Tropical Meridional Circulations: The Hadley Cell

Introduction

In the Tropical Climatology lecture, we hinted at there being a meridional (meaning north/south-oriented) overturning (meaning something with ascending and descending motions on its ends) circulation known as the Hadley cell. This lecture provides a deeper discussion of the Hadley cell, why it exists, what role it has in maintaining the Earth's energy and momentum balances, and how it can be quantified using simplified numerical models.

Key Questions

- What is the Hadley cell's basic meridional and vertical structure?
- What form of diabatic heating is most-commonly invoked as the Hadley cell's driving mechanism?
- What is the Hadley cell's primary role in global energy balance?
- How does the conservation of absolute angular momentum result in a subtropical jet stream?
- How is the Hadley cell's structure modified by land-sea temperature contrasts and the seasons?
- What commonalities do the thickness + Coriolis, Kuo-Eliassen, and shallow-water perspectives on Hadley cell structure share? How do they differ?

Introduction to the Hadley Cell

Assuming hydrostatic balance, a layer's thickness can be approximated using the mean virtual temperature within that layer, i.e.,

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

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

If we assume that R_D, g, p_2, and p_1 are all constant, the distance $z_2 - z_1$ between the two pressure surfaces p_1 and p_2 varies solely as a function of the mean virtual temperature in that layer.

Annually averaged net radiation is positive in the tropics and becomes negative at higher latitudes. Ignoring other processes that influence the virtual temperature for a moment, this results in higher layer-mean virtual temperatures in the tropics and lower layer-mean virtual temperatures at higher latitudes. This means that a given lower-tropospheric pressure surface is closer to the ground and a given upper-tropospheric pressure surface is further above the ground in the tropics than at higher latitudes. Thus, the near-surface conditions are characterized by relatively low pressure in the tropics and high pressure at higher latitudes, whereas the near-tropopause conditions are characterized by relatively high pressure in the tropics and low pressure at higher latitudes.

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) opposing the Coriolis force (right of motion in the Northern Hemisphere, left of motion in the Southern Hemisphere). For near-surface air parcels initially at rest, the pressure-gradient force accelerates them from high to low pressure, or from higher latitudes toward the Equator. In turn, upper-tropospheric air parcels initially at rest also accelerate from high to low pressure, except at these altitudes this is directed from the Equator toward higher latitudes. This results in air parcels that are deflected from east to west (i.e., easterly) near the surface and from west to east (i.e., westerly) near the tropopause. The magnitude of this deflection is less near the surface, where friction reduces the magnitude of the Coriolis force without changing the pressure-gradient force and thus leads to air converging into the area of near-surface lower pressures in the tropics. However, note that this is all a crude approximation since the Coriolis force becomes zero at the Equator itself.

Near-surface air directed toward the Equator from both hemispheres results in convergence near the Equator and divergence at higher latitudes near the surface. Conversely, upper-tropospheric air directed away from the Equator in both hemispheres results in divergence near the Equator and convergence at higher latitudes near the tropopause. From continuity, this means that there must be a deep layer of ascent near the Equator and a deep layer of descent at higher latitudes. It is this circulation – near-surface convergence, deep-layer ascent, and upper-level divergence near the Equator with near-surface divergence, deep-layer descent, and upper-level convergence at higher latitudes – that characterizes the Hadley cell.

Altogether, the Hadley cell is driven by the differential heating between the Equator and higher latitudes, itself driven by the previously discussed latitudinal variation in annual mean insolation, and is characterized by ascent in the relatively warm tropics, poleward transport of warm air near the tropopause, descent in the relatively cool higher latitudes, and equatorward transport of colder air near the surface. The Hadley cell is thus considered to be a *thermally direct* circulation, generically defined as one in which warm air rises.

The Hadley Cell and Global Energy Balance

The Hadley cell plays an important role in modulating global temperatures. Let us demonstrate this using a realistic thought experiment. Observations indicate that approximately 425 W m^{-2} of insolation (incoming solar radiation) reaches the top of the atmosphere in the tropics, with ~300 W m^{-2} (or ~70%) of this being absorbed (given an albedo of ~0.3) by the atmosphere and surface. Likewise, approximately 150 W m^{-2} of insolation reaches the top of the atmosphere near the poles, with ~50 W m^{-2} (or ~30%) of this being absorbed by the atmosphere and surface.

We can express the balance between radiation and temperature using the Stefan-Boltzmann equation:

(3)
$$F = \sigma T^4$$

This form of the Stefan-Boltzmann equation assumes that the Earth's emissivity is 1, indicating that Earth emits all of the radiation that it absorbs (defining a *blackbody emitter*). If we substitute 300 W m^{-2} and 50 W m^{-2} into this equation and solve for T (here defined as the *blackbody temperature*), we obtain:

- **Tropics**: T = 269.70 K
- **Poles**: T = 172.32 K

In other words, considering absorbed insolation alone, the pole-to-Equator temperature difference is 97.38 K! This far exceeds the 30-50 K temperature difference that is typically observed. Meridional heat transport by circulations such as the Hadley cell is one the primary ways by which this radiation-driven temperature difference is reduced.

To demonstrate this, we can represent Earth's energy budget in terms of its physical constituents:

$$E = c_P T + gz + L_v q + K + G + \Delta f \quad (4)$$

where the terms on the right-hand side of (4) represent sensible heating, potential energy, latent heating, kinetic energy, heat storage, and horizontal transport, respectively. The units on (4) are J kg^{-1}, or energy per unit mass. The horizontal transport Δf is comprised of three transports: 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 horizontal transport term helps to balance the global energy budget by transporting excess heat energy poleward. Broken down by components, mean transport across all latitudes is roughly:

- 30% due to *mean transport* by features like the Hadley cell, including the transport of water vapor from evaporation-dominant regions (subtropics) to precipitation-dominant regions (tropics).
- 40% due to *transient atmospheric eddies* such as midlatitude and tropical cyclones.
- 30% due to *stationary* eddies, such as the Gulf Stream and Kuroshio ocean currents.

In the tropics, however, the Hadley cell is responsible for most of the horizontal energy transport.

We can write the total zonally averaged meridional transport in terms of the mean, transient, and stationary features defined above as follows:

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

Here, mean transport indicates meridional transport by the Hadley cell and other large-scale climatological circulations, transient eddies represent features like synoptic-scale cyclones; and stationary eddies represent localized climatological features such as oceanic currents. Stationary eddies differ from mean transport in their horizontal scale (eddies are inherently smaller-scale than the mean larger-scale circulations) and their stationarity (the mean larger-scale circulations, while present all year, may change locations within a year). In the Northern Hemisphere tropics, total transport (of all quantities in (4)) is dominated by mean transports (Oort and Rasmussen 1971). At higher latitudes, however, the eddy terms become much larger, particularly the transient eddies that include both poleward-moving tropical cyclones and midlatitude cyclones.

Conservation of Absolute Angular Momentum

Earlier in this lecture, we noted that the Coriolis force causes flow moving away from the Equator in both hemispheres to be directed from west to east (or westerly). We can alternatively demonstrate this using the concept of *the conservation of absolute angular momentum*, which also allows us to demonstrate that this westerly flow *accelerates* as it moves away from the Equator.

Absolute angular momentum (AAM) is generically defined as the product of mass, rotational velocity, and the radius from the axis of rotation (for the Earth, the axis through the North and South poles):

(6) $$AAM = mr\omega$$

Here, m refers to mass. For the Earth, r is equal to $a \cos \Theta$ where a is equal to the radius of the Earth (6.3781×10^6 m) and θ is latitude. The rotational velocity ω is equal to the angular velocity of the Earth $r\Omega$ plus the zonal velocity u. Ω is defined as the rate of rotation of the Earth and is equal to 7.292×10^{-5} rad s^{-1}.

Substituting for r and ω and dividing by mass, we obtain the absolute angular momentum per unit mass:

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

Since Ω and a are constants, M varies only as a function of the zonal velocity u and latitude. In the absence of momentum exchange with other air parcels and/or with the surface due to friction, M is conserved (i.e., it does not change) along a moving air parcel.

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

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

Let's assume this air parcel originates at the tropopause, having reached there in the Hadley cell's ascending branch. Let's further assume that this air parcel will move from the Equator to 30°N in the Hadley cell's upper-tropospheric poleward branch. If M is conserved as this occurs, we can obtain the zonal wind u that the air parcel will have when it reaches 30°N:

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

Solving for u_{final}, we obtain 134 m s^{-1}! Because the distance from the axis of rotation (which extends between the North and South Poles) decreases as we move from the Equator to 30°N due to the Earth's curvature, u must increase for M to remain constant. If we displaced the parcel even further poleward, u would become even larger, a highly non-physical solution; however, the poleward extent to which the displaced air parcels reach is limited by the deflection associated with the Coriolis force, which itself is larger at higher latitudes.

The increase in zonal velocity that air parcels experience as they move poleward under the constraint of the conservation of absolute angular momentum is the driving mechanism for *subtropical jets*. 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), and absolute angular momentum loss to midlatitudes by atmospheric eddies.

Hadley Circulation Variability

The theory presented above assumes that the Earth is an aquaplanet, lacking land masses, but it reasonably captures the Hadley cell's basic structure. If we add continents, areas of near-surface high- and low-pressure form in preferred regions that are largely driven by land-sea temperature contrasts. Because land has a lower specific heat capacity than does water, it more readily warms in response to insolation. As a result, tropical land masses are typically slightly warmer than tropical oceans. In turn, this leads to slightly greater thickness over land as compared to water, leading to lower pressures over land and higher pressures over water.

The theory presented above is derived from the annually averaged radiation budget. However, as shown in our Tropical Climatology lecture, there is substantial seasonal variability in the Hadley cell's structure. The Hadley cell's ascending branch is typically found at the latitudes where the net warming is the greatest; this band is typically in whichever hemisphere is in summer. Conversely, the Hadley cell's descending branch is typically found at the latitudes where the net warming is the weakest; this band is typically in whichever hemisphere is in winter. In the annual mean, the Hadley cell's ascending branch skews slightly north of the Equator because there is more land in the tropics in the Northern versus the Southern Hemisphere. Further, this difference in continentality between hemispheres results in stronger ascent and descent in the Northern Hemisphere during local summer and winter, respectively.

The Hadley cell's structure also varies on climate time scales. There exists some evidence for an intensified Hadley circulation in both observations and future climate projections. Furthermore, several recent studies suggest that the Hadley cell's poleward extent has increased by 2-5° latitude since the 1970s, although the precise magnitude of this increase depends upon how the Hadley cell is defined. Climate-model projections indicate that the Hadley cell's poleward extent will continue to increase in the future. One explanation (Lu et al. 2007) is as follows. Assume that climate change warms tropical surface temperatures by 1°C. After an air parcel is lifted to saturation in the Hadley cell's ascending branch, ascent then follows a moist adiabat. However, the lapse rate along a moist adiabat is temperature-dependent due to latent heat release, resulting in a steeper lapse rate at colder temperatures relative to warmer temperatures. A warmer surface temperature results in a slower decrease of temperature with height after saturation, leading to even greater warming in the upper troposphere than was seen at the surface. The weaker lapse rate leads to increased static stability relative to the past climate, which can be shown to result in a subtropical jet that (on average) can reach to higher latitudes before becoming unstable and breaking down. 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: The Kuo-Eliassen Equation

The Kuo-Eliassen equation is a two-dimensional (horizontal: *y*, vertical: *p*) that can be used to describe the factors driving the meridional overturning circulation associated with the Hadley cell. This equation is given by:

(10)
$$A\frac{\partial^2\psi}{\partial y^2} + 2B\frac{\partial^2\psi}{\partial y\partial p} + C\frac{\partial^2\psi}{\partial p^2} = \frac{\partial H}{\partial y} + \frac{\partial M_S}{\partial p}$$

where *A* is the static stability and is related to how temperature (or formally, potential temperature) changes with height, *B* is the baroclinicity and is related to the horizontal temperature gradient's magnitude, *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 diabatic heating, $\frac{\partial M_S}{\partial p}$ is the vertical gradient of momentum sources and sinks (where M_s is approximated as $\frac{\partial u}{\partial y}$, with $u > 0$ for westerly flow), and Ψ is the streamfunction characterizing the meridional circulation, such that:

(11)
$$v = \frac{\partial\psi}{\partial p} \text{ and } \omega = -\frac{\partial\psi}{\partial y}$$

where v is the meridional (north-south) wind component and ω is the pressure vertical velocity.

From (10), we can describe a meridional circulation as being *driven* by momentum and heat forcing M_S and H and *modulated* by static stability, baroclinicity, and inertial stability A, B, and C. In the tropics, B is often negligibly small, such that (10) simplifies to:

$$A\frac{\partial^2\psi}{\partial y^2} + C\frac{\partial^2\psi}{\partial p^2} = \frac{\partial H}{\partial y} + \frac{\partial M_S}{\partial p} \tag{12}$$

Differentiating v with respect to p and ω with respect to y in (11) and substituting into (12), we obtain:

$$-A\frac{\partial \omega}{\partial y} + C\frac{\partial v}{\partial p} = \frac{\partial H}{\partial y} + \frac{\partial M_S}{\partial p} \tag{13}$$

It is useful to recall the Hadley cell's Northern Hemisphere basic structure. First, recall that ascent is found just north of the Equator and descent is found at higher latitudes. Thus, between these latitudes, $\frac{\partial \omega}{\partial y} > 0$: ω becomes more positive as you move northward along the positive y-axis. Next, recall that poleward flow is found near the tropopause and equatorward flow is found near the surface. Thus, between these altitudes, $\frac{\partial v}{\partial p} < 0$: v becomes more positive as you move toward lower pressure at higher altitudes. Noting that A and C are both positive-definite except in limited circumstances, the left-hand side of (13) is generally negative, such that $\frac{\partial H}{\partial y}+\frac{\partial M_S}{\partial p}$ is generally negative. This occurs when:

- $\frac{\partial H}{\partial y}$ is negative, indicating diabatic warming that is strongest near the Equator and becomes weaker at higher latitudes.

- $\frac{\partial M_S}{\partial p}$ is negative. In general, $\frac{\partial u}{\partial y}$ is positive near the surface (easterly wind at the Equator that becomes less easterly to the north) and near the tropopause (associated with the subtropical jet). In general, however, $\frac{\partial u}{\partial y}$ is more positive aloft than near the surface, resulting in a negative value for $\frac{\partial M_S}{\partial p}$.

Hadley Circulation Dynamics: The Shallow-Water Equations

The equatorial wave structures and characteristics were obtained from a simplified set of equations known as the shallow-water equations. If we apply a steady diabatic warming (mimicking, for instance, insolation or latent warming) centered at the Equator to these equations, we obtain the superposition of two equatorial waves: a Kelvin wave and an equatorial Rossby wave. Averaging this solution across longitudes (or zonally averaging the solution) produces a circulation that varies in the meridional directions and with height that resembles the Hadley cell (Gill 1980). However, it is too weak compared to observations, with horizontal and vertical velocities that are approximately 30% as strong as their observed counterparts and most-closely resembling an annually averaged Hadley cell.

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.*)
- Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447-462.
- 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.
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- 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.