallow nature. Here is a list of sampling issues with each signature:

• ZDR and CC rings are small, similar to the BWER in size. Radar needs to be close.

• The low CC updraft is small and the radar needs to be close.

• Significant severe hail signatures are large and relatively resistant to range degradation.

• The low CC inflow and ZDR arc are shallow and the radar needs to be close.

If you have passed the quiz, then you have successfully completed this lesson. If you have any questions, please contact us using any of the e-mail addresses listed on the bottom of the slide.

Welcome to the Radar & Applications Course, Convective Storm Structure and Evolution lesson on Supercell Archetypes.

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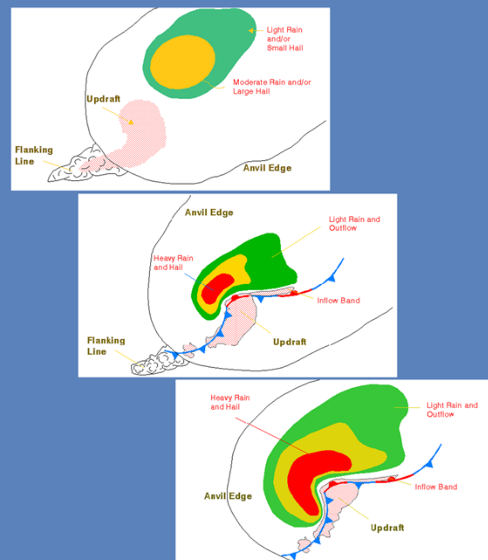
Motivation

- The broad diversity of supercell structures can make it a challenge to identify them in operations.
- This lesson covers:
  - Low Precipitation (LP)
  - Classic (CL)
  - High Precipitation (HP)
  - Left-moving (anticyclonic)
  - Mini



Even though all supercells contain mesocyclones whose source of vertical vorticity is derived from vertical wind shear, the broad diversity of supercell structures can make it a challenge to identify them in operations. Thus, this lesson will describe some of the ways that supercells may appear through radar and other data.

This lesson will cover the characteristics of supercells with different precipitation distributions around their updraft which are: Low Precipitation (LP), Classic (CL), and High Precipitation (HP). We will also cover left-moving (anticyclonic) supercells and mini superecells, and how their characteristics influence warning decisions.

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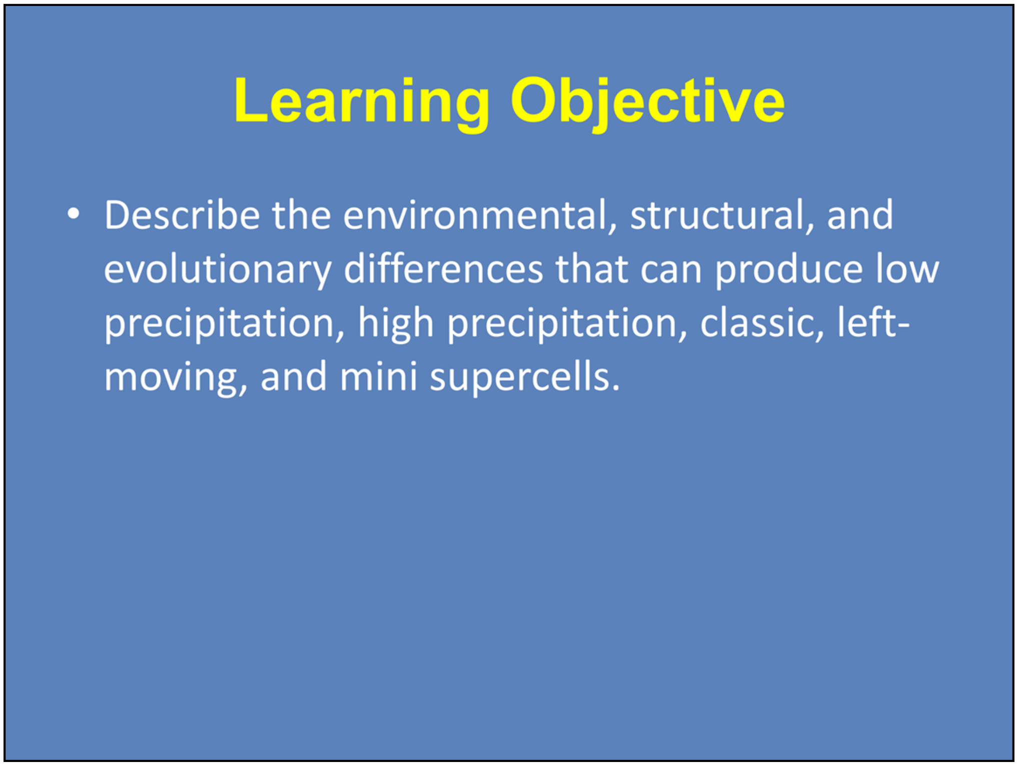
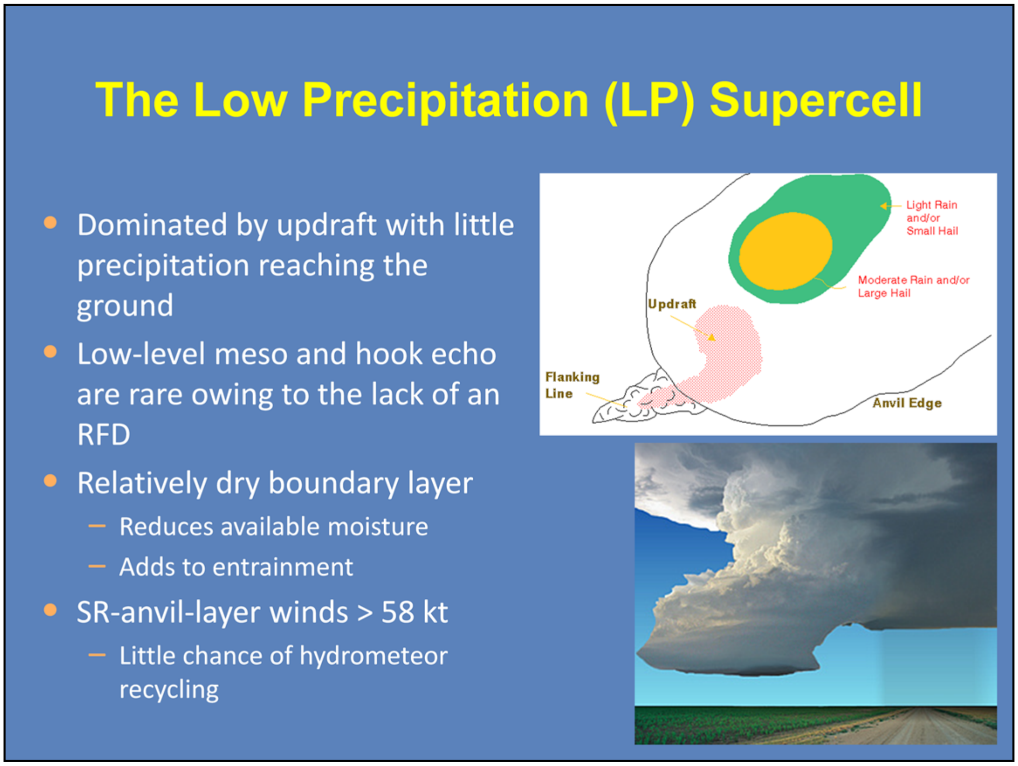


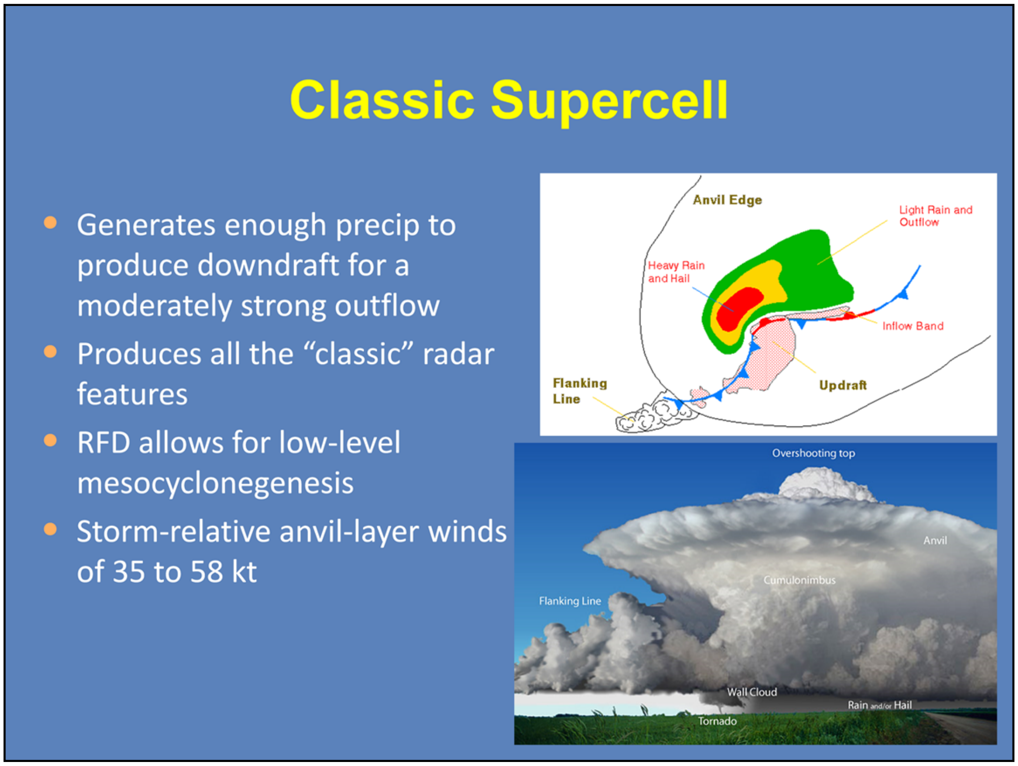
This lesson has one objective: **Describe the environmental, structural and evolutionary differences that can produce low precipitation, high precipitation, classic, left moving and mini supercells.**



Supercells are grouped into three different structural classes depending on the amount of precipitation contained within the core, and where the mesocyclone is located with respect to the main core.

Low Precipitation (LP) supercells are generally dominated by updraft with little precipitation reaching the ground. These storms are visualized by exposed "skeletal" updrafts and translucent to nearly transparent precipitation cores. The relative lack of precipitation leads to weak downdraft formation and thus these storms could be said to be outflow deficient. LP supercell updrafts often show significantly strong midlevel mesocyclones. However, low-level mesocyclones are rare owing to the lack of a well defined Rear-Flank Downdraft (RFD). There is rarely a hook echo, and most of the precipitation is carried well downstream of the updraft by the storm-relative upper-level winds. Maximum reflectivities in LP supercells are relatively weak with the maximum values likely produced by a few, large, dry hailstones. True LP supercells represent the dry extreme of the supercell spectrum and are quite rare.

LP supercells require significant instability and shear, but other conditions help to reduce precipitation efficiency. Relatively dry boundary air reduces available moisture and adds to entrainment, but LP storms can also exist where boundary layer moisture is high. **Additionally, very high storm-relative anvil-layer winds (>30 m/s or >58 kts) transport rising hydrometeors well away from the updraft before they descend out of the anvil layer (Rasmussen and Straka, 1998).** Hydrometeors may have little chance of recycling back into the updraft, especially if the midlevels are dry.



Classic (CL) supercells generate enough precipitation to be able to produce enough downdraft for a moderately strong outflow. These storms are associated with all the classic radar features of a supercell including an inflow notch, WER, BWER, hook, and mesocyclone. The RFD is stronger with a classic supercell than with an LP supercell and therefore, low-level mesocyclogenesis is more likely. The result is a greater threat of severe weather from winds and tornadoes.

Classic supercells occur in moister environments than are typical for LP supercells. Storm-relative, anvil-layer winds are likely to be lower for classic supercells (mainly between 18-30 m/s or 35-58 kts).

These supercells produce the majority of long-lived tornadoes. They are also the common cyclic tornado producer.

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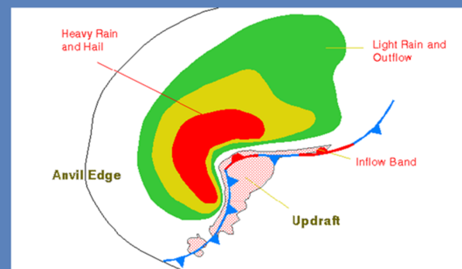
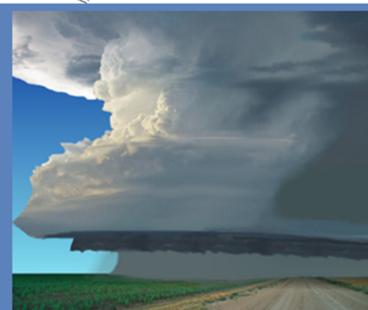
High Precipitation (HP) Supercell

- Most common supercell type
- Efficient precip producer
- Often produce strong downdrafts and outflows
- Large, high reflectivity hook
- Moist boundary layer
- Weak storm-relative, anvil-layer wind (< 35 kt) allows precip to reseed the updraft
- HP transition can be triggered by aggressive seeding





High Precipitation (HP) supercells are the most common of all supercells. They are highly efficient precipitation producers and often produce strong downdrafts and outflows. Large amounts of precipitation are available to wrap around the mesocyclone, producing a large, high reflectivity hook. Occasionally, the RFD gust front associated with the hook is intense enough to generate strong convection along its leading edge. The result is that the strongest core can be behind and to the right of the mesocyclone path. Occasionally, this process leads to HP supercells evolving into bow echoes.

HP environments typically show more boundary layer moisture than that of LP or even CL. However, high boundary layer moisture is not necessary for an HP. **Another possibility includes low anvil-level, storm-relative flow (<18 m/s or <35 kt) to allow precipitation to reseed the updraft improving precipitation efficiency.** A supercell can transition to HP if it is being seeded by aggressive cells on its flanking line or adjacent storms.

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High Precipitation (HP) Supercell Varieties

- Wide variety of possible HP configurations
  - But all have a meso well correlated to updraft and longevity
- Hard for spotters to observe mesocyclone
- HPs carry all threats of severe weather

30 Apr 2000 – Olney, TX - J. LaDue

There is a wide variety of possible HP supercell configurations, however, they all share traits common to all supercells – a mesocyclone well correlated with an updraft and longevity. The mesocyclone is usually well sampled by radar owing to the high reflectivities in the hook. However, spotters in the field often have a difficult time observing the mesocyclone area most favorable for tornadogenesis.

HP supercells carry all threats of severe weather including: Large hail, damaging winds, tornadoes, and flash flooding.

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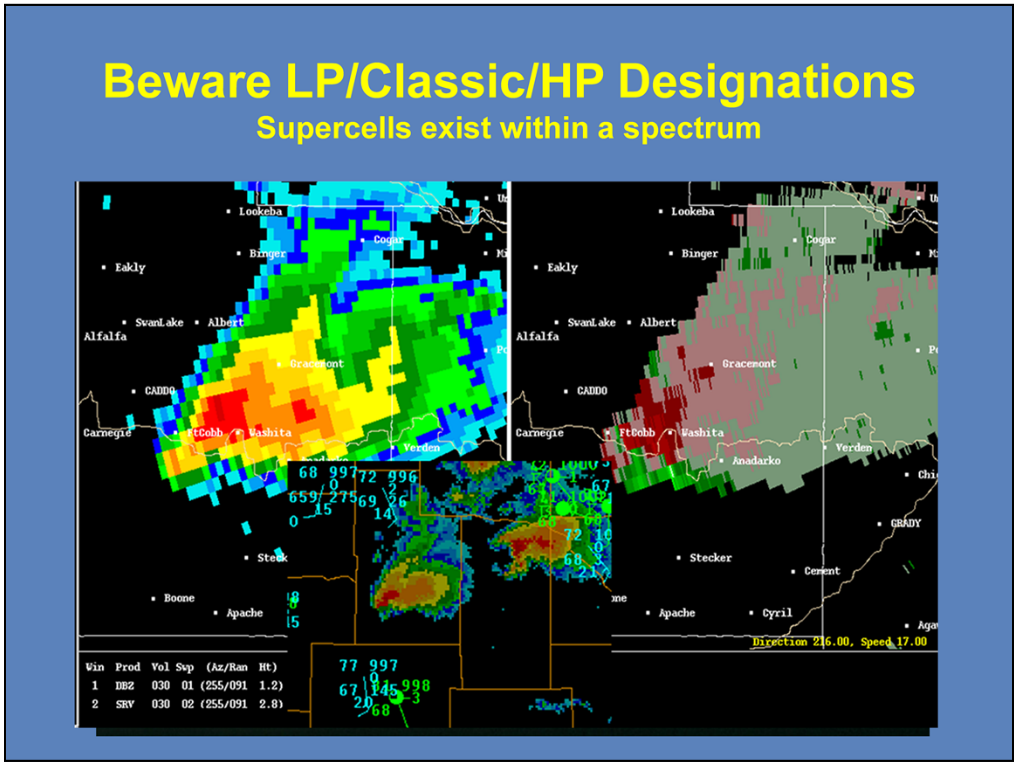


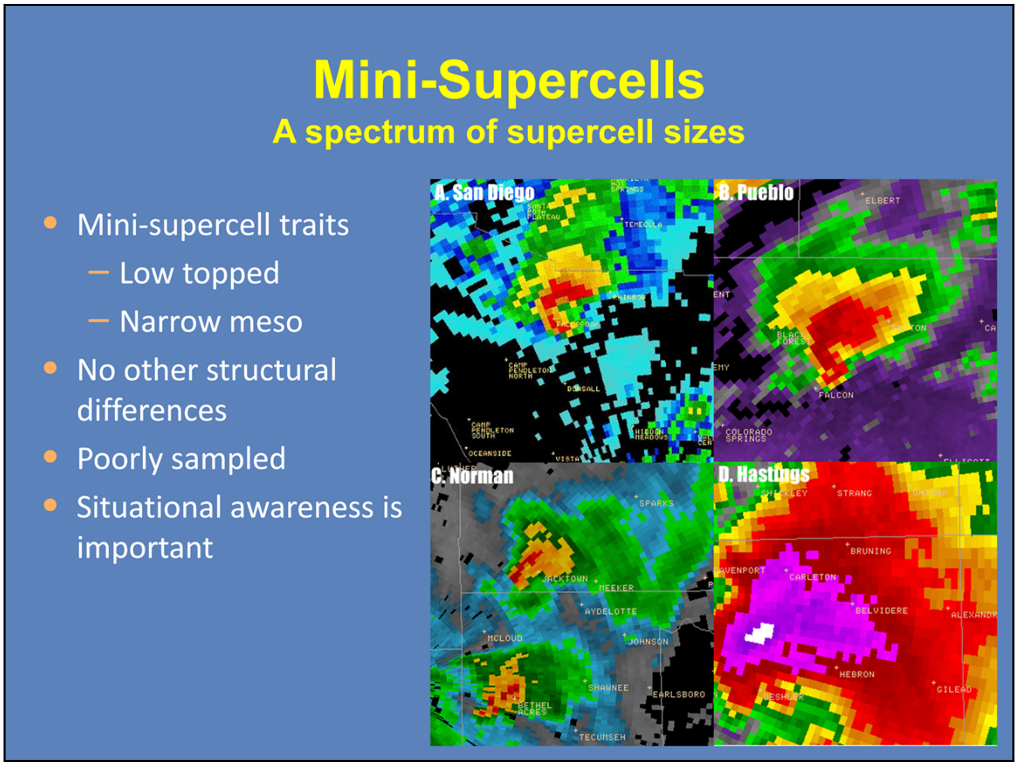
Having said all this, be cautious about spending lots of time trying to classify supercell type. There are no formal definitions, and many search papers refer to these supercell archetypes using different criteria. Supercells exist within a spectrum with no well-defined boundaries.

For example, the first designation of a Low Precipitation supercell was put forth in the 1970's (Davies Jones et al., 1976) as a name for storm they documented by radar with unusually low reflectivity. It was still a supercell and likely contained an active rotating updraft. Since their paper, the LP supercell has been photographically documented out in the field by others (Bluestein and Parks, 1983). Many storm spotters and storm chasers now label LP storms based on visual properties of a nearly transparent precipitation core and a fully exposed updraft tower. An example of what some spotters have designated as an LP storm is shown here.

The nearly transparent precipitation core in visual light can be deceiving. Much of the precipitation may be composed of large hail. In addition, there are precipitation shafts behind the lowered wall cloud under the right side of the updraft that may fall unnoticed by spotters.

The WSR-88D from 60 nm away showed that indeed the nearly transparent precipitation echoes were highly reflective. However, the radar was too far away to detect the hook echo. This supercell produced twenty tornadoes during its lifecycle, including ones shortly before and after these image were taken.

To summarize, when considering the potential hazards of a supercell, be careful not to base it too heavily on supercell classification. The storm in produced a tornado shortly after the image was taken.



As supercells can vary in the amount of precipitation falling around their mesocyclone, they can also vary in size. There can be low-topped supercells with wide mesocyclones, or high-topped supercells with narrow mesocyclones. **A mini-supercell is one that is both low topped and contains a narrow mesocyclone.**

There are no structural differences between mini- and normal-sized supercells. However, there are differences in the expected severe weather. Giant hail (>2.5" in diameter) is rare because of limited extent of the updraft into the hail growth zone, and smaller horizontal dimensions of the updraft. **Poor radar sampling of small mesocyclones means that it is more difficult to measure high rotational velocities.** To illustrate this point, the supercells in panels A-C of this image were tornadic even though their associated mesocyclones were very small and/or apparently weak. Therefore, it is important to identify mini-supercells and be aware that their apparently weak circulations (V_r < 30 kts) can still carry a significant tornado risk as document by Prentice and Grant in 1996.

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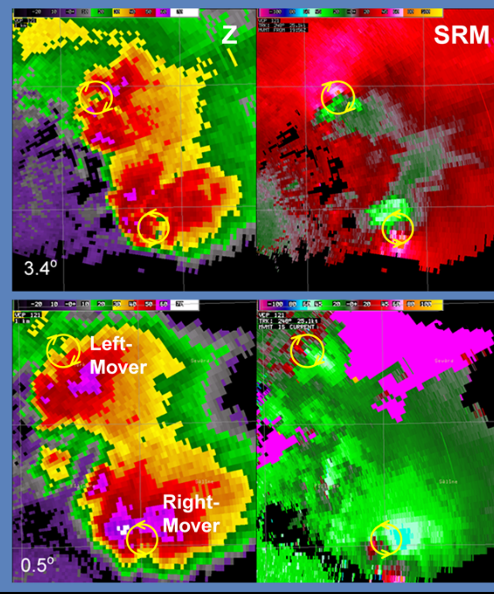
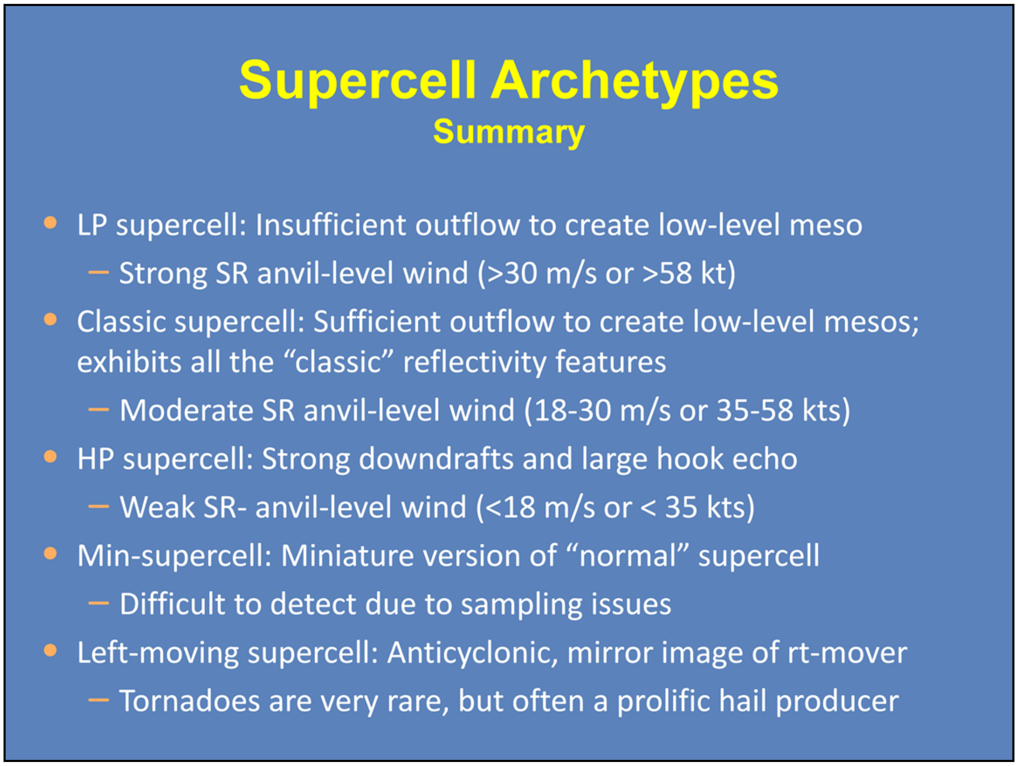
Left-Moving Supercells

- Rotate anti-cyclonically
- By-product of the original storm split
- Mirror image of the cyclonic right-mover
- Very few produce tornadoes
- Often prolific hail producers
- Updraft on the left-side of the core relative to storm motion



Left-moving supercells rotate anticyclonically (in the northern hemisphere) and are a by-product of the original storm split. They are structurally a mirror image of the cyclonic right-mover. Very few left-moving supercells produce tornadoes, and for reasons that are poorly understood, often produce long swaths of giant hail. As long as the hodograph is relatively straight, the left-mover can be as strong as the right-mover.

Left-moving supercells have their updraft of the left side of the reflectivity core relative to storm motion. Take a look at the left-mover in the 0.5 degree Base Reflectivity product. Notice the enhanced reflectivity gradient and concavity is located on the north side. The 0.5 degree SRM product shows anticyclonic shear associated with this updraft. Higher up, at 3.4 degrees, both the left- and right-movers have BWERs and rotation.



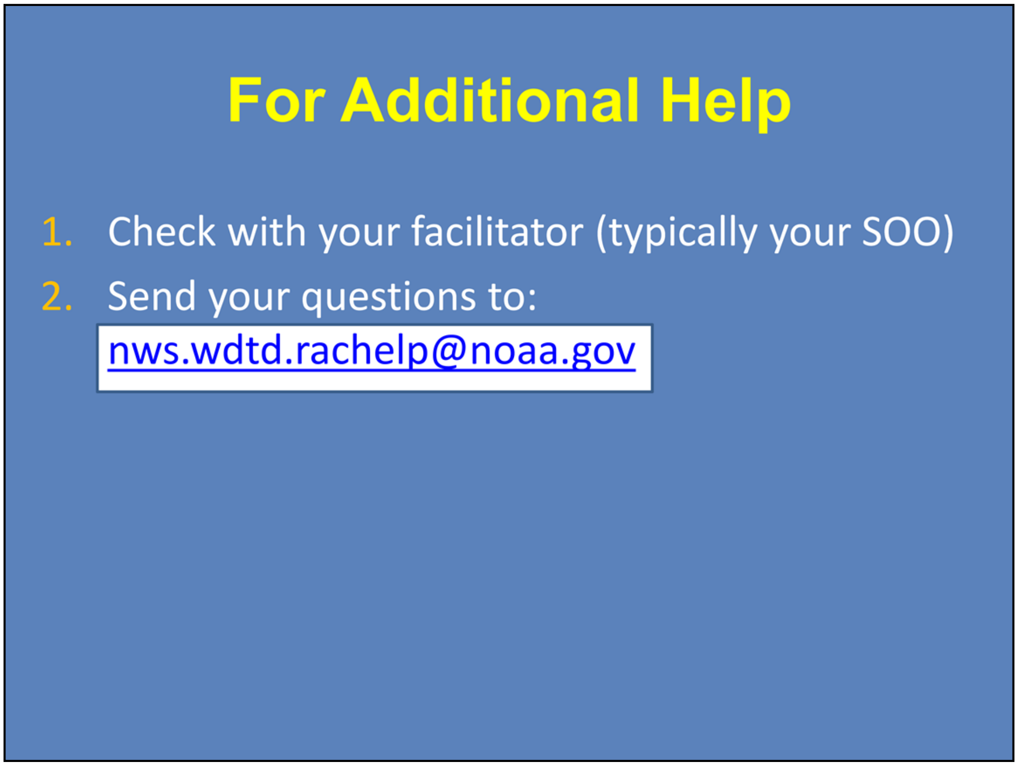
LP supercells exist with no real definition, and yet there is a consensus that LP’s are unable to form a hook echo and also produce insufficient outflow to create low-level mesocyclones. They typically exist in dry boundary layers and/or strong anvil-level storm relative flow, however just as often, it is the way LPs initiate that provide a clue as to their existence.

Classic supercells exhibit a small hook relative to forward flank core which is accompanied by sufficient outflow to create low-level mesocyclone. They appear to form most often with moderate anvil-level SR flow (18–30 m/s or 35-58 kts) but not always.

HP supercells exhibit a large hook, sometimes with most of core following the mesocyclone. Intense RFD outflows often accompany HPs. They typically form with weaker anvil-level SR flow (<18 m/s or <35 kts), or perhaps through multistorm seeding.

Mini-supercells are structured very similarly to their larger counterparts, however, you suffer the disadvantage of not being able to detect their features as readily unless the storm is close to your radar.

Left-moving, or anticyclonically-rotating supercells are structurally a mirror to their right-moving, or cyclonically-rotating counterparts. They rarely produce tornadoes, however, they are often prolific producers of hail. While they are rapid movers in most occasions in the Northern Hemisphere, some environments allow for the left-movers to be the slow movers. Their mesoanticyclones are currently undetectable by the MDA.



For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.

Welcome to the lesson entitled “Analyzing Tornadic Scale Signatures”. This lesson should last 30 minutes.

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Learning Objectives

1. Describe the necessary conditions for defining a Tornadic Vortex Signature (TVS) and a Tornado Signature (TS)
2. Understand the relationship between the TS and the TVS to the actual storm-scale circulation
3. Describe how to detect a dual-pol-based Tornado Detection Signature (TDS).

A WSR-88D with dual-polarization offers forecasters a broadened capability to infer ongoing or imminent tornadoes. Velocity data can detect a signature of a vortex that may be associated with a tornado. The signature may either be a Tornadic Vortex Signature (TVS) or a Tornado Signature (TS). Dual-pol data can detect lofted debris from columnar vortices connected to the ground. This detection is called a Tornado Debris Signature [TDS; Ryzhkov et. al (2005)].

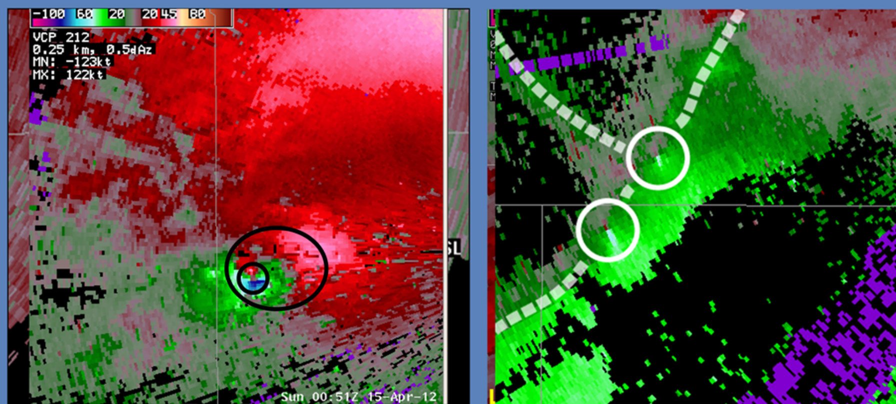
This lesson is in two parts. The first describes how to identify a TVS *vs.* TS and how to assess the potential to identify an actual tornado. We also discuss the types of true circulations that may manifest themselves as a TS and TVS. The second addresses the TDS and how to identify it.

• Describe the necessary conditions for Objectives defining a Tornadic Vortex Signature (TVS) and a Tornado Signature (TS).

• Understand the relationship between the TS and TVS to the actual storm-scale circulation.

• Describe how to detect a dual-pol-based Tornado Detection Signature (TDS).

Tornadogenesis and the TS/TVS



- Shear
- Vertical extent
- persistence

The tornadogenesis process traditionally manifests itself on radar as an increase in rotational velocity in the mid- and/or low-levels. The tightening of a region of circulation is common, which often leads to the development of a Tornado Signature (TS) and Tornadic Vortex Signature (TVS) by radar. On other occasions they appear to spring up out of nowhere. In reality, the range degradation of radar data prevents you from detecting the increased shear that exists just before the tightening process. Regardless of how it appears, the onset of a TS/TVS should be associated with the phasing of a strong updraft and increased low-level circulation. For the purposes of this lesson, we refer to the TS/TVS as one defined by the operator, not an algorithm.

The type of circulations that satisfy this category are possibly associated with tornadic rotation that **meets or exceeds established criteria for shear, vertical extent and persistence. A TVS/TS can be described as a tornadic velocity profile superimposed on a larger mesocyclone. However, a larger parent circulation is not required and sometimes the TVS/TS is the mesocyclone. Let's elaborate more specifically on the TVS/TS.**

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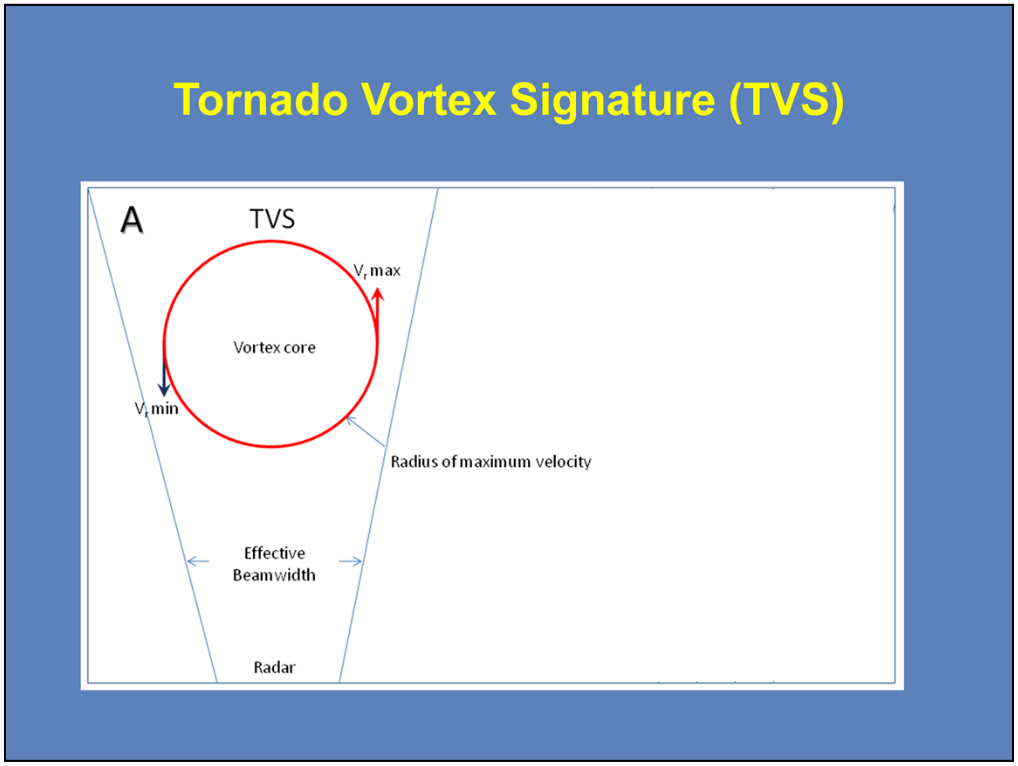

A TVS occurs when the core diameter of the tornado-scale circulation is smaller than the effective beam width of the radar. A TVS shows up as a signature where the radar detected maximum (Vr max) and minimum radial velocity (Vr min) are located on adjacent azimuths. That is unless the entire vortex core lies within a single beam. In such a case, the **Vr max, Vr min** would nearly cancel each other out leaving a nearly zero radial velocity and a very broad spectrum width.

Also, when the azimuthal sampling interval is significantly less than the effective beam width, as is the case with super-resolution data, then there should be a transition zone between **Vr max, Vr min** as the beam sampling becomes less independent of one another. However, in reality, that’s not frequent.

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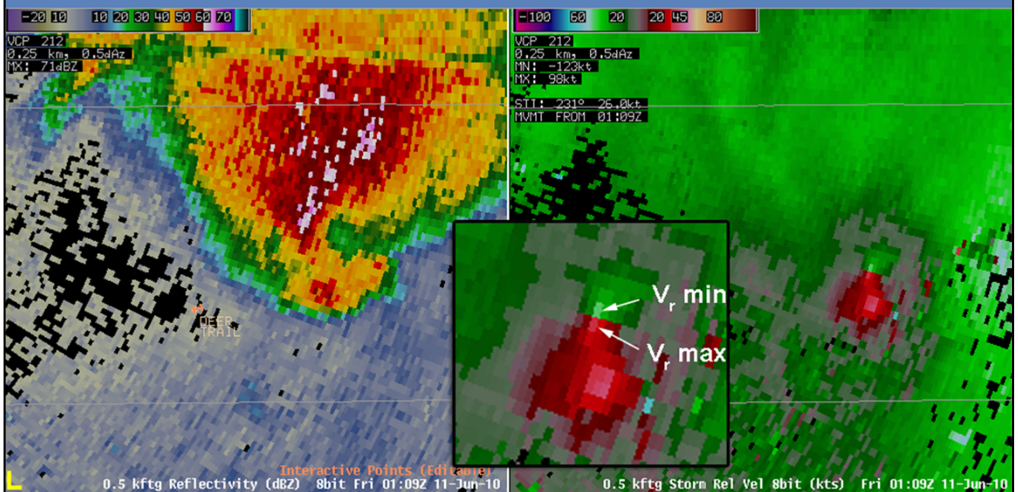
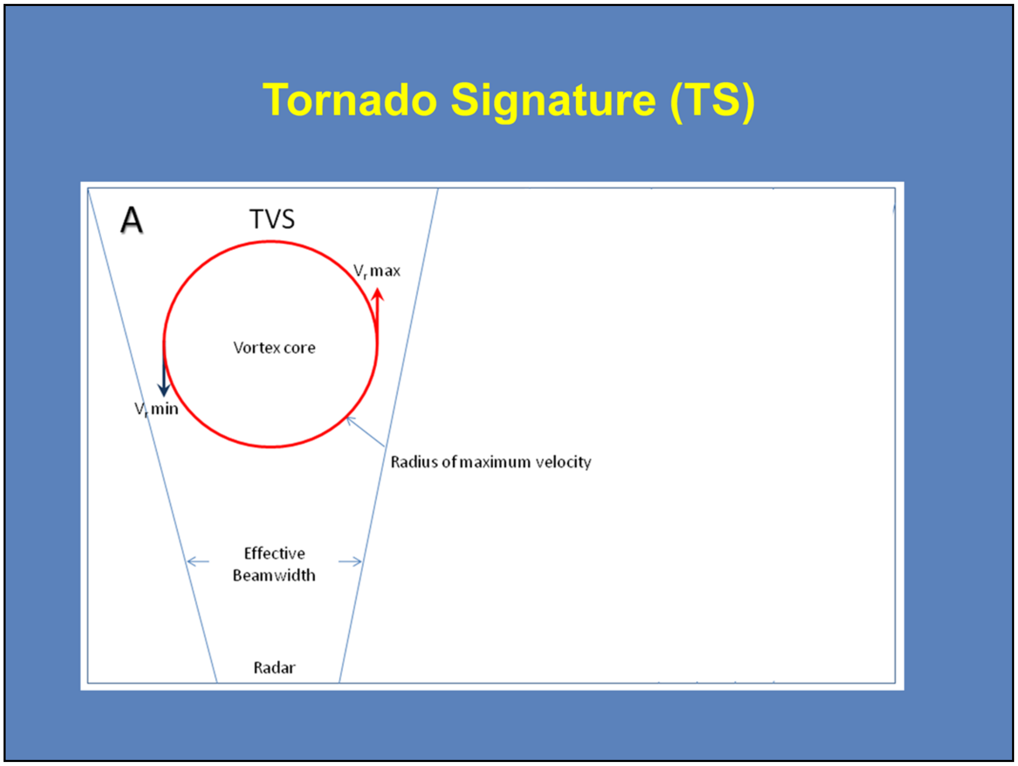
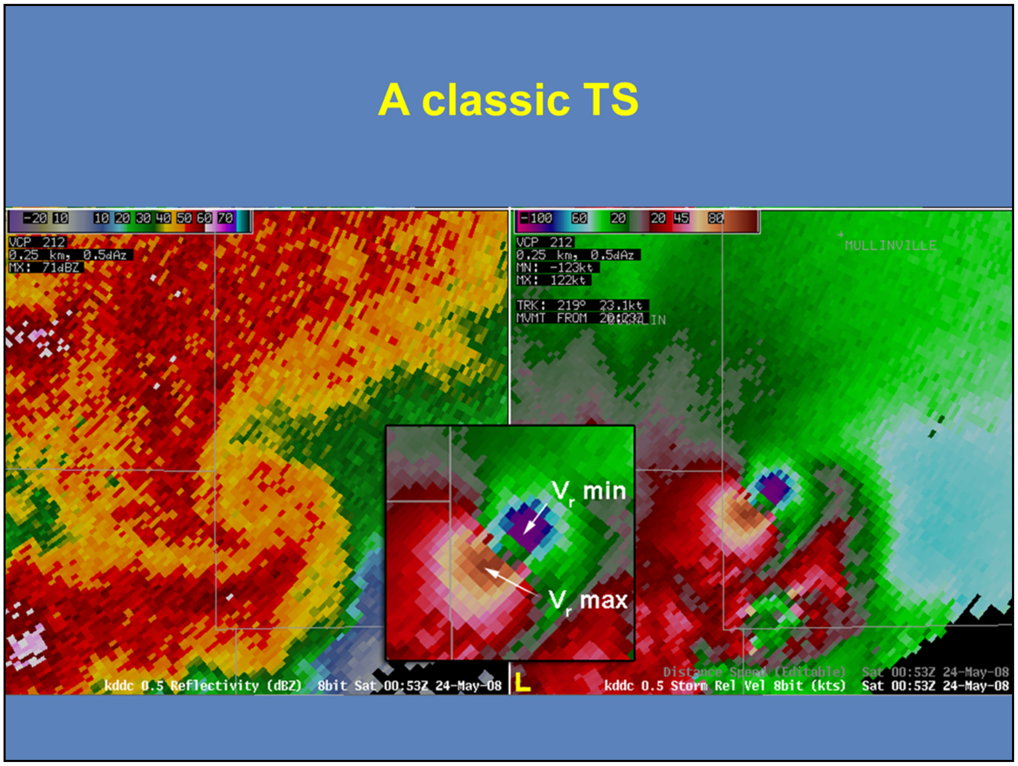
A classic TVS



This image shows an example of a TVS sampled with the 0.5o elevation angle from the Deer Trail, CO tornadic supercell on 11 June 2010 at 0109 UTC. The maximum radial velocity was 35 kts (18 m/s). The TVS did not have a radar detectable Rankine combined vortex structure, because the effective beam width was too large to sample the inner core. Instead, the KFTG WSR-88D only detected the potential flow increasing in speed as the distance to the vortex center decreased until Vr max, and Vr min were found on adjacent azimuths. This is a classic TVS. However, there was also a separate V min without a corresponding V max that was probably associated with a rear flank downdraft.



When the core diameter of a tornado-scale vortex is larger than the effective beamwidth, we call the vortex a TS. With more than one beam sampling the vortex core, Vr max and Vr min almost always appear separated by at least one radial.



Here is an example of a TS where Vr max and Vr min are separated by four effective beam widths or 1.15 nm (1.85 km). The maximum inbound velocity was 124 kts (64 m/s). Given these velocity figures and the reports from the ground, this TS represented a high-end large tornado.

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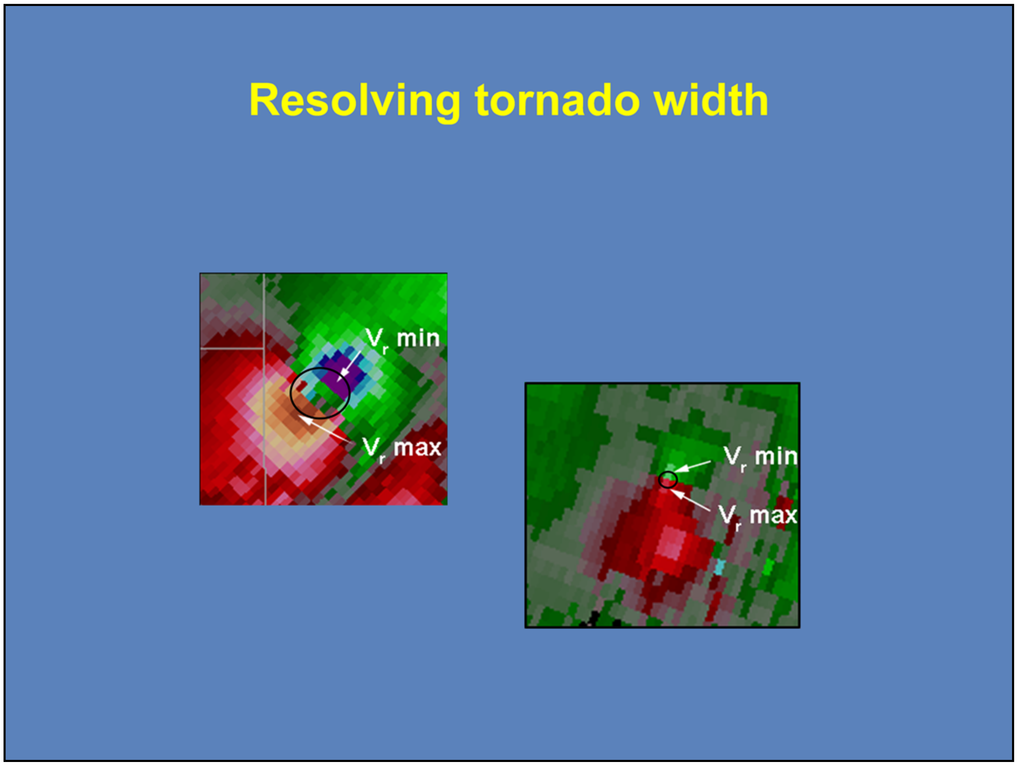
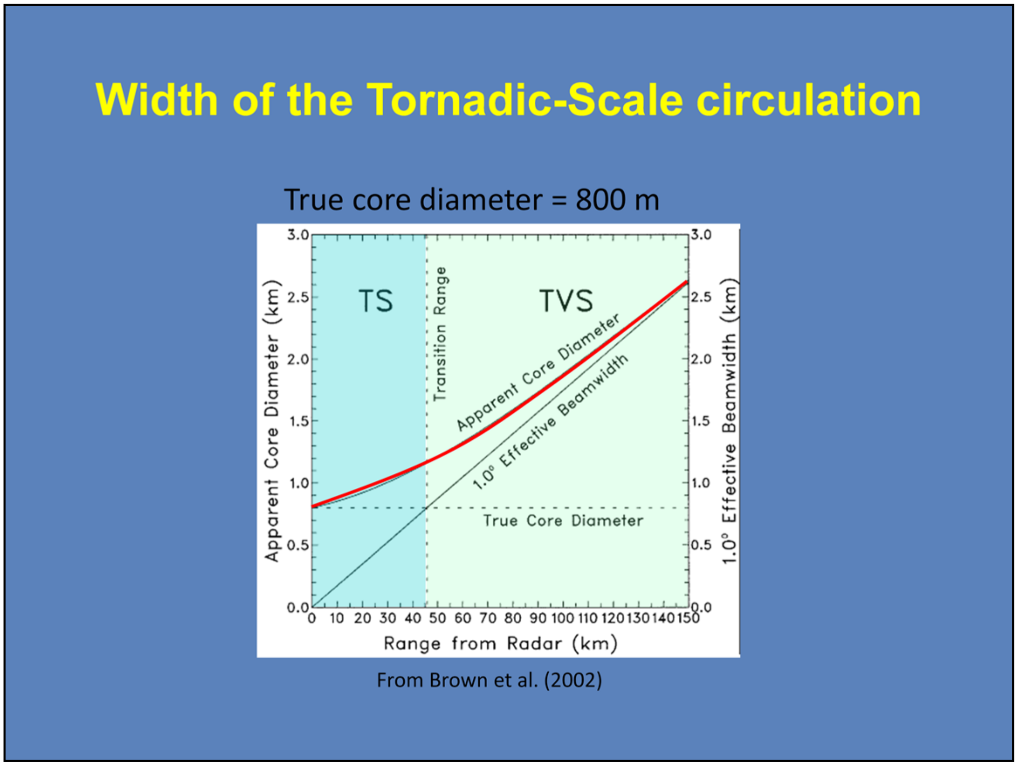


Since the radial velocity images show the same zoom magnitude, and each vortex was located 28 nm (51 km) in range, you may directly visualize the size differences between these two tornadoes. Even if they differ greatly in size and strength, both vortices exhibited an isolated velocity core leading up to a well defined Vr max, Vr min along with the maximum radial velocity gradient directed tangentially and counterclockwise. In other words, the detected velocity structure of both the TS and TVS was purely rotational.



Let me explain this graphic. Imagine a tornado whose true core diameter is 800 m and represented by the dotted line.

But note that as range increases from the radar represented by the X-axis, the 1.0 deg effective beamwidth increases in diameter.

Where the effective beamwidth is less than the true core diameter of the tornado, the radar should observe a TS. Beyond 45 km in range from radar, the effective beamwidth increases above the tornado diameter and this is where a TVS would be seen by radar. The transition range is that 'grey zone' between a TS and TVS. Again, this transition range is for this sized tornado only.

The existence of a TS does not guarantee that you are able to resolve the tornado width. The apparent core diameter of the TS immediately increases as range increases even if the true core diameter remains fixed. As this figure shows, there is no change in the how rapidly the diameter estimate increases with range even through the transition region from TS to TVS. Perhaps the apparent core diameter of the extremely large tornado would more closely match its true core diameter since its more than four radials wide, but even in this case, it is probably an overestimate.

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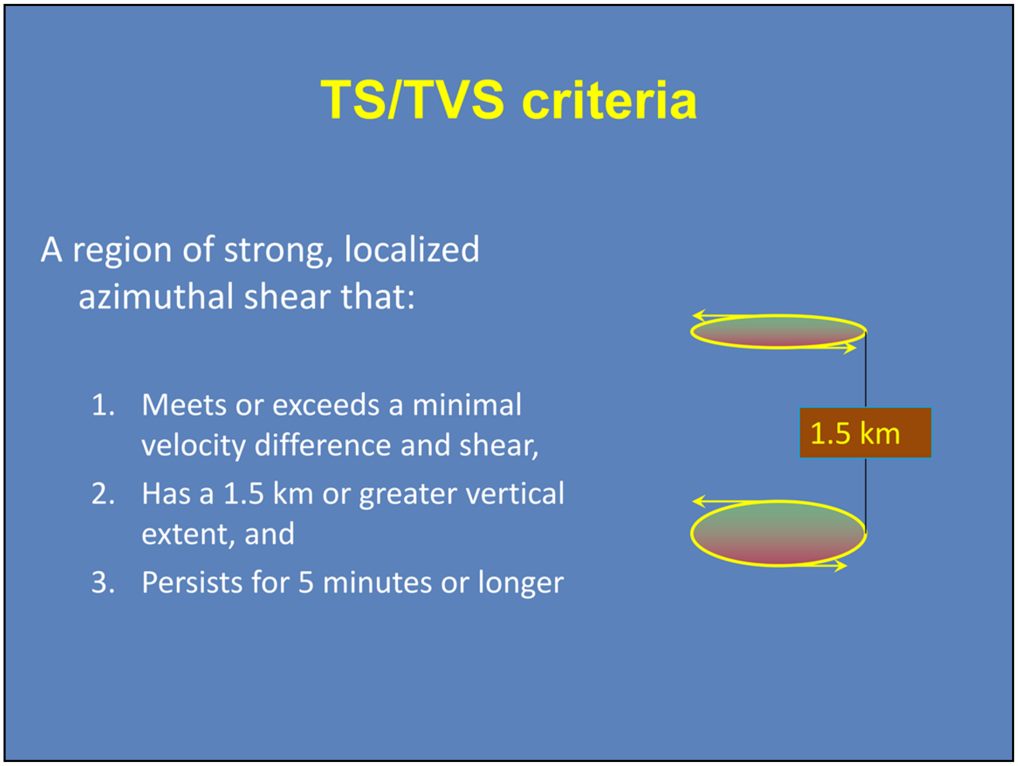


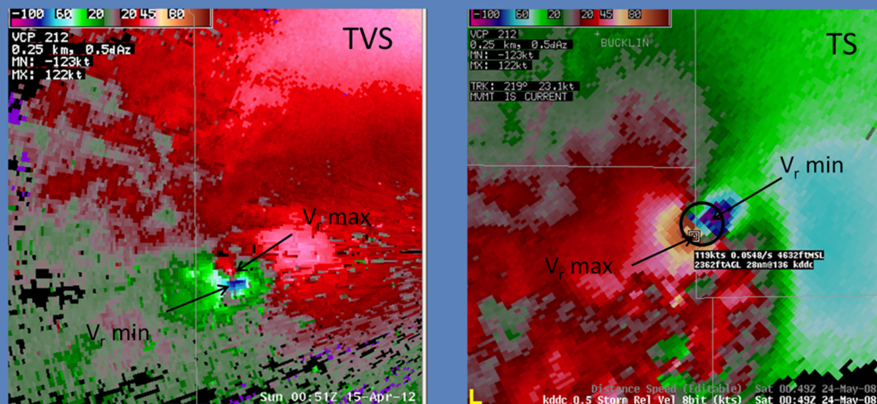
Some criteria are required to make sure that what circulation you are looking at is indeed a TVS. The three criteria are:

1. **A minimal shear:** There is no hard lower threshold in this criterion in the same way as an algorithm like the TDA would be assigned one. The minimal shear that an expert forecaster would define would depend on many things including the distance from the radar, the forecaster's assessment of the size of the vortex, near storm environment, and past experience. We will discuss what meaningful shears may be of significance.

2. **Vertical Extent:** At least some vertical continuity should be seen in a TS/TVS so that there is a high probability that an updraft is present in the circulation. For most events, the depth should be at least 1500 m (4900 ft.). Low topped supercells typically do not have deep TSs/TVSs, even if tornadic. Sometimes, and if detectable at all, only the lowest elevation angle contains a gate-to-gate rotational signature in tornadic low-topped supercells. Sometimes the vortex may appear as a TS and a TVS at adjacent elevation angles due to the vagaries of sampling and vortex structure. The vertical extent should include one or both manifestations as long as the true vortex appears to show vertical continuity.

3. **Persistence: In order to reduce the possibility of a circulation that randomly becomes vertically coordinated, you should ensure that the TS/TVS persists for at least five minutes**. However, mesocyclones can spin-up over a considerable depth in a very short time, and some legitimate TS/TVSs may become tornadic in less time. We suggest that if either signature forms in close proximity to a strong updraft signature, and a very supportive environment, persistence may not be a requirement to call it a TS/TVS and a tornado.

TS/TVS Velocity Difference



To measure TS/TVS strength objectively, we use the velocity difference (ΔV), where

$$\Delta V = V_{r\,max} - V_{r\,min}$$

The maximum azimuthal shear found in a TS/TVS is perhaps the most consistent method for evaluating its strength; however, a forecaster needs a method that can be done quickly using base data. The fastest method is to simply take the radial velocity difference from where the maximum radial velocity is located as long as it represents the vortex core perimeter.

Therefore, the velocity difference (ΔV or DV) is shown as equal to the difference between the max and min Vr, where Vr max, and Vr min **are defined in the images**. Notice that we do not use inbound vs. outbound velocities because Vr min may still be the same sign as Vrmax. Because the distance between Vr max, and Vr min increases as the distance to the radar increases, DV is not really equivalent to shear. However, to simplify the process, we still use DV and account for how decreasing resolution can affect the relationship between DV and shear (more on this later).

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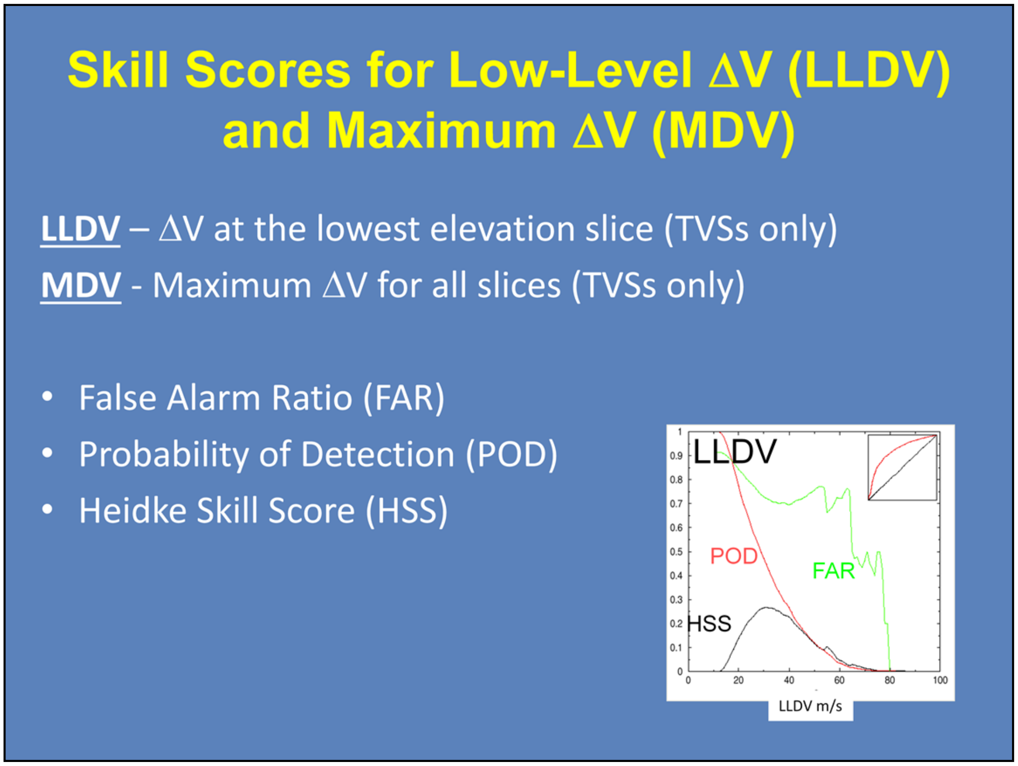

There are two DV calculations that are typically used:

- **The DV measured in the lowest slice is called the Low-Level Delta V, or LLDV**.
- **The maximum DV for all slices containing a TS/TVS is called Maximum Delta V (MDV)**.

To determine whether LLDV or MDV is large enough to satisfy part of the TS/TVS criteria depends on how useful it is to use these parameters in considering a tornado warning. Assuming a forecaster issues tornado warnings based solely on the presence of a TS/TVS, then the threshold LLDV and MDV are critically important to know. Unfortunately, there are many variables including storm type, environment, and distance to the radar, that impact and change these thresholds.

However, there is one way to provide some guidance to help comparing the likelihood that a certain LLDV and MDV is associated with a tornado. This guidance depends on incrementing the thresholds higher and higher and then look at how the False Alarm Ratio (FAR), Probability of Detection (POD), and Heidke Skill Score (HSS) change as the thresholds change using a large sample of TSs/TVSs of all storm types across the country. The HSS score compares FAR, POD, missed detections and correct nulls to show the best values for LLDV and MDV (TWG 2002). In other words, a forecaster's skill in issuing tornado warnings would peak when choosing the threshold values of LLDV and MDV where the HSS peaks.

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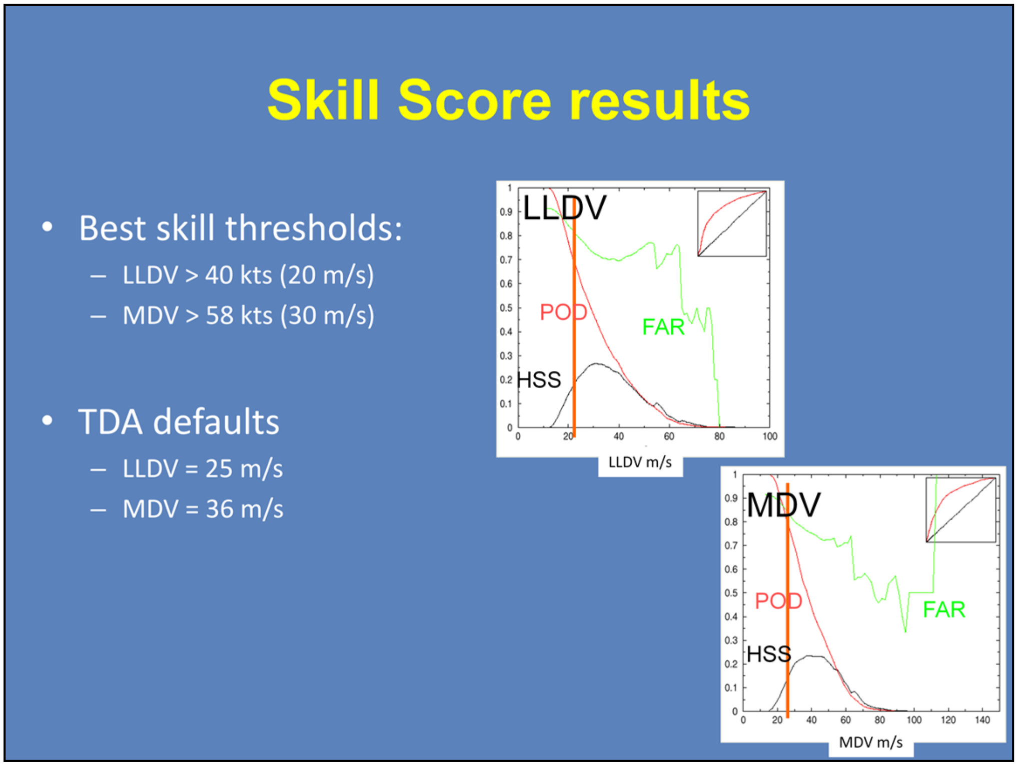

Results show significant skill score values are reached when LLDV exceed 20 m/s (40 kts) and MDV exceed 30 m/s (58 kts). The TDA default parameters are LLDV = 25 m/s and MDV = 36 m/s. As these values increase, the likelihood of a tornado also increases; however, a forecaster waiting for progressively higher values beyond where the HSS peaks before issuing a tornado warning suffers an increasing chance of missing a tornado.

As a note, the data used in TWG 2002 was done with legacy resolution velocity data and using the TDA to collect only instances of TVSs. Current indications are that super-resolution velocity data involving TSs and TVSs offer similar skill in tornado discrimination, however the HSS peaks roughly 5 m/s higher. Given that super-resolution data is likely to detect higher peaks in velocity, this result is not surprising.

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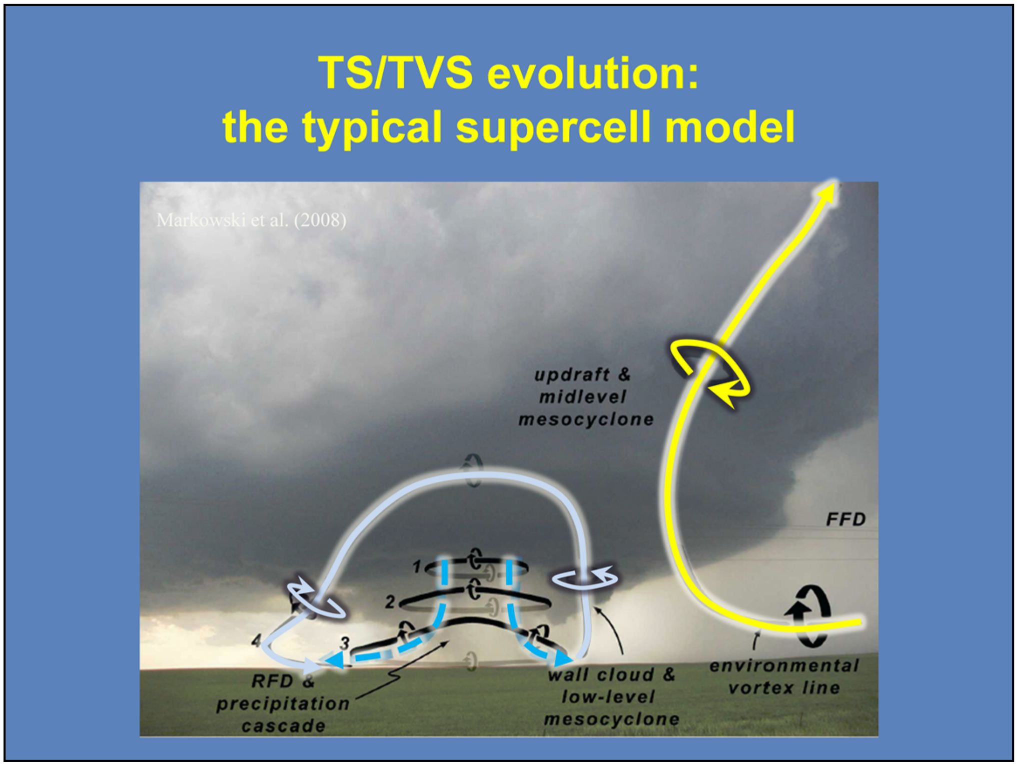
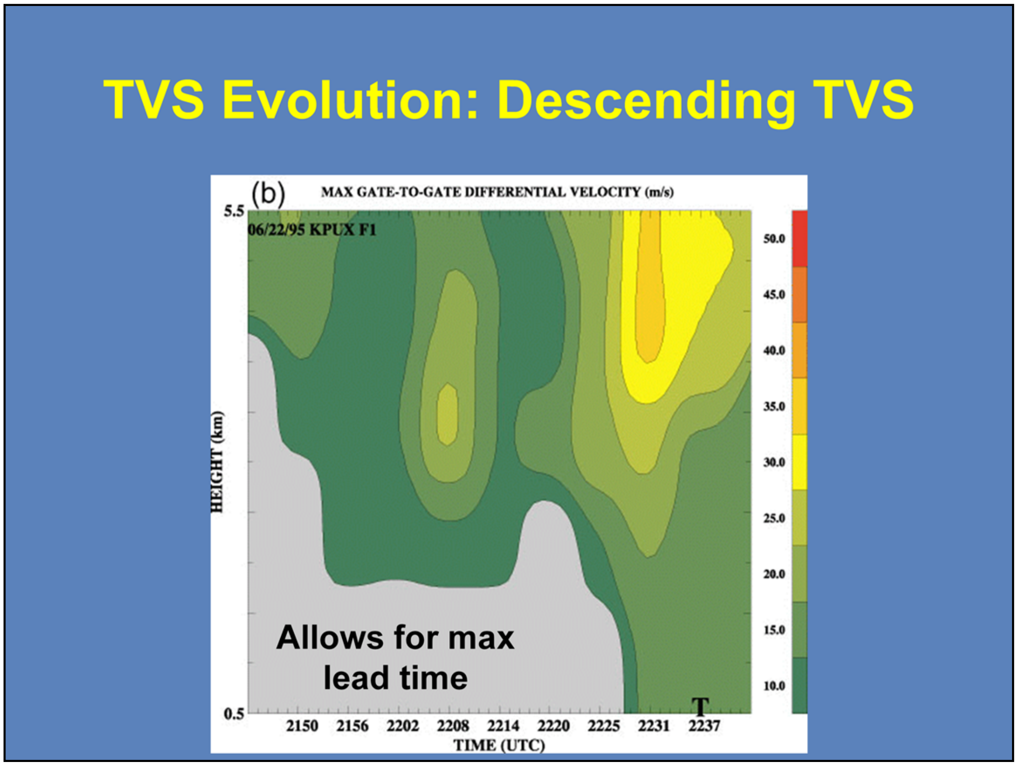


Traditional supercell mesocyclones often begin at midlevels as the updraft tilts environmental vorticity. As the updraft strengthens, the midlevel vortex may strengthen as well, possibly manifesting itself as a TS/TVS at those levels depending on the radar sampling.

At lower levels, the rear flank downdraft begins to generate horizontal vorticity around its exterior. As can be seen in this figure some of the vortex lines on the exterior of the RFD may get entrained into the main midlevel mesocyclone and updraft. As a result, a new low-level mesocyclone quickly develops at lower levels inside the wrapping RFD and under the updraft.

Since the low-level mesocyclone is feeding off of air of downdraft origins, it is often referred to as an occluded mesocyclone where the term occluded means the pre-storm air is no longer entraining directly into its base.



Following the mechanism of tornado formation in the last graphic, the WSR-88D may indicate that the mid-level TS/TVS is descending as the low-level mesocyclone strengthens. The process may or may not continue to intensify into a tornado, however to the warning forecaster, it may appear that the TS/TVS originates in the mid-levels and then descends to the ground over time. This process allows for the maximum lead time in a tornado warning.

And you can see that in this graphic as the envelopes of MDV descend with time until a tornado was reported at 2237 UTC.

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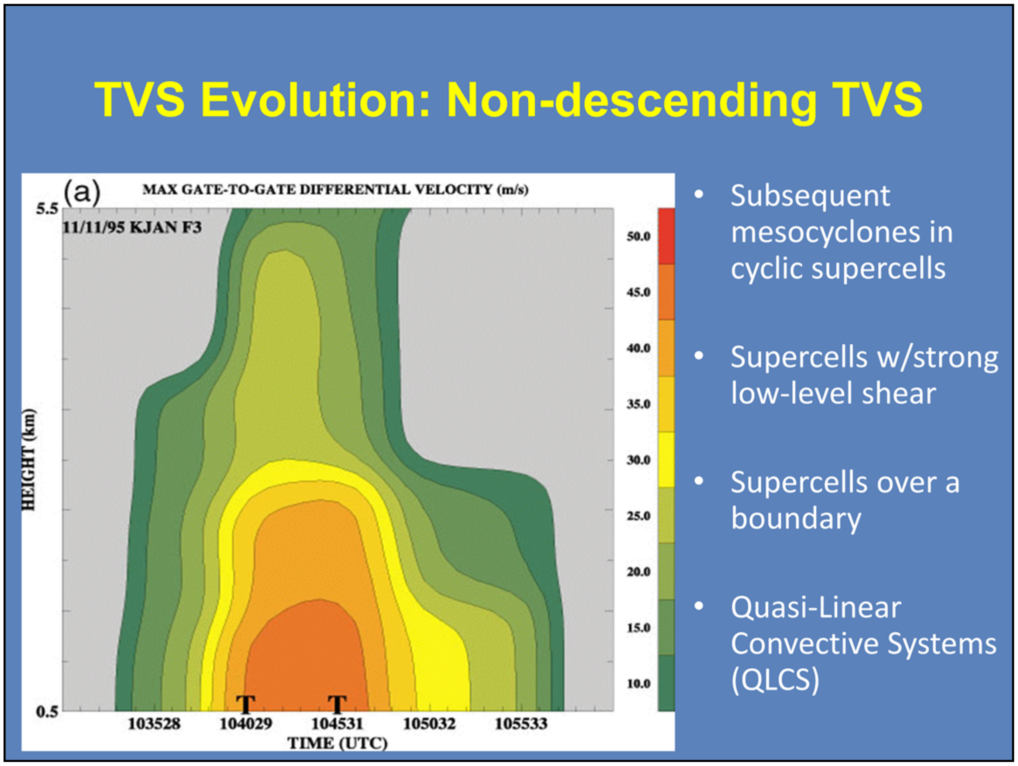
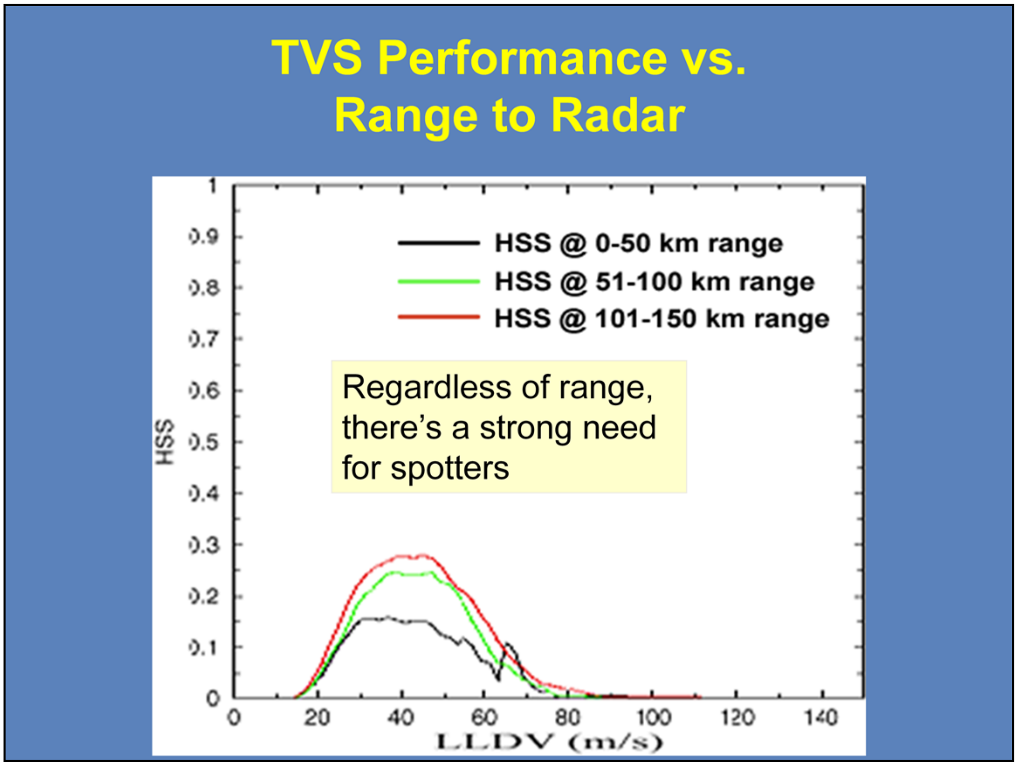
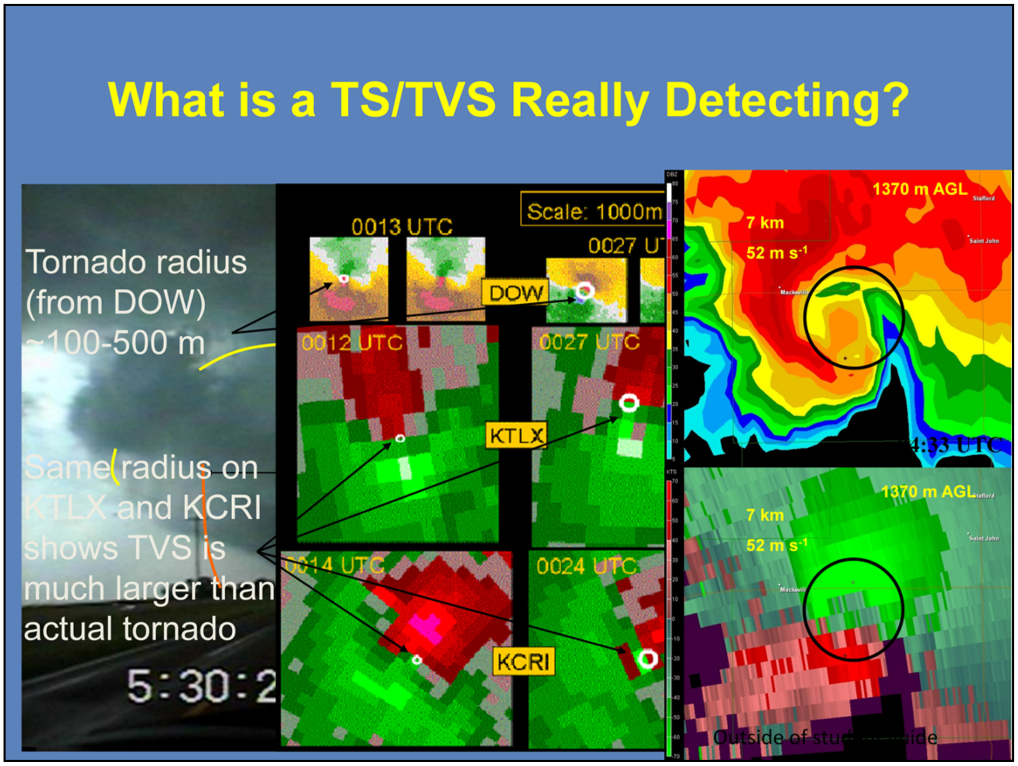


Non-descending TVS

Studies (Trapp et al., 1999; Wakimoto and Atkins, 1996) have indicated that not all mesocyclone induced TSs/TVSs descend from mid-levels to reach the ground. **About half originate at low-levels and then extend upward**. Often, this non-descending paradigm is associated with subsequent mesocyclones in cyclic supercells, or in supercells with very strong lowlevel shear, possibly from an outflow or other type of boundary. Non-descending TSs/TVSs occasionally originate within supercells above a boundary containing strong vertical vorticity (Wakimoto and Atkins, 1996). This is a critical observation, since low-level tornadogenesis can occur in moments. Warning lead time depends on monitoring the trend of the low-level TS/TVS shear, picking the right thresholds, and anticipating rapid tornadogenesis. Non-descending TSs/TVSs will be discussed further in the section on multicell squall lines.



TS/TVS detections are limited in range owing to degraded radar sampling with range. However, a comparison on the statistical performance of TSs/TVSs to detect tornadoes vs. range to radar indicates that there is little range degradation out to 150 km (~78 nm). These results show that other factors could be more important than radar range degradation - at least within the first 150 km. Therefore, there is a strong need for spotters regardless of range to the nearest radar.



The TS/TVS in the image is showing a small circulation that is still much larger than a significant close-range tornado as depicted by the high resolution Doppler on Wheels (DOW) data. Note that the tornado widths represented by the white circles were sampled by the DOWs and then pasted upon the most recent scan from KTLX and then KCRI (a testbed radar). Now you can see how much smaller the actual tornado widths were than the distance between the velocity peaks from the WSR-88Ds.

At low-levels, the TS/TVS most likely represents part of the intensifying mesocyclone inside the wrapping RFD. The RFD axis is usually closely aligned with the axis of the wrapping hook echo. The low-level flow inside the hook/RFD gradually accelerates with decreasing distance to the circulation center.

At far ranges, the TS/TVS may be more appropriately called a non-divergent mid-level mesocyclone. Since they are relatively common, nondivergent mid-level mesocyclones often appear as TVSs at far ranges and that is why we often limit TS/TVS to ranges less than a range of 80 nm (150 km).

In a few rare cases, the WSR-88D can resolve the tornado where the vortex core diameter is four or more effective beam widths wide. Such tornadoes are essentially low-level mesocyclones whose strength reaches tornadic values. These features manifest themselves as TSs and it is possible to have subvortices reveal themselves as TVSs within the larger TS. The Greensburg, KS tornado of 4 May 2007 exhibited TS characteristics from the KDDC radar with legacy resolution data, and within the TS, there were asymmetries in the velocity data that may have suggested such an occurrence (Lemon and Umscheid, 2008).

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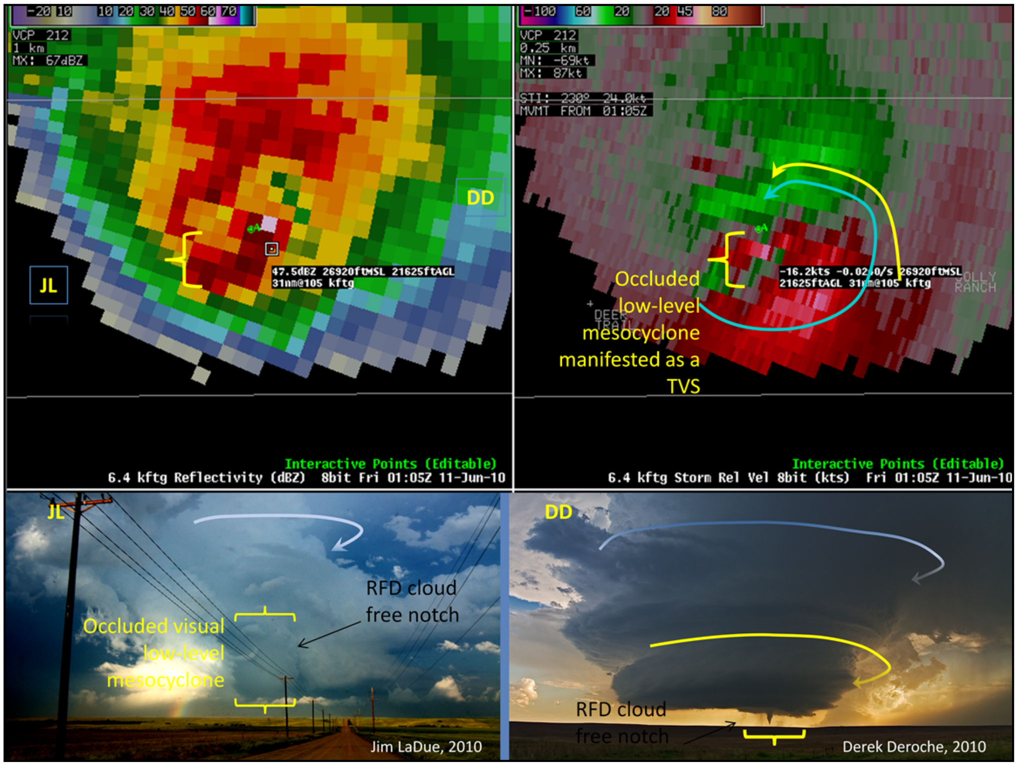
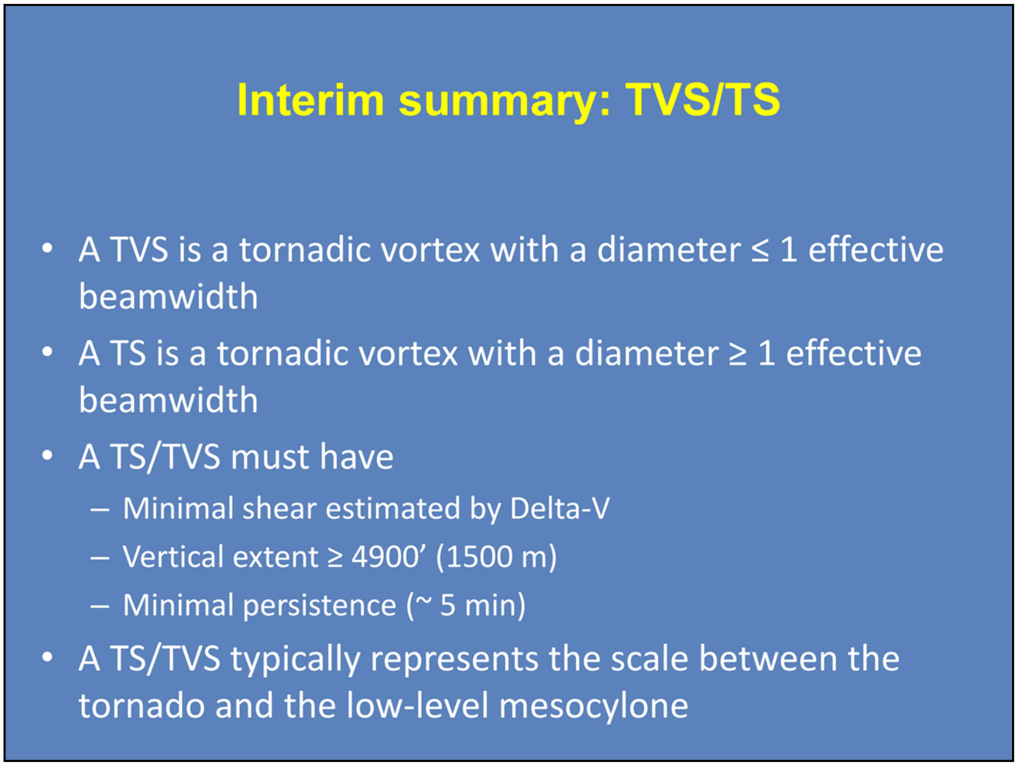


I like to discuss how the WSR-88D views the difference between the mid- and low-level mesocyclone. Here is an example taken from one of the Project VORTEX2 supercells near Deer Trail, CO on 10 June, 2010. Starting at low-levels, there is a classic TVS where Vr max and Vr min are located on adjacent gates, even with super-resolution data. This indicates that the tornadic vortex core was too small to be resolved. However, the potential flow outside the velocity peaks is part of the circulation outside the core and it is easily resolvable. The gate-to-gate LLDV for this TVS is about 55 kts at the lowest scan. Here we are looking at 0.9 deg in elevation

There are two pictures taken from either side of the supercell that will help to define the physical nature of this TVS. The first picture was taken west of the supercell where we can see many important features from the back side of the storm. The second picture was taken from the more traditional front side of the supercell. Overall the two images depict quite different scenes. The backside shot depicts very convective looking towers right from the cloud base to near the anvil with no obvious features that imply a rotating updraft. The front side shot, however, depicts an updraft that we normally associate with a supercell, a smooth circular updraft with circular banding. These views are typical for a supercell. More importantly is the prominent cloud-free notch extending up from cloud base to 1/3 the way up the convective tower as viewed from the backside. From the front, that dry slot appears left of the developing tornado. The low-level mesocyclone denoted by the potential flow in the velocity image is most likely that area including the dry slot, the tornado, and all the way to the north side of the updraft wall (inside the brackets). The inner core of the low-level mesocyclone, the tornado, is too small to be resolved so the velocity signature in radar is that of a TVS.

Going up in altitude to 11 kft AGL, we see the upward extending low-level mesocyclone (manifested as a TVS), and then strong inbound velocity to the north. This level would be above the cloud base as seen from front side visual image where the inbounds would visually appear as strong horizontal flow going left to right and then around the northern edge, and then to the back side of the updraft. The northern side of the updraft in the backside image is obscured by the bright updraft.

The upward extent of the low-level mesocyclone is visible even to 22 kft AGL. We also still see the inbounds on the northern side of the updraft but there is now a strong outbound to the south of the updraft. The blue streamlines depict approximately the altitude this flow field exists.



A TVS is a tornadic vortex with a diameter ≤ 1 effective beamwidth

A TS is a tornadic vortex with a diameter ≥ 1 effective beamwidth

A TS/TVS must have

- Minimal shear estimated by Delta-V
- Vertical extent ≥ 4900' (1500 m)
- Minimal persistence (~ 5 min)

A TS/TVS typically represents the scale between the tornado and the low-level mesocylone

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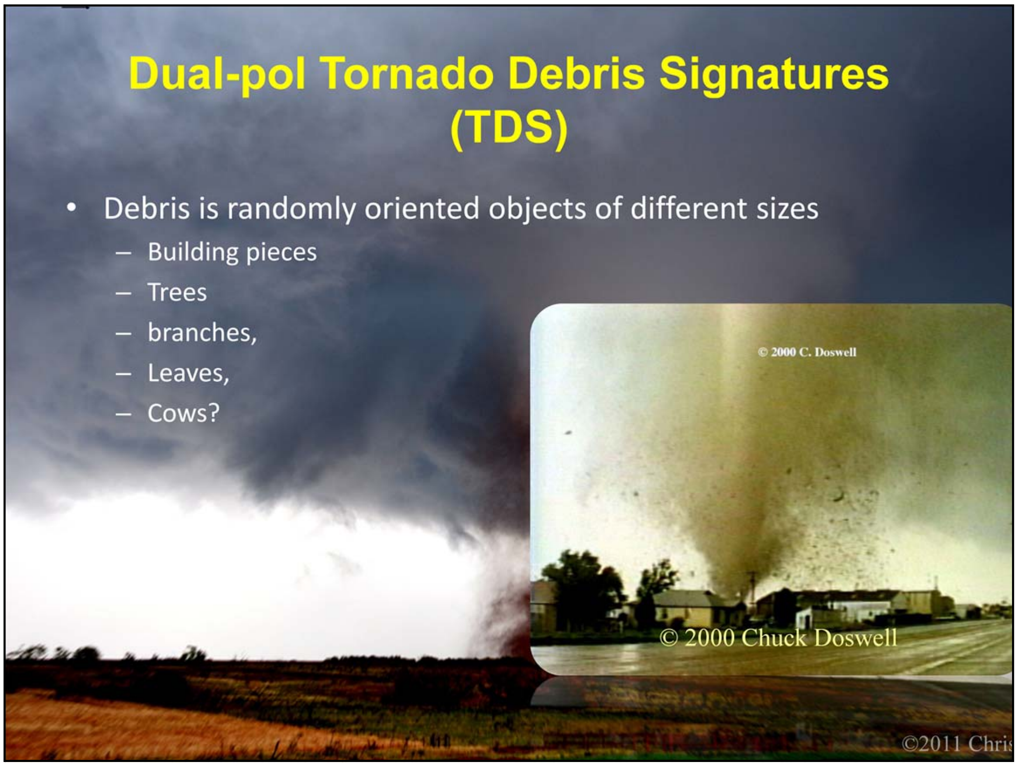

A valid identification of a Tornado Debris Signature (TDS) helps a warning forecaster identify that a tornado is most likely occurring and is producing damage. A valid signature is likely to be considered as close to an actual tornado detection as a

spotter report. With that being said, the process of identifying a TDS must be done carefully to avoid an incorrect identification.

The radar is detecting tornado debris that is comprised of large, randomly oriented objects ranging from leaves to building fragments. This example shows debris as a tornado went through Pampa TX on 08 June 1995.

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Dual-pol TDS characteristics

- CC is the best product
- Delay factor 5 – 10 min before detection
- Using CC alone is not sufficient

Because debris is randomly oriented, the dual-pol radar correlation coefficient product (CC) is by far the best product to discriminate debris from meteorological echoes. When analyzing a TDS, remember that debris was actually introduced to the circulation 5-10 minutes earlier. It takes time to loft and distribute the debris. And after the dissipation of the tornado, it takes time for the debris to settle out. However, identification of tornado debris with CC alone is not sufficient. Let's go through a method to make a good detection of a TDS.

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Dual-pol TDS detection method

	product	strategy	Verdict?
			Yes or no
1	Velocity	Is azimuthal couplet present?	
2	CC	Are values < 0.8 in or near the velocity couplet?	
3	Reflectivity	Are values > 35 dBZ where CC < 0.8?	
4	ZDR	Values near zero help confirm above 3 criteria	
5	verdict		

1. First identify a storm-scale vortex such as a mesocyclone and/or a TVS (or TS) located in the vicinity of an updraft as per the instructions in identifying a mesocyclone, TS and TVS. There is no lower bound velocity threshold but the rotational couplet should be pronounced. In some cases, a vortex may be unresolvable and spectrum width may show a local and very high peak.

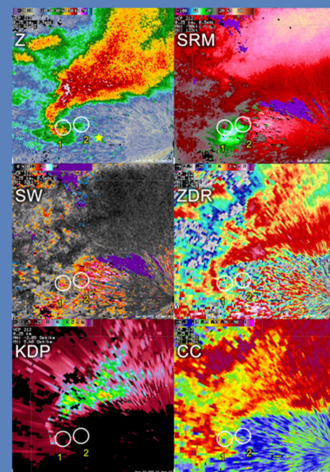
2. In the vicinity of a tornado vortex, look for a CC small minimum in CC. Typically, a value < 0.8 indicates a good potential for randomly oriented scatterers. This is not a hard threshold, however. Sometimes CC in a valid TDS may fall to 0.9 **only** when rain is mixed with debris. However, a TDS with a CC this high is rare.

3. If you have a localized CC minimum centered near a vortex, then check to see if the reflectivity is at least 35 dBZ. Lower reflectivities may result in untrustworthy CCs. In addition, the CC values may be the result of other non meteorological scatterers, such as insects or light suspended vegetation particles.

4. ZDR is typically near zero in valid tornado debris. However, the signature is not nearly as pronounced as CC. Nevertheless ZDR can be used as a confirmatory check.

Dual pol TDS case study

- 2012-April-15 0100 UTC
- Northwest of KVNX
- Two velocity couplets
  - Labeled 1 and 2
- Both in the same supercell



An example of two TDS candidates occurred with the 14 April 2012 Cherokee, OK supercell shown in this image. Follow along with the tornado identification methodology to find that two TVSs exist enclosed by Circles 1 and 2. Each vortex at the lowest scan has passed the vertical continuity check.

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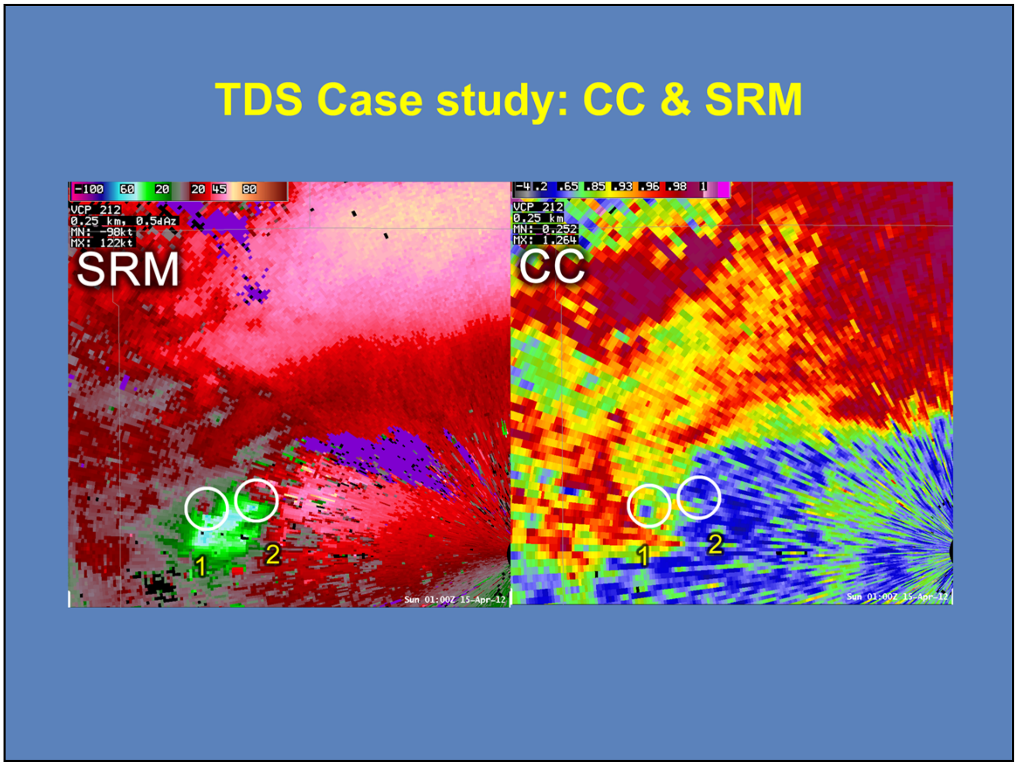
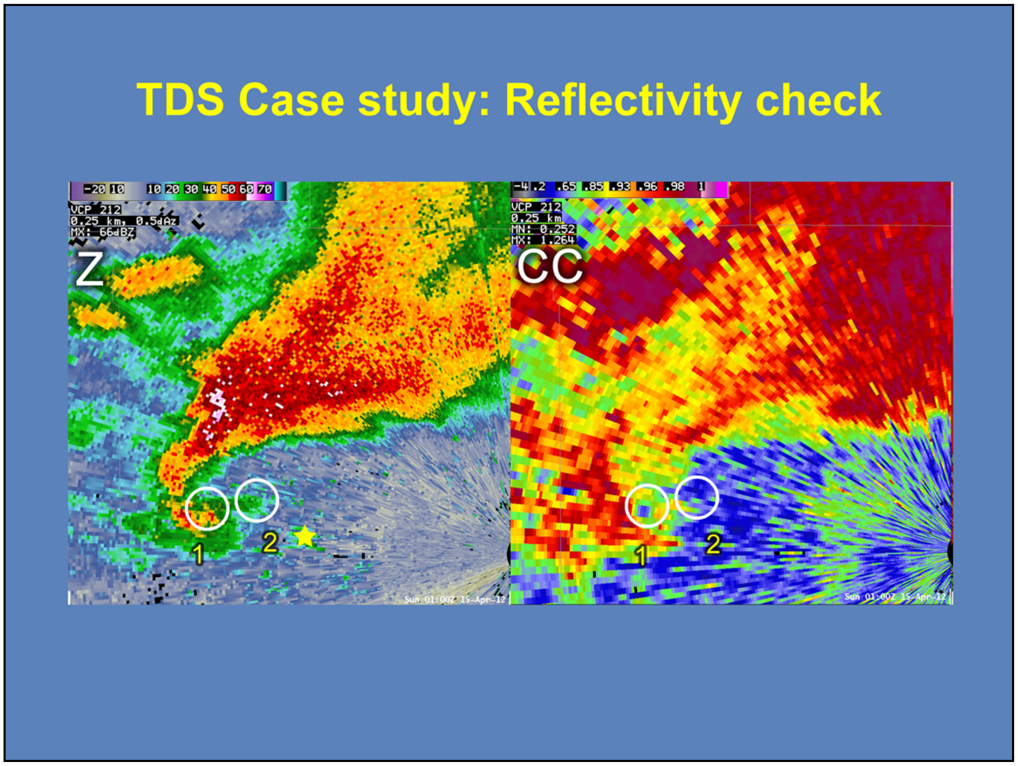


Both vortices exist along the edge of an RFD outflow and in the proper spot relative to the parent supercell. Going to the CC panel, note that there are low CC values within each circle. Circle 1 has a more pronounced CC minimum than Circle 2, while the low CC in Circle 2 is more in the low CC inflow.



The reflectivity panel shows that Circle 1 contains values above 20 dBZ and even some in excess of 40 dBZ. There may even be the suggestion of a debris ball, though not well defined. However, note that the lowest CC overlaps with 40 dBZ echoes in the southwest part of the circle. Circle 2 may have lower CC values but the reflectivity at the vortex center is just below 20 dBZ.

So, there is strong confidence that the low CC within Circle 1 is from tornado debris. However, confidence is low that the low CC values within Circle 2 are associated with any debris.

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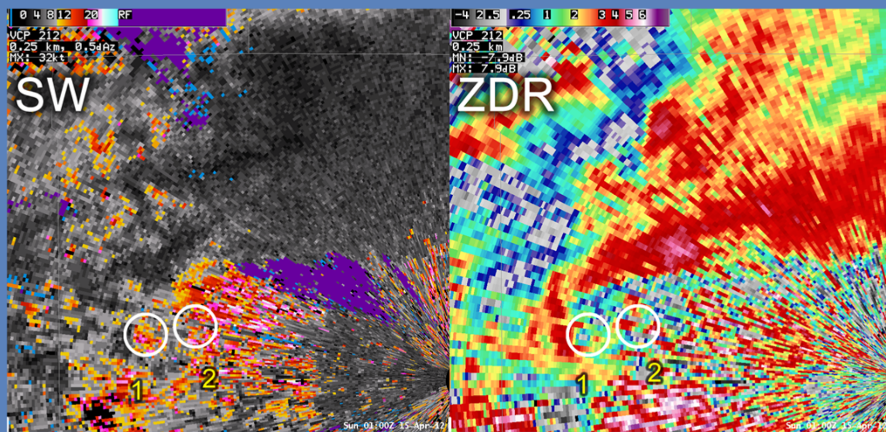
TDS Case study: ZDR and SW



Within Circle 1 there is a well defined ZDR minimum at the same location as the CC minimum. This adds confidence that there is a TDS in Circle 1. The ZDR in Circle 2 is mottled with a mix of very low and high values, similar to the pre-storm air. Confidence remains low for a TDS in Circle 2.

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Dual-pol TDS detection method

	product	strategy	Circ #1	Circ #2
			Yes/no	Yes/no
1	Velocity	Is azimuthal couplet present?		
2	CC	Are values < 0.8 in or near the velocity couplet?		
3	Reflectivity	Are values > 35 dBZ where CC < 0.8?		
4	ZDR	Values near zero help confirm above 3 criteria		
5	verdict			

Circle 1 contains a TDS. There is a TVS in the vicinity of a hook echo with a well defined CC minimum with sufficient reflectivity. Circle 2 shows no TDS signature. While the radar may be depicting debris, the signal cannot be separated out from the low CC non meteorological scatterers that exist around and within the inflow. This is the low CC inflow signature that is common in boundary layers with non-meteorological scatterers.

If there is a ‘No’ in any product in the first three rows (V/SRM, CC, Reflectivity) then you do not have confidence that you are seeing a Dual-pol TDS. The ZDR helps confirm the verdict but is not as strong of an influence.

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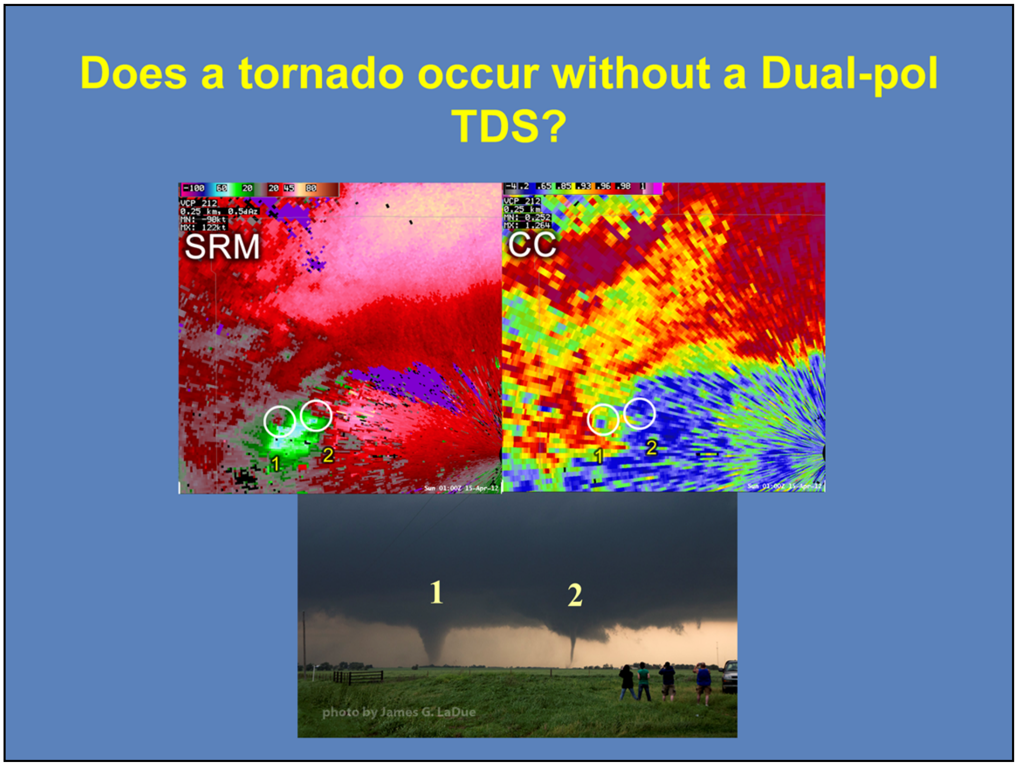

Does this mean that only one tornado is in existence? In fact it doesn't. In the photo, two tornadoes are traveling across the landscape west of Cherokee, OK.

The tornado on the left corresponds to Circle 1 while the one on the right corresponds to Circle 2. At this time the left tornado is larger and has been in existence for longer. Both of those will result in more debris later. The one on the right has just formed from the new mesocyclone and has yet to loft enough debris to raise the reflectivity sufficiently from the KVNX radar to help discriminate insects from debris at this range and height.

This example highlights one aspect of TDSs that is also common to many other radar signatures. That is the absence of a clear signature doesn't rule out the existence of the hazard for which the signature refers. But the presence of a TDS is as strong an indication of a tornado as a spotter report. The TDS should serve to raise confidence that tornado is or has been in progress. But warning issuance should never have to wait until a TDS occurs.

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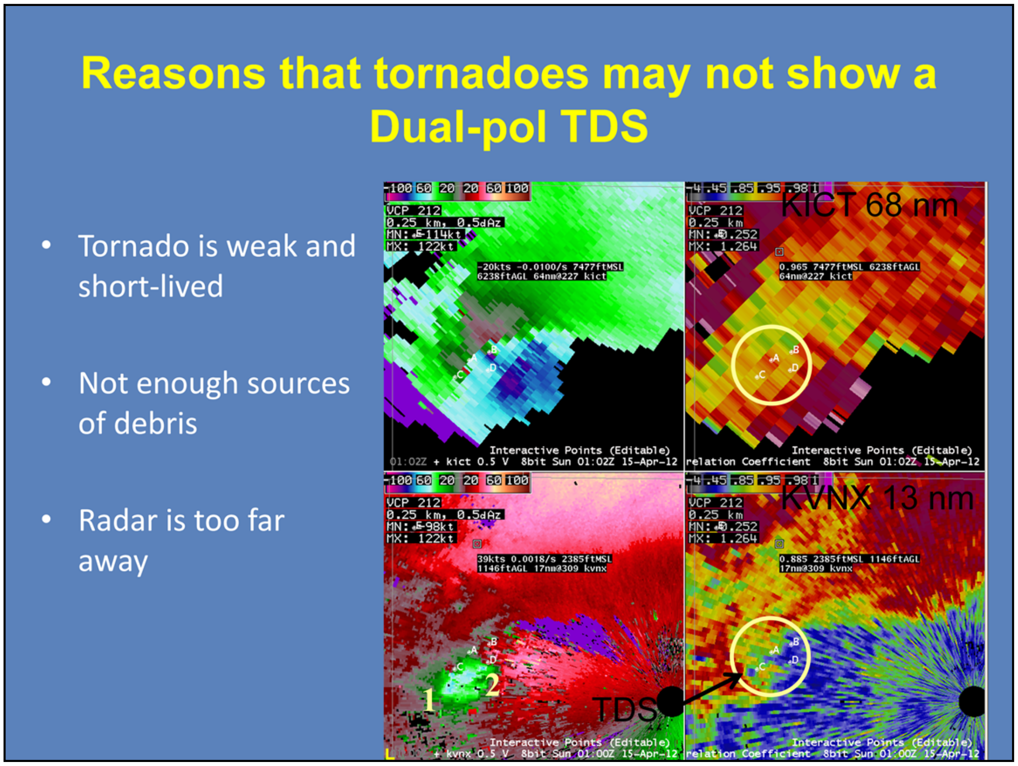
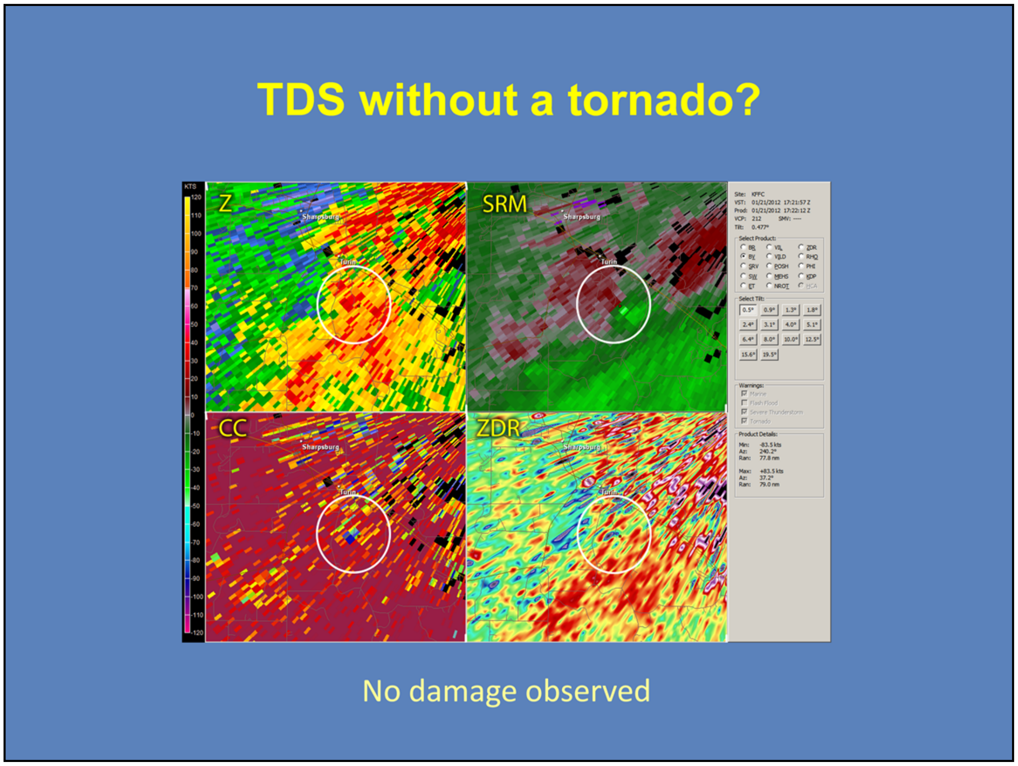


Three factors may lead to a tornado not exhibiting a Dual-pol TDS:

- The tornado is weak and short-lived. There is insufficient strength and time to loft detectable debris.
- There are not enough sources of debris. A tornado crossing an open dirt field is unlikely to generate as much detectable debris as a tornado of the same strength going through a town. Fine dust particles are too small for S-band radar detection in order to generate a TDS. A tornado must loft at least leaves, grass, and forest debris to generate a TDS.
- The radar is too far away. Range is everything. Weak tornadoes are unlikely to be detected more than 40 nm away. EF2 and greater tornadoes may generate a detectable TDS up to and possibly over 60 nm in range. The limiting factors are the height to which debris is lofted and the size of the debris footprint.

A great example of the distance problem comes up with the Cherokee storm when the same two tornadoes are viewed from KICT 68 nm away. The top two panels show the KICT radar where no CC < 0.8 was visible (top right panel) anywhere near the TDS that was visible from KVNX from 13 nm range. The letters represent the vortex locations from KICT (A and B), and KVNX (C and D, labeled 1 and 2).



Has a TDS been present without a tornado? The answer is possibly but very rarely. One such case occurred where a low CC bull's-eye was collocated with a rotational velocity couplet within high reflectivity in northern Georgia. NWS damage surveyors were unable to find significant tree damage. It is almost certain that a vortex was lofting light debris. The question is whether or not the vortex was strong enough to be defined as a tornado.

However, it is theorized that the main culprit in false detections seems to be when vortex signatures, low CC, and low ZDR values have been correlated within the weak reflectivity inflow notch ahead of a pendant, or hook echo. These associated vortex-like velocity signatures, in some instances, may have been side-lobe related, but in most instances, that is not the case. We simply do not know the origin of these “ghost-like” vortex signatures. However, the low values of CC and ZDR are easily explained in the updraft inflow notch of many supercells.

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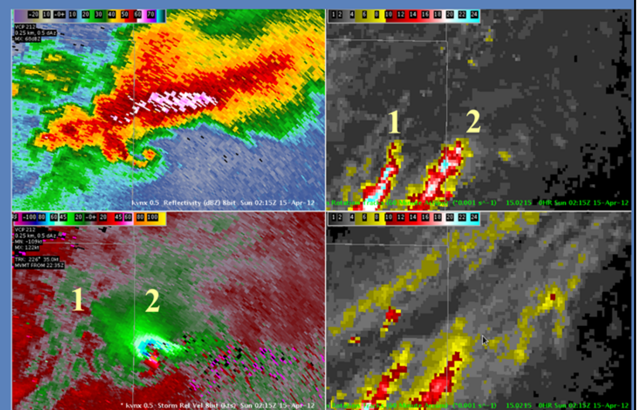
Multi-Radar/Multi-Sensor (MRMS) Mesocyclone/Tornado Signatures

- MRMS Low-Level (0-2 km) Rotation Tracks
  - Show low-level mesocyclone/possible tornadic circulation
  - ~0.015 s^{-1} indicative of possible tornado
- MRMS Mid-Level (3-6 km) Rotation Tracks
  - Show parent mesocyclone
  - ~0.01 s^{-1} indicative of significant mid-level rotation of possible tornado
- Both useful for tracking cyclic supercells



Another suite of radar products that can identify signatures near the tornadic-scale is Multi-Radar/Multi-Sensor (MRMS) Rotation Tracks products. Recall from the MRMS Products Course that these are time intervals of Azimuthal Shear and that they're available at low- and mid-levels. The Low-Level Rotation Tracks can identify low-level rotation and possibly even tornadic circulations at its higher values. Little research has been conducted to test the utility of this product for tornado detection, but if you see values greater than .015 s^{-1}, you should check to see if your other tornado signatures correspond with the azimuthal shear magnitude.

Mid-level Rotation Tracks show the mid-level, parent mesocyclone. Values greater than .01 s^{-1} may indicate strong mid-level rotation. Additionally, both Tracks products make cyclic supercells evident. In the example on the right, our storm from the TDS example created another set of twin tornadoes during its lifecycle. The separate tracks are visible on both the low-and mid-level Rotation Tracks.

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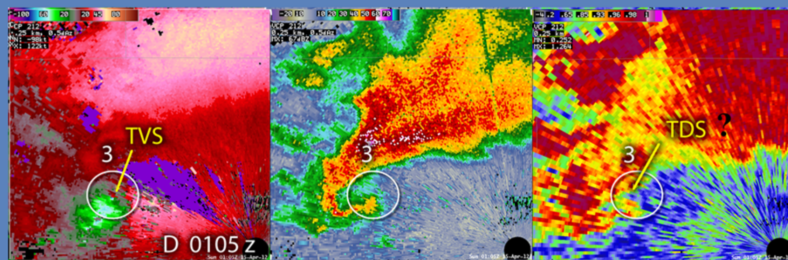
Tornado Lifecycle



- 0041 z - #2, convergent meso; #1 a TVS with TDS
- 0051 z - #2 a TVS with TDS; #1 dissipated, false TDS
- 0100 z - #3 convergent meso; #2 a TVS with TDS
- 0105 z - #3 a TVS with marginal TDS; #2 dissipated

http://www.wdtb.noaa.gov/courses/rac/severe/objects/CherMeso/

Now let's see how the tornado evolved over time for the same storm covered previously. In the 3 panel image a new rotational velocity couplet, mesocyclone (#2) begins with a strong convergent component as the old occluded rotational velocity couplet (#1) reveals itself as a TVS. The CC depicts a prominent TDS and in fact the reflectivity shows a donut hole at the tornado location.

Ten minutes later, the old TVS (#1) dissipated. What appears to be a continuing TDS is actually false. The reflectivity shows that the low CC bull's-eye is in very low reflectivity and there is no velocity couplet there. The new velocity couplet (#2) has become a TVS and is sporting a well defined TDS at the tip of the hook.

Ten minutes later, a new rotational velocity couplet (#3) has formed into a mesocyclone at the tip of the hook echo with a strong convergent couplet. Meanwhile rotational velocity couplet (#2) is still a TVS with an accompanying TDS.

Note that five minutes later, the rotational velocity couplet (#3) has become a TVS that is also accompanied by a TDS. This TDS is relatively marginal given the weak reflectivity but it has enough of a prominent CC minimum that debris could be the cause. Meanwhile, the rotational velocity couplet (#2) dissipated as it moved to the left of the track of the parent storm.

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Summary

- A vortex is depicted as a
  - TVS when: effective beamwidth > vortex core diameter
  - TS when: effective beamwidth < vortex core diameter
- Three criteria satisfy a TS/TVS
  - Vertical continuity, time continuity, a localized peak in azimuthal shear with little convergent component, range within 150 km
- TS/TVS strength
  - $\Delta V = V_{r\ max} - V_{r\ min}$ whether or not the velocity peaks lie on adjacent azimuths.

Tornado-scale vortices manifest themselves as a nearly pure rotational velocity couplet with an isolated maxima in **Vr max, and Vr min**. If the Doppler radar's effective beam width is the same size or smaller than the vortex core diameter, the vortex manifests itself as a Tornado Signature (TS). If the effective beam width is larger than the vortex core diameter, then a Tornadic Vortex Signature is (TVS) is the result.

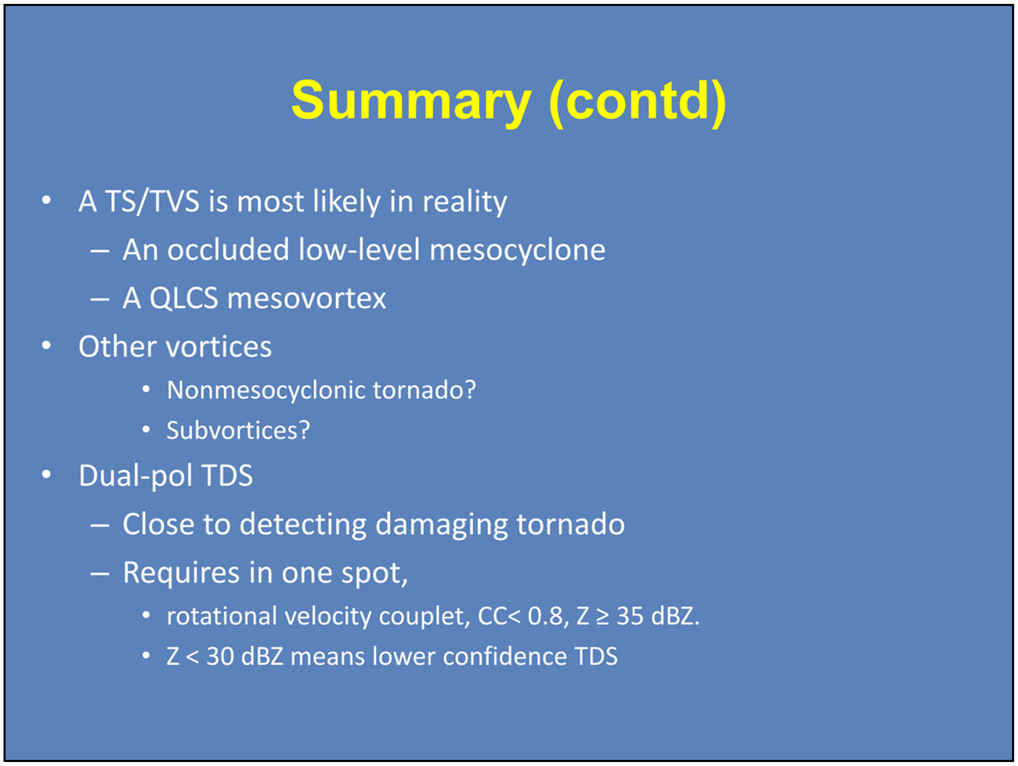
A TS exhibits both potential flow outside the vortex core, and solid body rotation within the core with Vr max, Vr min separated by at least one azimuth. This is the same as a Rankine combined vortex associated with nondivergent mesocyclones.

A TVS exhibits only potential flow and Vr max, Vr min on adjacent azimuths**.** Sometimes both Vr max, Vr min may be enveloped by the same radar beam resulting in a reduced mean velocity. In those cases, Vr max, Vr min may be separated by one effective beamwidth.

A valid TS/TVS needs exceedance thresholds in the velocity difference. However, the threshold is made to be flexible for you as a user, in order to accommodate variability in storm size and range degradation. Remember to discard hard and fixed rules on what constitutes a minimum velocity difference for an operator defined TVS.

A second criteria that is important is a TS/TVS should have some vertical continuity. We would like to see that continuity extend across at least two elevation scans. However, use 1500 m as a good starting point for vertical depth.

A third criteria is that the TS/TVS should persist for about five minutes. This criteria is not so hard since there are plenty of situations where a TVS may barely precede a tornado, if it does at all. Most often, when near the ground, a TVS or TS are ongoing tornadoes.



A TS/TVS may occasionally represent an occluded mesocyclone in a typical supercell at low levels where the radar is resolving a circulation somewhere in between the tornado scale and mesocyclone scale. Not all TVSs represent occluded mesocyclones, especially when considering non-supercell events.

Remember that you should not depend on a TVS as a primary consideration in a tornado warning. It is a signature that represents one of many cues in your decision making.

A TDS means that a tornado is almost certainly ongoing and capable of damaging any structures it impacts. However, **many tornadoes do not exhibit a TDS** due to range degradation, the strength and size of the tornado, and/or a lack of debris source.

1. Look for rotational velocity couplets in a velocity product

2. Identify an accompanying CC minimum, preferably less than 0.8

3. Check to see which CC minimum satisfies the minimum reflectivity threshold of 35 dBZ. Lower than 30 dBZ reflectivities mean a low confidence level of a TDS.

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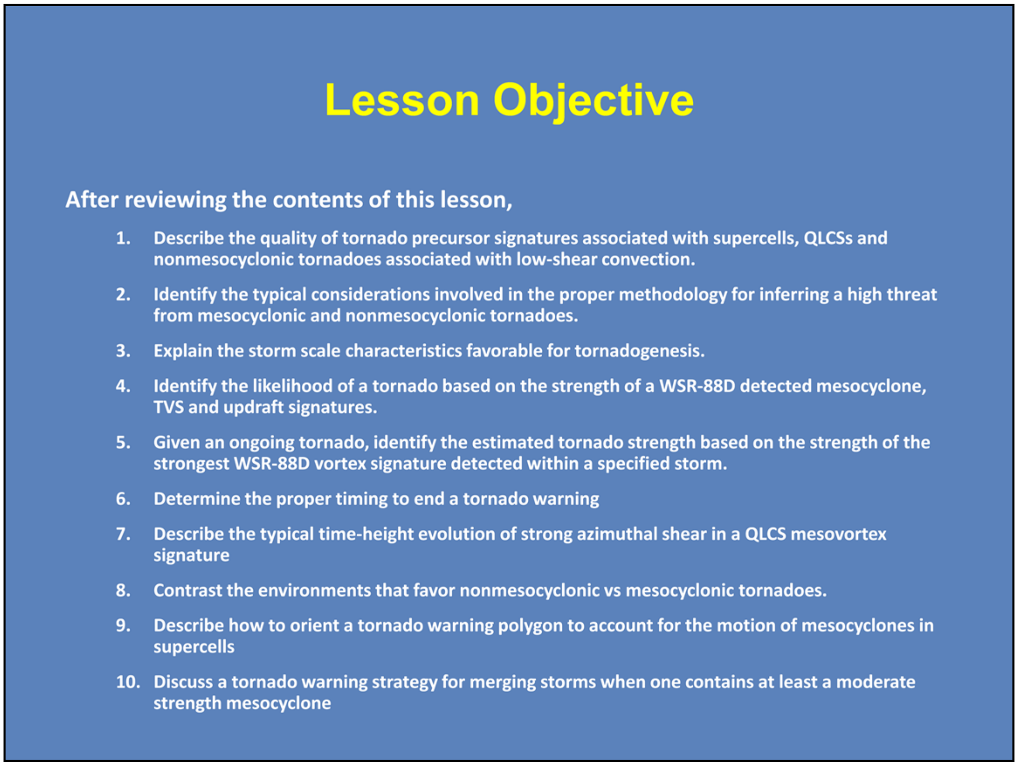
If you have passed the quiz, then you have successfully completed this lesson. If you have any questions, please contact us using any of the e-mail addresses listed on the bottom of the slide.


Radar & Applications Course
(RAC)
Convective Storm Structure and Evolution
Tornado Hazards
Warning Decision Training Division (WDTD)



The problem of deciding whether or not you should issue a tornado warning creates arguably more consternation than any other warning type. Trying to anticipate a tornado is difficult since science does not clearly understand how all the ingredients for a tornado come together. In fact, we may yet not know all the ingredients for a tornado. But, we are not completely at a loss. There are many sources of information available that can supply us with evidence that we can use in making accurate tornado warning decisions. In this lesson, we will describe these sources and give some advice that you can use in your warning operations.

This lesson covers ten objectives in which you may review here before going on with the lesson.

You'll get most of this lesson if you've had the lesson on Tornado signature identification.

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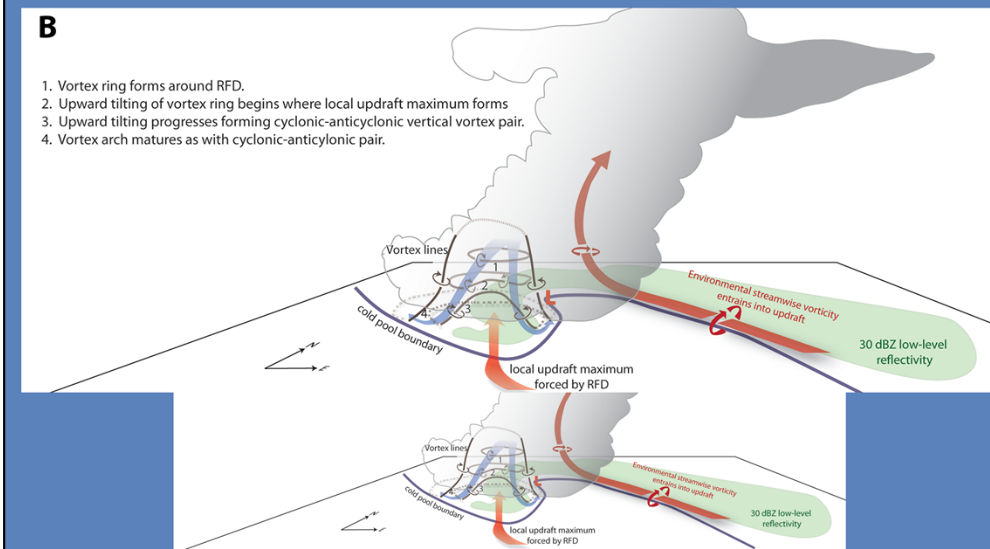
Tornado Types

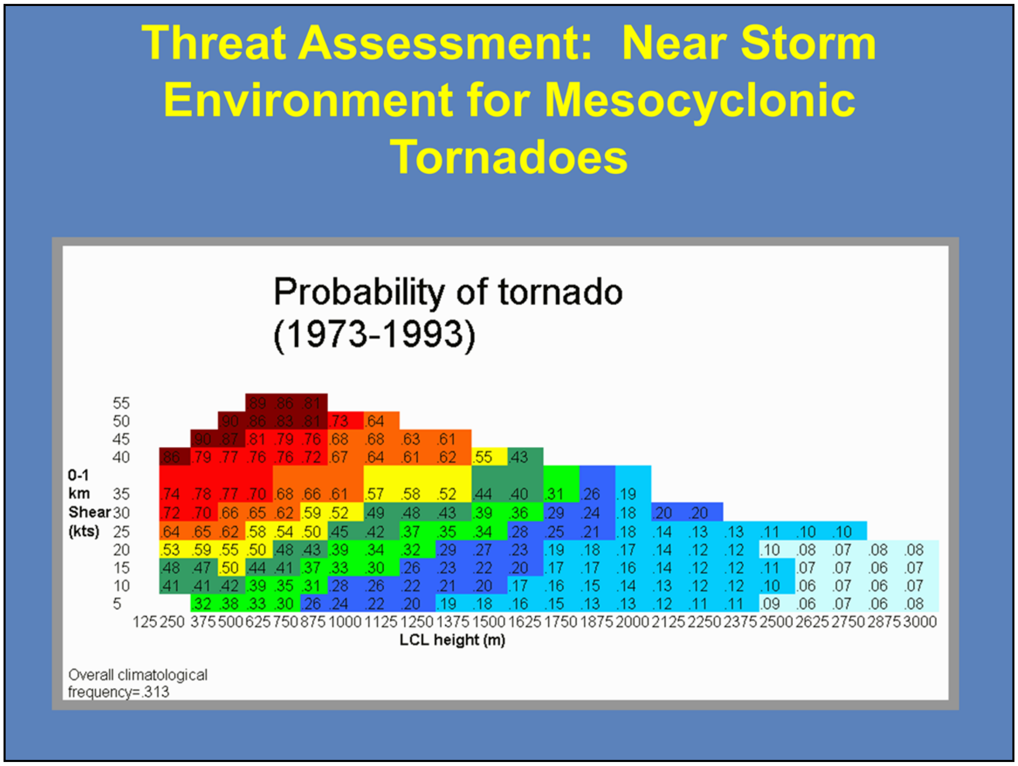


Your strategy for assessing a tornado threat depends greatly on how the threat is evolving. At its most basic level, a tornado requires a high concentration of vertical vorticity embedded in converging and ascending flow into a convective cloud base. The question is, “where did the vertical vorticity originate?”

The most basic origination is that vertical vorticity was already embedded in a line of convergence and waiting for a locally intense updraft to intensify it. This type of tornado has been called a **landspout**, but is more appropriately termed a nonmesocyclonic tornado. This type of tornado is the only kind that can occur in association with updraft dominated ordinary cells. Note that at the time of tornadogenesis, the parent storm is in the towering cumulus phase where little to no downdraft is present in the cell. At this time, the boundary layer is often characterized by a steep or dry-adiabatic lapse rate offering little resistance to ascending motion. Subsequently, appreciable precipitation and a downdraft will develop, and this often leads to the demise of the tornado.

The second type of tornado occurs in an environment that had little or no vertical vorticity previously to the storm. In order to get a tornado, a negatively buoyant downdraft must be involved in order to generate horizontal vorticity

that can then be tilted upward through interaction with a positive gradient of vertical velocity from the rear to the front. An updraft cannot accomplish this by itself. We call this a **mesocyclonic tornado** that typically accompanies supercells. Many tornadoes may derive their vorticity from both sources. As a side note, Quasi-Linear Convective Systems also fit into the 2nd tornado formation paradigm.



Most of the longest and most intense tornadoes accompany mesocyclones from supercells so it is important for development purposes that there is sufficient (> 15 m/s) deep layer shear and instability present in the environment. However, it is also true that most supercells are non-tornadic, so something more is needed to favor tornadogenesis.

Warning decisions should involve the environment (and its changes) just as heavily as they do radar. Research has built up evidence that **mesocyclone-induced tornado environments favor strong 0-1 km shear and a low LCL.** Low-level shear helps to strengthen low-level mesocyclones. The low LCL is associated with buoyant rear flank downdrafts, allowing vertical vorticity to be easily stretched (Markowski et al., 2002). **The LCL and shear should be in an environment that promotes strong low-level updraft acceleration (i.e., low CIN and strong low-level convergence)**.

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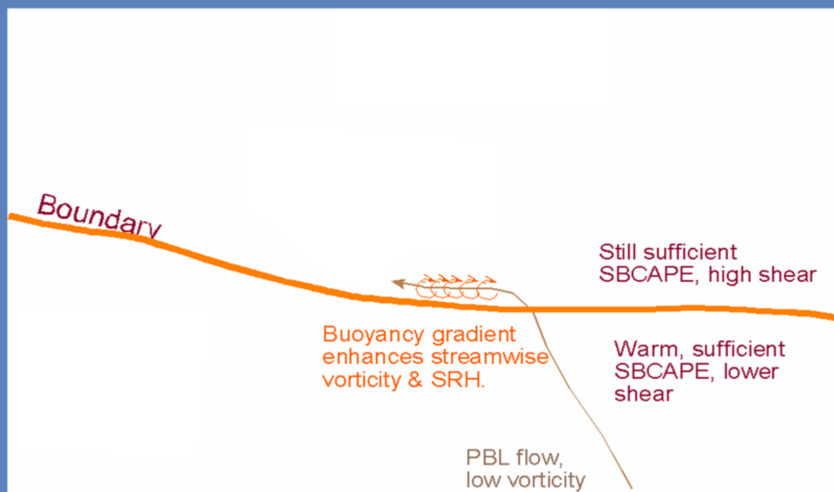
Boundaries and Mesocyclonic Tornadoes



Boundaries are regions that can be locally favorable for tornadic supercells. Look for these types of boundaries:

- **Subtle boundaries with backed winds and good SBCAPE providing a good clue of increased low-level shear, low LCL and little CIN , and**
- **Boundaries with strong vertical vorticity in supercell environments**.

Supercells can stretch environmental vertical vorticity along boundaries at least as effectively as pulse cells and provide little forewarning of tornadogenesis.

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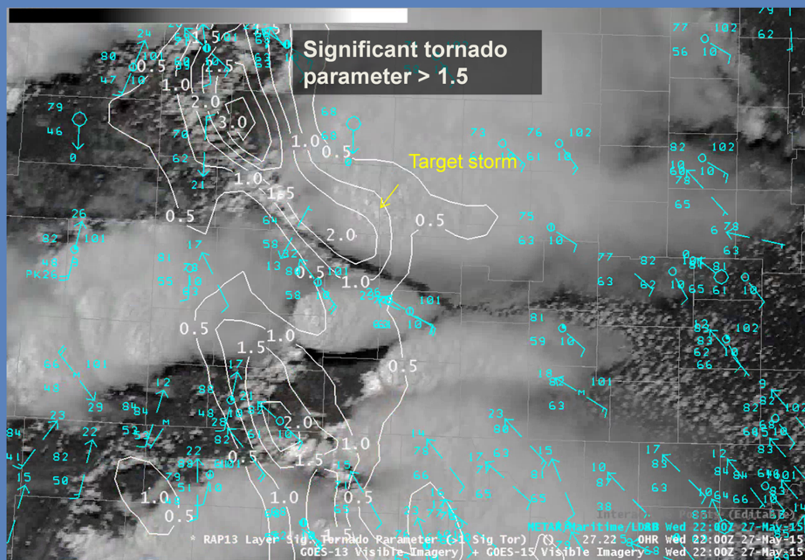
Example: Near Storm Environment to Target Storm



Frequently analyzing the near storm environment is an absolutely critical task in the warning team. If the warning forecaster doesn't do this then a mesoanalyst should be and then informing the warning forecaster. There is a significant amount of work done on showing the relationship between the favorability of the environment and the likelihood of a tornado given that a supercell is capable of producing any type of severe weather. This storm in the image is close to a region where the significant tornado parameter (SigTOR) is between 1.5 and 2. While there has been a lot of research done relating the SigTOR parameter with likelihood of tornadoes, remember to look at its components to determine how the SigTOR parameter is the number that it is. In this case the CAPE and deep layer shear were heavy contributors to the parameter while the MLCL was within acceptable limits and effective SRH was a little low. CIN is not a direct contributor to SigTOR but is very important controller in the storm's ability to stretch low-level vertical vorticity into larger values.

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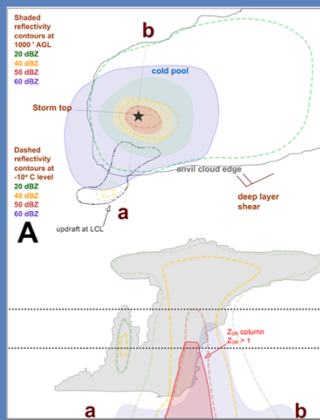
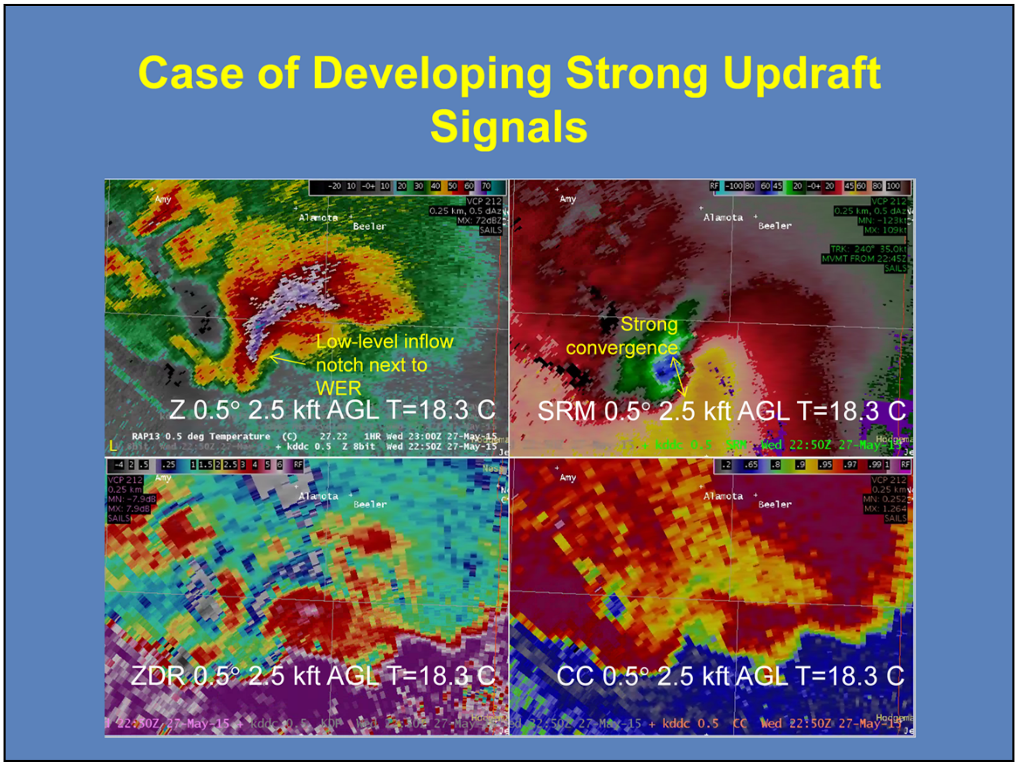
Strengthening Updraft Signals



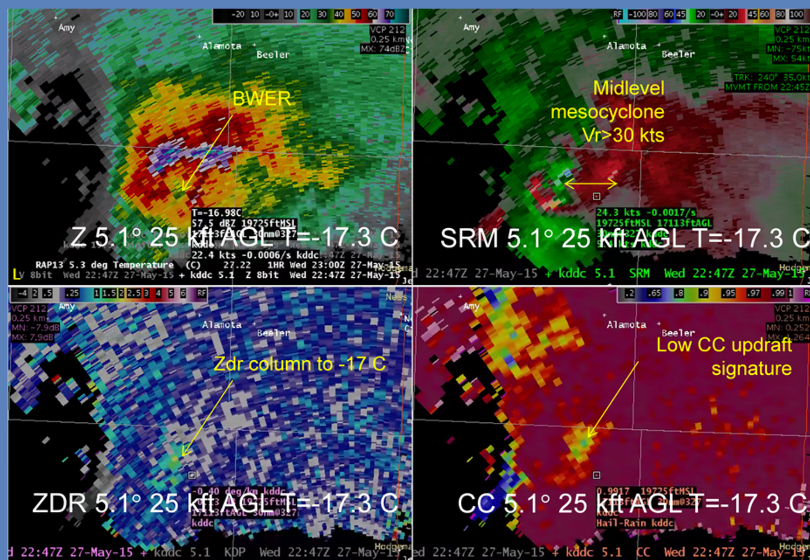
Evidence of a strong updraft in the lowest half of a storm provide even more support that a low-level circulation can be stretched into a tornado. Look for the classic reflectivity signatures, including the onset of a concavity in low-level reflectivity gradient, a strong echo overhang, displacement of the echo top over a WER, and evidence of change in storm motion.

BWERs are rare, but if visible, there is an enhanced threat of a tornado when coupled with a strong mesocyclone and/or TVS/TDS. However, you should not depend on a BWER to consider a tornado warning.

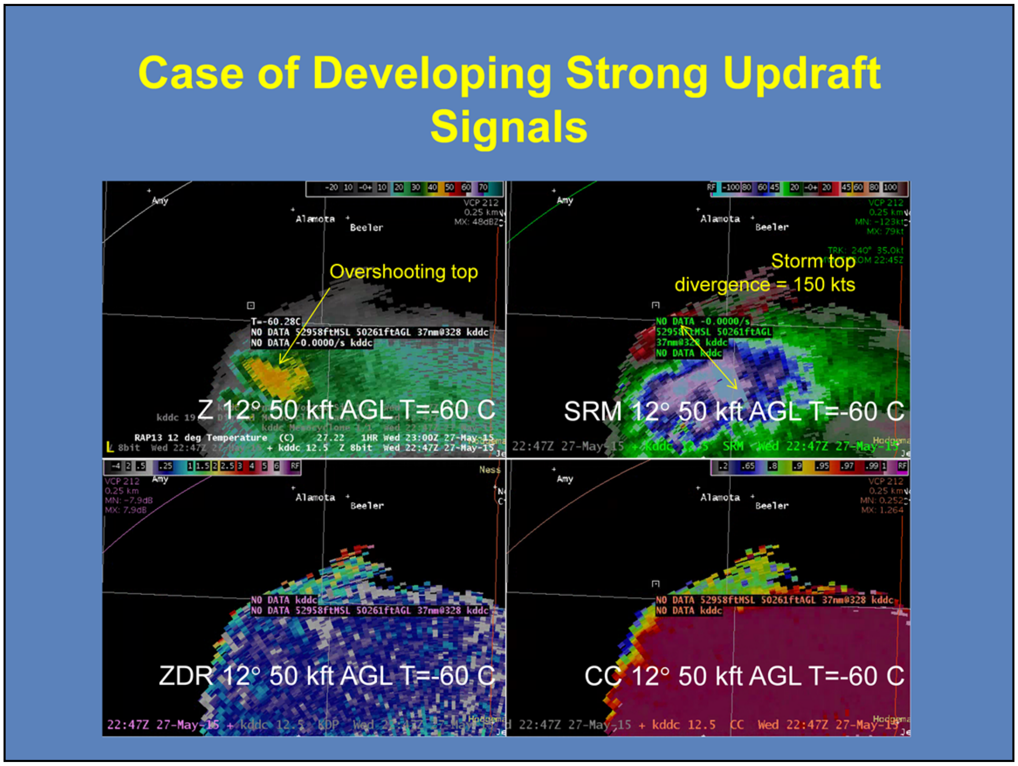


Here is an example of significant updraft signatures at low-levels. A low-level notch is evident in the upper left adjacent to the WER in the proximity of strong low-level convergence. These features indicate the updraft is significantly strong down to low levels which is favorable for stretching any vertical vorticity that can be found in the vicinity.

Case of Developing Strong Updraft Signals



The significant low-level updraft signatures underly a well-developed BWER and midlevel mesocyclone. A Zdr column to -17 C, along with a low CC updraft signature indicate the updraft is quite strong aloft and provides support to the low-level updraft in stretching vertical vorticity.



The storm top also indicates this updraft is powerful with a divergent velocity difference of 150 knots. That value is quite high as far as storms go, even for supercells.

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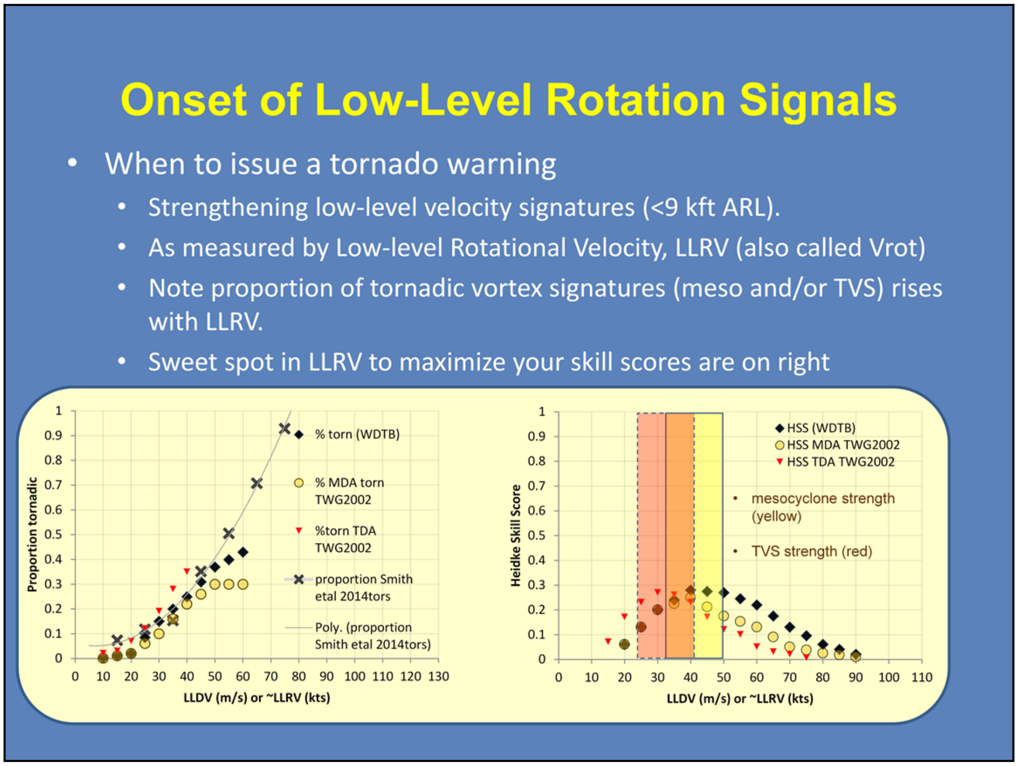

As low-level rotational velocity (LLRV) increases in any of the scans below 9 kft ARL, be aware of how well the mesocyclone/TVS/TS strength compares with the probability of tornado and the best skill in discriminating tornadic vs nontornadic stormscale vortex signatures. Several studies show similar themes. The stronger the LLRV, the more likely the vortex is tornadic. But the more important information is that there's a sweet spot where the Probability of Detection (POD), False Alarm Ration (FAR), and correct nulls point to an apex of overall skill in discriminating tornadic from nontornadic signatures. If it's the mesocyclone you're evaluating, a LLRV from 35-50 kts points to the best skill. For TVS detection, the LLRV is a bit lower, 25-40 knots. Why lower? It's likely because the velocity peaks in a more confined space of a TVS is more likely to miss higher velocity values elsewhere in the larger mesocyclone. The key point is that waiting for a vortex signature to increase beyond this sweet spot invites more missed events than you want. Likewise, issuing tornado warnings on vortices with weaker than optimal LLRVs invites too many false alarms.

A disclaimer here is that this sweet spot may move around depending on many factors including vortex diameter, distance from radar, and near storm environment. Distant or small circulations means a lower LLRV for the sweet spot in best skill. Also distant circulations are sampled higher in the storm and thus a not so favorable environment may disassociate the relationship between the strength of the elevated circulation with that at lower levels where it counts.

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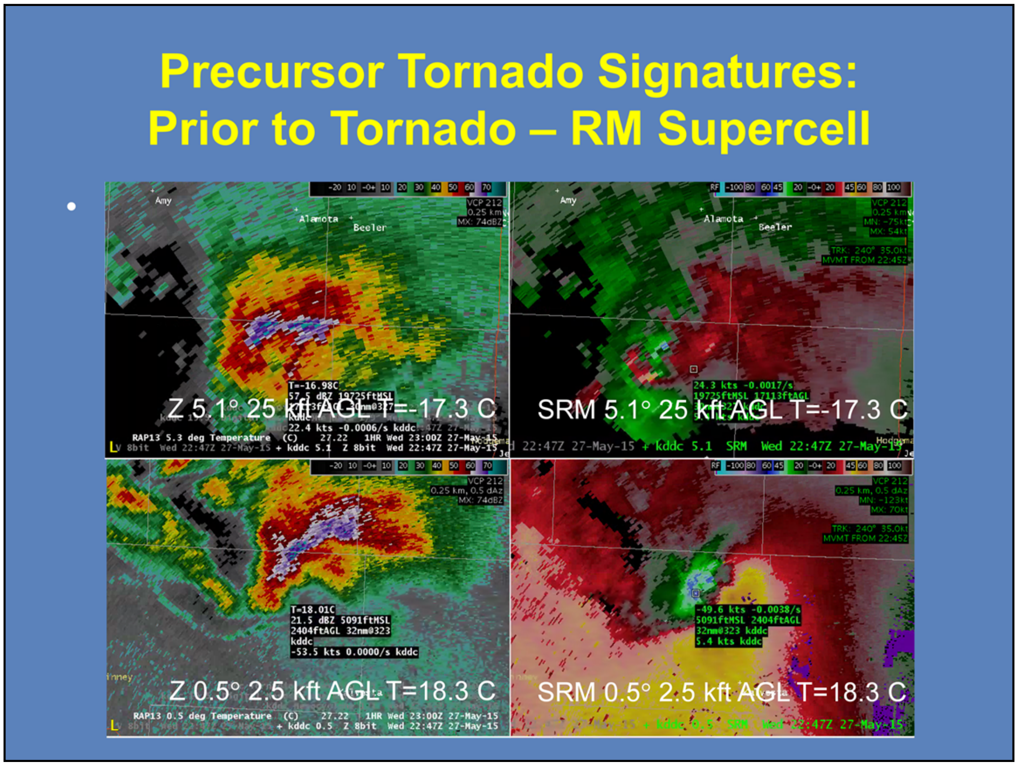
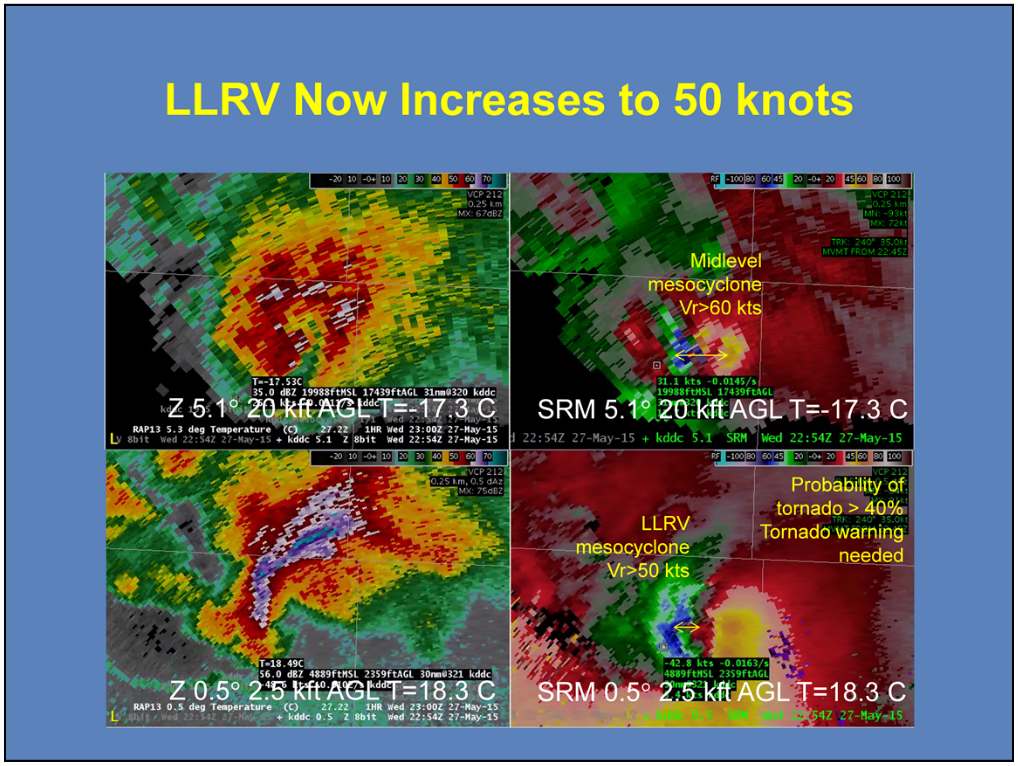


As a test case, let's go back to the same example storm that we know has strong updraft signatures. At this time, 2247 UTC, the storm has a midlevel mesocyclone that appears somewhat complicated by multiple centers. At lower levels the velocity field is dominated by strong convergence. Often that strong convergence can turn into tornadic rotation inside of a few minutes when it's located under the updraft signatures that this storm exhibits.



Five minutes later, the LLRV increases to 50 kts. The gates used in the calculation of LLRV appear at either end of the double ended arrow. A 50 knot LLRV equates to a near 40% probability that it's associated with a tornado. Given all the considerations, updraft strength, environment, deep, strong mesocyclone, this storm probably deserves a tornado warning.

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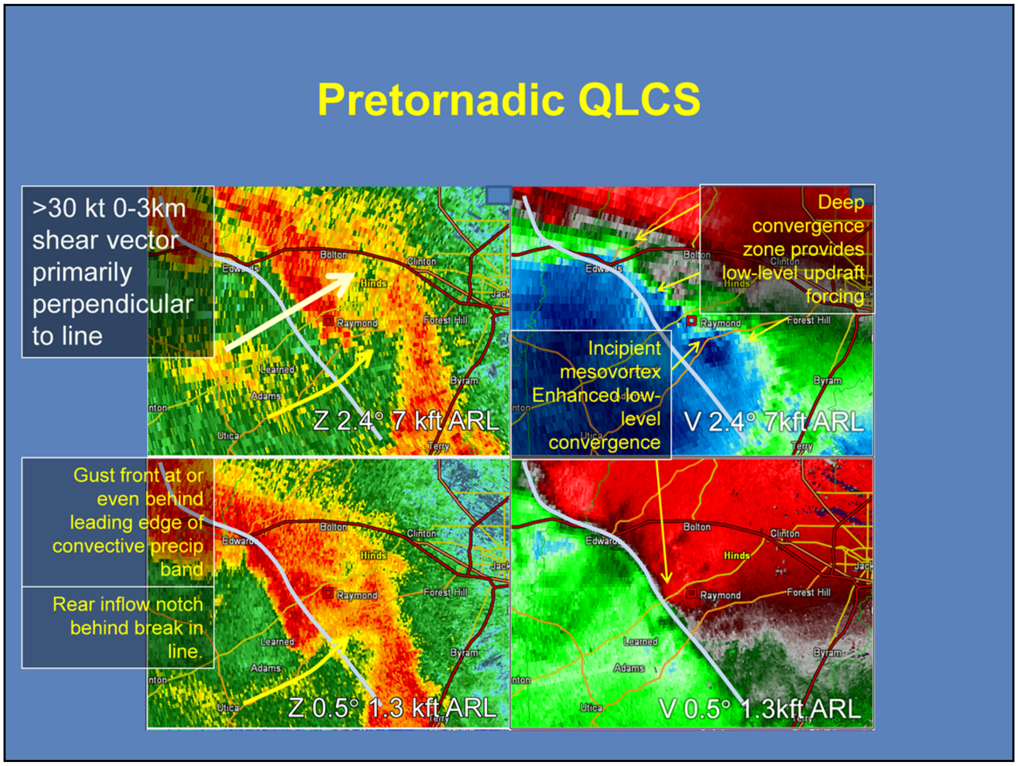

In a Quasi-Linear Convective System (QLCS), the tornado considerations are mostly the same except for one consideration. The 0-3 km shear should have a strong component orthogonal to the line orientation, as this one does. Otherwise, updraft signatures appear most strongly in the lowest 3 km with a deep convergence zone that's nearly vertical, even tilted forward because of storm motion. The gust front should be within or on the leading edge of the high reflectivities associated with convective precipitation. This means low-level convergence is directly underneath deep updraft. The quasi part of the QLCS refers to along-line variations such as rear- and front-inflow notches. If paired together, look out for mesovortex genesis. This case has a rear inflow notch associated with a break in the line.

With the favorable environment for tornadoes, mesovortices can quickly develop, often from the lowest scan, and then deepening with time. The lead time to tornadogenesis is consequently less than with supercells. This is a good time to have a tornado warning drafted up and ready to go.

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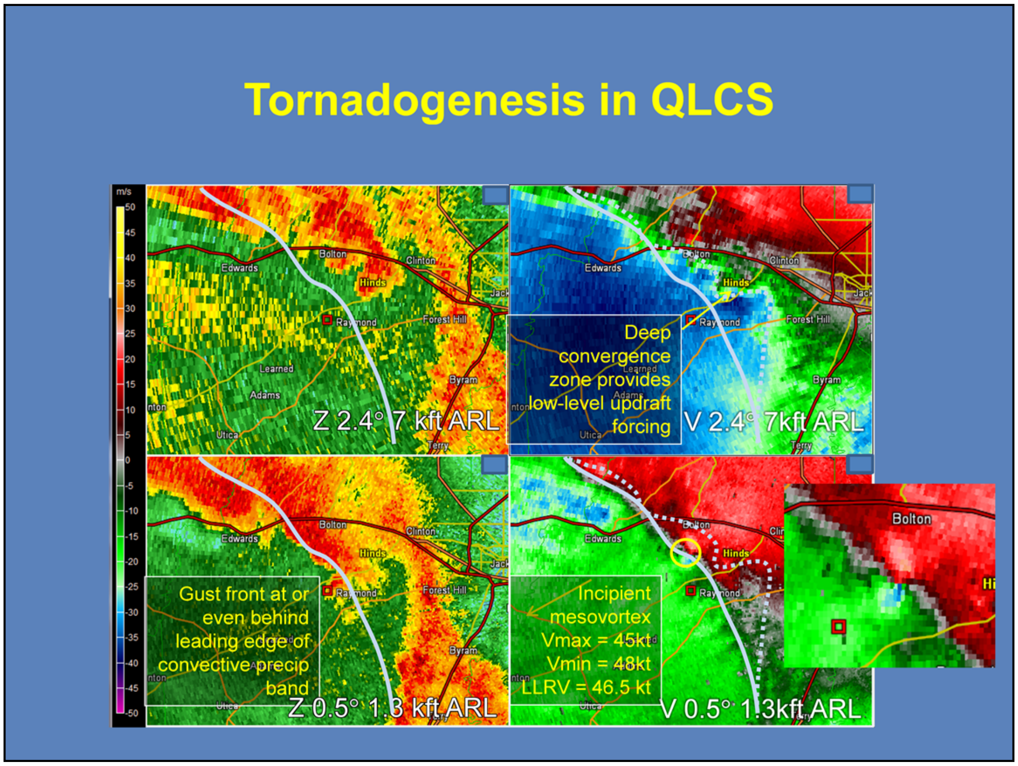

The next scan shows that one mesovortex intensified with strong velocities on both sides of the gust front. A notch of inflection in the gust front points to further indication of potential for this to become tornadic. The mesovortex (mesocyclone) circled has strengthened, with a LLRV of 46.5 knots, well within the sweet spot for a tornado warning. However, in this case, waiting for the LLRV to reach the sweet spot limited the lead time to only 2 minutes. The previously drafted warning should've been issued. The inset 2 minutes later shows an LLRV of 60 knots as the tornado developed.

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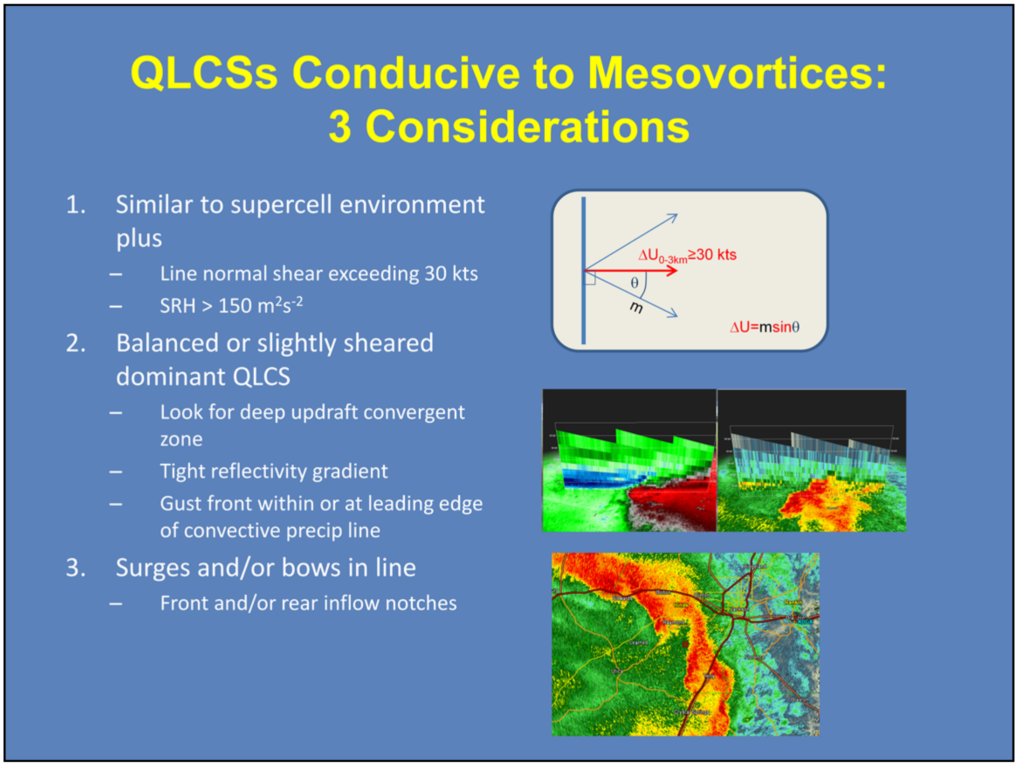

Tornadic QLCSs are preferred when the environment features strong deep layer shear, significant low-level storm-relative helicity, and line normal 0-3km shear greater than or equal to 30 knots.

Second, the QLCS should exhibit a balanced or slightly shear dominant appearance meaning that the gust front is deep, perhaps even leaning forward a bit, and is on the leading edge of the convective precipitation or embedded within (definitely not racing ahead of) the precipitation.

Third, surges and bows in the line featuring rear and/or front inflow notches is a common feature to keep in mind.

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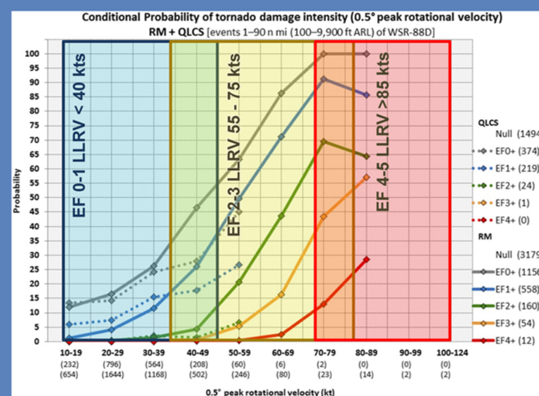
If a Tornado is Occurring, How Strong is It for Supercells?

- Using LLRV (also known as V_{rot} 0.5°)
  - Compare with graph on the right
- For supercells
  - EF0-1 – in blue
  - EF2-3 – in yellow
  - EF4-5 – in red
- Constraints
  - Beam elevation < 7 kft ARL
  - < ~70 nm
  - Pick highest value of 3 consecutive volume scans
  - Good velocity data



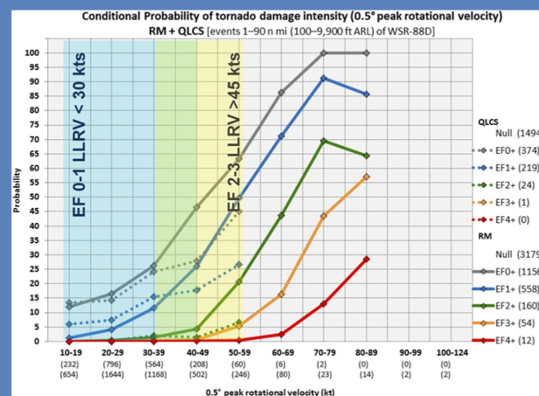
Thompson, R.L., B.T. Smith, J.S. Grams, A.R. Dean, J.C. Picca, A.E. Cohen, E.M. Leitman, A.M. Gleason, and P.T. Marsh, 2017: Tornado Damage Rating Probabilities Derived from WSR-88D Data. *Wea. Forecasting*, **32**, 1509–1528, https://doi.org/10.1175/WAF-D-17-0004.1

Now let's transition to what you need to look for to support your warning decision making with a confirmed tornado. Much of your messaging depends on your best estimate of tornado intensity and fortunately a mature suite of research is available from which you may draw. Compiled here to the most fundamental guidance is a relationship between LLRV and the probability of tornado intensity by EF-rating for supercells, especially right-moving ones (labeled RM). The strategy for coming up with an estimate is to ensure the WSR-88D is in SAILS mode so that you can choose the maximum value in the last 3 scans which would be within last 5 minutes of the current time. If the maximum LLRV falls lower than 40 knots, then expect a weak tornado. For LLRVs between 55 – 75 kts, expect an EF2-3, and for LLRVs greater than 85kts, expect a violent (EF4-5) tornado to be occurring at the time of observation.

Naturally there are significant overlaps where uncertainty is high. To minimize your uncertainty and potential for error, the tornado needs to be the main one in a mesocyclone, sampled by the nearest WSR-88D at altitudes lower than 10 kft above the surface (that is, less than 70 nm from the radar). Random sampling errors require that a few samples be taken and you take the maximum. Also, the velocity data must be good, or in other words, make sure there is no range folding, side lobe errors, or three body scatter spike (TBSS) interference. And there should be strong signal-to-noise ratio.

If a Tornado is Occurring, How Strong is It for QLCSs?

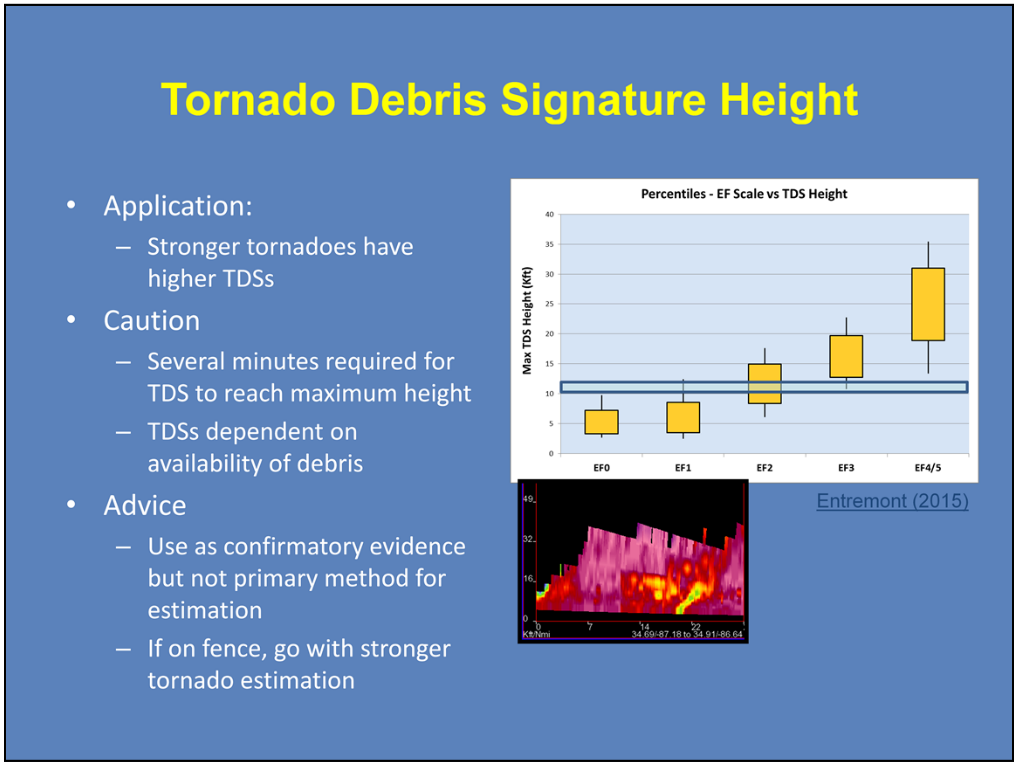
- Using LLRV (also known as V_{rot} 0.5°)
  - Pick highest value of 3 consecutive low-level scans
  - Compare with graph on the right
- For QLCSs shift down a bit
  - EF0-1 – in blue
  - EF2-3 – in yellow
- Constraints
  - Beam elevation < 7 kft ARL – suggest lower
  - < ~70 nm – suggest closer
  - Good velocity data



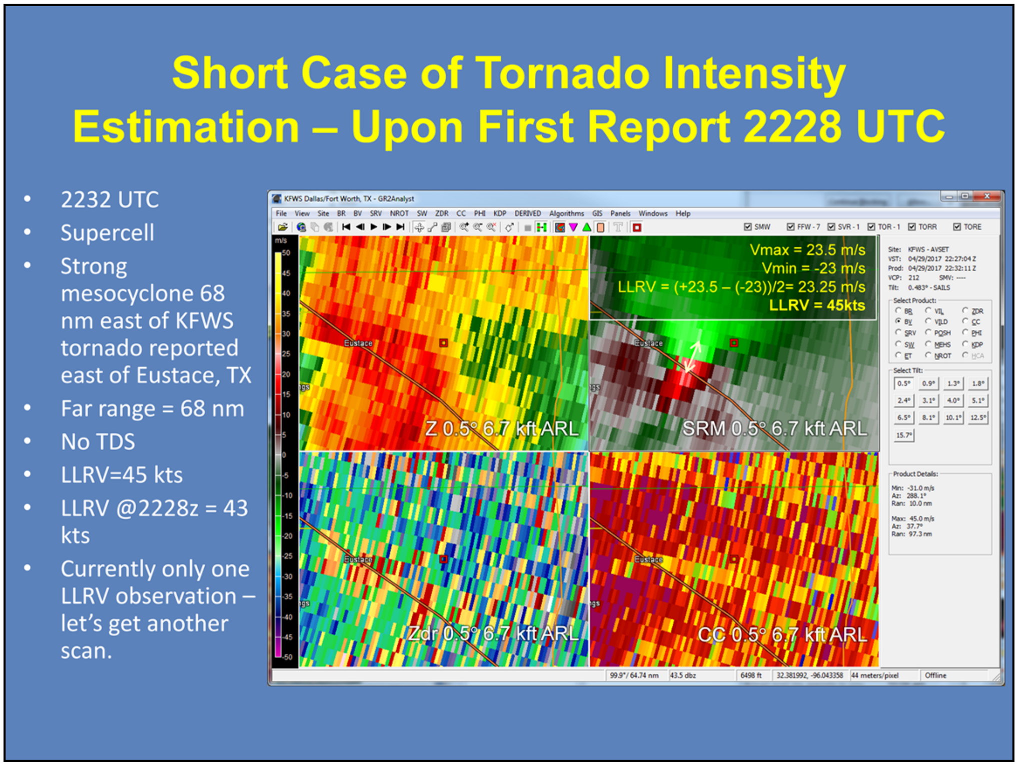
Thompson, R.L., B.T. Smith, J.S. Grams, A.R. Dean, J.C. Picca, A.E. Cohen, E.M. Leitman, A.M. Gleason, and P.T. Marsh, 2017: Tornado Damage Rating Probabilities Derived from WSR-88D Data. *Wea. Forecasting,* **32**, 1509–1528, https://doi.org/10.1175/WAF-D-17-0004.1

The tornado intensity to LLRV relationship is a bit weaker for QLCS tornadoes because of the more limited sample size and the shallower nature of the vortex signatures available to view. But similar to that of supercells, pick the highest LLRV of three low-level scans and then compare to the graph at the right. Weak tornadoes are often associated with LLRVs < 30 knots while strong ones are mostly above 45 knots. The sample in the work presented here does not have any violent tornadoes from QLCSs.

Similar to supercells, and perhaps even more important for QLCSs, make sure the event is close to a WSR-88D. And also make sure the velocity data is of sufficient quality.



Tornado debris signatures (TDS)s also have been shown to have a relationship between their maximum height and the reported tornado EF-rating. Stronger tornadoes have higher TDSs where TDSs are visible. The caution here is that TDSs grow with time and may take too long to reach a maximum altitude for effective warning communication. TDSs may not appear where tornadoes cross the surface lacking in debris sources such as sparsely vegetated terrain or water bodies. However they do provide confirmatory evidence to that of the velocity-based method to estimate tornado intensity. Also, if the LLRV straddles a zone of uncertainty between weak and strong tornadoes, the presence of a TDS has been shown to support raising the tornado intensity estimate.



Here's a short case on estimating tornado intensity from an event where a tornado report was received by an office at 2228 UTC. We are now at 2232 UTC and need to make an estimate of tornado intensity. This storm happens to be a supercell with a strong mesocyclone signature 68 nm east of the KFWS radar. This is relatively long range and so the estimate will be more uncertain than if the target is closer. But we will make an estimate nonetheless. The CC product shows no detectable TDS yet. At 2232 UTC the Vmax and Vmin have been sampled at the tips of the arrows indicated in the velocity panel yielding a LLRV of 45 kts. A LLRV was also sampled at 43 kts in the first scan of the tornado. Let's wait to get one more.

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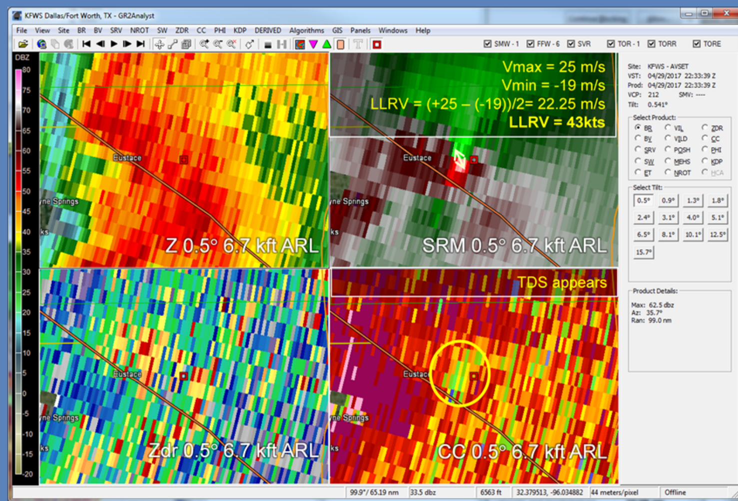
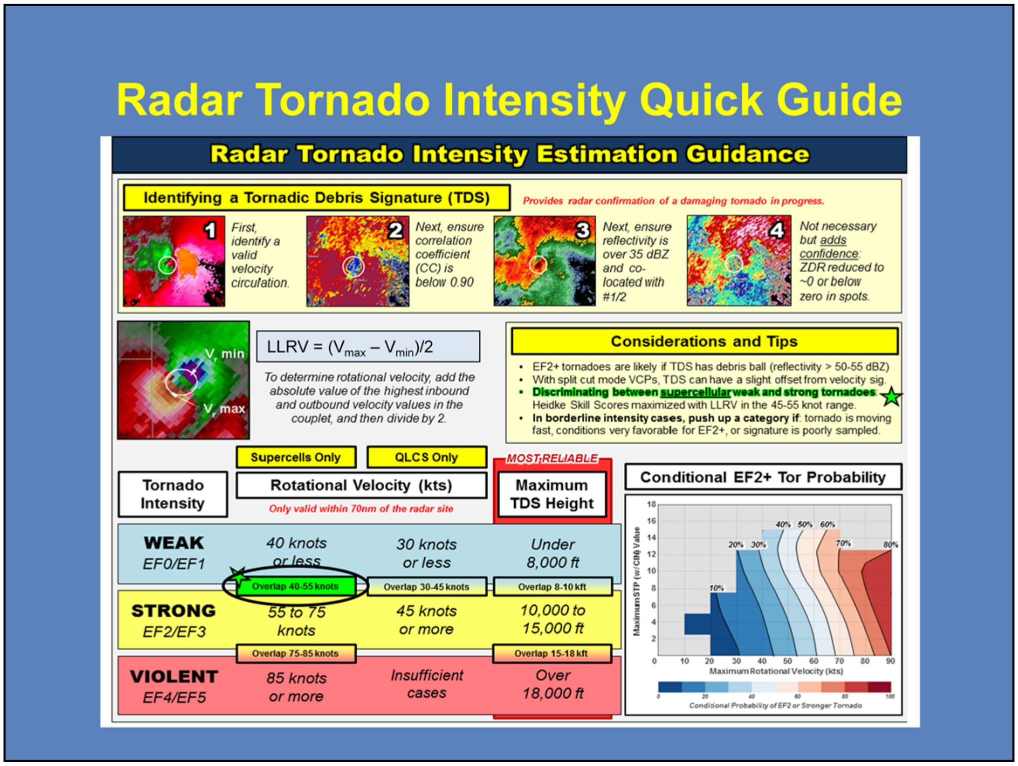
Short Case of Tornado Intensity Estimation

- LLRV @2228=43kts
- LLRV @2230=45kts
- LLRV @2233=43kts
- TDS appears
- now we will use LLRV=45 kts



Now at 2233, another LLRV sample yields a value of 43 kts. So over the past five minutes the maximum LLRV was 45 kts. We also note a significant CC reduction occurred in the vicinity of the vortex signature in an area with high reflectivity and thus we're confident a TDS appeared.



This guidance chart is made available to provide a quick reference assigning a tornado intensity to LLRV with a given storm type. Since we've got a supercell we associate a LLRV of 45 kts to tornado intensity and note we fall into one of those overlaps where we could have a weak or strong tornado. At this time, the TDS just appeared at a scan 6600 ft ARL and so this suggests weak tornado too. However, Thompson and co-authors also noted that if on the fence in the presence of a TDS, it may be a good idea to bump up the estimate. Now in the lower right we also notice that if the LLRV is on the fence, a strongly favorable environment, measured by the effective-layer Significant Tornado Parameter (STP), may also bump up the odds of increasing the categorical estimate.

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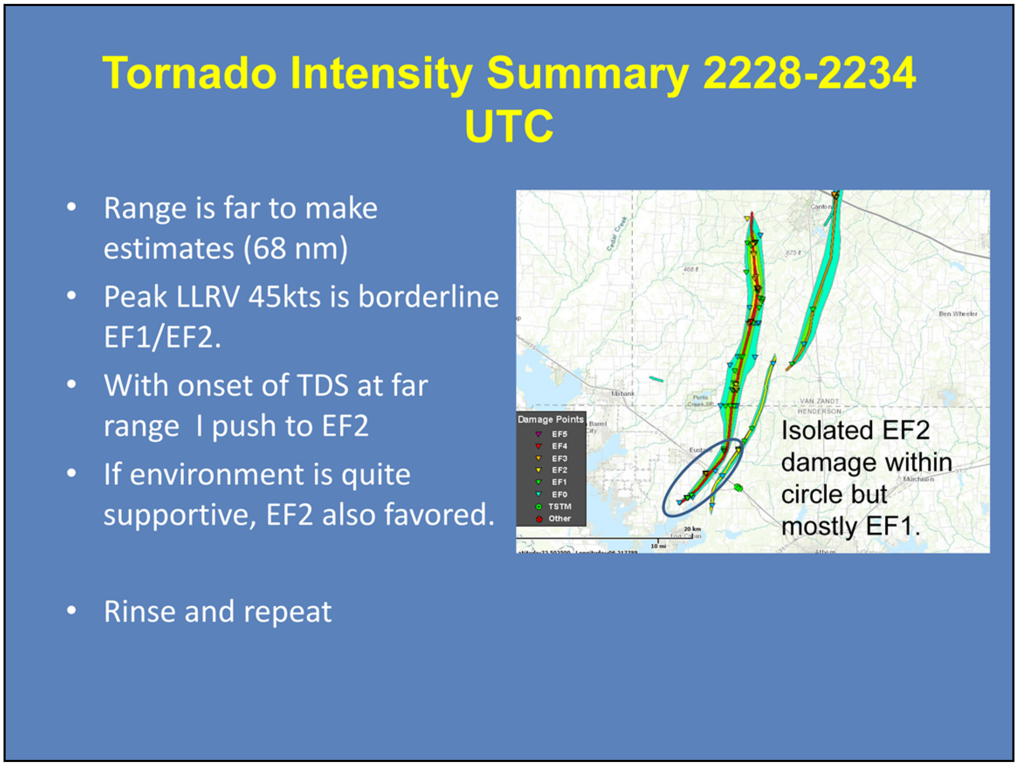


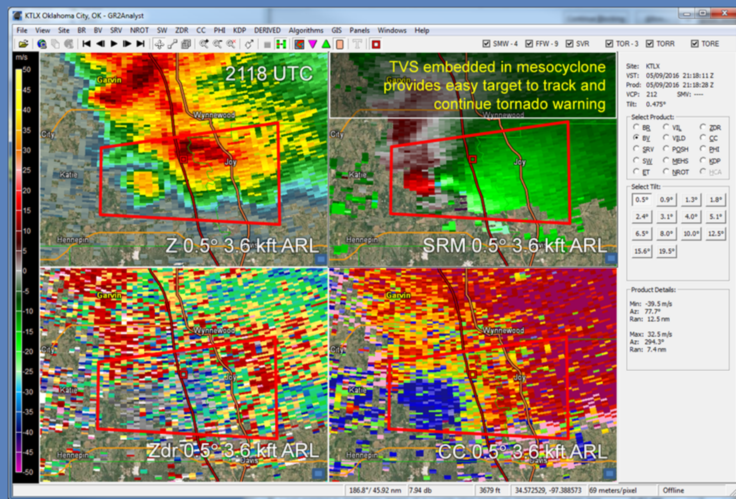
To summarize, the range is at the upper end of where skillful estimates of tornado intensity can be made. Since the velocity data looked good, the vortex signature clearly defined, an estimate is possible. We yield a borderline EF1/EF2 tornado and its possible to consider a reasonably large possibility of an EF2 tornado. Note that a strongly favorable environment may also support a stronger tornado.

As it turned out we sampled the beginning stage of the tornado within the ellipse overlaid on the damage survey. Within the ellipse most of the damage along the track centerline was EF1, but there was an isolated pocket of EF2 damage.

Now this was only a nowcast. Not much guidance exists in making a forecast and thus frequent updates to the warning are necessary as the tornado evolves quite quickly.

When to End a Tornado Warning? : Scenario 1, Part 1

- TVS/mesocyclone is tornadic



The final stage of a tornado warning is to determine when a tornado ends. Let's provide a case here to illustrate the considerations that must be in play in the last phase of a tornado warning, it's cancellation.

At this time we have a tornadic TVS embedded in a larger mesocyclone and thus the warning here is still valid.

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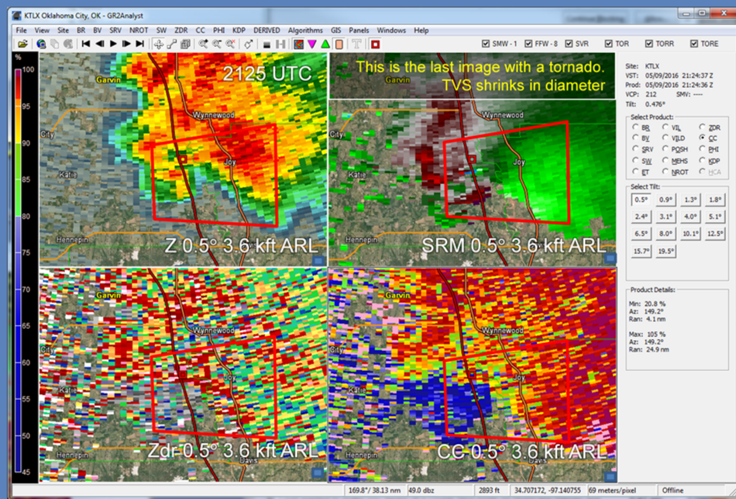
When to End a Tornado Warning?: Scenario 1, Part 2

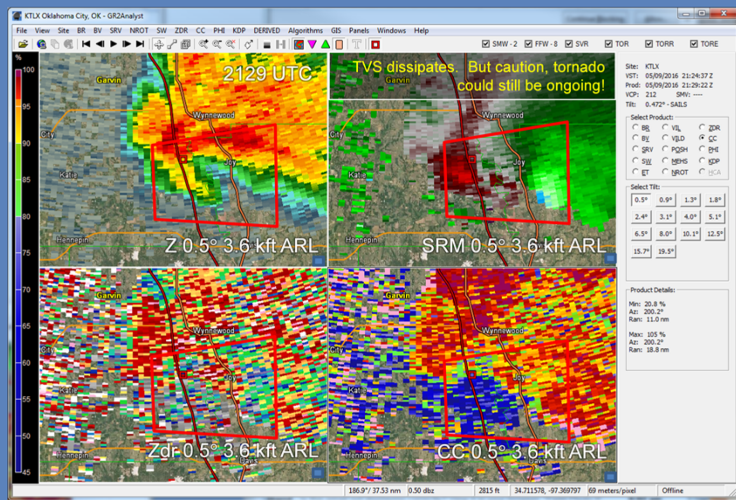
- TVS contracts in diameter, still tornadic
- Tornado warning continues



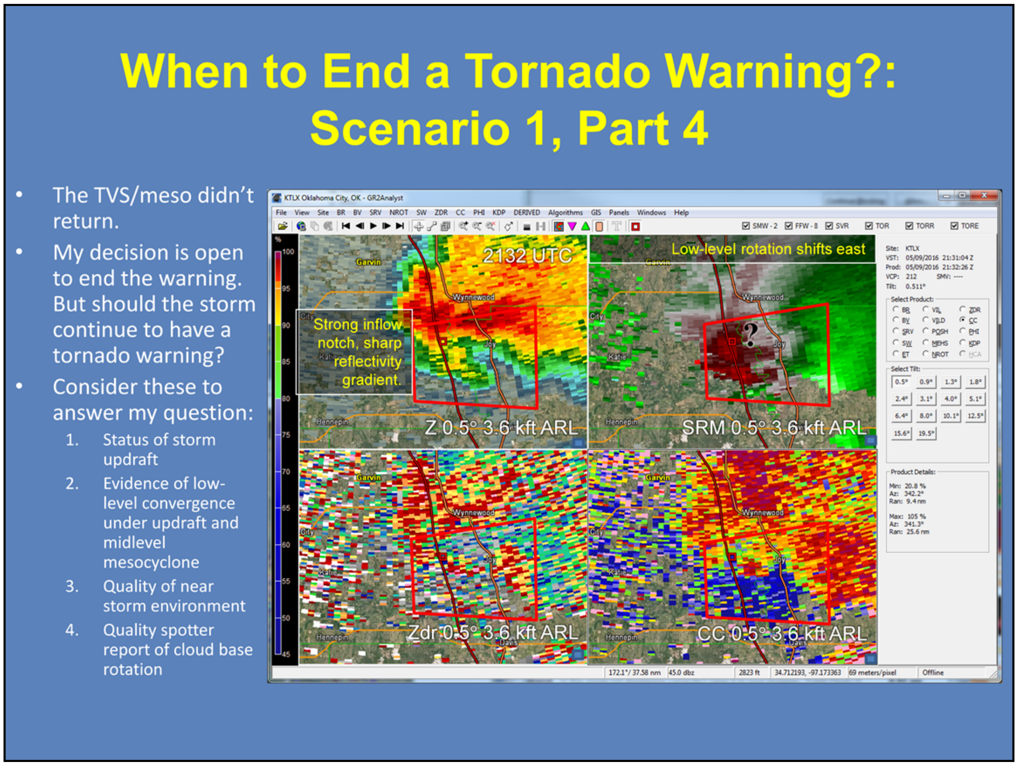
Seven minutes later the TVS contracts in size, a common evolution during the occlusion phase of a tornado. However, since the tornado continues accompanied by a TVS, the warning continues as well, though perhaps with its back end trimmed in a recent Severe Weather Statement.

When to End a Tornado Warning?: Scenario 1, Part 3

- TVS dissipates but tornado could still be ongoing!
- Tornado may be too narrow for radar to resolve.
- I wait several minutes to cover continued tornado threat and then decide.
- But after several minutes do I?



Four minutes later the TVS dissipated. But remember, a TVS doesn't mean the tornado dissipated. It only means the tornado may be too narrow for the radar to resolve. If I have no spotter information confirming tornado dissipation, I will wait several minutes to cover a continued tornado threat and then decide.



Three more minutes pass and there is no return of the TVS. Now I have a decision to make and one option is to end the warning. But a tornado warning is not a nowcast, it's also a forecast. Do I expect the parent storm to still have sufficient tornado potential to warrant a warning? To answer this question, consider the following: 1) status of the parent storm updraft, 2) evidence of low-level convergence and midlevel mesocyclone, 3) the quality of the near storm environment, and 4) any spotter reports of ongoing rotation.

From the lowest scan we still see a strong inflow notch, sharp reflectivity gradient, and we see low-level rotation shifting east.

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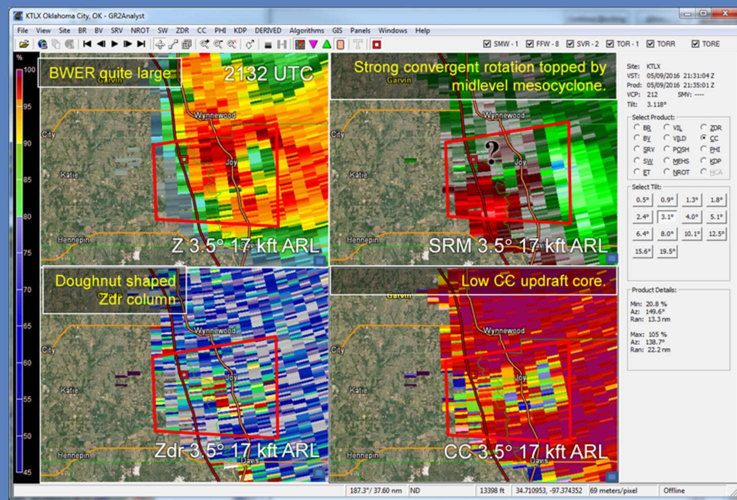
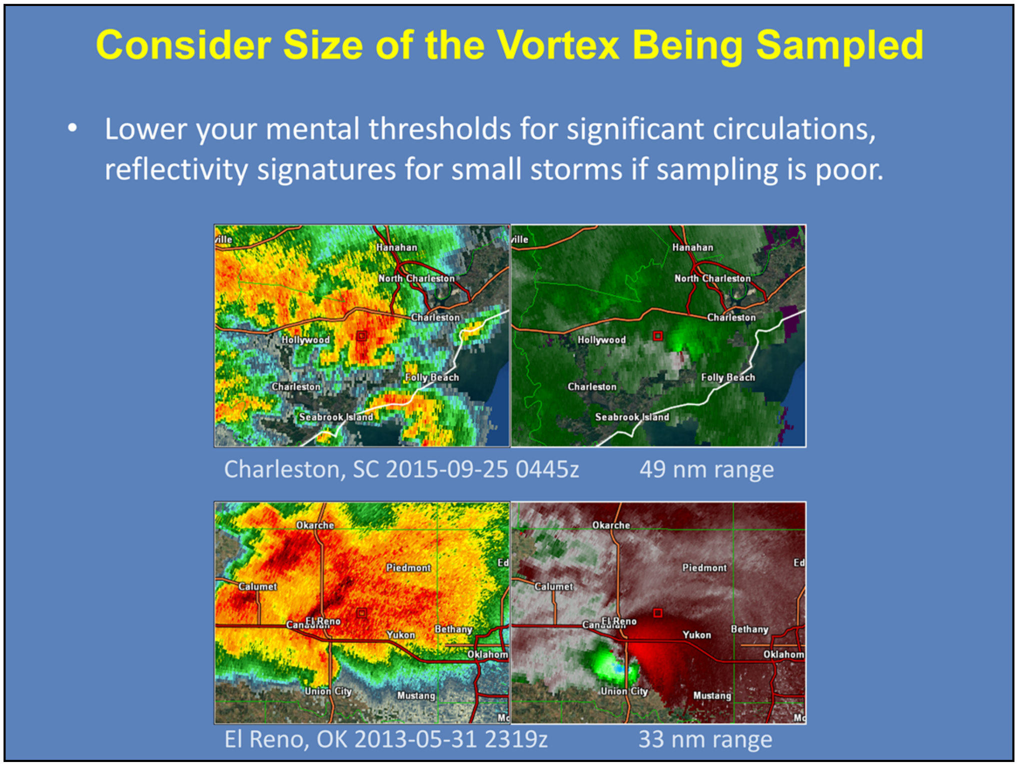
When to End a Tornado Warning?: Scenario 1, Part 4 (Con'td)

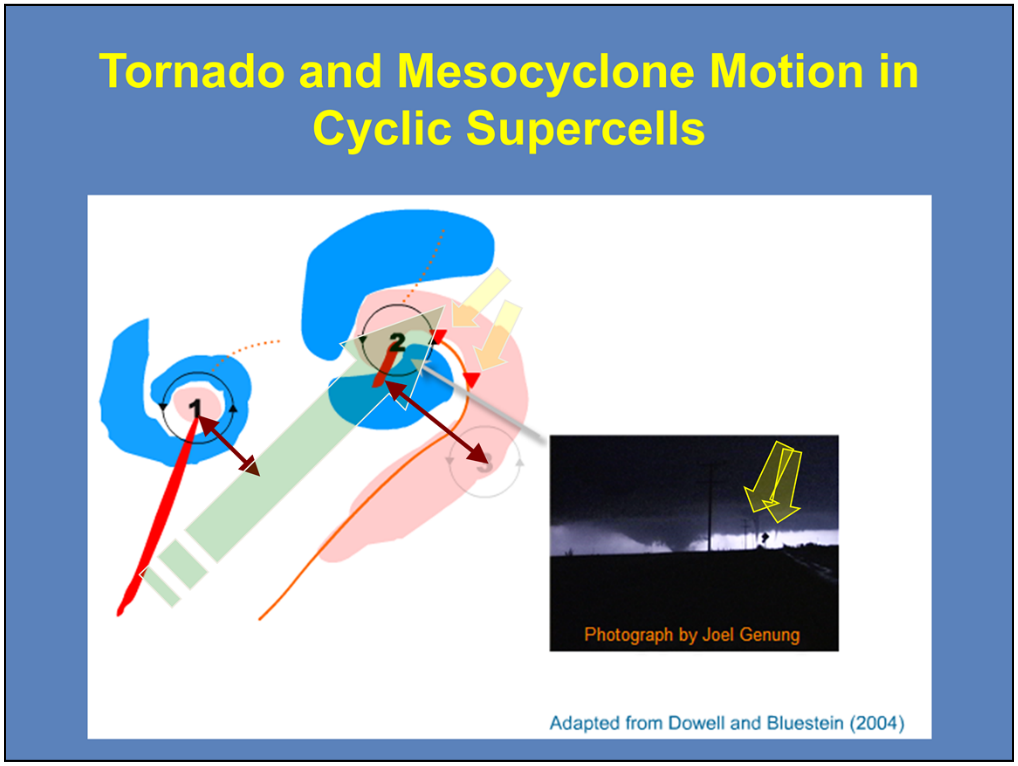
- Consider these to answer my question:
  1. Status of storm updraft
  2. Evidence of low-level convergence under updraft
  3. Quality of near storm environment
- A tornado warning should continue, whether to keep this running or to issue a new one.



Upstairs we see a large BWER, strong midlevel mesocyclone, a doughnut-shaped Zdr column, and a low CC updraft signature. All of these suggest that the updraft is healthy and the supercell has a strong mesocyclone overlaying low-level convergence and rotation. Now let's assume the environment is still favorable. All three suggest continue current warning or issue a new one.



The most important thing to remember is that ability for the WSR-88D to adequately sample a vortex is dependent on the ratio between the size of the vortex and the effective beamwidth. This means a well-sampled mini vortex is going to give you better details than a poorly sampled maxi supercell mesocyclone. Your thresholds for issuing a tornado should change accordingly. Lower them for poorly sampled storms and vortices, raise them when they're well sampled, given all else being equal. See the tornado signatures lesson on more details.



Account for differing tornado motions than storm motions, especially if a supercell is cycling. Recall from the lesson on supercells that older or decaying mesocyclones in cyclic supercells have significant motion to the left of the actual storm motion and newer mesocyclone motion. An old mesocyclone with a tornado can move up to 10 miles left of the main supercell path. Cyclic tornadogenesis appears to occur after the RFD to the right of a mature mesocyclone surges forward enhancing convergence at its leading edge of the cyclic mesocyclone section. The locally enhanced convergence helps initiate a local updraft which then tilts horizontal vorticity into the vertical. In a short time, a new mesocyclone forms on the head of the RFD surge while the old one continues motion to the left. The old mesocyclone often begins to move left of the main supercell either as a result of outflow from the main core pushing it rearward or from inflow locally pushing the RFD gust front backward. Meanwhile, more than one tornado may form on the RFD gust front, but outside the center of the low-level mesocyclone

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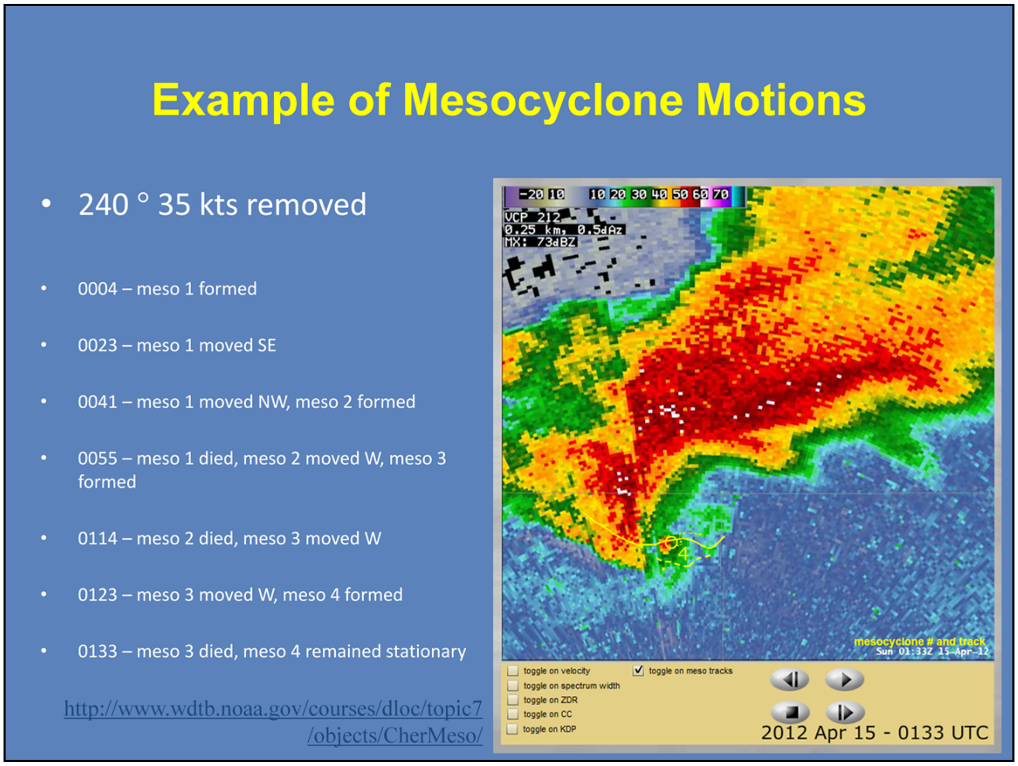

This example is not in the student guide but it clearly shows how differently the individual mesocyclones move relative to the parent storm. A popup display will appear that allows you to explore in more detail what I'm about to describe here. The link is a backup to the popup. We'll take a tour of the Cherokee storm from KVNX as it moved northeast at 35 kts. I used the feature-following zoom to remove the parent storm motion from the loop. In the discussion I refer to these circulations as mesocyclones. They are low-level mesos that sometimes appear as TVSs. I refer to their motions in a storm-relative sense. The storm starts far southwest of the radar (35 mi) and passes within 10 mi west of the radar before retreating to the northeast at 25 mi at the end of the loop.

At 0004 UTC , mesocyclone 1 formed looking like a TVS. It moved to the southeast by 0023 UTC. A Dual-pol TDS developed by 0037 UTC.

At 0041 UTC, meso 1 backtracked and moved northwest as the Dual-pol TDS indicated a tornado. Meso 2 developed. Meso 2 was well centered under the parent storm updraft and produced a Dual-pol TDS by 0051 UTC as well as a strong TVS.

At 0055 UTC, meso 1 died while meso 2 was a strong TVS with a Dual-pol TDS while it moved more quickly to the west. The tornado with meso 2 was the one labeled circulation #1 in the lesson on Tornado scale signatures. Meso 3 formed well east of #2 and on an eastward extension of the RFD.

At 0114 UTC, meso 2 died while meso 3 also moved westward while quickly producing a TS and a Dual-pol TDS. The TDS was quite strong at this time.

At 0123 UTC, meso 3 moved quickly westward and is on the backside of the reflectivity envelope. The TS contracted into a TVS and the Dual-pol TDS continued. Meso 4 developed at this time.

At 0133 UTC, meso 3 died while meso 4 developed a TVS and the largest, most pronounced Dual-pol TDS of the series. This tornado was large.

Note out of all of this how deviant the tornadoes happened to be relative to the main reflectivity core of the supercell. Some of these tornadoes were up to 6 miles west the main track. This happened because the storm-relative inflow overpowered the relatively weak RFD surges sending the low-level mesos westward. However, the meso 4 remained under the parent storm for awhile likely because its RFD surge was stronger and well balanced with the inflow. The bigger tornadoes often accompany mesocyclones that can keep this balance.

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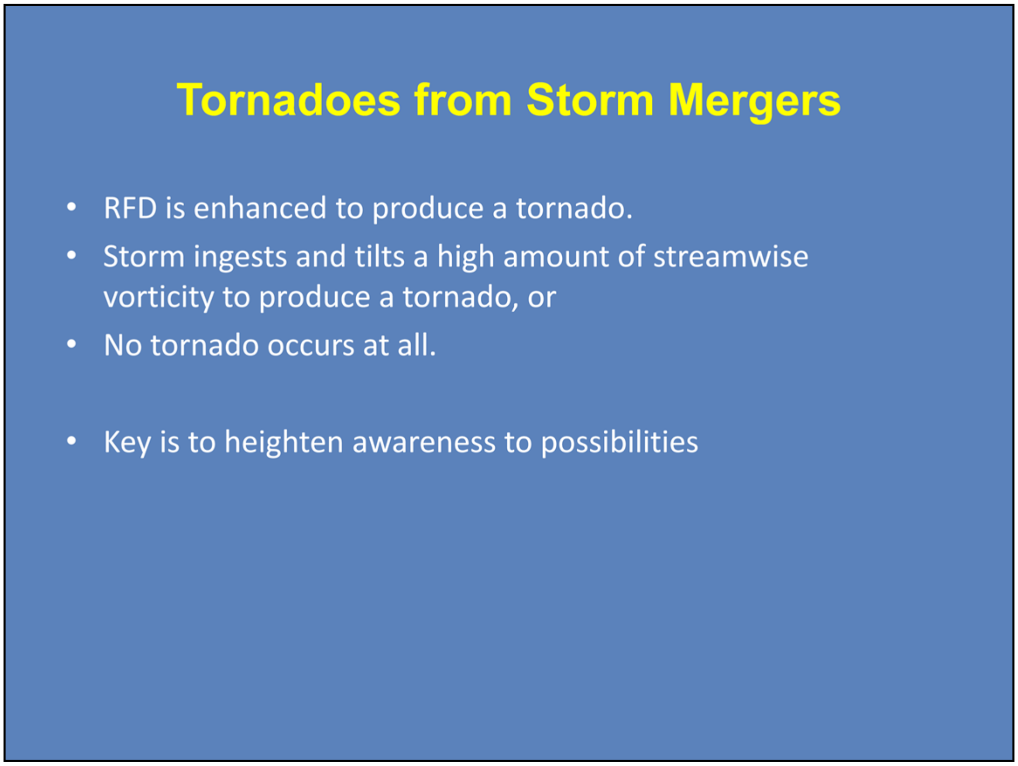

Be aware that mergers of a non-tornadic supercell with a gust front increases the possibility of a tornado. But a wide range of possibilities have been found to occur from this type of merger. Precisely what leads to tornadogenesis or failure is poorly understood. A non-tornadic supercell interacting with a neighboring storm may result in many evolutions:

RFD is enhanced to produce a tornado

Storm ingests and tilts a high amount of streamwise vorticity to produce a tornado, or

No tornado occurs at all.

Additionally, collisions between left- and right-moving supercells may or may not assist tornadogenesis. **The key lies in attaining a heightened awareness of the possibility of a tornado when you see an impending storm merge with another storm or gust front.**

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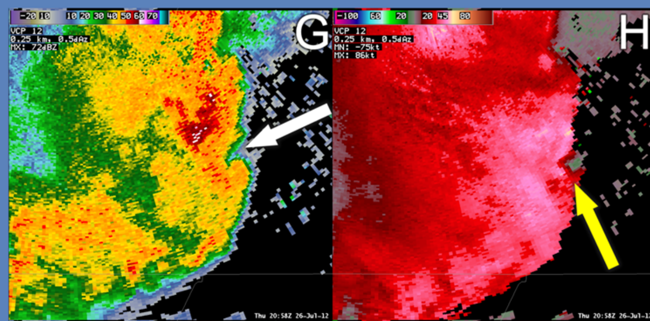
Squall Line Tornadoes



Because both bow echoes and supercells require strong vertical wind shear, supercells and severe bow echoes often occur in close proximity to one another, or evolve from one of these structures to the other during their lifetime. **Environments of bow echo tornadoes and supercell tornadoes are hard to distinguish.** Thus, it is important to examine storm structure of each individual cell within multicell structures, as supercell tendencies are frequently observed with well-organized multicell systems (i.e., those which develop in either sufficient deep shear and/or large CAPEs).

Well-defined front inflow notches often show up in reflectivity data to the north of a surging area of outflow prior to tornadogenesis. **In these cases, the vertical vorticity is enhanced by strong surface convergence and is at a maximum at low-levels. Thus, tornadoes in squall lines typically form much quicker (mean lead time of 5 minutes in the Trapp study) than with isolated supercells.**

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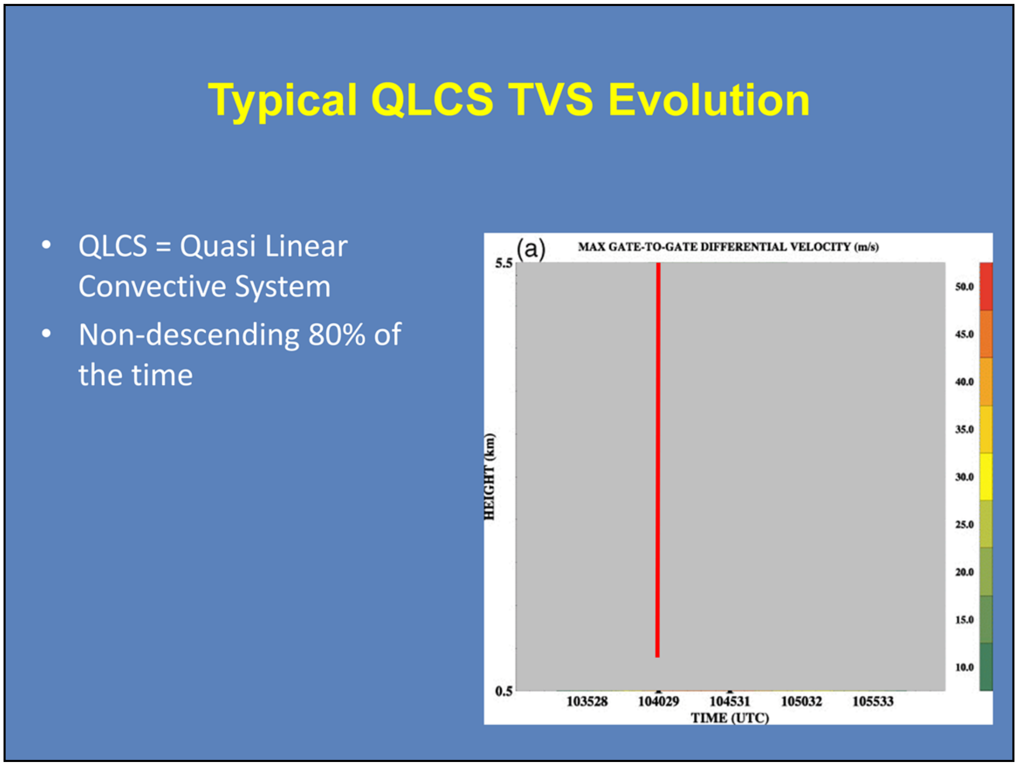
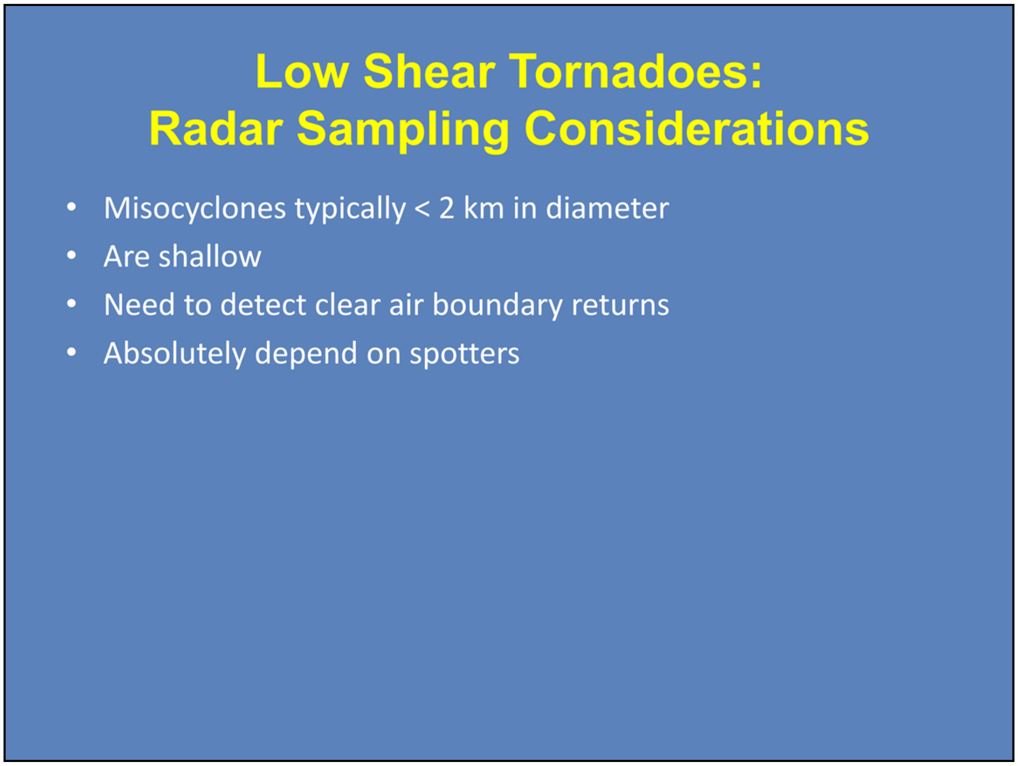


Tornadoes within quasi-linear convective systems tend to be associated with tornadic vortex signatures (TVS) that form from low-levels upward as opposed to some classic supercells which have midlevel circulations first and then build downward with time (Trapp et al, 1999). This non-descending paradigm for TVS evolution



Low shear, or nonmesocyclonic tornadoes have their own challenges in tornado warning considerations.

Because the origin of the rotation in these features is close to the ground, radar may easily overshoot the circulation unless it is within 50 km. The diameter and lifetime of these misocyclones are small (<2 km, 5-15 min) also limiting the ability of radar to resolve their velocity structures. This process creates a difficult job for Doppler Radar to provide adequate lead time of pulse storm tornadoes. Spotter reports of funnels in these situations should be taken very seriously (Lemon and Quoetone, 1995).

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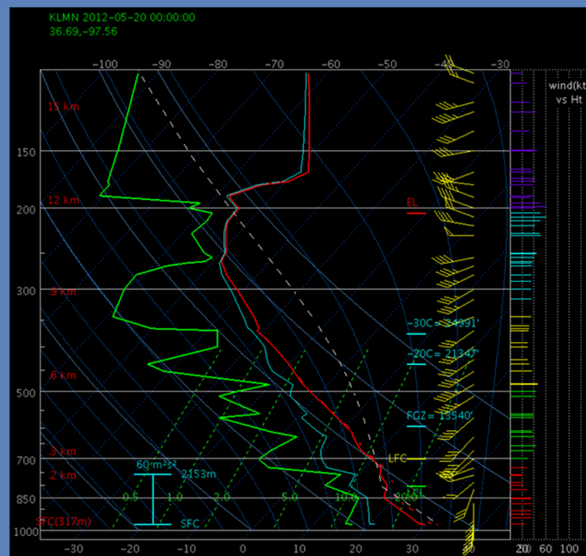
Low Shear Tornadoes: Sounding Considerations

- Steep lapse rates
- No CIN



Favorable conditions leading to the formation of a pulse storm tornado include:
An environment with steep lapse rates, strong surface heating and no CIN.

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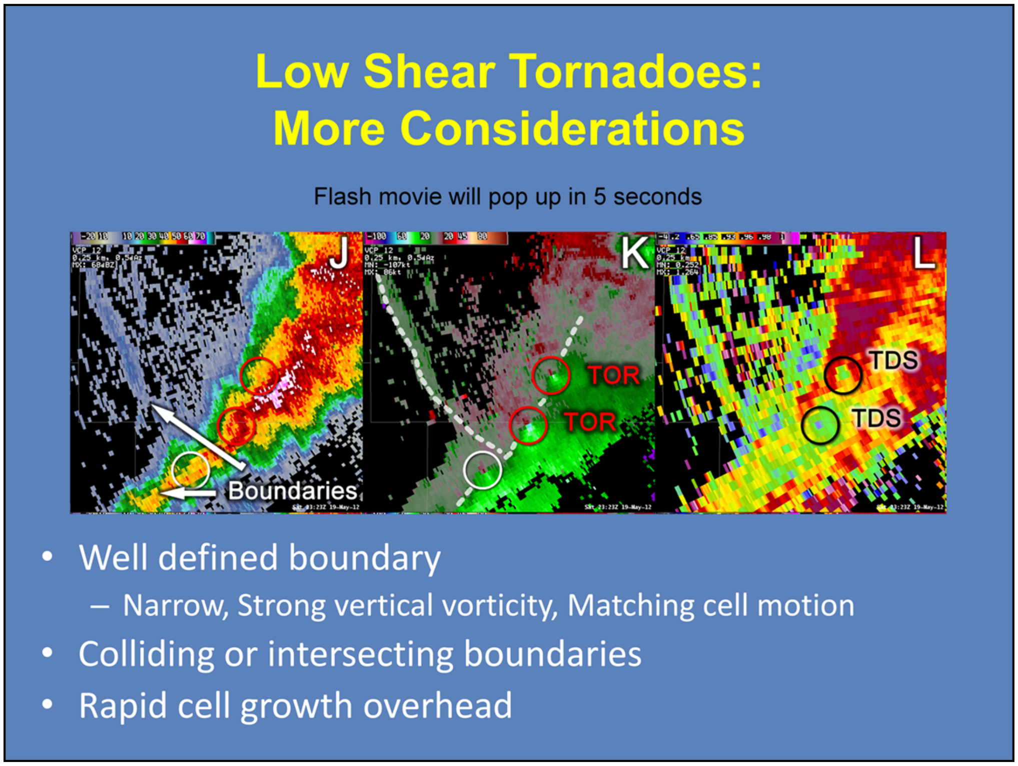
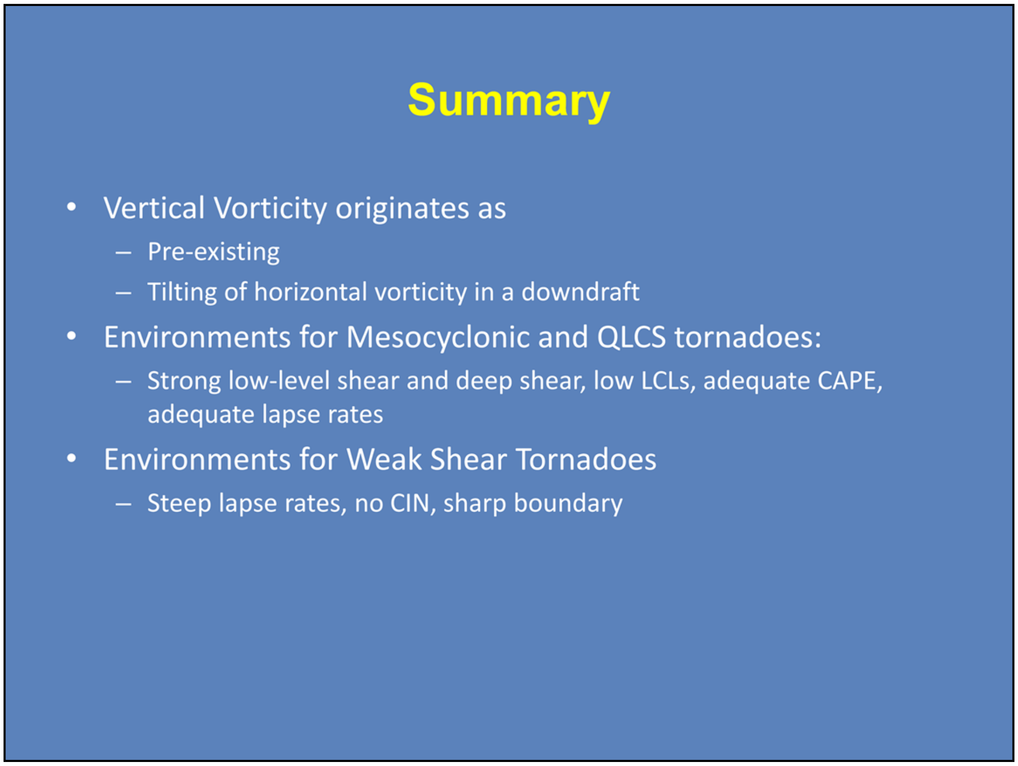


A popup window will display showing a case of a low-shear nonmesocyclonic tornado outbreak.

A well-defined boundary marked by a fine line in reflectivity, velocity discontinuity in base velocity, or a cumulus line visible from satellite. The boundary should have significant vertical vorticity. **Note that 10 m/s of shear across a 1 km wide boundary produces the same vorticity as a moderate mesocyclone (10-2 s-1).** Ideally, the boundary and cell motion should be nearly equivalent. Boundary collisions are also common regions of pulse storm tornadoes. An example of tornadoes forming along a boundary is shown in this image.

Colliding or intersecting boundaries with high potential for vertical vorticity production is an area to be closely monitored (Wakimoto and Wilson, 1989).

over the boundary is a strong clue. Note that many of these tornadoes are produced, at least initially, without nearby precipitation and may be hard to detect. At other times, thin lines close to the radar may attend the boundary. But in either case, visual observations can be critical in raising situational awareness. Be especially alert for tornadoes when the updraft forms an elevated reflectivity core.



Tornadoes derive their vorticity either from pre-existing vertical vorticity, or from tilting of horizontal vorticity into the vertical by a downdraft.

Three types of tornadoes we discussed in this lesson include mesocyclonic tornadoes, Quasi Linear Convective System (QLCS) tornadoes (also known as squall- line tornadoes), and nonmesocyclonic tornadoes.

Mesocyclonic and squall line tornadoes are favored in similar environments of strong low-level shear, strong deep layer shear, low LCLs, and sufficient CAPE. Nonmesocyclonic tornadoes require a sharp boundary with strong vertical vorticity, an uncapped atmosphere featuring significant CAPE, steep low level lapse rates, and a developing updraft riding the boundary.

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Summary (contd)

- Mesocyclonic Tornadoes
  - Strongest Precursor Signatures
    - Updraft, rotation, hook echo and RFD
- QLCS Tornadoes
  - More uncertain precursor signatures
    - TVS often non-descending, inflow notch, localized rear inflow
- Weakly Sheared Tornadoes
  - Most uncertain precursor signatures
    - Pre-existing boundary, developing updraft, shallow misocyclones
- Tornado warning considerations
  - Mesocyclone/TVS/TS LLRV positively related to tornado probability.
  - With ongoing tornado consult LLRV-tornado intensity estimates and make sure there is adequate data quality and vortex signature.
  - Tornado motion differs from parent storm motion. Make sure to track motion of tornado threat as opposed to a reflectivity storm centroid.
  - Keep tornado warning 15 minutes after vortex signature disappears to account for rope-out.

Mesocyclonic tornadoes in conjunction with supercells are easiest to anticipate as they are typically preceded by strong low- and midlevel updraft signatures, strengthening circulation aloft, and an onset of a hook echo and RFD. Confidence of a successful warning rises with the simultaneous presence of multiple signatures including TVS, mesocyclone, and BWER.

Squall line tornadoes typically do not have deep updraft signatures, however low-level updrafts can be very strong. They typically occur on the leading edge of a bowing line segment and may be preceded by the onset of a front inflow notch. TVS signatures likely suddenly appear in a non descending fashion. A parent mesocyclone is not a requirement.

Nonmesocyclonic tornadoes are difficult to detect via radar given the lack of a preceding deep circulation. However, given close enough proximity, radar can infer the presence of low-level circulations. They are more likely to occur as a young updraft phases with a pre-existing low-level circulation.

Recognizing the different types of conceptual models for specific tornadic storm development is crucial in effective tornado warning decision making. A thorough analysis of the environment in which tornadoes occur, in addition to the best possible radar interrogation strategies, must be the prime considerations for an effective tornado warning methodology.

Following are considerations in tornado warnings.

Mesocyclone/TVS LLRV is positively related to the probability of tornado. Consider the guidance charts in this, and other guides, for issuing tornado warning.

Likewise the mesocyclone/TVS/TS LLRV is positively related to tornado intensity given a mesocyclone or QLCS mesovortex is well-resolved and a tornado is in progress.

Tornado motion differs from parent storm motion. Make sure to track motion of tornado threat as opposed to a reflectivity storm centroid.

Keep tornado warning 15 minutes after vortex signature disappears to account for rope-out.

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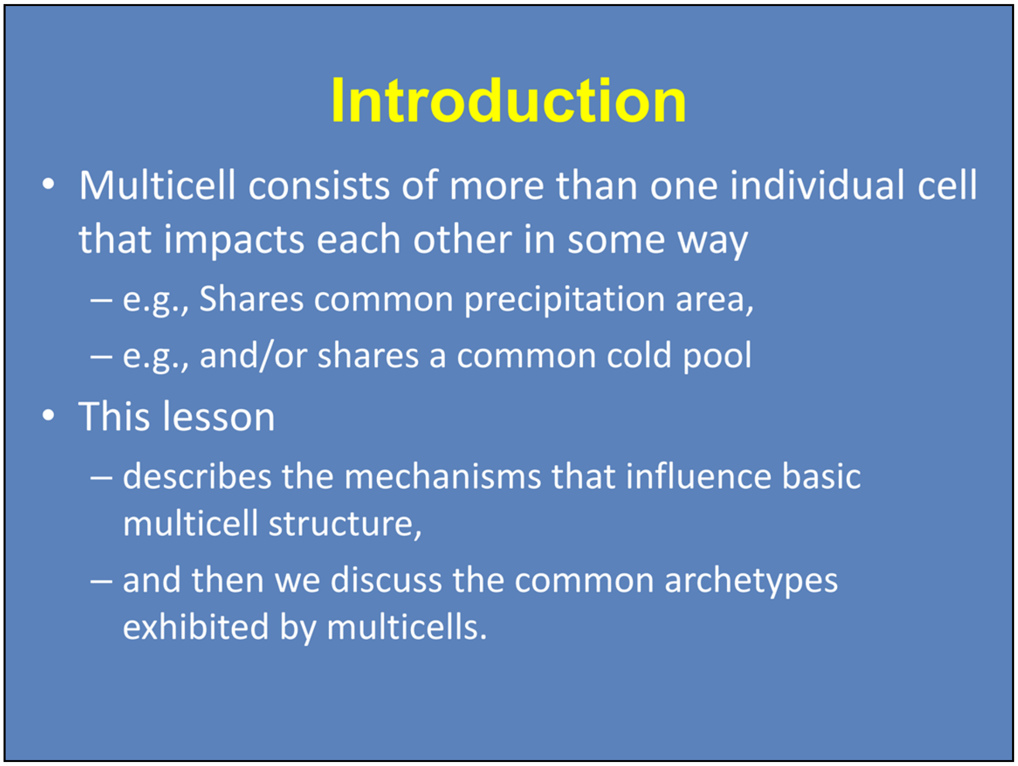
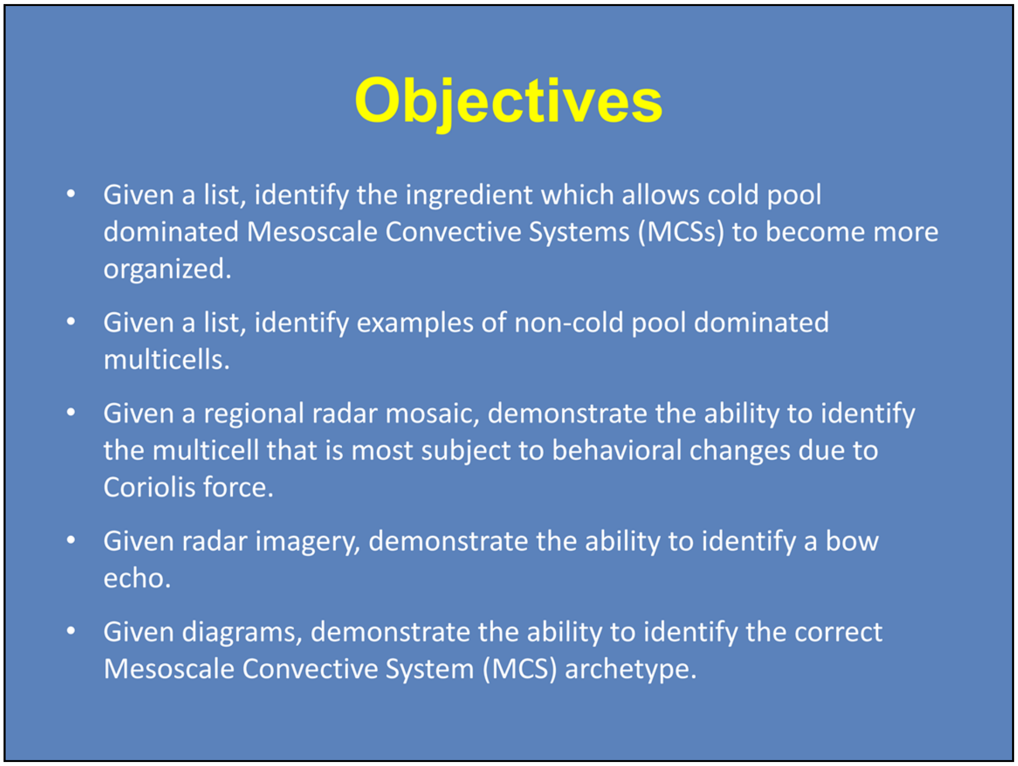
Welcome to the RAC Convective Storm Structure and Evolution lesson on Multicell Archetypes.

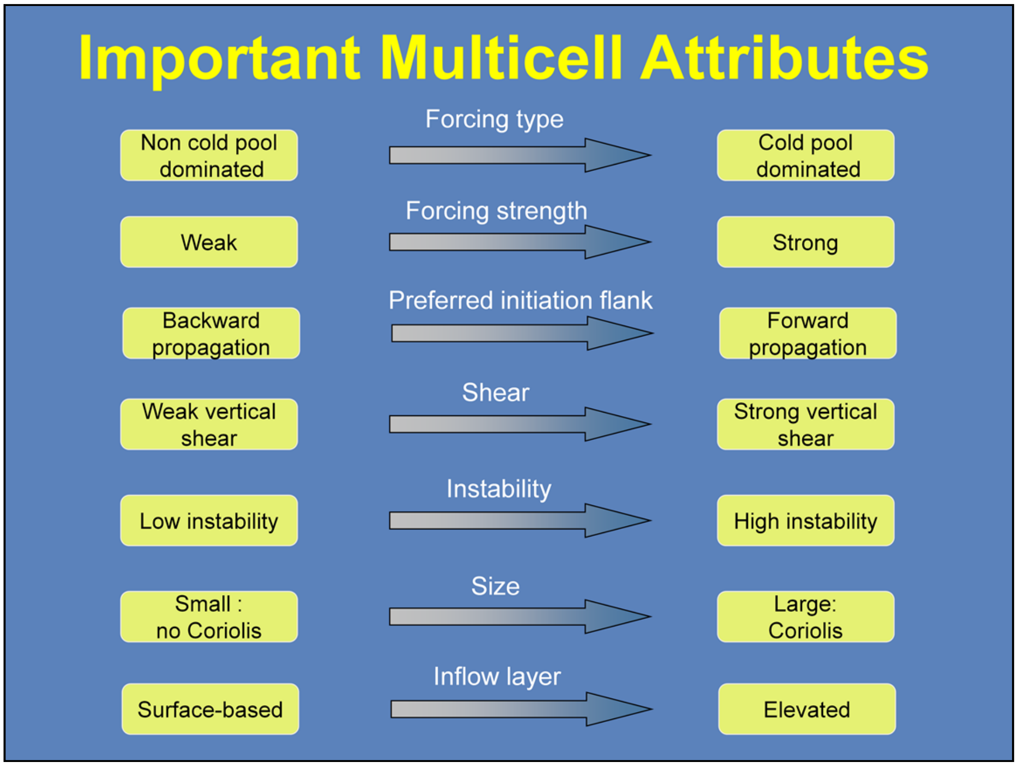


Multicell storms consist of individual cells, either ordinary or supercellular, in close enough proximity to affect each other in some way. For the purposes of this lesson, multicells are groups of two or more cells that at least share a common precipitation area and a cold pool. In nature, most Deep Moist Convection (DMC) becomes multicellular because there is typically more instability and forcing than one cell can alleviate. It is very rare for a single cell to be initiated in complete isolation from subsequent initiation; therefore, multicells are common in the broad parameter space of instability and vertical wind shear. However, the combination of forcing, vertical shear, and instability has an impact on the size and organization of multicell structures.

This lesson has two parts. The first part describes the mechanisms that influence basic multicell structure. The second part discusses the common archetypes exhibited by multicells.



Here are the objectives. Please take a moment to review them.



The categorization of multicells is quite complicated owing to the large variety of documented structures and forcing mechanisms that are immediately relevant to your severe weather forecasts. No one single multicell categorization scheme has been developed. Instead, let's describe the major characteristics which influence the type of multicell:

The type and orientation of forcing. Many multicells are driven primarily by their own cold pools. However, multicells can be dominated by non cold pool processes. The orientation of forcing may also influence the development of a multicell.

Strong forcing can certainly hasten the initiation of new cells more quickly than weak forcing (all other factors being equal).

The location of forcing relative to the multicell influences its motion.

The vertical wind shear of the environment interacts with the cold pool and influences the ability for new cells to generate. We explain this more in the lesson on multicell motion. The interaction of the vertical shear with updrafts within the multicell also influences the potential for new cell generation. This is especially true for multicells with weak cold pools or for multicells with embedded supercells. We explained this in the lesson on supercell dynamics.

The vertical stability profile influences the ease at which forcing can initiate new convection and the ability for a multicell to create a significant cold pool.

Multicell size and the influence of the Coriolis force are positively related to each other. As multicell size increases, so does the likelihood of developing a dominant cold pool, developing other internal circulations (such as a rear inflow jet), and the likelihood that the Coriolis force is a significant influence.

And lastly, **the parcel inflow layer of a multicell** influences its behavior depending to what degree the updraft/downdraft base is elevated.

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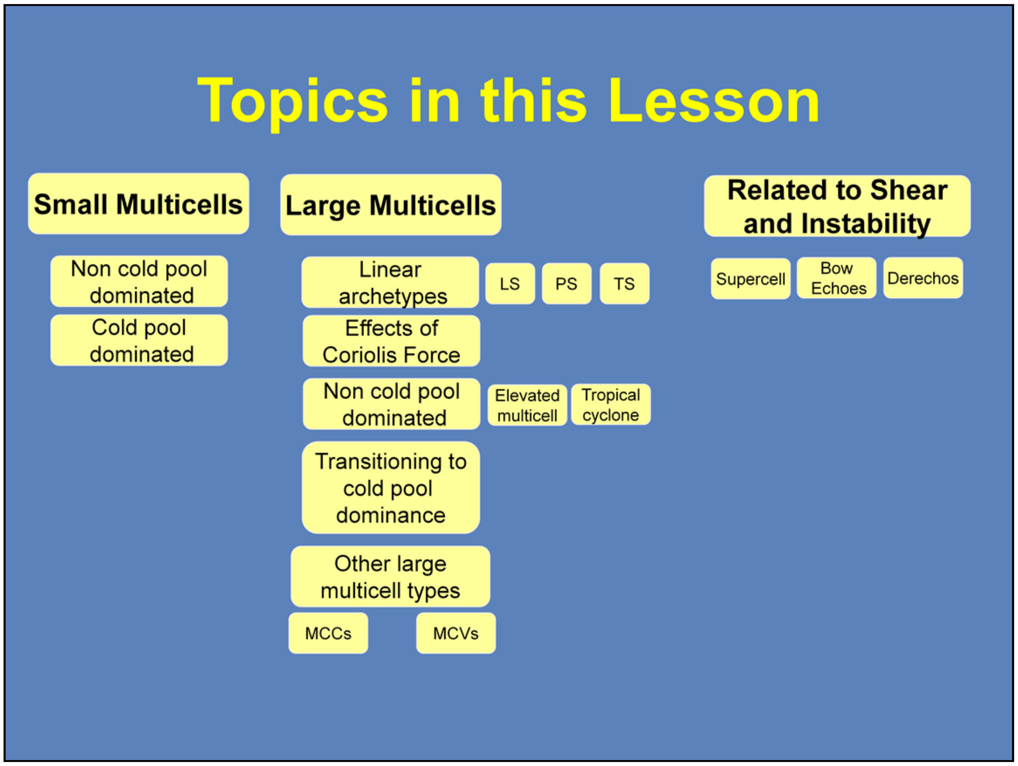
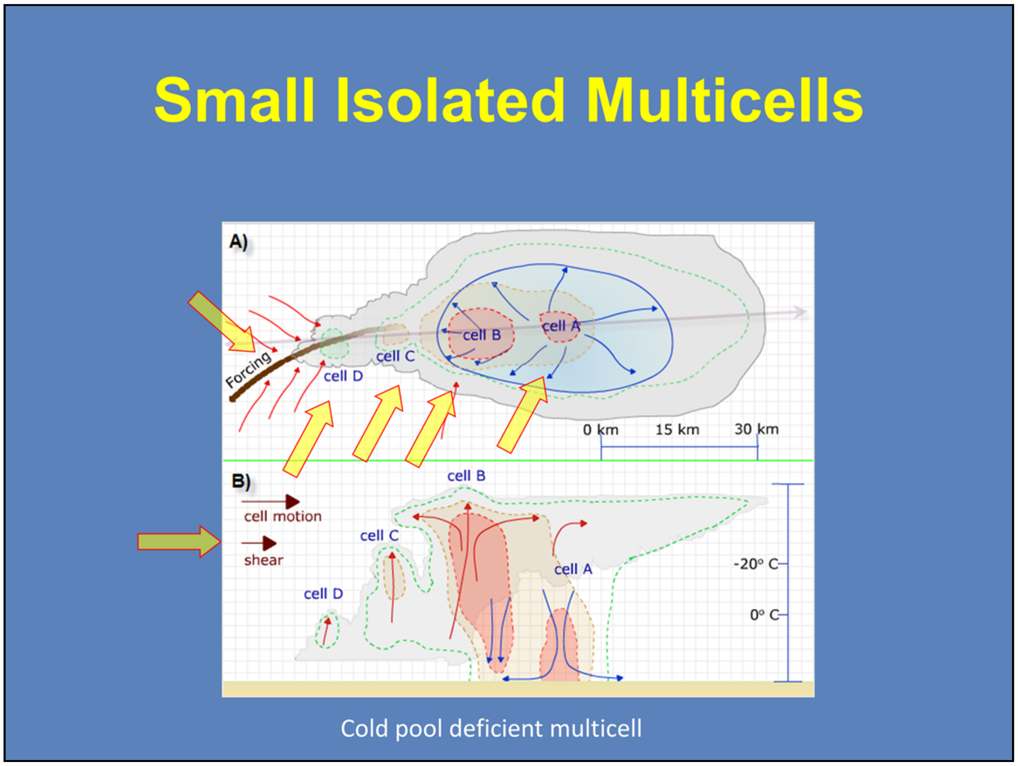


In this lesson we will talk about a variety of structures and archetypes. And we'll split them up according to size, because that's seems to have the greatest impact on how to discriminate multicell events. We'll start with small multicells and sub-categorize them by whether or not their evolution is dominated by their own cold pool.

Then we'll discuss large multicells. We'll talk about linear archetypes and a variety of sub-archetypes there. We'll also talk about the effect of Coriolis force on large multicells. Large multicells can also be non-cold pool dominated, so we'll talk about the elevated large multicells and even tropical cyclones. These large events can transition to cold-pool dominant events and we'll talk about the process that occurs there. And we'll also talk about other large multicell archetypes that have been defined using other instruments. Mesoscale Convective Complex's (MCCs) have been defined using satellite-based instruments. We'll also talk about Mesoscale Convective Vortices (MCVs).

All of these events are greatly impacted by their environmental shear and instability. For example, we can get supercells out of small multicell events. We can also get bow echoes and derechos.



The smallest (meso-g scale) multicells typically feature several individual cells in different stages in development. You may have seen these also called isolated multicells. New cells form before old ones dissipate but in close enough proximity to share cloud material, and precipitation. Typically there are a small number of cells in any one stage in the lifecycle as depicted in this schematic. Cell A represents the dissipating stage where the precipitation core is raining and fully occupied by downdraft. Cell B is in the mature stage where the heavy precipitation core is descending with downdraft and the updraft bubble has overshot its equilibrium level. Cell C and D represent the newest members of this multicell and are primarily updraft dominant. This type of multicell is too small to create a linear organization.

In the conceptual model shown in this figure, new cells are forming along an axis of forcing well outside of the cold pool boundary. The cold pool, while present, is too deficient to dominate the initiation of new cells. Small multicells are most likely to be cold pool deficient, and therefore, be dependent on external forcing and instability. While the forcing is depicted upstream of the steering layer flow, it does not have to be in order for this to be a non cold pool dominated small multicell.

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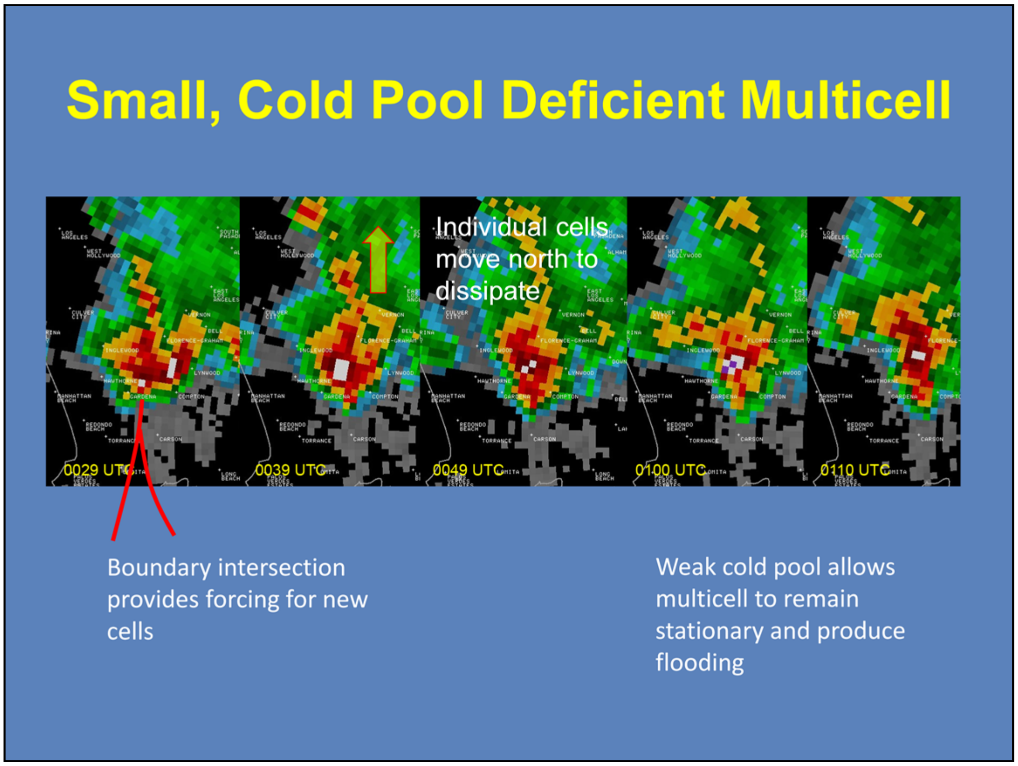

The multicell shown in this figure was primarily driven by a point source forcing manifesting as a nearly stationary interaction between a sea breeze front and a cumulus cloud roll. New cells formed at the intersection, then progressed north and dissipated on the north end of the multicell. The cold pool was too weak to force the multicell to move, and a backward propagating, flash flood producing multicell developed in south Los Angeles.

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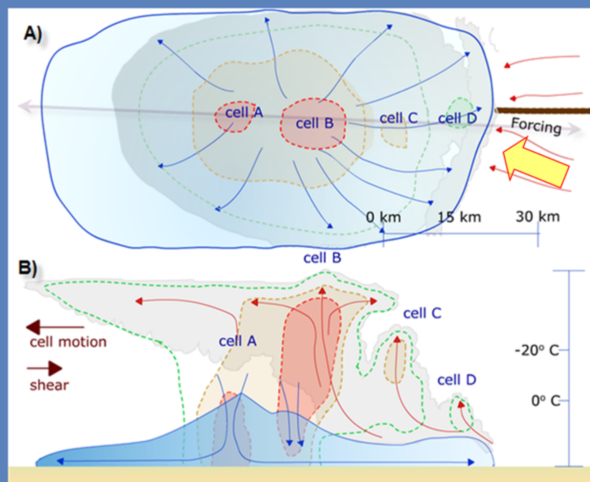
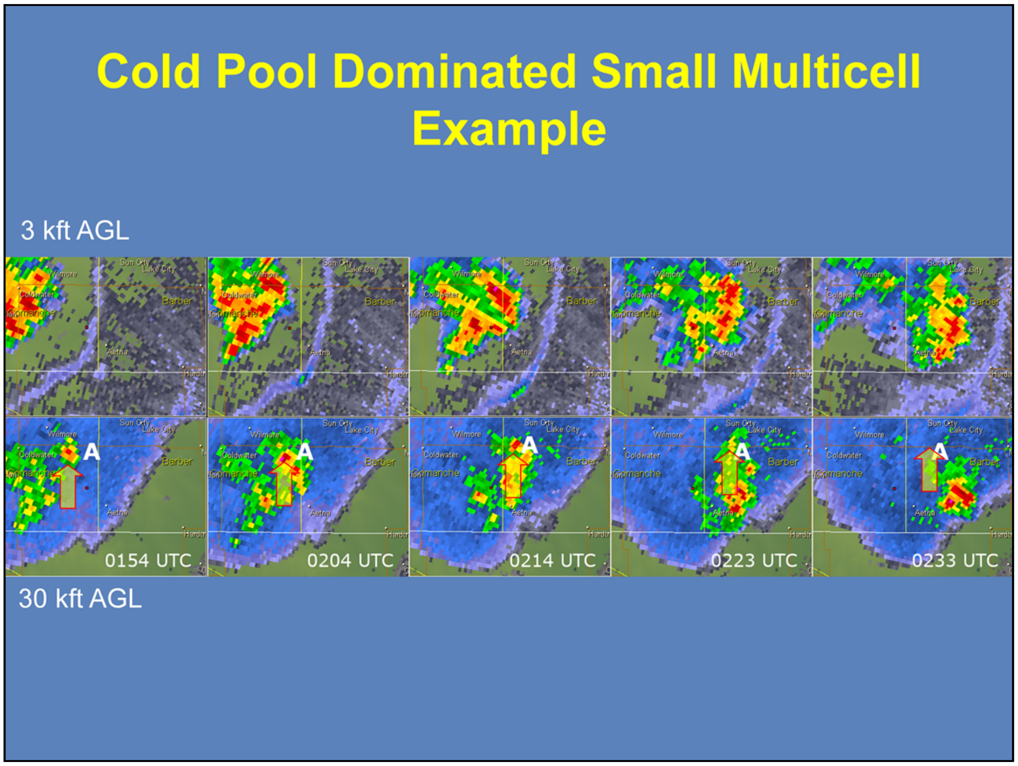
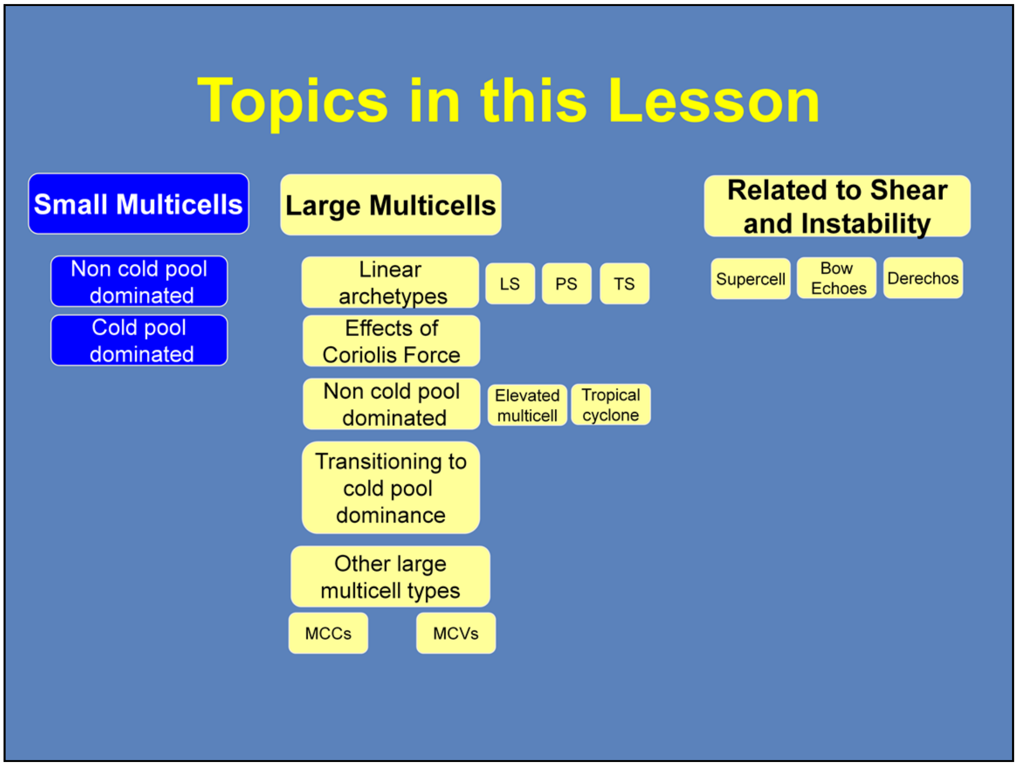
Cold Pool Dominated Small Multicell



A small multicell could wind up being dominated by a significant cold pool in an environment conducive to its formation. Here is a conceptual model of a cold pool dominated small multicell. New cell initiation is forced primarily by ascent from the leading edge of the cold pool. The gust front typically moves faster than individual cells, and therefore, this type of multicell outpaces individual cell motion. We call this forward propagation. Note that the cell motion label here is relative to the whole multicell. Also note the line of “forcing. It’s not necessary, but it does help to focus cell initiation in one spot.



Here is an example is of a small, cold-pool dominated multicell attempting to match speed with a surging gust front. Watch how cell A forms and quickly matures then dissipates as it moves to the east much more slowly than the gust front and the multicell.



Now, we'll talk about large multicells.

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Large Multicells

- From meso-β to meso-α scales
- Many cells in similar stages of development
- Mesoscale Convective System (MCS)
  - Exceeds 100 km in any direction
- Impact of Coriolis force can be significant

The dimensions of a large multicell can range into the meso-b and meso-a scales and persist for several or more hours. Large multicells typically contain many cells in similar stages of development.

A **Mesoscale Convective System (MCS)** is a multicell whose contiguous precipitation area exceeds 100 km (54 nm) in any direction. However a large multicell need not satisfy such criteria. Unorganized MCSs may exhibit multiple flanks of sporadic, relatively infrequent, new cell initiation. Meanwhile, organized MCSs may exhibit relatively frequent new cell initiation on a preferred flank. MCSs produce large anvil shields and subsequent areas of stratiform precipitation in addition to strong system wide circulations that influence its structure and evolution. Only in persistent MCSs does the Coriolis force become a significant influence in its evolution.

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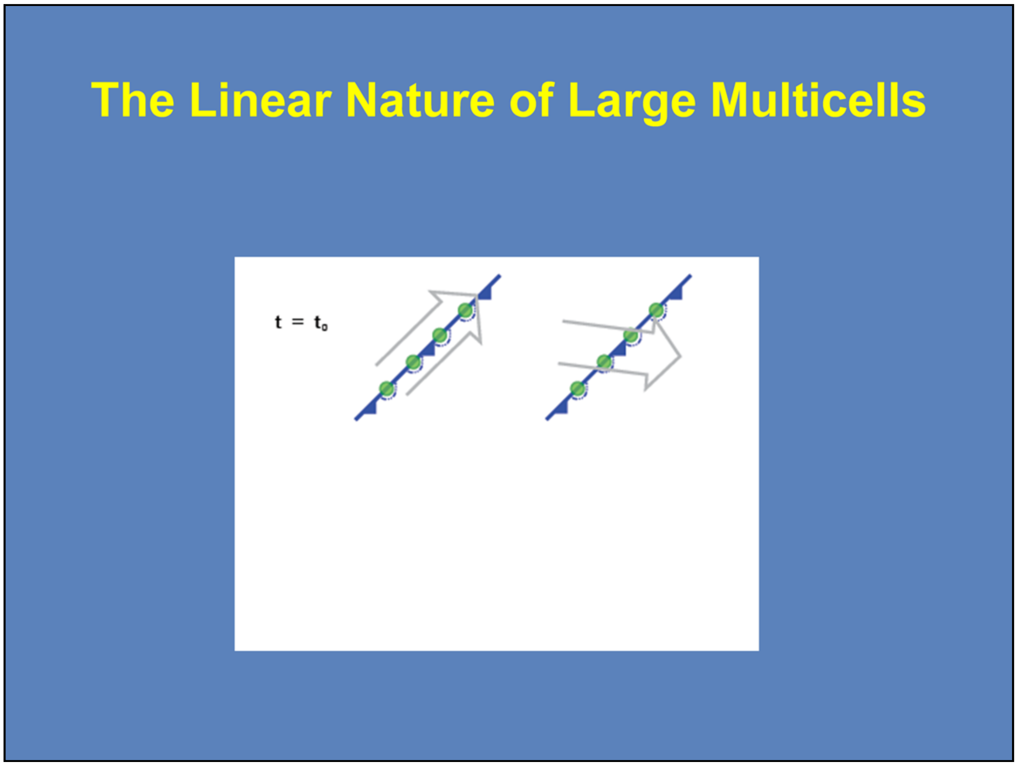
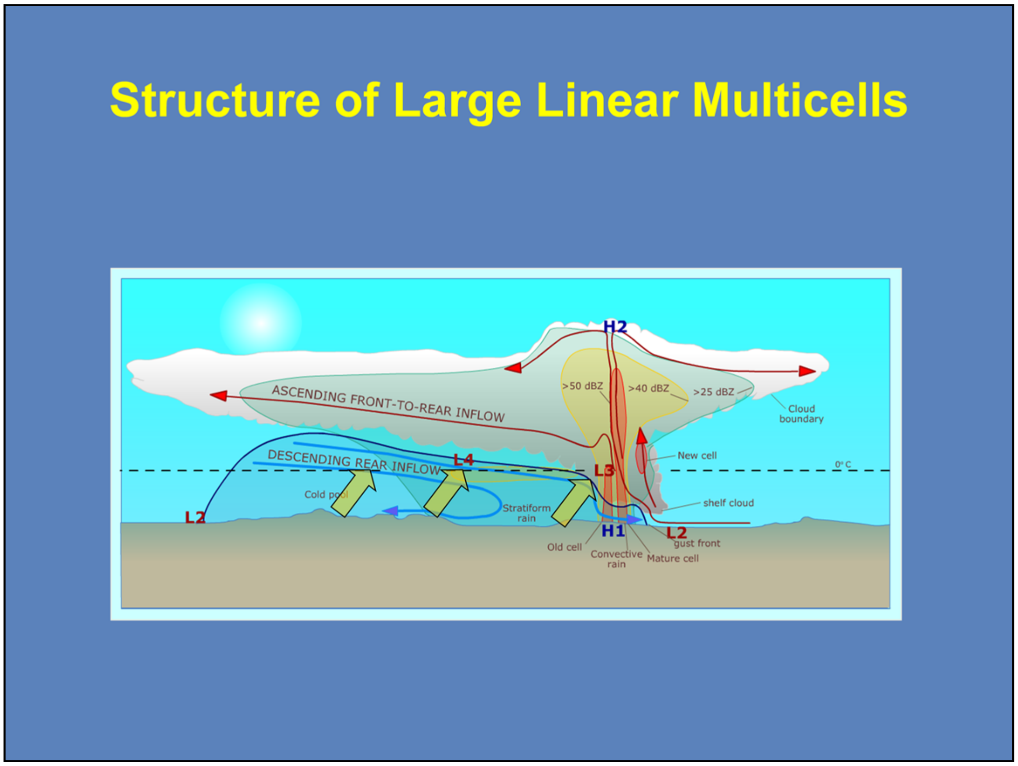


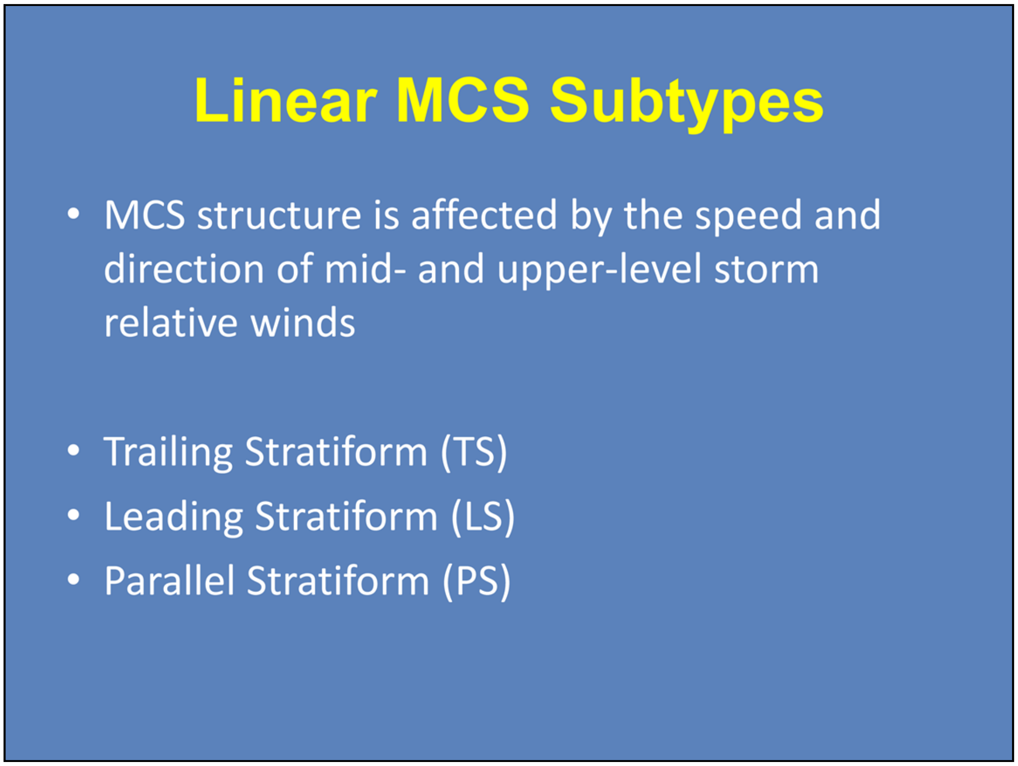
Large multicells are more apt to exhibit a linear nature to them reflecting the elongated lifting that commonly occurs along external forcing mechanisms (e.g., fronts), and internally generated cold pool boundaries. Fronts provide linear forcing, but multicells may not merge into a long line if the forcing is weak. However, if the deep layer shear is largely boundary-parallel, individual cold pools may more easily merge, reinforce the front, and enhance upscale growth into a long line.



In addition to the propensity for linear development, MCSs also develop significant system wide circulations as the large anvil and cold pool modulate the pressure field from the surface to upper levels. Large areas of stratiform rainfall fall from underneath the anvil.

Two of the more widely accepted conceptual models of the complex flow structure are from Smull and Houze (1987) shown here.

A **Rear-inflow jet (RIJ)** is a mesoscale region of strong winds that originate in the trailing stratiform rainfall region of a squall line near the top of the cold pool and are directed toward the leading edge**.** In the Smull and House (1987) model of a mature MCS, development of the RIJ is attributed to mid-level, mesoscale areas of low pressure (labeled L3 & L4). The mesolow “L3”, which forms immediately behind the leading line convection, is a hydrostatically-induced, negative pressure perturbation that develops under up-shear tilted warm, convective updrafts and above the evaporatively cooled downdrafts. Mid-level mesolow “L4” forms in the stratiform region in between the warm, buoyant air which gets pulled rearward past the cool, dry descending air flow. Note that the major difference between this figure and a small cold pool dominant multicell is the presence of a stratiform precipitation region, and attendant RIJ; both are features that more frequently accompany large MCSs.



Although MCSs develop a number of ways, typical mature systems contain both convective and stratiform precipitation regions. The eventual MCS type is determined to a large extent by the environmental conditions in which it develops and the strength of the system cold pool. Parker and Johnson (2000) studied numerous MCSs and determined the distribution of hydrometers and stratiform precipitation shapes were largely a result of mean storm-relative winds. **The speed and direction of the environmental mid- and upper-level winds relative to system motion affect the resulting evolution of the MCS.**

According to Parker and Johnson (2000), MCS squall lines evolve into three major archetypes: **1) trailing stratiform, 2) leading stratiform, and 3) parallel stratiform**. The main distinction arises from storm-relative flow fields.

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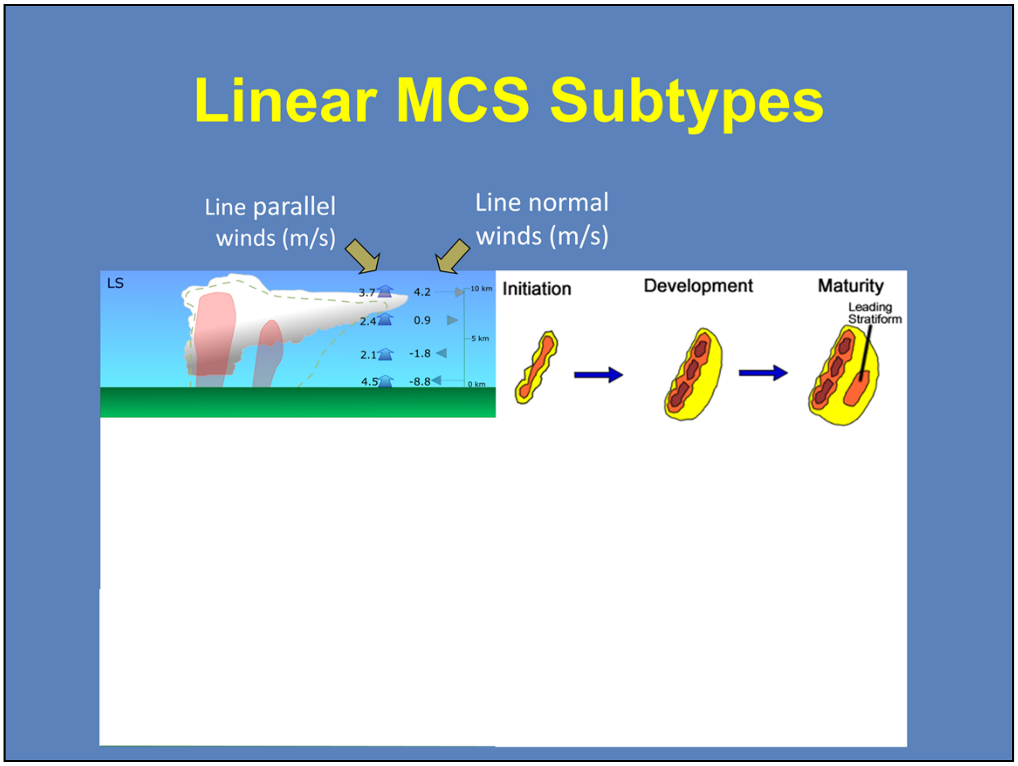

The Leading Stratiform (LS) precipitation squall line archetype, which is typically slower-moving than trailing stratiform systems, is characterized by stronger mid- and upper-level storm-relative flow (often described as rear-to-front flow) than any of the other types. It's not so much the line parallel wind, but rather the line-normal wind that carries the anvil debris ahead of the main convective line. As a result, they tend to produce the weakest cold pools and may sometimes not be cold-pool dominated.

The Parallel Stratiform (PS) MCS structures represent a slightly less common form. The anvil and attendant stratiform precipitation expand outward, both in front of and behind the intense convective line. They occur in situations with strong along-line storm-relative flow, especially in mid- to upper-levels. They produce strong cold pools and therefore it is difficult to maintain this archetype for long time periods.

The Trailing Stratiform (TS) squall line type has a sloped front-to-rear flow produced by stronger system-relative flow in low-levels (and subsequent stronger low-level convergence along the leading edge). The line normal wind is from front to rear, which produces clouds and precipitation behind the line, which generates a strong cold pool with a rear inflow jet and provides most of the forcing for its maintenance. Trailing Stratiform squall lines tend to move quite quickly.

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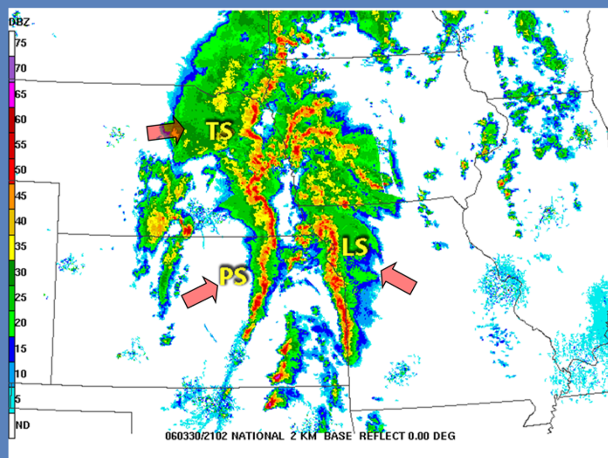
Linear MCS Archetypes Examples



This figure shows an example of all three linear MCS archetypes occurring simultaneously ahead of an ejecting, strong, upper-level shortwave trough. Each MCS formed from a different boundary. The Trailing Stratiform formed on a stationary front north of the surface low, the Parallel Stratiform formed on a dryline, and the Leading Stratiform formed on what may have been a residual outflow boundary with the cooler air to the east. This case was document by French and Parker in 2006.

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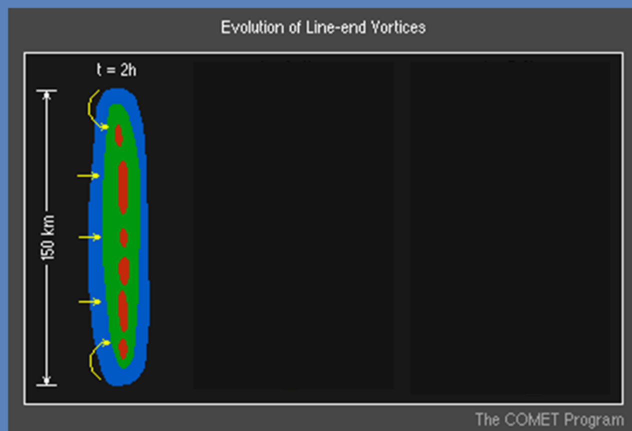
Effect of the Coriolis Force on Large Multicells



Large multicells, or Mesoscale Convective Systems (MCSs), are the convective events most subject to behavioral changes from the Coriolis force. As shown in figure, most linear MCSs develop book-end (or “line-end”) vortices through the tilting of horizontal vorticity, either from the environment or along the cold pool edge. Provided a sufficiently long lifespan, the cyclonic member of the system becomes reinforced by the Coriolis force while the anticyclonic member is weakened. The MCS becomes deformed and eventually exhibits a comma shaped configuration. A mid-level hydrostatic low under the anvil shield of the MCS also persists long enough to allow the Coriolis force to create a cyclonic circulation.

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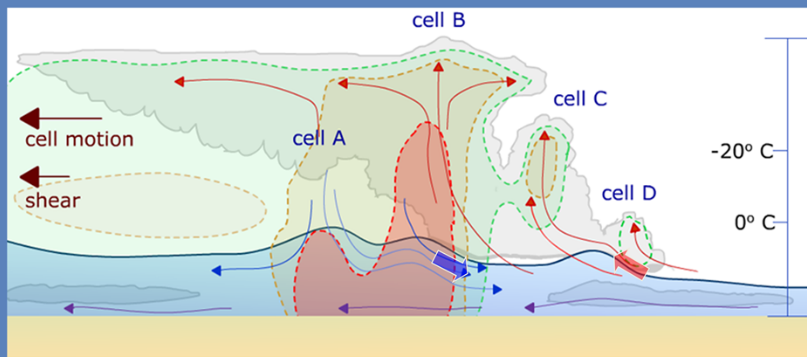
Large Non Cold Pool Driven Multicells: Elevated MCSs



While most of the large multicell archetypes that we have discussed are dominated by cold pools, there are a host of large multicells embedded in environments that do not allow the cold pool to become significant, or are embedded within environments that do not allow the cold pool to become a significant driver.

The elevated MCS is one type of multicell likely not to be affected by a cold pool. Elevated multicell deep, moist, convection implies that the updraft parcel roots for each cell has a source above the ground, and that the air near the ground is conditionally stable. The morphology of elevated multicells depends even more on the shape and intensity of the original forcing within the context of the vertical stability profile. Forcing mechanisms are more predominantly associated with elevated lifting such as differential vorticity advection, localized warm advection, elevated frontogenesis, or gravity waves. Elevated multicells produce downdrafts; however, the resulting gust front is ineffective at creating new surface-based convection as long as the near surface air has zero Convective Available Potential Energy (CAPE). Downdrafts penetrating into the stable layer may not reach the ground to create a gust front, and instead may crated gravity waves that help initiation new convection as you can see in this figure.

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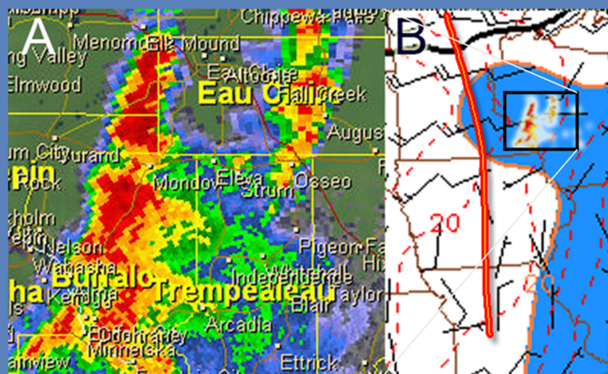
Elevated MCS Example

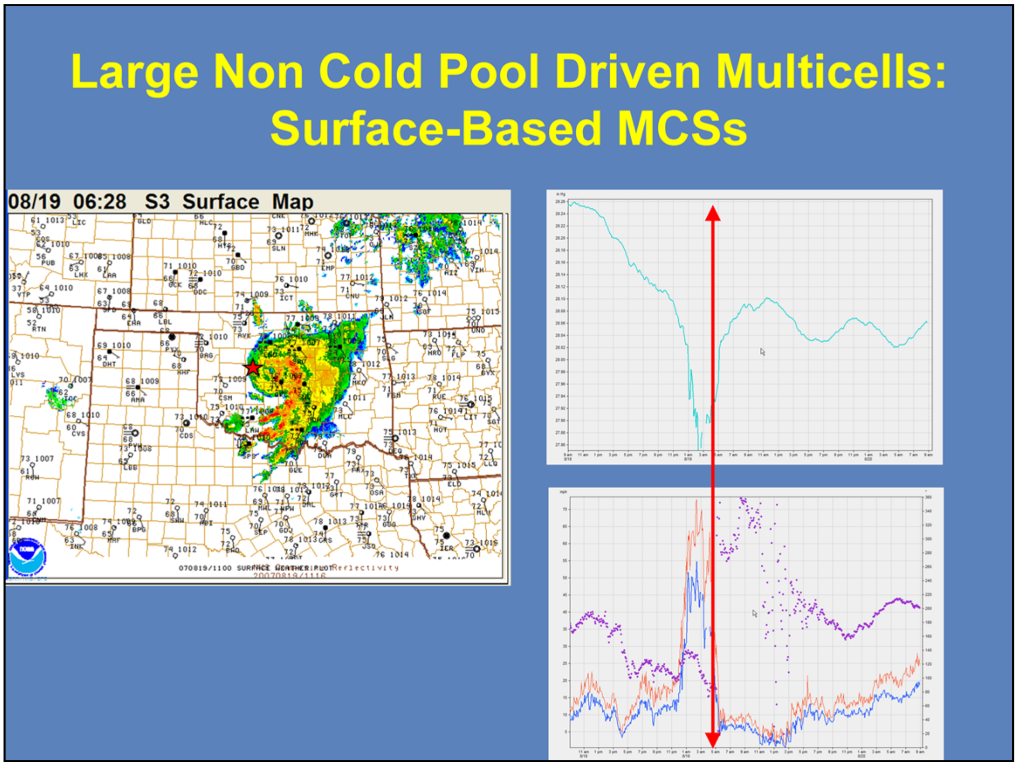


Lowest scan reflectivity from KARX

850 mb temperatures (dashed red), frontogenesis (blue shading)

Here is an elevated MCS example from July 2007 that shows a group of cells that initiated along a north-south axis which corresponds with a region of 850 mb warm advection and frontogenesis, and a region of elevated effective inflow base where buoyant parcel ascent was possible.

What happened to this MCS will be covered momentarily.



A surface-based MCS is also likely to be cold pool deficient in very moist environments with minimal DCAPE, or if there is significant DCAPE, a low-level heating source acts to modify any cooling.

An extreme example of this kind of multicell is a tropical cyclone. This is a warm core multicell whose pressure minimum under convective heating does not become concealed by a dense cold pool near ground, and the forcing is internally driven.

A tropical cyclone typically requires a constant heating source such as warm water in order to maintain itself and mitigate cold pools. However, similar structures have been found over land where cold pool production is weak. In this slide, the remnants of tropical storm Erin were reinvigorated as the circulation center redeveloped a new MCS. In turn, the MCS did not produce a cold pool, and a warm core low intensified right down to the surface.

Where the star is located, you can see the corresponding meteogram, with sea-level pressure in the top panel and winds in the bottom panel. Notice the wind maximum was co-located with the minimum of pressure, very typical of a tropical cyclone passage. The red arrow corresponds to the time of the radar image.

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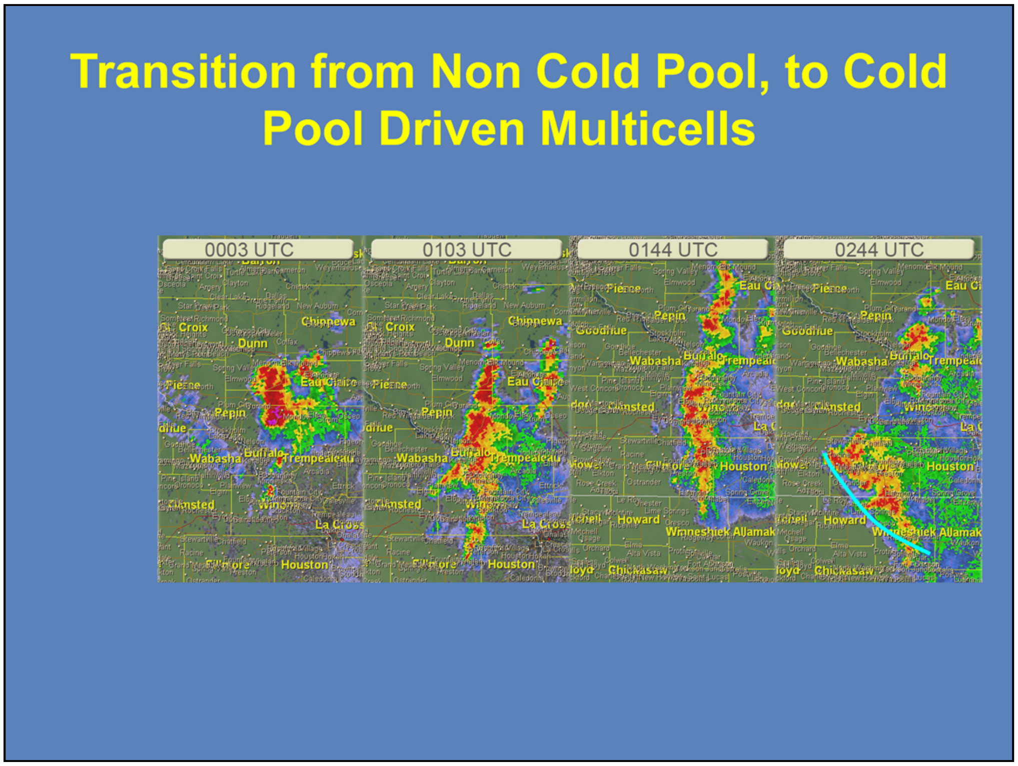

Let’s return to the elevated MCS from July 2007 and show what happened to it. Cold pools are prone to deepening as multicells grow upscale and/or persist. A multicell may transition from forcing-dominated to cold pool driven as lifting increases over the cold pool or by the gust front. This sequence shows our elevated multicell was dominated by frontogenesis and/or warm air advection forcing from 0003-0150 UTC. However, by 0244 UTC, a cold pool had formed which was large and powerful enough to force surface-based convection along the gust front boundary. As a result, the multicell propagated to the southwest.

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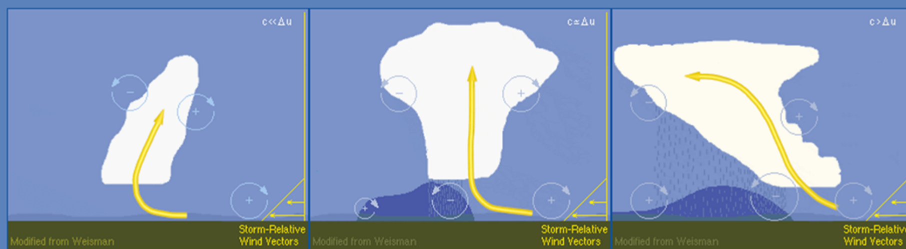
Transition from Non Cold Pool to Cold Pool Driven Multicells cont'd



Initial convection and weak cold pool	Cold pool forms and balances with inflow shear	Cold pool grows to dominated the multicell

Another common example of this transition occurs when a line of discrete deep, moist convection transitions into a squall line. In this case, discrete multicellular convection initiates from an external forcing mechanism and generates a cold pool that strengthens over time. Early on, multicell forcing may be dominated by either external mechanisms or updraft-induced dynamic pressure gradients. Eventually, the cold pool deepens and becomes the dominant forcing mechanism. How long this process takes depends on the strength of the cold pool compared to the vertical wind shear and the strength and orientation of the initial forcing.

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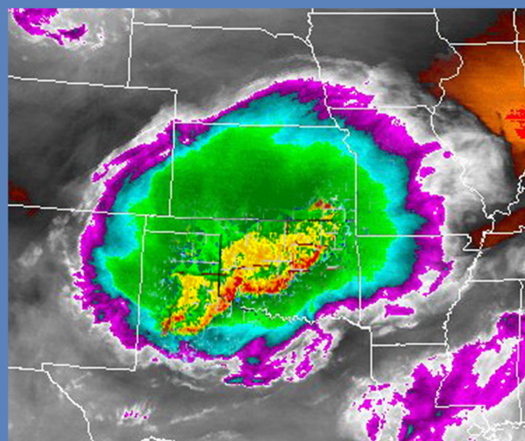
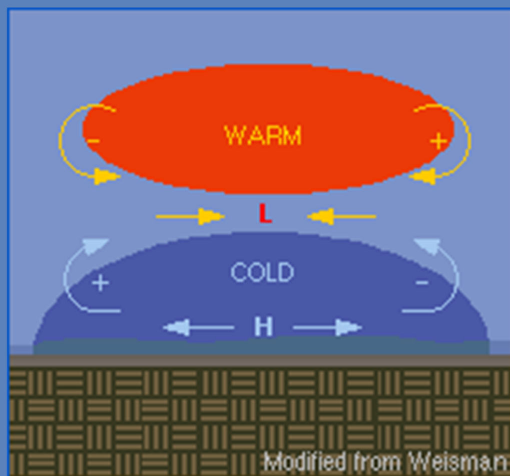
Mesoscale Convective Complexes (MCCs)



A Mesoscale Convective Complex (MCC) (defined by Maddox in 1980) is a subset of MCS that exhibits a large, circular (as observed by satellite) long-lived, cold cloud shield. The MCC’s circular anvil shield masks the linear nature of the active convection underneath.

Mesoscale Convective Vortexes (MCVs)



Regardless of the underlying linear shape of the intense convection, the anvil's circular shape may allow for a coherent Mesoscale Convective Vorticity (MCV) maximum to form in the midlevels as anvil-layer heating produces a hydrostatically induced low. MCVs tend to persist more when the vertical shear is not strong. Perhaps that is why they tend to form most often on the equatorward edge of the midlatitude westerlies, and in the tropics. While many MCS structures produce a midlevel hydrostatically induced low, the shape and size of MCC anvils makes them most likely to produce long lasting MCVs which can, in turn, produce a variety of multicell structures the following day.

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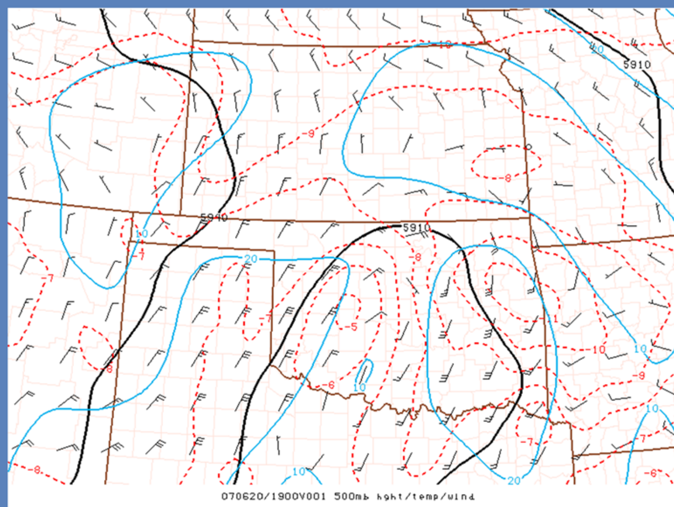
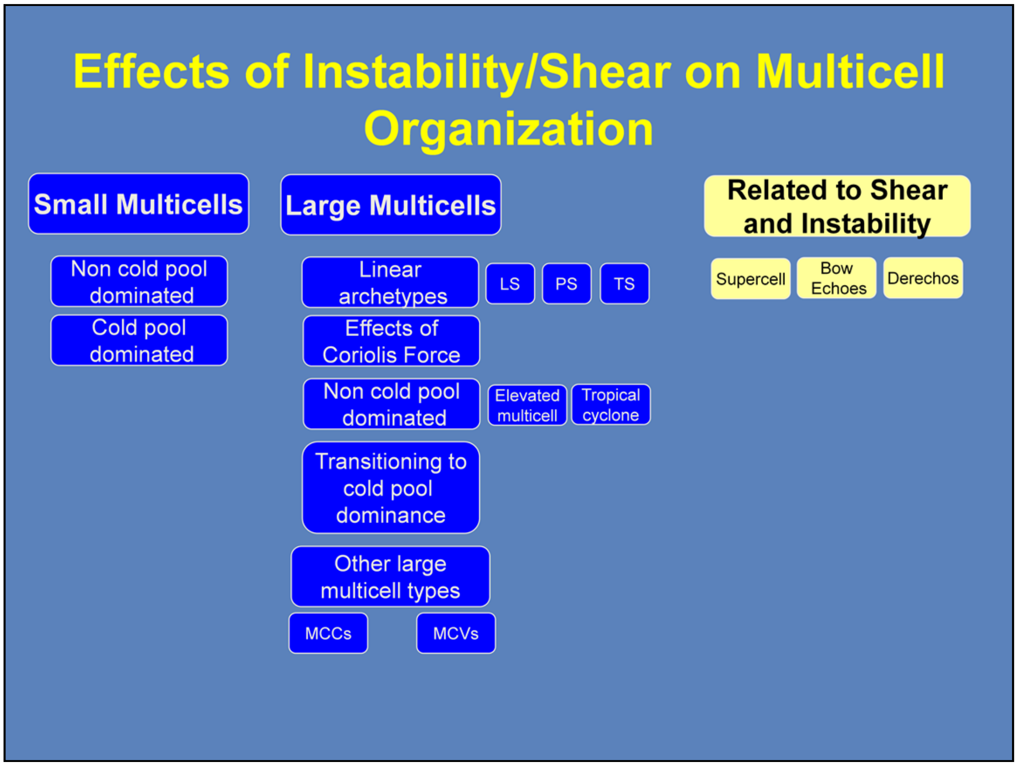
Mesoscale Convective Vortex (MCV) Example



The MCC two slides ago produced a substantial MCV that became detached from the westerlies and persisted for days producing repeated episodes of storms and flash flooding over the southern Plains.



Now that we've covered both small and large multicells, let's now talk about how they are impacted by shear and instability.

We first describe what it means for a multicell to be organized. More frequent and stronger initiation on a preferred flank of a multicell is one of the criteria for identifying a multicell as organized. In this section, we'll describe that process and what kind of structures result from that. We'll concentrate mostly on bow echoes and derechos because we discuss supercells in other lessons.

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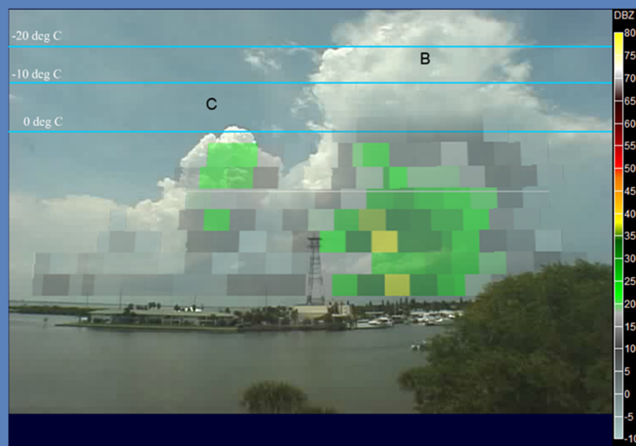
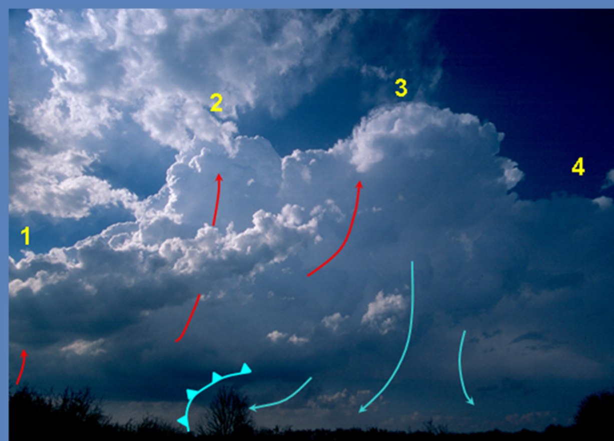
Weak Cold Pool Small Multicells – Weak Shear/Instability

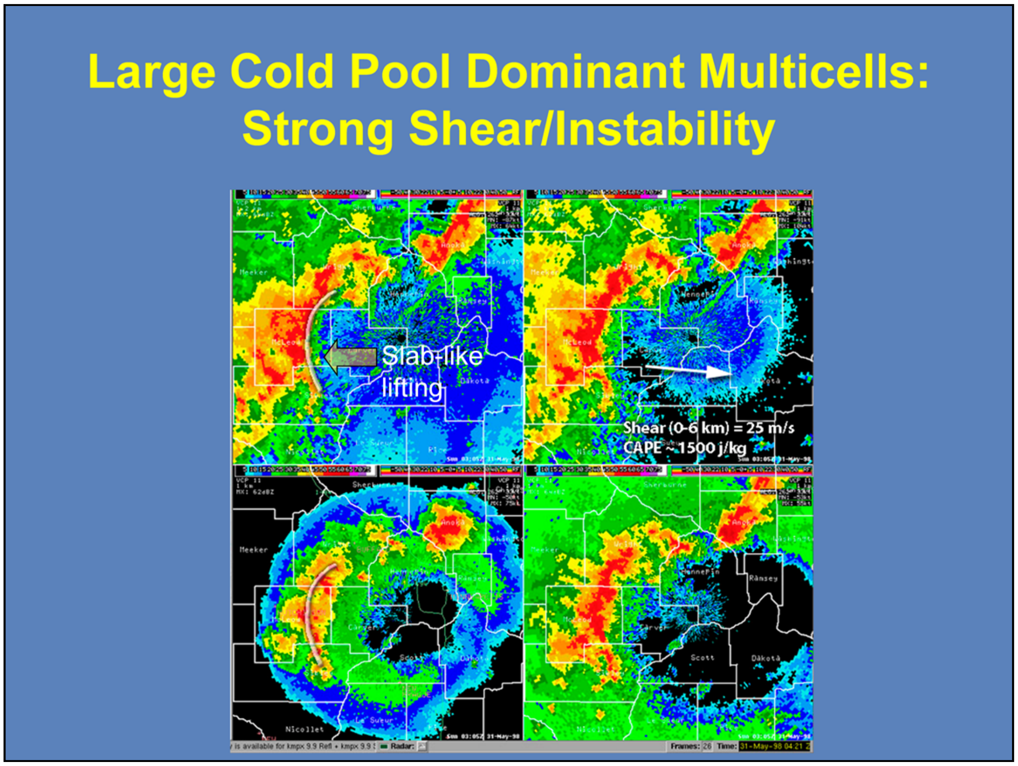


The weaker the instability and/or vertical shear, the more infrequent and weak cell regeneration. In this example, the individual cells are regenerating so infrequently that they are almost discrete.

Weak Cold Pool Small Multicells – Strong Shear/Instability

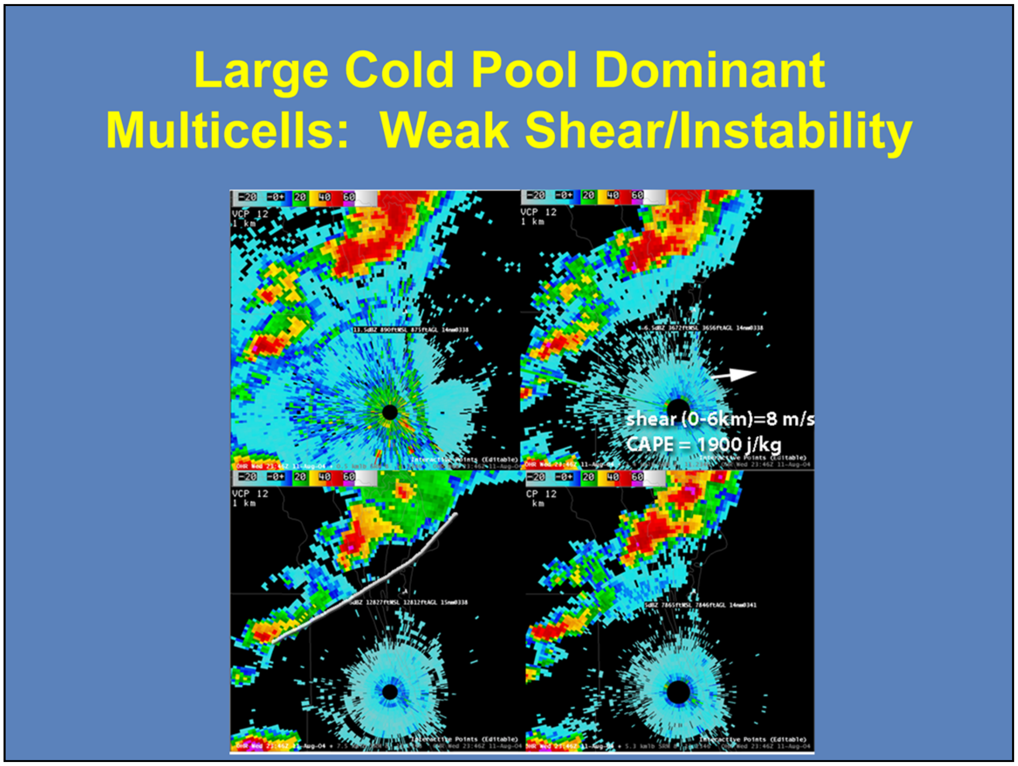


More organized small multicells exhibit an appearance of a more persistent, plume-like updraft, and adjacent heavy precipitation area as the rate of new cell initiation becomes more frequent. You may find that eventually the multicell acquires supercellular characteristics.



The impact of increasing vertical wind shear on a large, cold pool dominant MCS allows for the increased possibility of a continuous, stronger updraft along a preferred flank where a downshear component exists. An optimal state of shear would entail low-level shear well-balanced with the cold pool boundary-induced circulation, and/or sufficient deep-layer shear to allow the convective steering layer flow to match the gust front. Multicells satisfying these conditions exhibit slab-like lifting where you may have a difficult time discerning any discrete character to the cells. This image provides a good example of slab-like lifting along a gust front facing downshear. This event produced widespread significant wind damage.

We have a lesson on the environmental impacts on multicell organization in another lesson.



In this example, the difference between the shear and the cold pool strength has likely caused an imbalance which has limited lifting and the efficiency for creating new cells. You can see only scattered cells forming well behind the gust front. A shear/cold pool balance can be upset should the cold pool intensify such as changing of the instability while holding shear constant or changing the shear while holding instability constant.

The most important component of the shear that impacts gust front lifting is that which is perpendicular to the gust front. And here you can see that the shear is mostly parallel to the gust front with almost no perpendicular component.

Derechos

- Concentrated of severe wind reports 400 km in length, some wind gusts > 65 kts, no gaps > 3 hours.
- Are severe weather events, not a form of multicell
- Produced by strong shear/instability
- Produced by MCSs often containing bow echoes.

Strong shear interacting with a strong cold pool allows a cold pool forced multicell to become more severe and longer lasting, and can result in a **derecho**. A derecho is a widespread convectively induced straight-line windstorm that exhibits a concentrated area of damaging winds with a length of at least 400 km, shows an organized damage swath, contains gusts greater than 65 kt, and does not show gaps of more than three hours. These are severe weather events, not really a form of multicell, but they are produced by strong shear and instability combinations and are also produced by MCS's containing bow echoes.

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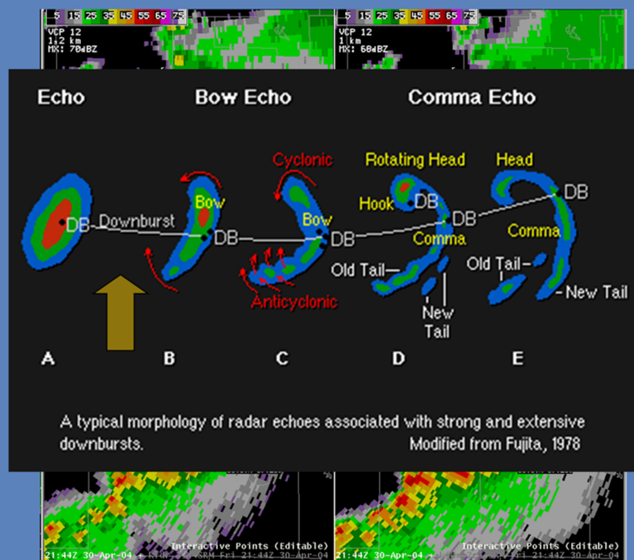
Bow Echoes



A bow echo is a bow-shaped, multicell line of convective cells that is often associated with swaths of damaging straight-line winds and sometimes tornadoes. The bow appearance occurs because the precipitation has been deformed into the characteristic shape by a rear-inflow jet (RIJ). Bow echoes are meso-b scale features and often have lifetimes between 3-6 hours. Severe bow echoes are strongly favored in high CAPE/shear environments. They represent almost continuous, slab-like lifting along a deep gust front of a cold pool dominated multicell.

Here is a four panel reflectivity image of an organized cold pool dominated multicell where the center picture has been taken from point A in the upper left panel. In this figure, an organized, small multicell is taking on a bowing configuration as strong environmental vertical shear interacts with the deep cold pool. If you were to assign a stage in this bow echo life-cycle, it would likely fall between stage A and B in this schematic. Note that the most obvious manifestation of the bow occurs typically well after the onset of severe surface winds.

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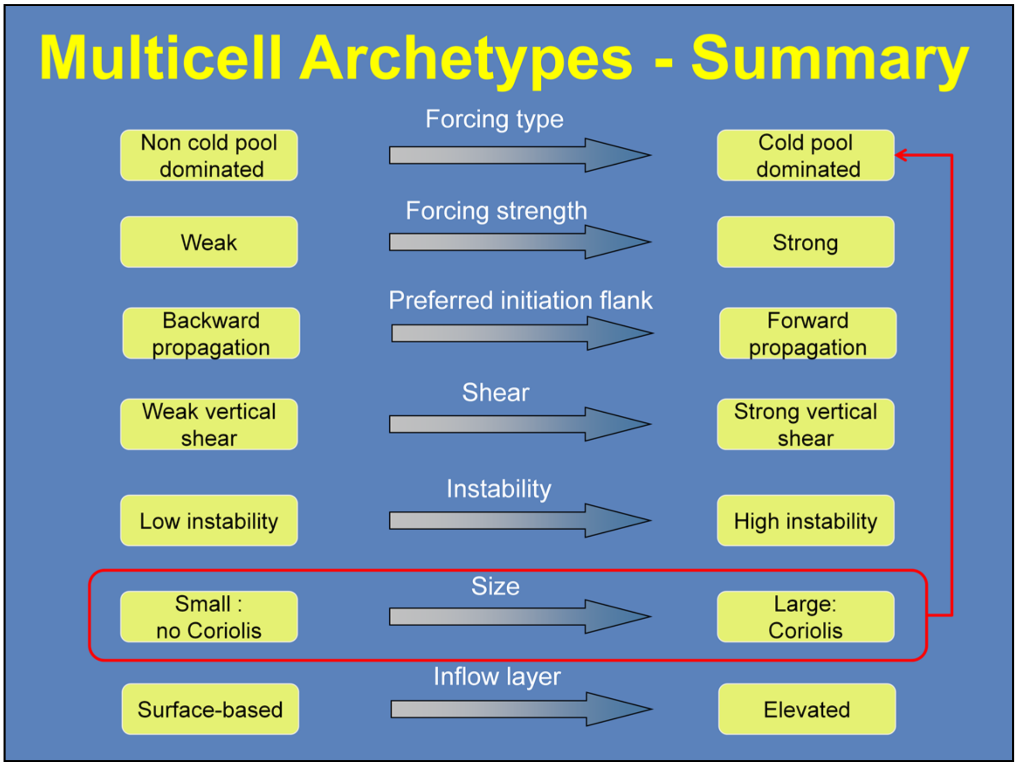


When you describe multicell archetypes, you may want to describe the multicell with fundamental attributes (much like adjectives) that help determine their behaviors. They include:

Small to large multicells, which represent a spectrum where large indicates an MCS potentially influenced by the Coriolis force. Large multicells are more likely to be cold pool dominant. Changing shear and instability impacts small and large multicells in different ways.

Cold pool dominant to non-cold pool dominant, multicells represent fundamentally different forcing mechanisms.

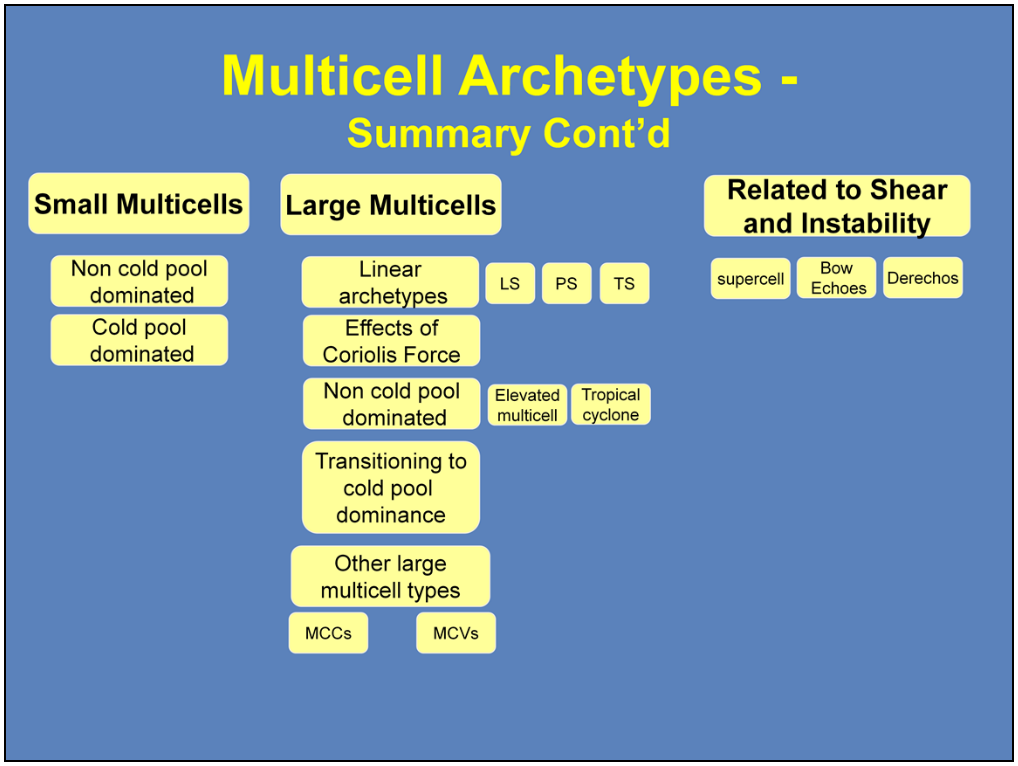
Non-cold pool dominant multicells are primarily forced by mechanisms external to the multicell (e.g., fronts, gravity waves, differential heating). Their shape, movement and size match closely with that of the forcing. Cold pools do not affect true elevated multicells or they are too weak to impact surface-based multicells.

The cold pool dominated multicells implies that they are forced by internally generated cold pools, though the strength of that forcing is modulated by the environmental vertical shear, stability, and the multicell itself. Multicells commonly evolve from non- cold pool to cold pool dominated modes as the multicell and cold pool grows with time.

The location of forcing relative to the multicell can determine its motion. If the forcing for new cells is to the rear (front) of the multicell then backward (forward) propagation is imminent.

The vertical wind shear and stability, assisted by strength of forcing, modulates the strength and frequency of new cell initiation. Small multicells may evolve into supercellular behavior with increasing shear. Large, cold pool dominated multicells may experience slab-like lifting with a balance between cold pool strength and shear. The severity of multicells is positively correlated with shear.

Finally, the behavior of the multicell is greatly influenced by its ability to tap the boundary layer.



The shape of the cloud and the precipitation distribution are other important descriptors of multicells that yield information valuable to a forecast process.

Most large multicells acquire some linear organization to them because of the shape of external forcing or by its own cold pool. Squall lines are an example of such accompanied by a brief burst of strong winds.

With linear large multicells, the stratiform precipitation distributions relative to the active convective line in large multicells lead to the following archetypes:

•Leading Stratiform precipitation (TS),
•Parallel Stratiform precipitation (PS),
•Trailing Stratiform precipitation (LS),

Trailing Stratiform MCSs tend to be cold pool forced, produce the most wind events of any MCS archetype and are the most common. Parallel Stratiform MCSs also produce intense cold pools, the updrafts tend to be more vertically oriented, and are responsible for both severe wind and heavy rainfall. Leading Stratiform MCSs are least likely to have strong cold pools.

Bow echoes form a common precipitation shape that evolve either from organized small multicells/supercells, and from larger linear multicells of the Trailing Stratiform or Parallel Stratiform variety. They feature a forward bowing of an intense convective line ahead of a focused rear inflow jet or downdraft outflow and bracketed by at least one bookend vortex. They form in strongly sheared environments. Derechos represent long-lived severe wind episodes and often contain one or more bow echoes.

Multicells can be organized around other common precipitation shapes. A tropical cyclone is a large multicell circularly organized around a common low in the absence of a significant cold pool. Most small, isolated multicells are too small to acquire a linear organization and so we often call these clusters. Large clusters do occur, especially with a nonlinear shaped forcing mechanism.

A Mesoscale Convective Complex (MCC) is a version of an MCS archetype with a roughly circular, and large anvil top. The shape of their anvils, and the relatively light vertical shear environment makes them especially conducive to forming Mesoscale Convective Vortices (MCVs) in midlevels, and anticyclones in upper levels.

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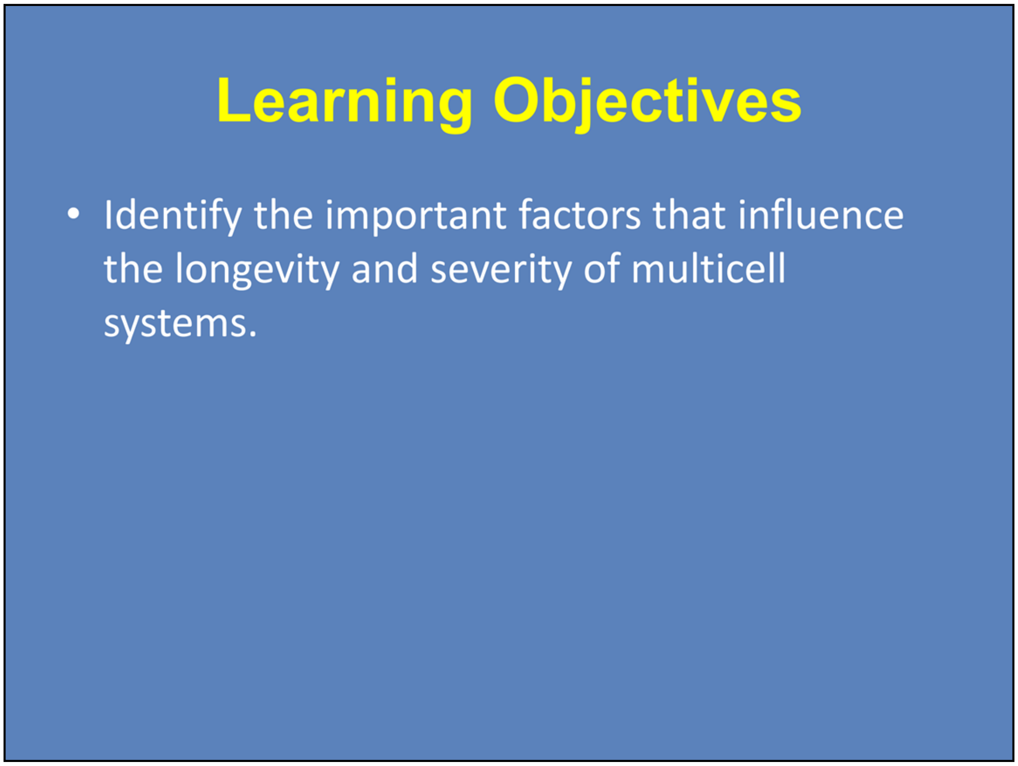
Welcome to the RAC Convective Storm Structure and Evolution lesson on Multicell Longevity and Severity. I'm Justin Gibbs of the Warning Decision Training Division.



Here are the learning objectives for this lesson.

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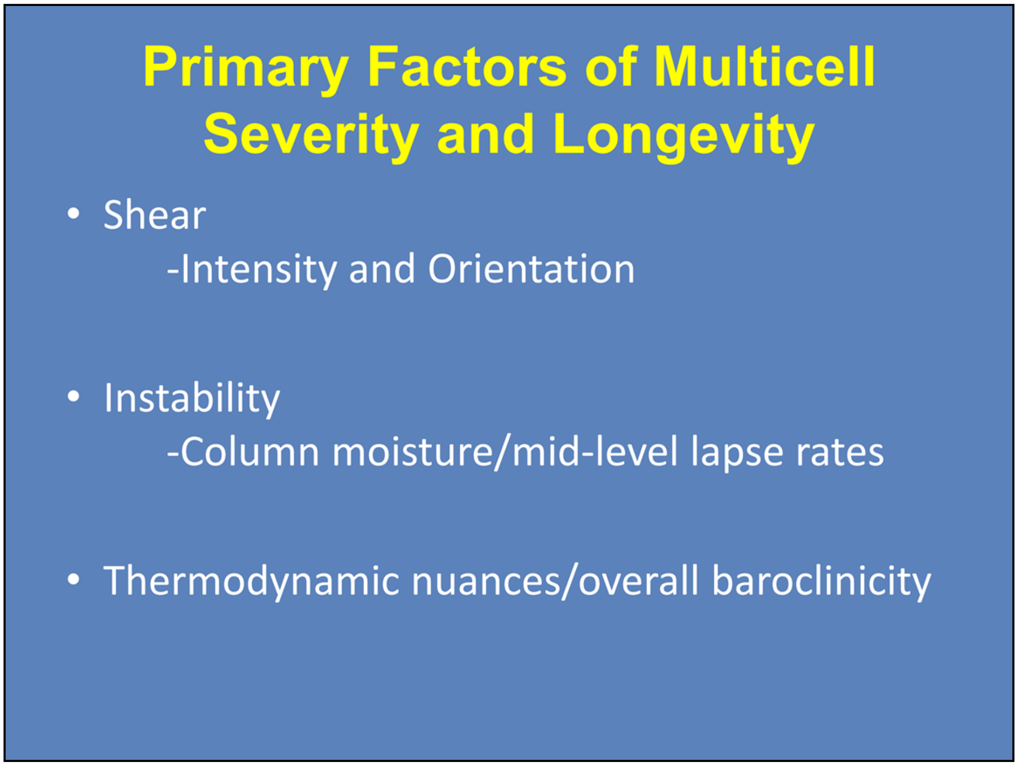

Like most convective processes shear, including the intensity and orientation of the shear

And instability are the main determinants of multicell or MCS severity and longevity

Other more nuanced factors of the envrionments overall baroclinicity and thermodynamic environment, like theta-e difference can also factor in.

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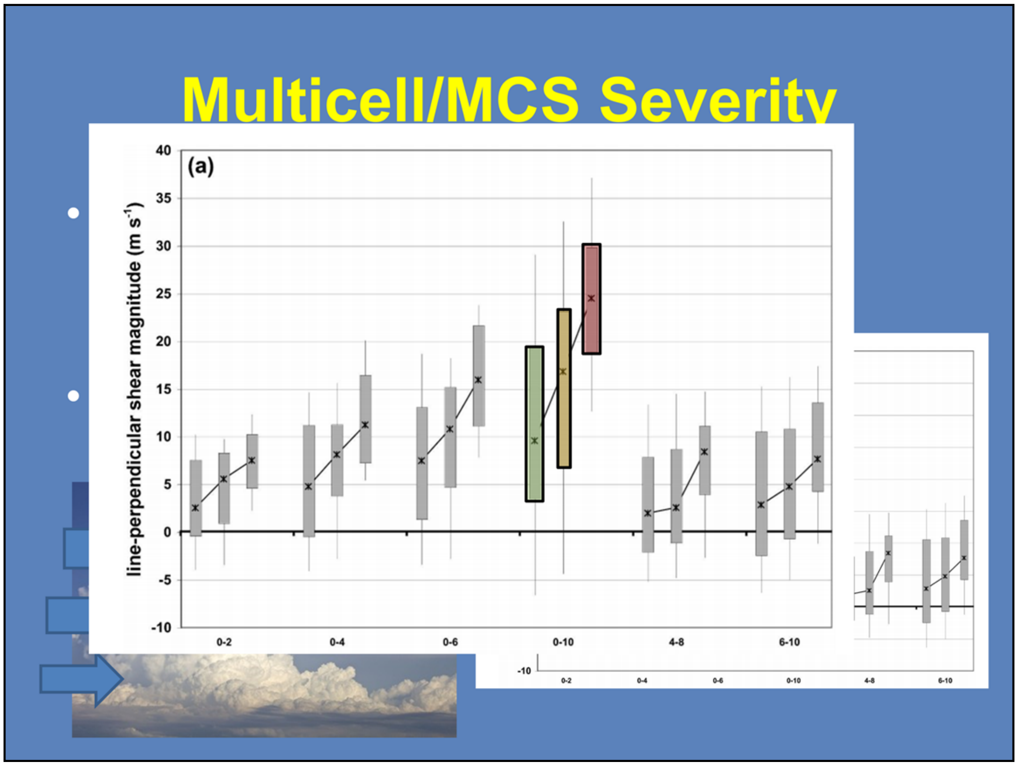


Assuming you have an environment where storms can develop, deep layer shear is the best discriminator between weaker and relatively strong convective systems.

Here,We can take a look at the line perpendicular shear magnitude with cases separated by very strong, or derecho MCS casesSevere, but not high end derecho cases

And sub severe convective systems. There is overlap but a useful signal of stronger shear in the 0-10 and 0-6 km layers showing stronger systems.

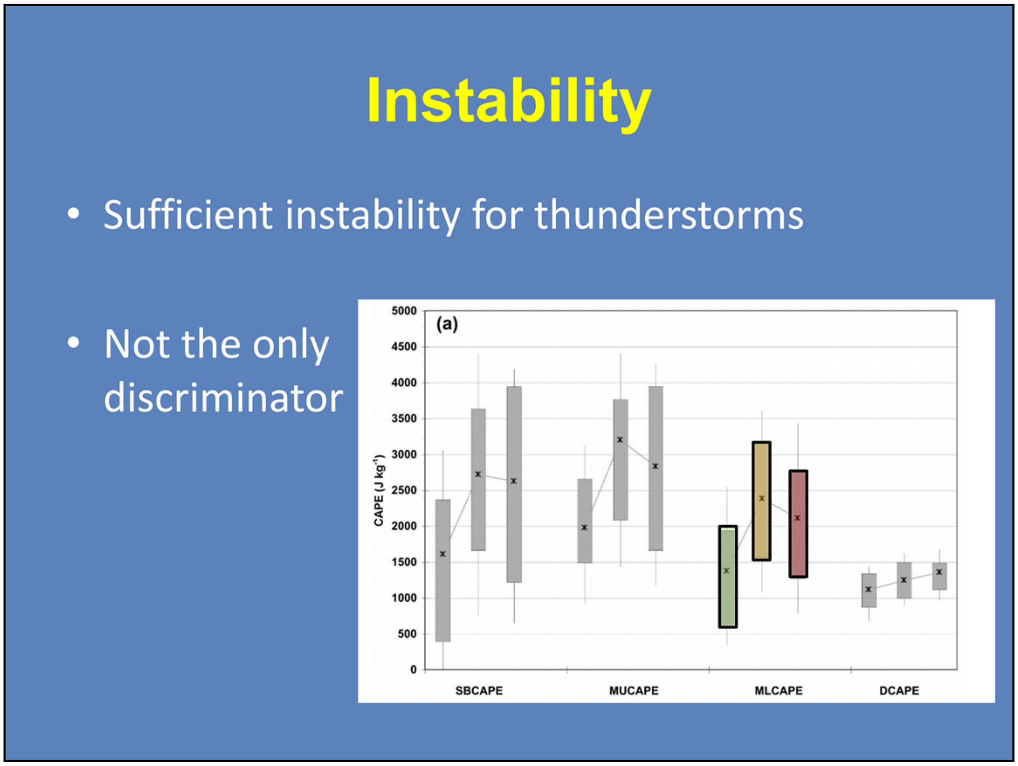
And this is the perpendicular shear, so shear at a 90 degree angle to storm orientation that is working most efficiently to separate the updraft and downdraft processes, and help form areas of mid level convergence, and rear inflow jets. So actual deep layer shear values will likely be higher, but you can evaluate the deep layer shear to see how well it is oriented perpendicular to the line, or anticipated line of convection.

-Cohen et al., 2007

-Evans and Doswell 2001

Gale Et al. 2002

Coniglio et al. 2004



Instability also obviously plays a role in convective intensity with

Derecho and

severe MCS cases showing stronger average instability than

Sub severe cases but

Its pretty clear that there is even more overlap here than the shear cases. So its not the only discriminator when trying to anticipate the severity and longevity of an MCS.

-Cohen et al., 2007

-Evans and Doswell 2001

Gale Et al. 2002

Coniglio et al. 2004

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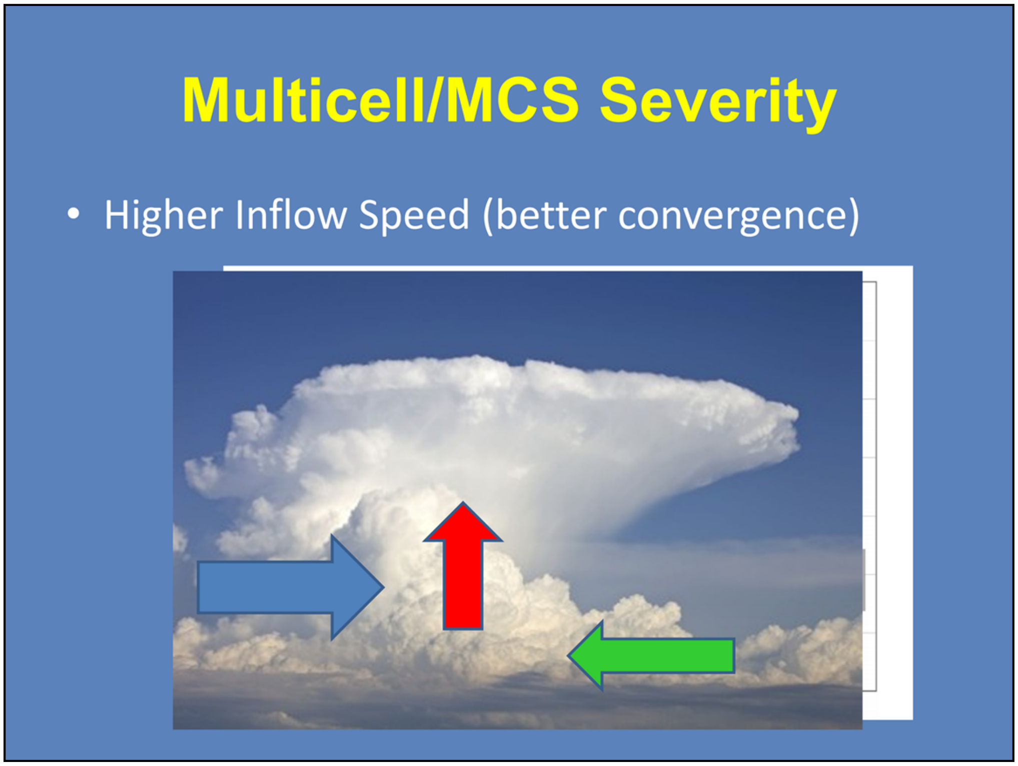

Higher inflow into the system, results in stronger convergence, and produces stronger, more organized systems.

You can see the system relative wind speed for Weak Severe And high end MCS systems show a tendency for storms with stronger system relative winds in the 0-1km layer to produce stronger systems.

So in a storm moving left to right the

Green arrow would be the system relative winds moving perpendicular into the storm, and interacting with the advective and propagating storm system to produce

Better convergence and uplift in the system itself. This can be achieved by higher ambient winds or faster MCS storm motion.

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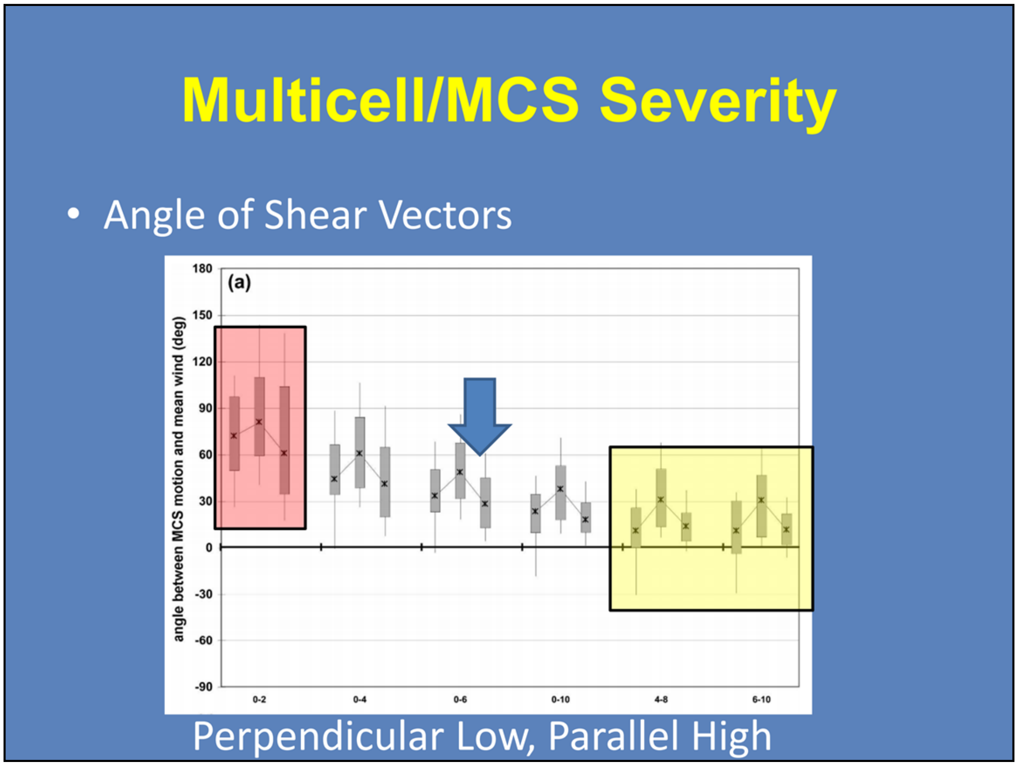

The angle of the shear vectors also has one other subtle component.

With systems tending to cluster around mostly perpendicular shear low and more parallel, but not completely parallel shear in the highest layers of the storm.

You can see that the severe, but not high end MCS systems in the middle stick out here. Which could potentially be explained by a less favorable thermodynamic environment.

Cohen et al. 2004

Corfidi et al 1996 and 2003

Coniglio et al 2004

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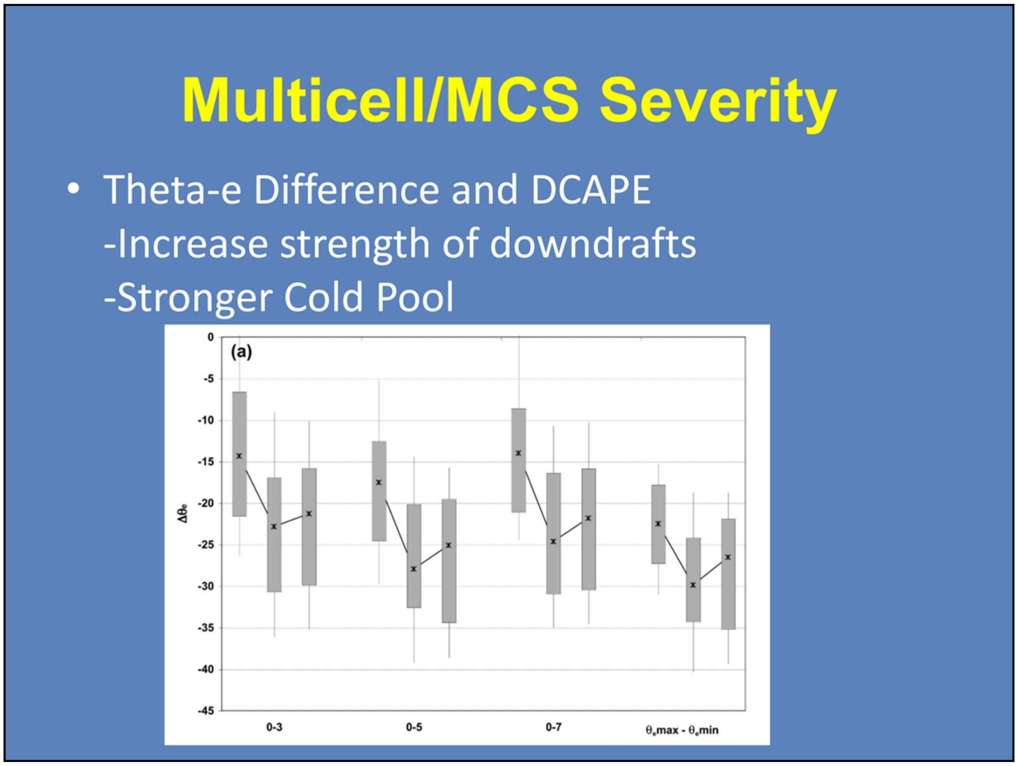
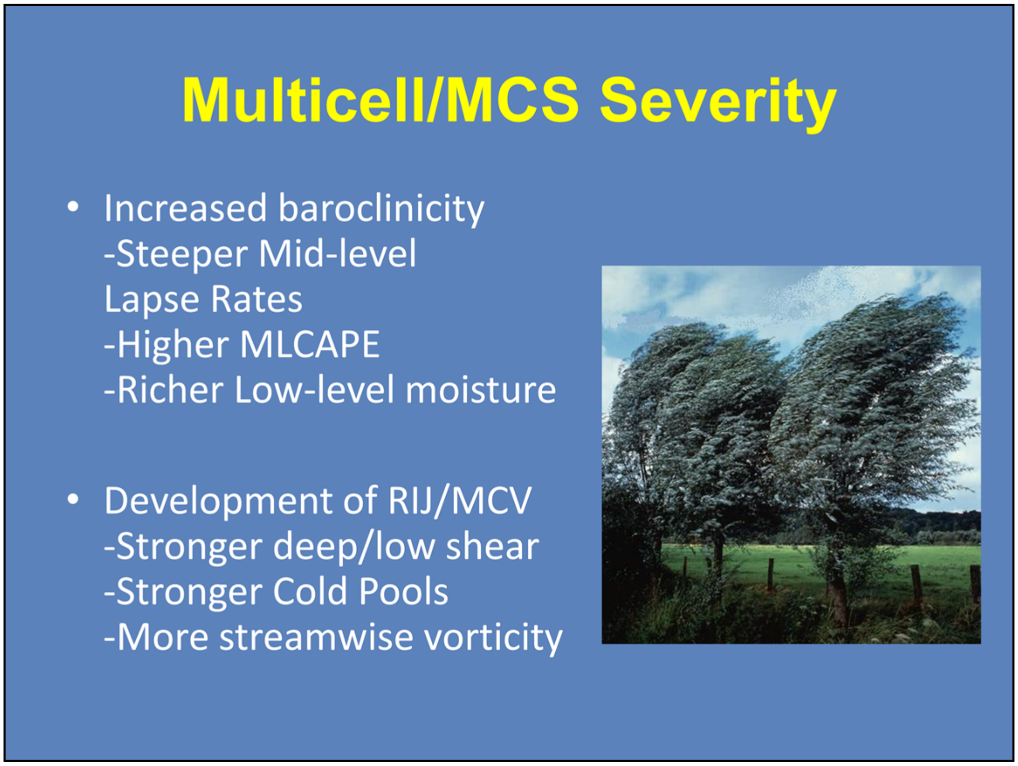


A nuance of the thermodynamic profile, higher theta-E difference, and DCAPE values tend to increase the strength of downdrafts and produce a stronger cold pool. This tends to increase storm organization and severity.

Cohen et al. 2004

Corfidi et al 1996 and 2003

Coniglio et al 2004



Overall you are diagnosing a developing squall line, in the environment out in front of it you are looking for increased baroclinicity, thermal differences. Steeper mid level lapse rates, higher mixed layer CAPE and rich low level moisture. In addition of course to a favorable shear profile.

For development of rear inflow jets or mesoscale convective vorticies, which can lead to locally more significant winds and impacts you would look for stronger deep layer and low layer shear, stronger cold pools and more streamwise vorticity that can be oriented into the updraft.

Atkins and St. Laurent, 2008

Evans et al. 2013

Wakimoto et al. 2006

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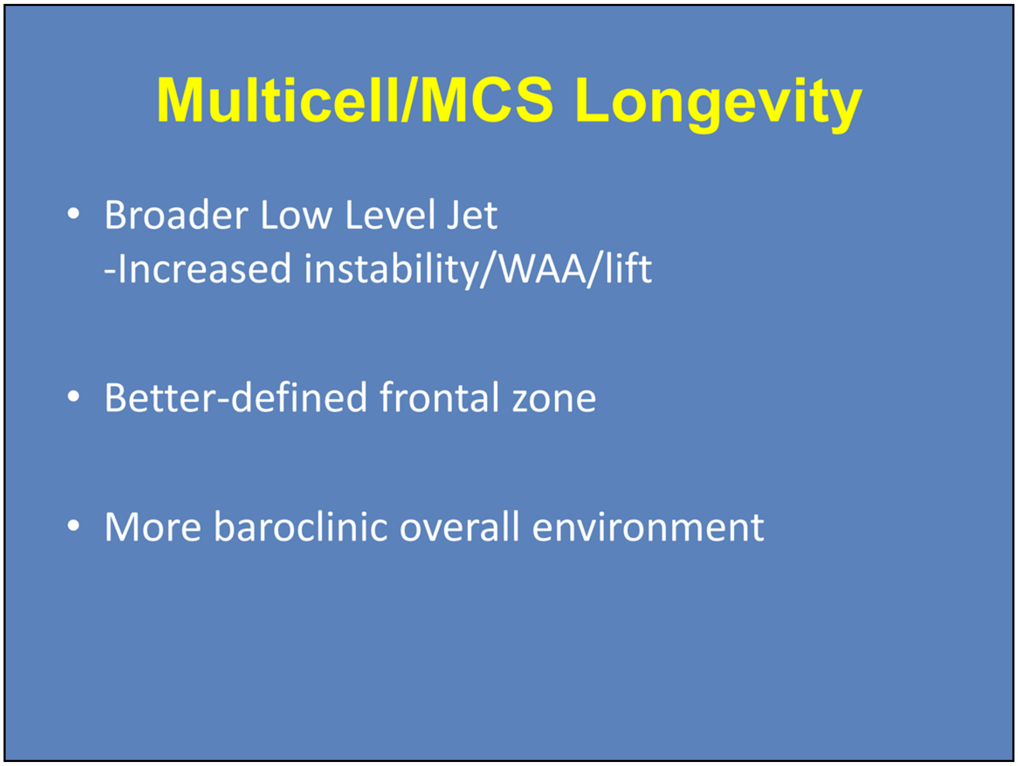

For a longer lived MCS you would want a broader low level jet, leading to increased instability warm air advection lift, basically helping to create a better thermodynamic environment across a broader area

A better defined frontal zone, like an east to west boundary

And just a more baroclinic overall environment that will lead to density differences that will help keep feeding the systems overall instability.

-Coniglio et al. 2010

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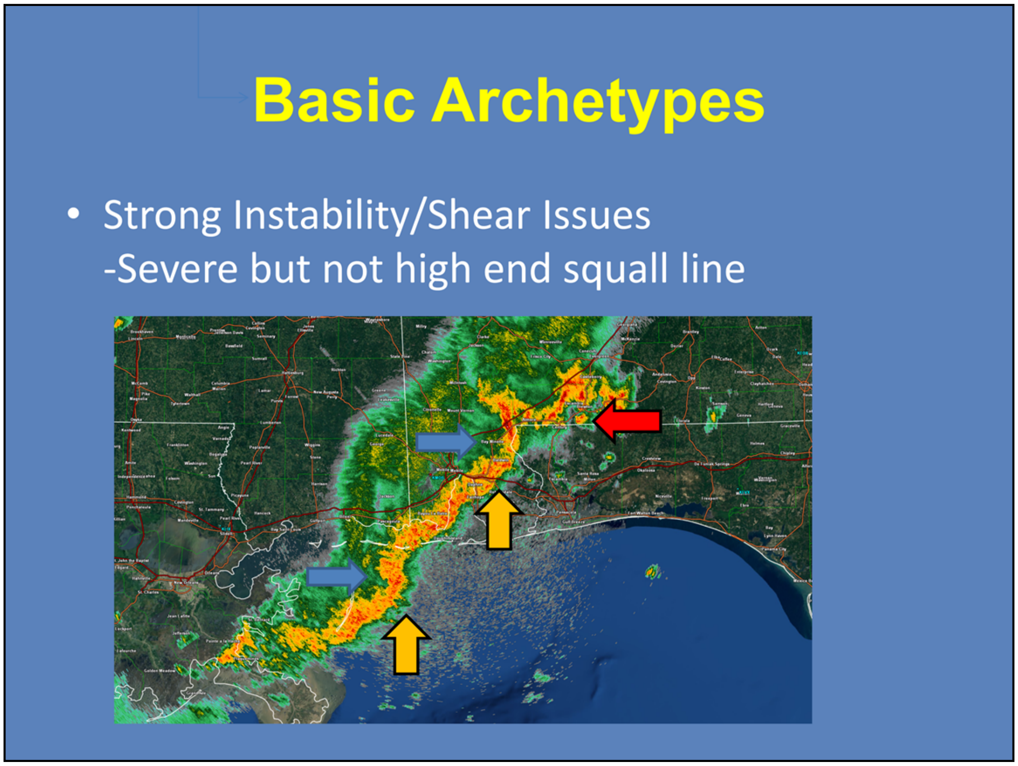

So looking at the basic archetypes of MCS and whats going on with them, in this example the area had strong late spring instability but not perfect upper level shear, along the natural baroclinic zone of the gulf coast. Its moving basically due east.

There are a couple of pretty well defined bow echoes and there is a pretty sharp reflectivity gradient along the leading edge of convection, the outflow from the storm doesn't look like its outrunning the convection.

This is the type of system that, especially in those bowing segments, may be producing severe winds, you would want to evaluate the depth of convection and see what the base velocities look like, as well as if there is any mid altitude radial convergence or other processes going on that would suggest severe potential.

I also see this area along the florida/alabama border that at a minimum is going to produce a thermal gradient or some sort of density diference for that bow to run along. That will increase streamwise vorticity and would be an area to watch closely for damaging winds or embedded tornadoes.

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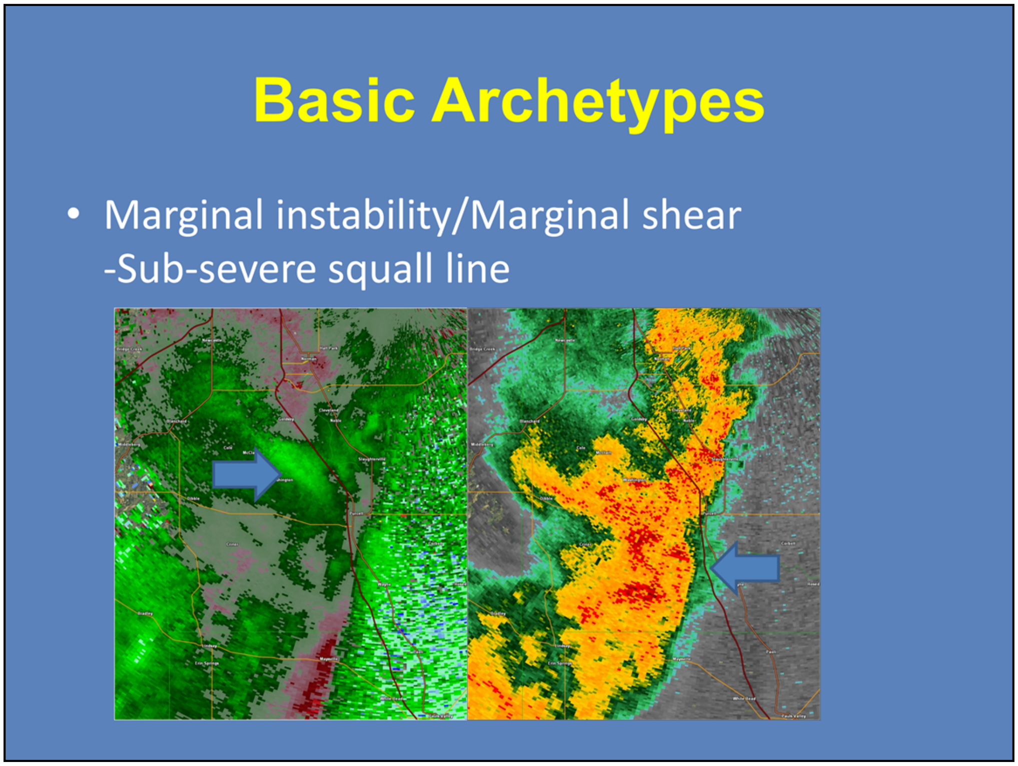

Heres an example where the shear and instability are both marginal. I suspect it's a little elevated as well as the boundary layer stabilizes. The velocity data, despite some beam angle issues are still quite weak,

some 25 to 30 knot inbounds but almost no velocity signal further down the line

Some reasonably tight reflectivity gradient but not any obvious signs of bowing or other reflectivity signatures that would make you concerned about local wind damage.

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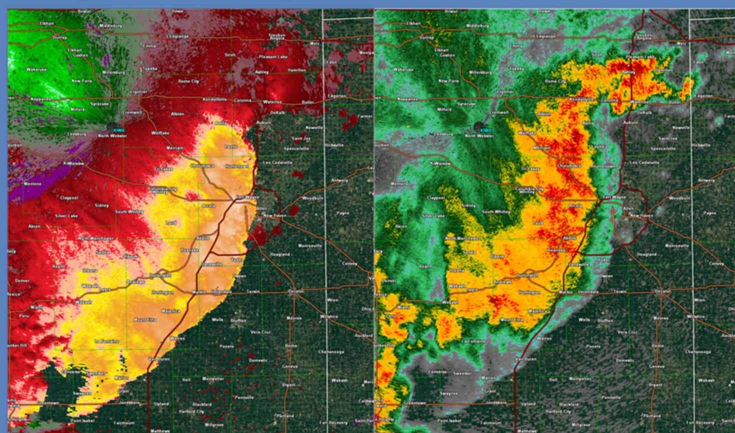
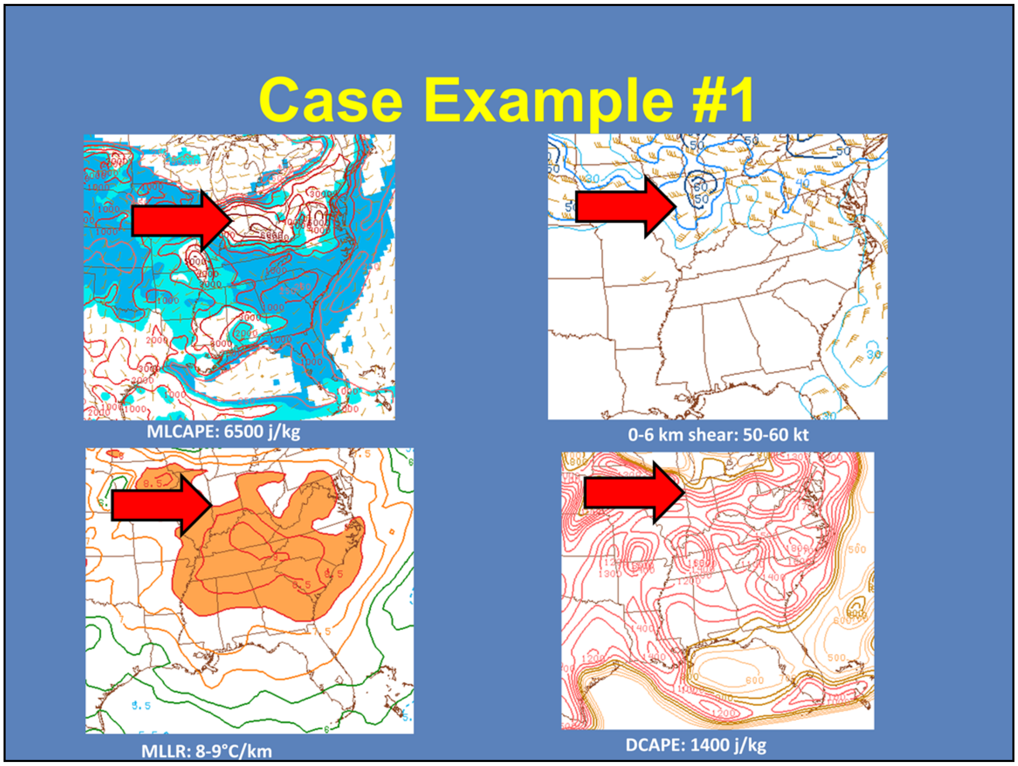
Basic Archetypes

- Strong Instability/Favorable Shear
  -High end/Progressive Derecho



Where in this case you have a broad bowing segement with 60-70-80 kts inbound that signals a high end progressive derecho in progress.



Looking more closely at this case at about the time of the last image the Mixed Layer CAPE values were extreme, 5500 to 6500 j/kg.

0-6km shear was very favorable at 50-60 kts and oriented perpendicular to the convection that was running northeast to southwest.

Mid level lapse rates were also extremely favorable at 8 to 9 degrees celsius per kilometer.

Downdraft CAPE was also strong at around 1400 j/kg

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Web Object

Address:
http://training.weather.gov/wdtd/courses/rac/severe/objects/iwx-shear/

With the advection, deep shear, and cell propagation all working together in such an extreme environment it created the perfect situation for wind damage. Which is exactly what happened on the afternoon of June 29, 2012 with thousands of wind damage reports across the Ohio Valley. Forward storm motion is 50 to 60 kts. The fort wayne airport recorded a 91 mph wind gust when this system stormed through.

-Mahoney et al. 2009

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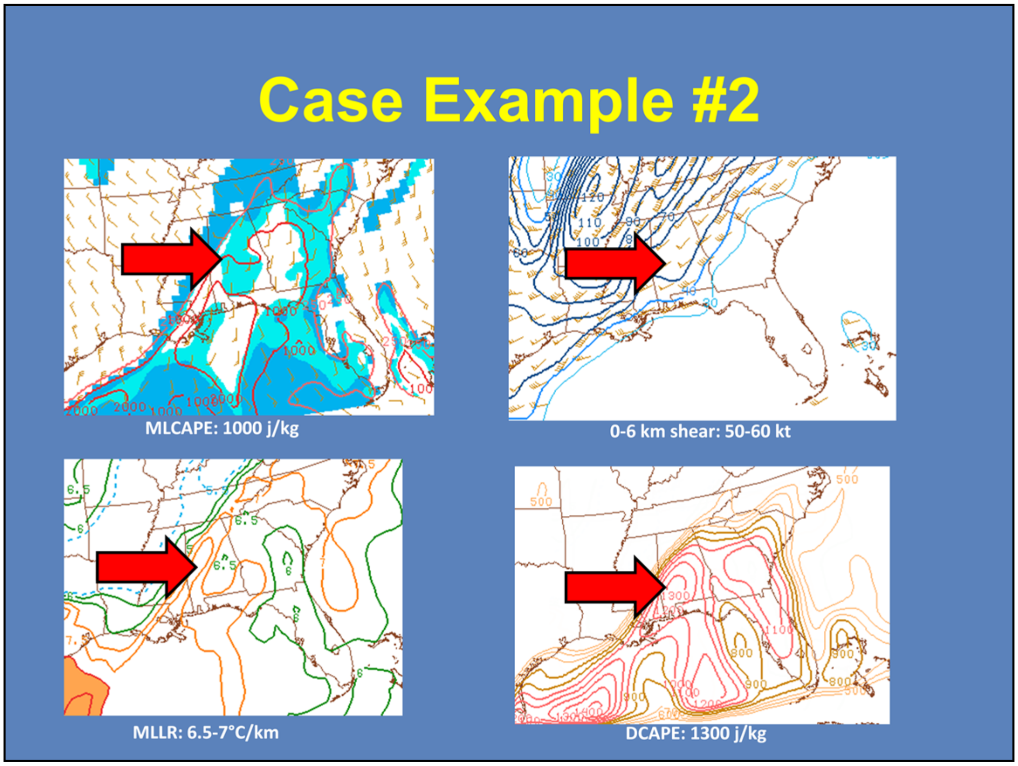

Looking at another case, the instability here is clearly lower with Mixed Layer CAPE values around 1000 j/kg

0-6 km shear is pretty good, with a pretty strong trough appearing to be approaching.

Mid level lapse rates are 6.5 to 7 degrees Celsius per kilometer, not unfavorable, but not as good as our extreme example earlier

And DCAPE values are pretty good, about 1300 j/kg

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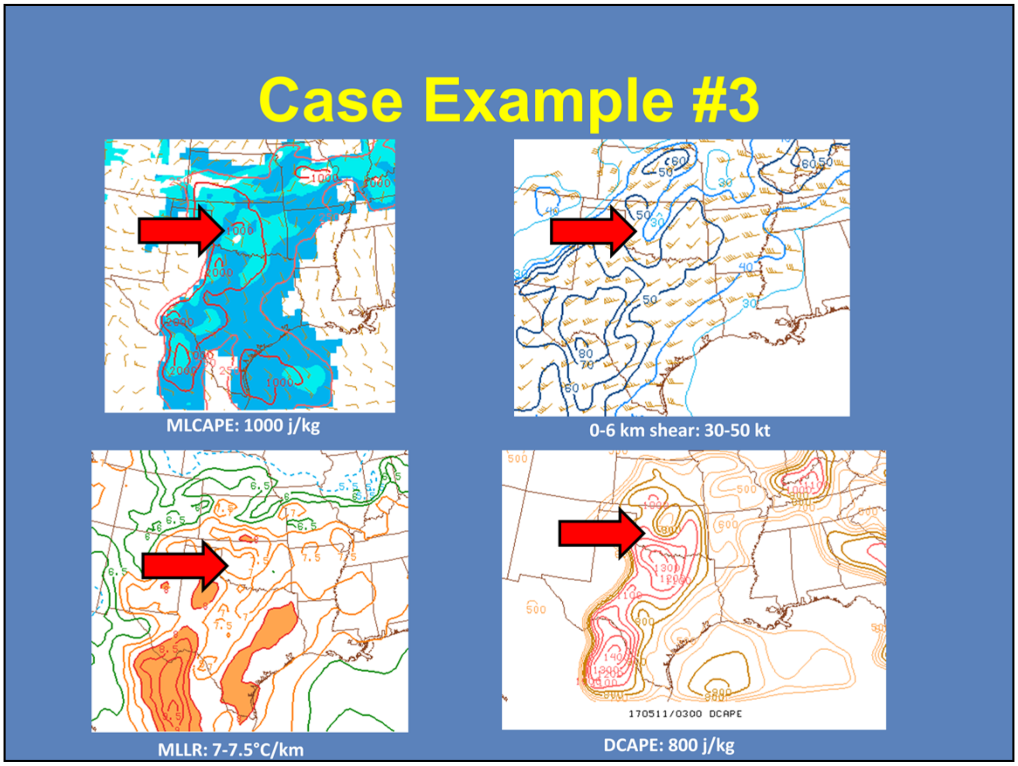
Web Object

Address:
http://training.weather.gov/wdtd/courses/rac/severe/objects/bmx-shear/

But in this case the deep shear is running basically parallel to the squall line, there are a few spots where the outflow seems to be outrunning the MCS and the cell propagation and advection are a little bit out of sync. Now this line is still producing damaging wind, but its more 60-70 mph and less of the 80-90-100 mph variety.



Another example, this time with even less Mixed Layer CAPE and very strong Convective Inhibition

Deep layer shear is 30 to 50 kts, not bad, but it will depend on the orientation of the storm.

Mid level lapse rates 7 to 7.5 c/km not bad.

Weaker DCAPE in this case about 800 j/kg.

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Web Object

Address:
http://training.weather.gov/wdtd/courses/rac/severe/objects/tlx-shear/

The result, despite near the red arrow maybe a decent bowing segment early it quickly fans out with the outflow shooting out ahead of the storm. The blue arrow points to where in that portion of the storm the outflow had more clearly separated from the convection. Without a change in the environment this MCS is on the decline quickly.

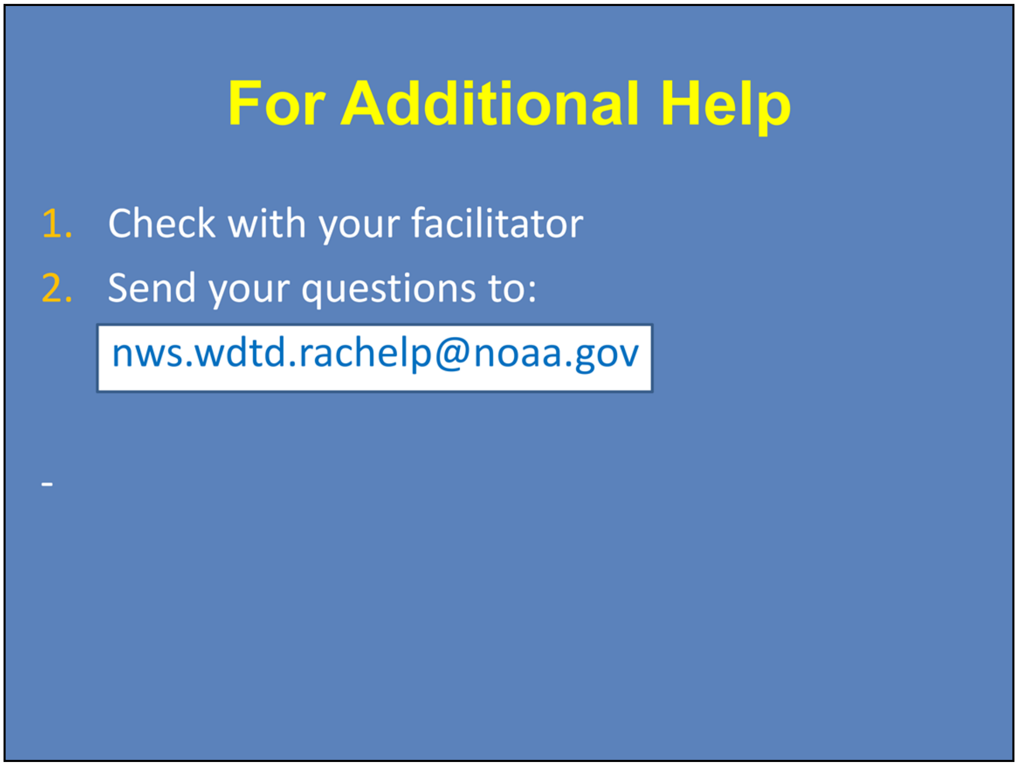
Primary Factors of Multicell Severity and Longevity

- Shear
  - -Intensity and Orientation
- Instability
  - -Column moisture/mid-level lapse rates
- Thermal/Shear nuances and overall baroclinicity

So in summary, its both pretty complicated and pretty simple. Its shear and

Instability. With perpendicular shear values in higher numbers in the deep layer leading to stronger systems. And higher values of instability generally leading to more vigorous updrafts and downdrafts and stronger systems.

But there are also more nuanced aspects to what sets weaker systems apart from stronger systems, such as theta-E difference, storm inflow winds, and overall baroclinicity including the mid level lapse rates.



For additional help, check with your facilitator or send your questions to the listserv e-mail address here.

Welcome to the RAC Convective Storm Structure and Evolution lesson on Multicell Motion.

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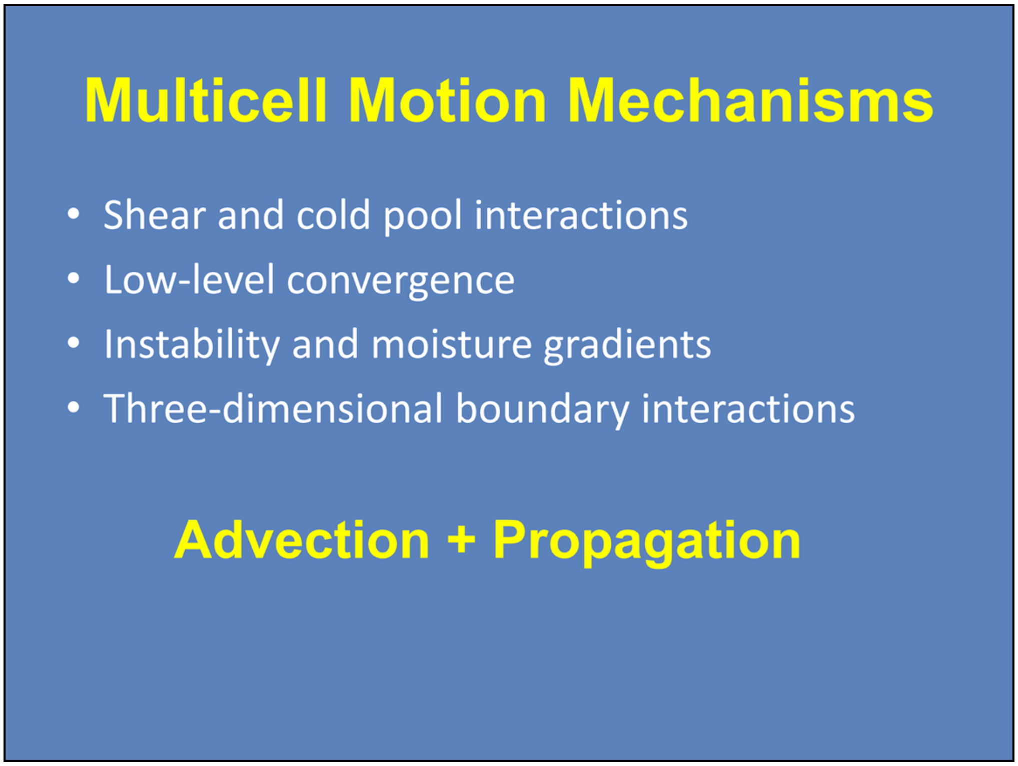
Learning Objectives

- Identify the primary vectors of motion of backbuilding and forward propagating multi-cell systems
- Identify what determines whether forward or backbuilding propagation is more likely
- Identify the role of boundary interactions on the motion of multi-cell systems
- Identify the role of instability gradients on the motion of multi-cell systems

This is the learning objective this lesson



Determining the motion of multicells is more challenging than for ordinary storms or supercells because there is usually quite a bit at work. Multicell motion is governed by shear and cold pool interactions, which

Influences the amount of low level convergence

Instability and moisture gradients can impact multicell storm motion

As can three-dimensional boundary interactions. The bottom line though multicell motion boils down to

Advection and propagation.

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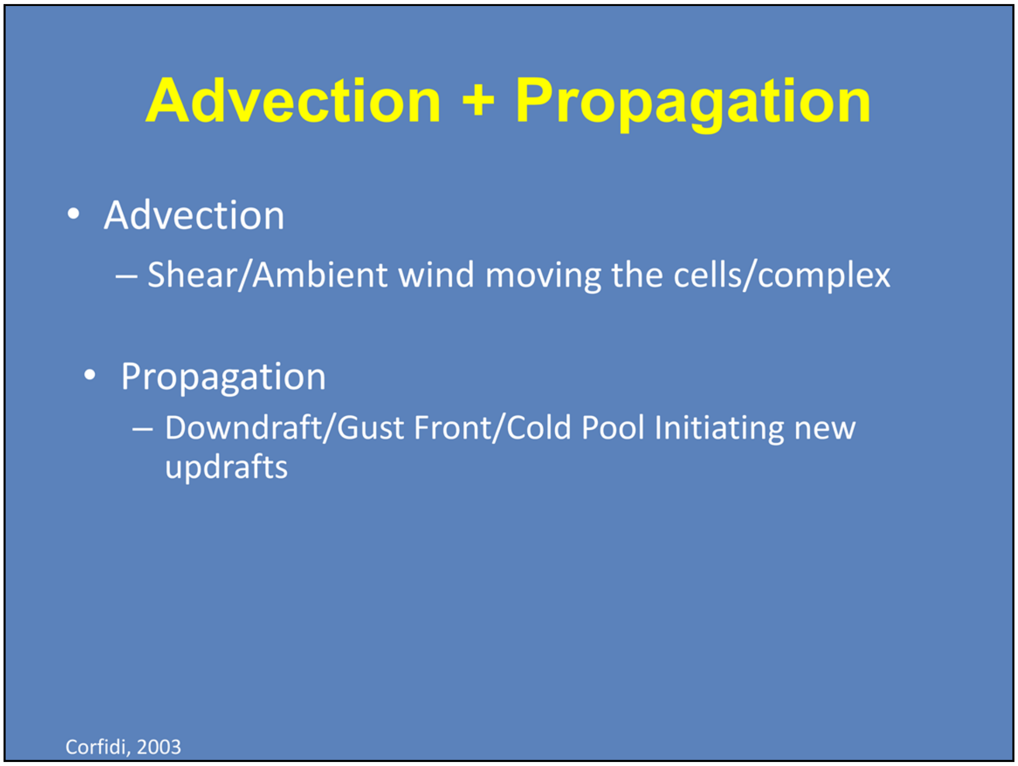
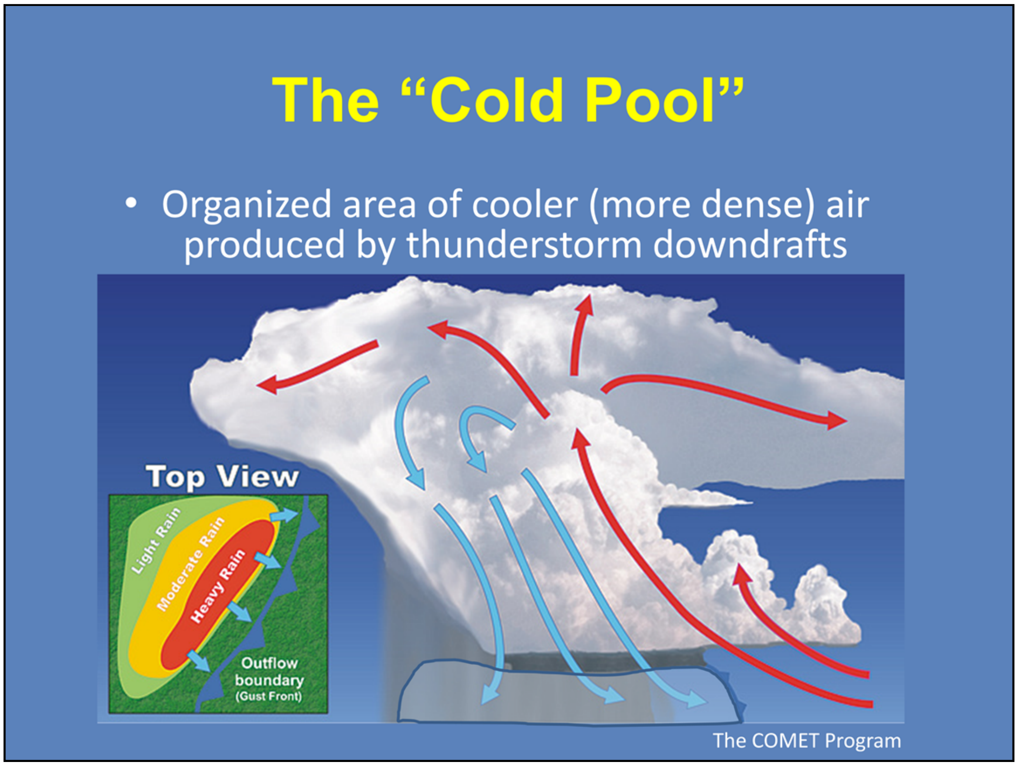


In the multicell or mesoscale convective system world Advection refers to shear and ambient winds moving and pushing the cells or MCS thorugh the atmosphere

While propagation is the apparent movement of the system due to downdraft and gust front or more completely, the cold pool initiating new updrafts.

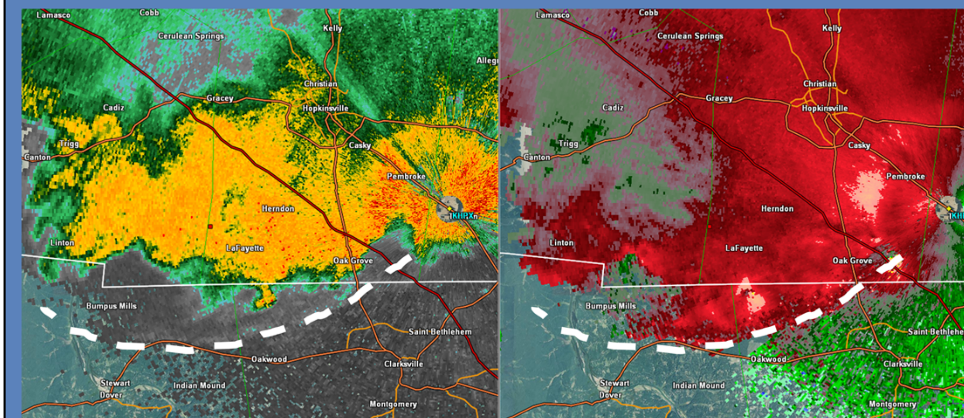
Corfidi, 2003



Just incase you are unfamiliar or forgot along the way, when we say "cold pool" we are just talking about the somewhat organized area of cooler, and more dense air produced by thunderstorm downdrafts.

Downdrafts from multiple cells can form a somewhat organized density current that helps drive storm and storm system motion.

The Cold Pool



On radar you can identify a cold pool, or outflow fairly easily. This image shows reflectivity on the left and storm-relative velocity on the right, so velocity from the perspective of the storm. Note that the inbounds are well ahead of the reflectivity/updraft region,

in this case the cold pool has moved out ahead of the convection.

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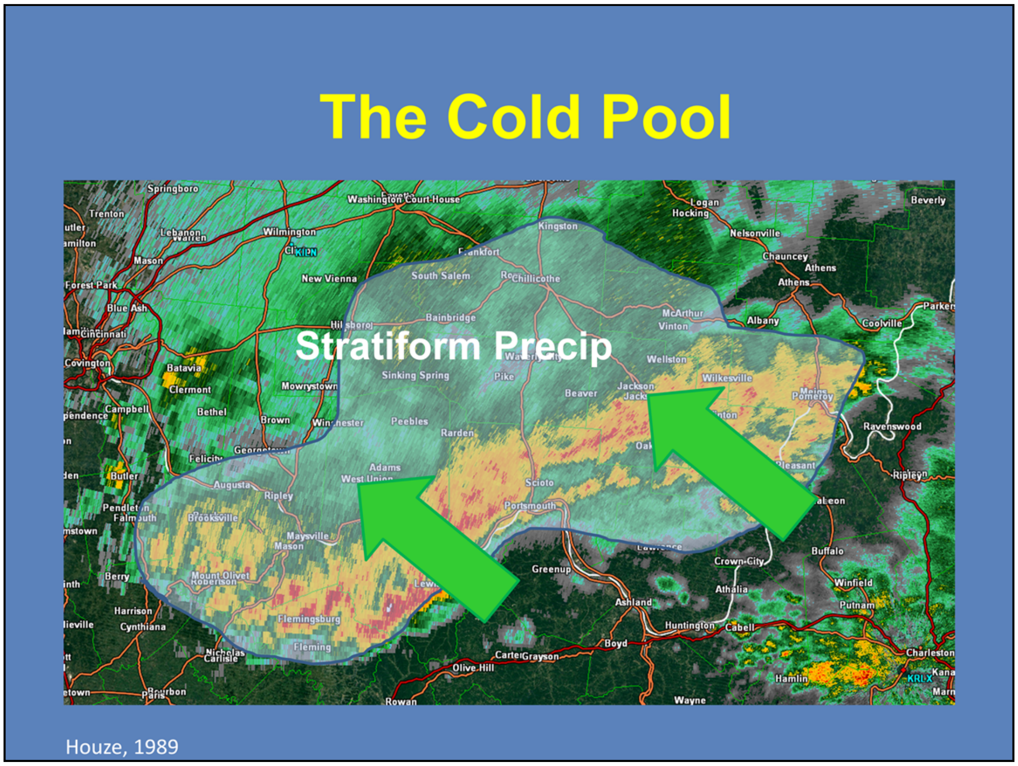
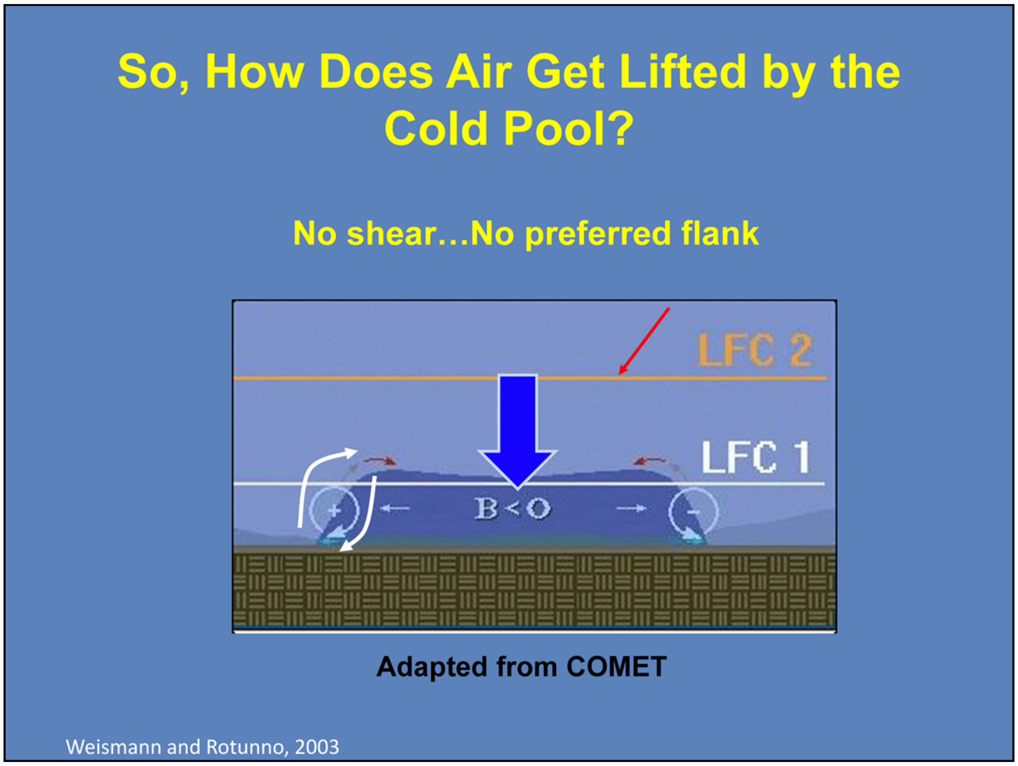


Stratiform precipitation can also help identify the location of the cold pool.

Here the approximate location of a mesoscale cold pool is outlined

as ambient winds move warmer air over the more dense cold pool stratiform precip develops.



New storm development favors the side of a multicell storm where the ambient winds are positive relative to the orientation of the boundary.

If no shear, or very minimal shear is present there will be no real preferred flank for new cell development.

A downdraft produces a new cold pool.

And the downdraft and cold pool produce low level convergence which can lead to new updrafts.
The height of the LFC, or height to which parcels will need to be lifted to develop new thunderstorms, will then determine whether or not new storm growth occurs. A lower LFC height, such as LFC1 labeled here would probably favor new storm growth with a cold pool at the depth shown.

But if it is at LFC2, new cell growth would appear unlikely in the absence of some other new forcing.

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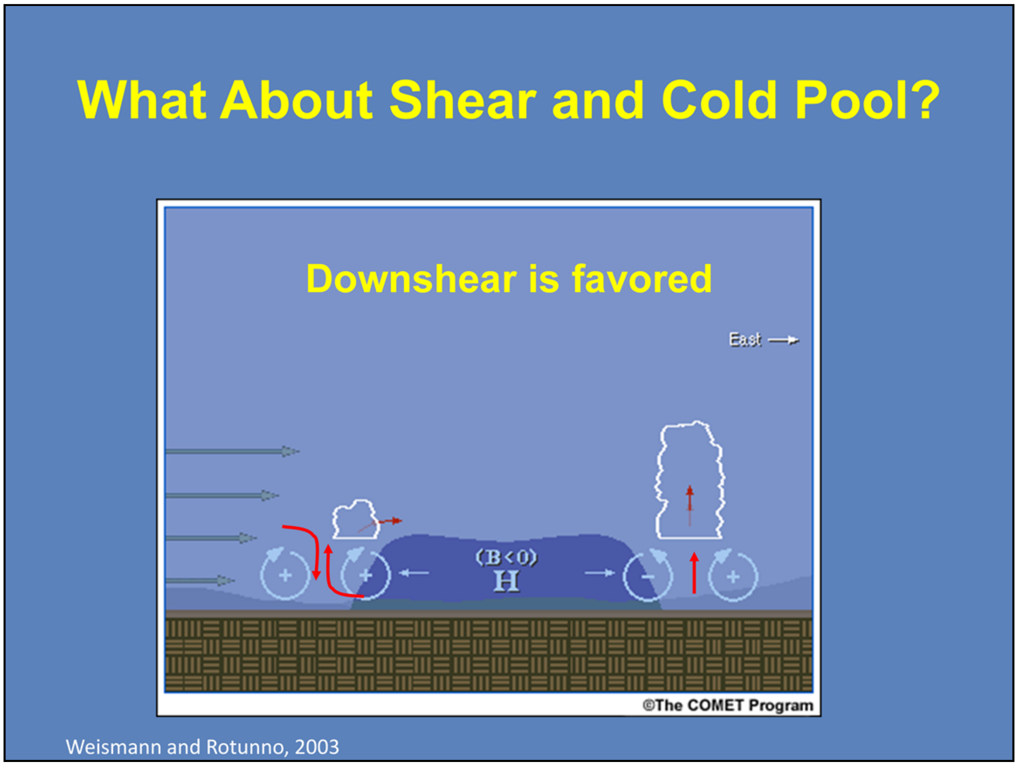
Web Object

Address:
http://training.weather.gov/wdtd/courses/rac/severe/objects/nqa/

Here is an example of a low shear multi-cell system. Note how the outflow/cold pool expand almost perfectly symmetrically, in some areas new cells go up, in others they don’t but there is no real preferred flank and this cluster never organizes in any appreciable way.



If we add shear things get a little more interesting. In this example we will see convection favored on the downshear side.

So in this image the cell on the right.

Cold pool winds, to the left of the cell with the negative sign, interact and converge with ambient winds to the right of the storm to favor development on that side.

On the left side the shear profiles are such that downward motion is favored on the left flank. Fast forward motion of the overall cold pool could help limit convergence on this flank, as could a convective feature like a rear-inflow-jet.

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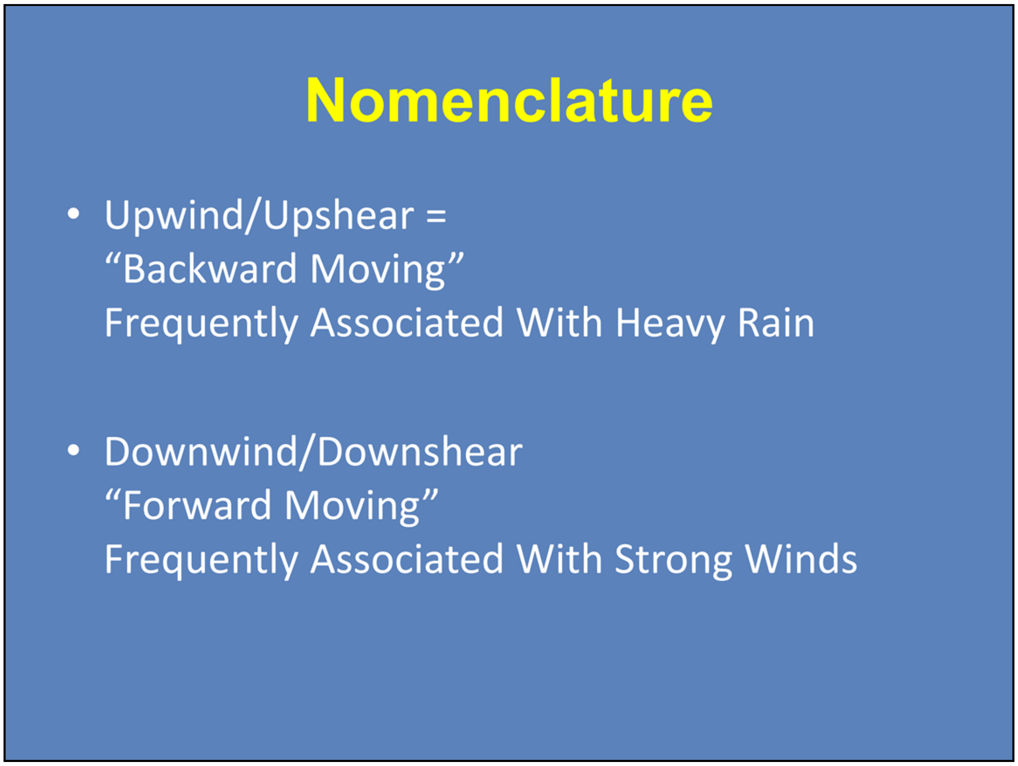

I always have a little trouble keeping MCS nomenclature straight. So just in case you do too. Upwind/Upshear systems are the type that appear "backward moving" or backward propagating. This results in slow ground-relative system motion and these systems are frequently associated with heavy rain, for more reasons than just the motion and we will get to that in a bit.

Downwind/downshear systems are forward moving, frequently associated with strong winds. The prototypical derecho or squall line would be a form of a downwind/downshear propagating multicell clusters.

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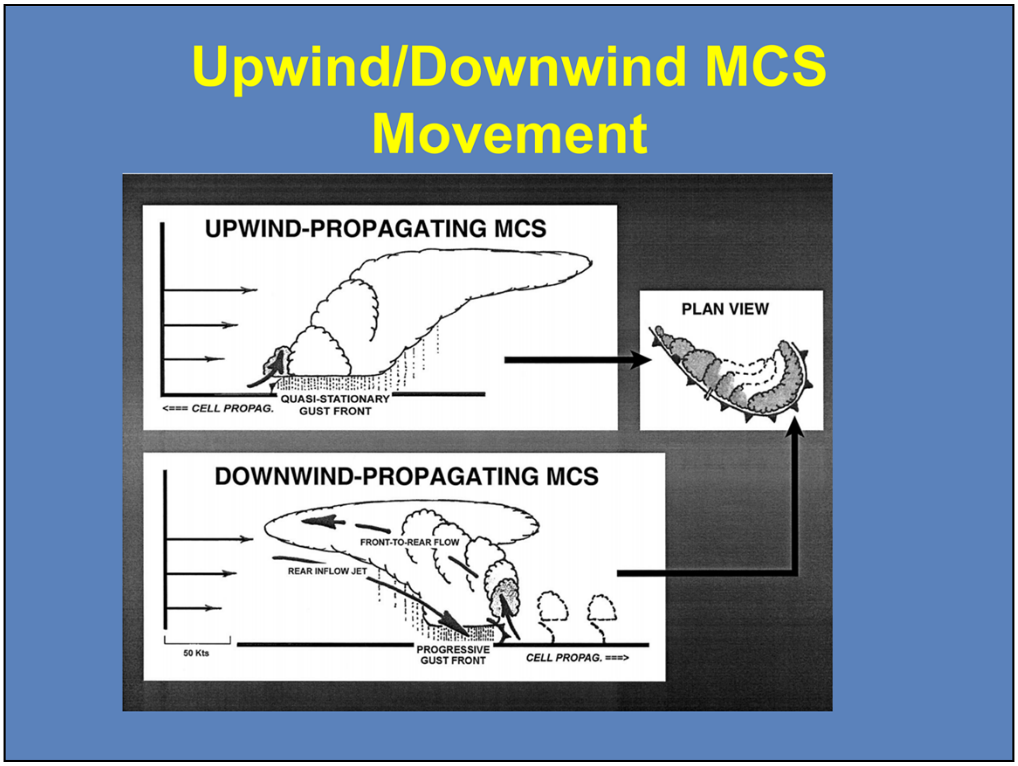
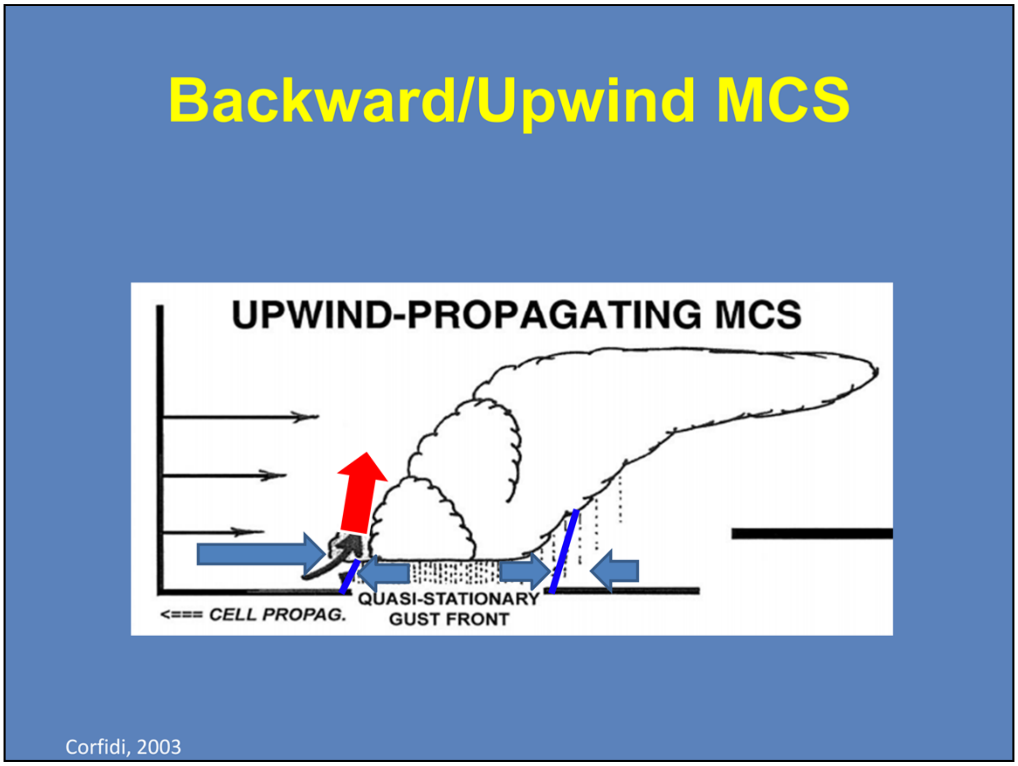


Heres a side view schematic of upwind and down wind MCS movement. The upwind example our rainier, slow moving MCS, while the downwind system being our more classic squall line.



Looking more closely in an upwind propagating MCS. We have our cold pool

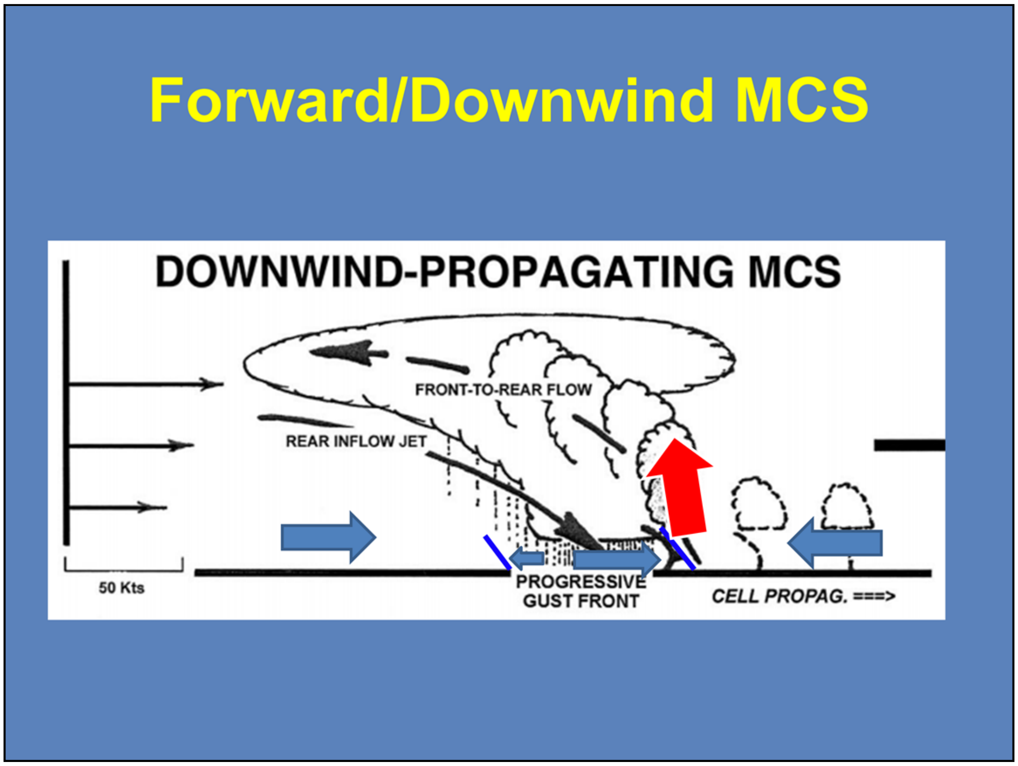
Roughly bounded here either side of the precipitation with outflow in both directions up against the cold pool boundary.

Shear to the left of this system might be relatively strong,

Compared with the ambient winds to the right, since its moving slowly on a quasistaitonary cold pool. Its not surging forward against the headwinds.

This will lead to the strongest convergence on the back side of the storm.

So you have minimal advection, in a system largely moved by propagation.



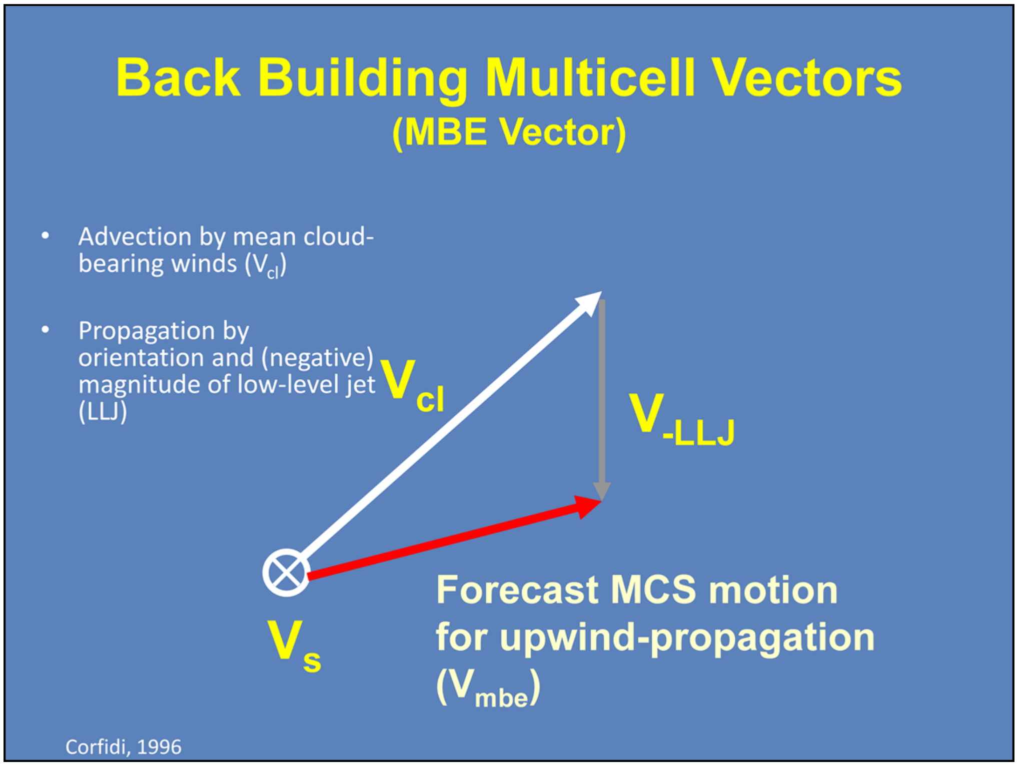
In the downwind example

You have the cold pool boundary

And your downdrafts may be close to the same magnitude but the forward motion of the gust front gives the front facing winds more push

So the winds push harder against the oncoming winds

And new updrafts are favored on the right side. So you have the advection and propagation of this storm moving it forward.



We can estimate multicell storm motion through two different approaches. For back building multi-cell systems we can use the MBE vector, MBE stands for Meso-Beta scale Element, since that's the scale these systems usually operate on.

You first take the advection of the system by taking the speed of the the mean cloud bearing winds, usually 850 to 300 mb.

Then for your propagation you take the negative vector, or the opposite of the low level jet. Usually your 850 mb winds.

The resultant vector will give you're your forecast MCS motion for back building or upwind propagating Multicell systems. Normally this is available as either MBE Vector or Corfidi-Upwind in AWIPS.

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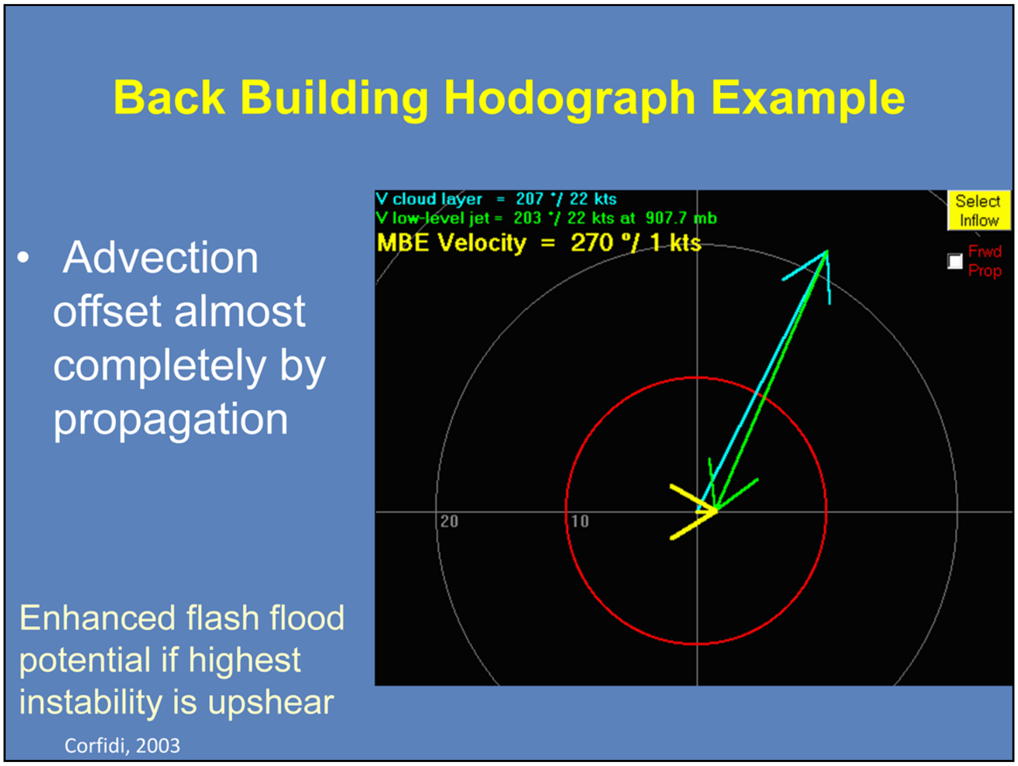
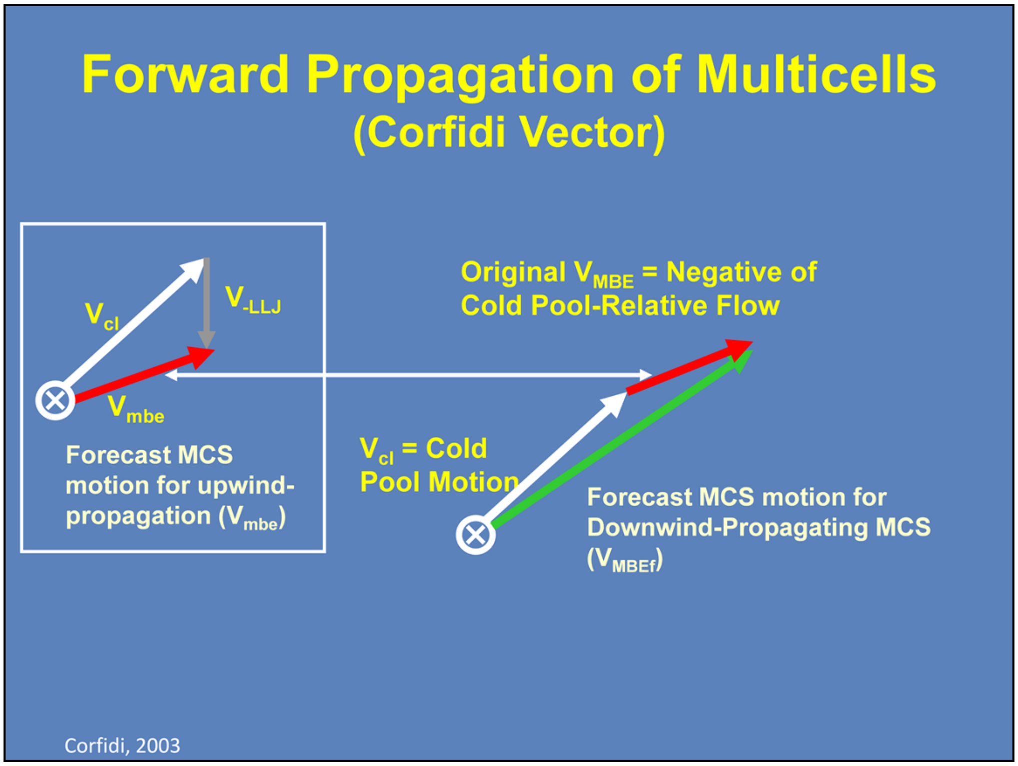


Heres an MBE vector example, in this case the mean cloud layer winds are 207 at 22 knots, and the low level jet is 203 at 22 knots,

So the forward advection of the system is almost completely offset by propagation.

The flash flood potential is higher if the highest instablity is upshear in the direction the system is working towards.



For forward moving, or downshear propagating systems you can use this technique.

We still use the original MBE vector

But we add it to the cold pool motion, which is roughly the mean winds of the cloud bearing layer. So it captures the same propagation as the original MBE vector but adds it to the larger scale advection, since propagation and advection are added in a forward moving MCS.

This vector is usually available as Corfidi-Downwind or Downwind MCS in AWIPS.

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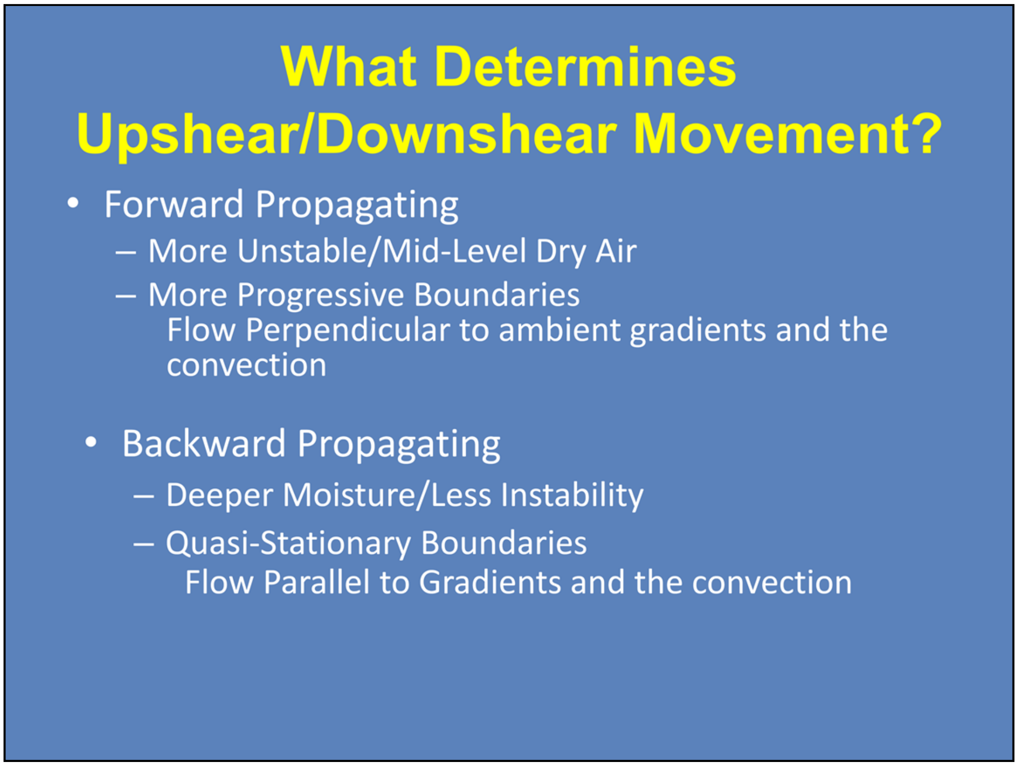


So what determines whether your system will follow upshear or downshear movement?

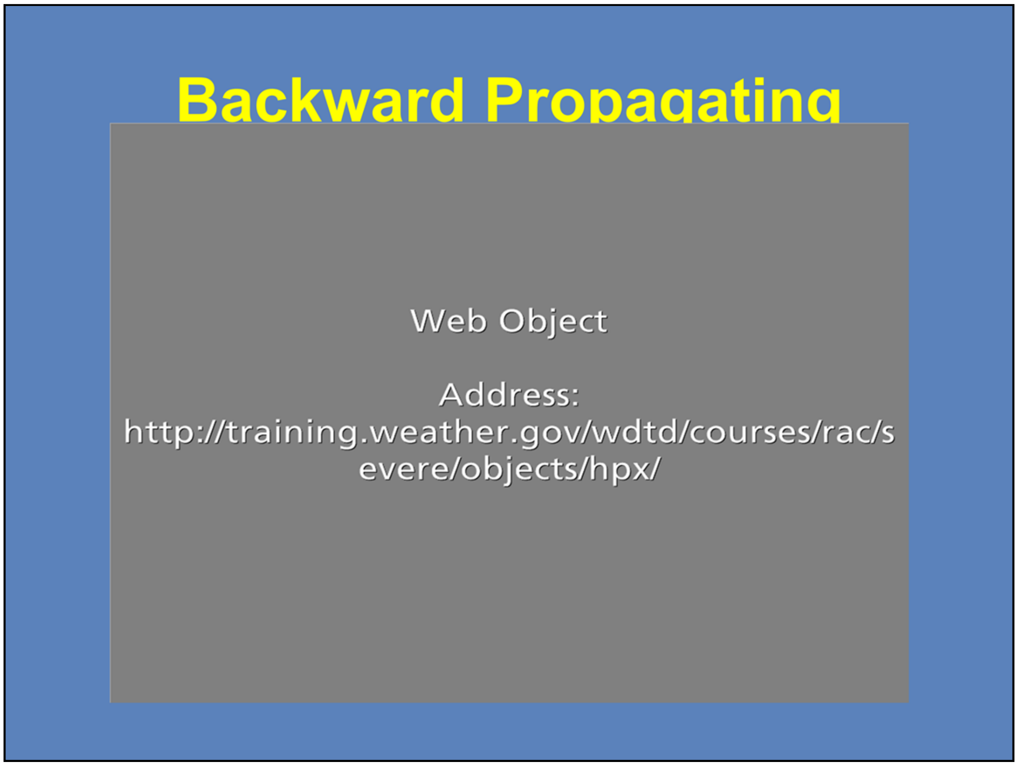
Forward propagating systems tend to occur in more unstable air, with drier mid level air that makes the cold pool stronger. They also tend to occur along more progressive boundaries which are a result of the fact the larger scale flow is perpendicular to ambient gradients and existing convection.

Backward propagating systems tend to occur in deeper moisture, with less overall instability, another reason they are a big flash flood concerns. They tend to be located along quasi-stationary boundaries due to the fact the flow is parallel to gradients and the convection.

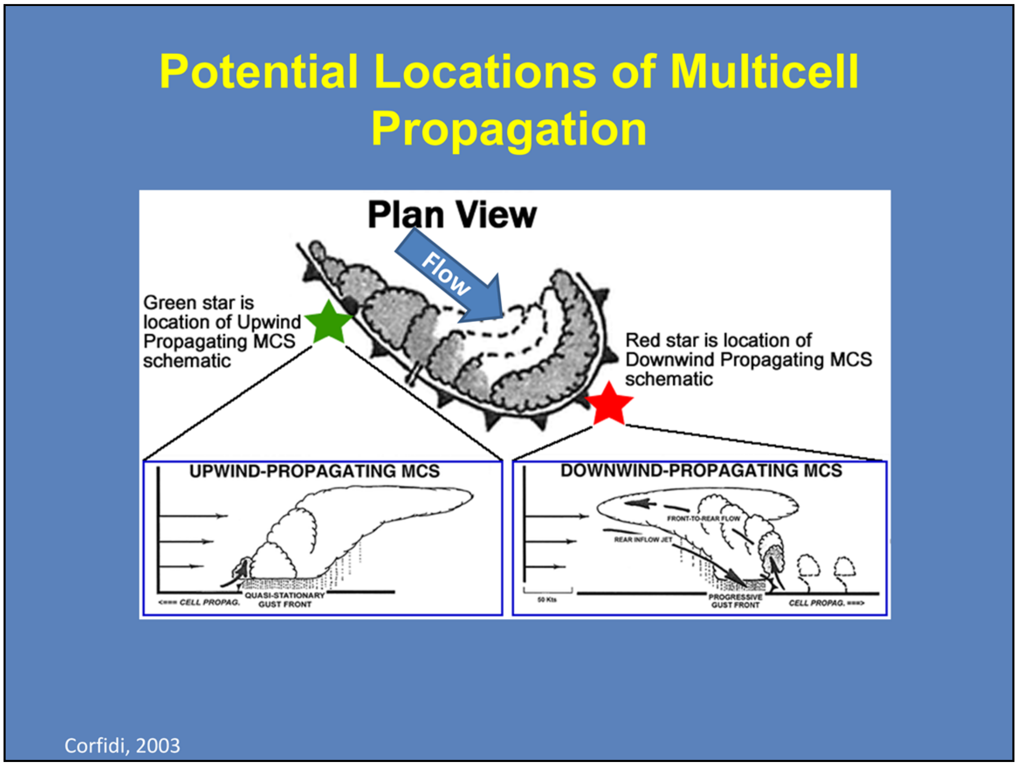
So it’s a combination of shear and instability, which of course are the processes producing advection and propagation.

Heres an example of a pretty solid forward propagating MCS.

Notice the sharp reflectivity gradient along the leading edge and stratiform precipitation behind the system. This complex produced several wind damage reports as it swept across Kentucky and West Virginia.



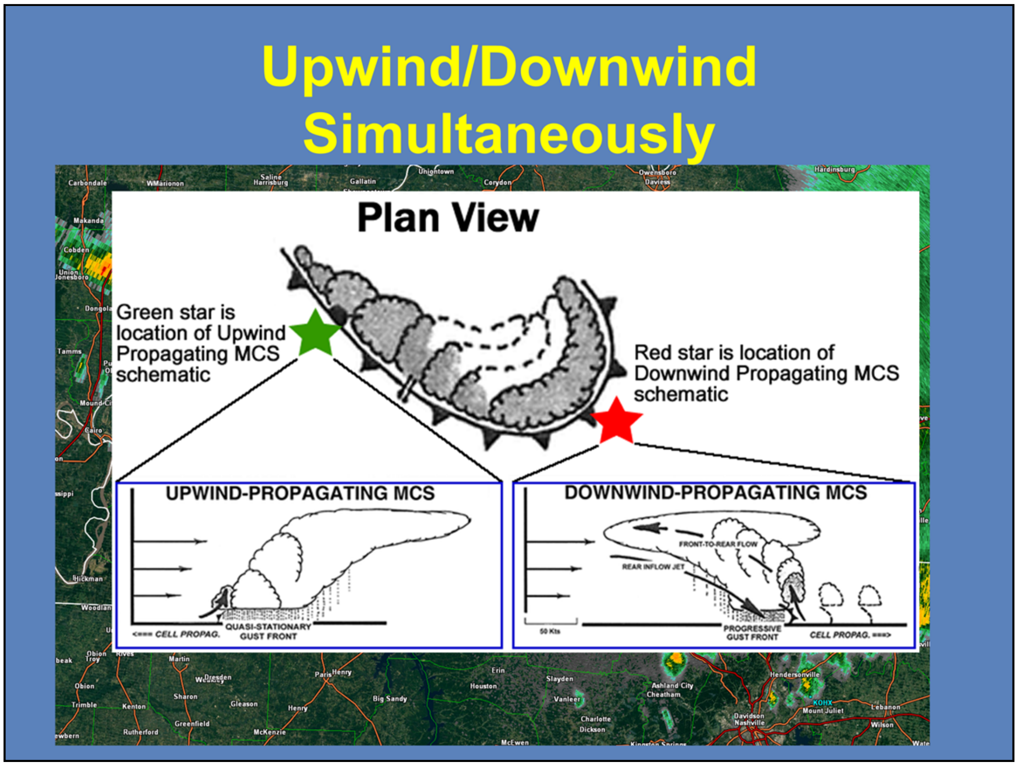
Here is a backward propagating MCS, the new cells developing on the left side of the complex. The mean flow is from the west to northwest and new cells are developing into the mean wind.



Can both occur at the same time?

Of course, it wouldn't be operational meteorology if it were cut and dried! In this case

With ambient flow out of the northwest the green star represents the location of upwind propagation, where the MCS and associated cold pool front is parallel to the flow and the Red star is where downwind propagation is occurring.



Here is an example of such.

The white dotted line represents the approximate location of the cold pool front.

The shear is from the northwest.

Convection here is forming downwind, or forward propagating

And here it is forming upwind, or backward propagating.

Looks a lot like this.

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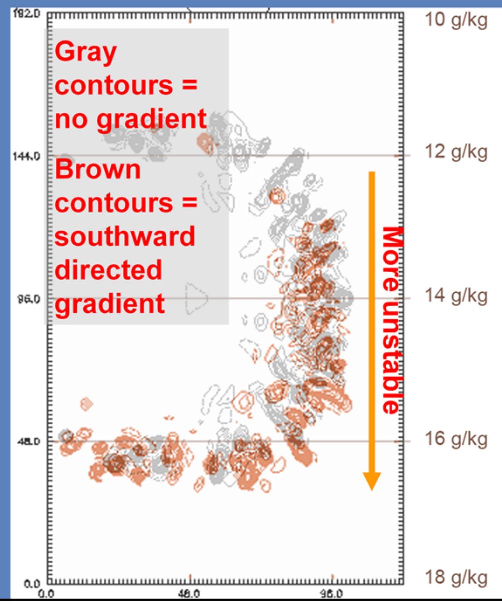
Theta-E Gradients

- Richardson’s (1999) results showed instability gradients have an effect on the propagation of a multicell line



Theta-e gradients, thermal or moisture boundaries, or both can cause changes in local convergence that can influence propagation, and overall storm motion as well. Development tends to head towards more unstable regions due to lower LFCs and stronger resultant downdrafts reinforcing the cold pool.

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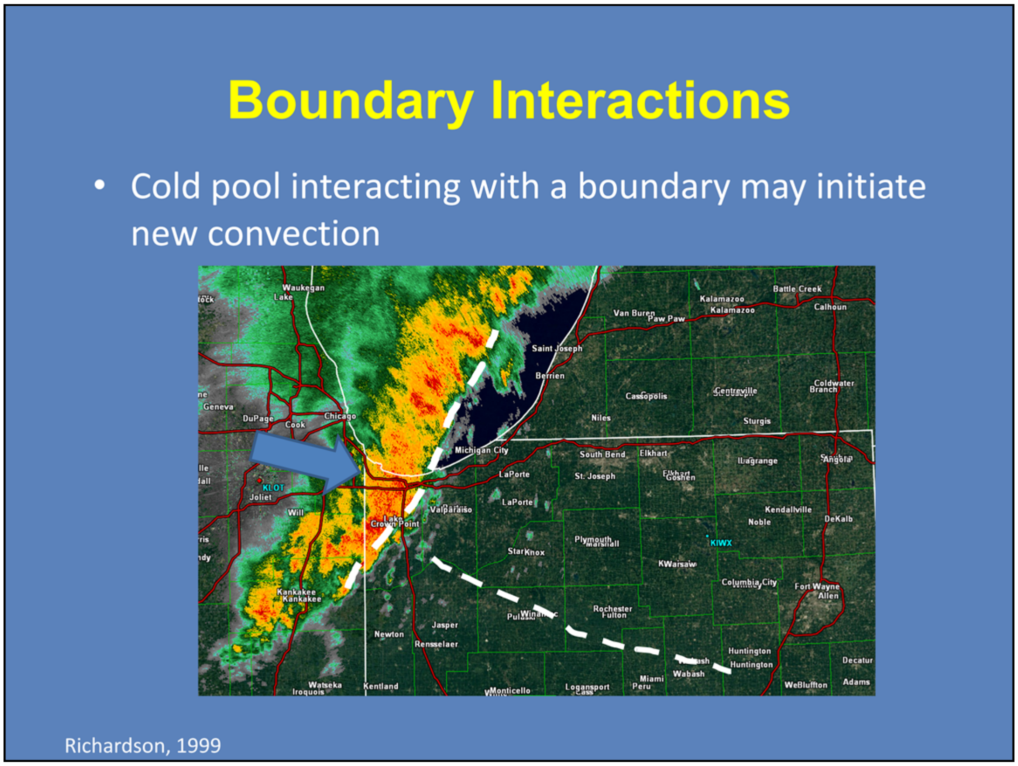

Local pre-existing boundaries can also impact storm motion with the propagation vector tending towards the location of the boundary. For instance in this example,

if the advection and propagation are fairly well balanced in this forward propagating squall line. A boundary

Here, could increase local convergence and cause new convection to form along the boundary shifting its motion, and ultimately leading it into a different airmass or shear environment.

Summary

1. Shear and cold pool interactions (instability)
2. Low-level convergence (shear)

    Advection + Propagation = Motion

3. Instability and moisture gradients
4. Three-dimensional boundary interactions

To summarize there are four key drivers to multi-cell motion.

Shear and cold pool interactions, which can be estimated by available instability

And low level convergence, which can be estimated by ambient wind shear.

Advection and propagation combined form the overall motion

More difficult to predict, but still detectable through mesoscale analysis, instability and moisture gradients can influence multicell motion, with development favoring more unstable regions.

And pre-existing boundaries can also influence storm motion with cold pool interaction along that boundary helping initiate new convection.

If you have passed the quiz, then you have successfully completed this lesson. If you have any questions, please contact us using any of the e-mail addresses listed on the bottom of the slide.

Check out the resources page to links for papers on the topic that can deepen your understanding.

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Welcome to the RAC Convective Storm Structure and Evolution lesson on Rear-Inflow Jets in Multicells. The morphology of the **rear-inflow jet (RIJ)** helps explain future evolution of a multicell and its potential to produce severe weather. This lesson describes the causes of a rear-inflow jet, its dynamics, favored environments, and the different manifestations of rear-inflow jets which impact the potential for severe weather.

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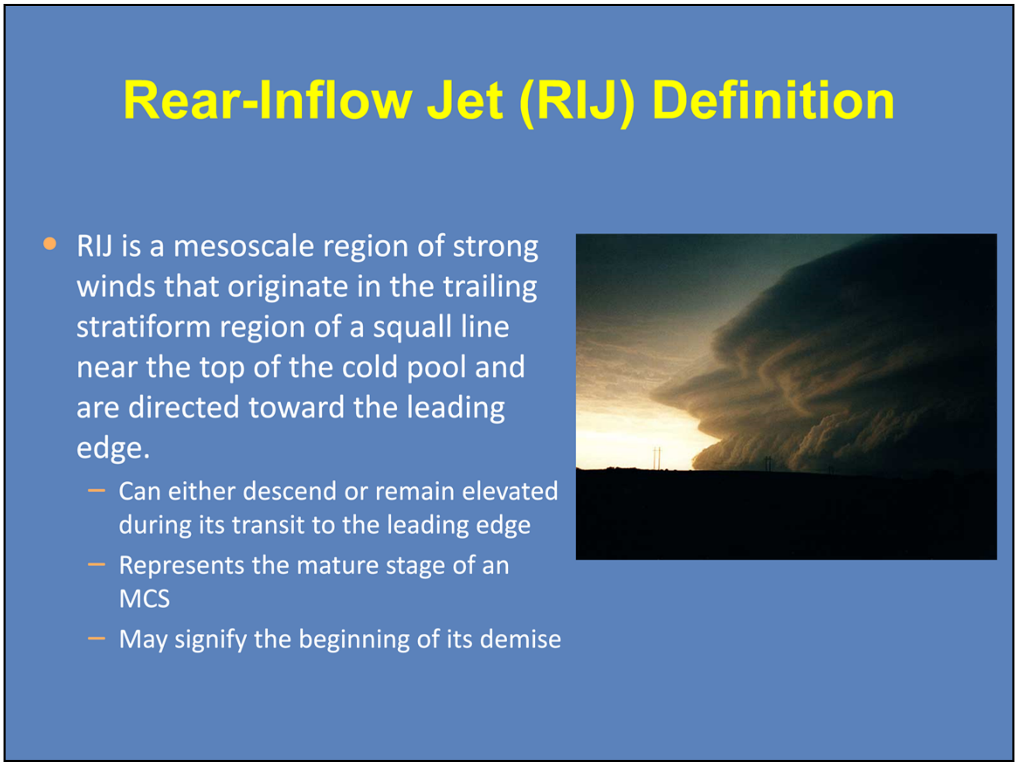
Learning Objective

- Describe the morphology and the influence of the Rear Inflow Jet (RIJ) on multicells.

The learning objective for this less is describe the morphology and the influence of the Rear Inflow Jet (RIJ) on multicells.



An important component of many **mesoscale convective systems (MCSs),** particularly the linear type, is a rear-inflow jet (RIJ). The RIJ is a mesoscale region of strong winds that originate in the trailing stratiform rainfall region of a squall line near the top of the cold pool and are directed toward the leading edge**. The RIJ can either descend or remain elevated during its transit to the leading edge.** It represents the mature stage of an MCS and may also signify the beginning of its demise. However, a large number of squall lines continue to show significant longevity and severity after the RIJ forms.

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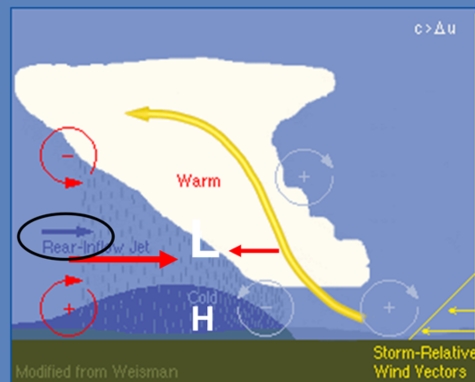
Dynamics of the Rear-Inflow Jet

How Does it develop?

- Latent heat release from anvil precipitate acts to hydrostatically lower the pressure beneath the anvil, but above the cold pool.
  - Hydrostatic high in cold pool
- Air begins to flow rear to front, initiating RIJ
- Strongest mid-level flow resides under thickest anvil
  - RIJ accelerates until it's just behind the updraft



From COMET(1998)

To explain the dynamics of the RIJ, we will start with the mature squall line schematic shown here. As a squall line matures, high-level anvil material begins to stream from the leading edge into the rear side of the squall line (represented by the yellow trajectory). Loaded with small- and medium-sized hydrometeors that have not fallen out of the leading edge, the anvil begins to precipitate, resulting in the region of trailing stratiform precipitation. The anvil material is also warming the upper-troposphere through latent heat released in the updraft along the leading edge. This heat acts to hydrostatically lower the pressure beneath the anvil but above the cold pool (marked by an 'L'). A hydrostatic high still exists near ground-level in the cold pool, dominating any pressure drop caused by anvil material aloft. In terms of pressure dynamics, the anvil-induced low induces air to laterally flow in from both the front and rear sides of the squall line (arrows to the right and left of 'L'). Air begins to flow rear to front, initiating the RIJ . Presumably, the strongest mid-level flow resides underneath the thickest part of the anvil just behind the deep updraft forced along the leading edge of the squall line. Therefore, the RIJ accelerates until it is just behind the updraft. We next discuss the strength of the acceleration and the factors that govern the slope of the RIJ.

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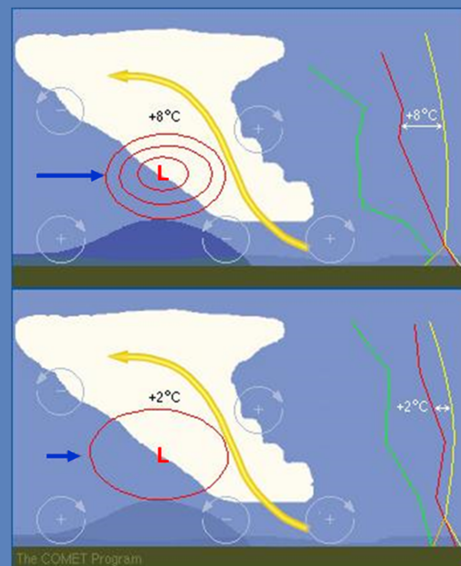
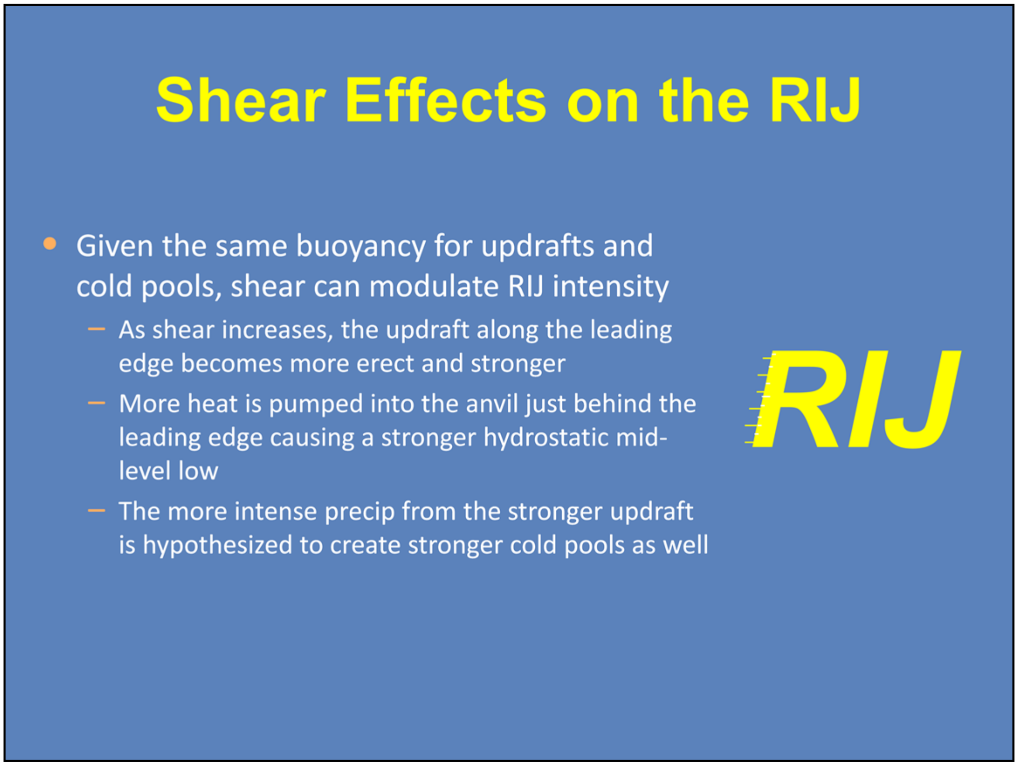
Buoyancy Effects on the RIJ

- Strength of Low depends on amount of net warming in the anvil
- Greater CAPE ….stronger RIJ



<advance>Let's talk about buoyancy effects on the RIJ. The strength of the low underneath the anvil depends on the intensity of the net warming in the anvil. Looking at this schematic, the squall line updraft with the greatest positive temperature excess is the one utilizing the greatest CAPE. Note the hypothetical sounding profile and the temperature excess of the updraft parcel to the environment. The result of higher CAPE is usually a stronger RIJ . Typically in large CAPE environments, lapse rates from the surface to mid-levels tend to be larger, promoting stronger cold pools. **From a vorticity argument, a stronger cold pool circulation to the rear of the squall line works with a more buoyant anvil aloft to generate strong mid-level horizontal inflow that forces the RIJ from the rear of the squall line.**



Given the same buoyancy for updrafts and cold pools, shear can modulate the intensity of the RIJ. According to numerical simulations, as shear increases, the updraft along the leading edge becomes more erect and stronger. More heat is pumped into the anvil just behind the leading edge causing a stronger hydrostatic low in the midlevels. The more intense precipitation from the stronger updraft is hypothesized to create a stronger cold pool as well.

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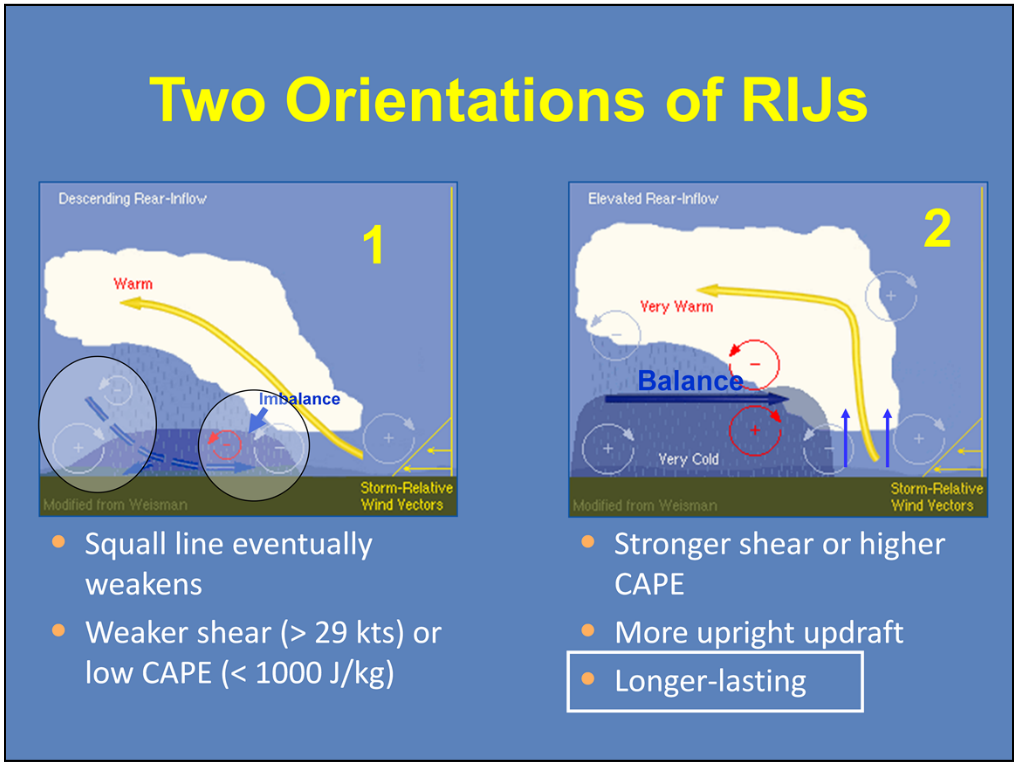

Weisman found that the longevity of the squall line may depend on the rear-to-front slope of the RIJ. Although there may be multiple slopes to the RIJs, there are two extremes: A descending RIJ and a non-descending RIJ.

A descending RIJ occurs when the vorticity generated just underneath the ascending front-to-rear updraft is weaker than the vorticity generated of the opposite sign on the rear edge of the cold dome. The imbalance between the two circulations can be seen to help force the RIJ downward towards the ground prior to reaching the leading edge of the gust front. The RIJ then reinforces the vorticity along the leading edge increasing the imbalance between the cold pool and environmental vorticity. The squall line is theorized to become increasingly sloped rearward and thus weakening. According to simulations by Weisman (1992), this situation occurs with weakening shear (less than 15 m/s, or 29 kts, over the lowest several km) or if the environmental CAPE is less than 1000 J/kg.

For a non-descending RIJ in Figure 2, as CAPE and/or shear increases , the vorticity underneath the rearward expanding anvil becomes much larger due to increased buoyancy. The counter-rotating vorticity along the back edge of the cold dome does not increase as much. This situation results in the increased buoyancy-induced vorticity under the anvil matching the cold dome vorticity to invoke a more horizontally oriented RIJ .

This non-descending RIJ progresses towards the leading edge of the cold pool with a horizontal vorticity structure that interferes with the spreading cold pool vorticity near the gust front. Thus, the strength of the gust front vorticity decreases, becoming more balanced with the environment, and the updraft retains an upright nature. **Squall lines with a non-descending RIJ tended to live longer than their descending RIJ counterparts.**

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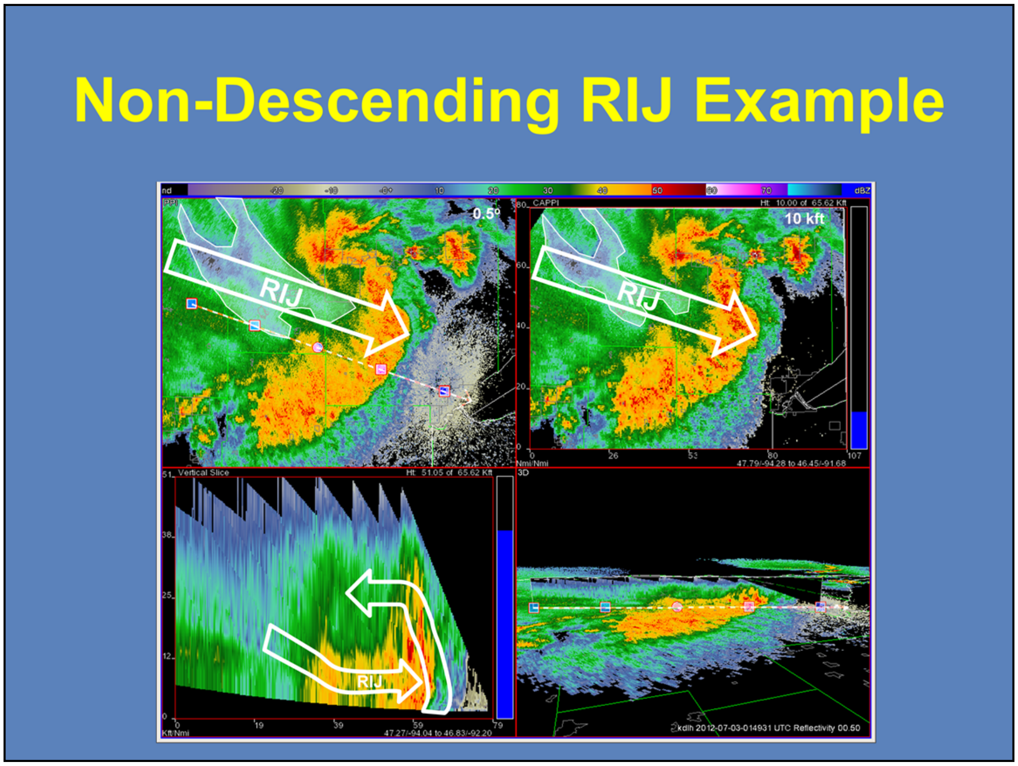
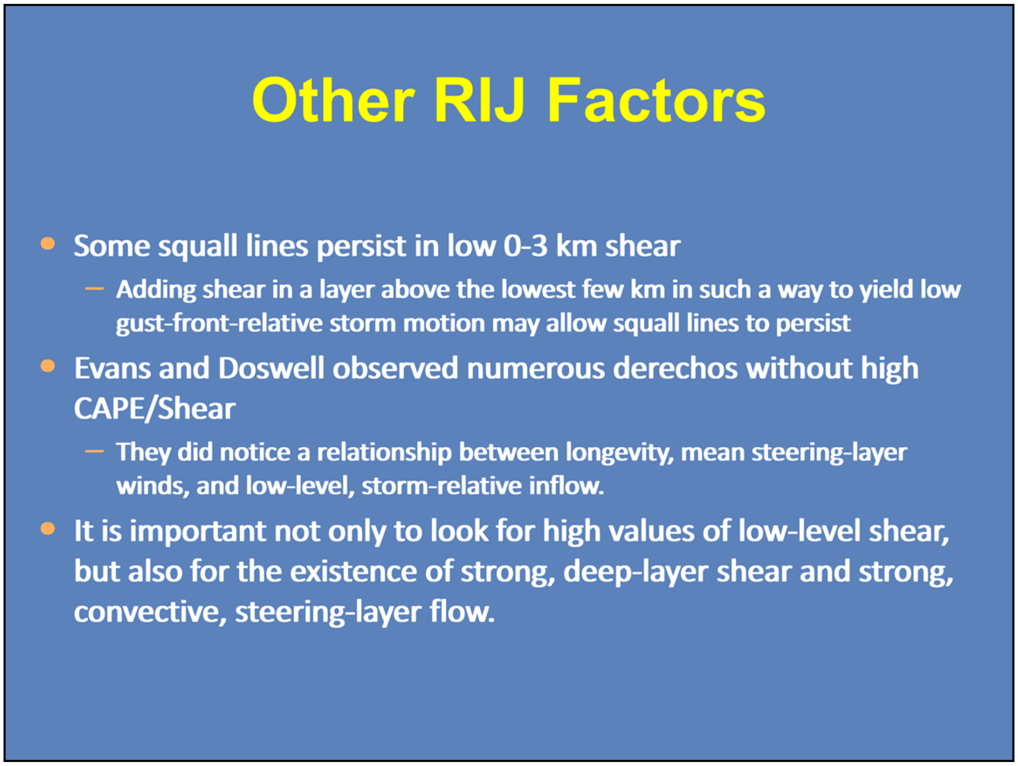


Here is an example of a non-descending Rear Inflow Jet as seen using the AWIPS Four Dimensional Stormcell Investigator tool. In this reflectivity display, notice the weak echo channel in the top two panels which reveals the RIJ location, the upright nature of the updraft as revealed by the cross section, and how the gust front is not easily evident because it hugs close to the strong reflectivity gradient. In this velocity display, notice in the vertical cross section how the RIJ doesn't descend all the way to the ground.

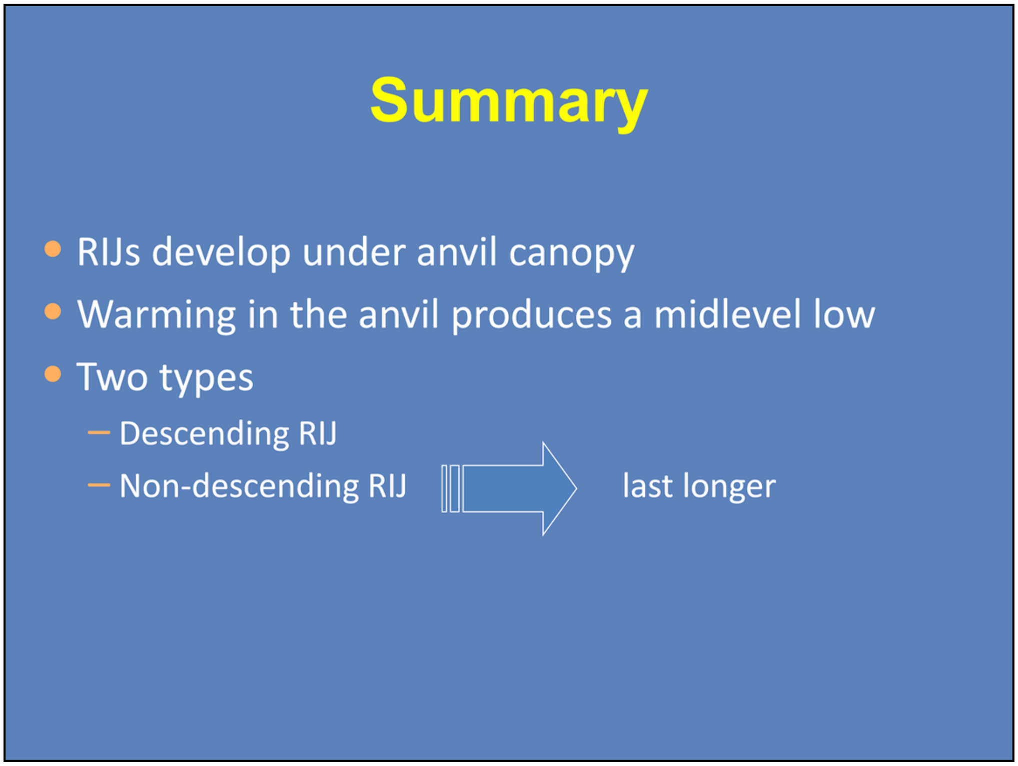


The role of a non-descending RIJ in squall line longevity put forth by Weisman (1992) may not adequately explain the longevity of some severe squall lines in environments exhibiting low values of 0-3 km shear. Other numerical experiments provide evidence that adding shear in a layer above the lowest few km in such a way to yield low gust front-relative storm motion may allow squall lines to persist longer than predicted by shear/cold pool balance theory. In addition, strong synoptic-scale mid-level winds may boost the initiation time and strength of the RIJ. An example would be a cold-season, pre-frontal squall line in the warm sector of a surface extra-tropical cyclone.

Evans and Doswell (2001) observed numerous cases of derechos without high values of either shear or buoyancy. They did notice a relationship between longevity, mean steering-layer winds, and low-level, storm-relative inflow. The latter relationship is likely due to the fact that derechos move quickly. In addition, strong RIJs may be the result of dynamics beyond that of balancing anvil-level buoyancy with cold pool strength. For example, small amounts of CAPE are sufficient to vertically mix strong, synoptic-scale mid-level winds down to the surface yielding a strong RIJ-like structure.

Therefore, it is important not only to look for high values of low-level shear, but also for the existence of strong, deep-layer shear and strong, convective, steering-layer flow.

As is often the case, the parameter space in which long-lived multicell squall lines are observed is often much larger than simulations suggest. The RIJ strongly influences the longevity of MCSs, but there are several other very important environmental and storm scale features that modulate how long an MCS will survive, as well as how severe the MCS will become.



In summary, rear inflow jets develop underneath a large anvil canopy as a midlevel low forms in response to upper-level warming within the anvil. The more CAPE that gets pumped into the anvil, the stronger the midlevel low becomes.

Two types of RIJs have been documented: The descending RIJ, which typically evolves as the cold pool forcing exceeds that of the buoyant anvil. The descent of the RIJ enhances the forward motion of the cold pool forcing an increasing tilt to the convective updrafts along the leading edge. And, a non-descending RIJ, which forms in an environment of increasing vertical wind shear. The anvil is typically more buoyant due to stronger updrafts. The RIJ forcing is balanced between the cold pool and anvil. Non-descending RIJs help to restrain the advance of the cold pool relative to the motion of the system as a whole, and a deeper outflow boundary is the result. Squall lines associated with nondescending RIJs typically survive for longer periods.

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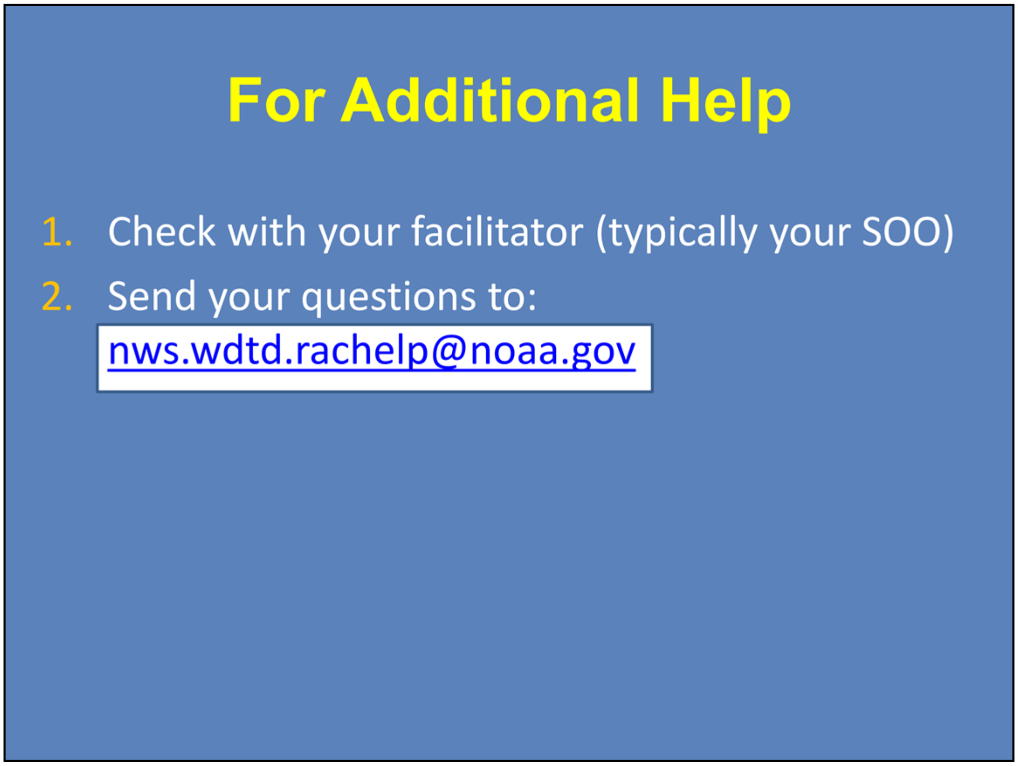

For additional help, check with your facilitator (typically your SOO) or send your questions to the listserv e-mail address here.

Welcome to the RAC Convective Storm Structure and Evolution lesson Bookend Vortices and Bow Echoes. I'm Justin Gibbs of the Warning Decision Training Division.

Learning Objective

- Identify the characteristics of bookend vortices and MCVs
- Identify the mechanisms involved in their formation.
- Identify operationally significant signatures associated with bow echoes and Rear Inflow Jets

Here are the learning objectives for this lesson.

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3-Dimensional Structures in Multicell/QLCS systems

- Elevated RIJs
- Bookend (Line-End) Vortices
- Bow Echoes
- Mesoscale Convective Vortices (MCVs)

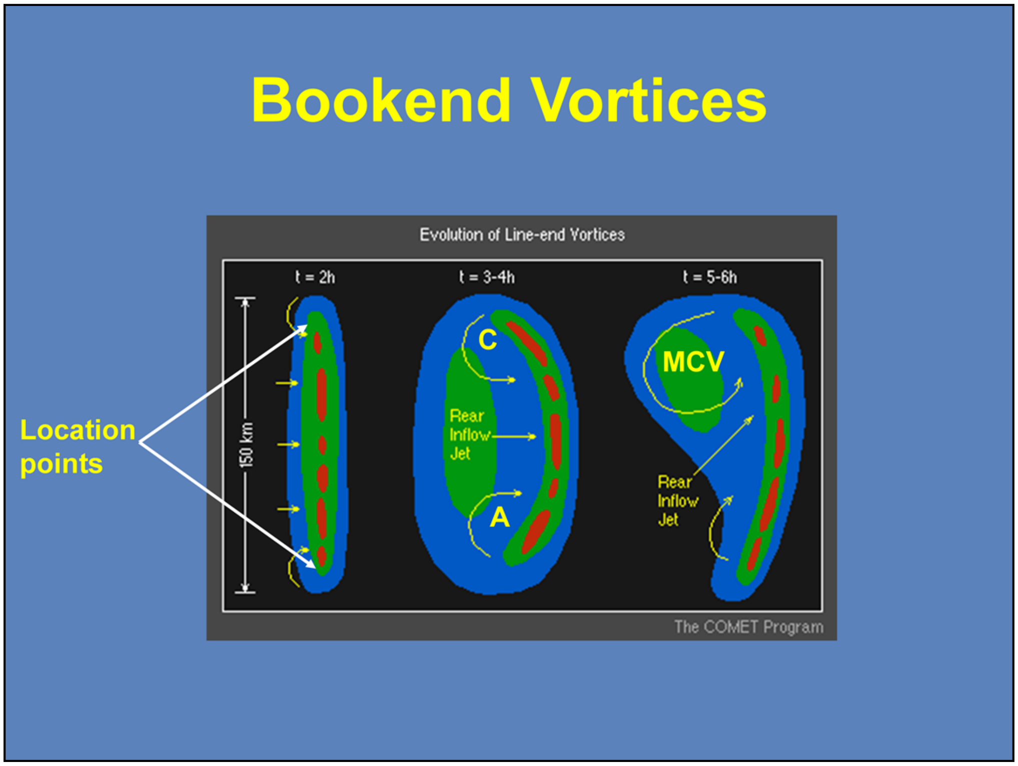
Organized multicell and QLCS systems can exhibit the following operationally significant features.

elevated RIJs (previously discussed),

bookend vorticies (also called line-end vortices or mesovortices), bow echoes,

and mesoscale convective vortices (known as MCVs).

These features are frequently associated with an increased risk for wind damage, and an increased probably for significant wind damage.



Bookend vorticies, evolve at the end of or breaks within a line.

The rear inflow jet pushes out a linear storm into a bow like feature and forms in this example, a cyclonic circulation at the top, and an anticyclonic circulation at the bottom.

This causes asymmetrical development through time because the Coriolis Force acts to increase convergence in the midlevels which helps to strengthen the northern cyclonic vortex and weaken the anticyclonic vortex. The dominant cyclonic vortex can last well beyond the lifetime of the originating convective system and grow upscale.

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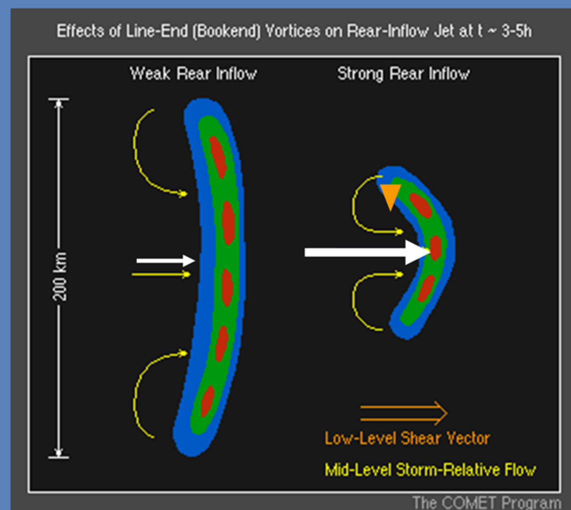
Effects of Line-end Vortices

- Increases strength of RIJ
- Increases downdrafts, localized damaging winds, and non-supercell tornadoes



Since the line-end vortices develop within the downdraft portion of the squall line, they can enhance the strength of the RIJ between the verticies, and produce severe weather. as a squall line evolves to a bow echo, the smaller the distance between the line-end vortices produced greater enhancement to the midlevel flow and RIJ (Weisman, 1992). Tornadoes often form just to the left of the maximum wind of the RIJ.

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Here is a real world example of a bookend vortex on the north side of a QLCS system. This feature persists for much of the duration of these images.

A tornado forms early in the animation, with local vorticity possibly from this area of convection intersecting the line perpendicular to its movement, a cell merger.

Shortly after the tornadic circulation becomes evident outbound velocities begin to show a rear inflow jet forming and expanding

With a rear inflow notch also becoming more apparent in reflectivity.

Another increase in low level circulation associated with an approaching cell merger occurs.

Followed shortly thereafter by another cell merger also associated with an inflow notch and outward extension of the reflectivity signature, associated with a strong circulation.

And the reformation of the bookend vortex.

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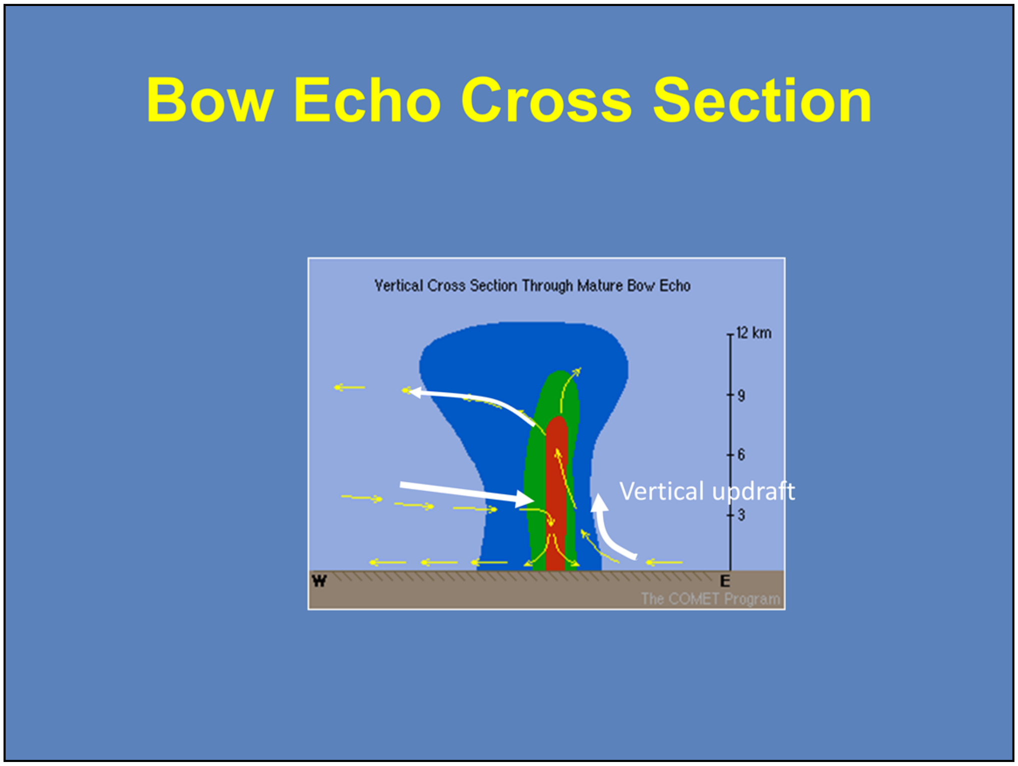
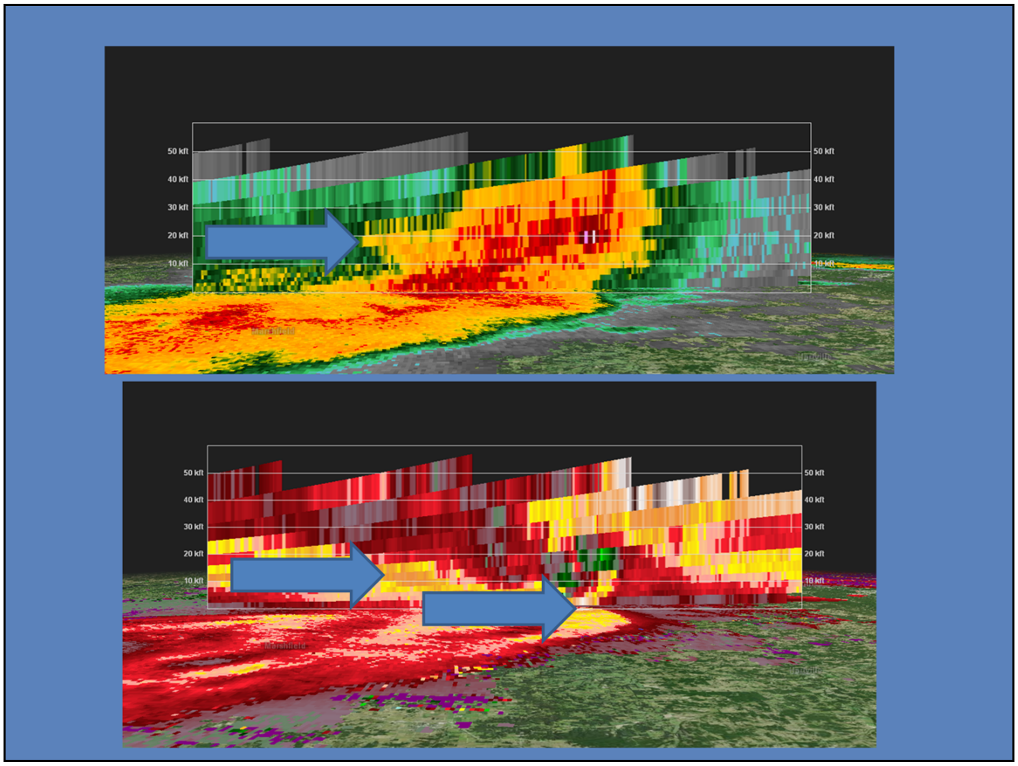


Vertical cross-sections in the core of mature bow echo simulations revealed a strong, vertical updraft at the leading edge of the system, a strong elevated RIJ moving in behind the updraft region before descending rapidly to the surface, and a front to rear flow near the top of the updraft which went back into the anvil and fed the trailing stratiform precip region.



Heres an example of what that would look like on the 88D cross section.

With the rear inflow notch in reflectivity.

And a broad area of very strong inbounds, 50 to 60 knots forming the rear inflow jet that reaches 80 to 90 kts on the lowest beam elevation under some inbounds indicating mid altitude radial convergence.

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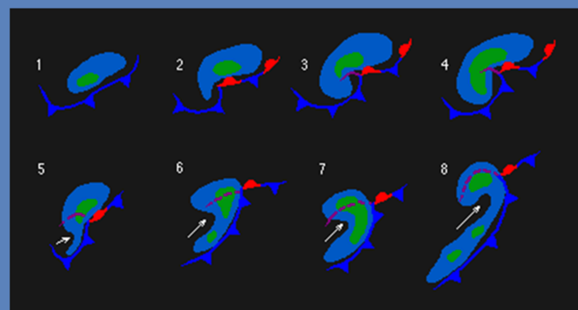
Bow Echo Transition

- Supercells to Bow Echoes
- Supercells within Bow Echoes



Evolution of an HP supercell to a bow echo. The arrows show the rear downdraft / outflow.
Modified from Moller et al., 1994

In an environment with supercells, especially with increasing large scale ascent, supercells can form into bow echoes and MCVs. The location of the supercell would be a favored area for severe weather within the resulting QLCS system.

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Web Object

Address:
http://training.weather.gov/wdtd/courses/rac/severe/objects/mcv/

MCVs are a frequent location of higher impact wind damage and tornadoes within QLCS systems.

Here notice the line kinks up a bit, starts to show some inflow notches and the outbound velocities begin to enhance.

That process continues and intensifies fairly rapidly, with outbound velocities around 80 knots

And an obviously rotating system with rainband like appendages.

The enhanced velocity signal continues for the remainder of the animation, about 45 minutes real time.

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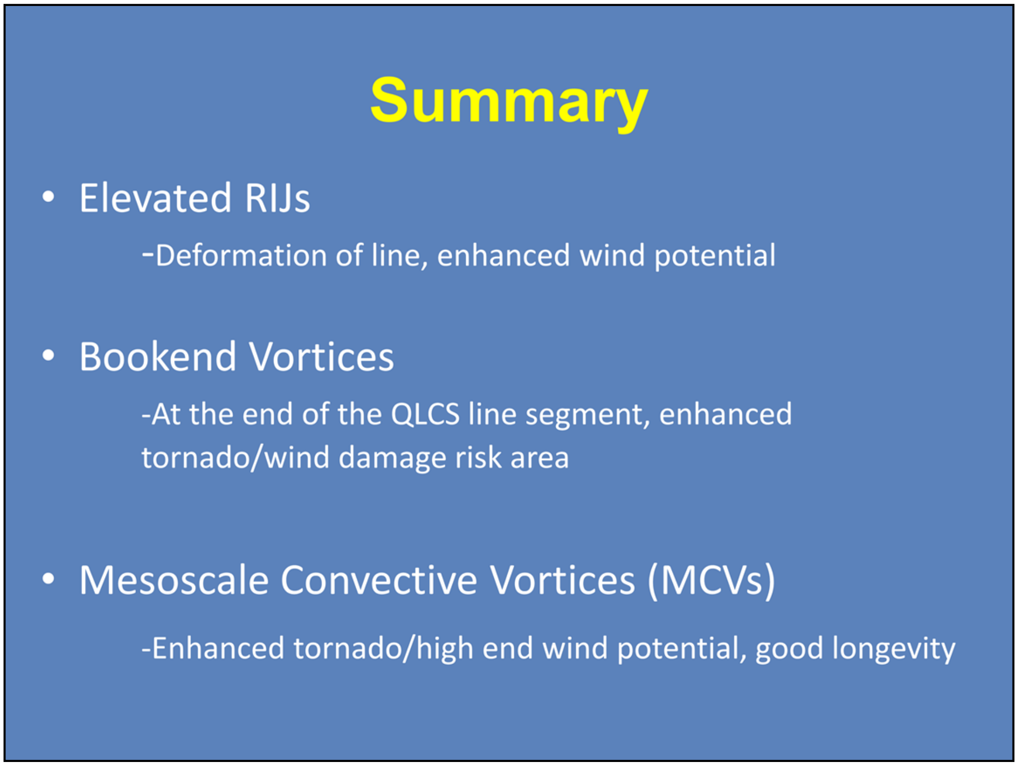
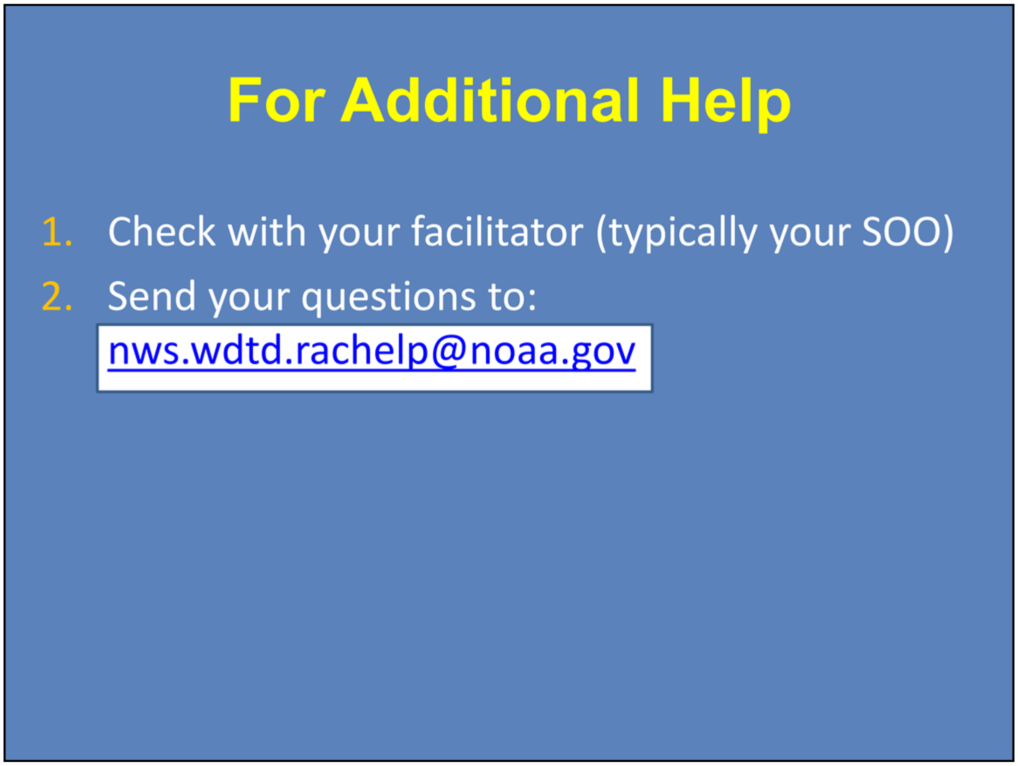


So to recap we talked a bit about elevated rear inflow jets, which can deform the QLCS system and create an enhanced wind potential.

Bookend vorticies, or line end vorticies, that form at either end of a qlcs line segment. They are often associated with at least a modestly enhanced tornado or wind damage potential.

And Mesoscale convective vorticies, or MCVs which are frequently associated with an enhanced wind damage, and high end wind damage potential as well as an increased threat for tornadoes. They also show a bit of resilience and have better longevity for that increased potential than QLCS systems without an MCV.



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Welcome to the RAC Convective Storm Structure and Evolution lesson on Multicell Severe Wind Hazards.

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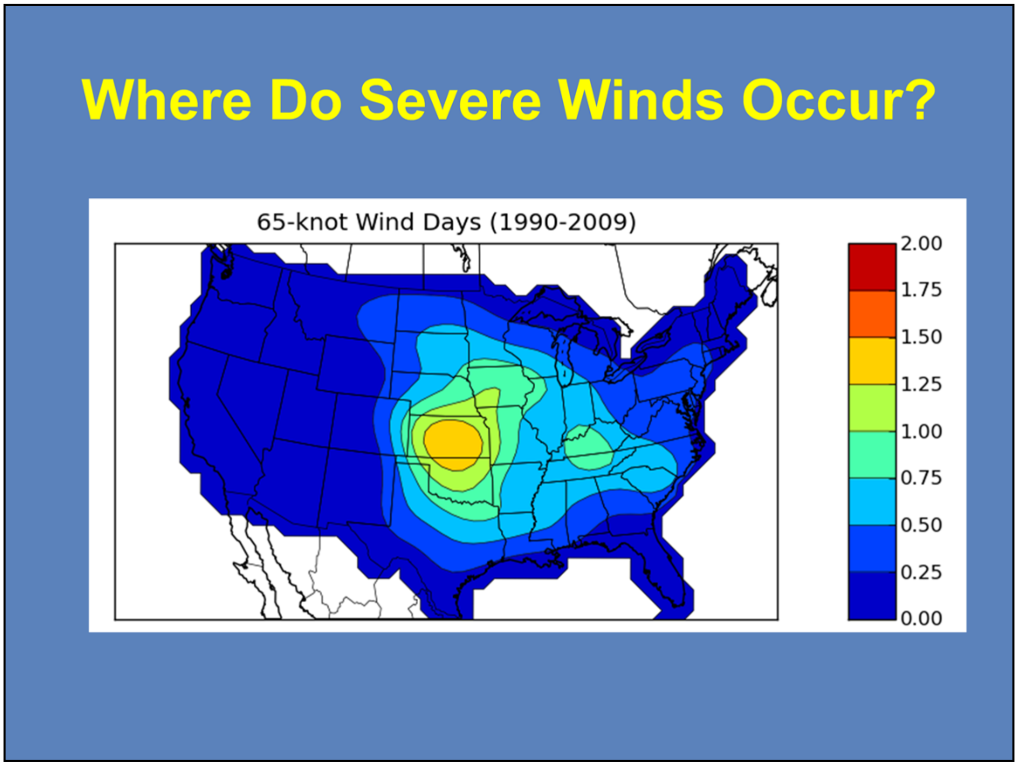
Learning Objective

- Recognize multicell storm signatures for monitoring and anticipating damaging straight line winds.

This is the learning objective for Lesson 19.



These are graphics from NSSL showing the mean number of severe thunderstorm wind days per year within 25 miles of a point. Note the three general frequency maxima: southern plains (OK/KS/MO/AR border region), Ohio /TN River Valley to the western Carolinas and one near DC and southern PA. There is a much weaker maxima in southern AZ due to the monsoons. When we fade to mean number of significant severe (> 65 kt) wind days per year, we see the maxima is centered squarely over KS and northern OK with a secondary maxima over TN/KY.

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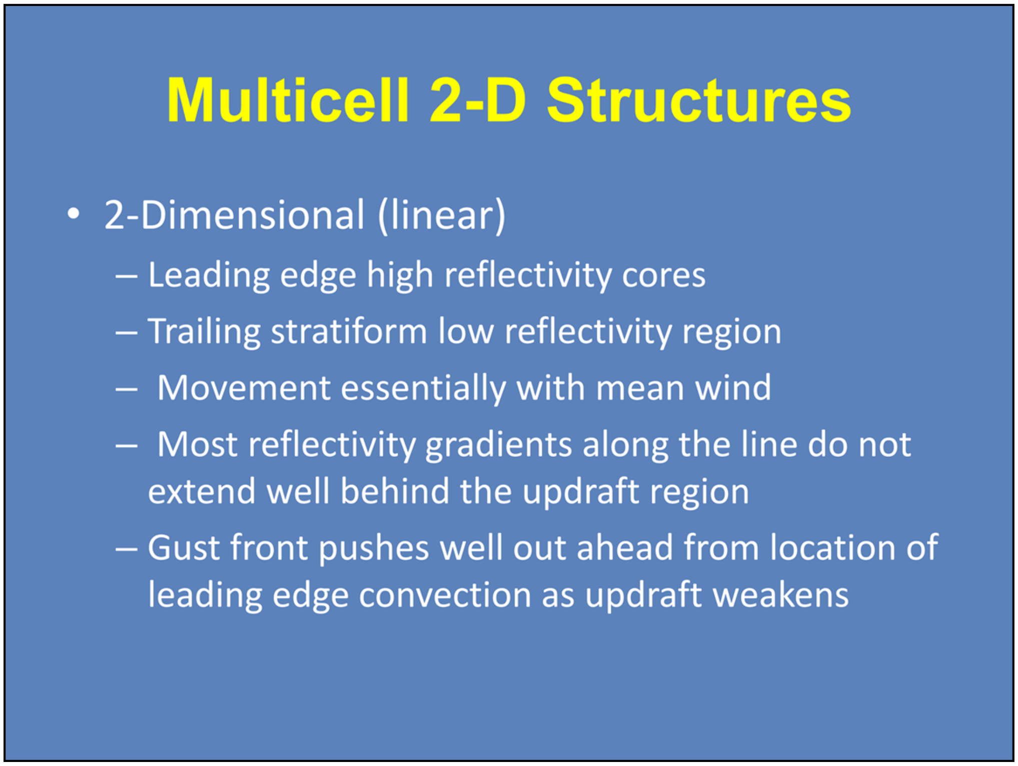

Storm signatures associated with damaging winds from multicells are typically associated with squall lines containing supercells and/or bow echoes. Squall lines can form a variety of ways. The structure depends largely on the shear profile. With weak to moderate shear and subsequent slower system motion, the structure of the multicell complex is typically 2-dimensional with none of the characteristics discussed in the previous lesson. These structures would include:

• Leading edge possessing the high reflectivity convective cores.

• Trailing stratiform possessing the low reflectivity regions.

• Movement of the line with mean wind.

• No bowing, or in other words, points along the line do not extend beyond the leading edge.

• Gust front pushes well out ahead of line.

I'll show a conceptual model of a plan view of this structure next.

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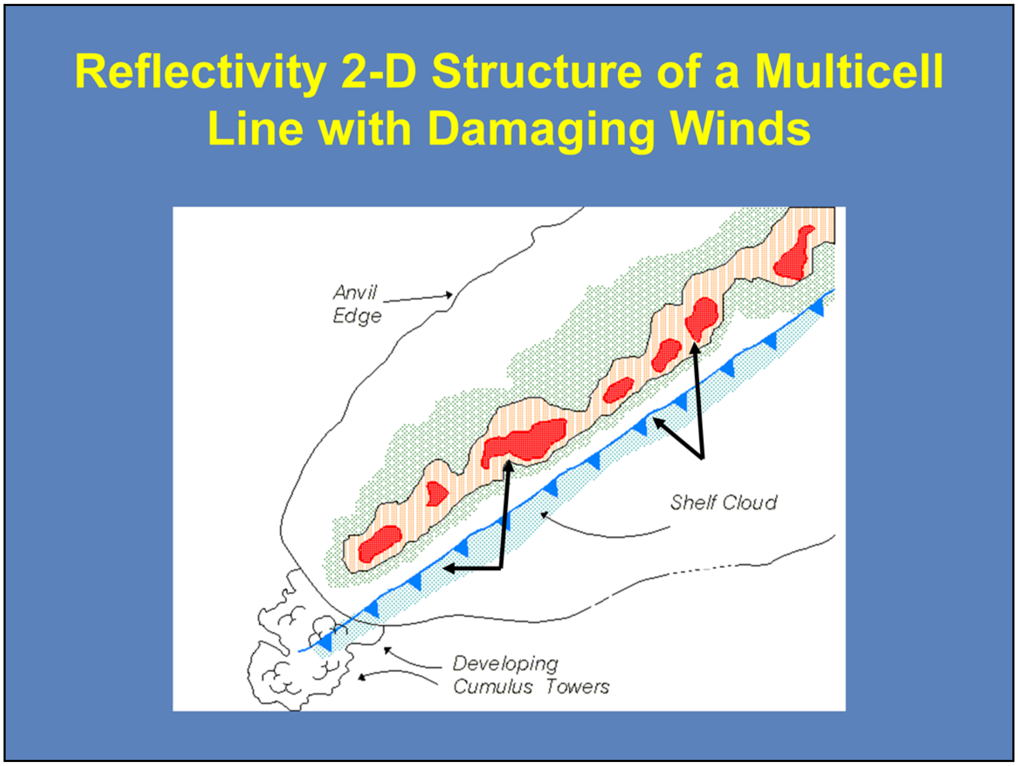

This is a conceptual model of a narrow multicell line possessing 2-D features. Note the position of the shelf cloud superimposed on the leading edge of the gust front, which, along with the regions around the stronger cores, is the most likely location for severe straight line winds. According to observations, flow is typically front to rear in cases such as these when the shear is not moderate to strong. Often this structure is observed in the early stages of squall line development.

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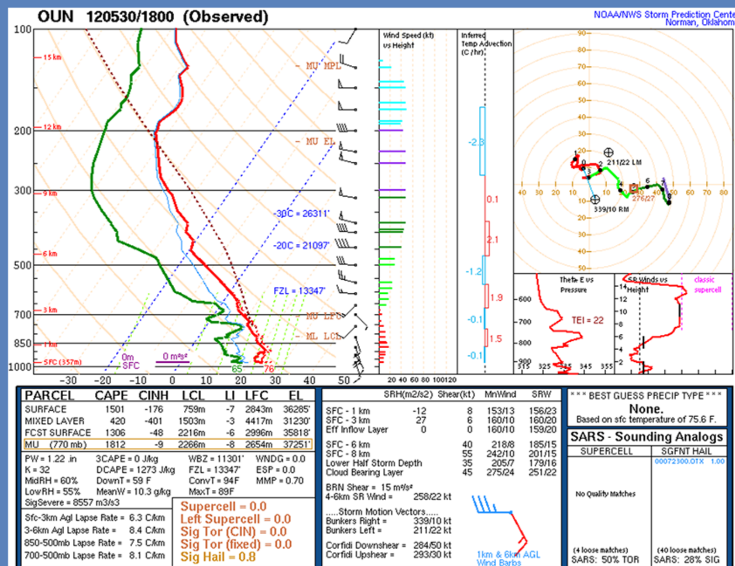
Sounding Example of Linear Multicell Line



This is an upstream proximity sounding from Norman, OK at 1800 UTC on May 30, 2012. Note the weak flow in low to mid levels, but flow increases above 400 mb. The bulk shear from 0 to 6 km is 40 kts with a mean wind in the cloud bearing layer (from) 275 deg at 24 kts. Sounding has a MUCAPE of 1812 j/kg. There's still some surface based CINH, so weak winds in the diminished inflow layer. Above the LFC, the winds are unidirectional so most of the vorticity that can be tilted by the updraft will be crosswise. The upper level shear suggests that given this environment, you could get both linear 2-D and 3-D structures in subsequent multicell thunderstorm development.

This is a long loop of lowest scan reflectivity, velocity and ZDR from KICT from 2251z to 0224z showing the evolution of a multicell. Note how for roughly the first hour in the loop lifting with storms developing along the line were relatively discrete and motion dominated by individual cell propagation. Many storms in the multicell cluster were moving east while others were beginning to sag south as a result of different boundary relative flow. There was a large storm which had developed out ahead of the line just east of KICT and was moving away from the primary line to the north of the radar. There was an outflow boundary extending E-W from the storm to the middle of the line. When the gust front travels at the same speed as the multicell line, the boundary-relative flow maximizes potential for new cell growth along the leading edge. And the updraft becomes more erect, especially in the center portion of the line just north of the radar. 3D structures will evolve as the updraft becomes more upright and erect. Otherwise, when the gust front pushes well out , you can expect a more sloped front to rear 2D structure. If you toggle to the V product, you can see where the intense winds occurred just north and east of Wichita where trees and power lines were downed due to straight line winds up to 70 kts occurred. Intense updrafts due to the increasing shear and moderate buoyancy in the environment also produced some large hail up to 2” in diameter which is not uncommon in multicell structures especially in the southern Plains. The large trailing stratiform shield becomes more apparent after 0200z as the center part of the line weakens and more intense updrafts develop and accelerate along the line out to the west. Eventually the western part of the line increased and intersected the central part of the line.

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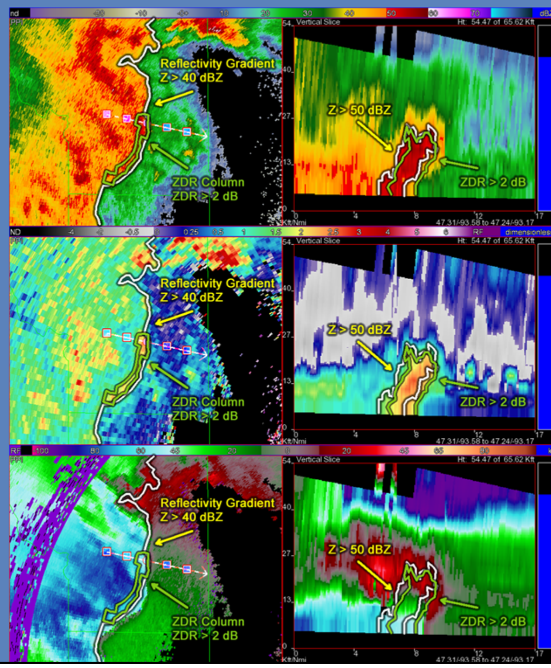
Multicell 3-D Structures

- Bows
- WERs/BWERs
- MARC

These are the 3-dimensional structures found in multicell squall lines and bows. Since we've already covered bow echo structures, I'm just going to show examples of a Weak Echo Region (WER) and a Mid-Altitude Radial Convergence (MARC) signature within a multicell.

WER Structure within a Multicell



The existence of a WER (or BWER – "Bounded" WER) in the multicell structure of a QLCS most often indicates an enhanced potential for damaging surface winds. Remember that these features are associated with strong, deep layer wind shear. This figure (which is Figure 7-200 in the Student Guide) shows an example of a WER structure from a multicell from DLH on the evening of 2 July 2012. Note that the intense updraft is located along the leading edge of the line where the mid-level overhang and WER are located and where the strong low-level reflectivity gradient is located. Moreover, this line-segment is also bowing. The ZDR column where values are > 2DB are noted in the middle left panel. The cross-section views (right column panels) show the tilted updraft and ZDR column overhang due to shear and the deeper convergence extending up through the updraft. This case illustrates the characteristics of multicells structures in very strong shear. In terms of reports, numerous trees and power lines were down in Itasca County, MN.

This is a FSI screen capture at 0135 UTC from KDLH on July 3, 2012 of a highly sheared QLCS structure with a WER structure. You can toggle to V, ZDR, and KDP to see the DCZ and the vertically erect updraft in the cross-section portion of the 4-panel. Toggle to the velocity product and you can note the strong rear-to front flow with convergence all the way to the leading edge in association of a descending RIJ.

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MARC Signature

- Precursor to descent of RIJ
- Persistent elevated convergence > 50 kts at 3 to 5 km AGL can provide lead time for first report of wind damage (before bow echo develops)
- Often part of a DCZ

Another radar signature which is often associated with severe winds in multicells is the Mid-Altitude Radial Convergence Signature, abbreviated MARC. Observations of a MARC have been noted by Przybylinski and others as a precursor to the descent of the elevated RIJ. Enhanced velocity differentials which signify areas of strong convergence are often located just downwind of high reflectivity cores along the leading edge of the convective line. Persistent areas of MARC greater than 50 kts at 3-5 km AGL can sometimes provide lead time for the first report of wind damage (often before a well-defined bow echo with bookend vortex develops). MARC signatures are often part of a Deep Convergence Zone (DCZ) found in some intense updrafts where the Velocity differences of 30-55 m/s are found in both multicells and supercells. Let’s look at an example of a MARC.

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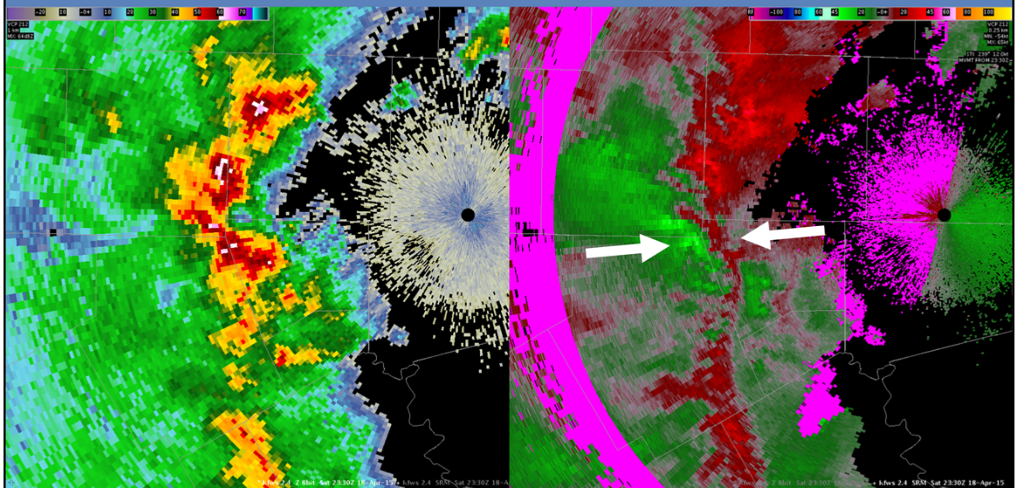
MARC Signature



MARC Signature from KFWD on 2330 UTC 18 April 2015

This is a 2.4 degree reflectivity and storm-relative velocity image of a Mid-Altitude Radial Convergence (MARC) signature. The white arrows indicate the location of the MARC signature. Remember, the MARC is detected via radial velocity data in the mid-levels (3-5 km AGL) within the intense reflectivity core of a squall line. Look for delta-V of at least 50 kts across the convergence axis. Advance to the next slide for a loop of this storm.

This is only a short loop but gives you the option to toggle back and forth from 2.4 deg Z and SRM, plus the 3 Dual-Pol products.

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Summary

- 2-D Features
  - Strong Reflectivity along leading edge
  - Gust front speed well-matched to system speed
- 3-D Features
  - Bow Echoes
  - WER/BWER
  - MARC

Multicell winds events arise from systems that occur on larger scales than individual downbursts. In the absence of mdt to strong shear, multicells usually possess 2-D characteristics such as strong reflectivities along the leading edge, and when the gust from remains close to the leading edge of the line.

When the multicell system is more mature, and when there is mdt to strong shear in the environment, 3-D features start to become more common such as bows, WERs/BWERs and MARCs. These all are classic signatures for severe winds in multicells (and in supercells). For a MARC signature, look for delta-Vs of at least 50 kts across the convergence axis through at least 3 km (~ 10 kft).

Warning Methodology

1. Introduction

1.1 Title

Notes:

Welcome to the Warning Methodology lesson.

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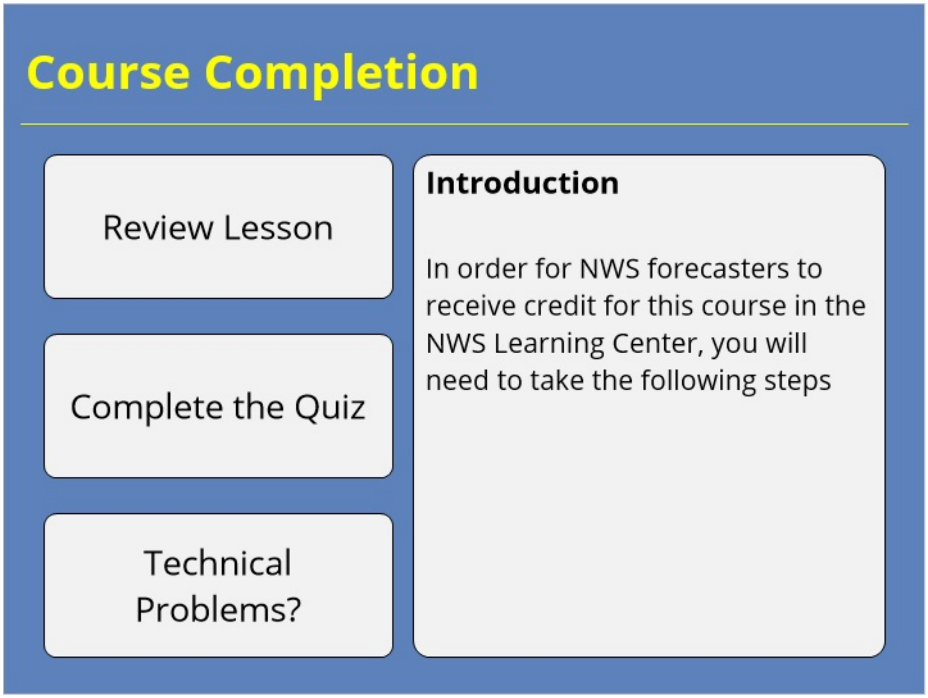
1.2 Course Completion Info



Notes:

If you are completing this lesson for credit, then view this interaction. When you are finished, click “Next” to continue.

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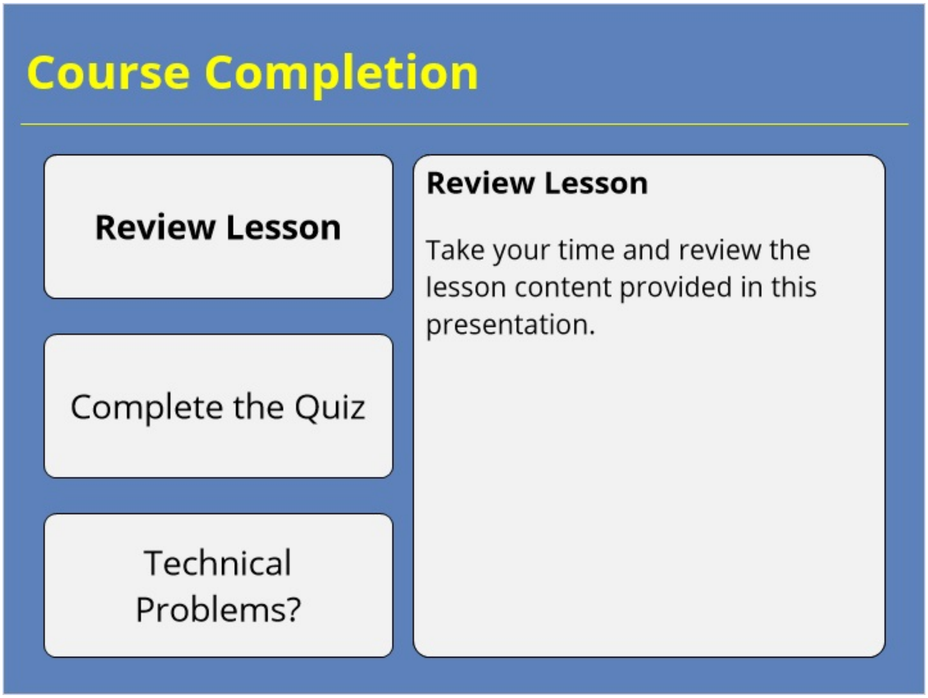
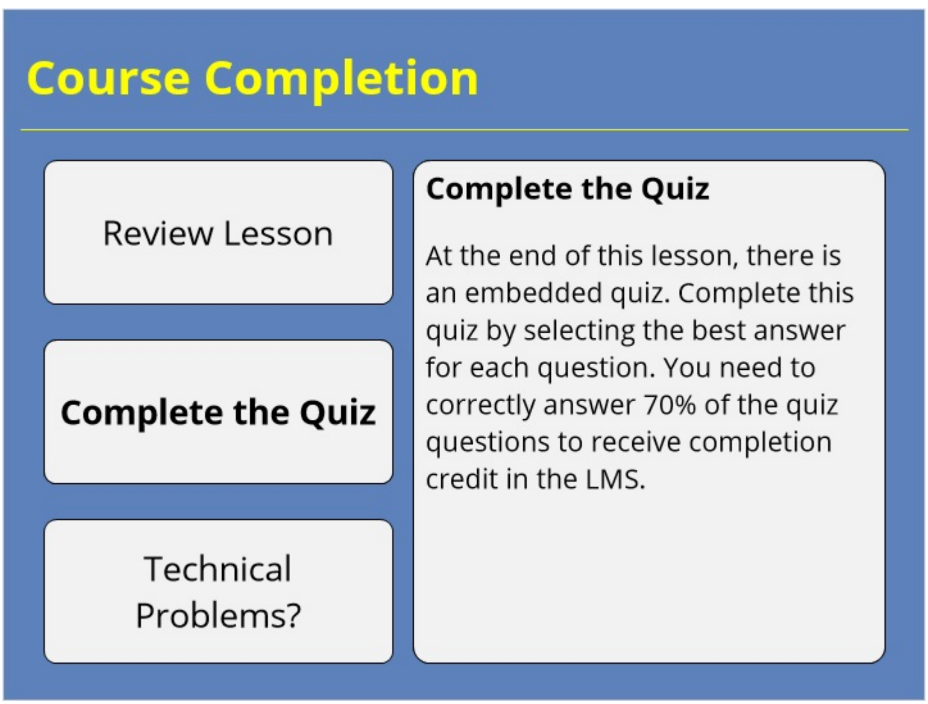
Review Lesson (Slide Layer)



Complete the Quiz (Slide Layer)



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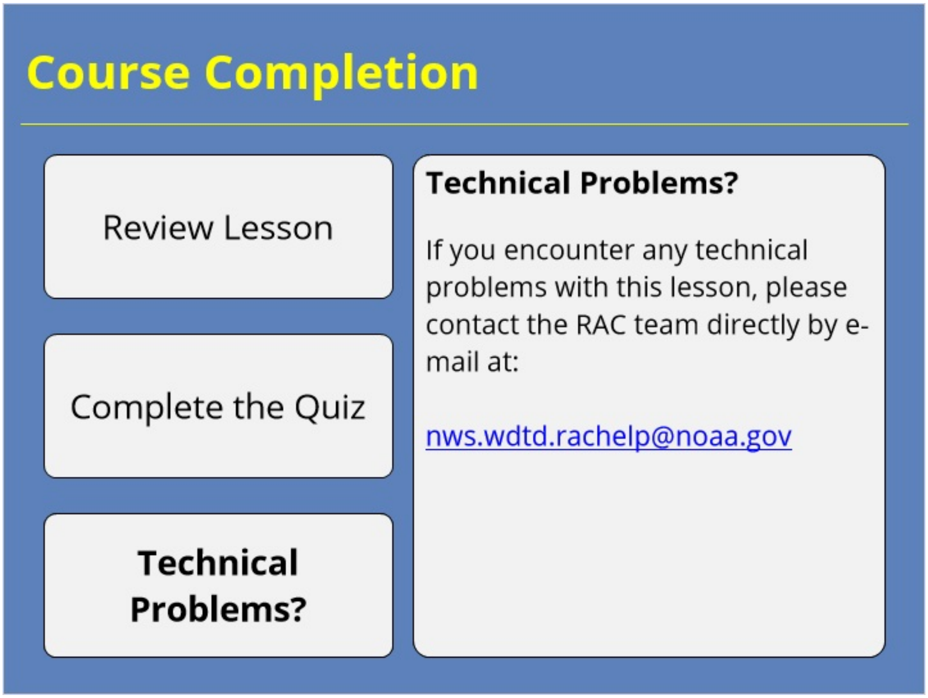
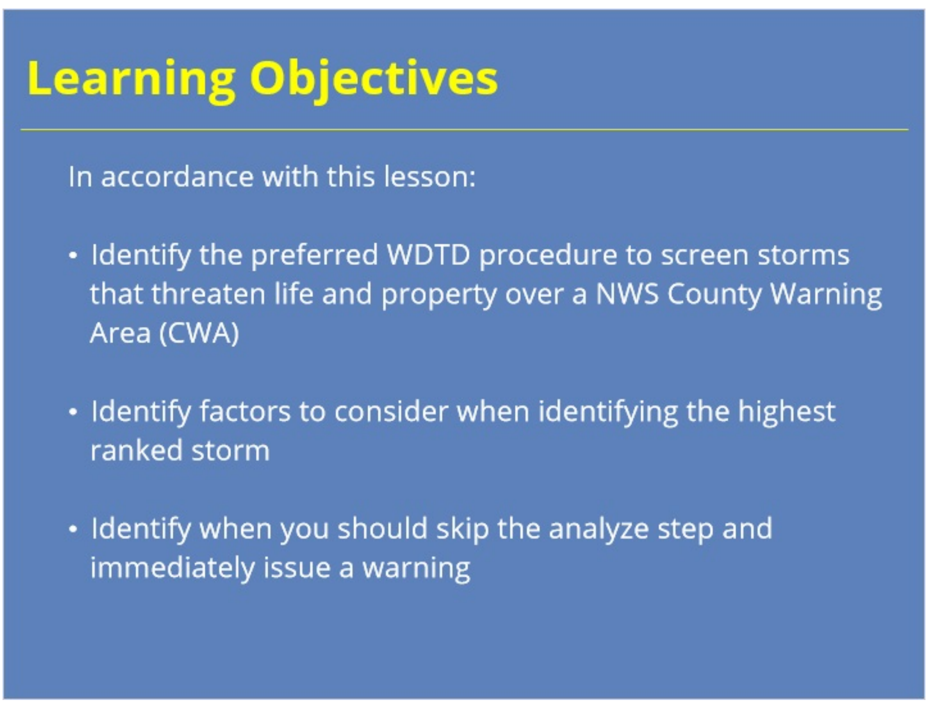
Technical Problems (Slide Layer)



1.3 Learning Objectives



Notes:

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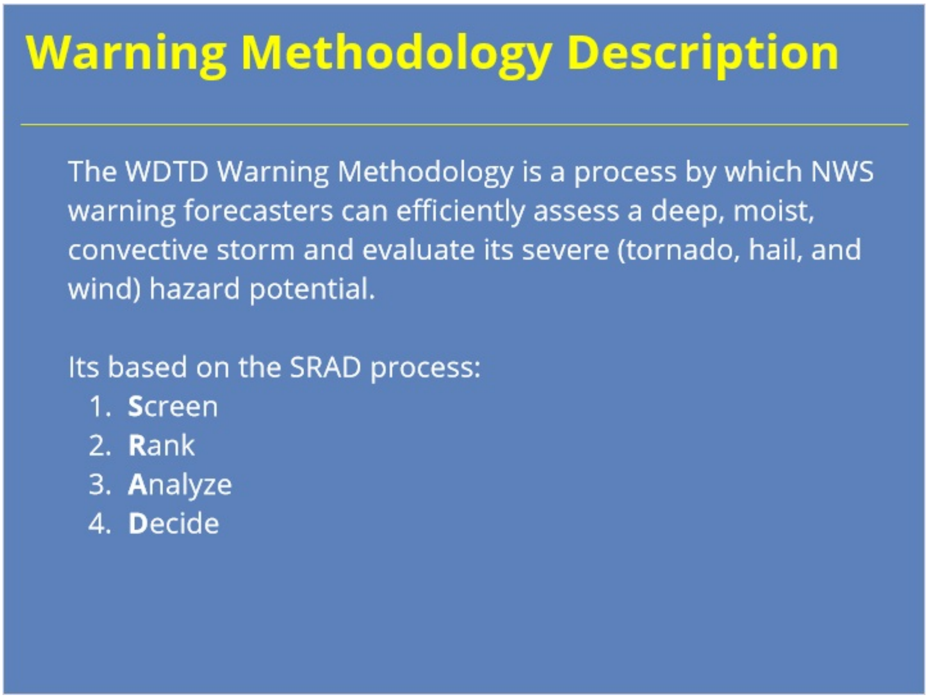
Here are the learning objectives for this lesson. Please take a moment to review them and then advance to the next slide where you are ready.

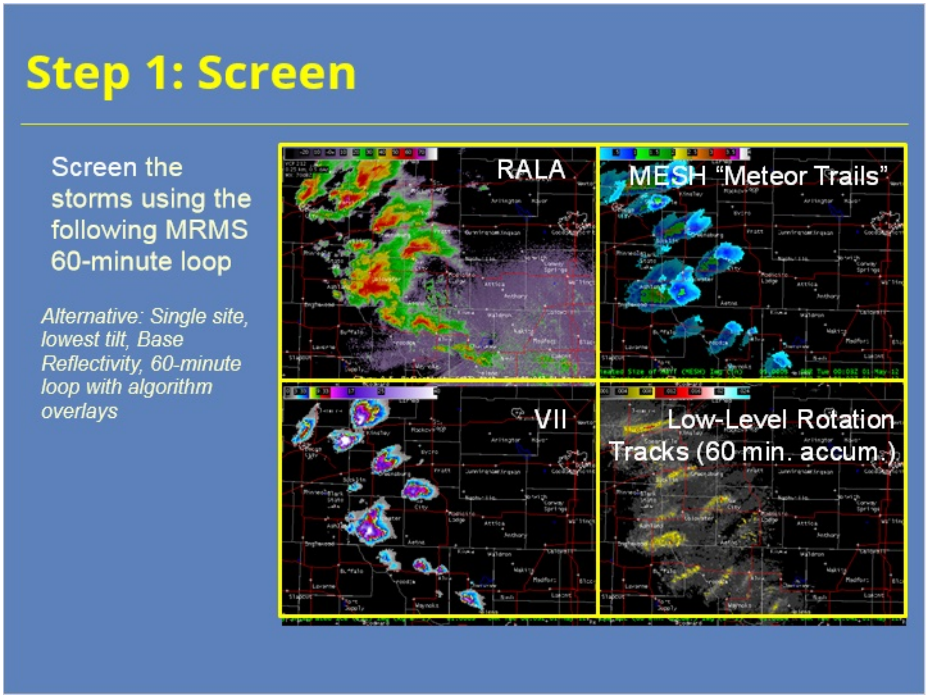
1.4 Warning Methodology Flow Chart



Notes:

The WDTD Warning Methodology is a process by which NWS warning forecasters can efficiently assess a deep, moist, convective storm and evaluate its severe hazard potential. Its based on the SRAD process: Screen, Rank, Analyze, and Decide.

1.5 Step 1: Screen



Notes:

Step 1 is to **Screen** the storms that threaten life and property over your County Warning Area (CWA). For severe (tornado, hail, and wind) hazards, load a 4-panel display showing a 60-minute loop of Multi-Radar/Multi-Sensor (MRMS) products: Reflectivity at Lowest Altitude (RALA), MESH "Meteor Trails" which consists of Maximum Estimated Size of Hail (MESH) and 60-minute MESH Tracks, Low-level Rotation Tracks (60 minute accumulation), and Vertically Integrated Ice (VII).

If MRMS product latency is an issue, then WDTD suggests using a single-site, lowest-tilt, Base Reflectivity, 60 minute time lapse loop with Meso, Hail, and TVS algorithm overlays.

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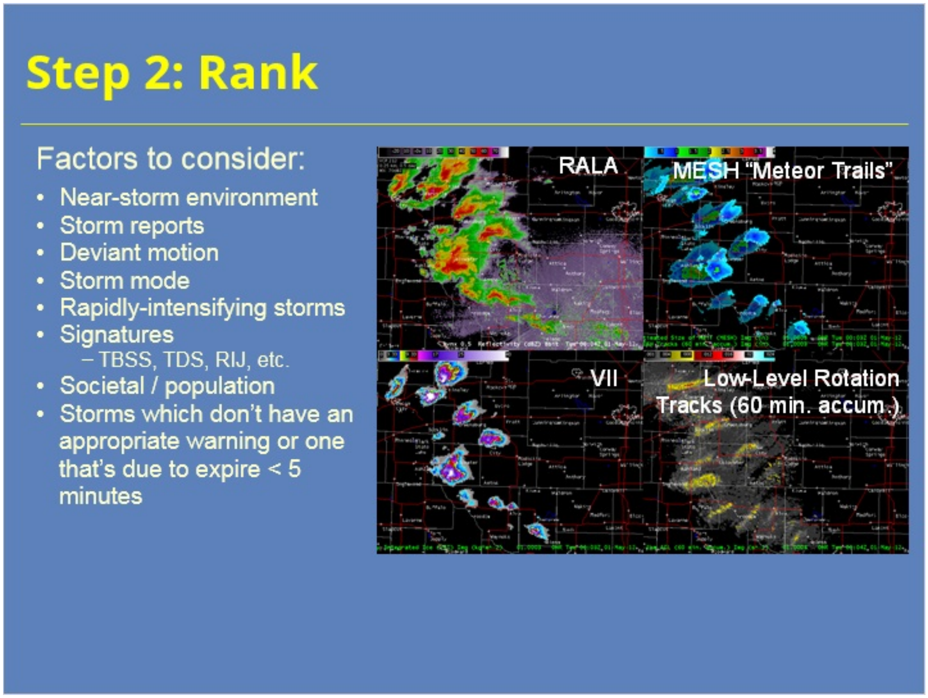
1.6 Step 2: Rank



Notes:

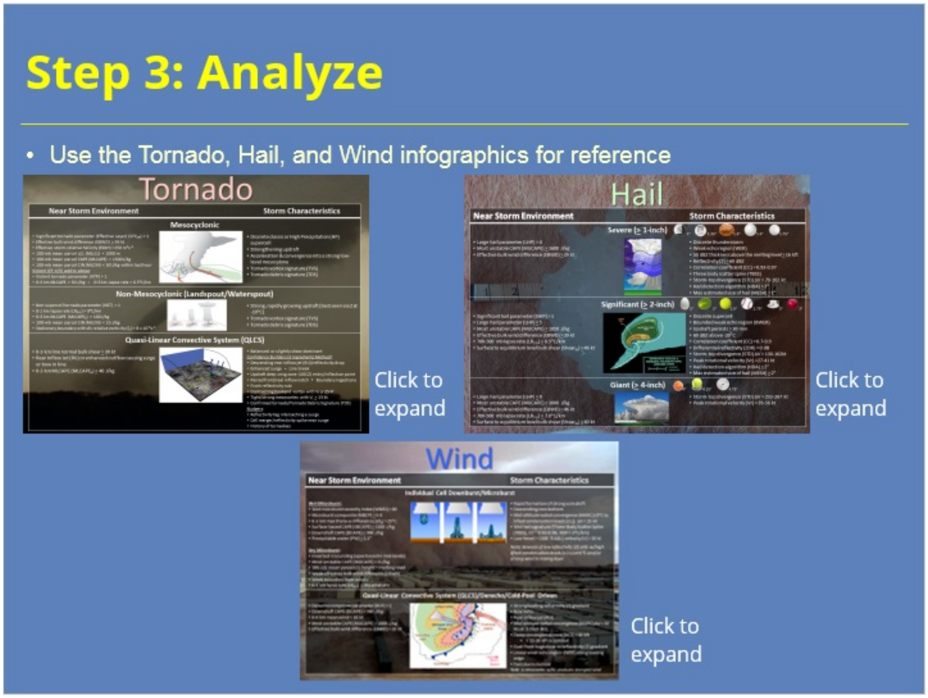
Identify the highest **Rank**ed storm. Factors to consider include: Near storm environment, storm reports, deviant motion, storm mode, rapidly-intensifying storms, signatures which suggest a severe storm (Inflow notch, three-body scatter spike (TBSS), hook echo, Tornado Debris Signature (TDS)), rear inflow jet (RIJ), etc.), societal/population considerations, and storms which don't have an appropriate warning or one that's due to expire soon.

1.7 Step 2: Rank (Cont'd)

Notes:

Go to step 4 to immediately issue a warning if the storm exhibits a high-confidence severe signature (such as TDS) and/or it has a high-confidence report, and it's unwarned, under-warned, or has a warning set to expire in less than 5 minutes. Otherwise, go to step 3.

1.8 Step 3: Analyze



Notes:

Step 3 is to **Analyze** the highest ranked storm's near storm environment and storm characteristics. Use the Tornado, Hail, and Wind infographics for reference.

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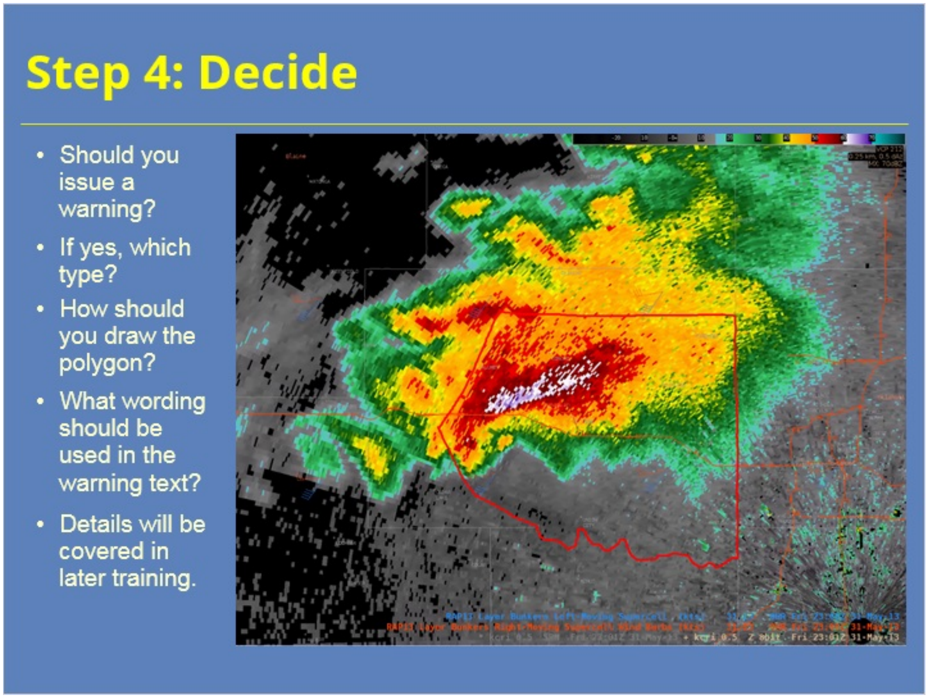
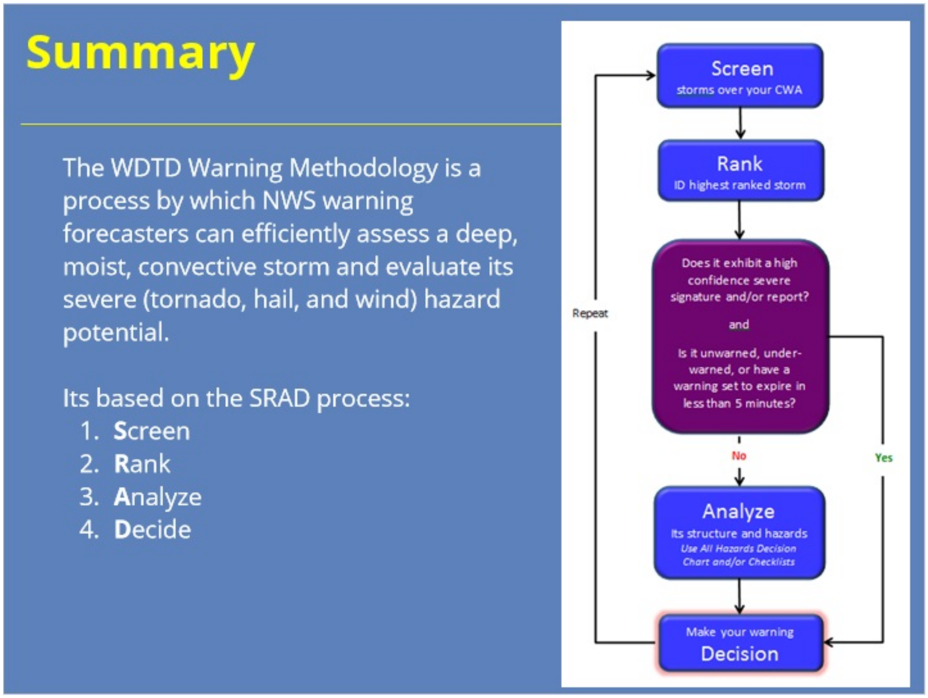
1.9 Step 4: Decide



Notes:

Step 4 is to **Decide**. Should you issue a warning? If yes, which warning type? How should you draw the polygon? What wording should be used in the warning text? Details on how to issue warnings will be covered in future training.

1.10 Warning Methodology Flow Chart

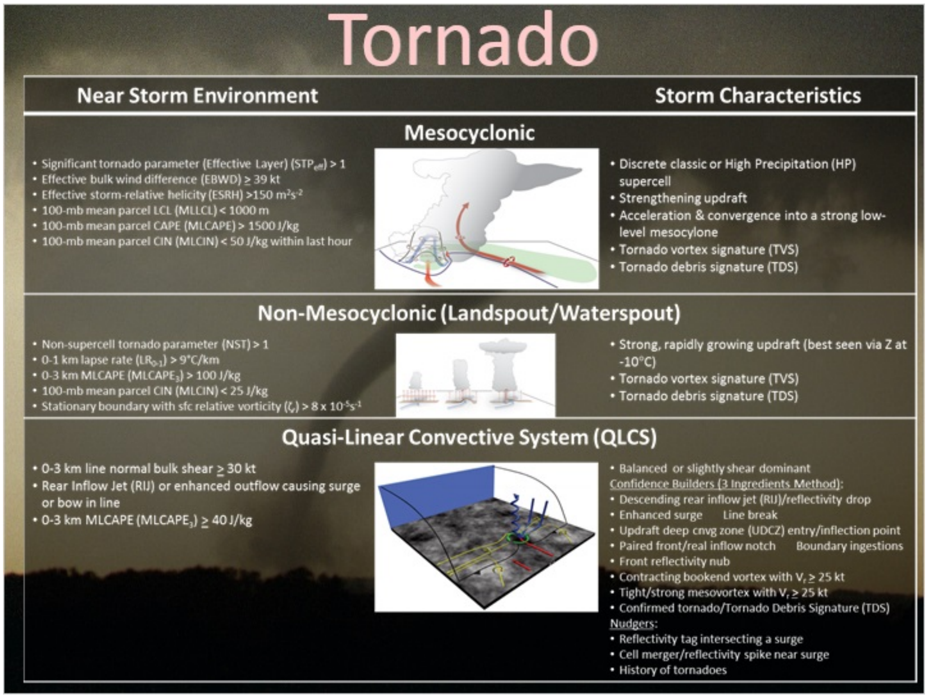


Notes:

The WDTD Warning Methodology is a process by which NWS warning forecasters can efficiently assess a deep, moist, convective storm and evaluate its severe hazard potential. Its based on the SRAD process: Screen, Rank, Analyze, and Decide. Here is a flow chart which illustrates the process.

3. Infographics

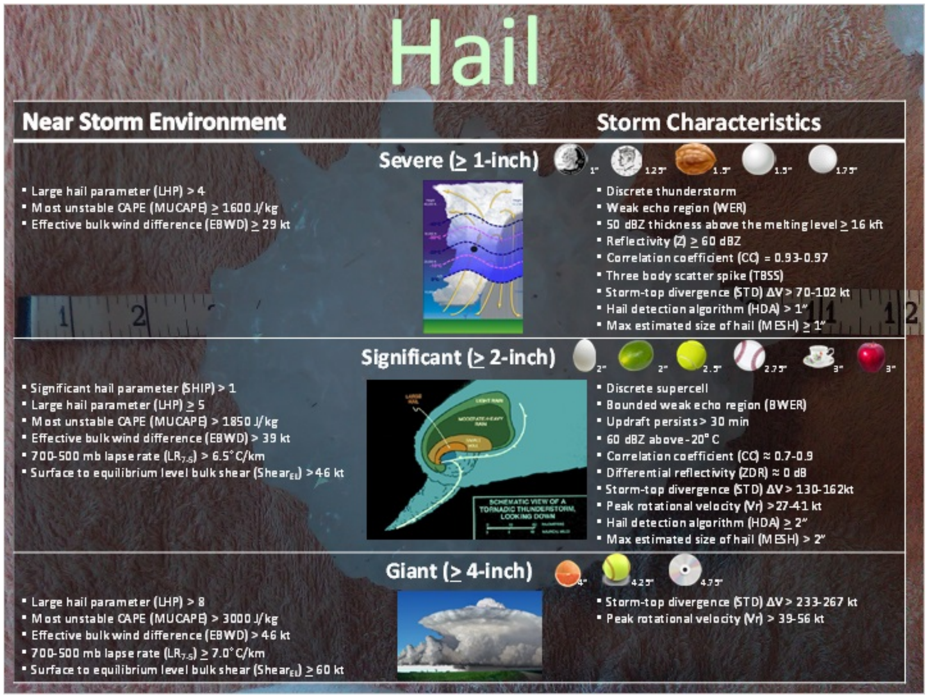
3.1 Tornado Infograpic



Notes:

This Tornado infographic provides both Near Storm Environmental and (primarily radar) Storm Characteristics for assessing a storm and evaluating its potential to produce mesocyclonic, non-mesocyclonic, and QLCS tornadoes.

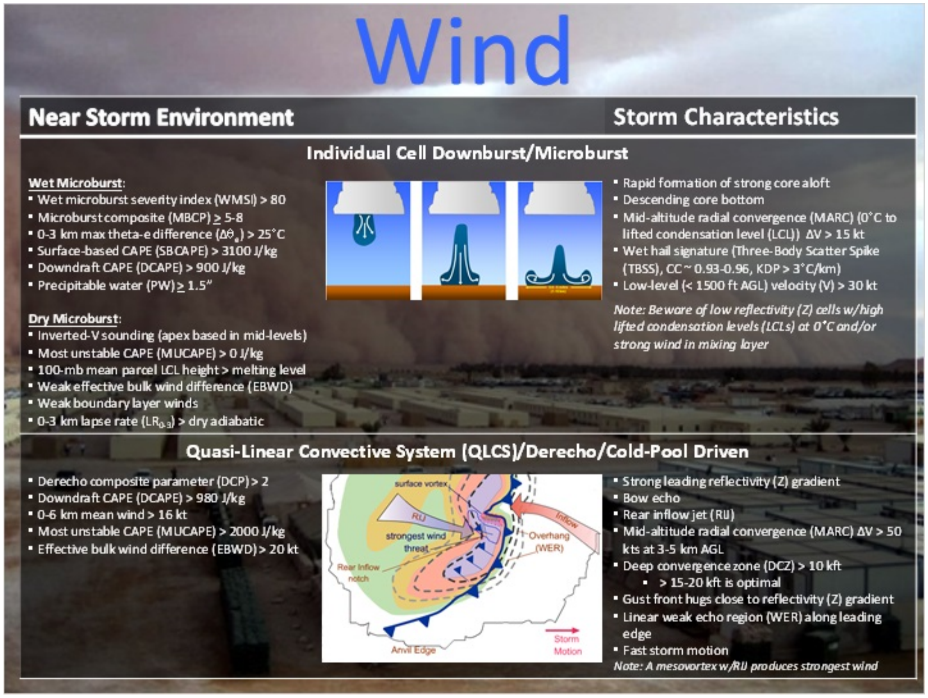
3.2 Hail Infographic



Notes:

This Hail infographic provides both Near Storm Environmental and (primarily radar) Storm Characteristics for assessing a storm and evaluating its potential to produce severe (≥1-inch), significant (≥2-inch), and giant hail (≥4-inch).

3.3 Wind Infographic



Notes:

This Wind infographic provides both Near Storm Environmental and (primarily radar) Storm Characteristics for assessing a storm and evaluating its potential to produce individual cell downburst/microburst and QLCS/derecho/cold-pool driven severe wind.

Tornado

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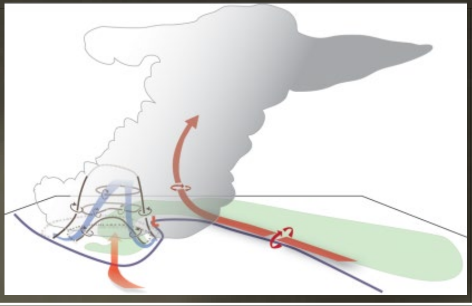
Near Storm Environment	Storm Characteristics

Mesocyclonic

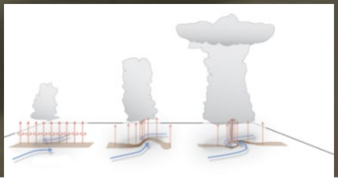
- Significant tornado parameter (Effective Layer) (STP_{eff}) > 1
- Effective bulk wind difference (EBWD) $\geq$ 39 kt
- Effective storm-relative helicity (ESRH) >150 m^2s^{-2}
- 100-mb mean parcel LCL (MLLCL) < 1000 m
- 100-mb mean parcel CAPE (MLCAPE) > 1500 J/kg
- 100-mb mean parcel CIN (MLCIN) < 50 J/kg within last hour



- Discrete classic or High Precipitation (HP) supercell
- Strengthening updraft
- Acceleration & convergence into a strong low-level mesocylone
- Tornado vortex signature (TVS)
- Tornado debris signature (TDS)

Non-Mesocyclonic (Landspout/Waterspout)

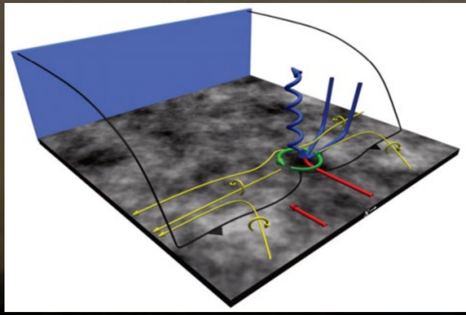
- Non-supercell tornado parameter (NST) > 1
- 0-1 km lapse rate (LR_{0-1}) > 9°C/km
- 0-3 km MLCAPE ($MLCAPE_3$) > 100 J/kg
- 100-mb mean parcel CIN (MLCIN) < 25 J/kg
- Stationary boundary with sfc relative vorticity (ζ_r) > 8 x $10^{-5}s^{-1}$



- Strong, rapidly growing updraft (best seen via Z at -10°C)
- Tornado vortex signature (TVS)
- Tornado debris signature (TDS)

Quasi-Linear Convective System (QLCS)

- 0-3 km line normal bulk shear $\geq$ 30 kt
- Rear Inflow Jet (RIJ) or enhanced outflow causing surge or bow in line
- 0-3 km MLCAPE ($MLCAPE_3$) $\geq$ 40 J/kg



- Balanced or slightly shear dominant

Confidence Builders (3 Ingredients Method):

- Descending rear inflow jet (RIJ)/reflectivity drop
- Enhanced surge • Line break
- Updraft deep cnvg zone (UDCZ) entry/inflection point
- Paired front/real inflow notch • Boundary ingestions
- Front reflectivity nub
- Contracting bookend vortex with $V_r \geq$ 25 kt
- Tight/strong mesovortex with $V_r \geq$ 25 kt
- Confirmed tornado/Tornado Debris Signature (TDS)

Nudgers:

- Reflectivity tag intersecting a surge
- Cell merger/reflectivity spike near surge
- History of tornadoes

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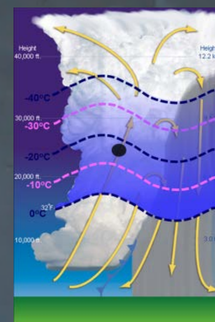
Hail

Near Storm Environment | Storm Characteristics

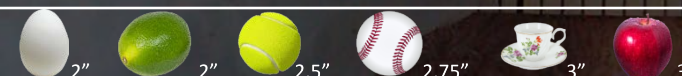
Severe (≥ 1-inch)

- Large hail parameter (LHP) > 4
- Most unstable CAPE (MUCAPE) ≥ 1600 J/kg
- Effective bulk wind difference (EBWD) ≥ 29 kt

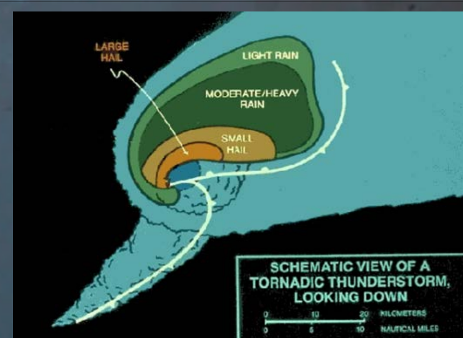


- Discrete thunderstorm
- Weak echo region (WER)
- 50 dBZ thickness above the melting level ≥ 16 kft
- Reflectivity (Z) ≥ 60 dBZ
- Correlation coefficient (CC) = 0.93-0.97
- Three body scatter spike (TBSS)
- Storm-top divergence (STD) ΔV > 70-102 kt
- Hail detection algorithm (HDA) ≥ 1"
- Max estimated size of hail (MESH) ≥ 1"

Significant (≥ 2-inch)

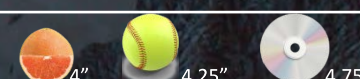


- Significant hail parameter (SHIP) > 1
- Large hail parameter (LHP) ≥ 5
- Most unstable CAPE (MUCAPE) ≥ 1850 J/kg
- Effective bulk wind difference (EBWD) ≥ 39 kt
- 700-500 mb lapse rate ($LR_{7\text{-}5}$) ≥ 6.5°C/km
- Surface to equilibrium level bulk shear ($Shear_{EL}$) ≥ 46 kt



- Discrete supercell
- Bounded weak echo region (BWER)
- Updraft persists ≥ 30 min
- 60 dBZ above -20°C
- Correlation coefficient (CC) ≈ 0.7-0.9
- Differential reflectivity (ZDR) ≈ 0 dB
- Storm-top divergence (STD) ΔV > 130-162kt
- Peak rotational velocity (Vr) >27-41 kt
- Hail detection algorithm (HDA) ≥ 2"
- Max estimated size of hail (MESH) ≥ 2"

Giant (≥ 4-inch)



- Large hail parameter (LHP) ≥ 8
- Most unstable CAPE (MUCAPE) ≥ 3000 J/kg
- Effective bulk wind difference (EBWD) ≥ 46 kt
- 700-500 mb lapse rate ($LR_{7\text{-}5}$) ≥ 7.0°C/km
- Surface to equilibrium level bulk shear ($Shear_{EL}$) ≥ 60 kt

- Storm-top divergence (STD) ΔV > 233-267 kt
- Peak rotational velocity (Vr) > 39-56 kt

Wind

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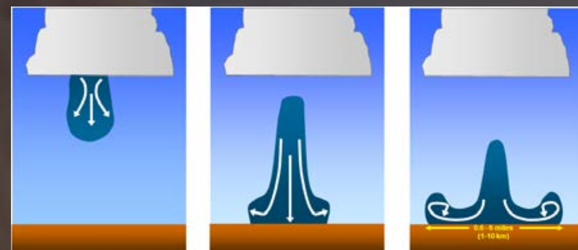
Near Storm Environment | Storm Characteristics

Individual Cell Downburst/Microburst

Wet Microburst:
- Wet microburst severity index (WMSI) > 80
- Microburst composite (MBCP) $\geq$ 5-8
- 0-3 km max theta-e difference ($\Delta\theta_e$) > 25°C
- Surface-based CAPE (SBCAPE) $\geq$ 3100 J/kg
- Downdraft CAPE (DCAPE) $\geq$ 900 J/kg
- Precipitable water (PW) $\geq$ 1.5”

Dry Microburst:
- Inverted-V sounding (apex based in mid-levels)
- Most unstable CAPE (MUCAPE) > 0 J/kg
- 100-mb mean parcel LCL height > melting level
- Weak effective bulk wind difference (EBWD)
- Weak boundary layer winds
- 0-3 km lapse rate ($LR_{0\text{-}3}$) $\geq$ dry adiabatic

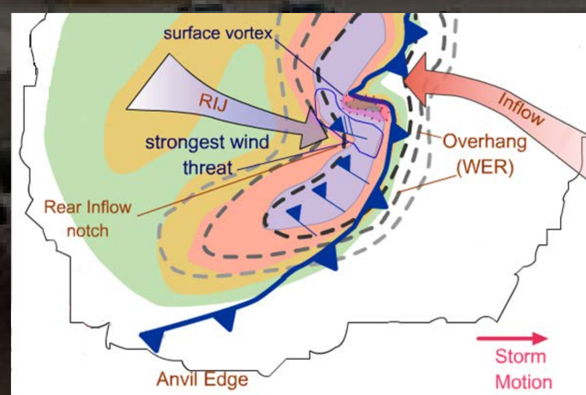


- Rapid formation of strong core aloft
- Descending core bottom
- Mid-altitude radial convergence (MARC) (0°C to lifted condensation level (LCL)) ΔV > 15 kt
- Wet hail signature (Three-Body Scatter Spike (TBSS), CC ~ 0.93-0.96, KDP > 3°C/km)
- Low-level (< 1500 ft AGL) velocity (V) > 30 kt

Note: Beware of low reflectivity (Z) cells w/high lifted condensation levels (LCLs) at 0℃ and/or strong wind in mixing layer

Quasi-Linear Convective System (QLCS)/Derecho/Cold-Pool Driven

- Derecho composite parameter (DCP) > 2
- Downdraft CAPE (DCAPE) > 980 J/kg
- 0-6 km mean wind > 16 kt
- Most unstable CAPE (MUCAPE) > 2000 J/kg
- Effective bulk wind difference (EBWD) > 20 kt



- Strong leading reflectivity (Z) gradient
- Bow echo
- Rear inflow jet (RIJ)
- Mid-altitude radial convergence (MARC) ΔV > 50 kts at 3-5 km AGL
- Deep convergence zone (DCZ) > 10 kft
    - > 15-20 kft is optimal
- Gust front hugs close to reflectivity (Z) gradient
- Linear weak echo region (WER) along leading edge
- Fast storm motion

Note: A mesovortex w/RIJ produces strongest wind

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Warning Methodology

Screen, Rank, Analyze, Decide (SRAD)

1. **Screen** the storms that threaten life and property over your CWA.

   - Severe Hazards (tornado/wind/hail): Load a 4-panel display showing a 60-minute loop of MRMS': Reflectivity at Lowest Altitude, Maximum Estimated Size of Hail (MESH) and 60-min MESH Tracks, 60-min 0-2 km Rotation Tracks, and Vertically Integrated Ice *(Note: An alternative could be a single-site lowest-tilt, Base Reflectivity, 60 minute time lapse loop with algorithm overlays. Use this alternative display if the MRMS products are experiencing latency.)*

2. Identify the highest **Rank**ed storm. Factors to consider include:

   - Near-storm environment
   - Storm reports
   - Rapidly-intensifying storms
   - Deviant motion (i.e., right-mover, left-mover)
   - Convective mode (ordinary cell, multicell, supercell, derecho, etc.)
   - Maximum Expected Size of Hail (MESH) value
   - Azimuthal shear / Rotation Tracks values
   - Signatures: Inflow notch, three-body scatter spike (TBSS), hook echo, Tornado Debris Signature (TDS), rear inflow jet (RIJ) etc.
   - Societal / population considerations
   - Storms which are under-warned or have a warning that's due to expire soon (<10 min)

   Go to Step 4 to immediately issue a warning for your highest ranked storm if:

   - It exhibits a high confidence severe signature (e.g., TDS) and/or it has a high confidence report, and
   - It's unwarned, under warned, or has a warning set to expire in less than 5 minutes.

   Otherwise, go to step 3.

3. **Analyze** the highest ranked storm's structure and hazards.

   - Use the "All Hazards Decision Chart" as a quick reference.
   - Use the Warning Decision Cycle checklists as detailed reference.
     - Updraft Strength
     - Tornado
     - Severe Hail
     - Severe Wind

4. Make your **Decision**. Consider the following factors when determining motion, duration, polygon orientation, and wording:

   - Tornado
     - Choose WarnGen Track type: "One Storm" and track the low-level vortex, but regard the parent storm's motion.
     - Be sure to account for possible mesocyclone occlusion(s) and motion uncertainty in your polygon (don't try to be too precise).

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- Capture multiple threats in close proximity with a single polygon when necessary.
- **Avoid:**
  - "Tornado Emergency" wording unless there is very high confidence of a significant (EF2+) tornado moving into an urban area.
- Non-mesocyclonic: Track the updraft interaction with the low-level boundary(ies).

- Severe Hail/Wind
  - Individual cell: Choose WarnGen Track type: "One Storm" and track the updraft/downdraft interface region; be sure to include both the updraft and downdraft regions in your polygon.
    - Supercell: Anticipate deviant motion; include the Rear Flank Downdraft (RFD) in your polygon.
  - Multicell: Choose WarnGen Track type: "One Storm" and track the area where cells mature; ensure polygon includes existing severe threat as well as anticipates new cell development.
    - Bow Echo/QLCS: Choose WarnGen Track type: "Line of Storms" and track the gust front; include trailing severe winds and hail in your polygon.

NOTE: One SRAD cycle (steps 1-4) should take about 5 minutes (with experience).

5. Repeat the SRAD process until no new warnings are required.

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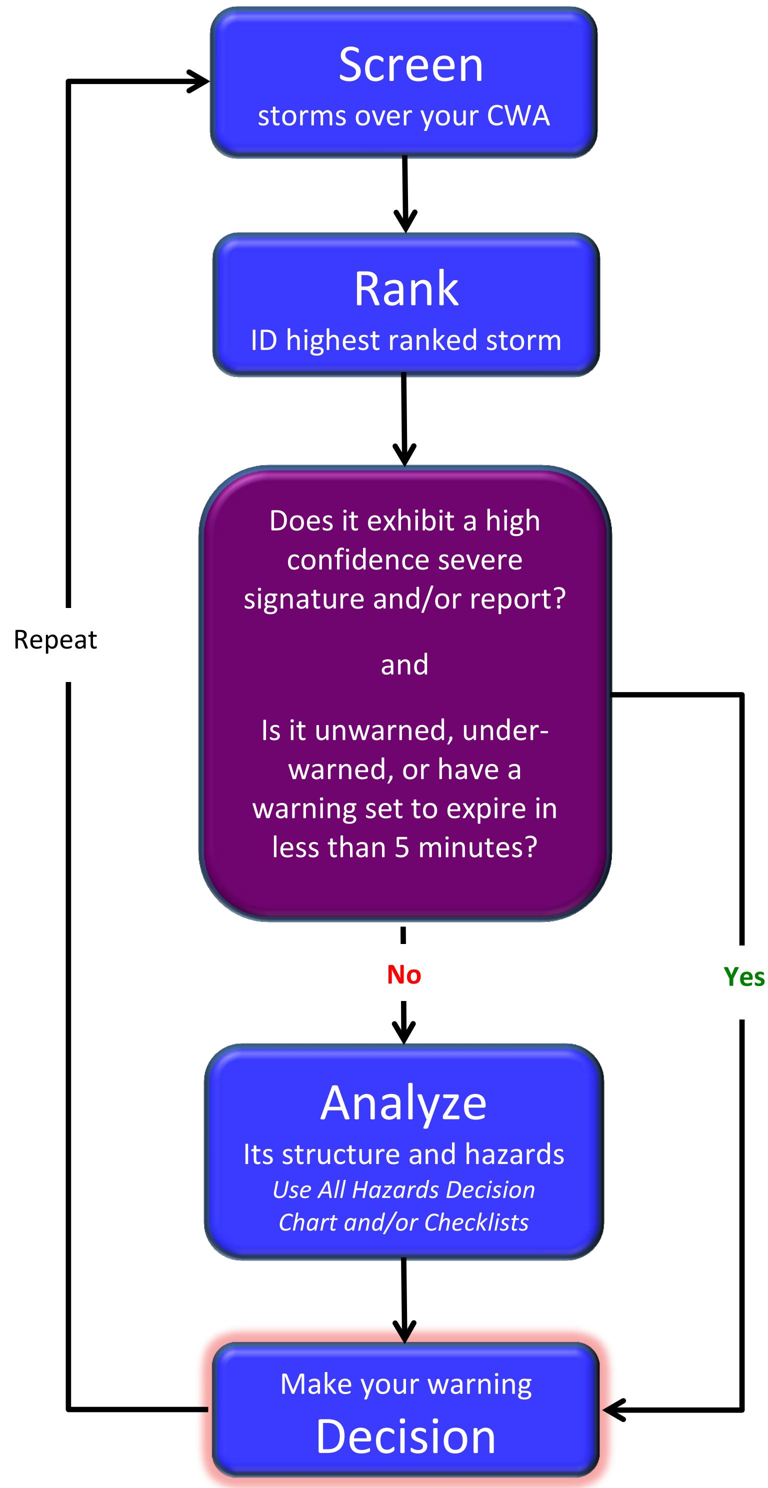
WDTD Suggested Warning Methodology:
Screen, Rank, Analyze, Decision (SRAD)

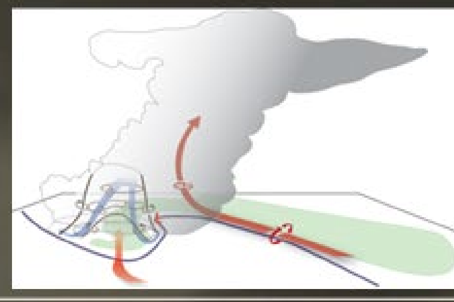


Tornado

Near Storm Environment	Storm Characteristics

Mesocyclonic

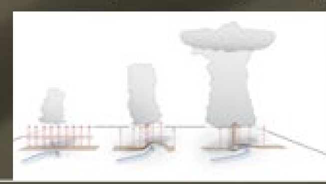
- Significant tornado parameter (Effective Layer) (STP_{eff}) > 1
- Effective bulk wind difference (EBWD) ≥ 39 kt
- Effective storm-relative helicity (ESRH) >150 m^2s^{-2}
- 100-mb mean parcel LCL (MLLCL) < 1000 m
- 100-mb mean parcel CAPE (MLCAPE) > 1500 J/kg
- 100-mb mean parcel CIN (MLCIN) < 50 J/kg within last hour



- Discrete classic or High Precipitation (HP) supercell
- Strengthening updraft
- Acceleration & convergence into a strong low-level mesocylone
- Tornado vortex signature (TVS)
- Tornado debris signature (TDS)

Non-Mesocyclonic (Landspout/Waterspout)

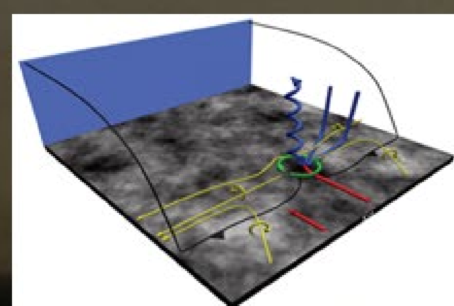
- Non-supercell tornado parameter (NST) > 1
- 0-1 km lapse rate (LR_{0-1}) > 9°C/km
- 0-3 km MLCAPE ($MLCAPE_3$) > 100 J/kg
- 100-mb mean parcel CIN (MLCIN) < 25 J/kg
- Stationary boundary with sfc relative vorticity (ζ_r) > 8 x $10^{-5}s^{-1}$



- Strong, rapidly growing updraft (best seen via Z at -10°C)
- Tornado vortex signature (TVS)
- Tornado debris signature (TDS)

Quasi-Linear Convective System (QLCS)

- 0-3 km line normal bulk shear ≥ 30 kt
- Rear Inflow Jet (RIJ) or enhanced outflow causing surge or bow in line
- 0-3 km MLCAPE ($MLCAPE_3$) ≥ 40 J/kg



- Balanced or slightly shear dominant

Confidence Builders (3 Ingredients Method):

- Descending rear inflow jet (RIJ)/reflectivity drop
- Enhanced surge
- Line break
- Updraft deep cnvg zone (UDCZ) entry/inflection point
- Paired front/real inflow notch
- Boundary ingestions
- Front reflectivity nub
- Contracting bookend vortex with V_r ≥ 25 kt
- Tight/strong mesovortex with V_r ≥ 25 kt
- Confirmed tornado/Tornado Debris Signature (TDS)

Nudgers:

- Reflectivity tag intersecting a surge
- Cell merger/reflectivity spike near surge
- History of tornadoes

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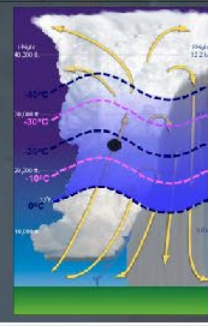
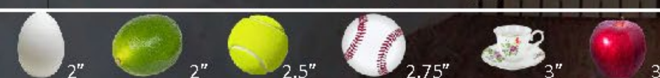
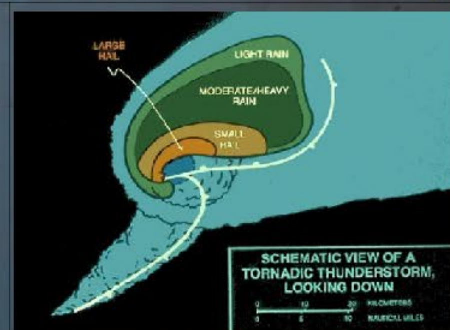
Hail

Near Storm Environment		Storm Characteristics
	Severe (≥ 1-inch) 	1" 1.25" 1.5" 1.5" 1.75"
• Large hail parameter (LHP) > 4 • Most unstable CAPE (MUCAPE) ≥ 1600 J/kg • Effective bulk wind difference (EBWD) ≥ 29 kt		• Discrete thunderstorm • Weak echo region (WER) • 50 dBZ thickness above the melting level ≥ 16 kft • Reflectivity (Z) ≥ 60 dBZ • Correlation coefficient (CC) = 0.93-0.97 • Three body scatter spike (TBSS) • Storm-top divergence (STD) ΔV > 70-102 kt • Hail detection algorithm (HDA) ≥ 1" • Max estimated size of hail (MESH) ≥ 1"
	Significant (≥ 2-inch) 	2" 2" 2.5" 2.75" 3" 3"
• Significant hail parameter (SHIP) > 1 • Large hail parameter (LHP) ≥ 5 • Most unstable CAPE (MUCAPE) ≥ 1850 J/kg • Effective bulk wind difference (EBWD) ≥ 39 kt • 700-500 mb lapse rate (LR_{7-5}) ≥ 6.5°C/km • Surface to equilibrium level bulk shear ($Shear_{EL}$) ≥ 46 kt		• Discrete supercell • Bounded weak echo region (BWER) • Updraft persists ≥ 30 min • 60 dBZ above -20°C • Correlation coefficient (CC) ≈ 0.7-0.9 • Differential reflectivity (ZDR) ≈ 0 dB • Storm-top divergence (STD) ΔV > 130-162kt • Peak rotational velocity (Vr) >27-41 kt • Hail detection algorithm (HDA) ≥ 2" • Max estimated size of hail (MESH) ≥ 2"
	Giant (≥ 4-inch) 	4" 4.25" 4.75"
• Large hail parameter (LHP) ≥ 8 • Most unstable CAPE (MUCAPE) ≥ 3000 J/kg • Effective bulk wind difference (EBWD) ≥ 46 kt • 700-500 mb lapse rate (LR_{7-5}) ≥ 7.0°C/km • Surface to equilibrium level bulk shear ($Shear_{EL}$) ≥ 60 kt		• Storm-top divergence (STD) ΔV > 233-267 kt • Peak rotational velocity (Vr) > 39-56 kt

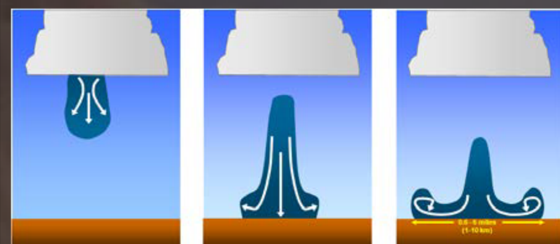
Wind

Near Storm Environment | Storm Characteristics

Individual Cell Downburst/Microburst

Wet Microburst:
- Wet microburst severity index (WMSI) > 80
- Microburst composite (MBCP) ≥ 5-8
- 0-3 km max theta-e difference ($\Delta\theta_e$) > 25°C
- Surface-based CAPE (SBCAPE) ≥ 3100 J/kg
- Downdraft CAPE (DCAPE) ≥ 900 J/kg
- Precipitable water (PW) ≥ 1.5"

Dry Microburst:
- Inverted-V sounding (apex based in mid-levels)
- Most unstable CAPE (MUCAPE) > 0 J/kg
- 100-mb mean parcel LCL height > melting level
- Weak effective bulk wind difference (EBWD)
- Weak boundary layer winds
- 0-3 km lapse rate (LR_{0-3}) ≥ dry adiabatic

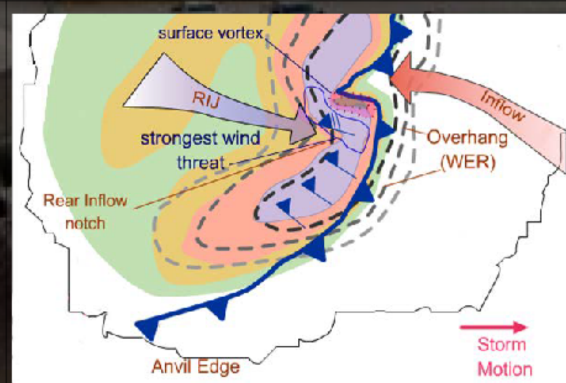


- Rapid formation of strong core aloft
- Descending core bottom
- Mid-altitude radial convergence (MARC) (0°C to lifted condensation level (LCL)) ΔV > 15 kt
- Wet hail signature (Three-Body Scatter Spike (TBSS), CC ~ 0.93-0.96, KDP > 3°C/km)
- Low-level (< 1500 ft AGL) velocity (V) > 30 kt

Note: Beware of low reflectivity (Z) cells w/high lifted condensation levels (LCLs) at 0°C and/or strong wind in mixing layer

Quasi-Linear Convective System (QLCS)/Derecho/Cold-Pool Driven

- Derecho composite parameter (DCP) > 2
- Downdraft CAPE (DCAPE) > 980 J/kg
- 0-6 km mean wind > 16 kt
- Most unstable CAPE (MUCAPE) > 2000 J/kg
- Effective bulk wind difference (EBWD) > 20 kt



- Strong leading reflectivity (Z) gradient
- Bow echo
- Rear inflow jet (RIJ)
- Mid-altitude radial convergence (MARC) ΔV > 50 kts at 3-5 km AGL
- Deep convergence zone (DCZ) > 10 kft
  - > 15-20 kft is optimal
- Gust front hugs close to reflectivity (Z) gradient
- Linear weak echo region (WER) along leading edge
- Fast storm motion

Note: A mesovortex w/RIJ produces strongest wind

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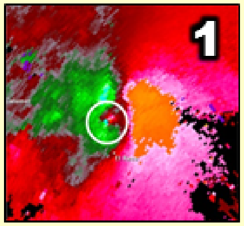
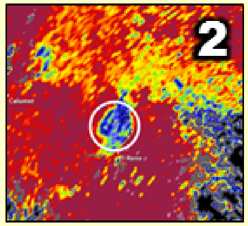
Radar Tornado Intensity Estimation Guidance

Identifying a Tornadic Debris Signature (TDS)

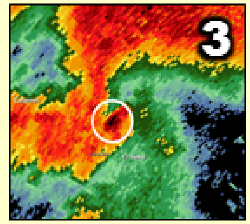
Provides radar confirmation of a damaging tornado in progress.



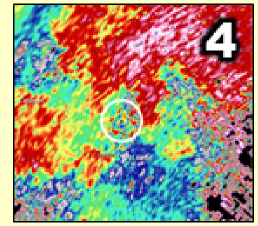
First, identify a valid velocity circulation.



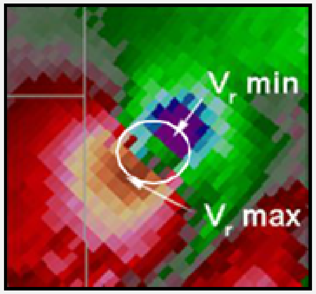
Next, ensure correlation coefficient (CC) is below 0.90



Next, ensure reflectivity is over 35 dBZ and co-located with #1/2



Not necessary but adds confidence: ZDR reduced to ~0 or below zero in spots.



$$V_{rot} = (|V_{in[max]}| + |V_{out[max]}|) / 2$$

To determine rotational velocity, add the absolute value of the highest inbound and outbound velocity values in the couplet, and then divide by 2.

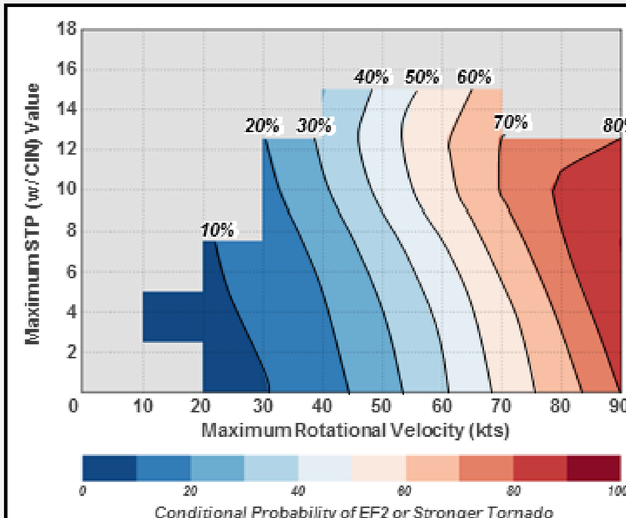
Considerations and Tips

- EF2+ tornadoes are likely if TDS has debris ball (reflectivity > 50-55 dBZ)
- With split cut mode VCPs, TDS can have a slight offset from velocity sig.
- **Discriminating between supercellular weak and strong tornadoes**: Heidke Skill Scores maximized with LLRV in the 45-55 knot range.
- **In borderline intensity cases, push up a category if**: tornado is moving fast, conditions very favorable for EF2+, or signature is poorly sampled.

Only valid within 70nm of the radar site

Tornado Intensity	Rotational Velocity (kts) – Supercells Only	Rotational Velocity (kts) – QLCS Only	Maximum TDS Height (MOST RELIABLE)
WEAK *EF0/EF1*	*40 knots or less*	*30 knots or less*	*Under 8,000 ft*
	Overlap 40-55 knots	Overlap 30-45 knots	Overlap 8-10 kft
STRONG *EF2/EF3*	*55 to 75 knots*	*45 knots or more*	*10,000 to 15,000 ft*
	Overlap 75-85 knots		Overlap 15-18 kft
VIOLENT *EF4/EF5*	*85 knots or more*	*Insufficient cases*	*Over 18,000 ft*

Conditional EF2+ Tor Probability



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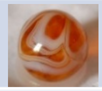
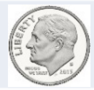
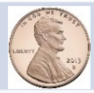
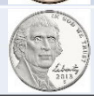
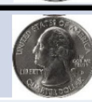
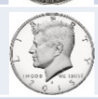
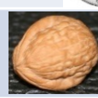
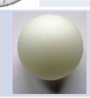
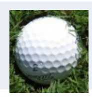
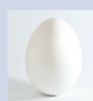
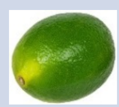
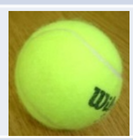
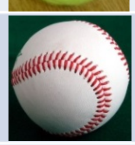
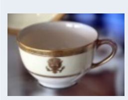
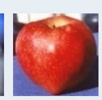
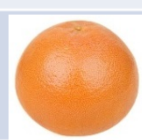
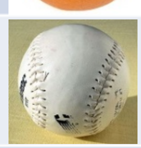
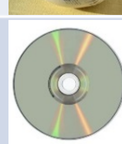
NWS Hail Size Chart

Description		Diameter	Updraft Speed
BB		< ¼"	< 24 mph
Pea		¼"	24 mph
Marble / Plain M&M	 	½"	35 mph
Dime		$^{7}/_{10}$"	38 mph
Penny		¾"	40 mph
Nickel		$^{7}/_{8}$"	46 mph
Quarter		*(Severe)* 1"	49 mph
Half Dollar		1 ¼"	54 mph
Walnut / Ping-Pong Ball	 	1 ½"	60 mph
Golf Ball		1 ¾"	64 mph
Hen Egg / Lime	 	*(Significant)* 2"	69 mph
Tennis Ball		2 ½"	77 mph
Baseball		2 ¾"	81 mph
Teacup / Large Apple	 	3"	84 mph
Grapefruit		4"	98 mph
Softball		4 ½"	103 mph
CD / DVD		4 ¾"	105 mph

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Radar Estimated Hail Type/Size

Storm-Top Divergence

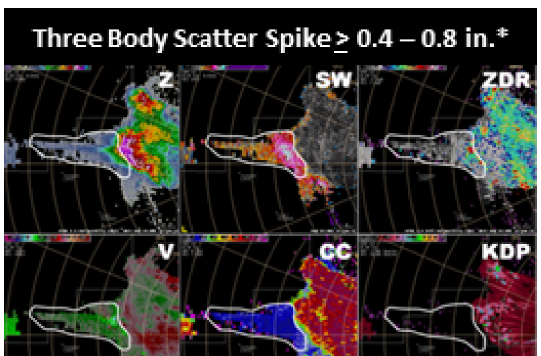
Peak ΔV (kts)	Max Hail Size (in.)
70-102	Quarter (1")
115-147	Golf ball (1 ¾")
174-207	Baseball (2 ¾")
233-267	Grapefruit (4")

Adapted from Witt and Nelson, 1991

Mesocyclone

Hail Size (in.)	Peak Rotational Velocity (kt)
1.75" to 2.00"	27-41
≥4"	39-56

Source: Blair et al., 2011



*Valid for S-band radar only, Source: Kumjian et al., 2010

DUAL-POL RADAR HAIL SIGNATURES

	Z: 45-59 dBZ = Hail poss; ≥60 dBZ = Hail likely	**ZDR**: -0.3 to 1 dB ≈ Dry or large hail; > 1 dB ≈ More liquid
	CC: 0.93 - 0.97 ≈ 1-2" hail; 0.70 - 0.90 ≈ ≥2" hail	**KDP**: <1°/km ≈ Mostly dry hail; >3°/km ≈ Rain/hail combo or melting hail
Hail Event Type	**Signature**	
Severe Hail (with little rain)	Z > 55 dBZ	ZDR < 1 dB
	CC ≈ 0.95-0.97	KDP < 1°/km
Severe Hail Mixed w/Rain	Z > 55 dBZ	ZDR ≈ 1-2 dB
	CC ~0.93-0.96	KDP > 0.5°/km
Sub-Severe Dry Hail	Z ≈ 45-55 dBZ	ZDR ≈ 0 dB
	CC > 0.98	KDP ≈ 0°/km
Sub-Severe Melting Hail	Z > 55 dBZ	ZDR > 2 dB
	CC ≈ 0.92-0.96	KDP > 4-5°/km
Significant (≥2") Hail	Z > 55 dBZ (>45 dBZ)	ZDR ≈ 0 dB or lower
	CC < 0.9 (possibly 0.7)	KDP not displayed