Tropical Meridional Circulations: The Hadley Cell

Introduction

Throughout much of the previous sections, we have alluded to but not fully described the mean meridional overturning circulation of the tropics known as the Hadley cell. Herein, we rectify this by describing the Hadley cell, why it exists, what its role is in maintaining energy and momentum balance within the Earth system, and how it can be quantified using a simple dynamical diagnostic model.

Key Questions

- What is the Hadley cell and why does it exist?
- What is meant by the conservation of absolute angular momentum and why is it important?
- Why is the Earth's meridional overturning circulation best represented by a "three-cell" (rather than "one-cell") model?
- What is the nature of the variability in the Hadley cell?

Introduction to the Hadley Cell

Under conditions of hydrostatic balance, the thickness of a layer can be approximated as a function of the mean virtual temperature within that layer, i.e.,

$$\partial z \approx -\frac{R_D \overline{T_V}}{g} \partial (\ln p) \tag{1}$$

$$z_2 - z_1 \approx -\frac{R_D \overline{T_V}}{g} \ln \frac{p_2}{p_1} \tag{2}$$

such that if we take R_D, g, p_2, and p_1 to all be constant, the distance $z_2 - z_1$ varies solely as a function of the mean virtual temperature within that layer. With greater relative heating at tropical latitudes and less relative heating at higher latitudes, the thickness of a given pressure layer (such as the tropopause, roughly 1000-150 hPa in the tropics) is higher in the tropics than at higher latitudes. This results in the development of a lower tropospheric, meridionally oriented horizontal pressure gradient between high pressure at higher latitudes and low pressure in the tropics. Conversely, near the tropopause, the resultant meridional pressure gradient is reversed.

The force balance corresponding to the *uncurved* flow (i.e., neglected centrifugal force) described above is characterized by geostrophic balance, with the pressure gradient force (directed from high to low) balancing the Coriolis force (directed to the right of the motion in the northern hemisphere). To first order, this results in the easterly near-surface flow in the tropics. It should be noted, however, that the Coriolis force is weak in the tropics and zero at the Equator, such that geostrophic balance is a crude approximation at best in the tropics. Near the surface, deviations from this flow occur due to frictional drag. As friction does not change the pressure gradient force, the corresponding force balance is characterized by a reduced Coriolis force. This results in the deflection of air parcels toward areas of low pressure.

This results in lower-tropospheric convergence near the Equator and lower-tropospheric divergence at higher latitudes (e.g., the subtropics). From continuity alone, this would imply ascent near the equator and descent in the subtropics. Taking the tropopause to be a rigid lid on vertical motions, deep-layer ascent leads to divergence aloft and deep-layer descent to convergence aloft. This implies poleward meridional motion aloft, with flow directed from the equator toward the subtropics. The Coriolis force acts to deflect this flow eastward, imposing a limit on the meridional extent to this overturning circulation. Alternatively, this circulation can be derived from the previously presented arguments of thickness and force balances.

Putting these pieces together, we are given a simplistic view of the meridional overturning circulation of the tropics. This circulation is driven by the differential heating between the equator and subtropics, itself driven by the previously discussed latitudinal variation in annual mean insolation, and is characterized by ascent where it is (relatively) warm and descent where it is (relatively) cool. In this regard, this circulation is a *thermally direct* circulation. The near-surface branch of this circulation is characterized by mean equatorward flow whereas its upper tropospheric branch is characterized by mean poleward flow.

The Hadley Cell and Global Energy Balance

We next turn to understanding the role of this circulation to the Earth system as a whole. Let us temporarily return to the concept of radiative balance. Here, we allow insolation to vary as a function of latitude so as to understand the impact of radiative balance on the mean temperature differential between the equator and poles. Observations indicate that approximately 425 W m^{-2} of insolation is incident at the top of the atmosphere in the tropics, of which, on average, 300 W m^{-2} of insolation per unit area per day is absorbed (given an albedo of ~0.3) by the atmosphere and surface. At the poles, approximately 150 W m^{-2} of insolation is incident at the top of the atmosphere in the tropics, of which, on average, 50 W m^{-2} of insolation per unit area per day is absorbed (given an albedo of ~0.7) by the atmosphere and surface.

As before, we can express radiative balance as a function of insolation via the Stefan-Boltzmann equation:

$$F = \sigma T^4 \tag{3}$$

Here, we have again taken the emissivity of the Earth to be equal to one, approximating the Earth as a blackbody emitter. If we substitute the net absorbed insolation per unit area for the tropics into (3) and solve for *T*, we obtain a mean equivalent blackbody temperature of 269.70 K. Similarly, if we substitute the net insolation for the poles per unit area into (3) and solve for *T*, we obtain a mean equivalent blackbody temperature of 172.32 K. Thus, from radiative balance based off of insolation alone, the pole-equator temperature difference is 97.38 K! (Note that we are more interested in the magnitude of this temperature difference than in whether the estimates for the temperature at the poles or equator are realistic, which they are not given that we are only considering absorbed incoming solar radiation in this example. If one considers more than just insolation, a similar magnitude with more realistic values may be obtained.)

In order to reduce the pole-equator temperature difference to a value that more closely matches observations – e.g., 30-50 K, depending upon summer versus winter – meridional redistribution of heat energy must be accounted for. ***What***, however, redistributes heat energy so as to weaken the excessively large pole to equator temperature differential obtained from radiative balance alone?

We can represent Earth's energy budget in terms of its physical constituents, such that:

$$E = c_P T + gz + L_v q + K + G + \Delta f \tag{5}$$

where the terms on the right-hand side of (5) represent sensible heating, potential energy, latent heating, kinetic energy, heat storage, and horizontal transport, respectively. The units on (5) are J kg^{-1}, or energy per unit mass. In other words, the Earth's energy budget is defined by the moist static energy, kinetic energy, and storage and transport mechanisms. The transport term Δf is comprised of three factors: sensible heat, potential energy, and kinetic energy in the atmosphere; water vapor (latent heat) in the atmosphere; and sensible heat in the ocean (e.g., by ocean currents).

The transport term helps to balance the energy budget across the globe by transporting excess heat energy poleward. Broken down by components, mean transport across the globe (all latitudes) is roughly:

- 15% due to meridional overturning circulations such as the Hadley cell
- 15% due to the transport of water vapor from evaporation-dominant to precipitation-dominant regions (e.g., from the subtropics to the intertropical convergence zone)
- 30% due to oceanic currents such as the Gulf Stream, Kuroshio current, and others
- 40% due to atmospheric eddies such as mid-latitude and tropical cyclones and other phenomena

Within the tropical latitudes, however, much of the transport is accomplished by the Hadley cell. We can define the total zonally averaged meridional transport of any quantity as follows:

(6) *total transport = mean transport + transient eddies + stationary eddies*

Mean transport reflects transport by the Hadley cell and other similar, large-scale climatological circulations; transient eddies reflect features such as synoptic-scale cyclones; and stationary eddies represent localized climatological features such as oceanic currents. Oort and Rasmussen (1971) demonstrated that in the northern hemisphere, during both winter and summer, the mean transport of momentum, sensible heating, potential energy, latent heating, and kinetic energy between 0-15°N is dominated by the mean transport term. At higher latitudes, however, the eddy terms become much larger, where mid-latitude cyclones are prevalent and act to transport energy meridionally.

Conservation of Absolute Angular Momentum

In addition to helping maintain global energy balance, tropical circulations arise out of and help to maintain global angular momentum balance. To demonstrate this, we turn to the concept of absolute angular momentum and its conservation. At the most basic of levels, absolute angular momentum (AAM) is defined as the product of mass, rotational velocity, and the radius from the axis of rotation (for the Earth, the axis aligned with the north and south poles). In mathematical terms, this is:

$$AAM = mr\omega \quad (7)$$

Here, *m* refers to mass. For the Earth, *r* is equal to *a cos Θ* where *a* is equal to the radius of the Earth (6.3781 x 10^6 m) and *θ* is latitude. The rotational velocity *ω* is equal to the angular velocity of the Earth *r*Ω plus the zonal velocity *u*. Ω is defined as the rate of rotation of the Earth and is equal to 7.292 x 10^{-5} rad s^{-1}. Per unit mass, absolute angular momentum can thus be expressed as:

$$M = (a\cos\theta)(\Omega a\cos\theta + u) \quad (8)$$

Since Ω and *a* are constants, *M* varies only as a function of zonal velocity and latitude. Along a moving air parcel, in the absence of the exchange of momentum with other air parcels and/or the surface such as by friction, we state that *M* is a conserved quantity.

Consider an air parcel at 0°N with no initial wind speed. Due to the rotation of the Earth, it still has an initial non-zero *M*, i.e.,

$$M = a * \Omega a = \Omega a^2 = 2.96 \times 10^9 \; m^2 \, s^{-1} \quad (9)$$

Let us consider an air parcel originating at the tropopause, with no zonal velocity due to it having just reached the tropopause in the ascending branch of the Hadley circulation. In the poleward branch of the Hadley circulation, let us presume that it moves from the equator to 30°N. If *M* is conserved, meaning that it is unchanged along the parcel's motion, then what is the zonal velocity of the initially zonally stationary parcel once it reaches 30°N?

$$2.96 \times 10^9 \; m^2 \, s^{-1} = (a\cos 30°)(\Omega a\cos 30° + u_{final}) \quad (10)$$

If we solve for u_{final}, we obtain a zonal velocity of 134 m s^{-1}! Because the distance from the axis of rotation decreases as we move from the equator to 30°N, the zonal velocity must increase to conserve *M*. Were we to displace the parcel even further poleward, the zonal velocity would grow even larger, a highly non-physical solution.

The conservation of absolute angular momentum also helps give rise to the subtropical jet stream, wherein parcels in the ascending branch of the thermally direct tropical meridional circulation accelerate eastward near the tropopause, due to Coriolis deflection, as they conserve absolute angular momentum. Note, however, that the subtropical jet is weaker than 134 m s^{-1} because of internal friction (such as is associated with small-scale turbulent eddies aloft), absolute angular momentum loss to the mid-latitudes by atmospheric eddies, and the extent to which the Coriolis force deflects air parcels sufficiently in a zonal direction such that they do not accelerate poleward any further.

What happens in the middle and polar latitudes? Note first that lower tropospheric winds in the middle latitudes are generally westerly while lower tropospheric winds in the polar latitudes are generally easterly. This implies higher pressures at the poles and in the subtropics and lower pressures in the middle latitudes. For convergent flow associated with low pressure, ascent results. For divergent flow associated with high pressure, descent results. Thus emerges a three-cell meridional circulation model, comprised of the Hadley tropical, Ferrel middle latitude, and polar circulations. Like the Hadley cell, the polar cell is a thermally direct circulation. Conversely, the Ferrel cell is a thermally indirect circulation, characterized by ascending cold air and descending warm air. The polar jet tends to be located on the interface of the Ferrel and polar cells and, like the interface between the Hadley and Ferrel cells, is the upper-tropospheric manifestation of the polar front air-mass boundary, with which it is in thermal wind balance. In the aggregate, these circulations act to deposit excess energy from the tropics at higher latitudes, significantly contributing to maintaining global energy balance.

Hadley Circulation Variability

The basic theory presented above was derived under the assumption of an aquaplanet, or an Earth without land masses, and captures the basic structure of the meridional circulations of our planet. However, when we consider the presence of continents, more complexity is introduced into the structures of these circulations. Instead of bands of high and low pressure everywhere along the interfaces of each cell, areas of lower-tropospheric high and low pressure form in preferred regions largely driven by land-sea contrasts and the corresponding differences in specific heat capacity of these two surface constituents, with cyclones (anticyclones) generally favored over land (water).

In our lecture on the "Climatology of the Tropics," we previously examined the mean structure of the vertical motion associated with the Hadley cells. The tropics (subtropics) are characterized by rising (sinking) motion, though the patterns are not mirrored about the equator, nor are they stationary throughout the year. Rather, there is substantial seasonal variability in the locations of the ascending and descending branches of the Hadley cell and in the intensity of these branches between the southern and northern hemispheres. The descending branch of the Hadley cell is relatively weaker (stronger) in the northern hemisphere during local summer (winter) as compared to its southern hemisphere counterpart. In a yearly mean, the ascending branch of the Hadley cell is generally displaced slightly northward from the equator.

A larger percentage of the tropical and subtropical surface is comprised of land (rather than water) in the northern relative to the southern hemisphere. Owing to considerations of specific heat capacity, for a given heat energy input, temperature over land changes both more rapidly and by a greater magnitude than over water. During local summer, this results in stronger zonally averaged tropical and subtropical heating in the northern hemisphere, leading to a poleward displacement of the upward branch and a weakening of the northern hemispheric downward branch of the Hadley cell. Conversely, during northern winter, this continentality results in stronger zonally averaged cooling in the northern hemisphere. This results in stronger wintertime descent in the subtropical descending branch of the northern hemisphere Hadley circulation. Thus, it is a combination of seasonality (i.e., the changing tilt axis of the Earth with respect to the sun) and the underlying distribution of land masses that contribute most strongly to variability within the Hadley cell.

The Hadley cell also exhibits variability on climate time scales. There exists some evidence for an intensified Hadley circulation in both observations and future climate projections. Further, several recent studies have suggested that the Hadley cell's poleward extent has increased since the 1970s by between 2-5° latitude, although the precise magnitude of this increase in part depends upon how the Hadley cell is defined. Climate model projections indicate that the Hadley cell's poleward extent will continue to increase into the future. Lu et al. (2007) explain this expansion as follows. Consider the case where climate change leads to a 1°C warming of the surface temperature within the tropics. After an air parcel is lifted to saturation within the ascending branch of the Hadley circulation, all further ascent follows a moist adiabat. However, the lapse rate along a moist adiabat is temperature-dependent due to latent heat released upon moist adiabatic ascent, resulting in a steeper lapse rate at colder temperatures relative to warmer temperatures. Thus, a 1°C warming at the surface translates to a greater than 1°C warming in the upper troposphere near the Equator as well as within the meridional flow extending poleward from the Equator. At any given tropical latitude, this leads to increased upper tropospheric static stability relative to the past climate. The baroclinic breakdown of the subtropical jet is one theory for why the subtropical jet's poleward extent is limited; its ability to breakdown is inversely related to the static stability. As a result, increased upper tropospheric static stability fostered by climate change results in the subtropical jet being able to extend further poleward before becoming unstable. The extent to which this will be realized in a future climate is somewhat uncertain, however, particularly given that climate models have not accurately represented the recent observed increase in the Hadley cell's poleward extent.

Hadley Circulation Dynamics

For more in-depth, comprehensive introductions to the dynamics of the Hadley cell, interested readers are referred to Held and Hou (1980) and Schubert et al. (1991). Herein, we consider the Kuo-Eliassen equation, an equation that is somewhat analogous to the non-linear Eliassen secondary circulation equation applied in an *r*-*z* coordinate system to the study of tropical cyclones and frontal circulations. For the study of the Hadley circulation, we apply this equation in a *y*-*p*, rather than *r*-*z*, coordinate system.

The Kuo-Eliassen equation is given by:

(11)
$$A\frac{\partial^2\psi}{\partial y^2}+2B\frac{\partial^2\psi}{\partial y\partial p}+C\frac{\partial^2\psi}{\partial p^2}=\frac{\partial H}{\partial y}+\frac{\partial M_S}{\partial p}$$

where *A* is the static stability and is related to the vertical potential temperature gradient, *B* is the baroclinicity and is related to the horizontal temperature gradient magnitude and (in as much as thermal wind balance holds in the tropics) vertical wind shear, *C* is the inertial stability and is related to both the Coriolis force and the meridional gradient of the zonal wind, $\frac{\partial H}{\partial y}$ is the meridional gradient of heating (both sources and sinks), $\frac{\partial M_S}{\partial p}$ is the vertical gradient of momentum sources and sinks, and Ψ is the streamfunction characterizing the meridional circulation, such that:

(12)
$$v=\frac{\partial\psi}{\partial p} \text{ and } \omega=-\frac{\partial\psi}{\partial y}$$

In the above, M_S is a function of the meridional gradient of the meridional flux (transport) of perturbation zonal velocity. However, herein, we shall assume that the meridional flow in the upper and lower tropospheric branches of the Hadley cell is constant in the meridional direction. We will also assume that the perturbation zonal wind is a perturbation from an initial state of rest. Thus, from these approximations, we state that M_S is proportional to the meridional gradient of the zonal velocity *u*, or $M_S \propto \frac{\partial u}{\partial y}$, where $u > 0$ indicates westerly flow.

From (11), we can describe a meridional circulation as being *driven* by momentum and heat forcing M_S and H and *modulated* by static stability, inertial stability, and baroclinicity A, B, and C. In the tropics, B is often negligibly small, such that (11) simplifies to:

$$A\frac{\partial^2\psi}{\partial y^2} + C\frac{\partial^2\psi}{\partial p^2} = \frac{\partial H}{\partial y} + \frac{\partial M_S}{\partial p} \tag{13}$$

Differentiating v with respect to p and ω with respect to y in (12) and substituting into (13):

$$-A\frac{\partial\omega}{\partial y} + C\frac{\partial v}{\partial p} = \frac{\partial H}{\partial y} + \frac{\partial M_S}{\partial p} \tag{14}$$

At this point, it is useful to recall the basic structure of the Hadley circulation in the northern hemisphere: ascent is found near the Equator, descent is found in the subtropics, poleward flow is found near the tropopause, and equatorward flow is found near the surface. Thus, over the tropics, $\frac{\partial\omega}{\partial y} > 0$ and $\frac{\partial v}{\partial p} < 0$. Note that except in very specific circumstances, A and C are both positive-definite. As a result, the left-hand side of (14) is negative such that $\frac{\partial H}{\partial y} + \frac{\partial M_S}{\partial p}$ is negative. This occurs for…

- $\frac{\partial H}{\partial y}$ is negative where heating is strongest near the Equator and weaker at higher latitudes
- $\frac{\partial M_S}{\partial p}$ is negative when primarily when $\frac{\partial u}{\partial y}$ is positive aloft, i.e., the upper tropospheric zonal velocity grows more westerly away from the equator.

From this, it is clear how insolation (leading to heating) maximized near the equator and the conservation of absolute angular momentum as parcels travel meridionally drive the base state of the Hadley cell circulation. Independent of the control exerted by static and inertial stability, Hadley cell intensity and/or structure are a function of the meridional distribution of heating and the vertical distribution of momentum forcing. The former are most often associated with the aforementioned land/sea and associated surface contrasts or with meridional variability in diabatic heating, such as may be associated with latent heat release in areas of deep, moist convection. The latter are most often associated with external (e.g., mid-latitude) forcing upon the tropical circulation.

For Further Reading

- Chapter 1, *An Introduction to Tropical Meteorology, 2nd Edition*, A. Laing and J.-L. Evans, 2016.
- Chapter 3, *An Introduction to Tropical Meteorology, 2nd Edition*, A. Laing and J.-L. Evans, 2016.
- Lecture Notes, The Zonally averaged Circulation, K. Emanuel, 2011. (*Full citation: Emanuel, K., 2005: 12.811 Tropical Meteorology. (Massachusetts Institute of Technology: MIT OpenCourseWare), http://ocw.mit.edu. License: Creative Commons BY-NC-SA.*)

- Held, I. M., and A. Y. Hou, 1980: Nonlinear axially symmetric circulations in a nearly inviscid atmosphere. *J. Atmos. Sci.*, **37**, 515-533.
- Lu, J., G. A. Vecchi, and T. Reichler, 2007: Expansion of the Hadley cell under global warming. *Geophys. Res. Lett.*, **34**, L06805.
- Oort, A. H., and E. M. Rasmussen, 1971: Atmospheric circulation statistics. NOAA Prof. Pap. No. 5, U.S. Dept. of Commerce, 323 pp. [NTIS COM-72-50295].
- Schubert, W. H., P. E. Ciesielski, D. E. Stevens, and H.-C. Kuo, 1991: Potential vorticity modeling of the ITCZ and the Hadley circulation. *J. Atmos. Sci.*, **48**, 1493-1509.