Mesoscale Meteorology: Mesoscale Convective Systems

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Formation and General Characteristics

Mesoscale convective systems are organized, persistent areas of multicell thunderstorms having a contiguous precipitation area of at least 100 km in length in at least one direction. Thus, MCSs are on the larger end, or meso-α-scale, of the mesoscale. On MCS spatiotemporal scales, the Coriolis force cannot be neglected. As we will see later in this lecture, the Coriolis force introduces three-dimensional structure where otherwise two-dimensional structure would result. There are several different phenomena that can be classified as MCSs, including squall lines, bow echoes, derechos, line echo wave patterns, and mesoscale convective complexes.

Mesoscale convective systems typically initiate during the local evening hours in response to the upscale growth of existing and/or widespread development of new deep, moist convection. They typically weaken during the local early morning hours due to increased environmental stability or decreased line-normal vertical wind shear (e.g., as associated with a veering and weakening lower tropospheric jet). In other words, MCSs weaken or dissipate as convective redevelopment along their advancing cold pools becomes progressively less likely. Mesoscale convective systems can form in a wide range of ambient environmental vertical wind shears, with 10-20 m s^{-1} in the lower troposphere most common, and most-unstable CAPE values. Later in this lecture, we will develop a relationship between the lower tropospheric vertical wind shear and cold pool intensity (itself a function of the environmental thermodynamic profile) to explain MCS maintenance.

Upscale growth is characterized by the merger of the cold pools from adjacent thunderstorms, such that a mesoscale corridor of focused lift to aid in subsequent convection initiation results. This is most common when the environmental vertical wind shear or upper tropospheric storm-relative winds have a significant component parallel to the boundary along which convection initiated, such that hydrometeor fallout is predominantly in the direction of a downshear thunderstorm. This can also occur when the environmental hodograph supports both supercell splitting and subsequent maintenance, such that interactions between left- and right-splits from adjacent thunderstorms are comparatively likely. This is most common for relatively straight hodographs with small clockwise curvature in the lower troposphere, such that the deep-layer vertical wind shear vector is roughly perpendicular to the boundary along which convection initiated.

The widespread development of new deep, moist convection predominantly occurs if large-scale forcing for ascent, such as that along an advancing cold front or with a broad lower tropospheric jet accelerating and ascending over a lower tropospheric baroclinic zone is strong enough to initiate convection over a wide region nearly instantaneously. The former is most common during the local afternoon and early evening, whereas the latter is most common in the local early to mid-evening.

Salient Structural Characteristics

Consider an MCS in an environment of predominantly lower tropospheric vertical wind shear. As the best balance between this vertical wind shear and that associated with baroclinic generation in the MCS cold pool is found when and where the environmental shear is perpendicular to the MCS, we can also assume the environmental shear is line-normal. Lower tropospheric air approaches the MCS from the front, is lifted to its LFC along the gust front, and ascends upward and tilts rearward over the cold pool. Ascent along the gust front is largely upright and is associated with deep, moist convection, whereas that rearward of the gust front is more gradual in nature and is associated with predominantly stratiform precipitation. Here, weak ascent promotes modest hydrometeor growth through water vapor condensation and deposition, with the fallout of these hydrometeors resulting in precipitation. Melting of these falling hydrometeors often results in the stratiform precipitation region being characterized by a bright band in radar reflectivity at the altitude of the 0°C isotherm.

Mesoscale convective system motion is given by the sum of individual cell motion, which is along the deep-layer mean wind vector, and the propagation vector representing downshear development along the advancing gust front. Thus, for predominantly lower tropospheric environmental vertical wind shear, MCS motion is typically faster than the environmental wind at all levels. Thus, system-relative environmental flow is directed from front-to-rear at all levels, such that hydrometeors preferentially fall out rearward of the gust front. Evaporating hydrometeors along the leading line and within the trailing stratiform region lead to the development of an expansive cold pool and mesohigh behind the leading line. Cold pool and mesohigh intensity are both proportional to the environmental moisture profile; cold pools tend to be stronger in drier environments and weaker in moister environments.

Despite predominantly front-to-rear-directed flow, MCSs are characterized by rear-to-front flow in the lower to middle troposphere, atop the cold pool. Rear-to-front flow develops as the leading-line updraft acquires an upshear tilt over the cold pool. The vertical shear of the ascending front-to-rear current is associated with negative horizontal vorticity, whereas the vertical shear along the back edge of the cold pool is associated with positive horizontal vorticity. Together, this results in flow accelerating from rear to front atop the cold pool. Alternatively, but equivalently, the release of positive buoyancy along the leading-line updraft results in hydrostatically-lowered pressures at and below the level of maximum warming, typically corresponding to altitudes just above the cold pool. With weak pressure anomalies at equivalent altitudes rearward of the leading line, this results in a forward-directed perturbation pressure gradient force that accelerates flow from rear-to-front.

Rear inflow gradually descends as it approaches the MCSs leading edge. Accompanying adiabatic warming reduces near-surface pressure hydrostatically as it does so. Where this warming exceeds the diabatic cooling and hydrostatic pressure increases associated with evaporating hydrometeors, a wake low may be found. This typically occurs rearward of the leading line and heaviest stratiform precipitation. Ahead of the intense leading-line updraft, compensating descent and accompanying adiabatic warming can also reduce near-surface pressure, resulting in a pre-squall mesolow. Note that these pressure anomalies are not in balance with the kinematic field; in general, the maximum near-surface divergence is found between the mesohigh and wake low, whereas maximum near-surface convergence is found in the rear of the wake low and along the gust front's leading edge.

Mesoscale convective systems can generally be viewed as quasi-two-dimensional features. Ascent along the gust front is often uniform along the line, giving rise to slab-like convection, particularly when persistent gust front lifting is sufficient to result in the formation and maintenance of a moist absolutely unstable layer (MAUL). However, in environments where the cold pool and lift along the gust front are relatively weak (e.g., higher ambient relative humidity), ascent may not be strong enough to bring air parcels to their LFC everywhere along the gust front. More localized forcing for ascent may be necessary to bring air parcels to their LFC, resulting in more cellular convection.

Archetypes

The archetypal MCS described above is a trailing stratiform MCS. Well over half of all MCSs are trailing stratiform MCSs. However, there exist four other distinct MCS archetypes. The first two can be distinguished from trailing stratiform MCSs by the middle-to-upper tropospheric system-relative wind and thus vertical wind shear in their environments. For upper tropospheric rear-to-front vertical wind shear (e.g., westerly wind increasing with height in the upper troposphere), system-relative winds aloft are directed from rear-to-front, resulting in hydrometeor fallout ahead of the gust front. This describes a leading stratiform MCS. Here, evaporative cooling typically reduces ambient temperatures more in the middle troposphere than at the surface, such that ambient buoyancy is not reduced compared to trailing stratiform MCSs. Depending on the lower-tropospheric hodograph, inflow for leading stratiform MCSs may not necessarily come from ahead of the system. In contrast, when the upper tropospheric vertical wind shear is along-line, such as may occur when the winds aloft veer or back from a predominantly westerly direction, system-relative winds aloft are directed along the line and hydrometeor fallout is predominantly along the line. This describes a parallel stratiform MCS.

There also exist two other MCS archetypes that are not as readily distinguishable from the others: the training line-adjoining stratiform and backbuilding MCS archetypes. Training line-adjoining stratiform MCSs are favored for predominantly along-line deep-layer environmental vertical wind shear, akin to parallel stratiform MCSs, but when lower-tropospheric environmental vertical wind shear is normal to an ambient boundary (e.g., a stationary front) along which convection may form. Thus, training line-adjoining stratiform MCSs are favored poleward of synoptic-scale boundaries. Backbuilding MCSs form when convection continuously redevelops upshear (e.g., where outflow boundary motion and the environmental vertical wind shear vector are antiparallel, opposite of our expectations for redevelopment) of existing convection. Backbuilding MCSs are typically quasi-stationary, with cell motion along the mean wind nearly equal to but in the opposite direction of upshear propagation fostered by upwind convective redevelopment.

RKW Theory for MCS Maintenance

In the late 1980s, Rich Rotunno, Joe Klemp, and Morris Weisman published a paper describing a theory for strong, long-lived MCSs that has come to be known as RKW theory. Two principles are at the heart of RKW theory:

- Squall line *longevity*, related to the ability for convection to regenerate along the gust front, and *severity*, formally in terms of lower tropospheric updraft intensity and not necessarily deep-layer updraft intensity or surface wind speeds, are related to updraft tilt. For greater updraft tilts, the buoyancy dilution by entrainment increases, thus slowing upward-directed parcel accelerations. Further, increased updraft width fostered by greater updraft tilt leads to a larger downward-directed perturbation pressure gradient force resulting from increased thickness in the layer of positive buoyancy. This also slows upward-directed accelerations. Updraft tilt is thus directly proportional to updraft intensity.

- Lower tropospheric updraft tilt is a function of the balance between the environmental line-normal vertical wind shear and the baroclinically-generated horizontal vorticity within the MCS cold pool. The depth of the environmental shear is typically assumed to be roughly equal to that of the cold pool, or on the order of 2-3 km.

The result of RKW theory is an analytic expression for this latter principle. Much of the resulting derivation follows that for density current dynamics considered a few weeks ago. Consider a two-dimensional (x, z) plane. In this plane, we begin with the tendency equation for the **j**-component of the horizontal vorticity η:

$$\frac{\partial \eta}{\partial t} = -\mathbf{v} \cdot \nabla \eta - \frac{\partial B}{\partial x} = -u\frac{\partial \eta}{\partial x} - w\frac{\partial \eta}{\partial z} - \frac{\partial B}{\partial x}$$

where we have assumed purely two-dimensional flow (i.e., $v = 0$) and have neglected friction. B represents buoyancy. For the Boussinesq approximation, substituting with the continuity equation allows us to rewrite the advection terms in flux form, i.e.,

$$\frac{\partial \eta}{\partial t} = -\frac{\partial (u\eta)}{\partial x} - \frac{\partial (w\eta)}{\partial z} - \frac{\partial B}{\partial x}$$

As before, we wish to integrate this equation over our two-dimensional volume, from x = R to x = L and from z = 0 to z = d. Unlike before, we do not assume no ambient flow. This modestly changes where the edges of this volume are placed. We place the left side of this volume well behind the leading edge of the density current and the right side of this volume well ahead of the leading edge of the density current. The left edge is placed where the horizontal velocity at the ground is equal to density current motion, such that there is no motion relative to the density current. The top edge of this volume is placed at a level where the motion with respect to the density current is zero both on the left and right edges. This is usually above the top of the density current, near the top of the environmental vertical wind shear layer.

If we integrate over this control volume, we obtain:

$$\begin{aligned}\frac{\partial}{\partial t}\left(\int_0^d\int_L^R \eta\,dxdz\right) &= -\int_0^d\int_L^R \frac{\partial (u\eta)}{\partial x}dxdz - \int_L^R\int_0^d \frac{\partial (w\eta)}{\partial z}dzdx - \int_0^d\int_L^R \frac{\partial B}{\partial x}dxdz \\ &= -\int_0^d (u\eta_R - u\eta_L)dz - \int_L^R (w\eta_d - w\eta_0)dx - \int_0^d (B_R - B_L)dz\end{aligned}$$

If we assume a steady-state solution and that the ambient buoyancy $B_R = 0$, with $w(z = 0) = 0$ by definition, we obtain:

$$0 = -\int_0^d u\eta_R dz + \int_0^d u\eta_L dz - \int_L^R (w\eta_d)dx + \int_0^d B_L dz$$

Since $\eta = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}$, for w = 0 at x = R and x = L, $\eta = \frac{\partial u}{\partial z}$, such that:

$$
\begin{aligned}
0 &= -\int_0^d u \frac{\partial u_R}{\partial z} dz + \int_0^d u \frac{\partial u_L}{\partial z} dz - \int_L^R (w\eta_d) dx + \int_0^d B_L dz \\
&= -\int_0^d u du_R + \int_0^d u du_L - \int_L^R (w\eta_d) dx + \int_0^d B_L dz \\
&= -\frac{1}{2}\left(u_{R,d}^2 - u_{R,0}^2\right) + \frac{1}{2}\left(u_{L,d}^2 - u_{L,0}^2\right) - \int_L^R (w\eta_d) dx + \int_0^d B_L dz
\end{aligned}
$$

Since we placed our control volume where $u_{R,d}$, $u_{L,d}$, and $u_{L,0}$ are all zero, this simplifies to:

$$0 = \frac{1}{2} u_{R,0}^2 - \int_L^R (w\eta_d) dx + \int_0^d B_L dz$$

If we assume an upright updraft, where the horizontal vorticity in the cold pool equals and opposes that in the ambient environment, then the first integral in the equation above is zero. In other words, for a control volume cutting across the gust front with equal parts ahead and behind, the integral goes to zero. Positive vertical velocity is collocated with positive horizontal vorticity ahead of the gust front, while it is collocated with negative horizontal vorticity of equal magnitude behind the gust front. In this case, our equation simplifies to:

$$0 = \frac{1}{2} u_{R,0}^2 + \int_0^d B_L dz$$

Rearranging this equation, we obtain:

$$u_{R,0}^2 = -2\int_0^d B_L dz \equiv c^2$$

This is equivalent to our derivation for density currents assuming no ambient flow. There, $u_{R,0}$ was a measure for the density current propagation velocity. If we take the square root of the above and retain the negative root (i.e., implying a balance between equal-yet-opposite forcings), we obtain:

$$u_{R,0} = -c$$

Since $u_{R,d} = 0$, $\Delta u = u_{R,d} - u_{R,0} = -u_{R,0}$, such that:

$$\Delta u = c \rightarrow \frac{c}{\Delta u} = 1$$

where Δu is a measure of the environmental line-normal vertical wind shear and c is a measure of cold pool depth and intensity. This defines the "optimal state" of RKW theory, where the ambient vertical wind shear and cold pool-generated horizontal vorticity have equal magnitude but opposite sign. In this case, the lower tropospheric leading-line updraft along the gust front is upright. Where $c > \Delta u$, the cold pool-generated horizontal vorticity is of greater magnitude, such that the updraft tilts rearward (upshear) over the cold pool. Where $c < \Delta u$, the environmental vertical wind shear is of greater magnitude, such that the updraft tilts downshear. MCS formation is characterized by the latter stage, with an MCS reaching the optimal state as falling hydrometeors evaporate and lead to cold pool formation and then becoming upshear-tilted as the cold pool intensifies. As we will see shortly, this upshear tilt is a prerequisite for the development of rear inflow.

The case where $c < \Delta u$ implies that $\int_L^R (w\eta_d)dx > 0$. In other words, the positive horizontal vorticity associated with the (assumed westerly) vertically-sheared environmental flow is larger than the negative horizontal vorticity generated within the cold pool, such that there is a net positive flux. The case where $c > \Delta u$ has the opposite interpretation.

To this point, we have assumed that the environmental vertical wind shear is contained entirely in the lower troposphere. However, MCS environments typically contain at least modest amounts of middle-to-upper tropospheric vertical wind shear. While this was not formally considered by RKW theory, we can obtain some rudimentary insight if we extend its principles to such sheared flows. Westerly vertical wind shear aloft can inhibit the upshear tilt over the cold pool that often develops, as its associated horizontal vorticity opposites that of the cold pool. Conversely, easterly vertical wind shear aloft can enhance upshear tilt over the cold pool, as its associated horizontal vorticity is of like sign to that within the cold pool.

Rear Inflow Jets and Line-End Vortices

Rear inflow jets develop in the lower-to-middle troposphere, atop the cold pool, after an upshear tilt develops. As noted above, this upshear tilt typically develops after the cold pool has become sufficiently deep and strong such that its associated baroclinically-generated horizontal vorticity becomes larger than that associated with the ambient vertical wind shear. In fact, this upshear tilt is a prerequisite for rear inflow jet development. We can understand rear inflow jet development through two complementary perspectives:

- The vertical wind shear of the ascending front-to-rear current is associated with negative horizontal vorticity. Conversely, baroclinic generation leads to positive horizontal vorticity along the back edge of the cold pool. The vertical structure of these anomalies results in rear-to-front flow. Note that this perspective implicitly assumes that the horizontal vorticity anomalies are of equal magnitude. Although the cold pool must be comparatively intense along its leading edge for the upshear tilt to develop, the cold pool is weaker and shallower along its back edge where evaporation is weaker. Thus, this assumption is reasonable.
- The ascending front-to-rear current is associated with positive buoyancy and thus diabatic warming as latent heat is released. From thickness arguments, this results in a negative perturbation pressure anomaly beneath the level of maximum warming, which is typically found above the top of the cold pool. For $p' \approx 0$ along the back edge of the cold pool, this establishes a rear-to-front directed perturbation pressure gradient force that accelerates the flow from rear-to-front atop the cold pool.

Rear inflow jet development can result in the reduction of updraft tilt. The vertically-sheared flow beneath the level of maximum rear-to-front flow has positive horizontal vorticity that counteracts the baroclinically-generated negative horizontal vorticity in the cold pool. In this context, the RKW optimal state can be expressed as:

$$\Delta u^2 + \Delta u_j^2 = c^2$$

where $\Delta u_j^2 = u_{L,d}^2 - u_{L,0}^2$, where the first right-hand-side term is no longer zero because of the rear inflow jet. Where the horizontal vorticity associated with the environmental vertical wind shear and the rear inflow jet equal and oppose that in the cold pool, leading-line updraft tilt is reduced. Thus, rear inflow jet formation may result in a quasi-optimal state being restored.

In the above, we have assumed a steady-state solution. As we know, however, the environment is far from steady, and the same can be said about the cold pool. Reducing environmental line-normal vertical wind shear and/or environmental buoyancy can reduce the buoyancy realized by leading-line updrafts, weakening the rear inflow jet and leading again to an upshear-tilted system. Research is underway to better understand the contributions of temporally-varying environmental conditions upon rear inflow jet evolution and MCS tilt and maintenance. Damaging surface winds can occur when the rear inflow jet descends to the surface, possibly as balance is disrupted in favor of the cold pool. Strong surface winds may also result from meso-γ-scale vortices along the gust front.

Rear inflow jets are often accompanied by line-end, or bookend, vortices along the lateral extents of the parent MCS. Consider a two-dimensional (x,z) MCS with negative horizontal vorticity in its cold pool, as above. On the system-scale, the ascending front-to-rear flow can tilt this horizontal vorticity into the vertical. This results in the formation of counterrotating vortices: cyclonic north and anticyclonic south. As the cold pool horizontal vorticity is predominantly crosswise, stretching does not amplify either vortex; rather, they move in tandem with the MCS. The flow around the counterrotating vortices superposes with and strengthens that of the rear inflow jet, with ~25% of rear inflow jet intensity attributable to these vortices. With time, planetary vorticity convergence (neglected in our supercell discussion) favors the cyclonic vortex, giving rise to the classic comma-shaped appearance of bow echoes. Bow echoes tend to be favored in environments with higher surface-based CAPE and lower tropospheric vertical wind shear; the former favors stronger cold pools, while the latter favors greater realized buoyancy within leading-line updrafts, each resulting in stronger rear inflow and bowing structure.

Mesoscale Convective Vortices

The vertical profile of diabatic heating within the stratiform rain region of an MCS is known as a stratiform heating profile. It is characterized by diabatic cooling due to evaporating hydrometeors at and near the surface, diabatic warming due to condensation, freezing, and deposition within the gently ascending front-to-rear flow in the middle-to-upper troposphere, and weak radiative cooling at cloud top. The persistence of this heating profile over a period of several hours can result in the formation of middle tropospheric mesoscale convective vortices (MCVs).

We can understand MCV development using isentropic potential vorticity thinking. Recall that the isentropic potential vorticity P on isentropic surfaces can be expressed as:

$$P = -g(\zeta + f)\frac{\partial \theta}{\partial p}$$

Under adiabatic, frictionless conditions, P is conserved following the motion. However, if diabatic heating and/or friction are important, P is not conserved. It is the non-conservation of P that leads to MCV development.

Neglecting friction and horizontal gradients in diabatic heating rate Q, an expression for the rate of change of P following the motion is given by:

$$\frac{DP}{Dt} = P\frac{\partial Q}{\partial \theta}$$

If you are interested in a full derivation of this expression, please see the Atm Sci 361 lecture on isentropic potential vorticity non-conservation. This equation is applied on isentropic surfaces; i.e., with θ as the vertical coordinate. In general, θ increases with height above the boundary layer. If we assume $P > 0$, then we simply need to evaluate how diabatic heating rate Q changes with height to evaluate how P changes following the motion.

Between the top of the cold pool and the ascending front-to-rear current, Q increases with height. This results in P increasing following the motion. From the definition of P, this is associated with both increased absolute vorticity $\zeta + f$ and static stability. Q decreases with height above the front-to-rear flow and immediately above the surface; here, P decreases following the motion, reducing both absolute vorticity and static stability at such altitudes.

Mesoscale convective vortices can aid in convective persistence and redevelopment. Consider an MCV translating with the deep-layer mean flow. For simplicity, we assume westerly-sheared flow, such that the MCV-relative flow is out of the east below the level of the MCV. Because the MCV is associated with enhanced static stability, isentropes bow upward into the MCV in the lower to middle troposphere. An easterly MCV-relative flow means that air parcels that conserve potential temperature will ascend east of the MCV and descent west of the MCV. From thermal wind, the westerly-sheared flow is associated with a meridional layer-mean potential temperature gradient, with isentropes that bow upward toward the north. Cyclonic flow around the MCV that conserves potential temperature will thus ascend east of the MCV and descend west of the MCV.

If this ascent can bring air parcels to their LFC, convective maintenance or redevelopment east of the MCV may occur. (Recall that lift can reduce CIN and increase CAPE through layer lifting, for example, to aid in this process.) This most commonly occurs in the local late afternoon and evening when the ambient surface-based CAPE is maximized. At night, convection preferentially develops near the MCV center, where large-scale convergence is maximized and cloud-top radiative cooling results in weak most unstable CAPE. Convective regeneration, wherever located, is required for the diabatic heating profile supporting MCV formation and maintenance to be maintained. When it does not occur and/or when the MCV is deformed by strong vertical wind shear, weakening and dissipation occur.