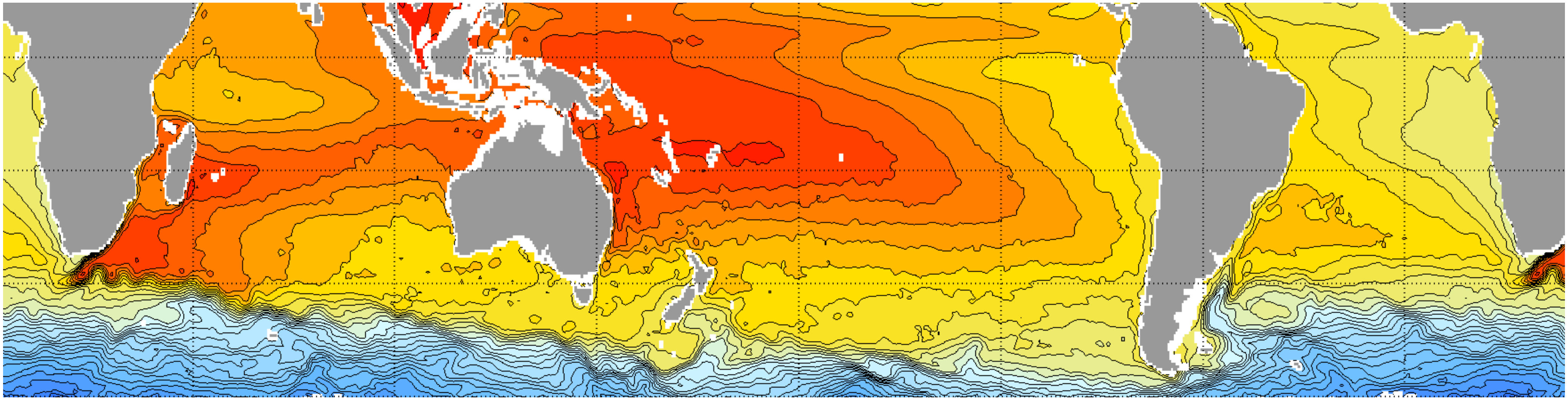
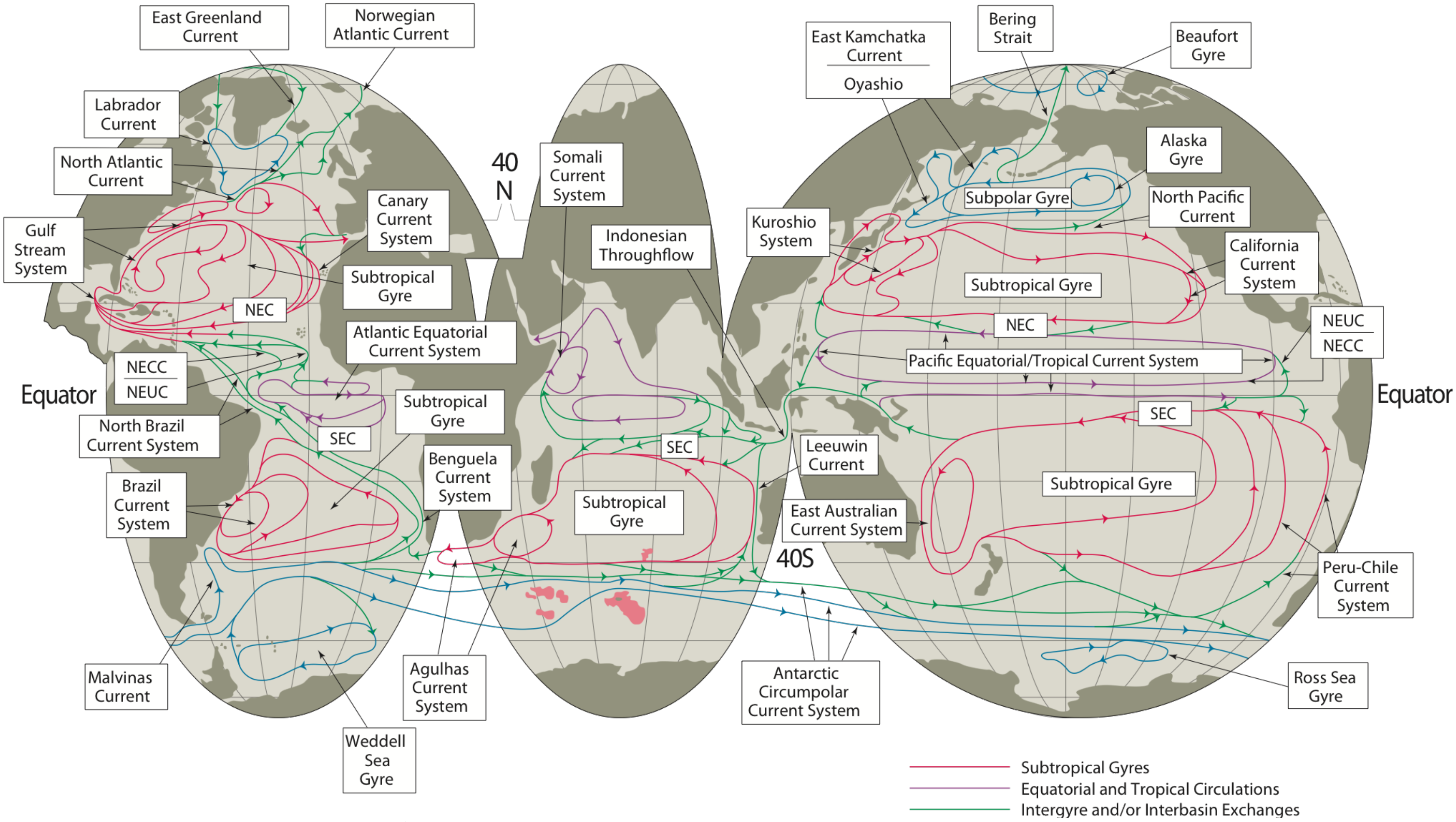


SIO 210: Global Ocean Circulation

- Summary of circulation
- Summary of water masses
  - Primary layers (4) for each ocean and sources
  - T/S
- Meridional overturning
- Global overturning
- Role of air/sea fluxes and diffusivity

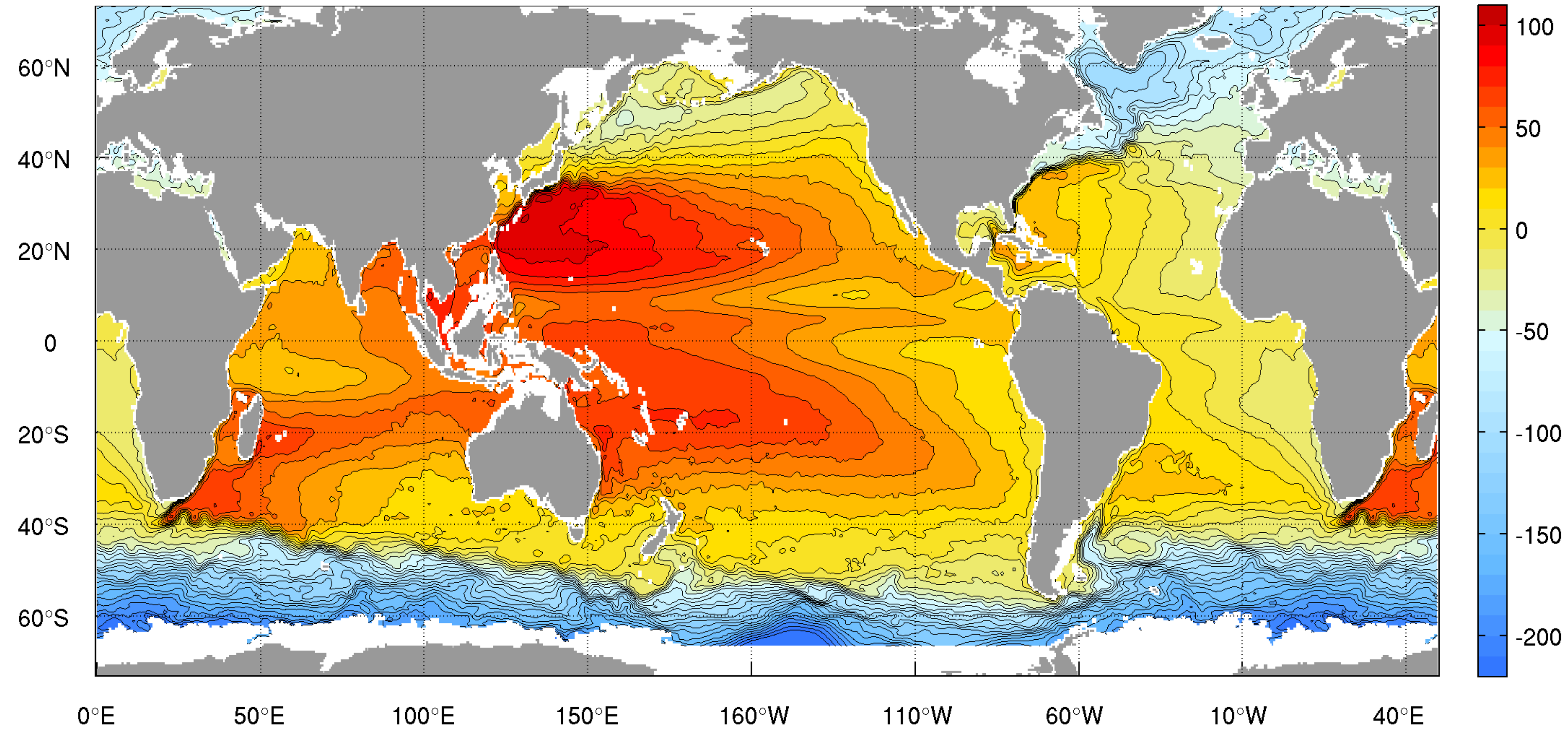
- Reading: parts of DPO Chapter 14

Upper ocean circulation



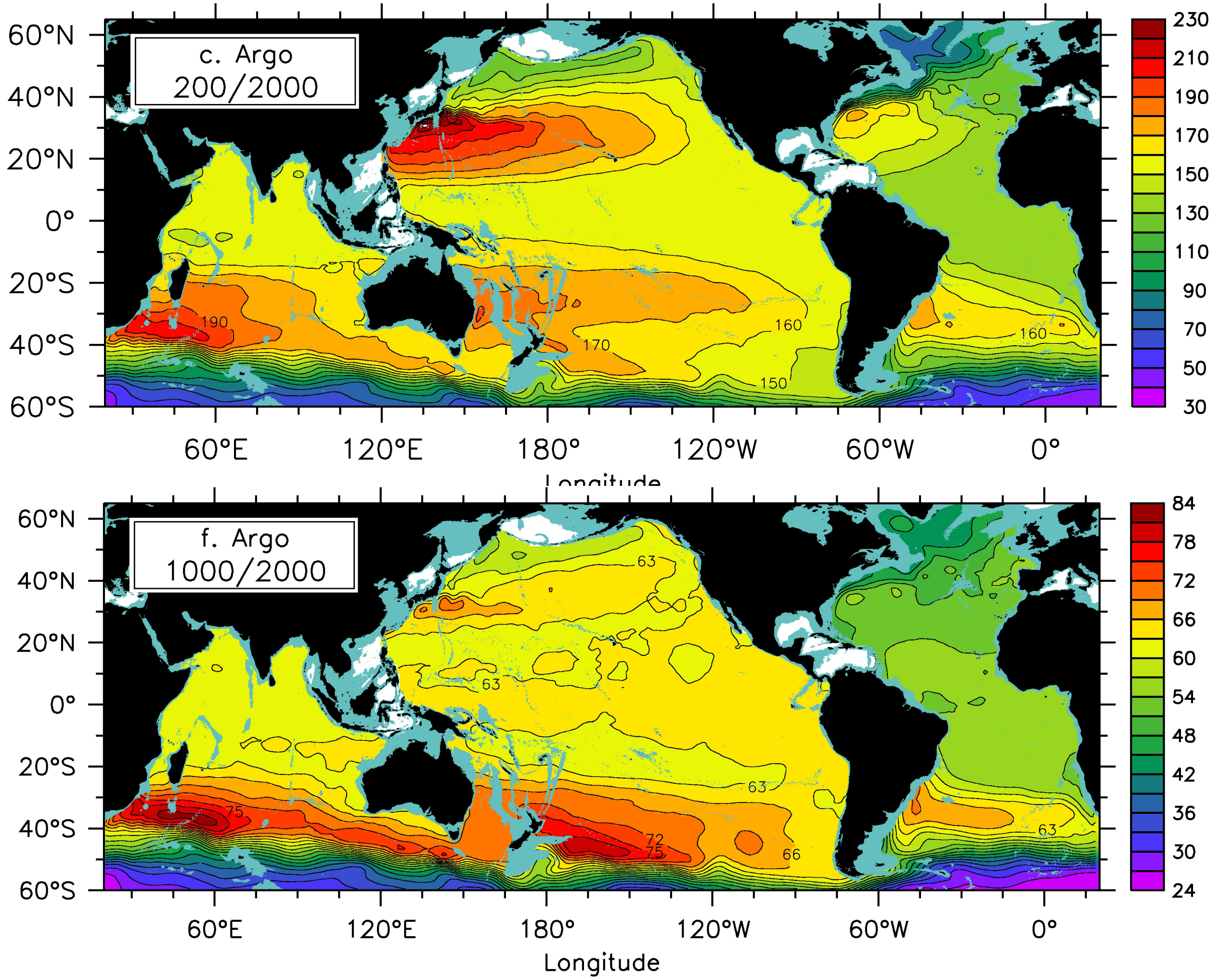
Schmitz (1995) (DPO Fig. 14.1)

Absolute surface height, related to surface geostrophic velocity



Maximenko et al., GRL, 2003 (DPO Fig. 14.2)

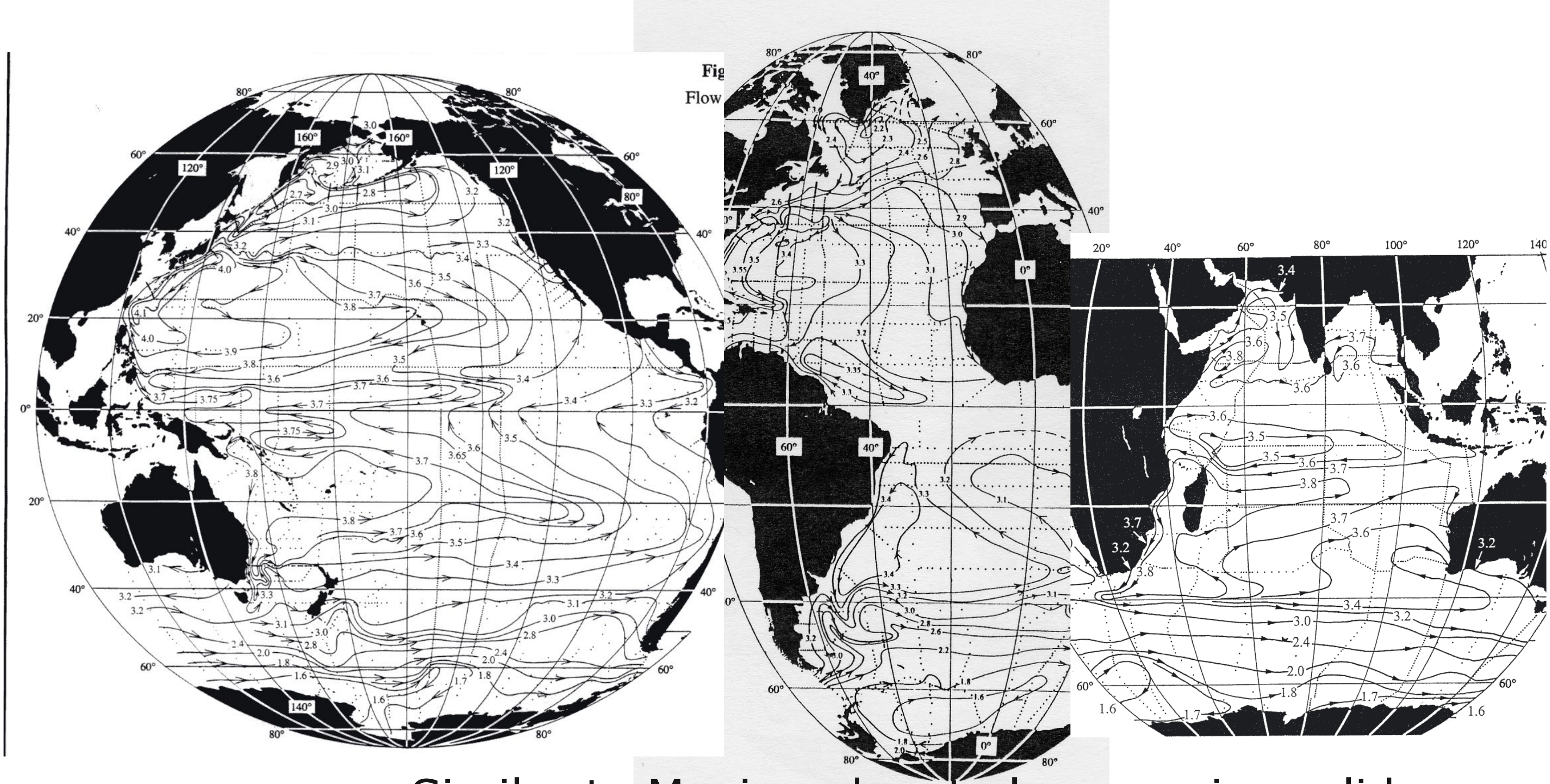
200 and 1000 dbar circulation rel. to 2000 dbar



Shrinkage of strong part of ST gyres to west and pole (into the WBC)

Roemmich and Gilson, 2009
DPO Fig. 14.3

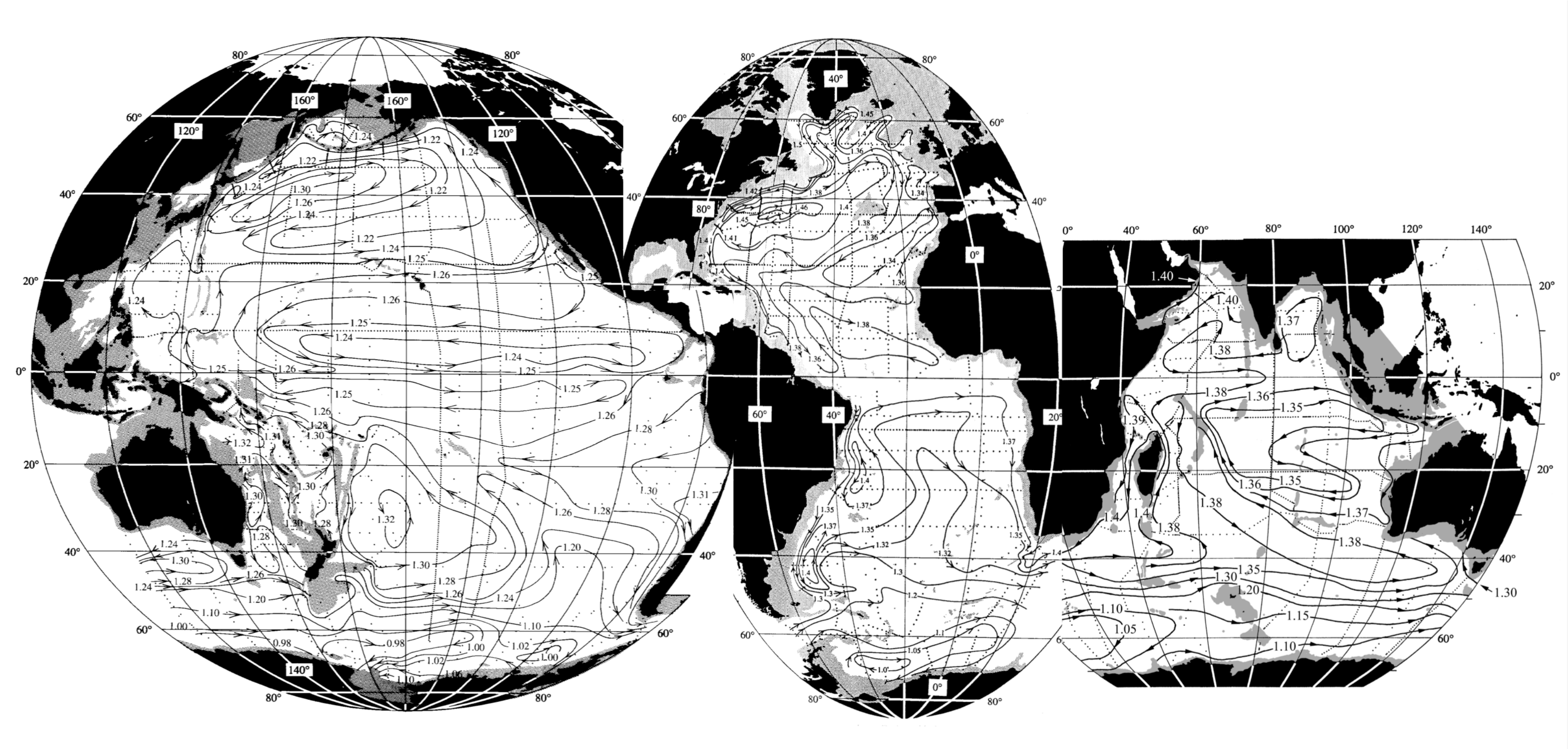
Surface circulation (absolute steric height from hydrography)



Similar to Maximenko et al on previous slide

(Reid, 1994, 1997, 2003)

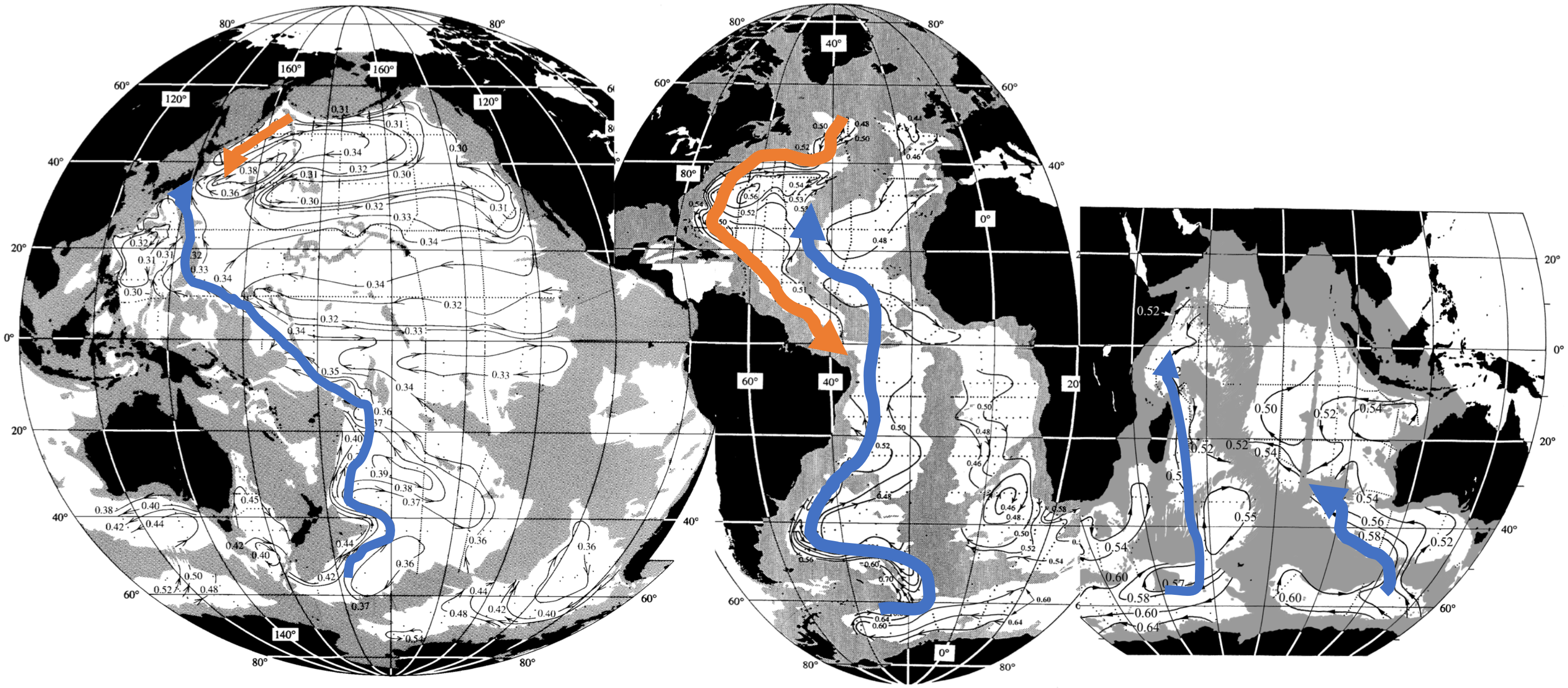
Circulation at 2000 dbar (adjusted steric height)



Greatly reduced subtropical gyres, continued ACC and equatorial zonal flows, weak circulations elsewhere.

DPO Fig. 14.4a

Circulation at 4000 dbar (adjusted steric height)



Below depth of NADW in S. Atlantic

Dominated by topography. Deep Western Boundary Currents, deep cyclonic flows in some isolated basins

(Reid, 1994, 1997, 2003)

DPO Fig. 14.4b

Global meridional overturning circulation

- Because sources of densest waters are in the northern N. Atlantic/Nordic Seas (NADW) and in the Antarctic (AABW), interesting to see how these fill the global ocean, upwell and return in upper ocean back to source regions

- Common to look at NET MERIDIONAL TRANSPORT across latitudes. Most useful to look at this in isopycnal layers (not depth layers), since flow is mostly along isopycnals.

- Also look at heat transport and freshwater transport this way.

- Choice of isopycnal layers is related to water masses. We can see signature of meridional overturn by looking at the intermediate, deep and abyssal water masses.

Calculation of meridional overturn

DPO Section 14.2.2

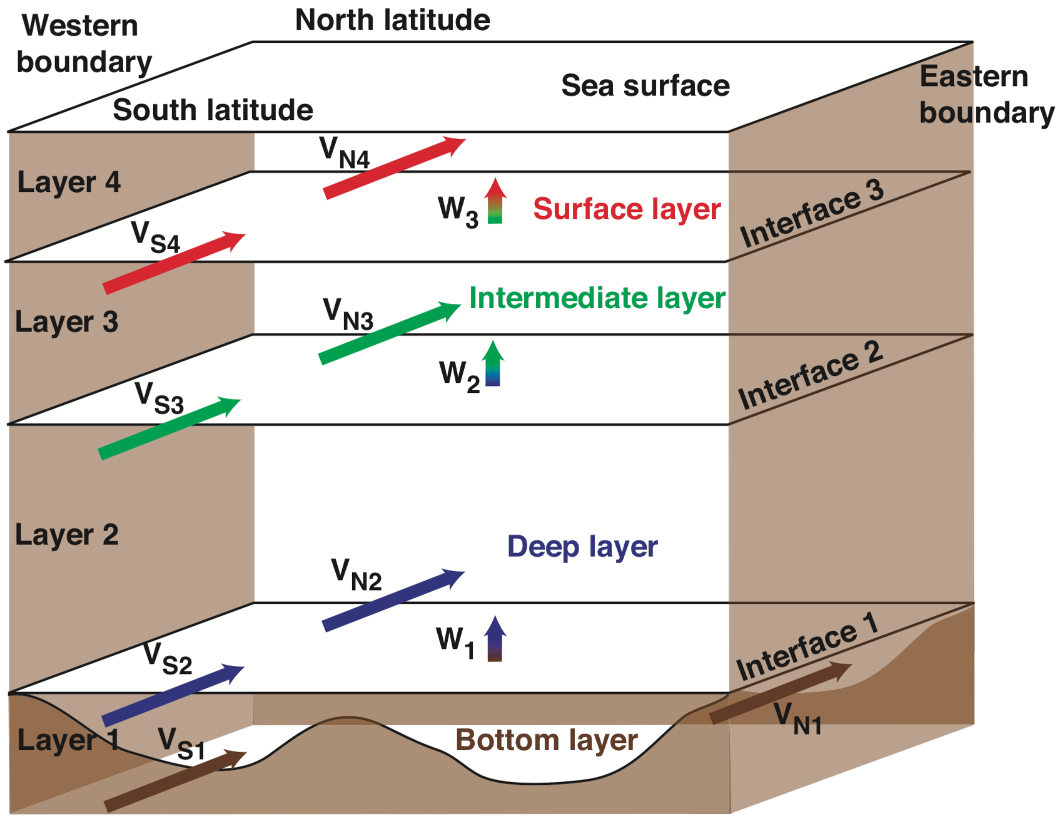
Use a zonal, coast-to-coast, top-to-bottom section

Compute meridional geostrophic velocities, and estimate meridional Ekman transport

Calculate zonally-integrated transports in layers (isopycnal layers or pressure layers).

Add Ekman transport to top layer.

Total transport through section should equal any leakages (such as about 1 Sv for Bering Strait)



Calculation of meridional overturning

(1) Total Mass transport in layer = 0

(2) Total vertical transport through interfaces calculated as difference between meridional transports and underlying vertical transport

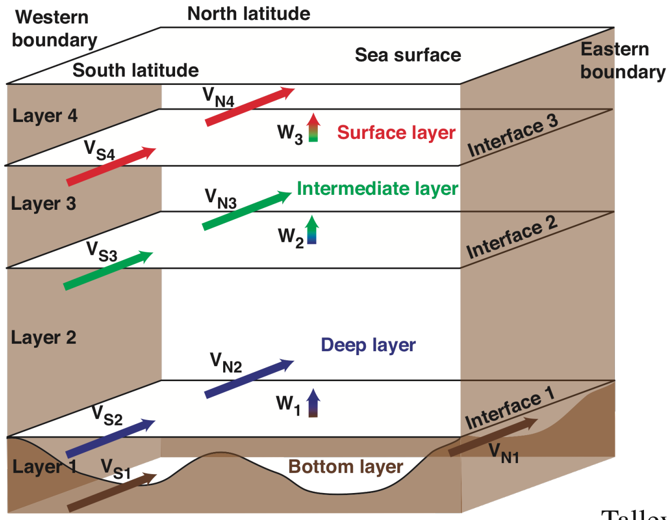
(3) Vertical velocity (average) = vertical transport divided by area of interface

Equation 14.1

(1) $M_{Ti} = V_{Ni} - V_{Si} + W_{i-1} - W_i = 0$

(2) $W_i = V_{Ni} - V_{Si} + W_{i-1}$

(3) $w_i = W_i/A_i$



DPO Fig. 14.5

Calculation of meridional overturning

Example:

(1) V_{S1} = 5 Sv of AABW moves northward through south face into bottom layer
V_{N1} = 4 Sv of AABW moves northward out of north face

(2) Therefore W_1 = 1 Sv of AABW must upwell into the next layer above.

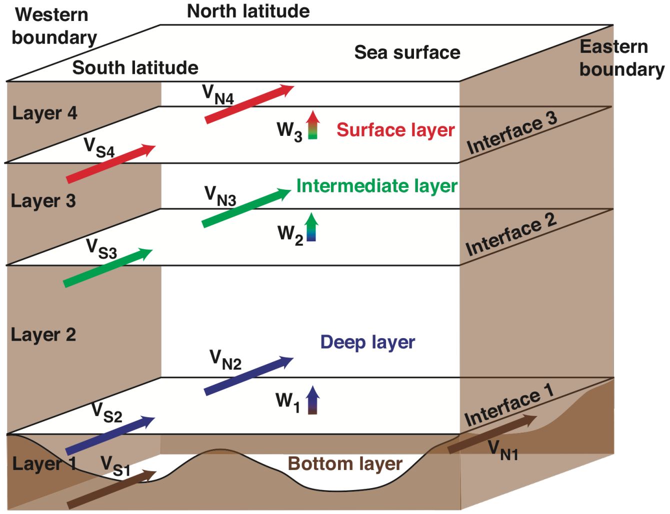
(3) If the area of the interface is 4,000 km x 1,000 km, the average vertical velocity is
$w_1 = (1\times10^6\ \mathrm{m^3/sec})/(4 \times 10^{12}\ \mathrm{m^2}) = 0.25\times10^{-6}\ \mathrm{m/sec} = 0.25\times10^{-4}\ \mathrm{cm/sec}$

Equation 14.1 (corrected sign)

(1) $M_{Ti} = V_{Ni} - V_{Si} + W_i - W_{i-1} = 0$

(2) $W_i = -(V_{Ni} - V_{Si}) + W_{i-1}$

(3) $w_i = W_i/A_i$

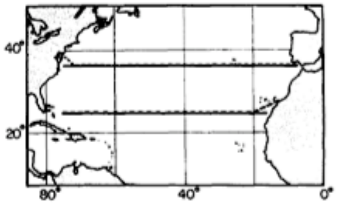


DPO Fig. 14.5

Calculation of meridional overturning transports

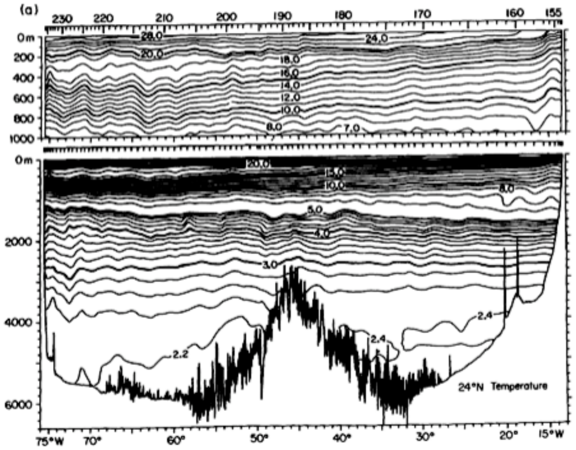
Example: 24° N and 36° N N. Atlantic

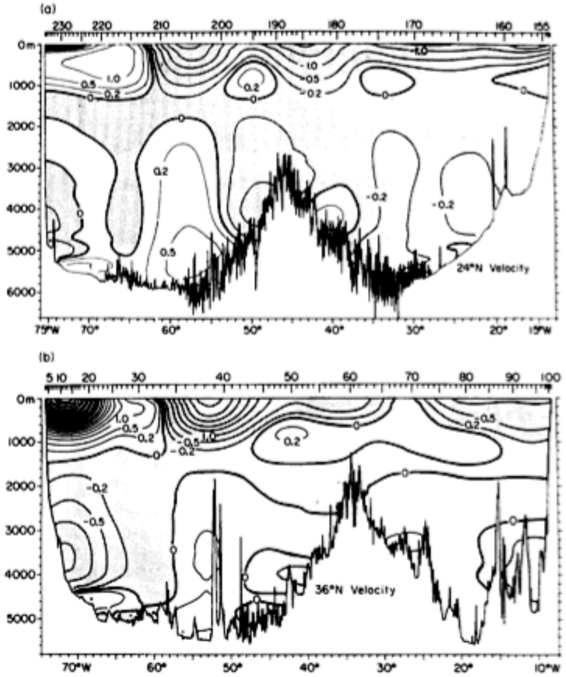
(Roemmich and Wunsch, JGR 1985)



Location

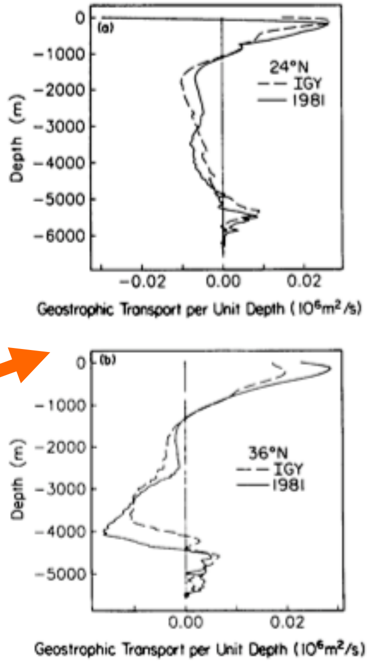
Temperature section





Smoothed geostrophic velocities

Fig. 9. (a) Horizontally smoothed geostrophic velocity (cm s⁻¹) at 24°N from 1981 data, based on the reference level calculation with constraints on total transport only. Southward velocities are shaded. The western end is on the left. (b) Same as (a) for 36°N.

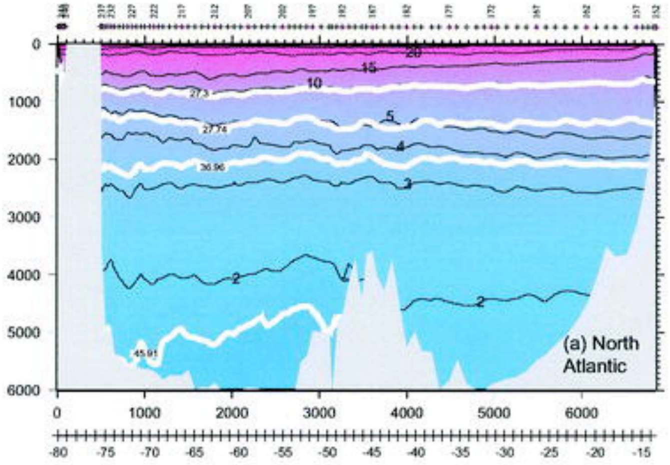


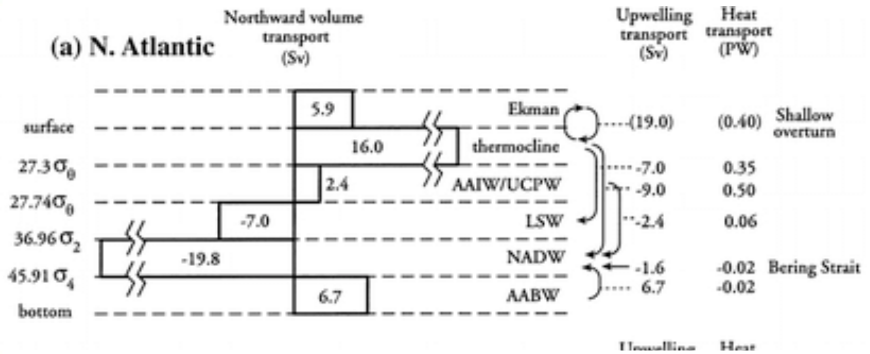
Meridional transports

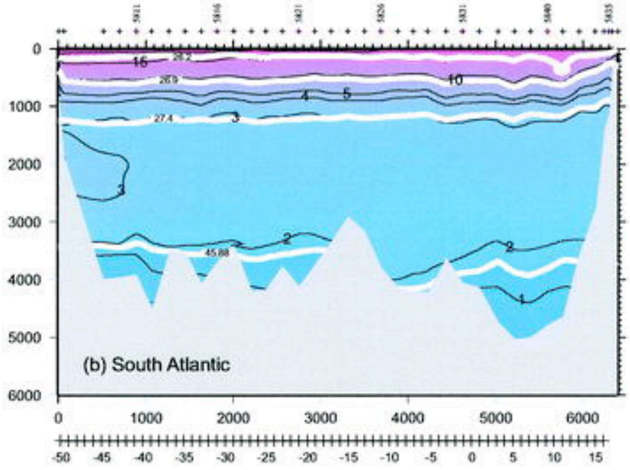
Fig. 6. (a) Geostrophic transport per unit depth across 24°N including Straits of Florida, based on the reference level calculation with constraints on total transport only. The solid line is from 1981 data and the dashed line from IGY data. (b) Same as (a) but for 36°N.

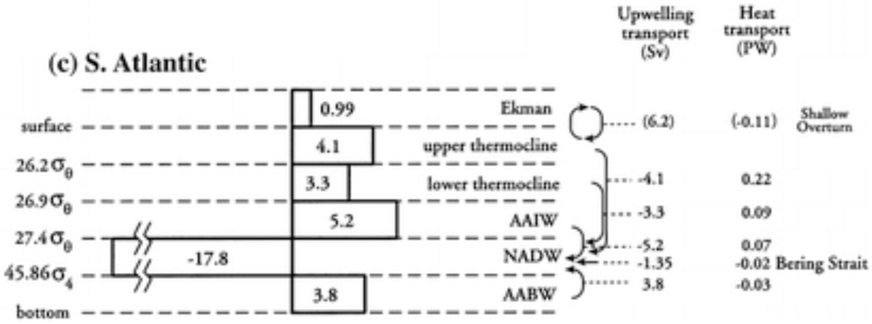
Calculation of meridional overturning transports

Example: Atlantic 24° N and 32° S (Talley, JPO, 2003)

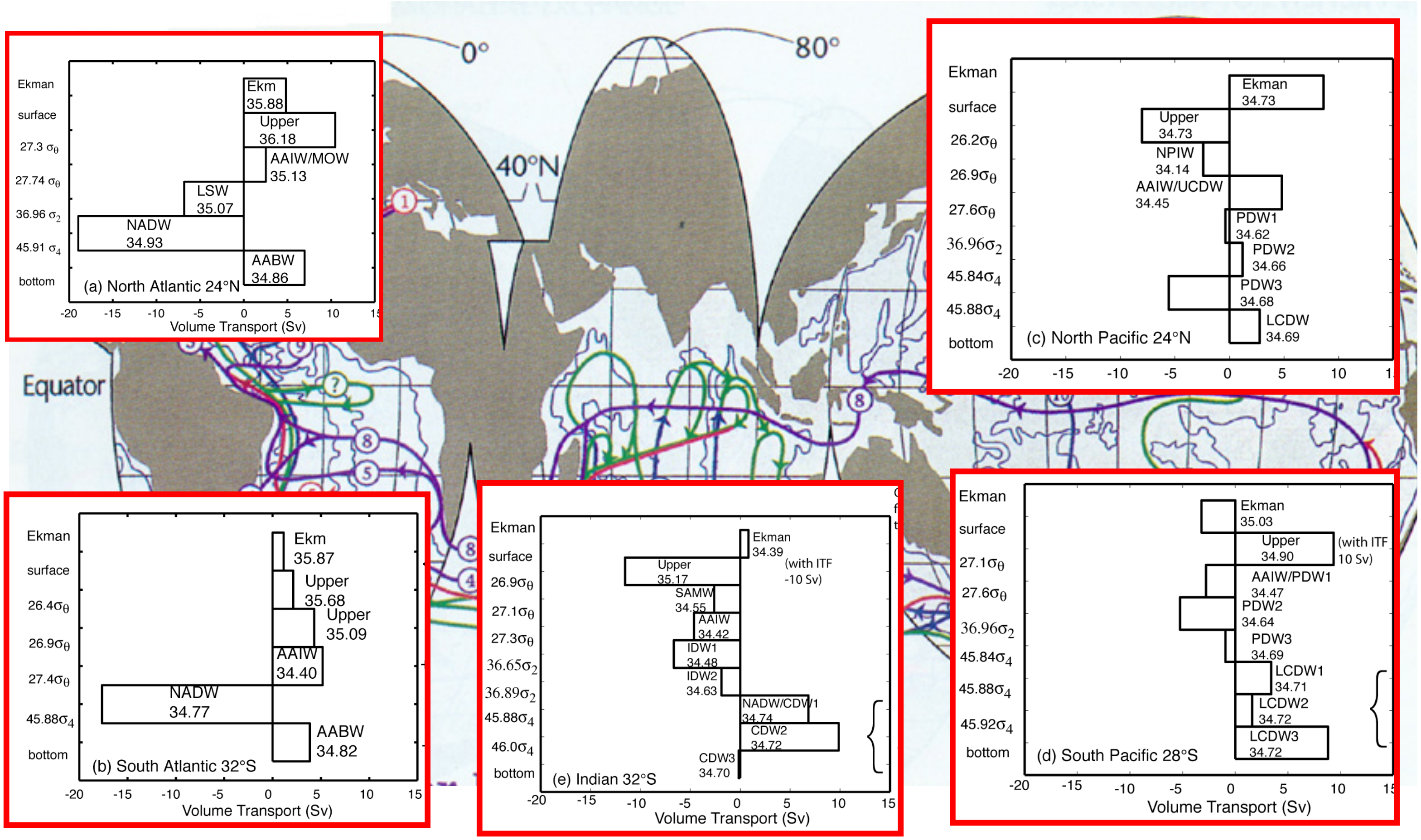








Net meridional transports in isopycnal layers



Global mass transport

(Ganachaud and Wunsch, 2000)

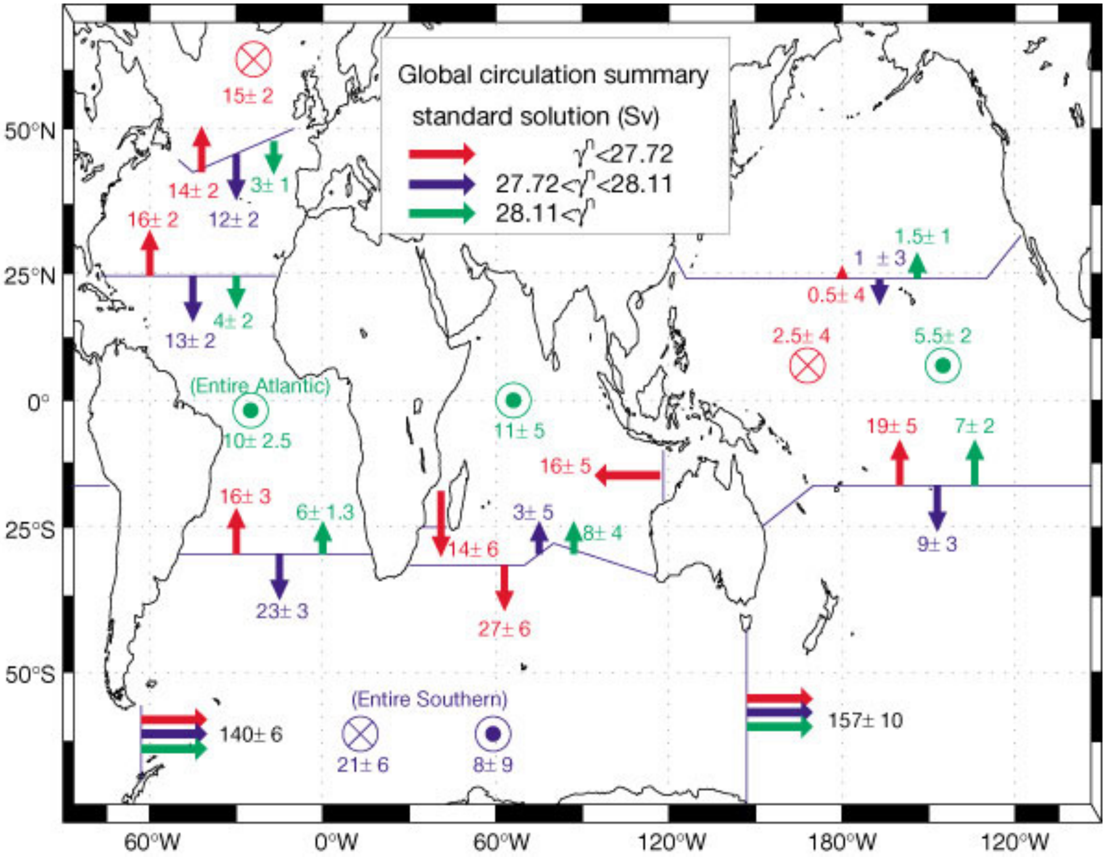


Figure 14.6 in DPO is more complicated, as it includes two additional transport analyses

Vertical velocities (model result)

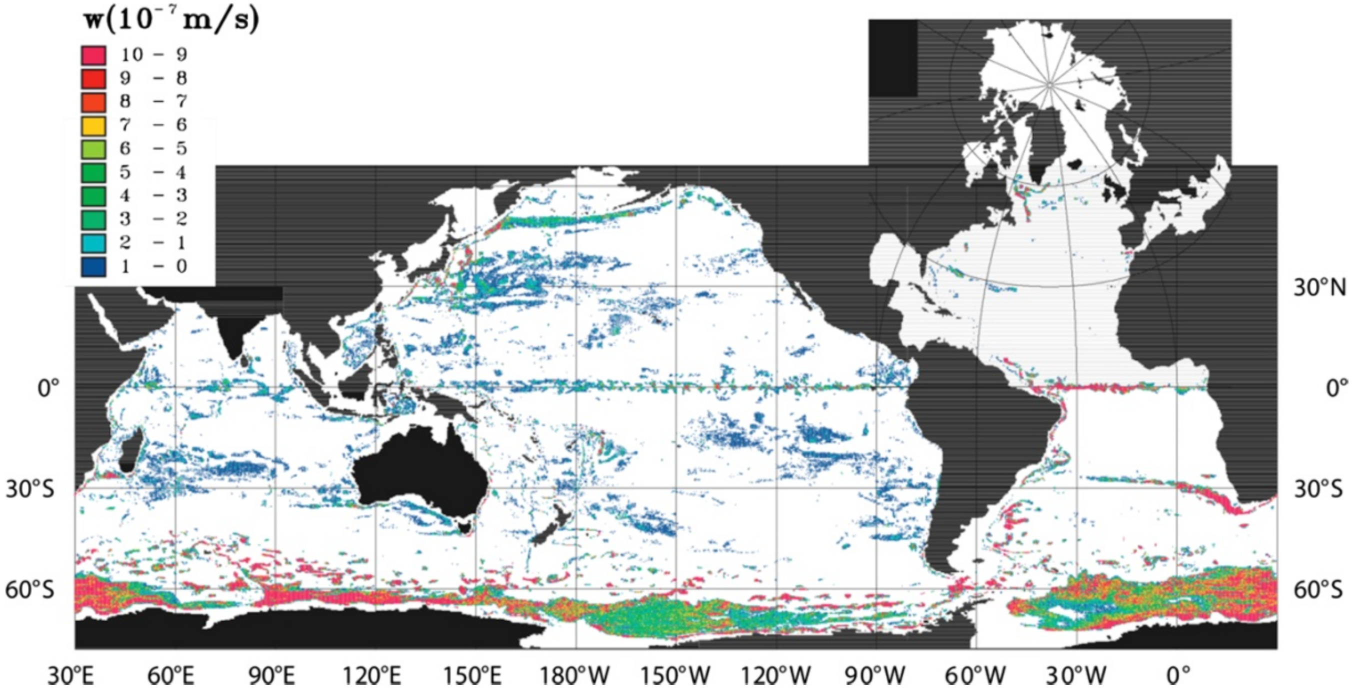


Figure 14.7
Modeled upwelling across the isopycnal 27.625 kg/m^3, which represents upwelling from the NADW layer. This figure can also be found in the color insert. Source: From Kuhlbrodt et al. (2007); adapted from Doos and Coward (1997).

Calculation: Meridional Overturning Streamfunction

To calculate a meridional overturning streamfunction Ψ (units are Sv):

Add layer meridional transports, from bottom to top, keeping track of value at each depth.

Best to have transports in relatively thin layers, and very best to have them in isopycnal layers.

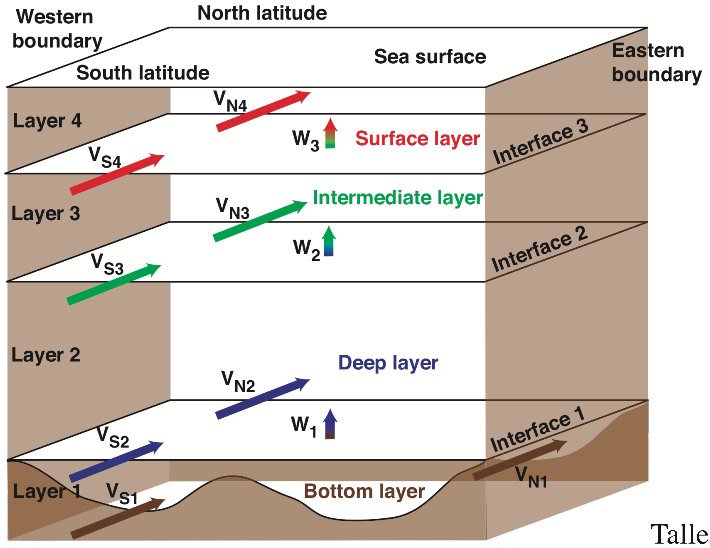
Useful to have transports at many latitudes (only available from models)

Equation 14.2

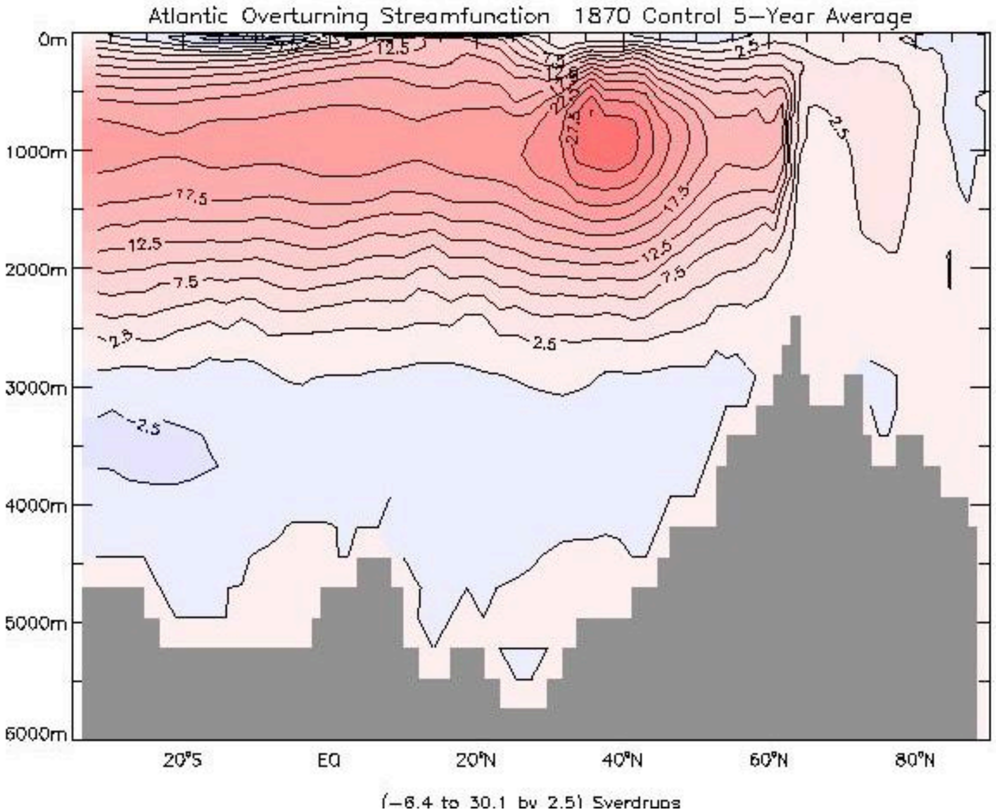
$$\boxed{\Psi_i = \sum_{i=1}^{N} V_i}$$

$$\Psi(z) = \int_0^z \int_{x_{west}}^{x_{east}} v(x',z')\,dx'\,dz'$$

$$\Psi(\rho) = \int_0^\rho \int_{x_{west}}^{x_{east}} v(x',\rho')\,dx'\,d\rho'$$

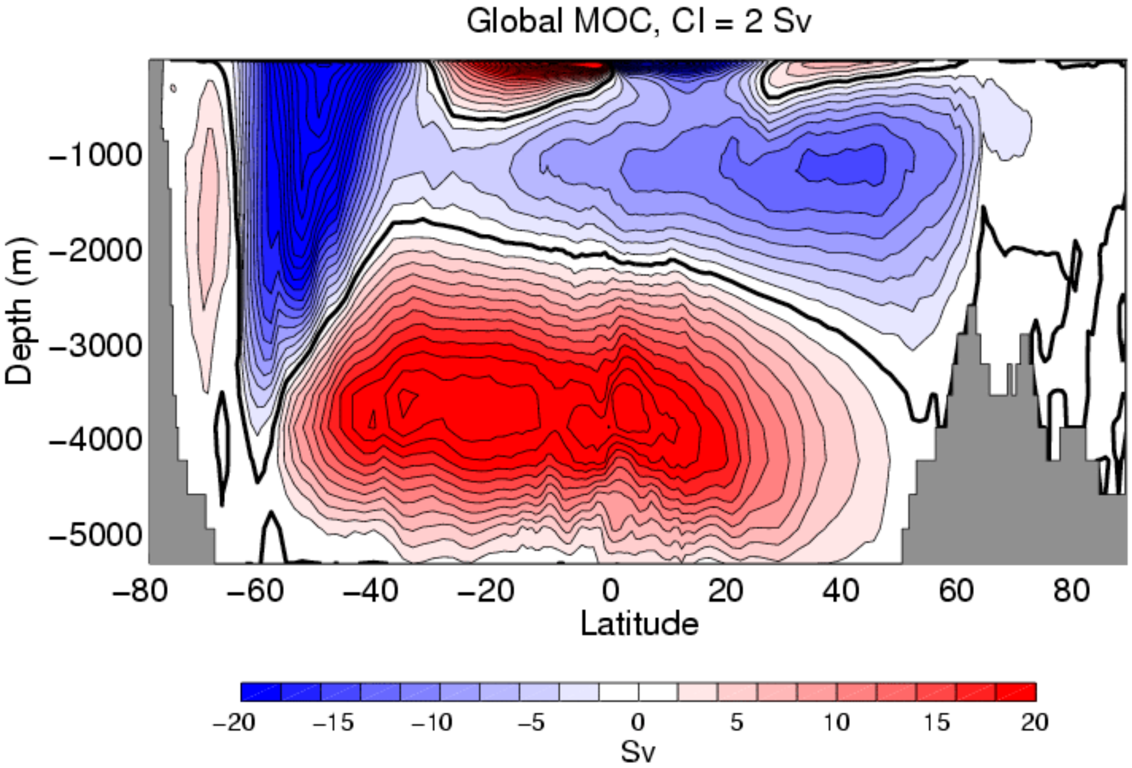


Meridional Overturning Streamfunction



Example of an Atlantic overturning streamfunction (from a circulation model). Numbers are transport in Sverdrups. From Gent (2000).

Meridional Overturning Streamfunction

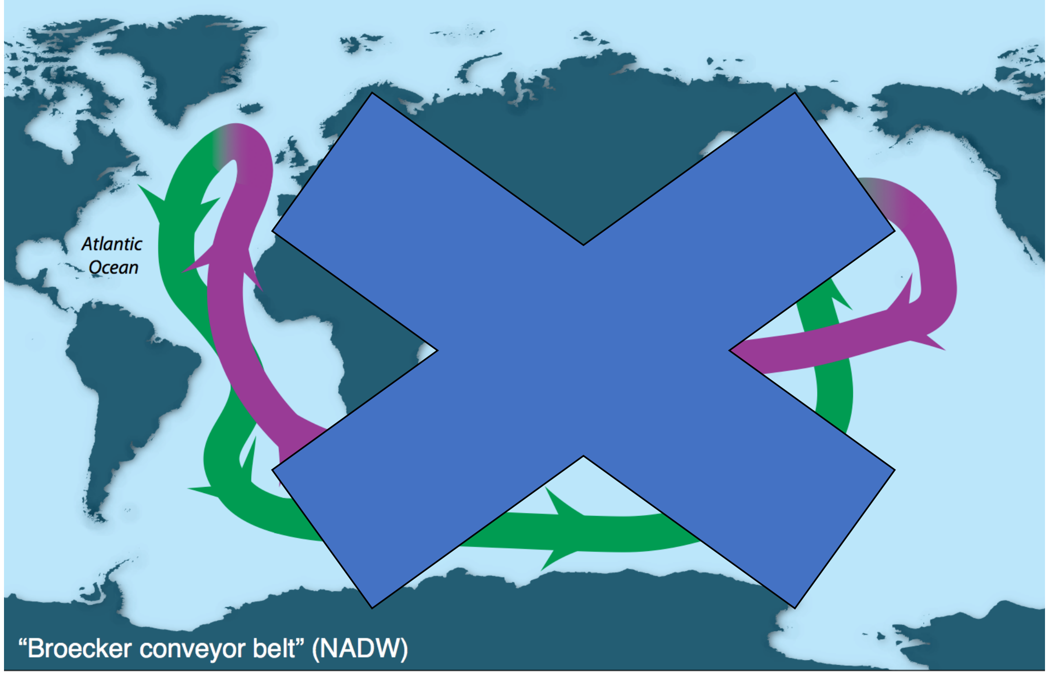


Global zonal average: Results from P-OMIP at GFDL

http://www.frontier.iarc.uaf.edu/pomip/results.php

Schematics of the overturn: based on water mass concepts and quantitative estimates of overturning transport

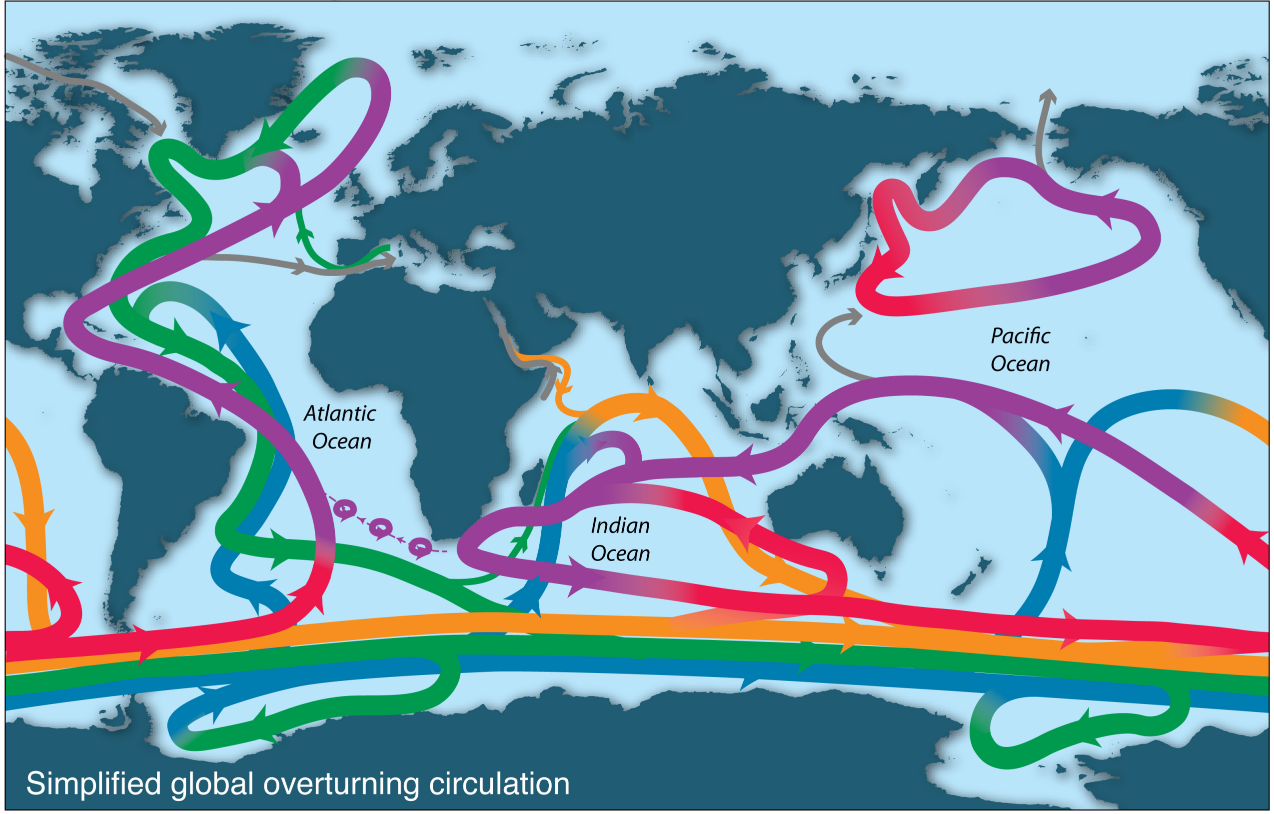
Broecker "conveyor belt"



Broecker (1981)
(DPO Fig. 14.10)

Simplification of NADW global overturning circulation ALONE
Missing: AABW global overturning circulation, actual ocean connections

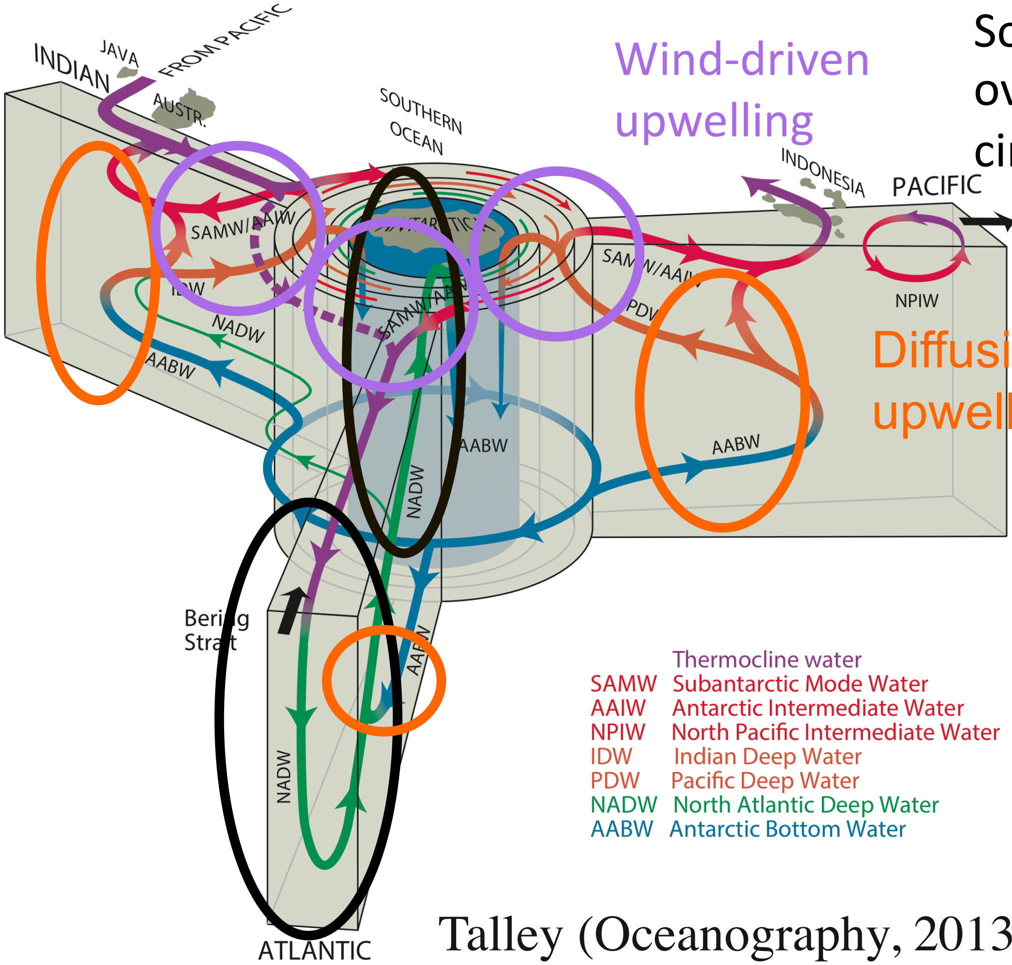
Global overturning schematic



More complete view based on major water mass transformations and quantitative overturning transports

Talley (2013) based on DPO Fig. 14.11a

Processes for the Global Overturning Circulation



Wind-driven upwelling

Brine rejection: Southern Ocean overturning circulation (AABW)

Diffusive upwelling

Deep convection N. Atlantic Meridional Overturning Circulation (AMOC) (NADW)

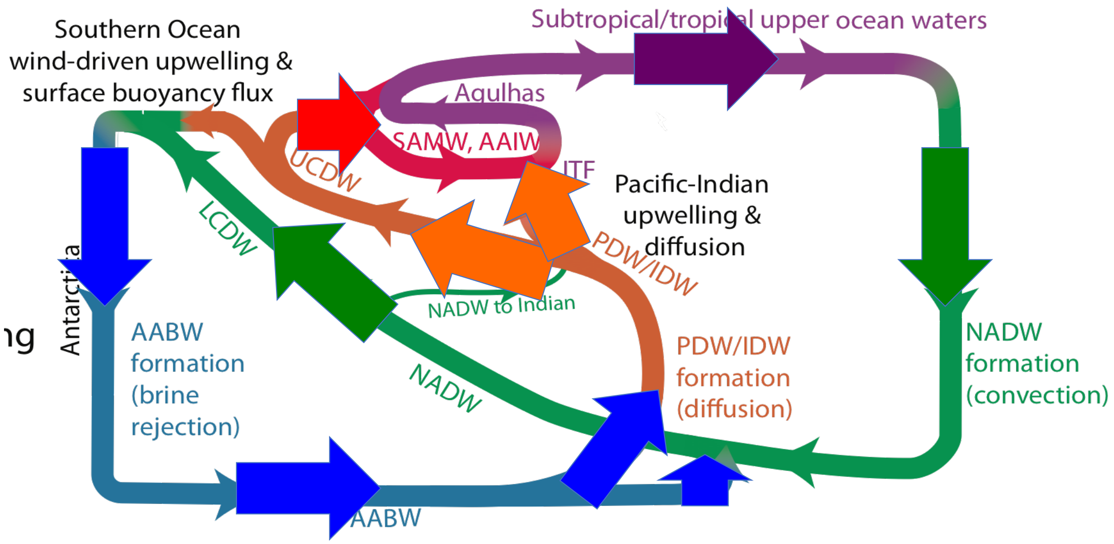
Talley (Oceanography, 2013)

Global Overturning Circulation schematic

Warm, salty, low nutrient thermocline/ surface water moves north and cools to make NADW (dense)

Cold, carbon and nutrient rich, old deep waters rise to S.O. surface

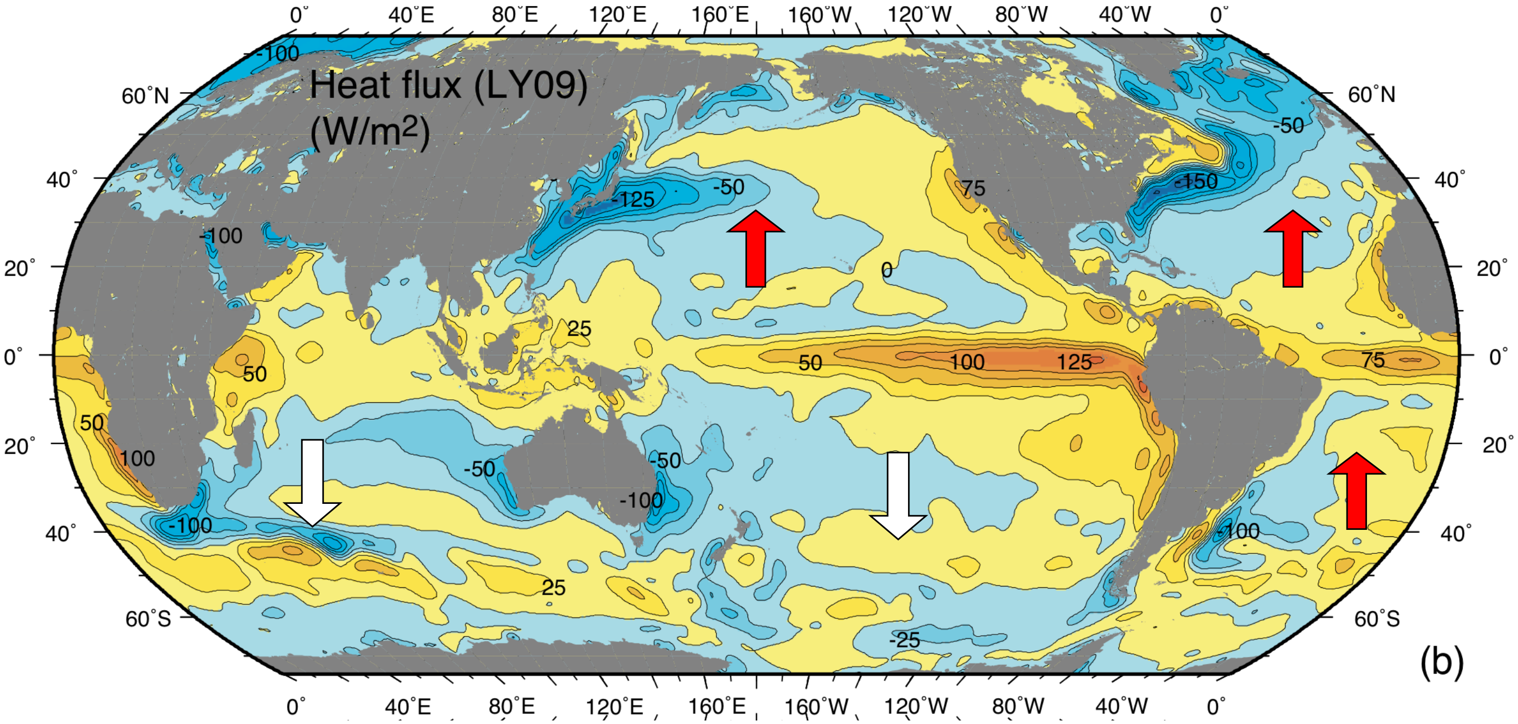
Split to make AABW (dense) and thermocline water (light)



DPO Fig. 14.11c

Talley (DPO, 2011; Oceanography, 2013)

Heat transported by ocean circulation (big arrows)



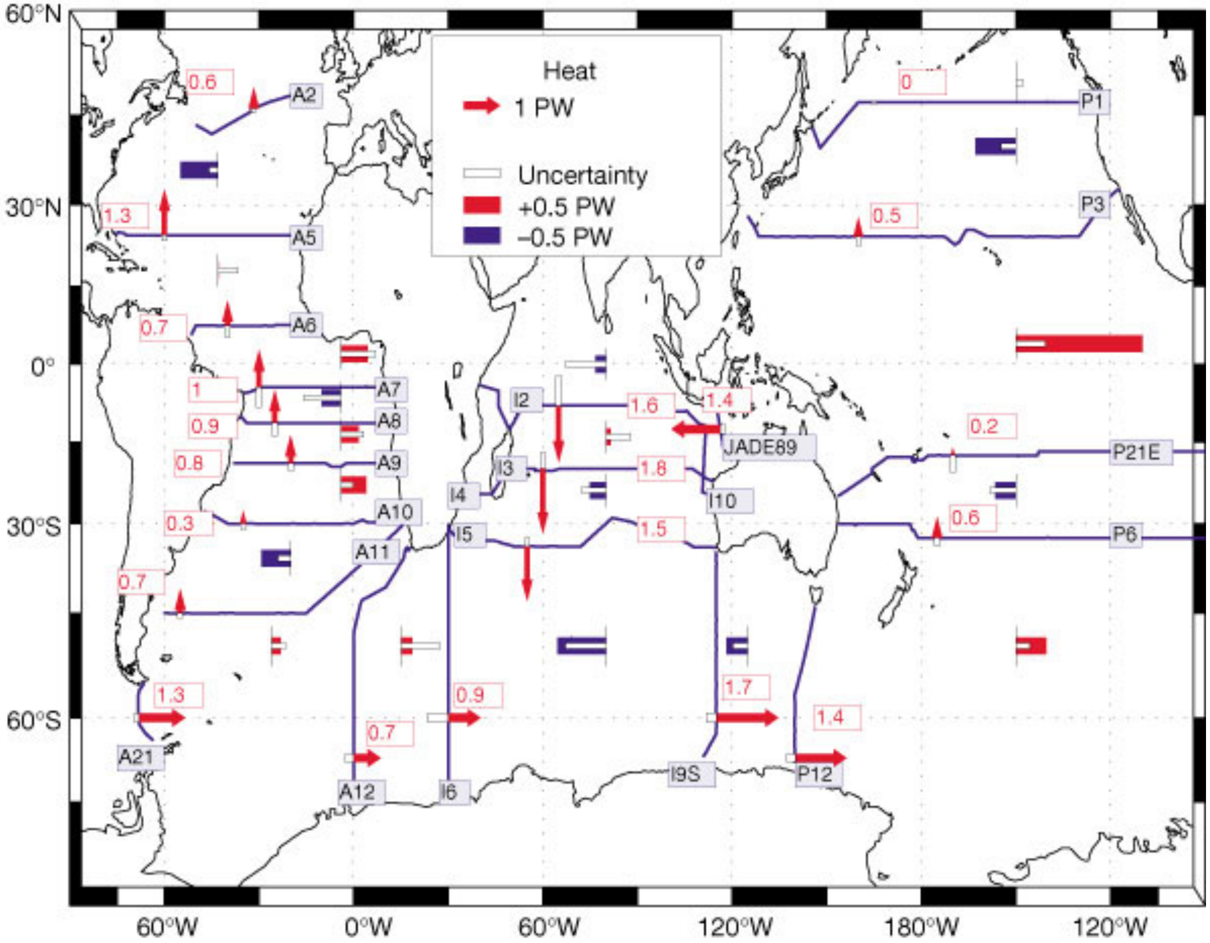
Heat transport: red is northward, white is southward.

Two components: (1) poleward due to subtropical gyre circulation (warm western boundary current plus cooler subducted water).

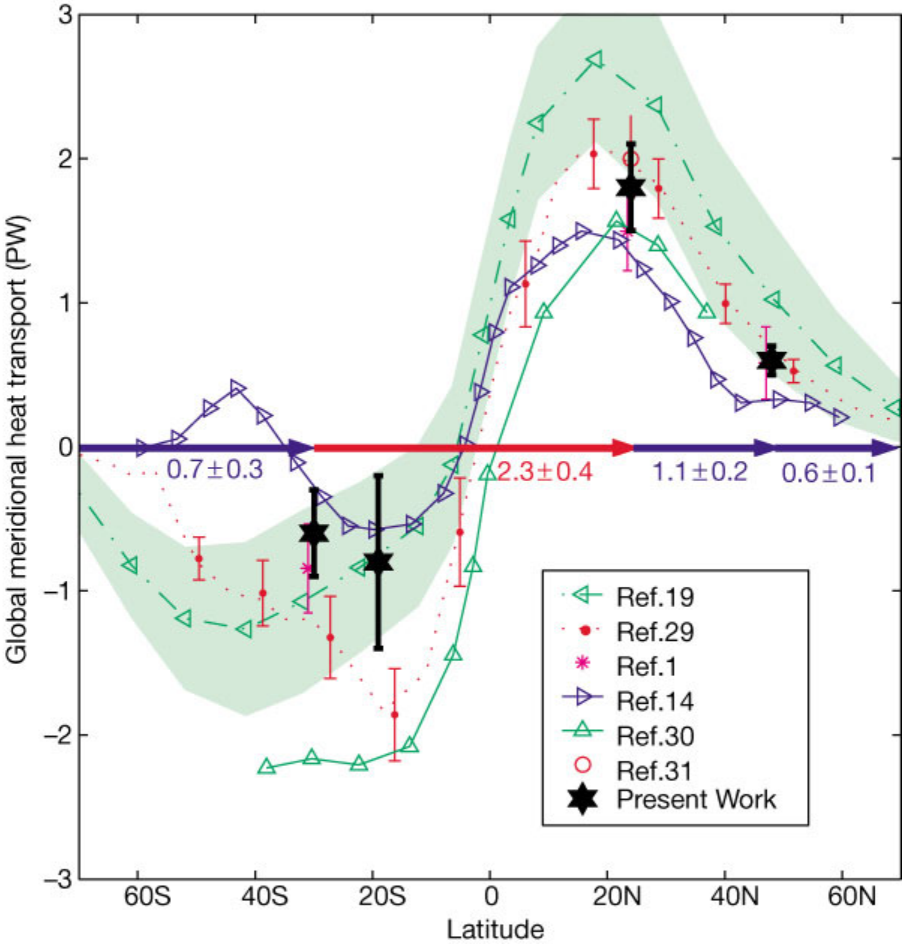
(2) MOC in Atlantic (warm surface waters plus cold NADW)

Air-sea heat flux: Yellow/orange - ocean gains heat. Blue - ocean loses heat.

Global heat transport
(Ganachaud and Wunsch, 2000)



Global heat transport (Ganachaud and Wunsch, 2000)



Water mass review: 4 layer view of the global ocean

(1) Upper layer: ventilated thermocline. Includes Mode Waters, Central Water, subtropical Underwater (salinity maximum water)

(2) Intermediate layer: Labrador Sea Water, Mediterranean Overflow Water, Red Sea Water, North Pacific Intermediate Water, Antarctic Intermediate Water

(3) Deep layer: North Atlantic Deep Water, Pacific Deep Water (also known as Common Water), Indian Deep Water, Circumpolar Deep Water

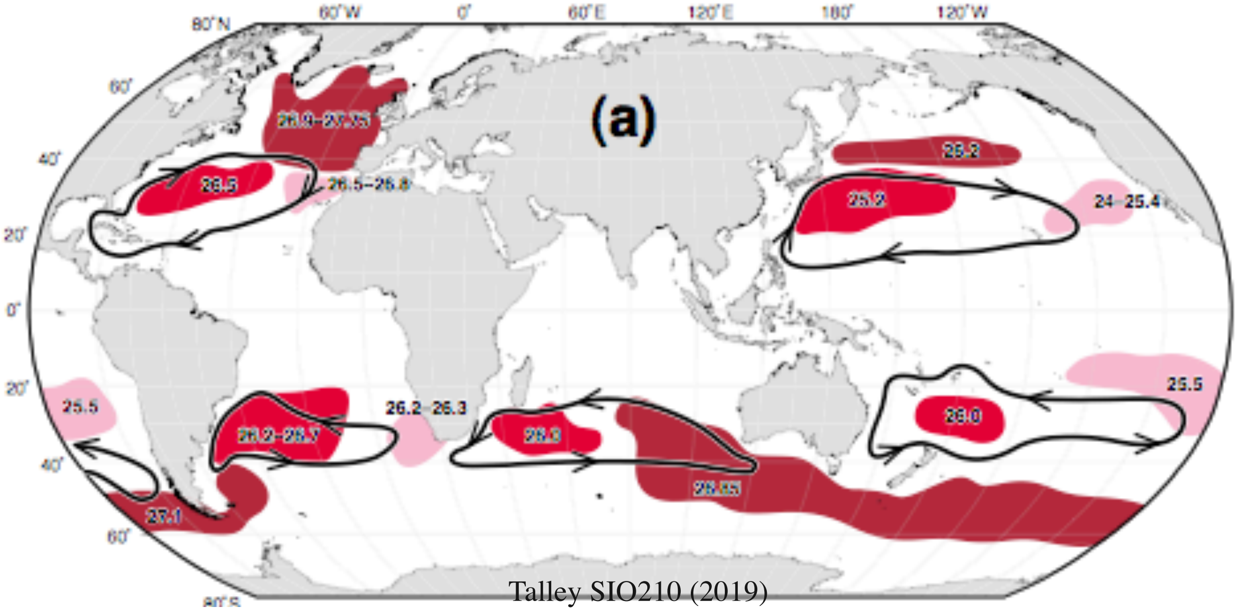
(4) Bottom layer: Antarctic Bottom Water (aka Lower Circumpolar Deep Water)

Remember these layer numbers!

(1) Upper ocean water masses

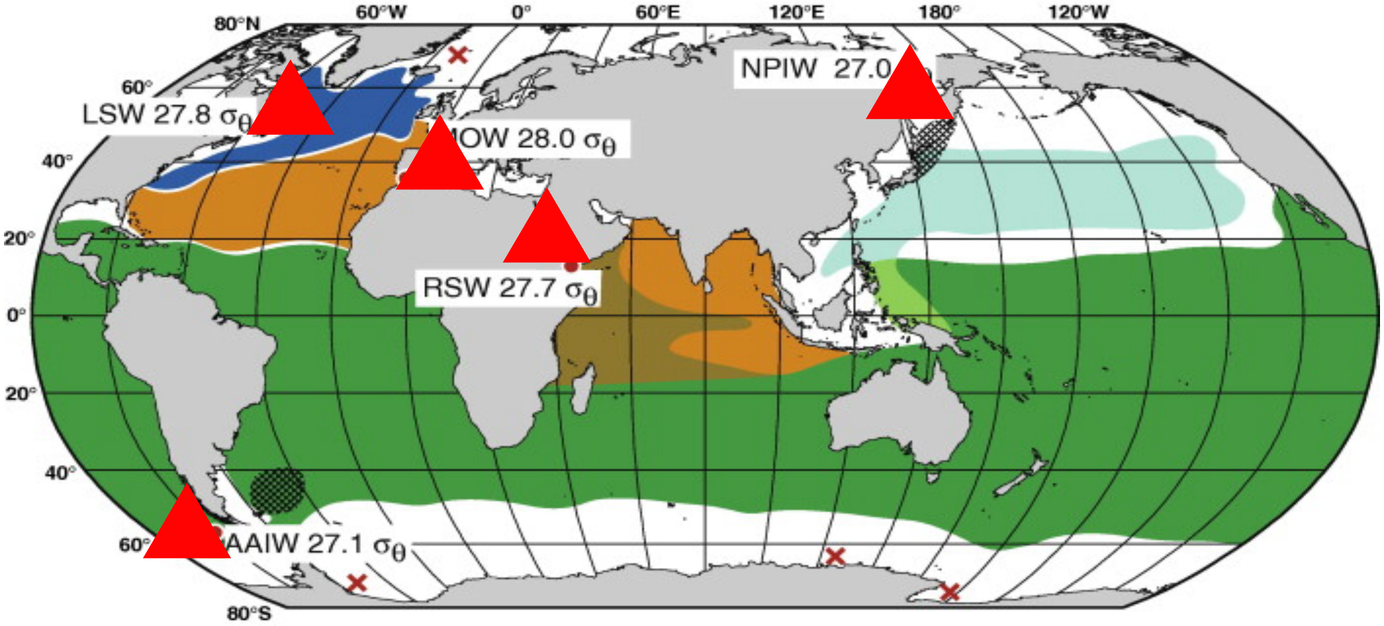
Central Waters (main subtropical thermocline, derived from broad subduction of subtropical surface waters) (not illustrated here)

- Subtropical Underwater (ST gyre, shallow salinity maxima, derived from subduction of saltiest ST surface water) (not illustrated here)
- Mode Waters (upper ocean, thick layers) (figure)



Hanawa and Talley (2001)

(2) Intermediate water summary

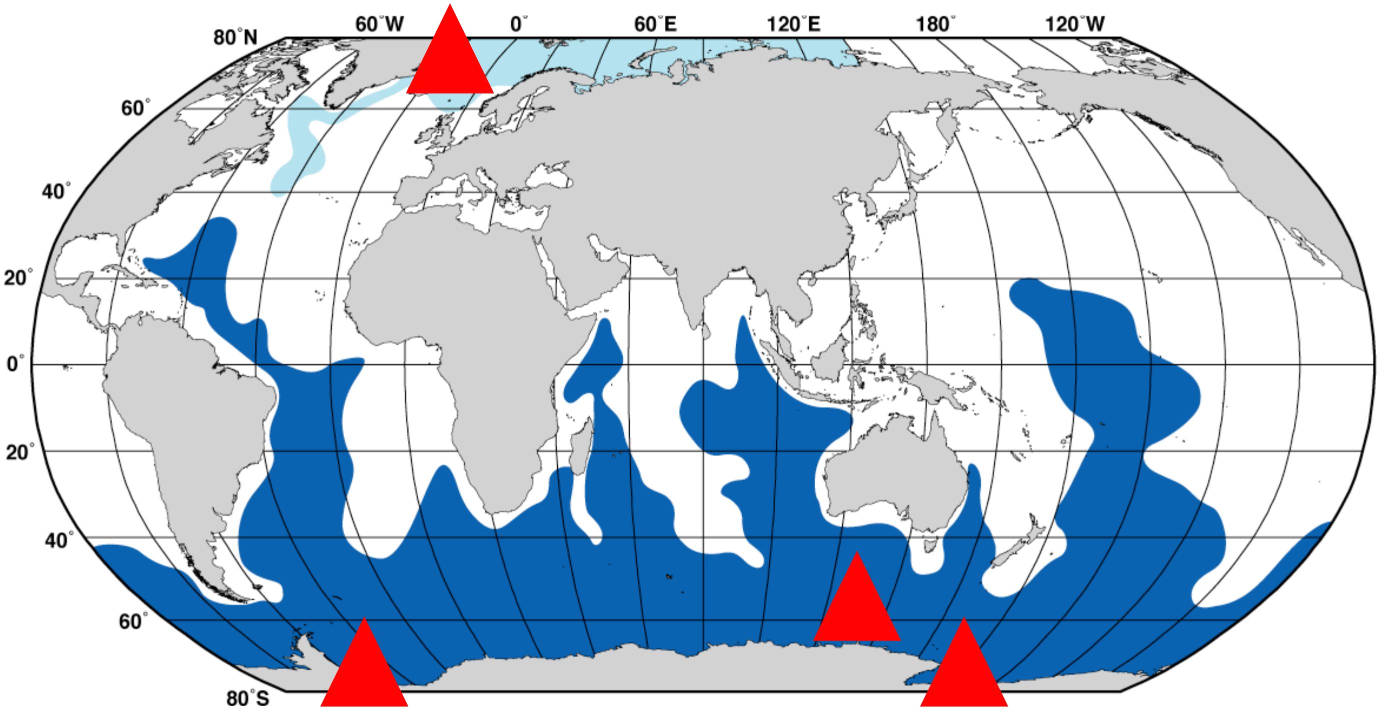


Low salinity: Labrador Sea Water, North Pacific Intermediate Water, Antarctic Intermediate Water

High salinity: Mediterranean Water, Red Sea Water

Talley (2008)

(3, 4) Deep water partial summary

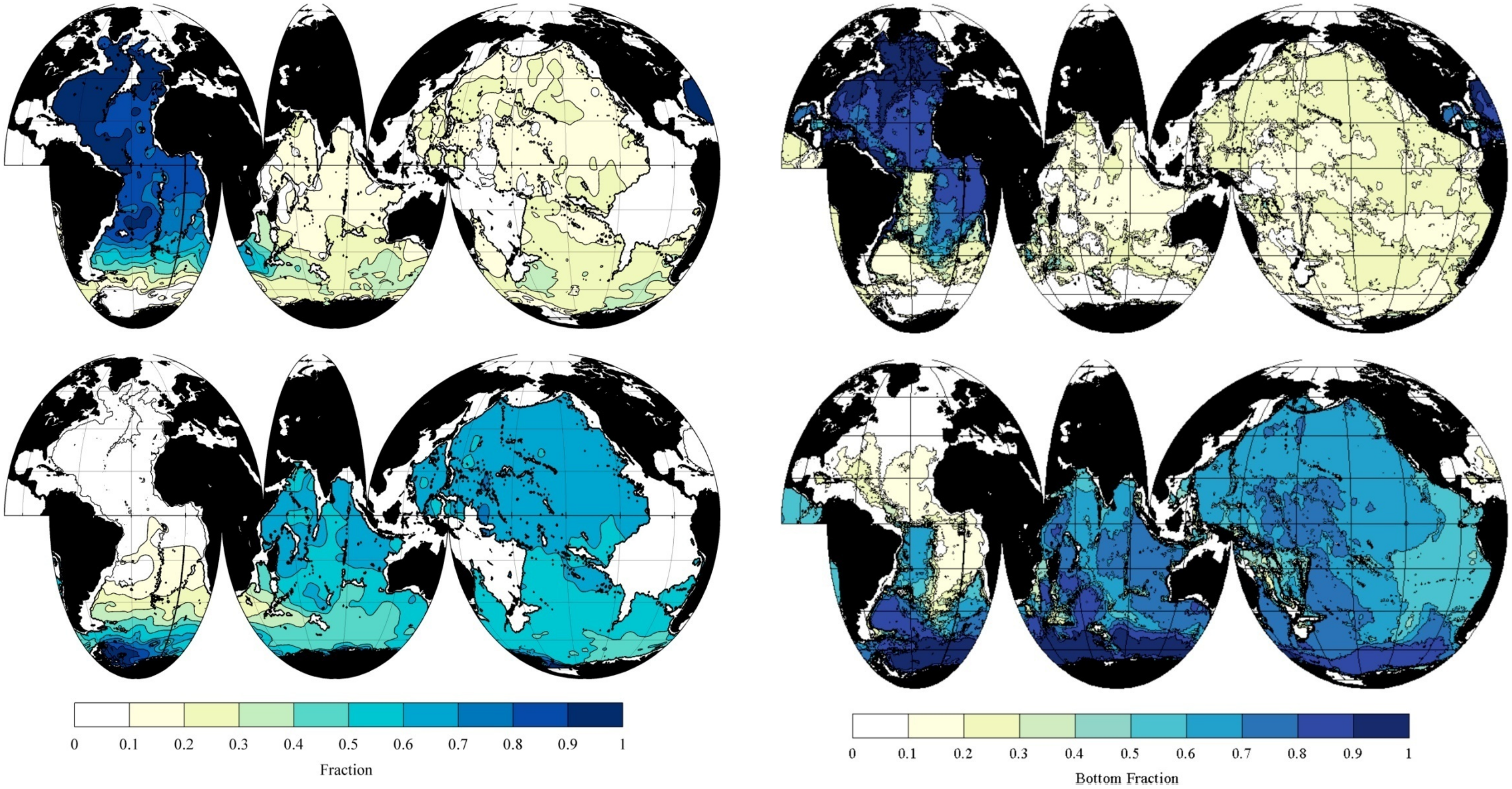


(3) Nordic Seas Overflow waters, contributing to NADW

(4) Antarctic Bottom Water in Weddell, Ross Seas and Adelie Coast

Talley (1997)

Fraction of NADW vs. AABW



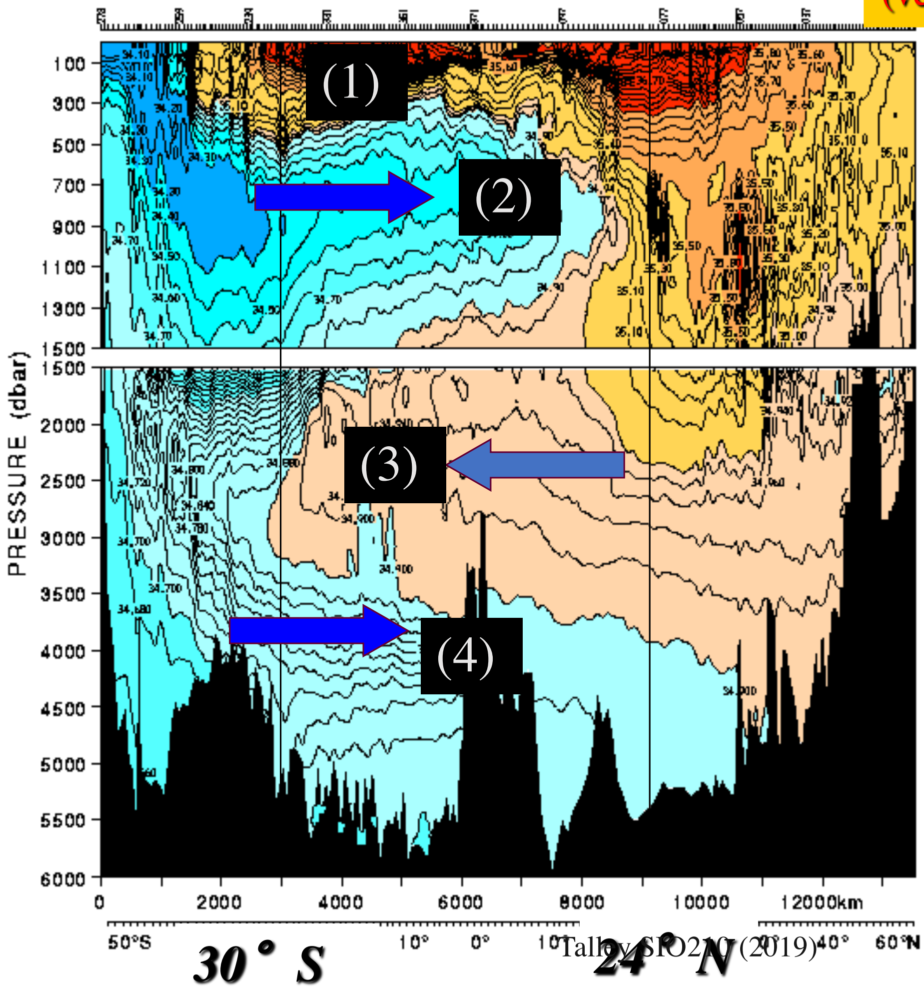
At about 2500-3000 m

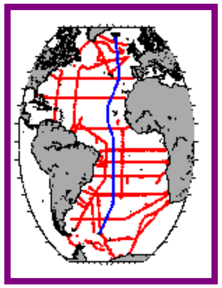
At the bottom

Johnson (2008) in DPO Figs. S14.4 and S14.5 (partially in fig. 14.15)

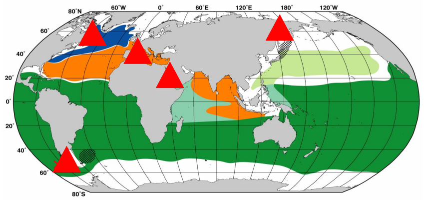
Atlantic Salinity (south-north section)

(1) surface waters
(ventilated thermocline)



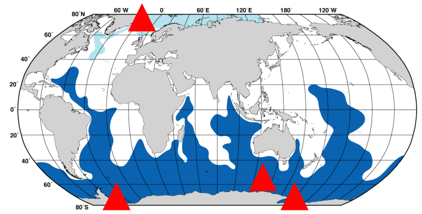


(2) Low salinity
Antarctic intermediate
water



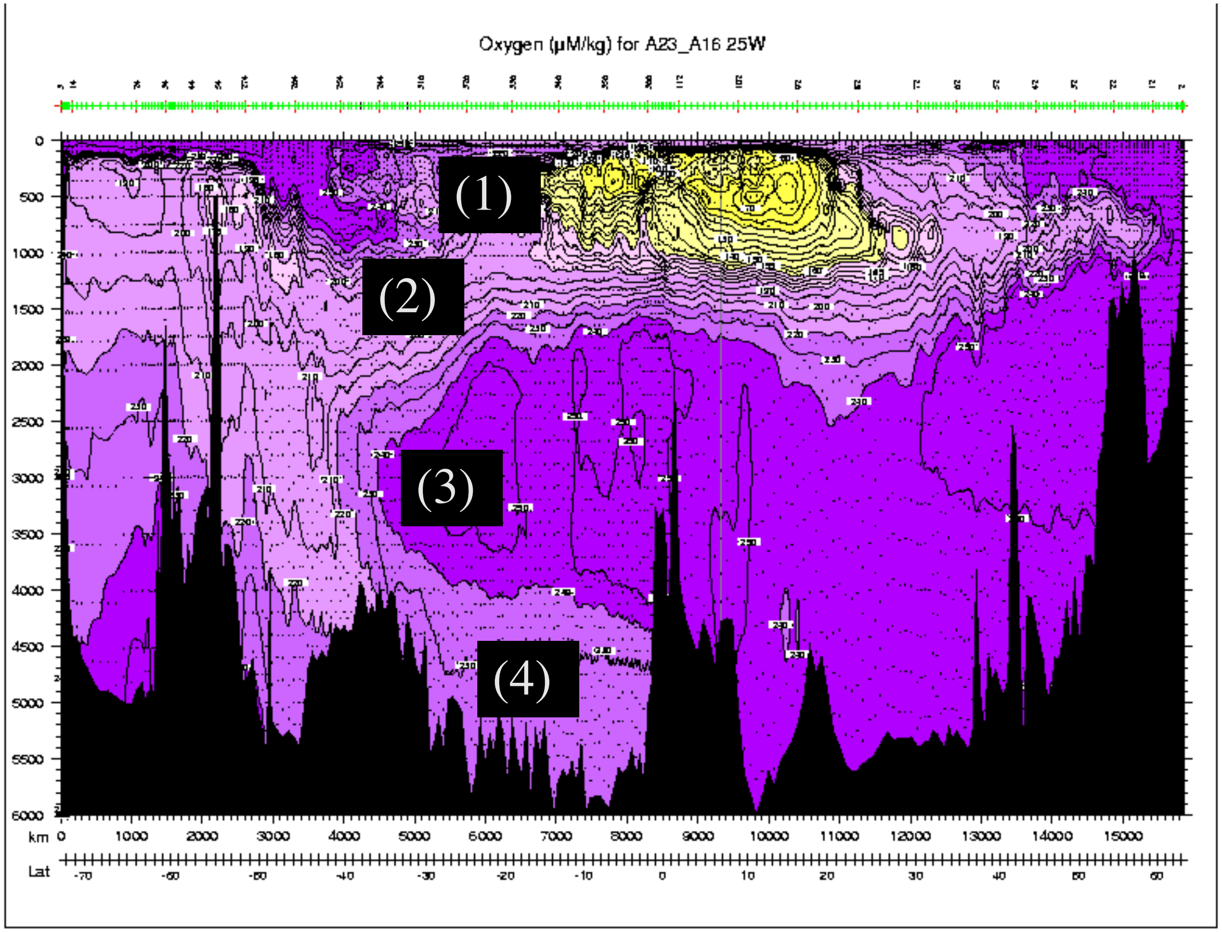
(3) High salinity

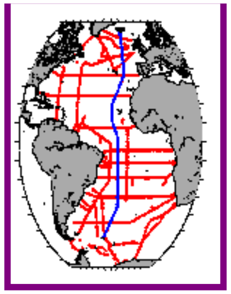
North Atlantic Deep
Water



(4) Low salinity
Antarctic bottom water

Oxygen in the Atlantic at 25W





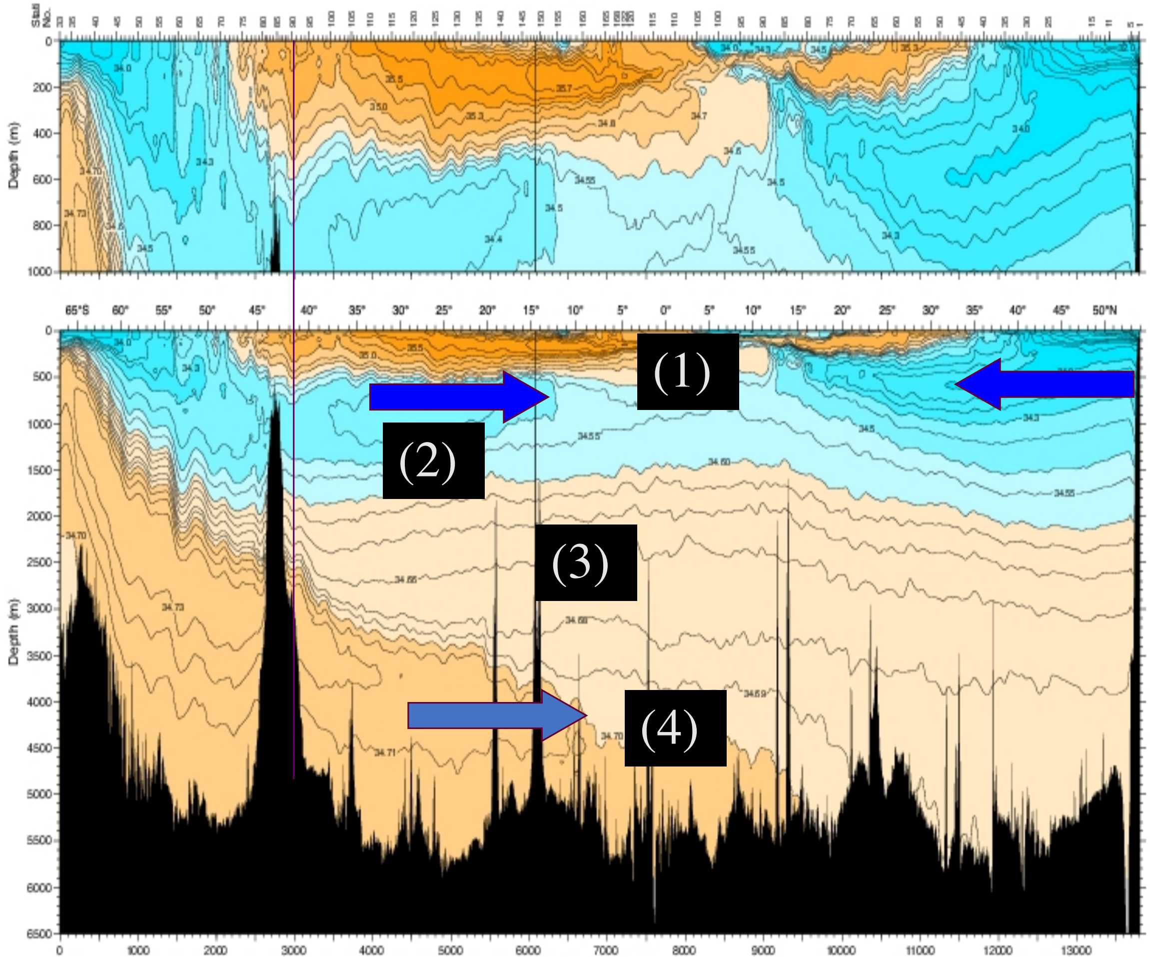
(1) Upper

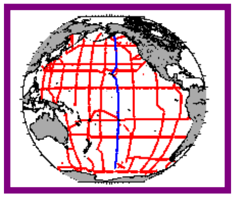
(2) AAIW
and LSW

(3) NADW

(4) AABW

Salinity in the Pacific (150° W)





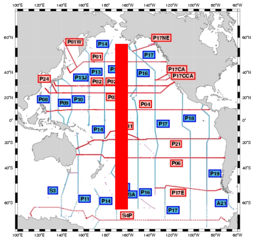
(1) Upper

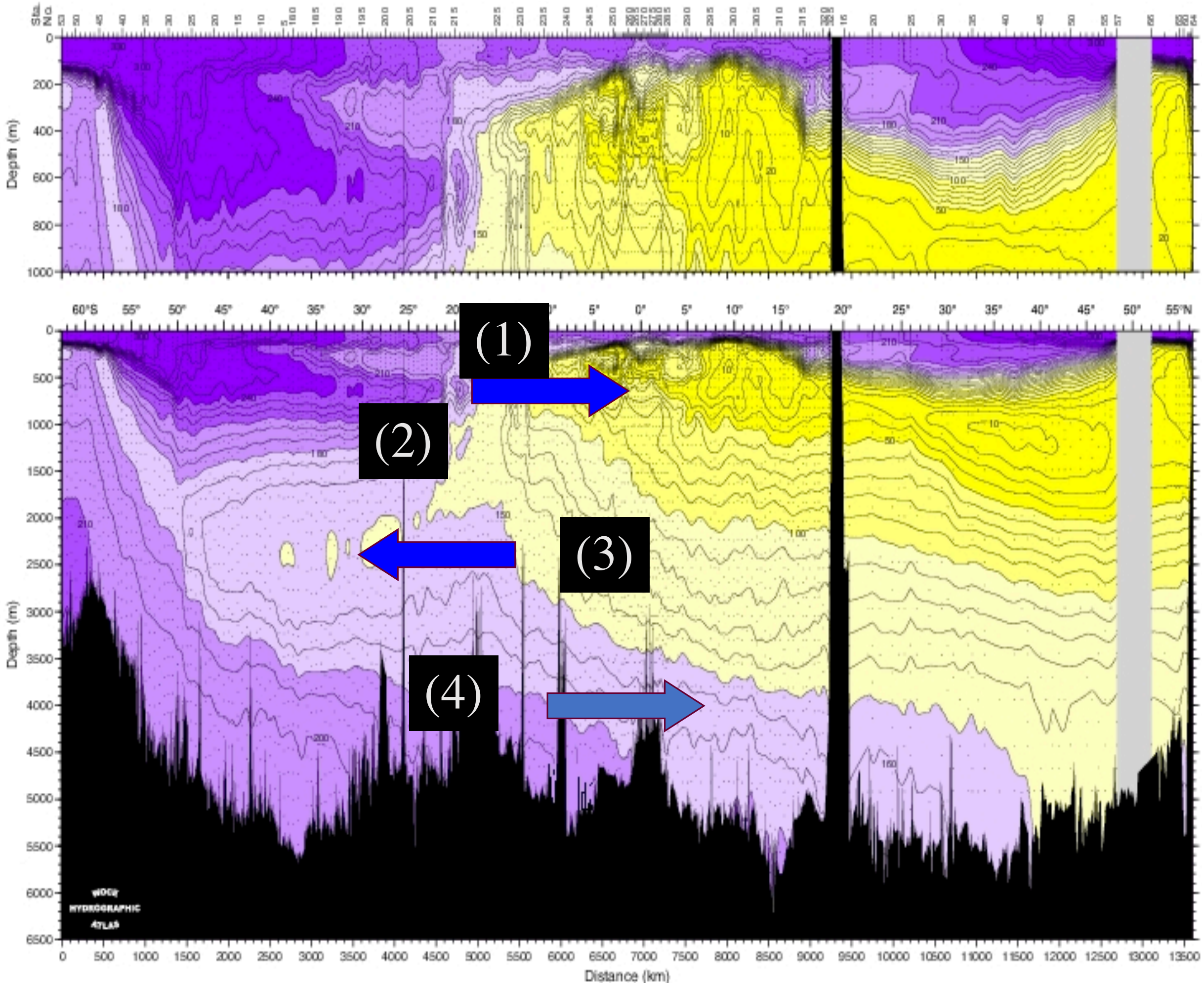
(2) AAIW and NPIW

(3) PDW

(4) LCDW (AABW)

Oxygen in mid-Pacific (150W)





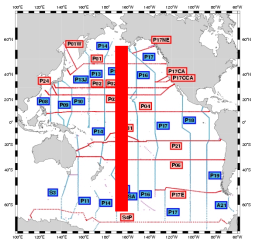
(1) Upper

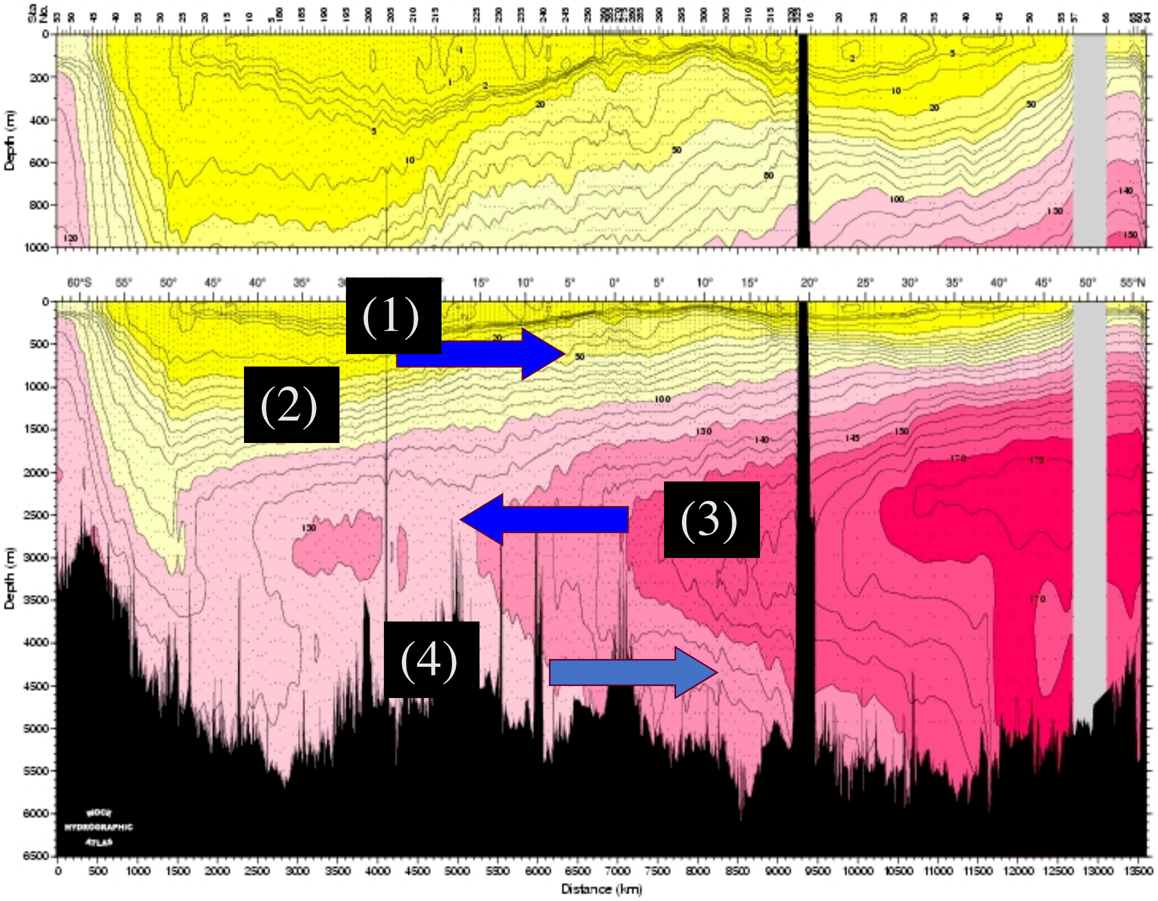
(2) AAIW and NPIW

(3) PDW

(4) LCBW (AABW)

Silicate in mid-Pacific (150W)





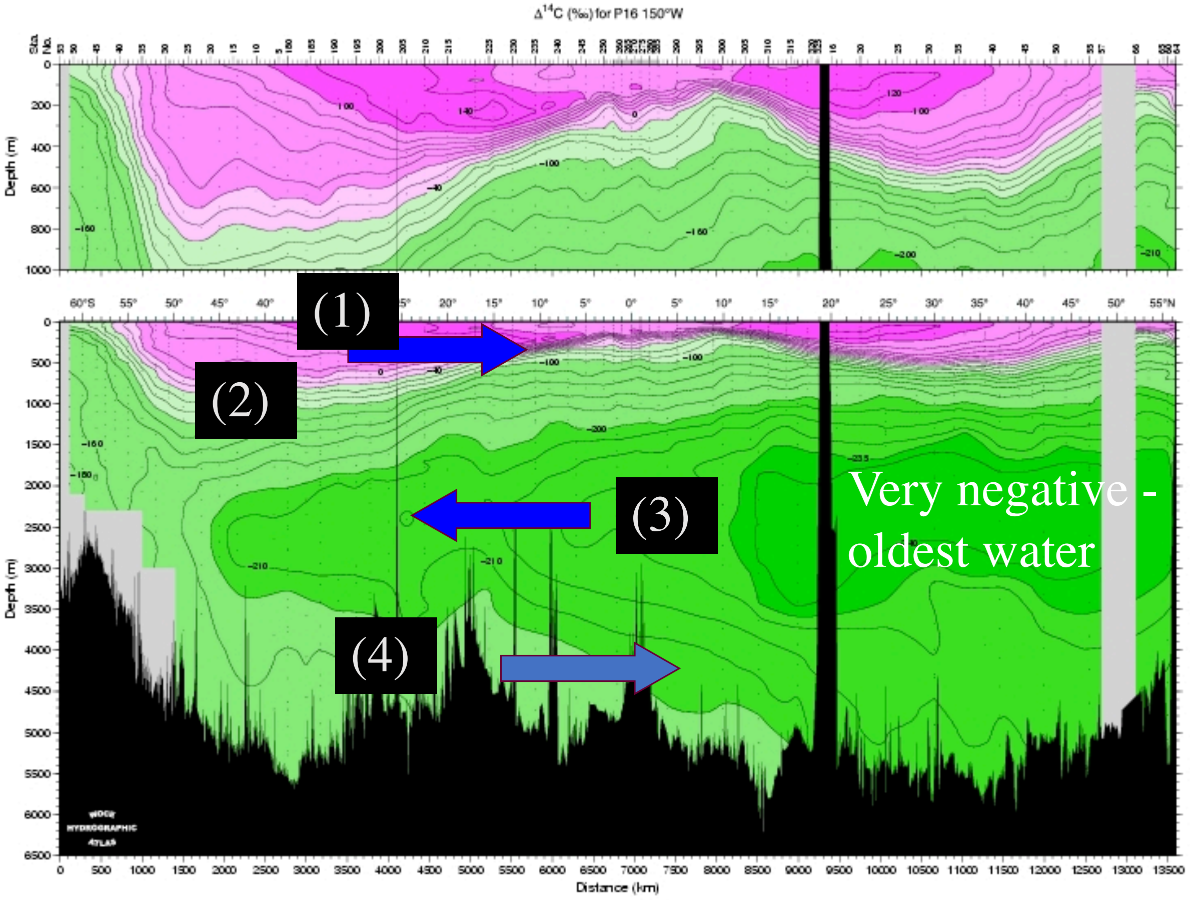
(1) Upper

(2) AAIW and NPIW

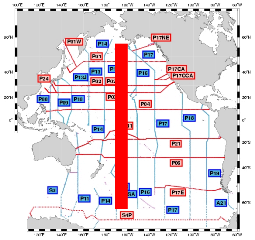
(3) PDW

(4) LCBW (AABW)

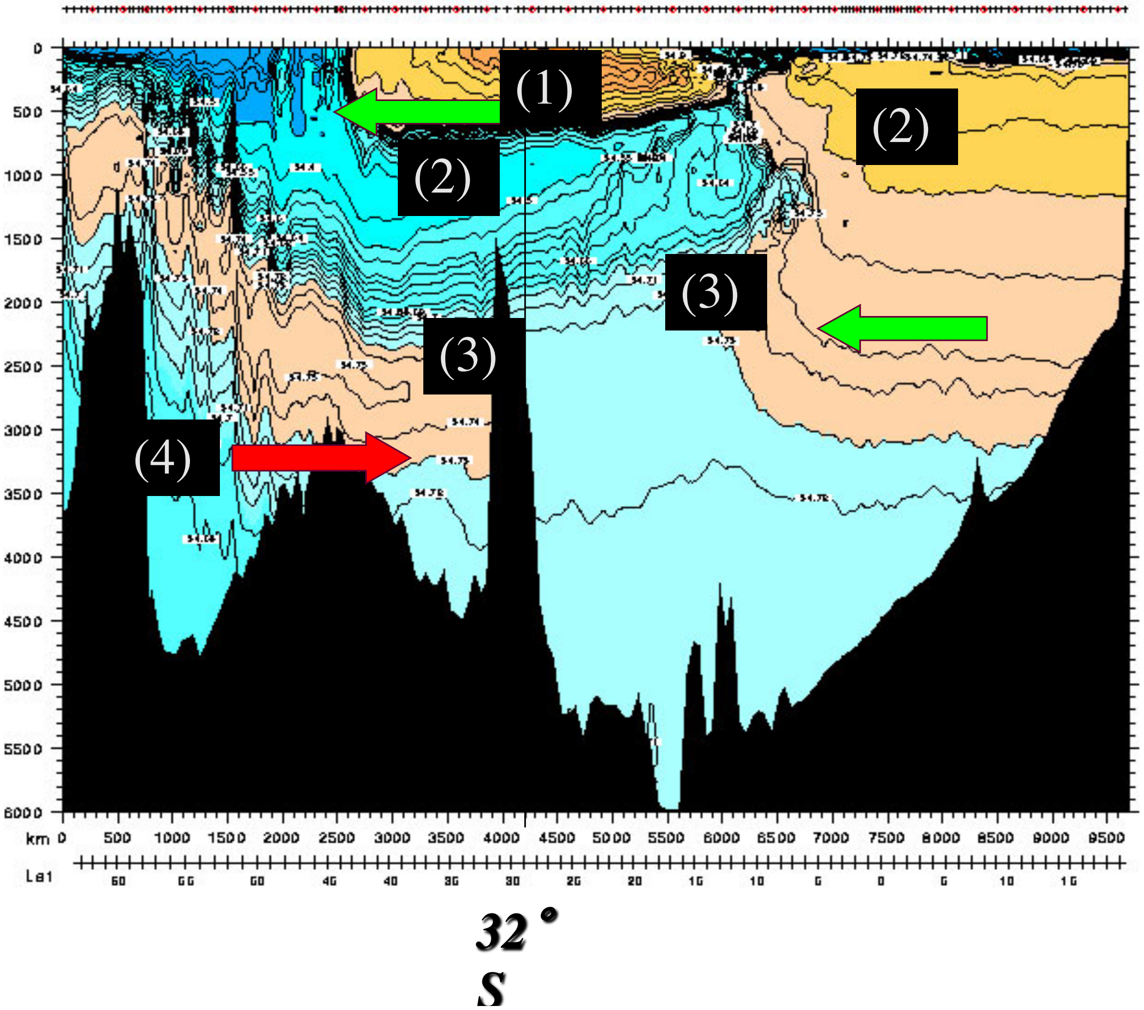
delC14 in mid-Pacific (150W)

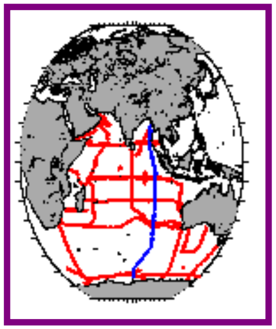


Very negative - oldest water



Eastern Indian water masses





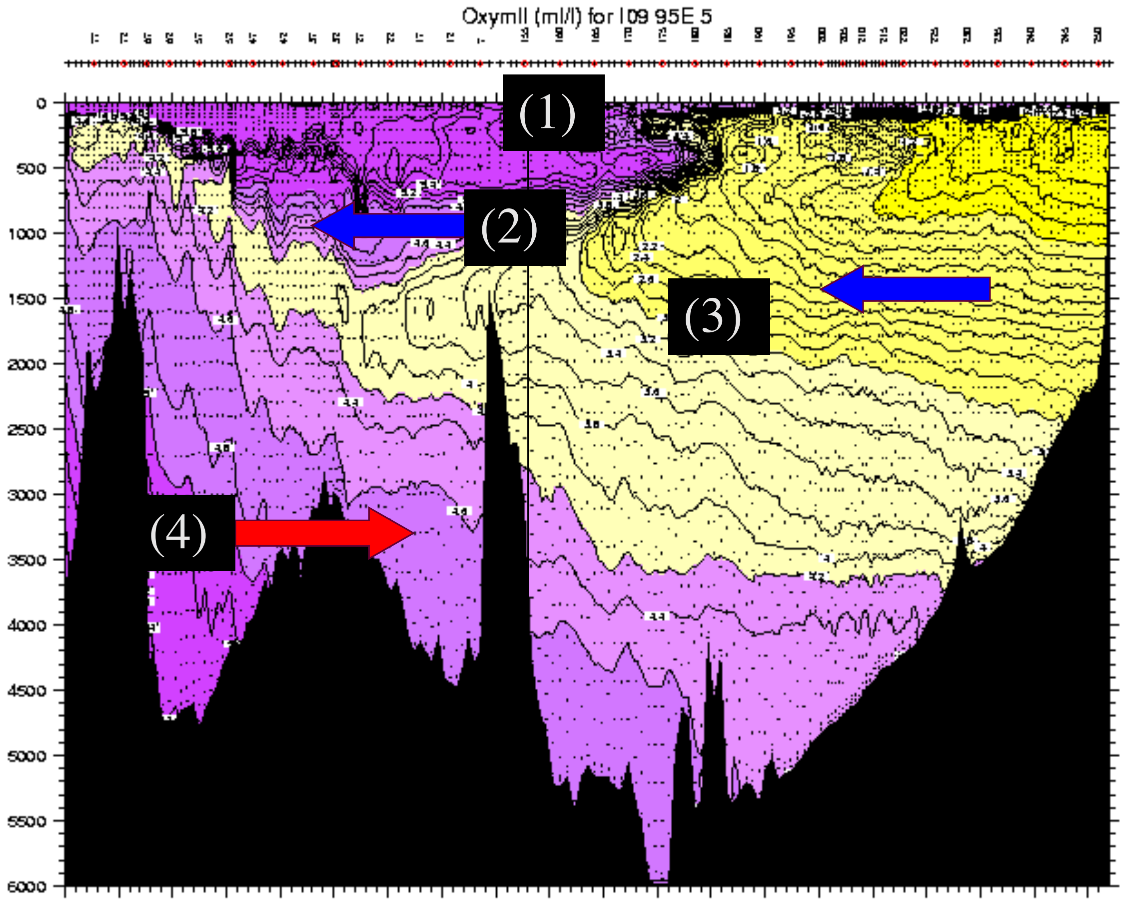
(1) Upper

(2) AAIW
and
RSW

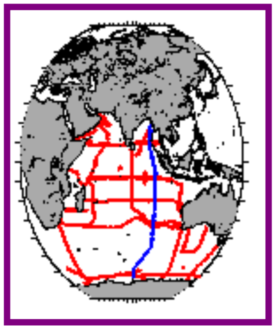
(3) NADW
and IDW

(4) LCBW
(AABW)

Eastern Indian water masses: oxygen



32° S



Lower oxygen: Red Sea Water and other northern Indian waters

Higher oxygen- Subantarctic Mode Water and Circumpolar Deep Water

Global deep water potential temperature-salinity

Worthington, 1982

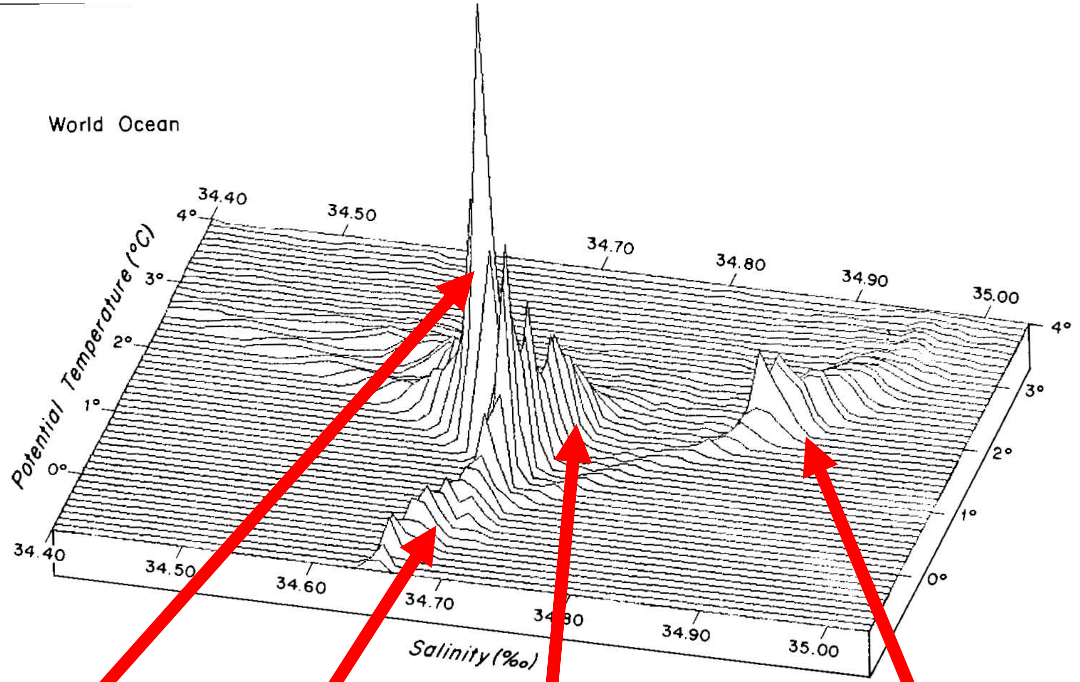


Figure 2.2 Simulated three-dimensional T-S diagram of the water masses of the world ocean. Apparent elevation is proportional to volume. Elevation of highest peak corresponds to 26.0 × 10^6 km^3 per bivariate class 0.1°C × 0.01‰.

Pacific
Deep Water

Antarctic
Bottom Water

Indian
Deep Water

North Atlantic
Deep Water