Mesoscale Meteorology: Convection Initiation and Convection Types

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Convection Initiation

Convection initiation (CI) describes the formation of deep, moist convection, whether it is severe or non-severe in nature. CI is triggered by a surface-based or elevated convergence mechanism in a moist environment that is positively-buoyant to upward parcel displacements over a great vertical depth. Thus, CI requires sufficient moisture, instability, and lift. In this sense, convective available potential energy (CAPE) is a necessary but insufficient condition for CI; CAPE cannot be realized unless a parcel can achieve its level of free convection (LFC), which requires sufficient lifting to overcome any convection inhibition (CIN) that may exist, while elevated boundary layer moisture content mitigates against the reduction in buoyancy associated with entrainment and mixing with drier environmental air as air parcels ascend toward their LFC.

Numerous meteorological phenomena exist that can provide the localized convergence requisite for CI. These include dry lines; frontal zones; elevated convergence zones (e.g., the nose of a low-level jet ascending over a lower tropospheric baroclinic zone); gust fronts; sea breezes; horizontal convective rolls; undular bores, gravity waves, and other wave-like phenomena; terrain-induced vertical circulations; and mesoscale boundaries resulting from horizontal variation in land-surface properties such as soil moisture and temperature. Interactions between multiple boundaries across horizontal scales often establish favored locations for CI. Lifting along these boundaries must be sufficiently large in magnitude and vertical extent to bring an air parcel to its LFC. In this sense, reduced CIN is a necessary but insufficient condition for CI. CI is not guaranteed even in the case of large CAPE and zero CIN, in large part due to the assumptions that enter in the formulations of CAPE and CIN through parcel theory.

CI is a scale interaction problem requiring the favorable superposition of the synoptic-, meso-, and micro-scales, with each scale providing a set of necessary but not sufficient conditions for CI. The synoptic- and meso-α-scales establish the large-scale thermodynamic and kinematic environment in which CI occurs. For example, CI is sensitive to regional variations in capping inversion strength atop the boundary layer, available buoyant energy, and moisture depth within the boundary layer. With respect to capping inversion strength, subsidence can increase inversion strength and mitigate CI potential, ascent can weaken a capping inversion through layer lifting, and horizontal advection can locally increase or decrease capping inversion strength. Persistent lower tropospheric moisture convergence can locally increase boundary layer moisture depth, reducing the buoyancy lost due to entrainment and mixing as an air parcel ascends toward its LFC.

The meso-β-scale contributes most strongly to horizontal variability in the large-scale environment whereas the meso-γ and microscales modulate the local boundary layer lifting, moistening, and environmental variability that are crucial to CI timing and location. The role of the smaller scales is such that CI does not occur everywhere within an environment that is otherwise favorable for it on the larger-scales. It thus should come as no surprise that CI is highly sensitive to the atmospheric state in which it occurs, particularly within the boundary layer. For example, CI is sensitive to the magnitude of the lower tropospheric vertical temperature and moisture gradients. Modifying these gradients by only 1°C and 1 g kg^{-1} over a short vertical distance can influence whether CI occurs! These values are approximately equal to the values of observational uncertainty and local boundary layer variability, resulting in CI being a particularly challenging mesoscale forecast problem.

Processes That Increase CAPE and/or Decrease CIN

If increased CAPE and decreased CIN are necessary but insufficient conditions for CI, it stands to follow that we are interested in the processes that can increase CAPE and/or decrease CIN. Here, we consider both CAPE and CIN for any parcel originating level; in other words, though we often think of surface-based CAPE and CIN, these insights are relevant no matter whether we consider a surface-based or elevated air parcel.

First, recall the definitions of CAPE and CIN:

$$CAPE = \int_{LFC}^{EL} B dz \approx g \int_{LFC}^{EL} \frac{T_v^{parcel} - T_v^{env}}{T_v^{env}} dz$$

$$CIN = -\int_{0}^{LFC} B dz \approx -g \int_{0}^{LFC} \frac{T_v^{parcel} - T_v^{env}}{T_v^{env}} dz$$

The lower bound on the integration for CIN refers to the parcel origination level, where a value of 0 indicates a surface-based air parcel. Both CAPE and CIN are vertically-integrated quantities. Their definitions arise from simplifying assumptions made in the definition of buoyancy B,

$$B = -\frac{\rho'}{\overline{\rho}} g \approx \left(\frac{T_v'}{\overline{T_v}} - \frac{p'}{\overline{p}} - r_h \right) g \approx \frac{T_v'}{\overline{T_v}} g \approx \frac{T_v^{parcel} - T_v^{env}}{T_v^{env}} g$$

The key assumptions are:

- Density variations result primarily from temperature and moisture variations (as manifest in virtual temperature) rather than from pressure variations or hydrometeor mass.
- The mean and perturbation quantities are taken to reflect environmental and parcel profiles of virtual temperature. This underpins *parcel theory*, which further assumes that air parcels are infinitesimally small air volumes that remain altogether isolated from the environment.

We will return to these shortcomings when describing the insufficiency of CAPE and CIN for CI. In a general sense, CIN decreases when the negative buoyancy and/or the vertical depth over which it is found decrease. This can occur by warming the virtual temperature of the ascending air parcel and/or cooling the environmental virtual temperature over the layer containing CIN. Similarly, CAPE increases when positive buoyancy and/or the vertical depth over which it is found increase. This can occur by warming the virtual temperature of the ascending air parcel and/or cooling the environmental virtual temperature over the layer containing CAPE.

How can we increase the virtual temperature of the ascending air parcel? In the context of parcel theory, this occurs by shifting the parcel ascent curves further to the right on a skew T-ln p diagram. The primary means by which this occurs are by warming and, to greater extent, moistening the air at the parcel origination level. In some cases, this may also change the altitude at which the lifting condensation level (LCL) is found; warming alone elevates it, moistening alone lowers it, while a combination of warming and moistening may result in comparatively less changes in LCL height.

For surface-based (or near-surface-based) air parcels, warming and moistening can occur by way of surface sensible and latent heat fluxes, and it should come as no surprise that surface-based CI most commonly occurs during the local late afternoon to early evening hours in the warm-season. However, if the ground surface is cooler or drier than the air, surface fluxes may lead to cooling and drying and thus reduction of CI potential. Horizontal advection may also locally warm and/or moisten air at the parcel origination level; as with surface fluxes, it can also locally cool and/or dry air at this level. Note, however, that horizontal advection cannot *create* local maxima – it can only transport them horizontally and/or vertically. For example, consider the water vapor mixing ratio budget equation:

$$\frac{\partial r_v}{\partial t} = -\mathbf{v} \cdot \nabla r_v - C + E \text{, such that } \frac{dr_v}{dt} = E - C$$

Here, water vapor mixing ratio can increase following the motion only by evaporation (e.g., surface latent heat flux into the air or the evaporation of precipitation falling from above) that exceeds the loss of water vapor to condensation. However, horizontal convergence can result in local maxima in the vertical depth over which relatively high water vapor mixing ratio is found, while high soil moisture content can result in local maxima in surface latent heat flux to the overlying air. In the former, vertical mixing triggered by surface sensible heat fluxes dilutes water vapor mixing ratio to a lesser extent than for a shallower depth; in the latter, vertical mixing is altogether weaker due to its connection to the Bowen ratio.

How can we decrease environmental virtual temperature? Shifting the environmental temperature curve further to the left on a skew T-ln p diagram, particularly in the middle to upper troposphere. While reducing environmental moisture can also reduce the environmental virtual temperature, the extent to which it does so is near-zero above the middle troposphere due to the near-zero mixing ratio values typically found at such altitudes. Further, as we will discuss later in this lecture, it also increases the potential that the ascending air parcel will mix with the drier environment, acting to reduce its buoyancy rather than increase it. Thus, to first order, we are interested in when the lapse rate over the CIN- and/or CAPE-bearing layers becomes steeper, such that environmental virtual temperature is smaller relative to the parcel virtual temperature ascending along a pseudoadiabat.

We can obtain an equation for the local lapse rate tendency from the first law of thermodynamics and the hydrostatic approximation. We start with the first law of thermodynamics:

$$Q = c_p \frac{dT}{dt} - \alpha \frac{dp}{dt}$$

Here, Q is the diabatic heating rate per unit mass. All other variables have their typical meaning. Note that:

$$\frac{dT}{dt} = \frac{\partial T}{\partial t} + \mathbf{v}_h \cdot \nabla_h T + w\frac{\partial T}{\partial z}$$

And, for pressure p that is a function of height z only,

$$\frac{dp}{dt} = \frac{dp}{dz}\frac{dz}{dt}$$

The first of these terms can be replaced by the hydrostatic equation, while the second defines w. Thus, substituting these expressions into the first law of thermodynamics, we obtain:

$$Q = c_p\left[\frac{\partial T}{\partial t} + \mathbf{v}_h \cdot \nabla_h T + w\frac{\partial T}{\partial z}\right] + gw$$

From this, we desire an equation for the local lapse rate tendency. First take the partial derivative of the above with respect to z and multiply the result by -1:

$$-\frac{\partial Q}{\partial z} = c_p\left[\frac{\partial}{\partial t}\left(-\frac{\partial T}{\partial z}\right) + \mathbf{v}_h \cdot \nabla_h\left(-\frac{\partial T}{\partial z}\right) - \frac{\partial \mathbf{v}_h}{\partial z} \cdot \nabla_h T + w\frac{\partial}{\partial z}\left(-\frac{\partial T}{\partial z}\right) - \frac{\partial w}{\partial z}\frac{\partial T}{\partial z}\right] - g\frac{\partial w}{\partial z}$$

Letting $\Gamma \equiv -\frac{\partial T}{\partial z}$ and dividing by c_p, noting that $\Gamma_d \equiv \frac{g}{c_p}$, we obtain:

$$-\frac{1}{c_p}\frac{\partial Q}{\partial z} = \frac{\partial \Gamma}{\partial t} + \mathbf{v}_h \cdot \nabla_h \Gamma - \frac{\partial \mathbf{v}_h}{\partial z} \cdot \nabla_h T + w\frac{\partial \Gamma}{\partial z} + \Gamma\frac{\partial w}{\partial z} - \Gamma_d\frac{\partial w}{\partial z}$$

If we solve this equation for the local lapse rate tendency, we obtain:

$$\frac{\partial \Gamma}{\partial t} = -\mathbf{v}_h \cdot \nabla_h \Gamma - w\frac{\partial \Gamma}{\partial z} + \frac{\partial \mathbf{v}_h}{\partial z} \cdot \nabla_h T + \frac{\partial w}{\partial z}\left(\Gamma_d - \Gamma\right) - \frac{1}{c_p}\frac{\partial Q}{\partial z}$$

From left to right, the right-hand side terms represent horizontal lapse rate advection, vertical lapse rate advection (equivalent to layer lifting of unsaturated layers), horizontal temperature advection by the vertically-sheared flow, stretching, and differential diabatic heating. On larger scales, horizontal lapse rate advection dominates; on smaller scales, the remaining terms can be important.

Note that the third term can be written as:

$$\frac{\partial \mathbf{v}_h}{\partial z} \cdot \nabla_h T = \frac{\partial \mathbf{v}_{hg}}{\partial z} \cdot \nabla_h T + \frac{\partial \mathbf{v}_{hag}}{\partial z} \cdot \nabla_h T = \frac{\partial \mathbf{v}_{hag}}{\partial z} \cdot \nabla_h T$$

The term involving the geostrophic wind is zero: its partial derivative with respect to z defines the thermal wind, which blows parallel to the isotherms or perpendicular to the horizontal temperature gradient. The dot product of two perpendicular vectors is zero.

Further, the first and third terms can be combined into a single term through the chain rule:

$$-\mathbf{v}_h \cdot \nabla\Gamma + \frac{\partial \mathbf{v}_h}{\partial z} \cdot \nabla_h T = -\frac{\partial}{\partial z}\left(-\mathbf{v}_h \cdot \nabla_h T\right)$$

This term represents differential horizontal temperature advection. The local lapse rate tendency is positive (e.g., temperature more rapidly decreases with height) for positive horizontal advection, particularly that which increases with height; dry adiabatic lifting of unsaturated layers (assuming $\Gamma < \Gamma_d$), particularly that which increases with height; and diabatic cooling increasing with height.

The Insufficiency of Increasing CAPE and Reducing CIN for CI

As previously noted, both increased CAPE and reduced CIN are necessary but insufficient for CI. Why? Let us start with the vertical momentum equation, neglecting the vertical component of the Coriolis force, friction, and terms associated with the non-spherical nature of the Earth:

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g$$

This equation states that vertical parcel accelerations following the flow result from an imbalance in the vertical pressure gradient force and buoyancy. We can illustrate this by first multiplying by density ρ:

$$\rho\frac{dw}{dt} = -\frac{\partial p}{\partial z} - \rho g$$

Define a horizontally-homogenous base state environment that is in hydrostatic balance, such that:

$$0 = -\frac{\partial \overline{p}}{\partial z} - \overline{\rho} g$$

Subtracting from the full vertical momentum equation, we obtain:

$$\rho\frac{dw}{dt} = -\frac{\partial p'}{\partial z} - \rho' g$$

Dividing through by ρ, we obtain:

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p'}{\partial z} - \frac{\rho'}{\rho} g$$

The second right-hand side term is simply the buoyancy B. Note, however, the perturbation vertical pressure gradient force that appears as the first right-hand side term. Vertical parcel accelerations result not from buoyancy alone but rather from a balance between these terms.

Let us consider a case where this matters. Consider the case of a positively-buoyant air parcel; i.e., one that is warmer than its environment. The thickness of the infinitesimally-small layer in which the air parcel is found is slightly larger than that of its environment. In geometric (z) coordinates, this implies perturbation high pressure atop the parcel and perturbation low pressure beneath the parcel, together resulting in a downward-directed perturbation pressure gradient force. While this force is typically small, it must be overcome for upward parcel accelerations to occur. Outside of buoyancy, how might this occur? Recall that horizontal convergence results in an accumulation of mass, such that pressure, a measure of the mass of the air above you, is locally elevated at the layer of maximized convergence. This results in an upward-directed perturbation pressure gradient that we typically see manifest as ascent – e.g., mechanically-forced lifting can aid is helping positively-buoyant air parcels to accelerate upward! Indeed, in the case of non-zero CIN, it is this forced lift that is often responsible for triggering CI.

Both CAPE and CIN are traditionally defined in terms of virtual temperature rather than the full buoyancy. Neglecting hydrometeor mass is typically justifiable before cloud development occurs; however, once cloud formation has occurred, hydrometeor mass becomes non-zero, reducing the true buoyancy. In some cases, particularly with limited buoyancy, this may result in CI terminating before it can result in a sustained updraft. The perturbation pressure at the level of the ascending parcel is typically small; as discussed above, however, its vertical variation can be important.

What else can be important? Recall that parcel theory assumes an air parcel of infinitesimally small volume that remains isolated from its environment, conserving θ and r_v before reaching saturation and conserving θ_e throughout, as it ascends. The chief shortcoming of this is that air parcels do not remain isolated from their environment as they ascend. Rather, they mix with, or *entrain* air from, their surrounding environment. This can occur both within the boundary layer before saturation is achieved and at higher altitudes where the parcel is presumably positively-buoyant. Since mixing ratio and temperature typically decrease with height, entrainment typically reduces the buoyancy of an air parcel that is warmer and moister than its surroundings.

Entrainment is particularly large when the environment is relatively dry (and, to lesser extent, cool) and when vertical wind shear magnitude is large over the ascent layer (to mechanically force the environmental air toward the updraft/cloud). Thus, deeper, moister boundary layers, such as may result from horizontal moisture convergence and ascent, are favored locations for CI: an ascending air parcel in such an environment experiences reduced entrainment compared to what it would experience upon ascending through a drier boundary layer. The use of so-called mixed-layer CAPE and CIN, wherein the ascending parcel's potential temperature and water vapor mixing ratio are estimated from their average values over the lowest 50-100 hPa, is intended to account for dilution due to mixing and entrainment within the boundary layer, but this is at best an approximation to the full physics of the problem.

Finally, there are shortcomings in the vertical profiles from which estimates of CAPE and CIN are commonly obtained. Observed soundings are generally obtained only twice per day, with only one of those times (0000 UTC in the United States) near the times at which CI most commonly occurs, and with spacing of several hundred kilometers. Given the sensitivity of CI to local boundary layer variability, soundings generally capture only the larger-scale environment in which CI occurs and not the finer-scale details important for its timing, location, and occurrence. While model-derived soundings do not have the timing or spacing issues of observed soundings, they are at best model-derived syntheses of observations. Model errors, of which there are many possible sources (e.g., numerical methods used to solve the primitive equations, parameterization of fine-scale processes not capable of being directly represented by the model, inaccuracies in underlying soil parameters, etc.), lead to model-derived soundings being approximations to the real atmosphere at best.

Convection Types

There are three necessary ingredients for deep, moist convection: lift, moisture, and instability. In the absence of vertical wind shear, the favorable superposition of these ingredients leads to *single-cell convection*. In the presence of moderate amounts of vertical wind shear, *multicell convection* results. Finally, in the presence of strong vertical wind shear, *supercell convection* results. Typical vertical wind shear magnitudes, usually considered over the 0-6 km layer, for each are on the order of < 10 m s^{-1}, 10-20 m s^{-1}, and > 20 m s^{-1}, respectively. Surface-based CAPE does not have a strong relationship with cell type; all can occur in low-, moderate-, and high-CAPE environments.

As the vertical wind shear magnitude increases, convection longevity and severity both typically increase. This contradicts our buoyancy-focused discussion above, where we said entrainment is particularly large when vertical wind shear is large. Thus, there is an implied balance: so long as CI can occur despite the deleterious effects of vertical wind shear-induced entrainment, vertical wind shear is favorable for convection longevity. The relevant dynamics, including the influence of vertical wind shear on propagation characteristics, will be discussed next week. Here, we focus on single-cell and multicell convection.

<u>Single-cell convection</u>

Single-cell convection is short-lived; after CI, precipitation forms, with the resulting hydrometeor mass reducing the buoyancy available to the updraft. A short time later, hydrometeors fall out, and the accompanying evaporative cooling helps to stabilize the boundary layer, further reducing the buoyancy available to the updraft and leading to its rapid dissipation. An outflow boundary moves away from the decaying storm. Absent environmental vertical wind shear to oppose the horizontal vorticity circulation along the outflow boundary's leading edge, lifting along the outflow boundary is weak and shallow and is generally unable to support subsequent CI events. In this sense, single-cell convection is sometimes referred to as pulse convection: it dissipates as quickly as it initiates. The expected duration of a single-cell thunderstorm can be expressed in terms of the time it takes air parcels to ascend to their equilibrium level and the time it takes falling hydrometeors to reach the ground from the equilibrium level, i.e.,

$$\tau = \frac{H}{w_0} + \frac{H}{v_t}$$

H is the height of the equilibrium level for a surface-based ascending parcel, w_0 is the mean updraft speed, and v_t is the mean hydrometeor terminal velocity. For $H \sim 10$ km and $w_0 \sim v_t \sim 5$ m s^{-1}, $\tau \sim$ 4000 s $\sim$ 66.67 min. This is in general agreement with, if somewhat of an overestimate of, observed single-cell convection longevity, which is typically on the order of 30-60 min.

From thermal wind, weak vertical wind shear implies a weak horizontal layer-mean temperature gradient and thus baroclinicity, such that large-scale forcing for ascent is typically weak in single-cell supporting environments. Consequently, single-cell convection is generally diurnal in nature; it occurs during the local mid-afternoon to early evening hours during peak surface heating (when CAPE is typically maximized and CIN typically minimized for surface-based parcels from surface sensible heating along). Lifting for CI is generally provided by a thermally-driven circulation such as a sea breeze, an ascending branch of a horizontal convective roll, or upslope flow along sharply-sloped terrain, which are all weak to non-existent outside of the peak of the diurnal cycle. In this sense, single-cell convection is sometimes referred to as air mass convection, as it is most common in synoptically-benign environments.

<u>Multicell convection</u>

Individual cells with multicell convection are, like their single-cell counterparts, short-lived. Here, however, environmental vertical wind shear can support subsequent CI events along the spreading outflow boundaries. Such CI events are favored along the outflow boundary in two locations. The first is where the buoyancy-driven horizontal vorticity in the density current best balances that of the environmental vertical wind shear. This is usually found where the environmental vertical wind shear vector is close to perpendicular to and in the same direction as outflow boundary propagation. The second is where the near-surface environmental storm-relative wind is close to perpendicular to and opposite outflow boundary propagation. For a straight hodograph, the two favored locations are in proximity to each other; for a curved hodograph, they can be somewhat separated. Mesoscale thermodynamic (e.g., CAPE and CIN gradients) and kinematic (e.g., the presence of larger-scale boundaries or sloped terrain along which lift may occur) can further modulate these locations.

Hydrometeor loading and evaporative cooling limit buoyancy for an individual cell in a multicell convection event, as occurs with single-cell convection. However, here, the ability for subsequent CI events to occur along the outflow boundary increases convective *episode* longevity. Thus, multicellular convection represents multiple regenerating single-cells along an outward-spreading outflow boundary. This can take the form of multiple discrete cells or a nearly contiguous linear band; the latter, or *mesoscale convective systems*, are discussed in more detail in a later lecture.

Given higher vertical wind shear magnitude than with single-cell convection, the environmental baroclinicity is somewhat higher in multicell than in single-cell convection environments. Thus, large-scale forcing for ascent is typically larger for multicell than single-cell convection, such that the initial CI event preceding multicell evolution can occur in concert with a wider range of meso- to synoptic-scale phenomena (boundaries) than with single-cell CI events. Subsequent CI events occur along the outward-spreading outflow boundary as earlier CI events reach maturity or begin to dissipate. Individual convective cells tend to *move* with the mean wind vector over their depth, while the entire multicellular system *propagates* both as a function of individual cell motion and as a function of the direction in which new CI events occur. Related to these ideas are the concepts of *continual* versus *discrete* propagation. Continual (discrete) propagation occurs for cell (storm) along the mean wind vector; when the cell and storm motion are both off of the mean wind vector, both continual and discrete propagation characteristics exist.