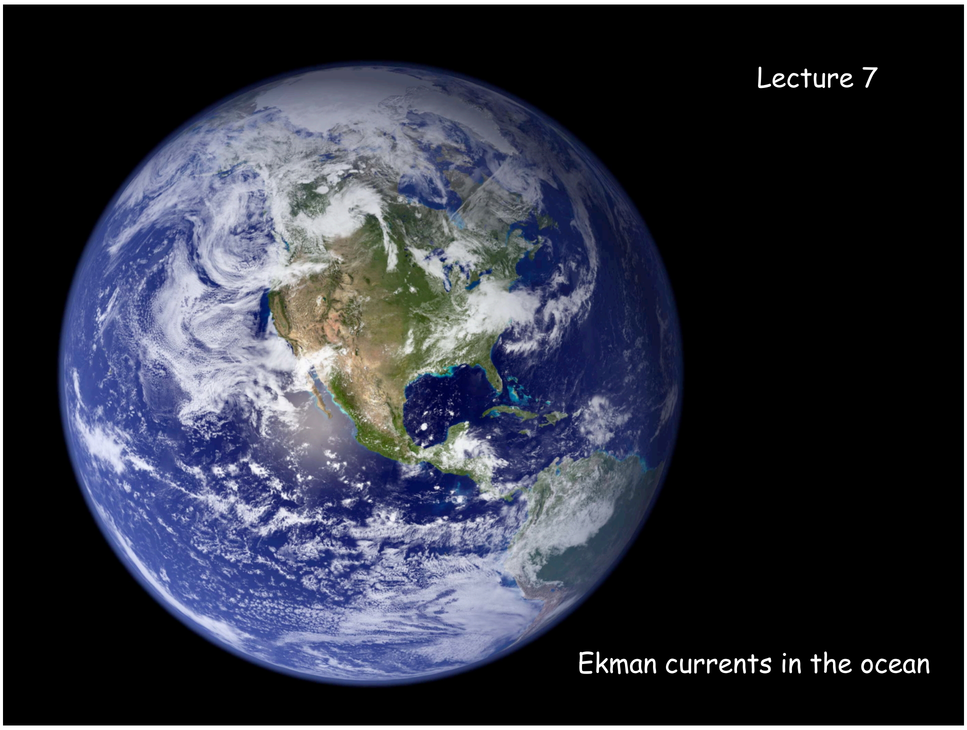
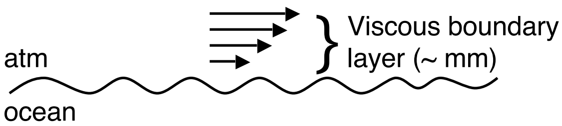


7.1 Surface wind stress

Wind stress τ_s = force per unit area exerted by the atmosphere on the ocean

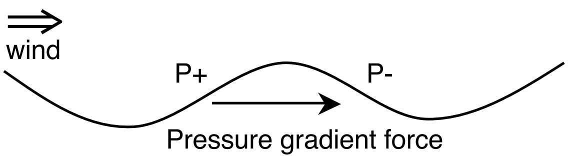
Wind stress consists of two components:

• viscous shear stress



Transfer of momentum downwards by moving air molecules.

• form stress



Wind pushes against surface waves

⇒ acceleration of fluid parcels.

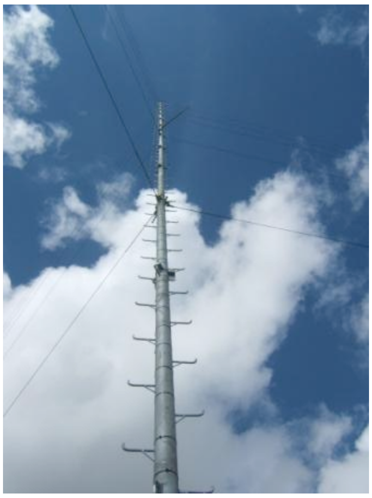
A bulk formula relates the net wind stress to the wind speed:

$$\tau_S = C_D \, \rho_A \, u_{10}^2$$

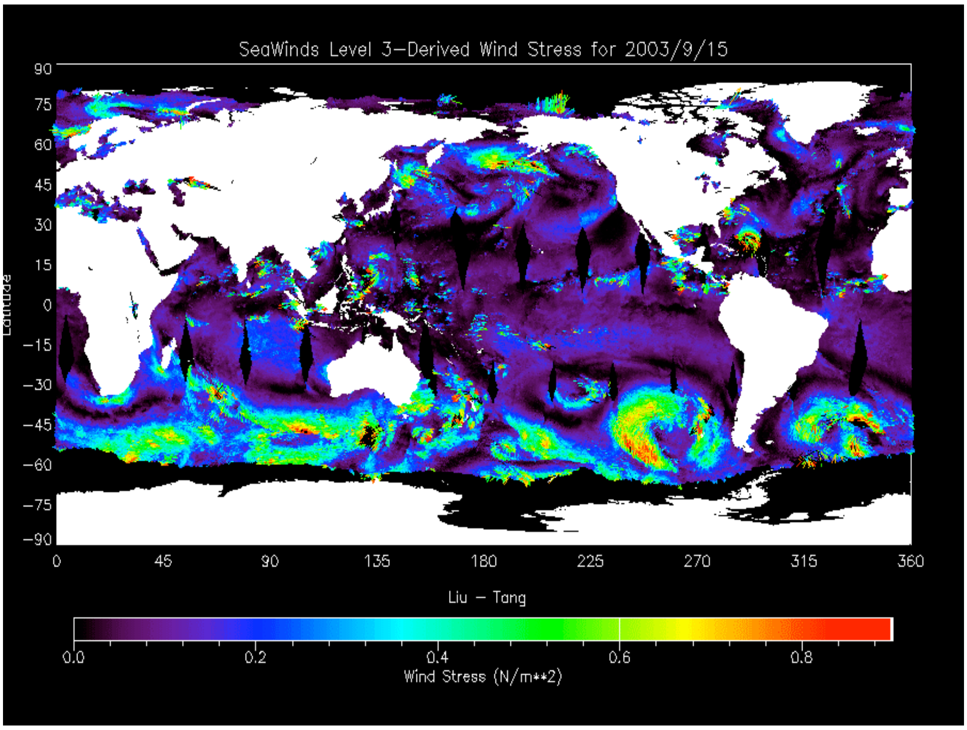
drag coefficient (~ 1.45 x 10^{-3})

density of air (~ 1.2 kg m^{-3})

wind speed at 10m above the sea surface



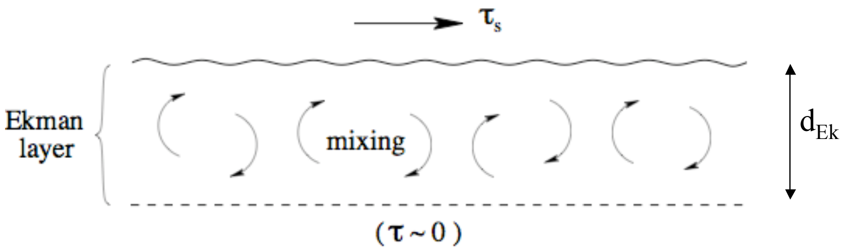
In practise C_D depends on the wind speed and the surface waves.



7.2 The Ekman Layer

Vagn Ekman (1874-1954)

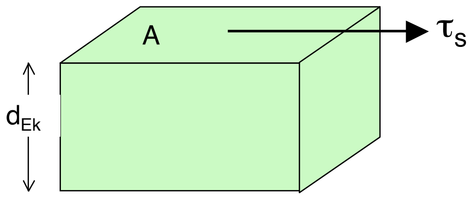
Wind stress directly accelerates fluid parcels lying at the sea surface. However, turbulence within the surface ocean effectively redistributes the momentum imparted by the wind over a layer of finite depth (typically 30-100m).



We will assume that the acceleration due to wind stress is uniform throughout the depth of the Ekman layer (d_{Ek}).

7.3 Ekman transports

Consider an element of the Ekman layer with surface area A:



Net force acting on element $= \tau_s A$

Mass of element $= \rho_0 d_{Ek} A$

$$\therefore \text{ Mean acceleration due to wind stress } = \frac{\tau_s}{\rho_0 d_{Ek}}$$

Within the Ekman layer, we need to modify geostrophic balance to include the acceleration by the wind stress:

$$-fv + \frac{1}{\rho_0}\frac{\partial P}{\partial x} = \frac{\tau_s^x}{\rho_0 d_{Ek}}$$

$$fu + \frac{1}{\rho_0}\frac{\partial P}{\partial y} = \frac{\tau_s^y}{\rho_0 d_{Ek}}$$

It is helpful now to separate the velocity into two components, a geostrophic velocity and an Ekman velocity:

$$\mathbf{u} = \mathbf{u_g} + \mathbf{u_{Ek}}$$

$\mathbf{u_g}$ is defined through geostrophic balance as the portion of the velocity which (when multiplied by f) balances the pressure gradient force. Therefore f $\mathbf{u_{Ek}}$ must balance the wind stress term:

$$u_{Ek} = \frac{\tau_s^y}{\rho_0 \, d_{Ek} \, f}$$

$$v_{Ek} = \frac{-\tau_s^x}{\rho_0 \, d_{Ek} \, f}$$

Integrating this over the Ekman layer ($U_{ek} = u_{Ek} d_{Ek}$ and $V_{ek} = v_{Ek} d_{Ek}$) gives the depth-integrated Ekman transport velocity:

$$U_{Ek} = \frac{\tau_s^y}{\rho_0 \, f}$$

$$V_{Ek} = \frac{-\tau_s^x}{\rho_0 \, f}$$

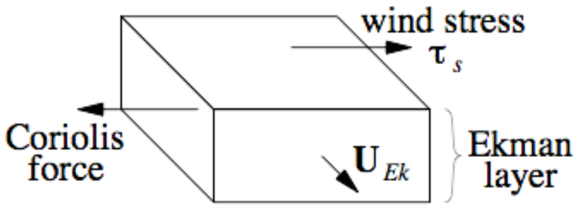
(units $m^2 \, s^{-1}$)

Hence the winds drive an Ekman transport that is directed at a right angle to the surface wind stress. It is directed to the right of the wind in the northern hemisphere and to the left of the wind in the southern hemisphere.

$$U_{Ek} = \frac{\tau_s^y}{\rho_0 \, f}$$

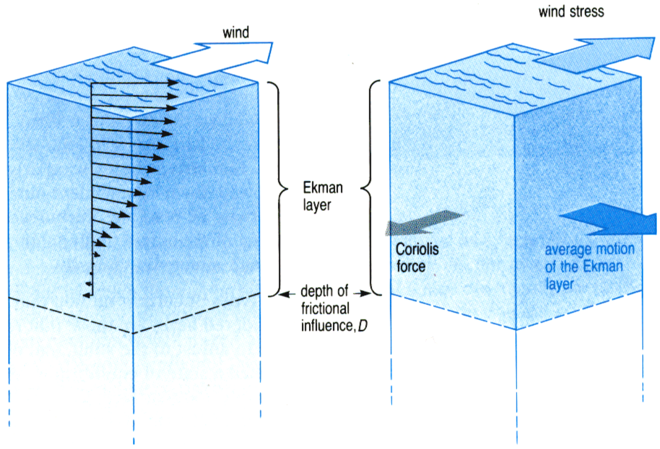
$$V_{Ek} = \frac{-\tau_s^x}{\rho_0 \, f}$$

There is a force balance between the surface wind stress and the depth-integrated Coriolis force within the Ekman layer.



Note that the Ekman transport is independent of the vertical structure (and even the depth!) of the Ekman layer.

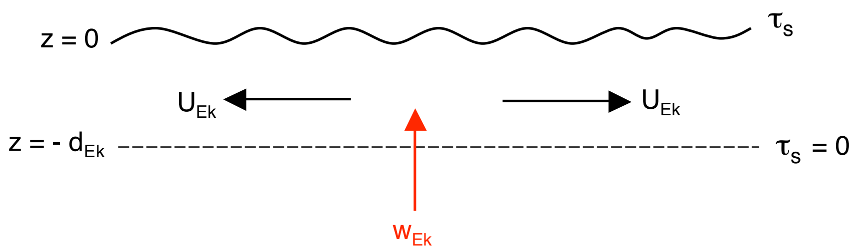
In fact, the theoretical Ekman layer has a rich vertical structure, with velocities spiralling and decreasing with depth…



The idealized profile is rarely observed in the turbulent open ocean. Our interest is in the net motion within the Ekman layer, which is independent of these details.

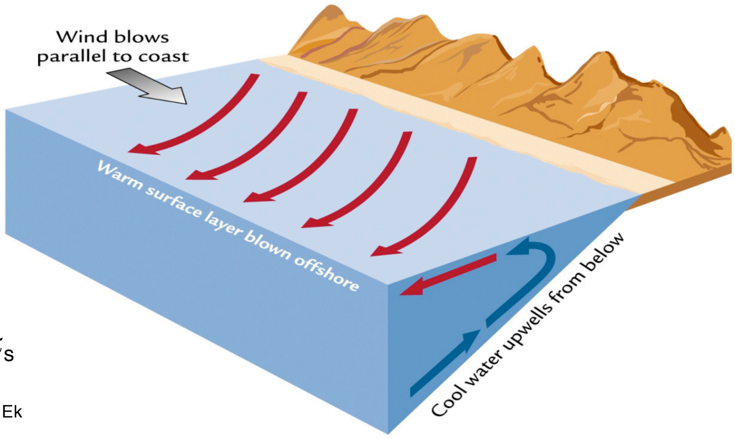
7.4 Ekman Upwelling

If surface Ekman currents converge or diverge, this results in upwelling into or downwelling out of the Ekman layer.



Examples

(a) Coastal upwelling



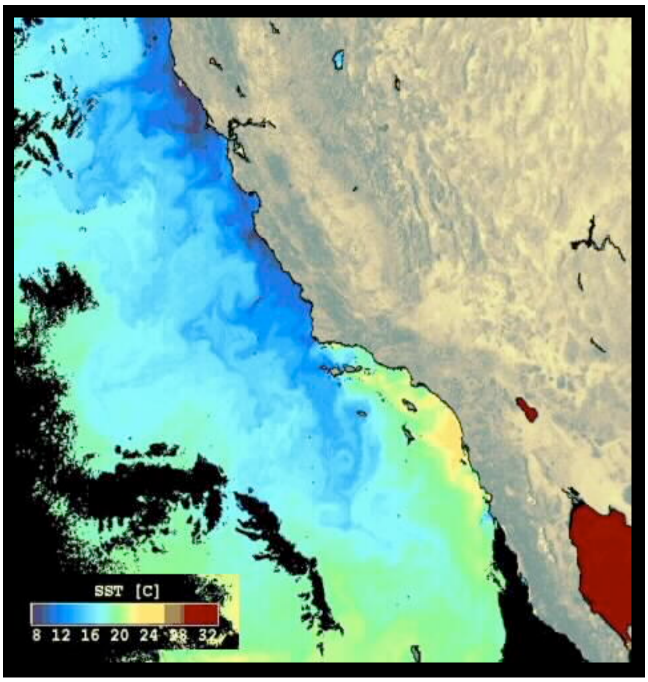
Equatorward τ_s

⇒ offshore U_{Ek}

⇒ coastal upwelling

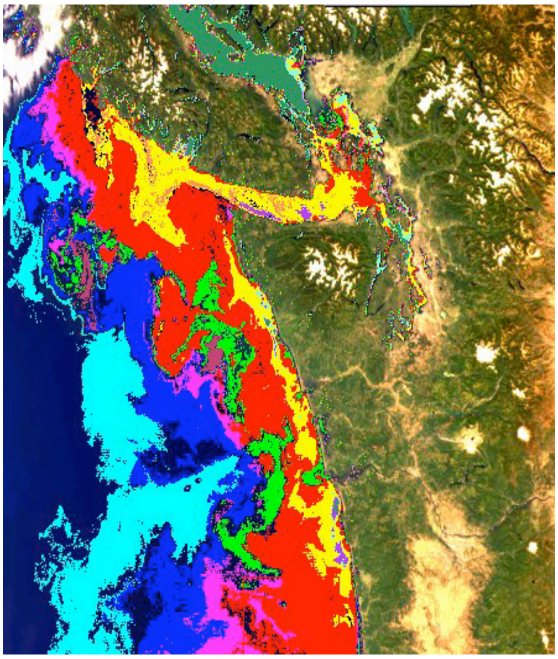
⇒ cold nutrient-rich water brought to the surface

⇒ many major fisheries located on eastern margins of ocean basins

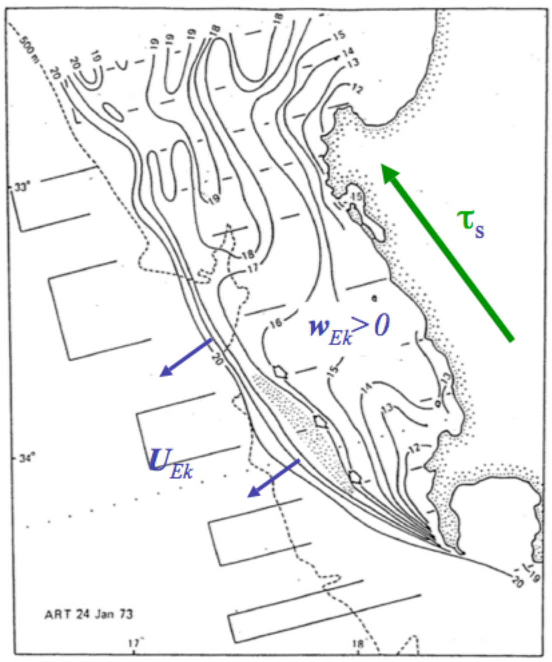


Satellite-derived sea surface temperature off California

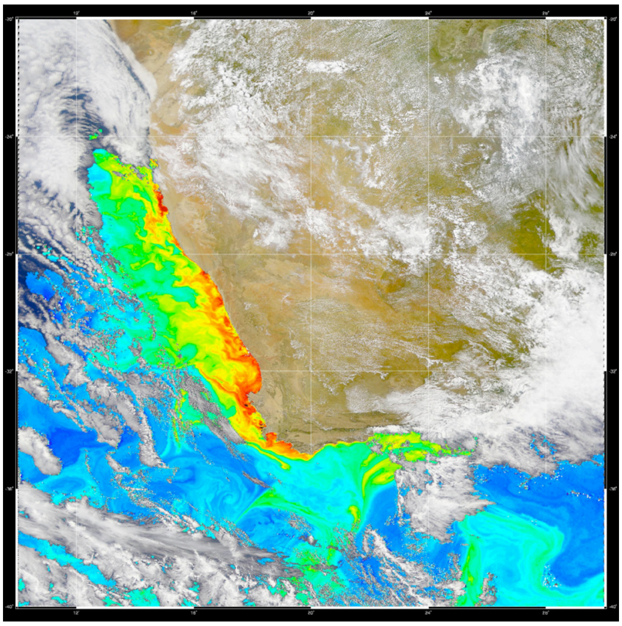
Upwelled water is nutrient rich.



Coastal Zone Colour Scanner (CZCS) image showing phytoplankton abundance off the west coast of Vancouver Island and Washington.



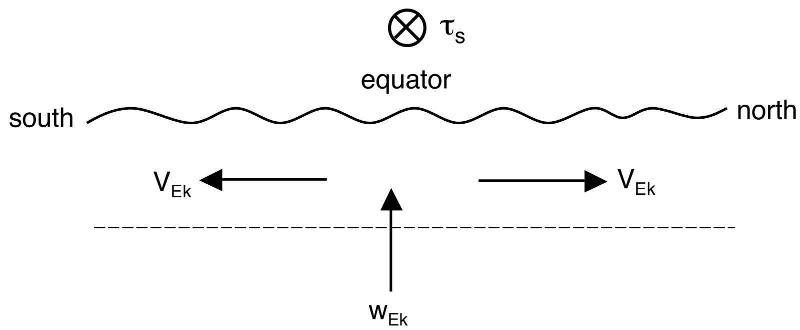
SST off west coast of Africa (Gill 1982)



Surface chlorophyll off west coast of South Africa

seawifs.gsfc.nasa.gov

(b) Equatorial upwelling

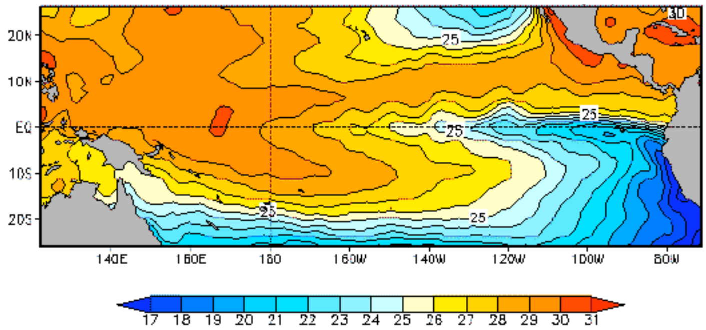


Easterly trade winds

$\Rightarrow$ $V_{ek} > 0$ north of equator and $V_{ek} < 0$ south of equator

$\Rightarrow$ Equatorial upwelling

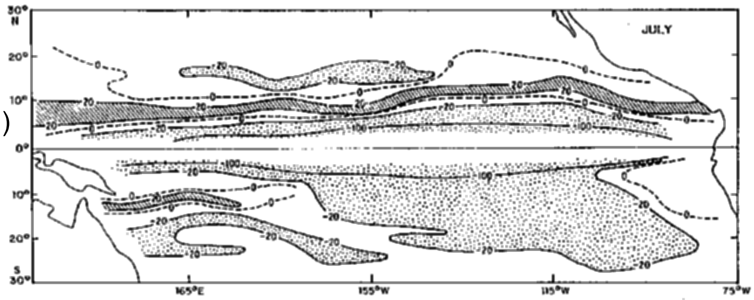
e.g this upwelling is the cause of the cold pool in the equatorial Pacific.

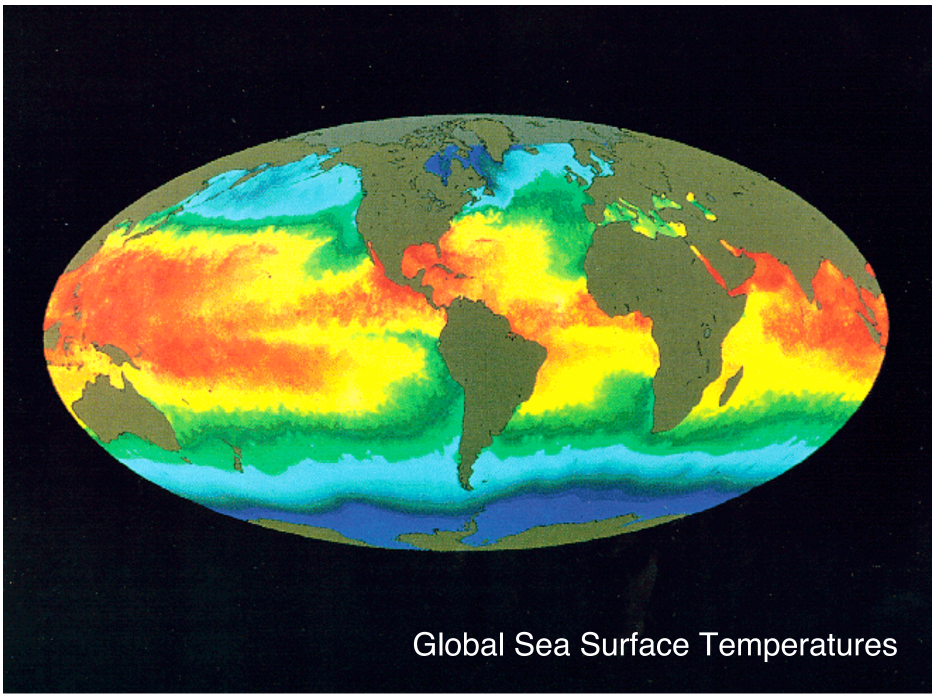


Tropical Pacific sea surface temperatures (°C)

Ekman downwelling (units 10^{-7} ms^{-1})

Gill (1982)





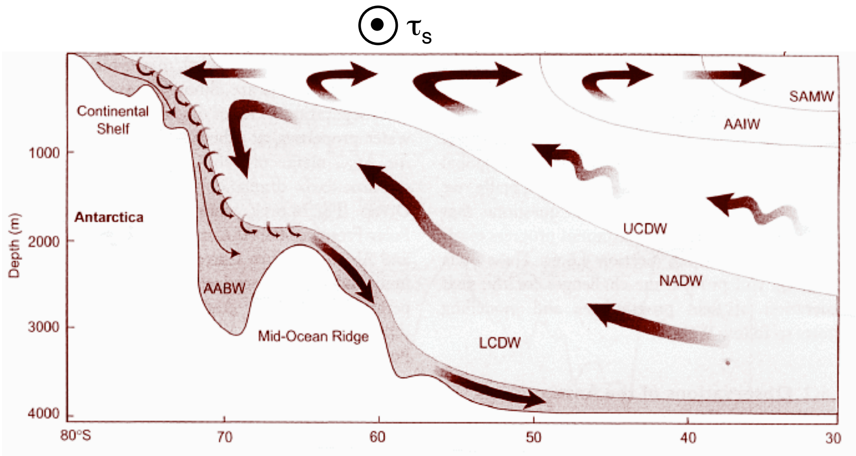
Global Sea Surface Temperatures

(c) Antarctic Circumpolar Current

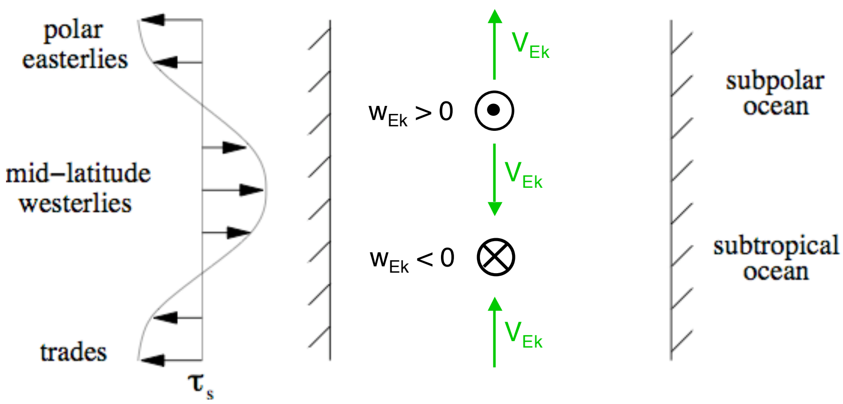
Westerly winds over the Southern Ocean

⇒ northward Ekman transport

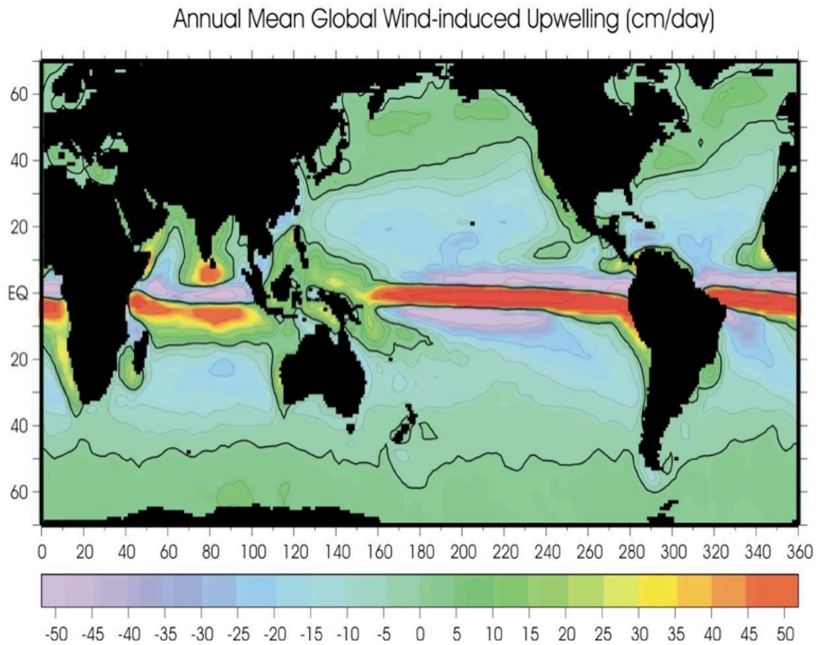
⇒ Ekman upwelling that tilts isopycnals, giving a strong geostrophic current



(d) Open ocean



In a typical ocean basin, convergence and divergence of Ekman velocities leads to Ekman upwelling in the subpolar ocean and Ekman downwelling in the subtropical ocean. This sets the interior ocean below in motion…



Xie and Hsieh (1995)

Summary of key points

- Surface winds input momentum from the atmosphere into a surface Ekman layer of the ocean.
- Horizontal Ekman transports are generated at a right angle to the surface winds through a balance between wind stress and the Coriolis force.
- Where the Ekman transports converge or diverge, vertical motion is generated through the base of the Ekman layer.

Ekman transport velocity ($m^2\ s^{-1}$)	$$U_{Ek} = \frac{\tau_s^y}{\rho_0 f}$$ $$V_{Ek} = \frac{-\tau_s^x}{\rho_0 f}$$

Exercise: Ekman transports

Assume that an easterly wind of 6 ms^{-1} blows above the equatorial Pacific.

- Estimate the surface wind stress
- Deduce the magnitude and direction of the Ekman transports at 4^oS and 4^oN in Sverdrups.
- Hence estimate the net volume of water upwelling between 4^oS and 4^oN.
- Estimate a mean value for the Ekman upwelling velocity between 4^oS and 4^oN, expressing your answer in meters per year.