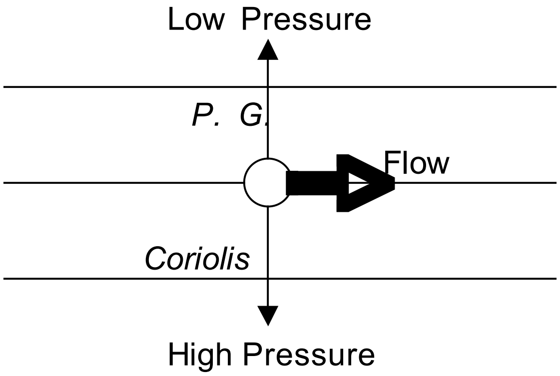
Bottom Boundary Layers

Friction occurs between the sea water in motion and the floor of the ocean, and this results in a frictional boundary layer above the sea floor. Structure of this boundary layer can be quite complex, depending on the character of the sea floor, strength of the flow and stratification. It can also be affected by the presence of suspended sediments in the water. A comprehensive treatment of the bottom boundary layer is beyond our scope (you will hear more about this in Marine Geology and Geophysics course); here we will heuristically describe the dynamics of the bottom boundary layer and its effect on the circulation in the interior.

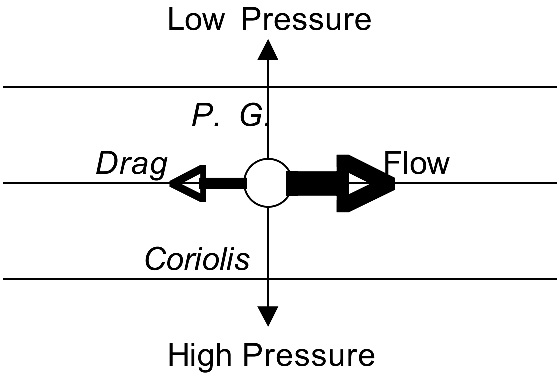
Force Balance in the Bottom Boundary Layer

In the grossest sense, we expect bottom friction to act as a *drag* on the interior flow. This means the frictional force will try to retard the flow, and thus will be directed *opposite to the direction of the flow vector*. Now let's imagine how this would affect the force balance in the bottom boundary layer.

We assume the interior flow to be in a geostrophic balance, so the flow is along isobars – lines of constant pressure. The Coriolis force thwarts the tendency for the water to flow from region of high to low pressure.



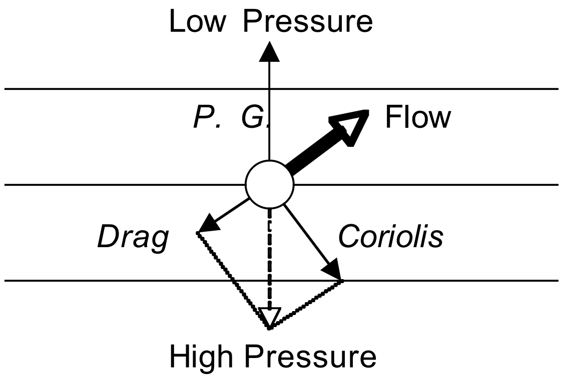
Now, add bottom friction to this balance. Since the flow is along the isobar, so would the frictional force. But this force would be opposed by no other force; so the balance would be broken, and the flow would slow down.



But! This would in turn reduces the Coriolis force, since it would have to be proportional to the speed of the flow. So the pressure gradient would outbalance the Coriolis force, and there would be acceleration towards the low pressure region. The flow would acquire a component down the pressure gradient.

A new force balance becomes possible, as depicted in the figure on the next page. The pressure gradient force is balanced by a combination of the Coriolis force and the bottom drag. This requires that the flow be no longer strictly along isobars, but have a component down the pressure gradient.

In this way, bottom friction breaks the grip of the Coriolis force and allows water to flow from high to low pressure.

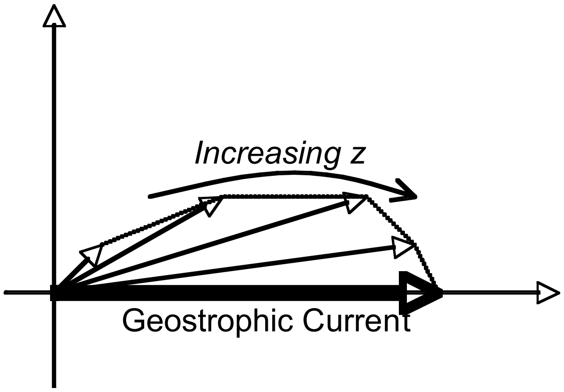


Exercise: Write down this force balance in terms of equations of motion. Convince yourself that, in the case of decreasing pressure in the y direction as depicted, the resultant flow has a positive y component. Derive an expression for this component in terms of the pressure gradient, the Coriolis parameter, and the drag coefficient.

The Bottom Ekman Layer

We can then expect that the bottom boundary layer associated with large-scale ocean circulation will contain transport down the pressure gradient. As in the case of the wind-driven surface boundary layer, the interior structure of this layer is complex; however a relatively simple solution can be obtained (not derived here), if we assume the flow to go to zero at the bottom and the turbulent viscosity coefficient A_z to be uniform. Similar to the surface Ekman layer, the solution features a spiraling current profile that eventually merges with the interior geostrophic flow, over a depth scale of $(2A_z/f)^{1/2}$, the same Ekman layer thickness.

The solution for the case of an interior geostrophic flow in the positive x direction is depicted in the figure below. Note that the flow in the spiral has a dominantly positive y component. This is as expected from the force balance argument as above. (The theoretical solution has a slight overshoot and a slight upgradient component just before the flow merges with the interior geostrophic flow.)

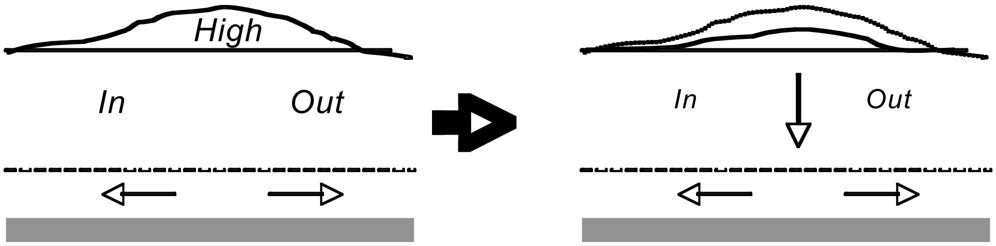


It is possible to integrate this solution in the vertical to obtain the transport in the bottom Ekman layer. The down-gradient component amounts to be $u(A_z/2f)^{1/2}$.

The Spin-Down

Consider a high pressure region due to a positive anomaly in sea surface elevation. The associated geostrophic flow will be clockwise in the Northern Hemisphere. In the bottom Ekman layer, the flow will have a downgradient component; so, there the flow is everywhere *outward* from the region of high pressure.

By conservation of mass, this implies a mass loss from the high pressure region. The sea surface will deflate, and the pressure at the center will decrease. The resultant decrease in pressure gradient will cause decrease in the geostrophic flow. An exactly analogous argument can be made for a low pressure region due to a negative sea surface anomaly.



As it turns out, for a given turbulent viscosity, this mechanism of slow-down by mass redistribution through the bottom Ekman layer is far more efficient than were the entire water column to slow down via turbulent viscosity (with the slow down time shortened by a factor of Ekman layer thickness divided by the depth of the ocean). This process of *spin-down* is the dominant process by which friction (eventually) slows down geostrophic circulation, or to limit it if it is actively driven. We will next consider an important example of an actively driven geostrophic flow, limited by the bottom Ekman layer.