The Zonal Walker Circulation

Introduction

Apart from temperature contrasts that are driven by differences in the heat capacities of land and water, we have largely considered temperatures along the Equator to be approximately the same at all longitudes. This is an oversimplification, however. There are three major sources of variable temperatures along the Equator:

1. Land-water temperature contrasts.
2. Ocean temperature differences between the eastern and western ends of equatorial ocean basins.
3. Diabatic heating differences between the eastern and western ends of equatorial ocean basins.

In the climatological mean, the latter two result in slightly warmer temperatures on the western sides of the equatorial ocean basins and slightly cooler temperatures on the eastern sides of the equatorial ocean basins. As we will describe shortly, these zonal temperature differences arise out of the Earth's general circulation and provide the diabatic forcing for the Walker circulation.

Key Questions

- What is the zonal Walker circulation?

- What are the thermocline and Ekman transport, and why are they important for understanding the oceanic heating that partially drives the Walker circulation?

- At a basic level, how do changes in the trade winds associated with El Niño and La Niña result in changes to the Walker circulation's structure and/or intensity?

- What equatorial wave modes comprise the warming-forced Walker circulation, and what is this circulation's basic lower-tropospheric structure?

What is the Walker Circulation?

The Walker circulation is a heating-driven zonal overturning circulation. It is named for Sir Gilbert Walker, who first recognized its existence from sea-level pressure differences between Darwin, Australia and Tahiti. In recent years, the term "Walker circulation" has come to refer to not just the equatorial zonal circulation in the Pacific Ocean but also the equatorial zonal circulations in the Atlantic and Indian Oceans as well.

Much like the Hadley cell, the Walker circulation is thermally direct, with ascent where it is relatively warm and descent where it is relatively cold. In fact, this heating distribution drives the Walker circulation! In the climatological mean, ascent is favored in the western equatorial ocean basins and descent is favored in the eastern equatorial ocean basins. Accordingly, annual precipitation is highest in the western equatorial ocean basins where air parcels ascend and lowest in the eastern equatorial ocean basins where air parcels descend. The corresponding lower- and upper-tropospheric zonal winds are easterly and westerly, respectively.

The Walker circulation is primarily driven by atmospheric and oceanic warming on the western edge of the equatorial oceans. This warming has three contributors:

1. Surface sensible warming of the equatorial land masses along the western edges of equatorial ocean basins: Indonesia (Pacific Ocean), Africa (Indian Ocean), and South America (Atlantic Ocean).
2. Atmospheric latent heating associated with thunderstorms. These thunderstorms form in response to the surface sensible warming of the equatorial land masses and, to lesser extent, the upslope flow associated with easterly trade winds impinging upon the eastern side of mountain ranges in Africa and South America.
3. Surface sensible warming associated with relatively warm waters in the western equatorial oceans. This warming results from zonal variability in ocean temperatures driven by easterly trade winds.

Let's consider the third warming contributor in a bit more detail.

The Thermocline and Ekman Transport

Just as tropospheric temperatures typically decrease with increasing height above the ground, *tropical* ocean temperatures typically decrease with increasing depth below the surface. The depths over which the ocean temperatures decrease most rapidly with increasing depth are known as the *thermocline*. The thermocline's depth varies from east to west in the tropics but typically is located 50-200 m below the surface. The 20°C isotherm often resides within the middle of the thermocline and is thus often used as a thermocline proxy.

At the ocean's surface, the surface winds attempt to tug the surface water along in the same direction. This is known as a *surface wind stress*. If this were the only process transporting surface water, the climatological easterly trade winds would tug the surface water from east to west in the equatorial oceans. This results in the *upwelling* of cooler subsurface water to the surface in the eastern equatorial oceans as the surface water is transported away and the *downwelling* of warmer surface water to greater depths in the western equatorial oceans as surface water accumulates on the western equatorial ocean boundaries. This alone can be used to understand the west-to-east oceanic warming distribution that drives the Walker circulation.

However, this is not the only process transporting surface water. The tugging described above is offset by friction, which causes the wind-induced ocean current to be weaker than the surface wind. In addition, the Coriolis force acts on the ocean current, with transport 90° right of the current north of the Equator and 90° left of the current south of the Equator. The resulting force balance leads to the near-surface ocean current being directed 45° right of the surface wind north of the Equator and 45° left of the surface wind south of the Equator. In turn, this near-surface current attempts to tug the water just below it in the same direction, with this tug also weakened by friction and deflected by the Coriolis force, and so on as you move deeper within the upper ~400 m of the ocean. This characterizes *Ekman transport* and is associated with a vertically integrated net upper-oceanic transport 90° right of the surface wind north of the Equator and 90° left of the surface wind south of the Equator. For easterly trade winds along the Equator, this means that upper-ocean transport is poleward *away* from the Equator in both hemispheres! This causes colder ocean water at lower altitudes to upwell, leading to relatively cool ocean-surface temperatures along the Equator versus those at higher latitudes in the tropics.

Walker Circulation Variability

The Walker circulation's dominant variability, specifically its equatorial Pacific component, is the El Niño-Southern Oscillation (ENSO). During El Niño, equatorial-wave forcing causes the easterly equatorial trade winds to weaken or even become westerly. This weakens or even reverses lower-tropospheric convergence in the western equatorial oceans and lower-tropospheric divergence in the eastern equatorial oceans. It also weakens or reverses near-surface ocean transports, leading to anomalously warm sea-surface temperatures in the eastern equatorial Pacific and anomalously cool sea-surface temperatures in the western equatorial Pacific. The Walker circulation weakens and its ascending branch shifts eastward as a result of these forcing shifts. The opposite is true during La Niña: the easterly trade winds strengthen, leading to a stronger Walker circulation with its ascending branch remaining in the western equatorial oceans. Smaller-scale variability in the Walker circulation is driven primarily by the Madden-Julian Oscillation (MJO), a thirty- to sixty-day oscillation between convectively active and convectively suppressed conditions in the tropics.

What Drives the Walker Circulation?

The Walker circulation arises in response to concentrated equatorial warming, often found in the western equatorial ocean basins. We can use thickness principles to demonstrate how this heating distribution results in a Walker-like circulation, as we did when we first described the Hadley cell. However, a more thorough treatment (Gill 1980) provides a more-accurate representation of the circulation's fine-scale structure. This can be done using the shallow-water equations that we have discussed in prior lectures.

The shallow-water equations for this exercise, including a prescribed heating, take the general form:

$$\frac{\partial u}{\partial t} - \frac{1}{2} y v = -\frac{\partial p}{\partial x} \quad (1)$$

$$\frac{\partial v}{\partial t} + \frac{1}{2} y u = -\frac{\partial p}{\partial y} \quad (2)$$

$$\frac{\partial p}{\partial t} + \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = -Q \quad (3)$$

In these equations, a coordinate (x, y) refers to a non-dimensional distance with x measured eastward and y measured northward from the Equator; u and v are measures of the zonal and meridional wind, respectively; and p is a measure of the pressure perturbation. This proportionality makes the system given by (1) – (3) appear more accessible than if the direct relationships were used. Q is proportional to the heating rate. (1) and (2) are the momentum equations of the system and (3) is a version of the continuity equation; note that the vertical velocity w is proportional to $\frac{\partial p}{\partial t} + Q$.

In order to study the response of the system to steady warming, dissipative processes must be included to maintain the steadiness of the system. Such dissipative processes include friction (Rayleigh friction) and cooling (Newtonian cooling). These are introduced into (1) – (3) by replacing $\frac{\partial}{\partial t}$ with ε, where ε represents the equal dissipation of momentum and heat within the shallow-water system. We thus obtain:

$$\varepsilon u - \frac{1}{2} y v = -\frac{\partial p}{\partial x} \quad (4)$$

$$\varepsilon v + \frac{1}{2} y u = -\frac{\partial p}{\partial y} \quad (5)$$

(6) $$\varepsilon p + \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = -Q$$

(7) $$w = \varepsilon p + Q$$

If we assume that the prescribed ε is small and of small meridional scale, it can be shown that the εv term in (5) can be neglected. This allows us to re-write (5) as:

(8) $$\frac{1}{2} yu = -\frac{\partial p}{\partial y}$$

This equation is equivalent to saying that *u* is in geostrophic balance with the meridional pressure gradient. Thus, our system is now comprised of the momentum equations (4) and (8) and continuity equation (6).

The heating *Q* can be considered in two forms: symmetric along the Equator (represented by wavenumber zero) or asymmetric across the Equator (wavenumber one, with an antisymmetric value of *Q* on the other side of the Equator). Here, we consider the solution to the system with symmetric heating *Q* on the Equator. In our study of the monsoons, we will return to this system to consider its solution for asymmetric heating *Q* across the Equator.

The shallow-water equation solution for symmetric warming is the superposition of a full wavelength of a Kelvin wave with the cyclonic half-wavelength of an $n = 1$ equatorial Rossby wave. This solution has the following characteristics:

- Lower-tropospheric winds that accelerate and converge into the prescribed warming.
- Ascent through the prescribed warming, with weak descent primarily confined to the western edges of the $n = 1$ equatorial Rossby wave's twin cyclones.
- Negative pressure anomalies (low pressure) that are strongest with the $n = 1$ equatorial Rossby wave's twin cyclones, with weaker negative anomalies along the Equator between these cyclones where their negative anomaly is partially offset by the positive anomaly with the Kelvin wave part of the solution.
- Solution magnitudes that dampen with increasing distance away from the prescribed warming, both poleward and to the east and west.

The prescribed warming is steady, such that these structures remain stationary even as individual equatorial waves would tend to propagate away from their initial forcing. This solution is insensitive to where, zonally, the prescribed warming is placed; it can be prescribed anywhere along the Equator to get this solution.

Why does the shallow-water system produce equatorially trapped Kelvin and equatorial Rossby waves and not also mixed Rossby-gravity or inertia-gravity waves when forced by symmetric or asymmetric warming? The Kelvin and equatorial Rossby waves are more resonant, or responsive, to prescribed heating (Matsuno 1966). Thus, these waves grow faster than the mixed Rossby-gravity and inertia-gravity waves.

For Further Reading

- Chapter 3, *An Introduction to Tropical Meteorology, 2nd Edition*, A. Laing and J.-L. Evans, 2016.
- Chapter 4, *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.