Interseasonal and Intraseasonal Variability

Introduction

In the mid-latitudes, the atmosphere exhibits strong synoptic-scale variability such that climatology is a poor representation of day-to-day weather. In the tropics, however, climatology does a good job (to first order) of approximating the day-to-day weather conditions at a given location. This makes localized variability more important for explaining departures from climatology within the tropics. These effects can be manifest on multiple spatial and temporal scales and impact energy, moisture, and momentum transport within the tropics and between the tropics and mid-latitudes. The best-known of these effects, manifest as an interseasonal oscillation, is ENSO. In addition to ENSO, however, there exist numerous modes of intraseasonal and interseasonal variability within the tropics, many with ties in whole or part to the heating distribution across the tropics. In the following, we consider four such modes of variability.

Key Concepts

- What is the Madden-Julian Oscillation and what are its impacts?
- What is the Quasi-Biennial Oscillation and what are its impacts?
- What are other modes of tropical variability and why are they important?

The Madden-Julian Oscillation

Background and Overview

The Madden-Julian Oscillation, or MJO, was first detailed in 1971 and affects phenomena on multiple scales ranging from the convective-scale to the planetary-scale. The time period of an MJO event is typically between 30-60 days and is manifest by a coherent cycle in sea-level pressure and atmospheric winds, particularly in the boundary layer and near the tropopause. These variations in sea-level pressure and atmospheric winds result in local variations in rainfall and cloud cover. In many regards, the MJO can be viewed as a local, progressive modifier to (or mode of variability in) the Walker circulation. The MJO reflects a coupled atmosphere-ocean mode of variability; however, herein, we focus almost exclusively on the atmospheric component of this variability. MJO events initiate roughly along the Equator near the Maritime Continent. They propagate eastward at approximately 5 m s^{-1}, thus taking approximately 30-60 days to circle the globe. An MJO event does not always continually circle the globe, however. Indeed, the amplitude of an MJO event can and often does weaken as the event progresses eastward, particularly near the Maritime Continent. MJO events are strongest in the Indian Ocean, Western Pacific Ocean, and near Australia during the local summer months, particularly in the Southern Hemisphere.

To first order, the climatology of the tropics is defined by lower-tropospheric easterly winds. As enhanced deep, moist convection associated with an MJO event passes a given location, these trades weaken or become westerly. The passage of the deep, moist convection heralds the onset of the “westerly” or active phase of the MJO, with accompanying lower-tropospheric zonal convergence and reduced surface pressure. As the MJO event progresses eastward, anomalous subsidence moves in, subsequently followed by strengthened easterly flow associated with lower-tropospheric zonal divergence and elevated surface pressure. The onset of the enhanced easterly trades heralds the arrival on the “easterly” or inactive phase of the MJO. The sense of the wind and pressure anomalies in the upper troposphere is opposite to that in the lower troposphere.

From shear vorticity arguments, enhanced easterlies along the Equator lead to enhanced anticyclonic shear to the north and south of the Equator. Conversely, enhanced westerlies along the Equator lead to enhanced cyclonic shear to the north and south of the Equator. Thus, the "westerly" (or active) phase of the MJO is typically associated with anomalous lower-tropospheric cyclones and anomalous upper-tropospheric anticyclones north and south of the equator. Conversely, the "easterly" (or inactive) phase of the MJO is typically associated with anomalous lower-tropospheric anticyclones and anomalous upper-tropospheric cyclones north and south of the equator.

MJO-Related Atmospheric Variability and Impacts

The signal of the MJO is not constant across seasons or years; rather, it is strongest during the Southern Hemisphere summer months and during neutral ENSO phases. It is weakest during strong El Niño and La Niña events. This enables us to infer that the MJO, ENSO, and Walker circulation are to some extent interrelated phenomena. Further, modulations in the start and end times of the Asian, African, and Australian monsoons may arise in part due to forcing from the MJO. The enhanced phase of the MJO can promote more-organized convective modes and an enhanced potential for tropical cyclone activity, the latter of which arises due to the MJO's impacts upon the background absolute vorticity within tropical cyclone formation regions. The anomalous easterly and westerly trades associated with the MJO impact surface ocean stress and may potentially be important for the excitation of equatorially trapped Kelvin waves and/or ENSO phase changes.

On the large-scale, similar to ENSO, the MJO can ignite a Rossby-wave train into the mid-latitudes from the Maritime Continent eastward. This can be understood in the context of two complementary perspectives: diabatic heating's effects on the vertical distribution of potential vorticity and upper-tropospheric meridional and zonal heat transport by the Hadley cell and subtropical jet, respectively. For a convective vertical diabatic heating profile, positive potential vorticity will decrease following the motion in the upper troposphere and increase following the motion in the lower troposphere. This results in subtropical ridge amplification following the motion, which is nominally poleward in the upper-tropospheric branch of the meridional Hadley cell circulation. Subsequent downstream heating transport, both meridional and poleward, by the subtropical jet and downstream development associated with subtropical ridge building can result in impacts far removed from the MJO event in both the subtropical and middle latitudes. Zhang (2013) summarizes many of the impacts of the MJO upon weather and climate within both the tropics and higher latitudes.

MJO Formation

Describing the formation of the MJO poses one of the biggest challenges in all of tropical meteorology. First attempts at describing the MJO were based off of Kelvin wave forcing, likely resulting from the eastward propagation of both Kelvin waves and the MJO. However, Kelvin waves (whether convectively coupled or dry) typically propagate more rapidly and have shorter longevity than the MJO. Later attempts at describing the formation of the MJO can be split into two camps: internally forced and externally forced. The former describes a system in which the MJO is driven by its own energy source of some form; the latter describes a system in which the MJO is driven by something external to the system, whether tropical or mid-latitude in nature. The majority of related theories in both camps rely on convective heating as the predominant driver of the MJO.

Internally forced MJO theories are based upon a local source of instability, tied to deep, moist convection, which supports MJO growth. The requisite moisture source for such instability and deep, moist convection is said to arise from either evaporation from the underlying ocean surface or advective processes. Such theories may be tied to Kelvin waves that support local convective growth on both MJO and Kelvin wave time scales. Externally forced MJO theories are based upon a non-local source of instability to drive deep, moist convection and the accompanying MJO signal. There exist many potential forcing mechanisms, such as the monsoon, stochastic processes associated with deep, moist convection, and eddy forcing from the mid-latitudes, but there as of yet exists a lack of strong evidence to support any MJO formation theory.

MJO Monitoring and Forecasting

In analyses of fields such as outgoing longwave radiation (as a proxy for convection) and zonal velocity, the MJO is most readily identified when the interannual (e.g., ENSO), annual, and seasonal cycles are removed from the data. The data that remain are largely functions of intraseasonal variability and associated meteorological phenomena. This filtering process enables us to identify the MJO in a simple, concise manner using these fields. An example is that discussed earlier to identify equatorial wave modes from noisy observational outgoing longwave radiation data, wherein the spatial scale, return period, and propagation characteristics can be used to isolate its impacts upon outgoing longwave radiation from those of both slower- and faster-evolving phenomena.

More complex analyses, such as outlined by Wheeler and Hendon (2004), enable the phase of the MJO to be concisely depicted in a "phase space" plot. In this specific example, an empirical orthogonal function (EOF) analysis is conducted on observed outgoing longwave radiation and upper- and lower-tropospheric zonal wind data. An EOF analysis is a mathematical tool that provides information relating to the modes of spatial variability of a given time series. The leading modes of the EOF analysis contain the most information about the variability in the system, often explaining 20-50% of the variance therein. However, an EOF analysis does not always provide you with a physically explainable mode of variability and care thus must be taken to interpret its output accordingly. It is also worth noting that substantial variability often exists outside of the first few modes of variability, or principal components, from an EOF analysis. Furthermore, such modes of variability themselves do not cause anything to happen; rather, they are numerical reflections of what is happening or the patterns that may cause something to happen.

In Wheeler and Hendon (2004), the first two leading modes of an EOF analysis of near-Equator meridionally averaged outgoing longwave radiation, 850 hPa zonal wind, and 200 hPa zonal wind are retained. Next, temporal variability on long time scales – annual to interannual – is removed from daily observed data, with the resulting fields subsequently projected upon the retained two EOF modes. Two numbers result, termed RMM1 and RMM2, that describe how well the observed data project against the retained two EOF modes. Large positive values for RMM1 and RMM2 mean that the observed data are strongly *directly* related to the spatial patterns associated with each of the two EOFs; large negative values for RMM1 and RMM2 mean that the observed data are strongly *inversely* related to the spatial patterns associated with each of the two EOFs. Eight phases, each describing a unique configuration of zonal wind and deep, moist convection, are used to describe the variability of the MJO system as obtained from RMM1 and RMM2.

Together, EOFs 1 and 2 explain 25% of the total variance. They are well-separated from the third EOF, which explains only 6.1% of the total variance. EOF1 is associated with enhanced convergence over the Maritime Continent. This leads to deep, moist convection and accompanying reduced outgoing longwave radiation. Lower tropospheric westerly (easterly) zonal wind anomalies are found across the Indian Ocean and Maritime Continent (Western Pacific Ocean). Anomalous upper tropospheric flow is opposite to that in the lower troposphere. EOF2 is associated with enhanced convergence and, thus, reduced outgoing longwave radiation across the Pacific Ocean. Lower tropospheric westerly (easterly) zonal wind anomalies are found across the eastern Maritime Continent and western Pacific Ocean (Central Pacific Ocean). Again, anomalous upper tropospheric flow is opposite to that in the lower troposphere.

Wheeler and Hendon (2004) found that 60% of the variance in EOFs 1 and 2 is concentrated in the 30-80 day range, or the temporal range of the MJO, enabling daily, unfiltered data to be projected onto the PCs of EOFs 1 and 2 to obtain information about the MJO. However, 40% of the variance in EOFs 1 and 2 is concentrated on shorter and, particularly, longer time scales such that other modes of variability may influence these analyses of MJO phase. PCs 1 and 2, described by RMM1 and 2, exhibit a lag temporal correlation with a maximum lag of RMM2 to RMM1 of about 9 days.

The Wheeler and Hendon (2004) MJO index is not the only such index that is used to identify and track the MJO. In recent years, other indices such as the OLR MJO index and its variants (Kiladis et al. 2014) and velocity potential MJO index (Ventrice et al. 2013) have been developed. The OLR MJO index and its variations rely on projecting observed, time-filtered outgoing longwave radiation data onto spatial EOF fields of time-filtered outgoing longwave radiation data, with subtle variations in the time filtering distinguishing between variations of this index. The velocity potential MJO index uses upper-tropospheric velocity potential – a measure of the divergent component of the horizontal wind – in lieu of outgoing longwave radiation and a lower-/upper-tropospheric zonal velocity EOF analysis. As is apparent, however, these indices closely resemble the Wheeler and Hendon (2004) index, again emphasizing the role of deep, moist convection and near-equatorial zonal velocity to MJO monitoring.

The Quasi-Biennial Oscillation

Introduction to the QBO

The discovery over the Quasi-Biennial Oscillation, or QBO, can be traced to two events. In 1883, the Krakatoa volcano in Indonesia erupted, transporting volcanic ash and dust into the stratosphere. The volcanic dust plume circled the globe in an east-to-west direction on easterly stratospheric winds. However, the first regular radiosonde observations a couple of decades later suggested westerly winds were present near and above the tropopause in the lower stratosphere. This apparent conundrum was resolved in the 1960s by the discovery of an oscillation in the zonal wind direction within the stratosphere. The period of the oscillation was found to be ~26 months, or approximately biennial.

The QBO is most evident within zonal wind fields, and most monitoring of the QBO relies on stratospheric zonal wind analyses. However, the QBO is also manifest in the stratospheric temperature field. Easterly wind phases of the QBO are generally stronger than westerly wind phases. The QBO is confined between 10 and 100 hPa. The QBO signal starts near 10 hPa (~30 km) and propagates downward with time at a rate of approximately 1 km per month (or ~3.86×10^{-4} m s^{-1}). Easterly winds propagate downward at a slower rate of speed than do westerly winds. This has impacts upon the duration of the easterly and westerly zonal winds in the upper and lower stratosphere: easterly (westerly) zonal winds are found over a longer period of time in the upper (lower) stratosphere. The QBO's maximum amplitude in the horizontal is found between 10-30°N/S and is found in the vertical near 20 hPa. The transition zone between decaying and developing QBO phases typically sets up between 30-50 hPa. There is considerable variability in the timing (i.e., duration) and amplitude of any given QBO event.

QBO Dynamics

Herein, we utilize the schematic adapted from Plumb (1977)'s work that is contained within the lecture materials to describe the dynamics of the QBO. This schematic should not be interpreted in a *z-x* framework; rather, it should be interpreted as depicting two concurrent forcings with coherent vertical structures at a given location. The QBO is driven by vertical momentum transport associated with equatorially trapped Kelvin and mixed Rossby-gravity waves. An equatorially trapped Kelvin wave moves eastward at some phase speed; herein, we will let that be equal to $+c$. With this propagation comes westerly, or positive, momentum. A mixed Rossby-gravity wave moves westward at some phase speed; herein, we will let that be equal to $-c$. With this propagation comes easterly, or negative, momentum.

An initial vertical profile of the zonal wind between the tropopause and stratopause is depicted in schematic panel (a). A narrow westerly jet is found in the lower stratosphere while a broad easterly wind maximum is found in the middle and upper stratosphere. As we discussed within our lecture on equatorial waves, the vertical wavelength of equatorial waves is sufficiently high so as to penetrate into the stratosphere. Thus, we see equatorially trapped Kelvin (mixed Rossby-gravity) waves transporting westerly (easterly) momentum upward.

A given jet maximum acts as a sort of barrier upon the vertical propagation of Kelvin and mixed Rossby-gravity wave momentum forcing. In our initial state, with an easterly zonal wind maximum above a westerly zonal wind maximum, easterly momentum propagates into the middle stratosphere whereas westerly momentum only propagates into the lower reaches of the lower stratosphere. As the westerly jet narrows in vertical extent due to the deposition of easterly momentum above it, the jet becomes unstable. Diffusion (i.e., mixing with the easterly momentum) then occurs, weakening and ultimately eroding the westerly jet. This leads to panel (b), where the stratosphere is dominated by easterlies. The dissipation of the jet in the lower stratosphere is critical to achieve an oscillatory state; without it, a nearly steady solution is obtained.

The removal of the westerly jet creates a situation by which equatorially trapped Kelvin waves are free to deposit westerly momentum through the stratosphere to the stratopause. This leads to a westerly jet that gradually intensifies and broadens in depth through the stratosphere. As this occurs, the easterly jet narrows in vertical depth as it is forced lower, ultimately leading to the inverse of the situation depicted in panel (a) (as seen in panel (d)). The easterly jet becomes unstable and diffusion/mixing again leads to the erosion of the zonal wind maximum. This enables easterly momentum associated with the mixed Rossby-gravity wave to penetrate through the stratosphere and be deposited near the stratopause. The cycle continues through panels (e) and (f) until a situation like panel (a) is obtained again.

QBO Impacts

The QBO influences tropical and mid-latitude weather patterns in both the troposphere and stratosphere. On a most basic level, the QBO impacts the transport of particulates such as volcanic dust within the stratosphere. The decay rate of such particulates depends on the phase of the QBO. On a more esoteric level, major stratospheric warming events preferentially occur during the easterly QBO phase. This is believed by some to be important for long-range forecasts of the polar vortex and mid-latitude tropospheric weather phenomena during the local winter months, although this remains an area in which further research is needed.

Tropical cyclone frequency in the Atlantic and western North Pacific basins increases during westerly QBO phases. Conversely, it increases during easterly QBO phases in the southwestern Indian Ocean. Westerly QBO phases are thought to reduce upper-tropospheric zonal wind shear in the Atlantic and western North Pacific, promoting enhanced upper-tropospheric divergence and, from mass continuity, deep-layer ascent that leads to deep, moist convection. This modulation of deep, moist convection is thought to also modulate rainfall across the Sahel region of western Africa.

Other Interseasonal Modes of Oscillation

The Atlantic Meridional Mode (AMM)

The AMM is the dominant source of coupled ocean-atmosphere variability in the Atlantic Ocean. The AAM is an interannual to decadal oscillation. The index describing the AAM is a function of a maximum covariance analysis of the anomalous SST, 10-m wind, and precipitation fields. However, the underlying physics and dynamics of the AAM remain a mystery.

Positive AMM phases are characterized by above-normal sea surface temperatures (SSTs) over the tropical North Atlantic and below-normal SSTs over the tropical South Atlantic. Negative AMM phases are characterized by the inverse of these conditions. Surface air pressure responds to these SST anomalies, as can be illustrated via thickness arguments, leading to below- (above-) normal surface pressure where above- (below-) normal SSTs are found. Accompanying surface wind anomalies flow from the anomalously colder to the anomalously warmer hemisphere. During positive AMM phases, this strengthens (weakens) the southeasterly (northeasterly) South (North) Atlantic trade winds, enhancing (reducing) upwelling and wind-induced evaporative cooling of the ocean in such locales. Further, positive AMM phases are associated with a northward displacement of the ITCZ in response to changes within the SST field. This often leads to drought over northern Brazil. The above-normal SSTs (along with reduced vertical wind shear that often accompanies positive AAM phases) lead to enhanced tropical cyclone activity across the North Atlantic. The opposite effects are observed with negative AMM phases.

The Atlantic Multidecadal Oscillation (AMO)

The AMO is a decadal variability of SST in the North Atlantic Ocean. Cool and warm phases of the AMO last for 20-40 years and are characterized by large-scale SST anomalies on the order of ±0.5°C. The AMO index itself is comprised of the running mean of the de-trended, area-averaged North Atlantic SST anomaly. Positive/warm (negative/cool) phases denote above-average (below-average) SSTs across the basin. It should be noted that whereas the AAM describes variability in the SSTs between the tropical North and South Atlantic Ocean, the AMO describes variability in the SSTs over the entire North Atlantic

Ocean. Current understanding into the dynamics of the AMO links its oscillatory nature to subtle changes in the overturning south-to-north oceanic circulation within the Atlantic Ocean.

The AMO influences tropical, subtropical, and mid-latitude weather patterns. Warm AMO phases are associated with enhanced numbers of major (Saffir-Simpson Scale Category 3+) hurricanes in the Atlantic. The frequency of occurrence of tropical storms and weaker hurricanes is not significantly impacted, however, by the AMO. Warm AMO phases are also associated with reduced rainfall across much of the United States and northeastern South America. Enhanced rainfall is noted across southern Alaska, northern Europe, western African, and the southeastern United States. Persistent Midwestern United States droughts, such as the 1930s Dust Bowl and the ongoing drought, have occurred during warm AMO phases. These impacts arise due to subtle shifts in the locations of subtropical ridges and upper tropospheric jet streams in response to oceanic forcing associated with the AMO.

For Further Reading

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