Module 3.1

Slide 1: This short video briefly goes over some resources that you can consult to prepare your tropical cyclone forecasts during lab periods. We'll look at a few different types of models, performance of models and some online resources you might find useful when forecasting.

Slide 2: Numerical models are either dynamical, meaning that they integrate physical equations in three dimensions through time to simulate a future state of the atmosphere, or statistical, meaning that they make empirically based predictions based on past outcomes or persistence. Statistical models are no longer used for track forecasts because dynamical models have superior performance; however, they are one important resource for intensity prediction. The table displayed currently shows several dynamical models used by the National Hurricane Center. Many of the models are global models, meaning that they model the atmosphere over the entire world. Horizontal grid spacing and the number of vertical levels determine the size of a discrete grid box in a model domain. The data assimilation system and cumulus parameterization are also listed. Most models have new output available every 6 hours, although output from the higher resolution model from ECMWF is only available every 12 hours, and only limited output fields are available to the public. Two non-global models are shown; they are HWRF and COAMPS. They are mesoscale models that run at high resolution and use global model fields as forcing.

Slide 3: Two types of ensembles are listed on this slide. Some are consensus ensembles, meaning that they take the average or weighted average track or intensity of several dynamic models to create a refined prediction. Global model ensembles are made of several runs of the same model initialized with different perturbations to the atmospheric state. Their purpose is to provide a sense of the uncertainty in a model track or intensity forecast that is based on the uncertainty in the imperfect time zero model analysis. This reduces forecasters' dependence on a single deterministic model. They capture a range of potential solutions of the future state of the atmosphere. This range is also referred to as model spread. In lab, we will be able to use global model ensembles output from GEFS—the GFS ensemble system, MOGREPS—the UK Met Office model ensemble system, and the Ensemble Prediction System in ECMWF. The U.S. Navy's mesoscale model, COAMPS, also produces an ensemble of solutions not shown here, but it is typically overspread, meaning that it expresses too much uncertainty.

Slide 4: Statistical models are no longer used for track forecasting; however, one, CLIPER5, which is based on climatology and persistence, is useful as a baseline for assessing the performance of dynamical models. Some other statistical models are used to assess wind radii and intensity. The errors of statistical models at intensity forecasting are still, even in 2020, comparable to errors of complicated dynamical models—even the ones like HWRF that are run on small scales. You may find yourself using SHIPS or DSHIPS in your forecasts in this class. DSHIPS is essentially SHIPS but includes a decay component for tropical cyclones moving over land. These models produce output every 6 hours.

Slide 5: The next few slides characterize track and intensity errors for North Atlantic and East Pacific tropical cyclones and compare them to official forecast errors from the National Hurricane Center. Track errors here are shown in nautical miles. Bolded values indicate models that performed better than NHC at the time represented by that column. Ensemble systems like the FSSE, TVCA, and HCCA often perform better than individual dynamical models. Of the dynamical models listed here, only HWRF and the two Navy models, NAVGEM and COAMPS, has larger than 190 nm error at 120-hour lead time.

Slide 6: The current figure shows the track forecast skill of several models relative to climatology. Essentially, this sort of says how much better a forecast model does than just using climatology as your best guess. Dashed lines represent consensus ensembles. On average, the European Center model is the best performing track forecast model and the Navy model is, while better than climatology, substantially worse than other dynamical models.

Slide 7: Track forecast errors have generally decreased over the past few years, especially at 4 to 5-day lead times. However, intensity forecast errors have not markedly improved. As seen in this chart, average intensity errors of dynamical and statistical models are comparable to one another.

Slide 8: Compared to climatology, the European Center model is actually *less* skillful at many lead times, while consensus models perform the best. Still average error of the best consensus intensity forecasts at 5 days was 20 kts.

Slide 9: Similar themes are seen in model performance over the East Pacific. The European Center model and consensus ensembles performed quite well in 2019, while the Navy global model in particular performed quite poorly.

Slide 10: Consensus models have performed well in the East Pacific over the last three years. The CMCI, represented by the yellow line, is the Canadian global model, and it also doesn't perform particularly well on average compared to other global models.

Slide 11: For some reason, NHC was outperformed at intensity forecasting in 2019 by many models, especially at long lead times. In fact, the best performing model at 5-day lead time for intensity was the DSHIPS, a statistical model.

Slide 12: In general, errors over the past 3 years have been similar to those over the Atlantic and a little less skillful relative to climatology percentage wise; however, the absolute error has been less over the Pacific.

Slide 13: We will discuss several resources in class for information you can use to help forecast. The first day in the lab will be challenging, but as the quarter progresses, you'll become more comfortable with the tools at your disposal. We'll use a combination of in situ aircraft data, satellite-based intensity estimates, model forecasts, remote sensing data, and more to make forecasts. These are some links to resources that you might find useful in real-time. Have a look through them and see what you can figure out with a little research. We'll collectively explore many of these resources during our forecasting lab periods.

<u>Module 3.2</u>

Slide 1: This module contains a description of mature tropical cyclone structure.

Slide 2: Tropical cyclones are among the most impactful weather phenomena occurring in the tropics. They are responsible annually for billions of dollars of damage globally and the loss of numerous human lives. Shown here is the spatial distribution of TCs throughout the world. TCs occur in the North Atlantic, East and Central Pacific north of the equator, West Pacific, South Indian Ocean, and in the Bay of Bengal and Arabian Sea. Most develop at low latitudes and eventually move poleward. Occasionally a tropical cyclone forms in the mid-latitudes, and very rarely, they occur in an entirely different basin, such as in the South Atlantic. One way of categorizing TCs by strength is using the Saffir-Simpson scale, which is based on maximum wind speed of a TC. Orange, red, and pink colors denote the tracks of the strongest TCs when they were most intense. Very strong TCs can happen in any basin where TCs occur, but they are most common over the warm waters of the West Pacific.

Slide 3: This figure is similar to the last, but it only shows genesis points and tracks for TCs in the 1970s and 1980s. The yellow shaded region represents the climatological contour of 26.5°C SST during the same period. While most tropical cyclones formed over the warm SSTs, a few did not, particularly over the North Atlantic.

Slide 4: Several requirements must be met for tropical cyclogenesis to occur. Many of these are the same as the general requirements for tropical convection to develop and persist. First, sufficient fluxes of energy from the ocean to the atmosphere must be possible. This requires a deep mixed layer of warm ocean temperatures—nominally 26°C or greater—although a few TCs have developed over cooler SSTs. Ample moisture in the location of the initial vortex must be present. This is because amplification of the TC vortex requires release of latent heat, which can only happen if convection grows deep. Dry air inhibits vertical growth of convection as discussed in previous modules. The atmosphere must also be sufficiently unstable to support convective growth. Some additional criteria are required for TC genesis. First, ample relative vorticity must be present in the environment. Second, vertical wind shear must be weak, nominally less than about 15 kts. Finally, the vortex must generally be about 5 degrees of latitude away from the equator, where the Coriolis parameter is not essentially zero.

Slide 5: Well-organized developing and mature tropical cyclones usually have a banded structure like that shown here. In the center of mature TCs is, of course, the eye, where subsiding air prevents formation of deep convection. The eye is surrounded by an intense ring or partial ring of convection known as the eyewall. Sometimes, two eyewalls are present. In such cases, the outer eyewall is also known as the secondary eyewall, and between the two eyewalls is the “moat”, a region of local subsidence where deep convection does not form. Rainbands then spiral outward from the center, sometimes several hundred kilometers away from the eye. The principal rainband is the largest, is directly connected to the core, and is sometimes contiguous with a secondary eyewall. The bands consist of convective precipitation, denoted by the dark shading, and weaker stratiform-like precipitation elsewhere in the rainbands.

Slide 6: From satellite imagery, the detailed structure of the inner core and rainbands is usually difficult to discern. This is because the entire storm is covered by cirrus cloud that expands laterally from the center. However, passive microwave imagery can detect some of the precipitation features, which is particularly useful over open ocean. Near land or from a TC-penetrating aircraft, radar can discern the detailed structure of TCs. Shown now is radar imagery of Hurricane Michael, just prior to its landfall along the Florida panhandle in October 2018. At this point, Michael was a mature and intense Category 5 hurricane. An eye is obvious. It is surrounded by a partial ring of intense convection, with the strongest rain appearing to the northeast of the eye. A primary rainband is visible. Several additional rainbands far from the eye are also present and resemble the schematic displayed on the previous slide.

Slide 7: A little while later, Michael made landfall, as seen in this image. The eye is completely surrounded by an eyewall of intense convection, and a disconnected secondary eyewall can be seen as well. Rain between the primary and secondary eyewalls was lighter, as denoted by the green colors. The primary rainband extends spirally outward from the secondary eyewall. This detail would be impossible to determine by simply looking at visible or infrared imagery.

Slide 8: The inner eyewall of a mature hurricane is shown here from the inside of Hurricane Katrina in 2005. The photo was taken from an aircraft that penetrated the eyewall of Katrina and sampled temperature, humidity, wind speed and direction, and pressure as well as made visual observations inside and surrounding the eye. Deep convection is absent in the eye; only shallow cumulus or stratocumulus is present. The eyewall cloud consists of deep, intense cumulonimbus clouds that slope upward and outward away from the eye. The shadow of the sun on one part of the eyewall can be seen in the right half of the image.

Slide 9: Photos like these are captured from surveillance aircraft that fly into developing and mature tropical cyclones that threaten land. The US Air Force Reconnaissance Squadron flies C-130 Hercules aircraft into tropical cyclones to collect radar and in situ observations. They are the primary reconnaissance aircraft that collect real-time observations used for characterizing tropical cyclones.

Slide 10: NOAA also sometimes flies P-3 aircraft, Miss Piggy and Kermit, into tropical cyclones. The P-3 usually flies higher than the C-130, which typically flies in the lowest 10,000 feet. The P-3 also contains three weather radars that can be used for operational characterization of a storm, or later, for research purposes.

Slide 11: A cross-section cut-out of an eye and eyewall is seen here, showing one quadrant of an idealized storm. The kinematic flow of the secondary circulation of the TC is shown with different arrows. In a future module and in homework, we will further explore the dynamics behind some of these motions. In a TC, the primary flow refers to the tangential flow around the cyclone’s main vortex. The secondary flow refers to the radial and vertical motion that accompanies the faster primary flow. Arrows denoting radial and tangential flow are shown in blue. The X denotes flow into the screen. The primary energy source to the TC core is frictional radial inflow in the boundary layer. Convergence near the eyewall drives upward motion in the eyewall cloud, which rises following a slanted surface of constant angular momentum while releasing latent heat as water vapor condenses. The latent heat release in the eyewall supports a contraction of the eyewall and radius of maximum wind, or RMW. Immediately along the inner edge of the sloped eyewall, saturated descent occurs. In the eyewall, forced descent occurs along a dry adiabat, forcing warming and stabilizing the atmosphere in the eye. As a result, deep convection is absent in the eye, although a shallow deck of stratiform cloud is present in the boundary layer. Light radially outward flow from the eye to the eyewall is present in the boundary layer.

Slide 12: The mean radial wind and primary circulation are shown here. Although the left panel is derived from observations of Atlantic hurricanes and the right panel from Pacific typhoons, the fundamental circulation structures are similar. Pressure is on the y-axis, and radius is on the x-axis. For radial wind, negative values represent flow inward toward the center. This occurs in the boundary layer, with a maximum magnitude around 950 mb. Radial outflow is seen in the upper troposphere, while the radial flow between the boundary layer and 300 mb is generally weak close the core. The tangential wind, or flow around the storm, is approximately in gradient wind balance to first order and becomes larger as one gets close to the center, and the altitude of the wind maximum increases moving inward. Positive values in the right panel represent the magnitude of cyclonic flow around the vortex. An upper-level anticyclone is common, with its maximum flow located far from the core of the cyclone. Angular momentum can be defined as shown. Following fluid motion, it is conserved, an important constraint on kinematics of the flow.

Slide 13: Isotherms of theta-e in a tropical cyclone show a warm core center—meaning that theta-e in the center exceeds theta-e immediately outside the center at the same pressure level. The radial gradient of theta-e in the inner core represents the magnitude of the warm core. The warm core is a result of dry adiabatic compression of subsiding air in the eye, and so the strength of the warm core decreases with height in upper levels of the storm. Above the 0°C level, isentropes slope out radially with height. The subsiding, warming air, causes a minimum in geopotential at the center of the storm, which helps to enhance the pressure gradient across the eyewall cloud. It is this pressure gradient that primarily controls the strength of the primary circulation.

Slide 14: Presence of a warm core is required for classification as a tropical cyclone. Cyclone phase space diagrams, like that shown, can help one diagnose quickly if a cyclone is fully tropical. The phase space diagram classifies a vortex based on its symmetry, and the strength of the warm core. This diagram shows an example of the evolution of Hurricane Sandy in the cyclone phase space, starting at A and progressing toward Z. The storm developed as a symmetric warm-core cyclone over the Caribbean before losing its symmetry (partly due to shear) as it moved northward. As the cyclone approached the mid-Atlantic coastline it transitioned into a cold-core, non-tropical cyclone.

Slide 15: The y-axis defines the thermal asymmetry near the cyclone center, while the strength of the warm core is defined here by the negative magnitude of the thermal wind vector in the lower troposphere. Consider the definition of thermal wind, shown at bottom. Recall from your dynamics class that the thermal wind describes a shear vector; in this case, it is the change in the magnitude of the geostrophic wind between 900 and 600 mb. The warm core cyclone—or having a temperature that decreases moving radially from the cyclone—imposes the condition that the geostrophic wind should decrease with height. In contrast, the primary circulation will increase in magnitude with height in a cold core cyclone.

Slide 16: This figure shows another cross section through a mature TC, with theta-e contours displayed on the left. The theta-e contours are parallel to contours of constant angular momentum, meaning that ascent in the eyewall is approximately isentropic. The right panel also shows temperature contours as dashed lines, and tangential wind contours as solid lines. Temperature lines also show weakening of the warm core with height. Note also how the radius from the center at which the maximum wind occurs is farther from the center high in the storm than in the low-troposphere. Furthermore, as seen in a previous slide, the maximum tangential wind occurs somewhere above 900 mb.

Slide 17: In a homework problem, you'll argue for why the eyewall must be sloped as pictured here.

Slide 18: Finally, the frameworks that we have viewed so far are for idealized axisymmetric tropical cyclones. However, in nature, tropical cyclones can rarely even be approximated as axisymmetric because vertical wind shear impacts the distribution of latent heating in the tropical cyclone. Consequently, the secondary circulation is often largely asymmetric, being stronger in one part of the cyclone than in another. A tropical cyclone can be broken up into four shear-relevant quadrants to describe this asymmetric circulation. In the eyewall, updrafts tend to first develop in the downshear half of the storm. They then rotate around the eye while precipitating. By the time the convection rotates around to the upshear side of the storm, convection has mostly dissipated and downrafts are present. Therefore, an eyewall often appears to be asymmetric, appearing sometimes as only part of a ring located at least in the downshear left part of the storm. The secondary circulation is strongest, therefore, in the downshear right quadrant, and diverging flow into the eye at upper levels is more common within the shear-relative left side of the storm. Motions tend to be comparably weak in the upshear right quadrant.

Slide 19: The behavior described is an average across many storms, and various storms have differences in their exact kinematic structure; however, the resulting rainfall pattern is often observed, as seen here again in the case of Hurricane Michael. The strongest radar echo in the inner eyewall is located to the left of the shear vector, with the downshear right being the quadrant with the weakest echo.

Module 3.3

Slide 1: This short module will describe and show examples of concentric eyewalls and replacement cycles that occur in intense tropical cyclones.

Slide 2: An example of part of an eyewall replacement cycle, also known as a concentric eyewall cycle, is shown here. In the left panel, we see high frequency passive microwave imagery of Hurricane Frances from 2004. Two eyewalls are apparent and are labeled. At this point in time, the inner, or primary, eyewall is weak but consists of a ring around a small eye. The weak ring is surrounded by a moat of warmer brightness temperatures. Further from the center, a solid ring of deep convection is present. It is the outer, or secondary, eyewall, and at early stages of its formation, it is often contiguous with the primary rainband. For comparison on the right is a mature tropical cyclone, Hurricane Katrina in 2005, that was not—at the time—undergoing an eyewall replacement cycle. A single ring of intense convection surrounded the eye, and a primary rainband extended radially outward from the inner core of the hurricane. Eyewall replacement cycles are generally easy to detect using high frequency passive microwave or radar data that has spatial resolution that is high enough to discern the structure of convection beneath the high cloud observed in IR or visible imagery. However, sometimes, replacement cycles become apparent on visible and IR imagery as well, especially if a large moat with reduced deep convection forms such as in the case here with Frances.

Slide 3: The reason or reasons that secondary eyewalls develop in the first place are not fully understood. A number of theories have been postulated, some providing a general mechanism for concentric eyewall formation, and others hypothesizing mechanisms for secondary eyewall formation in specific cases. We will not work through any of the incomplete list of mechanisms shown here in any detail during this course but will instead focus on developing a basic understanding of TC kinematics first. The rest of this module will show a few examples of eyewall replacement cycles.

Slide 4: Eyewall replacement cycles can be seen from radar as well; however, this is only observable continuously if the cyclone is located near land or if an aircraft equipped with radar flies through the storm. This example shows the evolution of eyewall replacement cycle observed by the ELDORA X-band radar on a NOAA P-3 flying through Hurricane Rita in 2005 during the RAINEX field campaign. In the upper left image, a primary eyewall is obvious, and it was surrounded by several high-echo regions in various pieces of rainbands. The upper right radar data was collected about 23 hours later and depicts a strong outer eyewall in a ring nearly 100 km wide in diameter that surrounded a weakened remnant of an inner eyewall. The secondary region of mesoscale subsidence, the moat, separates the two eyewalls. Eventually, the inner eyewall decayed, resulting in a very large eye surrounded by a single eyewall.

Slide 5: Three examples of eyewall replacement cycles are shown here for the same typhoon over a five-day period in SSMI passive microwave data at 85 GHz. Each cycle is outlined by a different colored box. This shows several examples of secondary eyewalls becoming the primary eyewall, contracting inward, then a new secondary eyewall developing and repeating the process.

Slide 6: Displayed currently are cross sections of wind speed obtained by aircraft through a very intense Atlantic hurricane, Gilbert, in 1988. The individual panels illustrate the evolution of tangential wind speeds over a 4-day long period. In the top panel, a large 50+ km wide eye is apparent and was surrounded by wind maxima of 40–45 m/s. The wind maxima corresponded in space with the eyewall. Over the course of about two days, the eyewall had contracted. The radius of maximum wind had shrunk to less than 15 km, and the maximum wind was nearly 80 m/s in the eyewall. A hint of an outer wind maximum began to appear about 100 km away from the center. 12 hours later, a clear outer wind maximum was present at 50–75 km radius. An additional day later, now looking at the fourth panel, the inner eyewall had weakened. Only a small wind maximum of about 25 m/s was present in the remnant of the inner eyewall, and the outer eyewall had become the primary wind maximum with a radius of maximum wind at 50–75 km and maximum wind speed of 40–45 m/s. This represented a significant weakening of the storm as the pressure gradient across the inner core weakened. However, the weakening is often temporary if the cyclone does not subsequently encounter unfavorable conditions, such as increased shear, cool ocean surface, dry air, or land. A day later, the new primary eyewall had contracted inward to a radius of about 40 km, and the wind speed had increased some as well. Many cycles occur much more quickly than the one shown here.

Slide 7: A schematic of a typical eyewall replacement cycle is shown here that summarizes what we have discussed so far. At top, time is on the x-axis, and a proxy for intensity is on the y-axis. This is an approximate evolution of cyclones experiencing replacement cycles occurring within a steadily favorable environment for intensification. Some intensification may occur after the outer wind maximum is first detected. This schematic indicates that about 9 hours later, weakening of the inner wind maximum will begin and that concentric rings will appear on passive microwave or radar imagery. Weakening may occur for about 16 hours before the cyclone begins to re-intensify. About 36 hours after the outer wind maximum first appeared, the inner eyewall is completely dissipated. Note that the time frame of evolution can vary significantly from storm-to-storm and may depend on both environmental characteristics outside the cyclone and dynamics internal to the storm.

<u>Module 3.4</u>

Slide 1: This module will provide a brief overview of the extratropical transition of tropical cyclones.

Slide 2: Tropical cyclones become extratropical—another term for post-tropical—as they move poleward and encounter increased environmental baroclinicity. The chart shown lists changes that a TC encounters and undergoes as extratropical transition occurs. As a TC encounters an environment with greater baroclinicity—essentially another way of saying that horizontal temperature gradients become non-negligible—it also usually encounters strong vertical wind shear, which cause the TC to take on a strongly sheared appearance, often with the strongest convection to the north and east of the center in the Northern Hemisphere. The TC core then tilts toward colder air with height and the primary source of kinetic energy in the cyclone transitions from release of latent heat (in the barotropic tropical environment) to baroclinic processes such as what you encountered in a mid-latitude dynamics course. Eventually, the TC completely loses its warm core, completing its transition to a post-tropical cyclone as a tilted, sheared, cold core system. The transitioned cyclone may then gradually decay, or it could intensify as a non-tropical low at middle to high latitudes.

Slide 3: Extratropical transition is very common, and you may encounter numerous real-time examples. One such example is displayed here on infrared satellite imagery. The top left panel is a West Pacific typhoon in the early stage of transition. The eye and convective banding in the warm core system are still obvious, but the storm has already encountered strong southwesterly shear, which can be seen by the advance of the upper level cloud field well ahead of the low-level center. 12 hours later, in the top-right panel, the warm core is essentially gone, and the circulation center denoted by the red plus is located far from the deepest convection. Frontal banding features are present, and the location of a weak cold frontal boundary is signified by a weak band of clouds extending southward of the circulation center. Another 12 hours later, the cyclone has fully transitioned into a partially occluded cold core cyclone, retaining none of its original tropical characteristics.

Slide 4: Transitioning tropical cyclones are generally advected poleward to the east of an upstream upper-level trough. A jet streak develops downstream of the trough axis between it and a downstream ridge. The ageostrophic flux of geopotential is denoted in the top panel by the red arrow; areas of convergence of the flux are shaded in red. At this time, advection of upper-level anticyclonic potential vorticity by the upper-level divergent outflow helps to amplify the downstream ridge. The area of latent heat release—in other words—the location of the clouds where kinetic energy in the form of upward motion is generated—transitions from the core of the tropical cyclone to along the downstream flank of the jet streak, which ultimately results in the strengthening of the mid-latitude Rossby wave feature. The schematic in the top panel is supported by plots of ageostrophic geopotential flux and baroclinic conversion to kinetic energy in a composite of transitioning West Pacific tropical cyclones. The kinetic energy budget is described by the equation at top right. The first right-hand side term describes the quantity plotted at bottom left. When vertically integrated to get a plot like that at bottom-right, the third and fifth terms may be dropped.

Slide 5: The conversion of baroclinicity to kinetic energy ahead of the tropical cyclone allows “predecessor rain events” to occur. The rain events occur along a baroclinic zone and downstream of a jet streak and are supported by moist poleward moving flow ahead of the upper-level trough. An example of a predecessor rain event in the central U.S. is shown at bottom right. These rain events can be responsible for significant flooding by saturating the soil prior to the arrival of the TC or transitioning cyclone itself.

Slide 6: Interactions between tropical cyclones and the mid-latitude Rossby wave train into which the TC moves are largely dependent upon the horizontal advection of potential vorticity at upper levels by the divergent component of the geostrophic wind. A strong interaction between a transitioning TC and the mid-latitude flow can amplify not only the downstream ridge immediately ahead of the TC, but also much of the downstream wave train. Such strong interactions have ramifications on downstream weather—especially in Europe or the US West Coast. Many atmospheric river events that occur in fall are associated with amplification of mid-latitude troughing associated with the advection of PV from the old TC into the Rossby wave train by the upper-level divergent flow.

Slide 7: An example of a recent extratropical transition of a well-documented Atlantic hurricane is shown here. Hurricane Matthew moved slowly northward along the US Southeast coast in October 2016. It resulted in over 400 mm of rainfall in some regions and caused catastrophic freshwater flooding along many coastal river basins. The plot shown here illustrates the evolution of theta-e and wind vectors offshore the Carolinas as Matthew approached. The lines represent theta-e isotherms, and the shading represents Pettersen frontogenesis. Blue shading depicts where the temperature gradient was increasing in magnitude.

Slide 8: We can zoom in on eastern North Carolina as Matthew approached. Plotted is frontogenesis at 900 mb. The equation for frontogenesis is shown displayed as well. Frontogenesis was primarily supported by deformation of the flow. Over the ocean, low-level air was warm and moist. Meanwhile, over land, to the northwest of the cyclone center, flow was northerly, relatively cool, and drier. Frontogenesis occurred at the boundary between the two airmasses. Over the 15-hour period shown, the magnitude of the temperature gradient between the two airmasses increases, providing additional forcing for isentropic ascent, which ultimately resulted in heavy precipitation focused along the front.

Slide 9: Radar reflectivity during the same period is shown here. Black contours indicate the region of largest frontogenesis. The heaviest rainfall occurred where a primary rainband collided with the developing cold frontal boundary. The warm, moist air rose isentropically as it moved inland in a conditionally unstable environment.

Slide 10: The purple colors in this figure show moisture convergence in a cross section across the frontal boundary, with the ocean denoted on the right side of each panel, and a location well inland in a cooler air mass located at the left-hand side of each panel. The lines represent theta-e isotherms, which were tilted toward the cold sector with height and converged over the 12-hour long period displayed. Moisture convergence was enhanced all along this tiltled boundary beneath 500 mb, with the greatest convergence occurring at low levels just ahead of the developing surface front—at the location where the primary rainband was responsible for the largest precipitation amounts. Effectively, as the TC underwent extratropical transition, the area of strongest precipitation behaved increasingly like a typical mid-latitude cold front; however, the air on the downstream side of the front was exceptionally warm and moist, which contributed to copious rainfall totals along a swath near where the front was located as the transitioning cyclone moved northward. This case was similar to many others previous observed over the U.S. East Coast and the West Pacific.

Module 3.5

Slide 1: In this module, we will cover some fundamentals of intensification of tropical cyclones. Much debate remains in the scientific community about many of the fundamental aspects of tropical cyclogenesis and intensification, and this is reflected in the slowness at which model forecasts of TC intensification have improved over the past couple of decades. This topic itself could merit an entire course, so devoting only a single module to TC intensification barely scrapes the surface of the broad, ongoing scientific debate about the dynamics responsible for TC intensification.

Slide 2: As we have seen in a previous module, one of the governing principles for TC dynamics is that angular momentum is conserved following fluid motions. The angular momentum is related to the radius from the center of rotation, the tangential velocity of the fluid around that center, and the latitude at which the motions are occurring. In order for flow moving inward radially toward a vortex center to conserve angular momentum, denoted here as *m*, its tangential velocity must increase. This implies that for a vortex to spin up, surfaces of constant angular momentum must move inward toward a vortex center.

Slide 3: We will discuss a couple of proposed mechanisms for TC spin up and intensification, concepts upon which numerous more complicated theories have been developed. The first is known as Conditional Instability of the Second Kind, or CISK.

Slide 4: The basic secondary circulation of a TC is shown here. Generally, angular momentum is conserved along the radial inflow and for ascent in the eyewall. However, in the boundary layer, which is located beneath the dashed line in this figure, angular momentum is not conserved following the radial inflow because friction reduces the fluid velocity, causing convergence of the radial inflow and supporting vertical motion. Essentially, the notion of CISK is that a positive feedback exists between in-cloud latent heating and boundary layer convergence that indirectly supports more latent heating.

Slide 5: The red area denotes where latent heat release occurs in cloud. This could be in an eyewall-like cloud or just in less organized deep convection located near the center of a developing vortex. The latent heating maximum occurs in the middle troposphere, with the maximum horizontal gradient in heating therefore occurring at the same level.

Slide 6: Inflow is present beneath the heating maximum and the strength of the inflow is dependent upon the magnitude of the radial gradient of heating.

Slide 7: Coriolis acting on the radial inflow increases the tangential wind.

Slide 8: And the radial inflow responds by becoming stronger. Friction increases as the radial inflow increases in magnitude, thus moisture convergence increases. Vertical motions strengthen, and latent heat release is enhanced. Thus, the radial gradient in latent heating increases, forming a feedback loop. As this occurs, the tangential wind is also in approximate gradient wind balance, and as the tangential wind increases, the flow moves inward to conserve angular momentum. This corresponds with an increase in the pressure gradient in the core of the storm.

Slide 9: The steps just outlined are listed here. One of the central arguments against CISK as a viable mechanism for explaining the intensification of TCs is that the closure for the CISK problem mathematically implies that the heating rate in convection is directly related to the moisture convergence in the boundary layer. However, the presence of water at spatial scales larger than that of convection does not itself cause convection. However, CISK assumes that the increased BL water supply essentially causes convection. Furthermore, CISK does not consider the important role of surface fluxes of energy from the ocean to the atmospheric boundary layer. An adjustment to CISK, known as the cooperative intensification mechanism, prevents a runaway feedback mechanism by allowing convection to stabilize the atmosphere and by gradually reducing flux between the ocean and atmosphere as the boundary layer becomes saturated. In its original form, CISK is not considered an appropriate explanation for TC spin up and intensification; however, it does successfully capture a couple of fundamental points: Spin up occurs as angular momentum surfaces are converged above the boundary layer and the boundary layer is a source of moist static energy to convective updrafts near the vortex center.

Slide 10: An alternative mechanism considers the role of surface moisture fluxes in intensification. It is known as wind-induced surface heat exchange, or WISHE. It describes a feedback between surface heat fluxes and wind speed. Recall from earlier in the quarter how heat fluxes between the ocean and atmosphere increase with increasing wind speed. Essentially, fluxes of energy to the atmosphere result in increases in wind speed, which result in enhanced flux of energy to the atmosphere.

Slide 11: Let’s examine this in more detail. First, let’s presume that the mixing ratio in the core region of the cyclone is increased for some reason. That reason might relate to surface fluxes of energy from the ocean to the atmosphere. The increase in moisture leads to an increase in the radial gradient of humidity and, thus, an increase in the radial gradient of theta-e in the boundary layer. Convection in the core of the vortex lifts the theta-e rich air from the boundary layer, increasing the radial gradient of theta-e in the core of the storm. This means that theta-e increases in the eyewall relative to surrounding areas. By thermal wind balance, this means that the magnitude of the vertical gradient of tangential wind must increase, leading to an increased tangential wind near the top of the boundary layer. Presumably, this increased momentum is mixed down to the surface via turbulence, where it corresponds with an increase in surface wind. Fluxes are directly related to the magnitude of the surface wind, so assuming no change in the sea surface temperature, the increased wind will increase surface fluxes of moisture, thereby continuing the feedback loop.

Slide 12: Recall our definitions of surface latent and sensible heat flux. Eventually, if the sea surface temperature and boundary layer temperature equilibrate, the flux will go to zero, which is one way the WISHE feedback process can be terminated. This might happen, for example, if a tropical cyclone remains over an area for a long period of time, thereby significantly cooling the sea surface. We'll return to explaining a governor on the intensity of a TC shortly.

Slide 13: Recent numerical simulations have shown that even with dramatically reduced surface fluxes, intensification can still occur. In this figure, three plots of TC maximum tangential winds are plotted. Three numerical experiments were run. In the first—denoted by the red line—fluxes were allowed to vary with winds as previously shown. In the other two experiments, wind speeds in the bulk flux equations were reduced to caps of 10 m/s and 5 m/s for the green and blue lines respectively. However, despite the cap in surface fluxes, rapid intensification still occurred in all three simulations. The maximum intensities by the end of the simulation were dependent upon the surface fluxes; however, the process of rapid intensification did not seem to depend much on the WISHE mechanism. Therefore, while surface fluxes are important for setting the maximum intensity of a cyclone, they are not necessary of first order importance for describing the rate of intensification to that maximum.

Slide 14: Vortical hot towers have been proposed as one alternative pathway through which vorticity is coalesced within a vortex core, possibly contributing to rapid intensification. The paper discussion for this week focuses on the evolution of local vorticity maxima in a numerical simulation of a developing TC.

Slide 15: The short animation here shows a TRMM overpass of Hurricane Rita over the Gulf of Mexico in 2005. The red colors indicate deep convection, some of which is located in the inner eyewall. These are often seen in developing TCs near the center of its vortex before intensification occurs and is indicative of the release of latent heat. In strongly rotating towers, gravity waves cannot easily disperse the heating outward, and the added temperature anomaly supports intense convection. Deep convection like that seen here are sometimes also associated with lightning in TCs.

Slide 16: Radar observations of one such tower are shown here. They are derived from aircraft radar data collected in developing Hurricane Ophelia in 2005. The bottom right shows several quantities. Vorticity is the colored shading, with red colors indicating largest values. The 35 dBZ reflectivity contour is outlined by the red line. Virtual potential temperature anomalies are denoted by white contours, and the two-dimensional wind vectors are denoted by arrows. This shows an example of deep convection in which the updraft is closely co-located with a maximum in vorticity. A large temperature anomaly—4 deg C or more—develops in the convection. Vertical cross-sections through the same cell are shown at top, with reflectivity shaded in the top panel and mass transport shaded in the middle panel. These reinforce the notion that this convective element was intense and vorticity-rich. In our paper discussion for this week, we will see how neighboring hot towers can merge in the early stages of TC development, contributing to the spin up of a larger vortex.

Slide 17: Finally, we'll wrap up with a discussion of how surface fluxes and boundary layer frictional flow might govern the maximum intensity of a tropical cyclone.

Slide 18: The secondary circulation of a tropical cyclone, when idealized, can be thought of as Carnot heat engine. The Carnot engine consists of four cyclical steps, and the maximum wind speed at the surface, is related to the difference between surface and outflow temperature and the difference between near-surface enthalpy and saturation enthalpy given the sea surface temperature. Eventually, frictional dissipation of energy in the boundary layer cancels out the energy flux from the ocean to the atmosphere. Thus, boundary layer friction acts as a governor on intensity in the WISHE mechanism.

Slide 19: The Carnot cycle consists of four phases: The first, isothermal expansion, occurs along the segment AB in the diagram. Along this segment, the temperature of radially inflowing air remains nearly constant. However, the pressure decreases moving toward the center. Following the ideal gas law, this means that the volume must increase, thus the expansion. Segment BC reflects the isentropic ascent in convection and divergent flow outward away from the vortex center at upper levels. This is the isentropic, or adiabatic, expansion stage of the Carnot cycle. During this period, theta-e is approximate conserved, and air expands as it rises. The temperature cools adiabatically and reaches some outflow temperature. The thermodynamic efficiency of the "engine" and the maximum wind speed in the boundary layer is related to the difference between this outflow temperature and the inflow temperature in the boundary layer. Segments CD and DA are isothermal and isentropic compression stages; however, the assumption that isothermal compression actually occurs in the lower stratosphere from C to D has been brought into question.

Slide 20: Maps of theoretical maximum potential intensity are created using this method and generally correspond closely with sea surface temperature. In this example, the ocean surface supports a theoretical maximum potential intensity of about 180 knots. Tropical cyclones rarely reach their maximum potential intensities, however, because environmental factors prevent continued intensification of a storm. Internal factors, such as eyewall replacement cycles, also govern intensity of a cyclone on time scales of hours to days.

Slide 21: An example of how minimum pressure varies as a function of boundary layer and outflow temperatures, given a low-level relative humidity of 80%, is shown here. As you can see, the minimum pressure generally decreases as the difference between the two temperatures becomes larger. Eventually, we reach simply theoretical intensities that correspond with temperatures that are not observed.

Slide 22: We end this discussion with a caveat. Winds in the boundary layer are not in gradient wind balance because they are subject to friction. This alters how surface fluxes and wind speeds in the boundary layer interact, especially in the inner core of a tropical cyclone, where the flow is strongest. Unbalanced flow in the boundary layer must be considered when considering intensification of a tropical vortex.

Module 3.6

Slide 1: This module covers some of the attributes of easterly waves, including African easterly waves and the African easterly jet. Easterly waves are disturbances in the mean flow of low-level easterly flow in the tropics. Easterly waves are responsible for the formation of many tropical cyclones in the Atlantic and Pacific basins, and are the precursors for a large majority of Category 3 or higher hurricanes in the Atlantic.

Slide 2: This is the first time we have seen in this course a diagram like this. It is called a wavenumber-frequency power spectrum diagram. The wavenumber is inversely proportional to wavelength. Low wavenumbers correspond to long wavelengths, and the zonal wavenumber refers to wavelength in the east-west direction only. A wavenumber 1 means that a single pair of a trough and ridge in a wave can fit in the circumference of Earth. Wavenumber 6, for example, indicates, that the wavelength is short enough so that 6 pairs of ridges and troughs can fit onto the circumference of Earth. In this case, the frequency refers to the phase speed of the wave divided by the wavelength. A plot like this projects fields associated with precipitation, like outgoing longwave radiation, or OLR, onto wavenumber and frequency space, such that the dark shading indicate pairs of wavenumber and wave frequency at which rainfall more often occurs in the tropics. Positive wavenumbers on this plot indicate features that propagate eastward. Negative wavenumbers are associated with westward propagating rainfall. The Kelvin wave and MJO part of the spectrum are shown at low wavenumber and frequency. We will discuss these later in the course. In the upper left of the plot is a broad shaded region labeled “TD”, which stands generically for tropical disturbances. This increased power represents westward propagating rainfall in the tropics that has a wave frequency of about one-quarter to one-fifth of a cycle per day, which corresponds to a wave period of roughly 2–4 days. The high wavenumber corresponds to relatively small disturbances. A wavenumber -15, for example, corresponds with westward propagating disturbances separated by about 2500 km. Given a wave period of about 3 days, this corresponds with a phase speed of about 10 m/s. Many of the disturbances that project onto this TD-labeled wavenumber and frequency combination are easterly waves, many of which originate over sub-Saharan West Africa.

Slide 3: These figures show the typical horizontal structure of easterly waves over the East and West Pacific. They are spatial regressions of OLR and 700 mb streamfunction relative to a lag equals zero days, which occurred at the time that the OLR minimum in an easterly wave was located at a pre-selected point. In the East Pacific, the middle panel, represents a lag of zero days, and the OLR minimum is at 10°N, 95°W, or in between the red arrows. The dark shading indicates where positive anomalies of OLR, or lower than normal precipitation, was present at the same time. At the same time, the streamfunction indicates the anomalous flow that is associated with an easterly wave. It is cyclonic around the OLR minimum, but very importantly, it is tilted toward the northeast with latitude. A weak signal in OLR and streamflow is apparent well downstream into the Central Pacific as well. The different panels show how the pattern evolves in time relative to the zero time. Thus, the top panel shows the wave anomalies 4 days before the maximum OLR anomaly, and the bottom panel represents the same 4 days afterward. The OLR anomalies propagate westward and poleward, while the streamfunction anomaly lessens in the Central Pacific. In the West Pacific, waves that propagate from the Central Pacific become highly amplified. The wave propagates toward the west near a latitude of 10°N, which is represented by the gray shaded area that is aligned with the blue lines that connect the various panels.

Slide 4: For the same regions, we can also look at the vertical structure of the wave at lag day zero. The top panels show the wave anomaly of OLR as a function of longitude. The OLR minimum represents where active convection in an easterly wave is present. The bottom three panels are vertical cross-sections of anomalous meridional wind, temperature, and humidity at 10°N. The darker, solid lines represent positive anomalies, and the lighter shaded, dashed lines represent negative anomalies. Some differences are seen between the East and West Pacific, but the basic structures are similar. The meridional wind anomaly indicates that vorticity is present in the wave, with a maximum vorticity in the middle troposphere. The maximum vorticity is a little lower in the West Pacific. Temperature anomalies are tilted with height. The result is that a positive temperature anomaly is located aloft over a negative temperature anomaly where the convection is present. This type of vertical structure for temperature anomalies is common where deep convection occurs in the tropics and is associated with the adiabatic motions in the gravity wave response to convective heating. The humidity anomaly is positive through the troposphere in the convection, with a maximum anomaly occurring between 700 and 850 mb. Thus, the convectively active part of an easterly wave is an area of moist air and enhanced vorticity, which is favorable for tropical cyclone development. The nearby regions to the west and east of the convective maximum are relatively cloud free and dry, with opposite signed vorticity and temperature anomalies present as well.

Slide 5: The track density of African Easterly waves is shown here, with red colors indicating the highest frequency of occurrence. Easterly waves originate in a few locations in sub-Saharan Africa, often focused on topographical features that we will see on a map soon. They move generally westward and slightly poleward as the move across West Africa and into the Atlantic. Some of the easterly waves consolidate into tropical cyclones and move poleward, while those that do not tend to move westward into the East Pacific, where they often move toward the northwest.

Slide 6: Not all easterly waves are African easterly waves, however. They can generate in the tropical North Pacific as well near northwestern South America and the Panamanian isthmus. This is an area with one of the highest occurrences of mesoscale convective systems in the world. These waves appear to be forced by topography in the region that produces a small region of vorticity that can grow upscale. A combination of observations and reanalysis illustrates how a region of positive relative vorticity develops in the Panama Bight, amplifies, then moves toward the northwest.

Slide 7: The currently displayed enhanced METEOSAT infrared satellite image alongside a water vapor absorption band shows several regions of enhanced convection over West Africa during a particularly active period. Each region of red indicates low brightness temperatures—or cold cloud tops, and each region is approximately located on the east side of a trough axis around 700 mb associated with an easterly wave. What is shown here is a sequence of easterly waves that occur in fairly rapid succession. Note how that trough axis is tilted eastward with height. As we will show soon, this is very important for amplification of the wave in an African easterly jet.

Slide 8: A drawing of a single trough axis is offered here that reinforces what we just saw. A trough that is tilted eastward with latitude is usually strongest around 700 mb and is accompanied by convergence upstream of the trough axis and divergence downstream. Therefore, the deepest convection occurs upstream of the trough axis.

Slide 9: The topographic map of Africa shown here depicts high elevation terrain with brown colors. African easterly waves are frequently observed to originate from the Ethiopian highlands, circled in the figure. Some easterly waves are also excited by terrain features in southern Chad and along the border with Cameroon and Nigeria. Easterly flow across the topography induces vorticity to form. The structure shown on the previous slide is advected westward and can be amplified by extracting energy from the African easterly jet, which is discussed in the next module.

Slide 10: Studies of the mean structure of African easterly waves extend back to the 1970s, such as the paper from which this figure was borrowed. This paper composited the features of African easterly waves by phase, which is denoted by the numbers as the bottom on the solid black line that follows the track of the average wave through space. These lines can be seen faintly in the background of the top panel. The top panel shows the structure of the anomalous wave component of winds at 700 mb. Although the background wind is easterly, the anomalous portion associated with the wave shows pairs of positive and negative relative vorticity that are tilted toward the east with latitude, much like we saw over the East Pacific.

Slide 11: They also looked at the anomalous wave horizontal circulation at multiple levels in different “phases” of the easterly wave at a location relative to what they defined as the disturbance center, denoted by the red plus symbol in each panel. Phase 4 on the x-axis is, based on the previous slide, the location where the maximum relative vorticity in the wave is located. The y-axes show the latitude relative to the latitude of the wave disturbance center, which varies by wave, but can be roughly estimated as 10–11°N. These figures show that vorticity is present up to at least 700 mb but it at its maximum around 850 mb, where a closed circulation may even develop. At 500 mb, the flow is far less rotational, and easterlies dominate along the wave trajectory and to its south.

Slide 12: The proximity of the waves to the Sahara Desert in North Africa is both a reason for their maintenance but can also be detrimental to their growth if dry Saharan air inhibits growth of deep convection. Dry desert air advected from the Sahara—which is at 15–20°N—into the track of easterly waves farther south is advected over the Atlantic by the mean easterlies and can become entrained within convective regions in easterly waves, preventing them from developing into tropical cyclones. The desert air is known as the Saharan Air Layer, or SAL, and is indicated in this figure by the yellow and orange colors that appear to be layered on top of the background shading for sea surface temperature. In particularly strong easterlies, the SAL can be advected all the way to North America. The dust carried in the SAL is often seen in visible satellite imagery, and channels centered on water vapor absorption bands can, when combined, quantify the dryness of the air. These events are impacted by the strength and location of the African easterly jet, which we describe next.

Slide 13: During the boreal summer, surface temperature over the Sahara can routinely exceed 40°C, while sea surface temperatures off the coast of western Equatorial Africa are much cooler, often around 25°C. This sets up a large pressure gradient between the ocean and a surface "heat low" over the desert.

Slide 14: However, the temperature profile over the desert follows a dry adiabat more closely than the marine atmosphere; therefore, the temperature gradient is dramatically reduced at 700 mb compared to low-levels and is actually usually reversed at 500 mb. Combined, this causes a shallow, large-scale, sea-breeze-like circulation to develop in response. We can consider the development of the easterly jet if we consider the direction of the thermal wind vector given the low-level temperature gradient from the ocean to the land.

Slide 15: Meridional, or latitudinal, cross sections of wind, temperature, vorticity, and moisture have also been documented. If we first look at zonal wind cross-sections relative to a reference latitude of 11°N, we see that low-level westerlies—denoted by the shaded area—are present in the very lowest levels, consistent with Coriolis acting upon the low-level component of the pressure gradient driven flow. Between 500 and 700 hPa, an easterly jet develops between the Sahara and the coast; this is required to maintain thermal wind balance. The maximum magnitude of the easterly jet, on average, is a little over 10 m/s and is centered a little above 700 mb. This jet can meander to the north and south, and its strength can change with time based on the contrast of temperature between the ocean and desert, which are implied by the bottom panel on this slide.

Slide 16: The absolute vorticity of the flow generally increases with latitude; however, a region exists where the meridional gradient is reversed. This region is shaded, and approximately coincides with the location of the jet. As we will see in the next module, this reversal is a requirement for barotropic instability of the flow, which is related to one mechanism through which easterly waves may be amplified.

Module 3.7

Slide 1: This module is a continuation of the previous and discusses the energetics of African easterly waves.

Slide 2: At the end of the previous module, we mentioned that a reversal in the typical meridional gradient of absolute vorticity, such that it decreases with latitude, is a necessary condition for barotropic instability. The meridional gradient of potential vorticity can be described by the Coriolis parameter *f*, and the streamfunction. If we assume the meridional flow is negligible and that easterly flow dominates, then we can express the PV gradient in terms of the mean zonal wind. Barotropic instability of the easterly flow, which can amplify the easterly waves, can only occur if this gradient is negative. The figure shown here shows mean PV over West Africa on an isentrope approximately located at 650 mb. The shaded region indicates where the meridional gradient of PV is negative, satisfying the condition for barotropic instability. It occurs where the easterly jet forms, downstream from the origin regions for easterly waves.

Slide 3: Maximum diabatic heating tends to occur to the south of the region of the negative PV meridional gradient. A positive vertical gradient of latent heating with height is a source of PV, meaning that convection acts as a PV source to the south of the region of negative PV gradient, helping to determine the location of the derivative reaching zero. The instability allows for the maintenance of easterly wave disturbances over West Africa.

Slide 4: The mean latitudinal distribution of PV at 650 mb is shown here. A similar reversal of the sign of the PV gradient is seen over northern Australia; however, we will focus on Africa in this module.

Slide 5: When this barotropic instability is present, the zonal mean kinetic energy can be converted into eddy, or wave, kinetic energy. Plotted here is the barotropic conversion, denoted by the expression highlighted in red. The x-axis is the difference in latitude from a reference of 11°N. Barotropic conversion has a maximum value at about 700 mb near the latitude of the easterly wave packet. The southwest to northeast orientation of the rotational flow in the wave, like we saw in the previous module, implies that easterly momentum is transported southward into the jet region.

Slide 6: Baroclinic conversion of zonal potential energy to eddy available potential energy is dominated by the term plotted here, which depends on the eddy covariance of the meridional wind with the temperature and the meridional gradient of temperature. This expression is positive at low-levels near the Sahara because northerly winds are often co-located with high temperatures, thus fluxing heat southward to the region of highest meridional temperature gradient.

Slide 7: However, the conversion of eddy available potential energy to eddy kinetic energy indicates an energy sink near the location of the jet and a small source of eddy KE near where the eddy available potential energy is generated. Combined, the effects of barotropic conversion of energy, in other words the transfer of energy from the mean flow to the eddies by the horizontal shear of the flow, seem to contribute most to the strength of the easterly jet, which is indicated by its co-location with the jet. However, more recent studies have suggested the that relative importance of barotropic and baroclinic instability in the jet may depend on the part the wave life cycle considered.

Slide 8: The top panel, like one from the previous module, indicates the track density of easterly waves in the Atlantic and Pacific, while the middle panel highlights areas where easterly waves form. Formation of easterly waves is preferred in areas where growth of the disturbances is dynamically supported. One of those areas, West Africa, is denoted by the dark shading in the bottom panel, indicating a positive growth rate of easterly wave amplitude. The other location where easterly wave growth occurs Is in the East Pacific near the Panama Bight.

Slide 9: Examining easterly wave growth in the East Pacific more closely, we look at budgets of eddy available potential energy first. The term plotted on the left, which is the first term on the right-hand side, describes the generation of eddy APE through spatial correlation between temperature and heating anomalies. This is a diabatic source of energy. The second term, plotted on the right, shows conversion of eddy APE to eddy kinetic energy. The final term describes baroclinic conversion of APE, but is an order of magnitude smaller than the two terms displayed. Also plotted are how the two terms vary with sub-seasonal variability of the zonal wind associated with features like the Madden-Julian Oscillation, which will we discuss in a future module. This variability occurs on timescales of 20 days or more. Generation and destruction of eddy APE are more common when background zonal winds at low-levels are less easterly than normal. Regardless of conditions though, the two terms largely cancel each other as the eddy APE produced by convection is converted to kinetic energy.

Slide 10: The largest sources of eddy kinetic energy in East Pacific easterly waves are both the eddy APE to kinetic energy conversion just shown, and displayed here in the second panel; and the barotropic conversion of zonal mean kinetic energy to eddy kinetic energy, as shown in the first panel. The first term is essentially the same as the that shown a few slides ago from West Africa, and we find here that it is also a source for amplification of easterly waves over the West Pacific.