14 THUNDERSTORM FUNDAMENTALS

Contents

Thunderstorm characteristics, formation, and forecasting are covered in this chapter. The next chapter covers thunderstorm hazards including hail, gust fronts, lightning, and tornadoes.

14.1. THUNDERSTORM CHARACTERISTICS

Thunderstorms are **convective clouds** with large vertical extent, often with tops near the tropopause and bases near the top of the boundary layer. Their official name is **cumulonimbus** (see the Clouds Chapter), for which the abbreviation is **Cb**. On weather maps the symbol represents thunderstorms, with a dot •, asterisk *, or triangle ∆ drawn just above the top of the symbol to indicate rain, snow, or hail, respectively. For severe thunderstorms, the symbol is .

14.1.1. Appearance

A mature thunderstorm cloud looks like a mushroom or anvil with a relatively large-diameter flat top. The simplest thunderstorm (see Figs. 14.1 & 14.2) has a nearly vertical stem of diameter roughly equal to its depth (of order 10 to 15 km). The large top is called the **anvil**, **anvil cloud**, or **thunderhead**, and has the official name **incus** (Latin for anvil). The anvil extends furthest in a direction as blown by the upper-tropospheric winds.

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Figure 14.1
Airmass thunderstorm having a single mature cell.

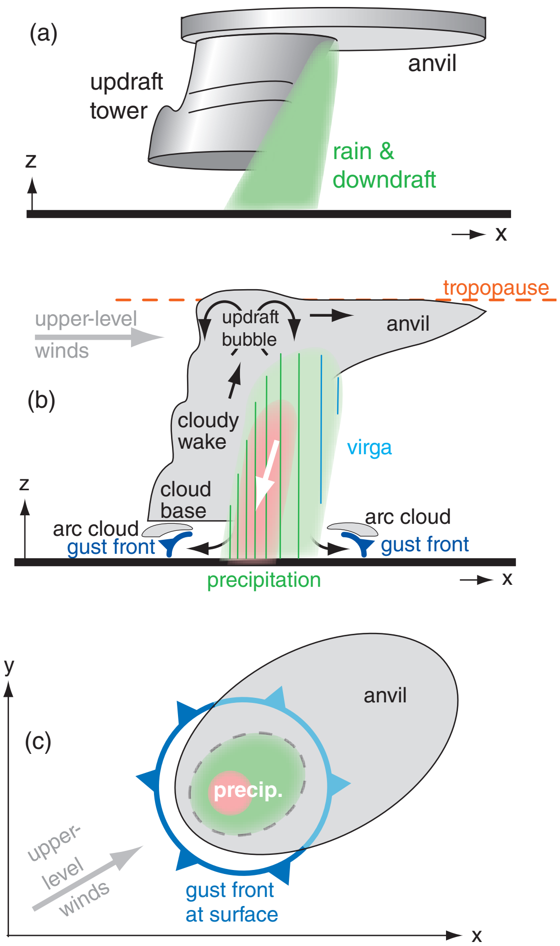

Figure 14.2
(a) Sketch of a basic (airmass) thunderstorm in its mature stage. (b) Vertical slice through the storm. Light shading indicates clouds, green and red shadings are moderate and heavy precipitation, and arrows show air motion. (c) Horizontal composite, showing the anvil at storm top (as viewed from above by satellite), the precipitation in the low-to-middle levels (as viewed by radar), and the gust front of spreading winds at the surface.

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Figure 14.3
Photo of supercell thunderstorm.

If the thunderstorm top is just starting to spread out into an anvil and does not yet have a fibrous or streaky appearance, then you identify the cloud as **cumulonimbus calvus** (see the Clouds Chapter). For a storm with a larger anvil that looks strongly **glaciated** (i.e., has a fibrous appearance associated with ice-crystal clouds), then you would call the cloud a **cumulonimbus capillatus**.

Within the stem of a mature thunderstorm is the cloudy **main updraft tower** topped by an **updraft bubble** (Fig. 14.2b). When this rising air hits the tropopause, it spreads to make the anvil. Also in the stem is a downdraft with precipitation. When the downdraft air hits the ground it spreads out, the leading edge of which is called the **gust front**. When viewed from the ground under the storm, the main updraft often has a darker cloud base, while the rainy region often looks not as dark and does not have a well-defined cloud base.

Not all cumulonimbus clouds have lightning and thunder. Such storms are technically not thunderstorms. However, in this book we will use the word **thunderstorm** to mean any cumulonimbus cloud, regardless of whether it has lightning.

More complex thunderstorms can have one or more updraft and downdraft regions. The most severe, long-lasting, less-frequent thunderstorms are **supercell thunderstorms** (Figs. 14.3 & 14.4).

14.1.2. Clouds Associated with Thunderstorms

Sometimes you can see other clouds attached to thunderstorms, such as a funnel, wall, mammatus, arc, shelf, flanking line, scud, pileus, dome, and beaver tail (Fig. 14.4). Not all thunderstorms have all these associated clouds.

Arc clouds (official name **arcus**, Fig. 14.2b) or **shelf clouds** form near the ground in boundary-layer air that is forced upward by undercutting cold air flowing out from the thunderstorm. These cloud bands mark the leading edge of **gust-front** outflow from the **rear-flank downdraft** (Fig. 14.4), usually associated with the **flanking line**. Often the undersides of arc clouds are dark and turbulent-looking, while their tops are smooth. Not all gust fronts have arc clouds, particularly if the displaced air is dry. See the Thunderstorm Hazards chapter.

The **beaver tail** (Fig. 14.4) is a smooth, flat, narrow, low-altitude cloud that extends along the boundary between the inflow of warm moist air to the thunderstorm and the cold air from the rain-induced **forward flank downdraft** (**FFD**).

A **dome** of **overshooting** clouds sometimes forms above the anvil top, above the region of strongest updraft. This is caused by the inertia of the upward moving air in the main updraft, which overshoots above its neutrally buoyant **equilibrium**

level (**EL**). Storms that have **overshooting tops** are often more violent and turbulent.

The **flanking line** is a band of cumuliform clouds that increase from the medium-size **cumulus mediocris** (**Cu med**) furthest from the storm to the taller **cumulus congestus** (**Cu con**) close to the **main updraft**. Cumulus congestus are also informally called **towering cumulus** (**TCu**). The flanking line forms along and above the gust front, which marks the leading edge of colder outflow air from the **rear-flank downdraft** (**RFD**).

If humid layers of environmental air exist above rapidly rising cumulus towers, then **pileus** clouds can form as the environmental air is pushed up and out of the way of the rising cumulus clouds. Pileus are often very short lived because the rising cloud tower that caused it often keeps rising through the pileus and obliterates it.

The most violent thunderstorms are called **supercell** storms (Figs. 14.3 & 14.4), and usually have a quasi-steady rotating updraft (called a **mesocyclone**). The main thunderstorm updraft in supercells sometimes has curved, helical **cloud striations** (grooves or ridges) on its outside similar to threads of a screw (Fig. 14.4a). Supercells can produce intense **tornadoes** (violently rotating columns of air), which can appear out of the bottom of an isolated cylindrical lowering of the cloud base called a **wall cloud**. The portion of the tornado made visible by cloud droplets is called the **funnel cloud**, which is the name given to tornadoes not touching the ground. Tornadoes are covered in the next chapter. Most thunderstorms are not supercell storms, and most supercell storms do not have tornadoes.

Attached to the base of the wall cloud is sometimes a short, horizontal cloud called a **tail cloud**, which points towards the **forward flank precipitation** area. Ragged cloud fragments called **scud** (**cumulus fractus**) often form near the tip of the tail cloud and are drawn quickly into the tail and the wall cloud by strong winds.

The wall cloud and tornado are usually near the boundary between cold downdraft air and the low-level warm moist inflow air, under a somewhat **rain-free cloud base**. In mesocyclone storms, you can see the wall cloud rotate by eye, while rotation of the larger-diameter supercell is slower and is apparent only in time-lapse movies. Non-rotating wall clouds also occur with non-rotating thunderstorms (i.e., non-supercell storms).

The **rain** portion of a thunderstorm is often downwind of the main updraft, in the forward portion of the storm. Cloud edges and boundaries are often not distinguishable in the rain region. Although thunderstorms are rare in cold weather, when they occur they can cause very heavy snowfall instead of

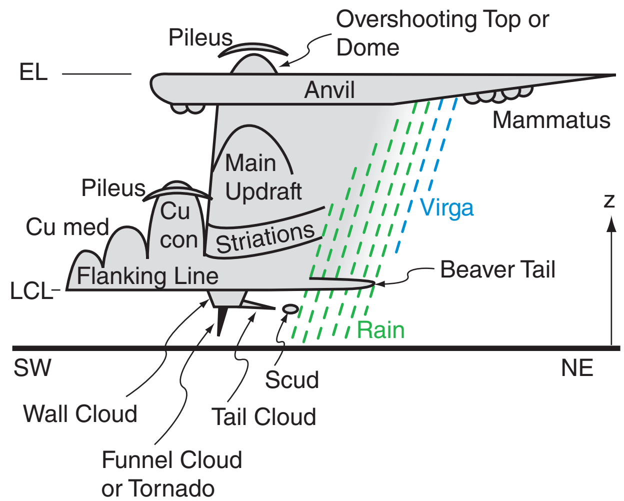

Figure 14.4a
Sketch of a classic supercell thunderstorm (Cb) as might be viewed looking toward the northwest in central North America. The storm would move from left to right in this view (i.e., toward the northeast). Many storms have only a subset of the features cataloged here. Cu med = cumulus mediocris; Cu con = cumulus congestus; LCL = lifting condensation level; EL = equilibrium level (often near the tropopause, 8 to 15 km above ground); NE = northeast; SW = southwest.

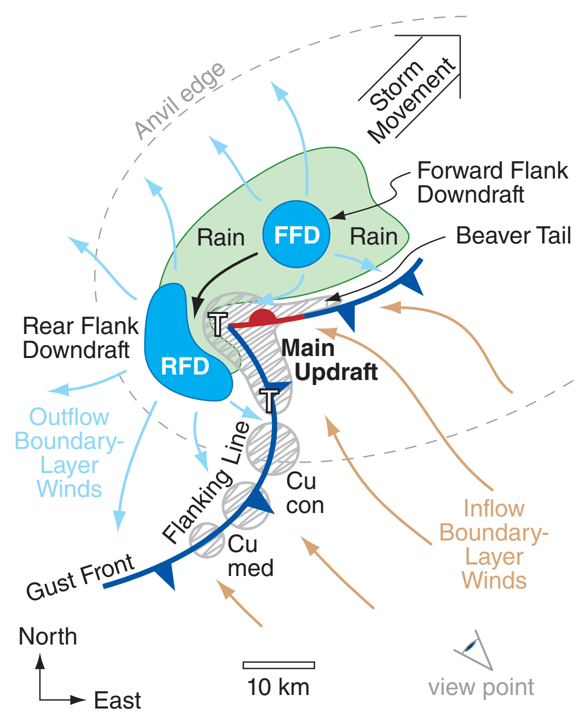

Figure 14.4b
Plan view (top down) sketch of a classic supercell thunderstorm. 𝕋 indicates possible tornado positions; RFD = Rear Flank Downdraft; FFD = Forward Flank Downdraft. Regions of updraft are hatched with grey; downdrafts are solid blue; rain is solid light green. Low surface pressure is found under the updrafts (especially near the "𝕋" locations), and high pressure under the downdrafts. Vectors are near-surface winds: cold are cyan, warm are tan. From the view-point location in the bottom right of Fig. 14.4a, you see a storm as sketched in Fig. 14.4a.

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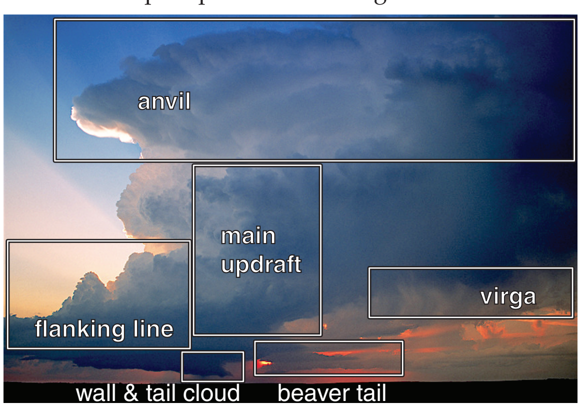
Figure 14.5
Mammatus clouds.

Sample Application
Identify the thunderstorm features, components, and associated clouds that are in Fig. 14.3.

Find the Answer
Given: Photo 14.3, reproduced here.
Find: Cloud features.

Method: Compare photo 14.3 to diagram 14.4a.



Base photo: © Gene Rhoden / weatherpix.com

Exposition: This cloud is a supercell thunderstorm, because it looks more like the cloud sketched in Fig. 14.4a than the simple single-cell air-mass thunderstorm in Fig. 14.2. The main updraft is tilted from lower left to upper right, a sign of wind shear in the environment. If a tornado exists under the wall cloud, it is too small to see in this figure. Given the orientation of the cloud features (similar to those in Fig 14.4a), the photographer was probably looking toward the northwest, and the thunderstorm is moving toward the northeast (to the right in this figure).

rain. Further downwind of the heavy rain area is a region of **virga**, where the precipitation (rain or snow) evaporates before reaching the ground.

Mammatus clouds look like sacks or protuberances hanging from the bottom of layered clouds, and are sometimes found on the underside of thunderstorm anvils (Fig. 14.5). No correlation exists between the presence of mammatus clouds and the intensity of thunderstorms (i.e., hail, tornadoes, etc.). Not all thunderstorms have mammatus clouds.

Mammatus forms because static instability in the anvil-base air drives convective overturning. The colder, droplet-laden cloudy air is sinking from the anvil cloud as mammatus lobes. Between each lobe is a return circulation of upward-moving warmer drier non-cloudy air from just under cloud base. Mammatus likelihood is greatest when the air just under the anvil is very warm and dry.

Two processes have been suggested for causing this static instability. One is infrared radiation: cloud droplets are better emitters of IR radiation than dry air, which causes the cloudy lobes to cool faster than the drier air below anvil base. The other process is **cloud-base detrainment instability**, which works as follows. Sinking air in the mammatus lobes are laden with cloud droplets or ice crystals. As some of the sinking air mixes out of the lobe (**detrains**) into the surrounding environment, evaporation of some of the water from the detrained air helps keep the mammatus clouds cooler than their environment, due to the latent heat of vaporization.

In an environment with weak wind shear, the **anvil cloud** in weak thunderstorms can be symmetrical, looking like a mushroom cap (Fig. 14.1). However, for larger, longer lasting thunderstorms in the presence of strong winds aloft, the ice crystals brought up by the main updraft will spread out asymmetrically (Fig. 14.2) in an anvil that can be 100 km wide (crosswind) and 300 km long (downwind).

If you want to chase and photograph isolated storms in N. America, the best place to be is well to the side of the storm's path. As most storms move northeast in N. America, this corresponds to your being southeast of the thunderstorm. This location is relatively safe with good views of the storm and the tornado (see "A SCI. PERSPECTIVE • Be Safe").

For non-chasers overtaken by a storm, the sequence of events they see is: cirrus anvil clouds, mammatus clouds (if any), gust front (& arc cloud, if any), rain, hail (if any), tornado (if any), and rapid clearing with sometimes a rainbow during daytime.

14.1.3. Cells & Evolution

The fundamental structural unit of a thunderstorm is a **thunderstorm Byers-Braham cell**, which is of order 10 km in diameter and 10 km thickness (i.e., its **aspect ratio** = depth/width = 1). Its evolution has three stages: (1) towering-cumulus; (2) mature; and (3) dissipation, as sketched in Fig. 14.6. Each individual cell lasts about 30 to 60 min.

The towering cumulus stage consists of only updraft in a rapidly growing cumulus congestus tower. It has no anvil, no downdrafts, and no precipitation. It is drawing warm humid boundary-layer air into the base of the storm, to serve as fuel. As this air rises the water vapor condenses and releases latent heat, which adds buoyancy to power the storm.

In the mature stage (Fig. 14.7), the thunderstorm has an anvil, both updrafts and downdrafts, and heavy precipitation of large-size drops. This often is the most violent time during the storm, with possible lightning, strong turbulence and winds. The precipitation brings cooler air downward as a downburst, which hits the ground and spreads out. In the absence of wind shear, two factors conspire to limit the lifetime of the storm (Fig. 14.6): (1) the storm runs out of fuel as it consumes nearby boundary-layer air; and (2) the colder outflow air chokes off access to adjacent warm boundary-layer air.

In the dissipating stage, the storm consists of downdrafts, precipitation, and a large **glaciated** (containing ice crystals) anvil. The storm rains itself out, and the precipitation rate diminishes. The violent aspects of the storm are also diminishing. With no updraft, it is not bringing in more fuel for the storm, and thus the storm dies.

The last frame in Fig. 14.6 shows the anvil-debris stage, which is not one of the official stages listed in severe-weather references. In this stage all that remains of the former thunderstorm is the anvil and associated virga, which contains many ice crystals that fall and evaporate slowly. No precipitation reaches the ground. The anvil remnant is not very turbulent, and drifts downwind as a thick cirrostratus or altostratus cloud as it gradually disappears.

Meanwhile, the cold outflow air spreading on the ground from the former storm might encounter fresh boundary-layer air, which it lifts along its leading edge (gust front) to possibly trigger a new thunderstorm (labeled "Cu con 2" in the last frame of Fig. 14.6). This process is called **storm propagation**, where the first storm (the "mother") triggers a daughter storm. This daughter storm can go through the full thunderstorm life cycle if conditions are right, and trigger a granddaughter storm, and so forth.

Some thunderstorms contain many cells that are in different stages of their evolution. Other storms have special characteristics that allow the cell to be maintained for much longer than an hour. The types of thunderstorms are described next.

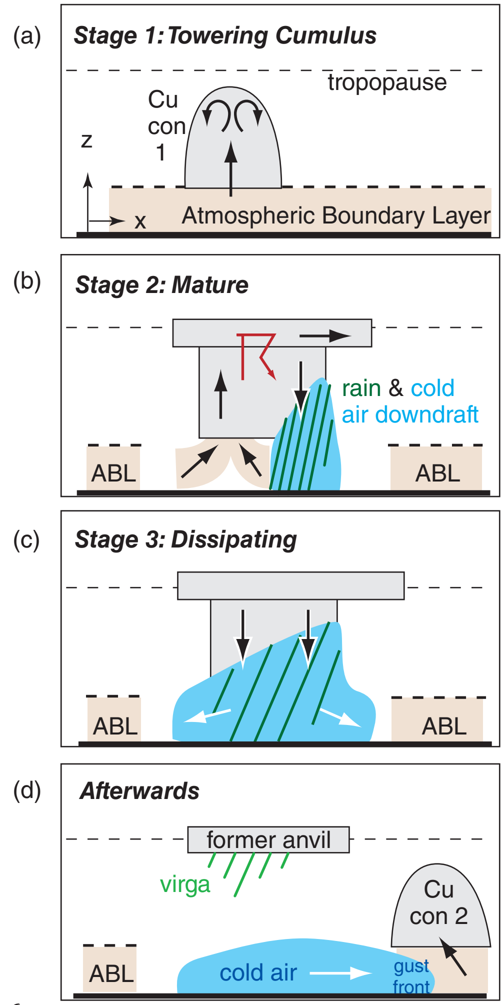


Figure 14.6
Phases of thunderstorm cell evolution. Light grey shading represents clouds, tan color is pre-storm boundary layer (ABL) of warm humid air that serves as fuel for the storm. Blue shading is colder outflow air. Diagonal dark green lines represent precipitation. Arrows show air motion. Cu con = cumulus congestus.

A SCIENTIFIC PERSPECTIVE • Be Safe

In any scientific or engineering work, determine the possible hazards to you and others before you undertake any action. If you are unable to increase the safety, then consider alternative actions that are safer. If no action is relatively safe, then cancel that project.

For example, chasing thunderstorms can be very hazardous. Weather conditions (heavy rainfall, poor visibility, strong winds, hail, lightning, and tornadoes) can be hazardous. Also, you could have poor road conditions and a limited ability to maneuver to get out of harm's way in many places.

Your safety and the safety of other drivers and pedestrians must always be your first concern when storm chasing. Some storm-chase guidelines are provided in a series of "A SCIENTIFIC PERSPECTIVE" boxes later in this and the next chapters.

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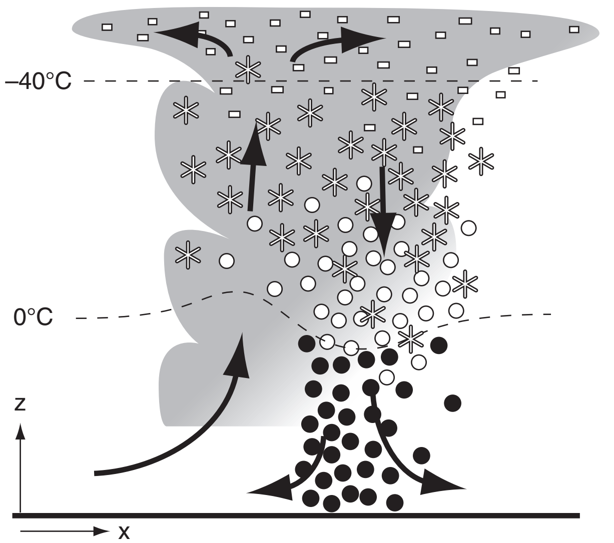


Figure 14.7
Airmass Byers-Braham thunderstorm in its mature stage during summer. Grey shading represents cloud droplets (except above the –40°C isotherm, where all hydrometeors are frozen), black dots represent rain, white dots are graupel, dendrite shapes are snow, and small white rectangles are small ice crystals. Arrows show winds, and dashed lines give air temperature.

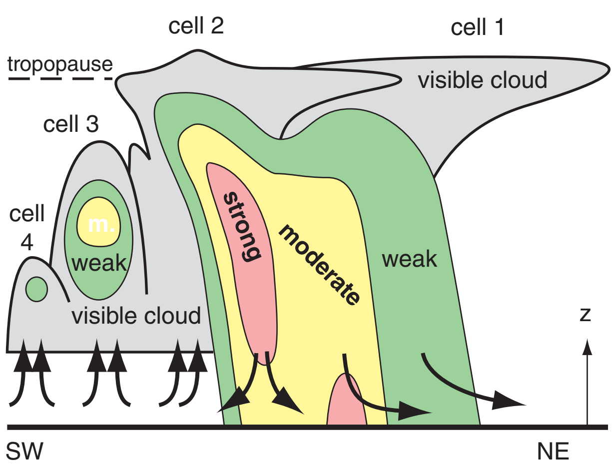


Figure 14.8
Multicell thunderstorm. Cell 1 is the oldest cell, and is in the dissipating stage. Cell 2 is mature. Cell 3 is in the cumulus stage. Cell 4 is the newest and smallest, in an early cumulus stage. Shaded regions indicate where we would see clouds and rain by eye. Green, yellow, and red shading indicate weak, moderate (m.), and strong radar echoes inside the clouds, which generally correspond to light, moderate, and heavy precipitation. Arrows show low-altitude winds.

14.1.4. Thunderstorm Types & Organization

Types of thunderstorm organization include:
- basic storms
  - weakly-forced (also called airmass storms)
  - multicell
  - orographic
- mesoscale convective systems (MCS)
  - squall line
  - bow echo
  - mesoscale convective complex (MCC)
  - mesoscale convective vortex (MCV)
- supercells
  - classic (CL)
  - low-precipitation (LP)
  - high-precipitation (HP)

14.1.4.1. Basic Storms

Weakly-forced (Airmass) Thunderstorms. (See Figs. 14.1, 14.2 & 14.7.) Satellite images of **airmass** thunderstorms show that several often exist at the same time. Sometimes these **weakly-forced storms** are scattered within the airmass analogous to mushrooms in a meadow. Each storm might be triggered by a thermal of rising warm air (heated at the Earth's surface) that happens to be buoyant enough to rise out of the top of the boundary layer. Other times these storms form at airmass boundaries. Each storm often has a Byers-Braham cell evolution very similar to Fig. 14.6. It can form in the middle of a warm humid airmass in late afternoon.

The lifetime of airmass thunderstorms are about the same as that of an individual cell (30 to 60 minutes), but they can spawn daughter storms as sketched in Fig. 14.6. Each can produce a short-duration (15 minutes) rain shower of moderate to heavy intensity that covers an area of roughly 5 - 10 km radius. Weakly-forced thunderstorms that produce short-duration severe weather (heavy precipitation, strong winds, lightning, etc.) during the mature stage are called **pulse storms**.

Multicell Thunderstorms. Most thunderstorms in North America are **multicell storms** (Fig. 14.8). Within a single cloud mass that by eye looks like a thunderstorm, a weather radar can detect many (2 to 5) separate cells (Fig. 14.9) by their precipitation cores and their winds. In a multicell storm, the different cells are often in different stages of their life cycle. Given weak to moderate wind shear in the environment, the downburst and gust front from one of the mature cells (cell 2 in Fig. 14.8) can trigger new adjacent cells (namely, cells 3 and 4).

In central North America where warm humid boundary-layer winds often blow from the southeast in the pre-storm environment, new cells often

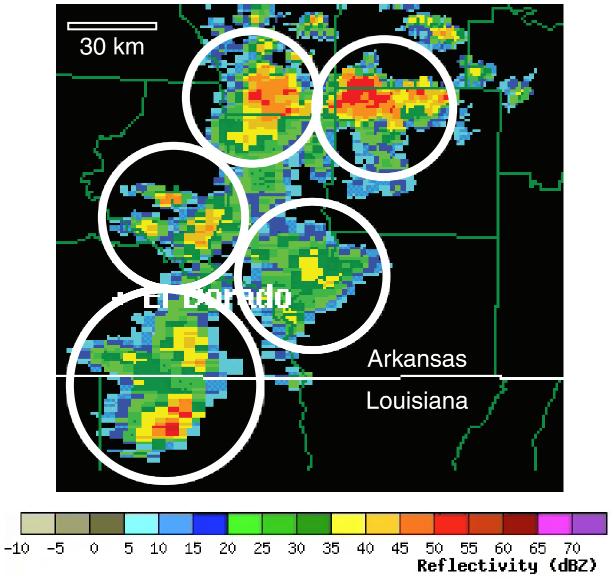


Figure 14.9.
Radar reflectivity PPI image of a multicell thunderstorm in south-central USA, 28 June 2001. Darkest red indicates > 50 dBZ; namely, the heaviest precipitation. Cells are circled. Courtesy of the US National Weather Service, NOAA.

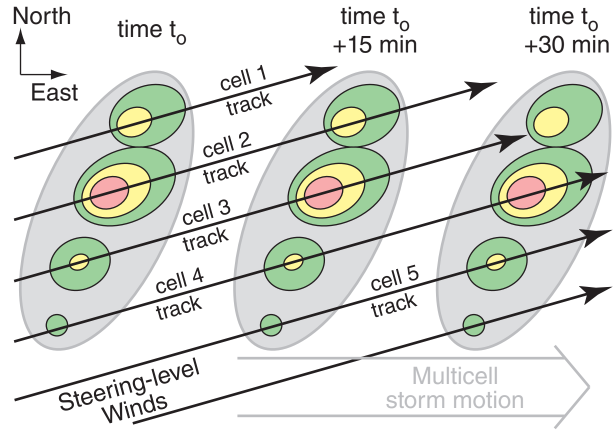


Figure 14.10
Diagram of multicell thunderstorm motion. Grey shading indicates visible clouds. Radar echo colors corresponds to Fig. 14.8. Black arrows show tracks of individual cells. Double-arrow in grey shows motion of the multicell cloud mass. Each light-grey large oval shows position of the cloud mass every 15 min.

form on the south flank of the storm (closest to the fuel supply), while the northeastward-moving mature cells decay as their fuel supply is diminished (Fig. 14.10). The resulting multicell thunderstorm appears to move to the right of the **steering-level winds** (normal or average winds in the bottom 6 km of the troposphere), even though individual cells within the storm move in the direction of the normal steering-level winds (Fig. 14.10).

<u>Orographic Thunderstorms</u>. Single or multicell storms triggered by mountains or hills are called **orographic thunderstorms**. With the proper environmental wind shear (Fig. 14.11), low-altitude upslope winds can continuously supply humid boundary-layer air. If the upper-level winds are from the opposite direction (e.g., from the west in Fig. 14.11), then these storms can remain stationary over a mountain watershed for many hours, funneling heavy rains into canyons, and causing devastating **flash floods**. One example is the Big Thompson Canyon flood in the Colorado Rockies, which killed 139 people in 1976 (see the Exposition in the Thunderstorm Hazards chapter).

Sometimes if winds are right, thunderstorms are triggered over the mountains, but then are blown away to other locations. For example, orographic thunderstorms triggered over Colorado and Wyoming in late afternoon can move eastward during the evening, to hit the USA Midwest (e.g., Wisconsin) between midnight and sunrise.

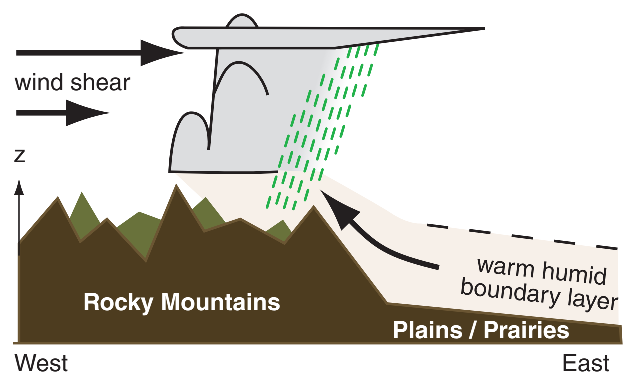


Figure 14.11
Orographic thunderstorm, stationary over the mountains.

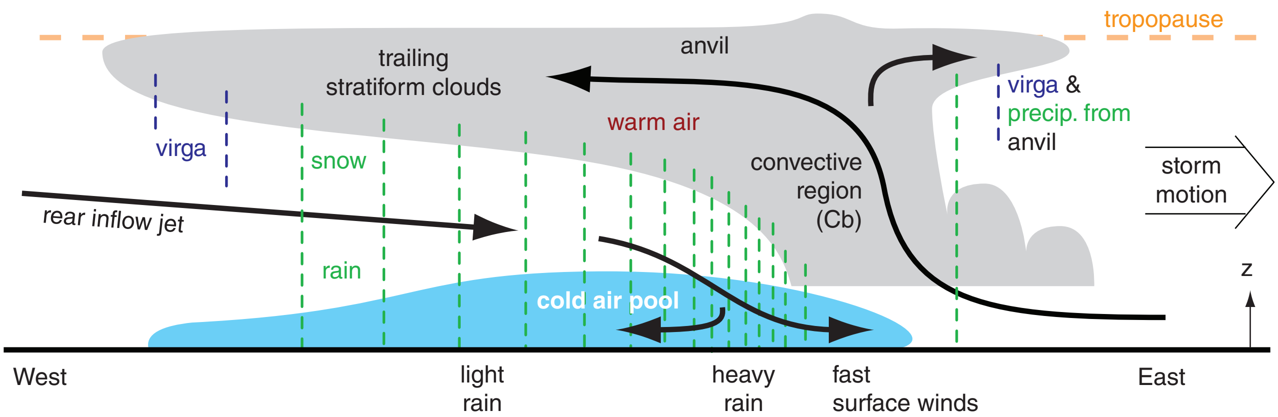


Figure 14.12
Vertical cross section of a mesoscale convective system (MCS). Grey shading indicates clouds. Blue shading indicates cloud-free cold air. Dark green dashed lines indicate precipitation.

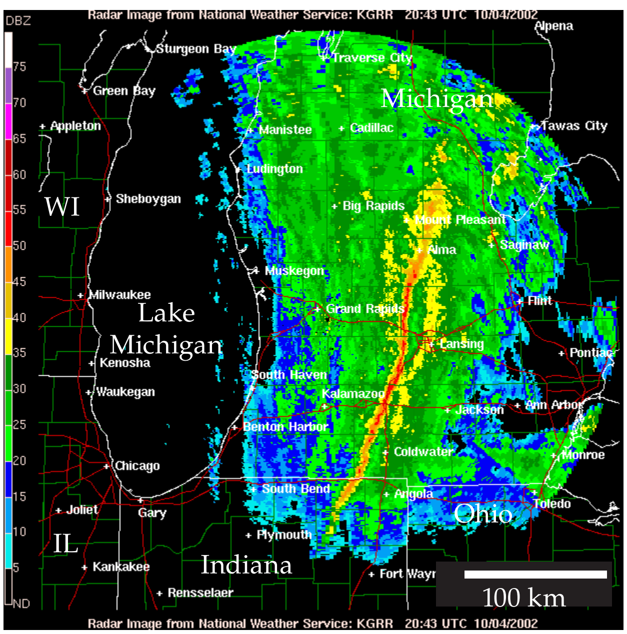


Figure 14.13.
Radar reflectivity PPI image of a mesoscale convective system (MCS) in southern Michigan, USA, on 4 Oct 2002. Red indicates a line of heavy thunderstorm rain associated with radar reflectivities > 50 dBZ. Ahead of and behind this squall line are broad regions of lighter, stratiform precipitation (green colors) with embedded convective cells (yellow). Courtesy of the US National Weather Service, NOAA.

14.1.4.2. Mesoscale Convective Systems

Another type of multicell-storm organization is a **mesoscale convective system** (**MCS**). This often has a narrow line of thunderstorms with heavy precipitation, followed by additional scattered thunderstorm cells (Figs. 14.12 & 14.13). The anvils from all these thunderstorms often merge into a single, large stratiform **cloud shield** that trails behind. This trailing **stratiform** region has widespread light to moderate precipitation that contributes about 1/3 of the total MCS rainfall.

These systems can be triggered by the terrain (many form just east of the Rocky Mountains), by synoptic systems (such as cold fronts), or by the gust fronts from a smaller number of thunderstorms. To survive, they need large convective instability and large wind shear over a deep layer of the environment. MCSs are responsible for 30 to 70% of precipitation over central N. America during the warm season. Some MCSs reach their peak intensity about midnight.

Ice crystals falling from the trailing stratiform clouds melt or partially sublimate, and the resulting rain then continues to partially evaporate on the way down. This produces a broad region of chilled air (due to latent heats of melting, sublimation, and evaporation) that descends to the ground and accumulates as a **cold-air pool** at the surface. The extra weight of the cold air produces a mesoscale high pressure (**meso-high**) at the surface in this region (Fig. 14.14).

Meanwhile, warm humid boundary-layer air ahead of the MCS is drawn into the storm by the combined effects of buoyancy and wind shear. This air rises within the embedded thunderstorm cells, and releases latent heat as the water vapor condenses. The resulting heating produces a layer of relatively warm air at mid and upper levels in the MCS (Fig. 14.14). The top of the anvil is cooled by IR radiation to space.

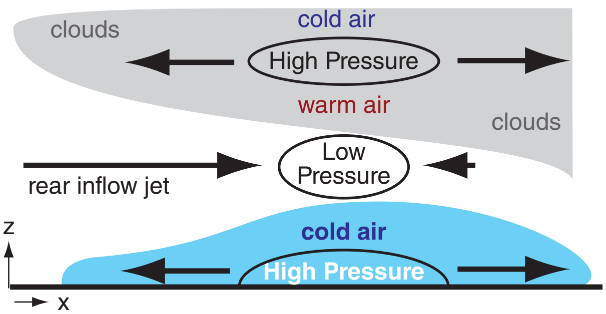


Figure 14.14
Zoomed view of the left half of Fig. 14.12, showing the rear portion of an MCS. Variation of pressure (relative to a standard atmosphere) with altitude caused by layers of warm and cold air in the stratiform portion of an MCS.

The resulting sandwich of relatively warmer and cooler layers causes a mesoscale low-pressure region (a **meso-low**) at mid altitudes (Fig. 14.14) and high pressure aloft (see Sample Application). The horizontal pressure-gradient force at mid altitudes causes air to converge horizontally. The thunderstorm updraft partially blocks mid-altitude inflow from the front of the storm, so the majority of inflow is from the rear. This is called a **rear inflow jet** (**RIJ**).

As the RIJ approaches the core of the MCS, it passes through the precipitation falling from above, and is cooled. This cooling allows the RIJ to descend as it blows toward the front of the storm, causing damaging **straight-line winds** at the ground. Large damaging straight-line wind events are called **derechos** (see a later INFO Box).

Sample Application

Plot MCS thermodynamics:

(a) Assume a standard atmosphere in the environment outside the MCS, and calculate $P_{std.atm.}$ vs. z.

(b) Inside the MCS, assume the temperature T_{MCS} differs from the $T_{std.atm.}$ by:
–12 K for $9 < z \le 11$ km due to rad. cooling at anvil top;
+8 K for $4 \le z \le 9$ km due to latent heating in cloud;
–12 K for $0 \le z \le 2$ km in the cold pool of air;
0 K elsewhere. Use P_{sfc} = 101.8 kPa in the MCS.
Plot $\Delta T = T_{MCS} - T_{std.atm.}$ vs. z .

(c) Calculate P_{MCS} vs. z in the MCS.

(d)Plot pressure difference $\Delta P = P_{MCS} - P_{std.atm.}$ vs. z

Find the Answer:
Given: T and P data listed above within MCS, and std. atm. in the environment outside the MCS.
Find & plot: ΔT (K) vs. z (km); and ΔP (kPa) vs. z (km)

Method:
(a) In a spreadsheet, use standard-atmosphere eq. (1.16) to find $T_{std.atm.}$ vs. z, assuming $H = z$.

(Continues in next column.)

Sample Application. *(Continuation)*

For this standard atmosphere outside the MCS, use hypsometric eq. (1.26b) to find P vs. z by iterating upward in steps of Δz = 250 m from $P_1 = P_{sfc}$ = 101.325 kPa at z_1 = 0. Assume $T_{v\,avg} = T_{avg} = 0.5 \cdot (T_1 + T_2)$ in eq. (1.26b) for simplicity (since no humidity information was given in the problem). Thus (with a = 29.3 m·K^{-1}):

$$P_2 = P_1 \cdot \exp[\,(z_1 - z_2)\ /\ (a \cdot 0.5 \cdot (T_1 + T_2)\,] \qquad (1.26b)$$

(b) Plot ΔT vs. z, from data given in the problem.

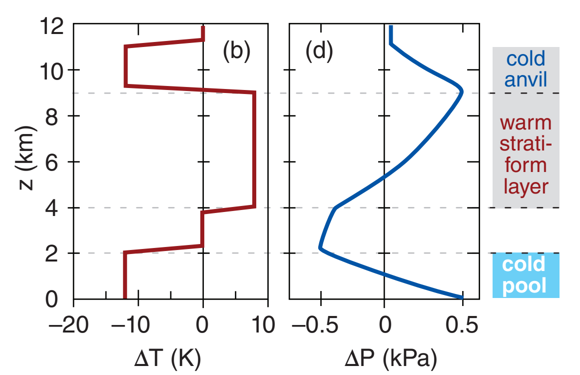
(c) Find $T_{MCS}(z) = T_{std.atm.}(z) + \Delta T(z)$. Do the similar hypsometric P calculations as (a), but inside the MCS.

(d) Find $\Delta P = P_{MCS} - P_{std.atm.}$ from (c) and (a), and plot.

Spreadsheet segments:

	(a)		(b)	(c)		(d)
	Std. Atmos.			MCS		ΔP
z (m)	T (K)	P (kPa)	ΔT (K)	T (K)	P (kPa)	(kPa)
0	288.2	101.3	–12	276.2	101.8	0.5
250	286.5	98.3	–12	274.5	98.7	0.4
500	284.9	95.4	–12	272.9	95.7	0.2
750	283.3	92.6	–12	271.3	92.7	0.1
. . .	. . .	. . .	. . .	. . .	. . .	. . .
2000	275.2	79.5	–12	263.2	79.0	–0.5
2250	273.5	77.1	0	273.5	76.6	–0.5
. . .	. . .	. . .	. . .	. . .	. . .	. . .
3750	263.8	63.7	0	263.8	63.3	–0.4
4000	262.2	61.7	8	270.2	61.3	–0.4
. . .	. . .	. . .	. . .	. . .	. . .	. . .
9000	229.7	30.8	8	237.7	31.3	0.5
9250	228.0	29.6	–12	214.0	30.1	0.5
. . .	. . .	. . .	. . .	. . .	. . .	. . .
11000	216.7	22.7	–12	204.7	22.7	0.0
11250	215.0	21.8	0	215.0	21.8	0.0

(Note: The actual spreadsheet carried more significant digits.)



Check: Units OK. Standard atmosphere P agrees with Table 1-5. Graphs are reasonable.

Exposition: High pressure at the surface causes the cold-pool air to spread out, as sketched in Figs. 14.12 & 14.14. The lowest pressure is above the cold pool and below the warm stratiform clouds. This low pressure (relative to the environment) sucks in air as a rear inflow jet. The deep layer of warm stratiform clouds eventually causes high pressure aloft, causing the cloudy air to spread out in the anvil. The cold anvil top then brings the pressure back to ambient, so that above the anvil there are no pressure differences.

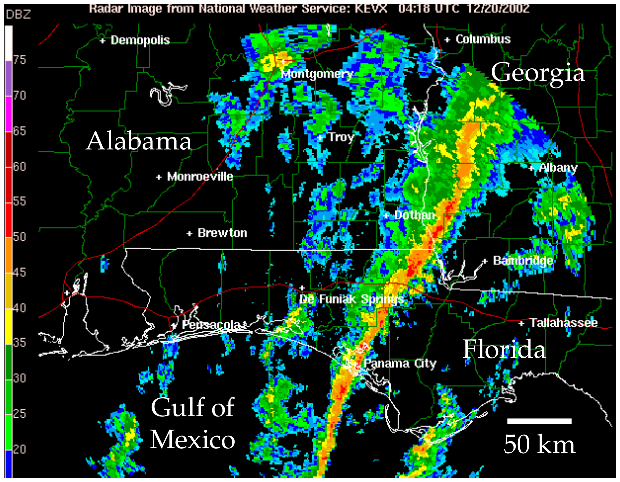


Figure 14.15
Radar reflectivity PPI image of squall line in Florida, Georgia and Alabama, USA, on 20 December 2002. Red corresponds to thunderstorms with reflectivity > 50 dBZ. Courtesy of the US National Weather Service, NOAA.

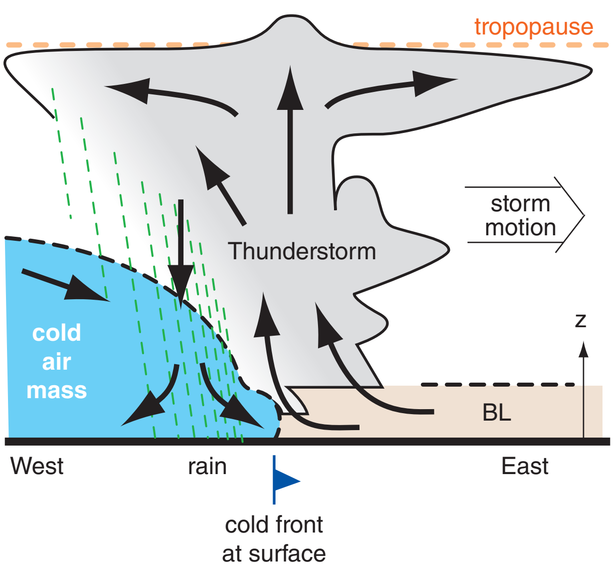


Figure 14.16
Vertical cross section through a squall-line thunderstorm, triggered by a cold front. Boundary-layer (BL, shaded tan) pre-storm air is warm and humid; cold air mass (shaded blue) is colder and drier. Thick dashed lines indicate base of stable layers such as temperature inversions. Thin dashed lines represent precipitation.

Squall Line. One type of MCS is a **squall line**, where a linear triggering mechanism such as a cold front, dry line or gust front causes many neighboring thunderstorms to merge into a long narrow line (Figs. 8.26 & 14.15). Squall lines can be many hundreds (even a thousand) kilometers long, but are usually only 15 to 400 km wide. They can persist for several hours to several days if continuously triggered by the advancing front. Heavy thunderstorm precipitation falls immediately behind the cold front at the surface, and sometimes has weaker stratiform precipitation behind it. Squall line storms can also have prodigious lightning, hail, damaging winds and tornadoes (although the tornadoes tend to be weaker than supercell-thunderstorm tornadoes).

Fig. 14.16 illustrates flow characteristics in a vertical cross section perpendicular to the squall-line axis. Environmental wind shear along the advancing cold front allows the thunderstorms to persist in the mature stage for the lifetime of the squall line. Cold, dry air descends from the west in the cold air mass from behind the squall line. This cold descent is enhanced by cool downburst air in the precipitation region. Both effects help advance the leading edge of the cold front at the ground. Warm, humid boundary-layer air blowing from the east ahead of the front is lifted by the undercutting denser cold advancing air, thereby continuously feeding the main updraft that moves with the advancing front.

A well-defined arc or shelf cloud often exists along the leading edge of the cold front, associated with violent straight-line winds. If the gust front from the downburst advances eastward faster than the cold front, then a second line of thunderstorms may be triggered along the gust front in advance of the cold front or the squall line might separate from the front.

Bow Echo. For MCSs with strong buoyancy in the updraft air and moderate to strong wind shear in the bottom 2 to 3 km of the environment, the cold pool becomes deeper and colder, and rear-inflow-jet (RIJ) wind speeds can become stronger. The RIJ pushes forward the center portion of the squall line, causing it to horizontally bend into a shape called a **bow echo** (Fig. 14.17). Bow-echo lines are 20 to 200 km long (i.e., smaller than most other MCSs) and are often more severe than normal MCSs.

Vorticity caused by wind shear along the advancing cold pool causes counter-rotating vortices at each end of the bow-echo line. These are called **line-end vortices** or **bookend vortices** (Fig 14.17), which rotate like an eggbeater to accelerate the RIJ and focus it into a narrower stream. For bow echoes lasting longer than about 3 h, Coriolis force enhances the northern cyclonic vortex in the

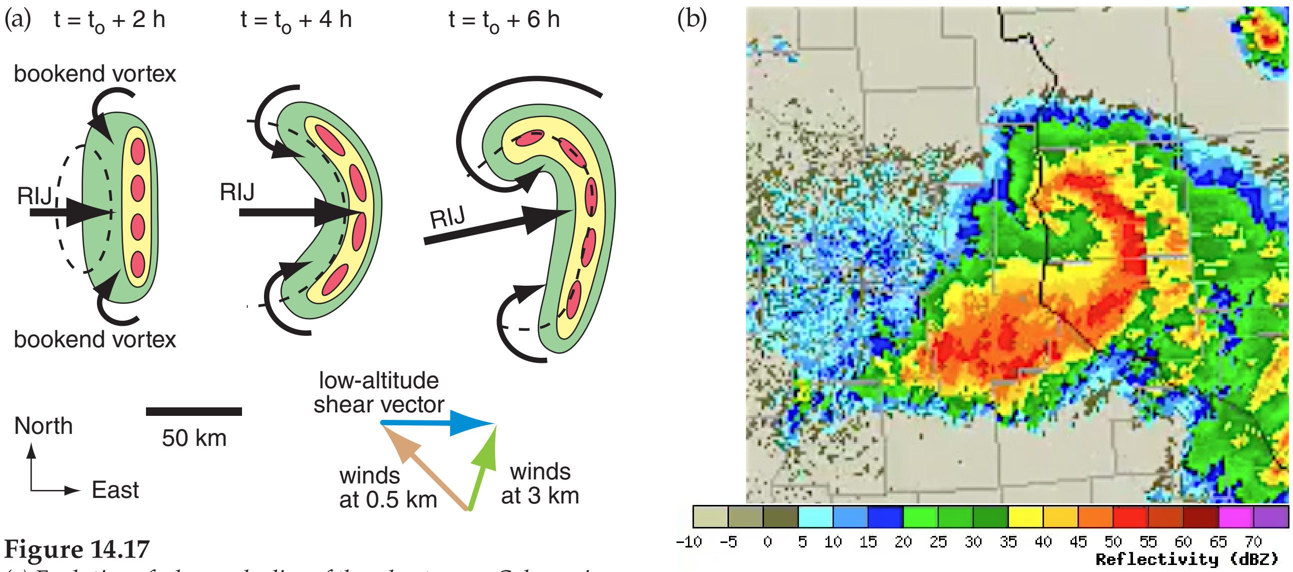


Figure 14.17
(a) Evolution of a bow-echo line of thunderstorms. Colors mimic radar reflectivity, dashed line outlines cold air pool at surface, RIJ is rear inflow jet, t is time, and t_o is formation time. (b) Radar image of a bow echo approaching Wisconsin, 14 Aug 2007.

Northern Hemisphere and diminishes the southern anticyclonic vortex, causing the bow echo to become asymmetric later in its evolution.

Strong straight-line winds near the ground from a bow echo can damage buildings and trees similar to an EF2 tornado (see the "Tornado" section in the next chapter for a description of tornado damage scales). If a sequence of bow echoes causes wind damage along a 400 km path or greater and lasts about 3 hours or more, with wind speeds ≥ 26 m s^{-1} and gusts ≥ 33 m s^{-1}, then the damaging wind event is called a **derecho** (see INFO Box 3 pages later).

<u>Mesoscale Convective Complex (MCC)</u>. If the size of the MCS cloud shield of merged anvils is large (diameter ≥ 350 km), has an overall elliptical or circular shape, and lasts 6 to 12 h, then it is called a **mesoscale convective complex** (**MCC**). An additional requirement to be classified as an MCC is that the top of the cloud shield, as seen in IR-satellite images, must have a cold brightness temperature of ≤ –33°C. Within this cold cloud shield must be a smaller region (diameter ≥ 250 km) of higher, colder cloud top with brightness temperature ≤ –53°C.

MCCs can produce heavy rain over large areas, with the most intense rain usually happening first. MCCs are most often observed at night over a statically stable boundary layer, which means that the fuel for these storms often comes from the warm humid air stored in the residual boundary layer from the day before. They can be triggered by weak warm-frontal zones and weak mid-tropospheric short waves, and are often associated with low-level jets of wind.

Sample Application

Estimate the speed of the rear inflow jet (RIJ) wind that is drawn into the MCS mid-tropospheric low-pressure region, for the previous Sample Application. The MCS has horizontal radius of 175 km, and is located in Oklahoma (latitude 35°N) .

Find the Answer

Given: ΔP = –0.5 kPa at height z = 2 km, from previous Sample Application.
R = 175 km = radius of meso-low in mid-troposphere, latitude ϕ = 35°N
Find: M_{RIJ} = ? m s^{-1}

Assume the RIJ is geostrophic ($M_{RIJ} = U_g$). To use eq. (10.26a) for the geostrophic wind, we need:
$\Delta P/\Delta y$, ρ and f_c .
Consider the region just south of the low center.
$\Delta P/\Delta y$ = (–0.5 kPa)/(175 km) = –0.00286 kPa·km^{-1}
Assuming std. atmosphere. Use Table 1-5 at 2 km
ρ = 1.0 kg·m^{-3} .
Use eq. 10.16 to find the Coriolis parameter:
f_c = (1.458x10^{-4} s^{-1})·sin(35°) = 0.836x10^{-4} s^{-1}.
Combine these data in eq. (10.26a):
U_g= –[(1.0 kg·m^{-3})·(0.836x10^{-4} s^{-1})]$^{-1}$·(–0.00286 Pa·m^{-1})
= 34.2 m s^{-1}
Thus,
$M_{RIJ} = U_g$ = <u>**34.2**</u> m s^{-1}

Check: Units OK. Magnitude reasonable. Sign OK.
Exposition: This implies a RIJ from the west, in the region just south of the mid-tropospheric meso-low center. If this RIJ descends down to the surface due to the influence of precipitation cooling, it could cause damaging straight-line wind gusts that exceed the threshold to be classified as a derecho.

One caution: the geostrophic wind assumes steady state. In a rapidly intensifying MCS, steady-state might not be the best assumption. If the wind were accelerating, then it could have an ageostrophic component pointing towards the low center.

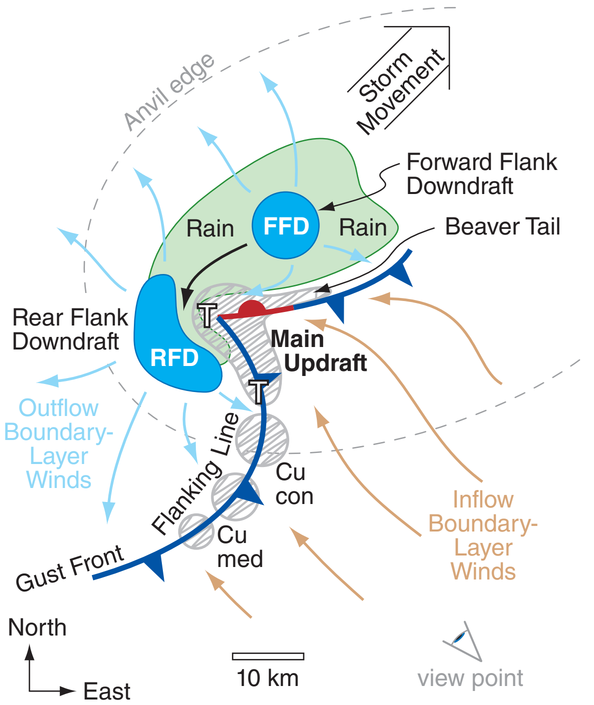


Figure 14.18
Plan view (top down) sketch of a classic supercell in the N. Hemisphere (copied from Fig. 14.4b). T indicates possible tornado positions; RFD = Rear Flank Downdraft; FFD = Forward Flank Downdraft. Regions of updraft are hatched with grey; downdrafts are solid blue; rain is solid light green. Low surface pressure is found under the updrafts (especially near the "T" locations), and high pressure under the downdrafts. Vectors are surface winds: cold are cyan, warm are tan colored.

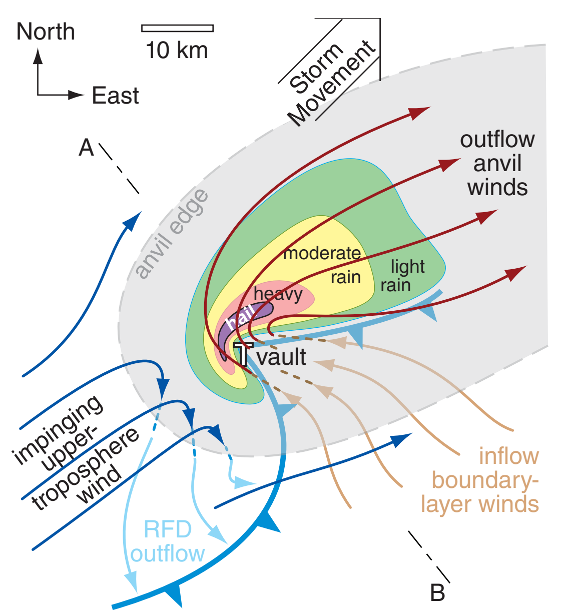


Figure 14.19
Plan view of classic (CL) supercell in the N. Hemisphere. Low altitude winds are shown with light-colored arrows, high-altitude winds with darker arrows, and ascending/descending air with short-dashed lines. T indicates tornado location.

<u>Mesoscale Convective Vortex (MCV)</u>. Between midnight and sunrise, as the static stability of the boundary layer increases, the thunderstorms within MCSs often weaken and die. Thunderstorms are more difficult to trigger then, because cold nocturnal boundary-layer air resists rising. Also, above the nocturnal layer is a thin residual layer that has only a small capacity to hold the fuel supply of warm humid air from the previous day's mixed layer. As the MCS rains itself out, all that is left by the next morning are mid-level stratiform clouds.

However, the mid-tropospheric mesoscale low-pressure region (Fig. 14.14) in the MCS has existed for about half a day by this time, and has developed cyclonic rotation around it due to Coriolis force. This mesoscale rotation is often visible (by satellite) in the mid-level (former MCS) stratiform clouds, and is called a **mesoscale convective vortex** (**MCV**). The stable boundary layer beneath the MCV reduces surface drag, allowing the rotation to persist for many more hours around the weak low.

Although these weakly rotating systems are not violent and often don't have a circulation at the surface, they can be tracked by satellite during the morning and early afternoon as they drift eastward. They are significant because they can enhance new thunderstorm formation via weak convergence and upward motion. Thus, by evening the dying MCV can change into a new MCS.

14.1.4.3. Supercell Thunderstorms

A violent thunderstorm having a rotating updraft and persisting for hours is called a **supercell storm** (Figs. 14.3 & 14.4). Its downdraft and surface features (Fig 14.18, a duplicate of Fig. 14.4b) are organized in a way to trigger new thunderstorms nearby along a **flanking line**, and then to entrain these incipient updrafts (cumulus congestus clouds) into the main supercell updraft to give it renewed strength. The main, persistent, rotating updraft, or **mesocyclone**, is 3 to 10 km in diameter, and can extend throughout the whole depth of the thunderstorm.

Supercell storms are responsible for the most violent tornadoes (EF3 - EF5, as defined in the next chapter) and hail, and can have damaging straight-line winds and intense rain showers causing flash floods. To support supercells, the pre-storm environment must have a deep layer of convectively unstable air, and strong wind shear in the bottom 6 km of the troposphere.

Although supercells can have a wide range of characteristics, they are classified into three categories: **low-precipitation** (**LP**) storms, medium-precipitation or **classic** (**CL**) supercells, and **high-precipitation** (**HP**) storms. Storms that fall between these categories are called **hybrid storms**.

Classic Supercells. Figures 14.3 and 14.4 show characteristics of a classic supercell. One way to understand how supercells work is to follow air parcels and hydrometeors (precipitation particles) as they flow through the storm.

Warm humid boundary-layer air flowing in from the southeast (Fig. 14.19) rises in the main supercell updraft, and tiny cloud droplets form once the air is above its LCL. So much water vapor condenses that large amounts of latent heat are released, causing large buoyancy and strong updrafts (often ≥ 50 m s^{-1}). Cloud droplets are carried upward so quickly in the updraft that there is not yet time for larger precipitation particles to form. Hence, the updraft, while visible by eye as a solid cloud, appears only as a **weak echo region** (**WER**) on radar PPI scans because there are no large hydrometeors. This region is also known as an **echo-free vault** (Fig. 14.19).

As the air reaches the top of the troposphere, it encounters stronger winds, which blows these air parcels and their cloud particles downwind (to the northeast in Fig. 14.19). Since the whole storm generally moves with the mean wind averaged over the bottom 6 km of troposphere (to the northeast in Fig. 14.19), this means that the air parcels in the top part of the storm are flowing to the storm's forward flank. This differs from MCSs, where the updraft air tilts back toward the storm's trailing (rear) flank.

While rising, the air-parcel temperature drops to below freezing, and eventually below –40°C near the top of the thunderstorm (similar to Fig. 14.7), at which point all hydrometeors are frozen. Various microphysical processes (cold-cloud Bergeron process, collision, aggregation, etc.) cause larger-size hydrometeors to form and precipitate out. The larger, heavier hydrometeors (hail, graupel) fall out soonest, adjacent to the updraft (to the north in Fig. 14.19). Medium size hydrometeors are blown by the upper-level winds farther downwind (northeast), and fall out next. Lighter hydrometeors fall more slowly, and are blown farther downwind and fall in a larger area.

In summer, all of these ice particles (except for some of the hail) melt while falling, and reach the ground as rain. The greatest precipitation rate is close to the updraft, with precipitation rates diminishing farther away. The very smallest ice particles have such a slow fall velocity that they have a long residence time in the top of the troposphere, and are blown downwind to become the anvil cloud.

Meanwhile, the falling precipitation drags air with it and causes it to cool by evaporation (as will be explained in more detail in the next chapter). This causes a downdraft of air called the **forward flank downdraft** (**FFD**, see Fig. 14.18). When this cold air hits the ground, it spreads out. The leading edge of

Sample Application

If a Doppler radar were located near the "view point" in the bottom right corner of Fig. 14.18, then sketch the resulting color patterns on the Doppler PPI wind display for near-surface winds.

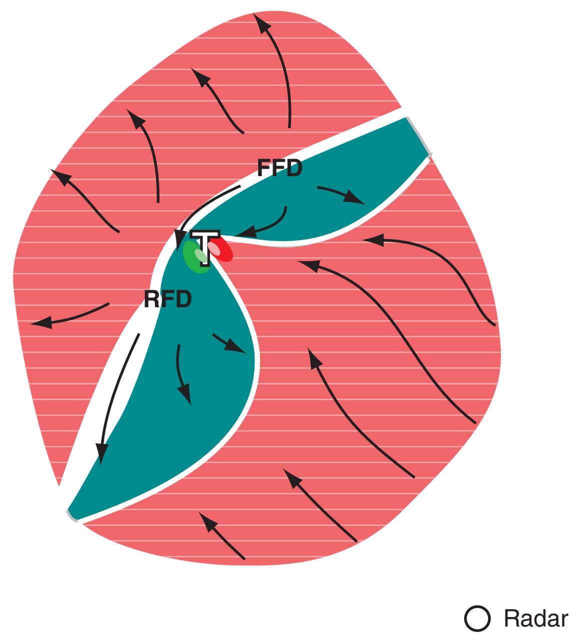
Find the Answer:

Given: Near-surface winds from Fig. 14.18 are copied below, with radar located southeast of the storm center.

Find: Sketch the Doppler wind display.

Assume sufficient number-density of insects to allow Doppler radar returns even in rain-free areas.

Hatched red corresponds to winds away from the radar. Solid turquoise corresponds to winds toward the radar. White is the zero isodop line. Brighter shading of the reds and greens near the "T" is to suggest a tornado vortex signature.



Check: Compares well with info in Chapter 8.

Exposition: Fig. 14.19 already shows the radar reflectivity display. By interpreting the reflectivity and Doppler info together, you can create in your mind a coherent picture of the storm.

The gust fronts show up clearly in the Doppler display. The tornado-vortex signature in the Doppler display and the hook echo in the reflectivity display both suggest the presence of a tornado.

Finally, real thunderstorms are embedded in an environment that is not usually calm, so there would also be red and blue colors outside of the regions sketched here.

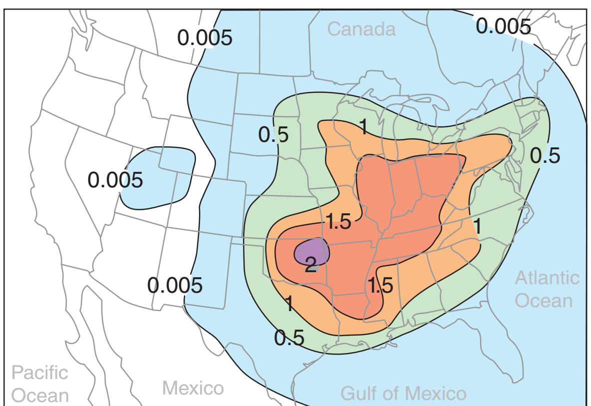
INFO • Derecho

A **derecho** is a hazardous event of very strong straight-line (non-tornadic) horizontal winds (≥ 26 m s^{-1}) often causing widespread (greater than 100 km wide and 650 km length) damage at the surface. It is associated with clusters of downbursts of air from a single moving mesoscale convective system (MCS), which causes localized destruction that sweeps through the event area during several hours to over a day.

Within the event area, the wind intensity and damage can be variable. Peak wind speeds over 65 m s^{-1} have been rarely observed. Derechos are associated with rear-inflow jets and bow echoes that propagate in the forward (downshear) direction.

The word "derecho" is based on a Spanish word for "straight-ahead" or "direct". In North America, derechos occur in the prairies east of the Rocky Mountains (see Fig. below), and were first reported over a century ago. Although derechos can occur during any month, they are most frequent in April through August, with peak frequencies during May and July.

In the USA, an average of 21 derechos occur each year, killing 9 and injuring 145 people annually. Most of the deaths are people in cars and boats, while most of the injuries are people in mobile homes and cars. Derechos can blow down trees, destroy mobile homes, and damage other structures and cars. Damage from derechos is sometimes mistakenly reported as tornado damage.



Climatology of Derechos (1986-2003). Avg. # events/yr.

Figure 14.a.
Derecho climatology for North America.

Based on data from Ashley & Mote, 2005: Derecho hazards in the United States. Bull. Amer. Meteor. Soc., ***86****, 1577-1592;, Corfidi, Coniglio, Cohen & Mead, 2016: A proposed revision to the definition of "Derecho". Bull. Amer. Meteor. Soc.,* ***97****, 935-949; and http://www.spc.noaa.gov/misc/AbtDerechos/derechofacts.htm .*

the spreading cold air is called the (forward-flank) **gust front**, and is often indicated with cold-frontal symbols. This gust front helps to deflect the boundary-layer air upward into the main updraft, and can create a **beaver-tail cloud**, as described earlier.

Due to rotation of the whole updraft (the origin of rotation will be discussed later), the precipitation region is often swept counterclockwise around the updraft for N. Hemisphere supercells. This creates a characteristic curve of the precipitation region that is seen on weather radar as a **hook echo** around the echo-free vault (Fig. 14.19).

In the S. Hemisphere (and for a small portion of N. Hemisphere supercells — see Fig. 14.63), the updrafts rotate clockwise. These supercells can still produce hook echoes, but with the hook projecting in a direction different than that sketched in Fig. 14.19).

The supercell storm is so tightly organized that it acts as an obstacle to fast winds in the upper troposphere. When these ambient winds from the west or southwest hit the supercell, some are deflected downward, creating the **rear-flank downdraft** (**RFD**). As the dry upper-tropospheric air moves down adjacent to the cloudy updraft, it entrains some cloud droplets, which quickly evaporate and cool the air. This negative buoyancy enhances the RFD such that when the air hits the ground it spreads out, the leading edge of which is a (rear-flank) **gust front**.

When warm, humid boundary-layer air flowing into the storm hits the rear-flank gust front, it is forced upward and triggers a line of cumulus-cloud growth called the **flanking line** of clouds. The inflow and circulation around the supercell draws these new cumulus congestus clouds into the main updraft, as previously described. At the cusp of the two gust fronts, there is low surface pressure under the updraft. This **meso-low** (see Table 10-6) and the associated gust fronts look like mesoscale versions of synoptic-scale cyclones, and are the surface manifestation of the **mesocyclone**.

Strong **tornadoes** are most likely to form in one of two locations in a supercell. One is under the largest cumulus congestus clouds just as they are being entrained into the main updraft, due to the rotation between the RFD and the updraft. These form under the rain-free cloud-base portion of the supercell, and are often easy to view and chase. The other location is under the main updraft at the cusp where the two gust fronts meet, due to rotation from the parent mesocyclone. Tornadoes are discussed in more detail in the next chapter.

Low Precipitation Supercells (Fig. 14.20). In drier regions of North America just east of the Rocky Mountains, LCLs are much higher above ground. However, mountains or dry lines can cause sufficient lift to trigger high-base supercell storms. Also, if strong environmental wind shear causes a strongly tilted updraft, then incipient precipitation particles are blown downwind away from the rich supply of smaller supercooled cloud droplets in the updraft, and thus can't grow large by collection/accretion. Most of the hydrometeors that form in these storms evaporate before reaching the ground; hence the name **low-precipitation** (**LP**) **supercells**.

With fewer surrounding low clouds to block the view, LP storms often look spectacular. You can often see the updraft, and can see evidence of its rotation due to cloud striations, curved inflow bands and possibly a rotating wall cloud. If present, the beaver tail is easily visible. Although rainfall is only light to moderate, LP storms can produce great amounts of large-diameter hail. Those LP supercells having exceptionally strong updrafts produce anvils that look like a mushroom cloud from a nuclear bomb explosion.

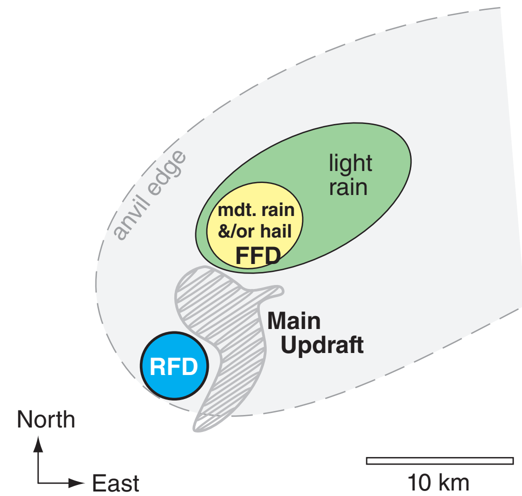


Figure 14.20
Low-precipitation (LP) supercell thunderstorm in the N. Hemisphere. FFD = forward flank downdraft; RFD = rear flank downdraft. Often little or no precipitation reaches the ground. The hatched grey region indicates updrafts.

High Precipitation Supercells (Fig. 14.21). In very warm humid locations such as southeastern North America, some supercell storms can produce widespread heavy precipitation. These are known as **high-precipitation** (**HP**) **supercells**. Precipitation falls from most of the cloud-base areas, often filling both the FFD and the RFD. So much rain falls upstream in the rear-flank region of the storm that the main mesocyclone is sometimes found downstream (forward) of the main precipitation region.

Some storms have curved inflow bands and cloud striations on the updraft that you can see from a distance. Also, the smooth-looking beaver-tail clouds are most common for HP storms, near the forward-flank gust front. Much more of the surrounding sky is filled with low and mid-level clouds, which together with low visibilities makes it difficult to get good views of this storm.

As this type of storm matures, the heavy-rain region of the hook echo can completely wrap around the updraft vault region. Storm chasers call this region the **bear's cage**, because of the high danger associated with tornadoes and lightning, while being surrounded by walls of torrential rain and hail with poor visibility and hazardous driving/chasing conditions.

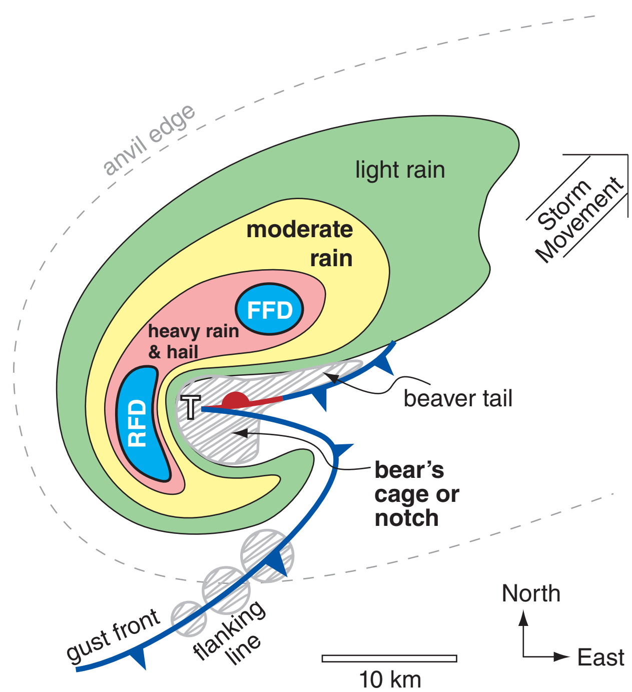


Figure 14.21
High-precipitation (HP) supercell thunderstorm in the N. Hemisphere. FFD = forward flank downdraft; RFD = rear flank downdraft. Hatched grey regions indicate updrafts.

Now that we are finished covering the descriptive appearance and types of thunderstorms, the next five sections examine the meteorological conditions needed to form a thunderstorm.

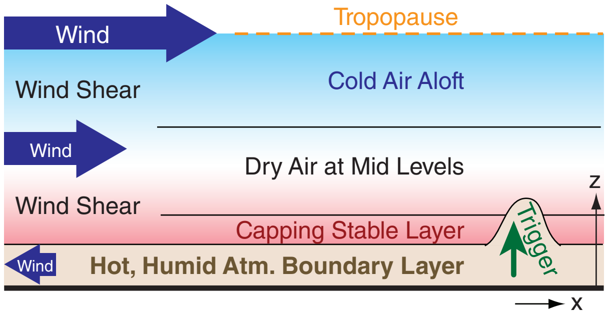


Figure 14.22
Atmospheric conditions favorable for formation of strong thunderstorms.

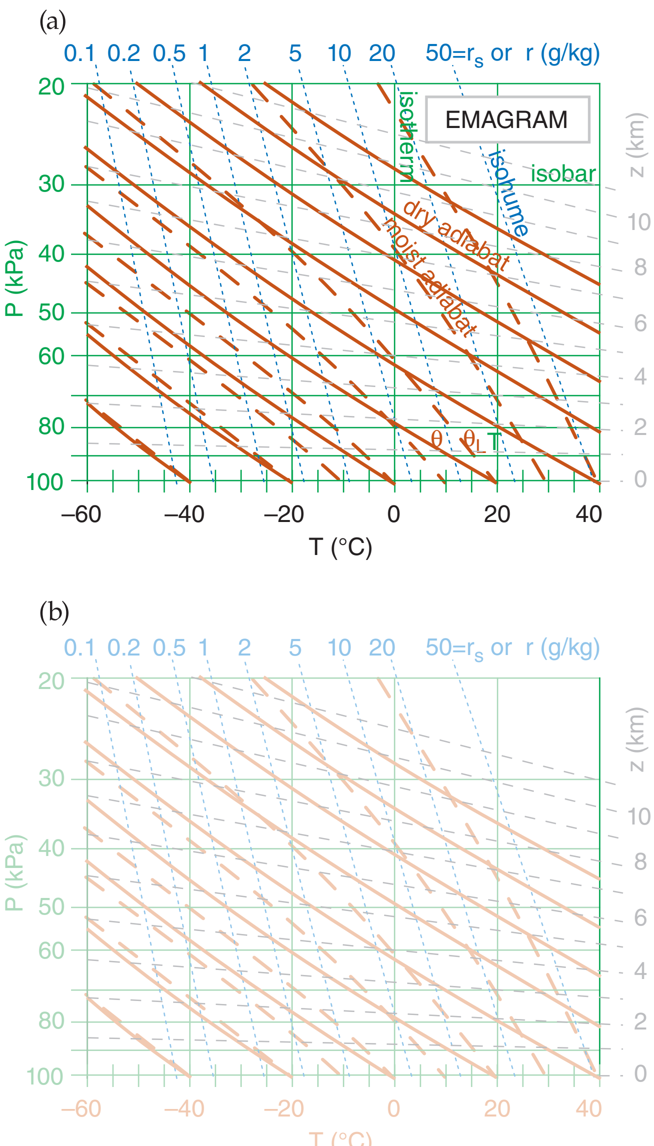


Figure 14.23
(a) Thermo diagram, with approximate height contours (z) added as nearly horizontal thin dashed lines (labeled at right). Sounding data plotted on this diagram will enable you to anticipate storm formation and strength. (b) Same as (a), but lighter colors to make it easier to write on.

14.2. THUNDERSTORM FORMATION

14.2.1. Favorable Conditions

Four environmental (pre-storm) conditions are needed to form the strong **moist convection** that is a severe thunderstorm:

1) **high humidity** in the boundary layer,
2) nonlocal conditional **instability**,
3) strong **wind shear**,
4) a **trigger mechanism** to cause **lifting** of atmospheric boundary-layer air.

Fig. 14.22 illustrates these conditions. Meteorologists look for these conditions in the pre-storm environment to help forecast thunderstorms. We will examine each of these conditions in separate sections. But first, we define key convective altitudes that we will use in the subsequent sections to better understand thunderstorm behavior.

14.2.2. Key Altitudes

The existence and strength of thunderstorms depends partially on layering and stability in the pre-storm environment. Thus, you must first obtain an **atmospheric sounding** from a **rawinsonde balloon** launch, **numerical-model** forecast, aircraft, dropsonde, satellite, or other source. Morning or early afternoon are good sounding times, well before the thunderstorm forms (i.e., **pre-storm**). The environmental sounding data usually includes temperature (T), dew-point temperature (T_d), and wind speed (M) and direction (α) at various heights (z) or pressures (P).

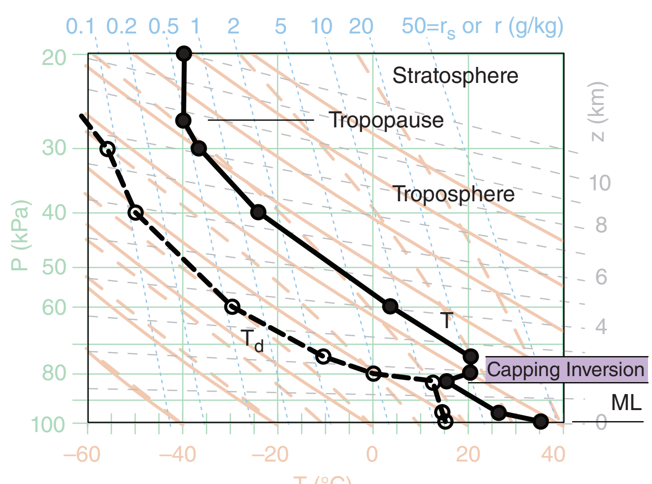


Figure 14.24
Environmental sounding of temperature (T) and dew-point temperature (T_d) plotted on a thermo diagram. The mixed layer (ML), capping inversion (Cap), tropopause, troposphere, and stratosphere are identified. This example shows an early-afternoon pre-storm environment.

Plot the environmental sounding on a thermo diagram such as the Emagram of Fig. 14.23, being sure to connect the data points with straight lines. Fig. 14.24 shows a typical pre-storm (early afternoon) sounding, where T is plotted as black-filled circles connected with a thick black solid line, and T_d is plotted as white-filled circles connected with a thick black dashed line. The mixed layer (ML; i.e., the daytime turbulently well-mixed boundary layer), capping inversion, and tropopause are identified in Fig. 14.24 using the methods in the Atmospheric Stability chapter.

Daytime solar heating warms the Earth's surface. The hot ground heats the air and evaporates soil moisture into the air. The warm humid air parcels rise as thermals in an unstable atmospheric boundary layer. If we assume each rising air parcel does not mix with the environment, then its temperature decreases dry adiabatically with height initially, and its mixing ratio is constant.

Fig. 14.25 shows how this process looks on a thermo diagram. The parcel temperature starts near the ground from the solid black circle and initially follows a **dry adiabat** (thin black solid line with arrow). The dew point starts near the ground from the open circle and follows the **isohume** (thin black dotted line with arrow).

Typically, the rising parcel hits the statically stable layer at the top of the mixed layer, and stops rising without making a thunderstorm. This capping temperature-inversion height z_i represents the average **mixed-layer (ML) top**, and is found on a thermo diagram where the dry adiabat of the rising parcel first crosses the environmental sounding (Fig. 14.25). In pressure coordinates, use symbol P_i to represent the ML top.

If the air parcel were to rise a short distance above z_i, it would find itself cooler than the surrounding environment, and its negative buoyancy would cause it sink back down into the mixed layer. On many such days, no thunderstorms ever form, because of this strong lid on top of the mixed layer.

But suppose an external process (called a trigger) pushes the boundary-layer air up through the capping inversion (i.e., above z_i) in spite of the negative buoyancy. The rising air parcel continues cooling until it becomes saturated. On a thermo diagram, this saturation point is the **lifting condensation level** (**LCL**), where the dry adiabat first crosses the isohume for the rising parcel.

As the trigger mechanism continues to push the reluctant air parcel up past its LCL, water vapor in the parcel condenses as clouds, converting latent heat into sensible heat. The rising cloudy air parcel thus doesn't cool as fast with height, and follows a **moist adiabat** on a thermo diagram (Fig. 14.25).

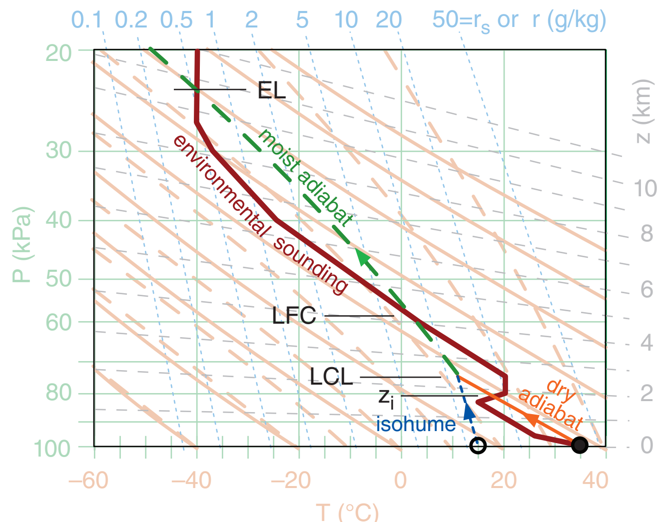


Figure 14.25
Afternoon pre-storm environmental air from near the surface (indicated with the circles), is hypothetically lifted as an air parcel to the top of the thermo diagram. Identified are the average mixed-layer height (z_i), the lifting condensation level (LCL), the level of free convection (LFC), and the equilibrium level (EL).

INFO • Cap vs. Capping Inversion

The strongly stable layer at the top of the daytime boundary layer (i.e., at the top of the mixed layer) is called the "**capping inversion,**" as sketched in the figure below (a modification of Fig. 14.25).

The region near the top of the mixed layer where the rising air parcel is colder than the environment is called the "**cap**." It is this region between z_i and the LFC that opposes or inhibits the rise of the air parcel, and which must be overcome by the external forcing mechanism. See the section later in this chapter on "Triggering vs. Convective Inhibition" for more details.

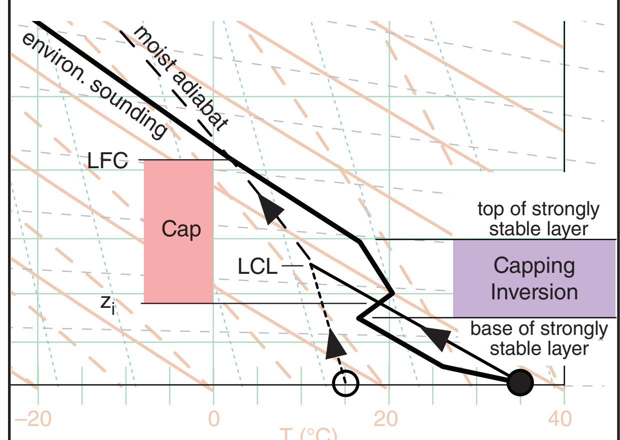


Figure 14.b
Idealized pre-storm environmental sounding (thick solid line), plotted on the bottom right portion of an Emagram. Circles and lines with arrows show the rise of an air parcel from near the surface.

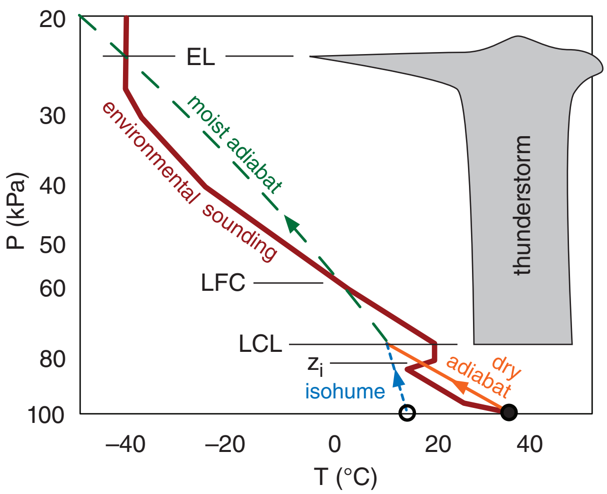


Figure 14.26
From Fig. 14.25, we infer that a thunderstorm could form with cloud top and base as sketched, if successfully triggered. Thunderstorm air parcels follow the thin diagonal solid and dashed lines, while the surrounding air is shown with the thick line.

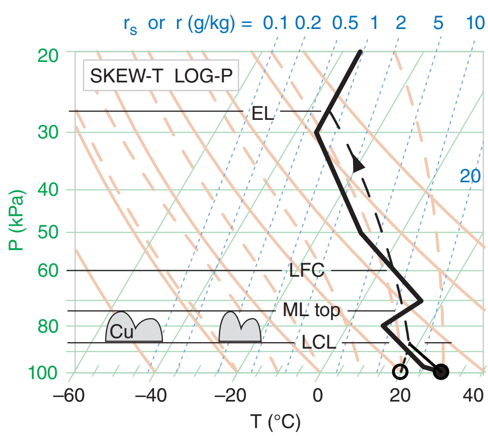
Sample Application

Plot this sounding on a full-size **skew-T** diagram (Atm. Stab. chapter), and estimate the pressure altitudes of the ML top , LCL, LFC, and EL for an air parcel rising from near the surface.

P (kPa)	T (°C)	T_d (°C)
100	30	20.
96	25	.
80	10	.
70	15	.
50	–10	.
30	–35	.
20	–35	.

Find the Answer

Given: Data in the table above.
Find: P_i = ? kPa (= ML top), P_{LCL} = ? kPa,
P_{LFC} = ? kPa, P_{EL} = ? kPa.



After plotting the air-parcel rise, we find:
P_i = **74 kPa**, P_{LCL} = **87 kPa**,
P_{LFC} = **60 kPa**, P_{EL} = **24 kPa**.

Exposition: The LCL for this case is below the ML top. Thus, the ML contains scattered fair-weather cumulus clouds (cumulus humilis, Cu). If there is no external trigger, the capping inversion prevents these clouds from growing into thunderstorms.

Even so, this cloudy parcel is still colder than the ambient environment at its own height, and still resists rising due to its negative buoyancy.

For an atmospheric environment that favors thunderstorm formation (Fig. 14.25), the cloudy air parcel can become warmer than the surrounding environment if pushed high enough by the trigger process. The name of this height is the **level of free convection** (**LFC**). On a thermo diagram, this height is where the moist adiabat of the rising air parcel crosses back to the warm side of the environmental sounding.

Above the LFC, the air parcel is positively buoyant, causing it to rise and accelerate. The positive buoyancy gives the thunderstorm its energy, and if large enough can cause violent updrafts.

Because the cloudy air parcel rises following a moist adiabat (Fig. 14.25), it eventually reaches an altitude where it is colder than the surrounding air. This altitude where upward buoyancy force becomes zero is called the **equilibrium level** (**EL**) or the **limit of convection** (**LOC**). The EL is frequently near (or just above) the tropopause (Fig. 14.24), because the strong static stability of the stratosphere impedes further air-parcel rise.

Thus, thunderstorm cloud-base is at the LCL, and thunderstorm anvil top is at the EL (Fig. 14.26). In very strong thunderstorms, the rising air parcels are so fast that the inertia of the air in this updraft causes an **overshooting dome** above the EL before sinking back to the EL. The most severe thunderstorms have tops that can penetrate up to 5 km above the tropopause, due to a combination of the EL being above the tropopause (as in Fig. 14.26) and inertial overshoot above the EL.

For many thunderstorms, the environmental air between the EL and the LFC is **conditionally unstable**. Namely, the environmental air is unstable if it is cloudy, but stable if not (see the Atmospheric Stability chapter). On a thermo diagram, this is revealed by an environmental lapse rate between the dry- and moist-adiabatic lapse rates.

However, it is not the environmental air between the LFC and EL that becomes saturated and forms the thunderstorm. Instead, it is air rising above the LFC from lower altitudes (from the atmospheric boundary layer) that forms the thunderstorm. Hence, this process is a **nonlocal conditional instability** (**NCI**).

Frequently the LCL is below z_i. This allows fair-weather cumulus clouds (cumulus humilis) to form in the top of the atmospheric boundary layer (see the Sample Application at left). But a trigger mechanism is still needed to force this cloud-topped mixed-layer air up to the LFC to initiate a thunderstorm.

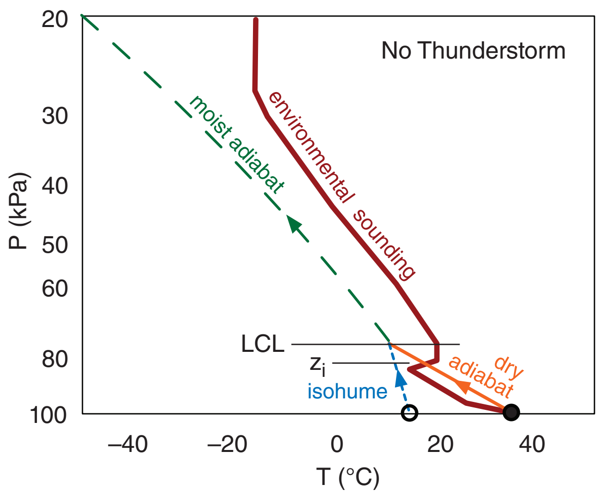


Figure 14.27
An environmental sounding that does not favor thunderstorms, because the environmental air aloft is too warm. Emagram.

In many situations, no LFC (and also no EL) exists for an air parcel that is made to rise from the surface. Namely, the saturated air parcel never becomes warmer than the environmental sounding (such as Fig. 14.27). Such soundings are NOT conducive to thunderstorms.

14.3. HIGH HUMIDITY IN THE ABL

One of the four conditions needed to form **convective storms** such as thunderstorms is high humidity in the atmospheric boundary layer (ABL). Thunderstorms draw in pre-storm ABL air, which rises and cools in the thunderstorm updraft. As water vapor condenses, it releases latent heat, which is the main energy source for the storm (see the Sample Application). Thus, the ABL is the fuel tank for the storm. In general, stronger thunderstorms form in **moister warmer ABL air** (assuming all other factors are constant, such as the environmental sounding, wind shear, trigger, etc.).

The **dew-point temperature** T_d in the ABL is a good measure of the low-altitude humidity. High dew points also imply high air temperature, because $T \geq T_d$ always (see the Water Vapor chapter). Higher temperatures indicate more sensible heat, and higher humidity indicates more latent heat. Thus, high dew points in the ABL indicate a large fuel supply in the ABL environment that can be tapped by thunderstorms. Thunderstorms in regions with $T_d \geq 16°C$ can have heavy precipitation, and those in regions with $T_d \geq 21°C$ can have greater severity.

Sample Application

How much **energy does an air-mass thunderstorm** release? Assume it draws in atmospheric boundary layer (ABL, or BL) air of T_d = 21°C & depth Δz_{BL} = 1 km (corresponding roughly to ΔP_{BL} = 10 kPa), and that all water vapor condenses. Approximate the cloud by a cylinder of radius R = 5 km and depth Δz = 10 km with base at P = 90 kPa & top at P = 20 kPa.

Find the Answer

Given: R = 5 km, Δz = 10 km, T_d = 21°C,
Δz_{BL} = 1 km, ΔP_{BL} = 10 kPa,
ΔP_{storm} = 90 – 20 = 70 kPa
Find: ΔQ_E = ? (J)

Use eq. (3.3) with $L_v = 2.5\text{x}10^6\ \text{J·kg}^{-1}$, and eq. (4.3)

$$\Delta Q_E = L_v \cdot \Delta m_{water} = L_v \cdot r \cdot \Delta m_{air} \qquad \text{(a)}$$

where $r \approx 0.016\ \text{kg}_{water}\cdot\text{kg}_{air}^{-1}$ from thermo diagram.

But eq. (1.8) $\Delta P = \Delta F/A$ and eq. (1.24) $\Delta F = \Delta m \cdot g$ give:

$$\Delta m_{air} = \Delta P \cdot A/g \qquad \text{(b)}$$

where A = surface area = πR^2.

For a cylinder of air within the ABL, but of the same radius as the thunderstorm, use eq. (b), then (a):

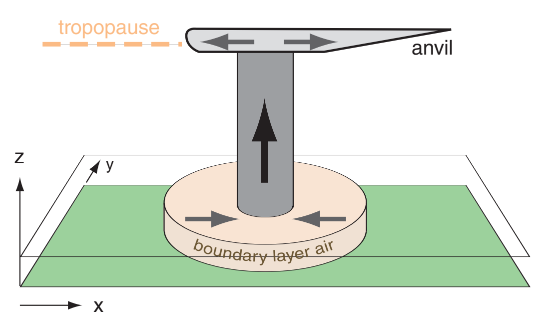
Δm_{air} = (10 kPa)·π·(5000m)^2/(9.8 m s^{-2})= 8x10^8 kg$_{air}$

ΔQ_E=(2.5x10^6 J·kg^{-1})(0.016 kg$_{water}$·kg$_{air}^{-1}$)·(8x10^8 kg$_{air}$) = 3.2x10^{15} J

Given the depth of the thunderstorm (filled with air from the ABL) compared to the depth of the ABL, we find that $\Delta P_{storm} = 7 \cdot \Delta P_{BL}$ (see Figure).

Thus, $\Delta Q_{E\ storm} = 7\ \Delta Q_{E\ BL}$

$\Delta Q_{E\ storm}$ = **2.24x10^{16}** J



Check: Physics OK. Units OK. Values approximate.
Exposition: A **one-megaton nuclear bomb** releases about 4x10^{15} J of heat. This hypothetical thunderstorm has the power of 5.6 one-megaton bombs.

Actual heat released in a small thunderstorm is about 1% of the answer calculated above. Reasons include: T_d < 21°C; storm entrains non-ABL air; and not all of the available water vapor condenses. However, supercells continually draw in fresh ABL air, and can release more heat than the answer above. Energy released differs from energy available (CAPE). For another energy estimate, see the "Heavy Rain" section of the next chapter.

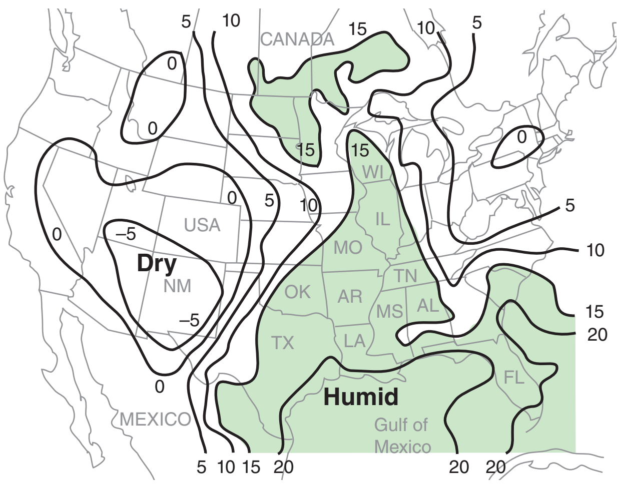


Figure 14.28
Surface weather map valid at 22 UTC on 24 May 2006 showing isodrosotherms (lines of equal dew-point, T_d, in °C) over N. America. Moistest air is shaded, and highlights the greatest fuel supply for thunderstorms. A dry line (sharp decrease in humidity) exists in west Texas (TX) and Oklahoma (OK).

Note, this is a different case study than was used in the previous chapter. The Extratropical Cyclone chapter used a Winter storm case, while here we use an early Spring severe-storm case.

On weather maps, lines of equal dew-point temperature are called **isodrosotherms** (recall Table 1-6). Fig. 14.28 shows an example, where the isodrosotherms are useful for identifying regions having warm humid boundary layers.

Most of the weather maps in this chapter and the next chapter are from a severe weather case on 24 May 2006. These case-study maps are meant to give one example, and do not show average or climatological conditions. The actual severe weather that occurred for this case is described at the end of this chapter, just before the Review.

An alternative moisture variable is **mixing ratio**, r. Large mixing-ratio values are possible only if the air is warm (because warm air can hold more water vapor at saturation), and indicate greater energy available for thunderstorms. For example, Fig. 14.29 shows **isohumes** of mixing ratio.

Wet-bulb temperature T_w , **wet-bulb potential temperature** θ_w , **equivalent potential temperature** θ_e , or **liquid-water potential temperature** θ_L also indicate moisture. Recall from Normand's rule in the Water Vapor chapter that the wet-bulb potential temperature corresponds to the moist adiabat that passes through the LCL on a thermo diagram. Also in the Water-Vapor chapter is a graph relating θ_w to θ_e.

For afternoon thunderstorms in the USA, prestorm boundary layers most frequently have wet-bulb potential temperatures in the θ_w = 20 to 28°C range (or θ_e in the 334 to 372 K range, see example in Fig. 14.30). For supercell thunderstorms, the boundary-layer average is about θ_w = 24°C (or θ_e = 351 K), with some particularly severe storms (strong tornadoes or large hail) having $\theta_w \ge$ 27°C (or $\theta_e \ge$ 366 K).

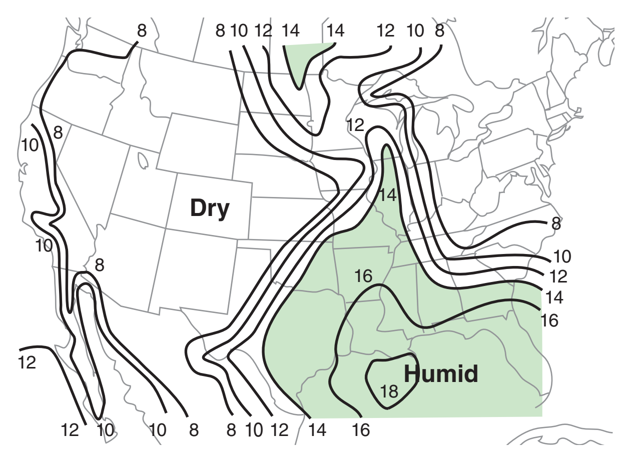


Figure 14.29
Similar to the previous weather map, but for isohumes of mixing ratio (g·kg^{-1}).

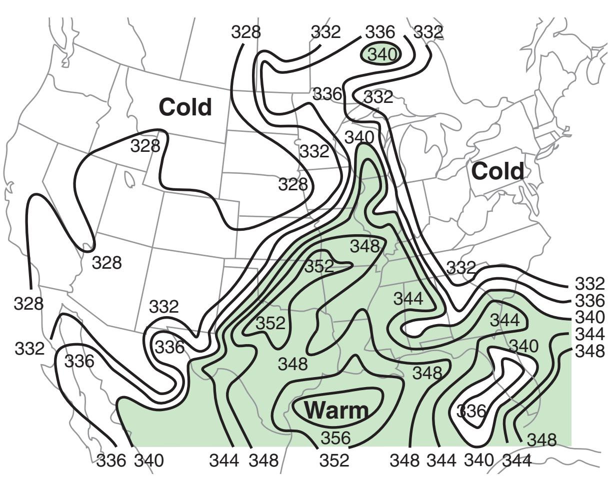


Figure 14.30
Similar to previous figure, but for equivalent potential temperature θ_e in Kelvin. Larger θ_e indicates warmer, moister air.

Precipitable water gives the total water content in a column of air from the ground to the top of the atmosphere (see example in Fig. 14.31). It does not account for additional water advected into the storm by the inflow winds, and thus is not a good measure of the total amount of water vapor that can condense and release energy.

Mean-layer lifting condensation level (**ML-LCL**) is the average of the LCL altitudes for all air parcels starting at heights within the bottom 1 km of the atmosphere (i.e., ≈ boundary layer). Lower ML-LCL values indicate greater ABL moisture, and favor stronger storms (Figs. 14.32 & 14.33).

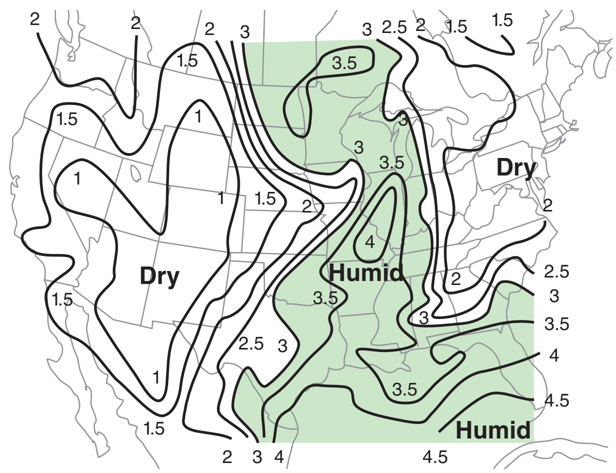


Figure 14.31
Similar to previous figure, but for precipitable water in cm.

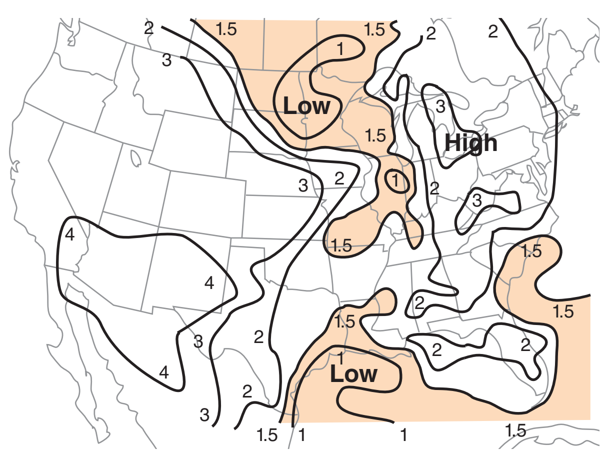


Figure 14.32
Similar to previous figure, but for mean-layer lifting condensation level (ML-LCL) heights in km above ground level. Lower ML-LCL heights (suggesting more intense storms) are shaded.

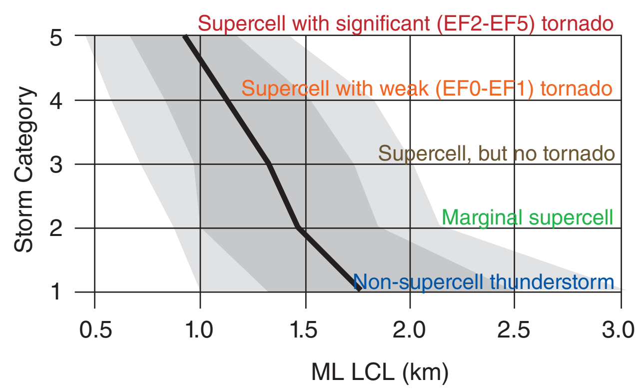


Figure 14.33
Statistical relationship between storm category (as labeled along the horizontal lines) and the mean-layer lifting condensation level (ML-LCL). A marginal supercell is one with weak (< 20 m s^{-1}) or short-duration (< 30 min) cyclonic shear. EF is the Enhanced Fujita scale tornado intensity. Thick line is median (50 percentile) of about 500 observed thunderstorms in the central USA. Dark grey spans 25th through 75th percentiles (i.e., the interquartile range), and light grey spans 10th through 90th percentiles. Lower ML-LCL values generally correspond to greater storm intensities. Caution: There is significant overlap of ML-LCLs for all storm categories, implying that ML-LCL cannot sharply discriminate between storm severities.

See the INFO box on the next page to learn about the median, interquartile range, and percentiles.

INFO • Median, Quartiles, Percentiles

Median, quartiles, and percentiles are statistical ways to summarize the location and spread of experimental data. They are a robust form of **data reduction**, where hundreds or thousands of data are represented by several summary statistics.

First, **sort** your data from the smallest to largest values. This is easy to do on a computer. Each data point now has a **rank** associated with it, such as 1st (smallest value), 2nd, 3rd, ... n^{th} (largest value). Let $x_{(r)}$ = the value of the r^{th} ranked data point.

The middle-ranked data point [i.e., at $r = (1/2)\cdot(n+1)$] is called the **median**, and the data value x of this middle data point is the median value ($q_{0.5}$). Namely,

$$q_{0.5} = x_{(1/2)\cdot(n+1)} \qquad \text{for } n = \text{odd}$$

If n is an even number, there is no data point exactly in the middle, so use the average of the 2 closest points:

$$q_{0.5} = 0.5\cdot [x_{(n/2)} + x_{(n/2)+1}] \qquad \text{for } n = \text{even}$$

The median is a measure of the **location** or center of the data

The data point with a rank closest to $r = (1/4)\cdot(n+1)$ is the **lower quartile** point:

$$q_{0.25} = x_{(1/4)\cdot(n+1)}$$

The data point with a rank closest to $r = (3/4)\cdot(n+1)$ is the **upper quartile** point:

$$q_{0.75} = x_{(3/4)\cdot(n+1)}$$

These last 2 equations work well if n is large (≥100, see below). The **interquartile range** (**IQR**) is defined as IQR = $q_{0.75} - q_{0.25}$, and is a measure of the **spread** of the data. (See the Sample Application nearby.)

Generically, the variable q_p represents any **quantile**, namely the value of the ranked data point having a value that exceeds portion p of all data points. We already looked at p = 1/4, 1/2, and 3/4. We could also divide large data sets into hundredths, giving **percentiles**. The lower quartile is the same as the 25th percentile, the median is the 50th percentile, and the upper quartile is the 75th percentile.

These **non-parametric statistics** are **robust** (usually give a reasonable answer regardless of the actual distribution of data) and **resistant** (are not overly influenced by **outlier** data points). For comparison, the mean and standard deviation are NOT robust nor resistant. Thus, for experimental data, you should use the median and IQR.

To find quartiles for a small data set, split the ranked data in half, and look at the lower and upper halves separately.

Lower half of data: If n = odd, consider the data points ranked less than or equal to the median point. For n = even, consider points with values less than the median value. For this subset of data, find its median, using the same tricks as in the previous paragraph. The resulting data point is the **lower quartile**.

Upper half of data: For n = odd, consider the data points ranked greater than or equal to the original median point. For n = even, use the points with values greater than the median value. The median point in this data subset gives the **upper quartile**.

Sample Application (§)

Suppose the z_{LCL} (km) values for 9 supercells (with EF0-EF1 tornadoes) are:

1.5, 0.8, 1.4, 1.8, 8.2, 1.0, 0.7, 0.5, 1.2

Find the median and interquartile range. Compare with the mean and standard deviation.

Find the Answer:

Given: data set listed above.

Find: $q_{0.5}$ = ? km, IQR = ? km,
Mean$_{zLCL}$ = ? km, σ_{zLCL} = ? km

First sort the data in ascending order:

Values (z_{LCL}):	0.5,	0.7,	0.8,	1.0,	1.2,	1.4,	1.5,	1.8,	8.2
Rank (r):	1	2	3	4	5	6	7	8	9
Middle:					^				

Thus, the median point is the 5th ranked point in the data set, and corresponding value of that data point is **median = $q_{0.5}$ = $z_{LCL(r=5)}$ = 1.2 km**.

Because this is a small data set, use the special method at the bottom of the INFO box to find the quartiles.

Lower half:

Values:	0.5,	0.7,	0.8,	1.0,	1.2
Subrank:	1	2	3	4	5
Middle:			^		

Thus, the lower quartile value is $q_{0.25}$ = 0.8 km

Upper half:

Values:	1.2,	1.4,	1.5,	1.8,	8.2
Subrank:	1	2	3	4	5
Middle:			^		

Thus, the upper quartile value is $q_{0.75}$ = 1.5 km

The IQR = $q_{0.75} - q_{0.25}$ = (1.5km – 0.8km) = **0.7 km**

Using a spreadsheet to find the mean and standard deviation:

Mean$_{zLCL}$ = **1.9 km** , σ_{zLCL} = **2.4 km**

Check: Values reasonable. Units OK.

Exposition: The original data set has one "wild" z_{LCL} value: 8.2 km. This is the **outlier**, because it lies so far from most of the other data points.

As a result, the mean value (1.9 km) is not representative of any of the data points; namely, the center of the majority of data points is not at 1.9 km. Thus, the mean is not robust. Also, if you were to remove that one outlier point, and recalculate the mean, you would get a significantly different value (1.11 km). Hence, the mean is not resistant. Similar problems occur with the standard deviation.

However, the median value (1.2 km) is nicely centered on the majority of points. Also, if you were to remove the one outlier point, the median value would change only slightly to a value of 1.1 km. Hence, it is robust and resistant. Similarly, the IQR is robust and resistant.

14.4. INSTABILITY, CAPE & UPDRAFTS

The second requirement for convective-storm formation is instability in the pre-storm sounding. **Nonlocal conditional instability** (**NCI**) occurs when warm humid atmospheric boundary layer (ABL) air is capped by a temperature inversion, above which is relatively cold air. The cold air aloft provides an environment that gives more buoyancy to the warm updraft air from below, allowing stronger thunderstorms. The "nonlocal" aspect arises because air from below the cap becomes unstable above the cap. The "condition" is that the ABL air must first be lifted past the cap (i.e., past its LCL and LFC) for the instability to be realized.

The capping inversion traps the warm humid air near the ground, allowing sensible and latent heat energy to build up during the day as the sun heats the ground and causes evaporation. Without this cap, smaller cumulus clouds can withdraw the warm humid air from the boundary layer, leaving insufficient fuel for thunderstorms. Thus, the cap is important — it prevents the fuel from leaking out.

14.4.1. Convective Available Potential Energy

Thunderstorms get their energy from the buoyancy associated with latent-heat release when water vapor condenses. The **Convective Available Potential Energy** (CAPE) is a way to estimate this energy using a thermo diagram. **CAPE** is proportional to the shaded area in Fig 14.34; namely, the area between LFC and EL altitudes that is bounded by the environmental sounding and the moist adiabat of the rising air parcel.

To explain this, you can use the definition of buoyancy force per unit mass $F/m = |g| \cdot (T_{vp} - T_{ve}) / T_{ve}$ that was covered in the Atmospheric Stability chapter. T_{vp} is the virtual temperature of the air parcel rising in the thunderstorm, T_{ve} is virtual temperature in the surrounding environment at the same altitude as the thunderstorm parcel, and $|g|$ = 9.8 m s^{-2} is gravitational acceleration magnitude.

Recall from basic physics that **work** equals force times distance. Let $(\Delta E/m)$ be the incremental work per unit mass associated with a thunderstorm air parcel that rises a small increment of distance Δz. Thus, $\Delta E/m = (F/m) \cdot \Delta z$, or:

$$\frac{\Delta E}{m} \cong \Delta z \cdot |g| \frac{(T_{vp} - T_{ve})}{T_{ve}} \tag{14.1}$$

Recall from Chapter 1 that virtual temperature includes the effects of both water vapor and liquid water. Water vapor is less dense than air ($T_v > T$), thus increasing the buoyant energy. Liquid- and

INFO • LCL, LFC & Storm Likelihood

The US Storm Prediction Center has found some clues for thunderstorm intensity based on the lifting condensation level (LCL) and the level of free convection (LFC):

- Tornadoes are more likely when $z_{LFC} < 2$ km agl.
- Thunderstorms are more easily triggered and maintained when $z_{LFC} < 3$ km agl.
- Deep convection is more likely when $z_{LFC} - z_{LCL}$ is smaller.

where agl = above ground level.

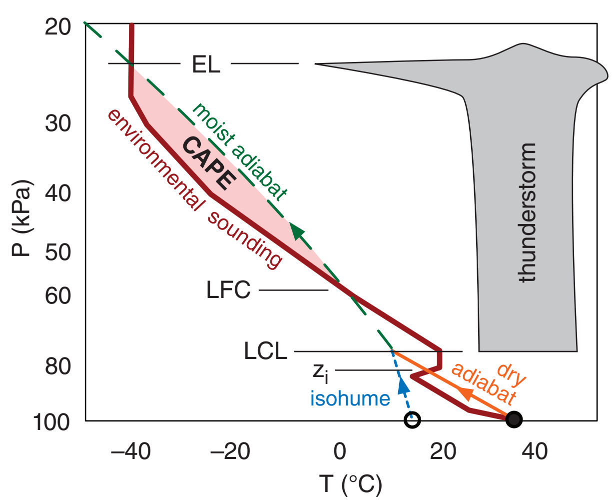


Figure 14.34
The pink shaded area indicates the surface-based Convective Available Potential Energy (CAPE) for an afternoon pre-storm environment.

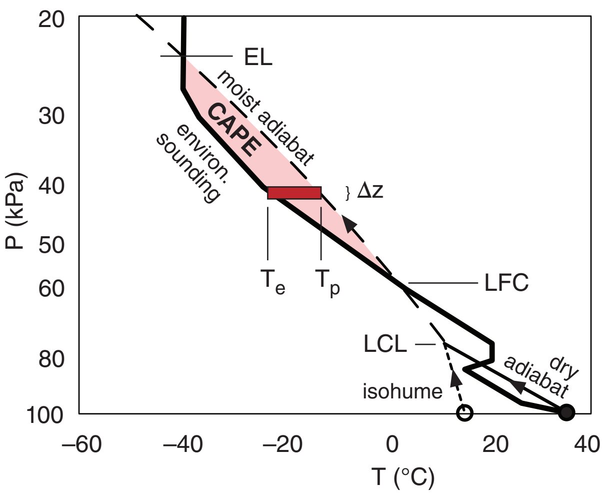


Figure 14.35
Dark shaded rectangle of incremental height Δz and width $T_p - T_e$ shows the portion of total CAPE area associated with just one thin layer of air.

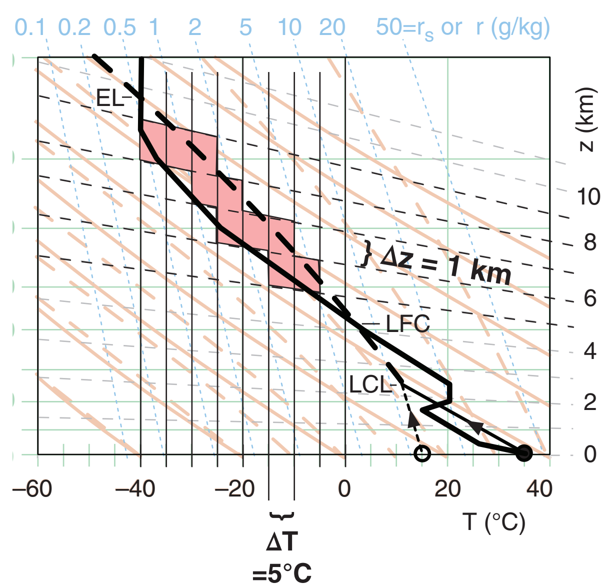


Figure 14.36
Approximating the CAPE area by non-overlapping tiles, each of area Δz by ΔT. In this example, Δz = 1 km, and ΔT = 5°C.

solid-water hydrometeors (cloud droplets, rain, and snow) falling at their terminal velocity are heavier than air ($T_v < T$), thus decreasing the upward buoyant energy. Both are important for thunderstorms, but are often difficult to determine.

Instead, you can use the approximation $T_v \approx T$, which gives:

$$\frac{\Delta E}{m} \cong |g| \frac{\Delta z \cdot (T_p - T_e)}{T_e} = |g| \cdot (Increm.Area) / T_e \qquad (14.2)$$

where T_p is air-parcel temperature, T_e is environmental temperature, and the incremental area (*Increm. Area*) is shown Fig. 14.35 as the dark-red rectangle.

Adding all the incremental rectangles between the LFC and the EL gives a total area (light shading in Fig. 14.35) that is proportional to CAPE.

$$CAPE \cong |g| \cdot (Total.Area) / T_e \qquad (14.3)$$

or

$$CAPE \cong |g| \sum_{LFC}^{EL} (T_p - T_e) \cdot \Delta z / T_e \qquad \bullet(14.4)$$

The units of CAPE are J·kg^{-1}. These units are equivalent to m^2·s^{-2} ; namely, velocity squared. The temperature in the denominator of eqs. (14.3 & 14.4) must be in Kelvin. Both numerator temperatures must have the same units: either Kelvin or °C.

The shape of the CAPE area is usually not simple, so calculating the area is not trivial. At severe-weather forecast centers, computers calculate CAPE automatically based on the pre-storm sounding. By hand, you can use a simple graphical method by first plotting the sounding and the surface-parcel rise on a thermo diagram, and then using whatever height and temperature increments are plotted on the background diagram to define "bricks" or "tiles", each of known size.

Namely, instead of using long narrow rectangles, each of different width, as sketched in Fig. 14.35, you can cover the complex shaded area with smaller, non-overlapping tiles, each of equal but arbitrary size ΔT by Δz (such as ΔT = 5°C and Δz = 1 km, see Fig. 14.36). Count the number of tiles, and multiply the result by the area of each tile ($\Delta T \cdot \Delta z$ = 5 °C·km in this example). When tiling the CAPE area, try to compensate for small areas that are missed by the tiles by allowing some of the other tiles to extend slightly beyond the boundaries of the desired shaded area by roughly an equal area (e.g., see the Sample Application). Smaller-size tiles (such as ΔT = 1°C and Δz = 0.2 km) give a more accurate answer (and are recommended), but are more tedious to count.

Sample Application

For the sounding in Fig 14.36, estimate the CAPE.

Find the Answer

Given: Fig 14.36.
Find: CAPE = ? J·kg^{-1}

Each tile has size ΔT = 5°C by Δz = 1000 m. The area of each tile is 5,000 °C·m, and there are 10 tiles. The total area is: *Area* = 10 x 5000 °C·m = 50,000 K·m , [where I took advantage of ΔT(°C) = ΔT(K)].

By eye, the average T_e in the CAPE region is about –25°C = 248 K. Use eq. (14.3):

CAPE = [9.8 m·s^{-2} / 248K]·(50,000 K·m) =
= 1976 m^2·s^{-2} = **1976 J·kg^{-1}**.

Check: Physics and units OK. Figure OK.
Exposition: This is a moderate value of CAPE that could support supercell storms with tornadoes.

On many thermo diagrams such as the emagram of T vs. ln(P) used here, the height contours and isotherms are not perpendicular; hence, the rectangles look like trapezoids (see the Sample Application that employed Fig. 14.36). Regardless of the actual area within each trapezoid in the plotted graph, each trapezoid represents a contribution of ΔT by Δz toward the total CAPE area.

To replace heights with pressures in the equation for CAPE, use the hypsometric equation from Chapter 1, which yields:

$$CAPE = \Re_d \cdot \sum_{LFC}^{EL} (T_p - T_e) \cdot \ln\left(\frac{P_{bottom}}{P_{top}}\right) \qquad \bullet(14.5)$$

where $\Re_d$ = 287.053 J·K^{-1}·kg^{-1} is the gas constant for dry air, P_{bottom} and P_{top} are the bottom and top pressures of the incremental rectangle, and the sum is still over all the rectangles needed to tile the shaded CAPE area on the sounding (Fig. 14.37). Again, the temperature difference in eq. (14.5) in °C is equal to the same value in Kelvin.

All of the CAPE figures up until now have followed a rising air parcel that was assumed to have started from near the surface. This is called **Surface-Based CAPE** (SBCAPE), which is often a good method for the mid-afternoon pre-storm soundings that have been shown so far.

CAPE values vary greatly with location and time. By finding the CAPE for many locations in a region (by using rawinsonde observations, or by using forecast soundings from numerical weather prediction models), you can write the CAPE values on

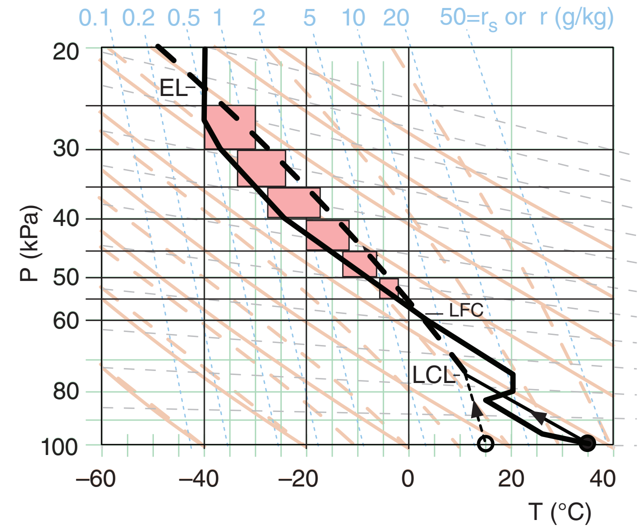


Figure 14.37
Approximating the CAPE area by non-overlapping tiles, each of area ΔP by ΔT. In this example, ΔP = 5 kPa, and ΔT varies.

Sample Application

For the sounding in Fig. 14.37, estimate the CAPE using pressure rather than height increments.

Find the Answer:

Given: Fig. 14.37.
Find: CAPE = ? J·kg^{-1}

Use eq. (14.5):
CAPE = [287 J/(K·kg)] · {
[–30°C – (–40°C)] · ln(30kPa/25kPa)] +
[–24°C – (–33°C)] · ln(35kPa/30kPa)] +
[–17°C – (–27°C)] · ln(40kPa/35kPa)] +
[–12°C – (–20°C)] · ln(45kPa/40kPa)] +
[–7°C – (–13°C)] · ln(50kPa/45kPa)] +
[–2°C – (–6°C)] · ln(55kPa/50kPa)] }

CAPE = [287 J·(K·kg)$^{-1}$] ·
{ 10K · 0.182 + 9.2K · 0.154 + 10K · 0.134 +
8K · 0.118 + 6K · 0.105 + 4K · 0.095 }

CAPE = [287 J·(K·kg)$^{-1}$] · 6.53 K = **1874 J·kg^{-1}** .

Check: Physics and units OK.
Exposition: Theoretically, we should get exactly the same answer as we found in the previous Sample Application. Considering the coarseness of the boxes that I used in both Sample Applications, I am happy that the answers are as close as they are. I found the height-tiling method easier than the pressure-tiling method.

a weather map and then draw isopleths connecting lines of equal CAPE, such as shown in Fig. 14.38.

Other parcel-origin assumptions work better in other situations. If the pre-storm sounding is from earlier in the morning, then the forecast max surface temperature for that afternoon (along with the dew point forecast for that time) is a better choice for the rising air-parcel initial conditions (Fig. 14.39). This is also a type of SBCAPE.

Another way is to average the conditions in the bottom 1 km (roughly 10 kPa) of the environmental sounding to better estimate ABL conditions. Thus, the initial conditions for the rising air parcel represent the mean layer (ML) conditions, or the mixed-layer (ML) conditions (Fig. 14.40). CAPE calculated this way is called **Mean Layer CAPE** (MLCAPE). Fig. 14.41 shows a case-study example of MLCAPE.

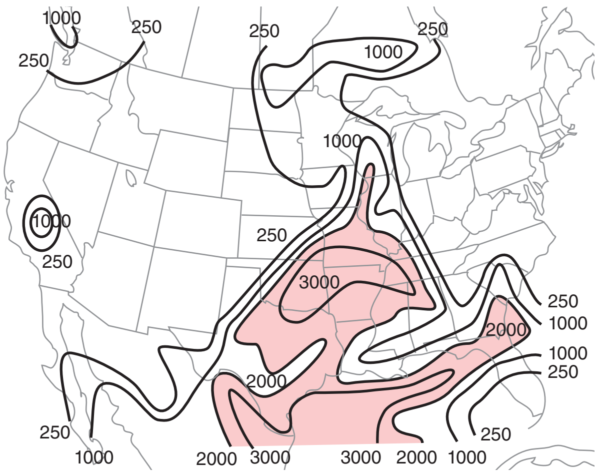


Figure 14.38
Weather map of surface-based CAPE (SBCAPE) in J·kg^{-1} over N. America. Valid 22 UTC on 24 May 2006. Shaded region highlights larger SBCAPE values, where more intense thunderstorms can be supported.

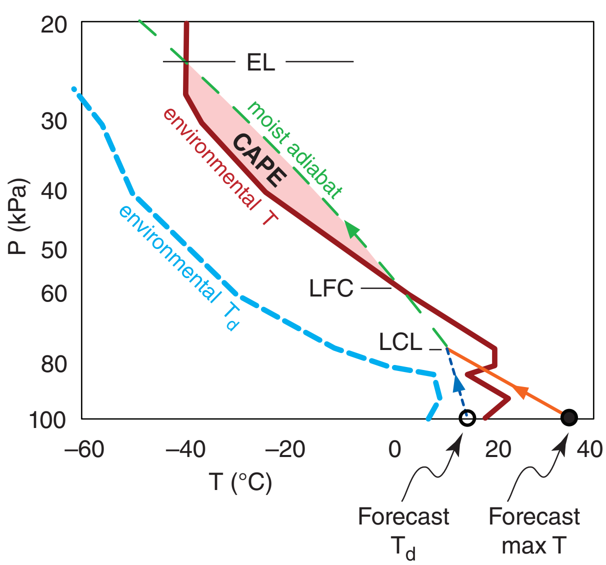


Figure 14.39
Shown is one method for estimating CAPE from an early morning sounding. Instead of using actual surface environmental conditions for the air parcel, you can use the forecast maximum near-surface air temperature (max T) and dew-point temperature (T_d) for later that day, and assume no change in environmental conditions above the ML.

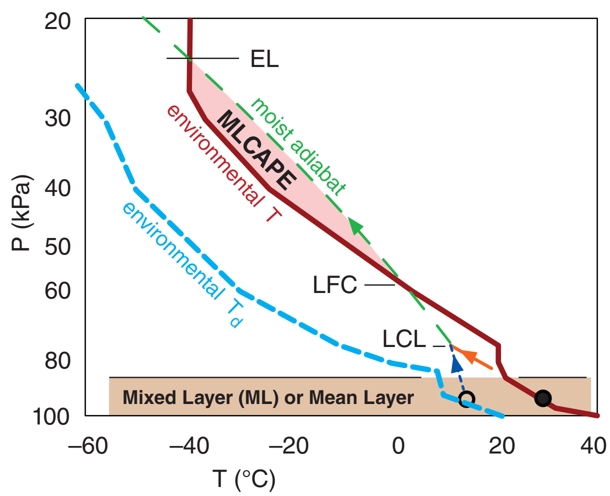


Figure 14.40
Another method for estimating CAPE is to use air-parcel initial conditions equal to the average conditions (filled and open circles) in the boundary layer (shaded). This is called the mean layer CAPE (MLCAPE).

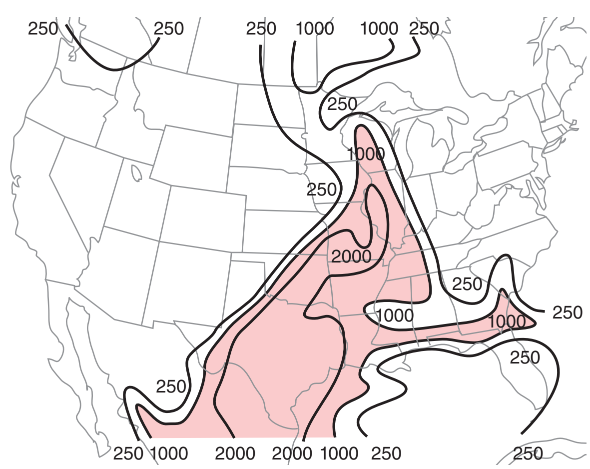


Figure 14.41
Weather map similar to Fig. 14.38, but for mean-layer CAPE (MLCAPE) in J·kg^{-1}.

Early studies of thunderstorm occurrence vs. MLCAPE lead to forecast guidelines such as shown in Table 14-1. However, MLCAPE is not a sharp discriminator of thunderstorm intensity, as shown by the large overlap of storm categories in Fig. 14.42.

Yet another way is to calculate many different CAPEs for air parcels that start from every height in the bottom 30 kPa of the pre-storm environmental sounding, and then select the one that gives the greatest CAPE values. This is called the **Most Unstable CAPE** (MUCAPE). This method works even if thunderstorm updrafts are triggered by an elevated source, and also works for pre-storm soundings from any time of day. Although it is too tedious to compute these multiple CAPEs by hand, it is easily automated on a computer (Figs. 14.43 & 14.44). MUCAPE is always ≥ SBCAPE. But MUCAPE is not a sharp discriminator of thunderstorm intensity.

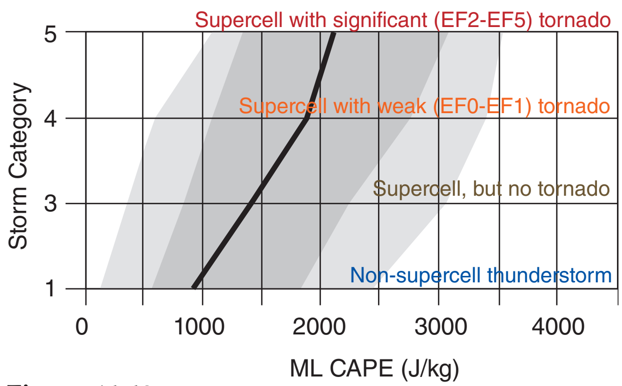


Figure 14.42
Statistics of thunderstorm intensity vs. MLCAPE, based on several hundred paired soundings and thunderstorms in central N. America. Dark line is median (50th percentile); dark shading spans the interquartile range (25th to 75th percentiles); light shading spans the 10th to 90th percentile range.

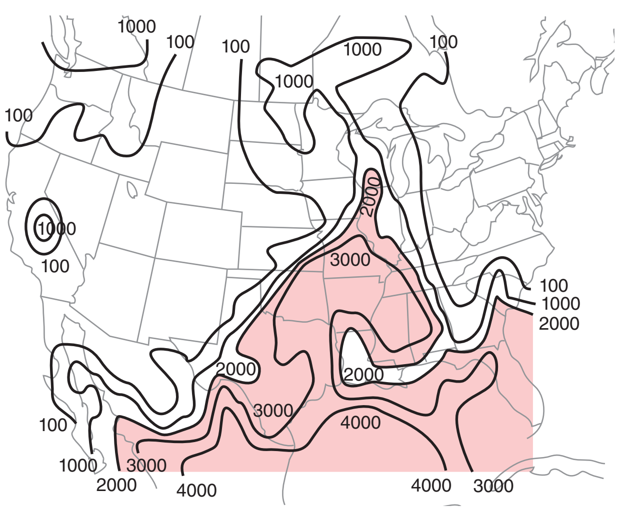


Figure 14.43
Weather map similar to Fig. 14.38 over the USA, but for most-unstable CAPE (MUCAPE) in J·kg^{-1}.

Table 14-1. Thunderstorm (CB) intensity guide.		
MLCAPE (J·kg^{-1})	**Stability Description**	**Thunderstorm Activity**
0 - 300	mostly stable	little or none
300 - 1000	marginally unstable	weak CB
1000 - 2500	moderately unstable	moderate CB likely; severe CB possible
2500 - 3500	strongly unstable	severe CB likely & possible tornado
3500 & greater	extremely unstable	severe CB & tornadoes likely

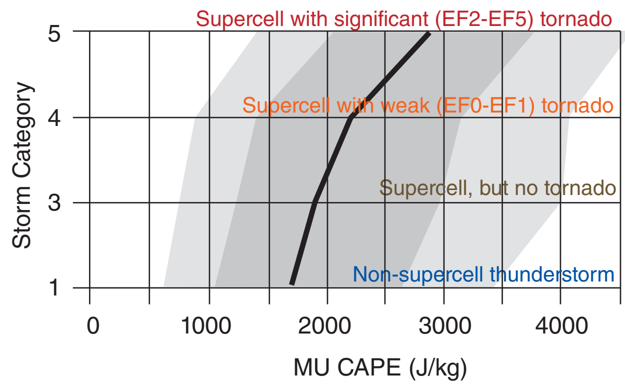


Figure 14.44
Statistics of thunderstorm intensity vs. MUCAPE, similar to Fig. 14.42.

Sample Application

Assume a thunderstorm forms at the dot in Fig. 14.43, where MUCAPE ≈ 2700 J·kg^{-1}. Estimate the possible intensity of this storm. Discuss the uncertainty.

Find the Answer

Given: MUCAPE ≈ 2,700 J·kg^{-1} at the dot (•) in Fig. 14.43, located in southern Illinois.
Find: Possible storm intensity. Discuss uncertainty.

Use Fig. 14.44 to estimate intensity. For MUCAPE of 2,700 J·kg^{-1}, the median line is about half way between "supercell with **weak tornado** (EF0-EF1)", and "supercell with **significant tornado** (EF2-EF5)." So we might predict a supercell with an EF1 - EF2 tornado.

Exposition: Although there is some uncertainly in picking the MUCAPE value from Fig. 14.43, there is even greater uncertainty in Fig 14.44. Namely, within the interquartile range (dark shading) based on hundreds of past storms used to make this figure, there could easily be a supercell with no tornado, or a supercell with a significant tornado. There is even a chance of a non-supercell thunderstorm.

Thus, MUCAPE is not capable of sharply distinguishing thunderstorm intensity.

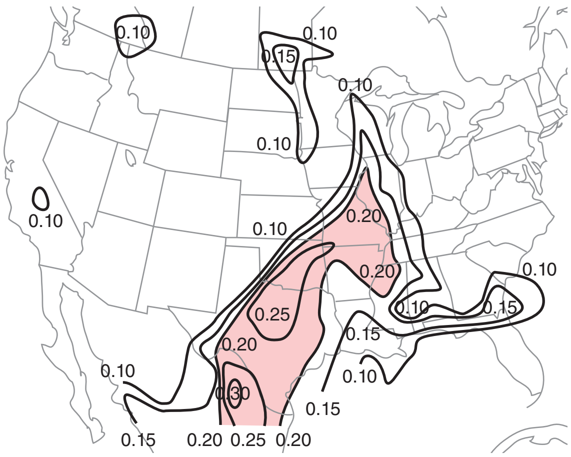


Figure 14.45
Weather map similar to Fig. 14.38 over the USA, but for normalized CAPE (nCAPE) in m·s^{-2}.

Sample Application
For the sounding in Fig. 14.36 (CAPE = 1976 J·kg^{-1}) find thunderstorm intensity, normalized CAPE, max updraft velocity, and likely updraft velocity.

Find the Answer
Given: Fig. 14.36, with CAPE = 1976 J·kg^{-1}
Find: Intensity = ? , $nCAPE$ = ? m·s^{-2},
w_{max} & $w_{max\ likely}$ = ? m s^{-1}

a) From Table 14-1, **moderate thunderstorms**.

b) By eye using Fig. 14.36, the bottom and top of the CAPE region are $z_{LFC} \approx 4.2$ km, and $z_{EL} = 9.6$ km.
Use eq. (14.6):
$nCAPE$=(1976 J·kg^{-1})/[9600–4200m]=**0.37**m·s^{-2}.

c) Use eq. (14.7):
$w_{max} = [2 \cdot (1976\ \text{m}^2\ \text{s}^{-2})\ ]^{1/2} = w_{max} = \underline{\mathbf{63}}$ m s^{-1} .

d) Use eq (14.8):
$w_{max\ likely}$ = (63 m s^{-1})/2 = **31** m s^{-1} .

Check: Physics and units OK. Figure OK.
Exposition: Because the LFC is at such a high altitude, the CAPE area is somewhat short and fat, as indicated by the small value of nCAPE. Thus, strong low-altitude updrafts and tornadoes are possible. This thunderstorm has violent updrafts, which is why aircraft would avoid flying through it.

The shape of the CAPE area gives some information about the storm. To aid shape interpretation, a **normalized CAPE** (**nCAPE**) is defined as

$$nCAPE = \frac{CAPE}{z_{EL} - z_{LFC}} \qquad \bullet(14.6)$$

where z_{EL} is height of the equilibrium level, z_{LFC} is height of the level of free convection, and the units of nCAPE are m·s^{-2} or J·(kg·m)$^{-1}$. Tall, thin CAPE areas (i.e., nCAPE ≤ 0.1 m·s^{-2}) often suggest heavy precipitation, but unlikely tornadoes. Short, wide CAPE area (i.e., nCAPE ≥ 0.3 m·s^{-2}) in the mid to lower part of the sounding can result in thunderstorms with strong, low-altitude updrafts, which cause vertical stretching of the air, intensification of rotation, and greater chance of tornadoes. See Fig. 14.45.

A word of caution: CAPE gives only an estimate of the strength of a thunderstorm if one indeed forms. It is a necessary condition, but not a sufficient condition. To form a thunderstorm, there must also be a process that triggers it. Often no thunderstorms form, even in locations having large CAPE. If thunderstorms are triggered, then larger CAPE indicates greater instability, stronger updrafts, and a chance for more violent thunderstorms. It is a useful but not perfect forecast tool, as demonstrated by the lack of statistical sharpness (Figs. 14.42, & 14.44).

14.4.2. Updraft Velocity

You can also use CAPE to estimate the updraft speed in thunderstorms. Recall from basic physics that kinetic energy per unit mass is $KE/m = 0.5 \cdot w^2$, where w is updraft speed. Suppose that all the convective available potential energy could be converted into kinetic energy; namely, CAPE = KE/m. Combining the two equations above gives

$$w_{\text{max}} = \sqrt{2 \cdot CAPE} \qquad (14.7)$$

which gives unrealistically large speed because it neglects entrainment of unsaturated air, frictional drag, and liquid-water loading. Studies of actual thunderstorm updrafts find that the most likely max updraft speed is

$$w_{\text{max likely}} \approx w_{\text{max}} / 2 \qquad \bullet(14.8)$$

Air in an updraft has inertia and can overshoot above the EL. Such **penetrative convection** can be seen by eye as mound or turret of cloud that temporarily overshoots above the top of the thunderstorm anvil. You can also see it in some satellite images. As stated before, such turrets or domes give a good clue to storm spotters that the thunderstorm is probably violent, and has strong updrafts. Such strong updrafts are felt as severe or extreme **turbulence** by aircraft.

14.5. WIND SHEAR IN THE ENVIRONMENT

Wind shear, the change of horizontal wind speed and/or direction with height, is the third requirement for thunderstorm formation. The wind shear across a layer of air is the vector difference between the winds at the top of the layer and winds at the bottom (Fig. 14.46), divided by layer thickness Δz. Shear has units of (s^{-1}).

Using geometry, shear can be expressed via its components as:

$$\frac{\Delta U}{\Delta z} = \frac{U_2 - U_1}{z_2 - z_1} \tag{14.9}$$

$$\frac{\Delta V}{\Delta z} = \frac{V_2 - V_1}{z_2 - z_1} \tag{14.10}$$

or

$$\text{Shear Magnitude} = \frac{\left[(\Delta U)^2 + (\Delta V)^2\right]^{1/2}}{\Delta z} \tag{14.11a}$$

Shear Direction:

$$\alpha_{shear} = 90° - \frac{360°}{C}\arctan\left(\frac{\Delta V}{\Delta U}\right) + \alpha_o \tag{14.11b}$$

where subscript 2 is the layer top, and subscript 1 is layer bottom. Eq. (1.2) was used to find the direction α of the shear vector, but with ΔU in place of U, and ΔV in place of V. α_o = 180° if $\Delta U > 0$, but is zero otherwise. C is the angular rotation in a circle (360° or 2π radians, depending on how your calculator or spreadsheet calculates the "arctan" function).

For thunderstorm forecasting, meteorologists look at shear across many layers at different heights in the atmosphere. To make this easier, you can use layers of equal thickness, such as Δz = 1 km. Namely, look at the shear across the bottom layer from z = 0 to z = 1 km, and the shear across the next layer from z = 1 to z = 2 km, and so forth.

When studying shear across layers of equal thickness, meteorologists often use the **vector wind difference** (ΔU, ΔV) as a surrogate measure of **vector wind shear**. We will use this surrogate here. This surrogate, and its corresponding **wind-difference magnitude**

$$\Delta M = \left(\Delta U^2 + \Delta V^2\right)^{1/2} \tag{14.12}$$

have units of wind speed (m s^{-1}).

Thunderstorms can become intense and long-lasting given favorable wind shear in the lower atmosphere. Under such conditions, humid boundary-layer air (thunderstorm fuel) can be fed into the moving storm (Fig. 14.47). Another way to picture this process is that wind shear causes the storm to move away from locations of depleted boundary-

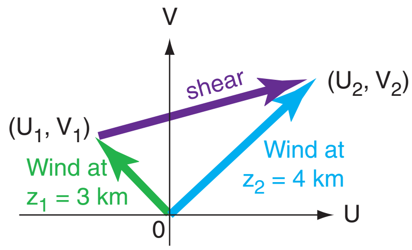
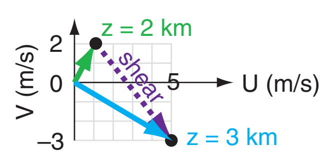


Figure 14.46
Wind difference (dark purple arrow) between two altitudes is the vector difference between wind (long cyan arrow) at the top altitude minus wind at the bottom (short green arrow). We will use vector wind <u>*difference*</u> *as a surrogate for vector wind* <u>*shear*</u>*.*

Sample Application

Given a horizontal wind of (U, V) = (1, 2) m s^{-1} at 2 km altitude, and a wind of (5, –3) m s^{-1} at 3 km, what are the wind-difference & shear magnitudes & direction?

Find the Answer:
Given: (U, V) = (1, 2) m s^{-1} at z = 2 km
(U, V) = (5, –3) m s^{-1} at z = 3 km
Find: ΔM = ? (m s^{-1}), α = ? (°)
Shear Magnitude = ? (s^{-1}),



Use eq. (14.12) to find the wind difference magnitude:
ΔU = (5 – 1) = 4 m s^{-1}. ΔV = (–3 – 2) = –5 m s^{-1}.
ΔM = [(4 m $s^{-1})^2$ + (–5 m $s^{-1})^2$ $]^{1/2}$ =
[16+25 m^2 $s^{-2}]^{1/2}$ = <u>**6.4**</u> m s^{-1}

Use eq. (14.11b) to find shear direction:
α = 90°–arctan(–5/4)+180° = 90°–(–51.3°)+180° = <u>**321.3°**</u>
which is the direction that the wind shear is coming <u>from</u>.

Layer thickness is Δz = 3 km – 2 km = 1 km = 1000 m.
Use eq. (14.11a) to find shear magnitude, rewritten as:
Shear Mag.= $\Delta M/\Delta z$ = (6.4 m s^{-1})/(1000 m) = <u>**0.0064 s**</u>$^{-1}$

Check: Units OK. Magnitude & direction agree with figure.
Exposition: There is both **speed shear** and **directional shear** in this example. It always helps to draw a figure, to check your answer.

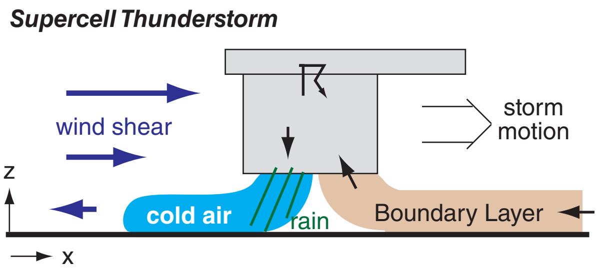


Figure 14.47
Wind shear allows long-lasting strong thunderstorms, such as supercell storms and bow-echo thunderstorm lines.

layer fuel into locations with warm, humid boundary layers. The storm behaves similar to an upright vacuum cleaner, sucking up fresh boundary-layer air as it moves, and leaving behind an exhaust of colder air that is more stable.

In an environment with wind but no wind shear, the thunderstorms would last only about 15 minutes to 1 hour, because the thunderstorm and boundary-layer air would move together. The storm dies after it depletes the fuel in its accompanying boundary layer (Fig. 14.6).

Shear can also control the direction that a thunderstorm moves. As will be shown later, a clockwise turning wind shear vector can create a dynamic vertical pressure gradient that favors thunderstorms that move to the right of the mean steering-level wind direction.

Mesocyclones can also be created by wind shear. A mesocyclone is where the whole thunderstorm rotates. A mesocyclone can be a precursor to strong tornadoes, and is discussed in detail in the next chapter. Strong shear also enhances MCSs such as bow-echo thunderstorms. Thus, there are many ways that **upper-air wind** (winds above the surface) and wind shear affect thunderstorms.

14.5.1. Hodograph Basics

One way to plot upper-air wind vs. height is to draw wind symbols along the side of a thermo diagram (Fig. 14.48). Recall from the Atmospheric Forces and Winds chapter that wind direction is from the tail end (with the feathers) toward the other end of the shaft, and more barbs (feathers or pennants) indicate greater speed. The tip of the arrow is at the altitude of that wind.

Another way to plot winds is to collapse all the wind vectors from different heights into a single polar graph, which is called a **hodograph**. For example, Fig. 14.49 shows winds at 1 km altitude blowing from the southeast at 15 m s^{-1}, winds at 3 km altitude blowing from the south at 25 m s^{-1}, and winds at 5 km altitude blowing from the southwest at 35

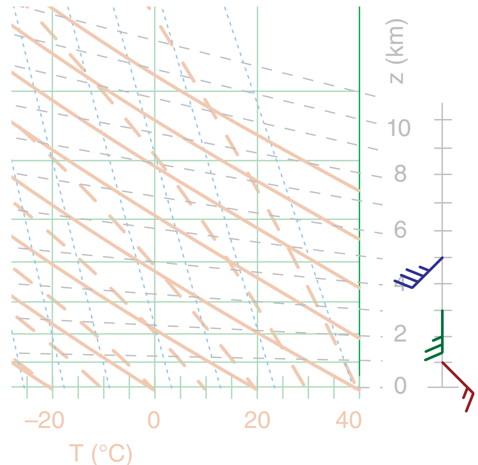


Figure 14.48
Winds at different heights, indicated with wind barbs adjacent to a thermo diagram. The same speeds are indicated by vectors in the next figure. Wind-barb speed units are m s^{-1}. (Look back at Table 10-1 for wind-barb definitions.)

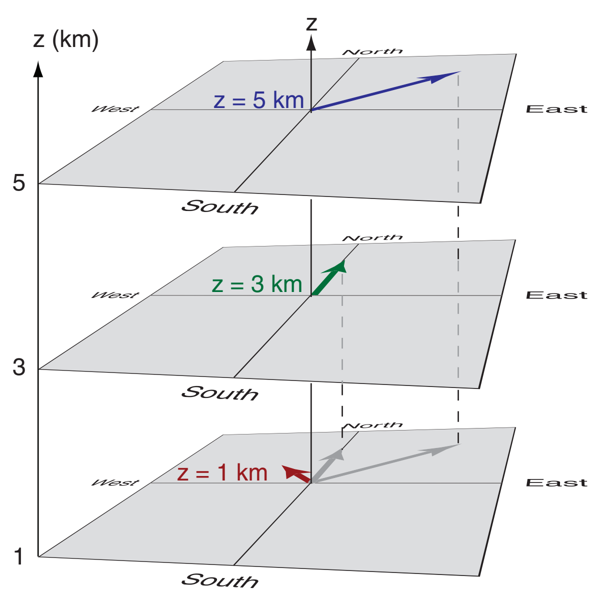


Figure 14.49
Projection of wind vectors (dark arrows) from different altitudes onto the bottom plane (grey arrows). These wind vectors correspond to the wind vectors in the previous figure.

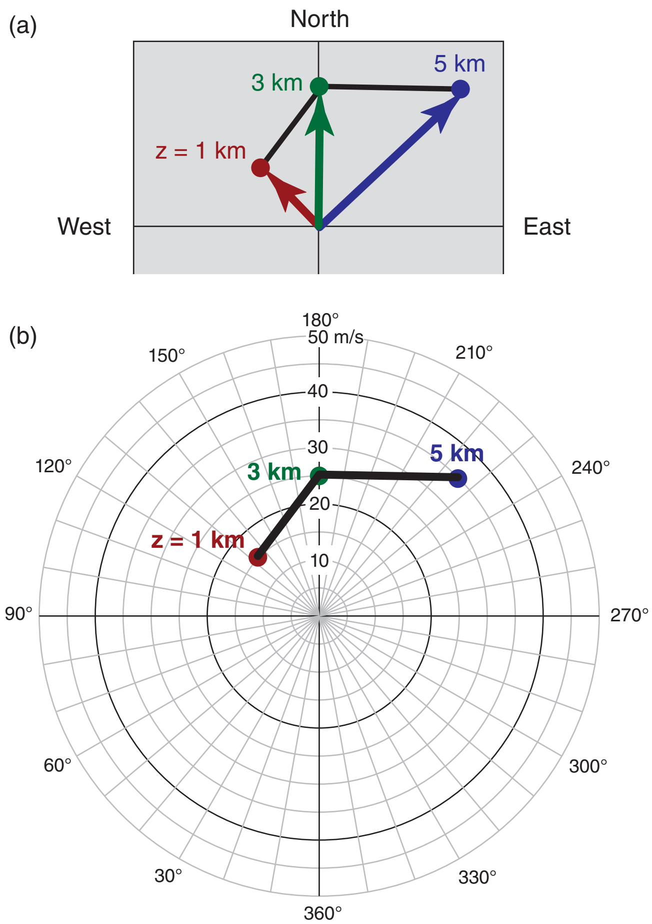


Figure 14.50
(a) Using the projected vectors from the previous figure, draw dots at the end of each vector, and connect with straight lines. When that line and dots are plotted on polar graph paper (b), the result is a hodograph.

m s^{-1}. Both **directional shear** and **speed shear** exist.

If you project those wind vectors onto a single plane, you will get a graph that looks like the bottom of Fig. 14.49. Next, draw a dot at the end of each projected vector (Fig. 14.50a), and connect the dots with straight lines, going sequentially from the lowest to the highest altitude. Label the altitudes above ground level (AGL) next to each dot.

Plot the resulting dots and connecting lines on polar graph paper, but omit the vectors (because they are implied by the positions of the dots). The result is called a **hodograph** (Fig. 14.50b). Fig. 14.51 is a large blank hodograph that you can copy and use for plotting your own wind profiles. Coordinates in this graph are given by radial lines for wind direction, and circles are wind speed.

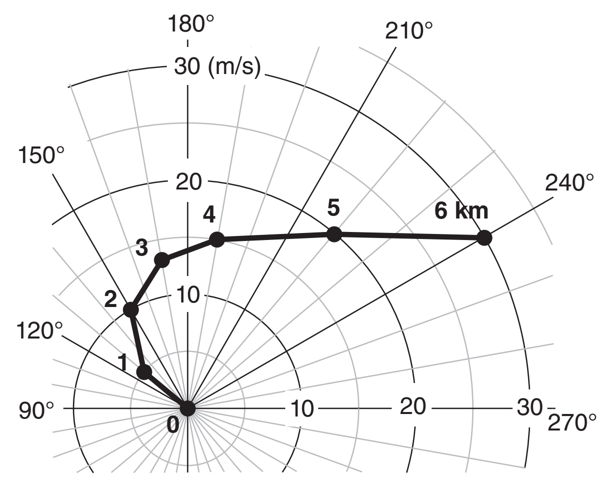
Sample Application

Use the table of winds below to plot a hodograph.

z (km)	dir. (°)	M (m s^{-1})	z (km)	dir. (°)	M (m s^{-1})
0	0	0	4	190	15
1	130	5	5	220	20
2	150	10	6	240	30
3	170	13			

Find the Answer

Plot these points on a copy of Fig. 14.51, and connect with straight lines. Label each point with its altitude.



Check: The wind at 6 km, for example, is from the west southwest (240°), implying that the tip of the wind vector would be pointing toward the east northeast. This agrees with the location of the 6 km dot.

Exposition: This profile shows both speed and directional shear. Also, these winds **veer** (turn clockwise) with increasing altitude. From the thermal wind Exposition in Chapter 11, recall that veering winds are associated with warm-air advection, which is useful for bringing warmer air into a thunderstorm.

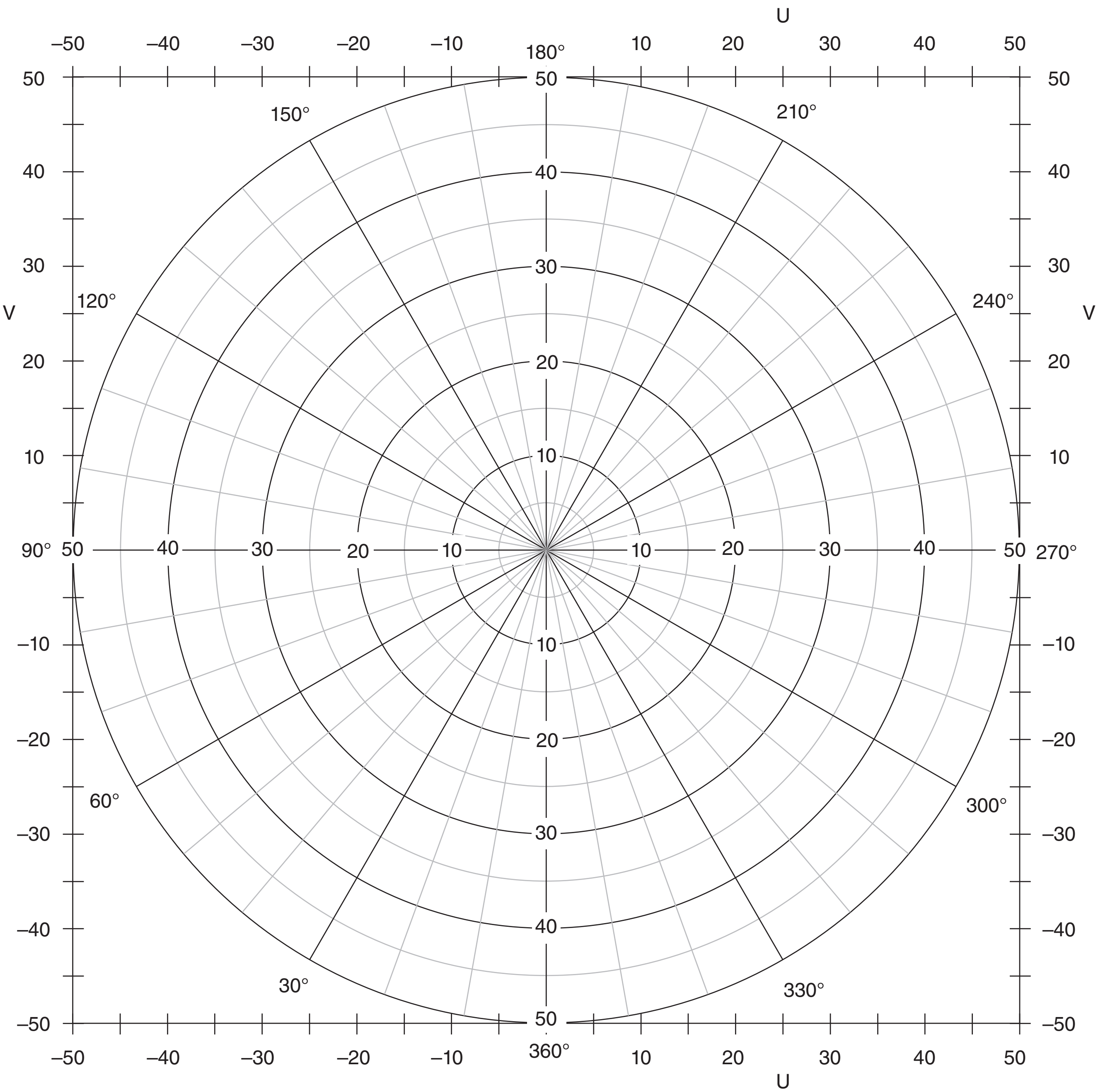


Figure 14.51
Blank hodograph for you to copy and use. Compass angles are direction winds are *__from__*. *Speed-circle labels can be changed for different units or larger values, if needed.*

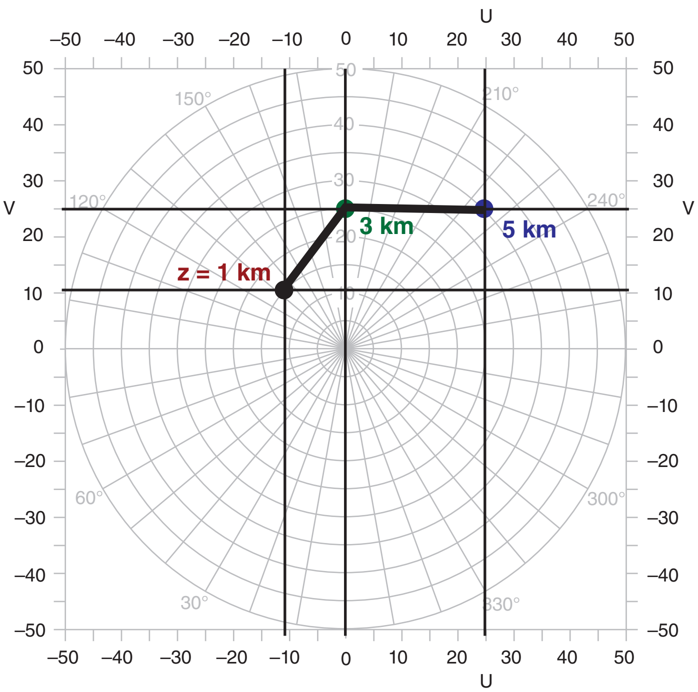


Figure 14.52
Example of determination of U and V Cartesian wind components from a hodograph. Wind speeds are in m s^{-1}.

Sample Application

For the hodograph of Fig. 14.52, find the (U, V) Cartesian coordinate for each of the 3 data points.

Find the Answer

Given: Hodograph Fig. 14.52 with data points at
z = 1, 3, and 5 km.
Find: (U, V) = ? (m s^{-1}) for each point.

Hint: The hodograph has the U velocities labeled at both top and bottom to make it easier to draw perfectly vertical lines through any data point. For the same reason, V velocities are labeled at both left and right.

Method: Draw horizontal and vertical lines through each data point, and then pick off the U and V values by eye. Results:

For z = 1 km: **U = –11** m s^{-1}, **V = 11** m s^{-1}
For z = 3 km: **U = 0** m s^{-1}, **V = 25** m s^{-1}
For z = 5 km: **U = 25** m s^{-1}, **V = 25** m s^{-1}

Check: Signs and magnitudes are consistent with directions and speeds.
Exposition: Be careful to use the proper units. The legend in the Fig. 14.52 caption tells us the units for wind speed.

CAUTION: Different agencies use different units (m s^{-1}, knots, miles·h^{-1}, km·h^{-1}) and sometimes forget to state the units in the hodograph legend or caption.

Notice that the compass angles labeled on the blank hodograph appear to be backwards. This is not a mistake. The reason is that winds are specified by where they come from. For example, the wind at 1 km altitude is from the southeast; namely, from a compass direction of 135° (Fig. 14.50). The south wind at 3 km is from 180°, and the southwest wind at 5 km is from 235°. Using the "from" direction and the wind speed to specify points, the result is a hodograph that implies wind vectors pointing from the origin to the correct directions in physical space. As another example, a wind from the west (direction 270° on the hodograph) gives a vector pointing from the origin toward the right (which is toward the east in physical space). We define physical space in this context as having north at the top of the page, east at the right, south at the bottom, and west at the left.

Often you need to use the U and V wind components to determine thunderstorm characteristics. You could calculate these using eqs. (1.3) and (1.4) from Chapter 1. Alternately, you can pick them off from the coordinates of the hodograph. As demonstrated in Fig. 14.52, for any hodograph point, use a straight-edge to draw a horizontal line from that point to the ordinate to determine V, and draw a vertical line from the point to the abscissa to get U.

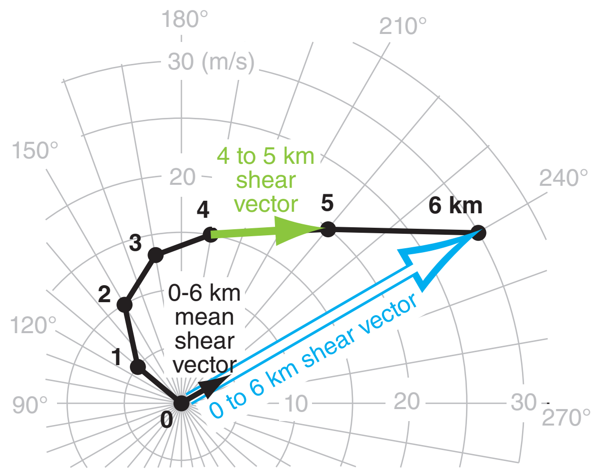


Figure 14.53
Example of mean shear vector between 0 to 6 km, shown by the small black arrow. There are six 1 km layers in this example. Vector wind differences are used as surrogates for shear.

Sample Application
Using data from the previous Sample Application, mathematically get the mean shear vector coordinates $(\Delta U, \Delta V)_{ms}$, magnitude ($\Delta M_{ms}$), and direction ($\alpha_{ms}$).

Find the Answer
Given: Wind vectors at 6 and 0 km altitude:
$M = 30$ m s^{-1} and $\alpha = 240°$ at $z = 6$ km,
$M = 0$ m s^{-1} and $\alpha = 0°$ at $z = 0$ km.
Let: subscript "*ms*" = "mean shear"
Find: $(\Delta U, \Delta V)_{ms} = ?$ (m s^{-1}),
$\Delta M_{ms} = ?$ (m s^{-1}), $\alpha_{ms} = ?$ (°)

First, find (U, V) components from eqs. (1.3) & (1.4):
$U_{6km} = -M \cdot \sin(\alpha) = -(30\text{m s}^{-1}) \cdot \sin(240°) = 26$ m s^{-1}
$V_{6km} = -M \cdot \cos(\alpha) = -(30\text{m s}^{-1}) \cdot \cos(240°) = 15$ m s^{-1}
and
$U_{0km} = -M \cdot \sin(\alpha) = -(0\text{m s}^{-1}) \cdot \sin(0°) = 0$ m s^{-1}
$V_{0km} = -M \cdot \cos(\alpha) = -(0\text{m s}^{-1}) \cdot \cos(0°) = 0$ m s^{-1}

Next, use eqs. (14.13 & 14.14):
$\Delta U_{ms} = (26 - 0 \text{ m s}^{-1})/6 = \underline{\mathbf{4.3}}$ m s^{-1}
$\Delta V_{ms} = (15 - 0 \text{ m s}^{-1})/6 = \underline{\mathbf{2.5}}$ m s^{-1}

Then, use $(\Delta U, \Delta V)$ in eq. (14.15):
$\Delta M_{ms} = [(\Delta U_{ms})^2 + (\Delta V_{ms})^2]^{1/2}$
$= [(4.3 \text{ m s}^{-1})^2 + (2.5 \text{ m s}^{-1})^2]^{1/2}$
$= [24.74 \text{ (m s}^{-1})^2]^{1/2} = \underline{\mathbf{5}}$ m s^{-1}

Finally, use $(\Delta U, \Delta V)$ in eq. (14.16):
$\alpha_{ms} = 90° - \arctan(\Delta V_{ms}/\Delta U_{ms}) + 180°$
$= 90° - \arctan(2.5/4.3) + 180°$
$= 90° - 30° + 180° = \underline{\mathbf{240°}}$

Check: Units OK. Agrees with Fig. 14.53 black arrow.
Exposition: The only reason the mean shear direction equaled the wind direction at 6 km was because the surface wind was zero. Normally they differ.

14.5.2. Using Hodographs

Wind and wind-shear between the surface and 6 km altitude affects the dynamics, evolution, and motion of many N. American thunderstorms. You can use a hodograph to determine key wind-related quantities, including the:
- local shear across a single layer of air
- mean wind-shear vector
- total shear magnitude
- mean wind vector (normal storm motion)
- right & left moving supercell motions
- storm-relative winds.

14.5.2.1. Shear Across a Single Layer

Local shear between winds at adjacent wind-reporting altitudes is easy to find using a hodograph. Comparing Figs. 14.46 with 14.50, you can see that the hodograph line segment between any two adjacent altitudes is equal to the local shear across that one layer. For example, the solid grey arrow in Fig. 14.53 (one of the line segments of the hodograph) shows the local shear across the 4 to 5 km layer.

14.5.2.2. Mean Wind Shear Vector

The **mean wind-shear vector** across multiple layers of air of equal thickness is the vector sum of the local shear vectors (i.e., all the hodograph line segments for those layers), divided by the number of layers spanned. Namely, it is a vector drawn from the bottom-altitude wind point to the top-altitude wind point (white arrow with grey border in Fig. 14.53), then divided by the number of layers to give the mean (solid black arrow). For Fig. 14.53, the mean 0 - 6 km shear vector is 5 m s^{-1} from the west southwest. This graphical method is easy to use.

To find the same mean wind-shear vector mathematically, use the U and V components for the wind at $z = 0$, and also for the wind at $z = 6$ km. The mean shear (subscript $_{ms}$) vector coordinates $(\Delta U, \Delta V)_{ms}$ for the 0 to 6 km layer is given by:

$$\Delta U_{ms} = (U_{6km} - U_{0km})/6 \qquad \bullet(14.13)$$

$$\Delta V_{ms} = (V_{6km} - V_{0km})/6 \qquad \bullet(14.14)$$

You can use these coordinates to plot the mean shear vector on a hodograph. You can use eqs. (14.15) and (14.16) to determine the magnitude and direction, respectively, of the mean shear vector. Namely:

$$\Delta M_{ms} = \left(\Delta U_{ms}{}^2 + \Delta V_{ms}{}^2\right)^{1/2} \qquad (14.15)$$

$$\alpha_{ms} = 90° - \frac{360°}{C} \cdot \arctan\left(\Delta V_{ms} / \Delta U_{ms}\right) + \alpha_o \qquad (14.16)$$

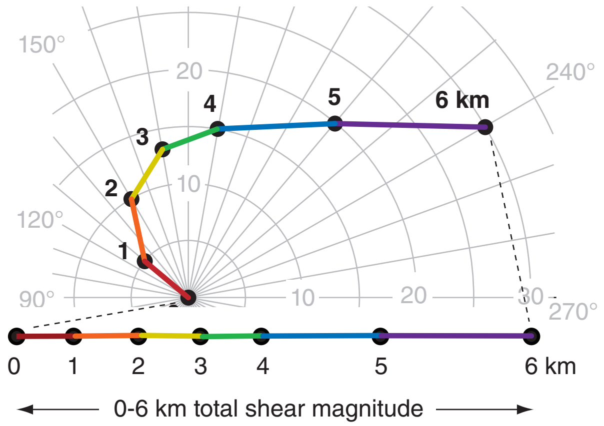


Figure 14.54
Example of total shear magnitude (rainbow-colored line), found as the sum of the shear magnitudes from the individual layers (colored line segments) between 0 to 6 km altitude. Vector wind differences are used as surrogates for shear.

where C = 360° or 2π radians (depending on the units returned by your calculator or spreadsheet), and α_o = 180° if $\Delta U_{ms} > 0$, but is zero otherwise. As before, these wind-difference values are surrogates for the true shear.

14.5.2.3. Total Shear Magnitude

The **total shear magnitude** (TSM) across many layers, such as between 0 to 6 km, is the algebraic sum of the lengths of the individual line segments in the hodograph. For a simple graphic method that gives a good estimate of TSM, conceptually lay flat the grey line from Fig. 14.54 without stretching or shrinking any of the line segments. The length of the resulting straight line (black line in Fig. 14.54) indicates the total shear magnitude. For this example, the total shear magnitude is roughly 45 m s^{-1}, comparing the length of the black line to the wind-speed circles in the graph. Again, wind difference is used as a surrogate for shear.

To find the total shear magnitude mathematically, sum over all the individual layer shear magnitudes (assuming layers of equal thickness):

$$TSM = \sum_{i=1}^{n} \left[\left(U_i - U_{i-1}\right)^2 + \left(V_i - V_{i-1}\right)^2 \right]^{1/2} \quad \bullet(14.17)$$

where n is the total number of layers, and subscripts i and i–1 indicate the top and bottom, respectively, of the i^{th} layer. For the Fig. 14.54 example, n = 6, and each layer is 1 km thick.

Although two different environments might have the same total wind shear (TSM), the distribution of that shear with height determines the types of storms possible. For example, Fig. 14.55 shows two hodographs with exactly the same total shear

Sample Application(§)

Mathematically calculate the total shear magnitude for the hodograph of Fig. 14.53 (i.e., using data table from the Sample Application 4 pages ago).

Find the Answer

Given: The data table from 4 pages ago, copied below.
Find: TSM = ? m s^{-1}

Use a spreadsheet to do these tedious calculations: First, use eqs. (1.3) and (1.4) to find U and V components of the winds. Next, find the wind differences $\Delta U = U_i - U_{i-1}$, and $\Delta V = V_i - V_{i-1}$. Finally, use these in eq. (14.17).

All speed and surrogate-shear units below are (m s^{-1}) except for direction α (degrees) and height z (km).

z	α	M	U	V	ΔU	ΔV	Shear
6	240	30	26.0	15.0	13.1	-0.3	13.1
5	220	20	12.9	15.3	10.3	0.5	10.3
4	190	15	2.6	14.8	4.9	2.0	5.2
3	170	13	-2.3	12.8	2.7	4.1	5.0
2	150	10	-5.0	8.7	-1.2	5.4	5.6
1	130	5	-3.8	3.2	-3.8	3.2	5.0
0	0	0	0.0	0.0	**Sum=TSM=**		**44.2**

Check: Units OK.
Exposition: The graphical estimate of 45 m s^{-1} by eye from Figure 14.54 was very close to the calculated value here.

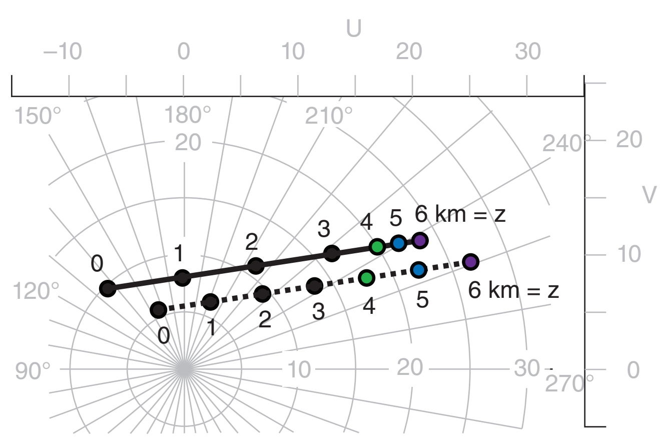


Figure 14.55
Two lines on this hodograph have exactly the same total length which means they have the same total shear magnitude. But that shear distributed differently with depth z for each line.

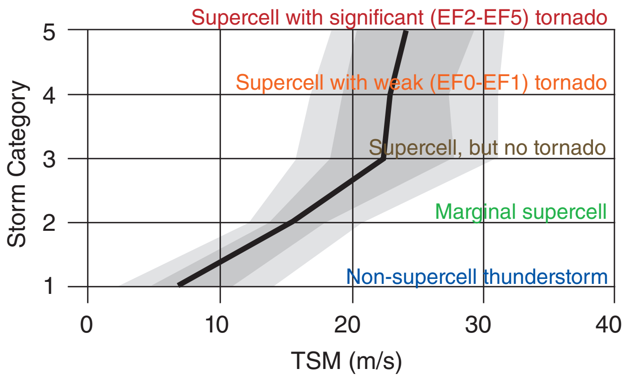


Figure 14.56
Observation frequency of thunderstorms of various intensities vs. environmental total shear magnitude (TSM) in the 0 to 6 km layer of atmosphere. Black line is the median (50th percentile) of several hundred observations in central N. America; dark grey shading spans 25th to 75th percentiles (the interquartile range); and light grey spans 10th to 90th percentiles.

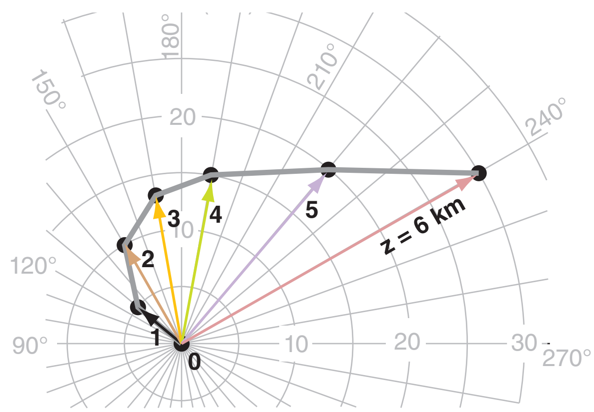


Figure 14.57
Wind vectors (colored) associated with points on a hodograph.

Sample Application
How many fence posts are needed to support a straight fence with 6 sections?

Find the Answer: 7 (see figure)

Exposition: Only 6 posts are needed for a closed loop.

magnitude, but the solid line has most of that shear in the bottom 3 km (as seen by the spacing between height (z) points along the hodograph), favoring **squall lines** and **bow echoes**. The dashed line has evenly distributed shear through the bottom 6 km, favoring **supercells**. Fig. 14.56 also shows that greater TSM values support supercells.

14.5.2.4. Mean Environmental Wind (Normal Storm Motion)

Airmass thunderstorms often translate in the direction and speed of the **mean environmental wind** vector in the bottom 6 km of the atmosphere. These are sometimes called **steering-level winds** or **normal winds**. To find the 0 to 6 km **mean wind** (i.e., NOT the shear) first picture the wind vectors associated with each point on the hodograph (Fig. 14.57). Next, vector sum all these winds by moving them tail to head, as shown in Fig. 14.58).

Finally, divide the vector sum by the number of wind points to find the **mean wind vector**, as indicated with the "X". [CAUTION: As you can see in Fig. 14.58, the 6 layers of air are bounded by 7 wind-vector points, not forgetting the zero wind at the ground. Thus, the vector sum in Fig. 14.58 must be divided by 7, not by 6.] For this example, the mean wind (i.e., the location of the "X") is from about 203° at 11 m s^{-1}. An arrow from the origin to the "X" gives the forecast **normal motion for thunderstorms**.

An easier way to approximate this mean wind vector without doing a vector addition is to estimate (by eye) the center of mass of the area enclosed by the original hodograph (shaded black in Fig. 14.59;

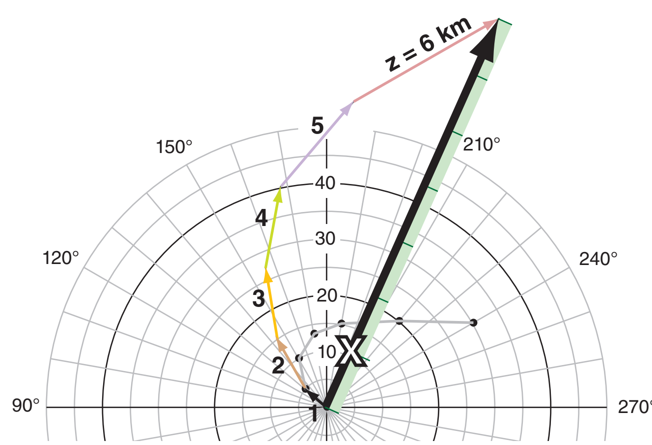


Figure 14.58
Individual wind vectors (thin colored arrows), added as a vector sum (thick black arrow), divided by the number of wind vectors (7 in this example, not forgetting the zero wind at the ground), gives the mean wind (X) in the 0 to 6 km layer of air. But a much easier way to find the "X" is shown in the next figure.

NOT the area enclosed by the re-positioned vectors of Fig. 14.58). Namely, if you were to: (1) trace that area onto a piece of cardboard, (2) cut it out, & (3) balance it on your finger tip, then the balance point (shown by the "X" in Fig. 14.59) marks the center of mass. "X" indicates the normal motion of thunderstorms, and is easy to estimate graphically using this center-of-mass estimate.

To mathematically calculate the **mean wind** (indicated with overbars) in the bottom 6 km of the atmosphere, first convert the vectors from speed and direction to their Cartesian components. Then sum the U velocities at all equally-spaced heights, and separately sum all the V velocities, and divide those sums by the number of velocity heights to get the components ($\overline{U},\overline{V}$) of the mean wind.

$$\overline{U} = \frac{1}{N}\sum_{j=0}^{N} U_j \qquad \bullet(14.18)$$

$$\overline{V} = \frac{1}{N}\sum_{j=0}^{N} V_j \qquad \bullet(14.19)$$

where N is the number of wind levels (not the number of shear layers), and j is the altitude index. For winds every 1 km from z = 0 to z = 6 km, then N = 7. [CAUTION: Don't confuse the mean wind given above with the mean shear of eqs. (14.13 & 14.14).] Then, use eqs. (1.1) and (1.2) [similar to eqs. (14.15 & 14.16), but for mean wind instead of mean shear] to convert from Cartesian coordinates to mean speed and direction ($\overline{M},\overline{\alpha}$).

In summary the normal motion of thunderstorms is ($\overline{U},\overline{V}$), or equivalently ($\overline{M},\overline{\alpha}$). This corresponds to the center-of-mass "X" estimate of Fig. 14.59.

A SCIENTIFIC PERSPECTIVE • Be Safe
(part 2)

Charles Doswell's web page recommends storm-chase guidelines, which I paraphrase here:

The #1 Threat: Being on the Highways
1. Avoid chasing alone.
2. Be very alert to standing water on the roads.
3. Avoid chasing in cities if at all possible.
4. Don't speed.
5. Pull fully off the road when you park.
6. Use your turn signals.
7. Slow down in poor visibility (rain; blowing dust).
8. Plan where to get fuel; don't let your tank get low.
9. Avoid unpaved roads (very slippery when wet).
10. Make your vehicle visible to other vehicles.

(continues in the next chapter).

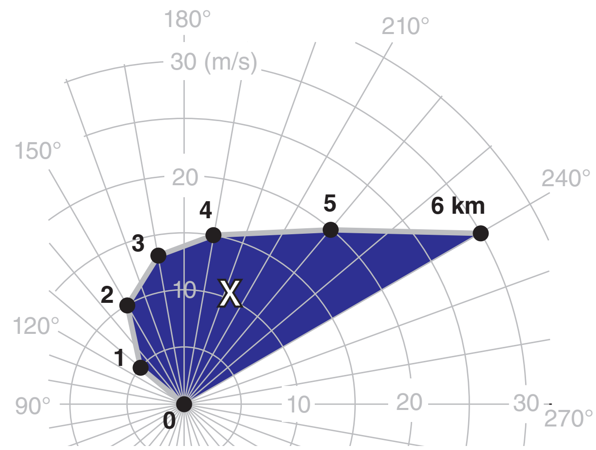


Figure 14.59
Mean wind ("X") in the 0 to 6 km layer of air, estimated as a center of mass of the area (dark blue) enclosed by the hodograph. "X" indicates the normal motion of thunderstorms.

Sample Application
Mathematically calculate the mean wind components, magnitude, and direction for the hodograph of Fig. 14.57 (i.e., using data table from the Sample Application about 5 pages ago).

Find the Answer:
Given: The data table from 5 pages ago, copied below.
Find: ($\overline{U},\overline{V}$) = ? (m s^{-1}), ($\overline{M},\overline{\alpha}$) = ? (m s^{-1} , °)

Use a spreadsheet to do these tedious calculations: First, use eqs. (1.3) and (1.4) to find U and V components of the winds. Next, average the U's, and then the V's, using eqs. (14.18 & 14.19). Finally, use eqs. (1.1) and (1.2) to covert to speed and direction.

All velocities below are (m s^{-1}), direction is in degrees, and height z in km.

z	direction	M	U	V
6	240	30	26.0	15.0
5	220	20	12.9	15.3
4	190	15	2.6	14.8
3	170	13	-2.3	12.8
2	150	10	-5.0	8.7
1	130	5	-3.8	3.2
0	0	0	0.0	0.0
Mean wind components (ms^{-1})=			**4.3**	**10.0**
Mean wind magnitude (m s^{-1}) =			**10.9**	
Mean wind direction (°) =			**203.5**	

Check: Units OK. Magnitudes and signs reasonable.
Exposition: The graphical estimate of mean wind speed 11 m s^{-1} and direction 203° by eye from Figs. 14.58 and 14.59 were very close to the calculated values here. Namely, expect normal thunderstorms to move from the south southwest at about 10.9 m s^{-1} in this environment. Thus, the center-of-mass estimate by eye saves a lot of time and gives a reasonable answer.

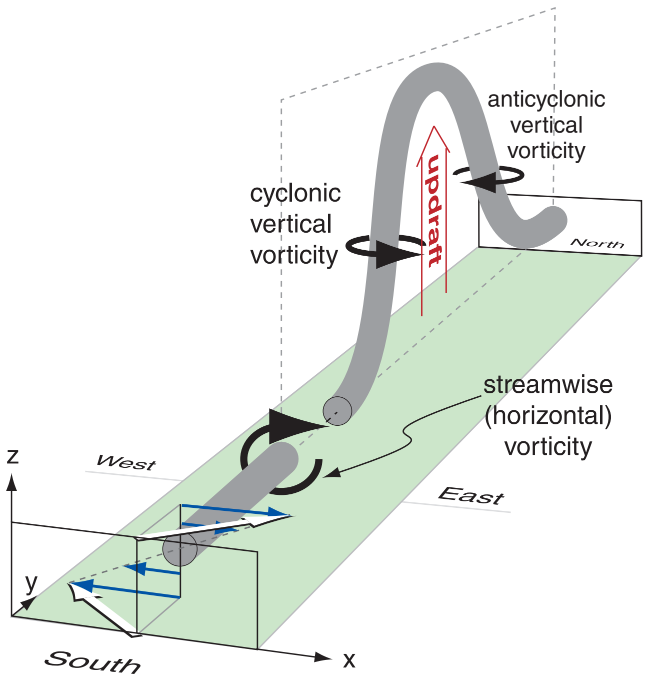


Figure 14.60
Counter-rotating mesocyclones formed when a convective updraft tilts the horizontal vorticity associated with environmental wind shear. Near-surface environmental winds from the southeast, and winds at 6 km altitude from the southwest, are white arrows. Dark cylinders represent vorticity axis.

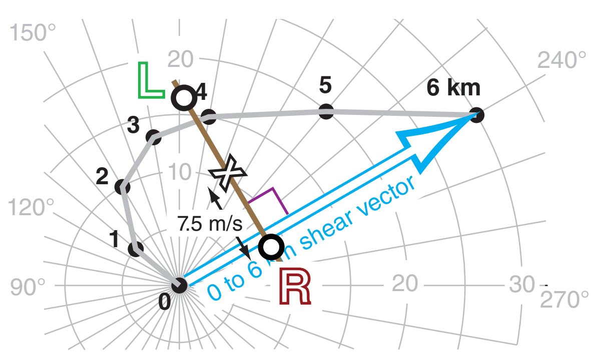


Figure 14.61
Bunker's internal dynamics technique for finding the movement of right-moving (R) and left-moving (L) supercell thunderstorms. (See text body for step-by-step instructions.)

14.5.2.5. Supercell Storm Motion

Supercell thunderstorms have rotating updrafts, and to get that rotation they need sufficient environmental wind shear. The environmental total shear magnitude typically must be TSM ≥ 25 m s^{-1} and be distributed relatively evenly across the lower troposphere (from the surface to 6 km) to support supercells. Total shear magnitudes less than 15 m s^{-1} often are too small for supercells. In between those two shears, thunderstorms may or may not evolve into supercells.

One reason that strong shear promotes supercells is that horizontal vorticity associated with vertical shear of the horizontal wind in the environment can be tilted into vertical vorticity by the strong convective updraft of a thunderstorm (Fig. 14.60). This causes counter-rotating vortices (mesocyclones) on the left and right sides of the updraft. These two mesocyclones are deep (fill the troposphere) and have diameters roughly equal to the tropospheric depth. One is cyclonic; the other is anticyclonic.

Sometimes supercell thunderstorms split into two separate storms: **right-moving** and **left-moving supercells**. Namely, the cyclonic and anticyclonic mesocyclones support their own supercells, and move right and left of the "normal" motion as would have been expected from the mean steering-level wind.

Bunker's **Internal Dynamics (ID)** method to forecast the movement of these right and left-moving supercells is:

(1) Find the 0 to 6 km mean wind (see "X" in Fig. 14.61), using methods already described. This would give normal motion for thunderstorms. But supercells are not normal.
(2) Find the 0.25 to 5.75 km layer shear vector (approximated by the 0 to 6 km layer shear, shown with the white arrow in Fig. 14.61).
(3) Through the mean wind point (X), draw a line (black line in Fig. 14.61) perpendicular to the 0 to 6 km shear vector.
(4) On the line, mark 2 points (white in black circles): one 7.5 m s^{-1} to the right and the other 7.5 m s^{-1} to the left of the "X".
(5) The point that is right (clockwise) from the mean wind "X" gives the average motion forecast for the right (R) moving storm.
(6) The point that is left (counter-clockwise) from the mean wind "X" gives the average forecast motion of the left (L) moving storm.

Not all storms follow this simple rule, but most supercell motions are within a 5 m s^{-1} radius of error around the R and L points indicated here. For the example in Fig. 14.61, motion of the right-moving supercell is 8 m s^{-1} from 246°, and the left mover is 17 m s^{-1} from 182°.

Only for straight hodographs (Fig. 14.62a) are the right- and left-moving supercells expected to exist simultaneously with roughly equal strength (Fig. 14.63a). Hodographs that curve clockwise (Fig. 14.62b) with increasing height favor right-moving supercell thunderstorms (Fig. 14.63b). For this situation, the left-moving storm often dissipates, leaving only a right-moving supercell.

Hodographs that curve counterclockwise (Fig. 14.62c) favor left-moving supercells (Fig. 14.63c), with the right-moving storm often dissipating. In the plains/prairies of North America, most hodographs curve clockwise (as in Figs. 14.61 and 14.62b), resulting in right-moving supercell thunderstorms that are about ten times more abundant than left-moving storms. The opposite is true in the S. Hemisphere for midlatitude thunderstorms.

In summary, supercell thunderstorms often do **not** move in the same direction as the mean environmental (steering-level) wind, and do not move at the mean wind speed. The methods that were discussed here are useful to forecast movement of storms that haven't formed yet. However, once storms form, you can more accurately estimate their motion by tracking them on radar or satellite. Regardless of the amounts of shear, CAPE, and moisture in the pre-storm environment, thunderstorms won't form unless there is a trigger mechanism. Triggers are discussed in the next main section (after the Bulk Richardson Number subsection).

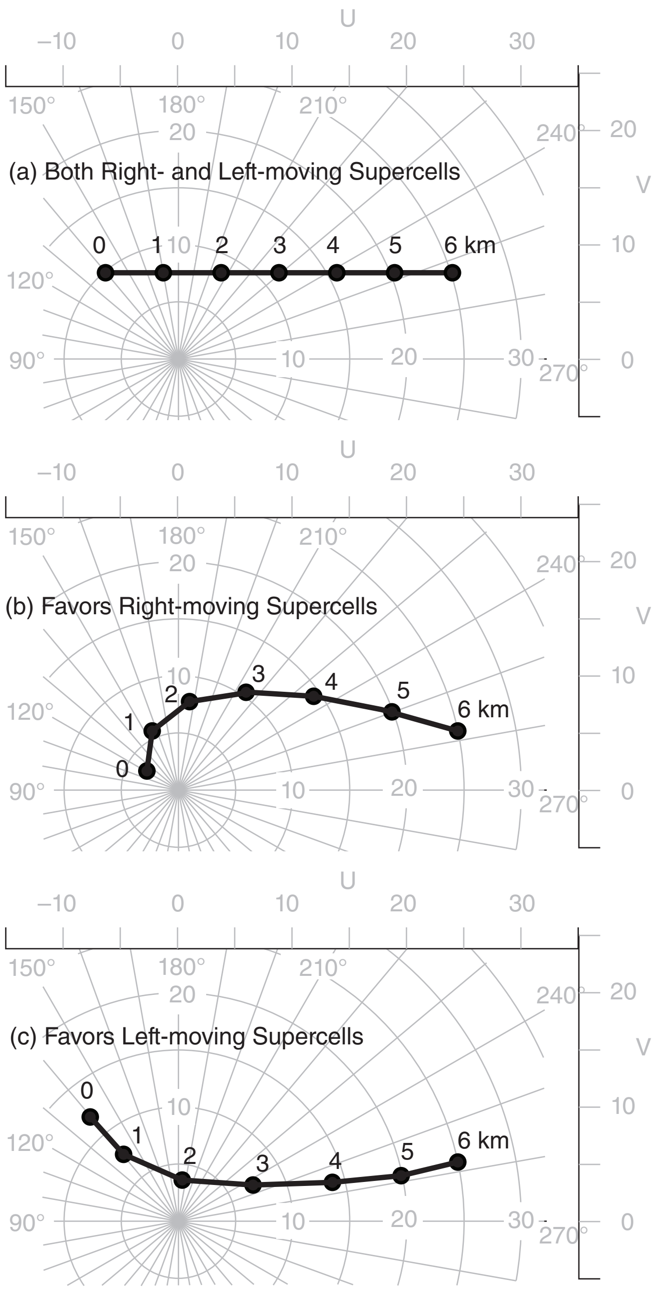


Figure 14.62
Shapes of hodograph that favor right- and left-moving supercell thunderstorms: (a) straight hodograph; (b) hodograph showing clockwise curvature with increasing height; (c) hodograph showing counter-clockwise curvature with increasing height.

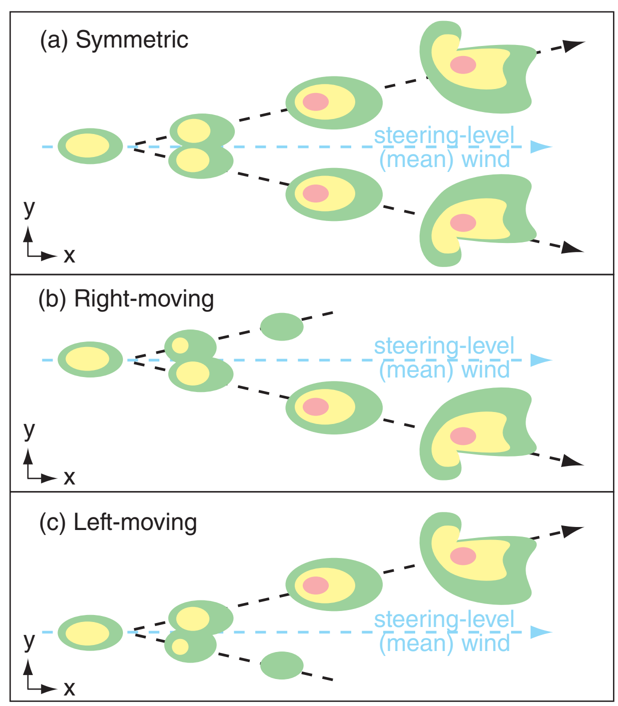


Figure 14.63
Sketch of sequence of radar reflectivity echoes over time (left to right) for: (a) symmetric supercell split; (b) dominant right-moving supercells; and (c) dominant left-moving supercells. Figures (a) to (c) above correspond to figures 14.62 (a) - (c). Note the different orientation of the hook echo in the left-moving supercell. The steering-level wind is the "normal" wind.

Sample Application

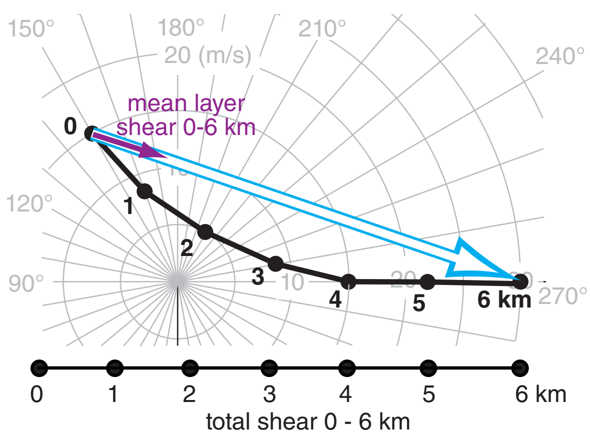
Given the wind profile in the table below:

z (km)	dir. (°)	speed (ms^{-1})	z (km)	dir. (°)	speed (m s^{-1})
0	150	15	4	270	15
1	160	8	5	270	22
2	210	5	6	270	30
3	260	8			

(a) Plot the hodograph;
(b) Graphically find the mean wind-shear vector for the 0 to 6 km layer;
(c) Graphically find the total shear magnitude.
(d) Does this hodograph favor bow echoes or supercells? Why?
(e) Graphically find the 0 to 6 km mean wind.
(f) Graphically find the motions for the right and left-moving supercell thunderstorms.
(g) Which is favored: right or left-moving supercell thunderstorms? Why?

Find the Answer

(a) Plot the hodograph (see curved black line below):



(b) The 0 to 6 km shear vector is shown with the white arrow. Dividing by 6 gives a mean layer shear (red arrow) of about **7 m s^{-1} from the west northwest**.

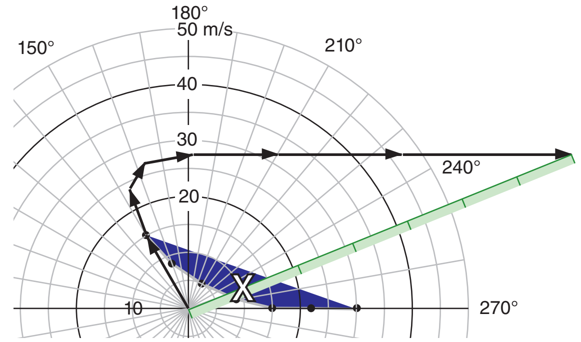
(c) The total shear magnitude, shown at the bottom of the hodograph above, is about **42 m s^{-1}**.

(d) **Supercells** are favored, because shear is evenly distributed throughout the bottom 6 km of atmosphere. Also, the total shear magnitude is greater than 25 m s^{-1}, as required for supercells.

(e) The answer is a vector, drawn from the origin to the "X" (see hodograph in next column). Namely, the mean wind is from about **250° at 10 m s^{-1}**. Vector method roughly agrees with center-of-mass method, done by eye.

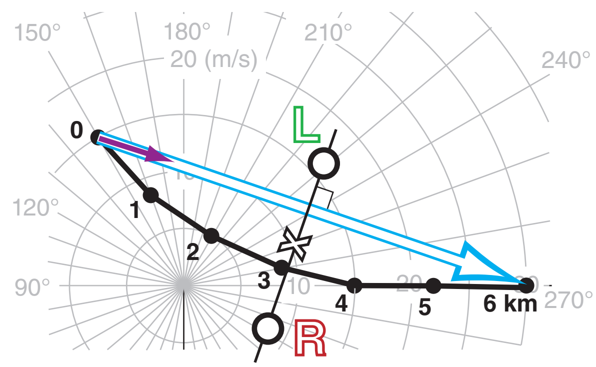
(Continues in next column.)

Sample Application *(continuation)*



Thus, for normal thunderstorms (and for the first supercell, before it splits into right and left-moving supercells) forecast them to move from the west southwest at about 10 m s^{-1}. For both this example and Fig. 14.59, the center-of-mass method gives a better estimate if we look at the hodograph area between z = 0.5 and 5.5 km.

(f) Bunker's Internal Dynamics method is shown graphically on the hodograph below. The left supercell (L) is forecast to move at **16** m s^{-1} **from 229°**, and the right supercell (R) at **8** m s^{-1} **from 298°**.



(g) **Left-moving supercells** are favored, because the hodograph curves counterclockwise with increasing height. This storm moves faster than the mean wind (i.e., faster than the "normal" thunderstorm steering-level wind).

Exposition: This hodograph shows **veering winds** (winds turning clockwise with increasing altitude), implying warm-air advection according to the thermal-wind relationship. Even though the wind vectors veer clockwise, the hodograph curvature is counterclockwise (implying dominant left-moving supercell storms), so be careful not to confuse these two characteristics.

Winds alone are not sufficient to forecast whether thunderstorms will occur. Other key ingredients are instability and abundant moisture, as previously discussed. Also needed is sufficient lifting by a trigger mechanism to overcome the capping inversion.

14.5.2.6. Bulk Richardson Number

Thunderstorm type depends on the amounts of both instability and wind shear in the pre-storm environmental sounding. The **bulk Richardson number** (**BRN**) is the ratio of nonlocal instability (i.e., the CAPE) in the mid to upper part of the troposphere to shear (ΔM) in the lower half of the troposphere:

$$BRN = \frac{CAPE}{0.5\cdot(\Delta M)^2} \qquad \bullet(14.20)$$

Table 14-2 summarizes the utility of the BRN. The BRN is dimensionless, because the units of CAPE ($J\cdot kg^{-1}$) are the same as ($m^2\cdot s^{-2}$) (see Appendix A).

In the denominator, the shear magnitude (ΔM) is given by:

$$\Delta M = \left[(\Delta U)^2 + (\Delta V)^2\right]^{1/2} \qquad \bullet(14.21)$$

where

$$\Delta U = \overline{U} - U_{SL} \qquad (14.22a)$$

$$\Delta V = \overline{V} - V_{SL} \qquad (14.22b)$$

Eqs. (14.22) are the shear components between the mean wind ($\overline{U},\overline{V}$) in the 0 to 6 km layer, and the surface-layer winds (U_{SL}, V_{SL}) estimated as an average over the bottom 0.5 km of the atmosphere. Fig. 14.64 shows how to estimate this shear magnitude from a hodograph.

The denominator of the BRN, known as the **bulk-Richardson-number shear** (**BRN Shear**), is

$$\text{BRN Shear} = 0.5 \cdot (\Delta M)^2 \qquad \bullet(14.23)$$

which has units of ($m^2\cdot s^{-2}$). BRN shear can help indicate which supercells might be tornadic (see Table 14-3). Although BRN shear is statistically sharper than ML-CAPE, MU-CAPE and ML-LCL in its ability to discriminate between thunderstorms and tor-

Table 14-2. Thunderstorm type determination using the bulk Richardson number (BRN)

BRN	Thunderstorm Type
< 10	(unlikely to have severe thunderstorms)
10 - 45	Supercells
45 - 50	Supercells and/or multicells
> 50	Multicells

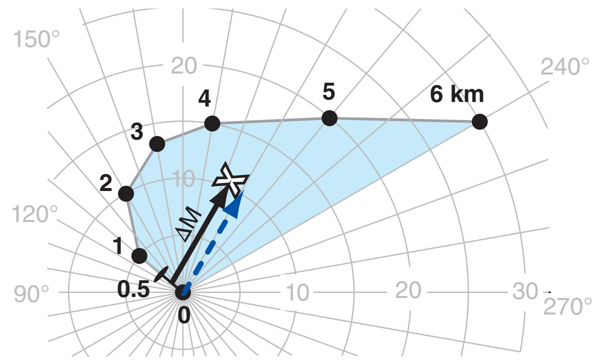
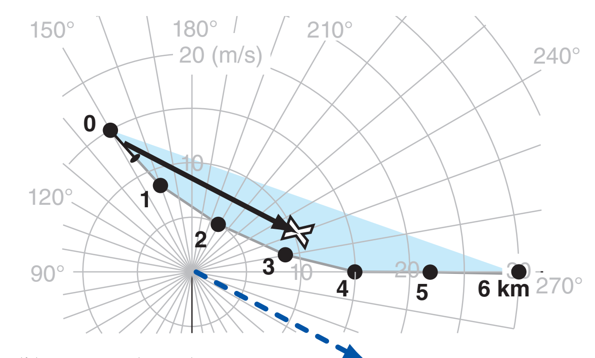


Figure 14.64
Solid black arrow shows shear ΔM between the average winds in the lowest 0.5 km of air, and the mean wind (X) in the lowest 6 km. Dashed arrow is just the solid arrow moved (without stretching) to the origin, so that the speed circles indicate the shear magnitude. $\Delta M = 11\ m\ s^{-1}$ in this example. ΔM is important in the bulk Richardson number (BRN), and BRN shear.

Sample Application
For the previous Sample Application, assume the associated CAPE = 3000 $J\cdot kg^{-1}$. (a) Find the BRN shear. (b) Find the BRN. (c) Are tornadic supercells likely?

Find the Answer
Given: CAPE = 3000 $J\cdot kg^{-1}$
hodograph = previous Sample Application
Find: BRN shear = ? $m^2\cdot s^{-2}$
BRN = ? (dimensionless)
Yes/no: thunderstorms? supercells? tornadoes?

(a) On the hodograph from the previous Sample Application, put the tail of a vector (solid black arrow) a quarter of the way from the z = 0 point to the z = 1 point on the hodograph). Put the arrowhead on the "X" mean wind. Then measure the vector's length (see dashed arrow), which is about $\Delta M = 18\ m\ s^{-1}$.
Then use eq. (14.23):
BRN Shear = $0.5\cdot(18\ m\ s^{-1})^2$ = <u>**162**</u> $m^2\ s^{-2}$



(b) Use eq. (14.20):
BRN = (3000 $m^2\ s^{-2}$) / (162 $m^2\ s^{-2}$) = <u>**18.5**</u>

(c) Assume CAPE ≈ MLCAPE. Then from Table 14-1:
<u>**Severe thunderstorms likely, possible tornado.**</u>
Next, use Table 14-2: Thunderstorm type = <u>**supercell**</u>.
Finally, use Table 14-3: <u>**Tornadoes are likely**</u>.

Table 14-3. If supercell thunderstorms form, then the bulk-Richardson-number shear (BRN Shear) suggests whether they might have tornadoes.

BRN Shear ($m^2 \cdot s^{-2}$)	Tornadic Supercells
25 - 35	less likely, but possible
35 - 45	likely
45 - 100	more likely

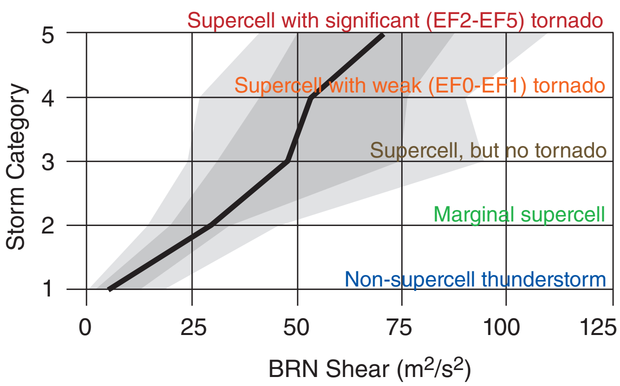


Figure 14.65
Statistics of bulk Richardson number shear (BRN shear) as a discriminator for thunderstorm and tornado intensity, based on several hundred storms in central N. America. Light grey spans 10th to 90th percentiles of the observations; dark grey spans the interquartile range of 25th to 75th percentiles; and the black line is the median (50th percentile).

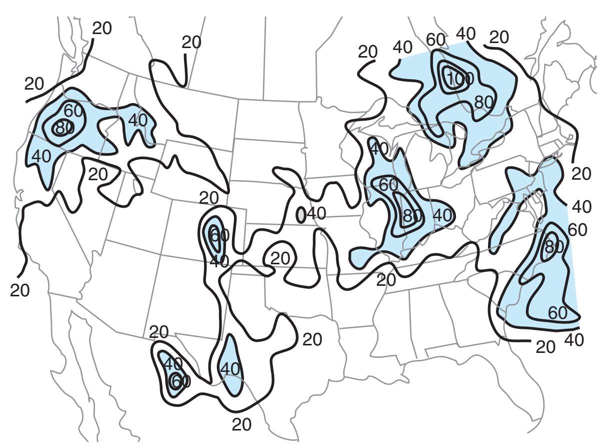


Figure 14.66
Weather map similar to Fig. 14.38, but for bulk Richardson number shear (BRN shear) in $m^2 \cdot s^{-2}$. Unlike the smooth variations in CAPE across N. America, the BRN shear values have a more local structure and are highly variable.

nadoes of different severity, there is still significant overlap in thunderstorm/tornado categories as was found by verification against several hundred thunderstorms in central N. America (Fig. 14.65).

A case study weather map of BRN shear is analyzed in Fig. 14.66. As before, you should NOT use BRN shear alone as an indicator of tornado likelihood, because additional conditions must also be satisfied in order to allow tornadic thunderstorms to form. Some of these other conditions are instability, and a trigger mechanism.

14.5.2.7. The Effective Layer

Most of the shear calculations shown so far have assumed that all thunderstorms are alike. Namely, all storms are roughly 12 km tall, and they all ingest the bottom 6 km of environmental air into the storm. That is why we have been plotting on the hodograph the 0 to 6 km winds. However, some storms are taller and some are shorter.

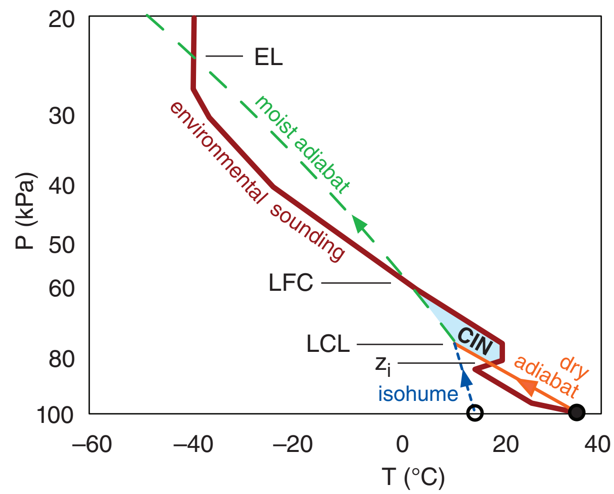
Also, for thunderstorms at night and early morning, a cold layer of nocturnal boundary-layer air might resist vertical motion because of its static stability, and thus not be ingested into the thunderstorm. These storms have **elevated inflows** with their **inflow base** z_{inflow_base} at some height above ground. But for most afternoon and evening thunderstorms, $z_{inflow\_base} = 0$.

An **effective thunderstorm depth** is defined as $\Delta z_{storm} = z_{EL} - z_{inflow\_base}$. This definition works for tall and short storms (with high and low equilibrium levels, z_{EL}, respectively), and for storms with zero and nonzero inflow-base altitude. An **effective-layer** bulk shear is then defined across the <u>bottom half</u> of the effective thunderstorm depth. This **effective bulk shear** would be used in place of the BRN shear previously discussed, to anticipate thunderstorm characteristics. Supercells are more likely when **effective bulk wind difference** is 12 to 20 $m \cdot s^{-1}$ or greater. The effective bulk shear is the same as the 0 to 6 km shear for average storms.

14.6. TRIGGERING VS. CONVECTIVE INHIBITION

The fourth environmental condition needed for thunderstorm formation is a **trigger** mechanism to cause the initial **lifting** of the air parcels. (Recall from section 14.2 that the other 3 conditions are high humidity in the boundary layer, instability, and wind shear.) Although the capping inversion is needed to allow the fuel supply to build up (a good thing for thunderstorms), this cap also inhibits thunderstorm formation (a bad thing). So the duty of the external trigger mechanism is to lift the reluc-

Figure 14.67 (at right)
Convective Inhibition (CIN) is proportional to the area (shaded light blue) where the rising air parcel is colder than the environment for an early afternoon pre-storm sounding.



tant air parcel from z_i to the LFC. Once triggered, thunderstorms develop their own circulations that continue to tap the boundary-layer air.

The amount of external forcing required to trigger a thunderstorm depends on the strength of the cap that opposes such triggering. One measure of cap strength is convective inhibition energy.

14.6.1. Convective Inhibition (CIN)

Between the mixed-layer top z_i and the level of free convection LFC, an air parcel lifted from the surface is colder than the environment. Therefore, it is negatively buoyant, and does not want to rise. For the air parcel to rise above z_i, the trigger process must do work against the buoyant forces within this cap region.

The total amount of work needed is proportional to the area shaded in Fig. 14.67. This work per unit mass is called the **Convective Inhibition** (**CIN**). The equation for CIN is identical to the equation for CAPE, except for the limits of the sum (i.e., z_i to LFC for CIN vs. LFC to EL for CAPE). Another difference is that CIN values are negative. CIN has units of J·kg^{-1}.

For the afternoon sounding of Fig. 14.67, CIN corresponds to the area between z_i and LFC, bounded by the dry and moist adiabats on the left and the environmental sounding on the right. For a morning sounding, a forecast of the afternoon's high temperature provides the starting point for the rising air parcel, where z_i is the corresponding forecast for mixed-layer top (Fig. 14.68). Thus, for both Figs. 14.67 and 14.68, you can use:

$$CIN = \sum_{z_i}^{LFC} \frac{|g|}{T_{ve}} (T_{vp} - T_{ve}) \cdot \Delta z \qquad \bullet(14.24)$$

where T_v is virtual temperature, subscripts p and e indicate the air parcel and the environment, $|g|$ = 9.8 m·s^{-2} is the magnitude of gravitational acceleration, and Δz is a height increment. Otherwise, CIN is usually found by summing between the surface (z = 0) and the LFC (Fig. 14.69).

$$CIN = \sum_{z=0}^{LFC} \frac{|g|}{T_{ve}} (T_{vp} - T_{ve}) \cdot \Delta z \qquad (14.25)$$

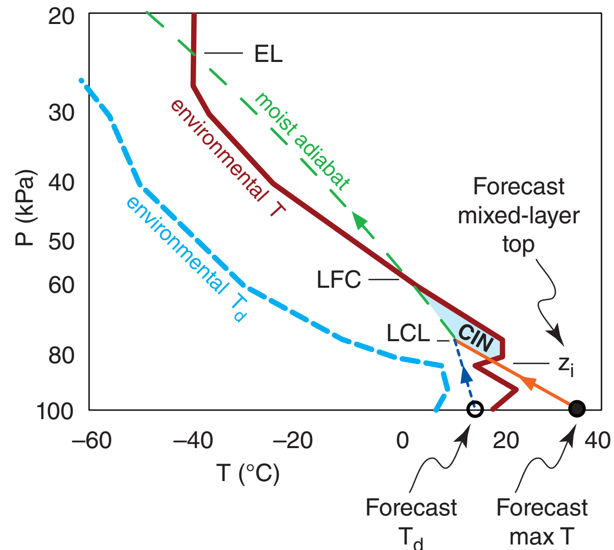


Figure 14.68
CIN based on max surface temperature and dew-point temperature forecasts and the early morning pre-storm sounding.

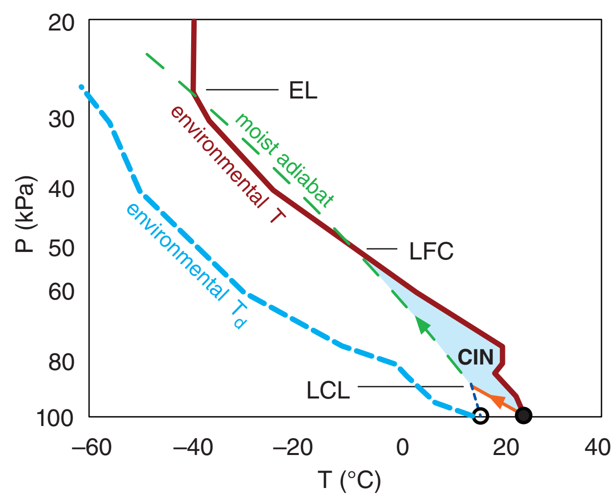


Figure 14.69
CIN between the surface and the LFC, for an early morning pre-storm sounding.

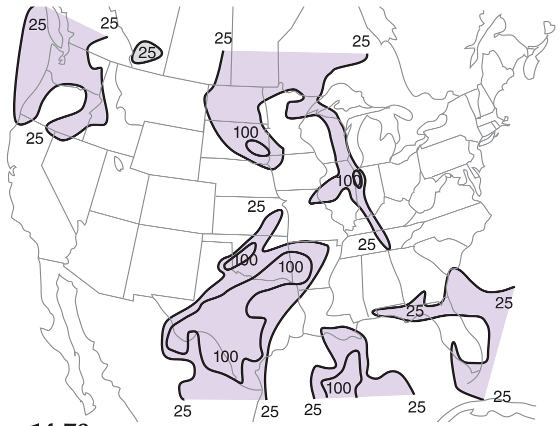


Figure 14.70
Weather map similar to Fig. 14.38 over USA, but for magnitude of mean-layer convective inhibition energy (ML CIN) in J·kg^{-1}.

Table 14-4. Convective Inhibition (CIN)

CIN Value (J·kg^{-1})	Interpretation
> 0	No cap. Allows weak convection.
0 to –20	Weak cap. Triggering easy. Air-mass thunderstorms possible.
–20 to –60	Moderate cap. Best conditions for CB. Enables fuel build-up in boundary layer. Allows most trigger mechanisms to initiate storm.
–60 to –100	Strong cap. Difficult to break. Need exceptionally strong trigger.
< –100	Intense cap. CB triggering unlikely.

Sample Application

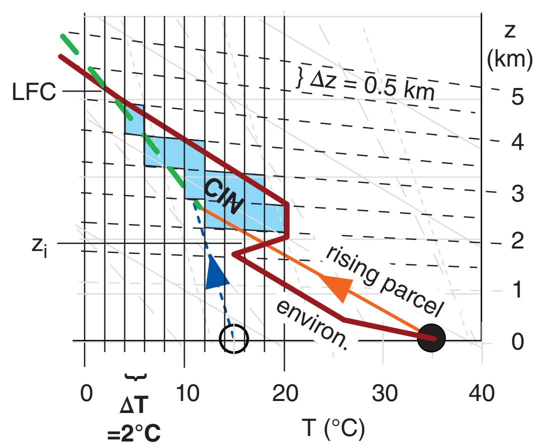
Estimate the convective inhibition (CIN) value for the sounding of Fig. 14.25. Are thunderstorms likely?

Find the Answer

Given: Sounding of Fig. 14.25, as zoomed below.
Find: CIN = ? J·kg^{-1}.

Cover the CIN area with small tiles of size Δz = 500 m and ΔT = –2°C (where the negative sign indicates the rising parcel is colder than the environmental sounding). Each tile has area = $\Delta T \cdot \Delta z$ = –1,000 K·m. I count 12 tiles, so the total area is –12,000. Also, by eye the average environmental temperature in the CIN region is about T_e = 12°C = 285 K.

Use eq. (14.26):
CIN = [(9.8 m s^{-2})/(285 K)]·(–12,000 K·m) = **–413 J·kg^{-1}**.
From Table 14-4, **thunderstorms are unlikely**.



Check: By eye, the CIN area of Fig. 14.25 is about 1/5 the size of the CAPE area. Indeed, the answer above has about 1/5 the magnitude of CAPE (1976 J·kg^{-1}), as found in previous solved exercises. Units OK too.

Exposition: The negative CIN implies that work must be done ON the air to force it to rise through this region, as opposed to the positive CAPE values which imply that work is done BY the rising air.

As before, if the virtual temperature is not known or is difficult to estimate, then often storm forecasters will use T in place of T_v in eq. (14.24) to get a rough estimate of CIN:

$$CIN = \sum_{z_i}^{LFC} \frac{|g|}{T_e} (T_p - T_e) \cdot \Delta z \qquad \bullet(14.26)$$

although this can cause errors as large as 35 J·kg^{-1} in the CIN value. Similar approximations can be made for eq. (14.25). Also, a **mean-layer CIN** (**ML CIN**) can be used, where the starting air parcel is based on average conditions in the bottom 1 km of air.

CIN is what prevents, delays, or inhibits formation of thunderstorms. Larger magnitudes of CIN (as in Fig. 14.69) are less likely to be overcome by trigger mechanisms, and are more effective at preventing thunderstorm formation. CIN magnitudes (Table 14-4) smaller than about 60 J·kg^{-1} are usually small enough to allow deep convection to form if triggered, but large enough to trap heat and humidity in the boundary layer prior to triggering, to serve as the thunderstorm fuel. CIN is the **cap** that must be broken to enable thunderstorm growth.

Storms that form in the presence of large CIN are less likely to spawn tornadoes. However, large CIN can sometimes be circumvented if an elevated trigger mechanism forces air-parcel ascent starting well above the surface.

The area on a thermo diagram representing CIN is easily integrated by computer, and is often reported with the plotted sounding. **[CAUTION: often the magnitude of CIN is reported (i.e., as a positive value).]** CIN magnitudes from many sounding stations can be plotted on a weather map and analyzed, as in Fig. 14.70. To estimate CIN by hand, you can use the same tiling method as was done for CAPE (see the Sample Application at left).

Thunderstorms are more likely for smaller values of Δz_{cap}, where $\Delta z_{cap} = z_{LFC} - z_{LCL}$ is the thickness of the nonlocally stable region at the bottom of the storm. A thinner cap on top of the ABL might allow thunderstorms to be more easily triggered.

14.6.2. Triggers

Any external process that forces boundary-layer air parcels to rise through the statically stable cap can be a trigger. Some triggers are:

- Boundaries between airmasses:
  - cold, warm, or occluded fronts,
  - gust fronts from other thunderstorms,
  - sea-breeze fronts,
  - dry lines.
- Other triggers:
  - the effects of mountains on winds,
  - small regions of high surface heating,
  - vertical oscillations called buoyancy waves.

If one airmass is denser than an adjacent airmass, then any convergence of air toward the dividing line between the airmasses (called an **airmass boundary**) will force the less-dense air to rise over the denser air. A cold front is a good example (Fig. 14.71), where cold, dense air advances under the less-dense warm air, causing the warm air to rise past z_i.

Surface-based **cold** and **warm fronts** are synoptically-forced examples of airmass boundaries. Upper-level synoptic fronts (with no signature on a surface weather map) can also cause lifting and trigger thunderstorms. Fig. 14.72 shows the frontal analysis for the 24 May 2006 case study that has been presented in many of the preceding sections. These frontal zones could trigger thunderstorms.

Sea-breeze (see the Regional Winds chapter) or **lake-breeze fronts** create a similar density discontinuity on the mesoscale at the boundary between the cool marine air and the warmer continental air during daytime, and can also trigger convection. If a cold **downburst** from a thunderstorm hits the ground and spreads out, the leading edge is a **gust front** of cooler, denser air that can trigger other thunderstorms (via the propagation mechanism already discussed). The **dry line** separating warm dry air from warm humid air is also an airmass boundary (Figs. 14.71 & 14.72), because dry air is more dense than humid air at the same temperature (see definition of virtual temperature in Chapter 1).

These airmass boundaries are so crucial for triggering thunderstorms that meteorologists devote much effort to identifying their presence. Lines of thunderstorms can form along them. Synoptic-scale fronts and dry lines are often evident from weather-map analyses and satellite images. Sea-breeze fronts and gust fronts can be seen by radar as con-

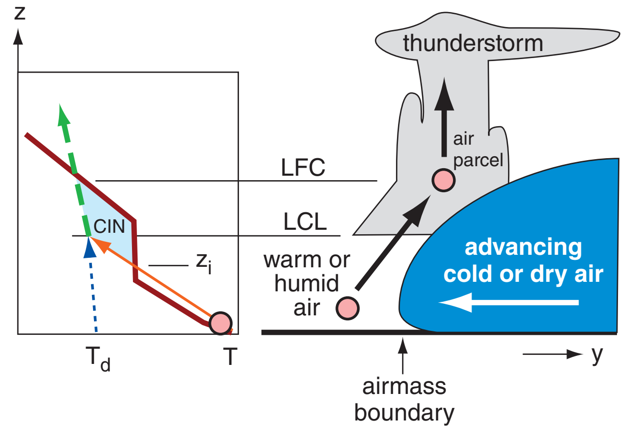


Figure 14.71
Warmer or moister (less-dense) air forced to rise over advancing colder or drier (more-dense) air can trigger thunderstorms if the air parcels forced to rise reach their LFC. The portion of thunderstorm clouds between the LCL and the LFC can have a smooth or laminar appearance. Above the LFC the clouds look much more convective, with rising turrets that look like cauliflower.

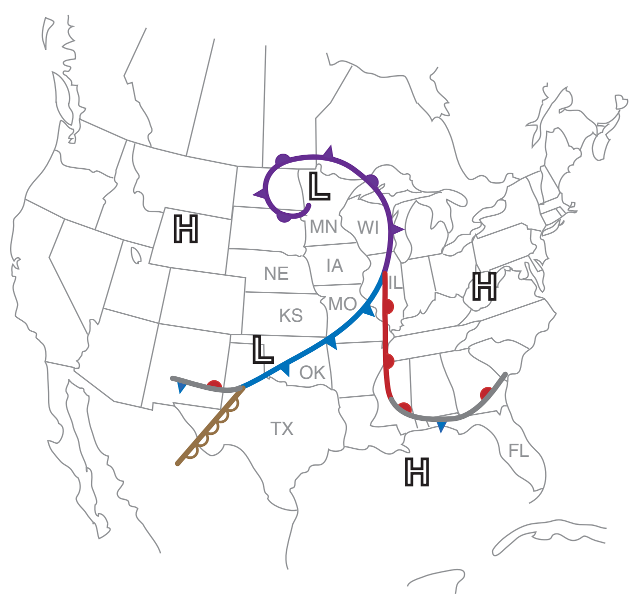


Figure 14.72
Synoptic frontal analysis over N. America at 22 UTC on 24 May 2006. An occluded front extends from a low (𝕃) in Minnesota (MN) into Illinois (IL), where the warm and cold fronts meet. The cold front extends from IL through Oklahoma (OK) and Texas (TX), and merges with a dry line in SW Texas. (See Chapter 13 section 13.1 "Info" for identification of US states.)

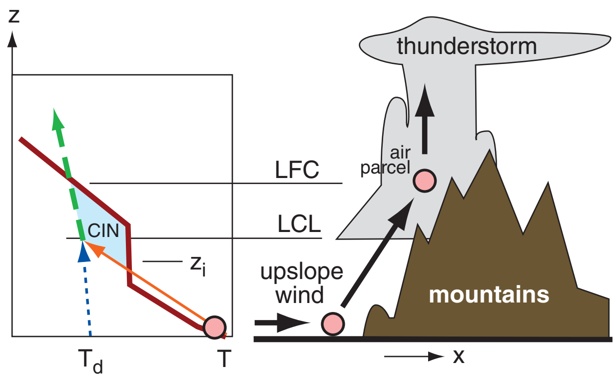


Figure 14.73
Wind hitting a mountain can be forced upslope, triggering thunderstorms called orographic thunderstorms.

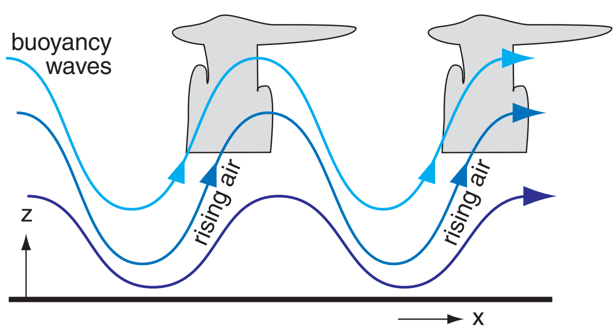


Figure 14.74
Thunderstorms generated in the updraft portions of buoyancy (gravity) waves.

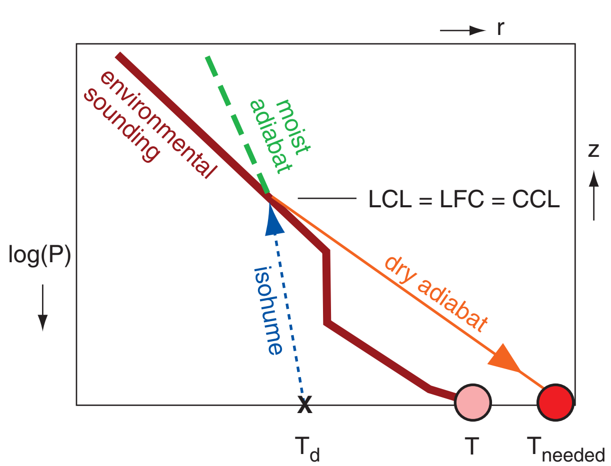


Figure 14.75
The convective temperature needed so that an air parcel has no CIN, and can reach its LFC under its own buoyancy. CCL = convective condensation level.

vergence lines in the Doppler wind field, and as lines of weakly enhanced reflectivity due to high concentrations of insects (converging there with the wind) or a line of small cumulus clouds. When two or more air-mass boundaries meet or cross, the chance of triggering thunderstorms is greater such as over Illinois and Texas in the case-study example of Fig. 14.72.

Horizontal winds hitting a mountain slope can be forced upward (Fig. 14.73), and can trigger **orographic thunderstorms**. Once triggered by the mountains, the storms can persist or propagate (by triggering daughter storms via their gust fronts) as they are blown away from the mountains by the steering-level winds.

Sometimes when the air flows over mountains or is pushed out of the way by an advancing cold front, vertical oscillations called **buoyancy waves** or **gravity waves** (Fig. 14.74) can be generated in the air (see the Regional Winds chapter). If the lift in the updraft portion of a wave is sufficient to bring air parcels to their LFC, then thunderstorms can be triggered by these waves.

Another trigger is possible if the air near the ground is warmed sufficiently by its contact with the sun-heated Earth. The air may become warm enough to rise as a thermal through the cap and reach the LFC due to its own buoyancy. The temperature needed for this to happen is called the **convective temperature**, and is found as follows:

(1) Plot the environmental sounding and the surface dew-point temperature (Fig. 14.75).
(2) From T_d, follow the isohume up until it hits the sounding. This altitude is called the **convective condensation level** (**CCL**).
(3) Follow the dry adiabat back to the surface to give the convective temperature needed.

You should forecast airmass thunderstorms to start when the surface air temperature is forecast to reach the convective temperature, assuming the environmental sounding above the boundary layer does not change much during the day. Namely, determine if the forecast high temperature for the day will exceed the convective temperature, and at what time this criterion will be met.

Often several triggering mechanisms work in tandem to create favored locations for thunderstorm development. For example, a jet streak may move across North America, producing upward motion under its left exit region and thereby reducing the strength of the cap as it passes (see the "Layer Method for Static Stability" section in the Atmospheric Stability chapter). If the jet streak crosses a region with large values of CAPE at the same time as the surface temperature reaches its maximum value, then the combination of these processes may lead to thunderstorm development.

Sample Application

Use the sounding in Fig. 14.25 to find the convective temperature and the convective condensation level. Are thunderstorms likely to be triggered?

Find the Answer

Given: Fig. 14.25.
Find: P_{CCL} = ? kPa, T_{needed} = ? °C , yes/no Tstorms

Step 1: Plot the sounding (see next column).

Step 2: From the surface dew-point temperature, follow the isohume up until it hits the sounding to find the CCL. Thus: P_{CCL} ≈ **63 kPa** (or z_{CCL} ≈ **3.7 km**).

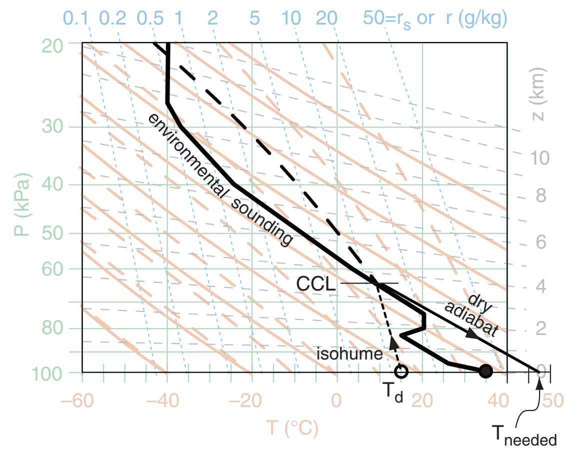
Step 2: Follow the dry adiabat down to the surface, and read the temperature from the thermo diagram.
T_{needed} ≈ **47 °C** is the convective temperature

Check: Values agree with figure.
Exposition: Typical high temperatures in the prairies of North America are unlikely to reach 47°C.

(continues in next column)

Sample Application *(continuation)*

Thus, the **triggering of a thunderstorm by a rising thermal is unlikely** for this case. However, thunderstorms could be triggered by an airmass boundary such as a cold front, if the other conditions are met for thunderstorm formation.



Finally, a word of caution. As with each of the other requirements for a thunderstorm, triggering (lift) by itself is insufficient to create thunderstorms. Also needed are the instability, abundant moisture, and wind shear. To forecast thunderstorms you must forecast all four of the key ingredients. That is why thunderstorm forecasting is so difficult.

14.7. THUNDERSTORM FORECASTING

Forecasting thunderstorms is not easy. Thunderstorm processes are very nonlinear (e.g., thunderstorms grow explosively), and are extremely dependent on initial conditions such as triggering, shear, and static stability. Individual storms are relatively short lived (15 to 30 min), and are constantly changing in intensity and movement during their lifetimes. Intense thunderstorms also modify their environment, making the relationships between pre-storm environments and storm evolution even more challenging to apply.

Much appears to depend on random chance, since we are unable to observe the atmosphere to the precision needed to describe it fully. Will a boundary-layer thermal or an airmass boundary happen to be strong enough to be the first to break through the capping inversion? Will a previous thunderstorm create a strong-enough downburst to trigger a new

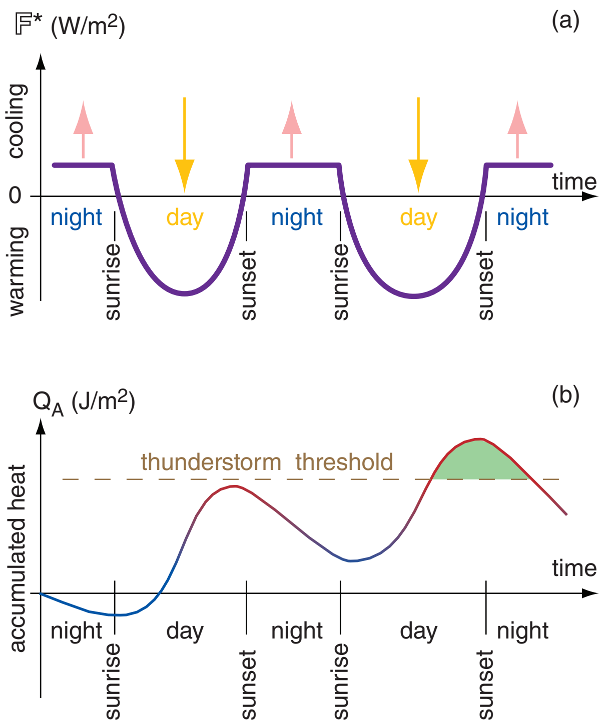


Figure 14.76
(a) Simplification of daily (diurnal) cycle of upward net radiation 𝔽, which combines solar heating during the day and infrared cooling day and night, for summer over land. Grey arrows represent direction of net flux at the surface. (b) Accumulated (sensible plus latent) heat Q_A in the boundary layer. During summer, days are longer than nights, allowing accumulated heat to increase from day to day. The dashed line thunderstorm threshold schematically represents the net effect of all the complex thunderstorm-genesis processes reviewed in this chapter. Green shading shows that thunderstorms are most likely to form around sunset, plus or minus several hours*

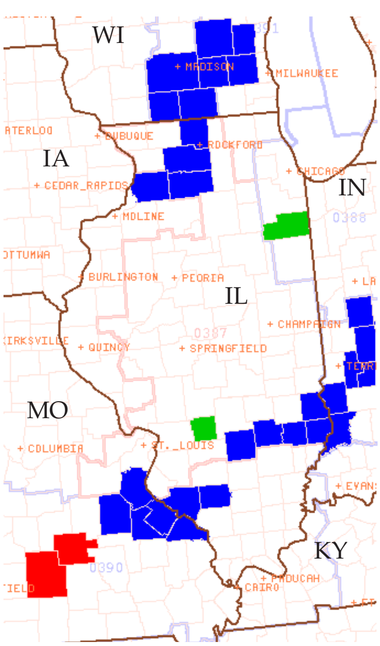


Figure 14.77
Map of tornado (red), flood (green), and severe thunderstorm (blue) ***warnings*** *associated with the 24 May 2006 case study in the USA. The colored boxes indicate which counties were warned. State abbreviations are: IL = Illinois, MO = Missouri, IA = Iowa, IN = Indiana, WI = Wisconsin, KY = Kentucky.*

thunderstorm with its gust front? Will high clouds happen to move over a region, shading the ground and thereby reducing the instability?

Tornadoes, hail, heavy rain, lightning, and strong straight-line wind events may or may not occur, and also vary in intensity, track, and duration within the constraints of the parent thunderstorm. Thunderstorms interact in complex ways with other existing thunderstorms and with the environment and terrain, and can trigger new thunderstorms.

Luckily, most thunderstorms over land have a very marked diurnal cycle, because atmospheric instability is strongly modulated by heating of the ground by the sun. Thunderstorms usually form in mid- to late-afternoon, are most frequent around sunset, and often dissipate during the night. As sketched in Fig. 14.76, the reason is that the greatest accumulation of heat (and moisture) in the boundary layer occurs not at noon, but about a half hour before sunset.

There are notable exceptions to this daily cycle. For example, thunderstorms can be triggered by the Rocky Mountains in late afternoon, and then propagate eastward all night to hit the Midwest USA with the greatest frequency around midnight to early morning.

14.7.1. Outlooks, Watches & Warnings

A thunderstorm is defined as "**severe**" in the USA if it has one or more of the following:

- tornadoes
- damaging winds (& any winds ≥ 25 m s^{-1})
- hail with diameter ≥ 1.9 cm

Severe weather (thunderstorm-caused or otherwise) also includes heavy rain that could cause flash flooding. All thunderstorms have lightning, by definition, so lightning is not included in the list of severe weather elements even though it kills more people than tornadoes, hail, or winds.

In the USA, severe convective weather forecasting is broken into three time spans: outlook, watch, and warning. These are defined as follows:

- **outlook**: a 6 to 72 h forecast guidance for broad regions. Types:
  - **convective outlooks**: very technical
  - **public severe weather outlooks**: plain language

- **watch**: a 0.5 to 6 h forecast stating that severe weather is favorable within specific regions called **watch boxes**. Watch types include:
  - **severe thunderstorm watch** (includes tornadoes)
  - **tornado watch**

• **warning:** a 0 to 1 h forecast stating that severe weather is already occurring and heading your way. Warned are specific towns, counties, transportation corridors, and small regions called **warning boxes**.
Warning types include:
- **severe thunderstorm warning**
- **tornado warning**

Forecast methods for **warnings** are mostly **nowcasts**. Namely, wait until the severe weather is already occurring as observed by human spotters or radar signatures, or find other evidence indicating that severe weather is imminent. Then anticipate its movement in the next few minutes, and warn the towns that are in the path.

These warnings are delivered by activating civil-defense warning sirens and weather-radio alert messages, by notifying the news media, and by talking with local emergency planners, police and fire agencies. Although the warnings are the most useful because of the details about what, where, and when severe weather will occur or is occurring (Fig 14.77), they are also the shortest term forecasts, giving people only a few minutes to seek shelter. The safest tornado and outflow wind shelters are underground in a basement or a ditch, or in a specially designed reinforced-concrete above-ground **safe room**.

Methods for forecasting **watches** are essentially the methods described earlier in this chapter. Namely, short term (0 to 24 h) numerical weather prediction (NWP) model output, soundings from rawinsondes and satellites, and mesoscale analyses of surface weather-station data are analyzed. These analyses focus on the four elements needed for thunderstorm formation later in the day: high humidity in the boundary layer; nonlocal conditional instability; strong wind shear; and a trigger mechanism to cause the initial lifting. The indices and parameters described earlier in this chapter speed the interpretation of the raw data. To handle some of the uncertainty in thunderstorm behavior, probabilistic severe weather forecasts are produced using ensemble methods (see the NWP Chapter).

Nonetheless, watches are somewhat vague, with no ability to indicate which specific towns will be hit and exactly when they will be threatened (the INFO box has an example of a tornado watch and the associated **watch box** graphic). People in the watch area can continue their normal activities, but should listen to weather reports and/or watch the sky in case storms form nearby. Watches also help local officials prepare for events through changes in staffing of emergency-management and rescue organizations (fire departments, ambulance services, hospitals, police), and deploying storm spotters.

INFO • A Tornado Watch (WW)

Urgent - immediate broadcast requested
Tornado watch number 387
NWS Storm Prediction Center, Norman OK
245 PM CDT Wed May 24 2006

The NWS Storm Prediction Center has issued a tornado watch for portions of
Eastern Iowa
Central into northeast Illinois
Far east central Missouri
Effective this Wednesday afternoon and evening from 245 PM until 1000 PM CDT.

Tornadoes...Hail to 3 inches in diameter... Thunderstorm wind gusts to 70 mph...And dangerous lightning are possible in these areas.

The tornado watch area is approximately along and 75 statute miles east and west of a line from 30 miles east northeast of Dubuque Iowa to 25 miles southwest of Salem Illinois [see Fig. below]. For a complete depiction of the watch see the associated watch outline update (WOUS64 KWNS WOU7).

Remember...A tornado watch means conditions are favorable for tornadoes & severe thunderstorms in & close to the watch area. Persons in these areas should be on the lookout for threatening weather conditions & listen for later statements and possible warnings.

Other watch info...Continue...WW 385...WW 386...

Discussion...50+ knot westerly mid level jet will spread across far southern Iowa/northern Missouri through the evening and increase shear atop moderate instability already in place across WW. Supercells and organized multicell clusters/lines are expected to evolve and shift from west to east through the mid evening. Large hail could become quite large with stronger cores...With additional isolated tornado threat accompanying more persistent supercells through the evening. Wind damage is also likely.

Aviation...Tornadoes and a few severe thunderstorms with hail surface and aloft to 3 inches. Extreme turbulence and surface wind gusts to 60 knots. A few cumulonimbi with maximum tops to 50,000 ft. Mean storm motion vector from 260° at 25 knots.

[Courtesy of the Storm Prediction Center, US Nat. Weather Service, NOAA.]

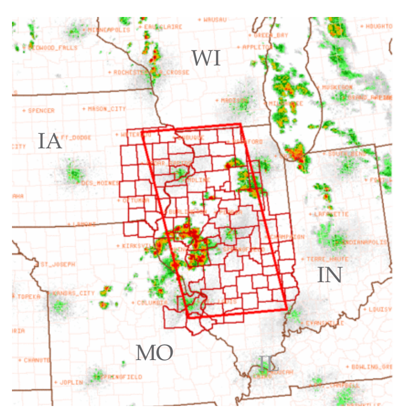


Figure 14.c
The large red parallelogram is a tornado watch box. Smaller red polygons are counties.

Table 14-5. Alphabetical listing of some thunderstorm (CB) stability indices, compiled from both this chapter and the next chapter. Asterisks (*) indicate indices covered in the next chapter.

Abbr.	Full Name	Use to anticipate
BRN	Bulk Richardson Number	CB type
BRN Shear	Bulk Richardson Number Shear	tornado likelihood
CAPE	Convective Available Potential Energy (SB=surface based; ML = mean layer; MU = most unstable; n = normalized)	CB intensity
CIN	Convective Inhibition	strength of cap
CPTP	Cloud Physics Thunder Parameter	lightning likelihood
DCAPE	Downdraft CAPE*	downburst and gust front intensity
EHI	Energy Helicity Index*	supercell intensity tornado intensity
LFC	Level of Free Convection	thunderstorm & tornado likelihood
ML LCL	Mean-layer Lifting Condensation Level	moisture availability & CB intensity
S	Swirl Ratio*	multi-vortex tornadoes
SCP	Supercell Composite Parameter*	supercell & tornado intensity
SHIP	Significant Hail Parameter*	large-hail likelihood
SRH	Storm-Relative Helicity*	mesocyclone rotation
STP	Significant Tornado Parameter*	tornado intensity
UH	Updraft Helicity*	supercell likelihood & tornado path length

Convective outlooks include a general statement about the risk that severe weather will occur in broad regions spanning roughly 350 x 350 km, many hours or days into the future. These risk levels are:

• **Slight** (SLGT): well-organized severe thunderstorms are expected in small numbers and/or low coverage. Specifically: 5 to 29 reports of hail ≥ 2.5 cm, and/or 3 to 5 tornadoes, and/or 5 to 29 wind events having wind speeds ≥ 25 m s^{-1}.

• **Moderate** (MDT): a greater concentration of severe thunderstorms, and in most situations, greater magnitude of severe weather. Details: ≥ 30 reports of hail ≥ 2.5 cm, or 6 to 19 tornadoes, or numerous wind events (30 that might be associated with a squall line, bow echo or derecho).

• **High** (HIGH): a major severe-weather outbreak is expected, with great coverage of severe weather and enhanced likelihood of extreme weather (i.e., violent tornadoes or extreme convective wind events over a large area). Details: ≥ 20 tornadoes with at least two rated ≥ EF3, or an extreme derecho causing ≥ 50 widespread high wind events with numerous higher winds (≥ 35 m s^{-1}) and structural damage reports.

14.7.2. Stability Indices for Thunderstorms

As you have already seen, meteorologists have devised a wide variety of indices and parameters to help forecast thunderstorm formation, strength, and type. Table 14-5 summarizes indices that were discussed earlier in this chapter as well as indices from the next chapter. Many of these indices integrate over large portions of the environmental sounding or hodograph, and are automatically calculated by computer programs for display next to the computer-plotted sounding.

Because no single index has proved to be the best, new indices are devised, modified, and tested every year. Some day a single best index might be found.

The use of indices to aid severe weather forecasting has a long tradition. Many older indices were devised for calculation by hand, so they use data at a few key altitudes, rather than integrating over the whole sounding. Table 14-6 lists some of the older indices and the associated forecast guidelines.

Table 14-7 compares the values of the different indices with respect to the forecasted weather elements. Beware that this table is highly simplified, and that the boundaries between different severity of storm are not as sharp as the table suggests. Nonetheless, you can use it as a rough guide for interpreting the index values that are often printed with plotted soundings or weather maps.

Table 14-6. Definition and interpretation of older thunderstorm stability indices. Notation: T = environmental temperature (°C). Td = environmental dew-point temperature (°C). $Tp_{s\text{->}e}$ = final temperature of an air parcel that started with average conditions at height s and then rose to ending height e following dry adiabat up to LCL, and moist adiabat above. M = wind speed (m s^{-1}). α = wind direction (degrees). Subscripts give pressure altitude. CB = thunderstorms.

Abbr.	Full Name	Definition	Values & Interpretation
K or KI	K Index	$K = T_{85kPa} + Td_{85kPa} + Td_{70kPa} - T_{70kPa} - T_{50kPa}$	< 20 CB unlikely 20 to 30 Chance of scattered CB 30 to 40 Many CB; heavy rain > 40 CB; very heavy rain
LI	Lifted Index	$LI = T_{50kPa} - Tp_{95\text{->}50kPa}$	> 2 CB unlikely 0 to 2 CB only if strong trigger -3 to 0 Weak CB possible -6 to -3 Moderate CB probable < -6 Severe CB likely
SSI	Showalter Stability Index	$SSI = T_{50kPa} - Tp_{85\text{->}50kPa}$	> 3 CB unlikely 1 to 3 Weak showers possible -3 to 0 Severe CB possible -6 to -4 Severe CB probable < -6 Severe CB likely
SWEAT	Severe Weather Threat Index	$SWEAT = 12 \cdot Td_{85kPa} + 20 \cdot (TT - 49) + 4 \cdot M_{85kPa} + 2 \cdot M_{50kPa} + 125 \cdot [0.2 + \sin(\alpha_{50kPa} - \alpha_{85kPa})]$ where TT = total totals index. Note: more rules set some terms=0	< 300 CB unlikely 300-400 Chance isolated severe CB 400-500 Severe CB likely; & tornado possible 500-800 Severe CB & tornado likely
TT	Total Totals Index	$TT = T_{85kPa} + Td_{85kPa} - 2 \cdot T_{50kPa}$	< 45 CB unlikely 45 to 50 Scattered CB possible 50 to 55 CB likely; some severe 55 to 60 Severe CB likely; tornado likely

Table 14-7. Approximate relationship between storm indices and storm intensity.

	Thunderstorm (CB) & Tornado (EF0 - EF5) Severity						
Index	**No CB**	**Ordinary CB**	**Marginal Supercell**	**Supercell, no tornado**	**Supercell & EF0-EF1**	**Supercell & EF2-EF5**	**Units**
BRN		150	70	30	30	30	
BRN Shear		7	30	45	55	70	$m^2 s^{-2}$
CAPE (ML)		950	1205	1460	1835	2152	$J \cdot kg^{-1}$
CAPE (MU)		1750	1850	1950	2150	2850	$J \cdot kg^{-1}$
CIN		18		35	12		$J \cdot kg^{-1}$
EHI (ML 0-1km)		0.1	0.5	0.8	1.4	2.1	
K	15	25	35	45			°C
LCL (ML)		1.75	1.47	1.34	1.18	1.00	km
LI	+1.5	–1.5	–4.5	–7.5			°C
SCP		0	1.1	3.5	5.9	11.1	
Shear (0-6km)		8	15	22	23	24	$m s^{-1}$
SRH (0-1km)		20	70	115	155	231	$m^2 s^{-2}$
SRH-effective		16	60	117	166	239	$m^2 s^{-2}$
SSI	+4.5	+1.5	–1.5	–4.5	–7.5		°C
STP		0	0	0.4	0.9	2.7	
SWEAT	300	350	400	450	500		
TT	42	47	52	57	62		°C

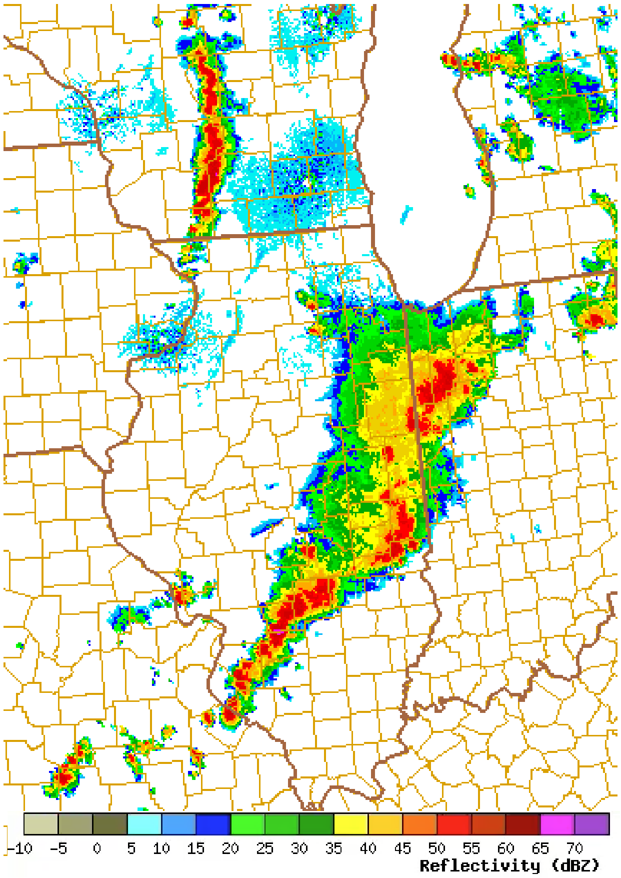


Figure 14.78
Weather radar composite reflectivity image of squall lines in Indiana (IN), Illinois (IL) and Wisconsin (WI), USA, at 2313 UTC on 24 May 2006. Dark red is 60 to 65 dBZ.

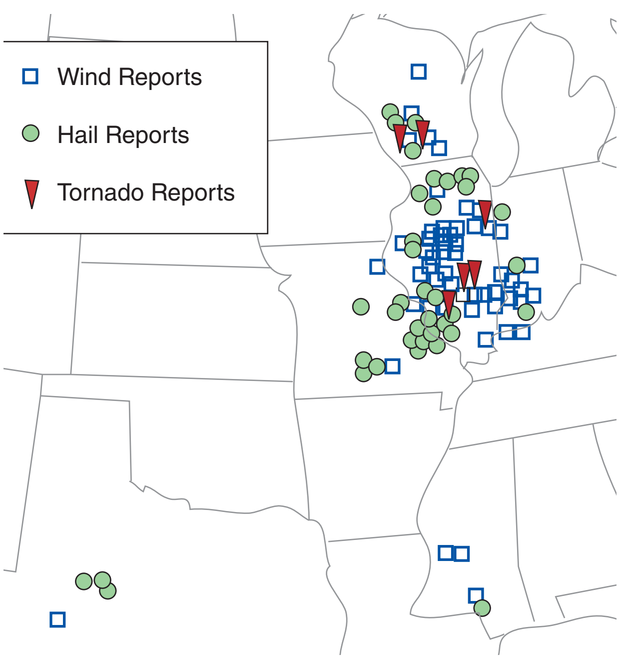


Figure 14.79
All storm reports during 24 May 2006 of damaging wind, hail, and tornadoes in the central USA.

14.8. STORM CASE STUDY

Throughout this chapter and the next chapter are examples of weather maps for the severe storm event of 24 May 2006. These maps presented indices and trigger locations that could be used to help forecast severe weather. The weather maps were created by the US National Weather Service and served over the internet in real time on that day, giving up-to-date weather information to meteorologists, public officials, storm chasers, radio and TV stations, and the general public.

On that day, warm humid air from the Gulf of Mexico was streaming northward into the Mississippi Valley, reaching as far north as the Midwest USA. This provided fuel in the boundary layer, ahead of a cold front that spanned from Illinois to Texas. An additional trigger was a west-Texas dry line that merged with the cold front. Cold dry air aloft coming from the west contributed to the instability of the soundings, and wind shear near Illinois enabled supercells to form.

The US Storm Prediction Center anticipated that severe weather would form that afternoon, and issued a series of watches and warnings that successfully covered the severe weather area. Most of the severe weather was expected to occur that day in the Midwest, although there was also a chance of storms in Texas near the dry line.

Fig. 14.78 shows a radar snapshot of the squall line that moved across the Midwest USA at 22 UTC (5 PM local daylight time). Fig. 14.79 shows a map of the resulting **storm reports**, which included:

- 6 tornado reports
- 97 damaging wind reports & 1 wind >32 m s^{-1}
- 84 hail reports + 3 reports of hail > 5 cm diam.

A total of 187 reports of severe weather were compiled for this storm case. This was a typical severe weather day.

Now that we are finished with the sections on the formation conditions and characteristics of thunderstorms we will proceed into the next chapter on Thunderstorm Hazards. These include heavy rain, hail, downbursts of wind, gust fronts, lightning, thunder, and tornadoes.

14.9. REVIEW

Thunderstorms (cumulonimbi) are violent convective clouds that fill the depth of the troposphere. Thunderstorms look like mushrooms or anvils. The most violent thunderstorms are called supercells, in which the whole thunderstorm rotates as a mesocyclone. Other storm organizations include airmass thunderstorms, multicell storms, orographic storms, mesoscale convective systems, squall lines, and bow echoes.

Four conditions are needed to form thunderstorms: high humidity; instability; wind shear, and a trigger to cause lifting. Thermodynamic diagrams and hodographs are used to determine the moisture availability, static stability and shear. A variety of stability and shear indices have been devised to aid thunderstorm forecasting. Trigger mechanisms are often mountains or airmass boundaries such as synoptic fronts, dry lines, sea-breeze fronts, and gust fronts, such as determined from weather radar, satellite, and surface weather analyses.

Thunderstorms are like gigantic engines that convert fuel (moist air) into motion and precipitation via the process of condensation and latent heat release. Once triggered, thunderstorms can often sustain themselves within a favorable environment for 15 minutes to several hours. They are most frequent in the late afternoon and evening over land. The explosive growth of thunderstorms, their relatively small diameters, and their sensitivity to initial conditions make it difficult to forecast thunderstorms.

14.10. HOMEWORK EXERCISES

14.10.1. Broaden Knowledge & Comprehension

B1. Search the internet for (and print the best examples of) photographs of:
- a. airmass thunderstorms
- b. multicell thunderstorms
- c. orographic thunderstorms
- d. squall-line thunderstorms
- e. supercell thunderstorms
  - (i) classic
  - (ii) low precipitation
  - (iii) high precipitation
- f. mammatus clouds
- g. wall clouds
- h. beaver tail clouds
- i. flanking lines
- j. tail cloud

(Hint: search on "storm stock images photographs".) Discuss the features of your resulting image(s) with respect to information you learned in this chapter.

B2. Search the internet for (and print the best examples of) radar reflectivity images of
- a. airmass thunderstorms
- b. multicell thunderstorms
- c. squall line thunderstorms
- d. mesoscale convective systems
- e. bow echoes

Discuss the features of your resulting image(s) with respect to information you learned in this chapter.

B3. Same as the previous question, but for radar Doppler velocity images.

B4. Search the internet for (and print the best examples of) satellite images of:
- a. airmass thunderstorms
- b. multicell thunderstorms
- c. squall line thunderstorms
- d. mesoscale convective systems
- e. dry lines

Do this for visible, IR, and water vapor channels. Discuss the features of your resulting image(s) with respect to information you learned in this chapter.

B5. Search the internet for plotted soundings. In particular, find web sites that also list the values of key stability indices and key altitudes along with the sounding. From this site, get a sounding for a location in the pre-storm environment toward which thunderstorms are moving. Print the sounding and the stability indices, and use the sounding and indices to discuss the likelihood and severity of thunderstorms. (Hint: To find regions where thunderstorms are likely, first search the web for a weather radar image showing where the thunderstorms are. Alternately, find via web search the regions where the weather service has issued thunderstorm watches.)

B6. Search the internet for plotted hodographs. (Hint: Some sites plot hodographs along with plotted soundings, while others have separate hodograph plots.) Find a web site that also shows key wind-shear parameters associated with the hodograph, and perhaps also gives expected storm motion and/or storm-relative winds. Do this for a location into which thunderstorms are moving (see hint from previous exercise). Print and discuss the results.

B7. Search the internet for (and print the best examples of) real-time weather-map analyses or forecasts (from numerical models) showing plotted fields of the four conditions needed for convective storm formation. (Hint, you might find these 4 conditions on 4 separate maps, so you should print and discuss all 4 maps.) Discuss.

B8. Search the internet for maps showing real-time thunderstorm and tornado watches. Print and discuss the results. If you also did the previous exercise, then discuss the amount of agreement of the watch areas with the regions that satisfy the conditions for thunderstorms.

B9. Use the internet to find thunderstorm tutorials. (Hints: some elementary tutorials are at the "university of Illinois online meteorology guide" site. More advanced tutorials are at the "ucar meted" site. Much miscellaneous info is at the "wikipedia" site.) After reading the tutorial, print and discuss one new aspect of thunderstorms that wasn't in this textbook, but which you find interesting and/or important.

B10. Search the web for newer thunderstorm indices than were discussed in this chapter. Summarize and print key information (e.g., definitions for, equations governing, advantages compared to older indices) for one of these new indices, and if possible show a real-time map of the values of this index.

B11. Search the internet for a recent sounding close to your present location. Get and print this sounding data in text form. Then manually plot the temperatures and dew points on a thermo diagram, and plot the winds on a hodograph. Manually calculate a few key thunderstorm indices, and make your own forecast about whether thunderstorms are likely.

B12. Same as the previous exercise, but for a sounding just ahead of a severe thunderstorm location.

B13. Search the internet for a "storm spotter glossary", and list 10 new thunderstorm-related terms and their definitions that were not discussed in this chapter, but which you feel are important. Justify your choice of terms.

B14. Search the internet for (& print good examples of) new research that is being done on thunderstorms or any of the topics discussed in this chapter.

B15. Search the internet for four key storm components that spotters analyze to estimate thunderstorm strength, type, and stage of evolution.

B16. Search the internet to explain the difference between "storm chasers" and "storm spotters".

B17. Search the internet to find which agency or agencies within your own country are responsible for making the national forecasts of severe thunderstorms. Print their web address, along with basic information about their mission and location. Also, find and print a list of qualifications they want in new meteorologists that they hire.

B18. For the following country, where on the internet can you find up-to-date tornado/thunderstorm watches/warnings?

a. Canada b. Japan c. China
d. Australia e. Europe f. USA
g. or a teacher-assigned country?

14.10.2. Apply

A1(§). Plot the pressure difference ($\Delta P = P_{MCS} - P_{std.atm.}$) vs. height within a mesoscale convective system (MCS), given the temperature differences ($\Delta T = T_{MCS} - T_{std.atm.}$) below. Also plot the given temperature difference vs. height. The surface pressure $P_{MCS\ sfc}$ under the MCS is also given below. For the standard atmosphere (std. atm.), assume the surface pressure is 101.325 kPa. Use a vertical resolution of Δz = 250 m.

Exercise Part	a.	b.	c.	d.
	ΔT (K)			
$0 \le z < 2$ (km)	–10	–8	–13	–15
$2 \le z < 4$ (km)	0	0	0	0
$4 \le z < 9$ (km)	+6.5	+5	+10	+9
$9 \le z < 11$ (km)	–10	–7.7	–15	–13
$11 \le z \le 12$ (km)	0	0	0	0
$P_{MCS\ sfc}$ (kPa)	101.8	101.7	101.9	102.05

A2. Using your answer for ΔP at 2 km altitude from the previous exercise, calculate the rear-inflow jet (RIJ) speed into the mid-tropospheric meso-low within an MCS. Assume the RIJ is geostrophic, for a latitude of 40°N. Assume the MCS has a horizontal radius (km) of:

a. 80 b. 100 c. 120 d. 140 e. 160
f. 180 g. 200 h. 220 i. 240 j. 260

A3. Given the following prestorm sounding. Plot it on a thermo diagram (use a skew-T, unless your instructor specifies a different one). Find the mixed-layer height, tropopause height, lifting condensation level, level of free convection, and equilibrium level, for an air parcel rising from the surface. Use a surface (P = 100 kPa) dew-point temperature (°C) of:

a. 22 b. 21 c. 20 d. 19 e. 18 f. 17 g. 16
h. 15 i. 14 j. 13 k. 12 l. 11 m. 10 n. 9
o. 8 p. 7 q. 6 r. 5 s. 4

Sounding:

P (kPa)	T (°C)
20	–45
25	–45
30	–40
40	–30
50	–19
70	+5
80	15
88	21
92	21
98	26
100	30

A4. Using the result from the previous exercise, forecast the intensity (i.e., category) of thunderstorm and possibly tornadoes that are likely, given the lifting condensation level you found.

A5. Find the median and interquartile range for the following data sets:

a. 6 3 9 7 2 1 6 0 8
b. 8 5 9 8 1 1 3 2 6 8 4 9 7 0
c. 3 5 2 7 9 4 7 7
d. 9 6 8 0 9 1 3 2 9 7 8
e. 5 8 2 2 1 6 4 3 7 9
f. 4 5 6 7 8 9 0 1 2 3
g. 1 2 3 4 5 6 7 8 9
h. 9 8 7 6 5 4 3 2 1
i. 0 0 0 0 3 6 8 3 4 8 9 0 6

A6. Find the mean and standard deviation for the data set from the previous exercise.

A7. For the sounding from exercise A3, find the value of surface-based CAPE, using:
(1) the height (z) tiling method.
(2) the pressure (P) tiling method.
Assume $T_v \approx T$. Also, on the plotted sounding, shade or color the CAPE area.

A8. Start with the sounding from exercise A3. Assume that the forecast high temperature for the day is 2°C warmer than the surface temperature from the sounding. Find the value of surface-based CAPE. Assume $T_v \approx T$.

A9. For the sounding from exercise A3, find the value of mean-layer CAPE. Assume $T_v \approx T$. For the air parcel that you conceptually lift, assume its initial temperature equals the average temperature in the bottom 1 km of the sounding, and its initial dew-point is the average dew-point temperature in the bottom 1 km of the sounding (assuming that the mixing ratio is constant with height in this 1 km mixed layer).

A10. Using the result from the previous exercise, forecast the intensity of thunderstorms and possibly tornadoes that are likely, and indicate the uncertainty (i.e., the range of possible storm intensities) in this forecast, given the MLCAPE you found.

A11. Forecast the thunderstorm and/or tornado intensity, and indicate the uncertainty (i.e., the range of possible storm intensities) in this forecast, given a ML CAPE ($J \cdot kg^{-1}$) value of:

a. 1000 b. 1250 c. 1500 d. 1750 e. 2000
f. 2250 g. 2500 h. 2750 i. 3000 j. 3250
k. 3500 l. 250 m. 500 n. 750

A12. Given the sounding from exercise A3, except with a surface temperature of T = 24°C at P = 100 kPa. Assume the surface dew point from exercise A3 defines a mixing ratio that is uniform in the bottom 2 layers of the sounding.

Find the most unstable CAPE considering air parcels that start their rise from just the bottom two levels of the sounding (i.e., for air parcels that start at P = 100 and 98 kPa). Assume $T_v \approx T$.

A13. Using the result from the previous exercise, forecast the intensity (i.e., category) of thunderstorm and possibly tornadoes that are likely, given the MUCAPE you found, and indicate the uncertainty (i.e., the range of possible storm intensities) in this forecast.

A14. Forecast the thunderstorm and/or tornado intensity, and indicate the uncertainty (i.e., the range of possible storm intensities) in this forecast, given a MU CAPE ($J \cdot kg^{-1}$) value of:

a. 1000 b. 1250 c. 1500 d. 1750 e. 2000

f. 2250 g. 2500 h. 2750 i. 3000 j. 3250
k. 3500 l. 3750 m. 4000 n. 4250 n. 4500

A15. Using the SBCAPE you found in exercise A7, and the sounding and key altitudes from exercise A3, find the value of normalized CAPE, indicate of the CAPE is tall thin or short wide, and suggest whether thunderstorms would favor heavy precipitation or tornadoes.

A16. Estimate the max likely updraft speed in a thunderstorm, given a SB CAPE (J·kg^{-1}) value of:
a. 1000 b. 1250 c. 1500 d. 1750 e. 2000
f. 2250 g. 2500 h. 2750 i. 3000 j. 3250
k. 3500 l. 3750 m. 4000 n. 500 n. 750

A17. Given a wind of (U, V) = (10, 5) m s^{-1} at z = 1 km. Find the wind difference magnitude, shear direction, and shear magnitude between z = 1 and 2 km, given winds at 2 km of (U, V) (m s^{-1}) of:
a. (20, –3) b. (25, –15) c. (30, 0) d. (25, 5)
e. (10, 15) f. (5, 20) g. (–15, 15) h. (–5, 0)
i. (–20, –30) j. (–5, 8) k. (–10, –10) l. (0, 0)

A18. Plot the following wind sounding data on a hodograph. (Make copies of the blank hodograph from Fig. 14.51 to use for all the hodograph exercises.)

z (km)	In each cell: wind direction (°), speed (m s^{-1})			
	Exercise			
	a	**b**	**c**	**d**
0	calm	100, 5	120, 8	150, 10
1	120, 5	120, 10	150, 5	180, 5
2	150, 8	160, 15	210, 5	240, 5
3	180, 12	220, 25	240, 10	260, 10
4	210, 15	240, 30	250, 15	260, 20
5	240, 25	250, 33	258, 25	250, 30
6	260, 40	250, 33	260, 35	240, 40

A19. On your plotted hodograph from the previous exercise, draw wind vectors from the origin to each wind data point. Discuss the relationship between these wind vectors and the original hodograph that connected just the tips of all the vectors.

A20. Using your hodograph plot from exercise A18, use the hodograph (not equations) to find the (U, V) components (m s^{-1}) at each altitude in the sounding.

A21. Graphically, using your hodograph plot from exercise A18, plot the local shear vector (m s^{-1}) across the z = 2 to 3 km layer.

A22. Same as the previous exercise, except solve for the local shear vector (m s^{-1}) mathematically.

A23. Graphically, using your hodograph plot from exercise A18, plot the 0 to 6 km mean shear vector (m s^{-1}).

A24. Same as exercise A23, except solve for the 0 to 6 km mean shear vector (m s^{-1}) mathematically.

A25. Using your hodograph plot from exercise A18, graphically find the 0 to 6 km total shear magnitude (m s^{-1}). Also, predict the likely thunderstorm and possibly tornado intensity based on this parameter, and indicate the uncertainty (i.e., the range of possible storm intensities) in this forecast.

A26(§). Same as the previous exercise, except solve for the 0 to 6 km total shear magnitude (m s^{-1}) mathematically. Also, predict the likely thunderstorm and possibly tornado intensity based on this parameter, and indicate the uncertainty (i.e., the range of possible storm intensities) in this forecast.

A27. Predict the thunderstorm and possibly tornado intensity based on the following total shear magnitude (m s^{-1}) across the 0 to 6 km layer, and indicate the uncertainty (i.e., the range of possible storm intensities) in this forecast.
a. 4 b. 6 c. 8 d. 10 e. 12 f. 14 g. 16
h. 18 i. 20 j. 22 k. 24 l. 26 m. 28 n. 30

A28. Graphically, using your hodograph plot from exercise A18, find the mean environmental wind direction (°) and speed (m s^{-1}), for normal storm motion.

A29(§). Same as the previous exercise, except solve mathematically for the mean environmental wind direction (°) and speed (m s^{-1}), for normal storm motion.

A30. Given the hodograph shape from exercise A18, indicate whether right or left-moving supercells would dominate. Also, starting with the "normal storm motion" from exercise A28 (based on hodograph from exercise A18), use the Internal Dynamics method on your hodograph to graphically estimate the movement (i.e., direction and speed) of
(1) Right-moving supercell thunderstorms
(2) Left-moving supercell thunderstorms

A31. Using your result from exercise A28 for normal storm motion (based on exercise A18), and assuming CAPE = 2750 J·kg^{-1}:

(1) calculate the shear ΔM (m s^{-1}) between the mean (0-6 km) environment winds and the average winds at z = 0.5 km.

(2) calculate the bulk Richardson number shear (BRN shear) in m^2·s^{-2}.

(3) calculate the bulk Richardson number (BRN).

(4) forecast the thunderstorm likelihood, thunderstorm type, and tornado intensity.

A32. Given the following value of bulk Richardson number, determine the likely thunderstorm type:

a. 7 b. 12 c. 17 d. 22 e. 27 f. 32 g. 37
h. 42 i. 47 j. 52 k. 57 l. 62 m. 67 n. 72

A33. Given the following value of bulk Richardson number (BRN) shear (m^2·s^{-2}), determine the likely thunderstorm and tornado intensity:

a. 5 b. 10 c. 20 d. 30 e. 40 f. 45 g. 50
h. 55 i. 60 j. 65 k. 70 l. 75 m. 80 n. 90

A34. Given the sounding from exercise A3, calculate the value of the convective inhibition (CIN) in J·kg^{-1} for an air parcel lifted from the surface. Assume $T_v \approx T$. Also, shade or color the CIN area on the plotted sounding.

A35. Given your answer to the previous exercise, interpret the strength of the cap and how it affects thunderstorm triggering.

A36. Calculate the mean-layer CIN using the sounding from exercise A3. Temperatures in the bottom 10 kPa of atmosphere are as plotted from the sounding, and assume that the mixing ratio is constant over this layer and equal to the mixing ratio that corresponds to the surface dew point.

A37. For the sounding from exercise A3:

(1) To what altitude must a surface air parcel be lifted by an airmass boundary or a mountain, to trigger thunderstorms?

(2) To what temperature must the surface air be heated in order to convectively trigger thunderstorms (i.e., find the convective temperature), and what is the value of the convective condensation level?

A38. What values of K index, Lifted Index, Showalter Stability Index, SWEAT Index, and Total-totals Index would you anticipate for the following intensity of thunderstorm (CB)?

a. no CB
b. ordinary CB
c. marginal supercell
d. supercell with no tornado
e. supercell with EF0 - EF1 tornado
f. supercell with EF2 - EF5 tornado

A39. What value of the Bulk Richardson Number, BRN-Shear, ML CAPE, MU CAPE, CIN, Energy Helicity Index, Lifting Condensation Level, Supercell Composite Parameter, 0-6 km Shear, Storm-Relative Helicity, effective Storm-Relative Helicity, and Significant Tornado Parameter would you anticipate for the following intensity of thunderstorm (CB)?

a. no CB b. ordinary CB
c. marginal supercell
d. supercell with no tornado
e. supercell with EF0 - EF1 tornado
f. supercell with EF2 - EF5 tornado

14.10.3. Evaluate & Analyze

E1. Identify as many thunderstorm features as you can, from the photo below. (Image "wea00106" courtesy of NOAA photo library).

E2. Why does a thunderstorm have a flat (anvil or mushroom) top, instead of a rounded top such as cumulus congestus clouds? (Ignore the overshooting dome for this question.)

E3. Almost all clouds associated with thunderstorms are caused by the lifting of air. List each of these clouds, and give their lifting mechanisms.

E4. Consider Figs. 14.4a & b. If you were a storm chaser, and were off to the side of the storm as indicated below, sketch which components of the storm and associated clouds would be visible (i.e., could be seen if you had taken a photo). Label the key cloud features in your sketch. Assume you are in the following direction from the storm:

a. northeast b. southwest c. northwest

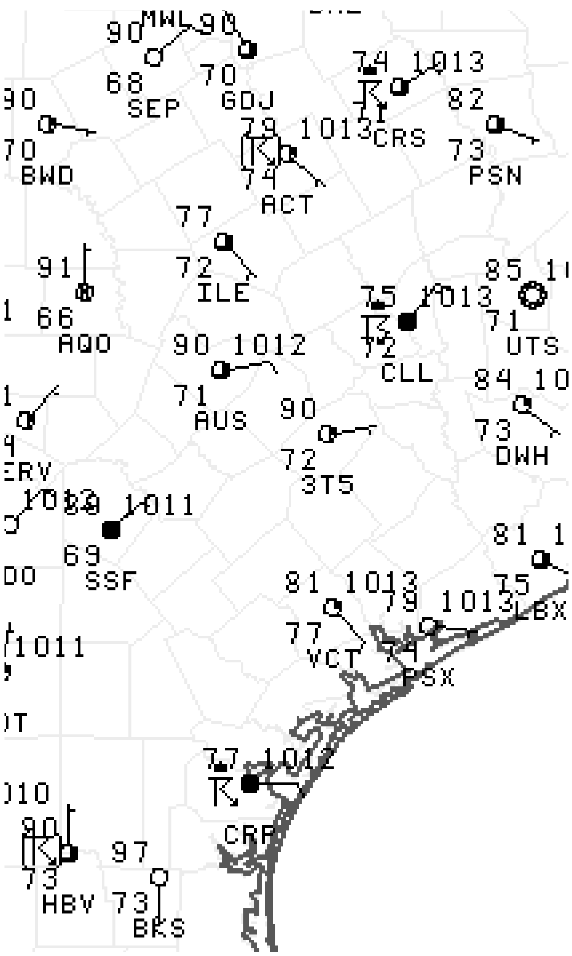
For example, Fig. 14.4a shows the sketch for the view from southeast of the storm.

E5. Consider Fig. 14.4b. If a Doppler radar were located near the words in that figure given below, then sketch the resulting color patterns on the Doppler PPI wind display for near-surface winds.

a. "FFD" b. "RFD" c. "Anvil Edge" d. "Cu con"
e. "Gust Front" f. "Beaver Tail"

g. "Outflow Boundary Layer Winds"
h. "Inflow Boundary Layer Winds"
i. "Main Updraft" j. "Storm Movement"
k. "Forward Flank Downdraft"

E6. Circle the stations in the weather map below that are reporting thunderstorms. (Image courtesy of the Storm Prediction Center, NWS, NOAA.)



E7. Can a thunderstorm exist without one or more cells? Explain.

E8. For each thunderstorm type or organization, explain which triggered mechanism(s) would have most likely initiated it.

E9. For each thunderstorm type or organization, explain how the phenomenon would differ if there had been no wind shear.

E10. What is a derecho, and what causes it? Also, if strong straight-line winds can cause damage similar to that from a weak tornado, what are all the ways that you could use to determine (after the fact) if a damaged building was caused by a tornado or derecho?

E11. Summarize the different types of supercells, and list their characteristics and differences. Explain what factors cause a supercell to be of a specific type, and where you would most likely find it.

E12. In a thunderstorm, there is often one or more updrafts interspersed with one or more downdrafts. Namely, up- and down-drafts are often adjacent. Do these adjacent up- and down-drafts interfere or support each other? Explain.

E13. List the 4 conditions needed for thunderstorm formation. Then, consider a case where one of the conditions is missing, and explain why thunderstorms would be unlikely, and what would form instead (if anything). Do this for each of the 4 conditions separately.

E14. If thunderstorms normally occur at your town, explain how the 4 conditions needed for thunderstorm formation are satisfied for your region. Namely, where does the humid air usually come from, what conditions contribute to shear, what trigger mechanisms dominate in your region, etc.

E15. If thunderstorms are rare in your region, identify which one or more of the 4 conditions for thunderstorms is NOT satisfied, and also discuss how the other conditions are satisfied for your region.

E16. If the cap inhibits thermals from rising, why is it considered a good thing for thunderstorms?

E17. If your national weather service were to make only one upper-air (rawinsonde) sounding per day, when would you want it to happen, in order to be most useful for your thunderstorm forecasts?

E18. In the section on convective conditions and key altitudes, one thermo diagram was presented with the LCL above z_i (the mixed layer top), while the Sample Application showed a different situation with the LCL below z_i. Both are frequently observed, and both can be associated with thunderstorms.

Discuss and justify whether it is possible to have an environment favorable for thunderstorms if:

a. z_i is above the LFC
b. z_i is above the tropopause
c. LCL is at the Earth's surface
d. LFC is at the Earth's surface
e. LCL is above the LFC

E19. The tops of thunderstorms are often near the tropopause.

a. Why is that?

b. Why are thunderstorm tops usually NOT exactly at the tropopause.

E20. Based on tropopause info from earlier chapters, how would you expect thunderstorm depth to vary with:
a. latitude
b. season

E23. Same as the previous exercise, but for the sounding given in the Sample Application after Fig. 14.26.

E24. Fig. 14.22 shows schematically the information that is in Fig. 14.25. Explain how these two figures relate to each other.

E25. Consider Fig. 14.22. Why is the following environmental condition conducive to severe thunderstorms?
a. cold air near the top of the troposphere
b. dry air near the middle of the troposphere
c. a stable cap at the top of the boundary layer
d. a warm humid boundary layer
e. wind shear
f. strong winds aloft.
(Hint, consider info from the whole chapter, not just the section on convective conditions.)

E26. Fig. 14.26 shows a nonlocally conditionally un-stable environment up to the EL, yet the sounding in the top of that region (i.e., between the tropopause and the EL) is locally statically stable. Explain how that region can be both stable and unstable at the same time.

E27. How much energy does an airmass thunderstorm release (express your answer in units of megatons of TNT equivalent)? Assume all the water vapor condenses. Approximate the cloud by a cylinder of radius 5 km, with base at P = 90 kPa and top at P = 30 kPa. The ABL air has depth 1 km, and dew-point temperature (°C) of:
a. 17 b. 16 c. 15 d. 14 e. 13 f. 12 g. 11
h. 18 i. 19 j. 20 k. 22 l. 23 m. 24 n. 25

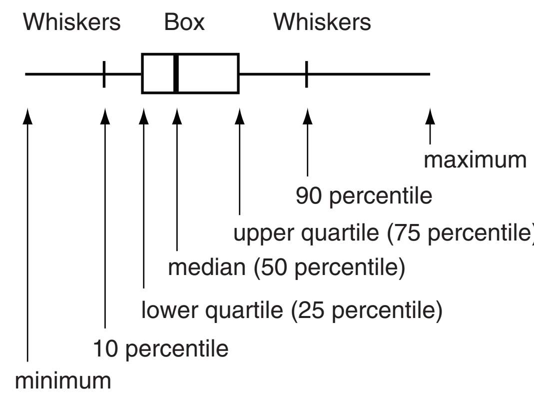
E28. Why do thunderstorms have (nearly) flat bases? (Hint, what determines the height of cloud base in a convective cloud?)

E29. Compare the different ways to present information about "high humidity in the boundary layer", such as the different weather maps shown in that section. What are the advantages and disadvantages of each?

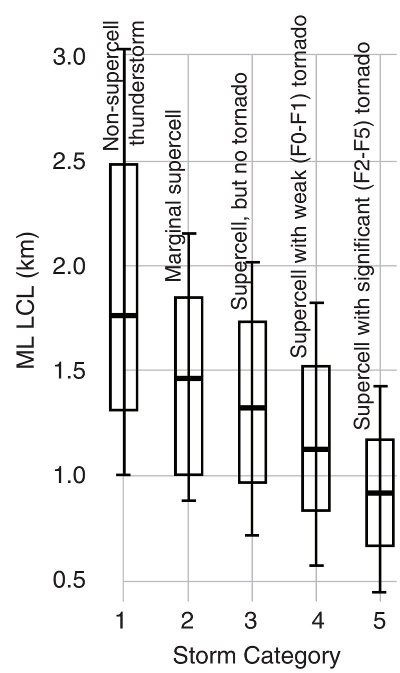
Normally, meteorologists need use only one of the humidity metrics to get the info they need on moisture availability. Which one moisture variable would you recommend using, and why?

E30. Look up the definitions of the residual layer and nocturnal stable layer from the Atmospheric Boundary Layer chapter. Consider atmospheric conditions (temperature, humidity) in the residual layer, relative to the conditions in the mixed layer from the afternoon before. Use this information to explain why strong thunderstorms are possible at night, even after the near-surface air temperature has cooled significantly.

E31. Often statistical data is presented on a **box-and-whisker diagram**, such as sketched below.



For example, the data such as in Fig. 14.33 is often presented in the journal literature as shown below (where the max and min values are not shown for this illustration, but are often given in the literature).



Create box-and-whisker diagrams, using data in this textbook for:

a. ML CAPE
b. MU CAPE
c. TSM
d. BRN Shear
e. 0-1 km SRH
f. 0-1 km EHI

E32. Look at a blank thermo diagram. At colder temperatures, the moist adiabats don't deviate far from the dry adiabats. Use this characteristic to help explain why CAPE is smaller for colder environments, and thus why strong thunderstorms are less likely.

E33. Thunderstorms are much more prevalent in the tropics than near the poles. Use all the info from this chapter to help explain this. Also relate it to info in other chapters, such as: radiation, global circulation, boundary layers etc.

E34. If you were given a file of ASCII text data of temperature and dew point vs. height in a sounding, show schematically (i.e., with a flow chart) a computer program that you could write to compute the CAPE. Don't actually write the program, just show the main steps, procedures, and/or data structures that you would use. (Hint: the method in Fig. 14.35 might be easier than the method in Fig. 14.36, especially for very small Δz.)

a. Do this based on eq. (14.2) [gives an approximate answer]

b. Do this based on eq. (14.1) [this is more accurate]

E35. Describe all the factors in a plotted sounding that could contribute to larger CAPE values.

E36. Compare and contrast the different parcel-origin methods for computing CAPE (SB, forecast SB, ML, and MU). Recommend which methods would be best for different situations.

E37. Derive eq. (14.5) from eq. (14.4) by using the hypsometric equation. Show your steps.

E38. Critically analyze eq. (14.6). Discuss the behavior of the equation as the height difference between the LFC and EL approaches zero.

E39. Consider CAPE in a prestorm sounding.

a. Can all the CAPE be consumed by the storm?

b. Do the CAPE eqs. have limitations? Discuss.

c. Are there reasons why thunderstorm vertical velocity might not get as large as given in eq. (14.8)?

E40. a. Is wind shear a vector or scalar quantity? Why?

b. Is it possible to have non-zero shear between two altitudes where the winds have the same speed? Explain.

E41. Suppose that Fig. 14.47 gives the winds relative to the ground. Redraw that figure showing the storm-relative winds instead. Also, explain why storm-relative winds are important for thunderstorm dynamics.

E42. On a hodograph, why is it important to list the altitude of the winds next to each wind data point?

E43. Draw a hodograph showing an environment where winds ____ with height. a. veer b. back

E44. On blank hodographs such as Fig. 14.51, why are the printed wind directions 180° out of phase with the normal compass directions?

E45. In Fig. 14.52, notice that the dark vertical and horizontal lines don't look straight. This is an optical illusion caused by the background circles of the hodograph. In fact, they are straight, as you can see by tilting the page so you are looking almost edge-on, sighting down each line. Based in this info, discuss why you should never draw the vertical or horizontal lines by eye, when trying to determine *U* and *V* components of wind.

E46. Three different types (i.e., ways of measuring) of shear vectors are shown in Fig. 14.53. Explain and contrast them.

E47. Create a table listing all the different types of shear that are discussed in this chapter, and add a column to the table that indicates what thunderstorm characteristics can be estimated from each shear type.

E48. A thunderstorm index or parameter is useful only if it can discriminate between different storm types or intensities. Using Fig. 14.56, critically evaluate the utility of TSM in estimating storm and tornado intensity.

E49. Explain the difference between the mean shear vector (as in Fig. 14.53) and the mean wind vector (as in Fig. 14.58), and discuss their significance.

E50. In Fig. 14.59, the mean wind is estimated as the center of mass of the shaded area. Is the actual mean wind exactly at the center of mass of the shaded area, or only close? [Hint, to help think about this question, create some very simple hodographs with simple shapes for the shaded area [such as a rectangle] for which you can calculate the exact center of mass. Then see if you can extend your argument to more complex, or arbitrary shapes. This scientific approach is called inductive reasoning (see p107).]

E51. Using information from Fig. 14.61, identify where in Fig. 14.60 are the right and left moving mesocyclones.

E52. Although Fig. 14.62c shows a hodograph that is concave up, the whole curve is above the origin. Suppose that, on a different day, the hodograph has the same shape, but is shifted to be all below the origin. Assume N. Hemisphere, mid-latitudes.

Would such a hodograph still favor left-moving supercell storms? Justify your answer.

E53. Both the bulk Richardson number (BRN) and the Supercell Composite Parameter (SCP) include CAPE and shear. However, BRN is CAPE/shear, while SCP is CAPE·shear (times a third factor that we can ignore for this Exposition). Which (BRN or SCP) would you anticipate to give the best storm intensity forecasts, and which is most physically justified? Explain.

E54. Figs. 14.67 - 14.69 show 3 different ways to calculate CIN. Discuss the advantages and disadvantages of each, and for which conditions each method is most appropriate.

E55. Table 14-4 relates CIN to the difficulty in triggering thunderstorms. For the “strong” and “intense” cap categories, can you think of a trigger mechanism that would be powerful enough to trigger thunderstorms? Explain.

E56. Figs. 14.71 and 14.73 show how thunderstorms can be triggered by warm, humid air forced to rise over an obstacle (cold air, or a mountain). In both situations, sometimes after the thunderstorms are triggered and develop to maturity, environmental wind shear causes them to move away from the triggering location. After they have moved away, what is needed in order for the supercell thunderstorm to continue to exist (i.e., not to die immediately after moving away from the triggering area)? Do you think this is possible in real life, and if so, can you give an example?

E57. Consider Fig. 14.74. The Regional Winds chapter states that buoyancy (gravity) waves need statically <u>stable</u> air to exist without damping. But thunderstorms need nonlocally conditionally <u>unstable</u> air to exist. These seems contradictory. Explain how it might be possible.

E58. Fig. 14.75 shows a way to estimate T_{needed} for convection, given T and T_d from a sounding earlier in the day. Although this is an easy method, it has a major flaw. Critically discuss this method. (Hint: what was assumed in order to use this method?)

E59. Step back from the details of thunderstorms, and look at the big picture. After the thunderstorm has occurred, some precipitation has fallen out, and there has been net heating of the air. Explain how these two processes affect the overall stability and entropy (in the sense of randomness) in the troposphere.

E60. Re-draw Fig. 14.76, but for a winter case over mid-latitude land where nocturnal cooling is longer duration than daytime heating, and for which the total accumulated heating during the day (because the sun is lower in the sky) is less than the accumulated cooling at night. Use you resulting figure to discuss the likelihood of thunderstorms in winter.

E61. Recall from other chapters that the ocean has a much large heat capacity than land, and thus has less temperature change during day and night. Redraw Fig. 14.76 for a location at a tropical ocean, and discuss how the resulting figure relates to thunderstorm occurrence there.

E62. Some of the older stability indices in Table 14-6 are very similar in basis to some of the newer indices in Table 14-5, as described throughout this chapter. The main differences are often that the old indices consider meteorological conditions at just a few levels, whereas the newer indices integrate (sum) over many levels. Create a table matching as many old indices to as many new indices as possible.

E63. Based on what you learned so far, which subsets of indices from Table 14-7 would you prefer to utilize in your own storm forecasting efforts, to forecast:

a. hail
b. heavy rain
c. lots of lightning
d. tornadoes
e. strong straight-line winds
f. downbursts

Justify your choices.

14.10.4. Synthesize

S1. Suppose that thunderstorms were typically 30 km deep. How would the weather and general circulation differ, if at all?

S2. How would general circulation differ if no thunderstorms exist in the atmosphere?

S3. Draw a supercell diagram similar to Fig. 14.4b, but for mid-latitudes in the southern hemisphere.

S4. Why do thunderstorms have flat, anvil-shaped tops, while cumulus congestus do not? Both reach their equilibrium level.

S5. Suppose that N. Hemisphere mid-latitude thunderstorms exist in an environment that usually has the same lower-tropospheric geographic arrangement of heat and humidity relative to the storm as our real atmosphere, but for which the winds aloft are from the east. How would thunderstorms differ, if at all?

S6. Suppose that thunderstorms never move relative to the ground. Could long-lasting supercells form and exist? Explain.

S7. In multicell storms, new cells usually form on the south side of the storm complex, closest to the supply of warm humid boundary-layer air. What changes in the environment or the thunderstorm might allow new cells to form on the north side of the storm? Assume N. Hemisphere.

S8. If orographic thunderstorms need the mountain-triggered lifting to be initiated, why can the storms persist after being blown away from the mountain?

S9. If the Earth's surface were smooth (i.e., no mountain ranges), then describe changes in the nature of thunderstorms. Could this alter the weather in your region? Explain.

S10. Suppose a 100 km diameter circular region of warm humid boundary layer air existed, surrounding by much colder air. If the cold air all around the circle started advancing toward the center of the circle and thunderstorms were triggered along this circular cold front, then describe the evolution of the thunderstorms.

S11. In Fig. 14.14 showing a vertical slice through a mesoscale convective system, the arrows show winds being drawn in toward low pressure in the mid troposphere, and other arrows blowing outward from high-pressure in the upper troposphere. Explain how this could happen, because normally we would expect winds to circulate around highs and lows due to Coriolis force, and not to converge or diverge (assuming no frictional drag because we are not in the boundary layer).

S12. Could a classic supercell change into a low-precipitation or a high-precipitation supercell? Explain what factors might cause this, and how the storm would evolve.

S13. If there was never a cap on the atmospheric boundary layer, explain how thunderstorms would differ from those in our real atmosphere, if at all.

S14. Start with the sounding in Fig. 14.24. Modify the dew point in the mid troposphere to create a new sounding that would support a layer of altocumulus castellanus (accus) clouds.

S15. Suppose that thunderstorm downdrafts could never penetrate downward through the cap at the top of the boundary layer. Explain how thunderstorms would differ, if at all?

S16. Suppose that shading of the ground by clouds would cause the Earth's surface to get warmer, not cooler during daytime. How would thunderstorms and climate differ, if at all?

S17. Suppose the atmospheric sounding over your town showed conditions nearly, but not quite, favorable for the existence of thunderstorms. If you could cause the surface energy balance over land to be partitioned differently between sensible and latent heat, how would you partition it in order to generate a thunderstorm? Note that given a fixed input of energy from the sun, increasing either the temperature or the humidity would decrease the other variable.

S18. When water vapor condenses in thunderstorms, suppose that the air cools instead of warms. Would thunderstorms occur? If so, describe their behavior.

S19. In the Exposition of rising air parcels in a nonlocally conditionally unstable environment, we assumed that the surrounding environment was not changing. But if there are many air parcels rising in a thunderstorm updraft, then there must be compensating subsidence in the environment that advects downward the temperature and moisture layers. How would this alter our description of the evolution of thunderstorms?

S20. Under what conditions would the median of a distribution exactly equal the mean? Under what conditions would half the interquartile range exactly equal the standard deviation?

S21. Consider Fig. 14.34. Suppose that when the air parcel is above its LFC, that it entrains environmental air at such a high rate that the air parcel follows a thermodynamic path that is exactly half way between the environment and the moist adiabats that passes through it at each height during its rise. Sketch the resulting path on a copy of Fig. 14.34, and discuss how thunderstorms would differ, if at all.

S22. Consider Fig. 14.34. Suppose that there is no frictional drag affecting the rising air parcel. All of the CAPE would lead to kinetic energy of the updraft. Once the air parcel reaches its EL, inertia would cause it to continue to rise (i.e., overshoot) until its negative potential energy (by being colder than the environment) balanced the initial kinetic energy. Assuming an isothermal sounding above the tropopause in that figure, determine exactly how high the overshooting air parcel would rise. Also, discuss whether such behavior is likely in real thunderstorms.

S23. Suppose that no wind shear existed in the environment. How would thunderstorms differ, if at all?

S24. Suppose that the wind profile of Fig. 14.53 (based on the wind data tabulated in a Sample Application a couple pages earlier) corresponded to the same environment as the thermodynamic sounding data in Fig. 14.24. Draw "phase-space" plots as follows:

a. T vs. U b. T vs. V c. T vs. M
d. T_d vs. U e. T_d vs. V f. T_d vs. M
g. T vs. T_d

(Hint: "phase-space" plots are explained in the Numerical Weather Prediction chapter as a way to help analyze the chaos of nonlinear-dynamics systems.)

Also, speculate on whether phase-diagram plots of real thunderstorm environments could be used to help forecast different types of thunderstorms.

S25. Suppose that moist adiabats curve concave upward instead of concave downward. How would thunderstorms be different, if at all?

S26. Suppose that the data plotted in the hodograph of Fig. 14.62b were everywhere below the origin of the hodograph. How would thunderstorms differ, if at all?

S27. Suppose that right-moving supercells altered the environment so that it favored left-moving supercells, and left-moving supercells altered the environment to favor right movers. How would thunderstorms differ, if at all?

S28. Suggest 3 or more thunderstorm trigger mechanisms beyond what was already discussed in this chapter.

S29. From the Precipitation chapter, recall the Wegener-Bergeron-Findeisen (WBF) cold-cloud process for forming precipitation. Describe the nature of thunderstorms if the WBF precipitation did not occur.

S30. Suppose that accurate thunderstorm (including hail, lightning, and tornado) warnings could be issued 2 days in advance. How would society and commerce change, if at all.