Part 8. Conceptual Models of Mesoscale Convective Systems (MCSs)

In this chapter we focus on thunderstorm systems, which are larger in horizontal scale than the multicellular, or supercell thunderstorms we discussed in the preceding chapters. We refer to these systems as *mesoscale systems* because they are too small to be captured by the routine upper-air sounding network (called synoptic observations), yet they are too large to be adequately sampled by special field experiments designed to look at individual thunderstorms. Thus, observation of mesoscale convective systems requires large-scale, coordinated field. Knowledge of mesoscale convective systems is still quite limited, since we have only been able to observe them for a short period of time during special thunderstorm-scale experiments. Moreover, because they contain distinct thunderstorm elements, they are a challenge to modelers, since the models must cover a domain of several hundred kilometers yet have fine enough resolution to simulate the thunderstorm elements properly.

**Squall-Line Thunderstorms**

*Ordinary Squall Lines*

Mesoscale convective systems organize into a variety of configurations, the easiest to recognize is the squall line. Squall-line thunderstorm systems occur throughout the tropics and mid-latitudes. An observer at the surface normally sees a sharp roll-like line of clouds followed by a sudden wind squall or gust of 12 to 25 m/s. Immediately behind the surface squall a heavy downpour starts, which may produce as much as 30 mm of rain in 30 minutes in the tropics. Often the heavy downpour is followed by several hours of steady rainfall from the stratiform anvil cloud that trails the squall front.

Figure 8.1 illustrates a low-level radar depiction of an ordinary tropical squall line. Tropical squall lines are somewhat bow-shaped rather than being perfectly straight lines. Tropical
squall lines are typically embedded in easterly flow with the strongest easterly winds at lower levels. As a result, the low-level surface squall travels faster than the upper-level cloud debris spewed from the rising convective towers at the squall line. Thus the highest radar reflectivities occur along the leading edge of the line, where convective updrafts produce intense rain showers. A vertical cross section through the squall line is illustrated in Figure 8.2. This shows the strong convective updrafts along the leading edge with pronounced flow from the front of the storm at low levels toward the rear of the storm at middle and upper levels. This front-to-rear flow transports moisture, cloud droplets, ice crystals and graupel particles toward the rear of the system. The faster-falling, heavily-rimed graupel particles settle out rather quickly just behind the convective line. The slowly
settling ice crystals and moist air lags behind the rapidly advancing squall line and forms the deep and widespread stratiform anvil cloud. Latent heat released during the vapor depositional growth of the snow crystals and freezing of supercooled raindrops warms the air in the middle and upper troposphere, causing slow, rising motion in those regions. The slow, rising motion causes adiabatic cooling, which produces more condensation, fueling further the formation of precipitation in the trailing stratiform anvil region.

The vertical divergence (or change in magnitude with height) of longwave radiation at the top of the stratiform anvil cloud also contributes to the upward motion in the stratiform anvil cloud. We have seen previously that all bodies emit radiation, and the amount of energy emitted is greater, the higher the temperature of the emitting body. For the temperature of the earth's surface and the air in the troposphere, the emitted electromagnetic energy is in the infrared region. An important property of infrared radiation is that it is not absorbed appreciably in cloud-free air. As a result much of
the infrared radiation emitted at the earth's surface passes through the atmosphere and escapes out to space (see Figure 8.3a). When clouds are present the situation changes rather dramatically, as clouds are excellent absorbers of infrared radiation. Hence, upwelling radiation emitted from the earth's surface is absorbed at the base of the stratiform anvil cloud, where it is re-emitted downward towards the earth's surface and upward to levels of the cloud above where it is further absorbed and re-emitted. Radiative warming of a layer in the atmosphere is caused by a net gain in radiant energy at a given level in the atmosphere. Thus, as illustrated in Figure 8.3b, at the base of the stratiform anvil cloud, more energy is received from the radiative energy upwelling from the earth's surface than is radiated downwards from cooler cloud layers immediately above. As a result, longwave radiation causes modest warming near the base of stratiform anvil clouds. Because clouds are excellent absorbers of infrared radiation, little radiative heating occurs through much of the cloud layer. Each level in the interior of a cloud is exposed to upwelling and downward radiation from cloud levels immediately above and below, which are nearly at the same temperature as

Figure 8.3. Illustration of infrared radiant energy absorbed and received at different levels in the atmosphere. Left panel illustrates a cloud-free atmosphere. Right panel illustrates a stratiform-anvil cloud layer. Length of arrows is proportional to the amount of radiant energy per unit area.
the given cloud level and therefore radiate about as much energy as the cloud level in question. Layers in the interior of a cloud, therefore, receive no significant net gain or loss of radiation and do not warm or cool appreciably. At cloud top, however, the situation changes dramatically. Here, large amounts of longwave radiation energy are emitted from cloud layers immediately below. The air above the cloud top is a poor absorber of infrared radiation, and therefore the amount of energy re-radiated downward from the cloud-free atmosphere is very little. As a result, the cloud layers near cloud top experience a net loss in radiation, producing a cooling of the cloudy air. A consequence of radiation cooling near cloud top and warming near cloud base is that the stratiform-anvil cloud becomes unstable, thus contributing to convective overturning in the cloud layer and further warming by latent heat release.

The destabilization of the stratiform-anvil cloud is greatest during the nighttime. This is because, during the daytime, short-wavelength solar radiation is absorbed at upper levels in the cloud. The amount of solar energy absorbed near cloud top, however, is not as great as the net loss in infrared radiation. This is because cloud droplets are good reflectors of solar radiation, so some of the solar energy is reflected back to space. Also, solar energy is not as efficiently absorbed as infrared radiation, contributing to a penetration of solar energy deeper into the cloud interior. Nonetheless some solar energy is absorbed near cloud top, during the day, warming the cloudy air and somewhat offsetting the cooling by infrared radiation. At night, the absence of solar radiation leads to greater net radiative cooling at cloud top and greater destabilization of the stratiform-anvil cloud.

The combination of destabilization of the anvil cloud by radiation and latent heating from condensation on cloud droplets, freezing, and vapor deposition growth of ice crystals creates warming in the middle and upper troposphere. As a result, gentle, rising motion is created. Cooling of air by melting of precipitation particles and evaporation in the sub-saturated layer beneath the melting level contributes to descending motion in the lowest layers. Between the rising air column at upper levels and the descending air column at low levels, the air column is stretched vertically, causing convergence or a layer of rear-to-front flow shown in Figure 8.2. A similar phenomenon occurs in the convective region along the squall front. There, buoyant air accelerates upwards at middle levels while, immediately behind the updraft region, strong, convective-scale downdrafts occur. Again,
This produces a vertical stretching of the air column at middle levels which, in turn, produces a region of low pressure between 2 and 4 kilometers. This low-pressure region also contributes to drawing middle level air from the rear of the squall line toward its front.

We see that an ordinary squall line is composed of two scales of motion: (1) the *cumulus scale*, having a horizontal dimension on the order of 2 to 10 km, and (2) the *mesoscale*, characterized by air motions on a scale of 20 to 200 km. The precipitation from squall lines reflects the presence of these two scales of motion, as about 60% of the precipitation is in the form of intense showers in the convective region and 40% is in the form of steady, stratiform rainfall from the trailing stratiform anvil region.

As a general rule, tropical squall lines are organized with a leading line of convective cells and a trailing, stratiform, anvil region. Occasionally, tropical squall lines exhibit both a trailing and leading stratiform region. Figure 8.4 illustrates a tropical squall line with a stratiform anvil that

![Figure 8.4](image)

*Figure 8.4.* Evolution of the vertical structure of squall system radar echoes. Cross sections are along propagation direction, with motion from right to left. Note the stratiform anvil cloud extending well in advance of the main convective cores. Shading thresholds are for the minimum detectable 24, 34, and 44 dB(Z). Small arrows denote front and back edge of convective zone. (From Houze, R.A., Jr., and E.N. Rappaport, 1984: Air motions and precipitation structure of an early summer squall line over the eastern tropical Atlantic. *J. Atmos. Sci.*, 41,).

extends well in advance of the convective line. It appears that such a leading stratiform region develops when the vertical shear of the horizontal wind is large at upper levels. We will see that this is a more common feature of middle-latitude squall lines.
The middle-latitude squall line shown in Figure 8.5 also exhibits a leading line of thunderstorms with a pronounced trailing anvil region. Typically, the low-level flow in the northern hemisphere is southeasterly to

Figure 8.5. Schematic cross-section through wake low (a) and surface pressure and wind fields and precipitation distribution during squall line mature stage (b). Dashed line in (a) denotes zero relative wind. Arrows indicate streamlines, not trajectories. (Adapted from Johnson, R.H. and P.J. Hamilton, 1988: The relationship of surface pressure features to the precipitation and air flow structure of an intense midlatitude squall line. *Mon. Wea. Rev.*, **116**, 444-1472.)
southwesterly, while the middle tropospheric flow is westerly and increases in speed with height through most of the troposphere. As in tropical squall lines, the trailing anvil region is characterized by a well-defined, front-to-rear flow in which the cloudy air is slowly rising with updraft speeds of a few tenths of a meter per second. This should be contrasted with the leading convective line, where updrafts are typically of the order of 10 to 20 m/s. Beneath the ascending front-to-rear flow is a descending rear-to-front flow that is driven in part by the evaporative cooling of raindrops settling from the stratiform cloud above, melting of precipitation, and by a downward-directed pressure gradient force. The causes of the downward-directed pressure force are still being debated by scientists, but it may be due to the horizontal spreading of a shallow layer of evaporatively cooled air beneath the convective and stratiform regions, which causes a net divergence of mass and lowering of pressure just above the cool-air layer. A similar phenomenon occurs in the convective regions, where downdrafts form by water loading, melting, and evaporation of precipitation, leading to 3 to 7 m/s downdrafts, which increase in speed toward the surface. Again, divergence of mass in the accelerating downdraft regions causes a lowering of pressure just behind the leading line of thunderstorms at the 2 to 4 km levels. This creates a pressure gradient force that causes dry air behind the convective line to feed and mix with the downdraft air, producing more evaporation and strengthening the low pressure region. Thus, thunderstorm downdrafts contribute to a horizontally-directed pressure gradient force, which is directed horizontally, forcing dry, cloud-free air to move from the rear of the system toward the convective line.

At the surface, the pressure field shown in Figure 8.5b is characterized by low pressure immediately ahead of the line. This is probably caused by slow, sinking motion in response to the strong upward motion in the convective region. The slowly sinking air warms adiabatically, lowering the surface pressure hydrostatically. Immediately behind the convective line is a pronounced region of high pressure which is caused by the evaporation and melting of strong convective precipitation in that region. Beneath the trailing stratiform anvil region is a low pressure region that is probably a result of adiabatic warming of the slowly sinking, cloud-free air in the rear-to-front flow.

A feature that distinguishes middle-latitude, continental squall lines from their tropical, oceanic counterparts is the strength of the convective-scale updrafts and downdrafts. In the tropics, where numerous thunderstorms
and mesoscale convective systems drive the stability of the atmosphere close to the wet adiabatic rate, clouds are not very buoyant and updrafts are typically 7 to 10 m/s. Also in the moist marine air, cloud-base heights are very low, being on the order of 500 to 600 meters above ground level. As a result thunderstorm downdrafts are shallow and do not obtain a great deal of negative buoyancy from evaporation of raindrops. Typically, downdraft strengths are only 2 to 3 m/s compared to 7 to 10 m/s in extra-tropical, continental storms. Thunderstorms over the High Plains of the United States, for example, often have cloud bases 2 to 3 km above a relatively dry sub-cloud layer. Downdrafts are therefore stronger and extend over greater depths. Also, because of the stronger surface heating over the interior continental regions and stronger wind shears in middle-latitudes, convective updrafts in squall-line storms can be very intense, being as high as 25 to 40 m/s. This means that the mass transports and resultant flow structures of middle-latitude, continental squall lines are more strongly influenced by thunderstorm-scale drafts and flow structures than their tropical cousins.

For example, consider the conceptual model of the ordinary squall line thunderstorm system shown in Figure 8.6. Like most ordinary squall lines, this squall line observed near Miles City, Montana, exhibited a bow-shaped leading squall front with thunderstorm cells aligned parallel to the squall front. Behind the leading squall line is a stratiform anvil region that extends several hundred kilometers. A two-dimensional cross-section perpendicular to the leading line through the cell labeled G1 is shown at the top of Figure 8.6. This cross section shows the characteristic ascending front-to-rear flow in the trailing anvil region and a descending rear-to-front flow. The three-dimensional flow field illustrated in Figure 8.6 reveals a more complicated structure with middle level air in front of the line flowing about the cell G1 and descending in the rear-to-front flow branch. Moreover, the air feeding the squall line is not a surface-based flow. Instead it glides over very stable-cool air left in the wake of an earlier supercell thunderstorm.
Figure 8.6. Schematic depiction summarizing the 2-D and 3-D flow features for the 2 August 1981 CCOPE squall line showing mesoscale outflow boundaries, surface streamlines (thin arrows), convective reflectivity structure (stippled), overriding flow (bold arrow), and storm relative flow (thin ribbons). The vertical cross-section corresponds to a representative depiction of the storm core G1 and shows reflectivity (thin solid lines), schematic storm relative flow (thin arrows), and location of the midlevel upshear inflow (shaded). The bold A refers to an isolated supercell system, G1 and F2 represent the cell groups along the squall line and the bold H represents the location of the surface mesohigh. Labeling of the storm relative flow ribbons refers to height AGL. (From Schmidt, J.M. and W.R. Cotton, 1989: A high plains squall line associated with a derecho. *J. Atmos. Sci.*, **46**, 281-302.)

Clearly, the flow field is much more complicated than the two-dimensional depictions of ordinary squall lines shown previously. Part of the complicated flow structure arises from the fact that this storm is not isolated, but is overtaking a more slowly moving supercell. Another reason for its more complicated flow structure is that the storm formed in an environment
with strong vertical shear of the horizontal wind through a large depth of the troposphere. This strong vertical shear was responsible for the formation of the supercell storm that preceded the squall line. Furthermore, the strong shear was responsible for creating the supercell-like storm in the squall line labeled G1. Cell G1 was a steady cell that lasted several hours, contained a rotating updraft structure, and exhibited a bounded weak echo region (BWER). Other cells along the line labeled F2 were ordinary thunderstorm cells having lifetimes of 35 min to one hour. The supercell-like storm G1 so dominated the squall line flow fields that much of the complicated flow structure in Figure 8.6 can be attributed to its presence.

This middle-latitude squall line was also noteworthy because it produced a nearly continuous swath of severe surface winds covering a four-state area from Montana to Minnesota. Such a severe, straight-line, wind-producing storm system is called a “derecho.” Derechos are noted for producing swaths of severe wind damage 100 km or more in width and extending 500-1000 km in length. The processes responsible for producing such extensive wind damage is unknown at this time.

Pre-Frontal Squall Lines

The largest and most violent squall lines are the pre-frontal lines that form in middle latitudes. Typically, they form along, or ahead of, a cold front associated with a vigorous, midlatitude cyclonic storm. Often, the line forms in the warm sector of the cyclone, where the low-level jet brings warm, moist air into the region ahead of the cold front. As noted previously, this region of the cyclonic storm is ideal for manufacturing severe storms; it contains plenty of fuel to sustain intense thunderstorms and strong vertical shear of the horizontal wind, which helps organize the thunderstorms into efficient machines. Scientists still don't understand why, under some conditions, a few isolated severe thunderstorms may form, while at other times a severe pre-frontal squall line may form that has a long dimension of over 1000 km and may produce severe weather over much of its length. Figure 8.7 illustrates a pre-frontal squall line observed on 20 May 1949. The squall line is ahead of the surface front and extends from the warm sector of the cyclone on its southern flank to north of the surface warm front. The squall line is depicted relative to the middle troposphere pressure chart (500 mb) in Figure 8.7a and the lower tropospheric pressure chart (850 mb) in Figure 8.7b. An example of a very long, pre-frontal squall line is
Figure 8.7. (a) 500-mb (middle troposphere) and (b) 850-mb (lower troposphere) charts 2100 CST 20 May 1949. Solid lines, contours (hundreds of ft); dashed lines, isotachs (knots); thin double lines, surface fronts. Blacked-in areas, rainfall in excess of 0.20 in per hour; inner light areas, 0.50 in or more. (From Newton, C.W. and H.R. Newton, 1959: Dynamical interactions between large convective clouds and environment with vertical shear. *J. Meteor.*, 16, 483-496.)

shown in infrared satellite imagery at two times in Figure 8.8. This particular squall line extends from southern Texas northeastward to the Illinois, Great Lakes region!
Figure 8.8. (a) Satellite infrared image showing the squall line at 2000 CDT fairly early in life. (b) As in (a) but at 2330 CDT showing increase in width of cloud shield. (From Srivastava, R.C., T.J. Matejka, and T.J. Lorello, 1986: Doppler radar study of the trailing anvil region associated with a squall line. *J. Atmos. Sci.*, 43, 356-377.)
A schematic cross section perpendicular to a severe pre-frontal squall line is shown in Figure 8.9. The flow structure is similar to the ordinary squall line shown in Figure 8.5 except that the storm is deeper and the updraft and downdraft strengths are much greater. Due to the strong shear in the pre-frontal air mass, the pre-frontal squall line exhibits a more pronounced leading stratiform-anvil than the ordinary squall line. At low levels, a layer of warm, moist air feeds the thunderstorm updrafts. Intense updrafts extend through much of the troposphere and even penetrate to great heights in the lower stratosphere. Some of the updraft air encounters large quantities of precipitation where water loading, melting, and mixing with dry air entering the rear of the storm contribute to negative buoyancy. This creates a vigorous convective-scale ‘up-down’ downdraft component and the updraft air is forced downward before it reaches the middle troposphere.

![Figure 8.9](image)

**Figure 8.9.** Schematic cross-section perpendicular to a severe, pre-frontal squall line. Slant-left shading represents depth of warm, moist air feeding the thunderstorms. Slant-right shading indicates layer of dry middle level air that becomes cold and negatively buoyant as cloud droplets and raindrops evaporate in it. (Adapted from Newton, C.W., 1966: Circulations in large sheared cumulonimbus. *Tellus*, 18, 699-712.)

At middle levels, dry air enters the storm from the rear, where it mixes with cloudy air. Raindrops then settle into the dry air causing evaporative cooling and general sinking motion. A feature of such intense, squall-line thunderstorms is the notable tilt of the updrafts upshear (in this case upwind), which leads to a more efficient storm system. This is because precipitation settling out of the updrafts falls into the downdraft regions,
thus increasing their strength rather than settling primarily into the updrafts and weakening them.

Frequently, the thunderstorm cells along the pre-frontal squall line are of the supercell variety, capable of producing tornadoes, grapefruit-size hailstones and wind damage. Because of the great length of the squall line, pre-frontal squall lines are responsible for the major outbreaks of severe weather. Fortunately, the conditions favorable for generating the severe, pre-frontal squall line occur rather rarely. It remains a challenge for forecasters to predict when such severe, pre-frontal squall lines will develop, how long they will last, the direction and speed they will move, and the locations of the severe thunderstorm cells. Moreover, because pre-frontal squall lines are so elusive and cover such a large region, it is difficult for meteorologists to implement a field experiment that can observe these storms. Our knowledge of these systems thus far has been a result of the chance passage of a small portion of a line through an instrumented network.

Cloud Clusters and Mesoscale Convective Complexes

Mesoscale convective systems are characterized by the individual thunderstorms organized in a variety of configurations, the squall-line organization being the most distinct. Cloud clusters are mesoscale systems, where the convection is organized in a rather random configuration. The larger cloud clusters may have squall lines and other thunderstorm units as part of the overall composition of the system. In some tropical clusters, lines of convection may be prevalent, but the orientation of the lines may be parallel to the wind shear vector rather than perpendicular to it as in most squall lines.

A distinct feature of the larger clusters is the large area of nearly circular, stratiform anvil cloud. A typical lifecycle of a tropical cluster begins with the formation of convective cells. This is followed by the mature stage, in which convective cells and a stratiform anvil cloud coexist with both the convective clouds and the stratiform clouds contributing nearly equally to surface precipitation. The dissipation stage of a cloud cluster is characterized by the dominance of the stratiform cloud and the slow decay of steady precipitation. Tropical, oceanic clusters exhibit a diurnal variation in rainfall with a maximum in the early morning hours and a minimum in the early evening. Radiative processes are believed to be responsible for the diurnal variation.
Cloud clusters are quite common in the tropics and are largely responsible for producing the bulk of tropical rainfall. They are particularly common in a preferred region of low-level convergence near the equator, called the intertropical convergence zone that shifts north and south of the equator depending on season. Some tropical clusters eventually develop into tropical cyclones or hurricanes. The few tropical clusters that develop into hurricanes are essentially indistinguishable from the more common clusters that have a much shorter lifecycle.

Mesoscale convective systems which are not organized in a squall line structure are also quite common in middle latitudes. They range in size from the scale of slightly larger than a multicellular thunderstorm, having three or four cells to the scale of a mesoscale convective complex or MCC. The MCC is defined in terms of the scale and duration of the storm as identified from infrared satellite imagery. Figure 8.10 illustrates a satellite image of an MCC at its mature stage. An MCC is defined as a mesoscale convective system that exhibits infrared cloud top temperatures colder than -32° C over an area exceeding 100,000 km², an interior cloud top temperature colder than -52° C with an area greater than 50,000 km² for a period of six hours or more. The MCC also has an eccentricity (minor axis/major axis) greater than 0.7 at the time the cloud shield exhibits its maximum extent. The eccentricity criterion eliminates the large, squall-line, mesoscale convective systems such as the prefrontal squall line as being MCC's. Smaller squall lines are often embedded beneath the nearly circular cloud shield that is characteristic of an MCC. In some cases, the organization of the thunderstorm cells is a consequence of the early genesis stages of the MCC. Figure 8.11 illustrates the evolution of an MCC whose roots can be traced to the Rocky Mountains. In this case, a line of cells formed along the length of the Colorado Rocky Mountains. This line of cells subsequently moved over the High Plains, where it encountered low-level moisture that is usually brought into the High Plains by a low-level jet that reaches maximum strength in the nighttime hours. At the same time, the mountain line of cells overtook a cluster of cells (labeled G, E, and D) that formed over the Plains. Thus, the system at its mature stage (see Figure 8.11e) can be seen to be composed of cells that formed initially over the mountains and cells that formed over the plains.
Figure 8.10. Enhanced infrared (IR) image of the United States at 0300 MDT 4 August 1977, from the GOES satellite at 70\deg W longitude. The stepped gray shades of medium gray, light gray, dark gray and black are thresholds for areas with apparent blackbody temperatures colder than -32, -42, -53 and -59\deg C, respectively. Temperatures progressively lower than -63\deg C appear as a gradual convective development of 4 August over the Colorado mountains. The intense meso-\alpha-scale convective complex (MCC) centered over eastern Kansas originated in the eastern Rockies and western plains the previous evening. (From McAnelly, R.L., and W.R. Cotton, 1986: Meso-\beta-scale characteristics of an episode of meso-\alpha-scale convective complexes. Mon. Wea. Rev., 114, 1740-1770.)
Figure 8.11. Schematic infrared (IR) satellite and radar analysis at 2-h intervals, from 01 to 11 UTC 4 Aug 1977, for the western MCC #1. The anvil cloud shields are indicated by the -32 and -53º C IR contours (outer and inner solid lines, respectively), re-mapped from satellite images at the labeled times. Darkly shaded regions (identified by letters) denote significant radar-observed, mesoscale convective features at about 25 min after the indicated whole hour, with the vectors showing their previous 2-h movements. The dashed line segments extending from the mesoscale convective features indicate flanking axes of weaker convection. In the more developed MCC stages, in (e) and (f), the light-shaded area within the dashed envelope indicates weaker, more uniform and widespread echo.

In other cases, MCC's develop a thunderstorm cell organization that resembles a miniature extratropical cyclone. Figure 8.12 illustrates a MCC, in which a north/south oriented squall line is joined by a nearly east/west

Figure 8.12. Mesoscale surface analysis of the mature system B at 0100 UTC. Pressure centers, streamlines, convergence lines (dotted), and fronts superimposed on reflectivity pattern (shaded) from Wichita radar, and the -54\degree C contour of the margin of the cloud shield (thin dashed).

oriented line of cells that forms a wave-cyclone pattern. Evidence of the wave-cyclone pattern can also be seen in the surface pressure field and in the upper-level wind fields. An idealized schematic of the flow structure of a MCC resembling a small wave-cyclone is shown in Figure 8.13. You are
Figure 8.13. Perspective drawing of an idealized, weakly rotating mesoscale convective complex as viewed from the rear. The apex region is located where the heavier rain is drawn. The cumulonimbus towers at right form on the mesoscale cold front. The mesoscale warm front and the forward part of the stratiform region are not shown in this view. (From Fortune, M.A., 1989: The evolution of vortical patterns and vortices in mesoscale convective complexes. Ph.D. Dissertation, Colorado State University, Dept. of Atmospheric Science, Fort Collins, CO 80523.)

viewing the storm from the rear, southwest flank. The airflow structure is composed of three dominant airstreams which we call, conveyor belts. Ahead of the storm, the low-level jet serves as a warm conveyor belt bringing warm, moist air from the south into the region of active convection. The warm, moist air rapidly ascends in cumulonimbus updrafts along the north-south oriented cold front. The warm conveyor belt also glides over the east-west oriented large-scale cold front, and some of it ascends rapidly in convective updrafts, while some of the air gently ascends in slantwise ascent.

Behind the east/west-oriented cold front is another low-level airstream that we call the cold, conveyor belt. This airstream is confined vertically between the ground and the gently ascending warm conveyor belt. The warm conveyor belt creates a strong inversion in
temperature, thus, serving as a lid to convective overturning in the cold conveyor belt airstream. The cool conveyor belt air is frequently quite moist as steady, stratiform rainfall settles into the layer and moistens it as it evaporates.

Figure 8.13 illustrates a third airstream which originates to the northwest of the storm at the 7-10 km level. This airstream, which we call the dry airstream descends as it enters the storm. It is the counterpart of the middle-level rear-to-front flow that we have seen is common in squall line systems.

Because the horizontal scale of an MCC is 300 to 400 km, air flowing into the storm in the warm conveyor belt responds to the effects of the earth's rotation and begins to turn to the right of its direction of motion in the northern hemisphere. Likewise, air entering the storm from the rear in the dry airstream curves to the right as it descends and spreads out in the surface-based cool pool. This produces a cyclonic or counterclockwise rotating flow in the northern hemisphere at middle levels. Near the top of the troposphere, upward moving air encounters the extreme stability of the tropopause and diverges outward from the storm center. The outward-moving air again travels several hundred kilometers from the storm center and experiences the effects of the earth's rotation. As a result the diverging air turns to the right of its direction of motion and turns anticyclonically or clockwise in the northern hemisphere.

Like squall lines, an MCC is composed of both convective updrafts and slowly ascending motion. Instead of primarily trailing the convective line as in squall lines, the MCC experiences slow ascent in the middle and upper troposphere in a broad region surrounding the main convective cores. The lifecycle of MCC's is such that early in their development, thunderstorm-scale motions prevail. The rainfall is quite showery at this point and there is a great likelihood that the system will produce severe weather, such as tornadoes, hail and flash floods. As the system reaches maturity and produces a stratiform-anvil cloud of maximum horizontal extent, the rainfall becomes light and steady, or more stratiform in character. At this point, the chance that the storm will produce tornadoes, hail, or very intense rainfall diminishes, although the storm produces the maximum volume of rainfall at its mature stage (see Figure 8.14). The storm system slowly decays, producing lighter and lighter rainfall. Some storms may last for 10 to 12 hours. The cyclonic circulation that
has developed early in the storm's lifecycle makes the storm act as a giant flywheel, slowly spinning down as the convection decays. Some storms

**Figure 8.14.** Bulk precipitation characteristics for 122 case composite MCC. Precipitation area (A, solid), average rain intensity (R, dashed), and volumetric rain rate (V, bars). (From McAnelly, R.L. and W.R. Cotton, 1989: The precipitation life cycle of mesoscale convective complexes over the Central United States. *Mon. Wea. Rev.*, **117**, 784-808.)

exhibit a circulation in the middle-level clouds (see Figure 8.15) after the deep convection has died and the upper-level stratiform anvil cloud has evaporated. In some cases, the middle-level, circulating cloud has been observed to spawn deep convection on the following day after the active
MCC has dissipated. Some systems that have formed over the Rocky Mountains have been observed to move eastward as far as the Atlantic Ocean, spawning new convection for two successive days.

![Visible satellite photograph](image)

**Figure 8.15.** Visible satellite photograph (1530 UTC, 14 May 1984) of decaying MCC showing circulation remaining long after active convection has ceased.

We have mentioned that MCCs produce sporadic bouts of severe weather. The dominant characteristic of MCCs is the rainfall they produce. Over the High Plains of the United States, MCCs are responsible for producing in excess of 50% of the rainfall that occurs during the growing season. They are, of course, also responsible for producing some of the most damaging flash flood events. What is surprising is that as many as 25% of all MCCs produce severe, straight-line, damaging winds or what we have called *derechos*. 
MCC’s are most common from May through the middle of August in the United States. Figure 8.16 illustrates that they are more common in the southern High Plains in April and May and move northward into the central High Plains in summer months. The tracks of MCCs illustrate the importance of the mountains in the genesis of many of the storms. Mountains are by no means necessary for the formation of MCCs, but they obviously play a role in their formation. Other regions in which MCCs are prevalent are in the lee of the Tibetan plateau in China, and in the lee of the Andes in South America.

**Genesis of Mesoscale Convective Systems**

The genesis of mesoscale convective systems requires sustained forcing from the larger-scale environment in order to trigger their formation. Generally, this is provided by low-level convergence of moist air. Some sources of low-level convergence are land- and sea-breeze circulations over tropical islands and peninsulas, mountain slope circulations, especially over large mountain ridges, convergence along large-scale, frontal boundaries in middle latitudes, and convergence along the inter-tropical convergence zone near the equator and in tropical easterly waves. The low-level convergence provides the fuel necessary to support the initiation of a number of cumulonimbus clouds in close proximity.

One of the first signs of the initiation of a mesoscale convective system is the merger of the anvil clouds emitted by the neighboring thunderstorms. At the same time, a cold pool of air forms near the surface due to cooling of the air by evaporation and melting of precipitation. Both processes appear to be important in the genesis of mesoscale convective systems. The merger of the anvil clouds results in sustained heating in the middle and upper atmosphere from the latent heat released by condensation on cloud droplets and by freezing and vapor deposition on ice crystals in the slowly rising air. The system will experience little warming, however, unless the stratiform anvil cloud is precipitating. This is because latent heating in rising updrafts will be compensated by evaporation of droplets in descending air or decaying cloud. The water precipitated from a storm will not be available for evaporation at middle and upper levels and thereby contribute to a net heating of the storm. Therefore, heating in the middle and lower troposphere is proportional to the precipitation rate from the storm.
Figure 8.16. MCC tracks for 1979 through 1983, for May, June, and July. Dotted lines indicate early storm development; the heavy lines define the mature MCC. The circle numbers correspond to the position of the maximum extent of the -32\degree C cloud shield from the infrared satellite data. The "x" corresponds to the "terminate" stage or the location where the MCC no longer met the requirements to be defined as an MCC. (From Cunning, J.B., 1986: The Oklahoma-Kansas Preliminary Regional Experiment for STORM-Central. *Bull. Am. Met. Soc.*, 67, 1478-1486.)
Latent heating due to condensation on cloud droplets and freezing and vapor deposition growth of precipitating ice crystals also occurs in the vigorous updrafts of the thunderstorms, but the heating is concentrated in small areas where the cells exist. Moreover, the heating is concentrated at low levels where the greatest amount of condensation and precipitation occurs. Thus, the widespread heating in the middle troposphere in the stratiform anvil clouds is more favorable for the formation of a middle-tropospheric, low-pressure region which drives a mesoscale circulation than does the localized heating in the thunderstorm updraft regions.

Near the earth’s surface, the cold pool of air formed by evaporation and melting of precipitation forms a bubble of high pressure. The cool air spreads outward from the raining regions in a shallow layer near the surface. At the interface between the cold pool and the warm, moist, environmental air, a horizontal pressure gradient is established, which is directed outward from the mesohigh. The convergence driven by this low-level pressure gradient lifts the low-level moist air. The meso-low aloft further lifts the moist air and draws it into the interior of the embryonic mesoscale system. As the volume of the evaporatively cooled low-level air gets larger, and the scale of the stratiform-anvil cloud with its associated latent heating and low pressure gets larger, air is drawn into the mesoscale system from increasingly larger distances. Eventually the air will be drawn in from large enough distances that it will experience a turning to the right (in the northern hemisphere) due to the earth’s rotation. At the same, time air drawn at middle levels into mesoscale and convective scale downdrafts also begins to turn to the right due to the earth’s rotation. Thus, the air drawn into the system will begin to turn cyclonically and the system will exhibit slow rotation.

The horizontal scale at which cyclonic rotation develops changes with latitude, since the Coriolis force varies with latitude. At low latitudes, the mesoscale convective system must be much larger or the inflow much stronger before the air drawn into the system will experience cyclonic rotation than at higher latitudes.

In the upper troposphere, the slowly rising air in the stratiform anvil cloud eventually encounters the tropopause, which is quite stable. The ascending air turns horizontally and spreads outward from the center of the system. Moreover, the rising air cools adiabatically and, because the air can retain little moisture at the cold temperatures in the upper troposphere, it
receives little heating from condensation and freezing of ice crystals. The upper part of the system thus becomes cooler than the surrounding environment, and a meso-high develops at the top of the system. The air thus spreads outward from the center of the storm at upper levels, and, if it spreads far enough, it will begin to turn to the right of its direction of motion (in the northern hemisphere) and form an anticyclone aloft. The upper part of the embryonic system becomes characterized by anticyclonic motion or clockwise rotation in the northern hemisphere.

The development of rotation is an important feature of a mesoscale convective system. Prior to developing rotation, the convective system is largely controlled by the static stability of the environment. Where static stability is weak, deep, intense convection will form. If the static stability increases, the convective system will immediately weaken and perhaps die. Moreover, in a non-rotating system, as precipitating, convective clouds heat the atmosphere, the rising air triggers gravity waves which propagate horizontally and vertically away from the cloud system. Thus, much of the heating and kinetic energy that is generated in the convective clouds is radiated away from the cloud system and does not contribute to the winds and circulations of the system in any consistent and sustained way.

As a mesoscale system develops rotation, however, it develops inertial stability, much like a flywheel in an internal combustion engine. As a consequence of its inertial stability a mesoscale convective system can move into a region of increased static stability and continue to survive as an organized system for an extended period, while a non-rotating system will rapidly die. Meteorologists identify the scale at which a mesoscale system develops enough rotation to be dominated by inertial stability as the Rossby radius of deformation, $\lambda_R$. $\lambda_R$ varies with latitude and static stability, being larger in lower latitudes than in higher latitudes. This means that an inertially stable mesoscale convective system must be much larger in tropical regions than in the extra-tropics.

Because gravity waves travel at a finite speed of 25 to 30 m/s in the troposphere, it takes a certain amount of time for them to radiate out of the domain of the mesoscale system. If a mesoscale system is larger than $\lambda_R$, then gravity waves emitted in the convective cells near the central core of the system will still be propagating within the system at the time that the winds begin responding to lower pressure in the system core and to the
earth's rotation. This means that the heating and kinetic energy associated with convective clouds will be more efficiently converted into a nearly-balanced, rotating circulation than in systems smaller than $\lambda_R$, where the heating associated with clouds is converted into gravity waves and radiated to space or long distances from the system.

Figure 8.17 illustrates the horizontal scale of various weather phenomena relative to the Rossby radius. In region I, we have

![Diagram showing scales of weather systems relative to Rossby radius](image)

**Figure 8.17.** Scale of weather systems relative to the Rossby radius of deformation. (Adapted from Frank, W.M., 1983: The cumulus parameterization problem. *Mon. Wea. Rev.*, **111**, 1859-1871.)

small-scale phenomenon, such as boundary layer turbulence and cumulus clouds. Region II is characterized by thunderstorms and smaller mesoscale systems, such as most squall lines and small clusters of thunderstorms. These are basically unbalanced systems which, as such, rapidly die when
convective forcing ceases. The systems in region III are characterized by balanced flow or inertially stable systems, so that a cessation in convection or other driving forces does not result in the immediate demise of the systems. The MCC fits in this category, as well as larger-scale systems, such as extra-tropical and tropical cyclones.

Remember, it is not the geometrical size of a system that determines whether or not a system is inertially stable. A large cluster of thunderstorms in the tropics may be unbalanced (lie in region II) whereas an MCC of similar geometric dimensions in middle latitudes will experience the effects of the earth's rotation and be a nearly balanced, inertially stable system.

Let us now consider the hypothetical genesis of a mesoscale convective system over the Rocky Mountains of the United States. We begin in the morning hours as air cooled near the surface of the earth becomes denser than air away from the mountains and gently slides down the mountain slopes in what is called a nocturnal drainage flow. For large mountain barriers, such as the Rocky Mountains, the descending drainage air moves far enough horizontally that it experiences the effects of the earth's rotation. In the northern hemisphere, the descending drainage air turns to the right, creating a weak, low-level anticyclone, or clockwise turning of the winds, about the mountain crest.

The first stage of genesis of a MCC begins at sunrise. Heating of the mountain slopes warms the air near the earth's surface causing gentle flow up the slopes; the counterpart to drainage flow. Cumulus clouds at this stage are ordinary cumulus and towering cumulus. Stage 2 (shown in Figure 8.18) is indicated by deep convection that first forms over the mountain ridgetop as the prevailing westerly flow, augmented by the slope circulation, advects moisture into the higher elevations, where it is lifted by the terrain to trigger cumulonimbi. As the deep convection moves eastward with the prevailing westerly flow, low-level, western slope moisture is carried along with it. When the cumulonimbus system migrates eastward it approaches a convergence zone. The convergence is caused by the intersection of eastern upslope flow with a mountain wave, caused by descent of stable air lifted on the windward slopes. All elements favoring explosive development of intense convection come together, including moisture, surface convergence,
Figure 8.18. Conceptual model showing flow field and position of convective elements at time deep convection forms (Stage 2). The surface topography is depicted by the black shading. The vertical axis is height in km above mean sea level (MSL) and the horizontal axis is west longitude. The stippled line represents the position of the plains inversion. Regions of cloud are indicated. Part a depicts the flow field with ground relative streamlines. Circles depict flow perturbation normal to plane. Part b depicts the pressure and temperature response. Pressure centers are depicted by solid closed contours and temperature by dashed contours. The length scale of 600 km (2LR) is indicated. (Tripoli, G., and W.R. Cotton, 1989a: A numerical study of an observed orogenic mesoscale convective system. Part 1. Simulated genesis and comparison with observations. *Mon. Wea. Rev.*, 117, 269-300.)
and pre-existing cumulus convection. As explosive development of an ensemble of mesoscale convection ensues, moisture advection from the eastern plains is enhanced, further invigorating convection.

Stage 3, (Figure 8.19) begins when an organized convective line forms along the eastern slopes. Until this time, the mountain and plains circulation was confined below 5 km MSL. Under dry conditions, when deep convection does not occur, the circulation remains below 5 km MSL and slowly migrates eastward in the presence of prevailing westerly flow. Under sufficiently moist conditions, deep convection organized in a mesoscale line develops and the circulation deepens to 12 km or the depth of the troposphere. In association with the deepened circulation cell, a pronounced thunderstorm outflow forms in the 10-12 km layer. The persistent upslope circulation forms a cyclone at low levels, and its outflow near the tropopause forms an anticyclone at upper levels.
Figure 8.19. Same as Figure 8.18, except for Stage 3. Flow is storm-relative. Individual parcel paths are given by dashed (updraft) and dotted (downdraft) lines. (From Tripoli, G., and W.R. Cotton, 1989a: A numerical study of an observed orogenic mesoscale convective system. Part 1. Simulated genesis and comparison with observations. *Mon. Wea. Rev.*, **117**, 269-300.)

Stage 4 (Figure 8.20) begins when the convective system moves from the mountains over the eastern plains, where a zone of suppression is encountered. (The zone of suppressed convection is a consequence of several topographically related phenomenon, which are beyond the scope of our presentation here.) As our embryonic mesoscale convective system moves onto the plains, it encounters greater sinking motion, as well as surface divergence, as the low-level air is forced up the mountain slope.
or away from the system core. Because the sinking motion warms the mid-level air and stabilizes the environment, upward motion within the system core collapses.

Figure 8.20. Same as Figure 8.18, except for Stage 4. (From Tripoli, G., and W.R. Cotton, 1989a: A numerical study of an observed orogenic mesoscale convective system. Part 1. Simulated genesis and comparison with observations. *Mon. Wea. Rev.*, 117, 269-300.)

The upper-level anticyclone and low-level cyclone persist, however, because they slowly adjust to Coriolis accelerations, forming a "flywheel" which we mentioned earlier. The collapse of the system also initiates a deep gravity wave of 150-200 km horizontal wavelength which propagates
laterally eastward and westward. The collapsing storm core, augmented by precipitation loading and evaporation, causes so much descending air that, after the condensate is exhausted, it warms adiabatically. The warmed core causes the upward motion to rebound, while at the same time the core has moved eastward out of the zone of suppression, where it encounters the influx of moisture-rich air from beneath the plains inversion. The convective system, therefore, intensifies and matures to mesoscale convective system proportions.

At this juncture, the system enters Stage 5 (Fig. 8.21). Until now, the system evolution is very site-specific and is a strong function of the particular properties of the Rocky Mountains' thermally driven slope/plains circulations and the availability of moisture unique to the larger-scale circulations of that region. Upon entering stage 5, the convective system enters a stage of its life cycle that is more generic in its mesoscale characteristics. As the system moves eastward, the mean-core circulation gradually strengthens as the kinetic energy generated by latent-heating is partially retained while the system strives toward a geostrophic balance. That only a minor fraction of the kinetic energy generated by latent heat is retained as a geostrophically balanced circulation is not surprising, since this is characteristic of systems smaller than the Rossby radius of deformation. Instead, the vast majority of energy is radiated vertically and horizontally as gravity wave energy.

Throughout stage 5, the convective core of the MCS remains localized at the western edge of the plains inversion. Short-lived convective cells residing over the plains inversion are triggered by gravity waves emitting from the system core, but they do not intensify, because they cannot tap the moist air below it. The western boundary of the plains inversion is continually eroded by the core circulation of the mesoscale system. Here, adiabatic cooling associated with the mesoscale ascent and turbulent mixing by the convection itself destroy the inversion interface. Precipitation is advected eastward, where its evaporation cools the air above the inversion and destabilizes the air on the westernmost edge of the inversion. Overall, the system moves at about the speed of the upper-tropospheric wind and exhibits a circulation that resembles the squall line described previously. Similar to the tropical and midlatitude squall lines we have examined, the
Figure 8.21. Same as Figure 8.18, except for Stage 5. (From Tripoli, G., and W.R. Cotton, 1989a: A numerical study of an observed orogenic mesoscale convective system. Part 1. Simulated genesis and comparison with observations. *Mon. Wea. Rev.*, **117**, 269-300.)

system forms a deep stratiform-anvil cloud aloft. A large part of the stratiform-anvil cloud extends in advance of the main convective core. The radiative effects of the anvil become more important as darkness approaches. Prior to this period, heating at cloud top from absorption of solar radiation largely offsets longwave radiation cooling at the cloud top. As night approaches however, longwave radiative cooling at cloud top and heating at the base of the stratiform layer lead to further destabilization of the
stratiform layer. The intensity of the mesoscale convective system, therefore, increases with peak vertical velocity occurring at around 8:00 p.m.

Stage 6 of the system (Figure 8.22) represents the transformation from an unsteady squall line-like system having a scale less than the Rossby radius of deformation to the more geostrophically balanced MCC system greater in scale than $\lambda_R$. Some significant transformations take place in the

Figure 8.22. Same as Figure 8.18, except for Stage 6. Also, the region of the low-level southerly jet is hatched. (From Tripoli, G., and W.R. Cotton, 1989a: A numerical study of an observed orogenic mesoscale convective system. Part 1. Simulated genesis and comparison with observations. *Mon. Wea. Rev.*, 117, 269-300.)
genesis of an MCC from an ordinary MCS as nighttime approaches. The surface begins to radiatively cool, and low-level conditional instability is reduced. Furthermore, as a low-level nocturnal inversion forms, convective updrafts no longer draw on surface air and, instead, begin to draw on air residing above the nocturnal inversion. At this point, the nocturnal, low-level jet of warm, moist air over the High Plains intensifies and fuels the convective updrafts. Heating in the convective regions and the stratiform anvil cloud further spins up the middle-tropospheric, cyclonic circulation and an MCC is born.