Introduction:

Satellite imagery: We will begin this class by looking at some geostationary satellite imagery. Specifically, we are looking at GOES-West imagery over the East Pacific. Tropical dynamics are unavoidably linked to moist convection, a variety of which is seen in this satellite loop from June 25, 2020. Many different features are present, but in this class, we will focus on the tropics, which are primarily confined to within 20 degrees of latitude from the equator. Just in this snapshot in time, you can see many different features that we will discuss in this class. For example, a tropical cyclone is present southeast of Hawaii. The intertropical convergence zone in the East Pacific is apparent, and it contains numerous mesoscale convective systems. Lines of cumulus clouds and cold pools are visible. Large stratocumulus decks are apparent of the west coasts of North and South America. In the far right of this disk image, you can see dust associated with a dry Saharan Air Layer reaching North America that generally inhibits convection. We will spend most of this class discussing various types of tropical convection. Moist convection is central to tropical dynamics; it both strongly influences tropical motions and is influenced by them. Convection is associated with some of the most high-impact weather events we will discuss. First, we will review some of the fundamental dynamics that control tropical motions.

Module 1.1:

Slide 1: In our first module, we will review fundamental equations for dynamics and discuss factors influencing the growth of tropical convection. We will then discuss one of those factors in more detail, the importance of environment moisture for cloud growth.

Slide 2: Synoptic-scale motions can be approximated in the mid-latitudes using QG theory. In the beginning, there were three equations of motion: the momentum equation, shown here in two-dimensional form, the mass continuity equation, and the thermodynamic equation, in which *Q* represents a heating rate in units of K or degrees C per unit time. One of the first things you learn in mid-latitude dynamics is that the atmosphere is approximately in geostrophic balance above the boundary layer. That means that the Coriolis “force” balances the pressure gradient force. This assumption cannot be made in the tropics, however. Why not?

Slide 3: Let’s look more closely at the momentum equation. We can’t assume geostrophic balance because f is approximately 0. Recall that f is proportional to the sine of the latitude, and the sine of zero is zero. Also recall that to be able to assume geostrophic balance, a very small Rossby number, say less than 0.1, is required. The Rossby number is just wind speed divided by *f* times a length scale, so as *f* goes to zero, the Rossby number gets big. On planetary spatial scales and away from the equator, we can roughly assume geostrophic balance, but most of the phenomena we discuss in this class are much smaller in spatial scale than O(10000km). This means that all three terms in the momentum equation are similar in magnitude and that none of them can be neglected.

Slide 4: Next, we’ll consider the thermodynamic equation. Diabatic heating is absolutely crucial to tropical dynamics. For many first-order approximations of QG theory, diabatic heating can be neglected, although it does become important when considering how the vertical gradient of heating affects the evolution of geopotential and when considering ageostrophic motions, such as along a front. When removing diabatic heating in QG theory, we are left in the thermodynamic equation with a balance between the Eulerian temperature tendency and 3D advection. In the tropics, however, upward motion is largely driven by release of latent heat associated with condensation and freezing of water. Furthermore, horizontal temperature gradients in the tropics are small, unlike in the mid-latitudes, where colliding air masses along fronts give rise to large temperature gradients that make the advective terms in the Lagrangian derivative large. In the tropics, u*dTdx and v*dTdy are small, and the time tendency of temperature is primarily a balance between diabatic heating and adiabatic cooling associated with upward motion. To first order, the two balance each other outside of the atmospheric boundary layer, and the time-tendency of temperature is zero; this is known as the **weak temperature gradient**, or WTG, approximation.

Slide 5: Several factors work jointly to impact the growth of tropical convection. A few are listed here. Perhaps the most important factor is availability of moisture, especially outside of the atmospheric boundary layer and below the 0°C level. A dry atmosphere prevents the growth of convection via entrainment. Static stability and surface fluxes of energy are other thermodynamic factors that affect the buoyancy of updrafts in moist convection. Wind shear and low-level convergence are dynamic factors that impact convection. Some recent research suggests that wind shear can dynamically alter the buoyancy of updrafts, but it may also alter the structure of isolated moist convection such as to make updrafts more likely to entrain environmental air and become less buoyant. Wind shear through a deep tropospheric layer is also involved in the organization of convection into laterally expansive mesoscale convective systems. Finally, low-level convergence in the lower troposphere can provide additional upward forcing for updrafts to overcome thermodynamic convective inhibition in the environment. For the remainder of this module, we will focus on the impact of moisture on vertical growth of convection.

Slide 6: The relationship between tropospheric moisture and precipitation is well documented. Two recent examples of the relationship in observations is shown here. The panel at left is based on ground-based radar and weather balloon data, and the panel to the right is based on satellite-derived retrievals of humidity and rain rate. On the x-axes are column relative humidity, which is described by the equation at the bottom. The column relative humidity is the integral of specific humidity through the atmosphere divided by the integral of saturation specific humidity, which is a function of temperature. If the atmosphere were completely saturated, CRH would be 1. The figures illustrate two important points:

Slide 7: 1) Below CRH of 0.6, significant rain rates are rare outside of isolated cumulonimbus clouds.

Slide 8: 2) An exponential pickup occurs around CRH of 0.7 to 0.8

Slide 9: and 3) While large rain rates can occur as CRH exceeds 0.8, there are many instances of little to no rain occurring in a moist environment. Therefore, a moist atmosphere is a necessary but insufficient condition for widespread heavy rain to occur. This might happen because other factors that are important for deep convection to form are unfavorable. Perhaps the environment is too statically stable, or perhaps there is insufficient low-level forcing to support deep convection. The reasons for this scatter remain an active topic of research.

Slide 10: Let’s take a look at a very idealized drawing of a cloud. This cloud has a parent updraft that extends through the center of the cloud. It is surrounded by a moist, but unsaturated, “shell”, where mixing between saturated cloudy air and unsaturated environmental air occurs.

Slide 11: Why does a dry environment inhibit the vertical growth of clouds though? The process through which environmental air from outside of a cloud is brought into a cloud is called **entrainment**, for which the AMS Glossary definition is shown here. Air inside cloud updrafts gets diluted with cooler and/or drier environmental air, making the updrafts less buoyant, thereby reducing the upward acceleration felt by updrafts. We’ll walk through the basics of what this process looks like mathematically next.

Slide 12: Let's consider the simple, idealized case of continuous, homogenous entrainment. This means that the entire cloud feels the effects of entrainment in the same way. This is obviously a gross oversimplification, but it is sufficient for describing how entrainment of dry air inhibits vertical growth of convection. At the bottom of the figure is a cloud that exists at time *t* with some mass *m* and any variable related to the enthalpy of the cloud denoted by the scripted H. Subscripts c and e, respectively mean cloud and environment. We could treat the scripted H as if it represents momentum, moist static energy, or moist enthalpy in the cloud. At the top of the figure is the same cloud at some time delta t later. It has a new mass that includes mass entrained from the environment and excludes mass detrained out of the cloud. The new enthalpy variable is the original plus whatever change is caused by entrainment. Finally, a source term includes in-cloud processes that would occur even in no entrainment/detrainment occurred. These include microphysical sources/sinks. For example, evaporation of liquid water would cause an increase in water vapor, or in terms of this equation, an increase in Hc.

Slide 13: If we take the limit of this equation as the change in time approaches zero, we get that the material derivative of our enthalpy variable equals the source plus a term that is proportional to the difference in enthalpy between the environment and cloud. The final equation tells us that moist enthalpy in a rising updraft changes based on the balance between entrainment and internal processes.

Slide 14: If we assume that a cloud updraft is saturated and has risen out of the boundary layer moist adiabatically, then it will generally be warmer and moister than the environment. Therefore, $h_c$ is greater than $h_e$, and entrainment will tend to make the moist enthalpy inside the cloud decrease. An environment that is cooler or drier than an updraft will tend to decrease the buoyancy of that updraft. Consider the definition of the buoyant force, shown here. Buoyancy is an acceleration that is proportional to the magnitude of the difference in density between some air parcel and surrounding air. The 3D momentum equation tells us that positive buoyancy increases upward acceleration. This just means that if an updraft is less dense than its surroundings, it will accelerate upward. You've probably heard the notion that warm air rises. It does so because it is less dense than its environment and experiences upward buoyant acceleration. Consider how a change in water vapor concentration changes rho-prime. A water vapor molecule is lighter than both molecular nitrogen and molecular oxygen, the two primary constituents in Earth's atmosphere. Therefore, a saturated cloud will always be less dense than surrounding air with the same temperature. That makes rho-prime negative, and when considering the negative sign in the equation for buoyancy, that saturated air parcel experiences upward acceleration. Entrainment is a process that happens at small spatial scales in individual clouds. Therefore, numerical models of the atmosphere cannot resolve the process explicitly. It must be parameterized, or described in a mean or ensemble sense across the entirety of some area much larger than an individual cloud. Because of the great sensitivity of cloud buoyancy to moisture, accurate representation of tropical rainfall in models is extremely sensitive to parameterizations of entrainment.

Module 1.2:

Slide 1: We will continue our discussion of thermodynamic factors that impact the vertical growth of moist convection by discussing how static stability impacts convection.

Slide 2: Recall from the previous modules that several processes can support or inhibit deep convection. Sufficient moisture is of paramount importance. It is generally difficult for tropical convection to become deep if the environment is insufficiently moist; however, sufficient moisture is insufficient alone to produce deep convection. It is also important to consider processes that affect the five factors listed at the top of this slide. For example, while low-level convergence can dynamically force convection, it can also cause moisture convergence, which helps to satisfy the moisture requirement for deep convection. Sea surface temperature and near-surface wind impacts surface fluxes as we will soon see. Adiabatically driven vertical motions affect static stability. Even moist convection itself is important for development of subsequent convection. The primary moisture source to the middle troposphere is convection, so in some sense, to get widespread moist convection in the tropics, you first need convection. This seems like a contradiction; however, in later modules, we will explain why this is sometimes the case.

Slide 3: For now, we will focus on static stability, an expression for which is shown here. The static stability parameter in the thermodynamic equation is proportional to the difference between the dry adiabatic lapse rate, which is close to 9.8 K/km, and the environmental lapse rate, which varies in space and time. We have defined $S_p$ to generally be positive, so that when the environmental lapse rate is less than dry adiabatic, positive $S_p$ corresponds with upward motion and warming by condensation. The difference in lapse rate can also be expressed as proportional to the vertical gradient in potential temperature. Essentially, cloudy air rising moist adiabatically out of the boundary layer will experience more positive buoyancy (or less negative buoyancy) as the environmental temperature decreases. An increased lapse rate means that the temperature of the environment cools more quickly with height. $S_p$ can also be defined as proportional to the negative vertical gradient of dry static energy. Dry static energy generally increases with height.

Slide 4: Let’s consider an example of a sounding we might see in the tropics. In the next module, we will review with more detail how to interpret soundings. This sounding is conditionally unstable, but is nearly moist neutral—a rather common situation over tropical oceans where the sea surface temperature is high, such as in the Indian Ocean, West Pacific, or adjacent seas.

Slide 5: The red line is another example of a conditionally unstable sounding. However, it is more statically unstable because the temperature decreases faster with height. Note that we aren’t yet considering moisture in our sounding at all.

Slide 6: Finally, suppose a parcel rises out of the boundary layer. A typical LCL over the west Pacific is 950 mb, and above that the parcel rises moist adiabatically along the black line, which starts near the surface with a similar temperature as the red and blue lines. The difference between the black line and the other two lines describes the instability. As you can see, the black line is much farther away from the red line than from the blue line, so the parcel will tend to be more buoyant in an environment that has a vertical temperature structure resembling the red line. This is related to the concept of convective available potential energy, or CAPE, which we will discuss more in a different module. Eventually, we will combine the effects of moisture and temperature into a single stability metric that is based on equivalent potential temperature instead of $T$ or theta, but the main takeaway here is that cooling the middle troposphere—say between 850 and 500 mb—more than the boundary layer promotes the vertical growth of shallow convection.

Slide 7: The effect of static stability on convective updraft intensity can be seen in this figure, which shows the flash density of lightning globally. Note how the majority of lightning occurs over continents. A few lightning hotspots are over central Africa, the Amazon and plains of southern South America, and basically any large land masses in the deep tropics, such as islands in Indonesia. In contrast, oceans experience relatively fewer lighting flashes. To first order, the reason for this involves the typical static stability profile over tropical land masses versus the ocean. Over the ocean, in places where the ocean is warm enough to support widespread deep convection, the environmental temperature profile often sits near a moist adiabat. Over land, the lapse rate tends to be steeper, meaning that the temperature decreases more quickly with height. The consequence of this is that updrafts over oceans tend to be weaker than those over land and therefore less likely to produce lightning. Nonetheless, we’ll see that some of the largest annual-averaged rainfall totals on Earth occur over the ocean. Surface fluxes, which we discuss in the next module, are responsible for this.

Module 1.3:

Slide 1: This module discusses how and why fluxes between the ocean and atmosphere are important for convection.

Slide 2: We also must consider surface fluxes when evaluating the potential for convective growth. Where flux of energy from the ocean to the atmosphere is low, deep convection tends to not occur. There are two types of surface fluxes we will discuss: Latent heat fluxes and sensible heat fluxes. The latent heat flux is that between the atmosphere and surface associated with the phase change of water. Evaporating ocean water into the atmosphere is one example of a positive flux of latent heat to the atmosphere. Eventually, the evaporated water will condense back into liquid water, and when it does, it will cause warming as water transitions into a lower energy state. The sensible heat flux is a conductive flux between the surface and the atmosphere. A warmer surface translates to a higher sensible heat flux. One way of defining the two terms is shown below. It is known as the "bulk" expression of the fluxes, which contrasts with a different method called the eddy covariance method. Note a couple of things: First, both fluxes are proportional to the near-surface wind speed, *U*. This means that fluxes go to zero where there is no wind near the surface. Second, the heat fluxes are proportional to the difference between properties of the sea surface, denoted by a subscript *s*, and properties of the atmospheric layer just above the ocean surface, often known as the surface layer. $q_s$ is computed as the saturation specific humidity that air would have if it were the same temperature as the ocean surface. The C-variables are known as bulk transfer or exchange coefficients, estimates for which are beyond the scope of this course.

Slide 3: The previous formulation indicates that fluxes vary linearly with *U*; however, reality is a bit more complicated. As wind increases, surface roughness increases, and this causes a change in the exchange coefficient. As seen in this plot, latent heat fluxes estimated from ship data increase nonlinearly as wind speed increases. However, the same general concept of fluxes increasing with increased wind generally holds true.

Slide 4: Consider some typical profiles of variables in a convective boundary layer. A few different layers within the atmospheric boundary layer are denoted. The surface layer is the thin layer closest to the ocean surface. The mixed layer makes up most of the boundary layer, and then above that is an entrainment zone, in which free tropospheric air is mixed with boundary layer air. A few different variables are plotted here. You can see that almost each is relatively constant in the mixed layer. The first is virtual potential temperature, which is a function of potential temperature and water vapor mixing ratio. Specific humidity tends to slowly decrease through the mixed layer, while wind speed increases from near zero right at the surface. The variable of interest that we will focus on next is the **buoyancy flux**. It is proportional to the eddy covariance of vertical velocity and virtual potential temperature and it decays with height quickly in the mixed layer. A positive buoyancy flux in the surface layer means that the ocean is transferring energy to the surface layer, such that it supports the upward acceleration of thermals that develop in the mixed layer and rise to become cloudy updrafts. A detailed explanation of the eddy covariance flux method for estimating fluxes will not be covered here but can be found in a course on boundary layer meteorology or air-sea interaction.

Slide 5: We will look a little more into the buoyancy flux however. Because it is related to vertical kinematic flux of $\theta_v$, and $\theta_v$ incorporates both moisture and temperature, the buoyancy flux is related to fluxes of latent and sensible heat. The eddy covariance expressions of turbulent latent and sensible heat fluxes are shown here. Exchange coefficients, wind, and the difference between ocean and atmospheric properties have been folded into the terms $w'q_v'$ and $w'\theta'$. Some algebra can lead you to re-expressing the buoyancy flux in terms of the latent and sensible heating fluxes. Note that for typical sea surface temperatures of about 300K, the coefficient $0.61*c_p*T/L_v$ is about 0.08. This has important implications for the relative importance of latent and sensible heat flux on buoyancy of rising thermals in the boundary layer through the buoyancy flux. For a typical value of q' of order 0.1 and θ' of order 1, the latent heat flux is roughly one to two orders of magnitude larger than the sensible heat flux. However, the contribution of the latent heat flux to the buoyancy flux is similar as that of the sensible heat flux. In other words, even though latent heat flux often dominates sensible heat flux in magnitude, they are similarly important for supporting buoyant updrafts that develop in the boundary layer.

Slide 6: Consider a very simple example. The blue line represents the ocean surface. The boundary layer above it is well-mixed, has a temperature of 28°C and a specific humidity of 18 g/kg. Let's consider two different cases with this setup: One in which near-surface wind is 5 m/s and the sea surface temperature, or SST, is 25°C; and another in which near-surface wind is only 2 m/s, but the SST is 30°C. In which case is the flux from the ocean to the atmosphere larger? Pause here for a moment and see if you can figure out the answer.

Slide 7: The case with the warmer ocean has the larger flux of heat to the boundary layer despite its lower wind speed. This is because the 30°C ocean is warmer than the 28°C boundary layer. In the case of the 25°C ocean, the atmosphere would actually flux sensible heat to the ocean. Based on the relationship between the buoyancy flux and heat fluxes, this means that a cool SST indirectly inhibits the upward acceleration of convection. It does so by reducing the convective instability of the environment by making the boundary layer cooler.

Slide 8: Next let's consider a different set of scenarios. On the right, we still have the 30°C SST with 2 m/s wind, but on the left with a 5 m/s wind, we now also have SST of 30°C. In which is the heat flux largest? Pause for a moment again to answer before moving on.

Slide 9: You hopefully figured out quickly that given the same SST, the scenario in which the near-surface wind is higher will experience the greater flux of energy from the ocean to the atmosphere. In other words, all else being equal, a strong, steady boundary layer wind can support growth of convection through enhancing surface fluxes. This is generally true for typical wind speed observed over tropical oceans, although the effect of wind on fluxes becomes more complicated at high wind speeds, such as those observed in tropical cyclones.

Slide 10: The importance of surface fluxes on precipitation is readily seen in any map of climatological annual mean rainfall on Earth. Seen here is a map of annual mean SST. You should be highly familiar with maps like this by the time this course is over. The highest SSTs in the world are typically found over the Indo-Pacific warm pool. Warm SSTs extend in a narrow band north of the equator across the Pacific, and very warm SSTs are locally found near Central America, the Caribbean Sea, and in the summer over the Gulf of Mexico and tropical Atlantic. Relatively cool SSTs at low latitudes are often found off the west coasts of continents, where cold upper-ocean currents advect water from higher latitudes toward the equator.

Slide 11: A map of climatological mean precipitation rate is shown here. The pink, purple, red, and black colors indicate where the most rainfall typically occurs. In the tropics, this corresponds closely to the regions where SST was highest. The intertropical convergence zone and South Pacific convergence zone are particularly visible here. A bullseye of rainfall over the interior Amazon is also visible to the east of the Andes. Mid-latitude storm tracks are also visible as well as the local influence of terrain, such as near the coast of southeastern Alaska or the west side of New Zealand. Note also the lack of rainfall in regions of the tropics where SST is low, such as off the west coast of equatorial South America or Africa. In such places were surface water is fairly cold, convection is generally not sufficiently buoyant to reach far upward into the troposphere. Therefore, clouds cannot transport moisture far into the free troposphere, meaning the environment is dry above the boundary layer. Furthermore, as will be discussed in a different module, free tropospheric air tends to subside where moist convection is not active in the tropics. This increases static stability of the environment above the boundary layer, further decreasing the likelihood that long-lived deep convection can develop where SST is low.

Module 1.4:

Slide 1: In this video, we will review how to read a Skew T-log P thermodynamic diagram. You'll also be familiarized with some important variables that we will use throughout the course, and we will put some of those variables into the context of the Skew T plot.

Slide 2: This is a Skew T-log P plot. It is called such because the pressure coordinate on the vertical axis is plotted logarithmically to be roughly proportional to height. The temperature axis is skewed from lower left to upper right as will see momentarily.

Slide 3: Zooming in on part of the plot, let's look over what the different lines represent. The horizontal lines represent pressure levels. In this plot, they are labeled every 50 mb and plotted in 10 mb increments. Soundings generally contain two lines: One, the bold red line here, represents the temperature, and the bold green line denotes the dew point observed.

Slide 4: The slanted, or skewed, temperature lines are highlighted here. The red dot represents a temperature of 20°C near 830 mb. The dewpoint at the same pressure level is about 19°C. The labels for each line are shown here and are plotted in increments of 1°C.

Slide 5: Although they are faint, if you look closely, you will see dashed green lines that also slant upward and to the right. One is highlighted here. These are lines of constant mixing ratio. The labels for these lines area shown at the bottom in various increments. At 880 mb, the mixing ratio is about 18 g/kg. You get this by following the green line. The saturation mixing ratio can be determined by where the temperature plot intersects lines of constant mixing ratio.

Slide 6: Dry adiabats, or lines of constant potential temperature are highlighted here. Potential temperature is defined as the temperature that a parcel would have if it descended adiabatically to 1000 mb. Therefore, dry adiabats and isotherms with the same value intersect at 1000 mb. In this example, the potential temperature at 910 mb is about 32°C, or 305K.

Slide 7: The purple line now represents a moist adiabat, or a line of constant equivalent potential temperature. They are labeled by the wet bulb potential temperature associated with each adiabat. The slope of the red temperature lines relative to the adiabats helps quickly determine the static stability of an observed profile in various layers of interest.

Slide 8: At high altitude, the moist and dry adiabats converge as the temperature lowers and the saturation specific humidity goes to zero.

Slide 9: You should be familiar with a few quantities. On the next slide, we'll look at how to find potential temperature (theta), equivalent potential temperature (theta-e), and saturated equivalent potential temperature (theta-es or theta-e*), the definitions of which are laid out here.

Slide 10: Suppose we have this hypothetical sounding with profiles of temperature and dewpoint and we are interested in the values of the three variables previously mentioned at 900 mb.

Slide 11: The dewpoint and temperature at 900 mb are denoted by blue dots. The temperature and dewpoint are respectively about 25°C and 21°C. Potential temperature is found by simply following the dry adiabat that the temperature profile intersects at 900 mb down to 1000 mb and reading off the temperature. It represents the temperature of dry air after it moves adiabatically up or down to 1000 mb. Once you become more familiar with Skew T plots, you can simply read off the temperature corresponding with the appropriate dry adiabat without following the adiabat all the way down to 1000 mb.

Slide 12: The equivalent potential temperature is the temperature that air would have if it were dry adiabatically lifted to saturation, then lifted moist adiabatically until no water vapor remained, then descended dry adiabatically to 1000 mb. It is a useful combined metric for describing the moisture and temperature of the environment and, like moist entropy, is conserved during reversible, adiabatic processes. To compute theta-e for this example, on a Skew-T plot, follow a dry adiabat upward from the temperature at 900 mb. Simultaneously, follow a line of constant mixing ratio up from the dewpoint at 900 mb. Note the moist adiabat where the two lines intersect, then follow that moist adiabat all the way up until it parallels the dry adiabats in the upper troposphere. This is where essentially all water vapor has been removed from the parcel. The value of the dry adiabat you end up on is the equivalent potential temperature. Note that the values of dry adiabats, particularly ones with high values, are also listed on the top and right axes of the plot.

Slide 13: The saturation equivalent potential temperature is simply a function of temperature and pressure. It describes the maximum possible equivalent potential temperature of air if it reached saturation at the specified temperature and pressure. For a temperature of 25°C at 900 mb, follow the moist adiabat all the way up until it parallels a dryadiabat. Like you did for theta-e, read off the value of the dry adiabat to get saturation theta-e.

Slide 14: Some other variables you may encounter include virtual potential temperature, which is the potential temperature needed by dry air to have the same density as air with some specified mixing ratio. You sometimes see theta_v used in expressions for buoyancy. The dry and moist adiabatic lapse rates describe the temperature change experienced by a rising parcel per unit altitude as it moves up a dry or moist adiabat respectively. The dry adiabatic lapse rate is about 9.8 K/km, and the moist adiabatic lapse rate varies with pressure, but is roughly 5 K/km in the lower troposphere. The environmental lapse rate simply describes the temperature change with height actually observed in the environment. Comparisons of the environmental lapse rate to the dry and moist adiabatic lapse rates help us assess environmental stability.

Slide 15: In this example, the slope of the red line denotes the environment lapse rate. In the blue box, bounded by 700 and 800 mb, the environmental temperature profile has a slope greater than that of the moist adiabatic lapse rate, but smaller than that of the dry adiabatic lapse rate. So, the environmental lapse rate is somewhere between 5 and 9.8 K/km—perhaps about 7 K/km.

Slide 16: For completely dry air, we can use vertical gradients of potential temperature to assess stability; however, for moist applications, we are concerned about the stability of moist, but not necessarily, saturated air. For this, we consider the vertical gradient of saturation equivalent potential temperature, which you may recall is a function of temperature and pressure. If the environmental lapse rate exceeds the dry adiabatic lapse rate, the environment is absolutely unstable; in this case the slope of the temperature profile would exceed that of a dry adiabat. If it less than the moist adiabatic lapse rate—meaning that the vertical gradient of theta_es is positive, then the environment is absolutely stable; in this case, the slope of the temperature profile would be less than that of even a moist adiabat. A moist neutral temperature profile is one that follows a moist adiabat, and a conditionally unstable environment is one in which the environmental lapse rate is between the dry and moist adiabatic lapse rates. Both absolutely unstable and conditionally unstable environments have negative vertical gradients of theta_es; however, the two stability metrics can be separated by assessment of the environmental lapse rate. A conditionally unstable environment is one in which a parcel can become unstable if it is lifted to saturation (the lifting condensation level) dry adiabatically and then onward moist adiabatically to the pressure level at which it is warmer than the environment. This level is known as the level of free convection. Finally, stability can be diagnosed in different layers. The same sounding can contain stable, conditionally unstable, and absolutely unstable layers.

Slide 17: Mean profiles of tropical potential temperature, theta-e, and theta-es are seen here. On average, you can see that motions below about 500 mb are stable to dry motions, but the environment would be considered conditionally unstable because theta-es decreases with height. The typical tropical atmosphere becomes nearly moist neutral in the mid to upper troposphere, and of course the atmosphere rapidly becomes stable above the tropopause.

The next module will focus on the interpretation of Skew-T thermodynamic diagrams.

Module 1.5:

Slide 1: This module is a continuation of the previous one. In this video, we will review a few more variables of importance then interpret a variety of tropical thermodynamic profiles plotted on Skew-T log-P diagrams.

Slide 2: Some additional quantities that will prove useful are shown below. The total precipitable water is the depth of liquid water that would be produced in a unit area if all the water in a column were condensed and rained out. It is related to the column relative humidity discussed in the first module. High values of TPW—say above 55 mm—correspond to values of column relative humidity that are sufficient for widespread, deep convection. Moist static energy is similar to theta-e; it is approximately conserved in adiabatic processes. It is the moist enthalpy plus geopotential. The gross moist stability, or GMS, is a metric that is sometimes used to assess the large-scale thermodynamic support for widespread deep convection. If GMS is negative, the environment is considered unstable. We won't see GMS on a Skew-T diagram, but it is a useful concept that will come up again later in the course, and I bring it up now because of its relationship to the MSE profile.

Slide 3: A typical MSE profile in the tropics is seen at left. It looks like the theta-e profile seen before and has a minimum somewhere in the middle troposphere. Negative GMS occurs when the net transport of MSE into a column is positive. The middle panel shows an example of convergence in the low-troposphere, where MSE is high, combined with middle tropospheric divergence, where MSE is low. This could occur in an environment with a low-level maximum in vertical motion, such as when shallow convection is prevalent. In this case, the column integrated divergence of MSE would be negative; there would be net convergence of MSE to the column, making it more unstable. The right-most panel shows a contrasting example of a vertical motion profile with an upper-tropospheric maximum, which is more indicative of deep convection. In this case, convergence occurs through a deep layer in the low to middle troposphere, so the net convergence of MSE into the column is lower than in the first example. Divergence occurs in the upper troposphere, where MSE is large. Therefore, net divergence of MSE in the column is positive, meaning that convection stabilizes the environment. It does so by transporting MSE from the boundary layer to the free troposphere.

Slide 4: Finally, convective available potential energy, or CAPE, is a useful metric to consider when consulting soundings. It is proportional to the difference in parcel and environment virtual temperatures, integrated from the level of free convection to the equilibrium level.

Slide 5: On our example sounding, the CAPE is the region between the black line and the red line. How exactly the CAPE is calculated depends on how the moist layer is treated. In this example, I have represented surface-based CAPE. That means that I connect a dry adiabat and a line of constant mixing ratio as if I were computing theta-e, then follow it up the moist adiabat until it reaches the equilibrium level, which is the level above the LFC where the parcel temperature has cooled to the environmental temperature. This sounding has very little CAPE, primarily because the boundary layer is quite dry. The LCL is the pressure level at which the dry adiabat representing the initial boundary layer parcel meets the mixing ratio line represented of the parcel. The LFC is the pressure level at which the parcel is no longer cooler than the environment as it rises.

Slide 6: If I add some moisture to the boundary layer, you can see the impact on CAPE. Suppose the blue line, instead of the green line represents dew point below 850 mb. In this case, the surface-based LCL is around 970 mb, the LFC around 800 mb, and the EL around 140 mb. This sounding has much more CAPE, meaning that an undiluted updraft that reaches the LFC could become particularly intense. This shows how moisture is not only important to convection because updrafts can entrain dry environmental air as they rise through the troposphere, but also that the moisture content of the boundary layer directly impacts the convective instability of the environment.

Slide 7: Let's look at a few examples of real soundings. In each, the rightmost thick black line is the temperature, and a thinner black line represents the moist adiabat that a surface-based parcel would follow if risen adiabatically. The first is a sounding from Koror, the capital of the small West Pacific island nation of Palau.

Slide 8: This sounding is near-saturated; we can tell this because the dewpoint closely parallels the temperature. In fact, the total precipitable water was an extremely juicy 67.9 mm! The sounding mostly follows a moist adiabat but is conditionally unstable between roughly 750 and 600 mb, so the CAPE was about 1800 J/kg. Although there is no set rule of thumb, CAPE over 2000 J/kg over tropical oceans is plenty sufficient to support deep convection—provided that sufficient moisture is present. This is in contrast with CAPE values over continental regions that can sometimes exceed 4000 J/kg during severe weather outbreaks. Such CAPE values are extremely rare over tropical oceans. In this case, the CAPE is distributed through a deep layer in the troposphere, meaning that it will support updraft velocities much less than what you might see in continental convection. Typically speaking, marine convection is less intense than continental convection, the latter of which produces the majority of lightning globally. The sounding is also saturated close to the 0°C level. If I saw this sounding with no context, I would assume it was launched into a stratiform precipitation region in a mesoscale convective system. Depending on the situation, I might expect subsequent soundings to be more stable.

Slide 9: This sounding is also from the tropical West Pacific, but it is much drier.

Slide 10: Note that the TPW is much lower than before. 43 mm is a low value for TPW. The CAPE was estimated to be about 1300 J/kg. Thus, the environment is buoyant enough to support vertical growth of an undilute parcel. However, this sounding indicates several dry layers. Two dry layers are seen in the lower troposphere, consistent with the low TPW. These dry layers prevent widespread growth of convection because updrafts entrain the air and become less buoyant. You will also note that the upper troposphere is also very dry. This is more indicative of the fact that deep convection is not present than it is of an environment unfavorable for convection. Cold upper tropospheric air can hold only limited water vapor before reaching saturation, meaning that the difference between water vapor in cloudy vs environmental air in the upper troposphere is small.

Slide 11: Here is an example of a coastal sounding in South Asia.

Slide 12: This is also a moist sounding as seen by the TPW of over 64 mm. The temperature profile in the boundary layer points to an unstable sounding. Usually, environmental lapse rates are much less than the dry adiabatic lapse rate over well-mixed marine boundary layers whose temperatures are controlled more by the underlying SST. Over land, boundary layer temperature is often less constrained and more impacted by insolation. When this sounding was made, convection had not yet occurred near the station, however, a warm, moist boundary layer and steep lapse rate below 850 mb certainly support deep convection. Note the CAPE of over 3200 J/kg.

Slide 13: Here is shown another tropical sounding, taken from the tropical East Atlantic.

Slide 14: The previous soundings were over regions in the Indo-Pacific tropical warm pool; however, this sounding is very different. It was taken over an island in a location where SSTs are typically cool; note the surface temperature of a little under 20°C. At about 850 mb, the temperature increases sharply, and the dewpoint drops off sharply. This is indicative of the bottom of a strong, deep layer of subsidence. Temperature increasing with height is called an inversion. Inversions indicate stability. They are often accompanied by dry layers that are indicative of descent. This sounding is not supportive of deep convection. CAPE is basically zero, and the TPW is restricted by the inability for deep convection to form and transport water upward.

Slide 15: The Amazonian rain forest can behave like a “green ocean” in terms of surface fluxes.

Slide 16: This sounding features a moist PBL and above 1000 J/kg CAPE, but it contains a dry layer centered near 700 mb that will inhibit convection. The stable, dry layer above 500 mb indicates subsidence. It’s possible that this thermodynamic profile could support some shallow and moderately deep convection rooted in the boundary layer, but deep convection will struggle to survive for long periods of time.

Slide 17: Finally, we will end with a word of caution about using just CAPE to assess the likelihood of strong convection. We’ve already mentioned that the CAPE needed over tropical oceans to support deep, widespread rainfall is much less than that often seen over continental regions during outbreaks of deep convection, such as the over the central US in springtime. We will next compare a couple of marine tropical soundings with continental soundings from Oklahoma. Here, we are looking at a sounding from Palau with a temperature profile that is nearly moist adiabatic. The CAPE, like before, is distributed through the troposphere in a deep layer. Total CAPE is about 2500 J/kg; however, the lower free troposphere is not particularly moist between 900 and 600 mb, so the TPW is only about 53 mm.

Slide 18: Next, look at a springtime sounding from Norman, Oklahoma. The CAPE is virtually the same as the Palau sounding. However, the lapse rate in the middle troposphere is steeper, and the CAPE is concentrated more in the middle to upper troposphere. In this example, more forcing is required so that convection can reach its LFC; however, convection that does get lifted to about 750 mb can accelerate upward more quickly than a hypothetical updraft in Palau that relatively gently rises from the boundary layer.

Slides 19 and 20: The last two slides illustrate a similar comparison: The Norman sounding actually has less CAPE than the Palau sounding. But undiluted parcels at Norman would be more buoyant than at Palau if they could penetrate the stable layer just above 700 mb. The main message here is that the vertical profile of temperature and moisture helps to govern the CAPE, and just consulting the value of CAPE will not necessarily tell you if rain should be expected. Second, the more slowly rising marine parcels are more susceptible to entrainment, so sufficient CAPE could be insufficient to promote deep convection if the atmosphere is too dry, such as in the example of the previous slide. Various researchers have shown little correlation between CAPE and tropical rainfall; although, some have found that the two may be weakly lag-correlated, with CAPE peaking before rainfall.

Module 1.6:

Slide 1: In this module, we will discuss the three modes of cumuliform convection in the tropics, separated by their vertical extent into the troposphere.

Slide 2: There are three modes of cumuliform convection observed in the tropics. The plot on the left depicts distributions of the number of convective cells seen as a function of their height. First, there are shallow cumulus clouds. They are generally 0–3 km in depth and are numerous. Shallow convection is pretty much ubiquitous through the tropical oceans except in regions that are very dry. Cumulus congestus are taller, more laterally expansive clouds that often consist of multiple updrafts that are part of the same cloud complex. They are non-precipitating clouds, with a modal height around 5 km, which is approximately the 0°C level in the tropics. Some non-precipitating clouds extend higher into the atmosphere, and some cumulonimbus clouds are the same height as a typical congestus cloud. Deep cumulonimbus can extend up to the tropopause. On the right are the heights of the tops of 20 dBZ radar echoes in convection over the Indian Ocean. This does not correspond to cloud top height, but it does show a bimodal distribution: One mode peaks at 2–3 km, and a second mode peaks at 5–6 km. These correspond to the congestus and deep cumulonimbus modes. Because the sensitivity of the TRMM radar was about 17 dBZ, neither shallow convection nor the tops of deeper convection could be seen.

Slide 3: Shallow convection, like that shown in the picture, is common in the tropics, and especially where sea surface temperatures are warm. Shallow moist convection can easily be triggered by turbulent motions in a warm, moist boundary layer and distributes moist static energy vertically through the boundary layer and exchanges more efficiently with the lower free troposphere than entrainment processes at the top of the boundary layer can alone. In locations where trade winds are present, these are sometimes called “trade cumuli” and are topped by a strong inversion through which the clouds cannot easily penetrate.

Slide 4: A minimum in tropical cloudiness tends to occur between about 600 and 800 mb. The figure shown currently illustrates the frequency of cloudiness as a function of height estimated from sounding data—in which a cloud was assumed if relative humidity in the sounding exceeded 93%. Based on this metric, clouds are present 4–6% of the time below 800 mb, and more frequently above 600 mb, but only about 2% of the time between 600–800 mb. Over long periods of time, these percentages also reflect the typical long-term areal coverage of cloudiness at these levels. The clouds below 800 mb represent commonly occurring shallow convection, and increased cloudiness above 600 mb probably corresponds to lateral expansion of clouds as altocumulus or altostratus in locations where moisture has recently been detrained into the environment by cumuliform convection. Clearly, something seems to prevent many shallow convective elements from getting higher than 800 mb.

Slide 5: This appears to happen because a climatological layer of negative environmental buoyancy is present above the boundary layer, at least over tropical oceans. These figures were calculated from two different numerical cloud-permitting simulations of tropical convection over the Indian Ocean. Blue regions indicate where the mean buoyancy in the model domain was negative. The simplified expression used to compute buoyancy is shown at bottom and is sensitive to the difference in temperature, pressure, and mixing ratio between the environment and clouds. In general, the temperatures of updrafts in this 700–850 mb layer are actually cooler than the surrounding environment by a small amount—less than 1K. This causes updrafts, on average to decelerate. Only the most intense updrafts can penetrate this layer to become moderately deep, congestus-like convection, and so a mode of convection that exists only within the boundary layer and part of the negatively buoyant layer exists.

Slide 6: Congestus convection, as shown here, tends to be more laterally expansive and deeper than the shallow boundary layer convection we just discussed. This type of convection transports moisture to higher altitudes, and it is also very sensitive to the vertical profile of moisture in the lower free troposphere. Its presence over several days can progressively moisten the environment, which reduces the dilution of moisture in convective updrafts and reduces the negative buoyancy imparted upon updrafts through the process of the entrainment. The modal height of such clouds is at about the level of the 0°C level. This is because locally, freezing releases latent heat and stabilizes the atmosphere near the 0°C level. Updrafts that are only marginally strong enough to penetrate the lower free troposphere given their initial velocity out of the boundary layer combined with entrainment of sub-saturated air are not able to penetrate this additional stable layer. Such clouds often diverge moisture into the environment near this stable layer, and thin altostratus or altocumulus cloud layers are often present as a result.

Slide 7: Convection with updraft velocity large enough to penetrate the stable layer above the boundary layer and stable layer at the 0°C level can grow into deep convection, which frequently precipitates. This rarely occurs in dry environments because entrainment of dry air causes deceleration of updrafts. Deep convection can penetrate the tropopause and reach as high as 20 km. It is responsible for most of the rain in the tropics, although this rain generally occurs over relatively short periods of time. Deep convection also grows laterally into sometimes expansive mesoscale convective systems, which we will discuss in detail in a future lecture series.

Slide 8: The current figure illustrates satellite-derived precipitation throughout the tropics, subtropics, and mid-latitudes. Most of the surface area on Earth does not experience rainfall at a given moment in time; although, satellite imagery would reveal that shallow convection is present over much larger regions than where precipitation occurs. Most of the colors seen in the tropics correspond with the locations of deep convection. It is in the deep convection where vertical motions are upward. Therefore, the transport of energy and moisture from the boundary layer to the middle and upper troposphere occurs almost exclusively through deep convection. While shallow convection is present in many other locations, the majority of the troposphere experiences gentle descent, or subsidence.

Slide 9: This diagram illustrates some idealized clouds in white with vertical motion vectors shown in blue. The red lines indicate arbitrary pressure or height levels. Most of the atmosphere is cloud-free; if this isn't already intuitive to you, it will be if you pay attention to cloud populations wherever you are for long periods of time. Outside of the clouds, air slowly descends.

Slide 10: Divergence occurs at the top of convection causing lateral distribution of moisture and energy, and convergence occurs near the base of the convection in the boundary layer and perhaps also in the low free troposphere. The lengths of the green arrows near cloud base in this diagram are intentionally shorter than those at the tops of clouds to indicate that the mass flux into cloud base should be the same as mass flux out of the cloud at top approximately. Remember, the density of air decreases exponentially with height.

Slide 11: In the regions of descent, the atmosphere warms adiabatically and it becomes slowly drier through vertical advection. As an atmospheric column dries, it becomes more transparent to terrestrial radiation emitted from below. Emission of longwave radiation to space causes cooling of the atmosphere. Drying the atmosphere increases the radiative cooling that happens in the troposphere because fewer absorption interactions occur. Inside the clouds, rising motion causes adiabatic cooling; however, the release of latent heat caused by phase changes of water balances the cooling. This balance between diabatic and adiabatic heating and cooling is known as radiative-convective equilibrium. On long time scales, we can say that the tropical atmosphere is in radiative-convective equilibrium. However, the atmosphere is not in RCE at all places all the time; it is deviations from this equilibrium state that drive moist convection.

Slide 12: Alterations to the large-scale subsidence can make the environment more or less favorable for convection by altering static stability. Consider the two plots shown here. On the left is a hypothetical change in large-scale vertical motion associated with some disturbance. It indicates that the disturbance causes a decrease in downward motion in the clear-air environment and that the biggest decrease occurs around 500 mb. The right-hand panel shows the changes in adiabatic heating rate, in blue, and radiative heating rate, in red caused by the change in vertical motion. Reducing the subsidence moistens the atmosphere, therefore, radiative heating increases. At the same time adiabatic heating decreases. The sum of the two causes the lapse rate in the lower troposphere to increase, making the environment more conducive for convection to penetrate the stable layer present above the boundary layer. As convection develops, large-scale mean vertical motion may become upward, but this only happens because convection is already occurring, stabilizing the environment back toward an equilibrium state. In other words, upward motion that occurs in datasets should only be used to indicate that convection is happening, not that the environment is becoming conducive for convection. The simultaneous discussion of large-scale and cloud-scale processes that we have had thus far has been convoluted, so in the next module we will discuss the difference and show a method for quantifying convective and radiative processes at various scales in observational or modeling datasets that divide the atmosphere into discrete spatial intervals.

Module 1.7

Slide 1: This module covers large-scale tropical heating and moisture budgets and the treatment of cloud-scale processes in gridded observational datasets and numerical models.

Slide 2: Suppose that we are interested in quantifying the adiabatic and diabatic heating or cooling that occurs over a large area, such as that pictured here. Processes that occur on the spatial scale of this entire image are slowly varying relative to the processes occurring within the millions of clouds present within the larger region. Thus, there is some sort of background state that the atmosphere is in, which is the large-scale state, and superimposed onto that is high amplitude variability that occurs at the mesoscale, on the order of 100 km or more, or at the cloud-scale on the order of 100 meters or more, and of course motions and processes at even smaller spatial scales that we won't spend much time discussing in this class.

Slide 3: In a numerical model or some sort of gridded observational dataset, the region would be divided into numerous discrete three-dimensional volumes. A hypothetical horizontal cross-section through one layer of such volumes is now pictured overtop of the map. Each square shown might represent one grid point in a model, for which the output of temperature, humidity, vertical motion, heating, or numerous other variables are expressed as a mean value within the box. However, even within each box, smaller-scale processes are prevalent. Suppose that each white box has a mean vertical motion vector that is upward, and all the rest of the boxes experience mean downward motion. This means that active convection is definitely present in the white boxes.

Slide 4: Furthermore, let's zoom in on one of the white boxes, now highlighted as red. The inset red box to the right is a hypothetical representation of what the cloud population really looks like. In this case, there are several convective elements.

Slide 5: Upward motion is present in the clouds, and downward motion is present outside of the clouds. Integrated over the area of the entire box, the mean vertical motion is upward because the strong vertical motions in the convection are both strong enough and expand over a large enough area to cause the local mass flux to be upward in the net even though the majority of the area within the box experiences subsidence. The vertical motion, or any other field, can be divided into a volume mean plus a perturbation. It is the volume mean that is reported as the value for each of the white boxes at left, but in reality, each volume is populated with numerous spatially varying perturbations from that mean.

Slide 6: We'll come back to those images shortly, but let's first define a couple of new terms. First, consider the budgets for the tendencies of the mean dry static energy and specific humidity within a grid box. The time tendency of dry static energy averaged within the box is related to flux convergence of dry static energy plus the sum of radiative and convective processes that occur within the box. As in the previous slide, the overbar represents a mean within a grid box. The time tendency of specific humidity is the moisture flux convergence plus mean evaporation, which adds to water vapor concentration, minus condensation rate, which quantifies removal of water vapor. The subscript $R$ denotes radiative heating, and the subscript $C$ denotes convective heating, which is related to the difference between the condensation and evaporation rate multiplied by latent heat of vaporization. Phase changes between liquid water and ice and between vapor and ice are neglected here. These equations state that condensation will cause warming and should be denoted as a heat source and moisture sink. Evaporation cools and is a moisture source. The third equation is simply mass continuity. We can define two terms, $Q_1$—the apparent heat source—and $Q_2$, the apparent moisture sink. $Q_1$ is the sum of the mean radiative and convective heating within the box plus the vertical flux convergence of eddy dry static energy within clouds. $Q_2$, shown here as an energy sink such that a positive value means vapor pressure decreases, is the sum of mean convective heating in the box and the vertical flux convergence of eddy moist static energy. Dividing the current formulation of $Q_2$ by $L_v$ would give us a rate of depletion of moisture instead.

Slide 7: Again, the eddy dry or moist static energy vertical flux convergence is essentially the vertical transport of energy within clouds.

Slide 8: Notice the separation of the variables underneath the overbar when defining $Q_1$ and $Q_2$. This allows us to separate out the eddy terms on the right-hand side.

Slide 9: We can further manipulate these expressions for heating. If we subtract $Q_2$ from $Q_1$, then move $Q_R$ to the other side of the equation, we get $Q_1$–$Q_2$–$Q_R$ equals the vertical flux divergence of eddy moist static energy. This equation basically says that the imbalance between convective heating and radiative cooling averaged inside some volume can be attributed to the vertical transport of MSE in clouds within the box that act to stabilize the environment.

Slide 10: If we look back at our hypothetical discretization of a sample environment, then the red terms in the equation at bottom refer to the grid box mean, and the white term represents processes occurring within the cloud.

xxxSlide 11: The same is true for the vertical transport of dry static energy, or moisture, which combined make up the moist static energy. In numerical models that are not run at high enough spatial resolution to explicitly resolve cloud-scale processes, these processes are often represented by cumulus parameterizations, which make various assumptions about the sensitivity of convection to thermodynamic properties of the atmosphere. This is common in global models of the atmosphere or models integrated over long periods of time such as climate models. Sometimes the cumulus parameterization is replaced with what is called super-parameterization, a more computationally expensive process in which the eddy terms are approximated by running a two-dimensional cloud-resolving model within each three-dimensional model grid box.

Slide 12: If we integrate mass-weighted $Q_1$–$Q_2$–$Q_R$ through the troposphere, we should find that it equals the total vertical flux of heat in a grid box. At the surface, the total energy flux from the surface to atmosphere is the sensible heat flux plus the latent heat flux.

Slide 13: We can also integrate $Q_1$ from the surface to the top of the troposphere to get column-integrated heat source. The total column-integrated heating, which is the total heating summed over a column of stacked boxes, is constrained by the sensible heat introduced to the column plus the total latent heat release caused by condensation. In a complete treatment, mean net freezing and deposition would also be included.

Slide 14: Likewise, we can integrate $Q_2$ through the column. As you might intuitively expect, the total moisture sink in the model is the sum of how much moisture is rained out, which is a sink because it represents condensed vapor that is not re-evaporated, minus the evaporation of water from the surface, which represents a source.

Slide 15: Put together, all the equations show how surface fluxes constrain precipitation and large-scale heating in the tropics. Although we won't do so here, a formulation for plume buoyancy can be expressed such that the tendency of large-scale buoyancy is directly related to the sum of boundary layer radiative heating and surface heat fluxes.

Slide 16: This figure shows what vertical profiles of $Q_1$, $Q_2$ and $Q_R$ look like over the Indo-Pacific warm pool where SSTs are high. The heat source and moisture sink are dominated by deep convection. Peak heating is usually above the 0°C level, which is denoted near 5 km in this figure. Here, peak heating is located near 7 km altitude. The magnitude of the moisture sink, converted to the same units is usually a little less in the upper troposphere. Radiative cooling is present throughout most of the troposphere in a cloud-free environment; however, clouds can produce much more locally complicated profiles of heating and cooling. If you compute $Q_1–Q_2–Q_R$ with these profiles, you will get a positive value in the upper troposphere, and a near zero to slightly negative value in the lower troposphere above the boundary layer. Thus, the largest vertical flux divergence of MSE occurs in the upper troposphere in deep convection.

Slide 17: Budgets like these are often computed from networks of closely located rawinsonde observations, such as in the network seen here from the DYNAMO field campaign in 2011. Each dot represents a different sounding site, and a gridded analysis of wind, temperature, humidity, vertical motion, and heating was created in 1 degree by 1 degree boxes within the quadrilaterals outlined between 70°E and 80°E.

Slide 18: These are just different examples of vertical profiles of several different quantities that can be derived from such datasets. Don't concern yourself about what the different colors mean. However, do notice that the peaks in $Q_1$ tend to correspond with peaks in vertical motion. This is expected based on thermodynamic equation. The other thing to notice is that sometimes $Q_2$ is negative. This means that clouds, which usually act as a moisture sink by condensing vapor and raining it out, act as a moisture source. This can happen when a substantial amount of condensate, rather than falling as rain, re-evaporates at higher altitudes and detrains into the environment. Such an example is seen by the black line in the $Q_2$ profiles shown here. This is related to moistening of the environment by congestus clouds, as was mentioned previously, and it will become an important consideration when discussing the transition from shallow to deep convection later in the course.