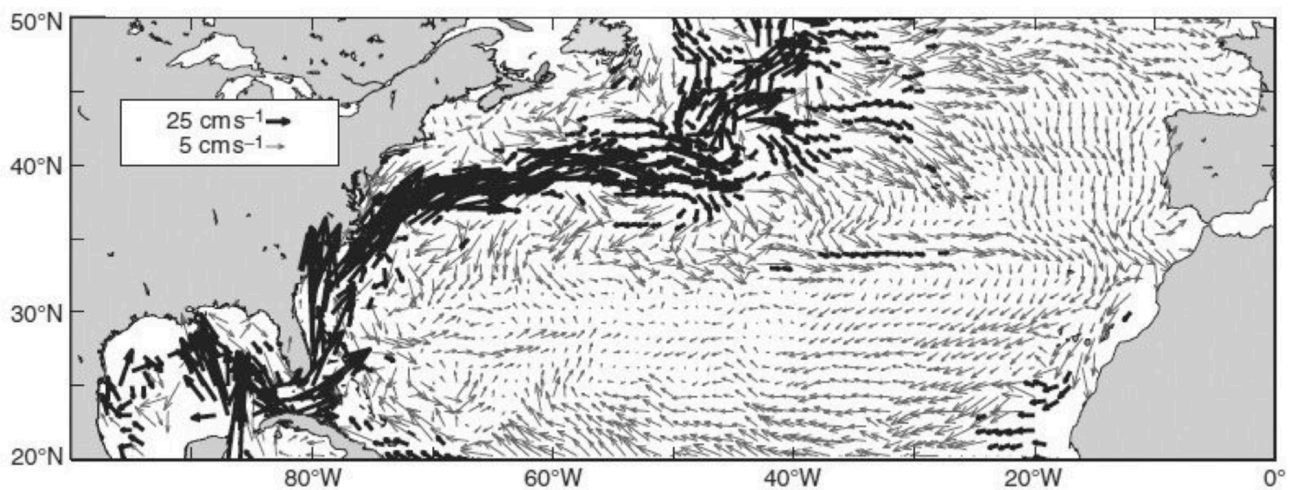
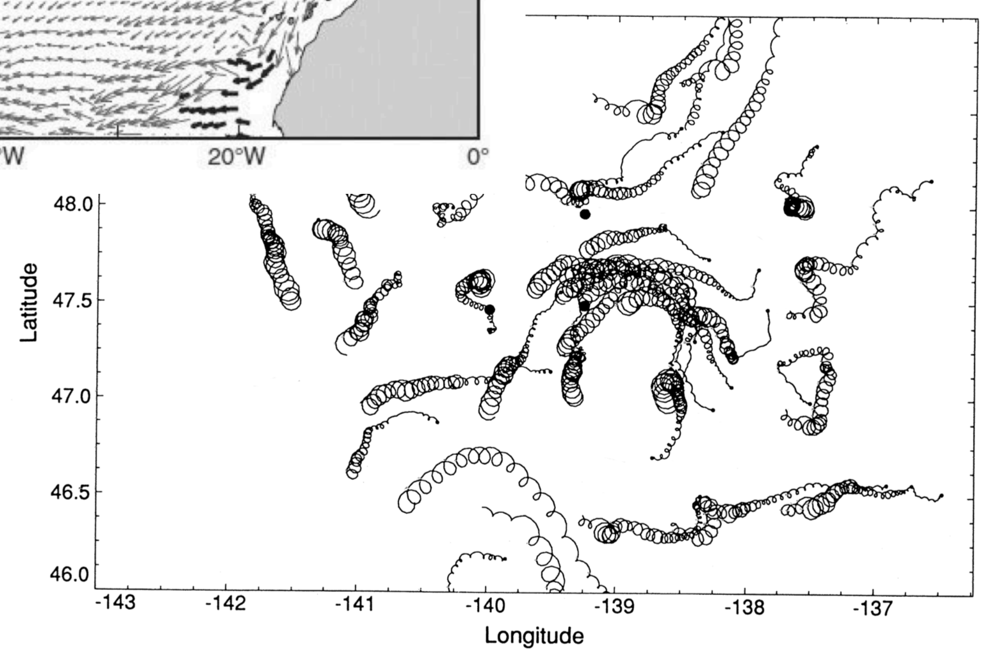
Week 5: Thermal wind, dynamic height and Ekman flow

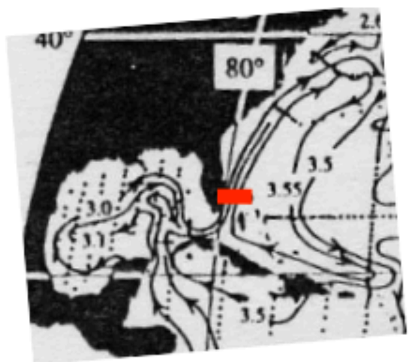


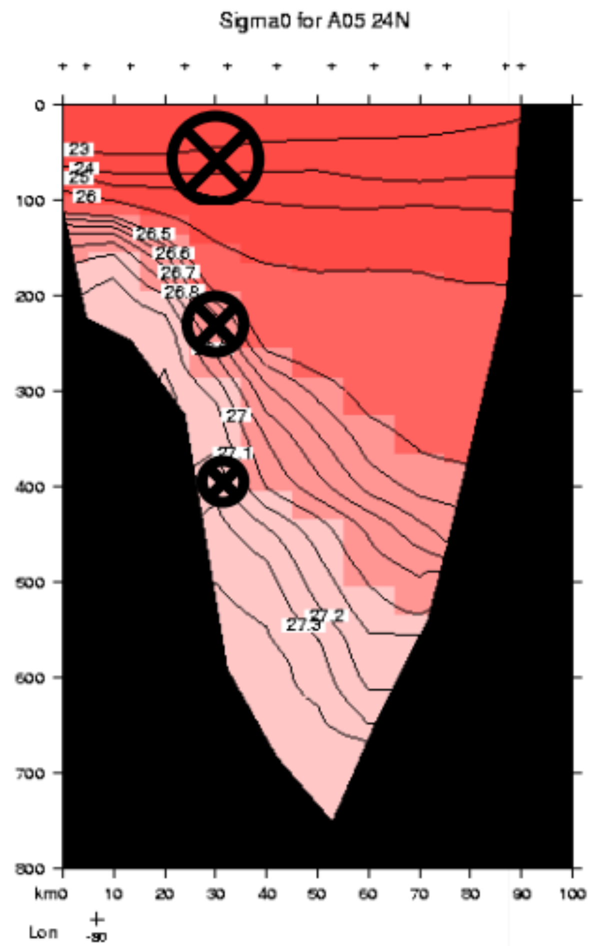


Term Project outline

- **Due date: October 4th**
- Develop an outline based on your questions.
- Based on the research questions, identify a few key papers.
- Develop a list of papers to read
  - Major journals (annual reviews, reviews of geophysics, Nature, Science, PNAS, etc). Citation search.
  - You just need to develop a list. You don’t have to read all of the references to write/submit the outline.
- Download PDFs. Start reading.
  - When reading references, focus on the main points
  - Write your own summary
  - You don’t need to understand every detail to get the big picture
  - Find more papers, building up your reference list (aim for 10+)

Observed density profile across the Florida Strait



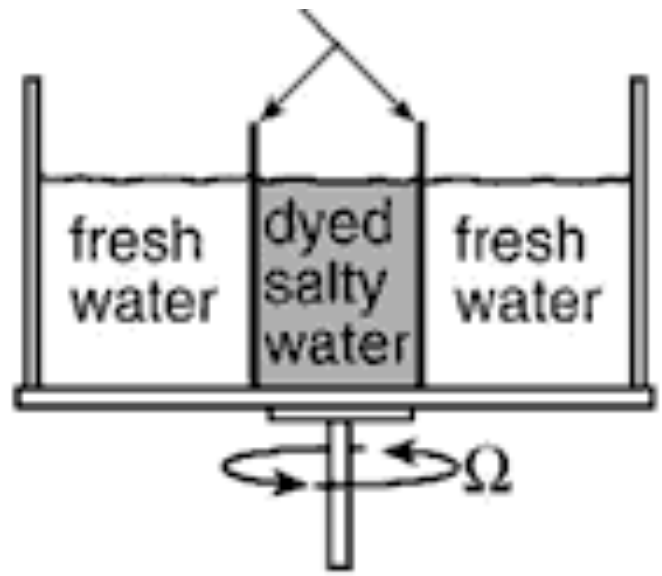


A tank experiment

Initially, metal container separates dyed salty (dense) water from the surrounding freshwater (less dense)

After the tank is in solid body rotation, the container is removed.

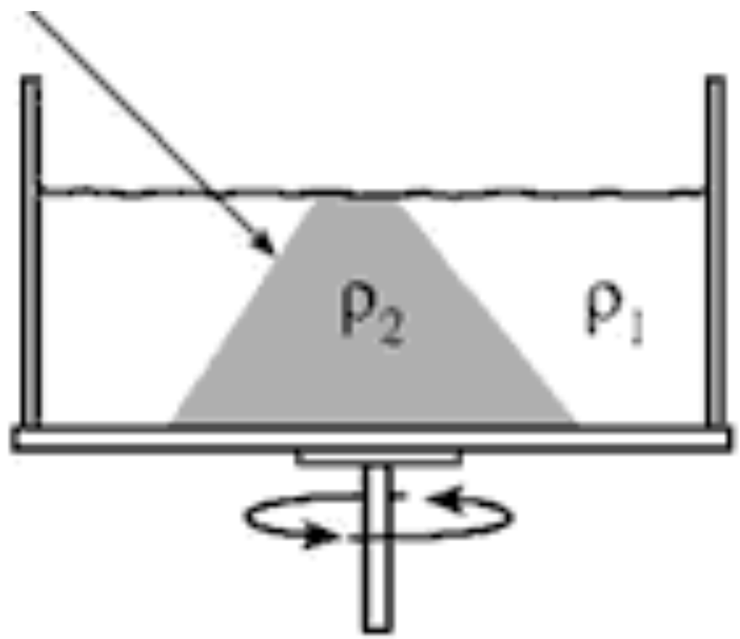
What will happen?



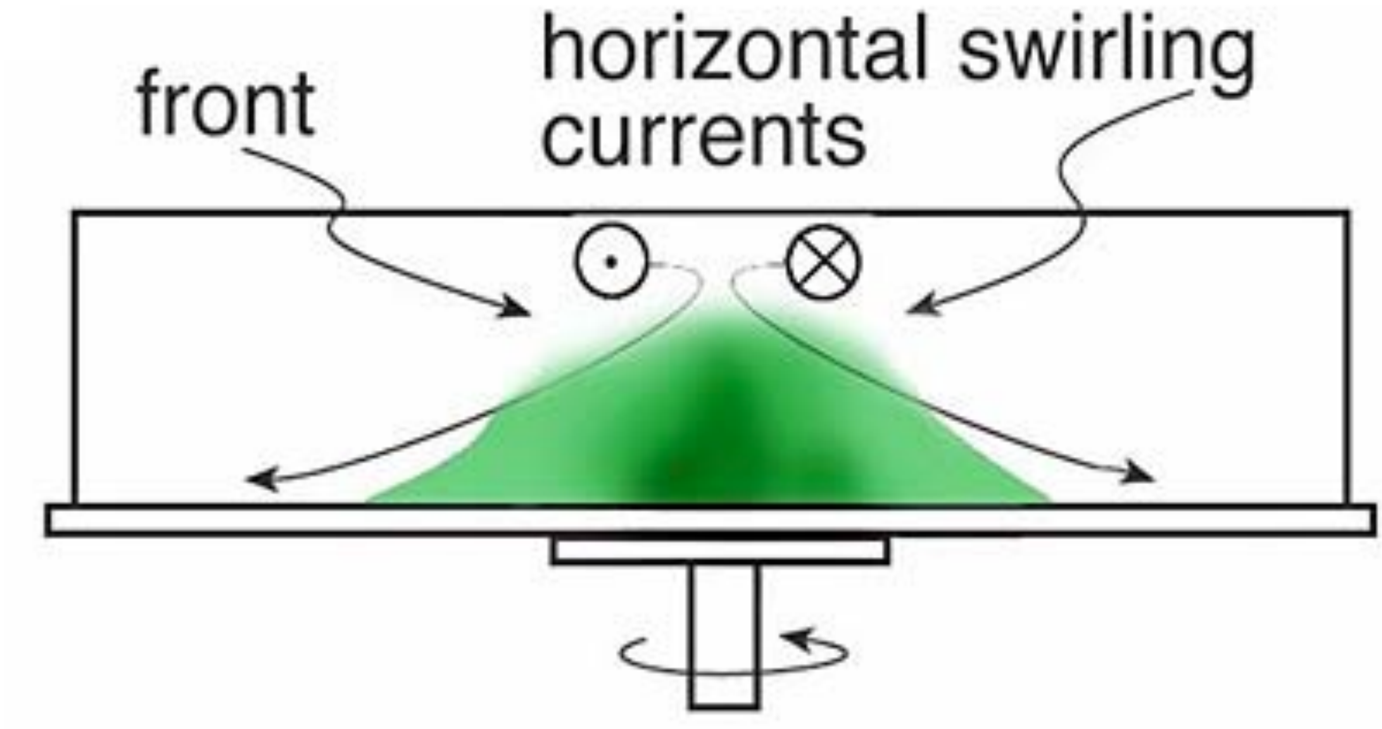
Evolution of circulation and isopycnal tilt

1. Gravity pull down the dense water downward
   Convergence at the top, divergence at the bottom

2. Coriolis effect defects the horizontal motion, the circulation starts to spin around the dense water



Isopycnal tilt and geostrophic circulation



"**Thermal wind balance**" = Coriolis force balancing the buoyancy force acting on the tilted isopycnal surface

Thermal wind balance

- Assume geostrophy in horizontal, and hydrostatic balance in vertical
- Eliminate pressure

$$\frac{\partial u_g}{\partial z} = \frac{g}{\rho f}\frac{\partial \rho}{\partial y},$$

$$\frac{\partial v_g}{\partial z} = -\frac{g}{\rho f}\frac{\partial \rho}{\partial x}.$$

If we measure $\rho(x,y,z)$ and the value of u and v at a given depth, we can calculate the geostrophic velocity

Thermal wind balance and geostrophic currents

The PGF is calculated as the difference of pressure between two stations at a given depth (relative to the geoid).

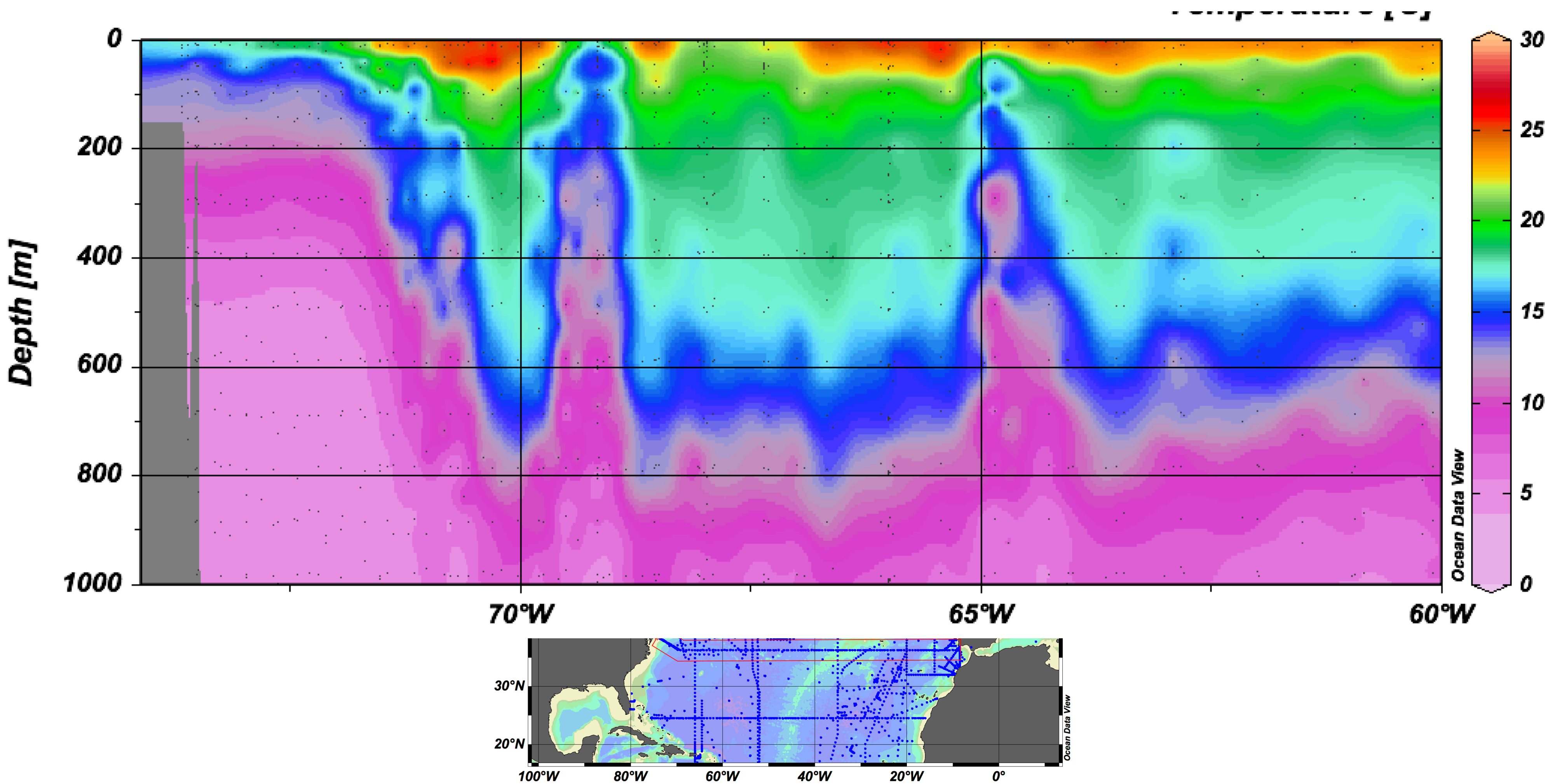
a) If the velocity is known at a given depth, then the PGF at that depth is also known (from geostrophy).

b) From the measured density profiles at the two stations, we can calculate how the velocity change with depth.

c) Using the known velocity from (a), which we call the reference velocity, and knowing how velocity changes with depth from (b), we can compute velocity at every depth.

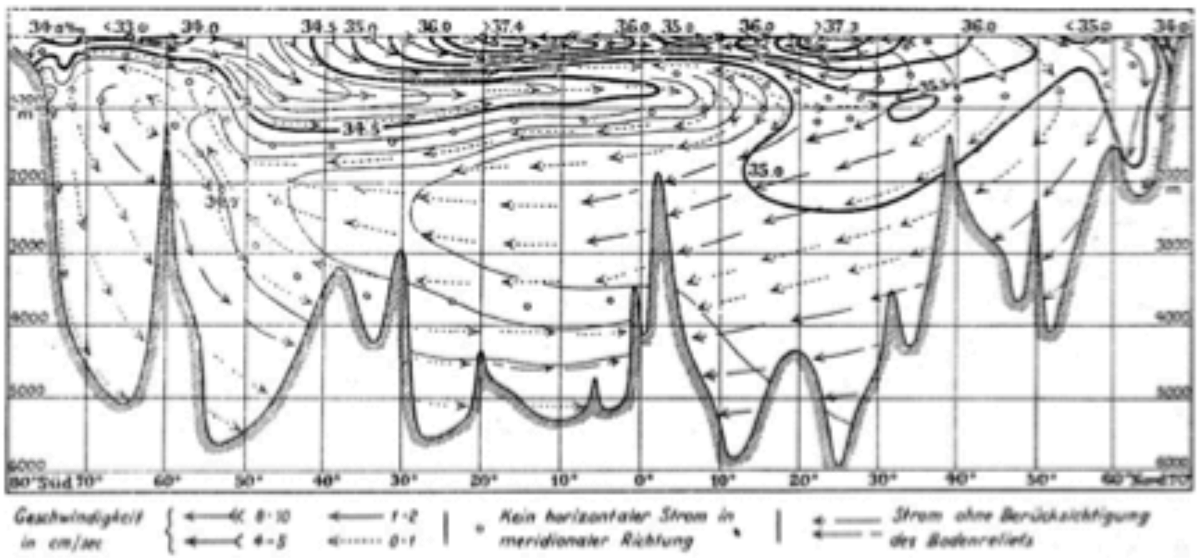
c-alt) ***If a reference velocity is NOT available***, assume deep water is not moving, approximating the reference velocity to be 0 (at arbitrarily set depth in the abyss = level of no motion).

Observed isopycnal tilt in the Gulf Stream



Meteor expedition (1925-27)

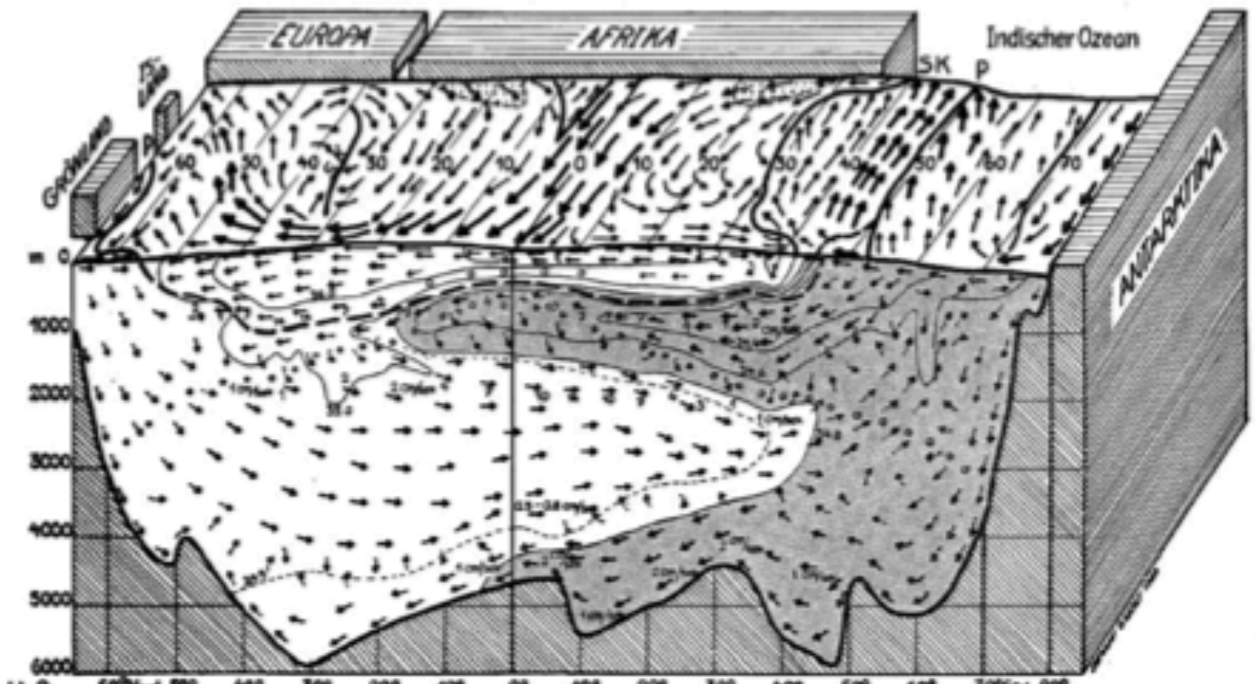
- Testing the thermal wind balance
- Alfred Merz



Merz, 1925

Three dimensional Atlantic circulation after Meteor expedition

G. Wust, 1935; 1949



bb. 2. Schematisches Blockdiagramm d. Oberflächenströmungen d. Atlant. Ozeans u. d. Tiefenzirkulation auf seiner Westseite.

Geostrophic calculation for the subsurface currents

- At the surface, horizontal SSH gradient tells us the geostrophic current (e.g. HW4).
- Thermal wind balance tells us how geostrophic current vary in depth
  → Vertical integration of the horizontal density gradient
- **Dynamic height / Geopotential anomaly**

Geostrophy in pressure coordinate

- Pressure decreases with height (hydrostatic balance).
  - We care about the horizontal variation of pressure that controls geostrophic circulation
  - If one follows a surface of constant pressure, high pressure region has a slightly shallower depth (greater z).

$$\left(\frac{\partial P}{\partial x}\right)_{y,z} = \rho\, g \left(\frac{\partial z}{\partial x}\right)_{y,P}$$

$$\left(\frac{\partial P}{\partial y}\right)_{x,z} = \rho\, g \left(\frac{\partial z}{\partial y}\right)_{x,P}$$

Geostrophy in pressure coordinate

- Using pressure as a vertical coordinate, the height of the pressure surface can be used to calculate PGF.

$$-fv_g = -g\frac{\partial z}{\partial x}$$

$$fu_g = -g\frac{\partial z}{\partial y}$$

Dynamic height

- Given the measurements of T and S, we can calculate the "dynamic height", i.e. steric height at a given pressure level.
  - α is the specific volume (inverse of density)

$$\frac{\partial P}{\partial z} = -\rho\, g$$

$$\frac{\partial Z}{\partial P} = -\frac{1}{\rho\, g} = -\frac{\alpha}{g}$$

$$\alpha' = \frac{1}{\rho(T,S,P)} - \frac{1}{\rho(0,35,P)}$$

→

$$\frac{\partial Z'}{\partial P} = -\frac{\alpha'}{g}$$

Dynamic height

- Given T and S, calculate density (and specific volume anomaly, α') at every pressure level.
- Then vertically integrate ($-\alpha'/g$) to get Z'

$$Z'(x,y,P) = -\int_{P_{ref}}^{P} \left(\frac{\alpha'}{g}\right) dP$$

- IF P_{ref} is chosen to be 2,000dbar, Z' is then called the "dynamic height with respect to the 2,000dbar reference level".
- Geopotential anomaly is defined as Φ=gZ'

Thermal wind calculation and dynamic height

To get the change in PGF between two pressure levels 1 to 2, calculate the geopotential anomaly (or dynamic height) change between the levels 1 and 2.

$\Delta Z = Z_2 - Z_1 =$ dynamic height difference

$f\,(v_2 - v_1) = g\partial\Delta Z/\partial x$
$f\,(u_2 - u_1) = -g\partial\Delta Z/\partial y$

If one set of the velocities are known, then the thermal wind balance will tell us the other set of the velocities are also known.

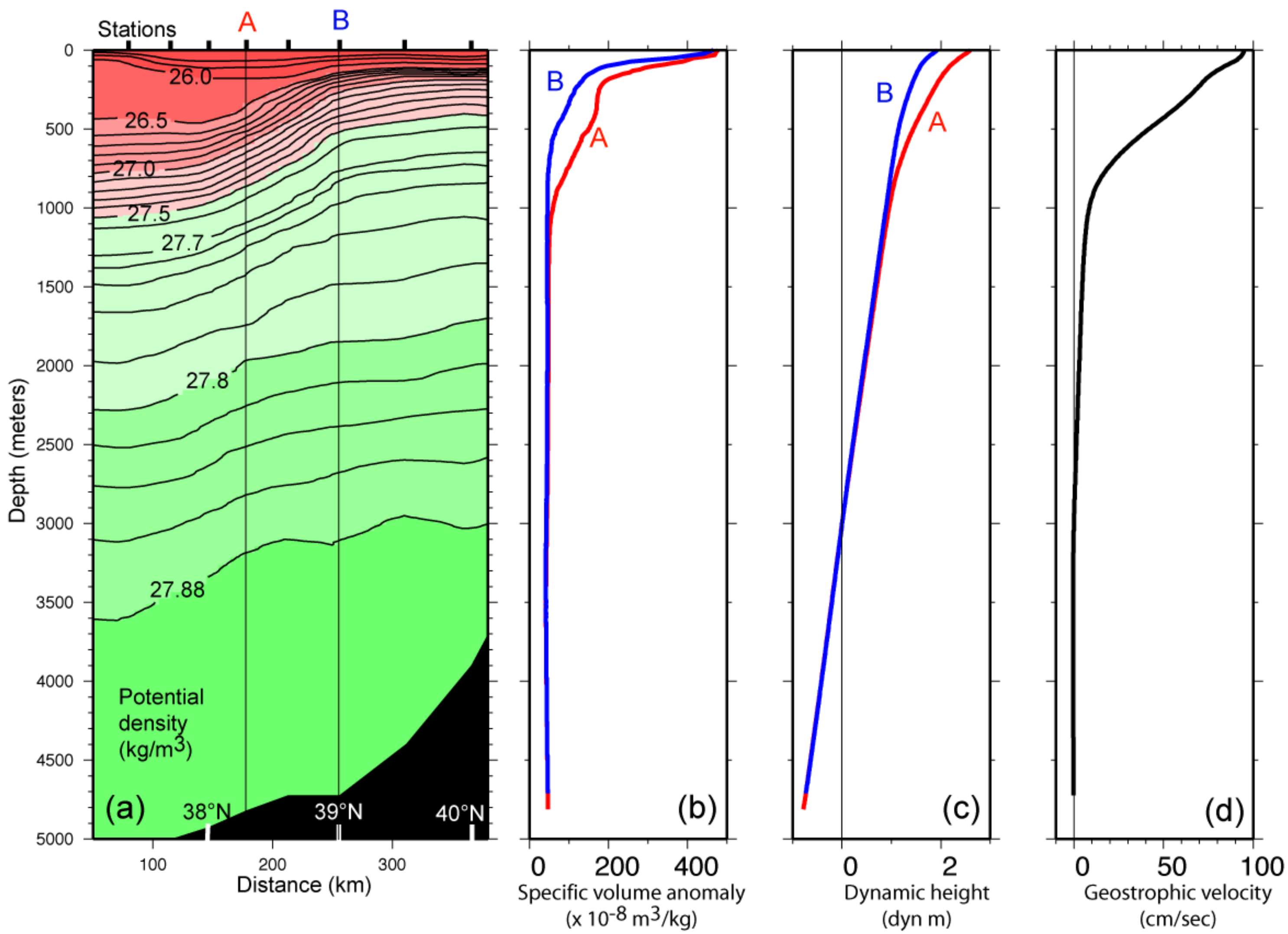
Reference level

Reference velocities:

Level of no motion: Old-fashioned - assume 0 velocity at some great depth, and compute velocities at all shallower depths. This is useful if you are focused on very energetic upper ocean flows, but not useful if you want to look at small deep flows.

Level of known motion: Much better - observe or determine through external means (use of tracers, mass balance, etc) a good guess at the velocity at some depth. This is essential if you want to study deep circulation.

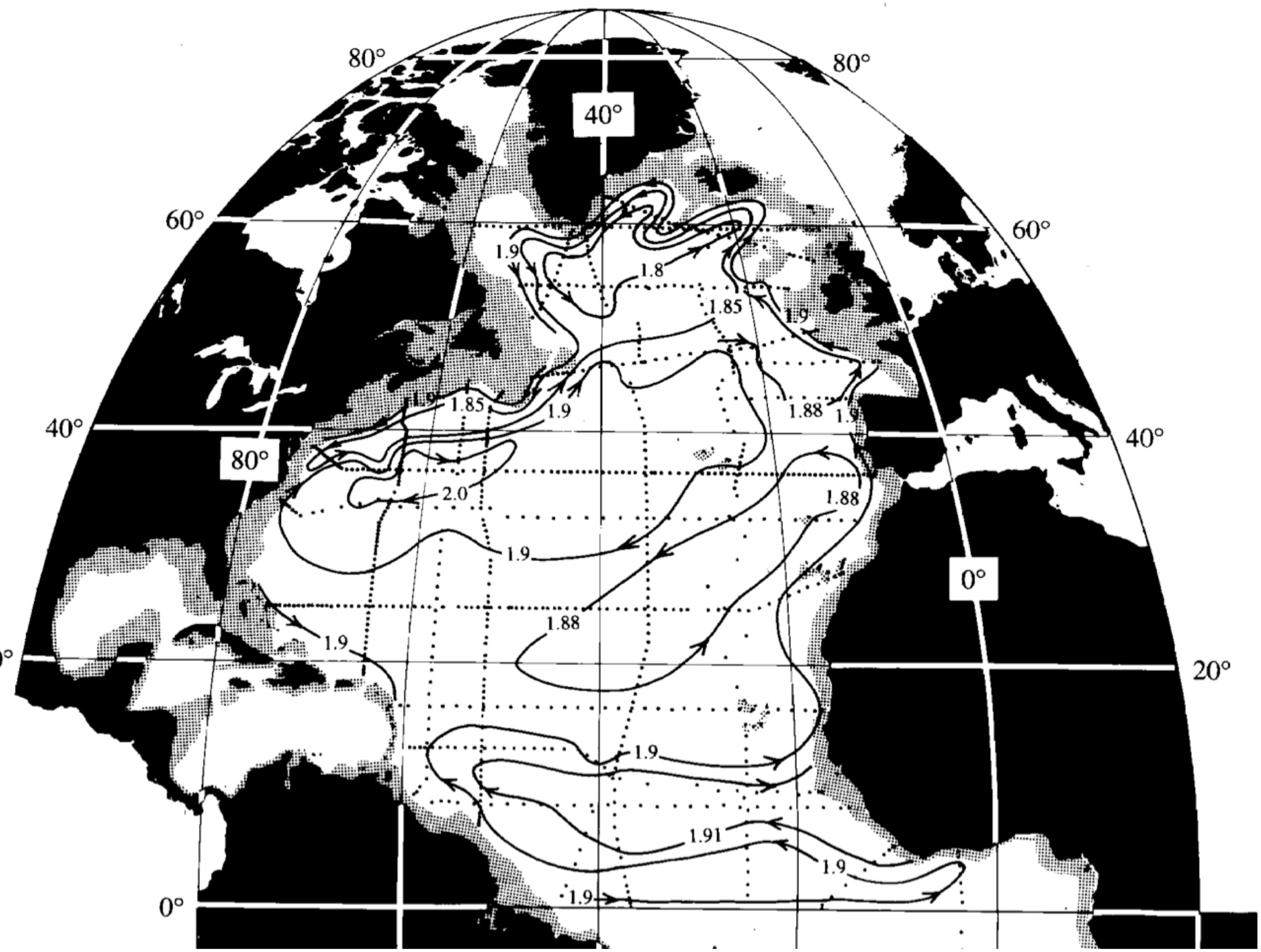
Gulf Stream density, dynamic height and geostrophic velocity



In this example, compute geostrophic velocities relative to 0 cm/sec at 3000 dbar. If we KNOW v at 3000 dbar, then just add it to the whole profile

DPO Fig. 7.11

Steric height at 1000 dbar in the N. Atlantic. Values are similar to sea surface height in meters. (Reid, 1994)



DPO Fig. S9.2b

Global SSH, Niiler et al., 2003

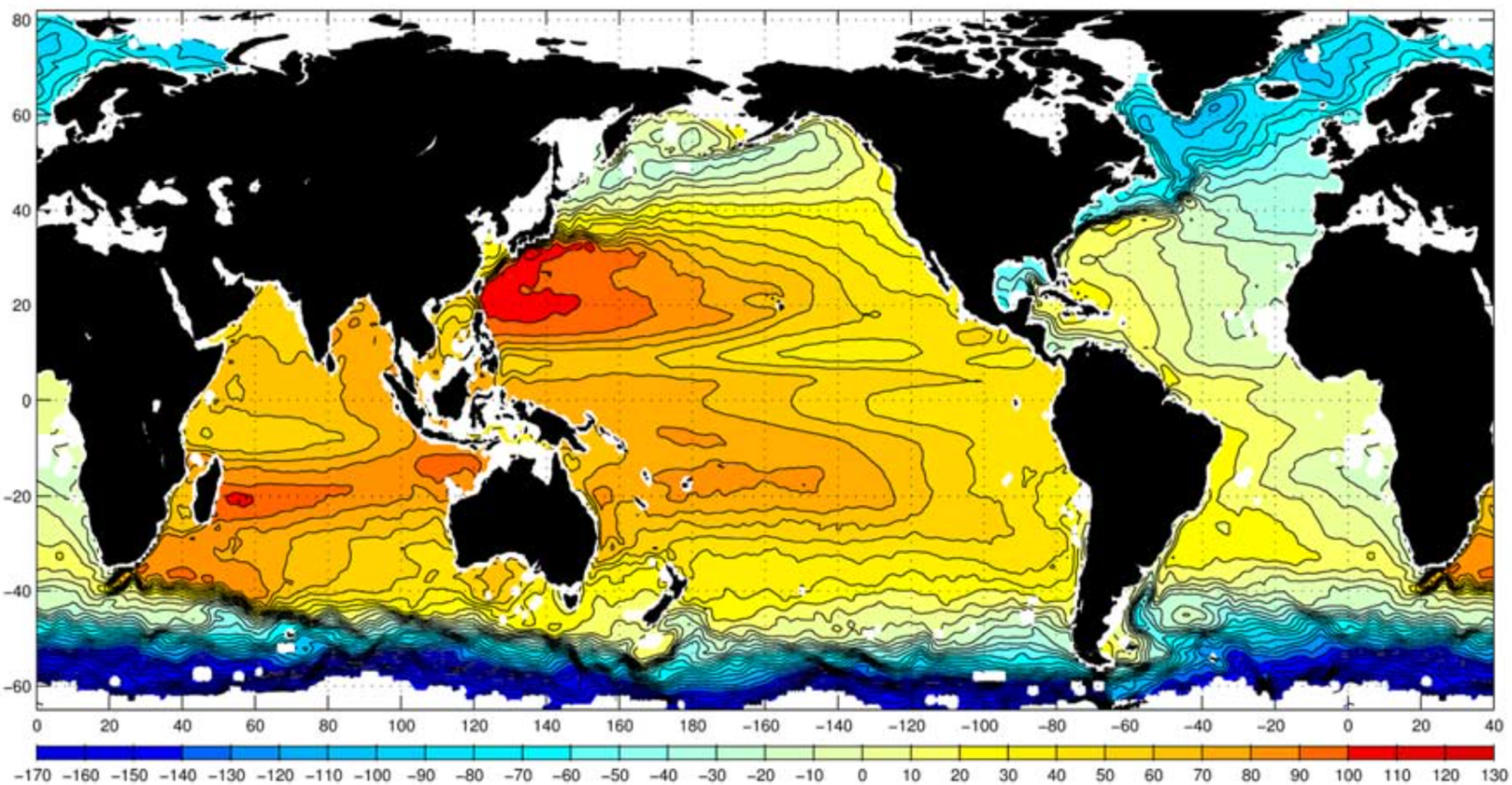
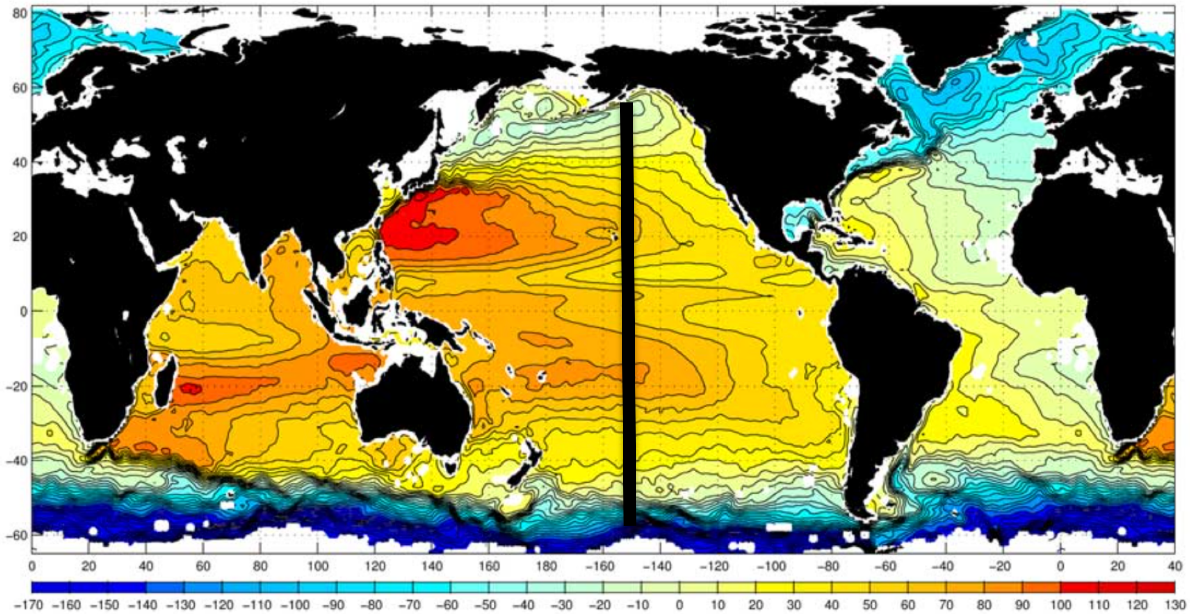


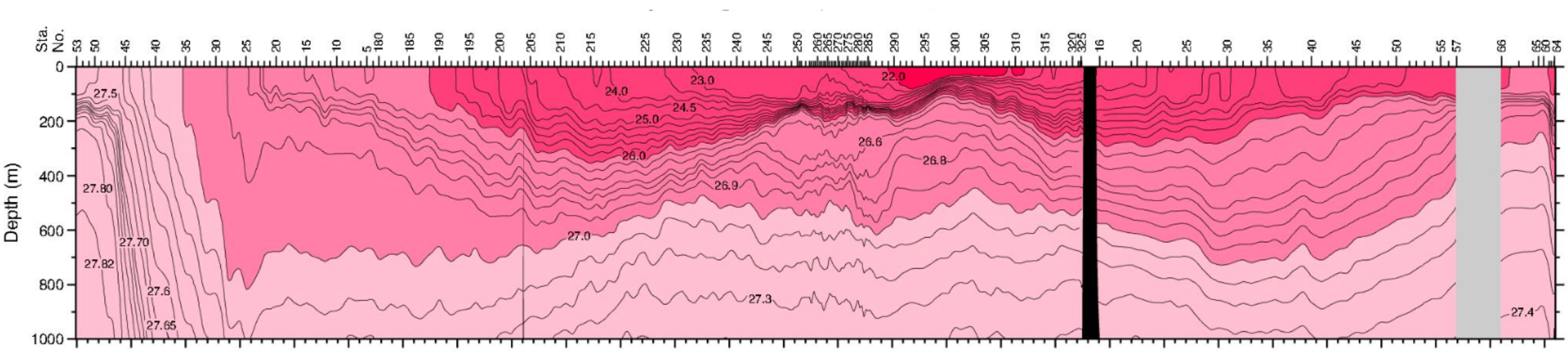
Figure 1. 1992–2002 mean absolute sea level η_0 obtained as described in this paper. Contour interval is 10 cm. Sea level is computed from the values of $\nabla\langle\eta\rangle$ estimated according to (2) on 1° spatial grid with the condition of zero global mean.

WOCE line P16

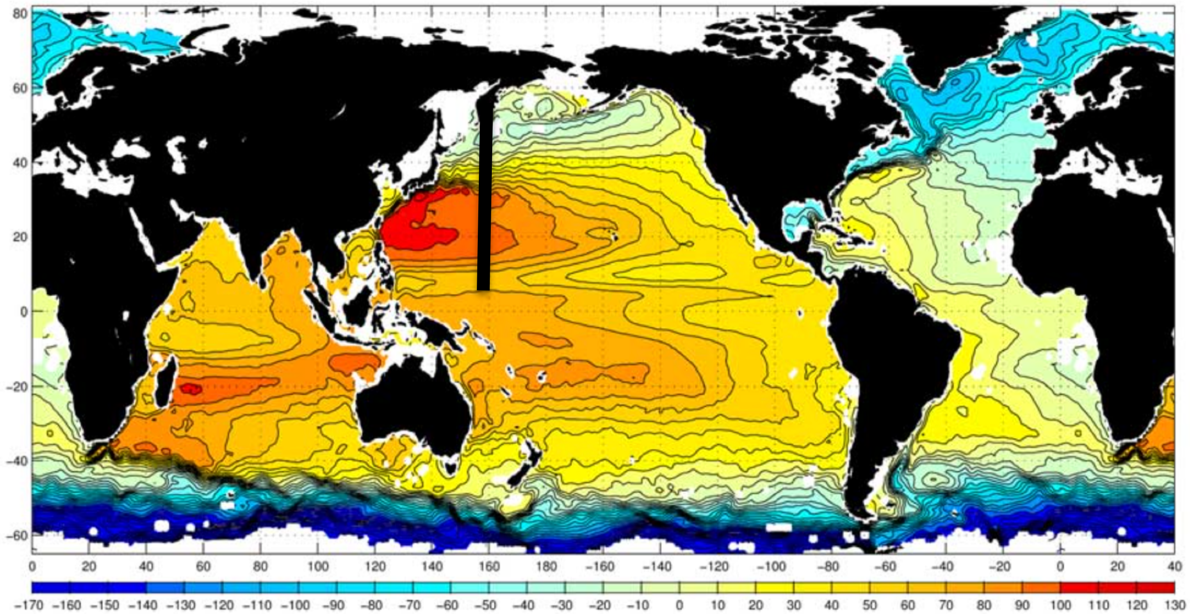


Relatively high SSH of the subtropics is reflected in the deep thermocline (i.e. high steric height).

Figure 1. 1992–2002 mean absolute sea level η_0 obtained as described in this paper. Contour interval is 10 cm. Sea level is computed from the values of $\nabla\langle\eta\rangle$ estimated according to (2) on 1° spatial grid with the condition of zero global mean.

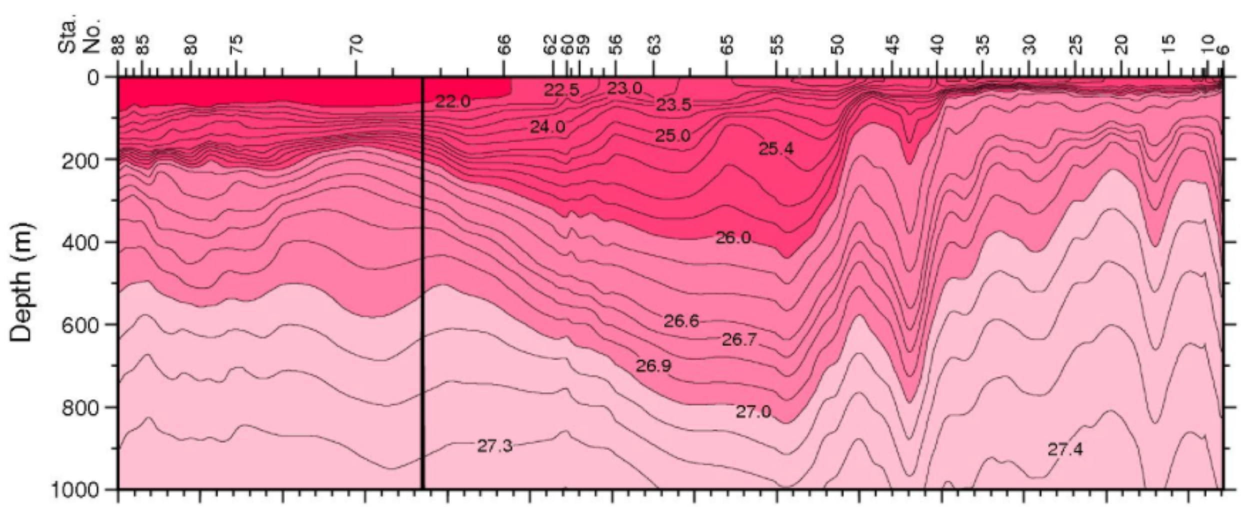


WOCE line P13



The high SSH of the western subtropics is reflected in the deep thermocline (i.e. high steric height).

Figure 1. 1992–2002 mean absolute sea level η_0 obtained as described in this paper. Contour interval is 10 cm. Sea level is computed from the values of $\nabla\langle\eta\rangle$ estimated according to (2) on 1° spatial grid with the condition of zero global mean.



Wind driven flow: Ekman circulation

Ekman flow: Coriolis and Frictional force by wind

X (east-west)

$$\frac{Du}{Dt} - fv = -\frac{1}{\rho}\frac{\partial P}{\partial x} + \frac{F_x}{\rho}$$

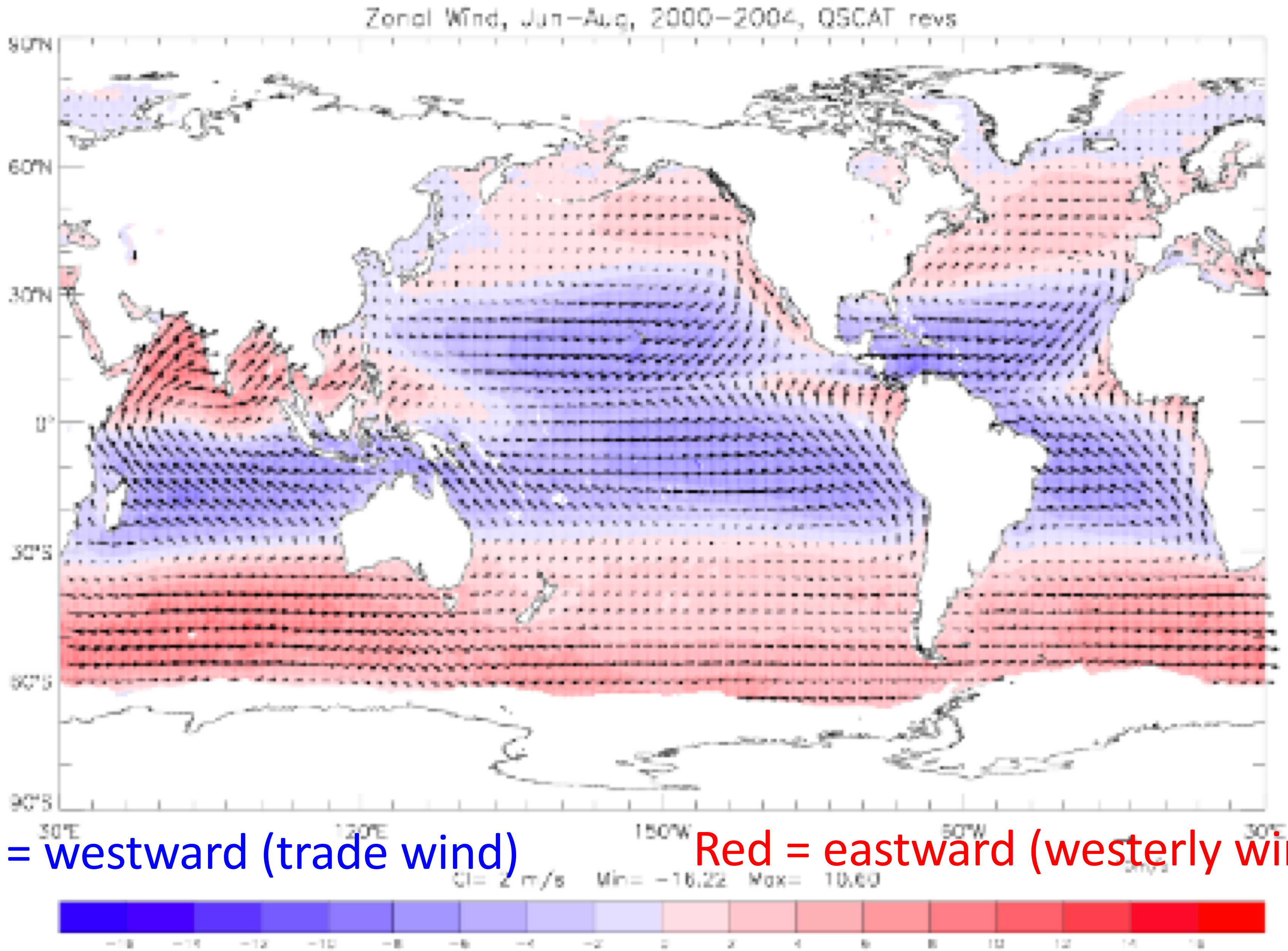
Y (north-south)

$$\frac{Dv}{Dt} + fu = -\frac{1}{\rho}\frac{\partial P}{\partial y} + \frac{F_y}{\rho}$$

Z (vertical)

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

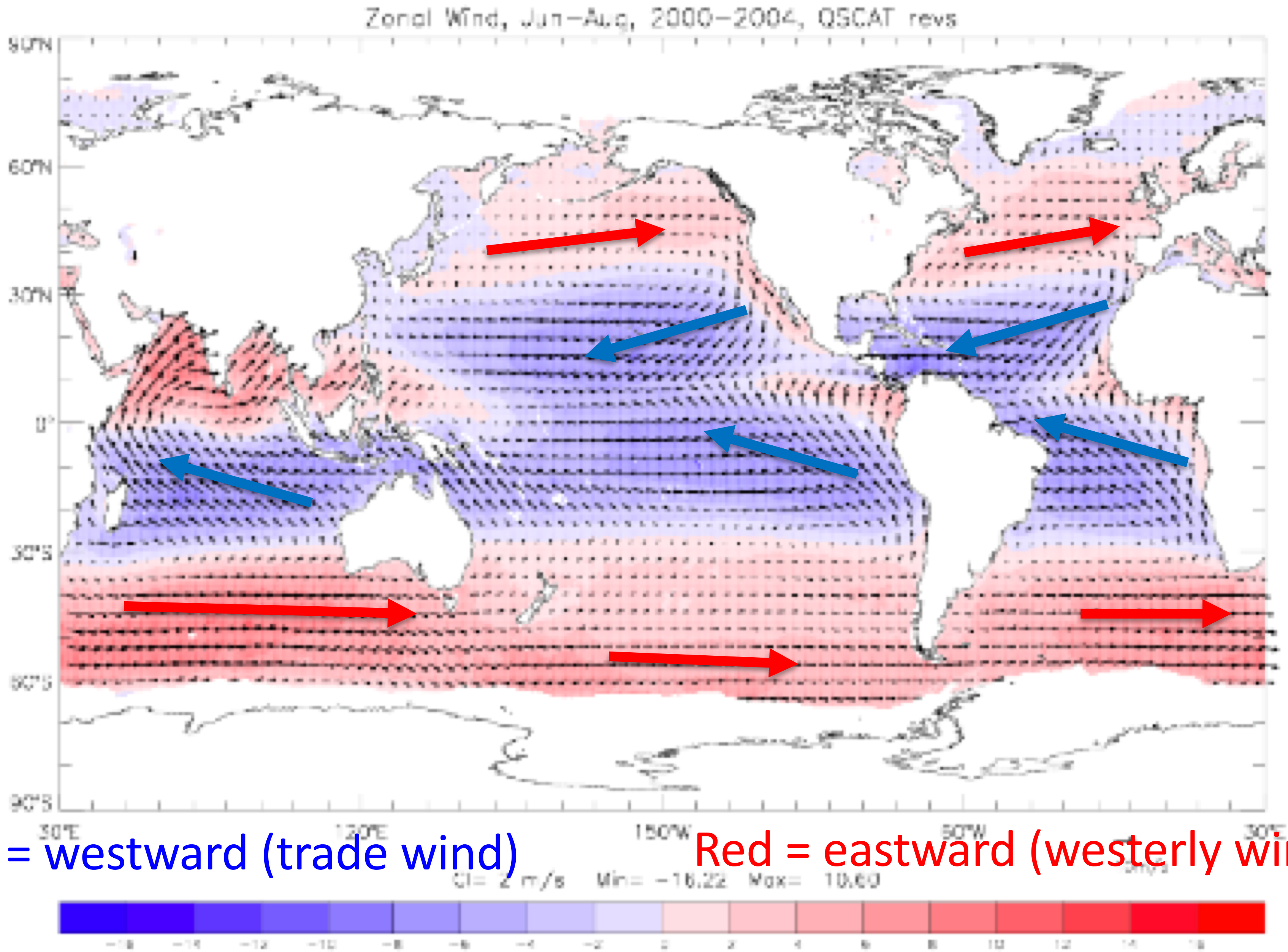
Observed wind stress



Blue = westward (trade wind)

Red = eastward (westerly wind)

Observed wind stress



Blue = westward (trade wind)

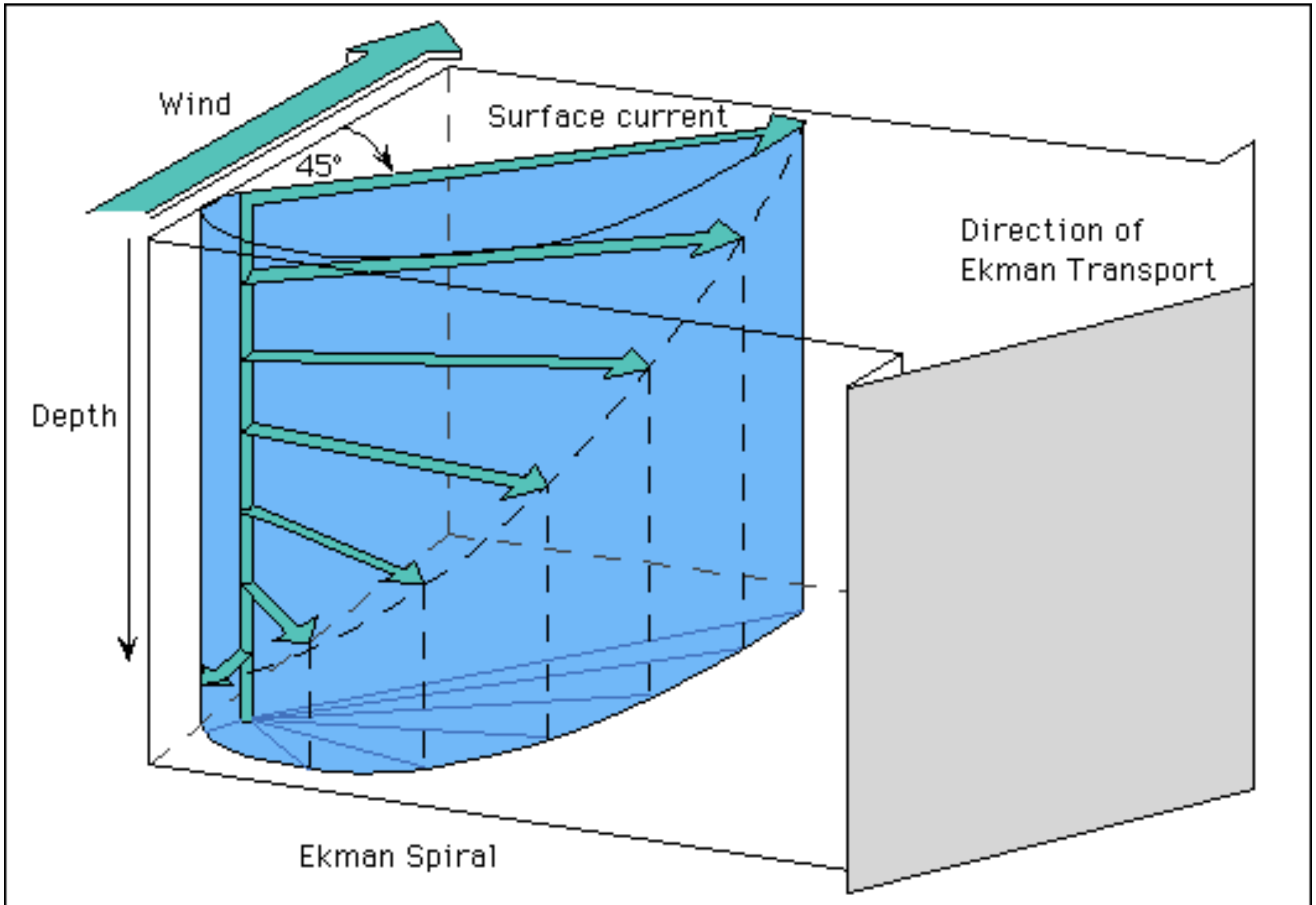
Red = eastward (westerly wind)

Ekman spiral

Vertical scale of the Ekman layer ~ 20-60m

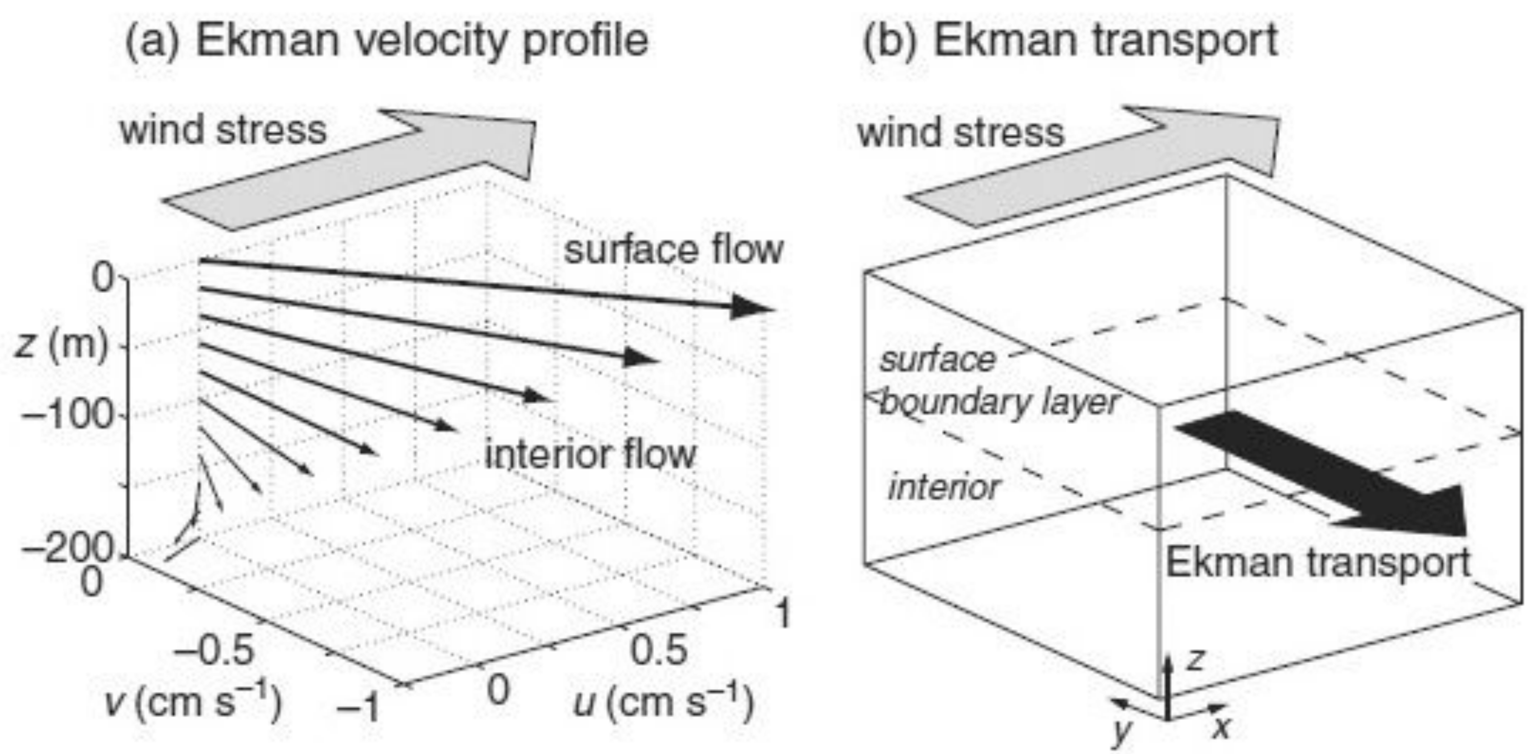
Top layer moves at 45 degrees to the right of the wind direction in NH.

This layer pushes the layer underneath. Layer by layer, it spirals downward clockwise.



Adapted from Thurman, Harold V. **Essentials of Oceanography, 5th ed.** Prentice-Hall, Inc., 1996.

Vertically integrated Ekman transport



Mathematical expression for Ekman layer

$$u_{ek} = \frac{1}{\rho f}\frac{\partial \tau_y}{\partial z},$$

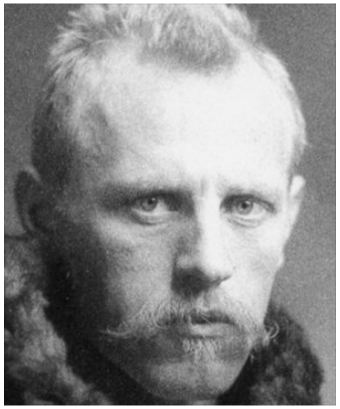
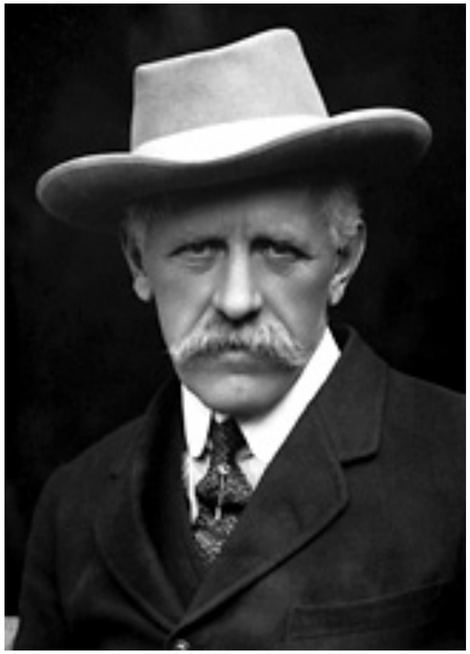
$$v_{ek} = -\frac{1}{\rho f}\frac{\partial \tau_x}{\partial z},$$

$$U_{ek} \equiv \int_{-D}^{0} u_{ek} dz = \frac{\tau_y^s}{\rho f},$$

$$V_{ek} \equiv \int_{-D}^{0} v_{ek} dz = -\frac{\tau_x^s}{\rho f},$$

Vertically integrated Ekman flow is 90 degree to the right of the direction of the wind

Wind-driven ocean current

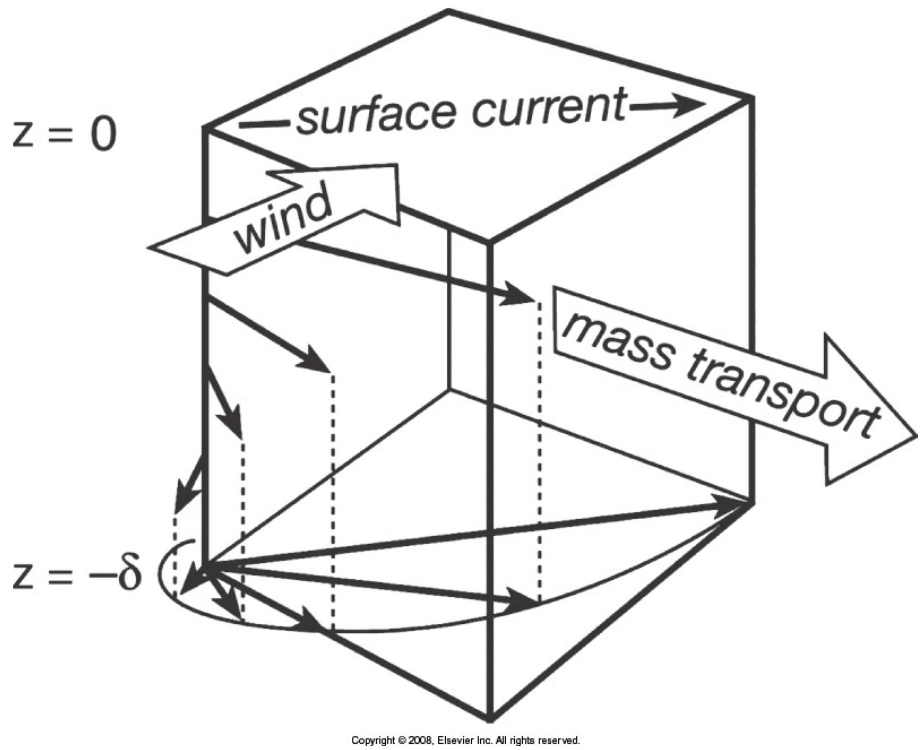



Fridtjof Nansen (1861-1930)
From Fram Museum and Nobel Foundation

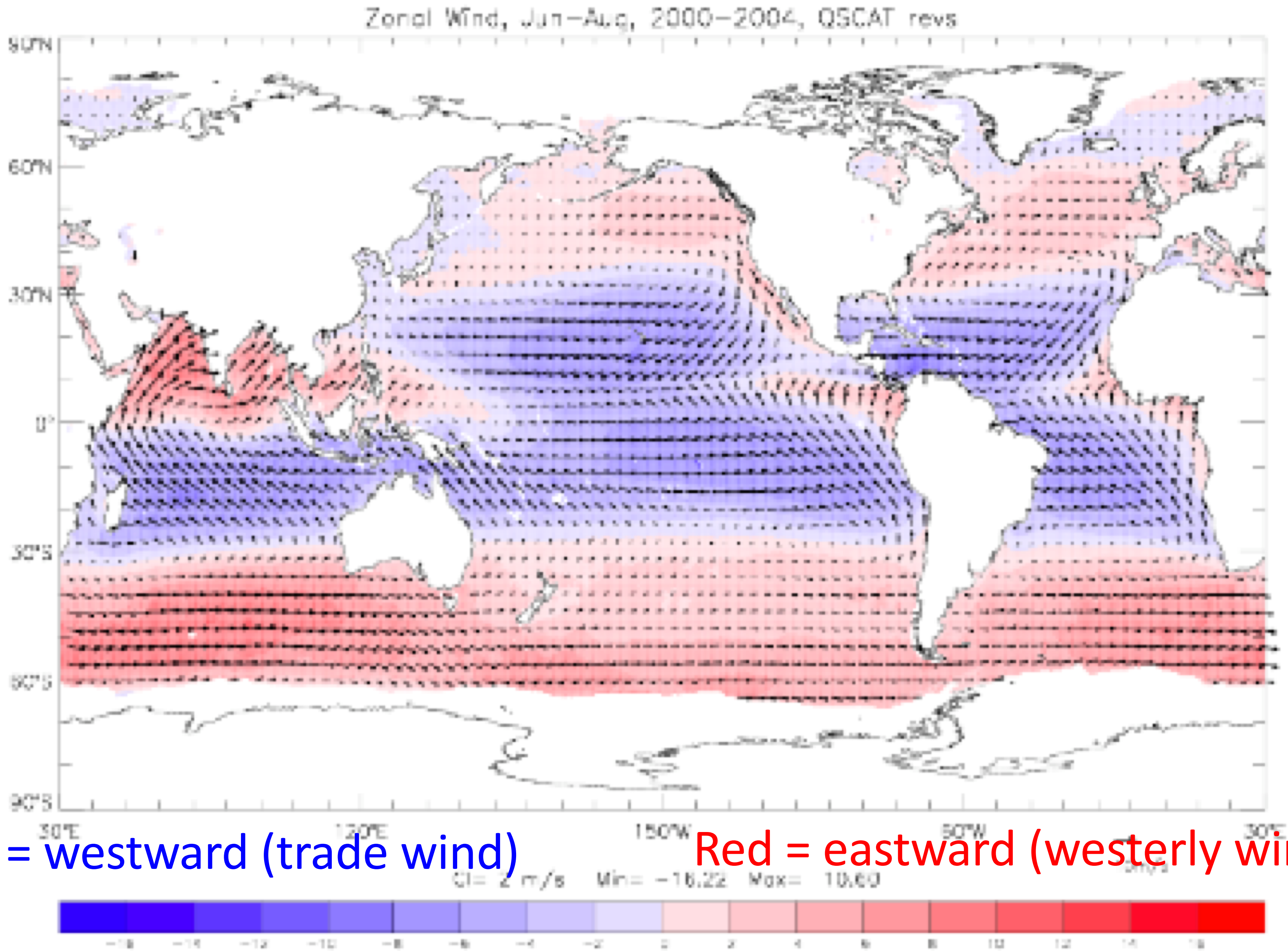
The Fram Expedition (Frist, 1893-1896)

Ekman current

Vagn Walfrid Ekman
(1874-1954)
Univeristy of Lund

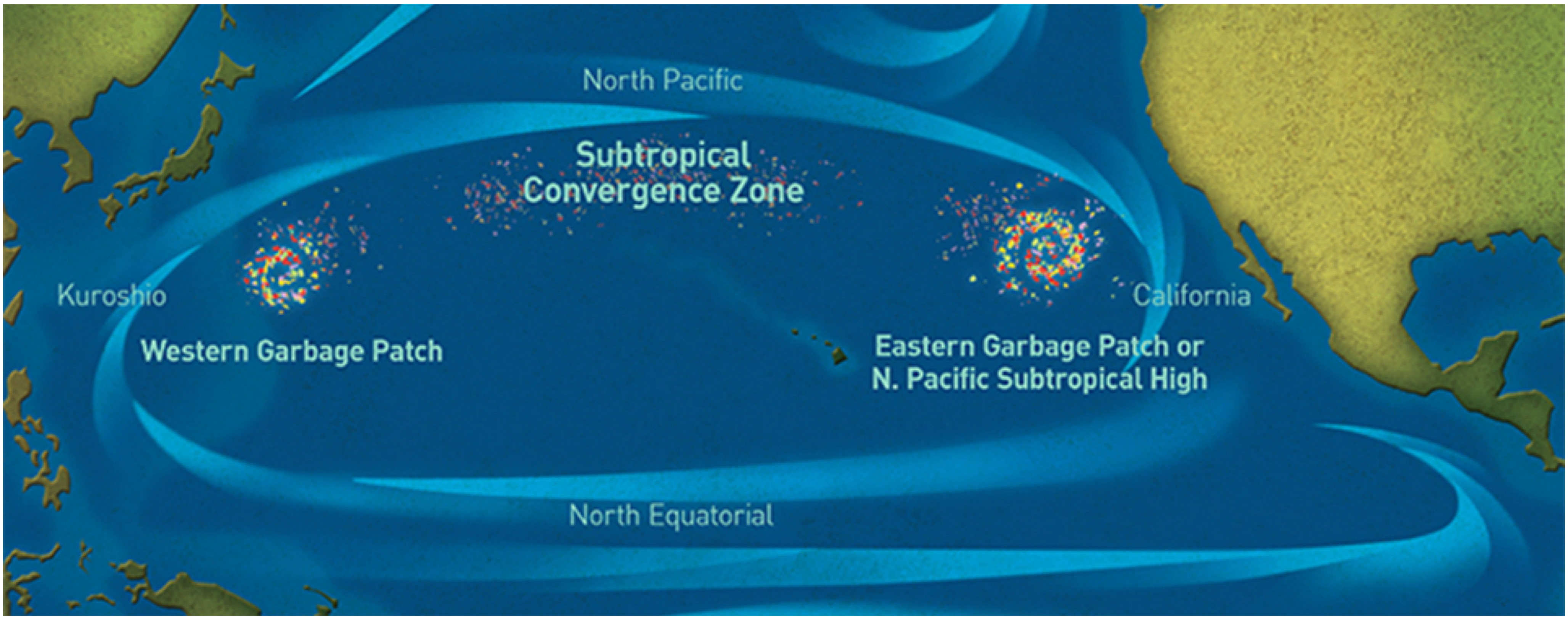


Observed wind stress



Marine debris

"The great garbage patch"



http://marinedebris.noaa.gov/

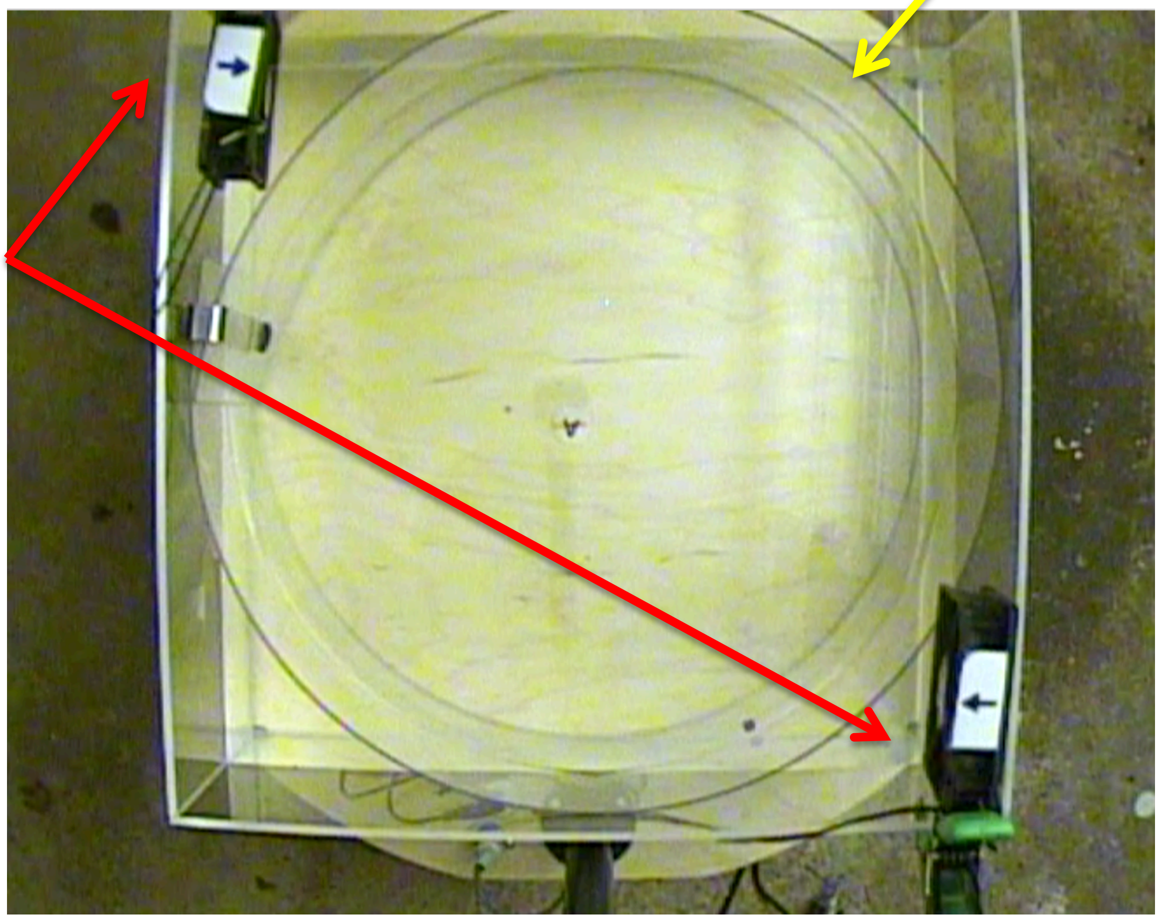
Tank experiment

NORTH

Round wall

Computer fan

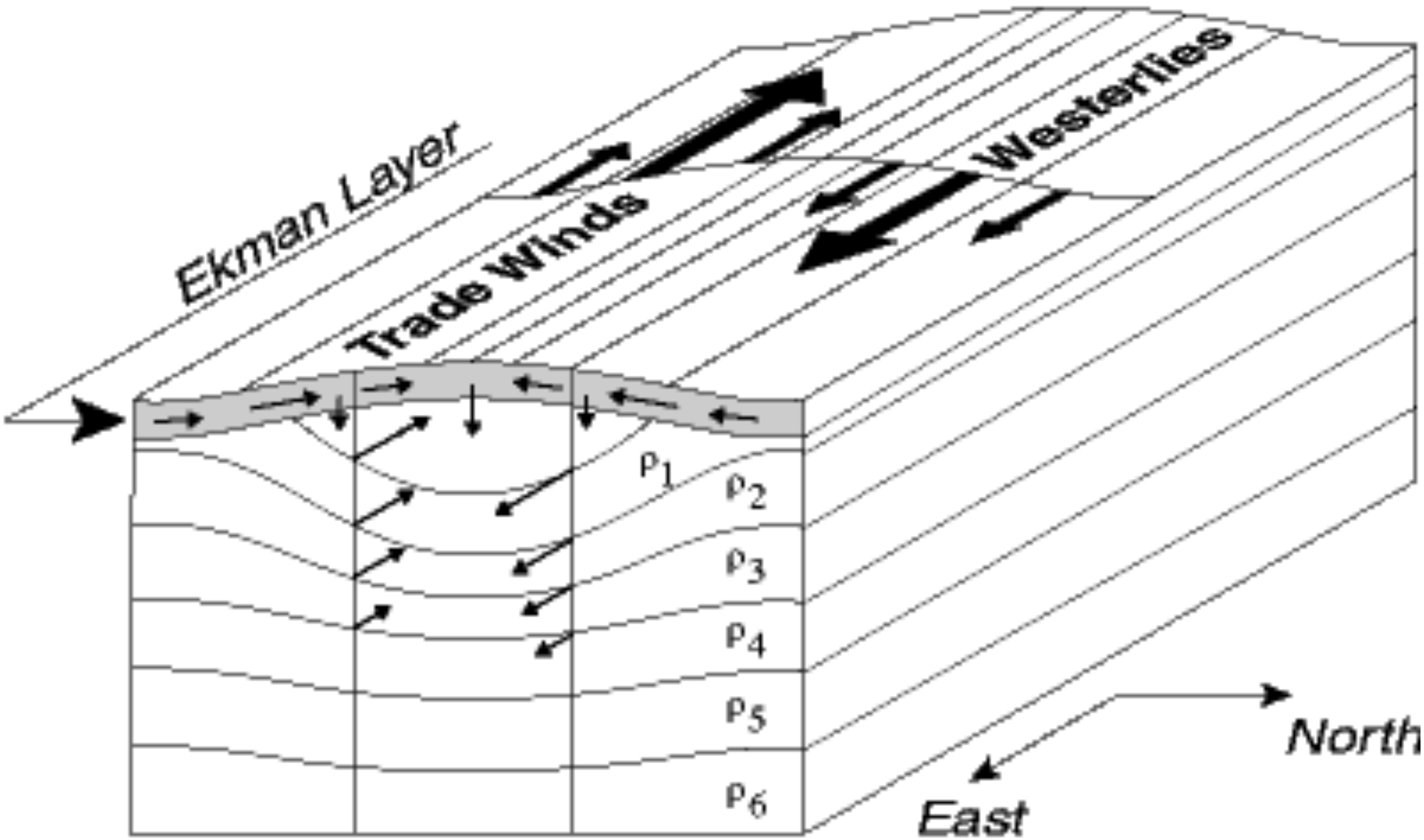
WEST



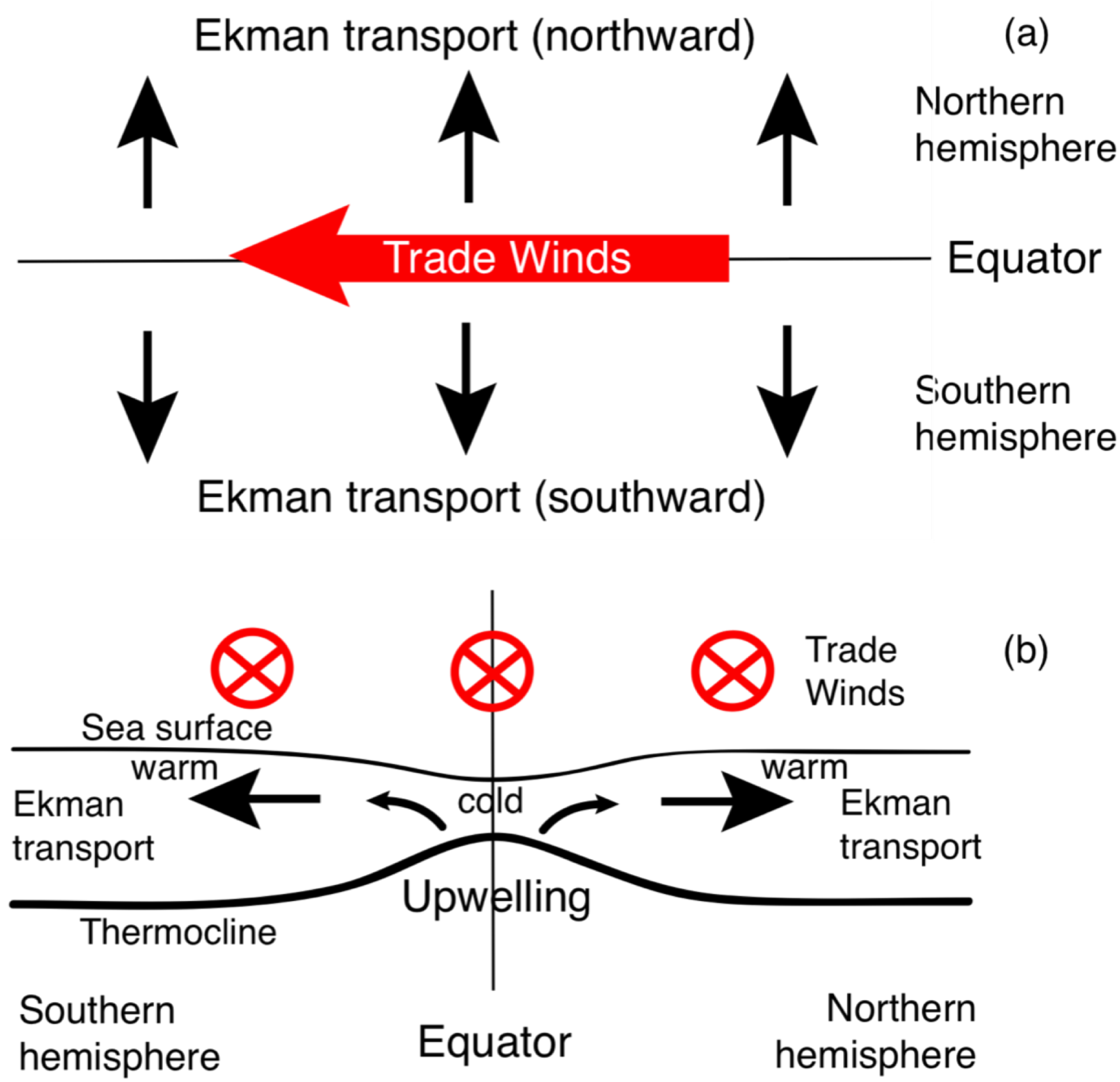
EAST

The view of the camera on top of the tank

Ekman convergence in the subtropics



Equatorial upwelling



Vertical motions induced by the Ekman flow

(a) schematic of coastal upwelling