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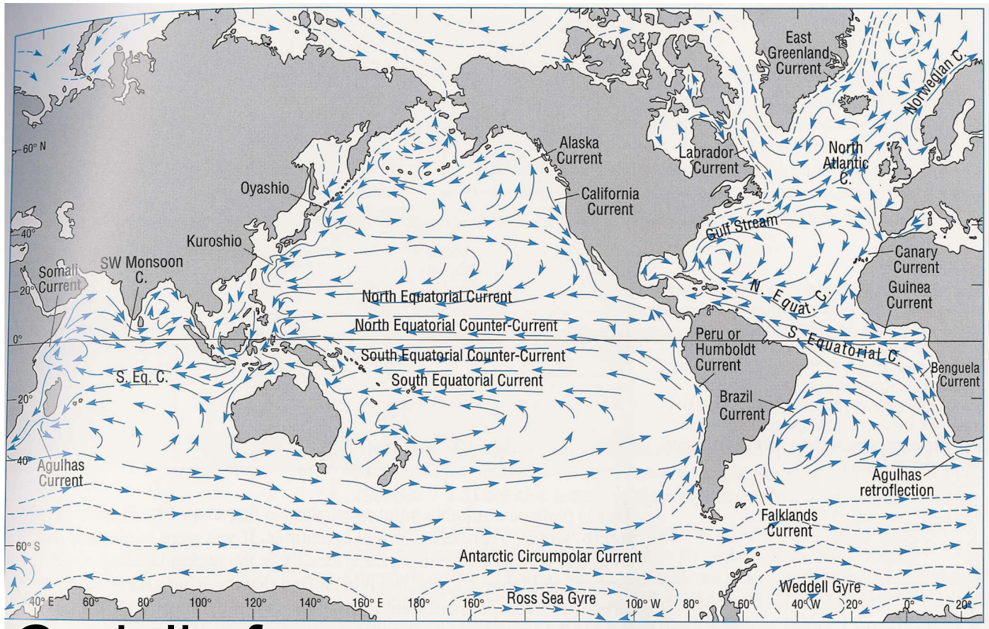
Ocean circulation



- ☐ Coriolis force
- ☐ Initial oscillation
- ☐ Geostrophic current
- ☐ Ekman layer
- ☐ Upwelling
- ☐ Global surface currents

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http://coda.oc.ntu.edu.tw/coda

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Ocean circulation and ocean climate

Yu-heng Tseng (YH) and Sze Ling Ho (SL)

10/3 Ocean circulation (dynamics and ocean currents) YH
10/17 Ocean circulation (Global Flux and Deep circulation) SL
10/31 Climate Impact from ocean circulation (ENSO and climate variability) YH
11/21 Biogeochemistry (carbon & others) SL
11/28 Sea level changes YH
12/12 Hydrology and ecosystem SL

http://www.sisal.unam.mx/labeco/LAB_ECOLOGIA/OF_files/54211042-Ocean-Circulation-Open-University.pdf

Education

1991-1995 (B.S.)	National Taiwan University, Mechanical Engineering
1997-1999 (M.S.)	Stanford University, Department of Mechanical Engineering
1999-2003 (Ph.D)	Stanford University, Department of Civil and Environmental Engineering.

Professional Experiences

2003-2004	Johns Hopkins University, Mechanical Engineering
2004-2006	Lawrence Berkeley National Laboratory, Computational Research Division
2006-2012	National Taiwan University, Department of Atmospheric Sciences
2012-2017	National Center for Atmospheric Research, Climate and Global Dynamics Laboratory
2017-	National Taiwan University, Institute of Oceanography
2021-	National Taiwan University, Ocean Center (Director)

Research interesting: turbulence and mixing, high-order numerical method, non-hydrostatic model, regional and global ocean circulation model, air-sea (ocean-atmosphere) interaction, climate model development and parameterization

Early ocean studies

- For much of human history, knowledge of the ocean and its currents was not recorded for many reasons - most sailors could not write and information was passed orally.
- European nations began systematic ocean explorations but kept information secret between the 15^{th}-18^{th} centuries.
- The Gulf Stream, was known as early as 151x (early 16^{th} century).

Bishop Resen of Copenhagen drew the first map of the Gulf Stream in 1605 – his map was based on records from trans-Atlantic voyages of explorer Martin Frobisher (1535-1594). Subsequent charts were published in 1678 by Athanasius Kircher and 1685 by Happelius.

Benjamin Franklin (1706-1790)

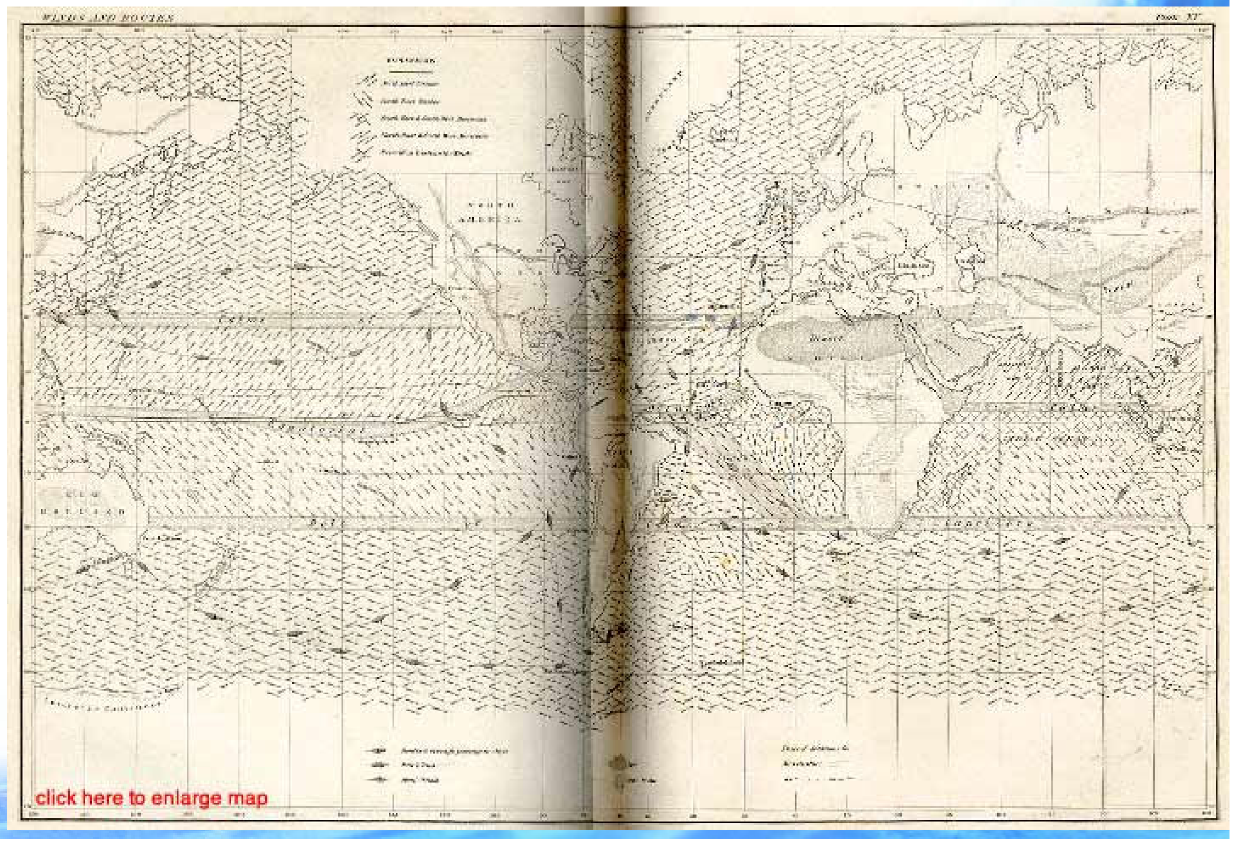
- Franklin served as colonial deputy postmaster general from 1753-1774.
- He found that British merchant ships arriving in the colonies from England took many days to weeks longer to make the voyage than American vessels.

Franklin's cousin Timothy Folger, a whaling ship captain told him of the Gulf Stream, the strongest surface current in the North Atlantic (10-12 km/hr or 6-7 miles/hr)
Franklin published his cousin's chart showing the location of the Gulf Stream and presented it to the British.

Matthew Fontaine Maury (1806-1873)

- The American Naval officer conducted the first systematic study of the ocean's surface currents and winds.
- Maury compiled information on currents and winds from logbooks of sailors' observations stored at the US Navy's Depot of Charts and Instruments.

He estimated current directions and speeds by analyzing deflections in ships courses caused by surface ocean currents. Combining thousands of observations, Maury constructed a map of average surface currents.



Coriolis Force: The force due to observing/predicting velocities in an accelerating (rotating) reference frame – the Earth

We return now to the acceleration term in the equation of motion - in an inertial reference frame, we fond that

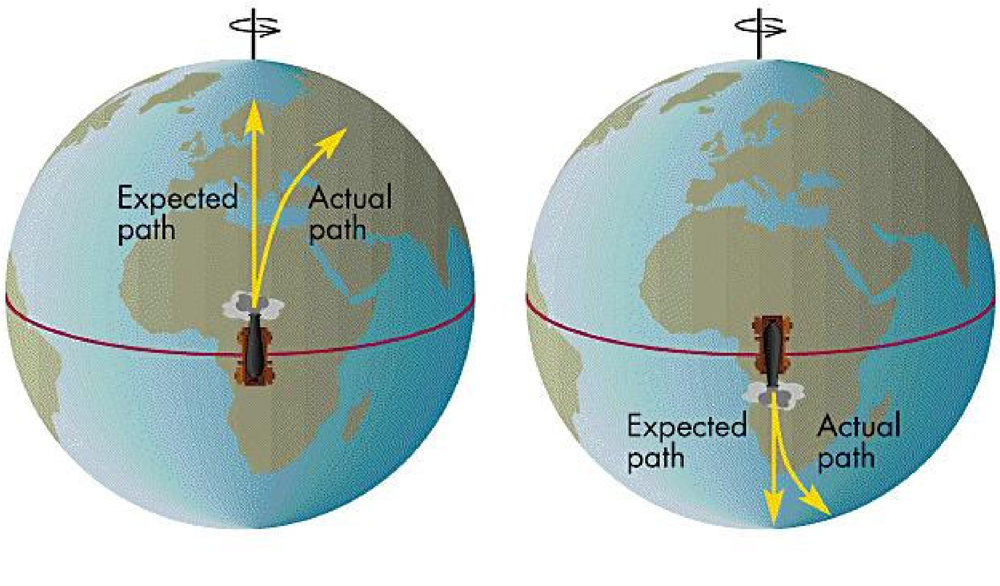
$$\frac{D\vec{u}}{Dt} = \vec{a} = \frac{\partial \vec{u}}{\partial t} + \vec{u} \circ \nabla \vec{u}$$

Fluid particle acceleration

Unsteady acceleration

Advective acceleration

How is this changed when we measure velocities in a rotating reference frame?



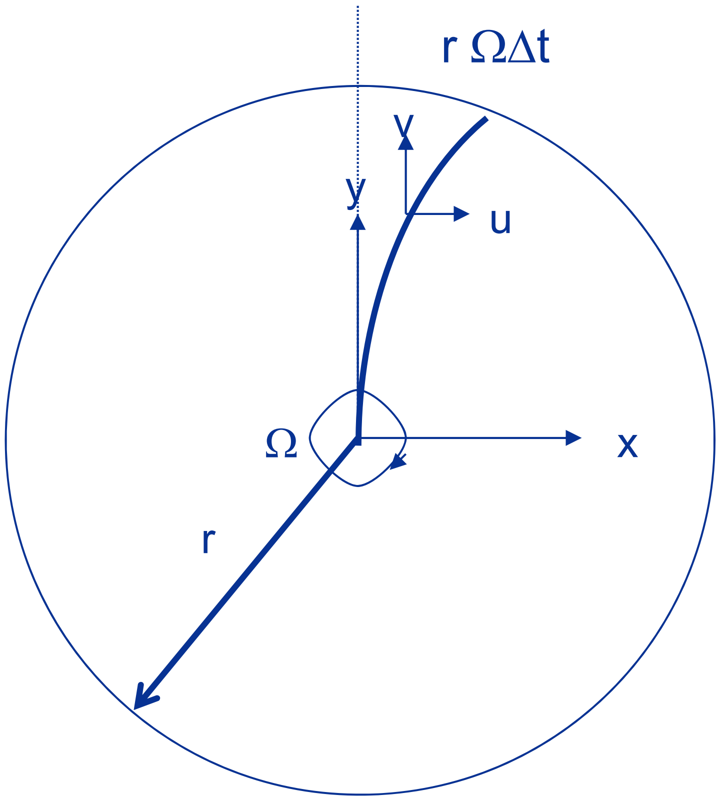
A Projectile fired northward

B Projectile fired southward

See also: http://www.ems.psu.edu/~fraser/Bad/BadCoriolis.html

https://www.youtube.com/watch?v=kCbMKSZZO9w

A simple kinematic argument: Let's look at a particle moving at steady velocity (in an absolute frame) on a rotating table



If we consider small changes in time

$$V\Delta t = r$$

$$\Delta x \approx r\Omega\Delta t = \Omega V \Delta t^2 = \frac{1}{2} a_x \Delta t^2$$

$$\Rightarrow \quad a_x = 2\Omega v$$ (Seen in rotating frame)

Thus we have an acceleration to the right, so to convert the acceleration Du/Dt in the rotating reference we need to write:

$$a_x = \frac{Du}{Dt} + \underbrace{2\Omega v}_{\text{Coriolis}}$$

More generally (after a lot of vector manipulations):

$$\vec{a} = \frac{D\vec{u}_{rel}}{Dt} + \underbrace{2\vec{\Omega} \times \vec{u}_{rel}}_{\text{Coriolis}}$$

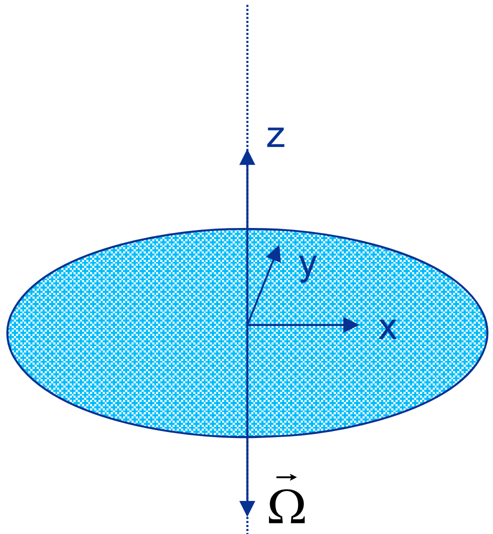
Vector cross product

This general relation emphasizes that Ω is a vector aligned with the axis of rotation of the coordinate system

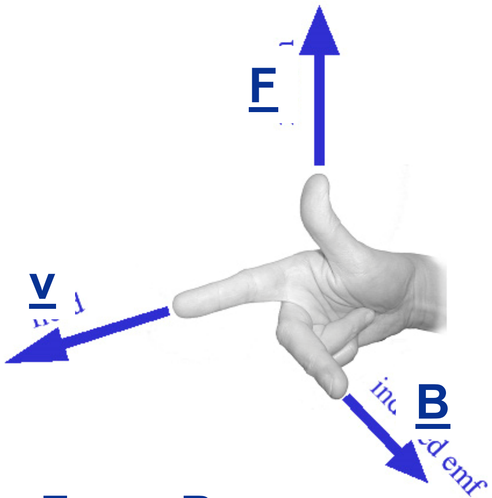
In the case $\vec{\Omega} = -\Omega\vec{k} = -\Omega(0,0,1)$

Sign convention is Ω positive if counterclockwise in a right hand coordinate system

Unit vector in z direction (positive out of page on previous page)



Note that rotation vector points down since it is negative – clockwise rotation



F=qvxB

Right hand rule (magnetic force)

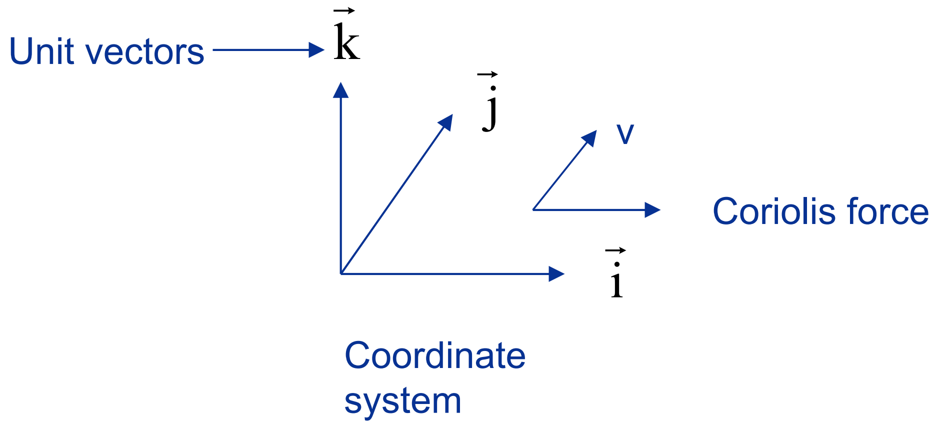
As we see in this simple example, it appears that the particle is being pushed to the right. If we move the acceleration to the “Force” side of the equation, we can write:

$$\frac{D\vec{u}_{rel}}{Dt} = -2\vec{\Omega} \times \vec{u}_{rel} + \ldots$$

Pressure gradients, gravity, shear stress gradients

Suppose

$$\vec{u}_{rel} = (0, v, 0)$$
$$\vec{\Omega} = (0, 0, \Omega)$$

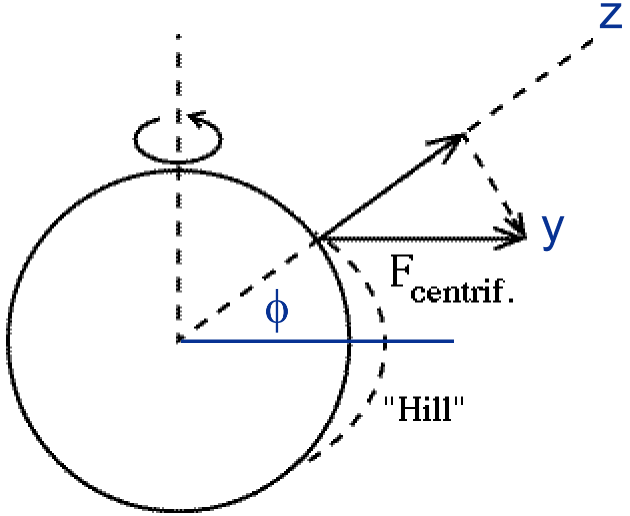


$$-2\vec{\Omega} \times \vec{u}_{rel} = -2\det\begin{vmatrix} \vec{i} & \vec{j} & \vec{k} \\ 0 & 0 & \Omega \\ 0 & v & 0 \end{vmatrix} = -2\left(\vec{i}(0-\Omega v) - \vec{j}(0-0) + \vec{k}(0v-0)\right) = 2\vec{i}\Omega v$$

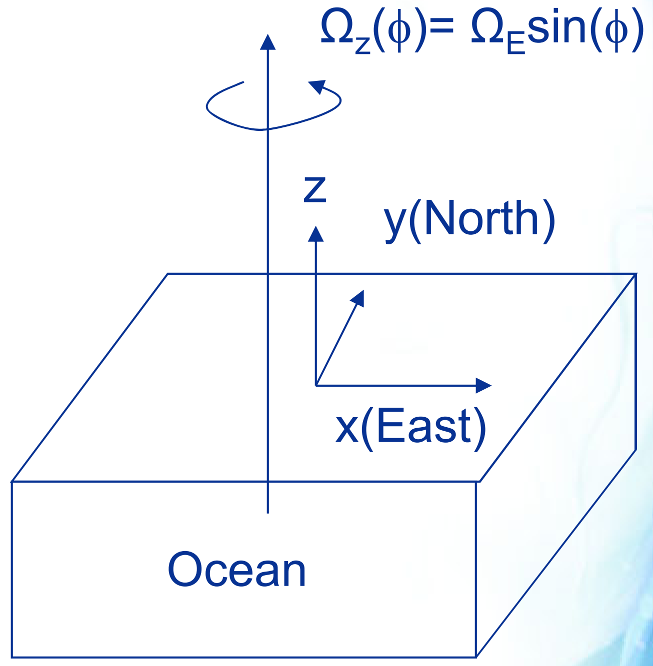
The Coriolis force acts to the right of the motion and is proportional to the velocity (for positive rotation)

Applying these ideas to flow on a rotating Earth...

Ω_E = 2πrad/24 hours



Local coordinate system (x,y,z) "glued on to Earth at latitude φ"



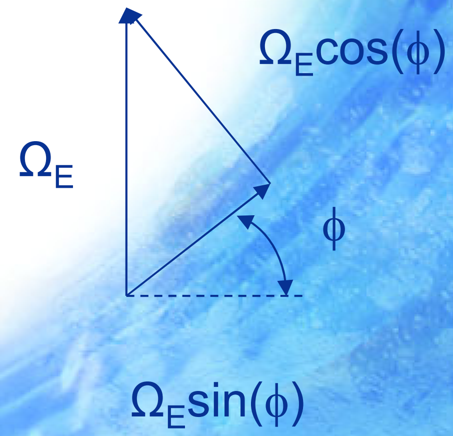
In this coordinate system, we have:

$$2\vec{\Omega} \times \vec{u}_{rel} \cong (+fv, -fu, 0)$$

$$f = 2\Omega_E \sin(\phi)$$

$$\vec{u}_{rel} = (u, v, w)$$

We generally ignore the vertical Coriolis force



Taking components of rotation

Note: Effects of rotation (Coriolis) – f varies with latitude!

$$f = 2\Omega_E \sin(\phi)$$

e.g in Taipei φ 25 deg N = 2π * 25/360 = 0.44 rad so sin(0.65) = 0.42

$\Omega_E = 2\pi/(24 * 3600) = 7.27 \times 10^{-5}$

$f = 2\Omega_E \sin(\phi) = 2\,(7.27 \times 10^{-5})(0.42) = 6.11 \times 10^{-5}$

Notes:

(1) $f = \pm\Omega_E$ at poles

(2) f > 0 in Northern hemisphere and f <0 in Southern hemisphere

(3) f = 0 at equator

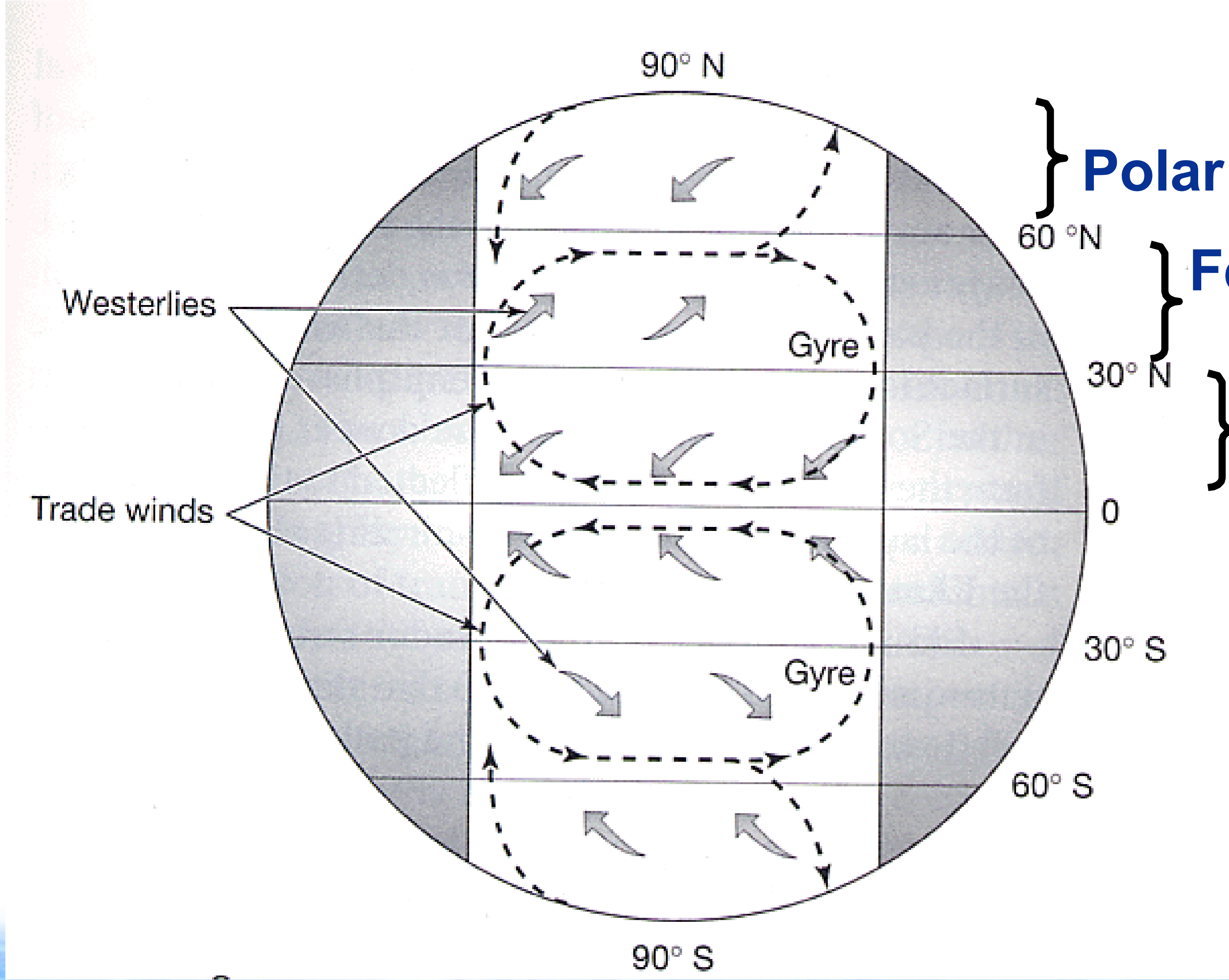
Common lingo:

Assume f is constant = "f - plane"

Assume f varies linearly with latitude $f = f_0 + \beta y$ – "Beta (β) plane"

$$\beta = \frac{\partial f}{\partial y} = 2(\Omega_E / R_E)\cos(\phi)$$

Winds and Surface Currents



Polar Cell

Ferrel Cell

Hadley Cell

Basic Ocean Structures

Warm up by sunlight!

☐ Upper Ocean (~100 m)

Shallow, warm upper layer where light is abundant and where most marine life can be found.

☐ Deep Ocean

Cold, dark, deep ocean where plenty supplies of nutrients and carbon exist.

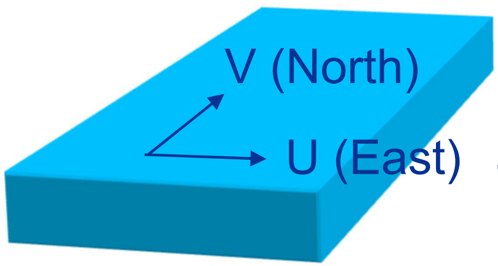
No sunlight!

Inertial oscillations: A simple flow in with fT ~ 1

(It was a dark and calm night...) We start by imaging an ocean in which nothing is happening. There are no pressure gradients, and everywhere things are rest. For a short time, a strong wind blows such that the upper ocean moves uniformly to the east as a slab with velocity $u = u_0$. Because the wind is everywhere the same, all fluid particles move in the same way. All that is left of the momentum equations are:

$$\frac{du}{dt} = 2\Omega\, v \sin\varphi = fv$$

$$\frac{dv}{dt} = -2\Omega\, u \sin\varphi = -fu$$



The first equation can be differentiated with time

$$\frac{d^2u}{dt^2} - f\frac{dv}{dt} = 0 \text{ with } \frac{dv}{dt} = -fu$$

These can be combined to give

$$\frac{d^2u}{dt^2} + f^2 u = 0$$

The solution to this (the basic oscillator) equation is:

$$u = A\cos(ft) + B\sin(ft)$$

where A and B depend on the initial conditions, i.e., A = u_0 and B is as yet undetermined.

We can find B by using the y momentum equation:

$$v = \frac{1}{f}\frac{du}{dt} = -u_0\sin(ft) + B\cos(ft)$$

$$v(0) = B\cos(0) = B = 0$$

Thus

$$u(t) = u_0\cos(ft)$$

$$v(t) = -u_0\sin(ft)$$

i.e., the velocity vector rotates clockwise with a period of

$$T=2\pi/f=12 \text{ hours}/\sin(\varphi)$$

a time known as the *inertial period*, or the period of precession of a Foucault pendulum. This period is also sometimes referred to as " pendulum day".

We can also compute the displacements by integrating the velocities in time. Since velocity is equal to the rate of change of particle position, i.e.,

$$\frac{dx}{dt} = u(t)$$

$$\frac{dy}{dt} = v(t)$$

So if we integrate these with respect to time, we find:

$$x(t) - x(0) = \int_0^t u_0 \cos(ft)dt = \frac{u_0}{f}\sin(ft)$$

$$y(t) - y(0) = \int_0^t - u_0 \sin(ft)dt = \frac{u_0}{f}[\cos(ft) - 1]$$

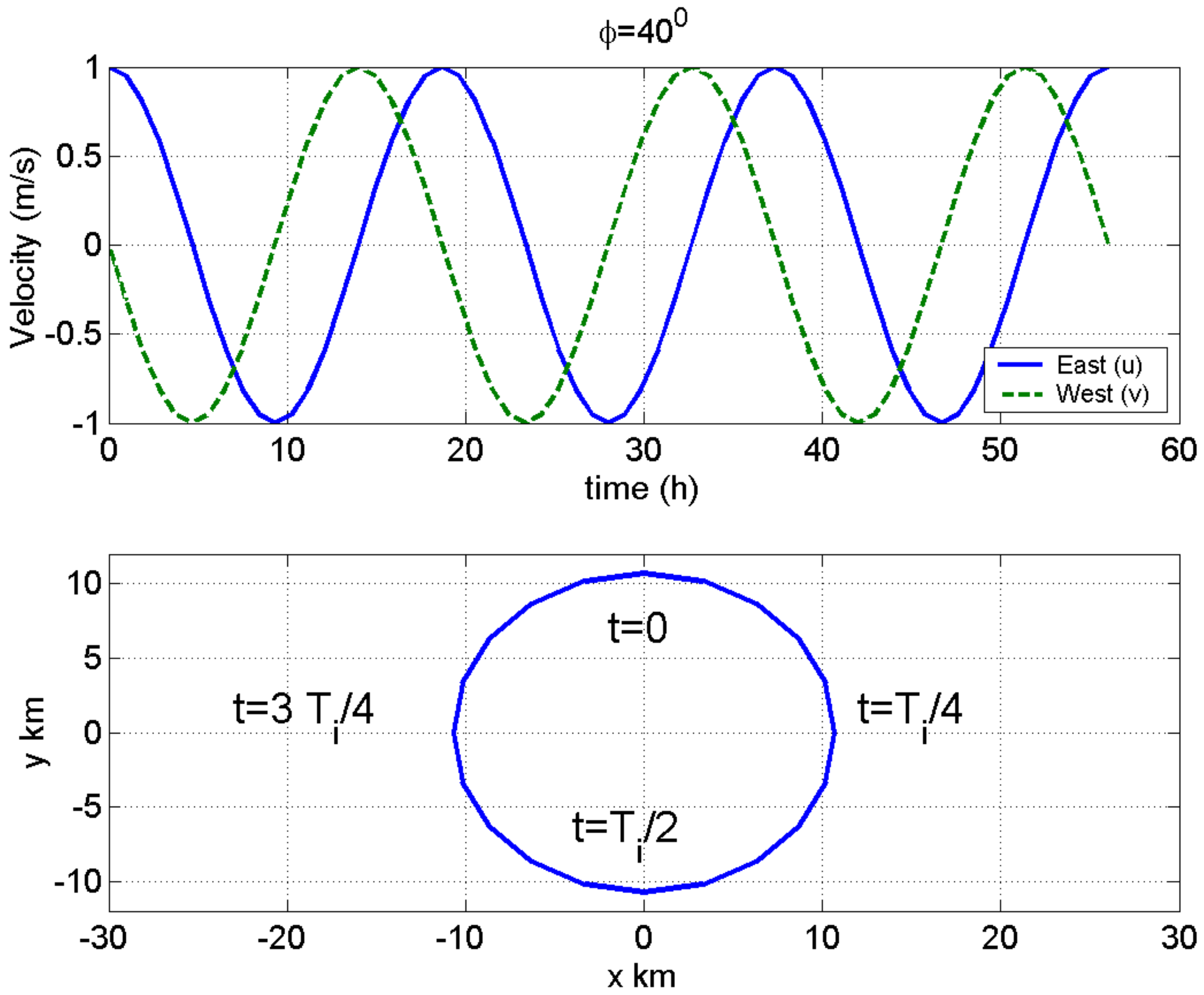
Suppose for definiteness sake, we choose x(0) = 0 and y(0) = u_0/f, then

$$x(t) = \frac{u_0}{f}\sin(ft)$$

$$y(t) = \frac{u_0}{f}\cos(ft)$$

Note that the particles move in (inertial) circles:

$$x^2 + y^2 = (\frac{u_0}{f})^2[\cos^2(ft) + \sin^2(ft)] = (\frac{u_0}{f})^2 = R_i^2$$



When it looks like....

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Upper ocean currents
39°10’ N
F=9.2 x 10^{-5}; $T_{inertial}$ 19 hours

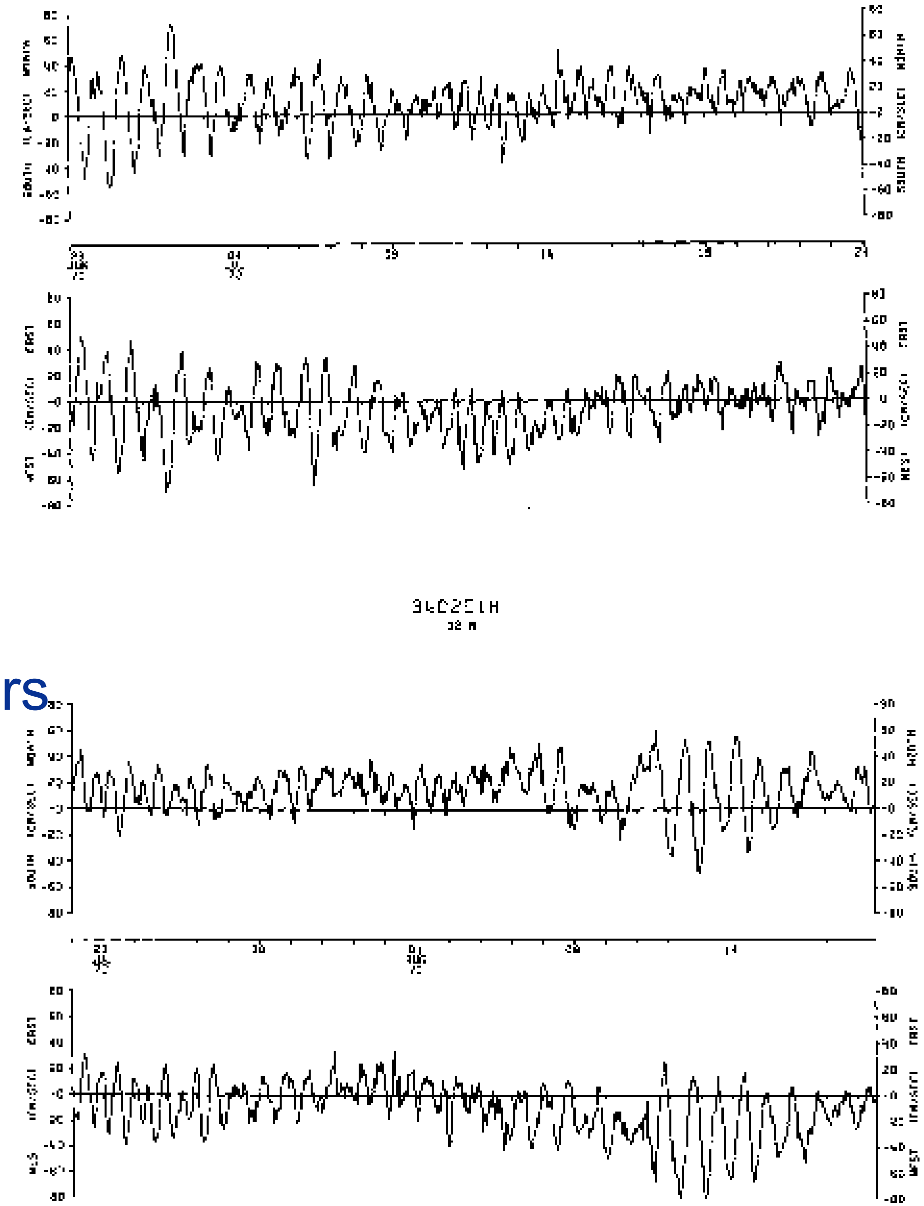


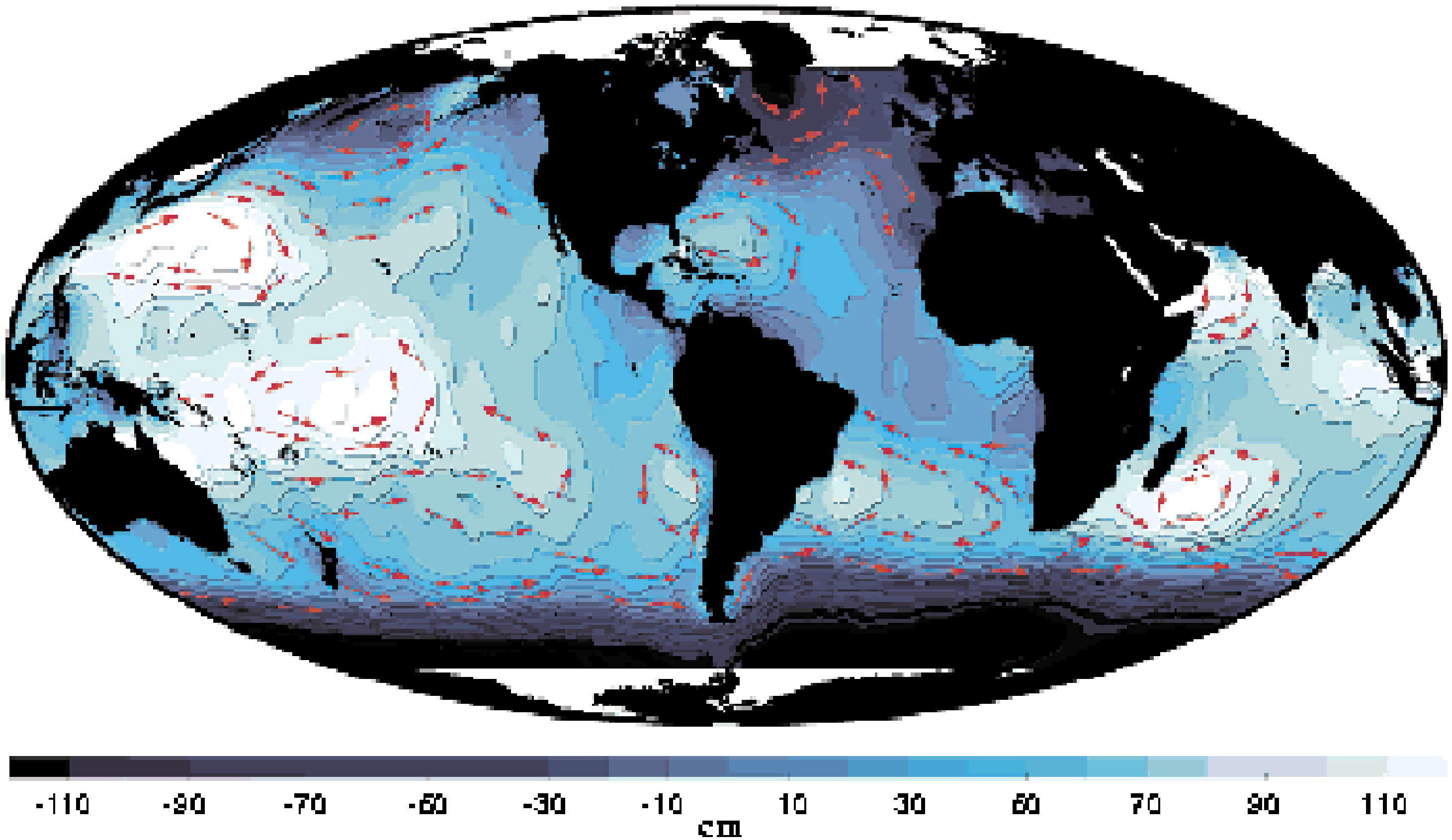
FIG. 3. Cartesian component plots at 12 m depth on mooring 340, using hourly averaged values from record 3401. Both inertial and semi-diurnal oscillations can be seen dominating the record at various times.

Properties of Near-Surface Inertial Oscillations
R. T. Pollard (1980)

https://www.youtube.com/watch?v=Dn4E3KdSUVg

Geostrophy

Sea surface dynamic topography as observed by TOPEX/POSEIDON



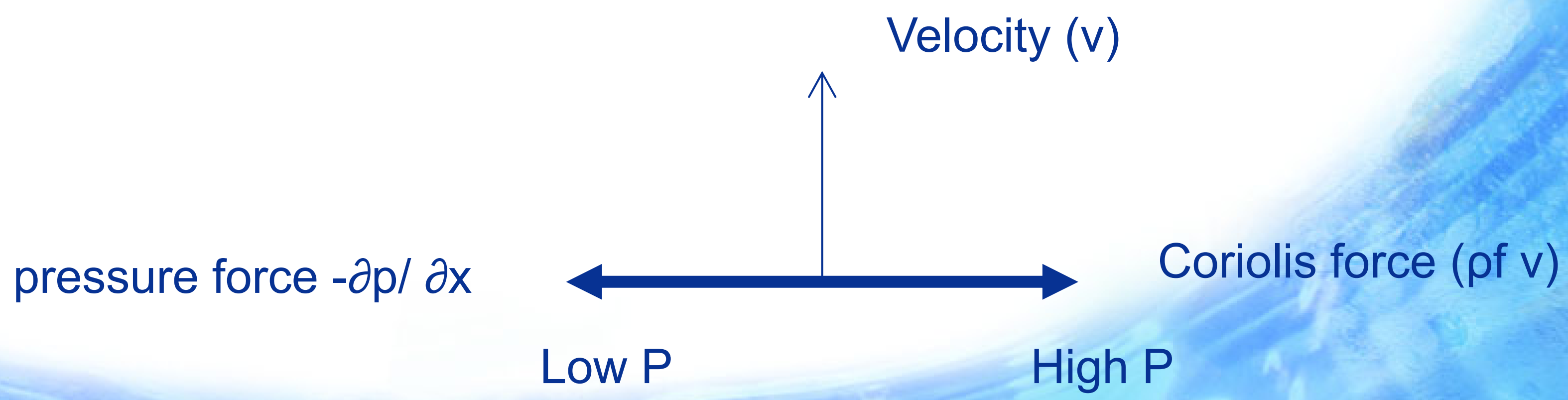
Geostrophic flows

Let us consider (for now) an idealized case with no wind stresses and also suppose that R_o = U/f L << 1 (i.e., is small), then, the (messy/complicated) momentum balance simplifies to which the Coriolis force is balanced by the force due to pressure variations:

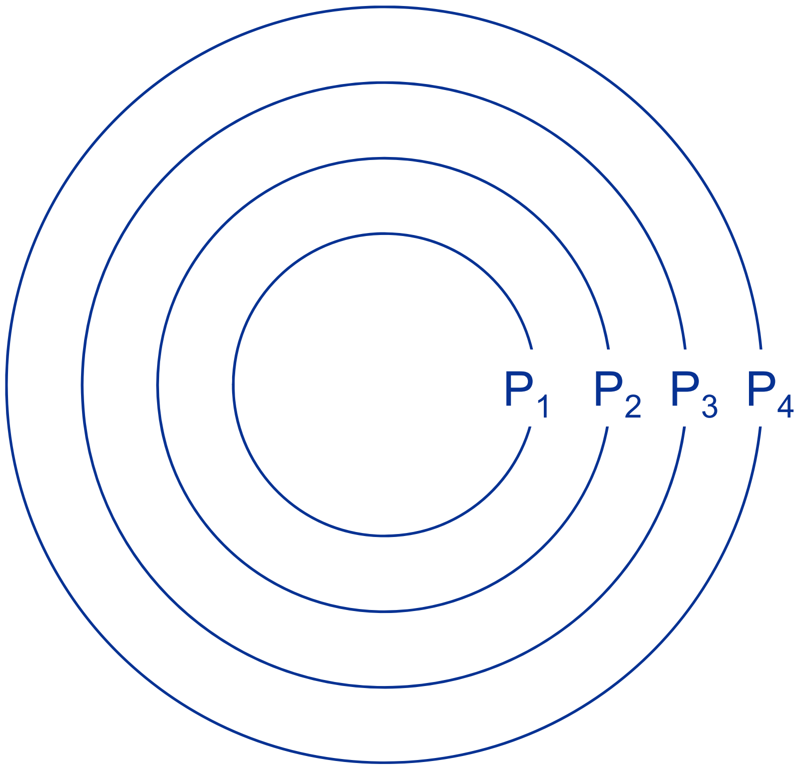
$$-\rho f\, v = -\frac{\partial p}{\partial x}$$ Zonal (east-west)

$$\rho f\, u = -\frac{\partial p}{\partial y}$$ Meridional (north-south)

Schematically, this would look like this:



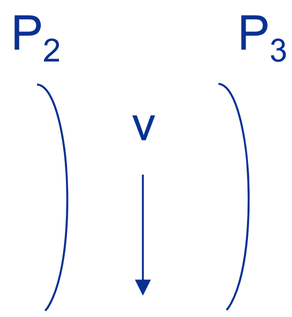
Imagine we look at a pressure field that looks like this (a “high”)



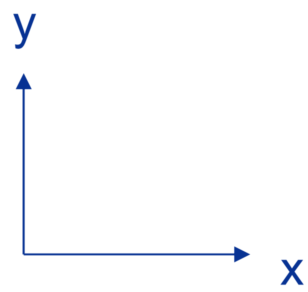
$P_1 > P_2 > P_3 > P_4$

What does flow look like?

Consider situation on east side of high



$$\frac{\partial p}{\partial x} < 0$$

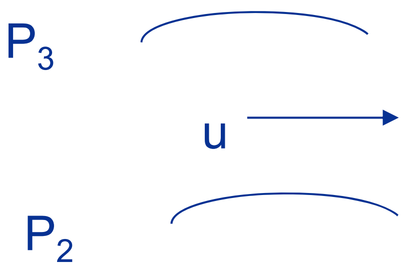


$$-\rho f v = -\frac{\partial p}{\partial x}$$

$$\rho f u = -\frac{\partial p}{\partial y}$$

$$\text{Since } \partial p / \partial x < 0, v < 0$$

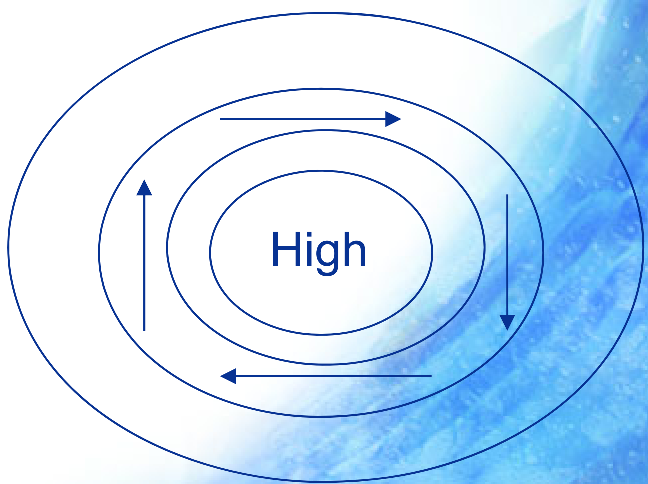
Whereas on north side of high



$$\frac{\partial p}{\partial y} < 0$$

$$\text{Since } \partial p / \partial y < 0, u > 0$$

Result (northern hemisphere):
Flow is clockwise around a high
(counterclockwise around low)

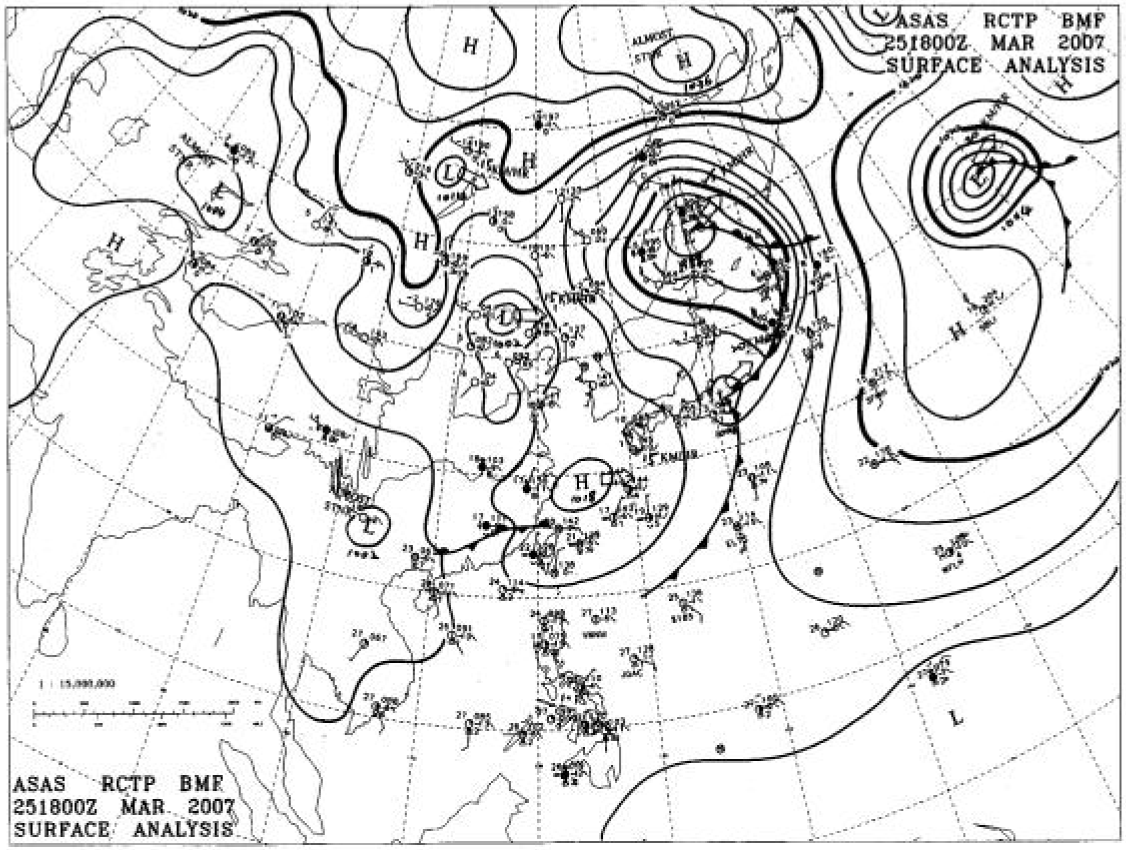


High = anticyclone
Low = cyclone

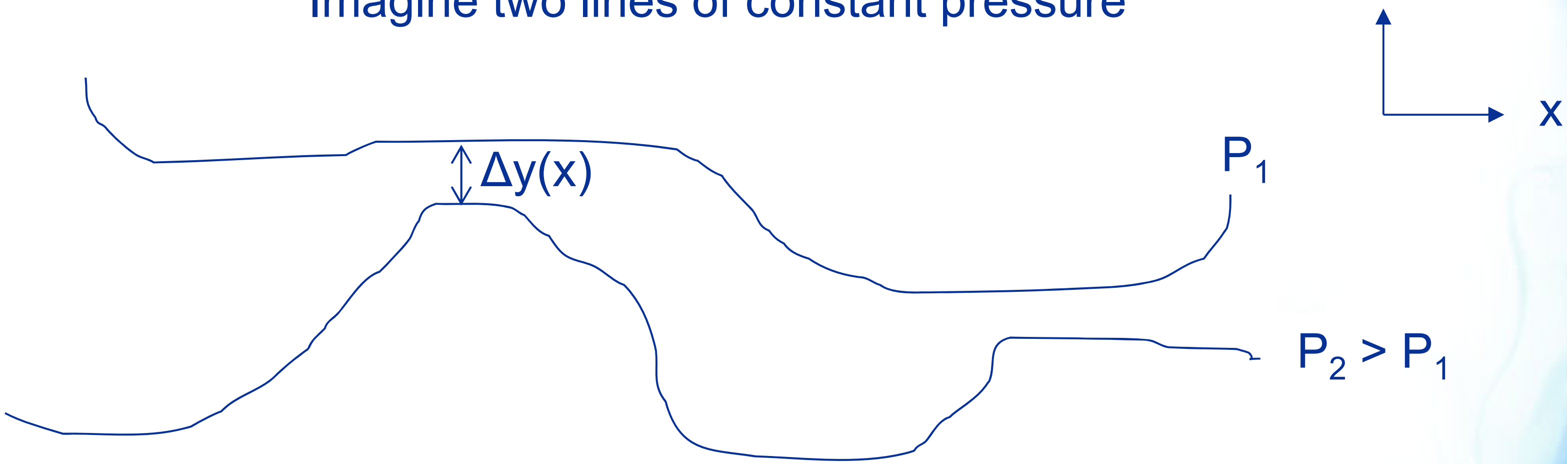
In general, flow is perpendicular to pressure gradient

$$(u,v)\circ\left(\frac{\partial p}{\partial x},\frac{\partial p}{\partial y}\right)=(u,v)\circ(-\rho fv,\rho fu)=-\rho fvu+\rho fuv=0$$

I.e., flow along lines of constant pressure!



Imagine two lines of constant pressure

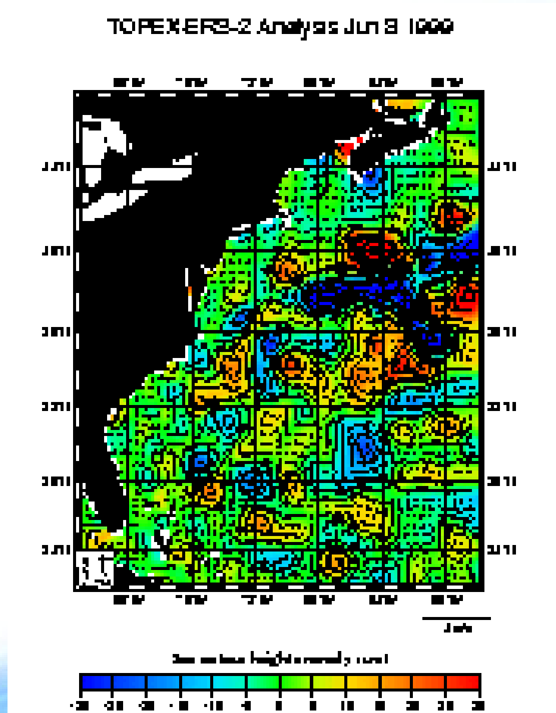


$$\frac{\partial p}{\partial y} \approx \frac{P_1 - P_2}{\Delta y} < 0$$

Note that where Δy is small, the pressure gradient is big and vice versa

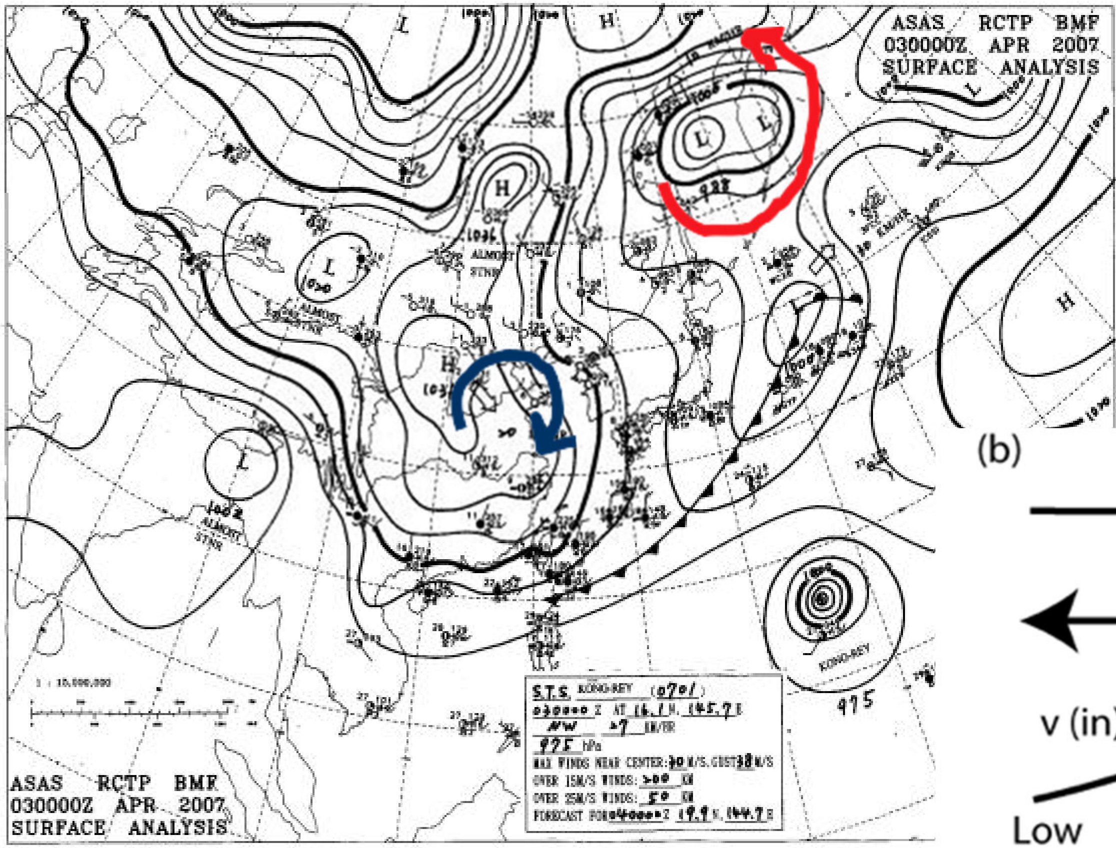
Since $u = -\frac{1}{\rho f}\frac{\partial P}{\partial y}$

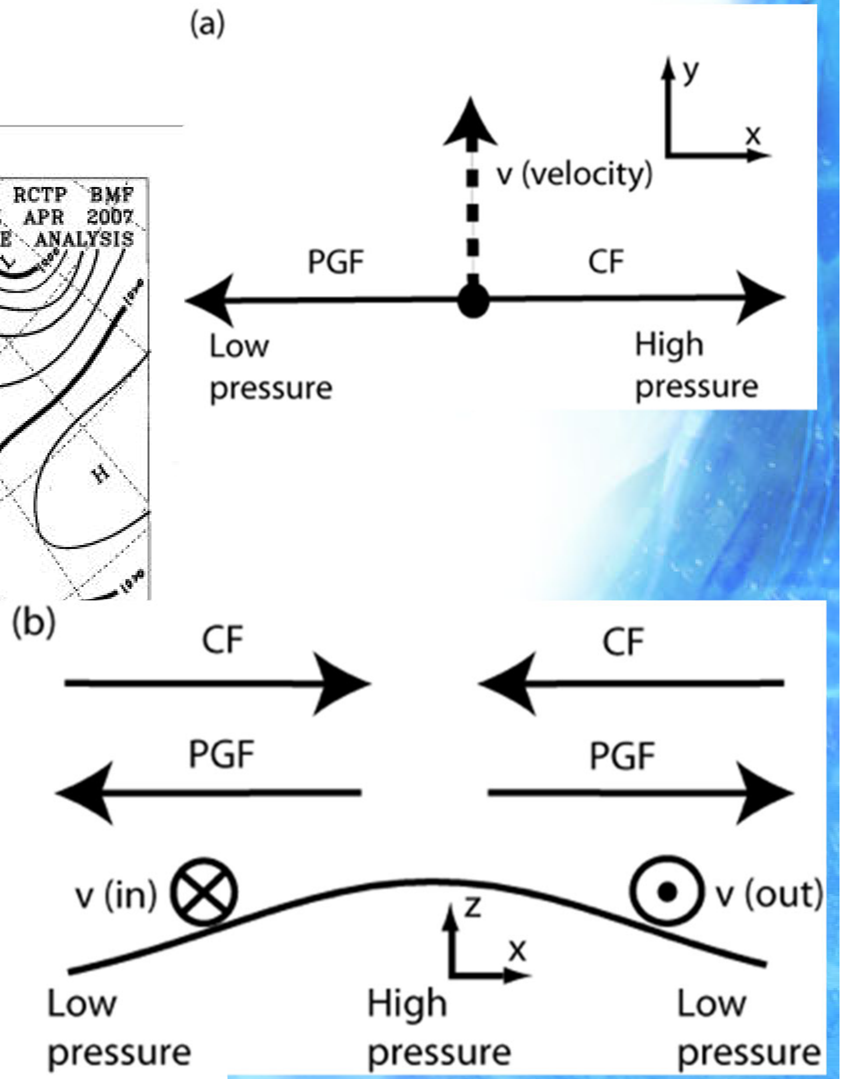
where Δy is small u is big and vice versa. Thus, where the lines of pressure are close together the flow is fast and where they are spread apart, the flow is slow.



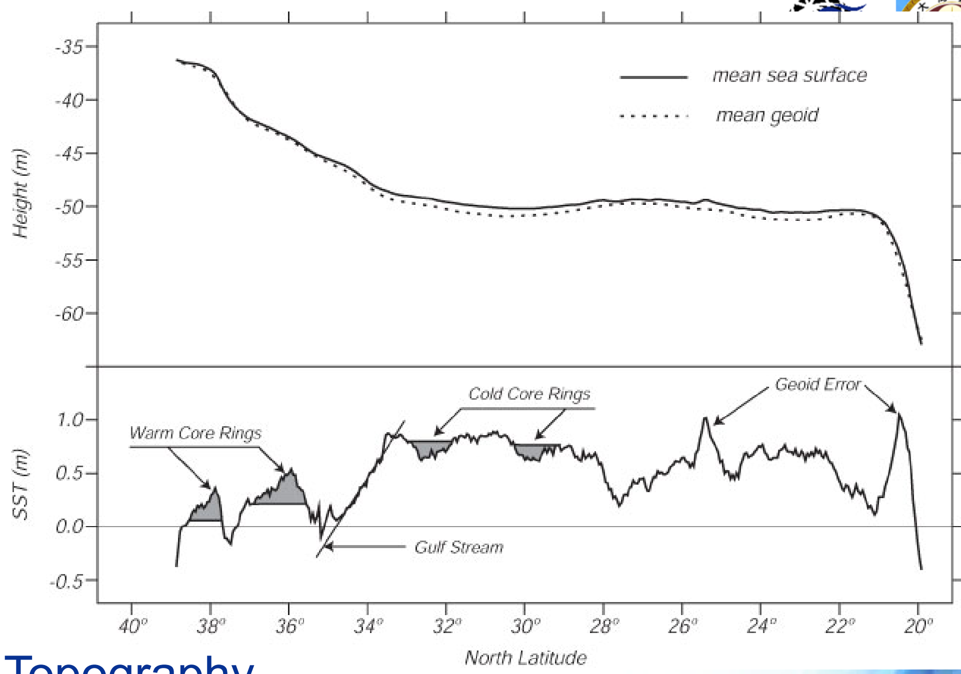
Another view

Cyclonic and Anticyclonic

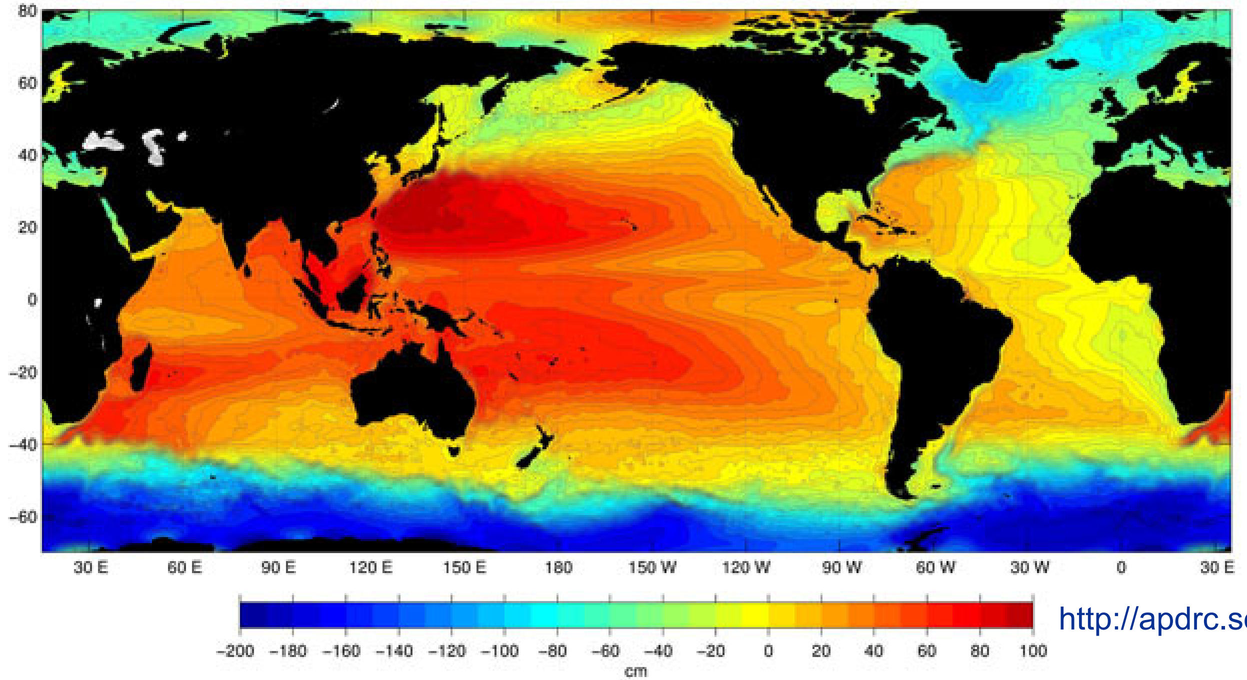




Altimetry and the Gulf Stream



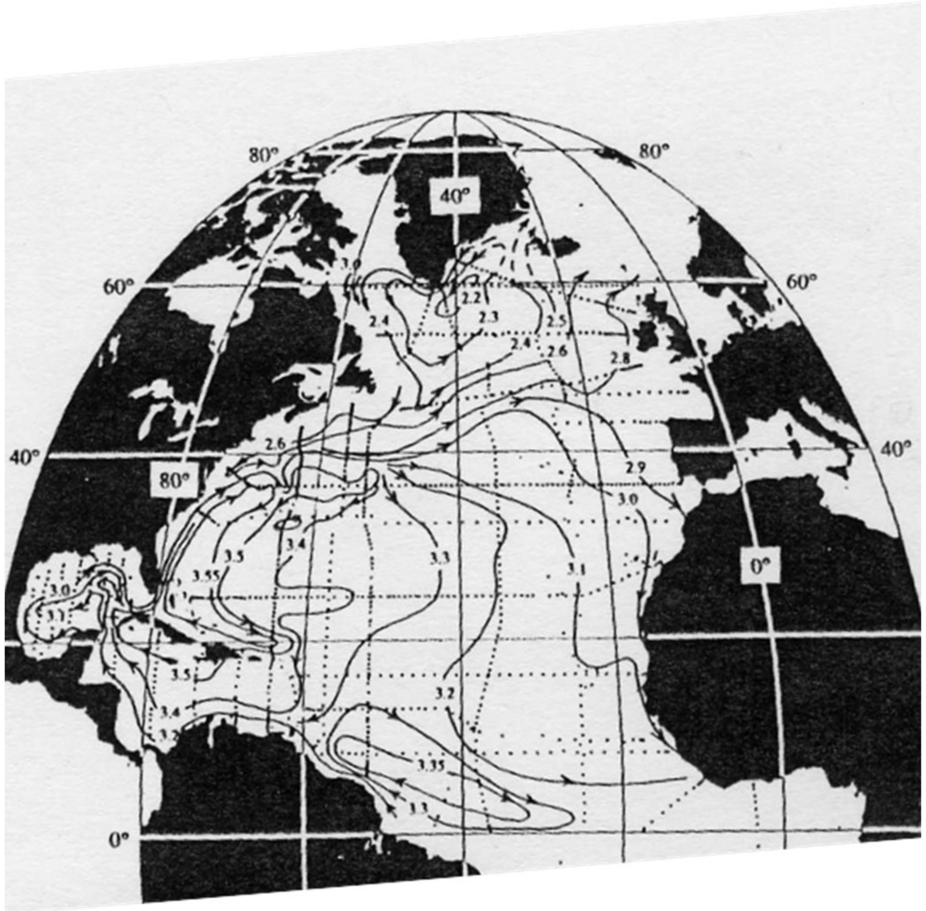
1992-2002 Mean Dynamic Ocean Topography

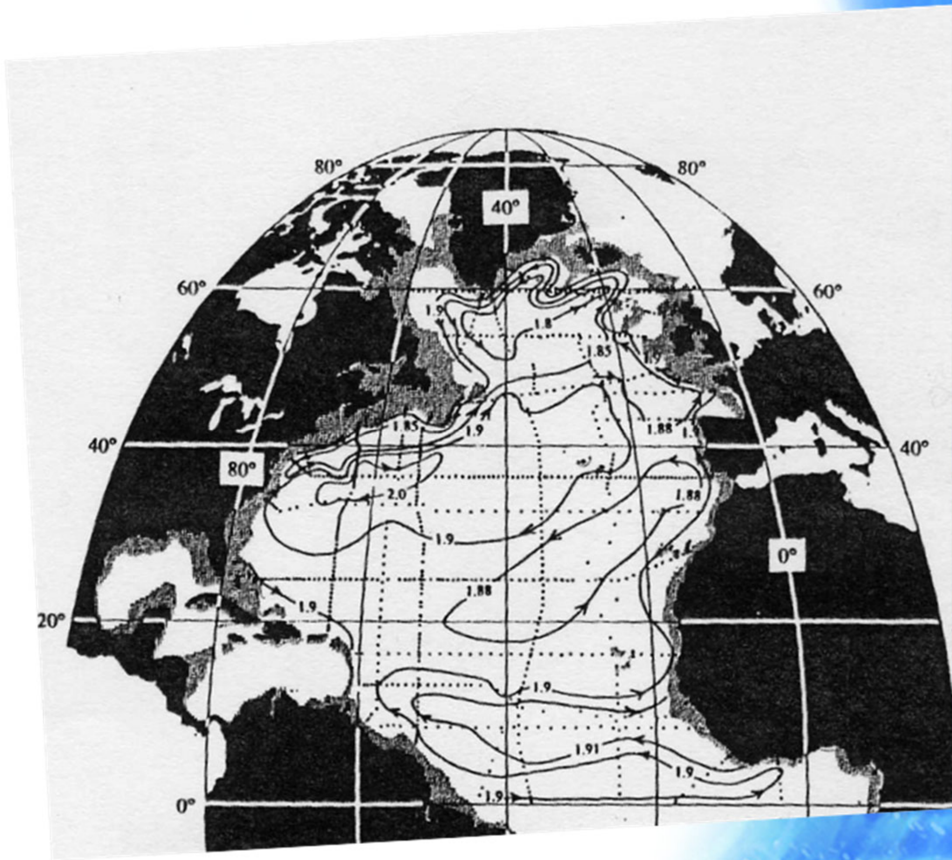


From Nikolai Maximenko (IPRC) and Peter Niiler (SIO)."

http://apdrc.soest.hawaii.edu/projects/DOT/index.html

North Atlantic steric height at sea surface and 1000 dbar (Reid, 1994)





Pacific steric height at sea surface and 1000 dbar (Reid, 1998)

Fig. 5a
Flow 0 db

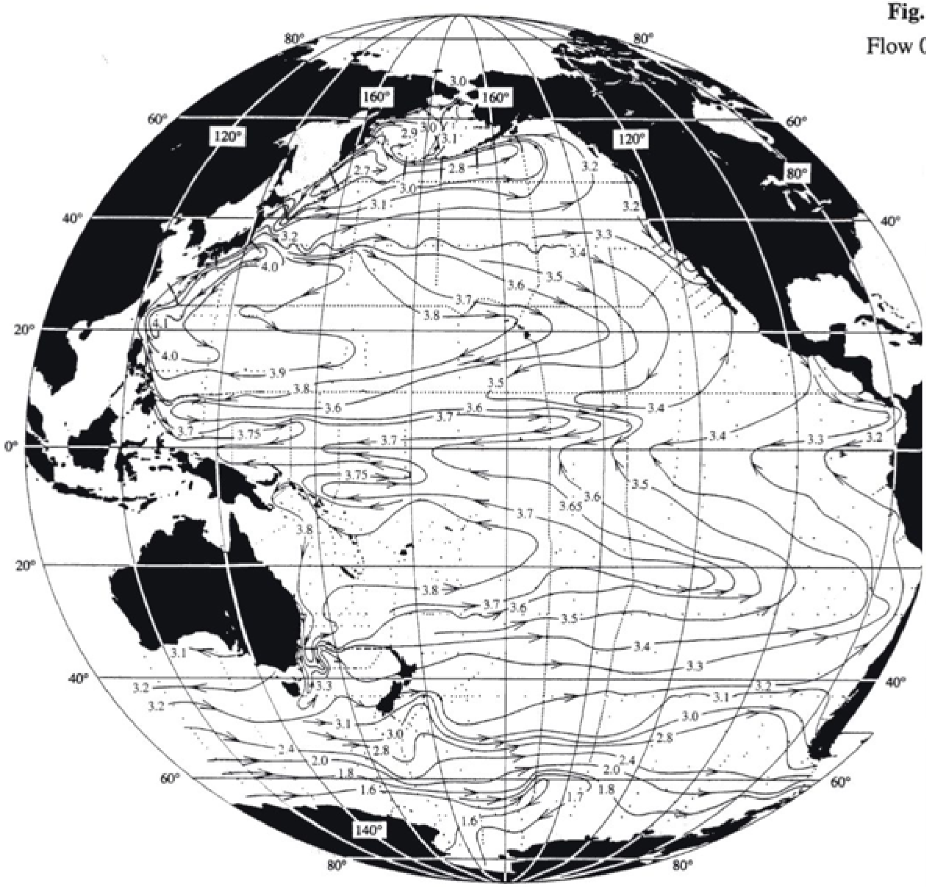
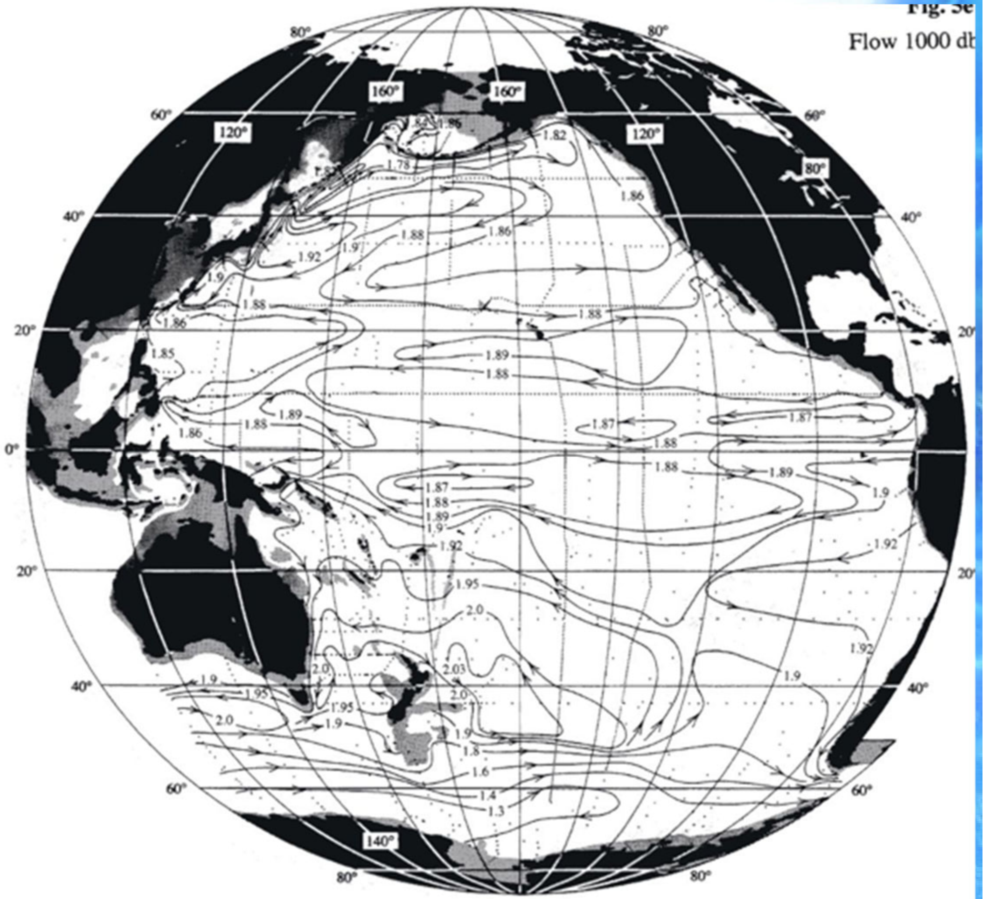


Fig. 5e
Flow 1000 db



https://www.youtube.com/watch?v=LWZEIVpuRxw

Winds on the ocean: The Ekman layer

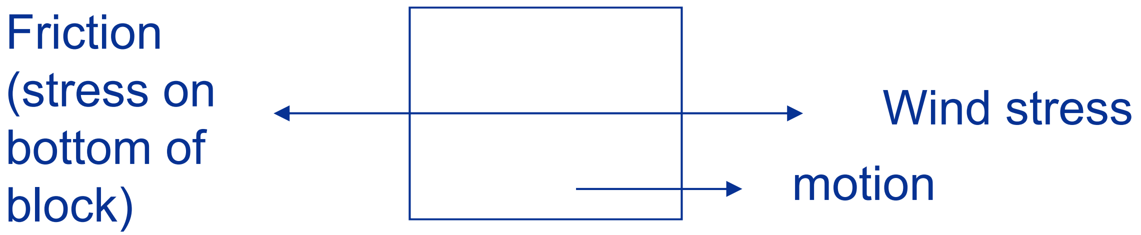
Vagn Walfrid Ekman (1874 - 1954)

The “Fram” frozen in the sea ice.

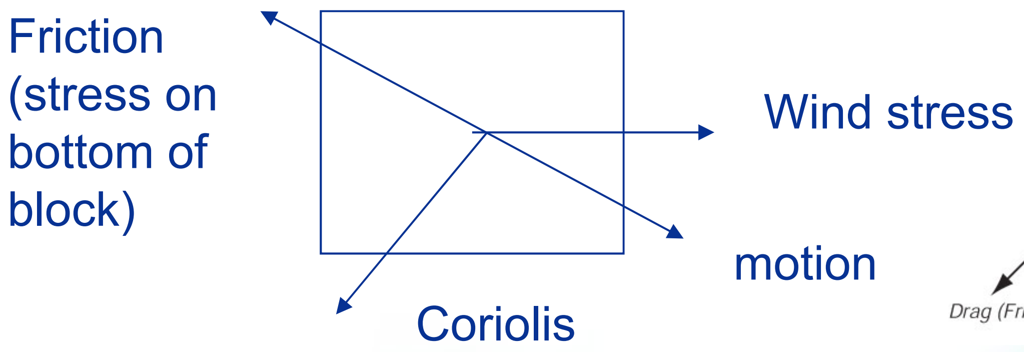
In 1893 the Norwegian adventurer Nansen loaded 6 years’ supply of food and 8 years’ supply of fuel on the specially made ship Fram, and left Oslo port, heading for the North Pole. Incredibly, his plan was to let his ship be frozen into the ice when the Arctic Sea froze, and drift with the ice, letting it carry him to the North Pole. As Nansen drifted according to plan in the Arctic Sea, he noticed something unusual. The ship did NOT drift downwind. It always drifted about twenty to forty degrees to the right of the wind’s direction ....

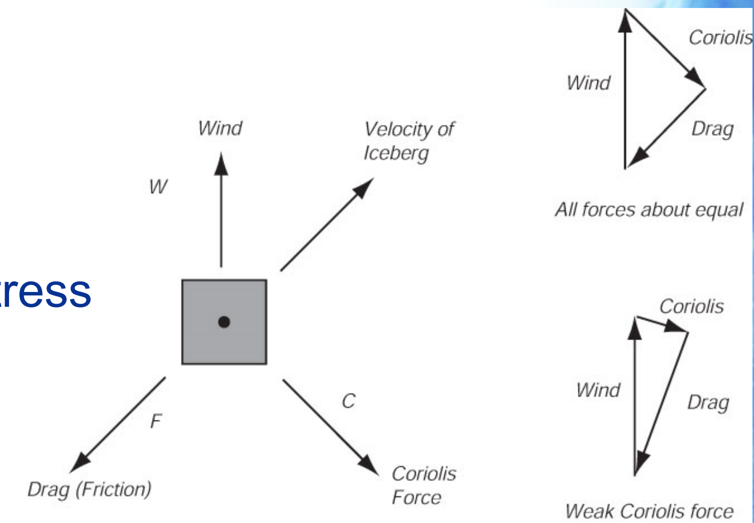
The Ekman layer – a first view of the upper ocean (theory)

We now consider how the ocean might be set in motion by the wind. To do this, we need to re-insert the shear stresses and (for now) remove the pressure gradient (as Ekman 1905). First, imagine the forces on a block of fluid if there was no Coriolis:



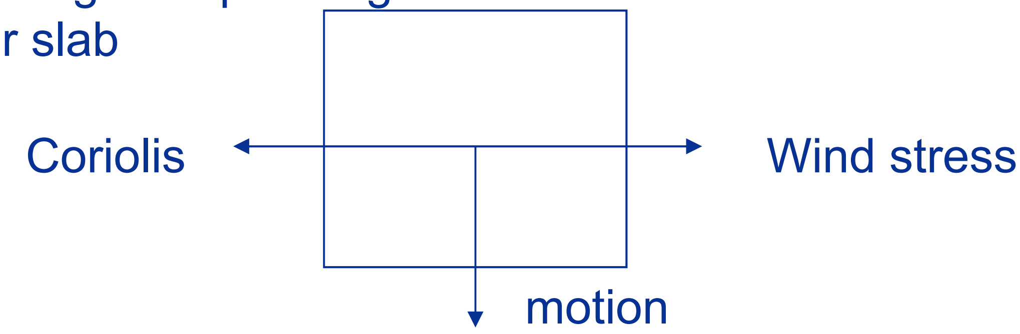
On the other hand if Coriolis is present





In order to balance Coriolis and wind stress, motion must be at some angle to the wind

Finally, imagine we go deep enough into the ocean to have the stress be zero on the bottom of our slab



Thus, we see that in an *integral* sense the motion we produce must be at right angles to the wind - thus motion is known as Ekman drift. We can see this behavior by considering our equation of motion (Coriolis balances friction):

$$-fv = \frac{1}{\rho_0}\frac{\partial \tau^{xz}}{\partial z}$$

$$fu = \frac{1}{\rho_0}\frac{\partial \tau^{yz}}{\partial z}$$

If we integrate these from the level where the stresses are zero to the surface we find that

$$-f\int_{-\infty}^{0} v\,dz = -fq_y = \frac{1}{\rho_0}\int_{-\infty}^{0}\frac{\partial \tau^{xz}}{\partial z}dz = \frac{\tau^{xz}(0)}{\rho_0}$$

$$f\int_{-\infty}^{0} u\,dz = fq_x = \frac{1}{\rho_0}\int_{-\infty}^{0}\frac{\partial \tau^{yz}}{\partial z}dz = \frac{\tau^{yz}(0)}{\rho_0}$$

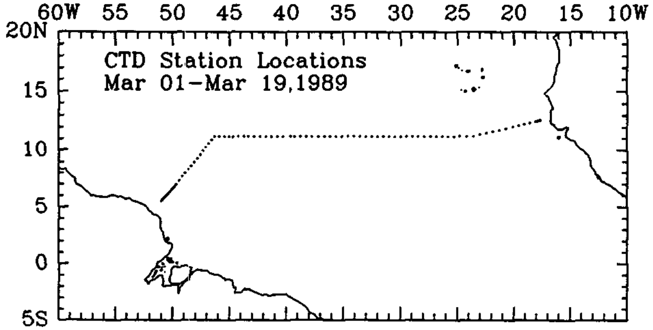
or that

$$q_y = \frac{-\tau^{xz}(0)}{f\rho_0} \quad \text{and} \quad q_x = \frac{\tau^{yz}(0)}{f\rho_0}$$

Note that transport is independent of detail of stress!

Chereskin and Roemmich (1991)-JPO:

A. Transect location
B. CTD section (temperature)
C. Geostrophic vel (from B – red to 250m)
D. Measured (smoothed ADCP)
E. Ageostrophic = D-C

A


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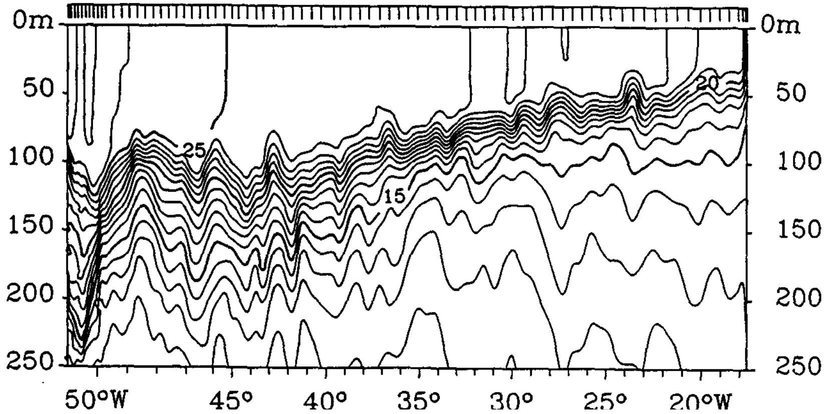

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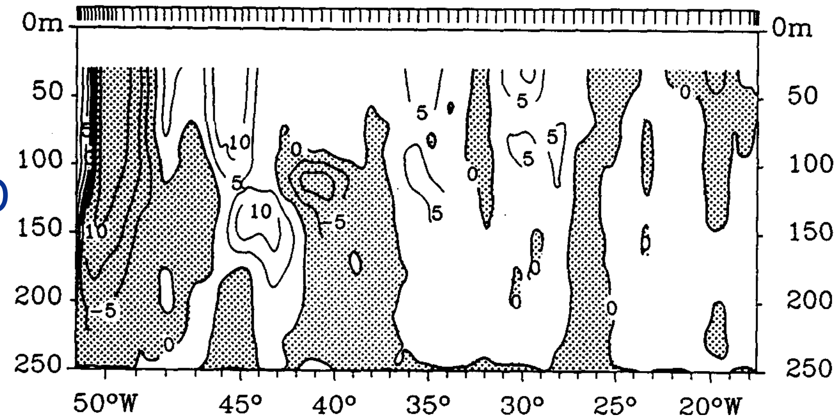
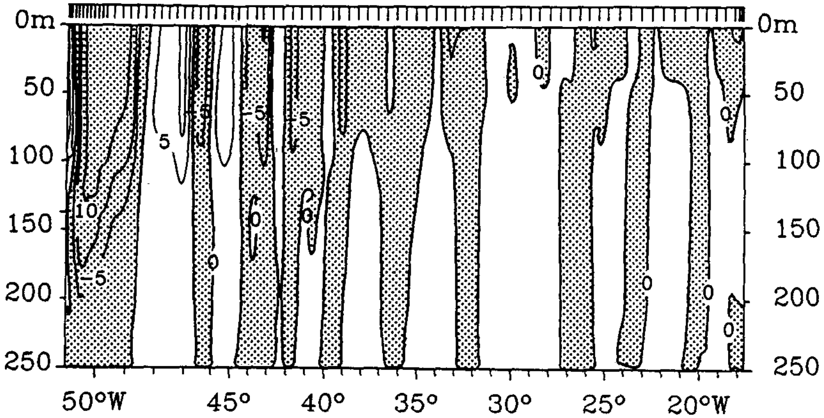


C


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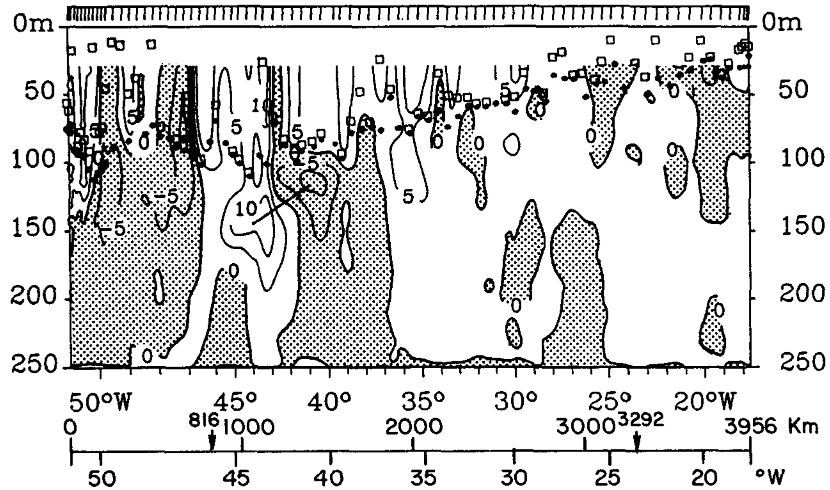


The comparison to Ekman transport theory

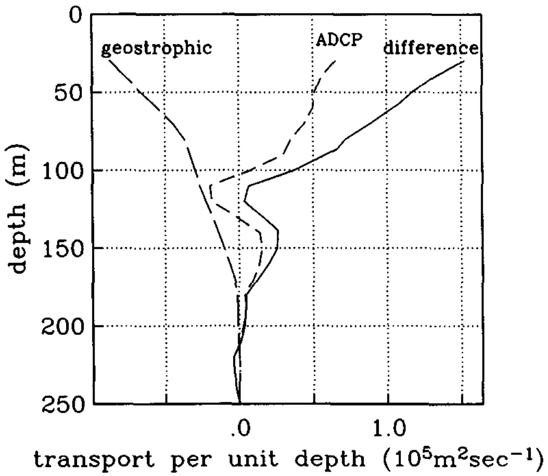


FIG. 3. Basin-integrated transport per unit depth (10^5 m^2 s^{-1}). Solid curve is the ageostrophic transport (ADCP − geostrophic). Large dashed curve is geostrophic transport; small dashed curve is ADCP transport.

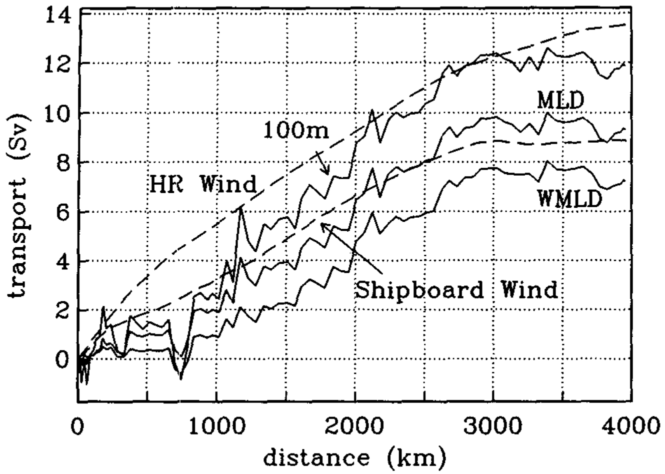


FIG. 6. Cumulative transport (10^6 m^2 s^{-1}) along the ship track (Senegal is on the right). Dashed curves represent the transport estimated from climatological (top) and shipboard (bottom) winds. Solid curves represent the transport estimated from ageostrophic velocity integrated to three depths (top to bottom): 100 m, the MLD, the WMLD.

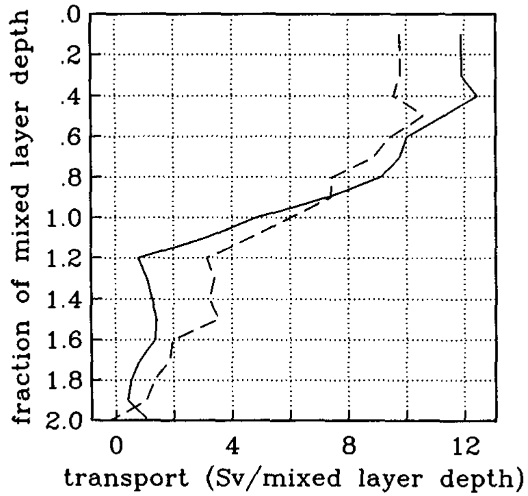


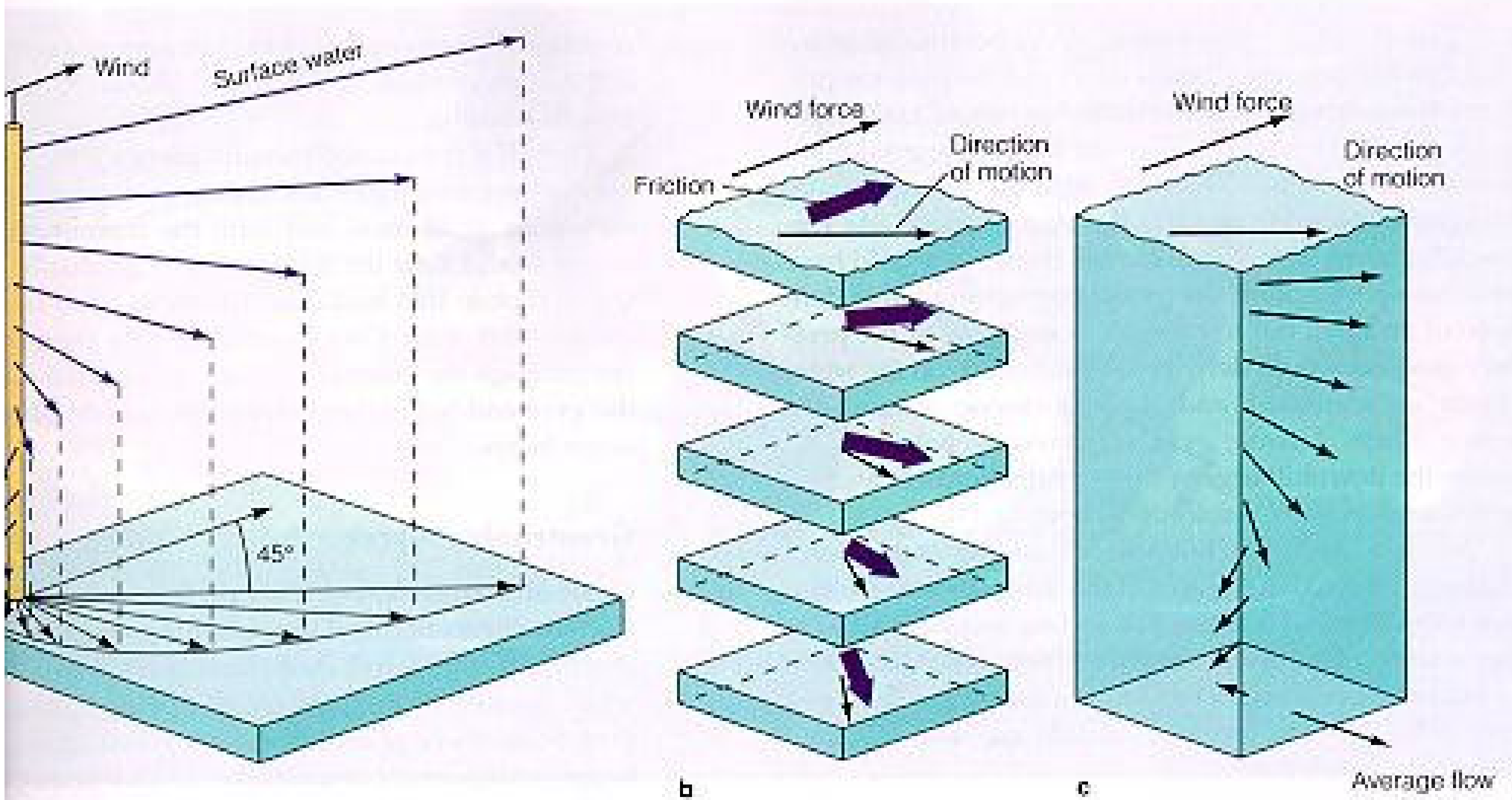
FIG. 5. Basin-integrated transport per unit of mixed layer depth (10^6 m^3 s^{-1}). The mixed layer depth for the solid (dashed) curve is defined as the depth where the temperature is 1.0 (0.1) °C cooler than the temperature at 6 m. Velocity at 0–20 m was set equal to velocity at 30 m (slab extrapolation). Vertical coordinate is fraction of mixed layer depth.

TABLE 1. Estimates of Ekman transport (units are 10^6 m^3 s^{-1}).

Estimates based on ageostrophic velocity	
Integral to 100 m*	12.0 ± 5.5
Integral to MLD	9.3
Integral to WMLD	7.0
Estimates based on in-situ wind stress	
Measured wind, March 1989*	8.8 ± 1.9
Beaufort wind, March 1989	14.5
Beaufort wind, February 1989	24.8
Estimates based on climatological wind stress	
Hellerman-Rosenstein March mean*	13.5 ± 0.3

* Estimates with error bars are considered "best" estimates.

The vertical structure of the ideal Ekman Layer



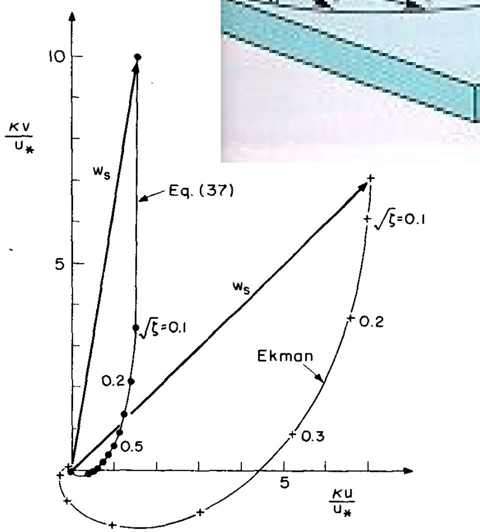
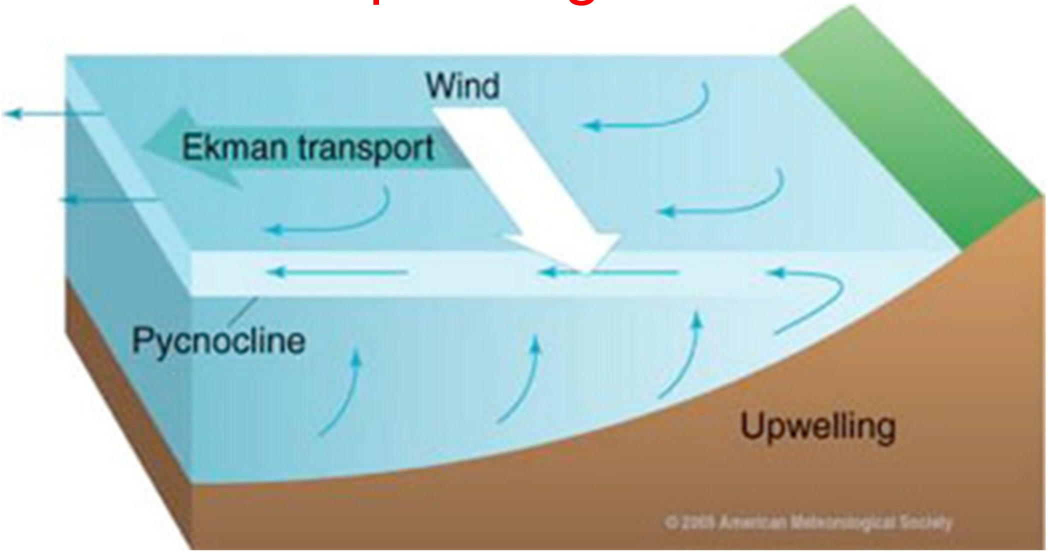


FIG. 1. Vertical velocity structure of a pure drift current in an infinitely deep homogeneous ocean of infinite lateral extent, comparing the turbulent Ekman spiral (●) and the classical Ekman spiral (+).

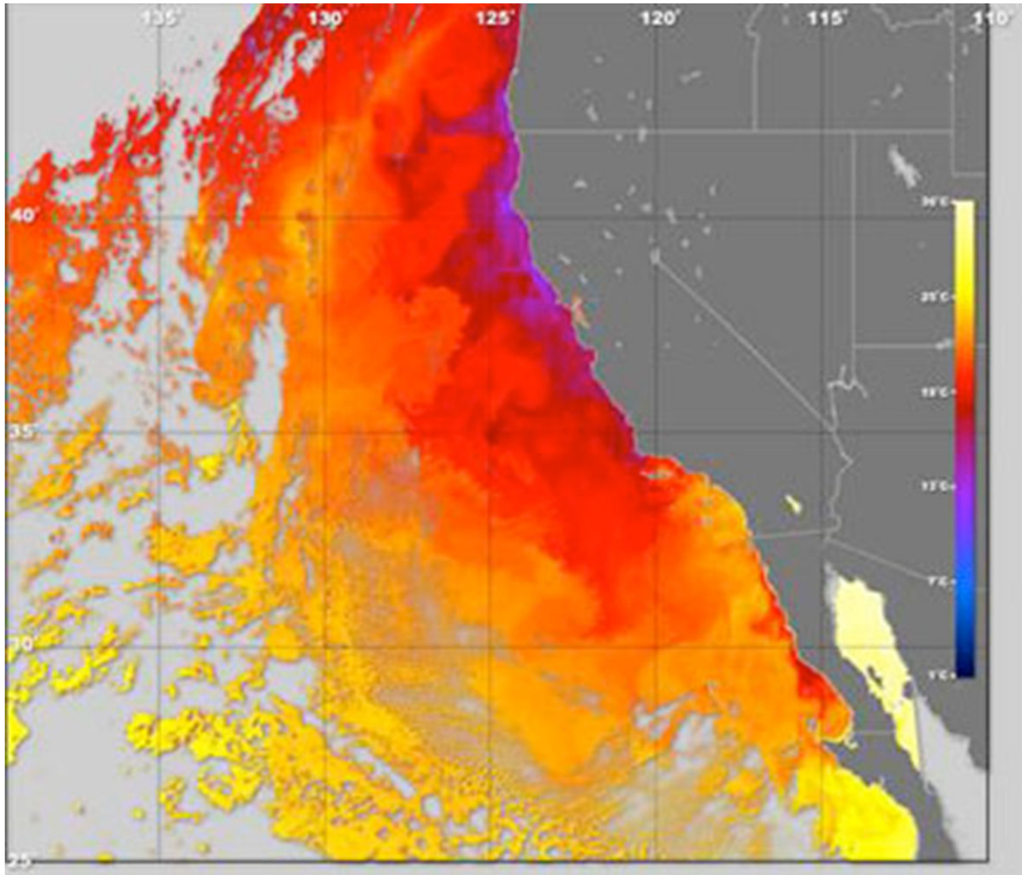
Effects of turbulent viscosity that increases linearly with distance, as calculated by Madsen (1977)

Upwelling



Where Ekman transport moves surface waters away from the coast, surface waters are replaced by water that wells up from below in the process known as upwelling.

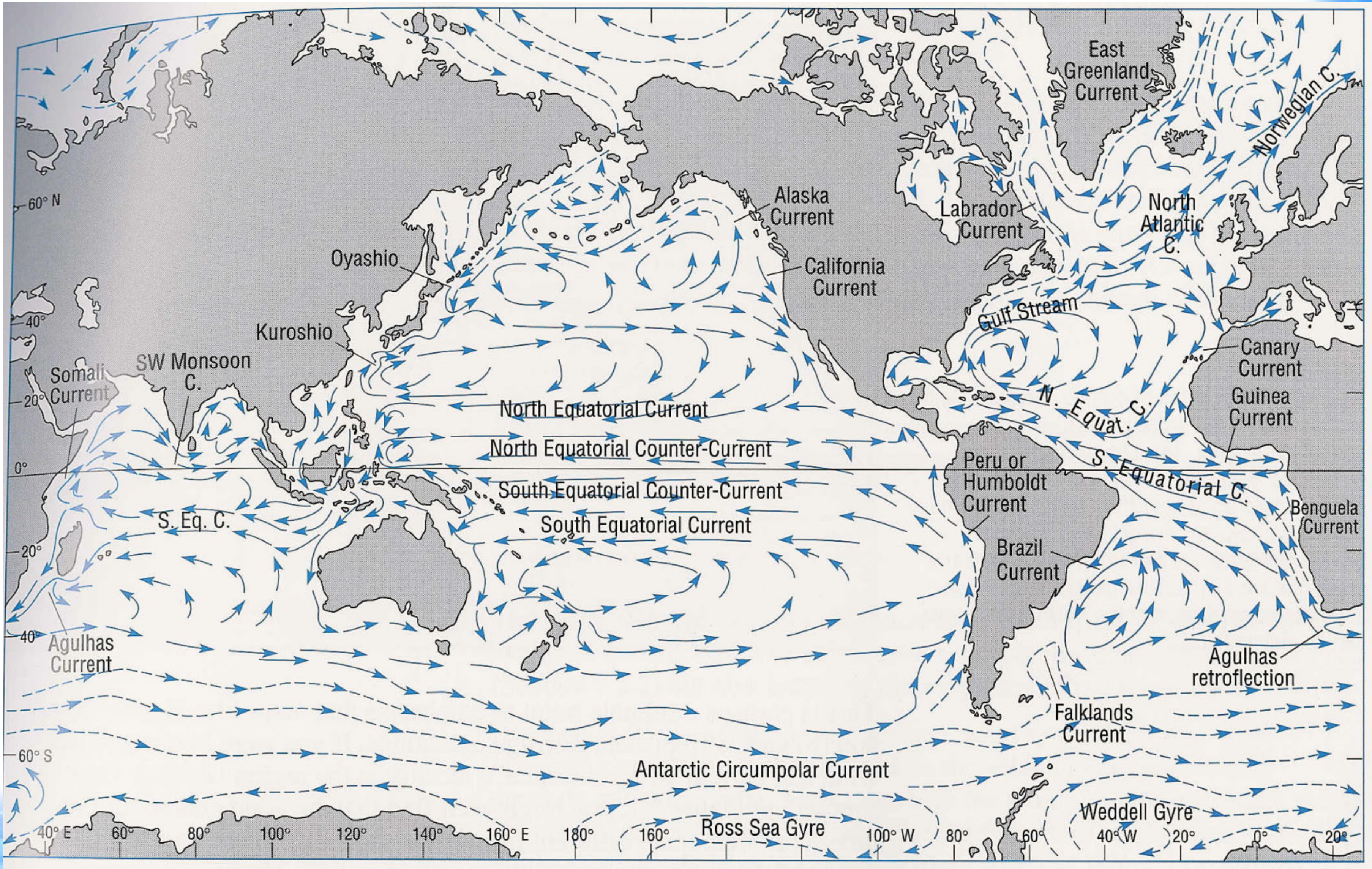
- the combination of persistent winds, Earth's rotation (the Coriolis effect), and restrictions on lateral movements of water caused by shorelines and shallow bottoms induces upward and downward water movements. The Coriolis effect plus the frictional coupling of wind and water (Ekman transport) cause net movement of surface water at about 90 degrees to the CW of the wind direction in the Northern Hemisphere.
- Coastal upwelling occurs where Ekman transport moves surface waters away from the coast; surface waters are replaced by water that wells up from below.



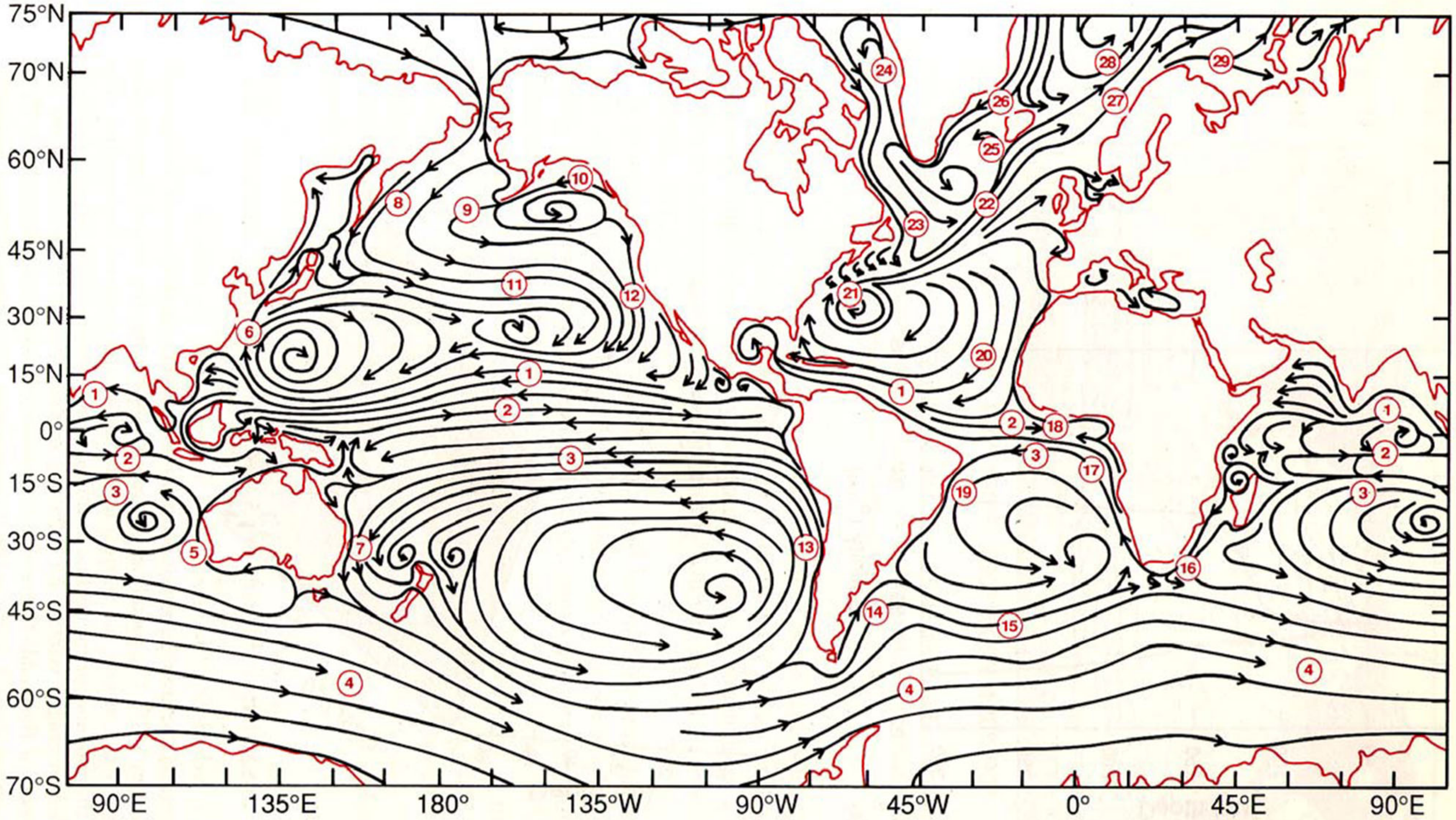
Eastern boundary current flowing southward along the California coast of the Western, United States. The image is created from Sea Surface Temperature (SST) data from the MODIS instrument, on the Aqua satellite and shows upwelling in the California Current system. The range of orange and purple colors, depict cooler water temperatures resulting from upwelling. The grey shapes, on the left are clouds.

https://www.youtube.com/watch?v=KBKmKI3tl4Q

Global Major Currents

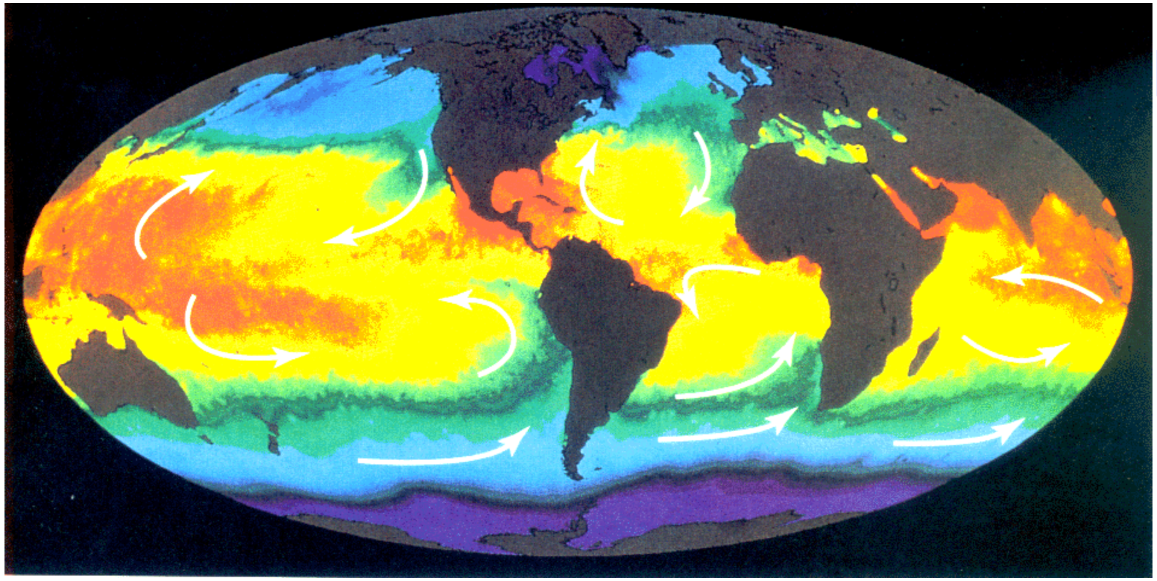


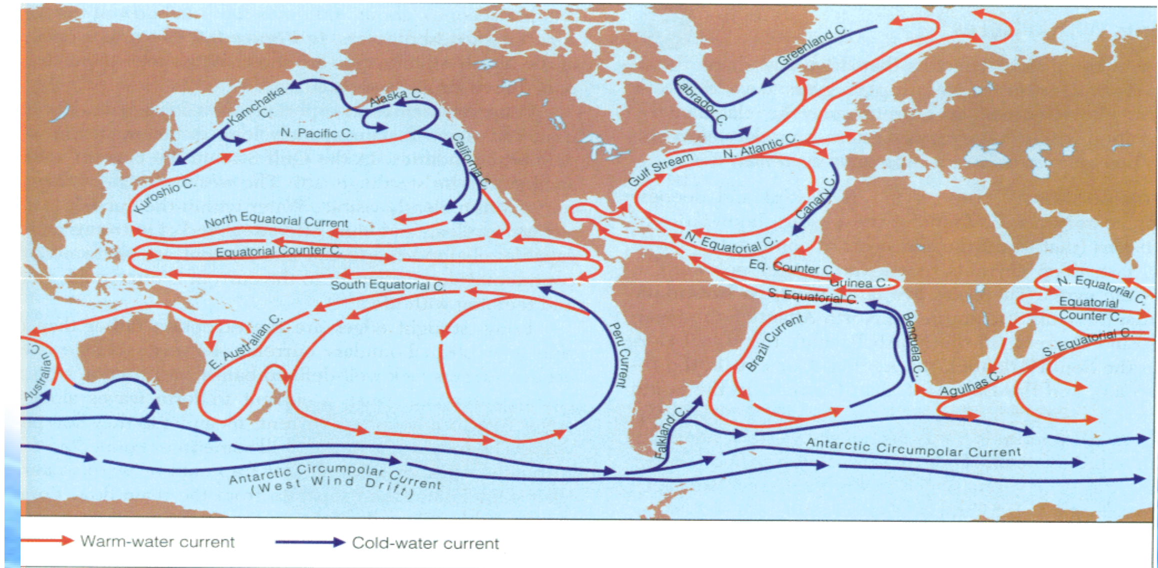
Global Surface Currents



1 North Equatorial Current
2 Equatorial Countercurrent
3 South Equatorial Current
4 West Wind Drift
5 West Australian Current
6 Kuroshio Current
7 East Australian Current
8 Oyashio Current
9 Aleutian Current
10 Alaska Current
11 North Pacific Current
12 California Current
13 Peru Current
14 Falkland Current
15 South Atlantic Current
16 Agulhas Current
17 Benguela Current
18 Guinea Current
19 Brazil Current
20 Canary Current
21 Gulf Stream
22 North Atlantic Current
23 Labrador Current
24 West Greenland Current
25 Irminger Current
26 East Greenland Current
27 Norway Current
28 Spitsbergen Current
29 North Cape Current

Six Great Current Circuits in the World Ocean





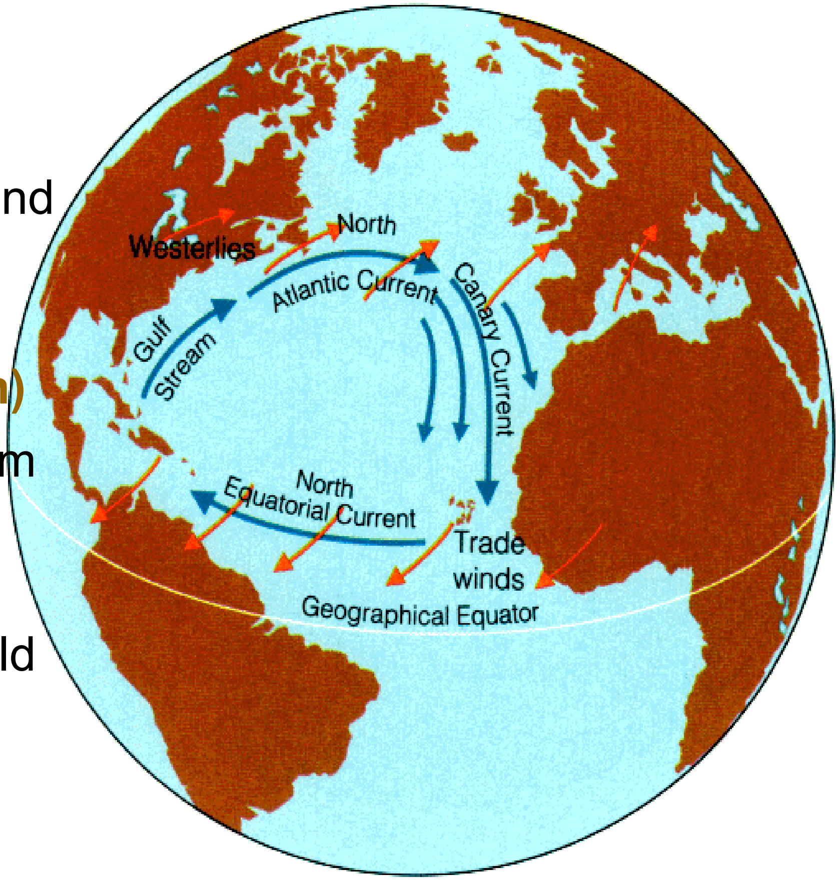
- **5 of them are geostrophic gyres:**
  - North Pacific Gyre
  - South Pacific Gyre
  - North Atlantic Gyre
  - South Atlantic Gyre
  - Indian Ocean Gyre
- **The 6th and the largest current:** Antarctic Circumpolr Current (also called West Wind Drift)

Characteristics of the Gyres

- **Currents are in geostropic balance**
- **Each gyre includes 4 current components:**

two boundary currents: western and eastern

two transverse currents: easteward and westward



Western boundary current (jet stream of ocean)

the fast, deep, and narrow current moves warm water poleward (transport ~50 Sv or greater)

Eastern boundary current

the slow, shallow, and broad current moves cold water equatorward (transport ~ 10-15 Sv)

Trade wind-driven current

the moderately shallow and broad westward current (transport ~ 30 Sv)

Westerly-driven current

the wider and slower (than the trade wind-driven current) eastward current

Volume transport unit:

1 sv = 1 Sverdrup = 1 million m^3/sec

(the Amazon river has a transport of ~0.17 Sv)

Major Current Names

☐ Western Boundary Current

Gulf Stream (in the North Atlantic)
Kuroshio Current (in the North Pacific)
Brazil Current (in the South Atlantic)
Eastern Australian Current (in the South Pacific)
Agulhas Current (in the Indian Ocean)

☐ Trade Wind-Driven Current

North Equatorial Current
South Equatorial Current

Eastern Boundary Current

Canary Current (in the North Atlantic)
California Current (in the North Pacific)
Benguela Current (in the South Atlantic)
Peru Current (in the South Pacific)
Western Australian Current (in the Indian Ocean)

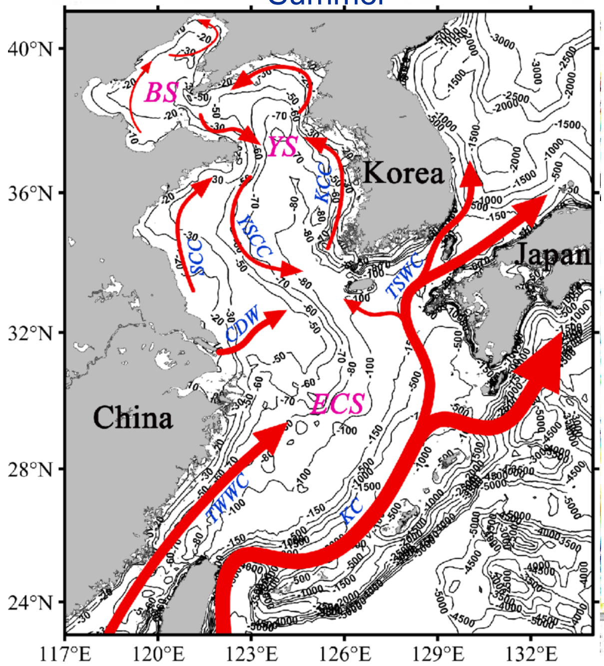
☐ Westerly-Driven Current

North Atlantic Current (in the North Atlantic)
North Pacific Current (in the North Pacific)

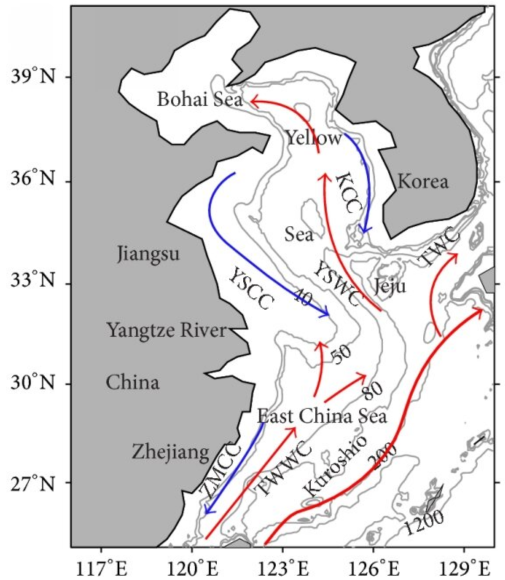
Asian Marginal Sea Circulation

Institute of Oceanography, National Taiwan University

Summer

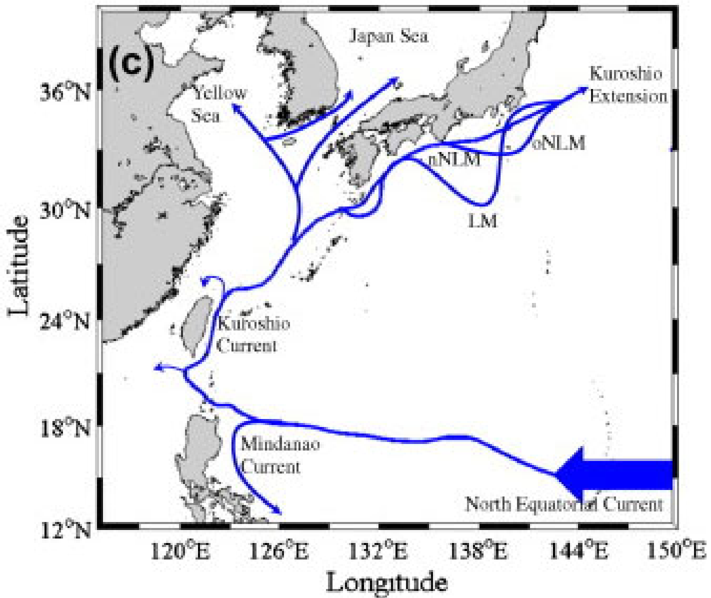


Winter

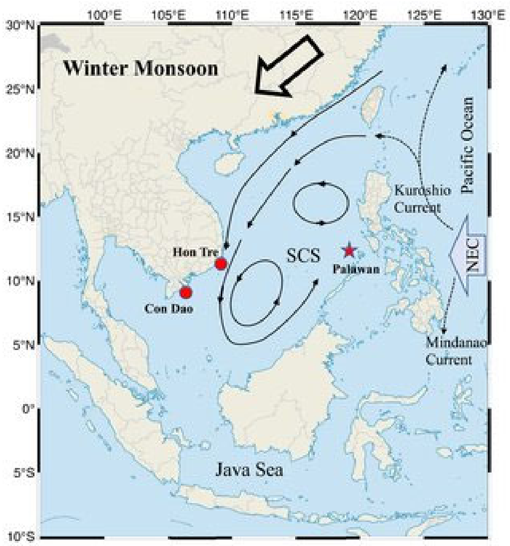


East China Sea and Yellow Sea

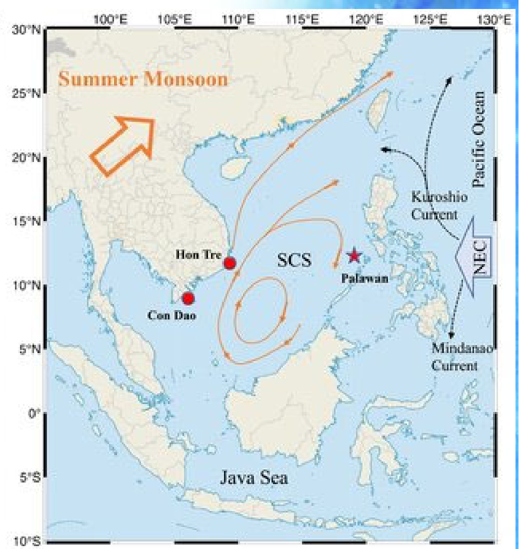
South China Sea



Winter

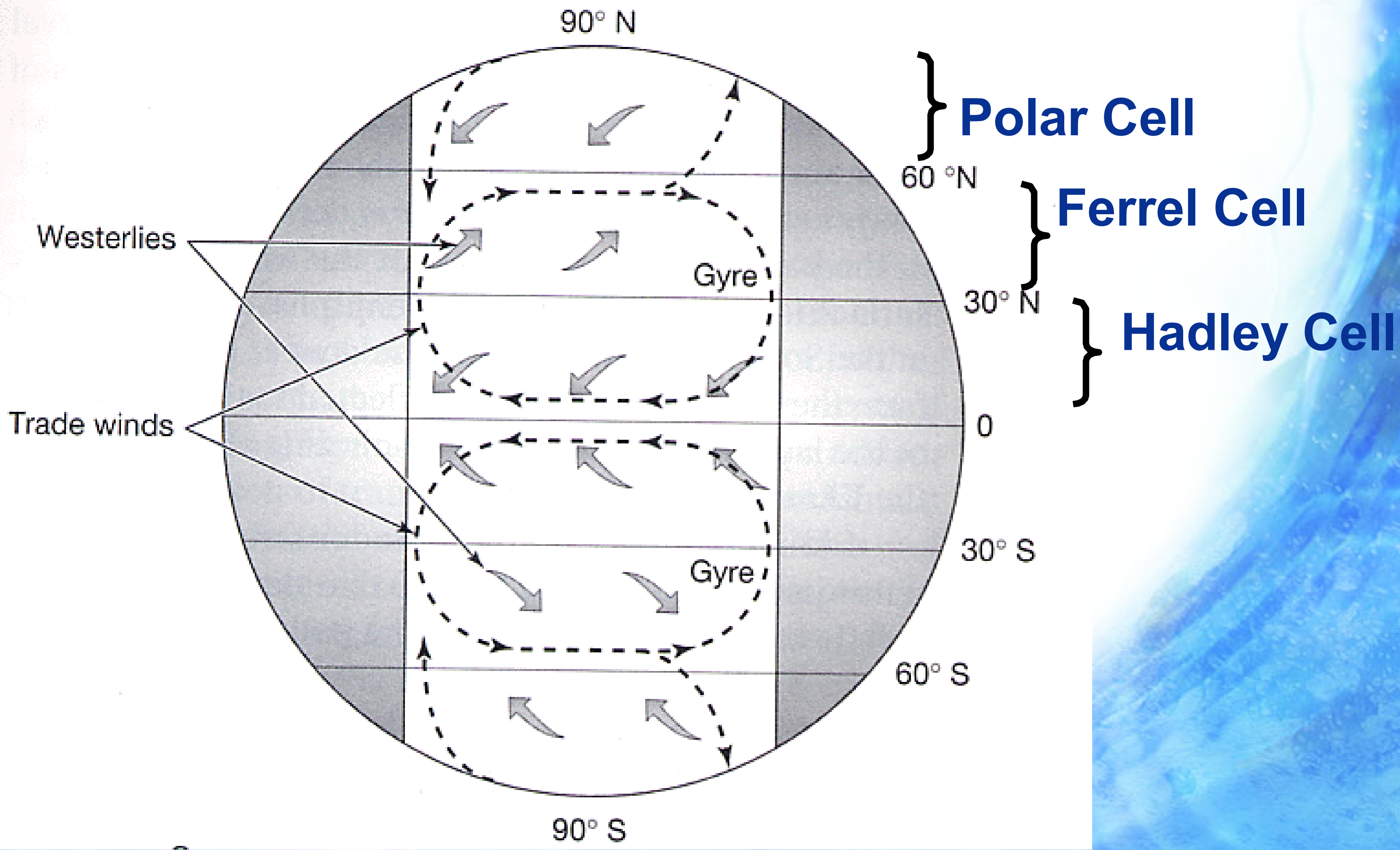


Summer

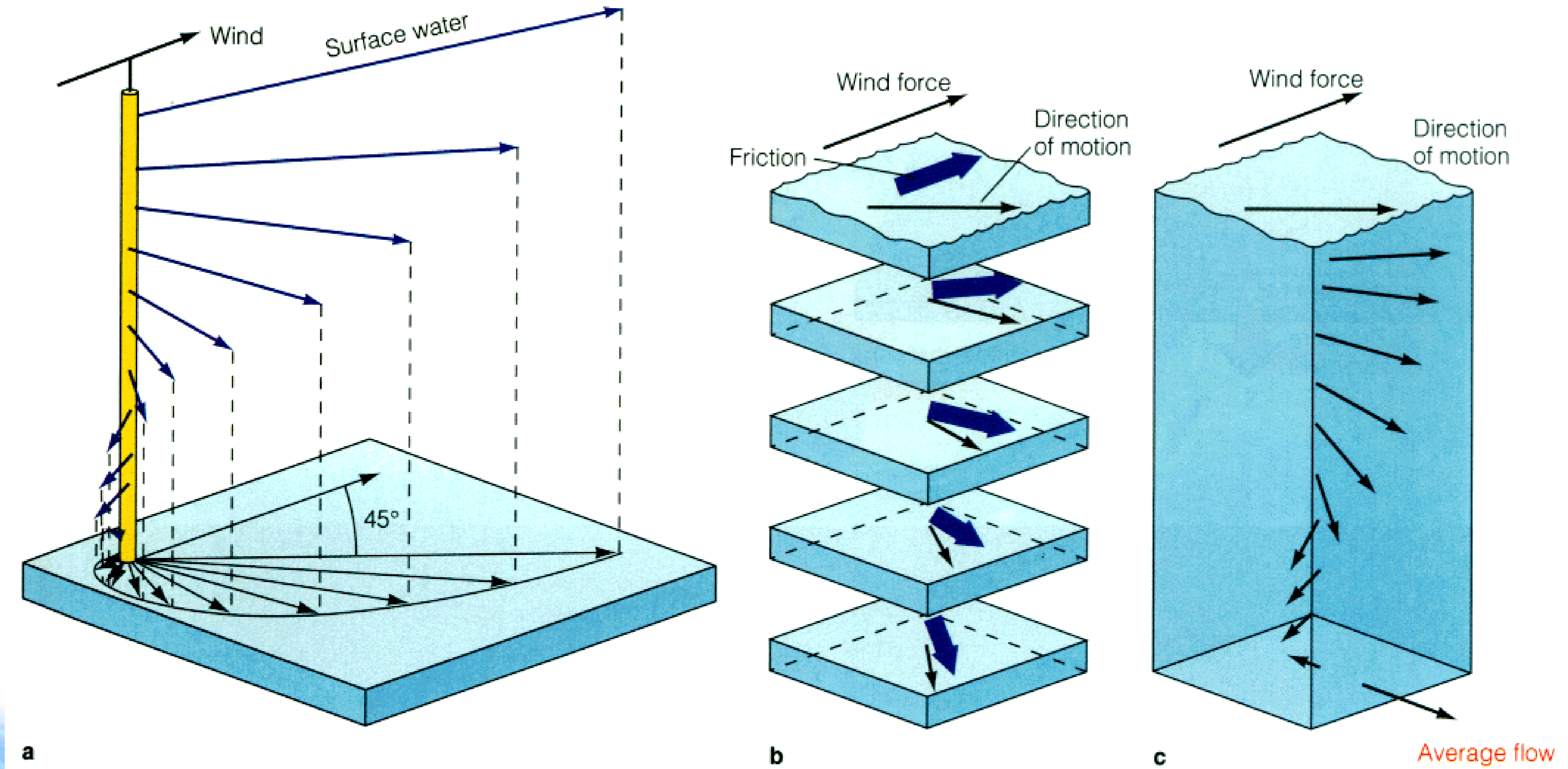


Summary

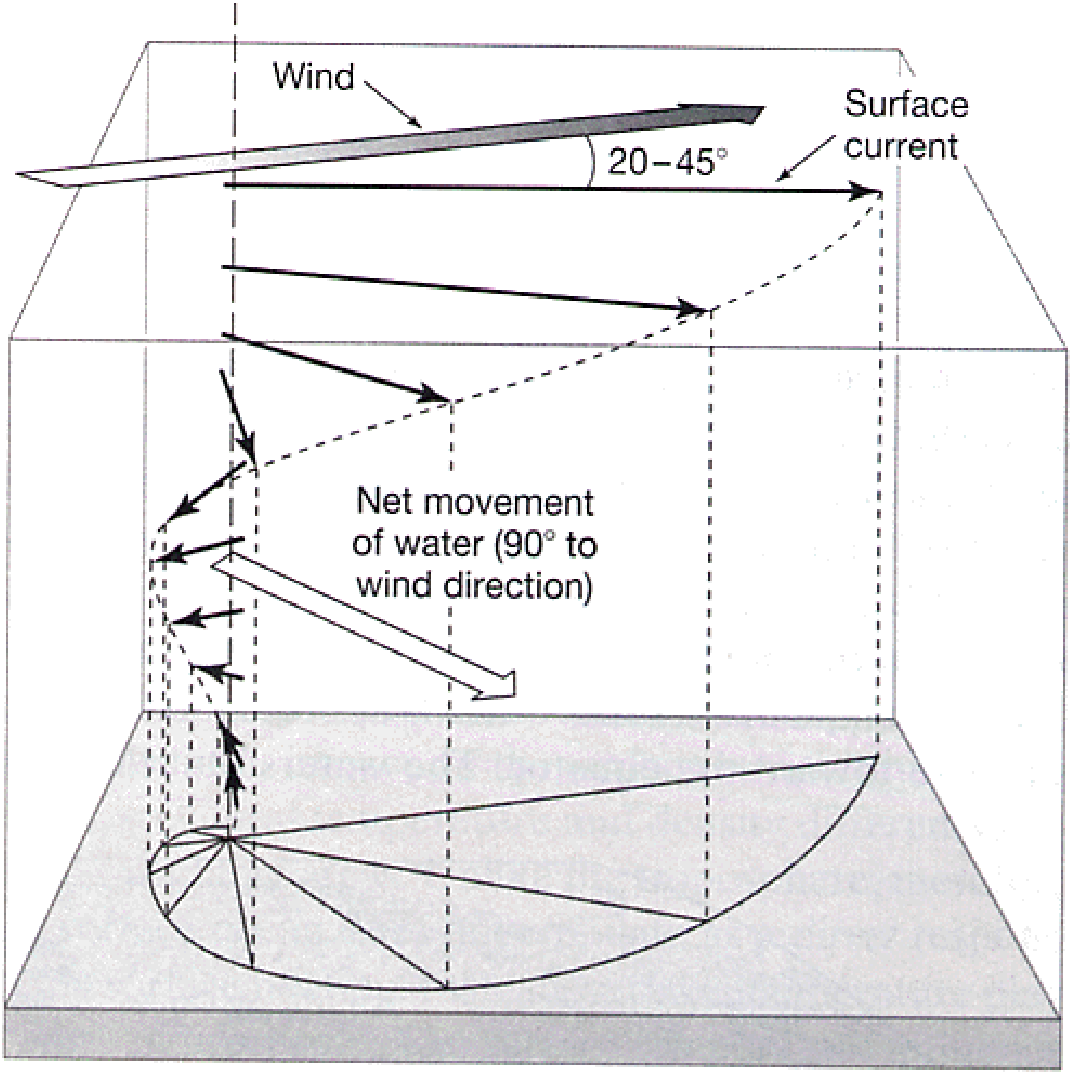
Winds and Surface Currents



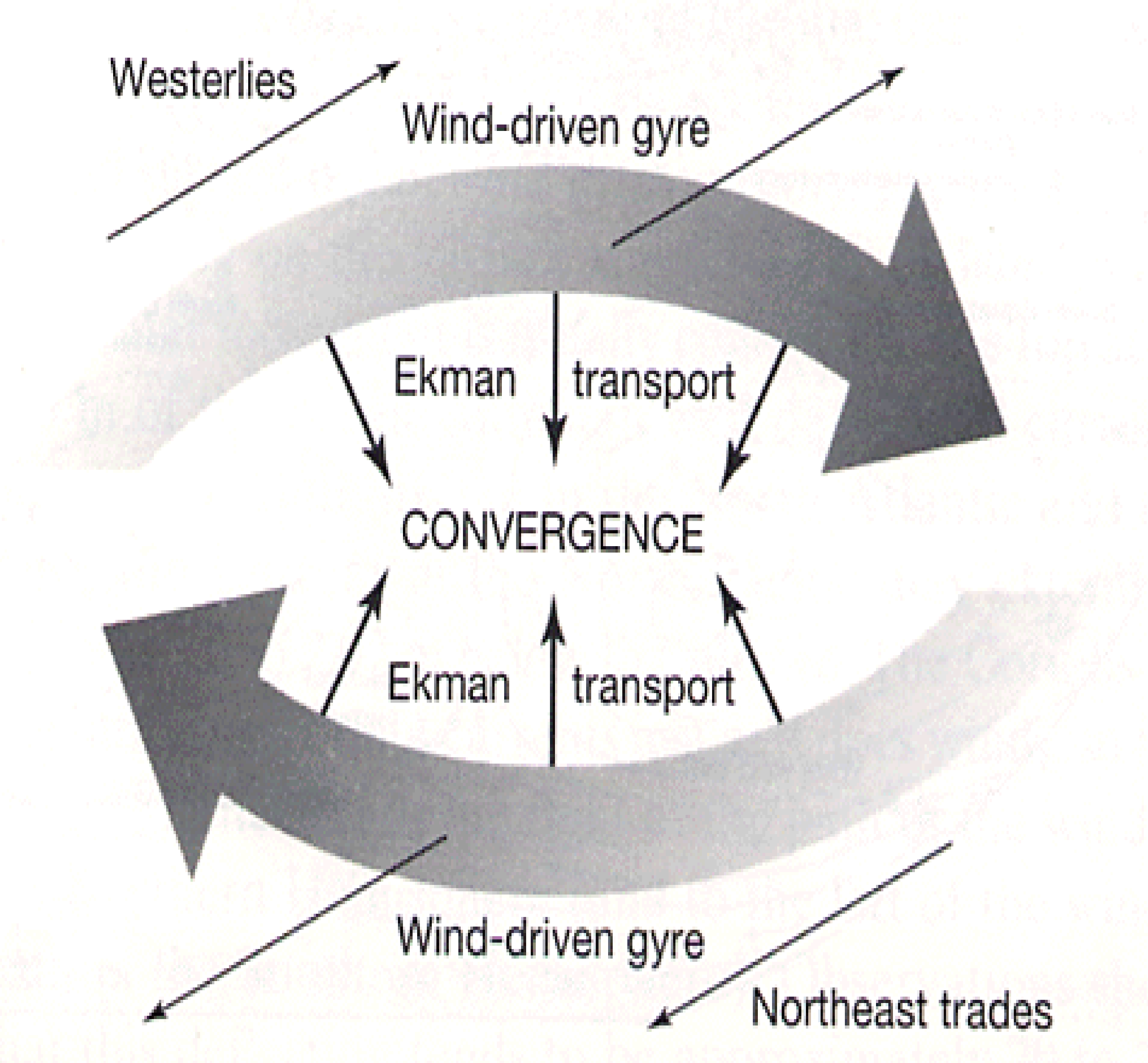
Step 2: Ekman Layer (frictional force + Coriolis Force)



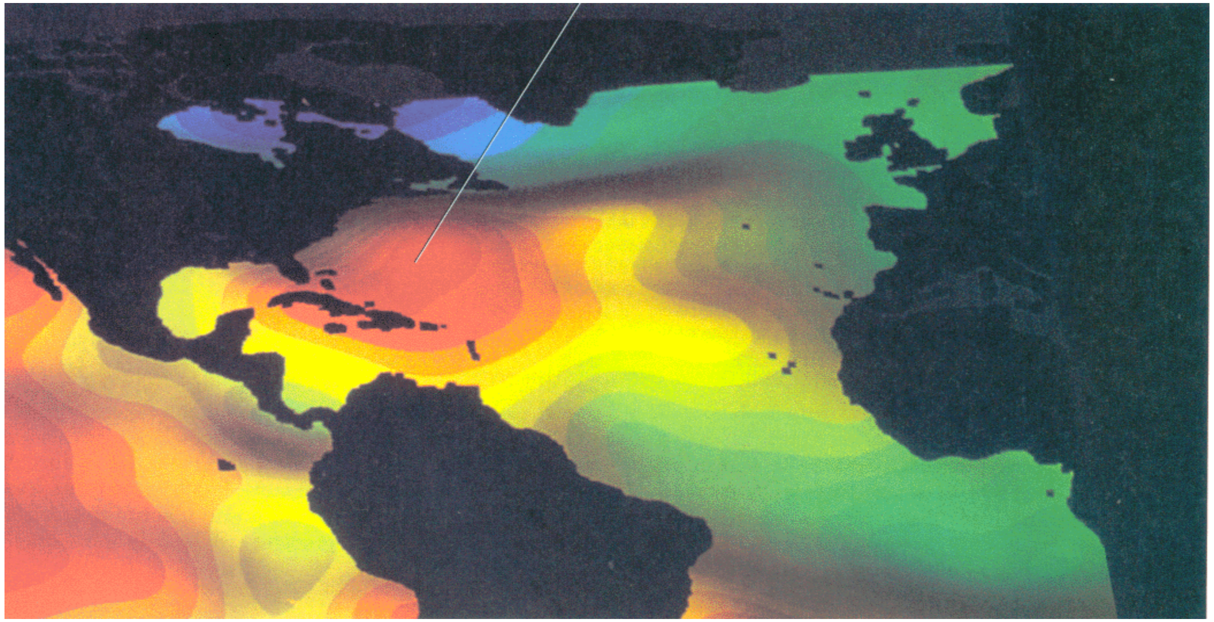
Ekman Spiral – A Result of Coriolis Force



Ekman Transport



Step 3: Geostrophic Current (Pressure Gradient Force + Coriolis Foce)



NASA-TOPEX Observations of Sea-Level Hight