Ocean dynamics: the wind-driven circulation

Weston Anderson

March 13, 2017

Contents

1 Introduction

In this set of notes we'll describe wind-driven circulations of the ocean, why we see western intensification of those currents (western boundary currents). For information on the Atlantic Meridional Overturning Circulation (AMOC) see the separate notes on climate in the Atlantic. For information on ENSO, see the separate set of notes devoted solely to ENSO and its teleconnections.

To begin with, let's remind ourselves of some of the momentum equation, which we'll be invoking throughout this section:

$$\frac{Du}{Dt} - fv = -\frac{\partial p}{\partial x} + \nu\nabla^2 u$$
$$\frac{Dv}{Dt} + fu = -\frac{\partial p}{\partial y} + \nu\nabla^2 v$$

Where the first term on the LHS is the advective term, the second term is the Coriolis force, the first term on the RHS is the pressure gradient force, and the final term is the wind stress (friction) term

In general we'll be talking about the balance between dominant terms in the momentum equation in three situations. The first is **geostrophic balance**, in which is the Coriolis force and pressure gradient terms balance against one another. The second is **Ekman transport**, which is the balance between the friction terms and the Coriolis force. The final balance is **sverdrup transport**, which refers to a balance between advection of planetary vorticity and the wind stress curl. This final balance is derived by transforming the momentum equation into a vorticity equation by taking the curl and applying continuity.

We'll generally take geostrophic flow as a given, as we do in the atmosphere. Figure 1 briefly reviews the balance of forces in geostrophic flow around a high pressure system in the ocean. As stated earlier, we're balancing the Coriolis force against the pressure gradient force:

$$-fv = -\frac{\partial p}{\partial x}$$

$$fu = -\frac{\partial p}{\partial y}$$

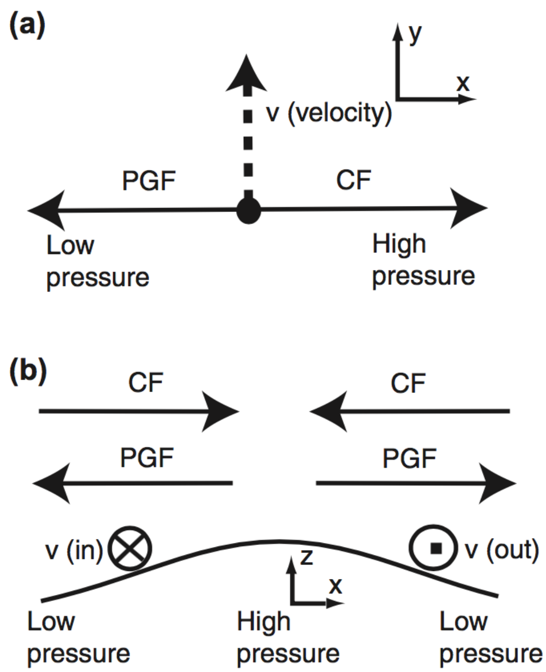

FIGURE S7.17 Geostrophic balance: horizontal forces and velocity. (a) Horizontal forces and velocity in geostrophic balance. PGF = pressure gradient force. CF = Coriolis force. (b) Side view showing elevated pressure (sea surface) in center, low pressure on sides, balance of PGF and CF, and direction of velocity v (into and out of page).

Figure 1: Figure credit: (Talley, 2011)

2 The wind driven circulation (Ekman Transport)

The Ekman layer refers to the upper 50m of the ocean that is affected by the surface wind stress. So to understand how surface wind stress affects ocean currents, we think about the depth-integrated transport over the entire Ekman layer (i.e. Ekman transport). The effect of Earth's rotation imparts two distinct features on Ekman transport: 1) the flow 'spirals' to the right with depth and 2) the net mass-transport is perpendicular to the right (in the NH) of the surface wind stress. Figure 2 demonstrates these two properties. In the southern hemisphere the spiral is reversed and the Ekman transport is to the left.

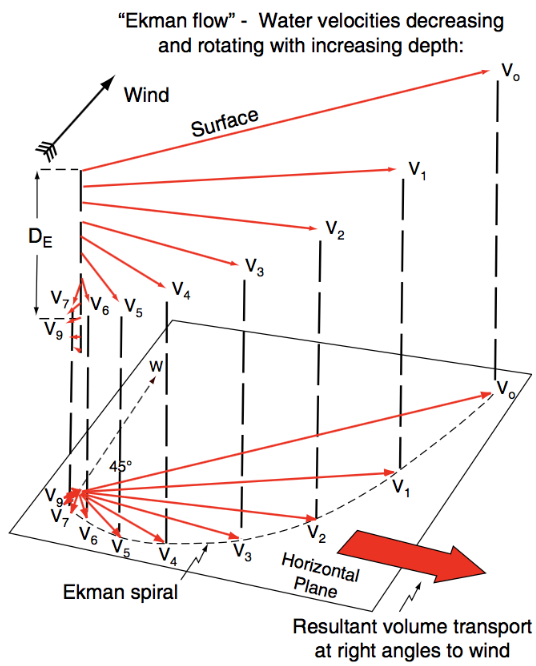

FIGURE S7.11 Ekman layer velocities (Northern Hemisphere). Water velocity as a function of depth (upper projection) and Ekman spiral (lower projection). The large open arrow shows the direction of the total Ekman transport, which is perpendicular to the wind.

Figure 2: Figure credit: (Talley, 2011)

We can apply the concept of Ekman transport in conjunction with equatorial trade winds to explain equatorial shoaling of the thermocline. When the trade winds blow along the equator, they force Ekman transport to the right in the NH and to the left in the SH, forcing divergence of surface water right along the equator, which creates upwelling (i.e. 'Ekman pumping'). When applied to along-shore flow, Ekman transport explains upwelling because it similarly creates divergence of surface waters. These features demonstrate both why eastern boundary currents are cold and nutrient rich, and why we have an equatorial cold tongue.

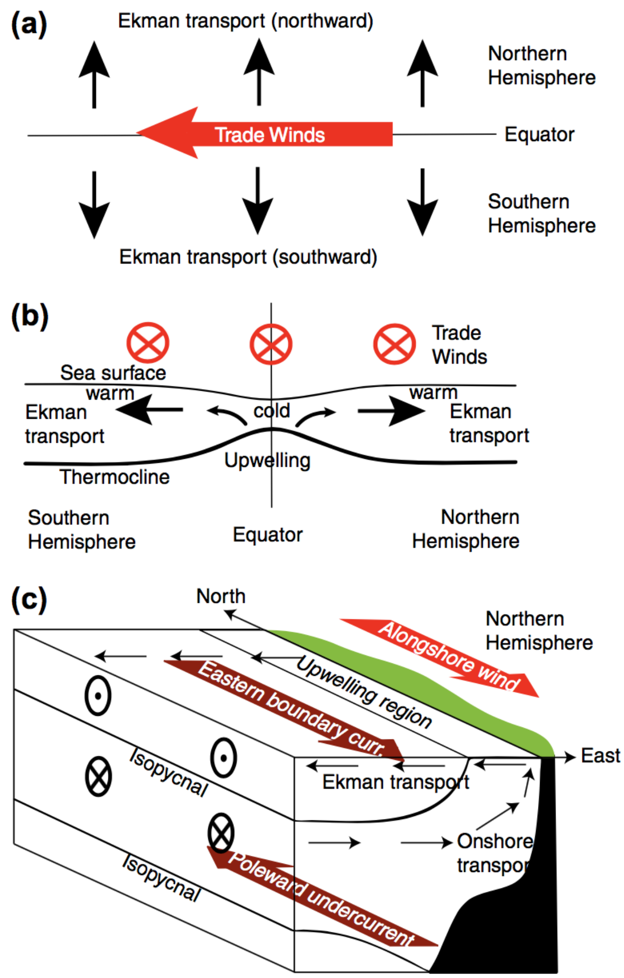

FIGURE S7.13 Ekman transport divergence near the equator driven by easterly trade winds. (a) Ekman transports. (b) Meridional cross-section showing effect on the thermocline and surface temperature. (c) Coastal upwelling system due to an alongshore wind with offshore Ekman transport (Northern Hemisphere). The accompanying isopycnal deformations and equatorward eastern boundary current and poleward undercurrent are also shown (see Section 7.9).

Figure 3: Figure credit: (Talley, 2011)

3 Sverdrup flow

In the interior of the ocean (or right near the equator) we consider the importance of Sverdrup transport, which refers to a flow resulting from a large-scale vorticity balance between the wind-stress curl and advection of planetary vorticity. To understand this balance we need to first consider potential vorticity:

$$Q = \frac{\zeta + f}{H}$$

Where H is the thickness of the fluid layer, ζ is the vertical component of the relative vorticity and f is the vertical component of the planetary vorticity (i.e. the Coriolis parameter). f is zero near the equator and maximum near the poles.

Now consider a fluid parcel that begins with no relative vorticity at a low latitude, and moves poleward. Because f increases as we move poleward the parcel must act to maintain constant potential vorticity (Q) by either: 1) inducing relative vorticity (i.e. change ζ) or 2) changing the height of the parcel (H). Relative vorticity is often small in the ocean interior, so parcels generally change H, as depicted below in Figure 4. This means that poleward-moving parcel tend to 'stretch' (increase H), while equatorward-moving parcels tend to 'compress'. Alternatively, if a parcel is 'stretched' by an outside force (e.g. Ekman pumping), then it will tend to move poleward and if it is compressed it will move equatorward.

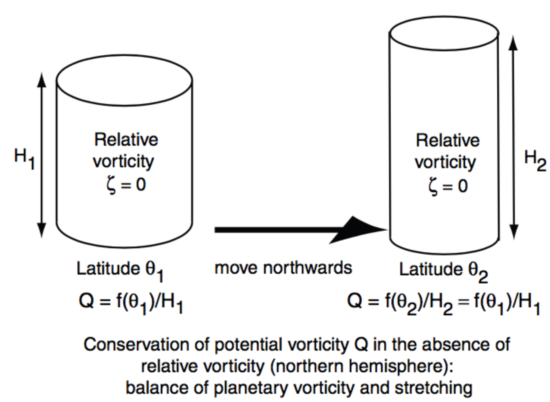

FIGURE S7.28 Conservation of potential vorticity: changes in thickness and latitude (Coriolis parameter f), assuming negligible relative vorticity (Northern Hemisphere).

Figure 4: Figure credit: (Talley, 2011)

Now if we take the geostrophic equations for the x- and y- momentum equations are used to forma. vorticity equation with $\beta = df/dy$:

$$f\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) + \beta v = 0$$

Then using the continuity equation to substitute into the parentheses:

$$f\left(-\frac{\partial w}{\partial z}\right) + \beta v = 0$$

We could alternately substitute in the wind stress curl for w in the above equation, which would demonstrate how divergent wind-stress curls create vortex stretching, which forces parcels of water to move to higher latitudes as a means of maintaining PV. Alternatively, when Ekman pumping forces compression of the water column (due to convergence) the parcels move equatorward. This is referred to as Sverdrup transport. Figure 5 conceptually ties together Ekman transport, the wind-driven gyres and Sverdrup transport.

FIGURE S7.32 Sverdrup balance circulation (Northern Hemisphere). Westerly and trade winds force Ekman transport creating Ekman pumping and suction and hence Sverdrup transport.

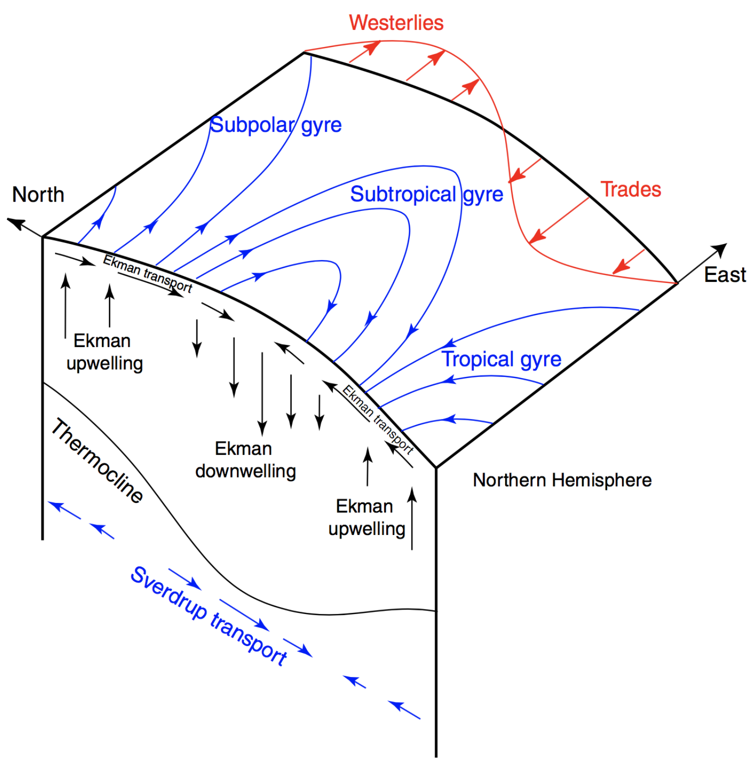

Figure 5: Figure credit: (Talley, 2011)

4 Western boundary currents (western intensification)

To understand why wind-driven currents are more intense on the western boundary of the ocean basin rather than the eastern boundary (see Figure 6), we again need to consider the conservation of potential vorticity.

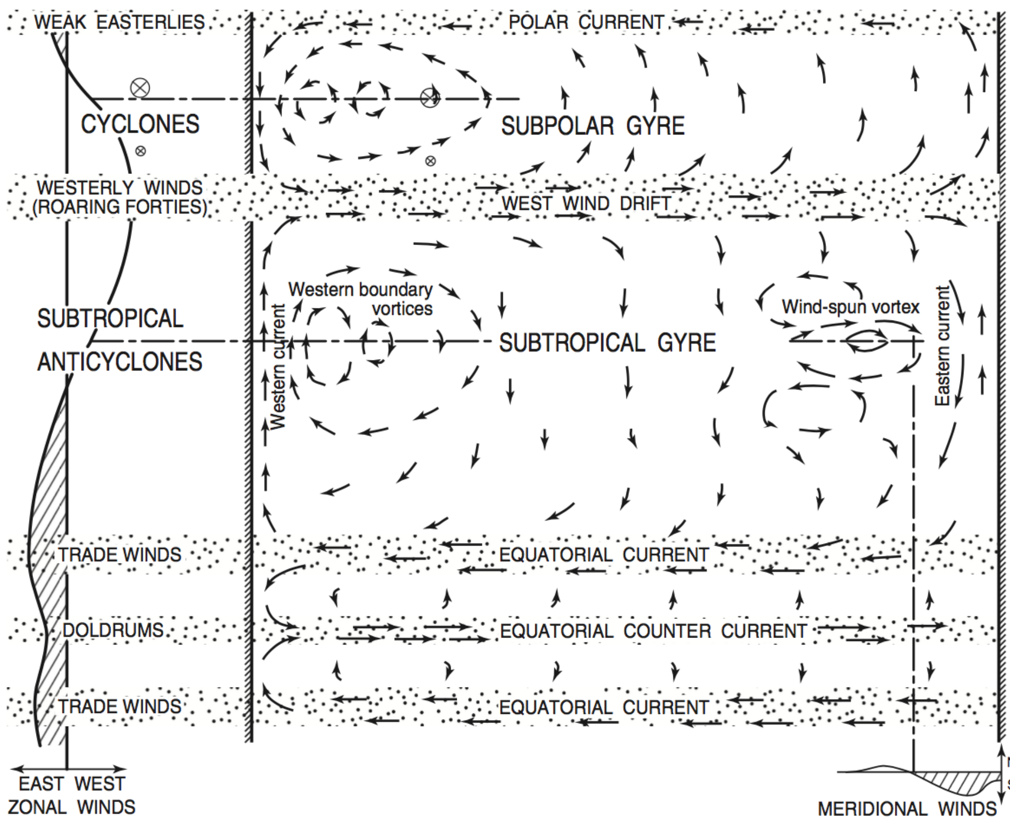

FIGURE S7.34 Munk's wind-driven circulation solution: zonal wind profiles on left and circulation streamlines in the center. *After* Munk (1950).

Figure 6: Figure credit: (Talley, 2011)

In the subtropical gyres we have Sverdrup transport to the south in the interior of the ocean, which is only partially compensated for (in a PV sense) by vortex stretching/compression. So there must be an alternative source of postive PV input into the system to balance the negative βv term of Sverdrup transport. This input is vorticity induced by boundary friction. Figure 7 demonstrates how a northward flow and a western boundary imparts positive vorticity to the system, while a northward flow and eastern boundary would impart negative vorticity. Because we need to add positive vorticity, the flow along the western boundary must intensify (as we see in western boundary currents such as the Gulf Stream and Kuroshio Current).

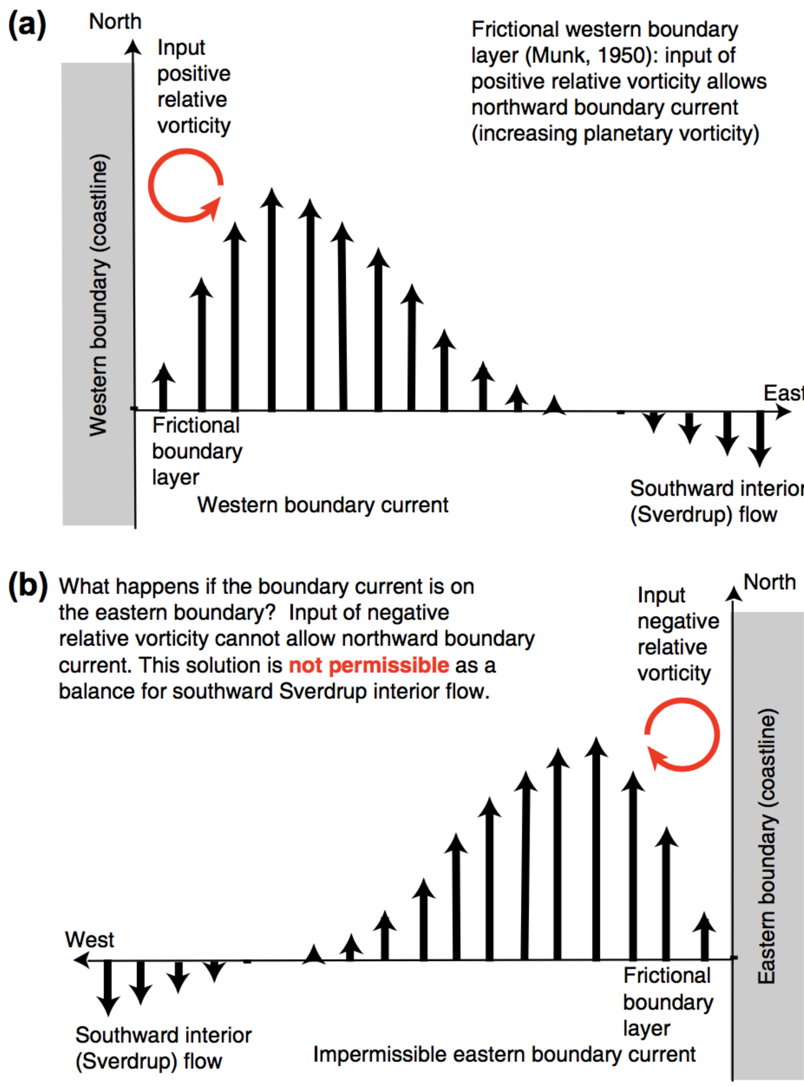

Figure 7: Figure credit: (Talley, 2011)

References

Lynne D Talley. *Descriptive physical oceanography: an introduction.* Academic press, 2011.