Module 4.1

Slide 1: This module will briefly overview two major thermally direct circulations in the tropics: the Hadley Cell and the Walker Circulation.

Slide 2: The primary driver of circulations on Earth is incoming solar radiation. More solar radiation enters at low latitudes, and the tropical atmosphere and surface experience a net radiative heating. In other words, more moist static energy enters via solar radiation than exits via outgoing longwave radiation. At higher latitudes north of about 45° latitude, the opposite is true. More energy leaves the atmosphere via longwave radiation than enters via insolation. The results in a meridional energy imbalance that requires transport of energy from the tropics to the poles. This requirement is the fundamental driver of global-scale meridional circulations.

Slide 3: A typical boreal summer SST distribution is shown here, similar to others you have seen. The warmest SSTs are near the equator and in the western portions of the Atlantic and Pacific Oceans. Insolation warms the ocean more in the deep tropics than at higher latitudes, and poleward upper ocean currents transport the warm water along continental coastlines in the West Pacific and Atlantic. The highest SSTs are located in the waters around the Maritime Continent in the Indo-Pacific and in the waters surrounding southern North America during boreal summer. Therefore, the deep, buoyant, marine convection is most likely to be supported in the tropics relative to other regions on Earth and moreso where SSTs are highest.

Slide 4: The convection acts as the upward branch of a meridionally overturning, thermally direct circulation called the Hadley cell. The Hadley circulation transports air rich with moist static energy from low-levels in the tropics to upper levels through convection, then transports the MSE rich air poleward at upper levels. The flow slowly descends as it moves poleward, and the large-scale subsidence regions at latitudes near roughly 20–30°N/S are near where many precipitation minima on Earth are located. Note that Coriolis acting on the low-level flow induces the background low-level easterlies that are present across much, but not all, of the tropics. The characteristics of the Hadley circulation, including its poleward extent, are governed by radiative-convective equilibrium and eddy flow in the midlatitudes. The Hadley cell has widened over the past several decades as both mean global temperature and the mean latitude of eddy-driven jets have increased.

Slide 5: As we just saw, a Hadley cell exists on both sides of the equator. Somewhere in the tropics, the two cells converge and drive their shared upward branch. This is where convection occurs most and is called the Intertropical Convergence Zone, or ITCZ. The ITCZ moves throughout the course of the year in some parts of the world. For example, in the Atlantic, and climatologically speaking, the ITCZ reaches its northernmost latitude in boreal summer and autumn before returning toward the equator during winter. Over Africa, the convection shifts between hemispheres. Similar behavior is seen over the Maritime Continent, although the ITCZ is also heavily affected by heating over nearby landmasses that drive monsoons. However, generally, convection in the ITCZ also occurs mostly in the summer hemisphere, while rainfall is persistent year-round over the Maritime Continent itself. A notable exception is the East

Pacific. There, the latitude of the ITCZ remains approximately fixed around 10°N over the course of the year because SSTs in the equatorial East Pacific are generally too low to support deep convection and there is no Southern Hemisphere SST maximum in the East Pacific. However, the frequency of deep convection is much less in the East Pacific ITCZ during boreal winter than summer. All put together, the migration of the ITCZ implies that the upward branch of the Hadley circulation also migrates seasonally and that the cell is often asymmetric about the equator.

Slide 6: The seasonal asymmetry of incoming solar radiation has implications for the strength of the circulation in each hemisphere. On the right shows results of simple numerical simulations of the Hadley circulation with latent heating centered on the equator. The circulation is symmetric about the equator. On the left is the resulting flow if latent heating is centered at 10°N, a common occurrence during boreal summer. In this case, the summer hemisphere's Hadley circulation is weakened, while the cross-equatorial branch is much stronger. We can see this in the streamfunction shown at top left. Low-level easterlies are present on either side of the upward branch in the symmetric heating scenario, and this is broadly true for the asymmetric heating case as well, except that a narrow region of surface westerlies can occur between the upward branch and the equator when asymmetric heating occurs. The westerlies may be found, for example, over the Indian Ocean during boreal summer and fall and flow into the frequently convecting Maritime Continent region.

Slide 7: In some locations, especially over the East Pacific, a shallow return flow in the Hadley cell is also observed. This happens in particular where horizontal SST gradients are large and can sometimes be seen in the East Atlantic. The shallow return flow is itself a thermally direct circulation driven by a pressure gradient force between the equator and the ITCZ. When convection in the ITCZ is weak and less mass is transported upward, the return flow is observed to be stronger. A mid-level inflow is sometimes seen above the 0°C level as well.

Slide 8: In idealized numerical simulations of a narrow tropical channel containing an ITCZ centered near 12°N, the shallow circulation within the larger Hadley cell has important implications for meridional MSE transport. The numbers are proportional to the flux of mass, water, or moist static energy into or out of the ITCZ via the low-level inflow of the Hadley cell, the shallow return flow above it, the mid-level inflow, and the upper-level outflow of the Hadley cell. The flux of MSE into the ITCZ region was an order of magnitude less from the northern cell. In the southern cell, about one-third of the incoming MSE from the south is exported through the shallow return flow; however, the same order of magnitude is imported through the mid-level inflow, such that the shallow circulations only have a small impact on the net poleward transport of MSE at upper-levels. As expected the vertically integrated transport of MSE was poleward both north and south of the ITCZ, with total transport being about 2.5 times greater in the southern hemisphere, and eddy transport being about 2.5 times greater in the northern hemisphere.

Slide 9: If we zoom back out to our map of SST, we notice that most of the deep tropics is covered by ocean. A few places—equatorial Africa, the Maritime Continent, and equatorial

South America—however, experience more shortwave heating. The Maritime Continent is also located in the middle of the warmest SSTs anywhere on Earth, and its mountainous terrain helps to force convection as local thermally direct sea-breeze circulations drive convergence near the bases of large topographical features. As a result, the Maritime Continent is climatologically a location of year-round precipitation and is where the upward branch of the zonal thermally direct Walker circulation is located in the Indo-Pacific.

Slide 10: The analytically derived response of the tropical atmosphere to equatorial and off-equatorial latent heating in convection can be considered. Suppose some heat source with a first baroclinic structure exists that is symmetric about the equator. This heat source can be denoted as Q. Shown on the left are the equations used to analytically determine the response of the circulation to tropical heating. They are just the horizontal momentum and continuity equations. The upward motion will be confined mostly to where that heating occurs. Additionally, a far-reaching symmetric circulation response to that heating will occur. We can think of the heat source as convection over the Maritime Continent. In response, a combined Kelvin and Rossby wave like response will occur and be anchored to the heating. These waves are discussed with more detail in a different lecture series. The phase speed of the Kelvin wave would be three times larger than that of the Rossby wave, and so the eastward extent of the zonal circulation response will extend well to the east of the heat source, while the western circulation is more limited zonally. The Rossby waves that develop to the west of the heat source help induce low-level westerly flow that is usually seen over the Indian Ocean.

Slide 11: The response is very different if the heat source is placed off the equator. If placed north of the equator, a zonal circulation still develops, but the western component of the circulation response in particular is asymmetric. Still a zonal thermally direct cell develops near the equator.

Slide 12: Meridional circulations also develop as a result of the heating. On the left is the meridional response to symmetric heating. The zonal wind and streamflow is as we expect and saw in a previous slide. With asymmetric heating, the circulation response only includes one primary Hadley cell, and surface westerlies are found between the rising motion and the equator as seen in a previous slide. Thus, both the major meridional and zonal thermally direct circulations in the tropics can be largely explained by the location of heating in the tropics. The heating is controlled largely by the distribution of landmasses and SSTs.

Slide 13: The zonally direct circulation is called the Walker circulation. The Pacific Walker circulation, seen at center in this figure, possesses the basic characteristics seen in the previous slides. The low-level flow is driven by pressure gradients that are induced by the zonal gradient in SST—and therefore boundary layer temperature. Other upward branches also exist, most notably over the Western Atlantic. These zonal circulations result in a reduction of background easterlies—or even mean westerlies—in the parts of the circulation to the west of deep convection. This is especially true over the Indian Ocean and sometimes also seen in the equatorial East Pacific. In the next module, we will discuss how variability in the strength of the Walker circulation has ramifications on weather globally.

Module 4.2

Slide 1: This module will qualitatively describe the El Niño-Southern Oscillation, or ENSO, and a related phenomenon over the Indian Ocean. Teleconnections associated with these phenomena are discussed in a different module.

Slide 2: Taking a look at a typical boreal summertime SST distribution, we notice a “cold tongue” of SST at the equator off the western coast of South America. It is the northern extent of cool SSTs along an equatorward cold-water current where stratocumulus cloud is present beneath a sharp inversion. Note also how this is an example of how ocean currents transport energy from the equator to the pole in the net just like circulations in the atmosphere. As a result of these ocean currents, a pressure gradient is present along the equator from east to west. Physically, this is the driver of the easterly flow across the Pacific. The easterly flow reinforces the ocean currents and helps to advect surface ocean water toward the west at the equator. This also induces upwelling off the western coasts of South America and equatorial Africa, especially south of the equator, transporting nutrient-rich water from deeper in the ocean and supporting vibrant marine ecosystems. When the temperature of the ocean water at the equator warms or cools relative to its climatological average, this upwelling is either enhanced or reduced. As the temperature gradient across the Pacific increases, so does the pressure gradient, and then so does the circulation. This variability, which usually occurs on irregular interannual timescales, is known as the El Niño-Southern Oscillation.

Slide 3: An example of a weak La Niña event is shown here for the same time period as for the SSTs shown in the previous figure. Shown here are the anomalies of SST relative to the seasonal climatological average over a 40-year long historical period. Notice how a “tongue” of cooler than normal SSTs is present near the equator in the Eastern Pacific. At the same time, warmer than normal SSTs are present over the Western Pacific near the Maritime Continent. Together, these enhance the cross-Pacific pressure gradient, which strengths the Walker circulation.

Slide 4: We can visualize this process in a nicer looking graphic from Australia’s Bureau of Meteorology. The current figure illustrates typical ENSO neutral conditions. The expected cross-Pacific SST gradient is seen. The gray lines indicate the Walker circulation, with convection occurring over the Maritime Continent. In the ocean, the thermocline is deeper over the West Pacific, and upwelling is persistent in the East Pacific.

Slide 5: During La Niña, this pattern is amplified. More rain occurs than normal over the Maritime Continent, low-level easterlies are stronger across the Pacific, and upwelling in the East Pacific is enhanced while the thermocline is located closer to the surface there.

Slide 6: During El Niño, the opposite happens: SSTs are warmer than normal in the East Pacific, and cooler than normal in the West Pacific. While the West Pacific is usually warmer, a weakening of the easterly trade winds occurs nonetheless. The depth of the thermocline becomes more uniform with longitude, and upwelling is reduced, resulting in an anomalous westerly current in the surface ocean and reduced upwelling in the East Pacific, which has important negative impacts on fisheries there. The Walker circulation weakens, and in the cases of strong El Niño events, can lose its familiar form of driving easterlies across the entire Pacific. More convection than normal occurs over the Central and East Pacific, while rainfall is reduced over the Maritime Continent and West Pacific. This can also be seen in anomalies of outgoing longwave radiation, in which positive anomalies will be detected where rainfall is suppressed. As we will see in another module, where latent heating occurs in the tropics has dramatic implications on mid-latitude weather. Indeed, ENSO is responsible for changes in weather patterns throughout the world.

Slide 7: The SST anomalies during the very strong 1997 El Niño event are shown here. The deep red colors over the equatorial East Pacific denote SST anomalies of up to 5 degrees, which is an unusually large anomaly.

Slide 8: The La Niña event that followed in 1998 also had large temperature anomalies—about 2–3 degrees in the East Pacific. Note how in both figures, the temperature anomaly, in this example, extends far into the Central Pacific.

Slide 9: Indices used for operational and historical purposes quantify the strength of an El Niño or La Niña event based on the SST anomaly somewhere in the East or Central Pacific. Different indices consider the temperature anomaly in different longitude bands across the equatorial Pacific. The Niño 3.4 region, highlighted by the bold box in the panel at left, correlates with changes in weather patterns over North America. A snapshot of time series of some Niño indices is shown at right for 2018 into 2019. This example shows weak temperature anomalies in the East Pacific in the Niño 1+2 region, and anomalies of generally less than 1 degree in the 3.4 region. These are indicative of a marginal to weak El Niño event. Some other examples of ENSO indices can be found at the link on this slide.

Slide 10: A similar phenomenon occurs over the Indian Ocean, and is known as the Indian Ocean Dipole, or IOD. It involves interannual variability in the strength of the Walker cell over the Indian Ocean and the magnitude of westerlies in the lower atmosphere.

Slide 11: As seen previously, during neutral states, convection occurs predominantly over the Maritime Continent, and SSTs are relatively uniform across the equatorial Indian Ocean from west to east, although perhaps a little cooler in the western part of the basin.

Slide 12: During the negative IOD phase, this circulation is enhanced with warmer SSTs over the Maritime Continent and cooler SSTs near Africa. This results in enhanced ocean upwelling near Africa, just as in the case of South America during La Niña.

Slide 13: During the positive phase, the circulation is weakened, with anomalous easterlies in the lower atmosphere and upper ocean, increased upwelling or reduced downwelling over the Maritime Continent, and a reduction of rainfall near the Maritime Continent and Australia. The IOD is usually linked closely to ENSO variability, with a positive phase being consistent with El Niño and suppressed convection over the Maritime Continent.

Slide 14: However, it does appear likely that the IOD is a dynamically separate event with variability driven by factors other than ENSO. The strong IOD event that occurred in 2019 in the absence of a strong ENSO event is one example of this. The 2019 event is shown here with positive temperature anomalies shown in red and negative anomalies shown in blue.

Slide 15: When quantifying the IOD based on the difference in temperature anomaly between the western and eastern Indian Ocean, the 2019 event was the strongest in recent history, as can be seen in the time series of the monthly dipole mode index, or DMI, seen at top right, which provides evidence that the IOD can form in the absence of forcing related to ENSO.

Module 4.3

Slide 1: Monsoons impact billions of people in South and Southeast Asia, the Maritime Continent, and Northern Australia annually. These large-scale rain events are responsible for much of the annual precipitation that occurs in many locations in the region. Essentially, the monsoon is the off-equatorial migration of the Intertropical Convergence Zone onto nearby landmasses. This module will discuss some of the basics of the South Asian-Australian monsoon.

Slide 2: Shown here again is satellite derived mean annual precipitation for Earth. It is maximum over the Indo-Pacific warm pool, where it is enhanced by warm sea surface temperatures and numerous topographical features. The black ellipse highlights the region where most of the monsoon rainfall occurs. Not all of the rainfall is associated with monsoons, but much of the rainfall over the continental landmasses in the tropics is. Note in contrast how dry the region north of the Himalayas is. Dry air, which is not supportive of the deep convection that develops during an active monsoon, is present there. The Himalayas act as a barrier against that dry air reaching lower elevations south of the range.

Slide 3: Typical onset dates of the summer monsoon are illustrated here. Generally speaking, the onset date is earliest near the equator and a broad, but not contiguous, area of precipitation progresses northward across India. First onset of the monsoon over most of India is climatologically in early to late June. Monsoon onset starts early over topographic features in southeast Asia and progresses toward the northeast into China later into the summer.

Slide 4: The fundamental forcing for the monsoon is the temperature contrast between the ocean and the land, much like we discussed for easterly waves that develop in the West African monsoon. The temperature difference drives a pressure gradient-driven flow toward the land, although rotational effects and the westerly Somali jet are also important components of the mean monsoon flow as we will see shortly. Anticyclonic flow develops over land in the upper troposphere, where the pressure gradient force is oriented generally from land to ocean. Prior to onset, the landmass can become very hot, setting up a strong pressure gradient.

Slide 5: After monsoon onset, the temperature gradient weakens as the ocean remains the same temperature but the land becomes cooler after rains begin and shortwave surface insolation is reduced. However, diabatic heating—both radiative and latent—help to reinforce the thermally direct component of the monsoon circulation after its onset.

Slide 6: The heaviest precipitation falls over eastern India and surrounding countries, including essentially all area immediately south of the Himalayas, which extend up to over 8000 meters high. However, locally enhanced precipitation is also observed where low-level westerlies encounter topography. This happens in the locations circled in black. Over these terrain features, and to their west, the monsoon circulation can modulate rainfall as mesoscale convective systems form near the topography and move westward, a feature that we saw earlier in the course is also diurnally dependent.

Slide 7: An active monsoon can be reliably identified based on the magnitude of the upper tropospheric temperature anomaly. During the summer monsoon, 200–500 mb temperature is greatest over the northern Indian subcontinent, as seen by the shaded region, which indicates where climatological mean temperatures are warmer than those at the same level over the equator. During boreal winter, the highest temperatures are located over the Maritime Continent. Therefore, the winter monsoon has essentially an opposite signed structure.

Slide 8: The mean low-level flow of the South Asian summer monsoon is shown here. Numerous features are present. First of all, note that the circulation is not simply a thermally direct circulation extending from the ocean to the warmer landmass. Importantly, the flow is influenced by the westerly Somali jet, which is governed largely by terrain in East Africa. The jet is the northern portion of moist cross-equatorial flow and is a major supply of moisture for the monsoon convection. The curvature of the cross-equatorial flow is expected for a steady state flow in quasi-geostrophic balance with a negative meridional pressure gradient present. Recall that we can only assert geostrophic balance on the largest scales in the tropics. In the absence of rotation, such as at the equator, the PGF is balanced by friction. Vortices are also found upstream of terrain features, such as to the west of the Western Ghats in the Arabian Sea, in the Bay of Bengal, or in the South China Sea. The location of the largest rainfall totals are shaded in gray and extend southward across India, Bangladesh, and Myanmar from the Himalayas. The red line indicates the approximate location of the monsoon trough. This is a region of enhanced relative vorticity and can lead to the development of tropical cyclones in the West Pacific, South China Sea, or Bay of Bengal. The monsoon trough is not fixed in space, and its extension into the West Pacific enhances low-level westerlies to the south of the trough axis, which enhances wind-driven surface fluxes and can be further favorable for deep convection.

Slide 9: An example of convection in the monsoon seen in visible imagery is shown here. Plenty of convection is active across South India and the Bay of Bengal, as well as Southeastern Asia. The trajectory of the flow is also apparent in the cloud field and is reminiscent of the previous schematic of low-level flow in the summer monsoon.

Slide 10: Locally, convection may be modulated near small topographical features impinged upon by the mean westerly low-level flow. This schematic is an example of how rainfall along the Western Ghats, along the western coast of the Indian Peninsula, is sensitive to the strength of the Somali jet. During strong winds, convection is strongly forced over the topography; however, dry air intrusions over the Arabian Sea prevent convection from forming offshore. During periods of weak westerlies, convection is not as strongly forced over land but is more likely to occur offshore.

Slide 11: Monsoon depressions are common in the Bay of Bengal and generally form along monsoon troughs. After developing, they tend to move toward the north and west and are additional features that contribute to heavy rainfall in Bangladesh and eastern India. Lines in this figure follow the tracks of monsoon depressions from the Bay of Bengal toward the west, and brighter colors denote deeper, stronger depressions. The exact mechanisms responsible for their northwestward motion are as of yet unclear.

Slide 12: During boreal winter, the monsoon circulation reverses as the landmass over Asia becomes relatively cold to the tropics and Northern Australia. The curved flow across the equator is again expected as Coriolis goes to zero. Topographically induced gyres may form in the South China Sea, and tropical cyclogenesis is also possible in regions of enhanced vorticity near Northern Australia. During the winter, the ITCZ shifts southward over the equator, and rainfall is more focused over the Maritime Continent, including Northern Australia. See the video linked at the bottom of this slide to watch a 3–4 minute-long NASA video showing various satellite data during the summer and winter Asian-Australian monsoons.

Slide 13: Monsoon activity is also subject to intraseasonal variability with timescales of 20–100 days. Intraseasonal oscillations are discussed with more detail in a different lecture series. However, intraseasonal variability can be separated into 8 phases. This figure shows anomalies of outgoing longwave radiation during the 8 phases during boreal winter on the left, and summer on the right. If we focus on Phase 2, we see blue colors over the Indian Ocean during both winter and summer. During winter, the blue area, which represents enhanced convection, propagates along the equator toward the east, impacting the Maritime Continent and Northern Australia. This is associated with the Madden-Julian Oscillation. During boreal summer, the blue region moves toward the northeast, eventually extending zonally from South Asia to the West Pacific, impacting Southeast Asia. This is known as the Boreal Summer Intraseasonal Oscillation. This pattern of enhanced and suppressed rainfall may be superimposed onto the background monsoon circulation, causing spatio-temporal variability in precipitation during the monsoon.

Module 4.4

Slide 1: This module will review some of the off-equatorial impacts to the global circulation caused by latent heating in the tropics. In particular, we will look at the impacts of variability in tropical convection caused by ENSO and the MJO on mid-latitude weather.

Slide 2: Again, we'll start by looking at a typical distribution of sea surface temperature, and although this is not a long-term average, it does capture the basic features of the SST distribution during boreal summer. The deepest, most persistent convection in the tropics typically occurs over the Indo-Pacific warm pool, and variability to this convection has impacts on mid-latitude dynamics far from where the tropical convection occurs. Interannual variability, such as ENSO, modulates the frequency of convection throughout the Pacific. For example, during El Niño, convection—and thus latent heat release in the atmosphere—shifts toward the east. On shorter sub-seasonal timescales, the MJO also modulates convection, particularly over the Maritime Continent, Indian Ocean, and West Pacific. The variability in tropical heating causes off-equatorial Rossby wave trains to develop, which we will now explore a little.

Slide 3: In a simple atmospheric model, an anomaly of diabatic heating can be inserted as an initial condition, and the response can tell us about the Rossby response to the heating. For example, consider the current figure from a 1982 paper. In the upper right of the figure are contours that indicate where latent heating was placed in the model. This is located over Southeast Asia and the West Pacific north of the equator. The heating had a first baroclinic-like vertical structure, with maximum heating near 400 mb. The model also included seasonally appropriate zonal winds as a function of latitude.

Slide 4: The steady-state response to that initial forcing is seen here. The approximate location of the initial heating anomaly is depicted by the red X, and the different panels indicate the perturbation geopotential at different pressure levels. A couple of things are obvious. First, a wave train develops and appears to propagate over the North Pole and into Europe. Second, the structure is similar at all levels, although its magnitude is largest in the upper troposphere. This indicates that the response is approximately barotropic.

Slide 5: Interestingly, the response to heating had seasonal dependence in the model. On the left are the geopotential anomalies caused by heating at 15°N and S in January, and on the right is the same for July. The geopotential response is maximized for heating in the same hemisphere during the winter.

Slide 6: Finally, the extratropical response is maximized for off-equatorial tropical heating. Heating at higher subtropical latitudes elicits a more muted response. Near equatorial heating also generates only a small extratropical response, largely because the vorticity generation by heating focused near the equator is small. Practically, this means that disturbances that modulate off-equatorial convection can alter mid-latitude geopotential—and therefore weather patterns. Variability in monsoonal convection is one example of an off-equatorial tropical source of heating that can modulatee extratropical geopotential. The Rossby wave component of the convectively active MJO also elicits off-equatorial latent heating. ENSO generally modulates heating on larger spatial scales for longer time periods than the MJO and also affects the propagation of Rossby wave trains out of the tropics. By shifting the pattern of geopotential globally, the tropical disturbances can impact weather in the subtropics and extratropics on timescales consistent with the tropical variability.

Slide 7: The barotropic Rossby wave train excited by tropical convection can also be represented analytically. A dispersion relationship describing the motion is shown at top with zonal and meridional wave numbers included. The M subscripts just indicate that a Mercator projection is used, and so the mean zonal wind and beta are scaled to the new projection. The group velocities in both the zonal and meridional wind direction are shown, and the magnitude of the group velocity for a stationary wave is shown at the bottom. This means that the group velocity for a stationary wave depends on the sign of the mean zonal wind, which is generally westerly aloft at mid-latitudes. Therefore, the energy in the Rossby wave train propagates faster than the phase of the wave and can cause amplification of existing downstream geopotential anomalies. The figure on the right shows how the phase of the wave propagates off the equator and toward the east with the mean flow that increases with latitude. The position of the anomalous troughs and ridges affect the mean storm tracks at mid-latitudes.

Slide 8: This quasi-stationary Rossby response is largely responsible for many of the extratropical teleconnections to tropical heating. Shown here are known climatological relationships between anomalous heating over the Pacific during El Niño and anomalous interannual variability in surface weather conditions in the Pacific and elsewhere. The sign of the changes can be approximately reversed for La Niña conditions.

Slide 9: During a strong El Niño event, anomalous divergence aloft and deep tropospheric diabatic heating may be found over the Central Pacific, such as in this example of the 1986–1987 El Niño. The equatorial anomaly excites a streamfunction that extends well off the equator as a wave train that roughly follows the red lines in middle and high latitudes. The negative streamfunction is consistent with negative geopotential perturbations. This means that the tropical heating anomaly drives a wave response that causes anomalous troughing as far away as the western Atlantic and southeastern United States. Since this plot shows the response during boreal winter, the signal is stronger in the Northern Hemisphere than in the Southern Hemisphere.

Slide 10: The extratropical response to the tropical heating anomaly can occur in just a couple of weeks as seen in this time series map of perturbation streamfunction for anomalous latent heating in the Central Pacific. A high amplitude wave train sets up over North America within two weeks that is consistent with the El Niño teleconnections observed in the United States.

Slide 11: The response is essentially of opposite sign for La Niña.

Slide 12: The MJO also causes variability in tropical precipitation, although it acts on much shorter time scales than ENSO. Nonetheless, sub-seasonal variability in the extratropical circulation is seen as a result of modulation in heating over the tropical warm pool. In this figure, each panel represents a different so-called “phase” of the MJO, and the colors indicate the anomalies of blocking frequencies over the North Pacific. Blocking refers to the formation of high amplitude ridges that reduce or eliminate the zonal flow that guides the typical storm track at mid-latitudes. Each panel therefore represents the change in blocking frequency from a long-term mean when convection in the MJO is located in a certain location consistent with the phase listed. For example, for Phases 3 and 4, recall that convection is more active than normal over the Maritime Continent. These plots show clear sub-seasonal variability in blocking frequency, with statistically significant results hashed. For example, during Phase 4, less than normal blocking—or more troughing than normal—is seen over the North Pacific. During Phase 8, the opposite is true, and more than typical blocking occurs. Anomalous troughing may be seen, for example, over the Northeast Pacific and Pacific Northwest when the MJO convection projects onto the first two principal components of sub-seasonal OLR variability in the tropics such that the MJO Phase is 5. See the module on MJO identification for more information about MJO phases.

Slide 13: Similar behavior is seen in the North Atlantic as well. Plotted here are blocking frequencies at 5, 10, and 15 days after Phases 3 (top) or 7 (bottom) of the MJO occurred. Generally, anomalous troughing is in place 1–2 weeks after Phase 3, and anomalous blocking is present in the same time period after Phase 7. Amplification of the signal is seen between 5 and 15 days after the anomalous heating in the MJO occurs because some time is required for the Rossby wave train to reach the North Atlantic, which explains the delay in the signal arriving. In fact, by the time the MJO is at Phase 7, convection is relatively suppressed over the Indo-Pacific, and the blocking signal seen over the Atlantic probably emanated from the convection several days prior when the MJO was at Phase 3.

Slide 14: This type of delayed response can be seen in Hovmöller-like plots that use MJO phase on the y-axis as a proxy for time assuming that MJO phase always progresses forward. We will see more plots like this in the paper discussion for this topic. Shading on this plot represents the anomalous frequency of atmospheric rivers at some time period listed on the x-axis after the phase shown on the y-axis. Plots are shown for the Pacific Northwest and for California. For example, the dark green color in the red circle means that more than normal atmospheric river activity occurs in British Columbia about 14–21 days after the MJO was in Phase 2. As the MJO convection moves eastward and its circulation eventually becomes decoupled from convection, the leftover, already produced Rossby response will continue to propagate so that the anomalous troughing and atmospheric river activity reaches the coast by the time the MJO is in Phases 5 or 6. Similar behavior is seen in California, although the negative signed response is higher in amplitude. This may be indicative of MJO-driven sub-seasonal variability of Santa Ana wind events.

Slide 15: Finally, just as a few more examples, the MJO can affect other types of weather events as well, presumably through the extratropical Rossby wave train that anomalous tropical heating excites. This figure was borrowed from a paper that examined sub-seasonal variability of severe weather over the Central United States. Similarly to the last plot, we can see how a lagged response exists between severe weather events or their thermodynamic and dynamic indicators and MJO activity. Currently, models do not capture this type of variability very well, meaning that the possibility still remains for gaining more skill at predictability for medium lead times of 2–3 weeks.