Tropical Climatology

Introduction

To first order, tropical meteorology can be defined by its climatology. In other words, tropical weather changes little between days, weeks, and in some cases even seasons. Thus, tropical weather conditions are often nearly identical to the tropical climatology, which is defined as the long-term average of these weather conditions. The same is not true at higher latitudes, however. Here, the weather is ever-changing thanks to transient features like Rossby waves, which characterize the large-scale trough-ridge pattern. These features dampen (or smear) when you average them over many years, such that higher-latitude weather is often quite different from its climatology. That aside, what does tropical climatology look like? And, what causes it to have its structure?

Key Concepts

- How can the tropics be defined? What are the strengths and weaknesses of these definitions?
- What are the predominant climate classifications within the tropics, and how do they differ from each other? How are these related to annual variability in tropical circulations?
- What are the mean structures of tropical vertical profiles of temperature, moisture, and wind? What is the nature of the variability in these profiles?
- What are the predominant diabatic processes present within the tropics?
- What are heat sources and moisture sinks? How are they related to tropical circulations?

Defining the Tropics

There is no widely accepted definition for the tropics. Most definitions, however, have some connection to tropical climatology. Below are several viable definitions for the tropics:

- The range of latitudes where the sun can be directly overhead. This encompasses latitudes between the Tropic of Capricorn (23.5°S) and Tropic of Cancer (23.5°N).
- The range of latitudes where the net annual incoming (or shortwave) solar radiation is greater than the net annual outgoing (or longwave) terrestrial radiation. This encompasses latitudes between ±35-40°N/S.
- The range of latitudes at which the Coriolis force, horizontal temperature differences, and horizontal pressure differences are relatively small or weak compared to middle and higher latitudes.
- The range of latitudes characterized by a tropical climate, with continuously warm temperatures all year and either a single, year-round wet season or separate wet and dry seasons.
- The range of latitudes characterized by tropospheric ascent, easterly boundary-layer (or low-level) flow, and lower surface pressures associated with a meridional circulation (the Hadley cell) that is driven by latitudinal variations in radiation. On average, this encompasses latitudes between ±30°N/S.

The ordering of these definitions is important. The first two relate to how the Earth's tilt changes throughout the year. The tropics are minimally affected by this changing tilt, however. This defines the climatological thermodynamic and kinematics characteristics of the tropics, which are given by the final three definitions.

Climatological Characteristics of the Tropics

The tropics are dominated by moist tropical (mT) air masses, classified with a preceding "A" in the Koppen climatological classification system. Locations adjacent to the equator are represented by the *Af* (tropical rainforest) and *Am* (tropical monsoon) Koppen classifications. Poleward of the equatorial latitudes, the *Aw* (tropical wet/dry, indicating distinct wet and dry seasons) Koppen classification is the predominant climate classification and is associated with the annual migration of the Hadley cell ascending branch. Portions of the tropical Americas and eastern tropical Africa are represented by the *H* (highland) Koppen classification owing to the presence of significant elevated terrain.

Three characteristics modulate the distribution of surface temperature within the tropics: net incoming solar radiation, the underlying surface type, and altitude above sea level. Areas near the equator receive greater annual net solar radiation than areas further from the equator. In the absence of other physical processes to compensate for this differential, annual mean surface temperatures are generally warmest near the equator. In addition, minimal variation in the solar declination angle (and thus the number of daylight hours per day) leads to surface temperatures near the equator changing little throughout the year.

Similarly, surface characteristics also modulate tropical surface temperatures. For instance, the annual mean surface temperatures over water are less variable than over land because water requires a greater amount of energy input to warm it by an equivalent amount (often considered to be 1°C, since that is how the *specific heat capacity* is defined). Furthermore, as seen in the climate classifications, lands closer to the equator are generally moister than those closer to the subtropics. Drier soils generally require less energy input to warm them by a specified amount than do their moister counterparts. Conversely, drier soils generally cool more rapidly at night and in the winter months than their moister counterparts. This contributes to both the diurnal (day vs. night) and annual (summer vs. winter) ranges of temperature over land being larger than over water at a given latitude. Finally, annual mean surface temperatures are higher near sea level and lower at higher altitudes.

The mean lower- and upper-tropospheric flow in the tropics is dominated by several circulatory phenomena. These include the subtropical anticyclones, found over land in the upper troposphere and over water in the lower troposphere, which are strongest during the local summer months; the Asian, Australian, and African monsoons; the tropical easterly jet associated with the Asian monsoon across the Indian Ocean and northern Africa; mid-oceanic upper-tropospheric troughs during local the spring and summer months; heat lows over arid land masses in the summer; and easterly near-surface trade winds contributing to the near-equatorial convergence zones.

Zonally averaged vertical motion within the tropics varies with the annual cycle of incoming solar radiation. Mean ascent is typically located between 5°S and 10°N, favoring the summer and fall hemispheres. Mean descent is typically located in the subtropical latitudes and is favored in the winter and spring hemispheres. April/May and October/November are the transition months and feature mean descent in the subtropics in both hemispheres and mean ascent slightly north of the equator. Locations predominantly characterized by mean ascent experience greater average annual rainfall (*Af* and *Am* classifications), whereas locations that alternate between mean ascent and mean descent during the year have distinct wet and dry seasons (the *Aw* classification).

Vertical Structure of the Mean Tropical Troposphere

Average tropical vertical sounding profiles are characterized by an atmospheric boundary layer of varying depth, the free troposphere above it, and the tropopause at an altitude of 15-18 km. Boundary layer depth is lowest (*O(100 m)*) over the open oceans where turbulent vertical mixing is limited and is at a maximum (*O(5 km)*) over relatively hot, dry surfaces where turbulent vertical mixing is maximized. Oftentimes, the boundary layer and free troposphere are separated by a temperature inversion. Such a temperature inversion is most prominent with greater distance from the equator and is often associated with subsidence associated with the Hadley cell's descending branch and the accompanying subtropical anticyclones.

Utilizing a ten-year composite of soundings from the West Indies (e.g., the Lesser Antilles), Jordan (1958) developed a climatological vertical sounding of the tropical North Atlantic and Caribbean Sea. As noted by Dunion (2011), this sounding has been extensively used as a reference for tropical soundings during the North Atlantic hurricane season and as an initial background state in numerous idealized model simulations. This mean sounding is characterized by relatively high moisture content and relative humidity within the boundary layer that decays at an approximately linear rate in the free troposphere. Temperature decreases at a rate approximately equal to or slightly greater than the moist adiabatic lapse rate. It is moderately unstable,

The mean vertical structure of the tropical North Atlantic and Caribbean Sea (Jordan 1958; Dunion 2011) is characterized by relatively high boundary-layer moisture content and relative humidity, both of which decrease approximately linearly with increasing altitude. Temperature decreases with increasing altitude at a rate approximately equal to or slightly greater than the moist-adiabatic lapse rate. The vertical profile is moderately unstable, with approximately 1700 J kg^{-1} of mixed-layer convective available potential energy and 25 J kg^{-1} of mixed-layer convective inhibition. Winds are predominantly easterly below 400-300 hPa and are maximized at the top of the boundary layer. This mean structure represents contributions from three distinct air masses: a moist tropical (MT; 66%) air mass, representative of the maritime (oceanic) tropics; a Saharan air layer (SAL; 20%) air mass, characterized by an elevated mixed layer that is initially generated over the Sahara desert and transported westward across the tropical North Atlantic; and a mid-latitude dry (MLD; 14%) air mass, representing periodic midtropospheric dry-air intrusions from midlatitudes into the tropics by features like shortwave and longwave troughs. SAL air masses are most common in early summer whereas MLD air masses are most common in spring and fall. Winds atop the boundary layer are strongest in SAL air masses owing to an easterly mid-tropospheric jet that often accompanies the SAL from Africa.

Diabatic Heating and Tropical Circulations

As we will demonstrate throughout the semester, diabatic heating drives tropical circulations. Before doing so, however, let's define what is meant by diabatic heating and the predominant forms it takes in the tropics.

Diabatic processes are those in which a substance such as an air parcel or water molecule exchanges energy with its surroundings because of a temperature difference between them. Diabatic heating characterizes the energy gained (warming) or lost (cooling) because of the diabatic process. We typically measure this energy exchange in terms of *specific enthalpy*, or the enthalpy per unit mass. Neglecting solid-phase water, specific enthalpy can be written as:

(1) $$h = c_p T + q_v L_v + constant$$

where c_p is the specific heat of the air mixture (dry air + water vapor + liquid water) at a constant pressure, T is temperature, q_v is the water-vapor mixing ratio (defined as the mass of water vapor divided by the total air mass), and L_v is the latent heat of vaporization. A similar equation with additional terms can be obtained if solid-phase water is not neglected.

There are three major types of diabatic heating of interest in the tropics:

- ***Latent heating***, characterizing the change in enthalpy that a water substance experiences when it changes phases (e.g., solid to liquid, or vice versa) at a constant temperature. This can occur at the surface, primarily in the form of surface evaporation as subsaturated air passes over water or moist ground, or within the atmosphere, where it can result from evaporation, condensation, sublimation, deposition, melting, or freezing.
- ***Sensible heating***, characterizing the change in enthalpy that air experiences when it encounters another substance (air, water, land, etc.) that has a different temperature. This predominantly occurs at the surface (e.g., land-atmosphere or water-atmosphere sensible heating).
- ***Radiative heating***, characterizing the change in energy or temperature due to the net absorption (warming) or emission (cooling) of shortwave (radiation emitted by the sun) and/or longwave (terrestrial radiation, or that emitted by the Earth and its atmosphere) radiation.

For example, cold, dry air passing over the relatively warm Lake Michigan waters in winter is characterized by two diabatic processes: *surface sensible heating*, whereby the near-surface air warms as energy is gained by the colder air from the warmer water, and *surface latent heating*, whereby the near-surface air moistens due to the evaporation of water from the underlying lake surface. Of greater relevance to this course, tropical cyclones are fueled by surface enthalpy exchange, which is characterized by both sensible (slightly warmer ocean than air) and latent (surface evaporation) heating. Radiative heating significantly influences night vs. day and winter vs. summer temperatures, with clouds constraining how much incoming shortwave radiation can be transmitted toward the surface and how much outgoing longwave radiation can be lost to space.

We can characterize these diabatic processes using the *apparent heat source* Q_1 and *apparent moisture sink* Q_2 (Yanai et al. 1973; Yanai and Tomita 1998). The vertically integrated apparent heat source $<Q_1>$ is given by:

(2) $$< Q_1 > = < Q_R > + L_v P + S$$

where $<Q_R>$ is the vertically integrated radiative heating rate (positive for warming), P is the precipitation rate (condensation – evaporation, with net condensation assumed to fall out as precipitation; this is a latent-heating metric), and S is the vertical surface sensible heat flux.

Likewise, the vertically integrated apparent moisture sink $<Q_2>$ is given by:

(3) $$< Q_2 > = L_v (P - E)$$

where E is the surface evaporation rate per unit area, commonly referred known as the vertical surface latent heat flux. The apparent moisture sink, which is positive for moisture reductions (as expected given the *sink* nomenclature), is governed by the balance between precipitation (moisture loss) and evaporation (moisture gain). A full derivation of (2) and (3) is provided in Appendix A.

Seasonal variations in $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ are presented in the accompanying lecture materials. During boreal (Northern Hemisphere) winter, $\langle Q_1 \rangle$ is large in the tropics and along the midlatitude north Pacific and north Atlantic storm tracks. In the tropics, $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ are both large, implying that latent heat release due to condensation is the major contributor to this warming. Because outgoing longwave radiation is small, much of this latent heat release can be attributed to thunderstorms (deep cumulus clouds), which prevent outgoing longwave radiation. Over the midlatitude storm tracks, while $\langle Q_1 \rangle$ remains large from west to east, $\langle Q_2 \rangle$ is negative to the west and positive to the east. This implies that surface sensible heating dominates early on in a storm's lifecycle but that latent heating becomes more important at later times. Conversely, $\langle Q_1 \rangle$ is negative in boreal winter over Northern Hemisphere continents and eastern portions of the subtropical north Pacific and north Atlantic oceans. Large negative values of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ indicate that radiative cooling (negative Q_R) exceeds sensible and latent heating, and that evaporation exceeds precipitation, in these areas.

During boreal (Northern Hemisphere) summer, the predominant heat sources (positive $\langle Q_1 \rangle$) in the tropics are located with the Asian monsoon, in the tropical western Pacific, and near Central America. These areas are characterized by large positive values of both $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ collocated with low values of outgoing longwave radiation. Thus, diabatic warming due to latent heat release in thunderstorms is the primary heat source in these regions during the local summer months. Heat sinks (negative $\langle Q_1 \rangle$) are located over much of the subtropical Southern Hemisphere as well as over the eastern north Pacific and north Atlantic Oceans. Large outgoing longwave radiation implies that radiative cooling (negative Q_R) is the primary cause of the diabatic cooling in these areas.

If $\langle Q_1 \rangle$ is applied as a forcing term to a simplified form of the equations that govern atmospheric motions, we can obtain a representation of the atmospheric flow resulting from such heating. $\langle Q_1 \rangle$ for January and July 1989 is presented in the accompanying lecture materials (Zhang and Krishnamurti 1996). These figures qualitatively resemble the Yanai and Tomita (1998) climatologies described above. These data are used to drive a diagnostic model based on a simplified form of the equations that govern atmospheric motions. This diagnostic model reasonably depicts the mean tropical circulation in both January and July. The subtropical highs of the Atlantic and Pacific oceans; the Mascarene high over the southwestern Indian Ocean; the heat lows over the Sahara Desert, Saudi Arabia, and southwestern United States; the monsoon trough over India and accompanying cross-equatorial monsoon current over the Arabian Sea; the tropical trade winds; and the tropical convergence zones are all well-captured by the diagnostic model. The greatest departures from observations are noted at midlatitudes, where it is hypothesized that processes unresolved by the diagnostic model may influence the general circulation. Thus, we are led to conclude that vertically integrated diabatic heating – or, more specifically, spatial variations therein (sometimes referred to as *differential heating*) – is the process that drives much of the observed circulation in the tropics.

How do tropical circulations result from diabatic heating? Assuming that horizontal temperature gradients in the tropics are weak – the so-called *weak temperature gradient approximation* – Sobel et al. (2001) show that the diabatic heating and horizontal divergence are directly proportional to each other. Diabatic heating forces horizontal divergence, which itself forces vertical motions (through the continuity equation), in turn impacting rotational flows (cyclones and anticyclones), with minimal changes to the horizontal temperature gradient (which is assumed to remain weak). We will examine these properties further in later sections.

Vertical Heating Profiles

In the foregoing discussion on heat sources and moisture sinks, we largely considered vertically integrated fields of these quantities and their respective forcing terms. However, the vertical structure of heating within the column can provide important information that, in the absence of other meteorological data, can enable us to infer the meteorological phenomenon or phenomena responsible for said heating.

The two most common vertical heating profiles are *stratiform* and *convective* heating profiles. Examples of both are presented in the lecture materials. The convective heating profile is characterized by strong diabatic warming throughout the depth of the troposphere with a maximum in the mid- to upper troposphere (where ascent is also maximized). The stratiform heating profile is characterized by diabatic warming maximized over a shallow layer containing stratiform clouds in the lower- to midtroposphere. Diabatic cooling is found above and below this layer and is characterized by radiative cooling above and evaporative cooling below. Stratiform precipitation and heating profiles are typically associated with decaying thunderstorms both in and beyond the tropics.

For Further Reading

- Chapter 1, *An Introduction to Tropical Meteorology, 2nd Edition*, A. Laing and J.-L. Evans, 2016.
- Dunion, J. P., 2011: Rewriting the climatology of the tropical North Atlantic and Caribbean Sea atmosphere. *J. Climate*, **24**, 893–908.
- Houze, Jr., R. A., 1997: Stratiform precipitation in regions of convection: a meteorological paradox? *Bull. Amer. Meteor. Soc.*, **78**, 2179–2196.
- Jordan, C. L., 1958: Mean soundings for the West Indies area. *J. Meteor.*, **15**, 91–97.
- Sobel, A. H., J. Nilsson, and L. M. Polvani, 2001: The weak temperature gradient approximation and balanced tropical moisture waves. *J. Atmos. Sci.*, **58**, 3650–3665.
- Yanai, M., and T. Tomita, 1998: Seasonal and interannual variability of atmospheric heat sources and moisture sinks as determined from NCEP-NCAR Reanalysis. *J. Climate*, **11**, 463–482.
- Yanai, M., S. Esbensen, and J.-H. Chu, 1973: Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *J. Atmos. Sci.*, **30**, 611–627.
- Zhang, Z., and T. N. Krishnamurti, 1996: A generalization of Gill's heat-induced tropical circulation. *J. Atmos. Sci.*, **53**, 1045–1052.