Climatology of the Tropics

Introduction

To first order, the meteorology of the tropics can be defined by its climatology. The same cannot be said about the middle and higher latitudes, where temporal averaging leads to the damping of the transient eddies (e.g., Rossby waves) responsible for much of the daily-to-weekly variability. Therefore, it is worthwhile to examine the climatology of the tropics before delving into the physics and dynamics of its salient features.

Key Concepts

- What are the climatological air masses and weather features of the tropics?
- What is the structure of the tropical mean vertical profile of temperature and moisture and what is the nature of the variability in this profile?
- What are heat sources and heat sinks, and how do they modulate the meteorology of the tropics?

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, signifying the presence of distinct wet and dry seasons) Koppen classification is the predominant climate classification within the tropics, as 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 at such locales. The physical basis behind the segregation of *Af/Am* and *Aw* climate zones is explored shortly.

Three characteristics modulate the distribution of surface temperature within the tropics: net incoming solar radiation, the underlying surface type, and altitude above sea level. As we will explore later in the semester, horizontal variations in heating play a large role in driving the circulation of the tropics, making it an important topic of consideration. Areas closer to the equator receive greater net annual solar radiation than those further away 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. Owing to the minimal variation in solar declination angle (and, subsequently, day length) at such latitudes, surface temperatures near the equator generally exhibit minimal variation throughout the year. Greater annual ranges in mean surface temperature are found at higher latitudes, where day length and solar declination angle are both more variable.

Similarly, surface characteristics also modulate tropical surface temperatures. For instance, annual mean surface temperatures over water are less variable than over land since water requires a greater amount of heat energy than land to warm it by an equivalent amount (e.g., 1°C). As seen in the climate classifications, land masses closer to the equator are moister whereas land masses closer to the subtropics tend to be drier. Drier soils and the soil types that accompany them generally require less heating to warm by a specified amount than do their moister counterparts. Conversely, however, they also generally cool more rapidly at night and in the winter months than their moister counterparts. This contributes to both the diurnal and annual ranges of temperature over land being larger than over water at a given latitude. Finally, as expected, annual mean surface temperatures are higher near sea level and lower at higher altitudes.

The mean lower- and upper-tropospheric flow within the tropics is dominated by the presence of several key circulatory phenomena. These include the subtropical anticyclones, found over land in the upper troposphere and over water in the lower troposphere and with greatest intensity during local summer; 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 spring and summer; the heat lows of the summertime arid land masses; easterly near-surface trade winds contributing to the near-equatorial convergence zones; and a monsoon depression circulation across the Asian continent during local summer. These features evolve with the annual cycle in response to local diabatic heating, as will be demonstrated in the "Heat Sources and Heat Sinks" section below.

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 immediately north of the Equator. The transitory nature of the regions of mean ascent and descent exerts a substantial influence on precipitation patterns across the tropics, particularly away from the equatorial latitudes. Thus, locations closer to the equator that experience mean ascent for a substantial portion of the year see greater average annual rainfall whereas locations closer to the subtropics have distinct wet (ascent) and dry (descent) seasons.

This can be seen from the monthly mean distributions of daily precipitation across the tropics. The intertropical convergence zone moves north and south with the region of zonally averaged ascent and is a focus for convection and precipitation. However, as noted above, near-equatorial regions tend to remain fairly wet throughout the year. The rainy monsoons of the Asian and Australian regions are established when large-scale descent is at a minimum, i.e., during local summer in each hemisphere. Subtropical precipitation is enhanced in the western north Pacific, eastern north Pacific, and western north Atlantic basins during local summer as a result of tropical cyclones influencing the local rainfall climatology. Such an enhancement is mitigated in the southern hemisphere during local summer owing to less frequent tropical cyclone activity there.

Vertical Structure of the Mean Tropical Troposphere

The mean vertical sounding profile in the tropics is characterized by a planetary boundary layer of varying depth, the free troposphere, and the tropopause at an altitude of 15-18 km. Boundary layer depth is at a minimum (*O(100 m)*) over the open oceans where mixing is limited and is at a maximum (*O(5 km)*) over relatively hot, dry surfaces where 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 descending branch of the Hadley cell meridional overturning circulation and the accompanying subtropical ridges of high pressure.

Utilizing a ten-year composite of available 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, with approximately 1700 J kg^{-1} of mixed-layer convective available potential energy and 25 J kg^{-1} of mixed-layer convective inhibition.

The mean vertical structure of the tropical North Atlantic and Caribbean Sea was re-examined by Dunion (2011). In a composite of approximately 6,000 July-October 1995-2002 soundings from the Caribbean Sea, the mean vertical structure of temperature and relative humidity closely resembles that of the Jordan (1958) sounding. In this composite, winds are generally easterly below the 300-400 hPa layer and are maximized near the top of the boundary layer. However, neither this sounding nor the Jordan composite sounding appropriately represent tropospheric vertical structure in this region at any given place and/or time. Rather, three distinct air masses influence the mean vertical structure of the tropical North Atlantic and Caribbean Sea: a moist tropical (MT) air mass, representative of much of the maritime tropics as a whole; a Saharan air layer (SAL) air mass, representative of the influences of northern Africa on the climatology of the tropical North Atlantic; and a mid-latitude dry (MLD) air mass, representative of periodic intrusions of dry air from the middle latitudes into the tropics by transient atmospheric eddies (e.g., shortwave and longwave troughs and associated cyclonic wave breaking events).

The MT air mass is present across the tropical North Atlantic approximately 66% of the time during the summer months. The SAL and MLD air masses are present approximately 20% and 14% of the time, respectively. SAL air masses are most common early in summer while MLD air masses are most common at the outset and conclusion to the local hurricane season. Neither the SAL nor the MLD air masses are conducive to tropical cyclone formation and/or intensification. The MT air mass is modestly moister and more unstable than the Jordan and Dunion composite mean structures while, as expected, both the SAL and MLD air masses are significantly drier and more stable than the Jordan and Dunion composite mean structures. The drying is most notable above the boundary layer. The mean temperature profiles associated with each of the three air masses are relatively similar. Mean winds are enhanced in the free troposphere within the SAL air mass as compared to the MT and MLD air masses owing to a an easterly mid-tropospheric jet that often accompanies the SAL from Africa.

Heat Sources and Heat Sinks

The spatial distribution of net heating across the tropics drives motion within the tropical latitudes. Before showing this using a simple dynamical model of the atmosphere, however, we first need to use basic principles of thermodynamics to introduce the concept of heat sources and heat sinks.

We first define dry static energy s and moist static energy h as follows:

(1) $$s = c_p T + gz$$

(2) $$h = c_P T + gz + L_v q$$

where c_P, g, and L_v are constants, T is temperature, z is height, and q is the mixing ratio of water vapor. The dry static energy is a measure of enthalpy plus potential energy and the moist static energy is a measure of dry static energy plus latent energy (e.g., energy released as water changes phases). Qualitatively, though not necessarily quantitatively, dry static energy may be viewed as akin to potential temperature while moist static energy may be viewed as akin to equivalent potential temperature.

For dry adiabatic processes, dry static energy is approximately conserved following the motion. For both dry and moist adiabatic processes, moist static energy is approximately conserved following the motion. However, the presence of diabatic processes leads to neither dry nor moist static energy being conserved. In particular, the first law of thermodynamics states that changes in dry static energy following the motion are a function of the heating rate. In the troposphere, this heating is a function of the net radiation (e.g., incoming minus outgoing) and net latent heating (particularly condensation and evaporation, as these are associated with much greater latent heats than phase changes between liquid and solid water).

In its simplest form, the *apparent heat source* Q_1 (positive for increased dry static energy) is defined by this heating rate, such that:

(3) $$Q_1 = \frac{Ds}{Dt} = Q_R + L_v(c - e)$$

where Q_R is the heating rate due to radiation, c is the rate of condensation per unit mass of air, and e is the rate of evaporation of cloud droplets per unit mass of air.

In (3), Q_1 is defined with respect to s, which is applicable independent of scale. However, we wish to instead derive an expression for $\bar{s}$, or the contribution to s from only large-scale processes (or those that we can readily measure, in contrast to the turbulent scales which we cannot readily measure). We can write any variable as the sum of a large-scale mean (overbar) and local perturbation (prime), such as:

$$s = \bar{s} + s'$$

Further, the total derivative can be written in flux form, where:

$$\frac{D(\;)}{Dt} = \frac{\partial(\;)}{\partial t} + \nabla \cdot (\mathbf{v}(\;))$$

If we write s, c, and e in (3) in terms of mean and perturbation quantities, expand the total derivative, apply Reynolds' averaging to separate the large (resolvable) and small (unresolvable) scales of motion, and apply Reynolds' postulates to simplify the result, we obtain:

(4a) $$Q_1 = \frac{\partial \bar{s}}{\partial t} + \nabla \cdot \left(\overline{\mathbf{v}'s'}\right) = \overline{Q_R} + L_v(\bar{c} - \bar{e})$$

The flux-form term in the middle of this equation has three components: two horizontal (*u's'* and *v's'*) and one vertical (*ω's'*). These represent the transport of perturbation dry static energy by turbulent eddies. If we assume that the horizontal components of this term can be neglected, we can rewrite (4a) as:

(4b)
$$Q_1 = \overline{Q_R} + L_v(\overline{c} - \overline{e}) - \frac{\partial}{\partial p}(\overline{s'\omega'})$$

where overbars denote horizontal averages and primes denote deviations from these horizontal averages. The third term on the right-hand side of (4b) reflects the large-scale average of the turbulent *vertical* sensible heat transport on the small, or unresolvable, scales of motion. In the aggregate, (4) shows that the apparent heating (on the large-scale) is a function of large-scale radiative heating, the large-scale release of latent heat due to net condensation, and the vertical sensible heat flux convergence (where *s'ω'* is the vertical sensible heat flux, or eddy/small-scale vertical transport of sensible heat). Because the left-hand side of (4b) is related to the dry static energy *s*, Q_1 can be related to changes in temperature or, through Poisson's law, potential temperature, following the motion.

If we vertically integrate the right-hand side of (4b) between the surface (p_{sfc}) and tropopause (p_{trop}), we obtain:

(5)
$$< Q_1 > = < Q_R > + L_v P + S$$

where *P* is the precipitation rate (net condensation excess) and *S* is the vertical surface sensible heat flux. The <> denote vertically integrated quantities. This equation enables us to readily demonstrate the impact of diabatic processes upon the vertically integrated horizontal heating distribution.

We can also define an equation of moisture continuity as follows (Yanai et al. 1973, Eqn. 7):

(6)
$$\frac{Dq}{Dt} = e - c$$

where the rate of change of water vapor mixing ratio *q* following the motion is a function of net evaporation. If we multiply (6) by L_v, re-order the right-hand side of the equation, make use of the flux form of the total derivative, and applying Reynolds' averaging, we obtain an expression for Q_2, the *apparent moisture sink*:

(7)
$$Q_2 = -L_v \frac{D\overline{q}}{Dt} = L_v(\overline{c} - \overline{e}) + L_v \frac{\partial}{\partial p}\left(\overline{q'\omega'}\right)$$

As the name moisture <u>sink</u> implies, Q_2 is positive for a reduction of *q* following the motion. The last term on the right-hand side of (7) represents the large-scale average of the turbulent vertical latent heat transport on smaller scales. Formally, this term is referred to as the vertical latent heat flux divergence.

Comparing (7) and (4b) to (1) and (2), it is clear that Q_1 reflects changes in dry static energy *s* and Q_2 reflects changes in moist static energy (by way of changes in water vapor mixing ratio *q*). Q_2 is a function of phase changes and the vertical convergence of latent heating that occurs on small scales.

Next, we vertically integrate (7) to obtain a relationship for $<Q_2>$ similar to that for $<Q_1>$ in (6):

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

where E is the surface evaporation rate per unit area, commonly referred known as the vertical surface latent heat flux.

If we combine (4b) and (7), we obtain:

(9) $$Q_1 - Q_2 - \overline{Q_R} = -\frac{\partial}{\partial p}\left(\overline{s'\omega'}\right) - L\frac{\partial}{\partial p}\left(\overline{q'\omega'}\right) = -\frac{\partial}{\partial p}\left(\overline{h'\omega'}\right)$$

where we apply the definition of moist static energy in (2) to obtain the final expression on the right-hand side. This equation has widely been used to measure the activity of cumulus convection, as noted by Yanai and Tomita (1998), wherein small-scale convective heating (the right-hand side of (9) above) can be computed from the large-scale apparent heat source, apparent moisture sink, and radiative heating rate. Similarly, if we combine (5) and (8), we obtain:

(10) $$< Q_1 > - < Q_2 >=< Q_R > + S + L_v E$$

Unique physical insights can be gained by comparing (4b) and (7) as well as (5) and (8), as noted by Yanai and Tomita (1998). If the heating over a given area is primarily due to condensation, wherein the terms in (5) and (8) prefaced with an L_v are dominant, $<Q_1>$ and $<Q_2>$ should be approximately equal to one another. However, if there is also strong sensible heating (S in $<Q_1>$) and/or evaporation from the surface (E in $<Q_2>$), then significant differences may arise between $<Q_1>$ and $<Q_2>$. The presence of large-scale and small-scale terms in (4b) and (7) enables considerations related to condensation to be further divided into stratiform (large-scale) and cumulus convective (small-scale) components. If condensational heating is mostly stratiform in nature, such as is common underneath subtropical ridges, then the vertical profiles of Q_1 and Q_2 are fairly similar to one another. However, if such heating is mostly driven by cumulus convection, the vertical profiles and levels of peak magnitude of Q_1 and Q_2 will differ from one another because of contributions from the flux divergence terms. In the absence of condensation and/or evaporation, then Q_2 is negligible and the vertical profile of Q_1 is modulated by turbulent sensible heat transport.

In the accompanying lecture materials, the seasonal variations of the global distributions of 15-year mean vertically integrated $<Q_1>$ and $<Q_2>$ are presented. During northern hemisphere winter, major heat sources are found along the north Pacific and Atlantic storm tracks as well as across South America, southern Africa, and from the Indian Ocean eastward to the south Pacific convergence zone. In the tropics, both $<Q_1>$ and $<Q_2>$ are large, indicating that latent heat release due to condensation is the major contributor to heating. With small outgoing longwave radiation, much of this latent heat release can be attributed to deep cumulus convection. Over the northern storm tracks, negative values of $<Q_2>$ over storm formation regions and positive values of $<Q_2>$ over downstream regions indicate that sensible heating (S in (5), with $<Q_1>$) dominates early in the mean storm lifecycle while latent heat release becomes more important at later times. Heat sinks are found over northern hemisphere continents and eastern parts of the subtropical north Pacific and Atlantic oceans. These heat sinks are associated with large negative values of $<Q_1>$ and $<Q_2>$, indicative of radiative cooling (negative Q_R) exceeding sensible and latent heating for $<Q_1>$ and evaporation exceeding precipitation for $<Q_2>$.

During northern hemisphere spring, much of Asia and North America become heat sources; however, north Africa and India remain heat sinks. Relatively high values of $\langle Q_1 \rangle$ coupled with small values of $\langle Q_2 \rangle$ imply that sensible heating of the underlying surface (the *S* term in (5) above) is the major contributor to heating over Asia and North America. Heat sources along the north Pacific and Atlantic storm tracks substantially weaken, as do their counterparts from the Indian Ocean to the south Pacific convergence zone. Heating and convection in the tropics are least vigorous during northern hemisphere spring as compared to the other seasons.

During northern hemisphere summer, the major heat sources in the tropics are located to the north of the equator. Principal heat sources are located with the Asian monsoon, the tropical western Pacific, and near Central America. Over land masses, $\langle Q_1 \rangle$ is large whereas $\langle Q_2 \rangle$ is small, implying dominant sensible heating. Over the tropics, however, both $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ are large and associated with small outgoing longwave radiation, implying dominant latent heating associated with deep cumulus convection. Cooling extends over much of the subtropical latitudes in the southern hemisphere as well as over the eastern north Pacific and Atlantic oceans.

Finally, during northern hemisphere fall, the major heat sources retreat southward toward the equator and weaken. Much of Europe, Asia, and North America become heat sinks. Central Africa and northern South America become heat sources owing to latent heat release associated with deep cumulus convection (large $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$, small outgoing longwave radiation).

Heating and Tropical Circulations

If the net vertically integrated diabatic heating rate is applied as a forcing term to the shallow water form of the equations of motion, it is possible to obtain a representation of the atmospheric flow resulting from such heating. In later lectures on the Walker circulation, we will go into more detail on the shallow water approximation and the analytical solution to the shallow water equations in response to prescribed heating. However, in this section, we instead focus upon demonstrating that diabatic heating drives the mean circulation of the tropics.

In the accompanying lecture materials, the net vertically integrated diabatic heating for two months, January and July 1989, is presented (Zhang and Krishnamurti 1996). Qualitatively, the net diabatic heating in each month closely resembles the global distributions of the vertically integrated apparent heat source from the climatology of Yanai and Tomita (1998). At any given location and over the entire tropics, there exist notable differences between the net diabatic heating rates in January and July. This follows naturally from the monthly and seasonal variability in the tropical mean circulation and its associated meteorological phenomena presented earlier. These net vertically integrated diabatic heating distributions are used to drive a diagnostic model based upon the shallow water approximation to the equations of motion.

As can be seen from a comparison of the predicted and observed monthly mean tropical circulation analyses, the simple diagnostic model driven by net vertically integrated heating reasonably depicts the mean tropical circulation during both months. The subtropical highs of the Atlantic and Pacific oceans; the Mascarene high over the southwestern Indian Ocean; the heat lows over the Sahara, 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 in the mid-latitudes, where it is hypothesized that other factors unresolved by the shallow water model may contribute to 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.

We can also demonstrate this connection by examining the shallow water equations themselves instead of a shallow water equation model's output, as in Sobel et al. (2001). Assuming that horizontal temperature gradients in the tropics are weak – the so-called *weak temperature gradient approximation* – scaling of the shallow water equations results in a direct proportionality between diabatic heating and horizontal divergence, with the latter the result of the former. From continuity, this horizontal divergence forces vertical motions, with tropospheric-deep ascent the result of diabatic warming by moist convection. This, in turn, promotes the amplification of pre-existing, lower-to-middle tropospheric cyclonic absolute vorticity, such as may be associated with a finite-amplitude disturbance, via vortex stretching and horizontal advection. In this sense, diabatic heating is directly responsible for the major tropical circulations yet has little effect on the magnitude of the horizontal temperature gradient, which is assumed to remain weak.

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 the stratiform and convective heating profiles. Examples of both are presented in the lecture materials. The stratiform heating profile is characterized by diabatic heating maximized over a shallow layer in the lower to middle troposphere near the altitude of the stratiform cloud deck. Above and below that layer, near-zero heating or even cooling (radiative above cloud top, evaporative below) are typically observed. In the tropics, stratiform precipitation (and thus stratiform heating profiles) is typically associated with older, decaying convection. The convective heating profile is characterized by strong diabatic heating throughout the depth of the troposphere with a maximum in the middle to upper troposphere. Both reflect the nature of the clouds and meteorological phenomena implied by their names.

Of course, not all heating in the atmosphere is in the form of diabatic heating associated with clouds and precipitation. Sensible heating, among others, is often characterized by strong near-surface heating that rapidly decays with increasing altitude (in the absence of diabatic processes). However, the convective and stratiform heating profiles are the most common manifestations of non-vertically integrated heating profiles.

For Further Reading

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