Lecture 6: The Ocean General Circulation and Climate

Chapter 7: The Ocean General Circulation and Climate

- Abstract
- 7.1. Cauldron of climate
- 7.2. Properties of seawater
- 7.3. The mixed layer
- 7.4. The wind-driven circulation
- 7.5. Theories for wind-driven circulations
- 7.6. The deep thermohaline circulation
- 7.7. Transport of energy in the ocean
- 7.8. Mechanisms of transport in the ocean

- ❑ Basic Structures
- ❑ Mixed Layer
- ❑ Wind-Driven Circulation
- ❑ Theories
- ❑ Thermohaline Circulation
- ❑ Ocean Transports

Basic Ocean Structures

Warm up by sunlight!

- ❑ **Upper Ocean (~100 m)**

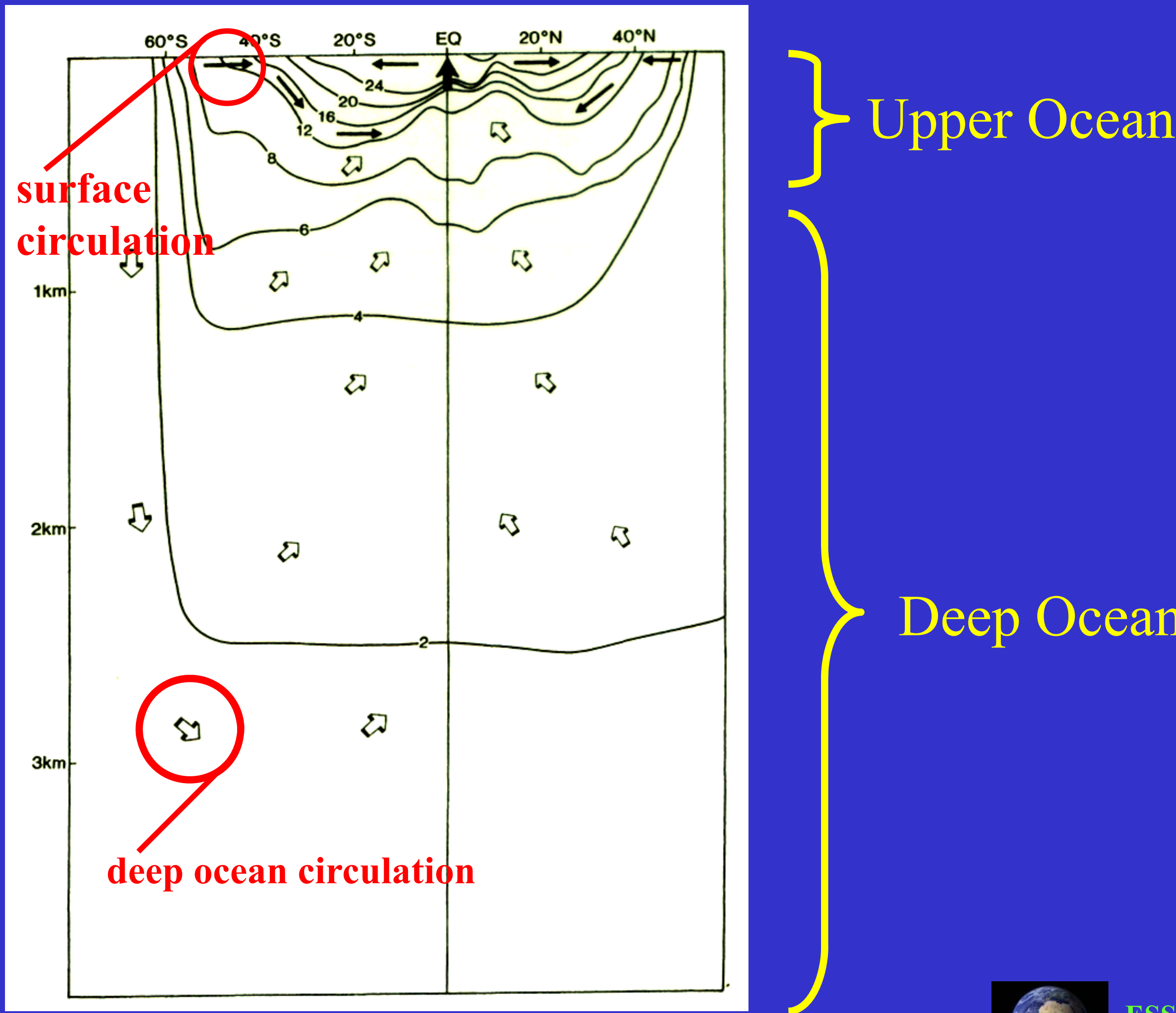
  Shallow, warm upper layer where light is abundant and where most marine life can be found.

- ❑ **Deep Ocean**

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

No sunlight!

Basic Ocean Current Systems



(from *"Is The Temperature Rising?"*)

The State of Oceans

- Temperature

  warm on the upper ocean, cold in the deeper ocean.

- Salinity

  variations determined by evaporation, precipitation, sea-ice formation and melt, and river runoff.

- Density

  small in the upper ocean, large in the deeper ocean.

Ocean Temperature

(from *Global Physical Climatology*)

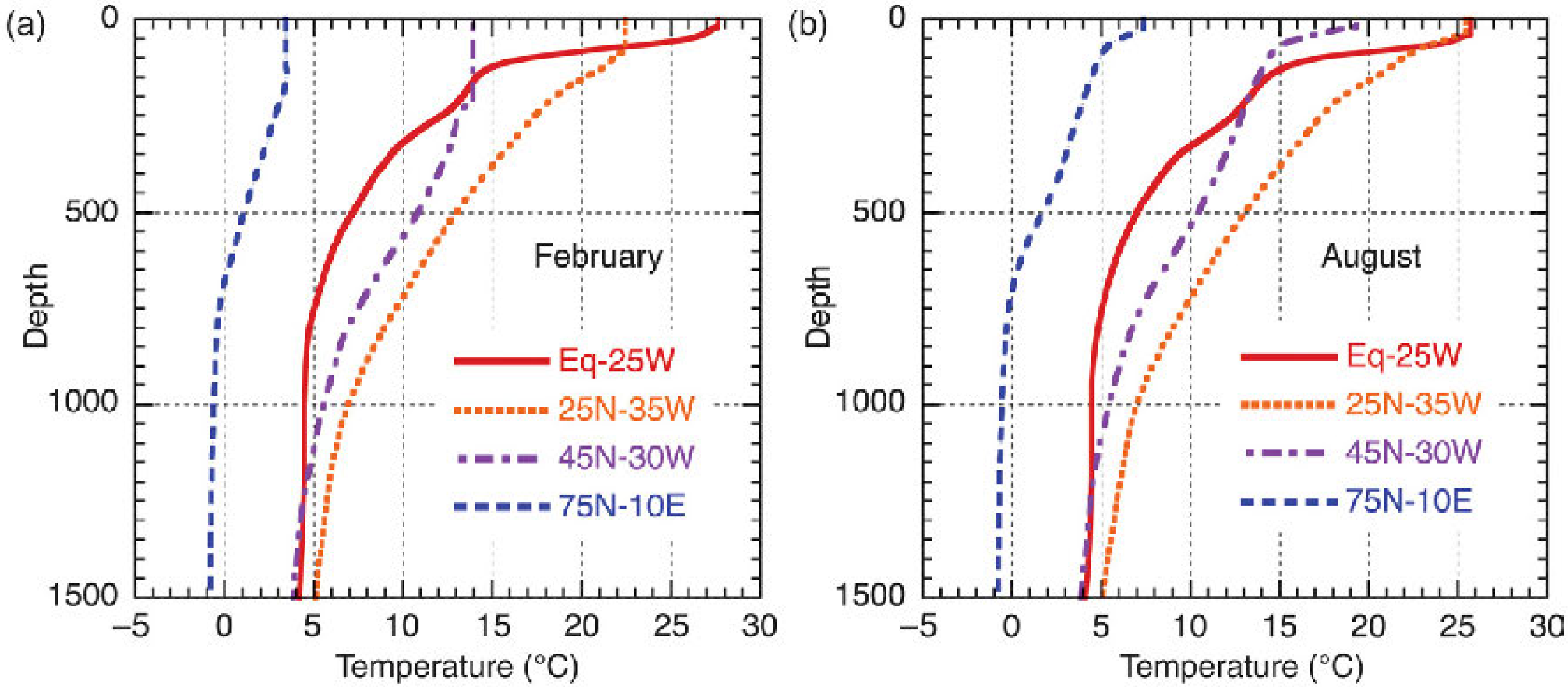
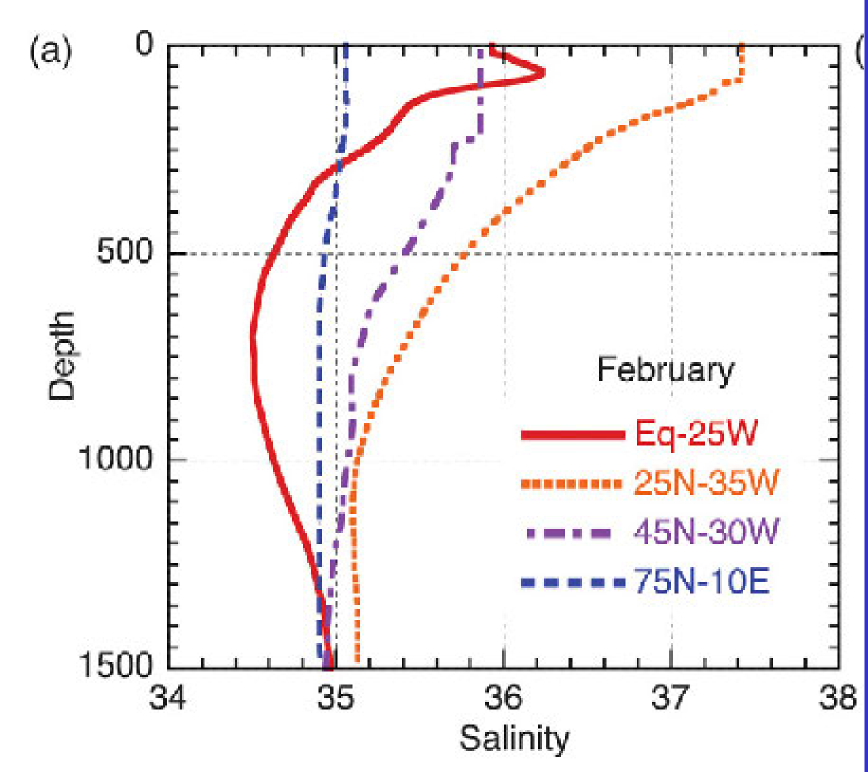


FIGURE 1.11 **Annual-mean ocean potential temperature profiles for various latitudes and as a function of depth in meters for (a) February and (b) August.** *MIMOC data.*

Temperature in the ocean generally decreases with depth from a temperature very near that of the surface air temperature to a value near the freezing point of water in the deep ocean

Ocean Salinity

(from *Global Physical Climatology)*

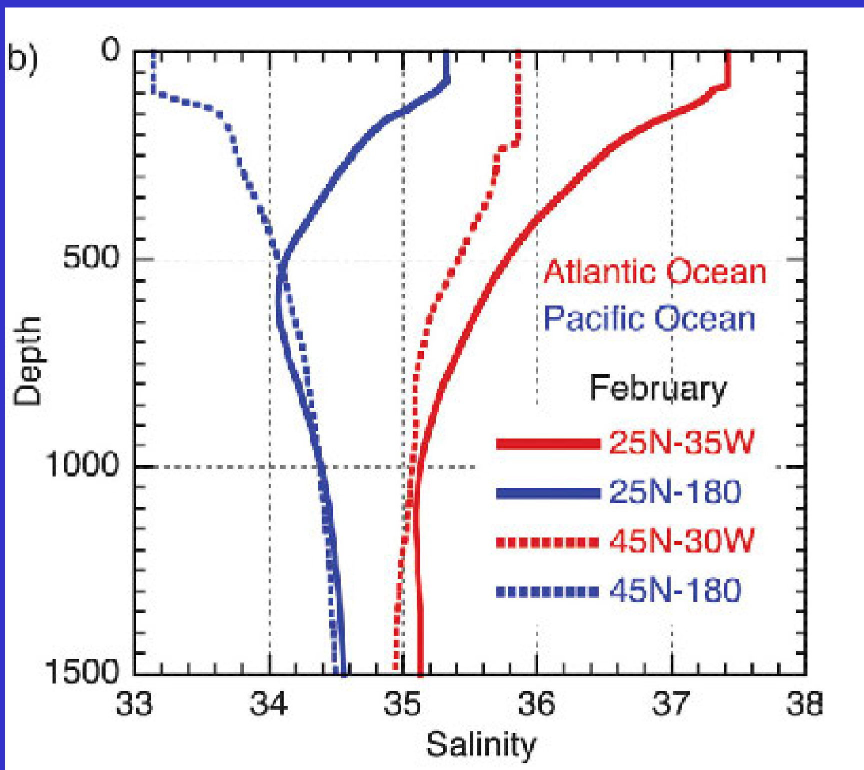


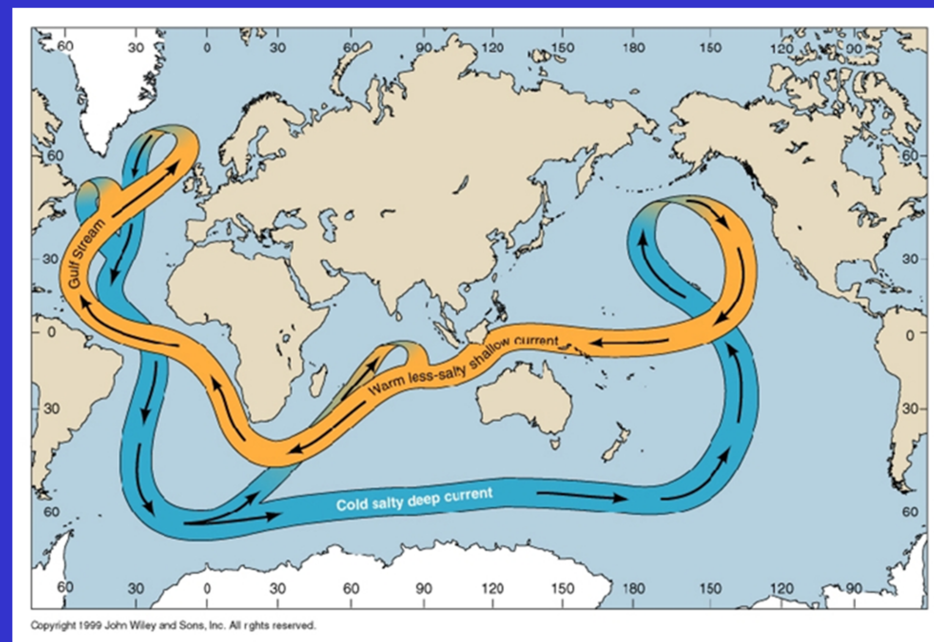
- Salinity of seawater is defined as the number of grams of dissolved salts in a kilogram of seawater.
- Salinity in the open ocean ranges from about 33 g/kg to 38 g/kg.
- Salinity is an important contributor to variations in the density of seawater at all latitudes and is the most important factor in high latitudes and in the deep ocean, where the temperature is close to the freezing point of water.

- Salinity of the global ocean varies systematically with latitude in the upper layers of the ocean.
- In the deep ocean, salinity variations are much smaller than near the surface, because the sources and sinks of freshwater are at the surface and the deep water comes from a few areas in high latitudes.

Ocean Salinity / Pacific vs. Atlantic

(from *Global Physical Climatology)*





- ❑ The Atlantic is much saltier than the Pacific at nearly all latitudes.
- ❑ For this reason the formation of cold, salty water that can sink to the bottom of the ocean is much more prevalent in the Atlantic than the Pacific.

Potential Temperature

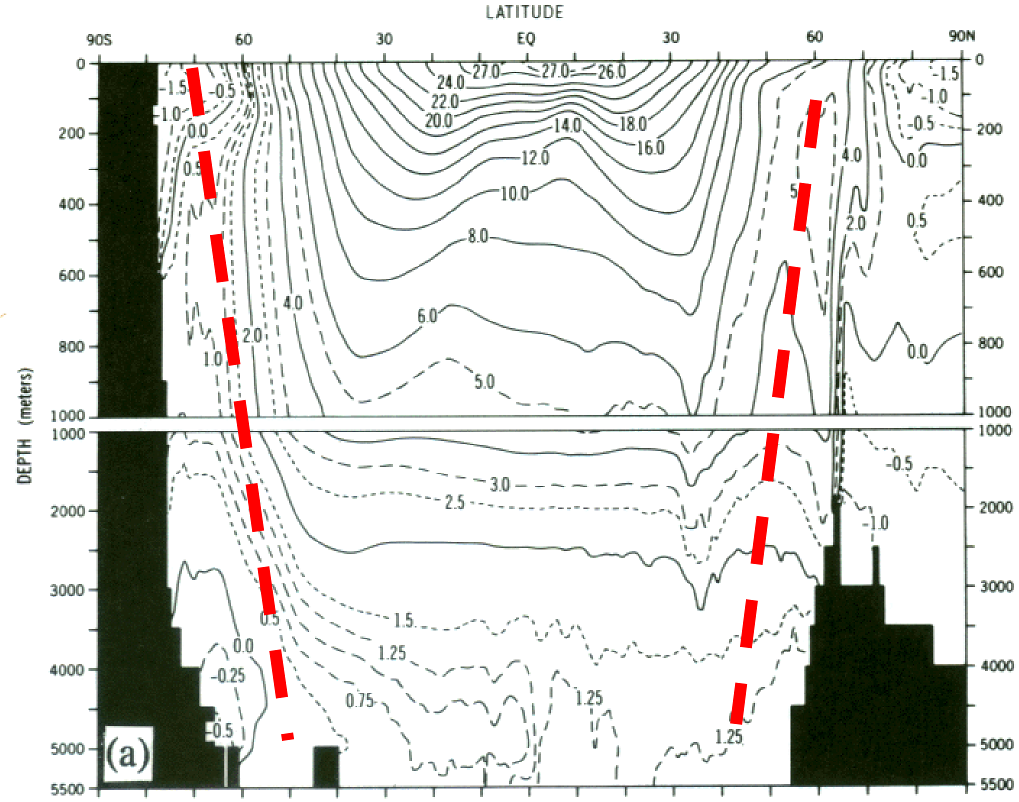


Fig. 7.1 Annual-mean zonal average for the global ocean of (a) potential temperature (°C), and (b) salinity [‰ (‰ = parts per thousand)], and (c) potential density ($\rho_t - 1000$, kg m^{-3}). [From Levitus (1982).]

(from *Global Physical Climatology*)

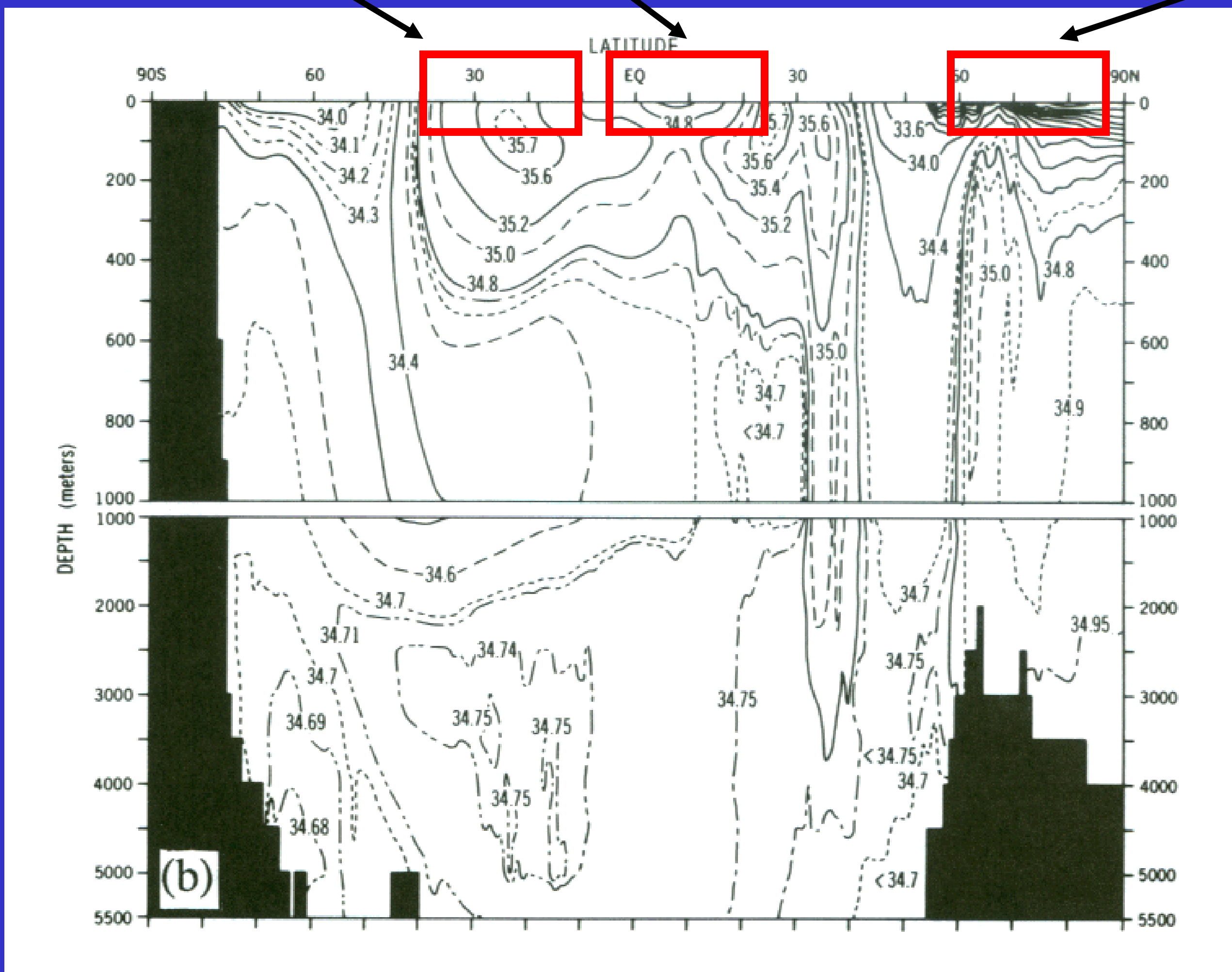
- ❑ Because water is slightly compressible, we define the potential temperature and potential density, which are the temperatures and densities at a reference pressure.
- ❑ Potential temperature is very close to temperature in the ocean.
- ❑ The average temperature of the world ocean is about 3.6°C.

Salinity

E > P

E < P

Sea-ice formation and melting

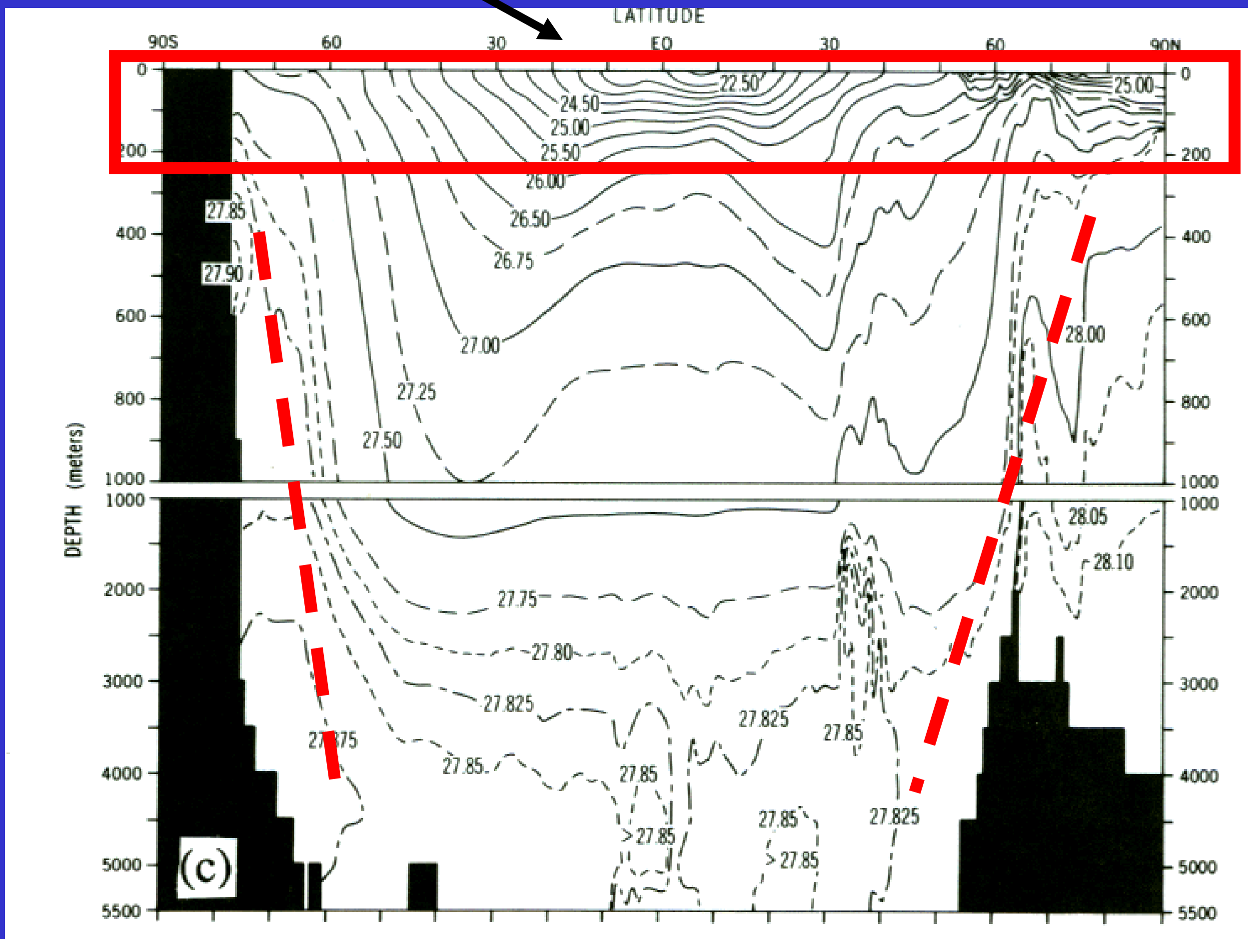


(from *Global Physical Climatology*)

- ❑ Salinity is the mass of dissolved salts in a kilogram of seawater.
- ❑ Unit: ‰ (part per thousand; per mil).
- ❑ The average salinity of the world ocean is 34.7‰.
- ❑ Four major factors that affect salinity: evaporation, precipitation, inflow of river water, and sea-ice formation and melting.

Low density due to absorption of solar energy near the surface.

Density

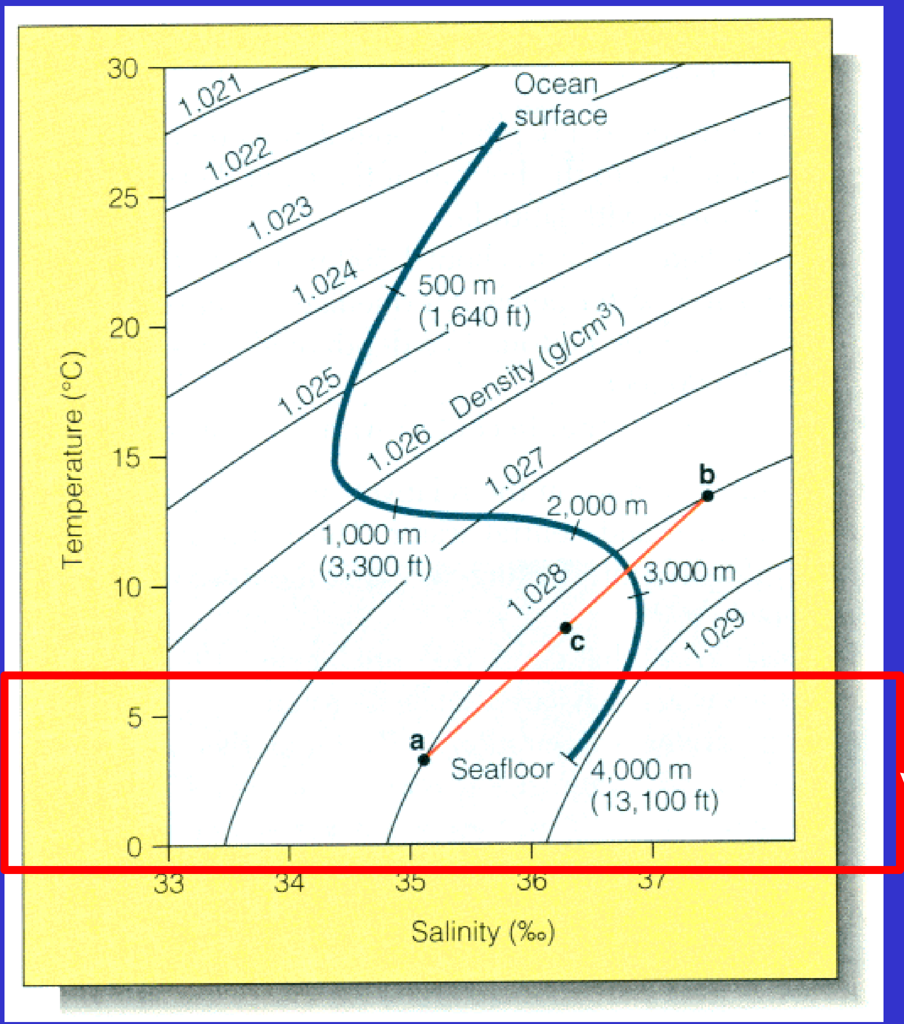


(from *Global Physical Climatology*)

- Seawater is almost incompressible, so the density of seawater is always very close to 1000 kg/m^3.
- Potential density is the density that seawater with a particular salinity and temperature would have at zero water pressure (or at surface air pressure).
- Potential density = density – 1000 kg/m^3.

ESS200
Prof. Jin-Yi Yu

Density and Temperature and Salinity



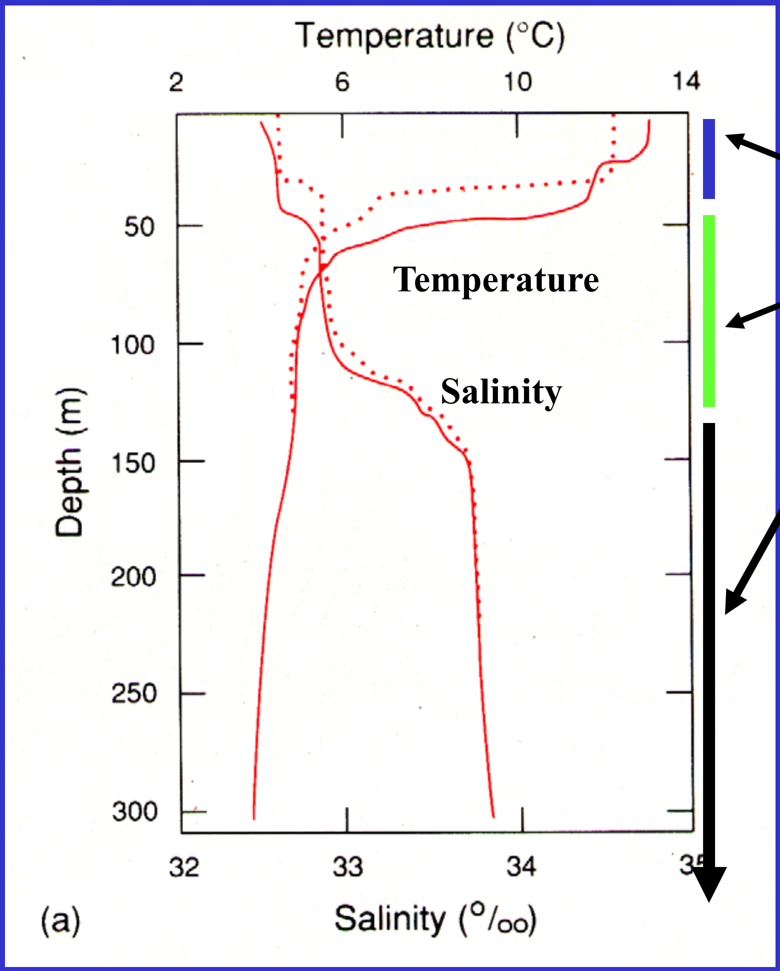
(Figure from *Oceanography* by Tom Garrison)

- Salinity and temperature variations have roughly equal importance for density variations in the ocean.
- The density of seawater is almost linearly dependent on salinity.
- However, the dependence of density on temperature does not have a simple linear behavior.
- For pure water, for example, the maximum density occurs at 4°C.

When the temperature of water approaches its freezing point, its density generally becomes less sensitive to temperature.

ESS200
Prof. Jin-Yi Yu

Vertical Structure of Ocean



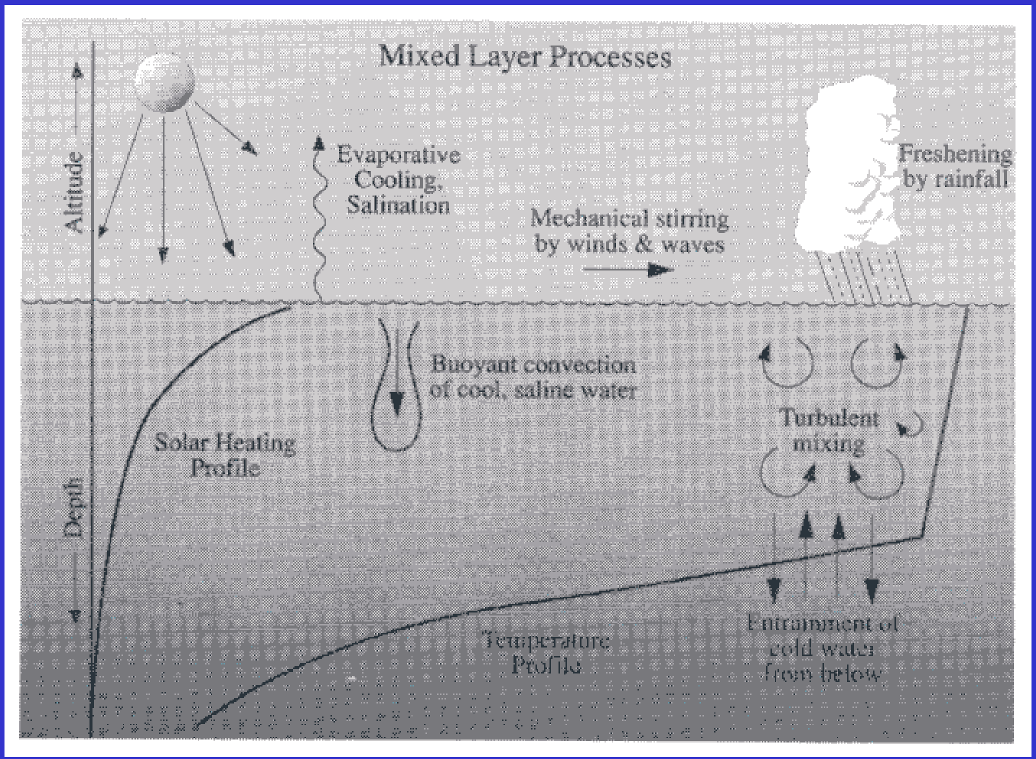
Mixed Layer: T and S well mixed by winds

Thermocline: large gradient of T and S

Deep Ocean: T and S independent of height
cold
salty
high nutrient level

(from *Climate System Modeling*)

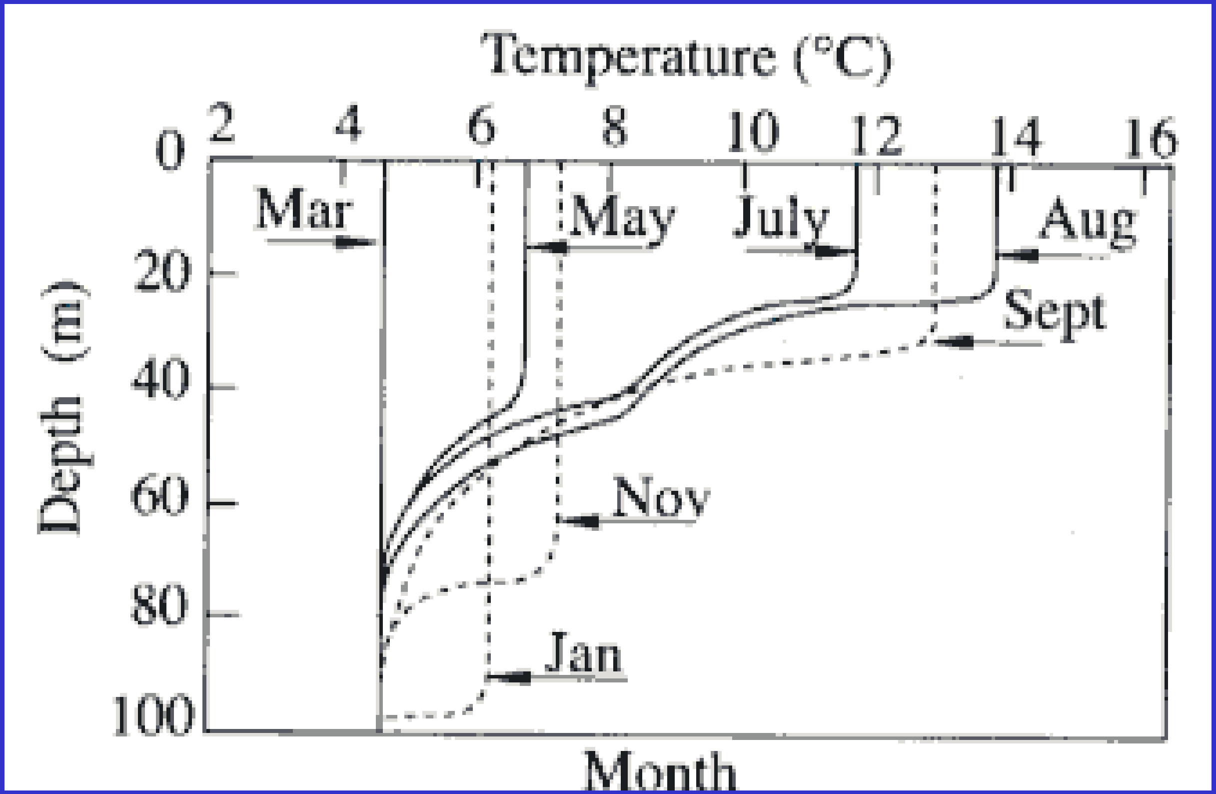
Mixed Layer Processes



(from *Global Physical Climatology*)

- ❑ The depth of the mixed layer is determined by (1) the rate of buoyancy generation and (2) the rate of kinetic energy supply.
- ❑ The atmosphere can affect the mixed layer through three processes: heating, wind forcing, and freshening (P-E).
- ❑ The global-average depth of the mixed layer is about 70 m.
- ❑ The heat capacity of the mixed layer is about 30 times the heat capacity of the atmosphere.

Seasonal Variation of Mixed Layer

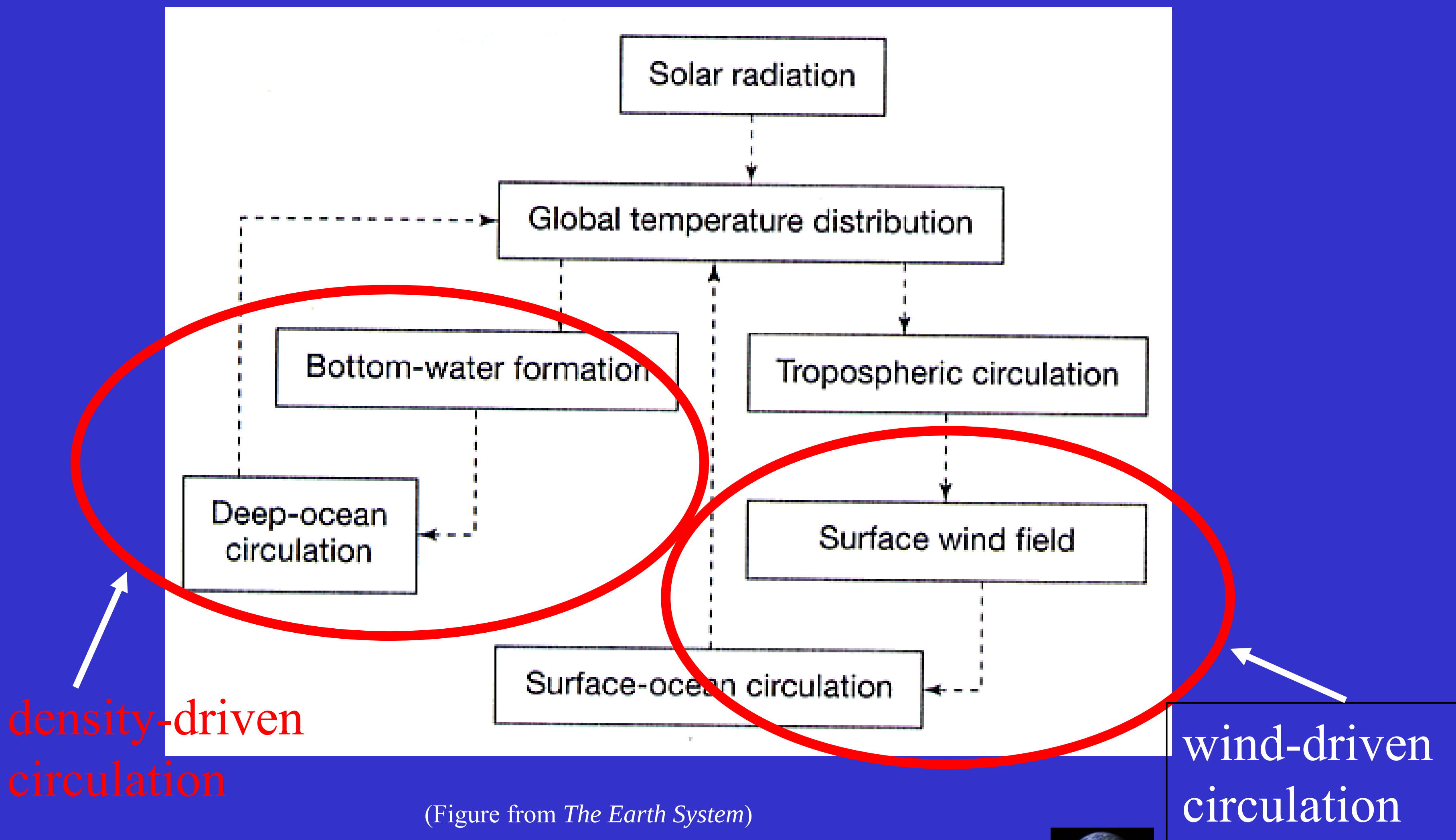


(from *Global Physical Climatology*)

- ❑ Summer: warm and thin.
- ❑ Winter: cold and deep (several hundred meters).

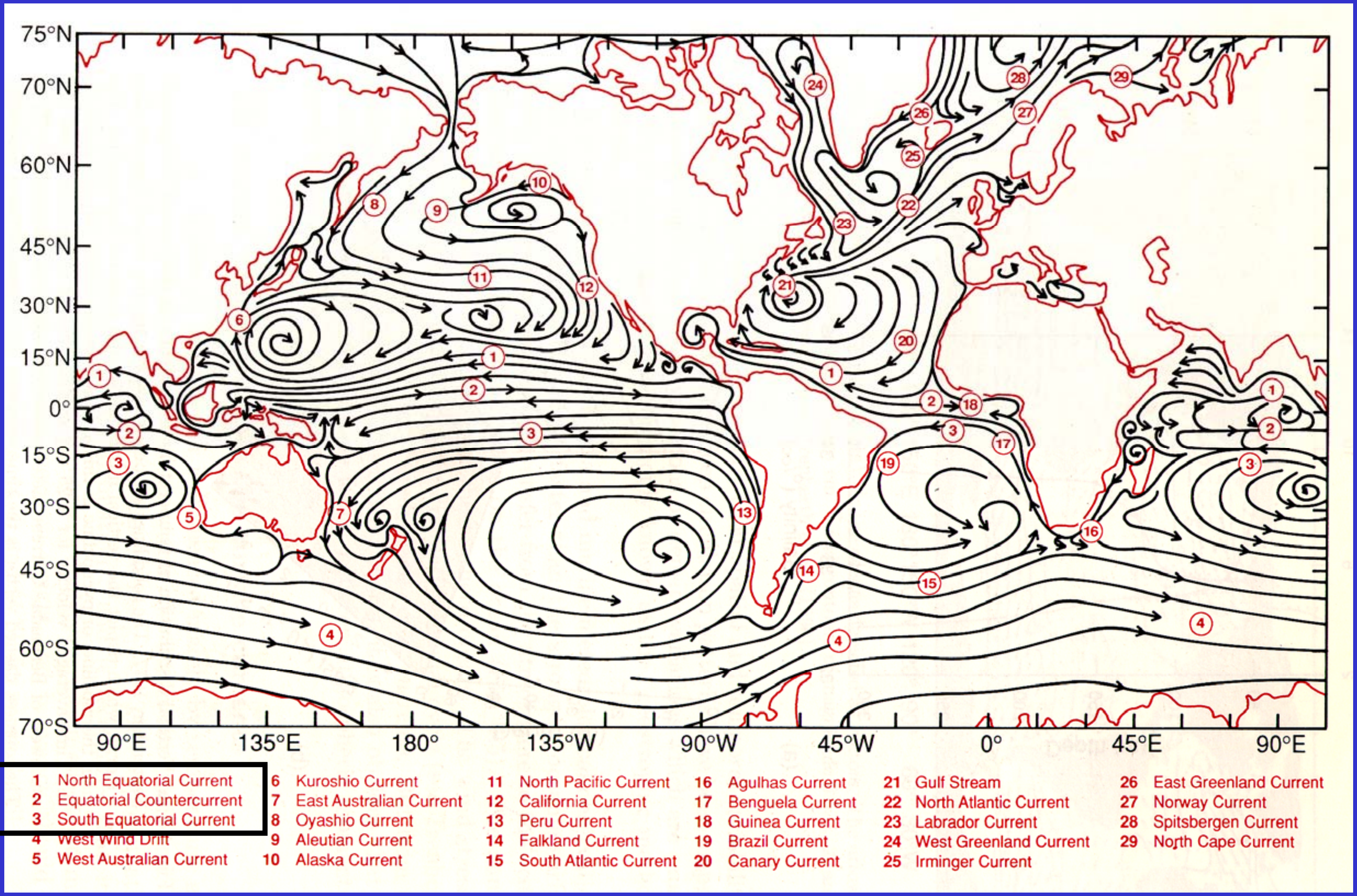
ESS200
Prof. Jin-Yi Yu

Two Circulation Systems of Oceans



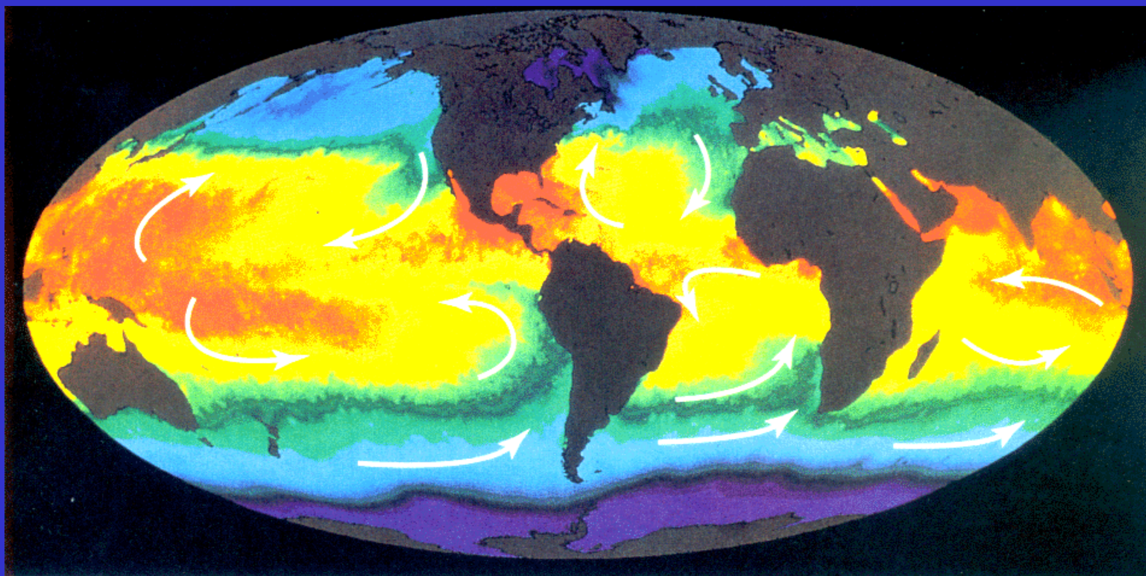
(Figure from *The Earth System*)

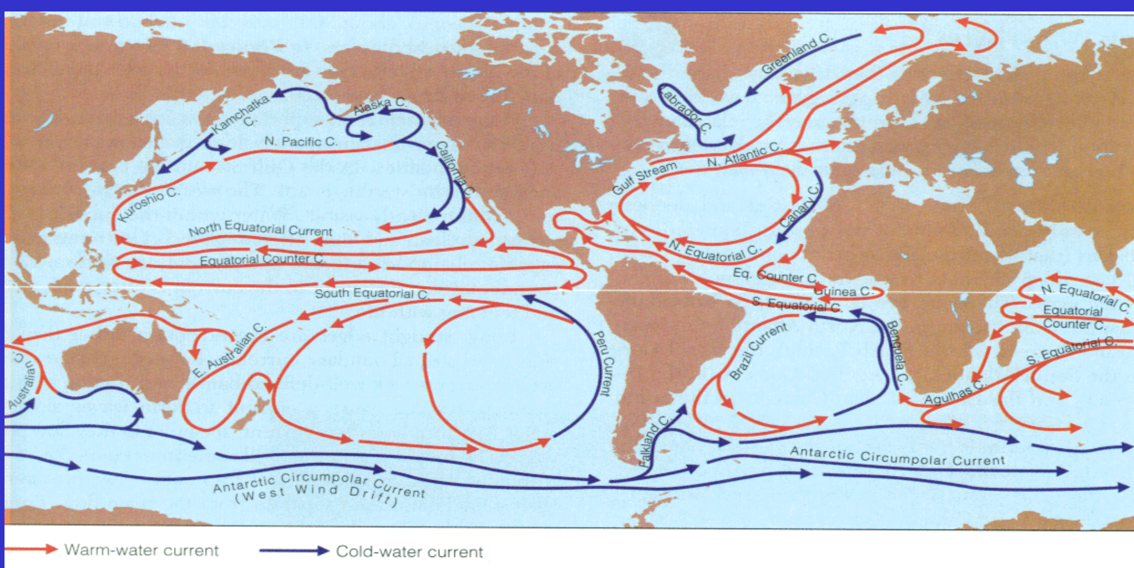
Global Surface Currents



(from *Climate System Modeling*)

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

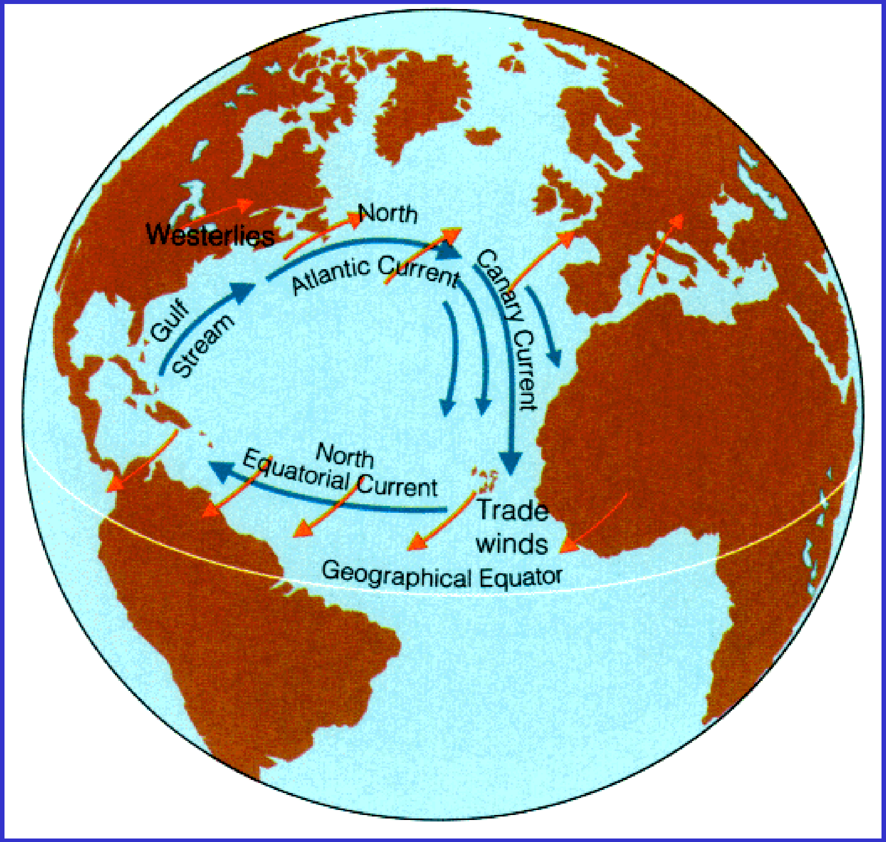
- **The 6th and the largest current:**
  - Antarctic Circumpolr Current
  - (also called West Wind Drift)

(Figure from *Oceanography* by Tom Garrison)

ESS200
Prof. Jin-Yi Yu

Characteristics of the Gyres

(Figure from *Oceanography* by Tom Garrison)



Volume transport unit:

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

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

- **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 polarward (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

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)

- **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)

- **Trade Wind-Driven Current**

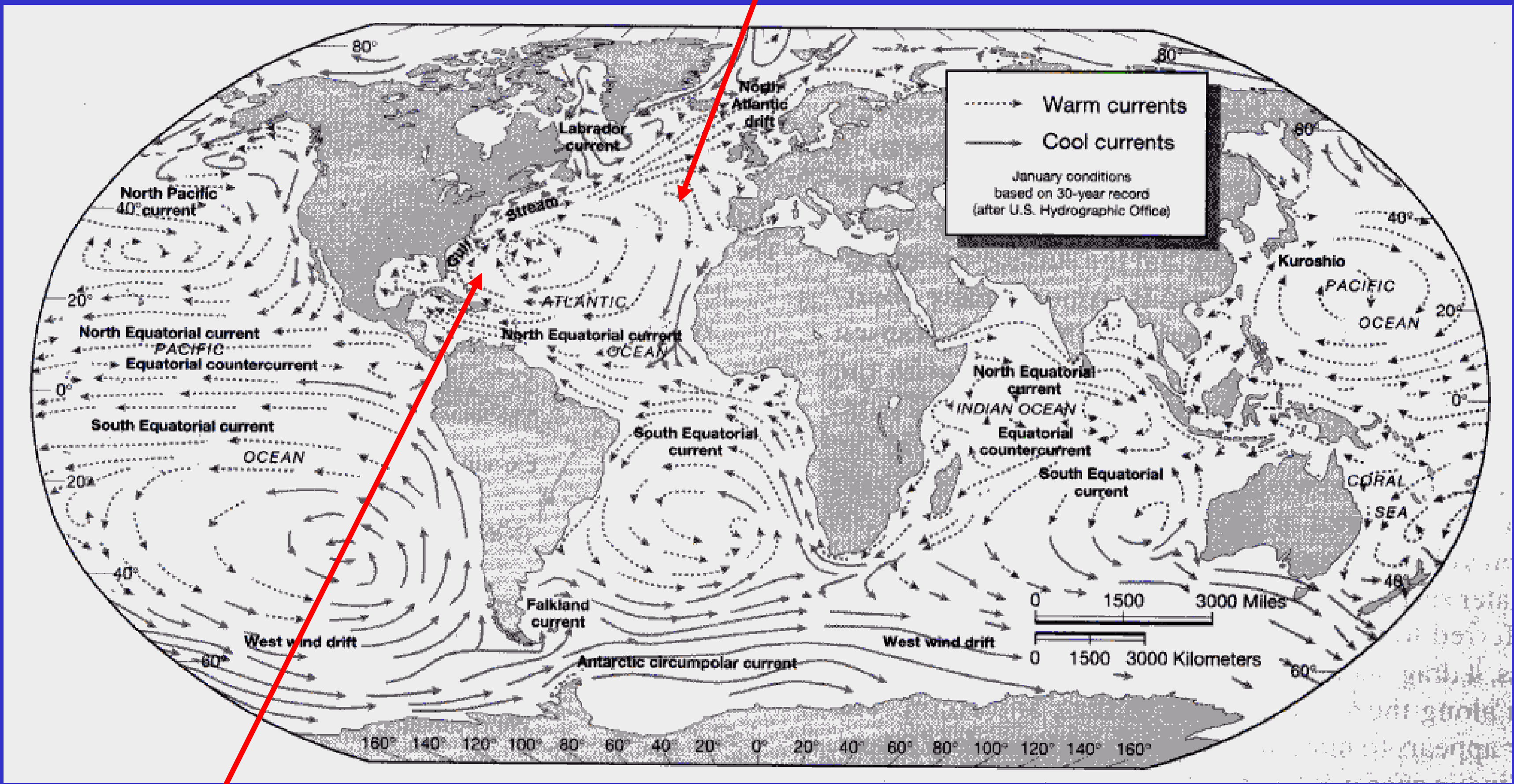
  North Equatorial Current
  South Equatorial Current

- **Westerly-Driven Current**

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

Boundary Currents

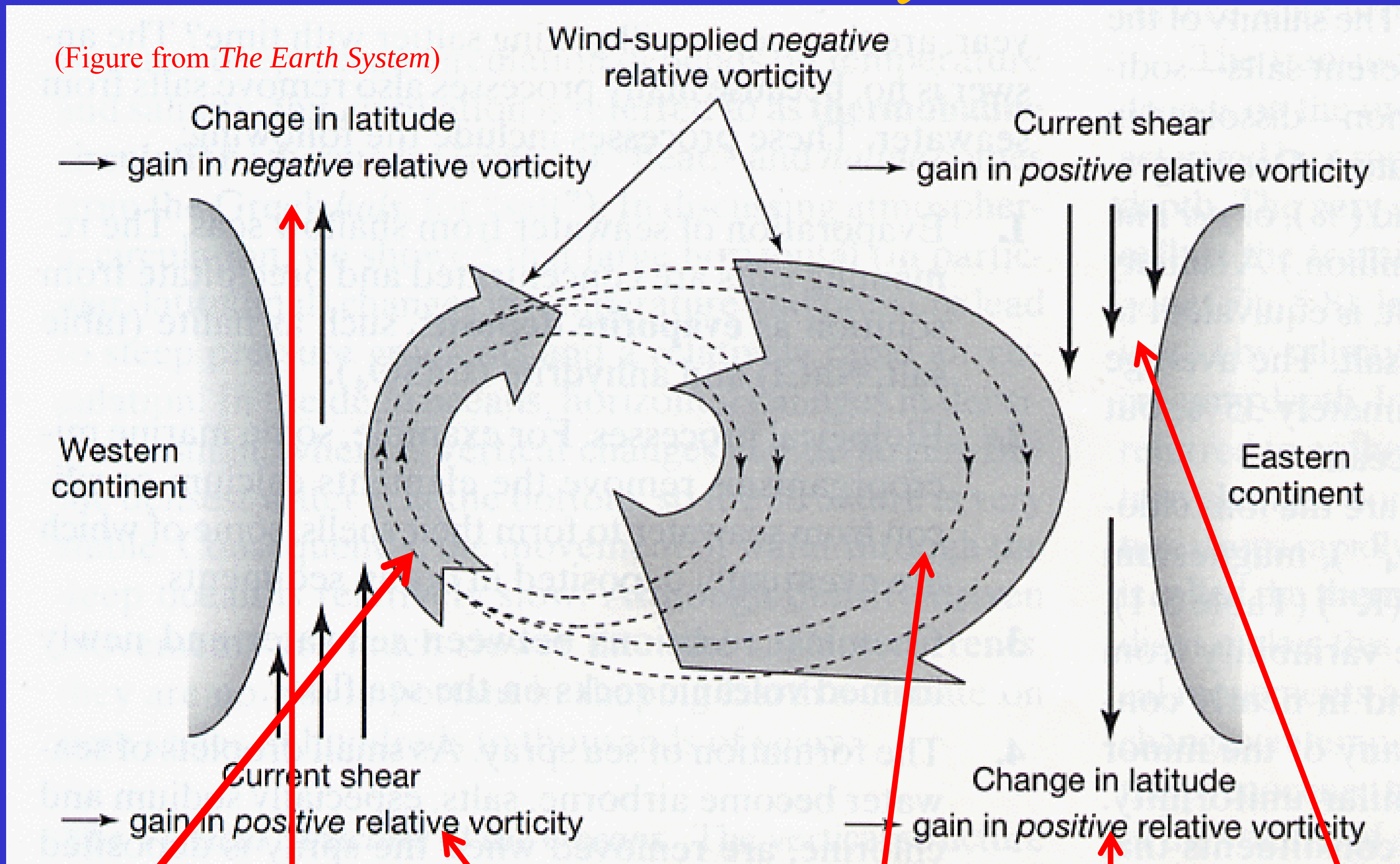
Eastern boundary currents: broad and weak



Western boundary currents: narrow and strong

Why Strong Boundary Currents?

A Potential Vorticity View



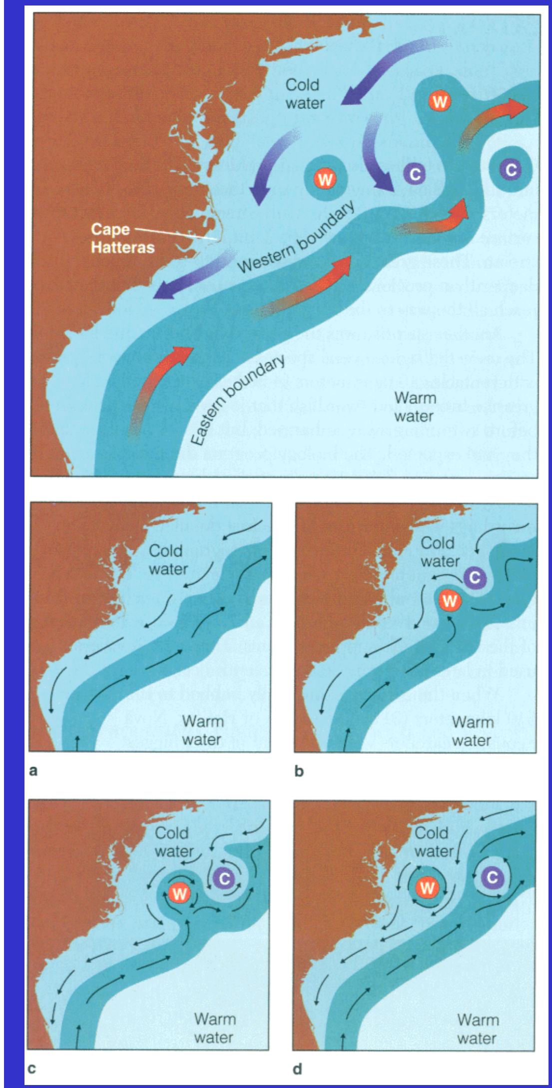
(Figure from *The Earth System*)

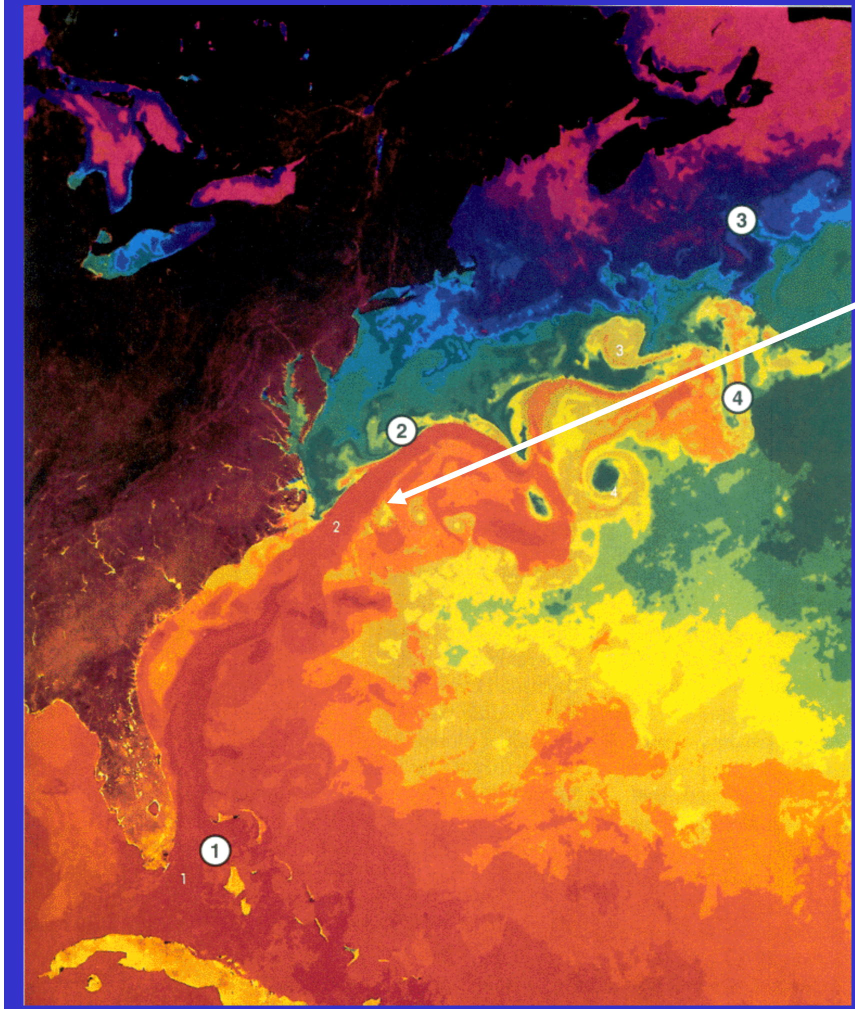
Goal: **Maintain the "steady state" of the negative vorticity induced by wind stress curve**

$\boxed{\xi -}$ = $\boxed{\xi -}$ **plus** $\boxed{\xi +}$ $\boxed{\xi -}$ = $\boxed{\xi +}$ **plus** $\boxed{\xi +}$

friction has to be big ➔ strong boundary current

Gulf Stream





A river of current

Jet stream in the ocean

- Speed = 2 m/sec
- Depth = 450 m
- Width = 70 Km
- Color: clear and blue

(Figure from *Oceanography* by Tom Garrison)

Eastern Boundary Current

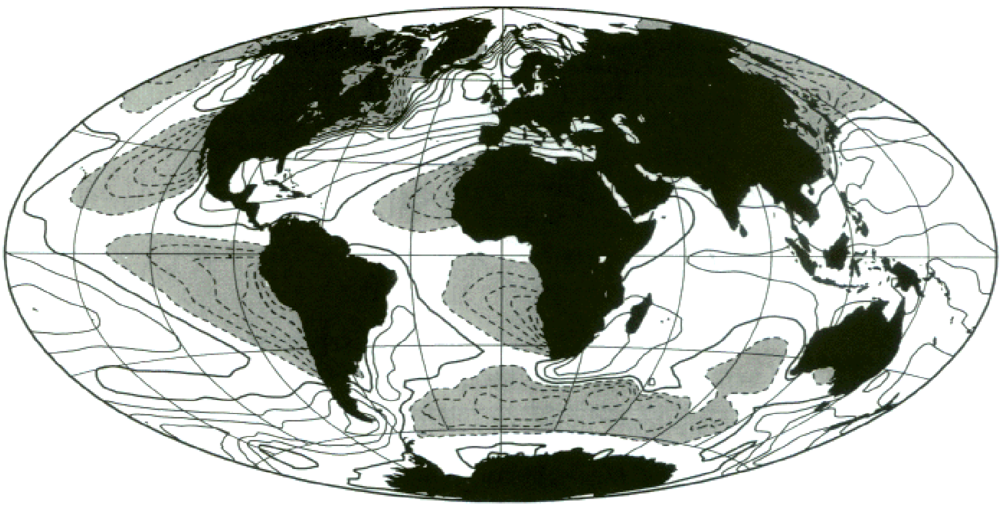
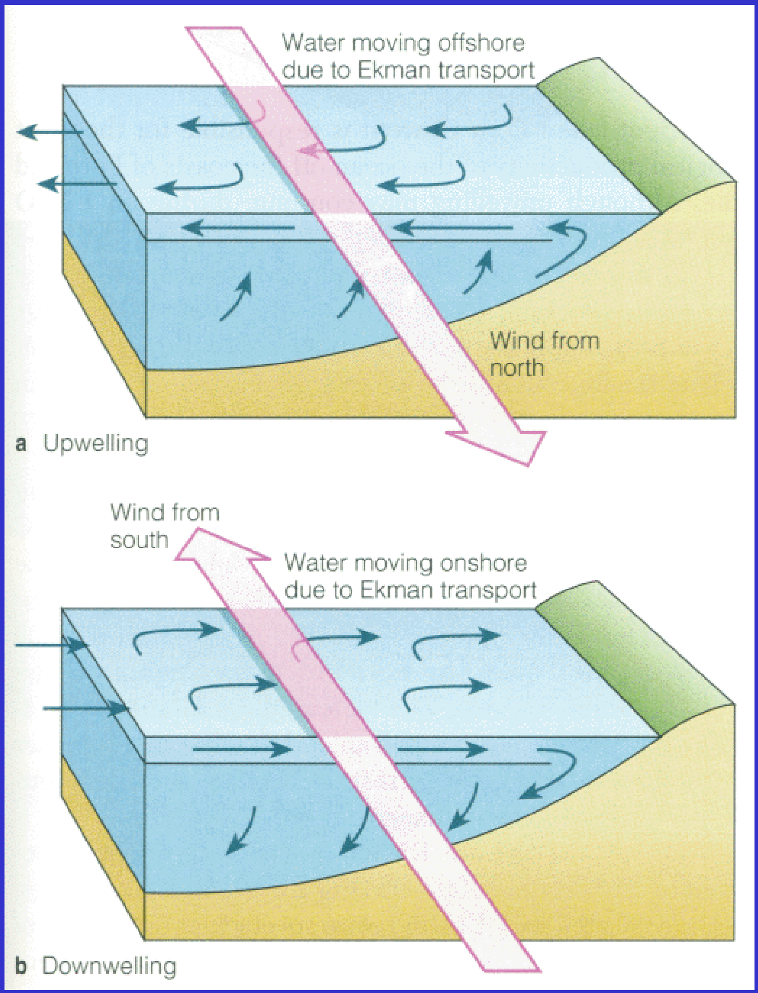


Fig. 7.11 The deviation of the July sea surface temperature from its zonal average at each latitude. Contour interval is 1°C, and values less than –1°C are shaded.

(from *Global Physical Climatology*)

- Cold water from higher latitude ocean.
- Costal upwelling associated with subtropical high pressure system.
- Atmospheric subsidence produce persistent stratiform clouds, which further cool down SSTs by blocking solar radiation.

Costal Upwelling/Downwelling



- ❑ A result of Ekman transport and mass continuity.

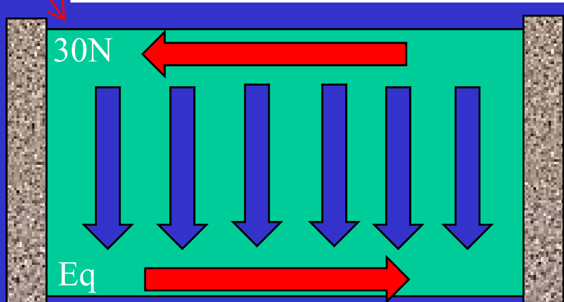
(Figure from *Oceanography* by Tom Garrison)

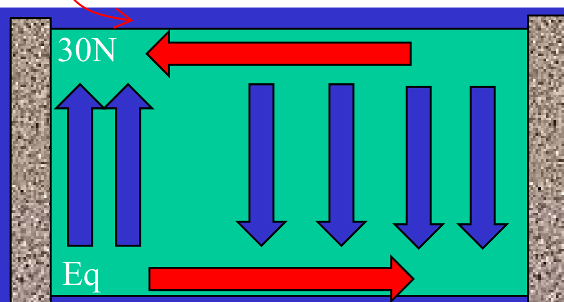
History / Wind-Driven Circulation

(from Robert H. Stewart's book on "*Introduction to Physical Oceanography*")

Table 9.2 Contributions to the Theory of the Wind-Driven Circulation

Fridtjof Nansen	(1898)	Qualitative theory, currents transport water at an angle to the wind.
Vagn Walfrid Ekman	(1902)	Quantitative theory for wind-driven transport at the sea surface.
Harald Sverdrup	(1947)	Theory for wind-driven circulation in the eastern Pacific.
Henry Stommel	(1948)	Theory for westward intensification of wind-driven circulation (western boundary currents).
Walter Munk	(1950)	Quantitative theory for main features of the wind-driven circulation.
Kirk Bryan	(1963)	Numerical models of the oceanic circulation.
Bert Semtner and Robert Chervin	(1988)	Global, eddy-resolving, realistic model of the ocean's circulation.





Surface Current – Geostrophic Gyre

- Ekman Layer

  Currents controlled by frictional force + Coriolis force

  → wind-driven circulation

  → Ekman transport (horizontal direction)

  → convergence/divergence

  → downwelling/upwelling at the bottom of mixed layer

- Thermocline

  downwelling/upwelling in the mixed layer

  → pressure gradient force + Coriolis force

  → geostrophic current

  → Sverdrup transport (horizontal)

Step 1: Surface Winds

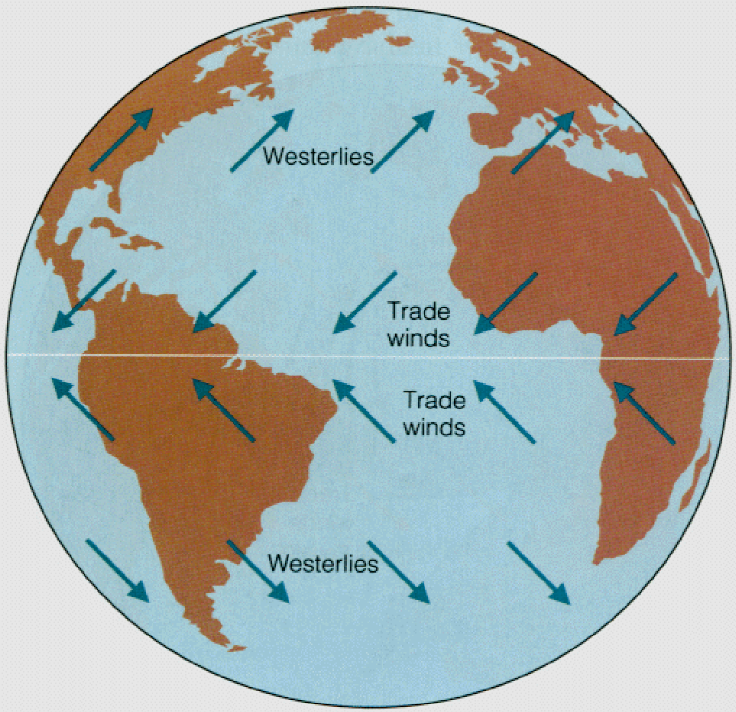


Figure 9.1 Winds, driven by uneven solar heating and Earth's spin, drive the movement of the ocean's surface currents. The prime movers are the powerful westerlies and the persistent trade winds (easterlies).

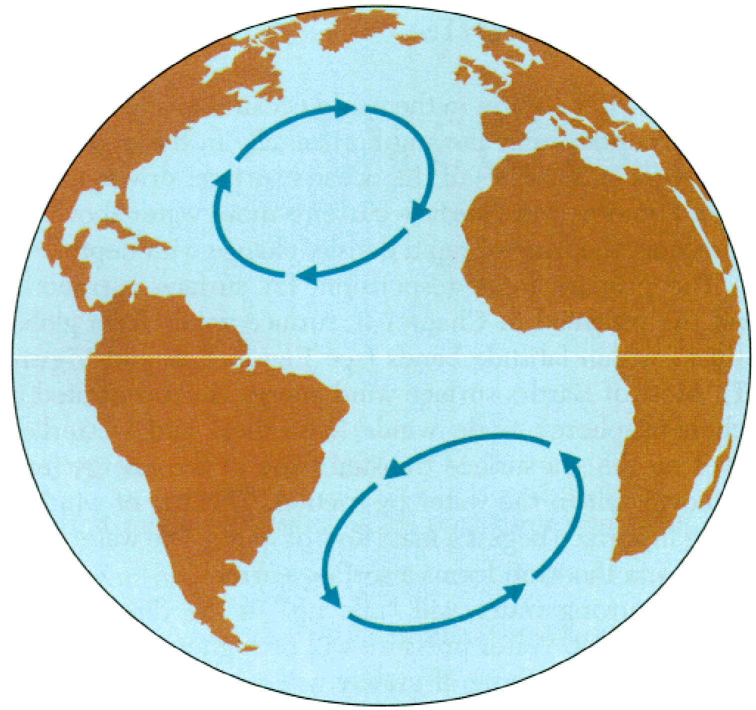
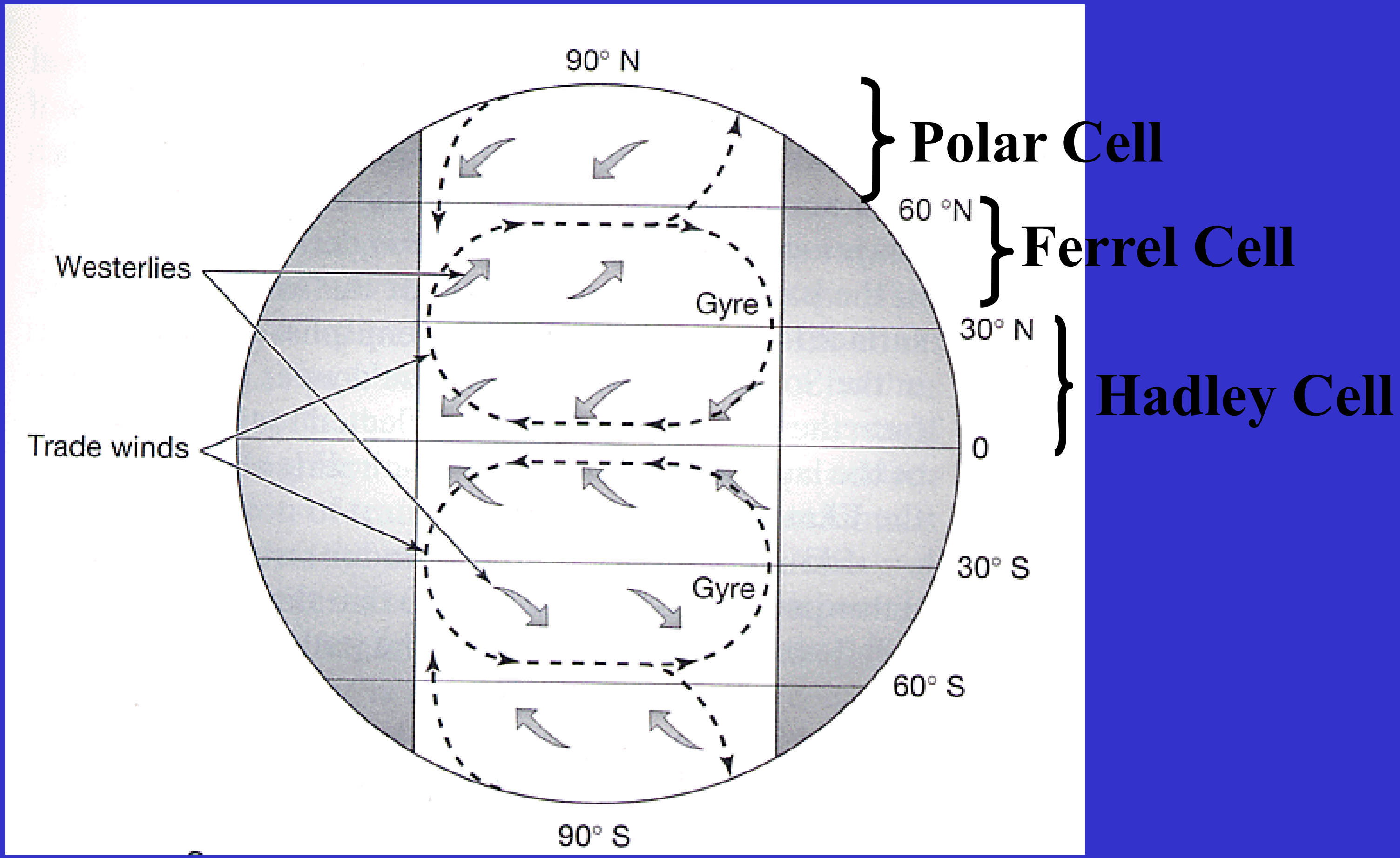


Figure 9.2 A combination of four forces—surface winds, the sun's heat, the Coriolis effect, and gravity—circulates the ocean surface clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere, forming gyres.

(Figure from *Oceanography* by Tom Garrison)

Winds and Surface Currents



(Figure from *The Earth System*)

Why an Angle btw Wind and Iceberg Directions?

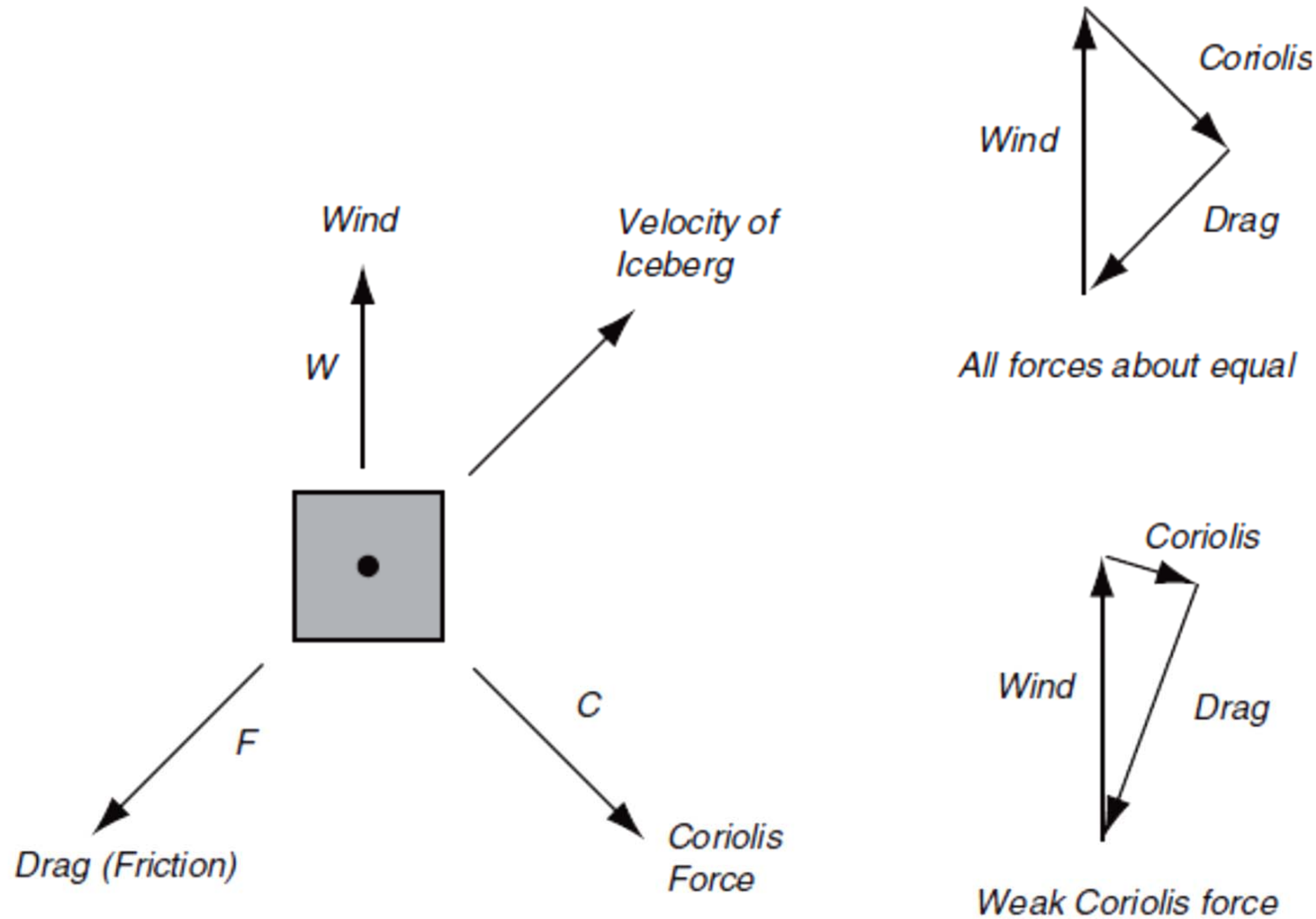
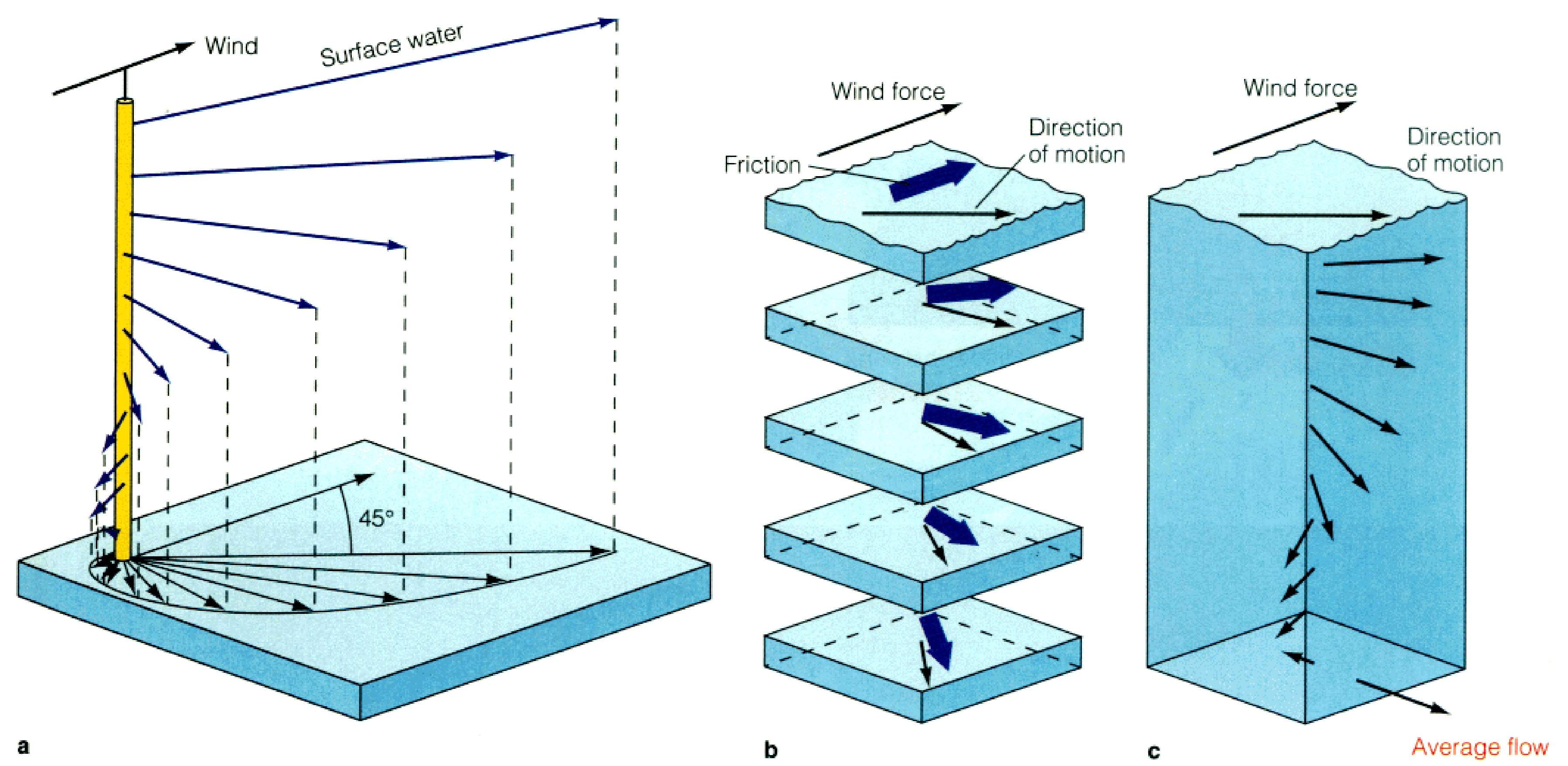


Figure 9.2 The balance of forces acting on an iceberg in a wind on a rotating Earth.

(from Robert H. Stewart's book on "*Introduction to Physical Oceanography*")

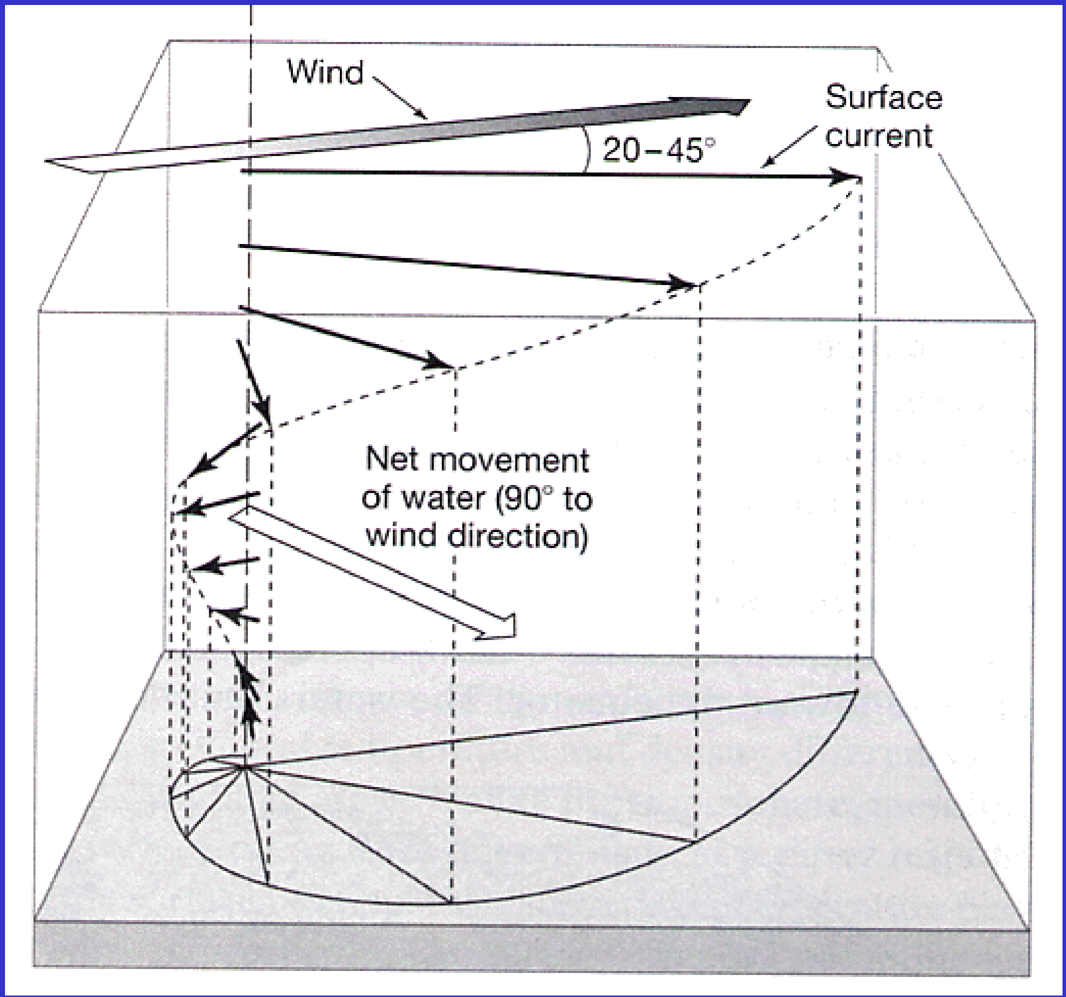
Step 2: Ekman Layer
(frictional force + Coriolis Force)



(Figure from *Oceanography* by Tom Garrison)

ESS200
Prof. Jin-Yi Yu

Ekman Spiral – A Result of Coriolis Force



(Figure from *The Earth System*)

ESS200
Prof. Jin-Yi Yu

In the Boundary Layer

For a steady state, homogeneous boundary layer

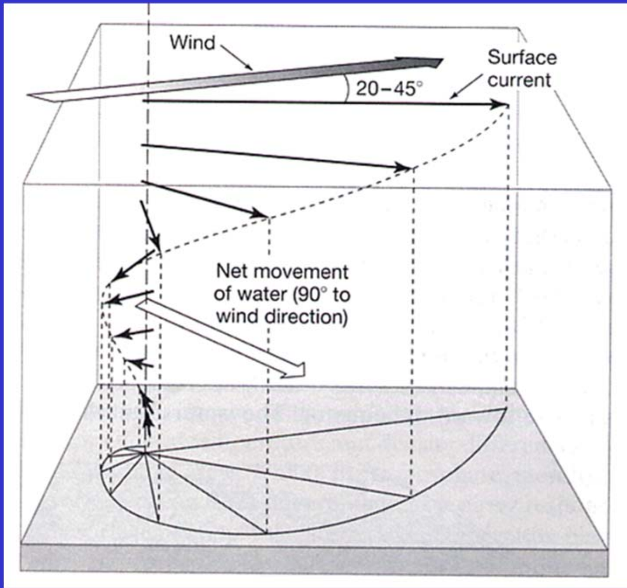
$$fv + A_z \frac{\partial^2 u}{\partial z^2} = 0$$
$$-fu + A_z \frac{\partial^2 v}{\partial z^2} = 0$$

Coriolis force balances frictional force

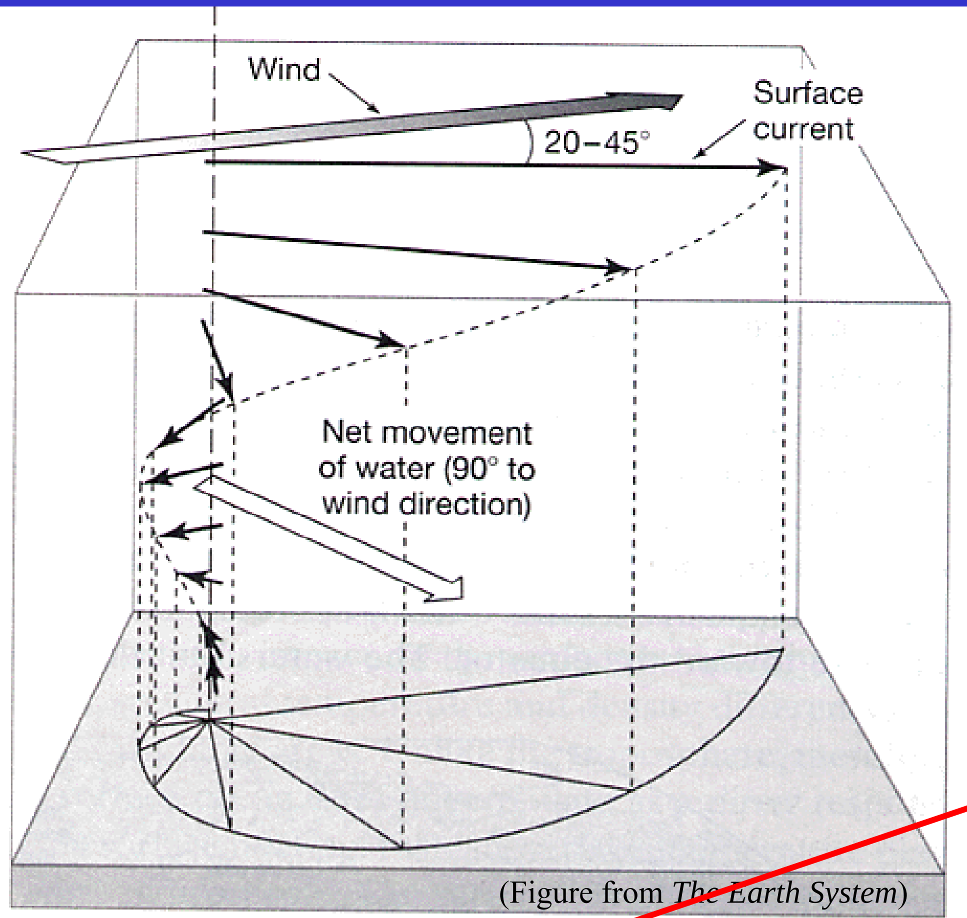
viscosity

$$u = V_0 \exp(az)\cos(\pi/4 + az)$$
$$v = V_0 \exp(az)\sin(\pi/4 + az)$$

$$a = \sqrt{\frac{f}{2A_z}}$$



Ekman Transport



(Figure from *The Earth System*)

$$fv + A_z \frac{\partial^2 u}{\partial z^2} = 0$$
$$-fu + A_z \frac{\partial^2 v}{\partial z^2} = 0$$

or

$$\rho f V + \frac{\partial T_{xz}}{\partial z} = 0$$
$$\rho f U - \frac{\partial T_{yz}}{\partial z} = 0$$

$$U_E = \int_{-\infty}^{0} u_E \, dz = \frac{\tau_y}{\rho_o f}; \qquad V_E = \int_{-\infty}^{0} v_E \, dz = -\frac{\tau_x}{\rho_o f}$$

How Deep is the Ekman Layer?

(from *Climate System Modeling*)

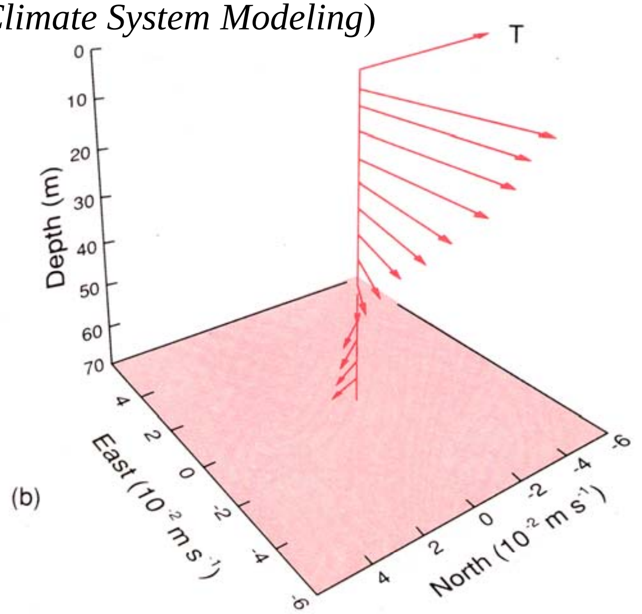


Fig. 4.4 (a) Vertical distribution of temperature and salinity at 50°N., 145°W. in early September, 1977. The solid lines are before a storm and the dotted lines are after a storm, which depict the vertical mixing above the seasonal thermocline. The main thermocline, or pycnocline in this area is between 110 m and 160 m depth. (b) Time-averaged velocity for a 25 day summer period at an open ocean site southwest of Bermuda. Current meter measured velocity is referenced to 70 m. The topmost dashed vector is the time-averaged wind stress (Price et al., 1986).

- $D \propto (\nu/f)^{1/2}$

ν = vertical diffusivity of momentum

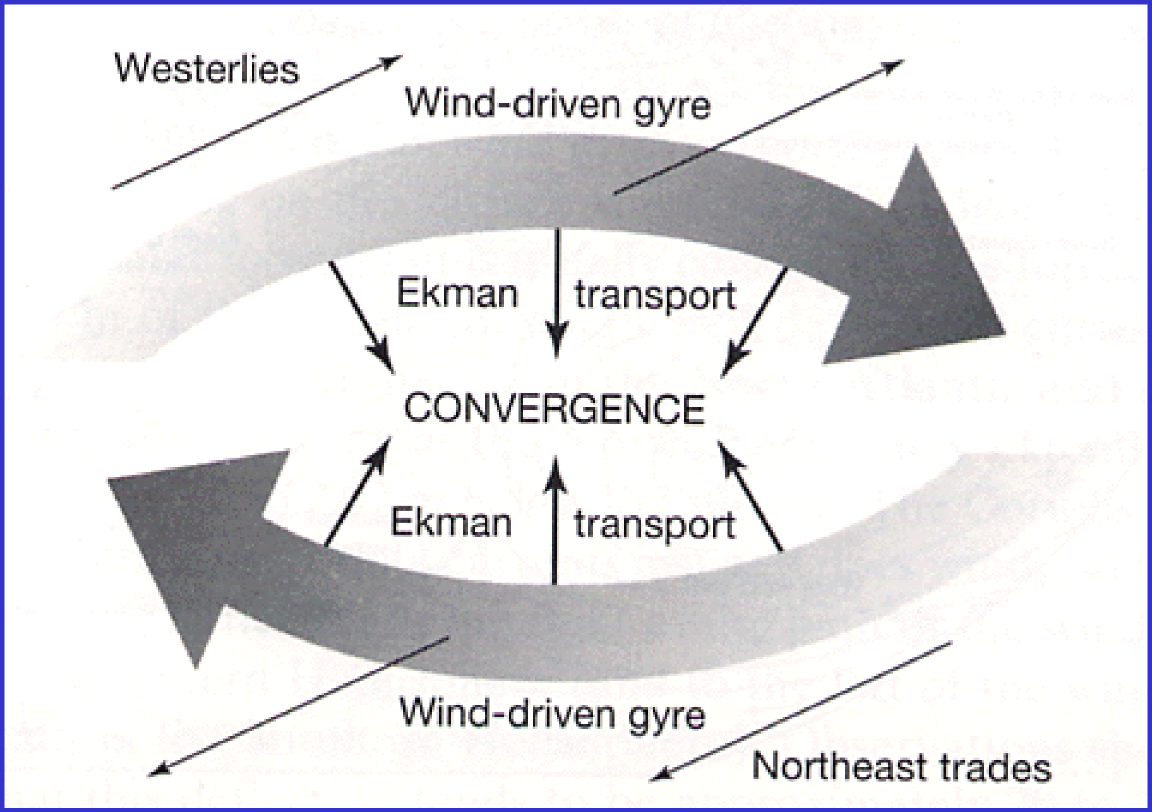
f = Coriolis parameter = $2\Omega\sin\phi$

The thickness of the Ekman layer is arbitrary because the Ekman currents decrease exponentially with depth. Ekman proposed that the thickness be the depth D_E at which the current velocity is opposite the velocity at the surface, which occurs at a depth $D_E = \pi/a$

$$\boxed{D_E = \sqrt{\frac{2\pi^2 A_z}{f}}}$$

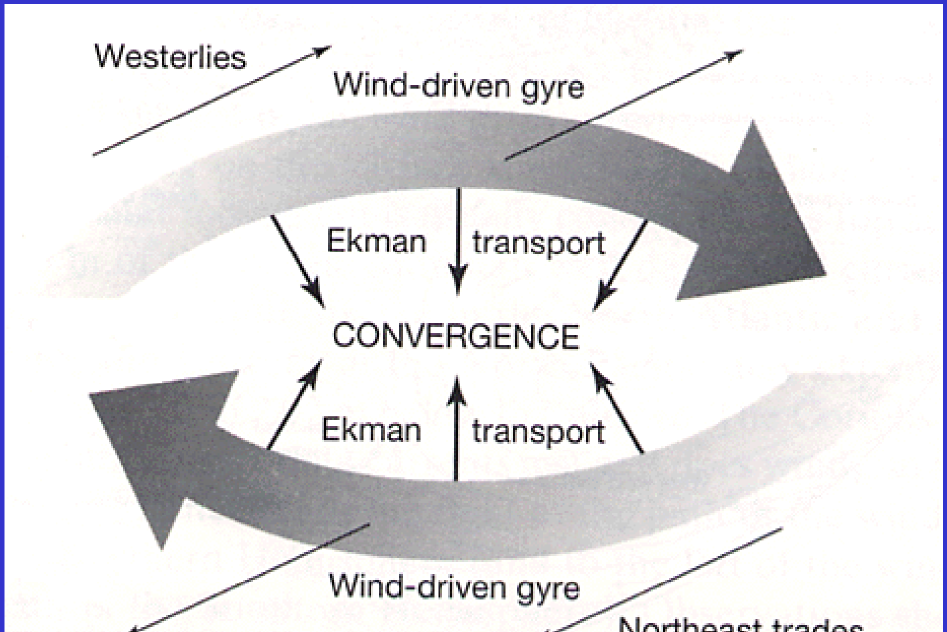
(from *Robert H. Steward*)

Ekman Transport

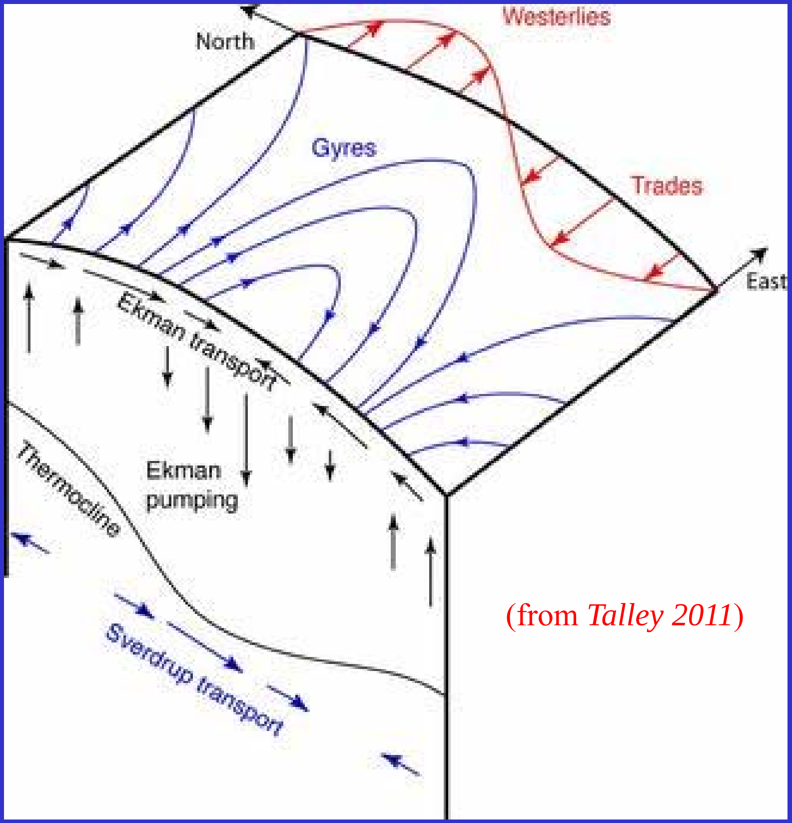


(Figure from *The Earth System*)

Ekman Transport and Ekman Pumping

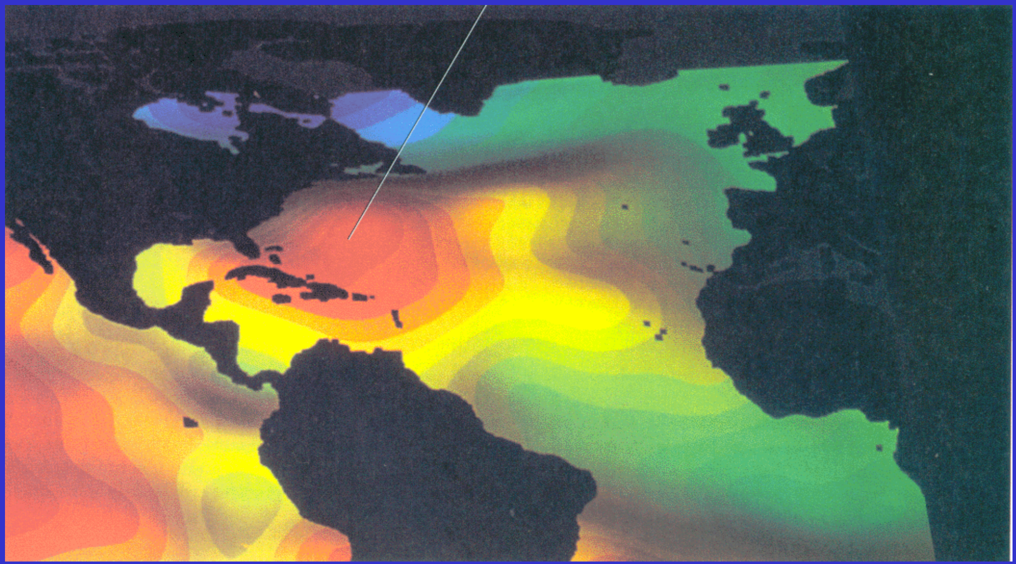


(from *The Earth System*)

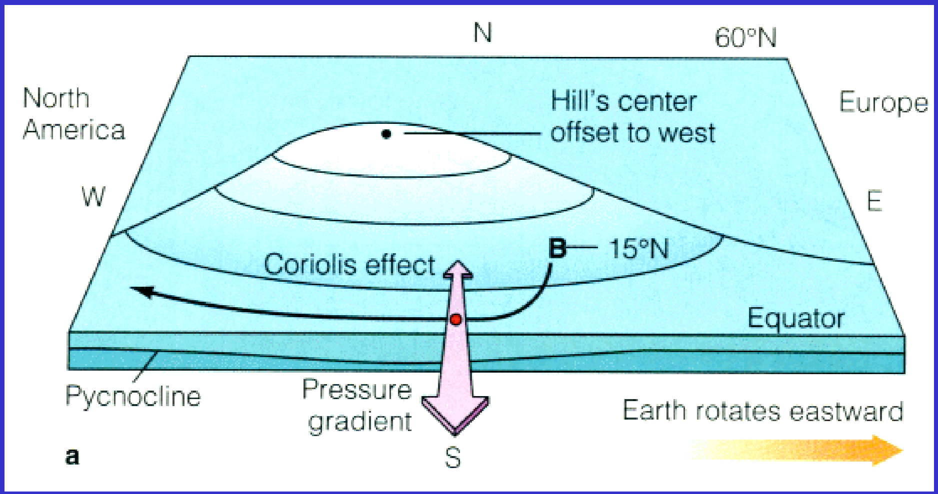


(from *Talley 2011*)

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



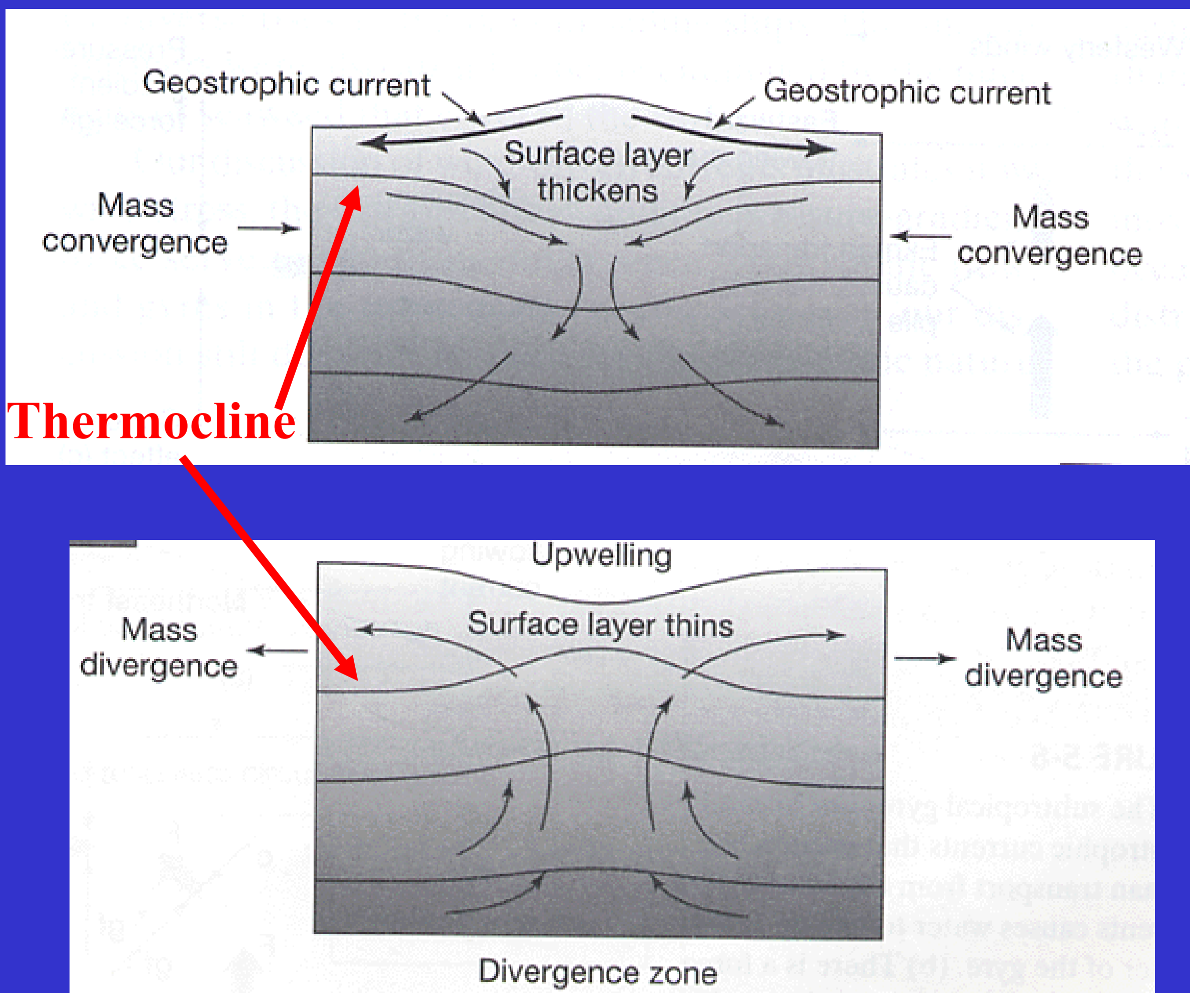
NASA-TOPEX Observations of Sea-Level Hight



(from *Oceanography* by Tom Garrison)

Ekman Transport → Convergence/Divergence

(Figure from *The Earth System*)



Surface wind + Coriolis Force

↓

Ekman Transport

↓

Convergence/divergence (in the center of the gyre)

↓

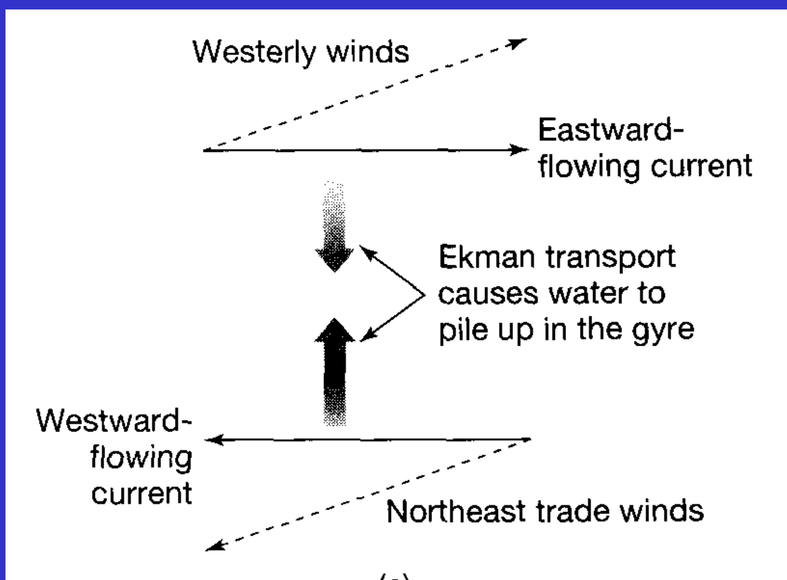
Pressure Gradient Force

↓

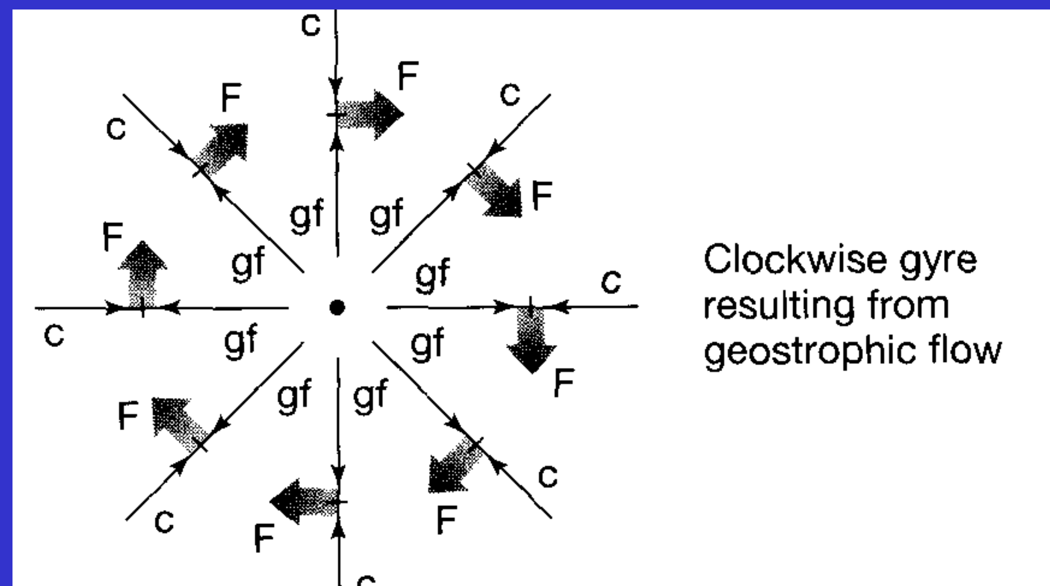
Geostrophic Currents

Geostrophic Current

Forces

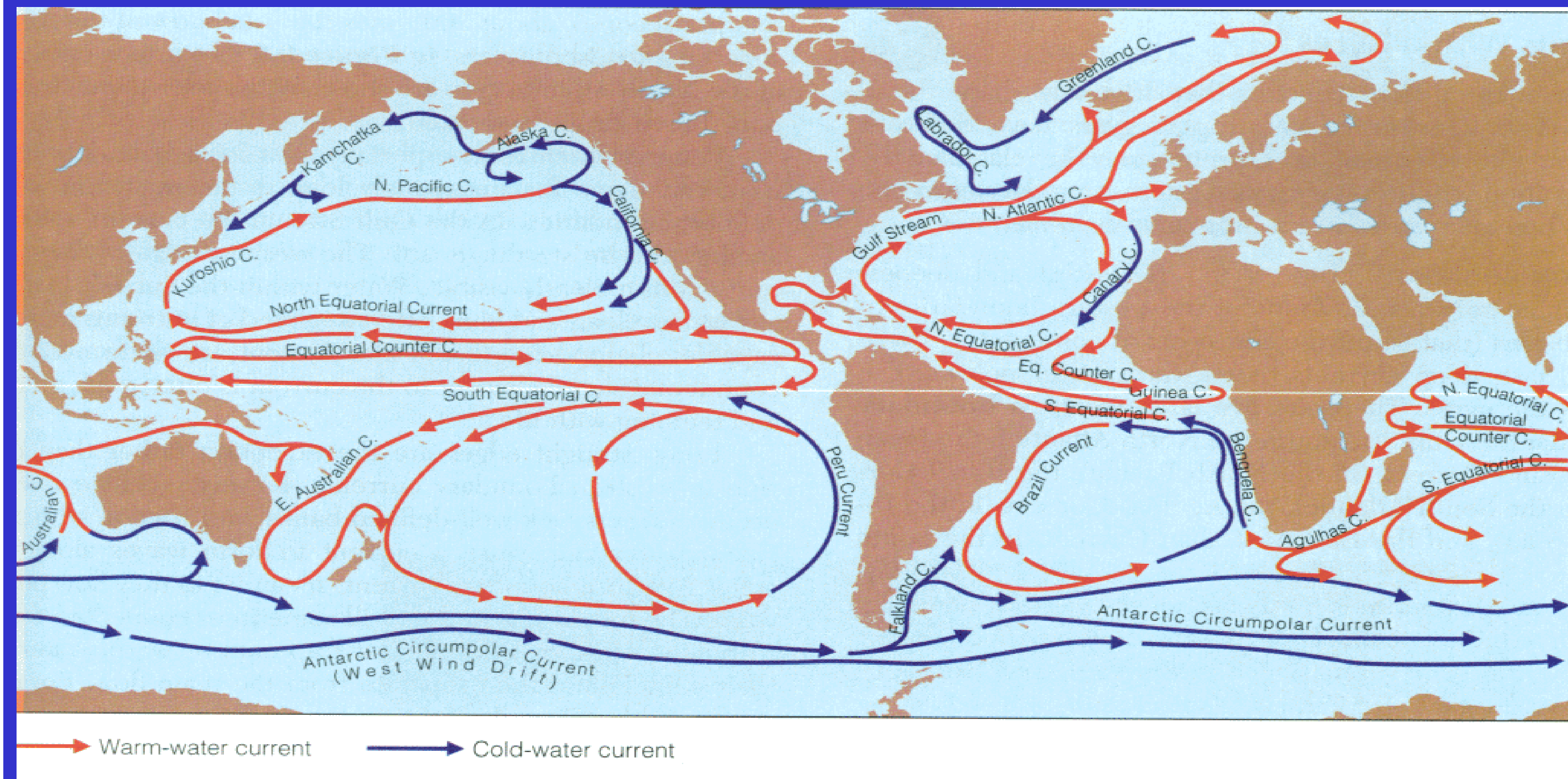


Geostrophic Gyre Currents



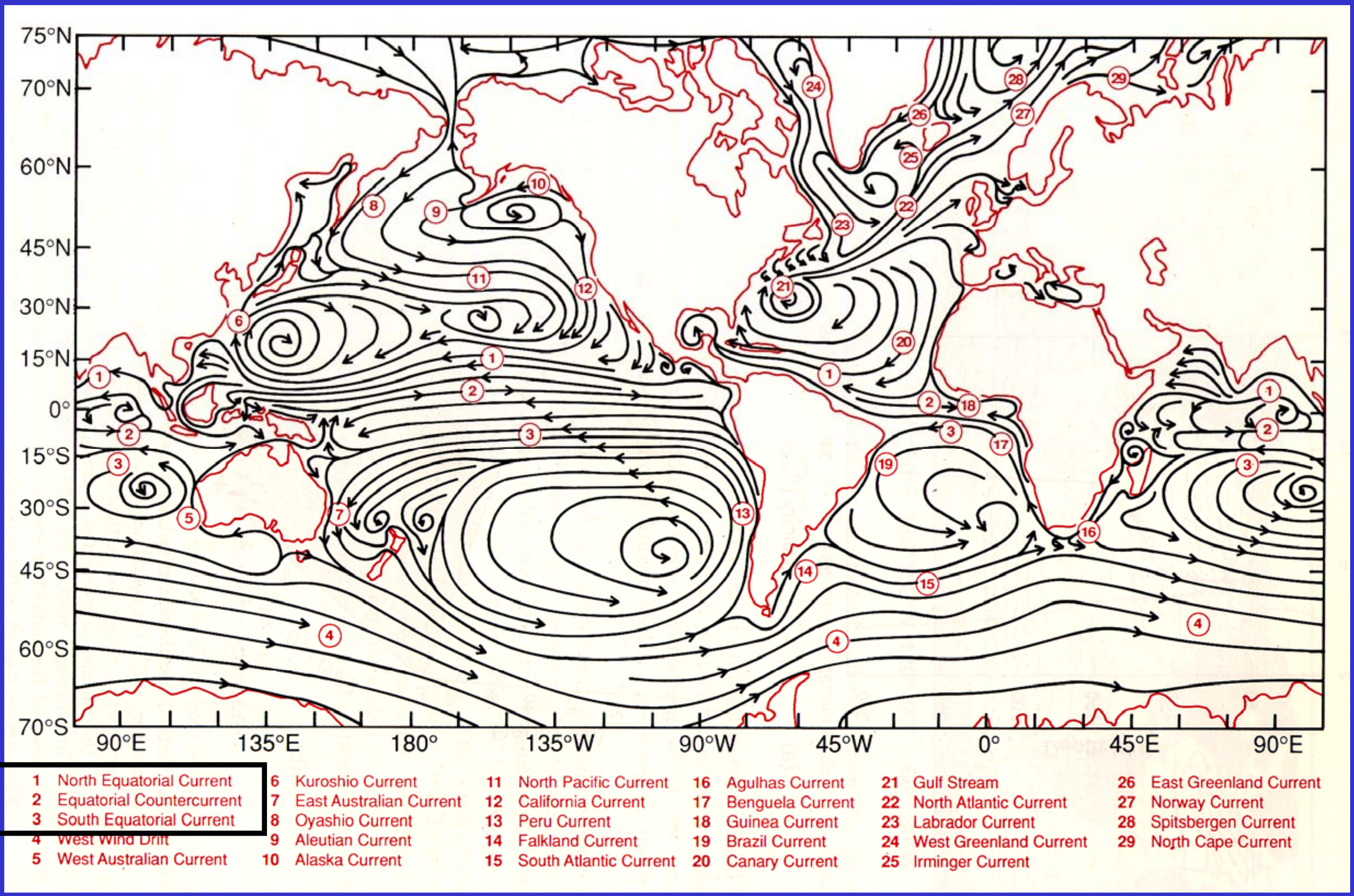
(Figure from *The Earth System*)

Global Surface Currents



ESS200
Prof. Jin-Yi Yu

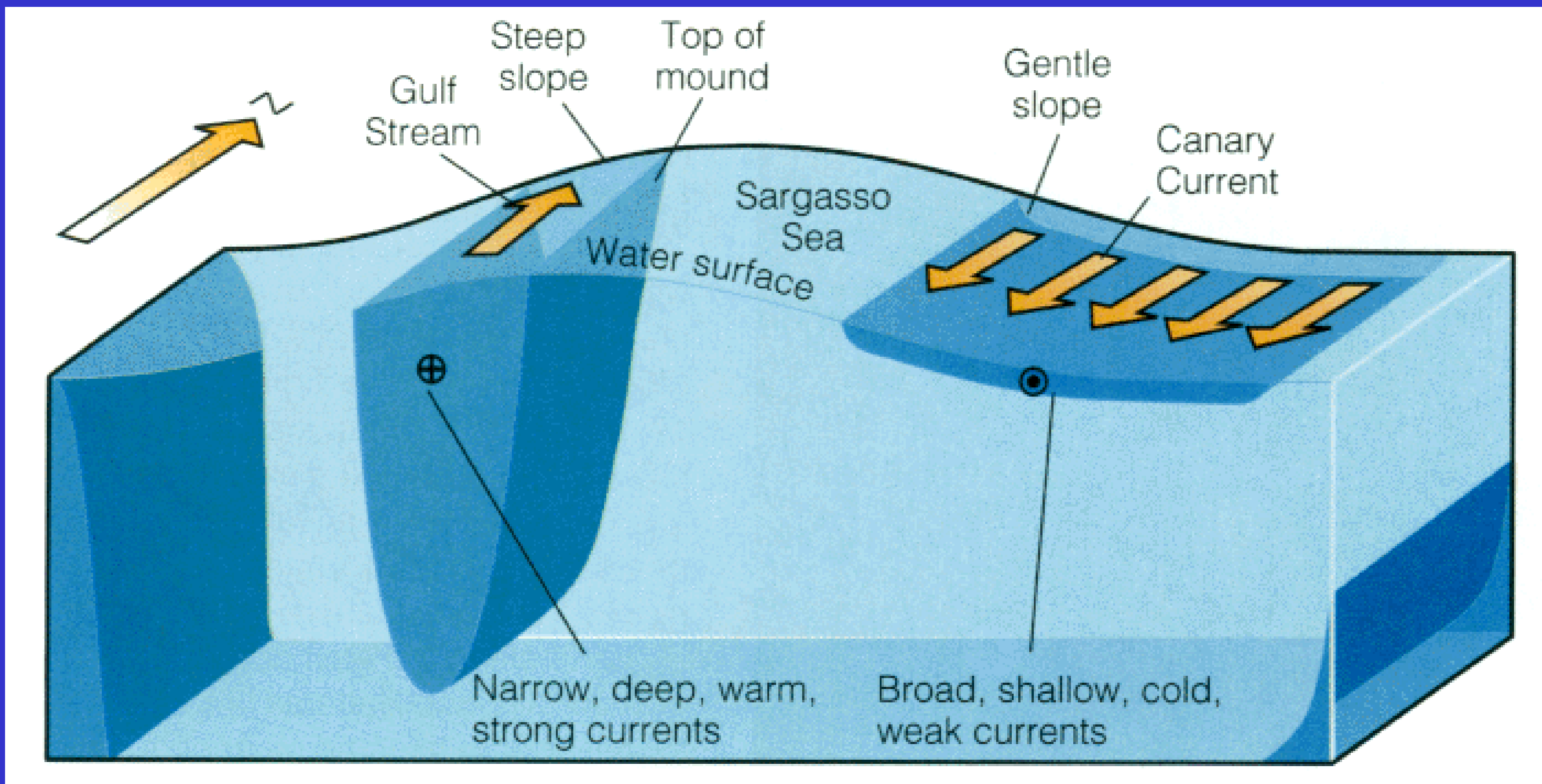
Global Surface Currents



(from *Climate System Modeling*)

ESS200
Prof. Jin-Yi Yu

Step 4: Boundary Currents



(Figure from *Oceanography* by Tom Garrison)

Theories that Explain the Wind-Driven Ocean Circulation

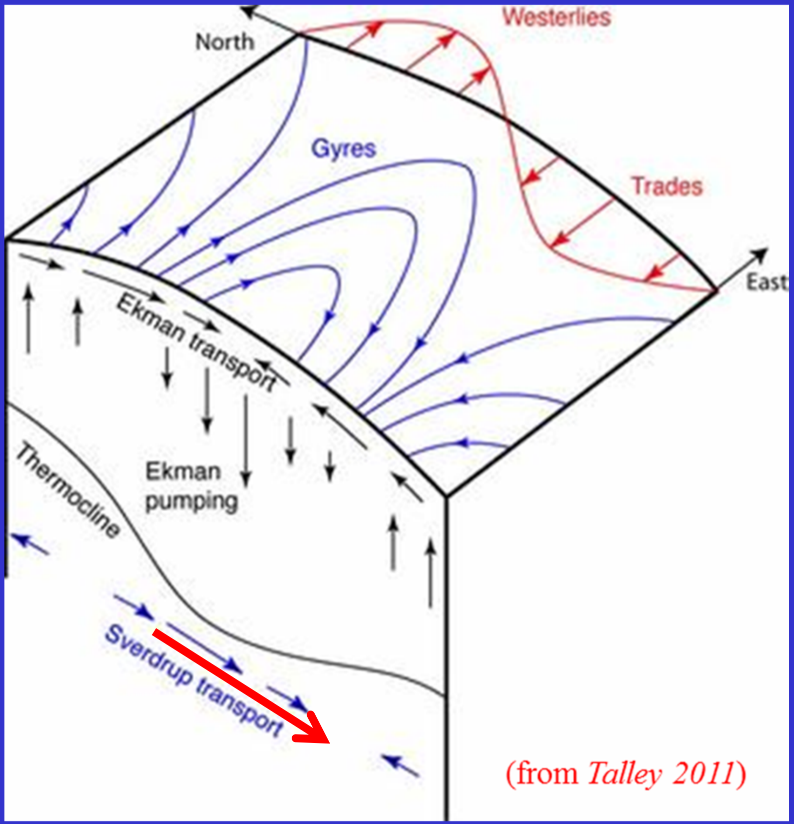
- Harald Sverdrup (1947) showed that the circulation in the upper kilometer or so of the ocean is directly related to the curl of the wind stress if the Coriolis force varies with latitude.
- Henry Stommel (1948) showed that the circulation in oceanic gyres is asymmetric also because the Coriolis force varies with latitude.
- Walter Munk (1950) added eddy viscosity and calculated the circulation of the upper layers of the Pacific.
- Together the three oceanographers laid the foundations for a modern theory of ocean circulation.

(from Robert H. Stewart's book on "*Introduction to Physical Oceanography*")

Sverdrup's Theory of the Oceanic Circulation

$$V = \hat{\mathbf{k}} \cdot \frac{\nabla \times \tau}{\beta}.$$

- The Sverdrup balance, or Sverdrup relation, is a theoretical relationship between the wind stress exerted on the surface of the open ocean and the vertically integrated meridional (north-south) transport of ocean water.



- Positive wind stress curl ➔ Northward mas transport

Negative wind stress curl ➔ Southward mass transport

Sverdrup Transport

$$\frac{\partial p}{\partial x} = f\,\rho\,v + \frac{\partial T_{xz}}{\partial z} \qquad \frac{\partial p}{\partial y} = -f\,\rho\,u + \frac{\partial T_{yz}}{\partial z}$$

$$\frac{\partial P}{\partial x} = \int_{-D}^{0} \frac{\partial p}{\partial x}\,dz, \qquad \frac{\partial P}{\partial y} = \int_{-D}^{0} \frac{\partial p}{\partial y}\,dz,$$

$$M_x \equiv \int_{-D}^{0} \rho\,u(z)\,dz, \qquad M_y \equiv \int_{-D}^{0} \rho\,v(z)\,dz,$$

vertical integration from surface (z=0) to a depth of no motion (z=-D).

$$\frac{\partial P}{\partial x} = f\,M_y + T_x$$

$$\frac{\partial P}{\partial y} = -f\,M_x + T_y$$

d/dy ($\frac{\partial P}{\partial x} = f\,M_y + T_x$) – d/dx ($\frac{\partial P}{\partial y} = -f\,M_x + T_y$) and use $\frac{\partial M_x}{\partial x} + \frac{\partial M_y}{\partial y} = 0$

$$\beta\,M_y = \frac{\partial T_y}{\partial x} - \frac{\partial T_x}{\partial y}$$

$$\beta\,M_y = \mathrm{curl}_z(T)$$

Sverdrup, Geostrophic, and Ekman Transports

$$\boxed{V = \hat{\mathbf{k}} \cdot \frac{\nabla \times \tau}{\beta}.}$$

- Continuity equation for an incompressible flow:

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

- Assume the horizontal flows are geostrophic:

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

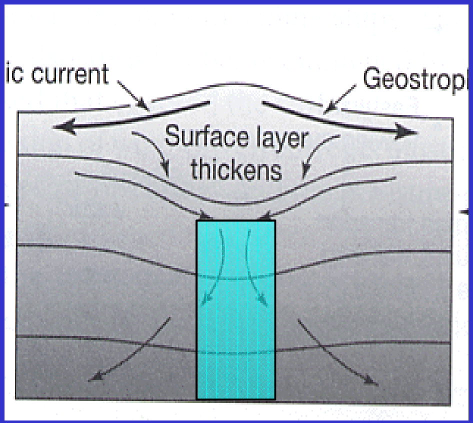
- Replace the geostrophic flow pressure gradients:

$$fu_g = -\frac{1}{\rho}\frac{\partial P}{\partial y}$$

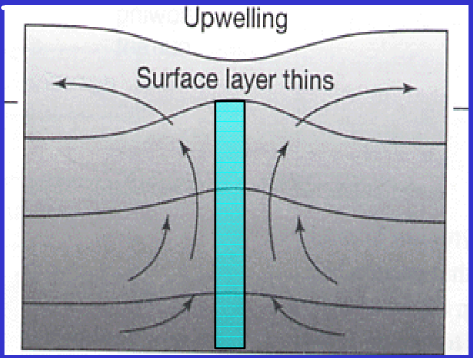
$$fv_g = \frac{1}{\rho}\frac{\partial P}{\partial y}$$

- The continuity equation becomes:

$$\frac{-\beta}{f} v_g + \frac{\partial w}{\partial z} = 0 \;\Rightarrow\; \boxed{\beta v_g = f\frac{\partial w}{\partial z}}$$



Ekman layer pumping
➔ vertical depth decreases
➔ move equatorward to conserve absolute vorticity.



Ekman layer suction
➔ vertical depth increases
➔ move poleward to conserve absolute vorticity.

$$\boxed{(\zeta + f)/h = \eta/h = \text{Const}}$$

Sverdrup, Geostrophic, and Ekman Transports

$$V = \hat{\mathbf{k}} \cdot \frac{\nabla \times \tau}{\beta}.$$

$$V_E = \int_{-\infty}^{0} v_E \; dz \; = -\frac{\tau_x}{\rho_o f}$$

$$U_E = \int_{-\infty}^{0} u_E \; dz \; = \frac{\tau_y}{\rho_o f}; \qquad V_E = \int_{-\infty}^{0} v_E \; dz \; = -\frac{\tau_x}{\rho_o f}$$

- Continuity equation for an incompressible flow:

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

- Assume the horizontal flows are geostrophic:

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

- Replace the geostrophic flow pressure gradients:

$$fu_g = -\frac{1}{\rho}\frac{\partial P}{\partial y}$$

$$fv_g = \frac{1}{\rho}\frac{\partial P}{\partial y}$$

- The continuity equation becomes:

$$\frac{-\beta}{f} v_g + \frac{\partial w}{\partial z} = 0 \quad \Rightarrow \quad \boxed{\beta v_g = f \frac{\partial w}{\partial z}}$$

- Integrate the equation from the bottom of the upper ocean (D_w) to the bottom of the Ekman layer (D_E):

$$\beta \int_{z=-D_w}^{z=-D_E} v \partial z = f\left[w_E - \cancel{w(-D_W)}\right]$$

assume zero

- Ekman pumping (w_E) is related to the convergence of the Ekman transport:

$$w(-D_E) = \frac{\partial}{\partial x}\left(\frac{\tau^y}{\rho f}\right) - \frac{\partial}{\partial y}\left(\frac{\tau^x}{\rho f}\right)$$

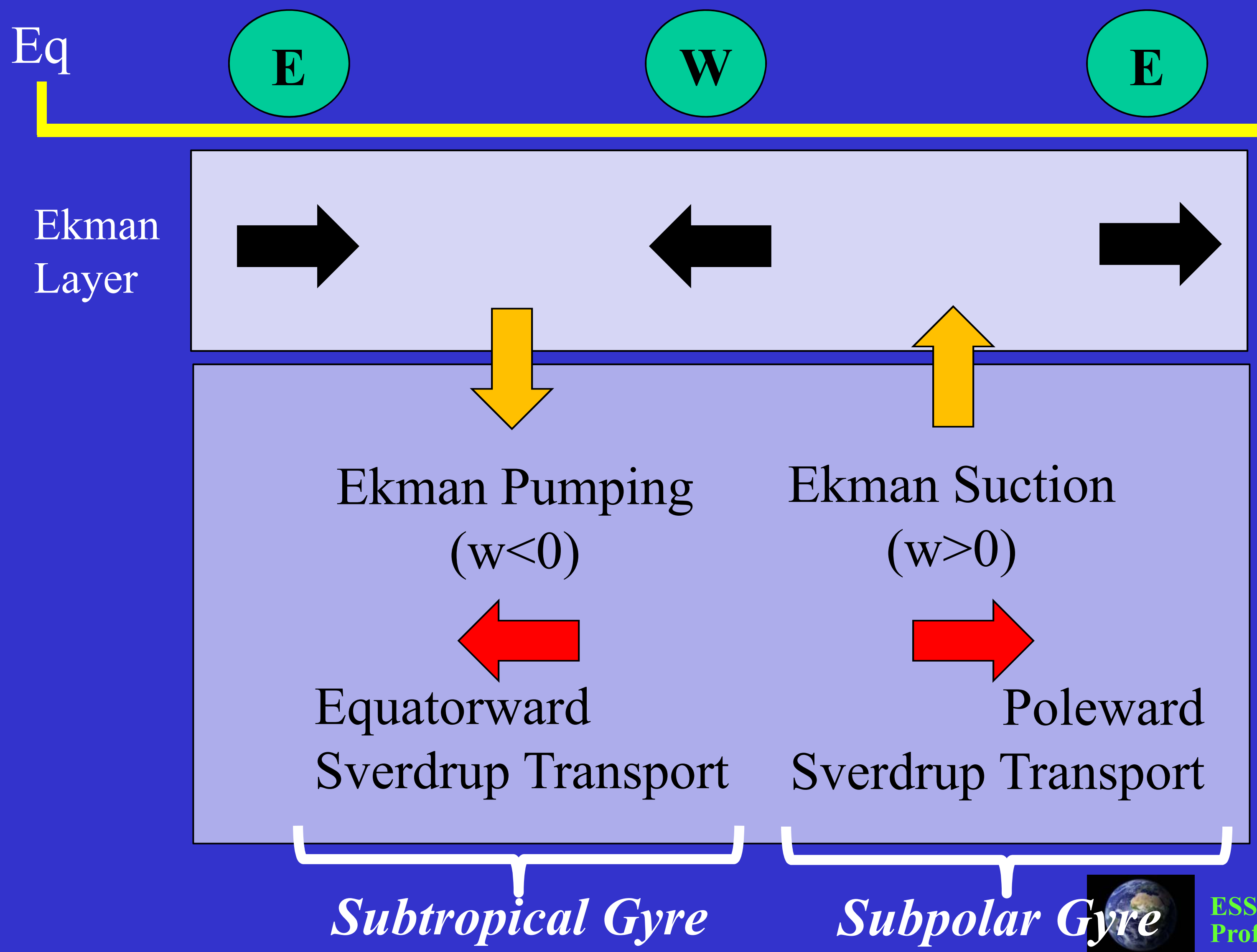
- Therefore, we obtain:

$$\underbrace{\int_{z=-D_w}^{z=-D_E} v \partial z}_{\text{geostrophic transport}} = \underbrace{\frac{1}{\rho\beta}\left(\frac{\partial \tau_W^y}{\partial x} - \frac{\partial \tau_W^x}{\partial y}\right)}_{\text{Sverdrup transport}} + \underbrace{\frac{1}{\rho f}\tau_W^x}_{\text{- (Ekman Transport)}}$$

- Therefore,

Sverdrup transport = Geostrophic transport + Ekman transport

Ekman and Sverdrup Transports



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

$$V = \hat{\mathbf{k}} \cdot \frac{\nabla \times \tau}{\beta}.$$

Ekman Pumping and Thermocline

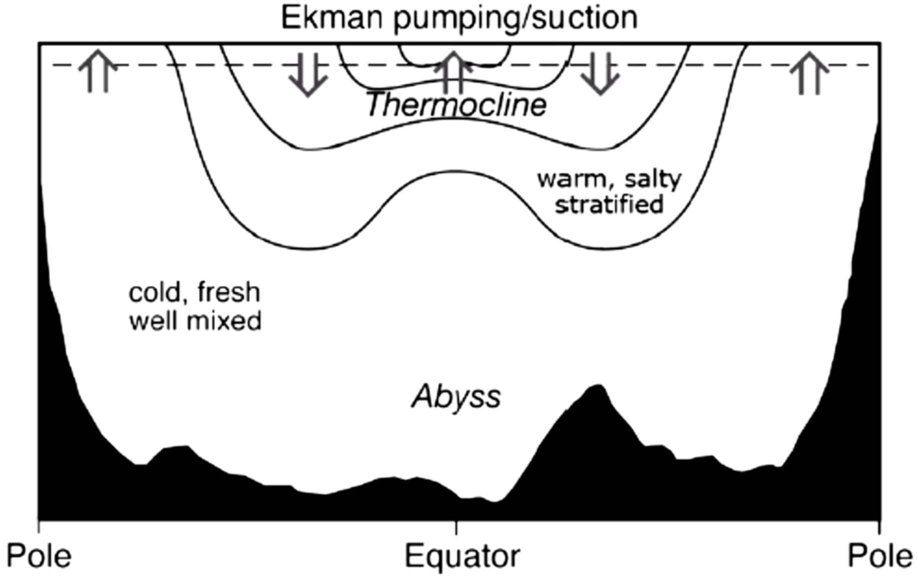


Figure 10.1: The ocean comprises a warm, salty, stratified lens of fluid, the thermocline, circulating on top of a cold, fresh, relatively well mixed, abyss. The surface layer, above the horizontal dotted line at a depth of about 100 m, is driven directly by the wind. The thermocline below is brought in to motion through a pattern of vertical velocity driven by the wind (Ekman pumping and suction) which induces flow in the ocean beneath.

(from John Marshall and R. Alan Plumb's Atmosphere, Ocean and Climate Dynamics: An Introductory Text)

Stommel's Theory of Western Boundary Currents

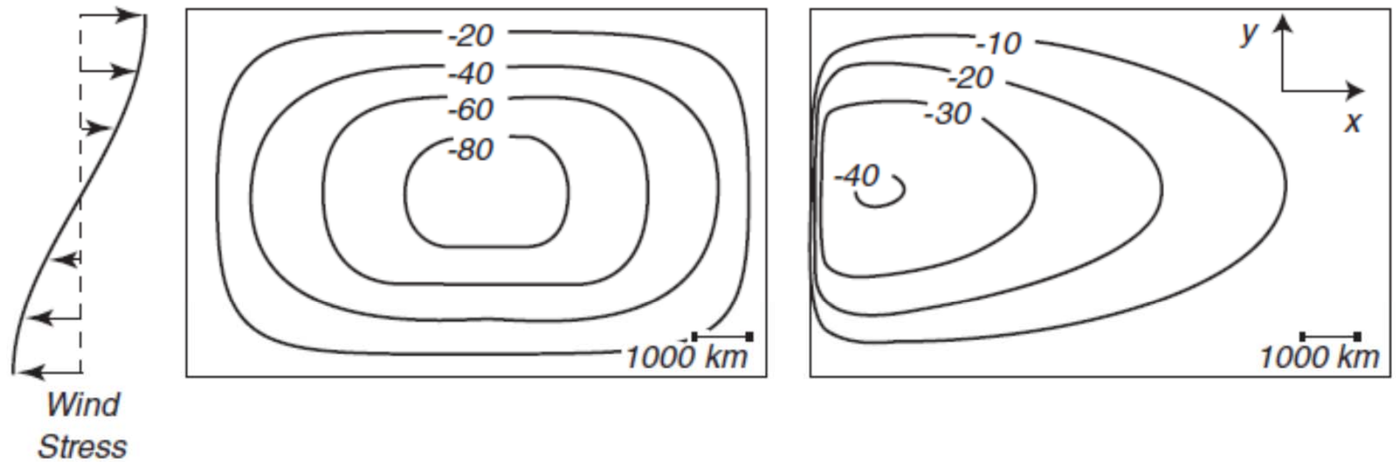


Figure 11.5 Stream function for flow in a basin as calculated by Stommel (1948). **Left:** Flow for non-rotating basin or flow for a basin with constant rotation. **Right:** Flow when rotation varies linearly with y.

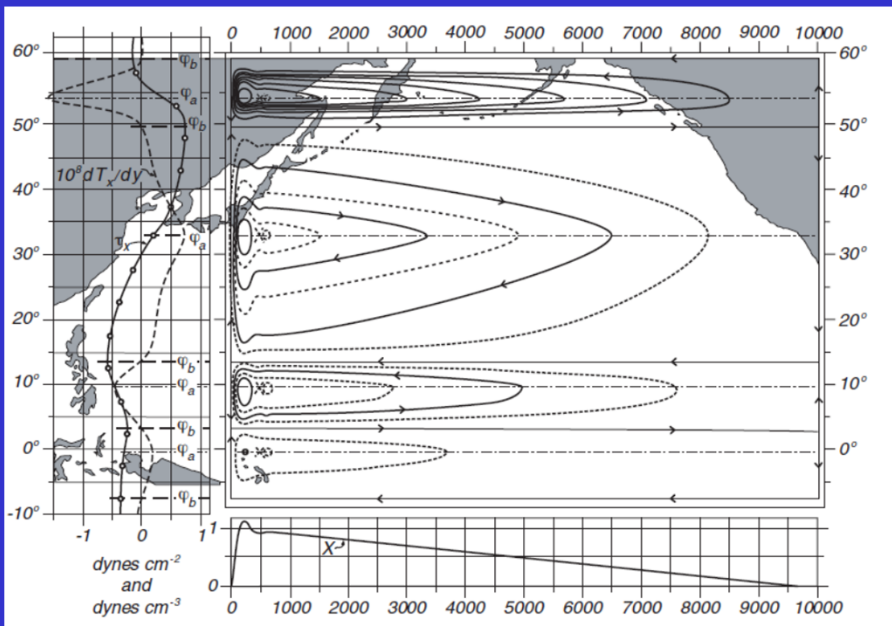
Stommel's Theory added bottom friction into the same equations used by Svedrup.

$$\frac{\partial p}{\partial x} = f\,\rho\,v + \frac{\partial T_{xz}}{\partial z} \qquad \frac{\partial p}{\partial y} = -f\,\rho\,u + \frac{\partial T_{yz}}{\partial z}$$

$$\underbrace{\begin{aligned} \left(A_z \frac{\partial u}{\partial z}\right)_0 &= -T_x = -F\cos(\pi\,y/b) \\ \left(A_z \frac{\partial v}{\partial z}\right)_0 &= -T_y = 0 \end{aligned}}_{\text{surface stress}} \qquad \underbrace{\begin{aligned} \left(A_z \frac{\partial u}{\partial z}\right)_D &= -R\,u \\ \left(A_z \frac{\partial v}{\partial z}\right)_D &= -R\,v \end{aligned}}_{\text{bottom stress}}$$

(from Robert H. Stewart's book on "*Introduction to Physical Oceanography*")

Munk's Theory of Western Boundary Currents



$$\frac{1}{\rho}\frac{\partial p}{\partial x} = f\,v + \frac{\partial}{\partial z}\left(A_z\frac{\partial u}{\partial z}\right) + A_H\frac{\partial^2 u}{\partial x^2} + A_H\frac{\partial^2 u}{\partial y^2}$$

$$\frac{1}{\rho}\frac{\partial p}{\partial y} = -f\,u + \underbrace{\frac{\partial}{\partial z}\left(A_z\frac{\partial v}{\partial z}\right)}_{\text{surface stress}} + \underbrace{A_H\frac{\partial^2 v}{\partial x^2} + A_H\frac{\partial^2 v}{\partial y^2}}_{\text{lateral friction}}$$

$$\underbrace{A_H\nabla^4\Psi}_{\text{Friction}} - \underbrace{\beta\frac{\partial\Psi}{\partial x} = -\,\mathrm{curl}_z T}_{\text{Sverdrup Balance}}$$

mass-transport stream function Ψ

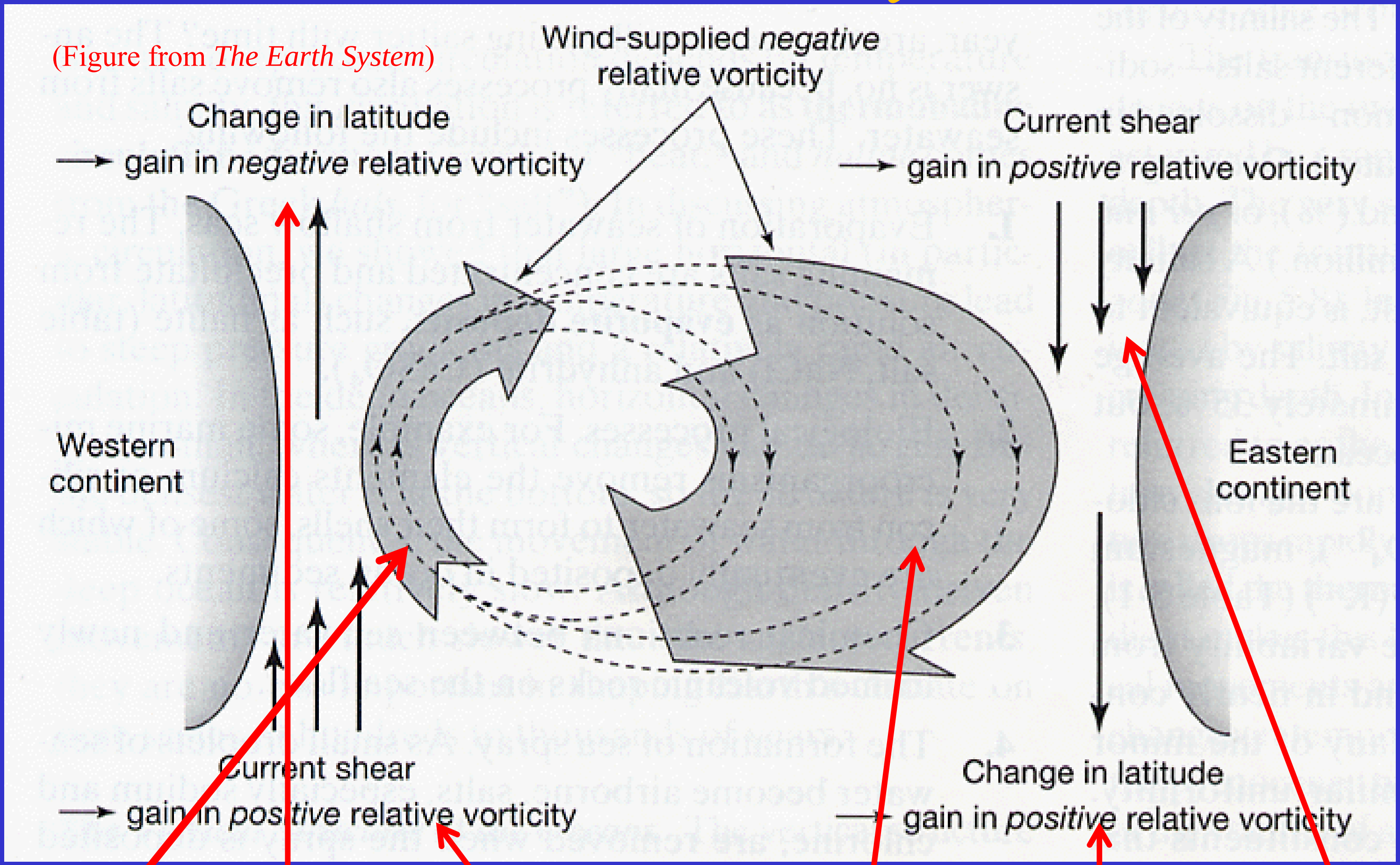
$$M_x \equiv \frac{\partial\Psi}{\partial y}, \qquad M_y \equiv -\frac{\partial\Psi}{\partial x}$$

- Munk (1950) built upon Sverdrup's theory, adding lateral eddy viscosity, to obtain a solution for the circulation within an ocean basin.
- To simplify the equations, Munk used the mass-transport stream function.

(from Robert H. Stewart's book on "*Introduction to Physical Oceanography*")

Why Strong Boundary Currents?

A Potential Vorticity View



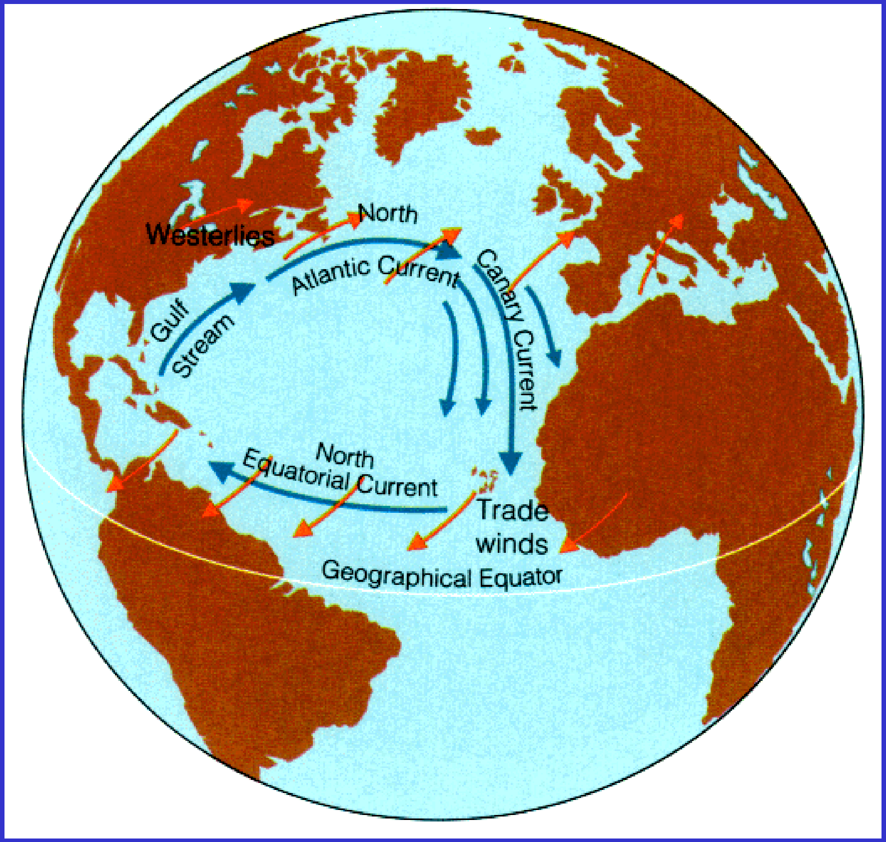
Goal: **Maintain the "steady state" of the negative vorticity induced by wind stress curve**

$\xi-$ = $\xi-$ **plus** $\xi+$ $\qquad$ $\xi-$ = $\xi+$ **plus** $\xi+$

friction has to be big ➔ strong boundary current

Characteristics of the Gyres

(Figure from *Oceanography* by Tom Garrison)



Volume transport unit:

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

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

- **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 polarward (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

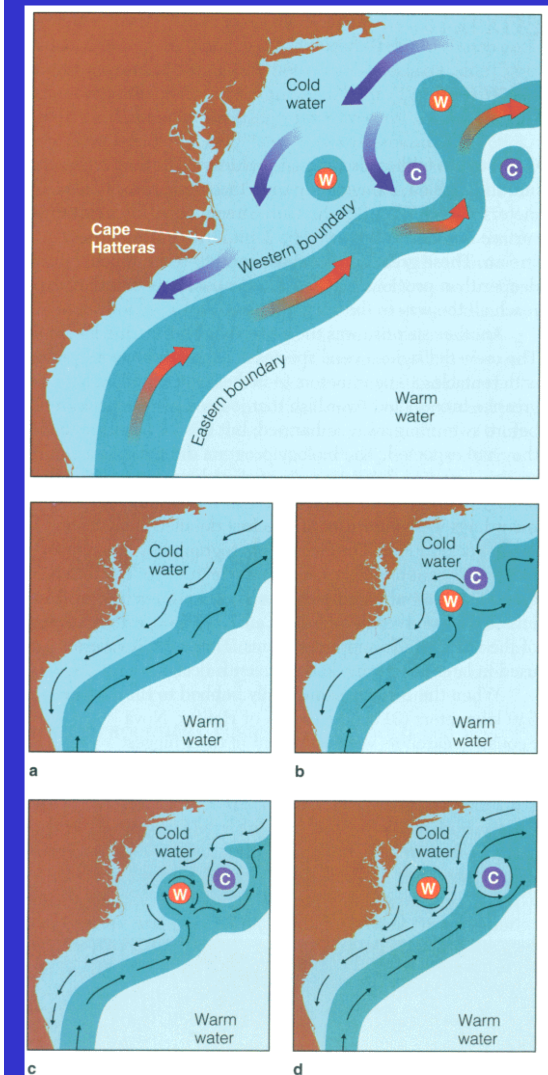
the moderately shallow and broad westward current (transport ~ 30 Sv)

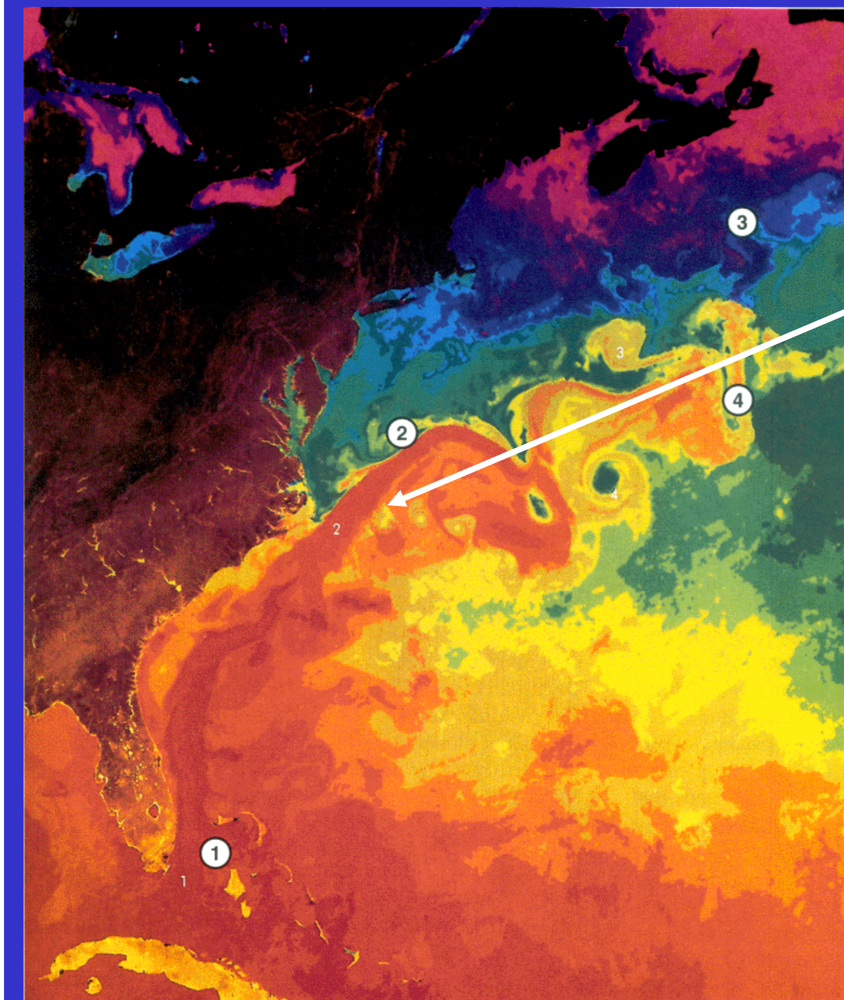
Westerly-driven current

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

ESS200
Prof. Jin-Yi Yu

Gulf Stream





A river of current

Jet stream in the ocean

- Speed = 2 m/sec
- Depth = 450 m
- Width = 70 Km
- Color: clear and blue

(Figure from *Oceanography* by Tom Garrison)

Equatorial Current System

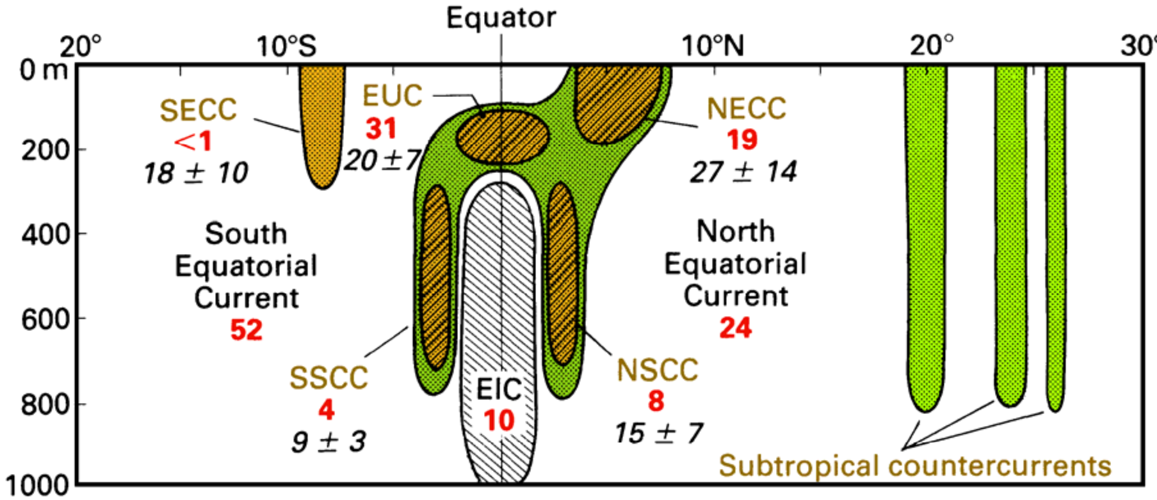


Fig. 8.7. A sketch of the structure of the equatorial current system in the central Pacific Ocean (170°W). Eastward flow is coloured. All westward flow north of 5°N constitutes the North Equatorial Current, westward flow south of 5°N outside the EIC represents the South Equatorial Current. EUC = Equatorial Undercurrent, EIC = Equatorial Intermediate Current, NECC and SECC = North and South Equatorial Countercurrents, NSCC and SSCC = North and South Subsurface Countercurrents. Transports in Sverdrups are given for 155°W (bold figures; based on observations from April 1979 - March 1980) and 165°E (italics, based on January 1984 - June 1986).

- The ***Equatorial Counter Current***, which flows towards the east, is a partial return of water carried westward by the North and South Equatorial currents.

Equatorial Under Current

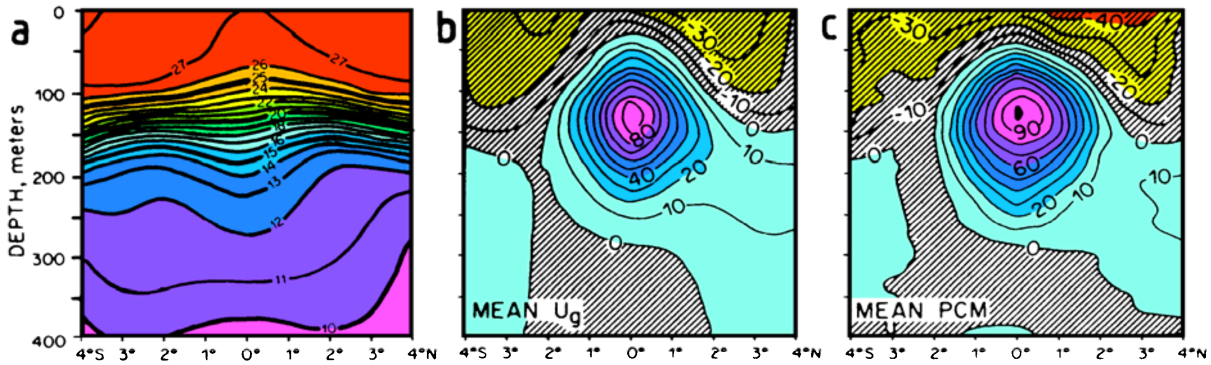
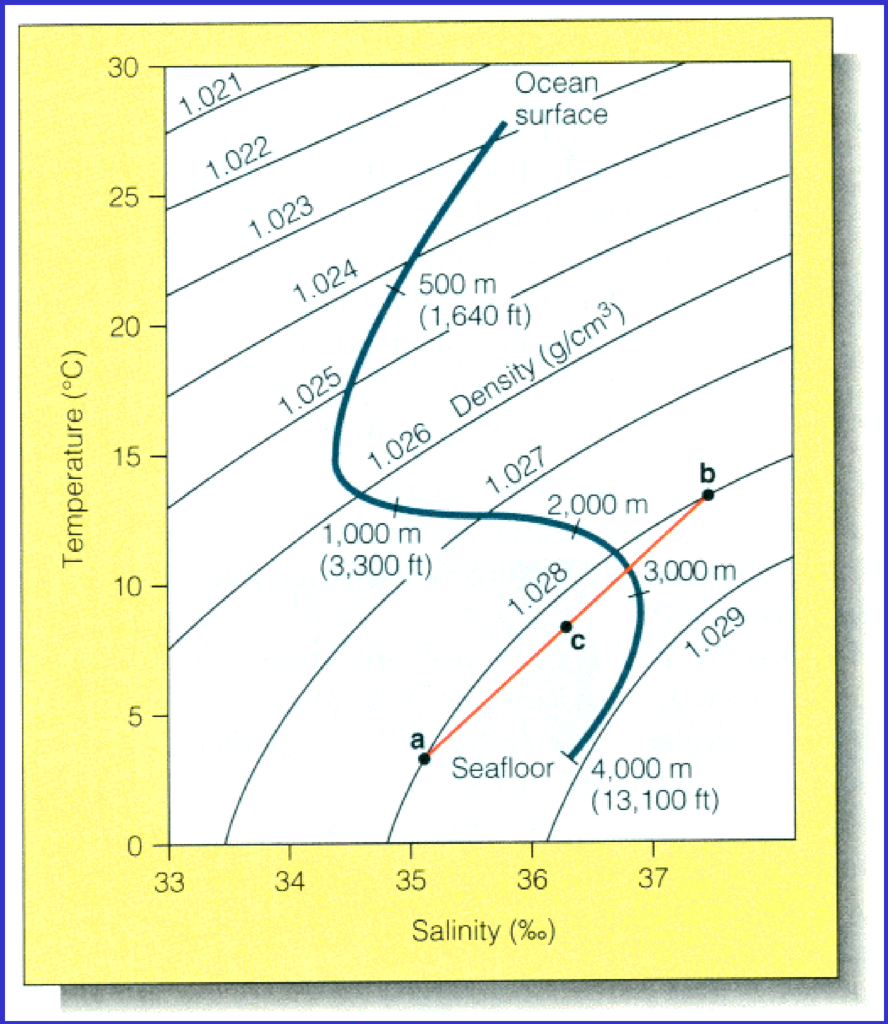


Fig. 8.8. The Equatorial Undercurrent during February 1979 - June 1980 near 155°W. (a) Mean temperature (°C), (b) mean geostrophic zonal velocity (10^{-2} m s^{-1})X, (c) mean observed zonal velocity (10^{-2} m s^{-1}). Note the spreading of the isotherms at the equator. From Lukas and Firing (1984).

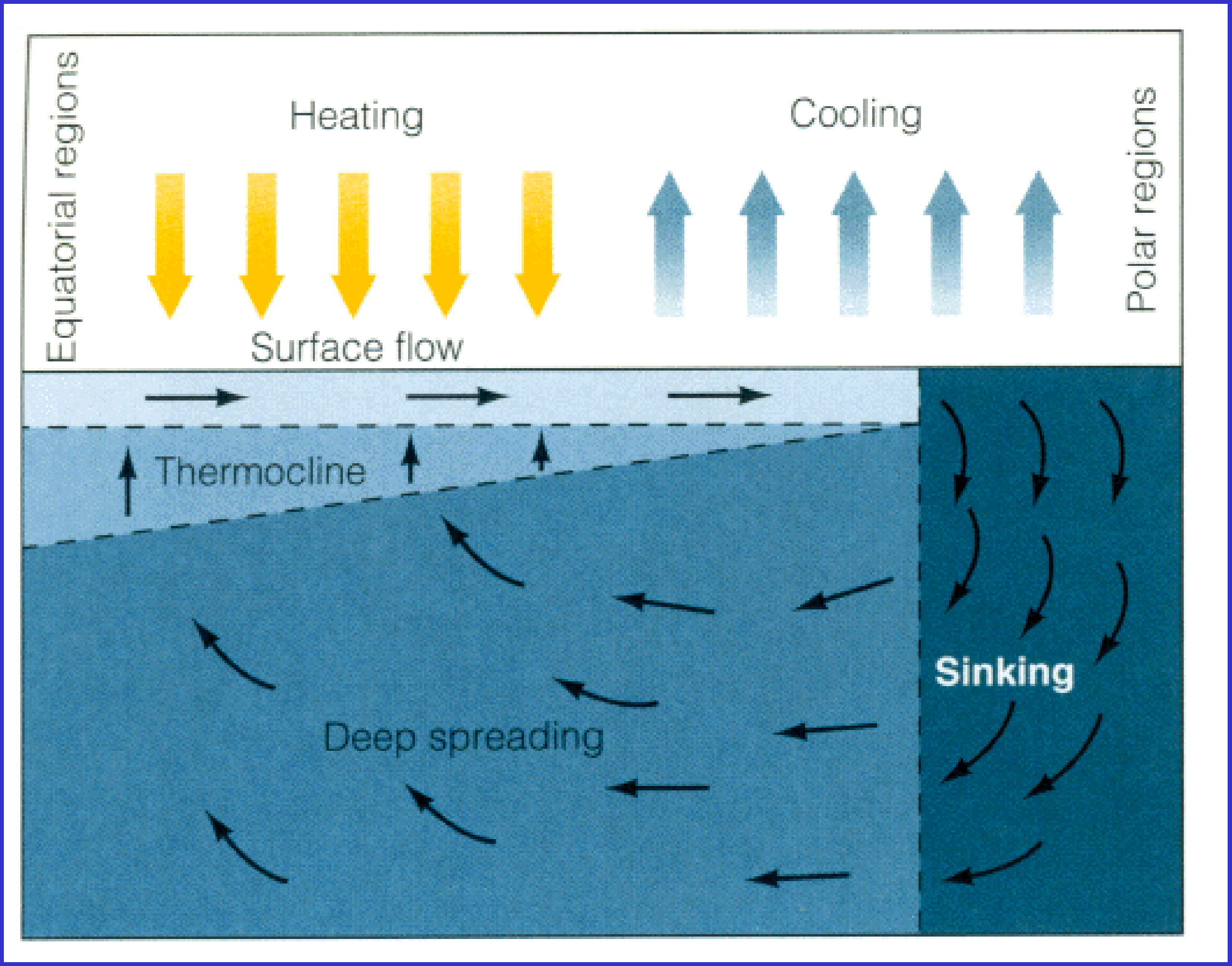
- The most prominent of all eastward flows is the ***Equatorial Undercurrent*** (EUC).
- It is a swift flowing ribbon of water extending over a distance of more than 14,000 km along the equator with a thickness of only 200 m and a width of at most 400 km.
- The current core is found at 200 m depth in the west, rises to 40 m or less in the east and shows typical speeds of up to 1.5 m s-1.

Deep Ocean Circulation: Density-Driven



(Figure from *Oceanography* by Tom Garrison)

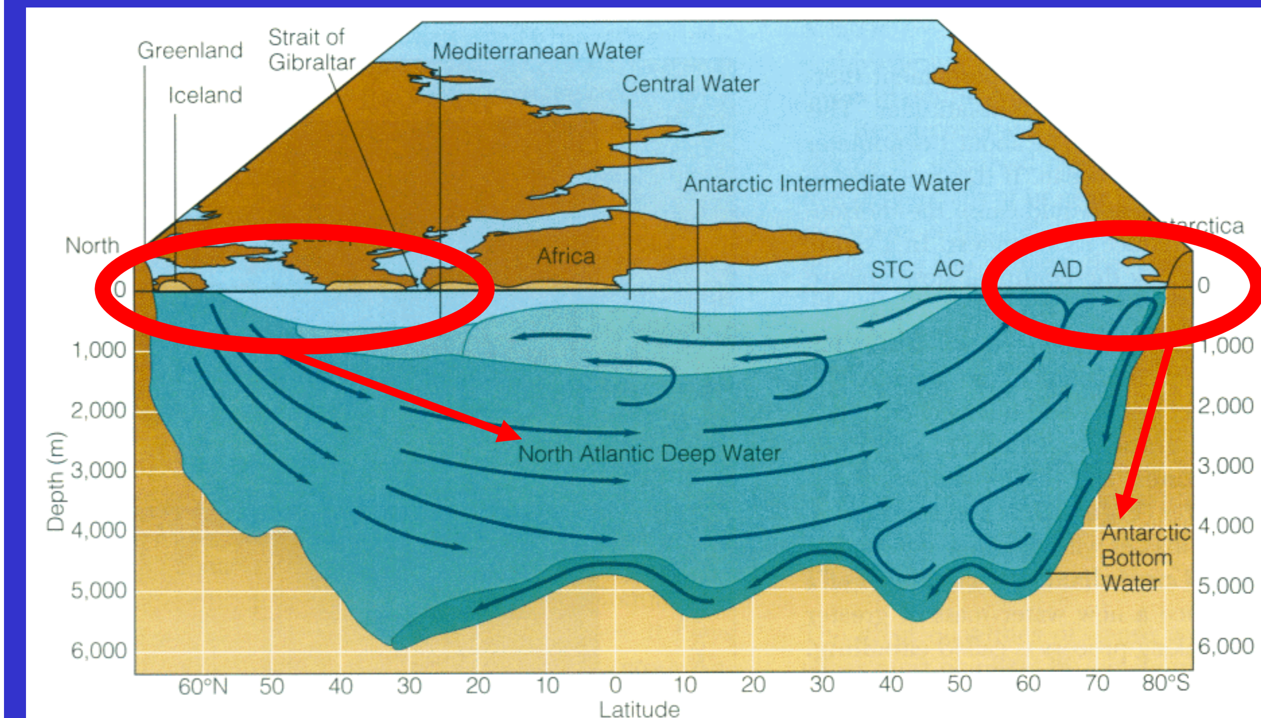
Thermohaline Circulation



(Figure from *Oceanography* by Tom Garrison)

ESS200
Prof. Jin-Yi Yu

Two Regions of Deep Water Formation



(Figure from *Oceanography* by Tom Garrison)

- ❑ Antarctic Bottom Water

  Salinity = 34.65‰

  Temperature = -0.5˚C

  Density = 1.0279 g/cm^3

  Formed at Weddell Sea

  Related to ice formation

  During Winter

- ❑ North Atlantic Deep Water

  Due to winter cooling and evaporation.

ESS200
Prof. Jin-Yi Yu

Two Processes to Increase Salinity in High Latitudes

- **Evaporation**: Extremely cold, dry winter air enhances evaporation from the relatively warm ocean → increase salinity in the ocean.
- **Formation of Sea Ice**: When sea ice forms, salts are left in the ocean → increase salinity

Ocean Water Mass

- We define a water mass as a body of water with a common formation history.
- An example of water mass formation is the cooling of surface water near the Antarctic continent, particularly in the Weddell Sea, which increases the density and causes the water to sink to great depth.
- All water which originates from this process shares the same formation history and is called Antarctic Bottom Water.
- It is found in all oceans well beyond its formation region, extending even into the northern hemisphere.
- Common names of known water masses usually relate to their major area of residence.
- Unfortunately, this can give rise to ambiguity since the same name may be used for a well defined water mass or simply for water found in a certain region.
- To avoid this confusion we adopt the convention that water masses are always identified by capitals.
- For example, "Bottom Water" can stand for Antarctic, Arctic, or other Bottom Water but always refers to a water mass, while water found at the bottom of an oceanic region may be referred to as "bottom water" without implying that it is a known and well defined water mass.

(from Regional Oceanography)

Ocean Water Mass

Surface Water
to a depth of about 200 meters

Central Water
to the bottom of the main thermocline

Intermediate Water
to about 1500 meters

Deep Water
below intermediate water but not in contact with the bottom

Bottom Water
in contact with sea floor

- Ocean water masses possess distinct, identifiable properties and don't often mix easily when they meet.
- In stead, they usually flow above or below each other.
- Ocean water mass can retain their identity for great distance and long periods of time.
- Oceanographers name water masses according to their relative position.

Ocean Water Mass

The properties of Central Water in the Coral Sea correspond closely to those in its formation region, indicating that little mixing with other water masses occurred along its way.

The intermediate and deep water masses are not present with their original T-S values; their properties are modified by mixing with water above and below.

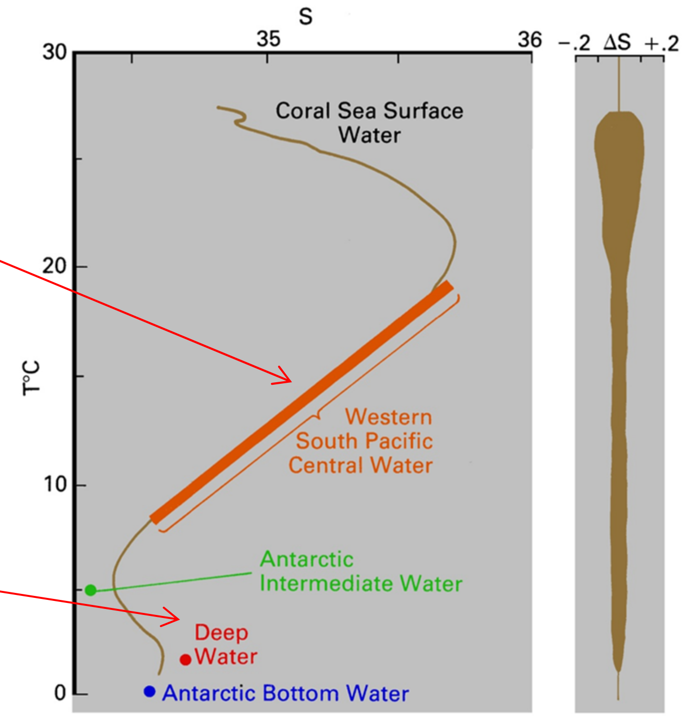
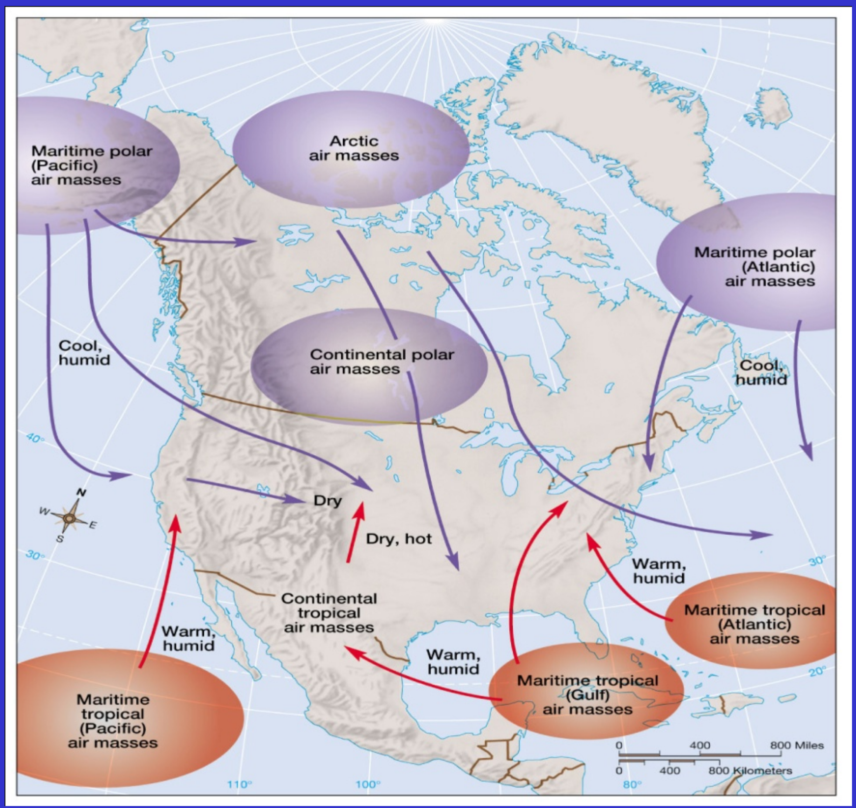


Fig. 5.1. Mean T-S diagram and standard deviation ΔS of salinity (for given temperatures) in the eastern Coral Sea, in comparison to water mass definitions in the south Pacific Ocean. Large dots and the heavy line indicate water mass properties in the formation regions, which for all but Surface Water are located far outside the Coral Sea. The standard deviation was determined by comparing stations in the region with a space average and does not include variability in time. Similar standard deviations can be derived for temperature and other properties. Based on Tomczak and Hao (1989).

- ❑ As the water masses spread across the ocean they mix, and several water masses are usually present at an oceanic location.
- ❑ It is possible to determine the percentage contribution of all water masses to a given water sample, because the water mass elements retain their properties, in particular their potential temperature and their salinity, when leaving the formation region.
- ❑ Water masses can therefore be identified by plotting temperature against salinity in a so-called T-S diagram.

(from Regional Oceanography)

Five Types of Air Masses



- Theoretically, there should be 6 types of air masses (2 moisture types x 3 temperature types).
- But mA-type (maritime Arctic) does not exist.
- cA: continental Arctic

  cP: continental Polar

  cT: continental Tropical

  mP: maritime Polar

  mT: maritime Tropical

Distribution of Ekman Pumping

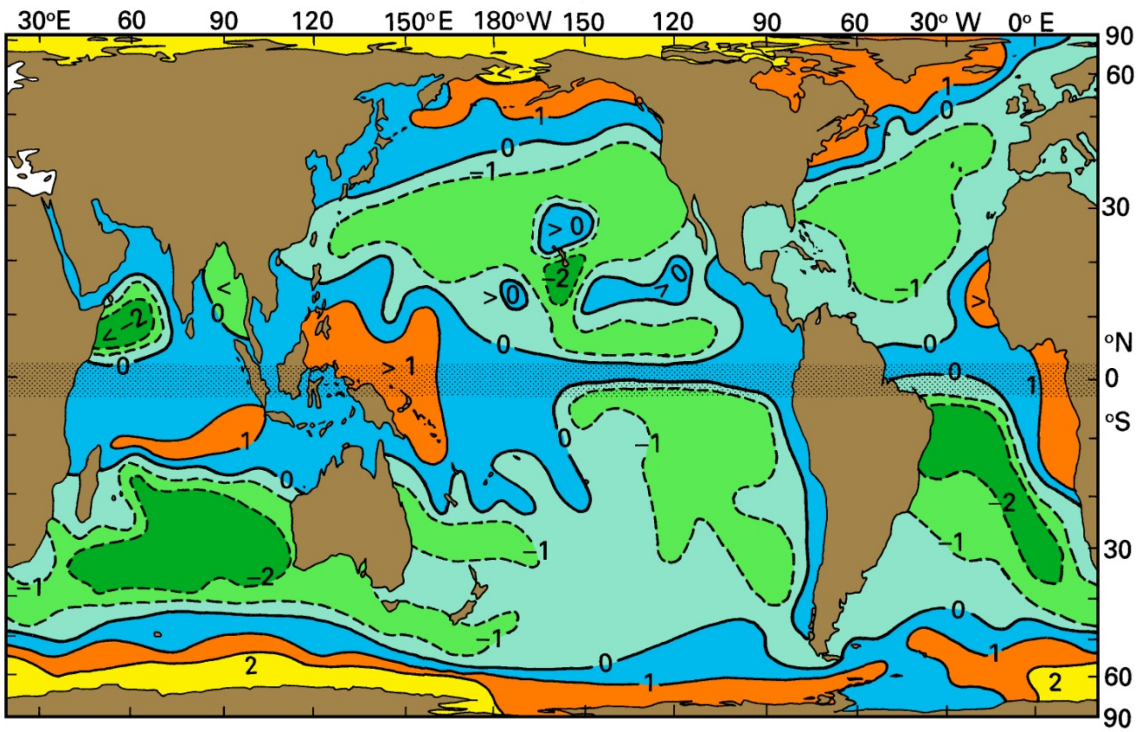


Fig. 4.3. Annual mean distribution of curl($\boldsymbol{\tau}/f$), or Ekman pumping, calculated from the distribution of Fig. 1.4 (10^{-3} kg m^2 s^{-1}). Positive numbers indicate upwelling. In the equatorial region (2°N - 2°S, shaded) curl($\boldsymbol{\tau}/f$) is not defined; the distribution in this region is inferred from the dynamical arguments of Fig. 4.1 and is not quantitative.

(from Regional Oceanography)

Subduction

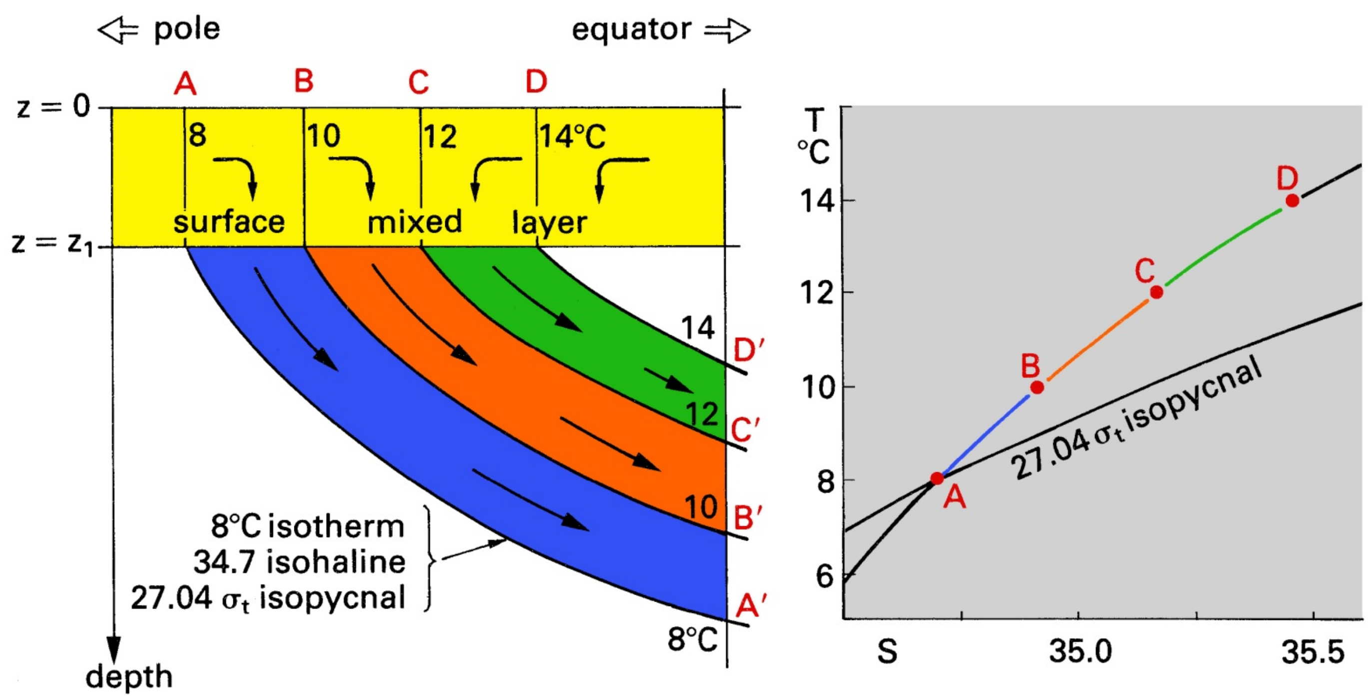


Fig. 5.3. Sketch of water mass formation by subduction in the Subtropical Convergence. The T-S diagram shows both the meridional variation of temperature and salinity between stations *A* and *D*, and the vertical variation equatorward of station *D* from the surface down along the line *A'B'C'D'*. For more detail, see text.

(from Regional Oceanography)

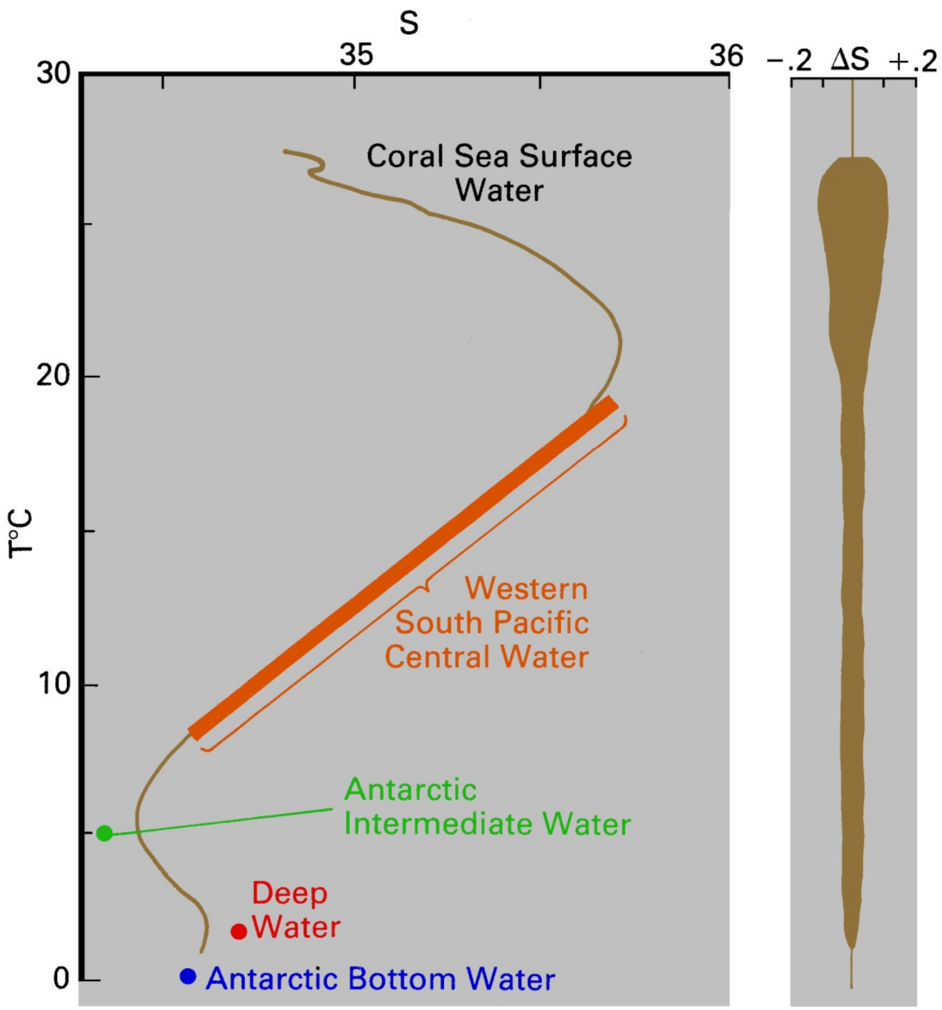
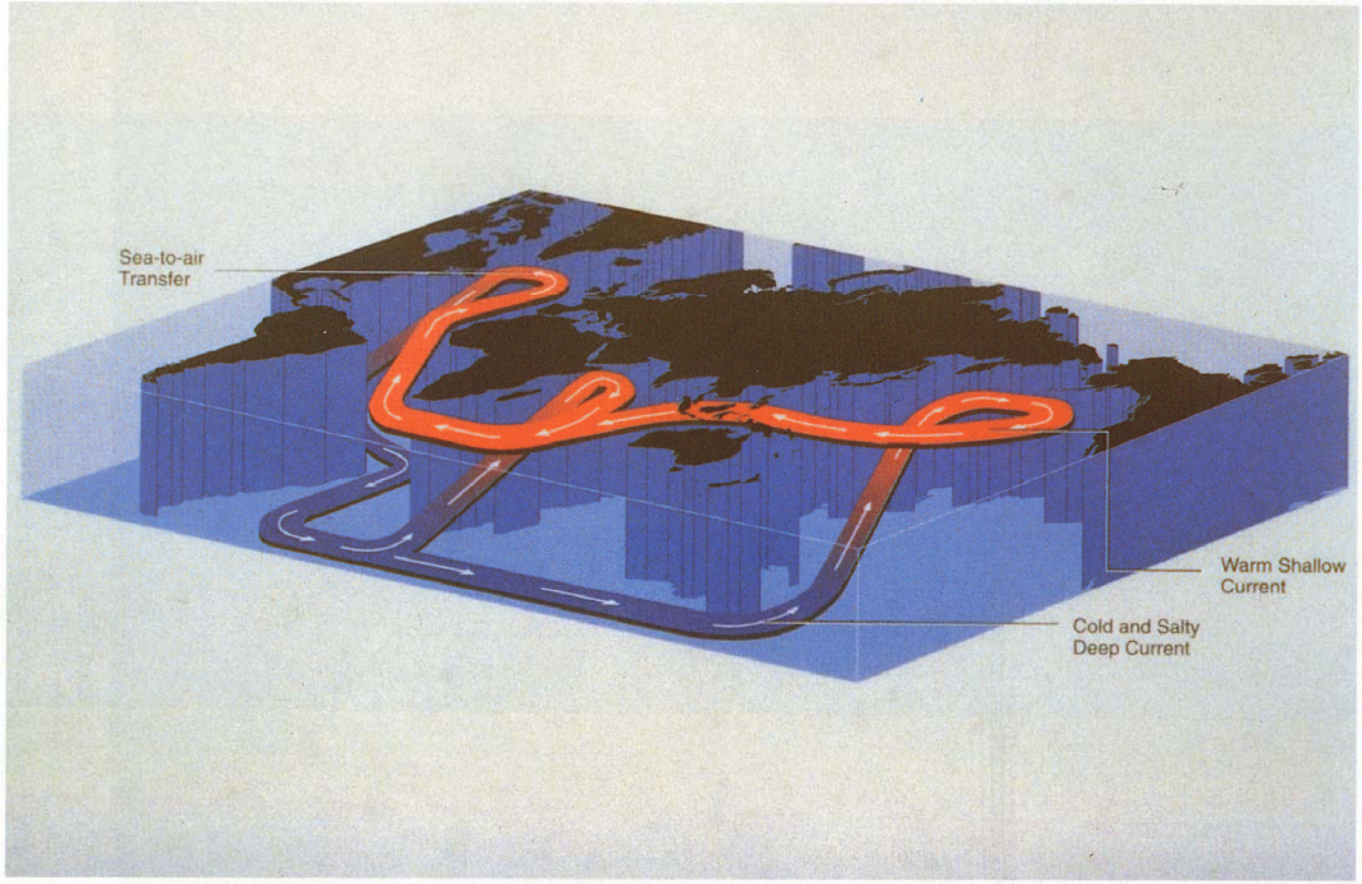


Fig. 5.1. Mean T-S diagram and standard deviation ΔS of salinity (for given temperatures) in the eastern Coral Sea, in comparison to water mass definitions in the south Pacific Ocean. Large dots and the heavy line indicate water mass properties in the formation regions, which for all but Surface Water are located far outside the Coral Sea. The standard deviation was determined by comparing stations in the region with a space average and does not include variability in time. Similar standard deviations can be derived for temperature and other properties. Based on Tomczak and Hao (1989).

Thermohaline Conveyor Belt



(Figure from *Climate System Modeling*)

Thermohaline Circulation

- Thermo → temperature
- Haline → salinity

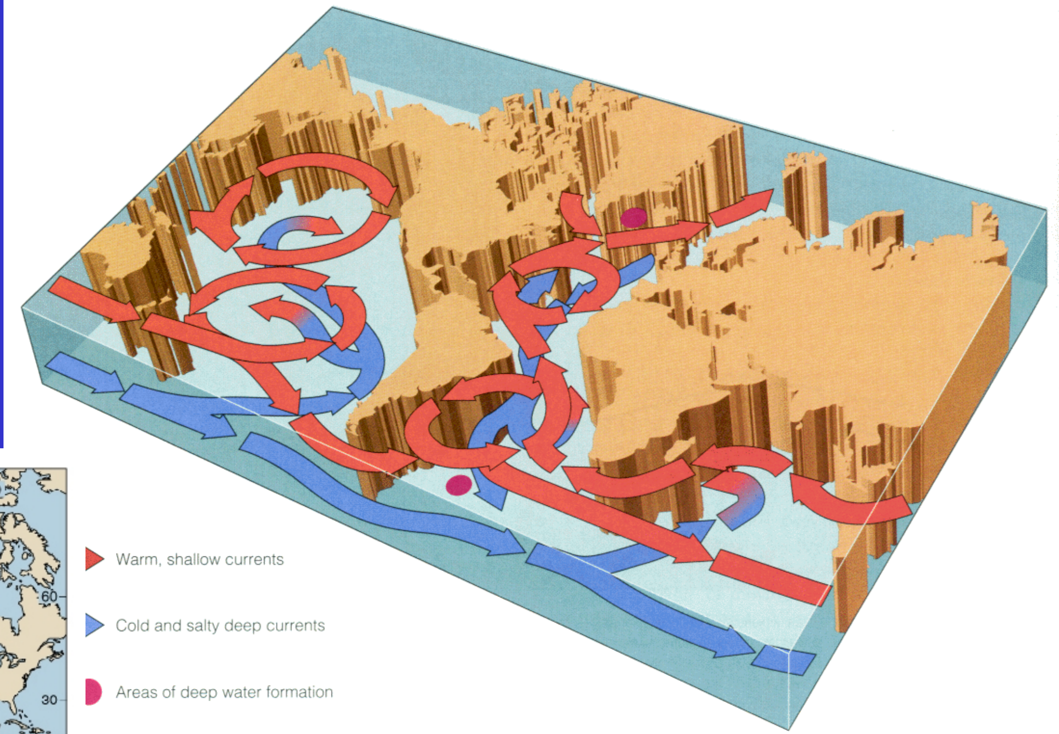
Density-Driven Circulation

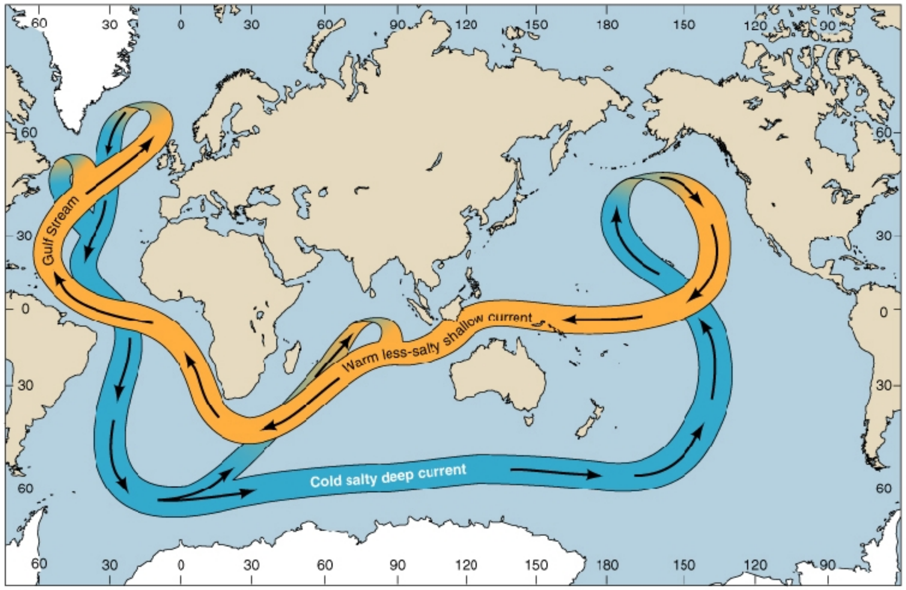
Cold and salty waters go down
Warm and fresh waters go up

Thermohaline Conveyor Belt

❑ Typical speed for deep ocean current: 0.03-0.06 km/hour.

❑ Antarctic Bottom Water takes some 250-1000 years to travel to North Atlantic and Pacific.





(Figure from *Oceanography* by Tom Garrison)

ESS200
Prof. Jin-Yi Yu

It Takes ~1000 Years for Deep Ocean Waters to Travel Around…

- ❑ If we date a water parcel from the time that it leaves the surface and sink into the deep ocean

➔ Then the youngest water is in the deep north Atlantic, and the oldest water is in the deep northern Pacific, where its age is estimated to be 1000 year.

The Most Unpolluted Waters are..

the waters in the deep northern Pacific.

- The man-released CFC and the chemical tritium and C^{14}, which were released through atmospheric atomic bomb test in the 1950s and 1960s, entered the deep ocean in the northern Atlantic and are still moving southward slowly.

- Those pollutions just cross the equator in the Atlantic ➔ They have not reached the deep northern Pacific yet!!

Global Warming and Thermohaline Circulation

- ***If the warming is slow***

  The salinity is high enough to still produce a thermohaline circulation

→ The circulation will transfer the heat to deep ocean

→ The warming in the atmosphere will be deferred.

- ***If the warming is fast***

  Surface ocean becomes so warm (low water density)

→ No more thermohalione circulation

→ The rate of global warming in the atmosphere will increase.