Dynamics of the Ocean

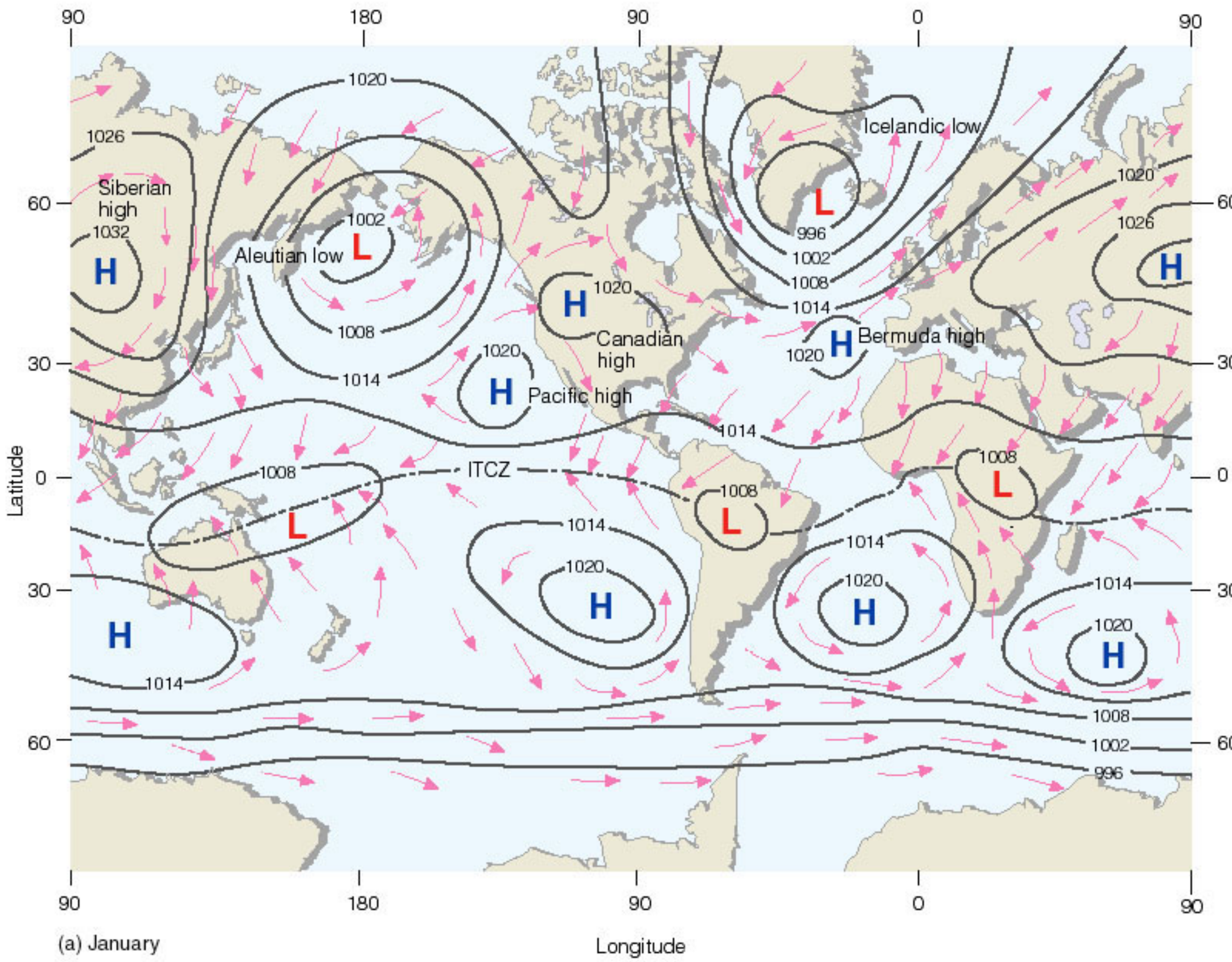
NOAA Tech Refresh

20 Jan 2012

Kipp Shearman, OSU

January Average Surface Map

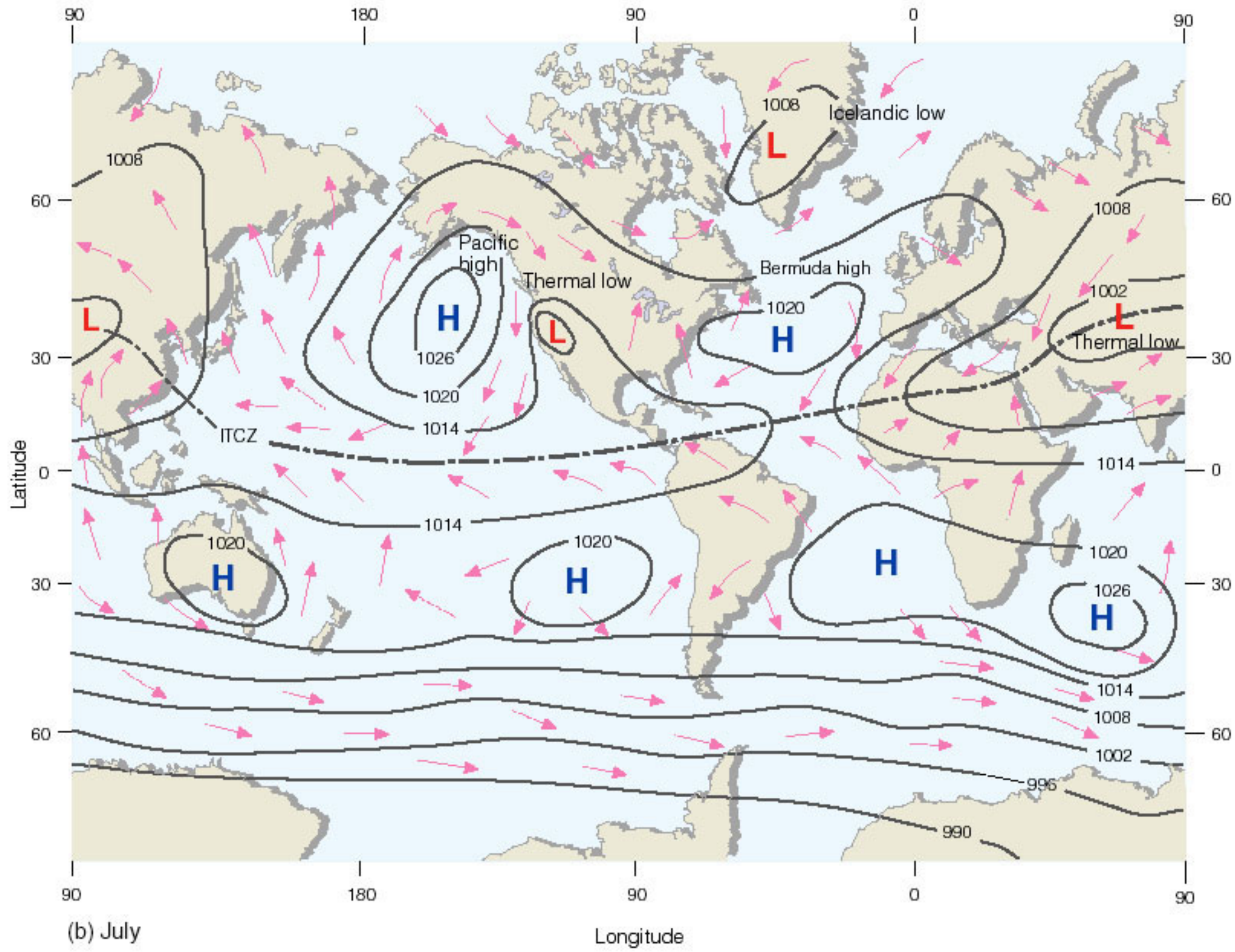
Northern Hemisphere land masses are dominated by high pressure (on average) during winter. For example, the **Siberian High**



The low pressure in the North Pacific (**Aleutian Low)** and in the North Atlantic (**Icelandic Low)** these ocean basins. Notice also the highs that are to the south (**Pacific High, Bermuda High)**. Note the position of the **ITCZ** (center of tropical convection and the base of the Hadley cell).

July Average Surface Map

The surface flow in the Southern Hemisphere is much smoother and less wavy due to less prominent land masses.



Notice: **Pacific High** pressure dominates the North Pacific during the summer. The **Bermuda High** is also more prominent during summer (it is this feature that steers hurricanes in the Atlantic). These high pressure systems also shift as the **ITCZ** moves northward

Outline

- Momentum Equations!
- When is Coriolis important?
- Geostrophic Balance
- Ekman Balance

Geostrophy

Most Important Balance Ever!

$$\frac{Du}{Dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv + F_z$$

$$\frac{Dv}{Dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu + F_z$$

$$\frac{Dw}{Dt} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + F_z$$

A **B** **C** **D** **E**

A $\rightarrow$ acceleration

B $\rightarrow$ pressure gradient force

C $\rightarrow$ Coriolis force $f = 2\Omega \sin\phi$

D $\rightarrow$ gravitational force

E $\rightarrow$ other (friction, tidal, wind forcing, etc.)

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- Momentum Equations!
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When is Coriolis important?

Standard answer: Compare terms in the momentum equations.

$$\frac{\text{Advection}}{\text{Coriolis}} = \frac{\left(u\dfrac{\partial u}{\partial x}\right)}{(fv)} \sim \frac{\left(U\dfrac{U}{L}\right)}{(fU)}$$

$$\sim \frac{U}{fL} = \text{Ro} = \text{"Rossby Number"}$$

When Ro >> 1, Coriolis is NOT important

When Ro $\leq$ 1, Coriolis is important

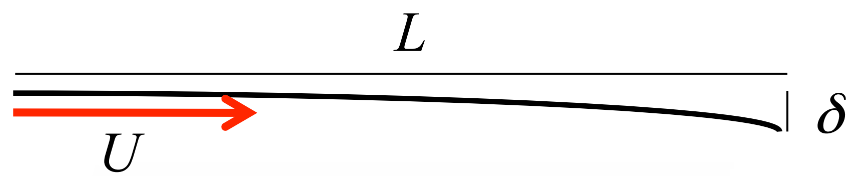
Coriolis Deflection

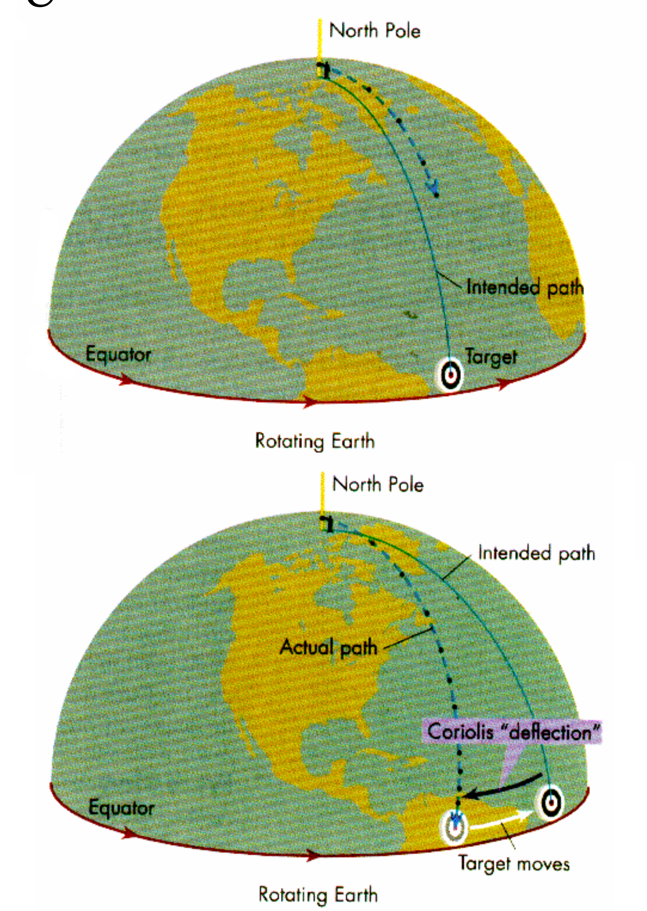
$$t = \frac{L}{U}$$

$$\delta = \tfrac{1}{2} a t^2$$

$$= \tfrac{1}{2}(fU)\left(\frac{L}{U}\right)^2$$

$$= \tfrac{1}{2}\frac{fL^2}{U}$$





http://abyss.uoregon.edu/~js/glossary/coriolis_effect.html

When is Coriolis important?

Answer: When Coriolis deflection is "big".

One definition of "big"…

First Recall: $\delta \approx \dfrac{fL^2}{U}$

Compare δ to L :

$$\frac{L}{\delta} = \frac{U}{fL} = \mathrm{Ro}$$

When $\mathrm{Ro} >> 1$, Coriolis is NOT important

When $\mathrm{Ro} \leq 1$, Coriolis is important

Coriolis Effect

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A **B** **C** **D** **E**

A $\rightarrow$ acceleration
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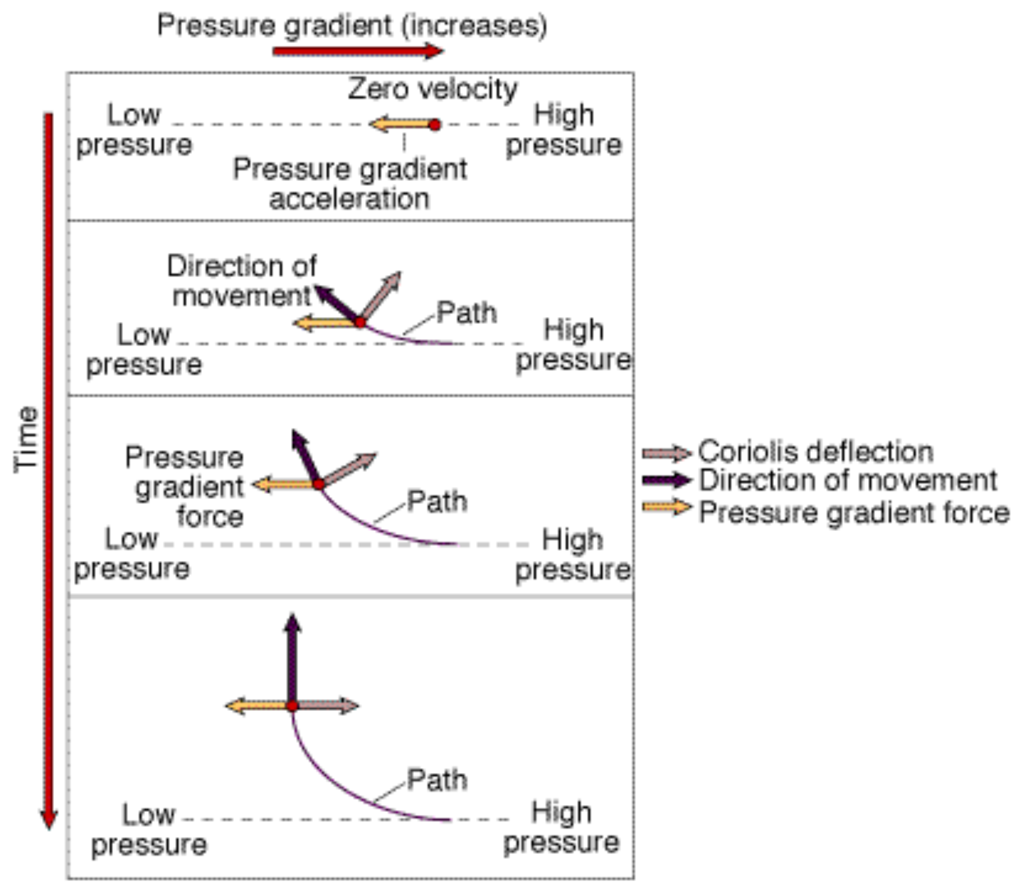
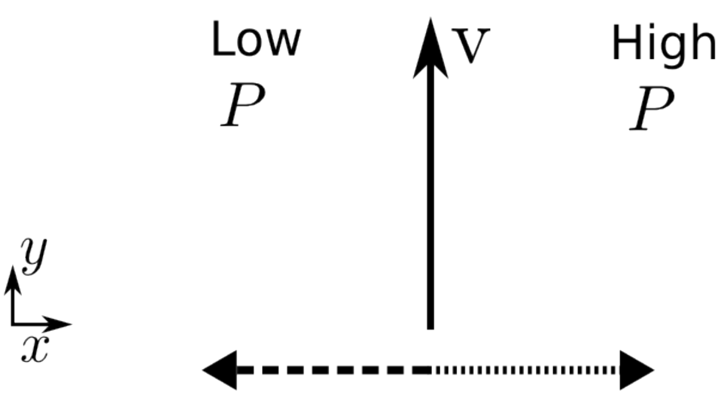
E $\rightarrow$ other (friction, tidal, wind forcing, etc.)

Geostrophic Balance

- Most common force moving water is PRESSURE (P) difference (gradient), which forces water in the direction from High to Low water pressure.
- But now, with rotation, as soon as particle starts to move down Pressure gradient, a Coriolis force (CF) at right angles starts to build; the stronger the flow, the stronger the force to the right (*in the northern hemisphere*).
- Eventually, CF and P are balanced, so particle has no force acting (continues at same velocity).
- In northern Hemisphere, particles move with high pressure on the right
- Flow is not down P gradient, but along it.



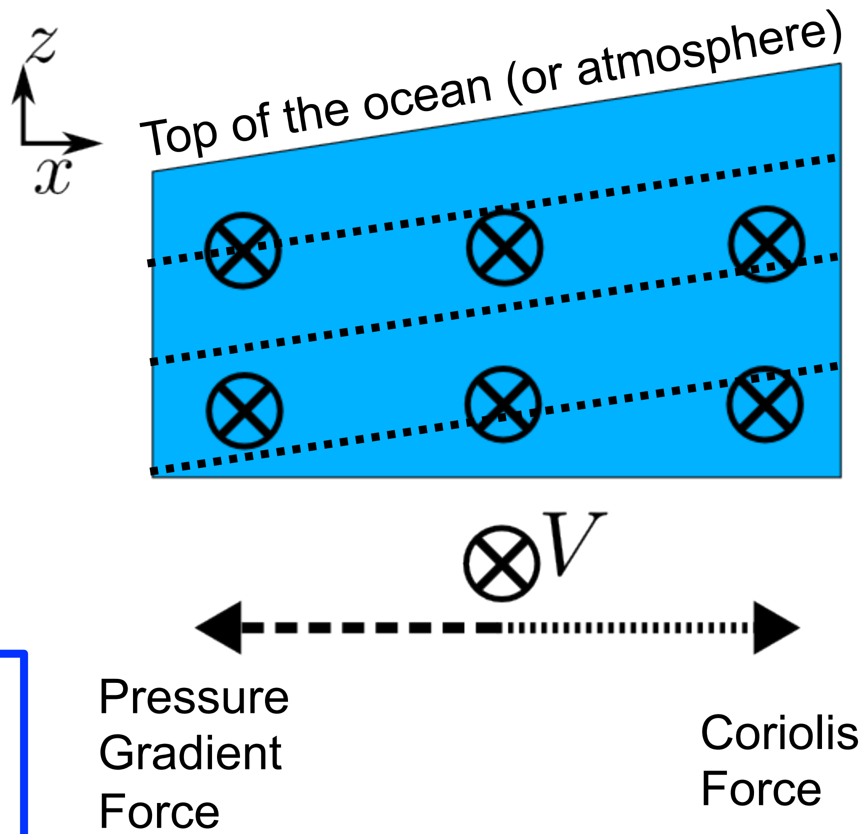
Geostrophic Balance



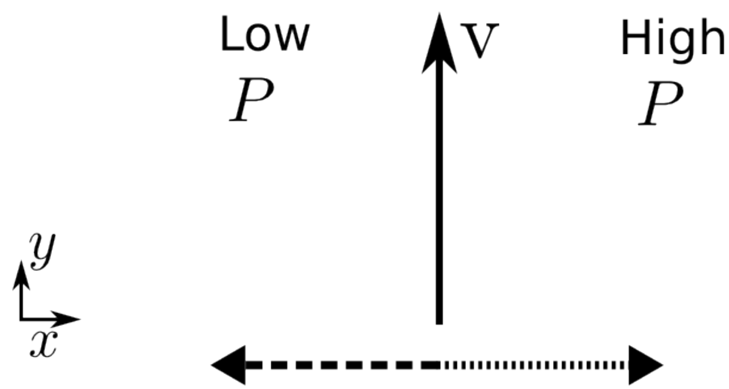
$$\frac{Du}{Dt} = \frac{-1}{\rho_\circ}\frac{\partial P}{\partial x} + f\mathrm{v} = 0$$

- **High Pressure to RIGHT of velocity in northern hemisphere**
- **High Pressure to LEFT of velocity in southern hemisphere**

Barotropic Pressure Gradient

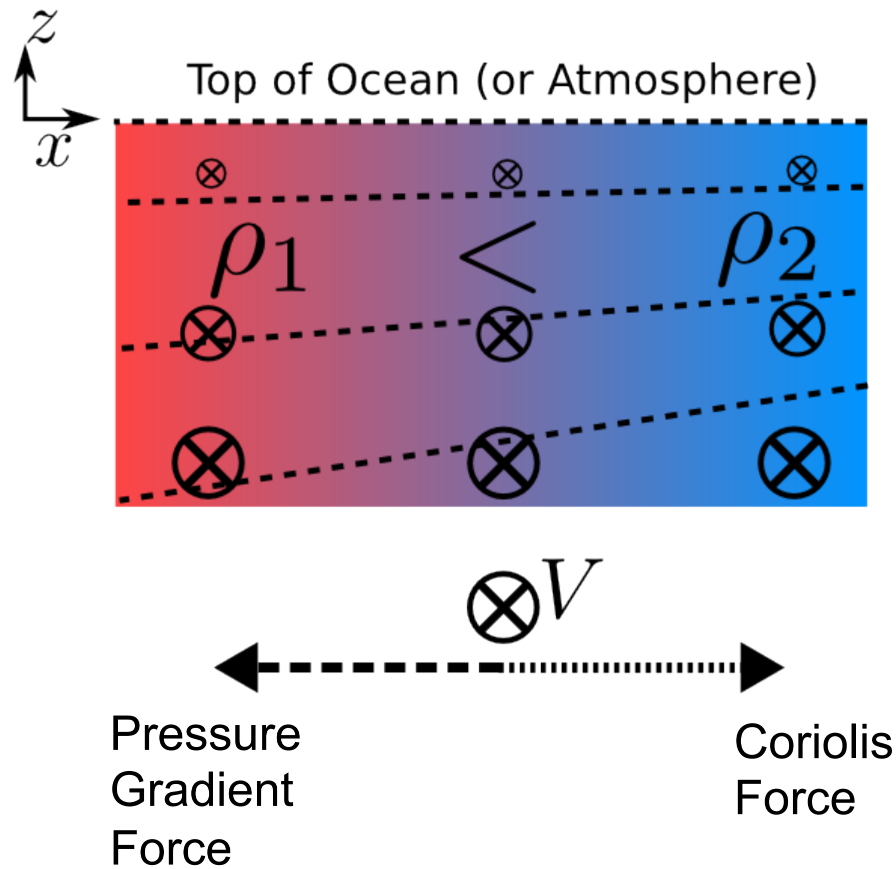


Geostrophic Balance



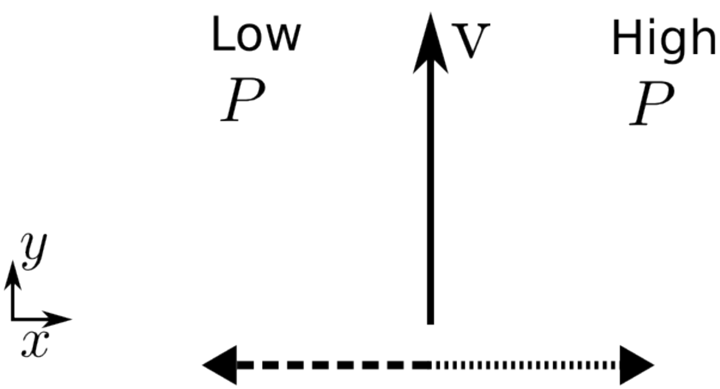
$$\frac{Du}{Dt} = \frac{-1}{\rho_\circ}\frac{\partial P}{\partial x} + f\mathrm{v} = 0$$

Baroclinic Pressure Gradient

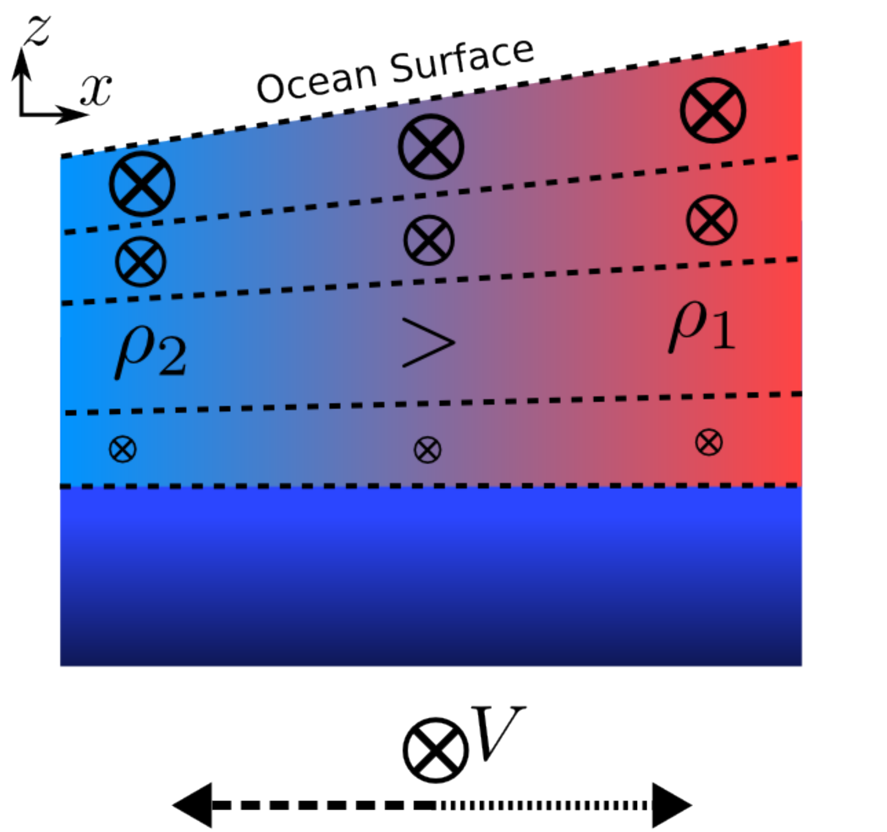


Geostrophic Balance

Barotropic + baroclinic pressure gradient

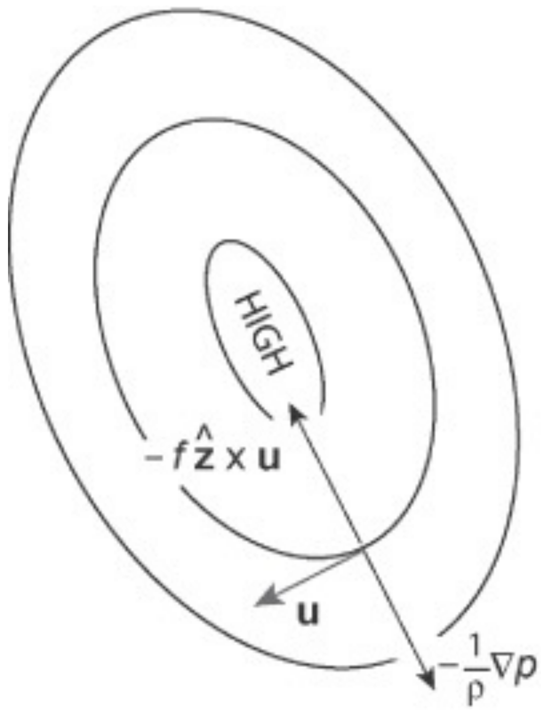


$$\frac{Du}{Dt} = \frac{-1}{\rho_\circ}\frac{\partial P}{\partial x} + f\mathrm{v} = 0$$

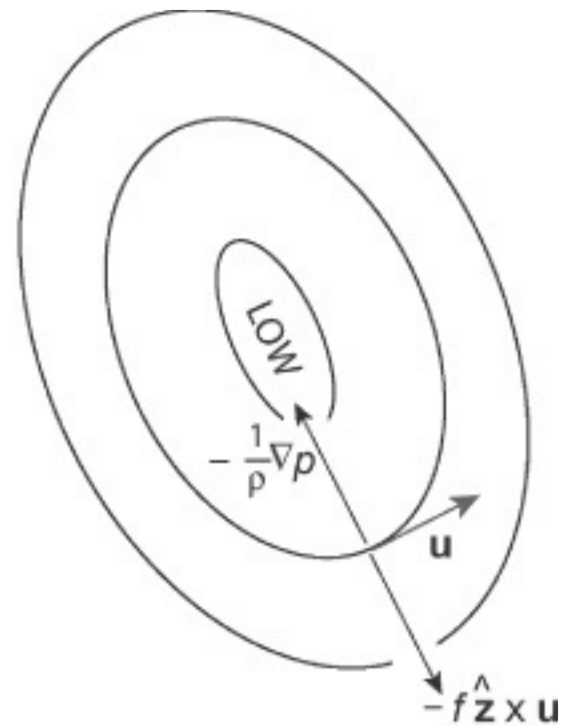


Drawn for northern hemisphere

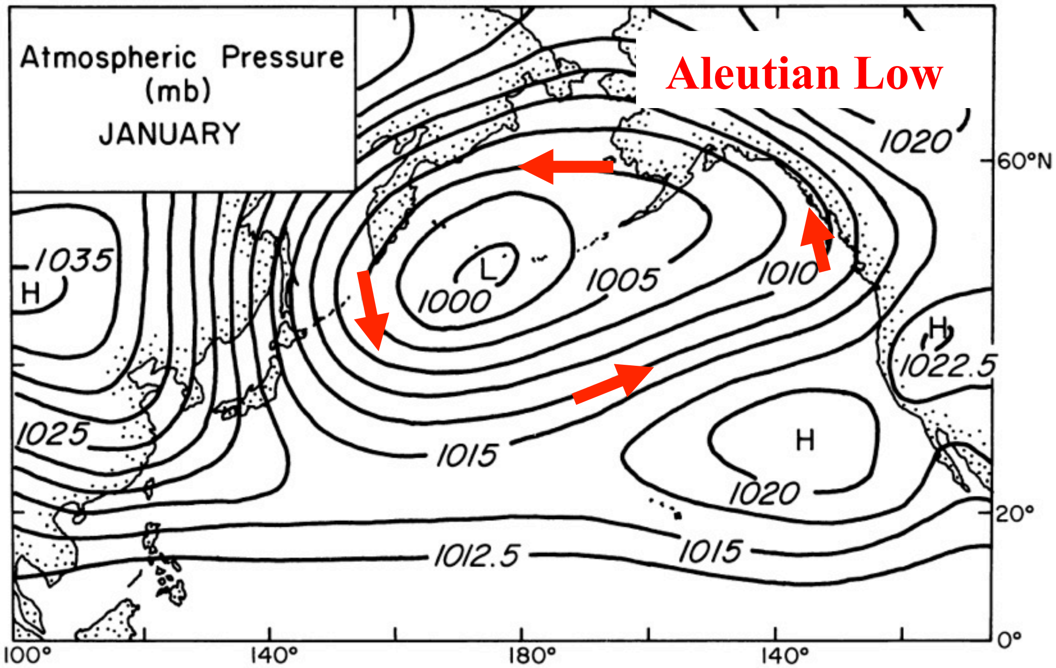
Coriolis effect on circulation around low and high pressure systems

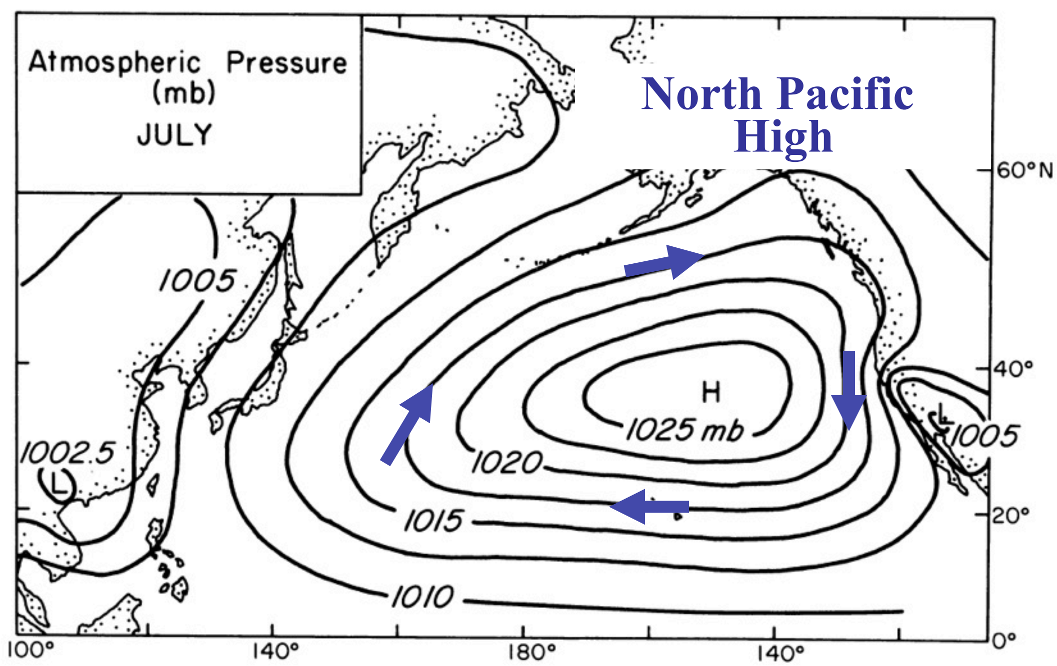


High pressure
Clockwise (N. Hemi.)
Anticyclonic



Low pressure
Counterclockwise (N. Hemi.)
Cyclonic





Wind direction with respect to atmospheric pressure in different season for North Pacific

- Big seasonal changes in the atmosphere
- Winds reverse direction

High pressure is to the right of the direction of the wind.

Huyer (1983)

Within the last 15 years, we can measure the sea surface height using satellite altimetry.

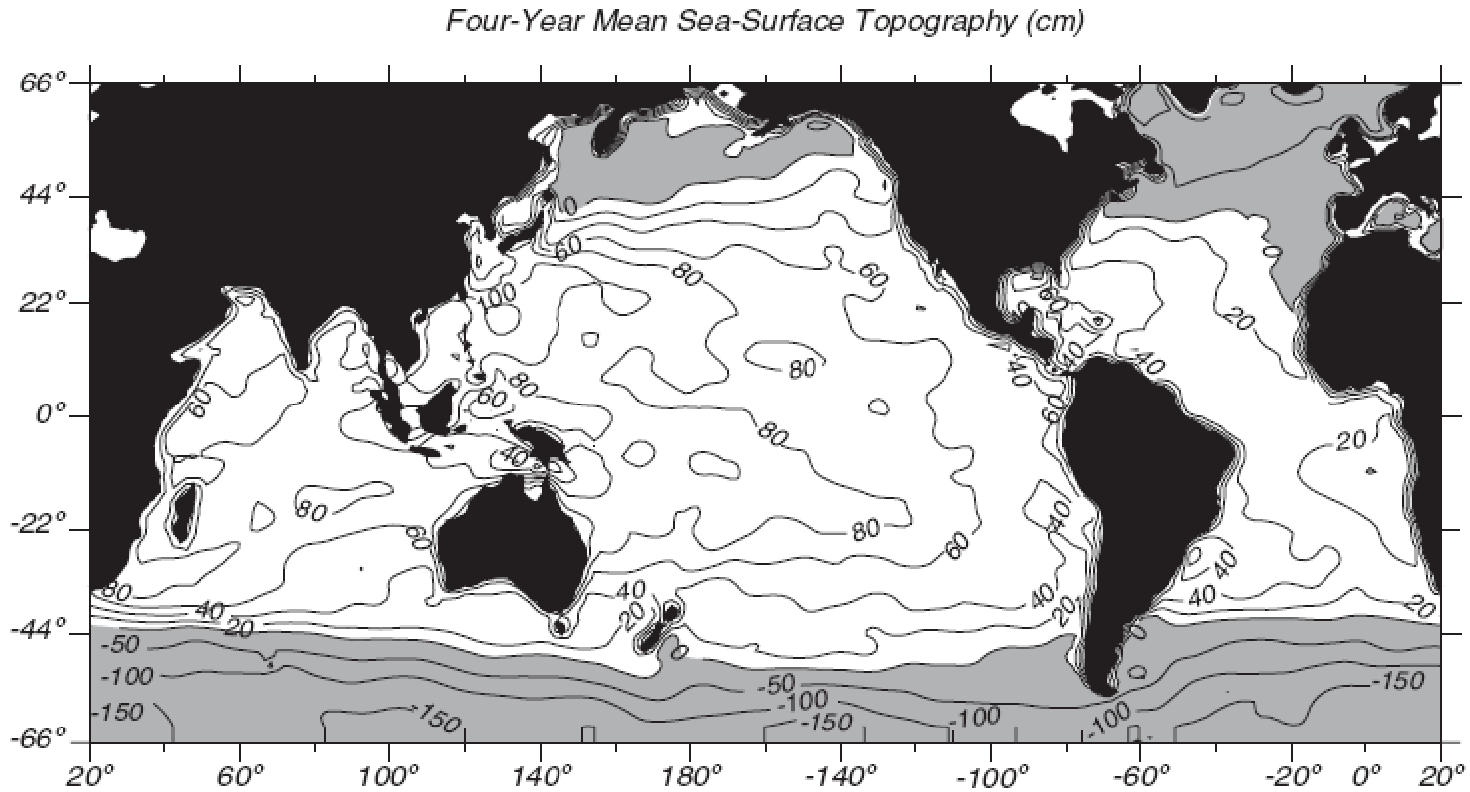
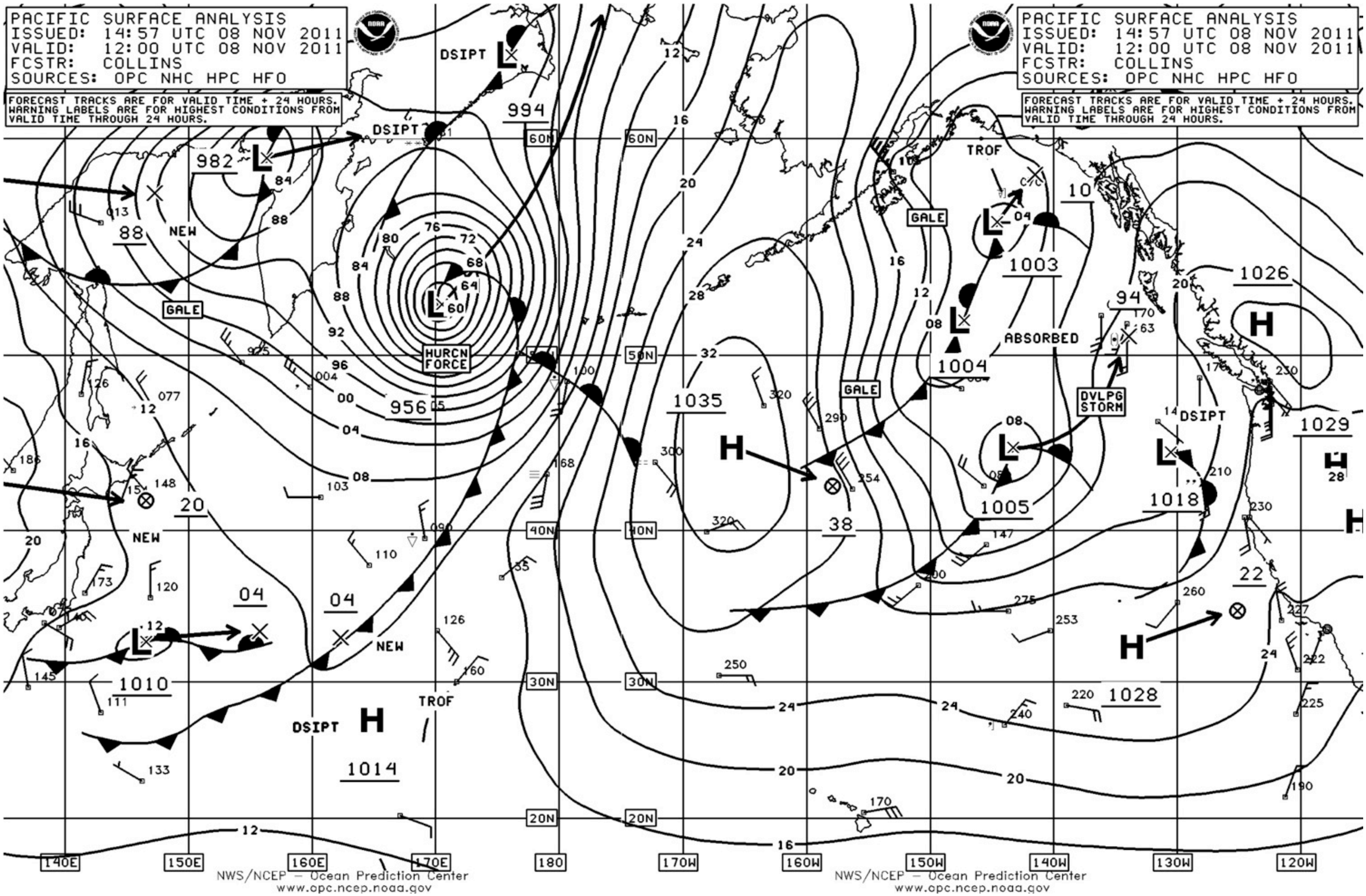


Figure 10.5 Global distribution of time-averaged topography of the ocean from Topex/Poseidon altimeter data from 10/3/92 to 10/6/99 relative to the jgm–3 geoid.


PACIFIC SURFACE ANALYSIS
ISSUED: 14:57 UTC 08 NOV 2011
VALID: 12:00 UTC 08 NOV 2011
FCSTR: COLLINS
SOURCES: OPC NHC HPC HFO
FORECAST TRACKS ARE FOR VALID TIME + 24 HOURS. WARNING LABELS ARE FOR HIGHEST CONDITIONS FROM VALID TIME THROUGH 24 HOURS.
DSIPT
994
982
NEW
GALE
HURCN FORCE
956
1035
1004
1003
1005
1026
1029
1018
1028
1010
1014
ABSORBED
DVLPG STORM
TROF
60N
50N
40N
30N
20N
140E
150E
160E
170E
180
170W
160W
150W
140W
130W
120W
NWS/NCEP – Ocean Prediction Center
www.opc.ncep.noaa.gov

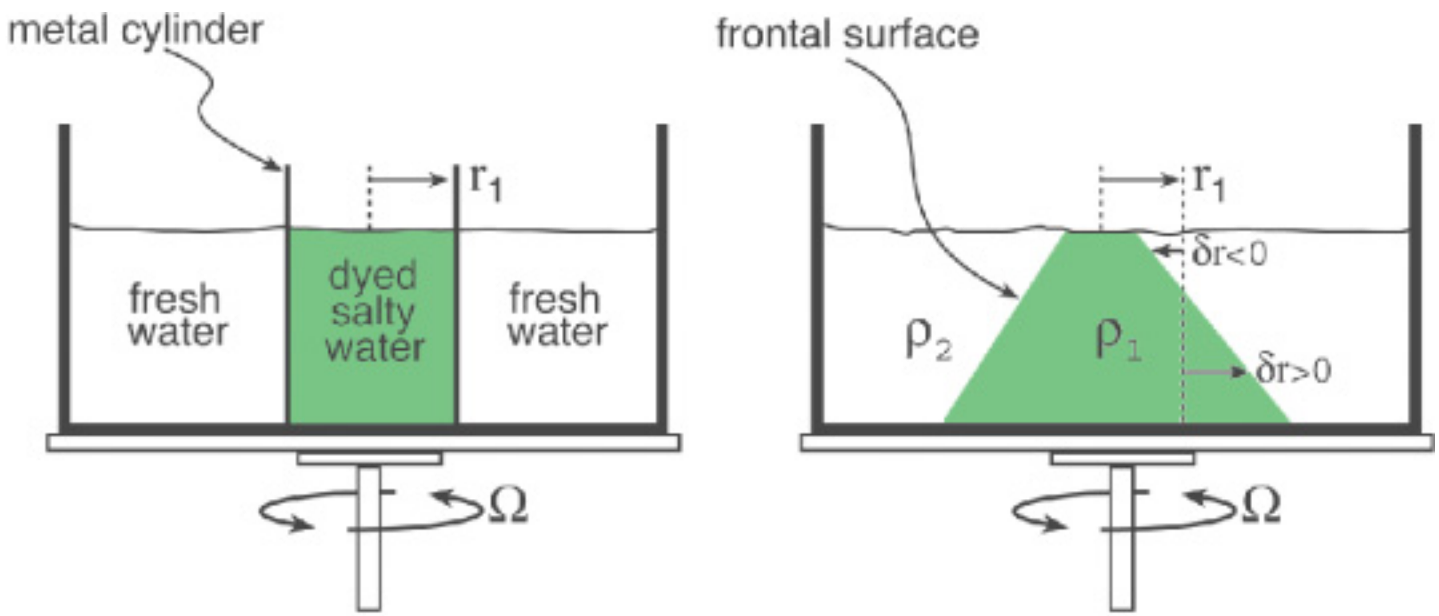


Figure 7.15: We place a large tank on our rotating table, fill it with water to a depth of 10 cm or so, and place in the center of it a hollow metal cylinder of radius $r_1 = 6$ cm, which protrudes slightly above the surface. The table is set into rapid rotation at 10 rpm and allowed to settle down for 10 minutes or so. While the table is rotating, the water within the cylinder is carefully and slowly replaced by dyed, salty (and hence dense) water delivered from a large syringe. When the hollow cylinder is full of colored saline water, it is rapidly removed to cause the least disturbance possible—practice is necessary! The subsequent evolution of the dense column is charted in Fig. 7.16. The final state is sketched on the right: the cylinder has collapsed into a cone whose surface is displaced a distance dr relative to that of the original upright cylinder.

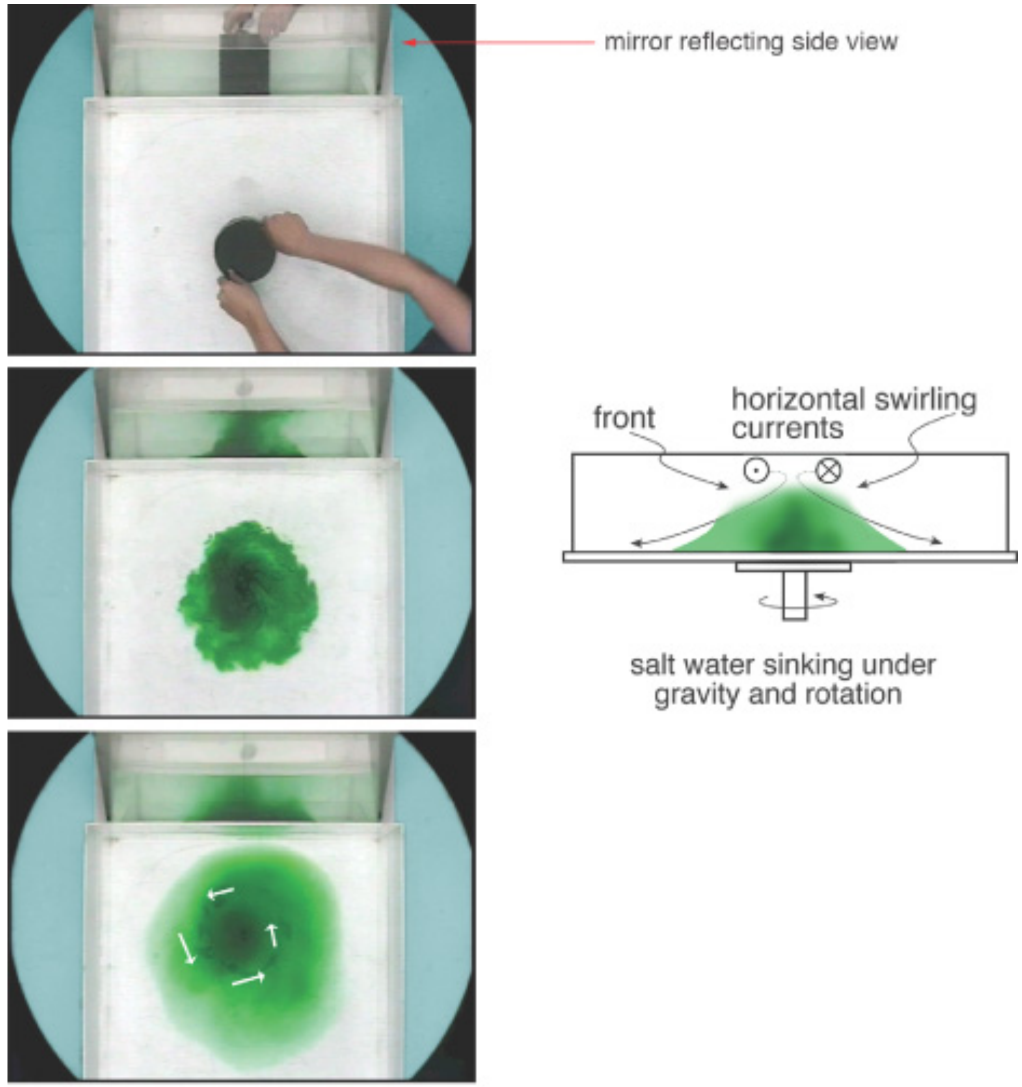
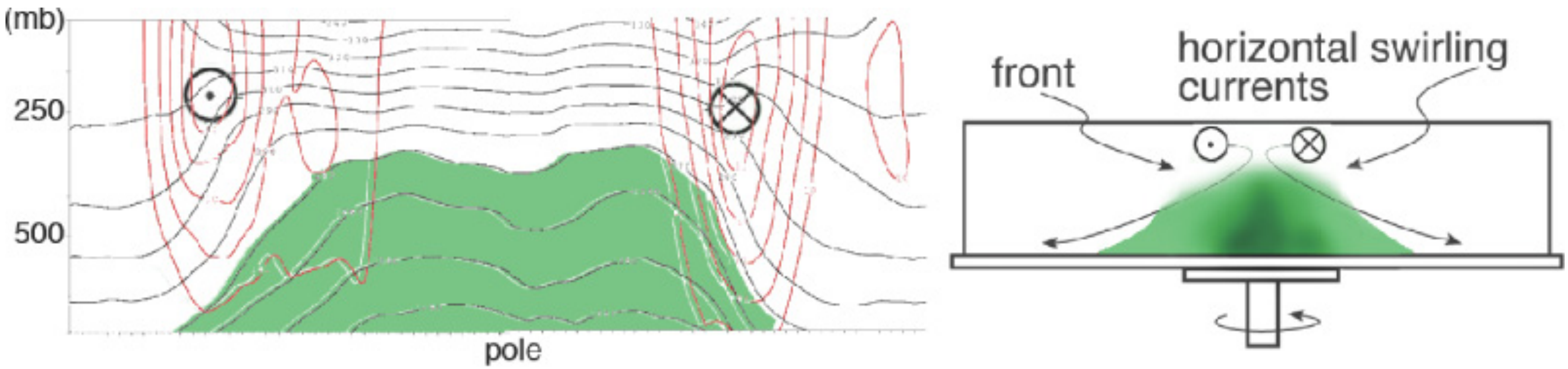


Figure 7.16: Left: Series of pictures charting the creation of a dome of salty (and hence dense) dyed fluid collapsing under gravity and rotation. The fluid depth is 10 cm. The white arrows indicate the sense of rotation of the dome. At the top of the figure we show a view through the side of the tank facilitated by a sloping mirror. Right: A schematic diagram of the dome showing its sense of circulation.

Eddies!



- Dense core – cyclonic rotation
- Light core – anticyclonic rotation

GFD Trivia: Geostrophic Flow is Non-divergent

$$u_g = -\frac{1}{f\rho}\frac{\partial P}{\partial y}, v_g = \frac{1}{f\rho}\frac{\partial P}{\partial x}$$

$$\frac{\partial u_g}{\partial x} + \frac{\partial v_g}{\partial y} + \frac{\partial w}{\partial z} = 0$$

$$\cancel{-\frac{1}{f\rho}\frac{\partial^2 P}{\partial y \partial x}} + \cancel{\frac{1}{f\rho}\frac{\partial^2 P}{\partial x \partial y}} + \frac{\partial w}{\partial z} = 0 \qquad \Rightarrow w = 0!$$

Thermal Wind

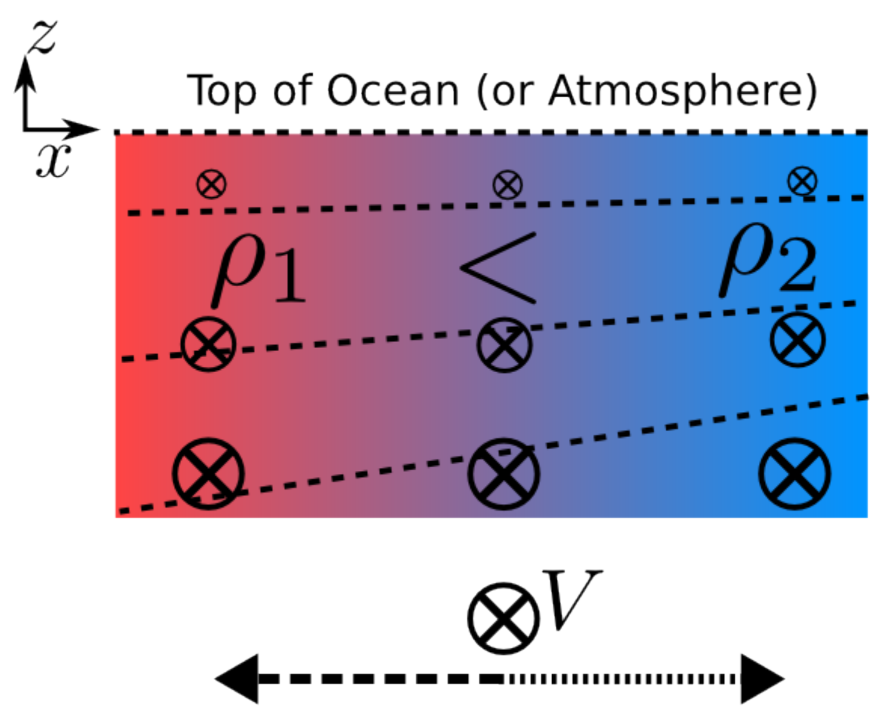
$$\frac{Du}{Dt} = \underbrace{\frac{-1}{\rho_\circ}\frac{\partial P}{\partial x}} \underbrace{+ f\text{v}} = 0$$

$$\frac{\partial}{\partial z}\left(\frac{1}{\rho}\frac{\partial P}{\partial x}\right) = \frac{\partial}{\partial z}(f\text{v})$$

$$\frac{1}{\rho}\frac{\partial}{\partial x}\frac{\partial P}{\partial z} = f\frac{\partial \text{v}}{\partial z}$$

$$\frac{1}{\rho}\frac{\partial}{\partial x}(-\rho g) = f\frac{\partial \text{v}}{\partial z}$$

$$\frac{g}{\rho f}\frac{\partial \rho}{\partial x} = -\frac{\partial \text{v}}{\partial z}$$



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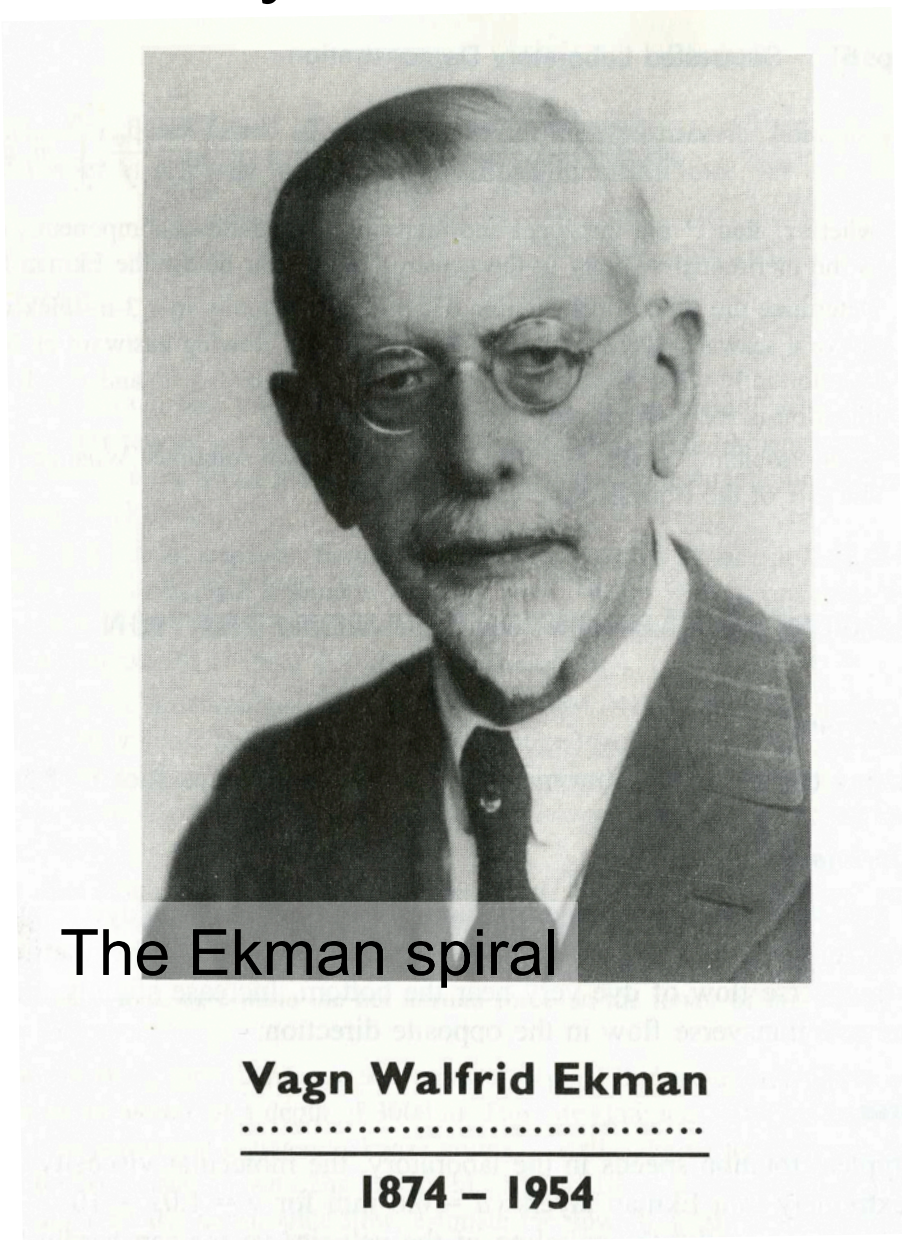
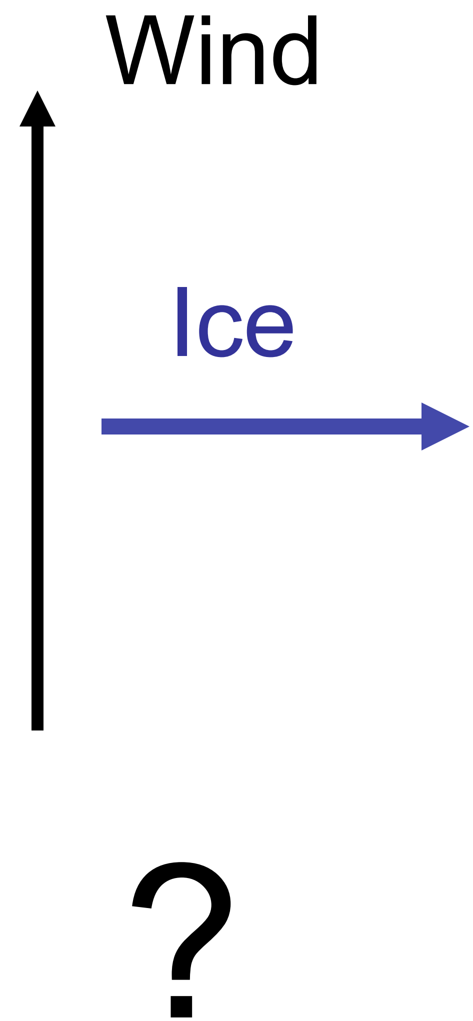
The mysterious world of …





Equations of motion

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} = -\frac{1}{\rho}\frac{\partial p}{\partial x} \qquad + fv + \frac{1}{\rho}\frac{\partial \tau_x}{\partial z}$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} = -\frac{1}{\rho}\frac{\partial p}{\partial y} \qquad - fu + \frac{1}{\rho}\frac{\partial \tau_y}{\partial z}$$

$$\frac{\partial w}{\partial t} + u\frac{\partial w}{\partial x} + v\frac{\partial w}{\partial y} + w\frac{\partial w}{\partial z} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$$

We are interested in the balance between Coriolis and wind stress.

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} = -\frac{1}{\rho}\frac{\partial p}{\partial x} \qquad + fv + \frac{1}{\rho}\frac{\partial \tau_x}{\partial z}$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} = -\frac{1}{\rho}\frac{\partial p}{\partial y} \qquad - fu + \frac{1}{\rho}\frac{\partial \tau_y}{\partial z}$$

$$\frac{\partial w}{\partial t} + u\frac{\partial w}{\partial x} + v\frac{\partial w}{\partial y} + w\frac{\partial w}{\partial z} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$$

Balance between Coriolis and Wind stress

$$0 = +fv + \frac{1}{\rho}\frac{\partial \tau_x}{\partial z}$$

$$0 = -fu + \frac{1}{\rho}\frac{\partial \tau_y}{\partial z}$$

$$0 = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g$$

Wind stress is parameterized

A_z is vertical diffusivity.

Oceanic value ~ 10^{-2} m^2s^{-1}

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0,\ \tau_x = \rho A_z \frac{\partial u}{\partial z},\ \tau_y = \rho A_z \frac{\partial v}{\partial z}$$

Ekman Transport

Ekman Transport on x $= U_E = \int u\,dz$

Unit: m^2s^{-1}

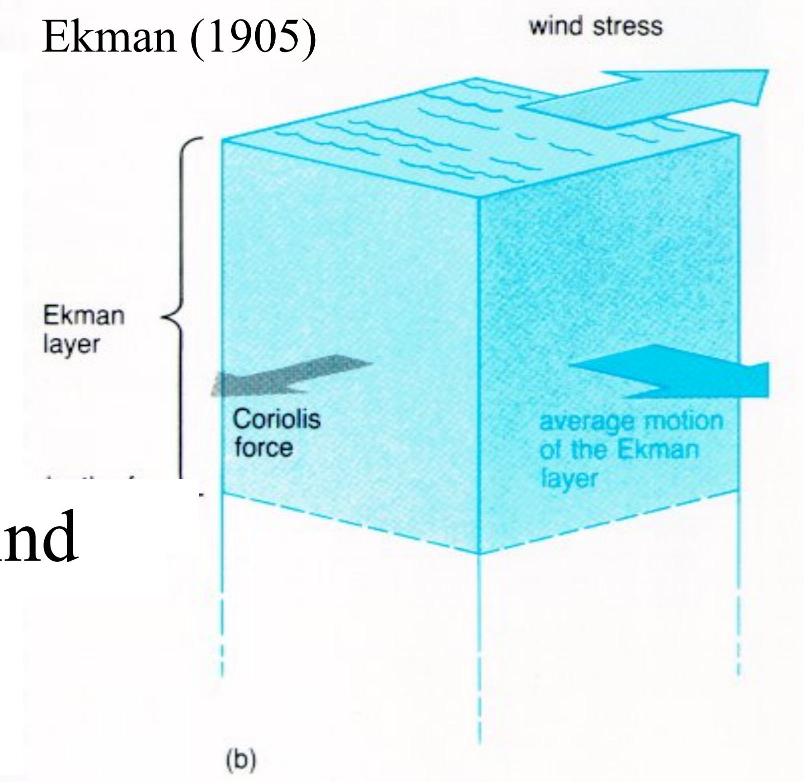
Ekman Transport on y $= V_E = \int v\,dz$

Ekman Transport:

Due to wind and the Earth's rotation

Always to the right of the wind in Northern Hemisphere.

Ekman (1905)



$$U_E = \frac{\tau_y}{\rho f}$$

$$V_E = \frac{-\tau_x}{\rho f}$$

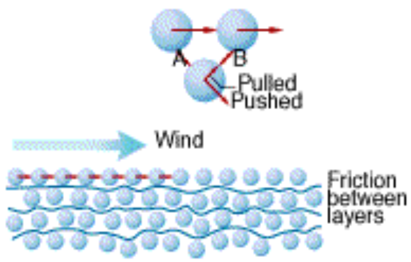
$\left(\tau_x, \tau_y\right)$ = Wind stress on x and y

ρ = Density of Sea Water

f = Coriolis parameter (Earth rotation rate)

Wind Forcing at the Ocean Surface

- Wind-forcing can generate *currents and waves*, as wind transfers some of its momentum into the ocean
- Wind acts via friction at the surface: wind stress τ



Stresses have units of N/m^2, (force/area), like pressure. Stresses are forces parallel to a surface, pressure is force perpendicular to the surface.

- Force/Area depends on the square of the wind speed u, and it points in the same direction as the wind:

$$\tau \propto u^2$$

$$\vec{\tau} = \rho_a C_D |\vec{u}| \vec{u}$$

$$C_D = \text{drag coefficient} \approx 1.4 \times 10^{-3}$$

$$\rho_a = \text{density of air} \approx 1.3\, kg/m^3$$

Example: 20kt wind ≈ 10 m/s → 0.18 N/m^2 = 1.8 $dyne/cm^2$

Vertical structure of u and v (Ekman spiral)

$$u = \frac{\pm\tau^{sy}}{\rho\delta|f|} e^{\frac{-z}{\delta}} \left[\cos\left(-\frac{z}{\delta}\right) - \sin\left(-\frac{z}{\delta}\right)\right]$$

$$v = \frac{\tau^{sy}}{\rho\delta|f|} e^{\frac{-z}{\delta}} \left[\cos\left(-\frac{z}{\delta}\right) + \sin\left(-\frac{z}{\delta}\right)\right]$$

<u>Don't memorize u and v.</u>

$$\delta = \sqrt{\frac{2A_z}{|f|}}$$

δ is Ekman depth: Decay depth of Ekman spiral. Depth of frictional influence.

<u>You want to understand the meaning of this depth.</u>

Ekman Spiral

1. Ocean at surface is dragged by wind, but then acted on by Coriolis Force. Current at surface are 45° to right of wind in Northern Hemisphere.

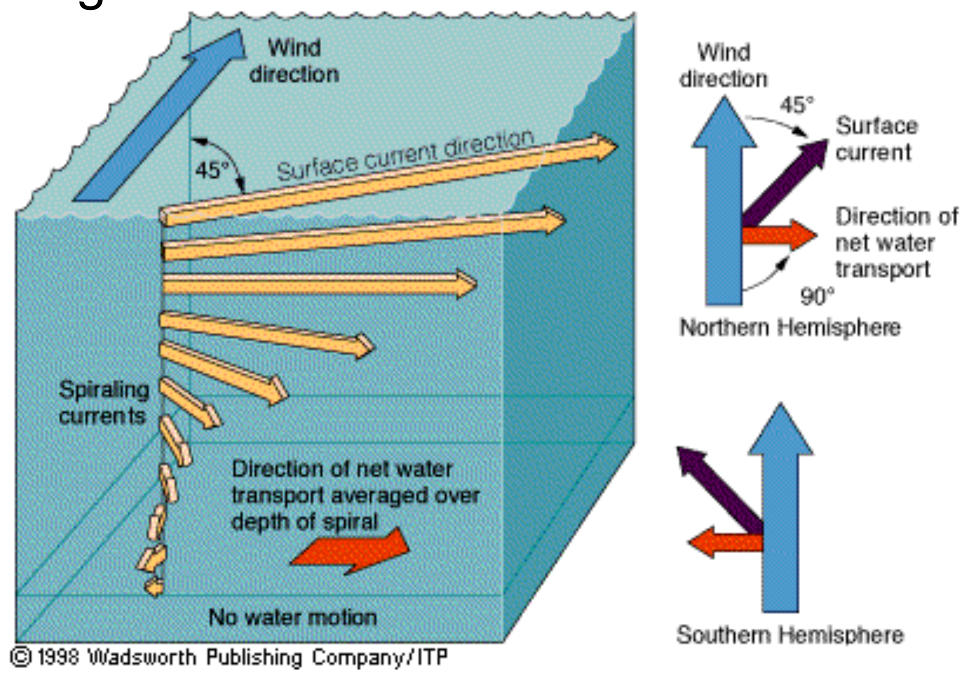
2. Friction transmits stress (drag) downward within water column: upper layer rubs on layer below and moves it. But response will be weaker (frictional losses) and further to the right.

3. Process continues down through water column.

4. Creates a spiral, decaying with depth. This is the ***Ekman spiral***.

5. Typical decay depths are 10-30 m.

$$\delta = \sqrt{\frac{2A_z}{|f|}} \approx \sqrt{\frac{2\times 10^{-2}}{|10^{-4}|}} \approx 15m$$



What does the real oceanic surface boundary layer look like?

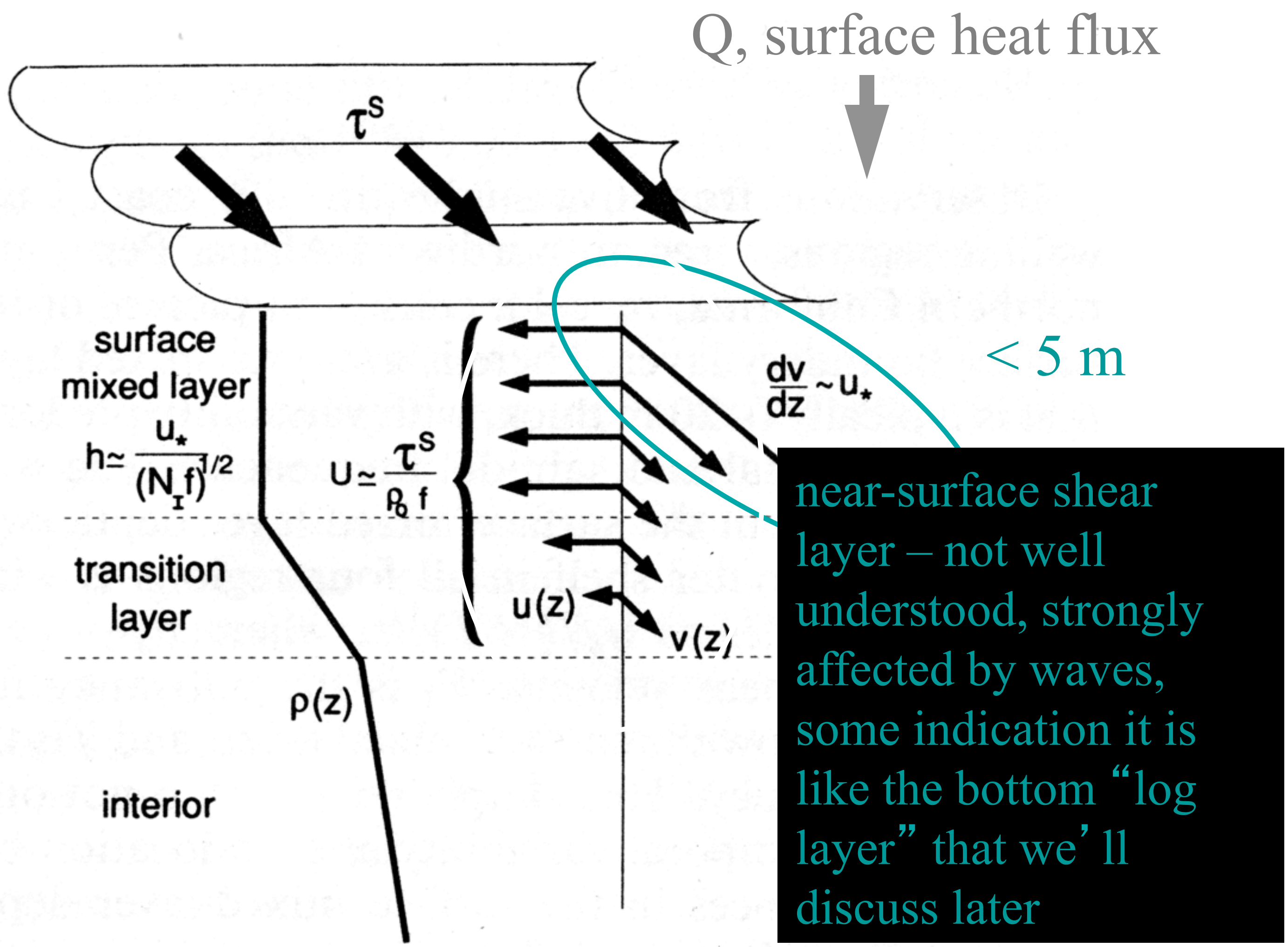
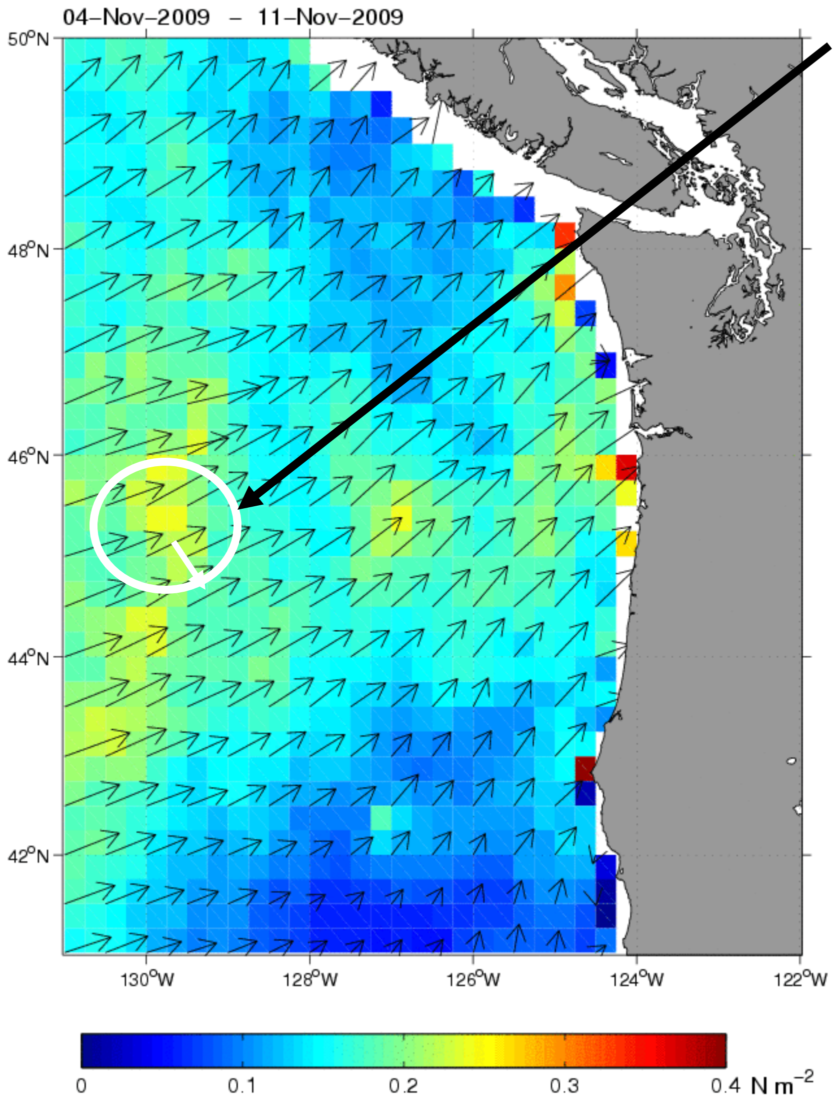


FIG. 20. Schematic summarizing some of the characteristics of the surface boundary layer in a coastal upwelling region: $u_* = (\tau^S/\rho_0)^{1/2}$ is the shear velocity and U is the cross-shelf transport in the surface mixed layer plus the transition layer.

Lentz, 1992

Example: Calculate Ekman Transport on y

Wind data by QuickSCAT from OrCOOS (
http://agate.coas.oregonstate.edu/data_index.html)



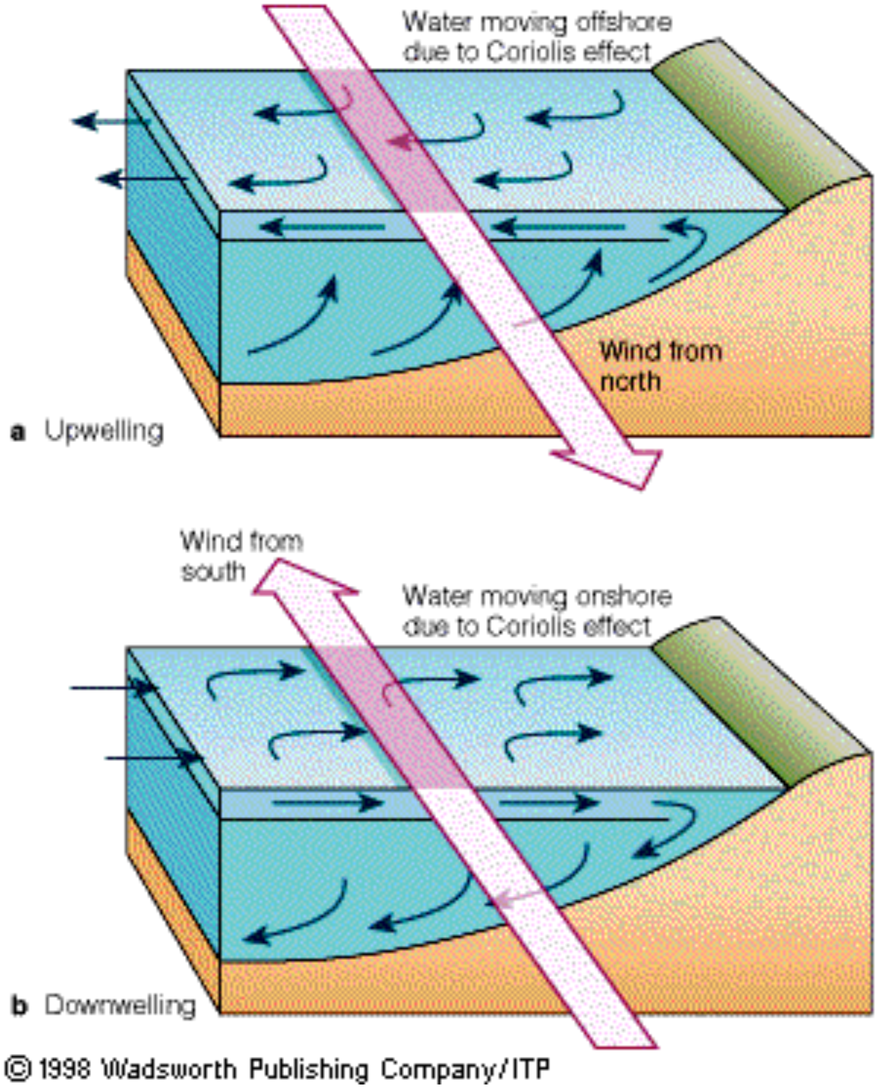
$\tau_x = 0.2\ \mathrm{N/m^2}$
$\tau_y = 0.1\ \mathrm{N/m^2}$
$\rho = 1025\ \mathrm{kg/m^3}$
$f = 2\Omega\sin\varphi = 1.03\times10^{-4}\mathrm{s^{-1}}$

$U_E = \tau_y/\rho f\ = +1\ \mathrm{m^2/s}$
$V_E = -\tau_x/\rho f\ = -2\ \mathrm{m^2/s}$

Summary

- ***Ekman spiral*** is due to wind stress and the Earth's rotation which is decaying with depth. Decay depth is ***Ekman depth***.
- Current at surface are 45° to right of wind in Northern Hemisphere.
- Vertical integration of ***Ekman spiral*** is **Ekman transport (U_E and V_E).**
- **Ekman transport** is always to the right of the wind in Northern Hemisphere.

Upwelling/Downwelling driven by the presence of a coastal boundary:

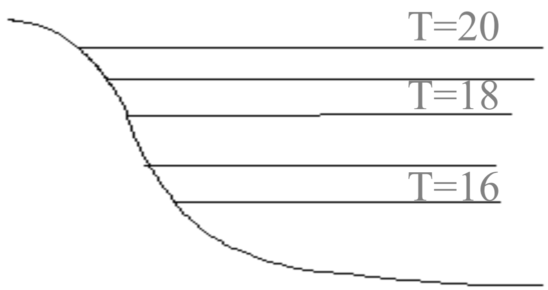


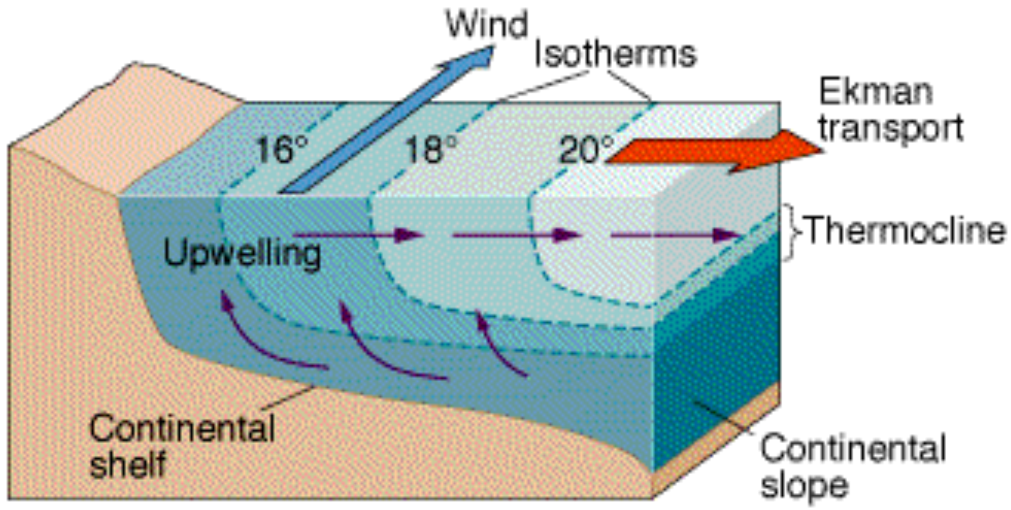
Coastal Upwelling:
Wind to South. Ekman transport in surface layer is to right of wind (West). Flow is divergent at the coast. Deeper water is ***upwelled*** into near-surface.

Primarily seen during spring/summer off Oregon coast.

Coastal Downwelling:
Wind to North. Ekman transport in surface layer is to right of wind (East). Flow is convergent at the coast. Deeper vertical velocity is downward.

Upwelling/Downwelling with Stratification



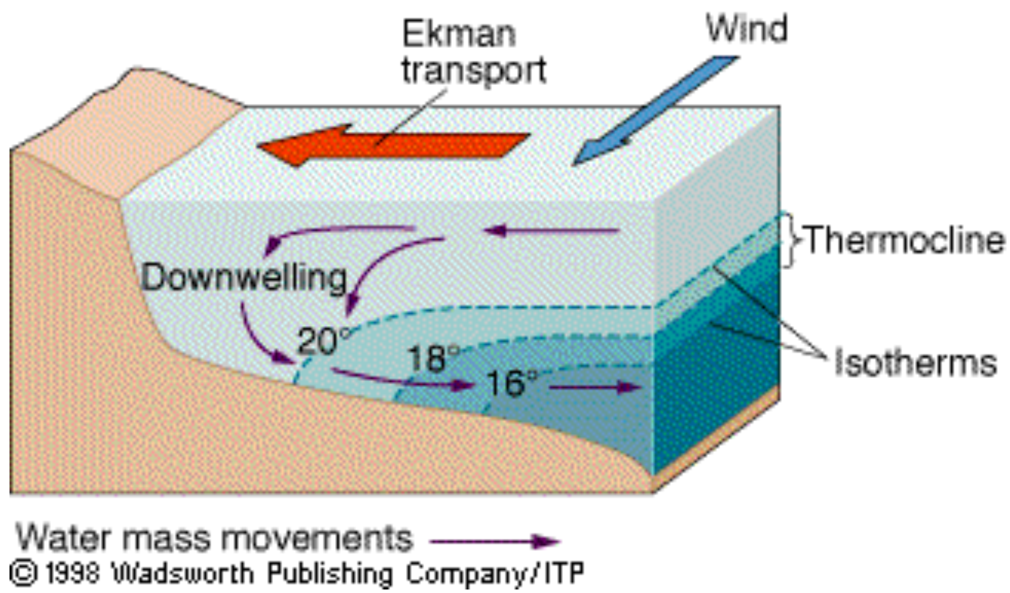


Upwelling

- Cold deep water brought to surface near coast
- Nutrients (max near bottom) brought up to surface
- Creates fronts in T,S

Downwelling

- Surface water transported to coast
- Warm surface water forced downward



Coastal Upwelling: Sea Surface Temperatures

Coldest temperatures near coast.

Surface water at the coast came from deeper in the water column.