Module 2.1

Slide 1: This lecture series will introduce you to a few basics of convection that develops on the mesoscale in the tropics. Clouds organized at the mesoscale are responsible for the majority of observed precipitation in the tropics. In the first module, we will start off with a discussion of convective and stratiform precipitation.

Slide 2: As discussed in the previous lecture series, cumuliform convection occurs with a continuum of depths up to clouds that extend from the boundary layer to the tropopause. A population of shallow convection overcomes inhibition and dry air to eventually grow into deep convection, and as divergence aloft from deep convection causes moistening of the near-cloud environment, anvil clouds that are laterally expansive compared to the parent convective elements from which they develop begin to form. Systems of numerous individual convective elements that are connected to a stratiform region are known as a mesoscale convective system, or MCS. MCSs are typically 100 km or greater along one horizontal dimension. They transport large amounts of water to the upper troposphere and through its extensive anvil clouds, distribute it far from where convection initially developed. As a result, MCSs are a significant contributor to upper tropospheric water vapor, promoting the development of tropical cirrus clouds.

Slide 3: MCSs occur throughout the tropics, subtropics, and mid-latitudes. They frequently occur over both continents and oceans. The current figure highlights where MCSs occur in the tropics. The top three panels indicate the locations of MCSs, defined by size, in boreal winter, and the bottom three panels indicate the same during boreal summer. MCSs are common over the Indo-Pacific warm pool year around. During boreal winter, they frequently occur over southern Africa and central South America. Large MCSs are common over the Maritime continent during boreal winter. During boreal summer, MCSs are more common in the Northern Hemisphere. Many more occur in the ITCZ in the Atlantic, East Pacific and northwestern South America than during winter. Small MCSs are more common in southern North America as well and are associated with the North American “monsoon”. MCSs are frequently observed in southeast Asia and western Africa during boreal summer as well, largely associated with summer monsoons in those locations.

Slide 4: MCSs with a variety of structures exist, and the shape of an MCS is largely dependent on the background environmental structure, including wind shear. A canonical MCS is often represented as a system with a leading convective line and trailing stratiform region, which we will discuss more soon. The schematic shown shows such a system. A convective region is horizontally narrow but contains intense updrafts that can overshoot the tropopause. The heaviest precipitation in an MCS occurs near these updrafts. Moisture advected out of the convective regions supports the development of anvil clouds both ahead, which is to the left in this figure, and behind the convective cell. The blue region indicates where relatively light stratiform precipitation occurs. Above the 0°C level in stratiform, deposition occurs and releases latent heat. Melting occurs as ice hydrometeors fall through the 0°C level, which causes a radar brightband to appear due to the difference in dielectric constants between liquid water and ice. Evaporation of liquid water occurs below the 0°C level, which consumes energy from the atmosphere, causing cooling of the ambient air. The precipitating regions advect moist air at upper levels into the surrounding environment, forming non-precipitating anvil clouds, denoted here by purple circles. A leading anvil, which is in front of the convection, is usually small, but the trailing anvil can extend rearward of convection for hundreds of kilometers and persist for hours to days after precipitation dissipates. Anvil clouds enhance upper-level shortwave heating, experience longwave heating near their bases, and all the clouds experience longwave cooling right at their tops. The net radiative heating that occurs in anvils reinforces the latent heat release at upper levels in stratiform regions. The diabatic heating helps to reinforce upper level upward motion on the mesoscale.

Slide 5: Examples of anvil clouds seen in West African MCSs are shown here. In the top left is reflectivity from a vertically pointing cloud radar as an MCS passes over it. The MCS is moving from right to left, so the leading anvil is present to the left of the gray box. The leading anvil is close to deep convection (denoted by the echo free region beneath the gray box that indicates strong attenuation of the radar beam) and is much higher in altitude than the larger trailing anvil at the rear of the system. At the bottom panels, derived longwave and shortwave heating are shown, and the total heating is shown at top right.

Slide 6: Three idealized profiles of latent heating in convective and stratiform precipitation are shown in the top panel. One is for shallow convection, which in this case refers to moderately deep convection like that in the congestus mode—except that this convection is cumulonimbus. It has a peak around 2–3 km and no heating above about 6 km. The deep convective heating profile extends through the depth of the troposphere and has a peak near 6 km. These peaks in heating denote the levels at which the most condensation occurs in cloud. The stratiform heating profile has a peak positive value in the upper troposphere and a peak negative value in the lower troposphere, with a change in sign around the 0°C level. The heating and cooling are dominated by deposition and evaporation, respectively. The exact depths and shapes of real heating profiles are much more complicated, but these simple functions capture heating to first order in the three major types of precipitating tropical convection. They can be combined to produce linear combinations of the idealized profiles that characterize the mean net heating profile within an MCS that depends on the relative precipitation rates in deep convection and stratiform. The deep convective profile looks like the first baroclinic mode in the troposphere, while stratiform and shallow convective heating look like opposite signed second baroclinic modes in the troposphere. We will describe the baroclinic modes a little more shortly.

Slide 7: These types of profiles can be confirmed using a numerical model, as shown in this figure, which further divides convection into a few additional categories and identifies the convection based on the simulated low-level reflectivity field. Mean vertical velocity and latent heating are shown respectively in the left and right panels for different categories of rainfall. In the model, deep convection tends to still have heating throughout most of the troposphere, although cooling below 800 mb may be associated with heavy rain in convective downdrafts. Stratiform precipitation shows heating aloft and cooling with a peak below the 0°C level, but latent heating in the model shows near zero heating below 800 mb, which may or may not be realistic. The shallow convective heating profile is also denoted by the light blue colors and includes cooling above the convection that is likely attributed to longwave cooling. In summary, the modeled profiles resemble the idealized profiles to first order but show some interesting additional complexities.

Slide 8: Previously, we referred to convective heating profiles as resembling baroclinic modes of the troposphere. This encourages a short discussion of what barotropic and baroclinic modes are, although we will discuss them in more detail mathematically when we discuss equatorial waves later in the quarter. The modes refer to gravity wave modes of motion that occur in an idealized multi-layer fluid. A barotropic mode, which is known as the external mode, describes the vertically averaged motion within a fluid. In the barotropic mode, density is only controlled by pressure. You can visualize this by a wave in an n-layered shallow fluid (approximating the troposphere as the shallow fluid) in which the crests and troughs are aligned with depth in the fluid. In baroclinic modes, density of the fluid is controlled not only by pressure but also temperature, humidity, or salinity. You can imagine the first baroclinic mode as a wave in which the crests are spread out and the troughs are close together, or vice versa. In the troposphere, the crests spread out are denoted by hypothetical contours of theta-e, representing deep warming through the troposphere, which resembles the latent heat release in deep convection. Cooling throughout the troposphere is the negative first baroclinic mode, which is somewhat similar to mean radiative cooling throughout the atmosphere, although the vertical structure of radiative cooling in clear air has maxima in the upper and lower troposphere.

Slide 9: Here, we visually compare the first baroclinic mode to the second. The second baroclinic mode might consist of heating aloft that spreads out crests through the top of the fluid but compresses them in lower layers of the same column, or vice versa. The positive second baroclinic mode has heating aloft and cooling in the lower troposphere; this resembles the combined heating profiles caused by diabatic heating in stratiform and non-precipitating anvils. The opposite case with heating in the lower troposphere resembles shallow convective heating topped by radiative cooling aloft.

Slide 10: Convective and stratiform regions can be challenging to detect in satellite imagery. The example shown here is a visible satellite image of an MCS located near Panama. The deepest convection is not immediately obvious, although some deep convection might be suspected north of Panama.

Slide 11: The infrared brightness temperatures at the same time highlight the coldest cloud tops in red colors. These are likely close to the convective precipitation regions because the coldest cloud tops occur as a result of deep convection reaching high altitude and promoting stratiform cloud growth aloft. Farther from the convective clouds, the tops of clouds are typically lower as hydrometeors subside. Combined with clouds being optically thinner farther from the source of convection, the higher brightness temperatures indicate where stratiform precipitation or non-precipitating anvil is located.

Slide 12: Precipitation is difficult to detect using visible and IR radiances alone; however, radar can detect echoes of various strength that occur beneath the cloud tops seen by a satellite. The current figure shows radar echoes from the tropical warm pool in a variety of different scenarios. The top left shows isolated convection that is characterized by several small, scattered cumulonimbi. The top right shows a more convectively active state where convective echoes (those with the high reflectivity approximately greater than 40 dBZ) are larger, and surrounding weak echo regions where hydrometeors from convective echo were transported laterally. The bottom left illustrates an example of an eastward moving leading squall line of intense convection accompanied by a trailing stratiform region of weaker echo. The bottom right depicts a large stratiform region with embedded convective elements located within it somewhat randomly. These variety of structures are often difficult to identify easily by simply consulting satellite imagery.

Slide 13: As you may have gathered from previous slides, a cloud population tends to evolve from one that consists predominantly of shallow convection first. Deep convection then develops, followed by lateral expansion of those deep convective elements. Eventually, the formation of broad stratiform regions signifies the mature stage of the convective lifecycle. This progression can be seen in the figure shown here, which shows radar-derived time series of rainfall associated with each type of convection as a function of time relative to the peak rainfall time of several convective events.

Slide 14: The evolution resembles the progression seen in the drawing at the beginning of this module.

Slide 15: The evolution from convection to stratiform occurs on many different time scales also. Previously, you just saw how convection evolves on an approximately 2-day long period for a convective event that was regional in spatial scale. The current figure shows the evolution of convection and stratiform in a larger-scale phenomenon, the Madden-Julian Oscillation, over a 30-day long period. If you just focus on the black line, you’ll see that the convective rain rate maximizes on average at the time that the peak precipitation occurs, while stratiform precipitation amplitude is largest a few days after. The general lag of stratiform behind convective rainfall makes sense because deep convection is a perquisite for the formation of stratiform.

Module 2.2

Slide 1: In this module, we will look through some basics of MCS kinematic and microphysical structure.

Slide 2: We'll start off from a complicated looking diagram from the 1970s, before extensive observations of MCSs had been collected. The figure is rather remarkable because its fidelity to realistic MCS structure is quite good even though many of the features intuitively indicated here were not explicitly observed yet. Shown here is a canonical MCS with a leading line of convective echo trailed by a broad stratiform region. These generally occur wind when substantial deep layer vertical wind shear is present in the environment.

Slide 3: Convective updrafts are found within the red oval, and stratiform regions are behind this convection, roughly in the blue oval. Note that the convective updrafts are not exactly co-located with the heaviest precipitation, which is located to the aft of the convection. Thus, the strongest radar echoes are not exactly co-located with where the strong upward motion is, which explains some of the discrepancy between idealized and modeled profiles of vertical motion and latent heating in convective elements seen in the previous module.

Slide 4: Such mesoscale systems often have a prominent overturning circulation in the stratiform region. Aloft, a front to rear ascending outflow is present, and below the 0°C level, a descending rear-to-front inflow occurs, transporting low theta-e air from aloft down toward the surface in stratiform regions. The descent is accompanied by adiabatic warming and diabatic cooling driven by evaporation of rain drops.

Slide 5: This figure from a later paper is based on observations of canonical MCSs and shows the same salient features as that predicted prior. The shaded regions indicate where enhanced radar echoes are found: in the convective region and along the radar brightband in the stratiform precipitation. More recent updates of this figure show the anvil cloud height sloping downward with distance from the convective cores, like that seen at bottom left.

Slide 6: As seen previously, separating the deep convection from stratiform is difficult with visible imagery alone, but…

Slide 7: The coldest cloud tops can be easily detected using infrared brightness temperatures.

Slide 8: However, the structure of clouds beneath the cloud top is still invisible without passive or active microwave instruments. For example, radar can detect the size of the largest drops present in convection; larger drops typically indicate the presence of stronger updrafts. Shown here are examples of two MCSs. On the left is a broad stratiform region with embedded convection scattered within it. On the right is an example of an MCS with a leading convective line moving toward the southwest. The case on the right occurs in an environment with stronger vertical shear and winds aloft were likely southwesterly. Unlike satellite imagery, radar can only see precipitating regions, and without cloud radar, the non-precipitating anvil clouds are not fully, if at all, detected. A combined analysis of radar and satellite imagery is best used to characterize MCS structure, which can vary substantially as seen in this image.

Slide 9: Convection can take on a variety of organizational patterns, which is what the current figure expresses based on observations of mid-latitude continental MCSs. The canonical MCS that many schematics are based off of is that with a leading convective line and trailing stratiform region, denoted on this figure as TS. However, depending on the wind profile, stratiform can be seen parallel to the convective line or even ahead of it.

Slide 10: In the tropics, it is especially common to observe MCSs that take on no obvious structure; in other words, the convection appears to be randomly distributed within stratiform. This happens frequently in low-shear environments, which are common in the tropics. The “particle fountain” conceptual model, shown here, visualizes how many convective cells distribute hydrometers and moisture to upper parts an MCS, resupplying the upper-level stratiform and anvil cloud with moisture to persist. Such an MCS decays when the atmosphere becomes too stable for deep convection to develop. Indications of atmospheric stabilization will be warming of the middle to upper troposphere and cooling of the lower troposphere.

Slide 11: For the canonical leading line-trailing stratiform type of MCS, observations have shown the rearward ascending outflow and descending inflow mentioned earlier. The descending inflow sometimes drives formation of a cold pool in the boundary layer, which is a density current that we will describe more in another module. Conservation of vorticity drives these major flows in the stratiform region.

Slide 12: We can show this by starting with the vorticity equation for two-dimensional flow. Theta* represents a perturbation from an environmental mean. Given the small spatial scale, we can neglect the Coriolis effect. We can define a streamfunction such that its vertical derivative is the zonal wind and negative the zonal derivative is the vertical velocity. If we assume a steady state, meaning that the Eulerian derivative of vorticity is zero, and define relative vorticity as du/dy minus dv/dx, then we can plug the streamfunction in place of relative vorticity. The resulting equation tells us that the stream function is governed by large-scale environmental flow and the horizontal distribution of buoyancy.

Slide 13: Streamlines of storm-relative inflow are shown here. Vorticity is conserved following the streamlines in a sheared, unstable environment. Three types of circulations follow from conservation of vorticity: a “jump updraft”, which ascends moving from lower right to upper left, an overturning updraft, which sometimes breaks off from the jump updraft, and an overturning downdraft which feeds into a low-level cold pool. The updraft corresponds roughly with the convective region and front-to-rear ascending region, and the overturning downdraft corresponds with the descending rear-to-front inflow jet.

Slide 14: We can manipulate the vertical momentum equation while neglecting Coriolis and friction. I won’t take you through the derivation, but you end up with the Bernoulli equation in Boussinesq flow. If you move the buoyancy term to the right-hand side, what this equation tells you is that along a streamline, the CAPE released can be converted not only to vertical motions, but also to horizontal motions and enthalpy of the fluid. As a result, buoyantly driven motions are not concentrated just within updrafts in an MCS but are distributed throughout the system.

Slide 15: The microphysical structure of MCS scan also be examined in the context of the flow within a canonical MCS. Shown here is the microphysical makeup of the convective regions of MCSs, with the jump updraft and overturning updrafts shown by black arrows. The heaviest rain occurs near the jump updraft and is flanked by regions of moderate or light rain. A small area of rimed ice aggregates and graupel can be found above the 0°C level near the jump updraft. A region of melting, wet ice aggregates are presented in a brightband region rearward of the jump updraft.

Slide 16: In the stratiform region, the ascending front to rear flow is seen aloft, with the overturning downdraft seen below. Part of the descending downdraft breaks off into a cold pool that promotes convergence beneath or near the base of the jump updraft, reinforcing upward motion. Broad regions of light rain are topped by wet, melting aggregates. Occasional convection will promote the growth of some larger rimed particles. In the upper levels of both convective and stratiform precipitation, dry (non-melting) ice particles are present, and particularly in stratiform regions, small ice crystals are present. These ice crystals slowly fall through sedimentation and are often left behind as remnants of anvil cloud after the precipitating part of the MCS dissipates.

Module 2.3

Slide 1: This module will discuss the diurnal cycles of convection observed in parts of the tropics.

Slide 2: When convection is active, its precipitation is observed to vary based on the time of day, which is related to the variability in shortwave insolation that occurs over a 24-hour period. This figure illustrates the main mechanisms involved in the diurnal cycle of convection. During nighttime, two mechanisms dominate. The first is by altering lapse rate. During night, the bottoms of clouds warm because they absorb longwave radiation emitted upward from the surface below and emitted downward from cloud above. At the tops of clouds, radiative cooling occurs, because longwave radiation escapes to space and is only absorbed from emissions below. The low-level warming and upper-level cooling steepens the lapse rate in the troposphere, making the environment more unstable and promoting more intense updrafts.

Slide 3: The other growth mechanism occurs as a result of a local pressure gradient that develops at low-levels between clouds and their clear-air environment. While the bases of cloud warm due to absorption of longwave radiation, the clear-air environment cools by emitting longwave radiation to space. The cooler region induces a local high pressure, and the resulting pressure gradient drives low-level flow from the environment toward the convection, inducing low-level convergence, which was one of the ingredients we listed at the beginning of the course as a promoter of updraft growth. During the daytime, shortwave radiation heats deep convection, although not uniformly: Upper reaches of deep convection are warmed more, although this is partially cancelled out by longwave cooling that still occurs during the day. The vertical profile of heating induces two circulations: A deep circulation that drives inflow at mid-levels, and a shallow circulation that detrains moisture into the environment gradually and drives low-level convergence that promotes continued growth of shallow convection.

Slide 4: The vertical structure of radiative heating in non-precipitating anvil clouds reiterates these points. The figure displayed illustrates radiative heating rates in anvils, such as 0 on the y-axis equals cloud base, and 1 denotes cloud top. Solid black lines denote daily-mean longwave heating profiles for anvils with a variety of thickness. Dashed lines denote shortwave heating. You can see how longwave heating profiles denote heating at low levels of the clouds and cooling at upper levels—or at least less cooling at the base of a cloud than at its top for when considering thin cirrus. This promotes destabilization of the atmosphere by increasing lapse rate. During nighttime, when only longwave radiative transfer occurs, longwave cooling can cause a diurnal maximum in cloud depth, size, and precipitation. During the day, heating occurs through the cloud and tends to cancel out most of the longwave cooling aloft in anvils. Therefore, solar radiation supports stabilization of the atmosphere after being destabilized at night.

Slide 5: The diurnal cycle of precipitation and clouds are shown here for over tropical oceans during large-scale conditions that are—for the top panel—convectively active—and for the bottom panel, convectively inhibited. The blue dashed line in each panel denotes the rain rate as a function of local time of day, shown on the x-axis. In the deep tropics, daytime is typically something like 6AM to 6PM. The green shading indicates the magnitude of low-tropospheric humidity, and the red shaded region indicates the magnitude of the SST, which varies between 28.5 and 29.5°C in this schematic. During convective enhanced periods, the rainfall maximum occurs in early morning before daybreak and is accompanied by the deepest convection observed during the 24-hour period. Convection then decays after sunrise, but as convection dissipates, more solar radiation reaches the surface, warming the sea surface and enhancing surface fluxes while the troposphere—absent of as much convection—slowly dries. The energy-rich boundary layer promotes the growth of new shallow convection at night, which gradually moistens the lower troposphere. The low troposphere reaches maximum humidity a couple of hours before the maximum in rate rate; in other words, the low troposphere humidity is required before the deep convection can develop. In the bottom panel, during convectively suppressed conditions, the SST varies more with time and similarly reaches a late afternoon maximum. However, two muted peaks in rain rate are observed: One occurs as a result of increased convection promoted by the SST maximum. SST decreases after dark and convection weakens, but the convection that does develop moistens the lower troposphere, which allows for a secondary overnight maximum in rain rate to occur in the early morning hours. Of course, other variability occurs on top of the diurnal cycle, so convection may not follow this schematic every day, but the graphic does describe the long-term average behavior of convection over tropical oceans.

Slide 6: The relationship between rainfall, cloud depth and coverage, and sea surface temperature can be seen in model simulations as well. At top is total area coverage plotted in the red line and areal coverage colored as a function of height. At about 1 km altitude, a maximum in cloud areal coverage occurs in late afternoon and mostly consists of boundary layer convection. It corresponds with a maximum in sea surface temperature, seen on the bottom plot. The maximum cloud coverage at 3 to 4 km altitude occurs a few hours later and is followed shortly thereafter by maximum precipitation.

Slide 7: Simulated timeseries of Q2 profiles at top show that clouds cause moistening of the low free troposphere, denoted by the blue colors. The moistening is maximized in the model in early evening around 1800 local time, and a maximum in specific and relative humidity follows a couple of hours later as seen on the bottom plot. The anomalous humidity profiles seen at bottom in shaded colors are tilted to the right on the time axis, meaning the humidification occurs at higher altitudes *after* it occurs at lower altitudes. In other words, shallow convection mostly confined to the boundary layer moistens the troposphere where it resides before deeper convection develops.

Slide 8: Diurnal variability is also frequently seen in tropical cyclones. The diurnal cycle often manifests itself in the form of lower brightness temperatures—meaning colder and higher cloud tops—during nighttime. Cloud regions are often observed to expand outward away from the center of the cyclone during morning. This figure shows one such example by looking at infrared brightness temperatures. The bottom panels show the difference in brightness temperature between mid-afternoon and early morning, and the black rings indicate radii from the hurricane’s center. A ring of cold cloud tops can be seen expanding outward from the center in the early day into mid-afternoon.

Slide 9: Another example in a more sheared hurricane is shown here.

Slide 10: The daily outward expansions of low brightness temperatures, the shaded regions on the plots here, are seen in the Hovmöller plots for various storms in 1999. Time increases moving down the y-axis, and the x-axis shows increasing radius moving to the right.

Slide 11: The red arrows show examples of lower brightness temperatures moving down and right on the plot, meaning outward—or to larger radii—at later times.

Slide 12: Model simulations of idealized, steady-state tropical cyclones provide some insight into the mechanisms responsible for the apparent outward expansion of clouds in a diurnal cycle. Shown in the top plots are radiative heating and streamfunctions during nighttime (03L) and daytime (14L). Blue shading indicates radiative cooling, while red denotes radiative heating. During the night, deep upward motion is present near the core of the cyclone, and two overturning circulations are present. The bottom left panel is a time series of anomalous radiative heating and vertical motion relative to the long-term mean as a function of time of day. Upward motion is maximized at low-levels during the night and at upper-levels during the day as part of a radiatively-driven overturning circulation seen in the top right panel in which upward motion is partially driven by shortwave heating aloft in deep cumulonimbus clouds.

Slide 13: Plots of theta_v and streamfunction as a function of radius, like the previous Hovmöller plots we saw but turned 90 degrees, indicate that thermodynamic and dynamic features in the TC propagate outward with speeds of about 10 meters per second, which corresponds closely with the observed outward propagation speed of brightness temperatures. It is hypothesized that the outward propagation is driven by inertia-gravity waves forced by radiative heating of upper-level cloud; however, the propagation of the wave does not appear to be dynamically coupled to precipitation, meaning that it does not reinforce or get reinforced by latent heating associated with condensation.

Slide 14: Different diurnal variability is seen over land, and the diurnal variability over land can furthermore impact that seen over nearby oceans. Seen here are observed diurnal cycles of rainfall during boreal summer over the Philippines. The amplitude of the diurnal variability is much larger over land than ocean, as seen in the leftmost panel. This occurs because terrain-induced precipitation is strongly forced by diurnal variability. The time of day of maximum precipitation as a function of space is shown in the middle panel. Rainfall peaks over land in later afternoon. Convection then begins to move offshore, occurring at later hours overnight over adjacent seas. For example, mesoscale convective systems often from near land during late afternoon and propagate offshore over the South China Sea. Thus, the maximum rainfall over the sea is typically at nighttime, although factors other than just diurnally varying radiation may influence when and where rain occurs. The diurnal cycle is so pronounced over land in some places because convection is forced topographically. During the day, the land warms much faster than the ocean. As a result, a sea breeze circulation—driven by pressure gradients—develops. The pressure gradients drive flow onshore during the daytime and forces convergence near the bases of steep terrain.

Slide 15: Similar variability is seen in this figure over the Maritime Continent. The figure is plotted such that time proceeds in a clockwise direction, with midnight local time in the bottom right panel. Peak rainfall over the land occurs in the late afternoon or early evening, after daytime insolation has warmed the land surface. Convective elements and associated stratiform regions then move offshore during the overnight hours, causing rain to maximize over adjacent seas overnight. A daytime minimum in rainfall is seen around midday. Large-scale dynamics also impact diurnal variability and affect the propagation of various equatorially trapped features. If convection becomes “trapped” over land and fails to propagate offshore, the upward transport of moist, warm boundary layer air from over the ocean is limited, reducing the amount of water vapor and upper-level cloud present, which impacts radiative heating profiles—further feeding back onto the diurnal cycle.

Module 2.4

Slide 1: This module will discuss some of the characteristics of cold pools, which thus far, we have discussed in the context of downdrafts in mesoscale convective systems. Cold pools are one mechanism for forced low-level convergence that can support the vertical growth of convection in the tropics.

Slide 2: Essentially, a cold pool is a density current, also called a gravity current. Dense, generally relatively cold air, advances into warm air like a mesoscale front, forming a mesoscale high near the base of the jump updraft in an MCS and producing overturning flow and a mesoscale low near the head of the density current. A secondary pressure maximum occurs at the rear of the cold pool head. Cold pools also frequently occur outside of MCSs. They can form as low theta-e air accompanying downdrafts in isolated convection propagate outward within the boundary layer from where the downdraft encounters the surface. Convergence at the leading edge, or nose, of the cold pool forces upward motion which, if strong enough, can reach the LCL and form an arc cloud and possibly deeper cumulonimbus clouds.

Slide 3: The advance of gravity currents is driven by the pressure gradient force. Given a simple two-dimensional cross-section in the x- and z-directions only, and assuming a steady state, we can determine the propagation speed by starting with the horizontal momentum equation. Given some depth of the cold pool—we'll call it little *h*—and a density that is larger than environmental density by delta rho, the forward propagation speed of the gravity current is related to the phase speed of a gravity wave modified by the density difference. The difference in density across the boundary of the cold pool is controlled by both temperature and moisture, and so we consider the difference in theta-e between the cold pool and its environment. A large difference in density, or a cold or relatively dry current, will propagate more quickly into the environmental air. Gravity currents also propagate more quickly as they grow deeper.

Slide 4: We can also manipulate the horizontal momentum equation to show that following the trajectory of a parcel in the cold pool, that the sum of the kinetic energy associated with motion and enthalpy are conserved. If we integrate along *x* to some distance well ahead of the cold pool where the ambient flow is toward the cold pool at velocity capital *U*, and from a point immediately ahead of the gust front, we find that the pressure perturbation results from conversion of kinetic energy to enthalpy. Thus, the pressure perturbation at the leading edge of the cold pool is not hydrostatically driven. Furthermore, the meso-low at the top of the head is also non-hydrostatic, and while not shown here, is related to rotation at the top of the cold pool head. The meso-high at the rear of the head is hydrostatically induced, meaning that it is related to the column-integrated mass at that location and above.

Slide 5: We can also, from the 3D momentum equation with anelastic assumptions, divide the pressure perturbation into its hydrostatic contribution, the buoyancy pressure perturbation, and a dynamic pressure perturbation. The total pressure perturbation is seen at top. It is dominated by the buoyant pressure perturbation in the rear of the gust front. However, right at and just in front of the gust front, the dynamic pressure perturbation dominates as discussed on the previous slide.

Slide 6: Numerous cold pools can co-exist in the same environment, and their collision enhances convergence and promotes deep convection. On the left is an example of a rainfall-induced cold pool. Low theta-e air descends and spreads out through the boundary layer. Shallow convection forms at the edge of the cold pool, where high theta-e environmental air is forced over the head of the advancing gravity current. The image at right illustrates two intersecting cold pools. Shallow convection will form where a cold pool expands outward, but convergence is enhanced where two cold pools intersect. Convection is preferentially forced in locations where more low-level convergence occurs, and as the convection intensifies, inflow of high environmental theta-e air further fuels the updrafts, cutting off moist static energy from shallow convection elsewhere along the cold pool boundaries.

Slide 7: The typical updraft velocities along simulated tropical cold pool boundaries is shown here. The average updraft velocities in isolated vs intersecting cold pools are similar, as denoted by the red and blue boxes. However, the blue and red arrows denote extreme events. The strongest intersecting cold pools are stronger than the most intense updrafts in isolated cold pools. Therefore, the most intense convection that is triggered by cold pools typically occurs where two or more gravity currents collide.

Slide 8: Finally, some examples of cold pools on radar imagery are shown. Radar reflectivity is shown at a low elevation angle so that we can see echo in the boundary layer close to the radar and low free troposphere over about 75 km from the radar. The figure proceeds in time from a to b to c to d. At time a, a cold pool is present inside the white dashed circle. The cold pool appears as a clear echo region. The cell with the white arrow produces a cold pool, which can be seen in panel b, which was about 30 minutes later. Some of these cold pools expand laterally outward at later times, and the cluster of cold pools to the northeast of the radar site appear to merge together. New convective elements can be seen developing along the edges of these cold pools.