Monsoons

Introduction

Of all tropical phenomena, the rainfall (or, sometimes, the lack thereof) associated with monsoons has the greatest impact on the greatest number of people. In particular, the Asian monsoon, the most well-known and well-studied of all monsoons, directly or indirectly impacts nearly half of the world's population, much of which lives in third-world nations such as India and Indonesia. Excessive rains with strong monsoons can bring flooding and loss of life and property; the lack of rains with weak monsoons can result in reduced yields for the crops that are vital to providing both food and income to the masses. Independent of this societal significance, the monsoon is a complex meteorological phenomenon that is impacted by phenomena on multiple temporal and spatial scales. In the accompanying lectures, we discuss the monsoon in great detail, explaining why it is structured and how it evolves as it does. In so doing, we examine the structure and characteristics of the Asian and other monsoons of the world.

Key Concepts

- What is a monsoon?
- Where do monsoons typically occur?
- Why are they important to both the climatology of the tropics and billions of people worldwide?
- What drives a monsoon and how is this impacted by phenomena on intraseasonal to interseasonal timescales?

Introduction to Monsoons

To first order, the term *monsoon* refers to the seasonal reversal of the prevailing surface winds over southern Asia and the Indian Ocean. Accompanying these shifts in the prevailing surface winds are modulations in rainfall activity: convergent onshore flow leads to enhanced precipitation during local summer while divergent offshore flow leads to reduced precipitation during local winter. There is significant underlying structure to this system, however, as well as substantial variability in this structure on intraseasonal to interseasonal time scales. Further, monsoons are not exclusively present over southern Asia and the Indian Ocean; they also occur, in some form, over Australia and the Maritime Continent, Africa, and the Americas.

A monsoon is characterized by convergent, cyclonic lower-tropospheric flow and divergent, anticyclonic upper-tropospheric flow in the summer hemisphere. The pattern is reversed in the winter hemisphere. Cross-equatorial lower-tropospheric flow is oriented from the winter to the summer hemisphere whereas it is oriented from the summer to the winter hemisphere in the upper troposphere. Ascent (descent) is promoted in the vicinity of the convergent cyclonic (divergent anticyclonic) lower-tropospheric flow; thus, monsoons are associated with enhanced (reduced) rainfall in the summer (winter) hemisphere and thus contribute strongly to the annual climatology of the tropics in regions where monsoon circulations are present. Each monsoon system has unique structural characteristics that will be described in greater detail shortly. The rainy periods of monsoons are typically associated with "active" and "break" periods that modulate their effects on intraseasonal time scales.

As with many other phenomena of the tropics, heating (or, more specifically, spatially variable heating) drives the monsoon. We now shift into describing both simple and complex models for the evolution and structure of the monsoon, focusing in particular on describing the Asian/Indian Ocean monsoon. Subsequently, each of the monsoons, their relevant characteristics, related definitions, and variability inherent to each will be described.

Models of Monsoon Evolution and Structure

Simple Conceptual Model: Thickness Arguments

To first order, monsoon evolution can be conceptualized in a manner similar to that for the Hadley circulation. In the following, we will be Northern Hemisphere-centric; however, note that similar arguments may be made for a Southern Hemisphere-centric monsoon view. Let there be a landmass to the north (~10-20° latitude) of the Equator and a body of water along and to the south of the Equator. The seasonal cycle dictates that as you move from local spring to local summer, the sun angle becomes more (less) direct to the north (south). This promotes greater (lesser) insolation (incident at the top of the atmosphere, at least) to the north (south). Recall that land requires less heat energy to warm or cool by a given amount (e.g., 1°C) than water. Furthermore, recall that specific land types require more or less heat energy to warm or cool by this given amount than do other land types.

The land-sea contrast and seasonal cycle combine to establish a north-south gradient of temperature and heating, wherein warmer (cooler) temperatures are found over land (water) in summer. From thickness arguments, this leads to greater (lower) thicknesses over land (water) that results in low pressure in the lower troposphere and high pressure in the upper troposphere over land. Conversely, over water, the pressure anomalies are of opposite sign, with high pressure in the lower troposphere and low pressure in the upper troposphere. Flow down the pressure gradient, or from high to low pressure, results in lower tropospheric cross-equatorial flow from water to land (or from the winter to the summer hemisphere). It also results in lower-tropospheric divergence, descent, and reduced precipitation over water and lower-tropospheric convergence, ascent, and enhanced precipitation (especially as associated with moisture transport from the ocean) over land. As the seasons progress from summer to fall and beyond, this north-south/land-sea heating contrast weakens and eventually reverses, weakening and terminating the monsoon.

In the above, we have neglected the effect of the Coriolis force on the winds in both hemispheres. Incorporating this portion of the force balance to our conceptual model results in a deflection of the winds, such as is depicted within the lecture materials, resulting in curved flow in association with the monsoon. Further modifications to this basic structure arise in response to deep, moist convection and its impacts on both local and large-scale heating (e.g., such as highlighted in our earlier discussions of the Yanai heat source and moisture sink). From this model, one can infer that the monsoon manifests itself as the summer hemisphere ascending and winter hemisphere descending branch of the “equatorially displaced” Hadley circulation.

More Complex Conceptual Model: Shallow-Water Equations

As with the Walker circulation, we can turn to the shallow-water equation set of Gill (1980) to describe how the monsoon arises as a function of heating. In the Walker circulation example, we assumed that heating was symmetric along/about the Equator. This led to a solution featuring the superposition of an equatorially trapped Kelvin wave with an $n = 1$ equatorial Rossby wave. The zonally integrated solution to this system was used to obtain a Hadley-like meridional circulation while the meridionally integrated solution to this system was used to obtain the zonal Walker circulation. In this framework, the Hadley-like circulation was approximately 20% as strong as the Walker circulation.

For the monsoon, we instead prescribe heating that is asymmetric along/about the Equator. We proceed from a Northern Hemisphere-centric perspective, prescribing heating over a limited zonal area located north of the Equator. As with the symmetric heating case, the solution to the asymmetric heating case features the superposition of two equatorial waves. The first of these waves is an $n = 0$ mixed Rossby-gravity wave, the structure of which we discussed in our lectures on equatorial waves. The second of these waves is an $n = 2$ equatorial Rossby wave, the structure of which is depicted within the lecture materials. The $n = 2$ equatorial Rossby wave has similar pressure anomaly patterns about the Equator as does the $n = 0$ mixed Rossby-gravity wave; however, the nature of the horizontal flow is different, particularly near the Equator. With the $n = 2$ equatorial Rossby wave, there exists westerly flow immediately north of the Equator for lower pressures in the northern hemisphere. Conversely, there exists easterly flow immediately south of the Equator for higher pressures in the southern hemisphere. The inverse of these flow configurations exists if the sign of the pressure anomalies is flipped.

The superposition of these two equatorial wave solutions for the lower troposphere during Northern Hemisphere summer is given in the lecture materials. Again, recall that these solutions are quasi-steady, wherein we assumed a steady heat forcing ($-Q$) offset by Newtonian cooling and Rayleigh friction (ε) so as to highlight the mature structure of the system. In reality, we know that a given system is never truly steady in nature; in the example of the monsoon, the forcing varies as a function of the seasons as well as multiple modes of intraannual and interannual variability. This solution is characterized by low (high) pressure in the lower troposphere north (south) of the Equator. Ascent (descent) is maximized southeast (northeast) of the pressure minimum (maximum). Said ascent is located near the location of the heating source. It is thus southeast of the lower-tropospheric pressure minimum where deep, moist convection would be expected to preferentially form. Indeed, it is here and with the convergent lower-tropospheric cyclonic circulation itself that deep, moist convection is favored with monsoon circulations. The extension of this solution to southern hemisphere summer simply requires a mirroring (or flipping) of the solution about the Equator.

The resultant zonally integrated solution is also given within the lecture materials. Note that the meridionally integrated solution is generated entirely from the symmetric part of the forcing as the anomalies associated with the asymmetrically forced solution cancel out in the meridional integration process. Lower-tropospheric zonal flow near the Equator is westerly in the heated Northern Hemisphere. It is easterly near the Equator in the Southern Hemisphere. This antisymmetry about the equator is different than that observed with the zonally integrated solution for the symmetric heating case. The upper-tropospheric flow is of the opposite sign/direction to that in the lower troposphere in both hemispheres. Cross-equatorial flow is from the Southern to the Northern Hemisphere in the lower troposphere and from the Northern to the Southern Hemisphere in the upper troposphere. In both cases, this characterizes flow down the pressure gradient toward lower pressures. Given greater latent heat release aloft in deep, moist convection, it also describes a circulation that exports heat from the warmer summer hemisphere to the cooler winter hemisphere. The net picture is of a Hadley-like circulation displaced northward off of the Equator toward the heating source. The strength of this circulation, as measured by the magnitude of the streamfunction describing it, is ≈60-65% of the zonal Walker circulation and better resembles observations.

Monsoons of the World

Asian Monsoon

The Asian monsoon, sometimes referred to as the Indian monsoon, helps to pull the intertropical convergence zone (ITCZ) northward during Northern Hemisphere summer. Note that the Asian monsoon and ITCZ are separate phenomena; rather, the ITCZ just happens to evolve with the Asian monsoon over India and southeast Asia. The salient characteristics of the Asian monsoon are depicted within the lecture materials. In the lower troposphere, these include:

1. The monsoon low over India and southeast Asia, associated with intense heating.

2. Onshore, cross-equatorial flow. This cross-equatorial flow is driven in large part by meridional temperature contrasts between a warm surface in the summer hemisphere and a cooler surface in the winter hemisphere. The thermal contrast forces a meridional pressure gradient; incorporating friction into the force balance results in cross-equatorial flow down the pressure gradient from high to low pressure (i.e., from where it is cool to where it is warm).

3. The Somali jet, arising in part by the channeling of the cross-equatorial flow by the high mountains of east Africa.

4. The Mascarene high in the southwestern Indian Ocean, associated with relatively cool surface conditions and large-scale subsidence.

In the upper troposphere, these include:

1. From mass continuity, an upper-tropospheric high above the monsoon low.

2. Updrafts in regions of deep, moist convection and lower tropospheric convergence, particularly across the eastern Indian Ocean, India, and western Maritime Continent.

3. The tropical easterly jet south of the upper-tropospheric high across the northern Indian Ocean. We will discuss the tropical easterly jet in more detail in a subsequent lecture.

The arrival of the Asian monsoon, or monsoon onset, is characterized by the arrival of persistent rains. Onset occurs first in the south, in the southern Bay of Bengal, in late April and progresses steadily northward toward Japan through early July. From June through August, monsoon onset also progresses eastward into the western North Pacific. This eastward extension is sometimes referred to as the East Asian and/or Western North Pacific monsoon. The monsoon ends, or retreats, with the seasonal cycle. Enhanced precipitation associated with the monsoon is favored in two locations: over strongly heated land masses (such as Indonesia and India) and over the warm eastern equatorial Indian Ocean. If the oceanic precipitation is particularly intense, such that it is associated with strong rising motion, the necessary compensating subsidence is often found over land, acting to suppress convection there.

To first order, monsoon strength is a function of the magnitude of the large-scale temperature gradient. In India, Southeast Asia, and the Indian Ocean, this temperature gradient is relatively strong and largely meridional in nature between the elevated heat source (Tibetan Plateau) and relatively cool South Indian Ocean. Over the Maritime Continent and western North Pacific Ocean, the monsoon is typically weaker owing to counteracting meridional (warmer Pacific ocean/cooler Australia) and zonal (warmer southeast Asia/cooler Pacific Ocean) temperature gradients. Here, mesoscale to synoptic-scale phenomena such as the monsoon trough and Mei-yu/Baiu front exert a significant control on the location of the heaviest precipitation associated with the monsoon. These phenomena will be discussed in more detail shortly.

As previously noted, monsoon evolution is significantly modulated by the seasonal cycle. It is also regulated by cross-equatorial oceanic transport. As discussed in an earlier lecture, the oceanic transport vector is influenced by the Coriolis force such that it points approximately 45-90° to the right (left) of the mean surface wind in the Northern (Southern) Hemisphere. During the Asian monsoon, Ekman transport results in southward transport of oceanic heat that acts to weaken the meridional sea surface temperature gradient across the Indian Ocean, thereby weakening the cross-equatorial monsoon flow and monsoon circulation as a whole.

Mei-yu/Baiu Front: The Mei-yu/Baiu front is a semi-permanent (during local summer), quasi-stationary weak frontal zone that extends from eastern China near 25°N east-northeastward into the North Pacific. This front first becomes established in mid-May and shifts northward through early to mid-summer before weakening. The front is the focal point for persistent heavy precipitation produced by episodic mesoscale convective systems (MCSs) that form over eastern China. Such MCSs preferentially form in this location given the presence of a lower-tropospheric jet that brings warm, moist air northward to the region from the South China Sea and Bay of Bengal. Isentropic ascent atop the Mei-yu/Baiu front, lower-tropospheric convergence on the nose of the lower-tropospheric jet, and upper-tropospheric divergence promote ascent through the troposphere that, in a moist, unstable environment, promotes deep, moist convection initiation, upscale growth, and organization. The majority of the rainfall falls on the warm (i.e., equatorward) side of the Mei-yu/Baiu front.

Monsoon Trough: Typically, the monsoon trough is oriented from west-northwest near the Philippines to east-southeast north of the Maritime Continent. It owes its origins to shear vorticity arguments: near-equatorial lower-tropospheric winds in the far western North Pacific are westerly in nature, as associated with the Asian monsoon, while lower-tropospheric winds in the subtropical western North Pacific are easterly in nature, as associated with the oceanic subtropical anticyclone and tropical trade winds. Cyclonic vertical vorticity is maximized on the easternmost flank of the monsoon trough where confluence is maximized between the Asian monsoon and easterly trades of the tropical and subtropical western North Pacific Ocean. Owing to the presence of enhanced cyclonic vertical vorticity in a moisture-rich environment, tropical cyclones commonly form on this easternmost flank of the monsoon trough. Intraseasonal variability in the orientation, strength, and structure of the monsoon trough all may modulate its impacts on precipitation and tropical cyclone activity.

Australian/Maritime Continent Monsoon

The Australian/Maritime Continent monsoon is the seasonal opposite of the Asian monsoon. Monsoon onset first occurs over Malaysia in late August and progresses south and east through local spring and summer, reaching its southernmost extent over northern Australia in early February. This monsoon is driven by temperature contrasts between strong heating over the Maritime Continent and deserts of northern Australia and cooling over Asia and the northwestern Pacific Ocean. As before, this meridional temperature contrast drives cross-equatorial flow, here from the Northern to the Southern Hemisphere. It should be noted, however, that the influence of Coriolis upon the monsoon circulation is weaker than for the Asian monsoon given the close proximity of the Maritime Continent and northern Australia to the Equator. Despite this, the Coriolis force is sufficiently strong so as to promote the northward transport of oceanic heat that acts to weaken the meridional sea surface temperature gradient between the Maritime Continent and western North Pacific Ocean, thereby weakening the cross-equatorial monsoon flow and monsoon circulation as a whole over time.

West African Monsoon

The West African monsoon (WAM) is characterized by rainy periods in both the Northern and Southern Hemisphere across interior Africa. However, given Africa's proximity to the strong monsoons of Asia and Australia/Maritime Continent, the WAM and its accompanying cross-equatorial flow is not as well-defined. Precipitation is enhanced across interior Africa where warm, dry air from the Sahara intersects relatively moist air from the south (e.g., South Atlantic Ocean). During Northern Hemisphere summer, this enhanced precipitation is found along the intertropical convergence zone, or near 10-15°N, whereas during Northern Hemisphere winter, this enhanced precipitation is found across south-central Africa. Thus, in some regards, the WAM can be viewed as an enhancement to the ITCZ across Africa during Northern and Southern Hemisphere summer. Features such as the African easterly jet, African easterly waves, and Saharan air layer are often found in conjunction with the WAM during Northern Hemisphere summer.

North and South American Monsoons

Weak, non-classical monsoons exist in both North and South America. The North American monsoon (NAM) is driven by heating over the deserts of western Mexico and the southwestern United States. Such heating establishes what is called a heat low at the surface. Convergent flow into the heat low advects in moist, tropical air from the Gulf of California and tropical eastern Pacific Ocean. Above the boundary layer, moisture is also advected into the region from the Gulf of Mexico. Monsoon onset is typically between mid-June and mid-July and lasts for two to three months until September. The extent of the monsoon is typically defined by the area over which the majority of the annual precipitation (35-40% to 60+%) is associated with the NAM. Conversely, the South American monsoon (SAM) is driven by heating over the moist interior of South America. The tropical Atlantic Ocean and Amazon River basin form the primary moisture sources for enhanced precipitation in association with the SAM. Monsoon onset first occurs in August near the equator in northwestern South America and progresses southeastward with time through to Brazil in September and October.

Monsoon Variability

Extreme variability in the strength of the monsoon and its accompanying rainfall can lead to devastating floods and droughts across some of the most vulnerable third-world regions of the world. Such variability is found on both interannual (1-10 yr) and intraseasonal (3 day-1 mo) time scales. In this section, we discuss some of the leading modes of such variability, focusing most closely on variability within the Asian, Australian, and West African monsoons.

On interannual time scales, ENSO exerts a strong influence on the Asian and Australian monsoons. During El Niño events, with the ascending branch of the Walker circulation shifted eastward into the central Pacific Ocean, rainfall tends to be reduced in association with the Asian monsoon. During La Niña events, with the ascending branch of the Walker circulation strengthened in the western Pacific Ocean, rainfall tends to be enhanced in association with the Asian monsoon. The same general pattern holds for the Australian monsoon: enhanced rains tend to be associated with La Niña events while reduced rains tend to be associated with El Niño events. Note the use of the word "tends" in the preceding sentences. Not all El Niño or La Niña events will necessarily be associated with reduced or enhanced rainfall, respectively; other modes of variability also exert a significant control on rainfall and thus monsoon strength.

The most important influence on the intensity of the West African monsoon is the thermal contrast between the Gulf of Guinea (in the tropical Atlantic) and the heated North African continent. Warmer waters in the tropical Atlantic and Gulf of Guinea (e.g., as associated with a negative Atlantic Meridional Mode – AMM – phase) reduce the magnitude of the land-sea temperature gradient, thus acting to reduce the onshore flow into and subsequent precipitation across western Africa. (Recall that the negative AMM phase has enhanced precipitation across the near-equatorial South Atlantic, consistent with its modulation of the monsoon over land.) In positive AMM phases, the magnitude of the land-sea temperature gradient is enhanced, thereby enhancing both onshore flow and precipitation across western Africa.

On annual time scales, variability in upper oceanic temperatures across the Indian Ocean exerts a strong influence on the Asian and Australian monsoons. This variability, manifest as a dipole in sea surface temperature, is commonly referred to as the Indian Ocean Dipole (IOD). The IOD has a typical period of approximately one year with a peak magnitude in the late fall to early winter. When warmer waters are found in the western Indian Ocean, heavy rains tend to be located from eastern Africa toward India. When warmer waters are found in the eastern Indian Ocean, heavy rains tend to be located across southeast Asia and Indonesia. These impacts follow from the forcing exerted by sea surface temperatures upon the near-surface kinematic and mass fields (e.g., through sensible heating of the overlying air), promoting lower pressure, convergence, and deep-layer ascent over regions of warmer sea surface temperatures.

Intraseasonal variability in the monsoon is often manifest via active and break periods of the monsoon. Active periods are associated with an enhanced monsoon circulation whereas break periods as associated with a weakened monsoon circulation. For the Asian monsoon, active periods are characterized by above-normal rainfall over land whereas break periods are characterized by above normal rainfall over the near-equatorial oceans. Below-normal rainfall during active and break periods are found over the near-equatorial waters and over land, respectively.

The MJO is the leading mode of intraseasonal variability in the Asian and Australian monsoons. When the active or westerly phase of the MJO is found across the Indian Ocean or Maritime Continent (e.g., phases 2-5), monsoon onset and/or an active monsoon are promoted. When the inactive or easterly phase of the MJO is found across the Indian Ocean or Maritime Continent (e.g., phases 1 and 6-8), monsoon onset is suppressed and a break in the monsoon may be promoted. Since the MJO is typically strongest during Southern Hemisphere summer, the impact of the MJO is somewhat greater upon the Australian monsoon than upon the Asian monsoon. The impact of the MJO on these monsoons tends to vary annually as a function of the strength of the MJO itself. Modest impacts of the MJO, as manifest via equatorial wave excitation and propagation, may also be observed with the West African monsoon. Indeed, owing to the nature of the monsoon circulation as a superposition of multiple equatorial wave modes, it stands to follow that the excitation and subsequent evolution of equatorial waves may exert a strong influence upon monsoon precipitation on the time scale of an individual wave.

For Further Reading

- Chapter 3, *An Introduction to Tropical Meteorology, 2nd Edition*, A. Laing and J.-L. Evans, 2016.
- Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447-462.
- Matsuno, T., 1966: Quasi-geostrophic motions in the equatorial area. *J. Meteor. Soc. Japan*, **44**, 25–43.