**Coastal upwelling circulation**

We found that in the northern hemisphere, the transport in the surface Ekman layer is to the right of the wind. At the bottom, there is a flow in the boundary layer that is to the left of the interior flow. Let's quantify the consequences of these boundary layer flows in two examples in coastal upwelling. (Next week we will look at the wind-driven circulation.)

Upwelling is the process by which deep water is brought to the surface. It occurs wherever there is a divergence in the flow at the surface. Continuity (mass conservation) requires upward vertical flow to replace the water lost by surface divergence. The divergence can occur as the result of spatially varying winds or at a land boundary. For a uniform flow to the south along the coast and uniform winds, the maximum upwelling occurs at the coast. What feeds the upwelling and what are the consequences for the interior flow? For simplicity, we will assume that there are no alongshore gradients in any of the quantities.

z

Constant density

wind

N

E

x
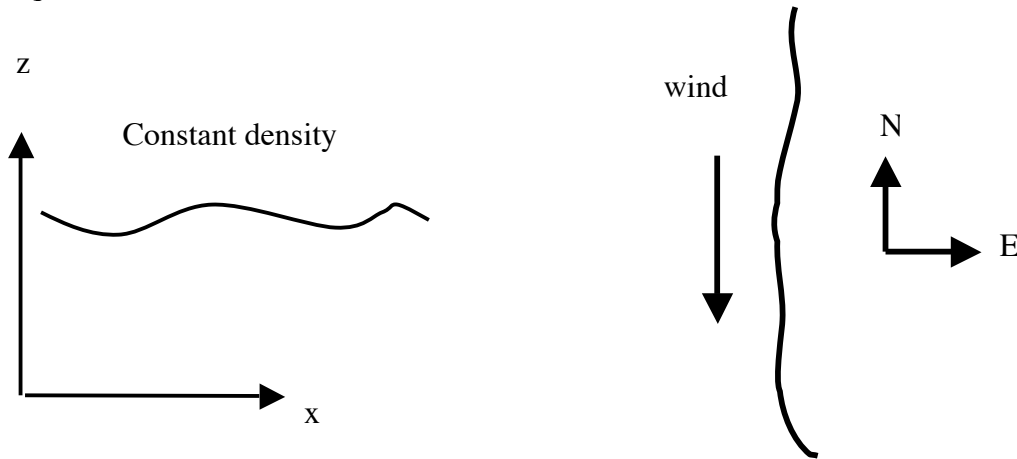
1)   After the wind starts to blow, the Ekman layer is set up.

The Ekman layer takes about a day to spin up (actually the timescale is ~ 1/f) with net transport to the right (N. Hemi) of the wind, away from the coast.

Alongshore wind stress (out of the page)

z

x
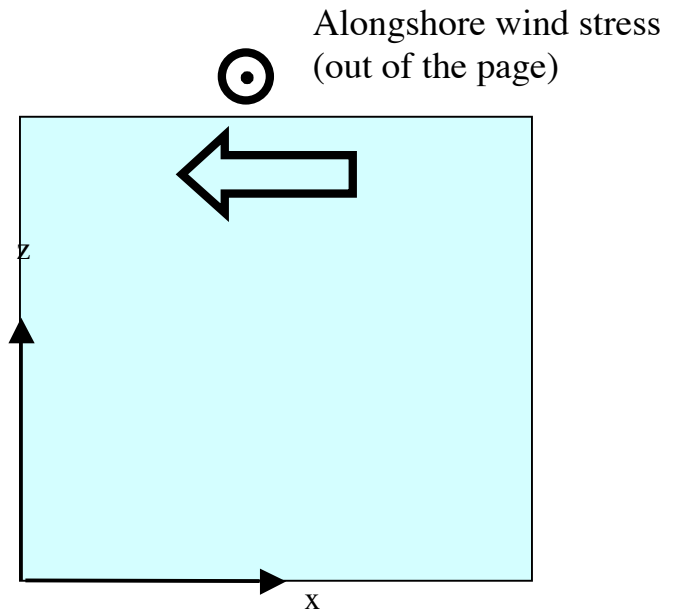
2)   The offshore flow causes a divergence, dropping sea level near the coast. The lowered sea level creates a pressure gradient perpendicular to the coast. To conserve mass, there must be an upward vertical velocity along the coast.

What is the momentum balance in the interior of the flow? Note that the alongshore pressure gradients are zero (by assumption). The cross shore momentum balance will just be geostrophic

$$-fv = -\frac{1}{\rho}\frac{\partial p}{\partial x}$$

The alongshore momentum balance will be

$$\frac{\partial v}{\partial t} + fu = 0$$

As the water is carried offshore, the pressure decreases nearshore (dp/dx decreases). This generates an alongshore geostrophic flow (to the south) that accelerates, coincident with a cross shore flow in the interior that feeds the upwelling. As the alongshore flow accelerates, an onshore bottom boundary layer transport also develops.

3) The flow finally reaches a steady state when the bottom stress matches the wind stress. Another way to think about this is that the mass transport in the bottom boundary layer balances the mass transport in the top Ekman layer. We can get an estimate for the strength of the flow by matching the wind stress and the bottom boundary stress

$$\tau_{wind} = \rho C_D u^2$$

For a wind-stress of 0.1 pascals (N/m^2), we have an alongshore flow of 0.31 m/s. (Recall that for the bottom boundary layer, we use the density of seawater.)

What happens at the bottom? There is a bottom Ekman layer. You can apply everything you know about surface Ekman layers if you look at the ocean from the bottom up, in a coordinate frame moving with the flow. The bottom stress opposes the flow (northward in this example), and the bottom Ekman layer transport is to the right of the bottom stress. The sense of the Ekman spiral is reversed from that in the surface layer, i.e., the Ekman transport is to the left of the flow, or in this case, towards the coast.
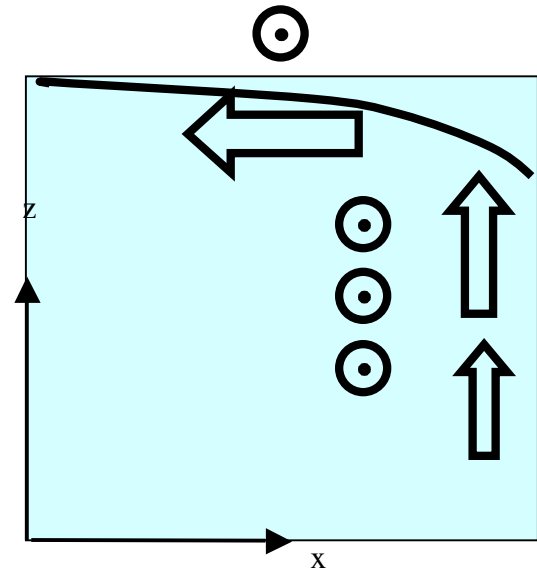
At the coast, when the convergence in the bottom Ekman layer supplies enough water to balance the offshore flow in the surface Ekman layer, a steady state is reached. Sea level ceases to drop at the coast and the alongshore geostrophic flow stops accelerating.
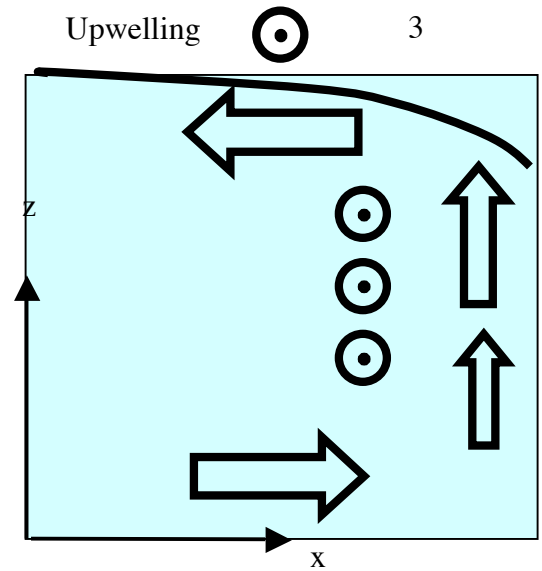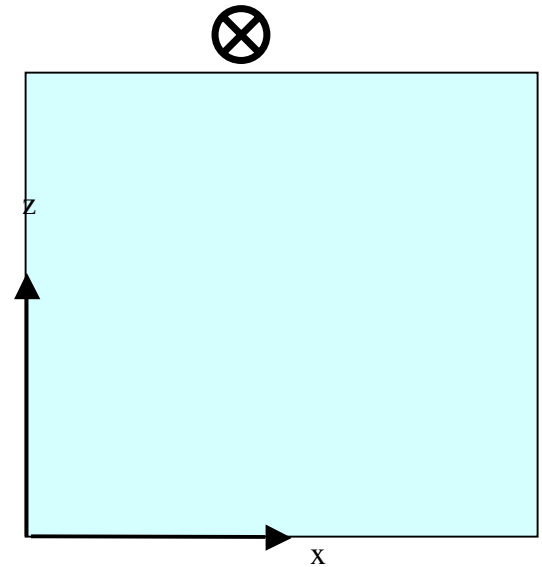
In-Class Problem:

What happens when the wind blows the other direction?

Repeat the arguments above for this direction and sketch in the circulation pattern and the sea surface.

What is the direction of the Ekman transport? What happens right at the coast? Is there vertical motion? Why?

Is there convergence or divergence in the flow field? What happens at the bottom?

Upwelling in a Baroclinic System

Now let's think about the effects of stratification and a more realistic continental slope and shelf break. The slope of the isopycnals is in thermal wind balance with the equatorward geostrophic flow. With sustained wind-induced upwelling, a temperature front forms some distance offshore, geostrophically associated with a coastal jet (in the neighborhood of 1 m/s in the California Current.)

Along the west coast of North America, this upwelling is enhanced after the spring transition, when the wind changes from southerly to northerly. Often a front forms near the shelf break, and the resulting current becomes unstable (via something called **baroclinic instability**) and eddies and filaments form.

The accompanying figure is of sea surface temperature (SST) for the California Current taken from an infrared radiometer on a satellite.

Also see the web site at http://bonita.mbnms.nos.noaa.gov/sitechar/phys21.html for more pictures of satellite sea surface temperature associated with these eddies.

In a stratified system, isopycnal slopes will tend to reduce the surface pressure gradient with depth (i.e., the flow is baroclinic.) The geostrophic flow changes with depth and we have to modify our ideas about what happens in the bottom boundary layer. The upwelling system draws its water from isopycnals at

Figure 3: 1-km AVHRR SST for June 02, 2003 from Ted Strub's archive at http://coho.coas.oregonstate.edu/globec/avhrr.html.

100-200 meters depth. Offshore of the shelf break, this subsurface onshore flow is geostrophic. Thus, in the open ocean there is a north-south pressure gradient with high on the right (or more generally, on the equatorward side.) When this flow begins to encounter the bottom, bottom stress becomes important, resulting in a flow partially down the pressure gradient in the *poleward* direction, called the **eastern boundary undercurrent**. It is typically found near the shelf break, over the continental slope. For further description of this see Knauss page 143-146 and associated pictures.
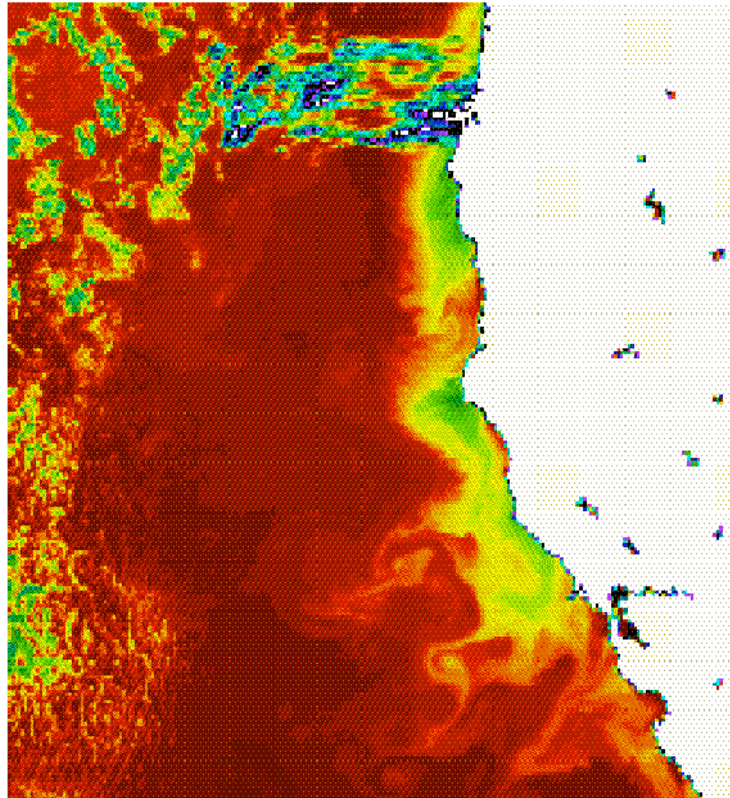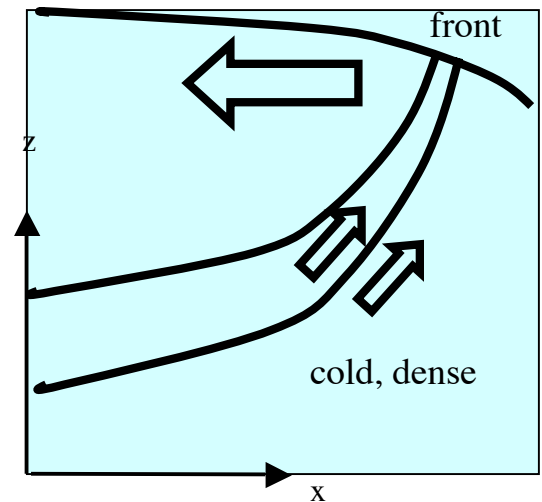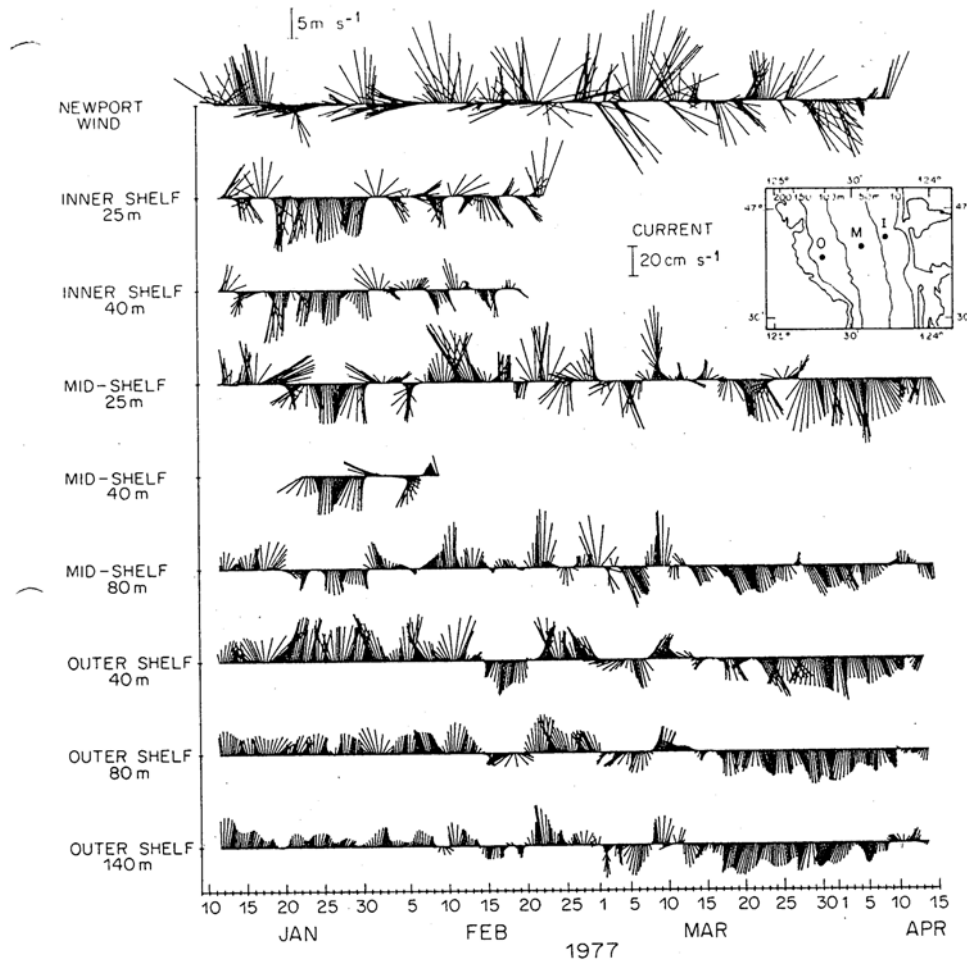
For the coast off Washington, the spring transition in the wind field happens in mid-March and can be seen in the following figure from Hickey (1988), both in the winds and in the alongshore current structure.

Fig. 2.6 Selected subtidal vector velocity time series on a section <u>across</u> the Washington shelf during winter-spring 1977. Data have been rotated into a local isobath frame of reference. Mooring locations are shown on the inset map. Wind at Newport, Ore. (~45° N) is also shown. The data demonstrate the existence of significant cross-shelf and vertical differences during winter and early spring.

In the 1980s a large experiment was conducted in the California Current off northern California to examine the sizes of the terms in the momentum equations: the Coastal Ocean Dynamics Experiment (CODE). The two figures below were taken from Huyer (1984) and Kosro (1987) articles in the CODE volume of articles, also published in JGR. The first shows the relationship between the isotherms and the winds: warm water moves offshore in response to downwelling-favorable winds and onshore when winds decrease or reverse.
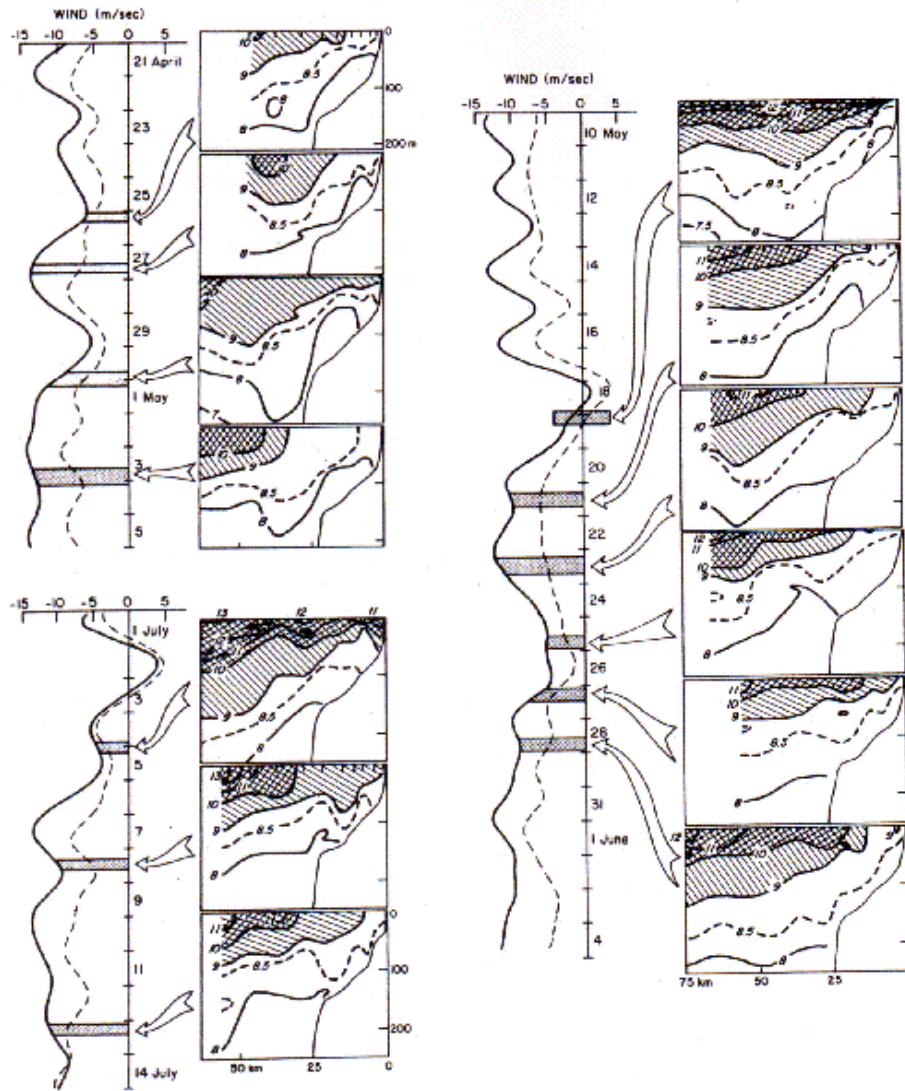
FIG. 5. Sequences of repeated temperature sections with the low-passed alongshore component of the wind at Buoy 46013 (solid) and at Pt. Arena (dashed), during periods of variable winds: 21 April–5 May, 10 May–4 June and 1–14 July 1981.