The wind-driven circulation

Winds exert stress on the surface, which, as we discussed before, may be estimated using the bulk aerodynamic formula

$$\vec{\tau} = \rho_{air} C_D |u| \vec{u}$$

This drives circulation in the ocean. Comparing the surface wind stress pattern and the upper ocean currents, one sees many similarities.

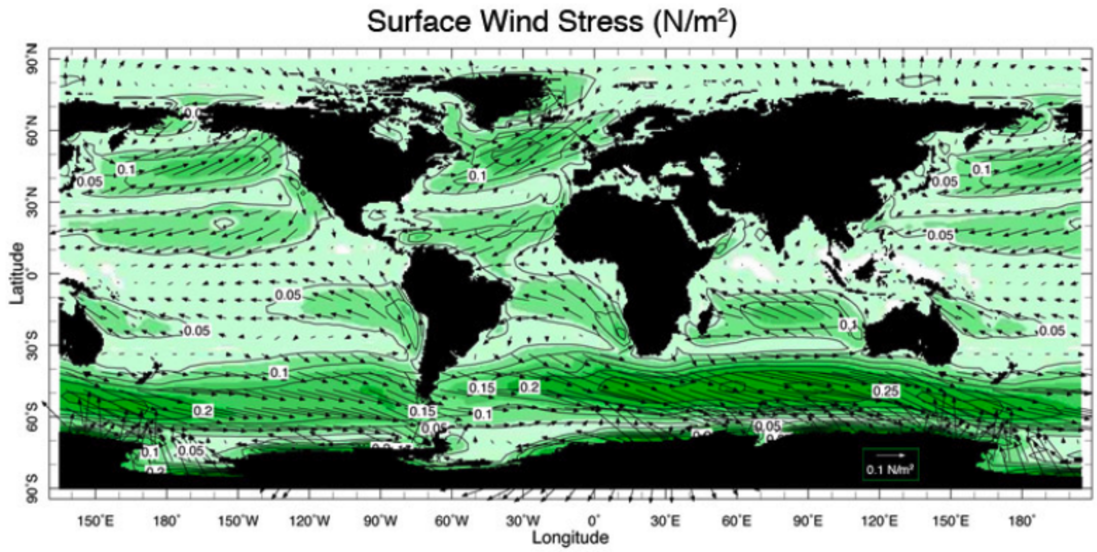
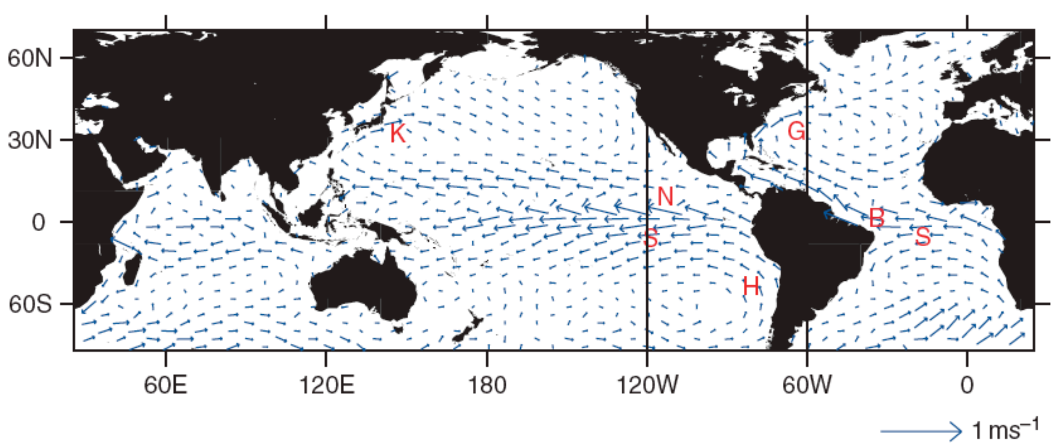
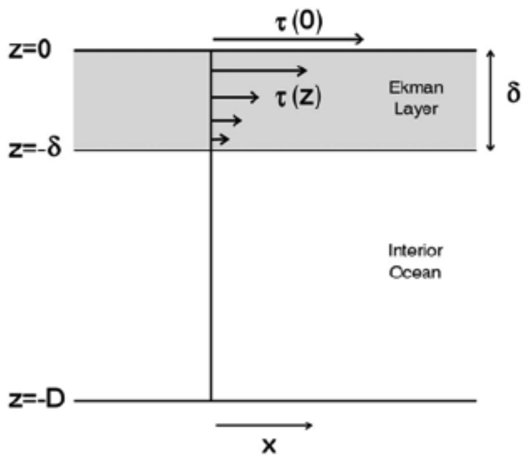


Figure 10.2: Annual mean wind stress on the ocean. A contour of 1 represents a wind-stress of magnitude 0.1 $\mathrm{N\,m^{-2}}$. Stresses reach values of 0.1 to 0.2 $\mathrm{N\,m^{-2}}$ under the middle-latitude westerlies, and are particularly strong in the southern hemisphere. The arrow is a vector of length 0.1 $\mathrm{N\,m^{-2}}$. Note that the stress vectors circulate around the high and low pressure centers shown in Fig.7.27, as one would expect if the surface wind, on which the stress depends, has a strong geostrophic component.



The explanation seems simple: where you exert an eastward stress, water flows towards the east. There are however a few issues with this. First, the stress decays quite fast with depth so the net force is only felt in the very top layer of the ocean, but the currents extend through the thermocline. Second, we are on a rotating planet, and things don't actually move in the direction that you push it. Third, the Equatorial Counter Current is one counter example to the above explanation.

Ekman layer:



The wind stress exerted at the surface decays quite quickly with depth so that the net force associated with wind stress is only significant over a shallow layer (10-100m). This is the Ekman layer. In this layer, we have

$$-fv + \frac{1}{\rho_{ref}}\frac{\partial p}{\partial x} = \frac{1}{\rho_{ref}}\frac{\partial \boldsymbol{\tau}_x}{\partial z} \quad ; \quad fu + \frac{1}{\rho_{ref}}\frac{\partial p}{\partial y} = \frac{1}{\rho_{ref}}\frac{\partial \boldsymbol{\tau}_y}{\partial z} \tag{1.1}$$

By making some assumptions about the turbulence, Ekman determined the structure of the current within this layer, known as the Ekman spiral.

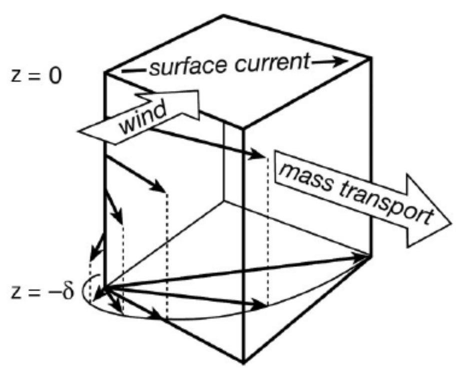


Figure 10.5: The mass transport of the Ekman layer is directed to the right of the wind in the Northern Hemisphere — see Eq.(10.5). Theory suggests that horizontal currrents, u_{ag}, within the Ekman layer spiral with depth as shown.

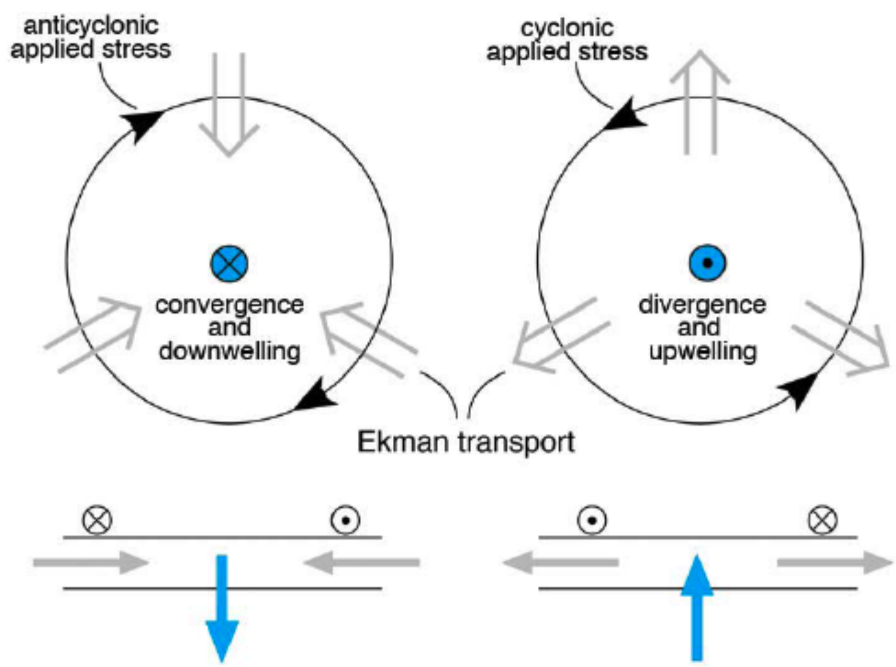
We may bypass this step by integrating across the Ekman layer to get:

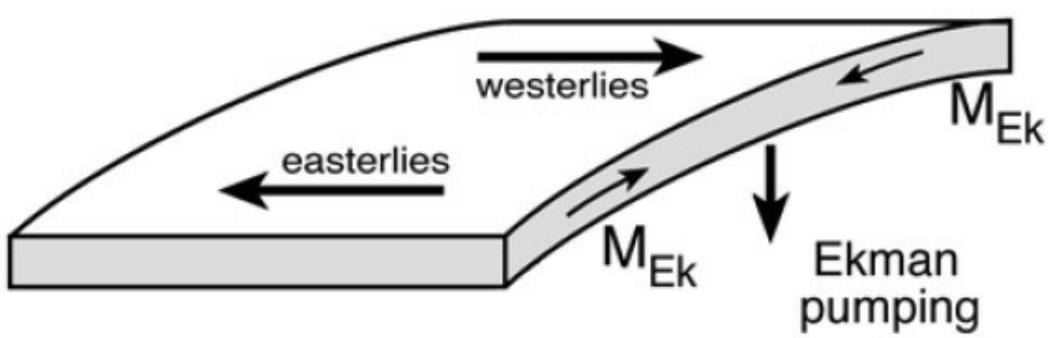
$$-f\delta V_{ag} = \frac{1}{\rho_{ref}}\tau_x$$

$$f\delta U_{ag} = \frac{1}{\rho_{ref}}\tau_y$$

where V_{ag}, U_{ag} are the ageostrophic winds averaged over the Ekman layer.

From this, we see that with anticyclonic/cyclonic wind stress, we have Ekman convergence/divergence. This is completely analogous to what we saw in the atmosphere.





Near the surface, horizontal divergence is dominated by the ageostrophic flow (geostrophic flow can be divergent when f is not a constant but that divergence is small near the surface). And this divergence/convergence depends on the curl of τ_{wind} /f. Typically τ_{wind} varies much more than f so the pattern of Ekman pumping $w_{\_Ek}$ is largely set by variations in the wind stress. Therefore, water will converge towards regions of anticyclonic wind stress and diverge away from regions of cyclonic wind stress. For the observed wind stress pattern, this would cause sea level to change by tens of meters per year.

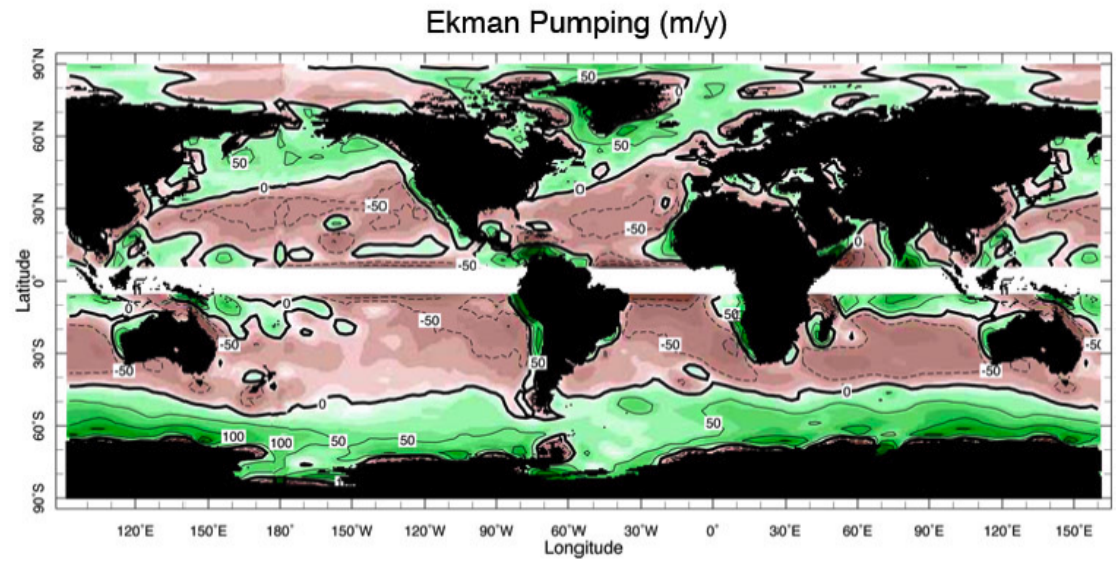


Figure 10.11: The global pattern of Ekman vertical velocity $(\mathrm{m\,y^{-1}})$ computed using Eq.(10.7) from the annual mean wind-stress pattern shown in Fig.10.2. Motion is upward in the green areas, downward in the brown areas. w_{Ek} is not computed over the white strip along the equator because $f \longrightarrow 0$ there. The thick line is the zero contour. Computed from Trenberth et al (1989) data. The broad regions of upwelling and downwelling delineated here are used to separate the ocean in to different dynamical regimes, as indicated by the colors in Fig.9.13.

We obviously don't observe the ocean surface in the subtropics to increase by 30m/yr (at least 10000 times that from global warming). Note that the Ekman layer depth doesn't increase, even though the convergence happens within this layer. This is because the Ekman layer depth is determined by the wind. When you make it too thick the lower part will no longer directly feel the wind stress. So we are increasing the height of the water column in the interior of the ocean.

What would happen now?

Water would want to flow outward. This implies downward velocity within the column. In this sense, wind stress is pumping water downward (Ekman pumping). This brings down the isopycnals. The wind stress is doing work in order to pump down warmer (or more precisely, lighter) water.

On the rotating Earth, having water flowing outward will modify its vorticity and generate anticyclonic motion. This is same as the spin-down problem that we talked about before. In other words, even though the effect of wind stress decays over the thin Ekman layer, its effect is felt down below through Ekman pumping and the divergent flow in the interior ocean.

If we continue to apply the anticyclonic wind stress, it will continue to pump water down, and water will continue to flow outward, and the vorticity will become more and more anticyclonic. (The ocean floor could balance the vorticity tendency from the Ekman pumping. But in the observed ocean, the stress at the ocean floor is quite weak).

The way to achieve steady state so that the vorticity does not become more and more anticyclonic is through meridional advection of planetary vorticity. Recall that the planetary vorticity is f, and the relative vorticity is ζ, and the absolute vorticity is $f+\zeta$. The ratio of ζ/f is also measured by the Rossby number. And in the ocean $\zeta \ll f$. We know that f increases northward. Therefore, fluid parcels at northern latitudes have larger absolute vorticity. When they are advected to more southern latitudes, because the local planetary vorticity is smaller, conservation of absolute vorticity (assuming the depth of the fluid doesn't change) implies an increase in the relative vorticity ζ.

Therefore, because of the variation of f with latitude, meridional advection provide a vorticity tendency of βv, where $\beta = df/dy$. And in steady state, this vorticity tendency balances that from the surface stress.

Considering the vorticity balance of the entire column (including both the Ekman layer and the interior ocean) and assume bottom friction is weak, we have:

$$\beta V = \frac{1}{\rho_{ref}} \left(\frac{\partial \tau_{windy}}{\partial x} - \frac{\partial \tau_{windx}}{\partial y} \right) \tag{1.2}$$

where $V = \int_D^0 v dz$ is the depth integrated meridional velocity. This is known as the Sverdrup relation, after the famous oceanographer. Indeed, ocean transports are measured in Sverdrups ($10^6 m^3 s^{-1}$). Transport of the subtropical gyre is ~50Sv. In comparison, that of the Amazon River is 0.2Sv.

Western boundary currents

The Sverdrup relation tells us that where you have anticyclonic wind stress, water transport is equatorward. In stead state, this transport must be returned poleward somewhere. Recall that the Sverdrup relation only works away from ocean boundaries (so lateral friction may be neglected). So this transport must be returned at the narrow boundaries. Why it occurs over the western boundary can be deduced by requiring vorticity balance for the whole ocean basin.

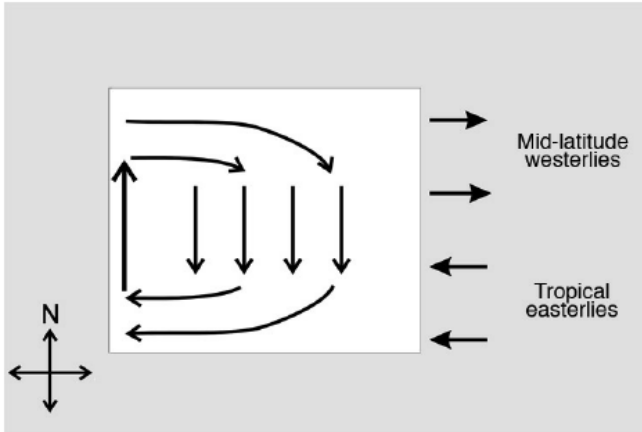


Figure 10.12: Schematic diagram showing the sense of the wind-driven circulation in the interior and western boundary regions of subtropical gyres.

So the main point is that the gyre circulation that we see are driven by wind stress, which through Ekman pumping and suction, also determines the free surface height and the horizontal density variations in the ocean. Thus, that the subtropical ocean surface is higher is not because water there is warmer. Instead, water is warmer and the ocean surface is higher there because of the wind stress.

And it's not the wind stress itself, but the curl of the wind stress that matters. To see this more clearly, consider the following two cases: both with constant density and without bottom friction, and with constant f. In the first case, we apply a rotational wind stress, and in the second case, we apply a divergence wind stress. In both cases, there is an Ekman layer within which the surface stress is felt direct and an interior.

In the first case, suppose the wind stress is cyclonic, water in the Ekman layer will flow outward (Ekman suction). (You can decompose this into a stage, where water acquires cyclonic circulation, and a stage, where due to the Coriolis force associated with the cyclonic circulation, water flows outward). This creates a low pressure at the center and drives an inward flow below the Ekman layer, which increases the pressure at the center and also accelerates water in the cyclonic sense. At steady state, we need to have force balance in both the radial and tangential directions. In the radiation direction, the balance will be between the pressure gradient forcing, which is inward, and the Coriolis forcing associated with the cyclonic flow. In the tangential direction, the pressure gradient force is zero (for constant density fluid, pressure gradient force is zero when integrated over a loop). While in the Ekman layer, the Coriolis force associated with the outward flow may balance the force due to the applied wind stress, in the interior, there is no wind stress to balance the Coriolis forcing associated with the inward flow and water will continue to accelerate until it acquires a rotation rate that is the same as that of the wind so the wind stress vanishes. Another way to think about this is that, in steady state, the radial inflow in the interior must balance the radial outflow in the Ekman layer (so water level doesn't change). When integrated over the depth of the water (both the interior and the Ekman layer), the Coriolis force in the tangential direction is zero so there is nothing to balance the impose wind stress. Therefore, the rotational component will keep increasing and eventually water will acquire the same rate of spin as the wind so there is no more wind stress (or when friction at the side boundaries balances on the torque that is being applied by the wind stress).

In the second case, there will be no divergent current, a steady state is achieved with the Coriolis force associated with a rotational flow balancing the wind stress and pressure gradient force. There are no forces in the tangential direction. Suppose the wind stress is pointed inward towards the center. In the Ekman layer, force balance in the radial direction is among pressure gradient force (outward), wind stress force (inward), and the Coriolis force associated with a cyclonic flow (outward). In the interior, there is force balance between the Coriolis force associated with an anticyclonic flow (inward) and pressure gradient force (outward). Note that thermal wind is not satisfied here because of the significant shear stress in the Ekman layer.

Another useful thought experiment is that of uniform wind stress, where you can just pile up water against one coast and achieve a steady state of no motion.

Now we can understand why there is an Equatorial Counter current that flows against the wind stress.

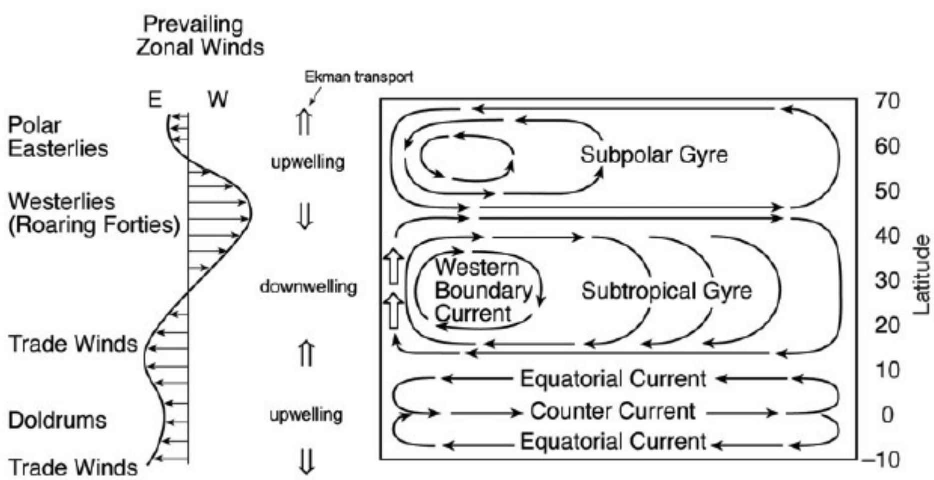


Figure 10.20: Schematic diagram showing the classification of ocean gyres and major ocean current systems and their relation to the prevailing zonal winds. The pattern of Ekman transport and regions of upwelling and downwelling are also marked.

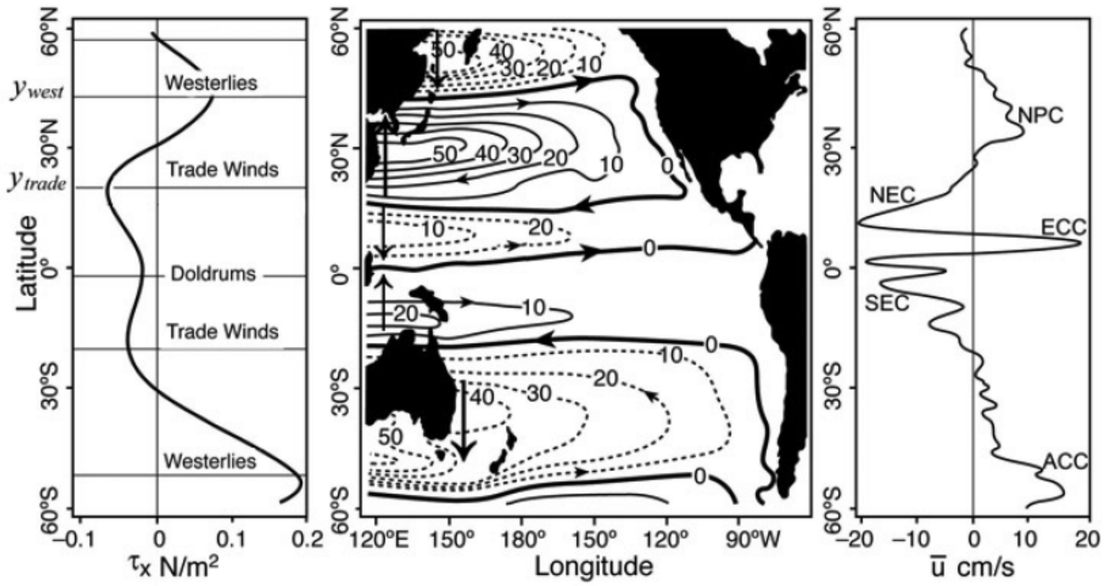


Figure 10.21: (left) The zonal-average of the zonal wind-stress over the Pacific ocean. (middle) The Sverdrup transport streamfunction (in $Sv = 10^6\,\mathrm{m^3\,s^{-1}}$) obtained by evaluation of Eq.(10.20) using climatological wind-stresses, Fig.10.2. Note that no account has been made of islands — we have just integrated right through them. The transport of the western boundary currents (marked by the $N \leftrightarrow S$ arrows) can be read off from Ψ_{west_bdy}. (right) The zonal-average zonal current over the Pacific obtained from surface drifter data shown in Fig.9.14. Key features corresponding to Fig.9.13 are indicated.

Why are there Doldrums? The explanation has to do with eddy-mean flow interaction. As we discussed before that in regions where baroclinic eddies are generated, as they propagate away, they converge zonal momentum to the source region. This balances surface friction on the surface westerly. Similarly, the tropical easterlies are maintained by waves that break in the tropical latitudes. Deeper in the tropics, however, the waves have a harder problem to get to. So there is less wave absorption and weaker surface easterlies.

Baroclinic instability

Ekman suction and pumping increase the slope of the isopycnals and eventually will lead to baroclinic instability. This explains the abundance of eddies in the ocean. Their sizes are ~100km because of the smaller deformation radius in the ocean. And their lifetime is long (L/U). Even though L is a factor of 10 smaller than that in the atmosphere, U is a factor of 100 smaller.

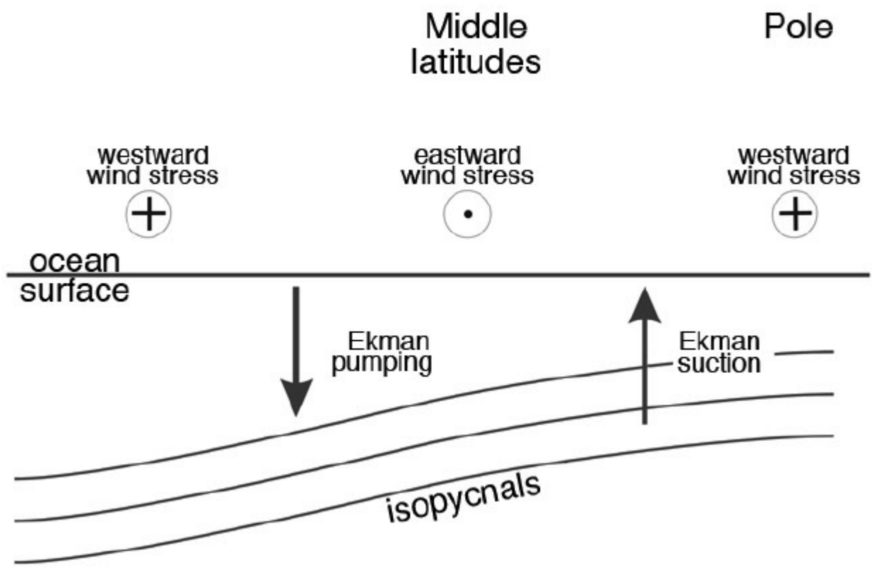


Figure 10.25: A schematic of the mechanism by which a large-scale sub-surface horizontal density gradient is maintained in the middle-latitude ocean. Ekman suction draws cold, dense fluid up to the surface in subpolar regions; Ekman pumping pushes warm, light fluid down in the subtropics. The resulting horizontal density gradient supports a thermal wind shear. Its baroclinic instability spawns an energetic eddy field which tends to flatten out the horizontal gradients.

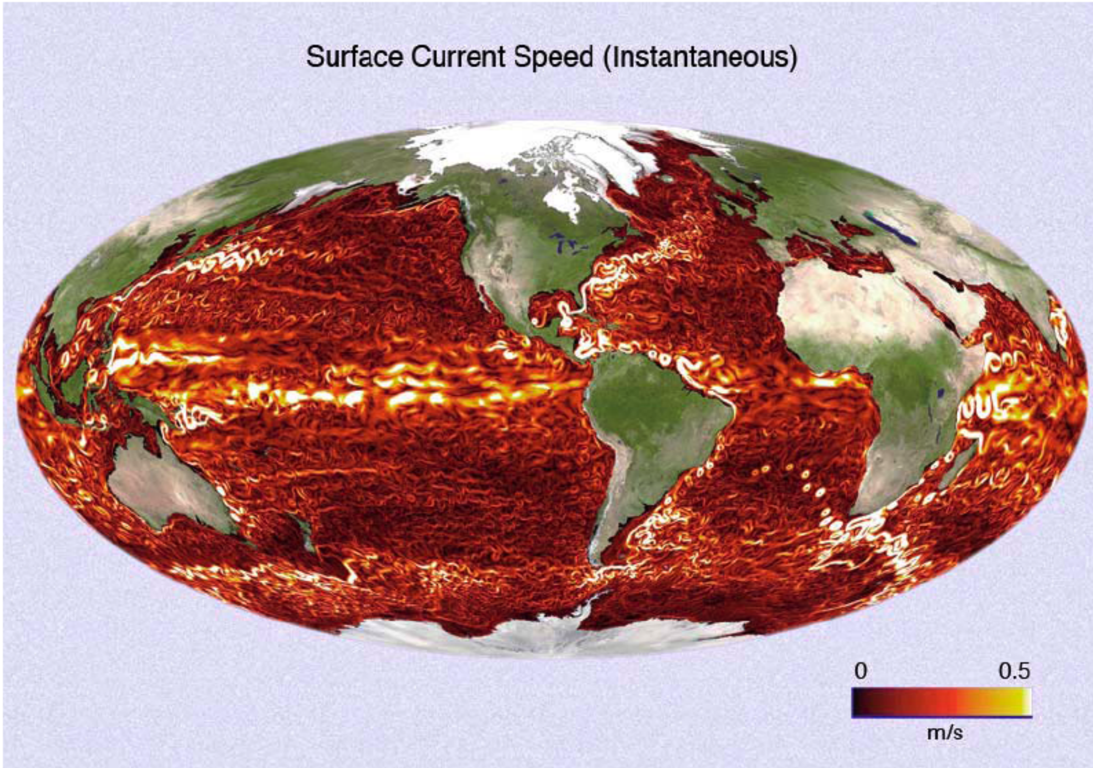


Figure 9.24: Instantaneous map of surface current speed from a global 'eddy-resolving' numerical model of ocean circulation. The scale is in units of $\mathrm{m\,s^{-1}}$. Modified from Menemenlis et al (2005).

The meridional transport by ACC cannot be explained by the Sverdrup relation as there is no zonal boundaries, and the baroclinic eddies play a central role in its meridional transport.

In the tropics, because of the mean easterly trade winds, ocean surface is higher on the west side of the Pacific Ocean and the Atlantic Ocean, and the thermocline is deeper. The pressure gradient force associated with the free surface height gradient balances the wind stress. The shallower thermocline over the eastern Pacific is essential to the El Nino phenomenon. One can also see a feedback here: when the trade winds are strong, then the thermocline in eastern Pacific becomes even shallower and hence colder. This increases the east-west SST gradient and enhances the trade winds. This is known as the Bjerknes feedback and is central to El Nino as well.

Because of the shallow thermocline over the eastern Pacific and Atlantic, coastal upwelling associated with Ekman divergence cause the cold SST (and more nutrient) in these regions. This effect is further spread over a larger region through the advection by the currents. The SST is in general colder in the southern hemisphere subtropical Pacific ocean. This is thought to be related to the orientation of the continents and their effect on coastal upwelling.