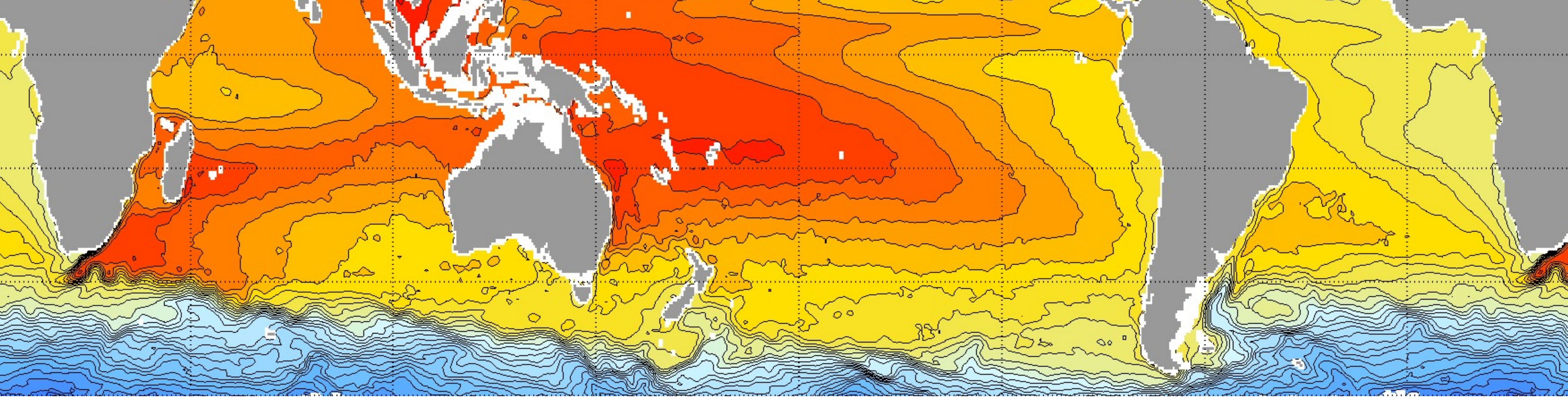


Diagnosing the mean global ocean overturning circulation and property budgets from hydrographic section data

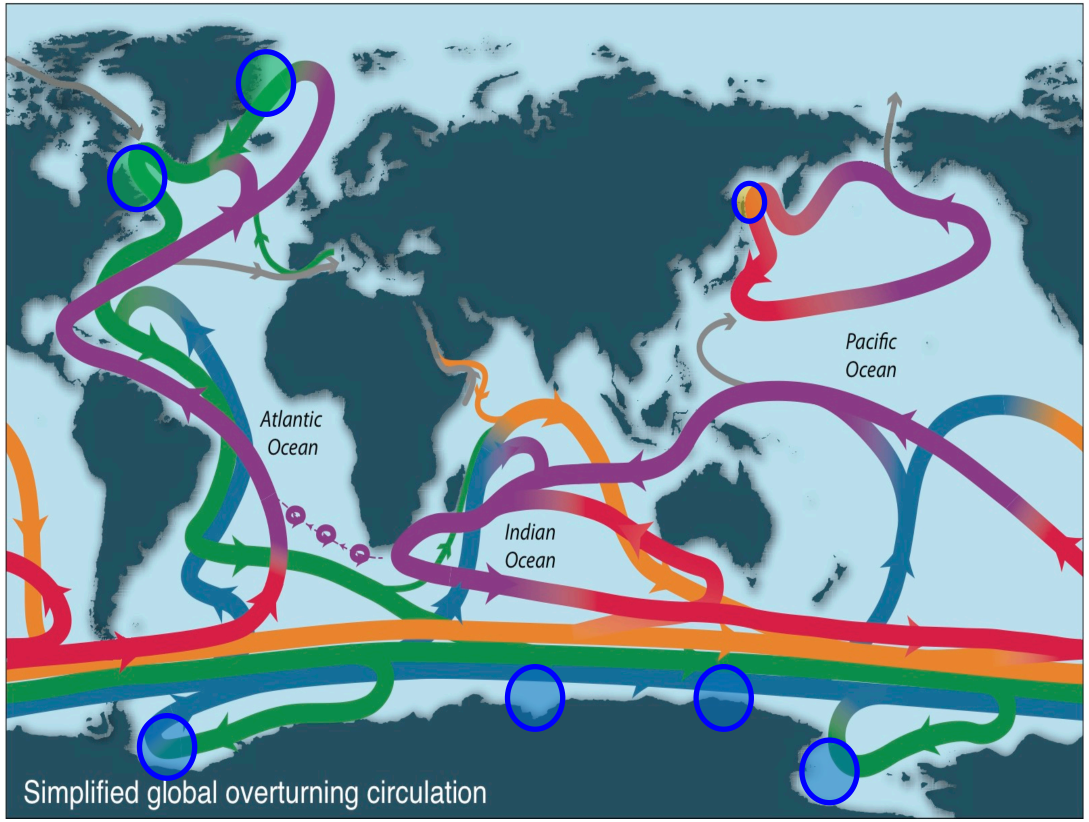
Lynne Talley, Professor

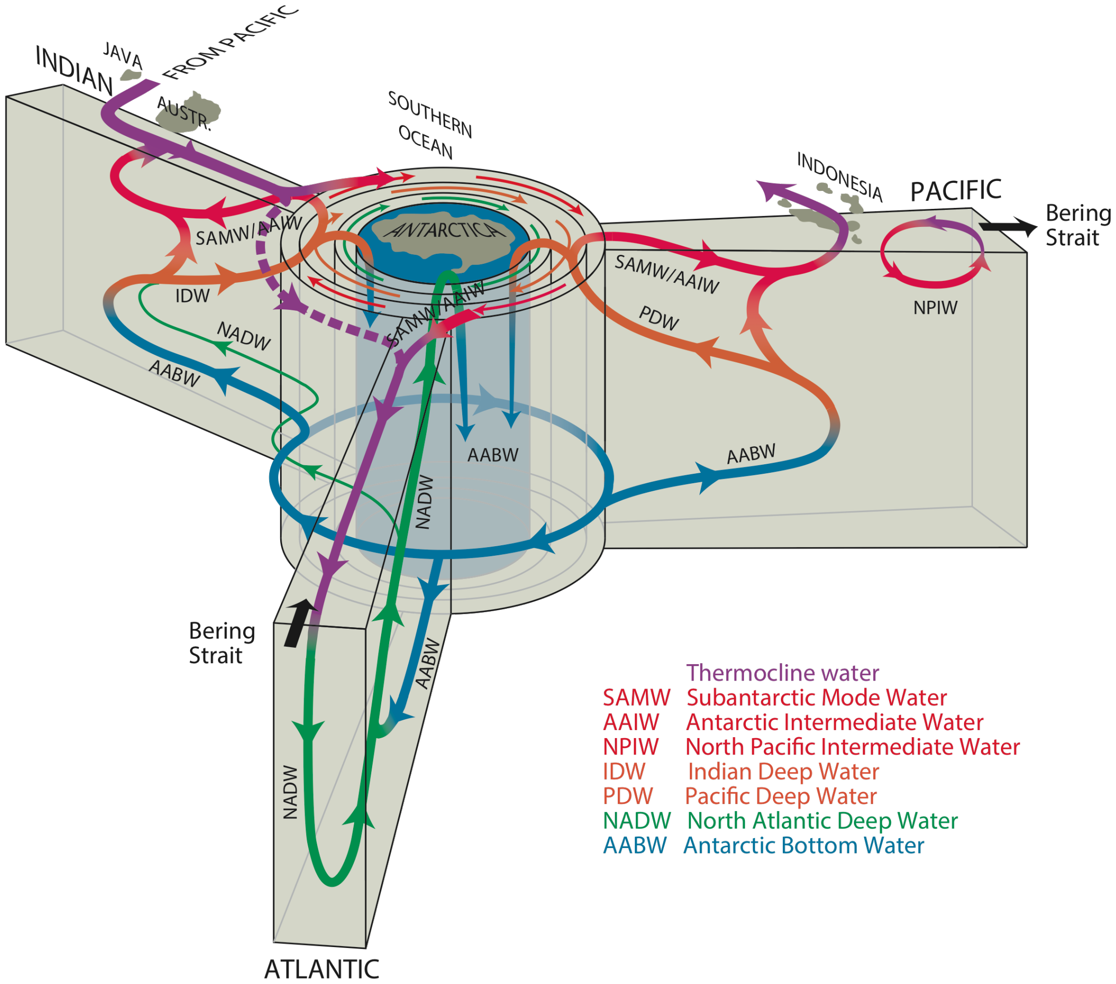
Scripps Institution of Oceanography, UCSD

7th Summer School on Theory, Mechanisms and Hierarchical Modelling of Climate Dynamics: ICTP Summer 2026
Estimating Ocean Transports: Single Sections, Box Models and Reanalysis Products

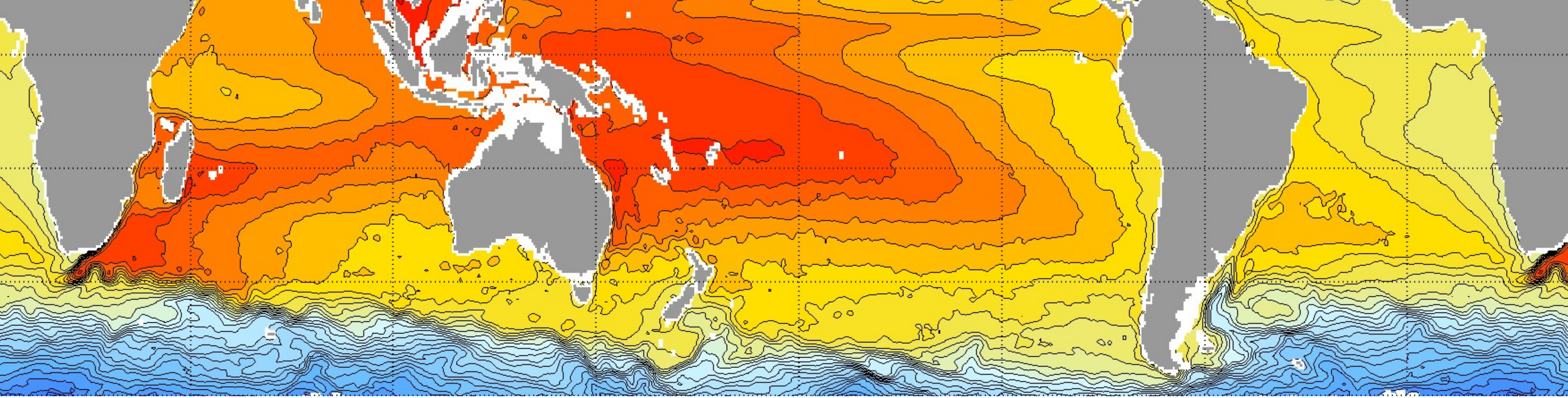
July 6, 2026

Goal: Building the Global Overturning Circulation





Talley (DPO 2011; Oceanography, 2013)



Diagnosing the mean global ocean overturning circulation and property budgets from hydrographic section data

Materials:

Reid (Progress in Oceanography, 1994, 1997, 2003)

Introduction to Descriptive Physical Oceanography textbook, 6th edition (Talley et al., 2011)

Ganachaud and Wunsch (2000) (global inverse)

Talley et al. (J. Clim. 2003) (overturning streamfunction)

Lumpkin and Speer (JPO 2007) (global inverse)

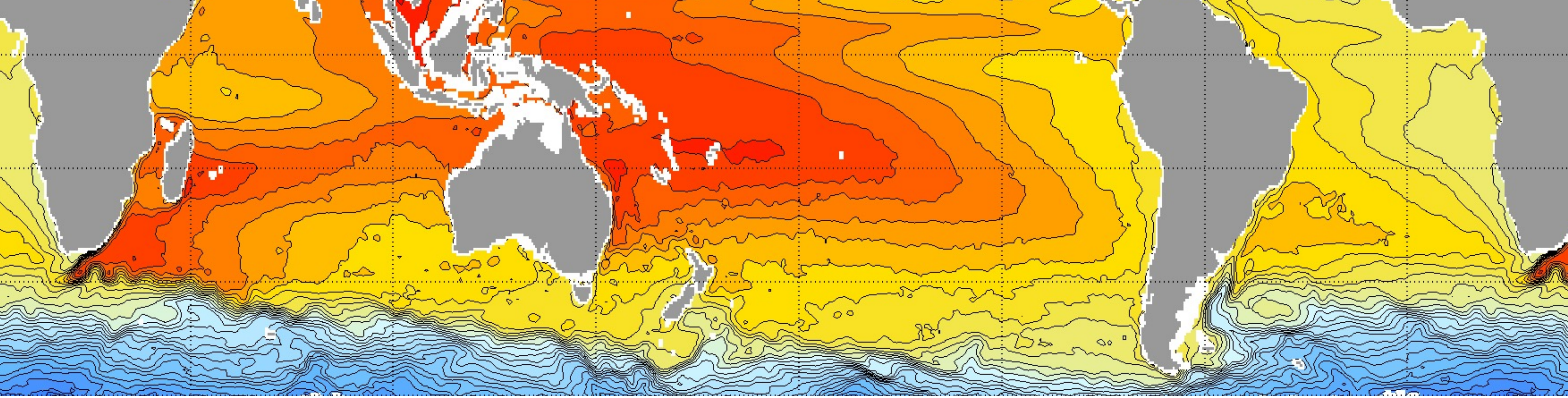
Talley (PiO 2008) [freshwater transports]

Macdonald et al. (PiO 2009) Pacific WOCE inverse

Talley (Oceanography 2013) [global overturning circulation]

Various papers covering hydrographic inverse model results

New work to be submitted on (observed) dynamics of global-scale overturning



Dynamics: Geostrophy and Ekman flow, reference velocities
Result: Observed global ocean circulation
Result: Meridional water mass distributions
Method: Transport and Overturning circulation calculations
Result: Global overturning circulation mass transports
Result: Global heat, freshwater transports
Method: Budgets and implications for diffusivity
Result: Diapycnal diffusivity estimates
Dynamics: overturning circulation at large spatial scales
Result: Demonstration of large-scale force balance for GOC

Elements for calculating ocean transports, specifically for inverse models:
Geostrophic velocity and Ekman transport

Complete force balance with rotation (equations)

Inertial motion

Geostrophic flow

x: $\partial u/\partial t + u\, \partial u/\partial x + v\, \partial u/\partial y + w\, \partial u/\partial z - fv =$
$- (1/\rho)\partial p/\partial x + \partial/\partial x(A_H \partial u/\partial x) +$
$\partial/\partial y(A_H \partial u/\partial y) + \partial/\partial z(A_V \partial u/\partial z)$ (7.11a)

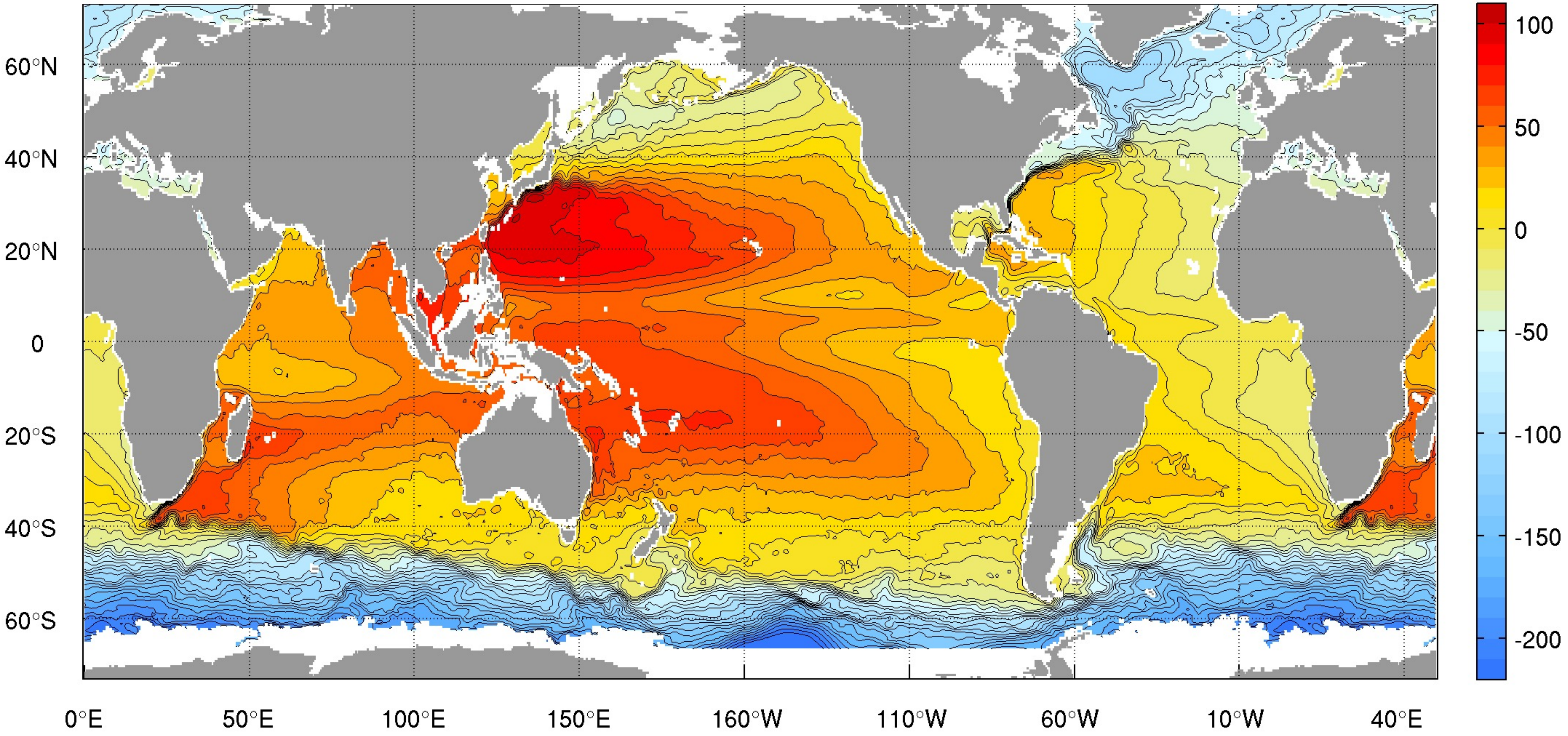
y: $\partial v/\partial t + u\, \partial v/\partial x + v\, \partial v/\partial y + w\, \partial v/\partial z + fu =$
$- (1/\rho)\partial p/\partial y + \partial/\partial x(A_H \partial v/\partial x) +$
$\partial/\partial y(A_H \partial v/\partial y) + \partial/\partial z(A_V \partial v/\partial z)$ (7.11b)

z: $\partial w/\partial t + u\, \partial w/\partial x + v\, \partial w/\partial y + w\, \partial w/\partial z$ (+neglected small Coriolis) =
$- (1/\rho)\partial p/\partial z - g + \partial/\partial x(A_H \partial w/\partial x) +$
$\partial/\partial y(A_H \partial w/\partial y) + \partial/\partial z(A_V \partial w/\partial z)$ (7.11c)

Hydrostatic

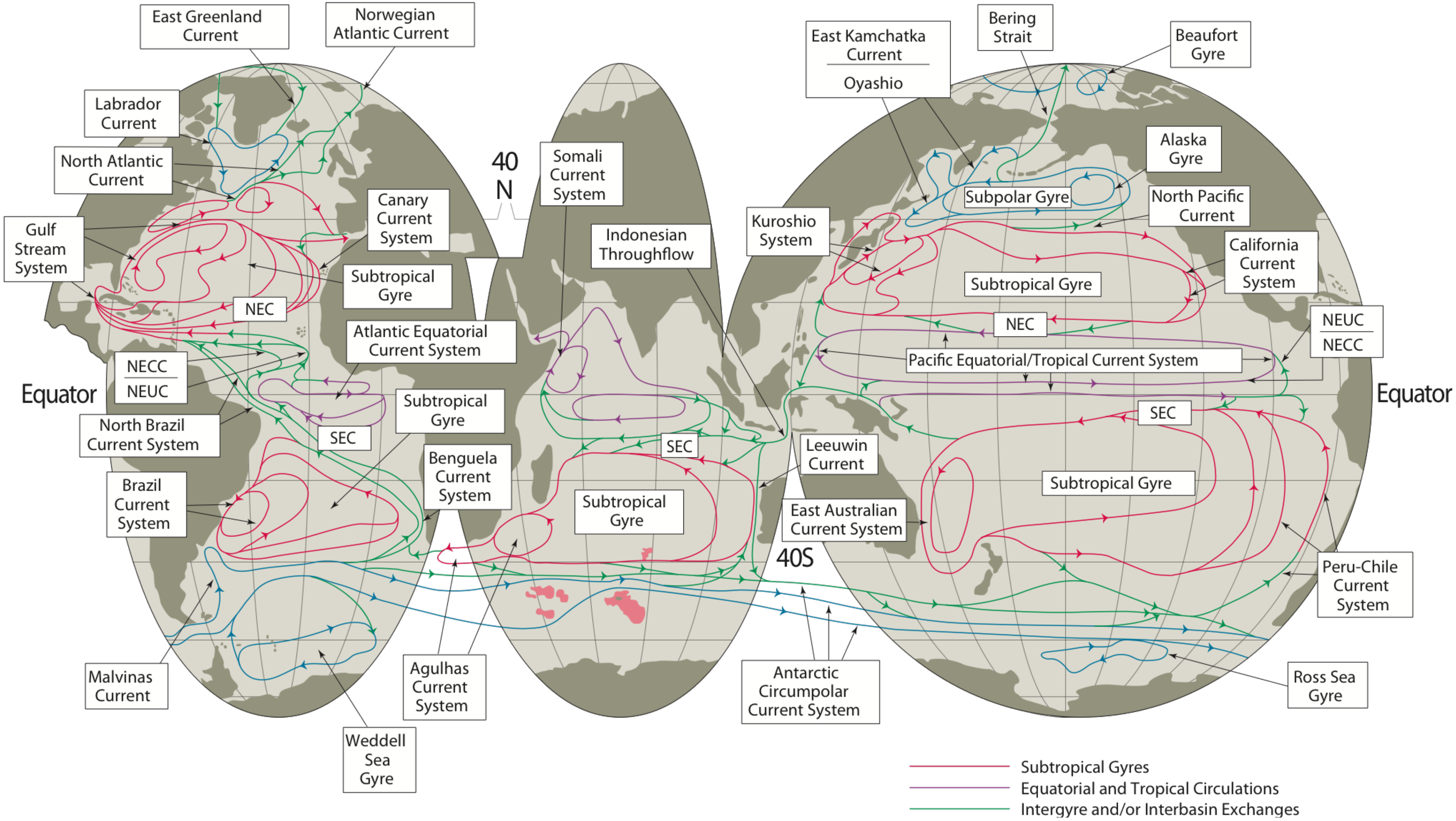
(where g contains both actual g and the effect of centrifugal force)

Absolute surface height, related to surface geostrophic velocity



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

Upper ocean circulation: current systems



Schmitz (1995) (DPO Fig. 14.1)

Velocity profiles from thermal wind

Thermal wind: The vertical shear in geostrophic velocity is proportional to the horizontal density derivative.

$$- f\, \partial v/\partial z = (g/\rho_o)\, \partial \rho/\partial x$$

$$f\, \partial u/\partial z = (g/\rho_o)\, \partial \rho/\partial y$$

Integrate in z

$$- f \int (\partial v/\partial z)\, dz = (g/\rho_o) \int (\partial \rho/\partial x)\, dz$$

$$f \int (\partial u/\partial z)\, dz = (g/\rho_o) \int (\partial \rho/\partial y)\, dz$$

Velocity at one depth z_2 relative to that at depth z_1

$$\boxed{\begin{aligned} - f\, (v_2 - v_1) &= (g/\rho_o)\, \partial/\partial x\left(\int \rho\, dz \right) \\ f\, (u_2 - u_1) &= (g/\rho_o)\, \partial/\partial y\left(\int \rho\, dz \right) \end{aligned}}$$

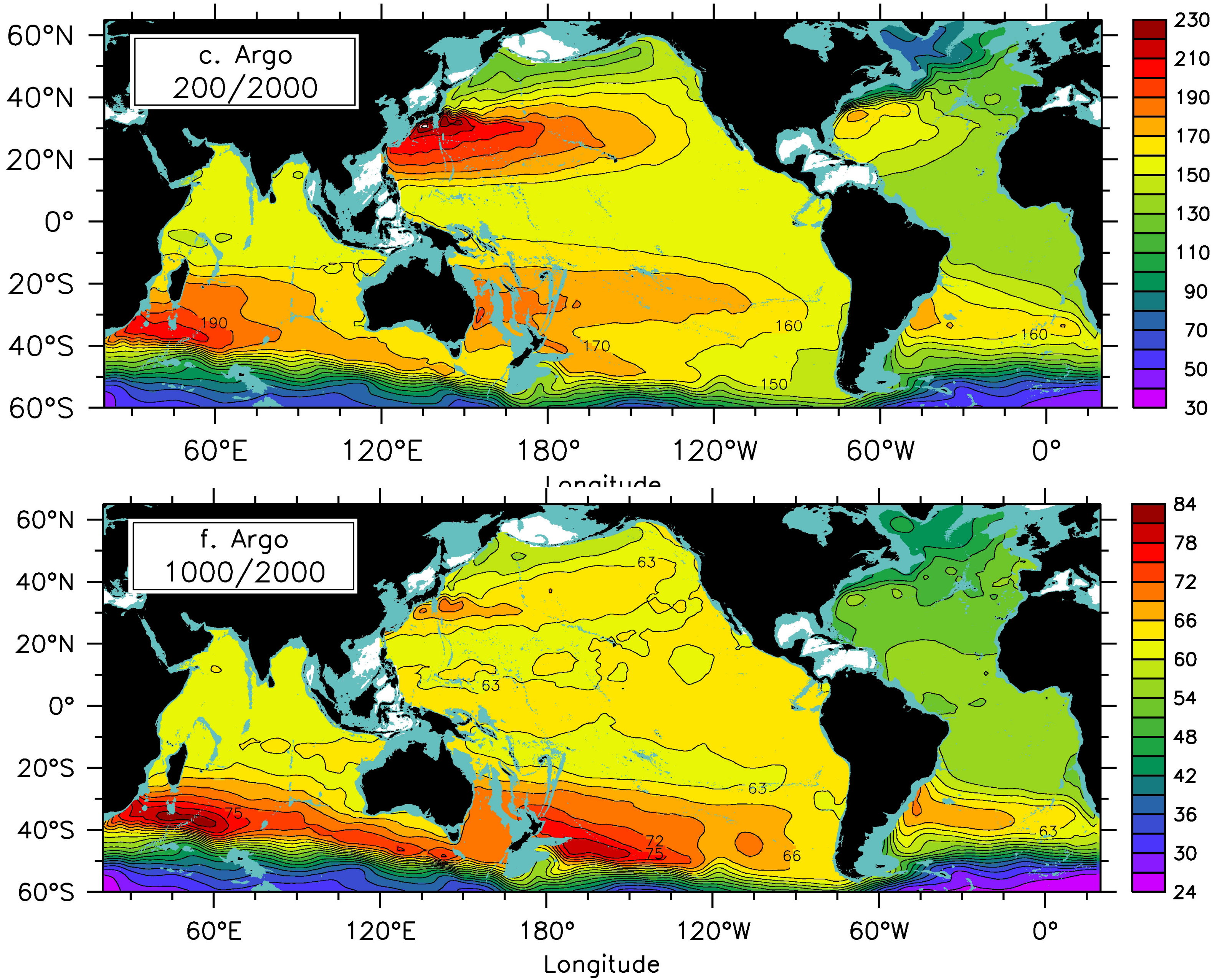
Reference velocity for thermal wind balance: your objective in this summer school through inverse modeling!

How to choose reference velocities v_1 and u_1

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 for weak deep flows, hence not useful for Global Overturning Circulation.

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

200 and 1000 dbar circulation rel. to 2000 dbar (level of NO motion example)



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

Roemmich and Gilson, 2009
DPO Fig. 14.3

Complete force balance with rotation

Ekman balance

x: $\partial u/\partial t + u\,\partial u/\partial x + v\,\partial u/\partial y + w\,\partial u/\partial z - fv =$
$- (1/\rho)\partial p/\partial x + \partial/\partial x(A_H \partial u/\partial x) +$
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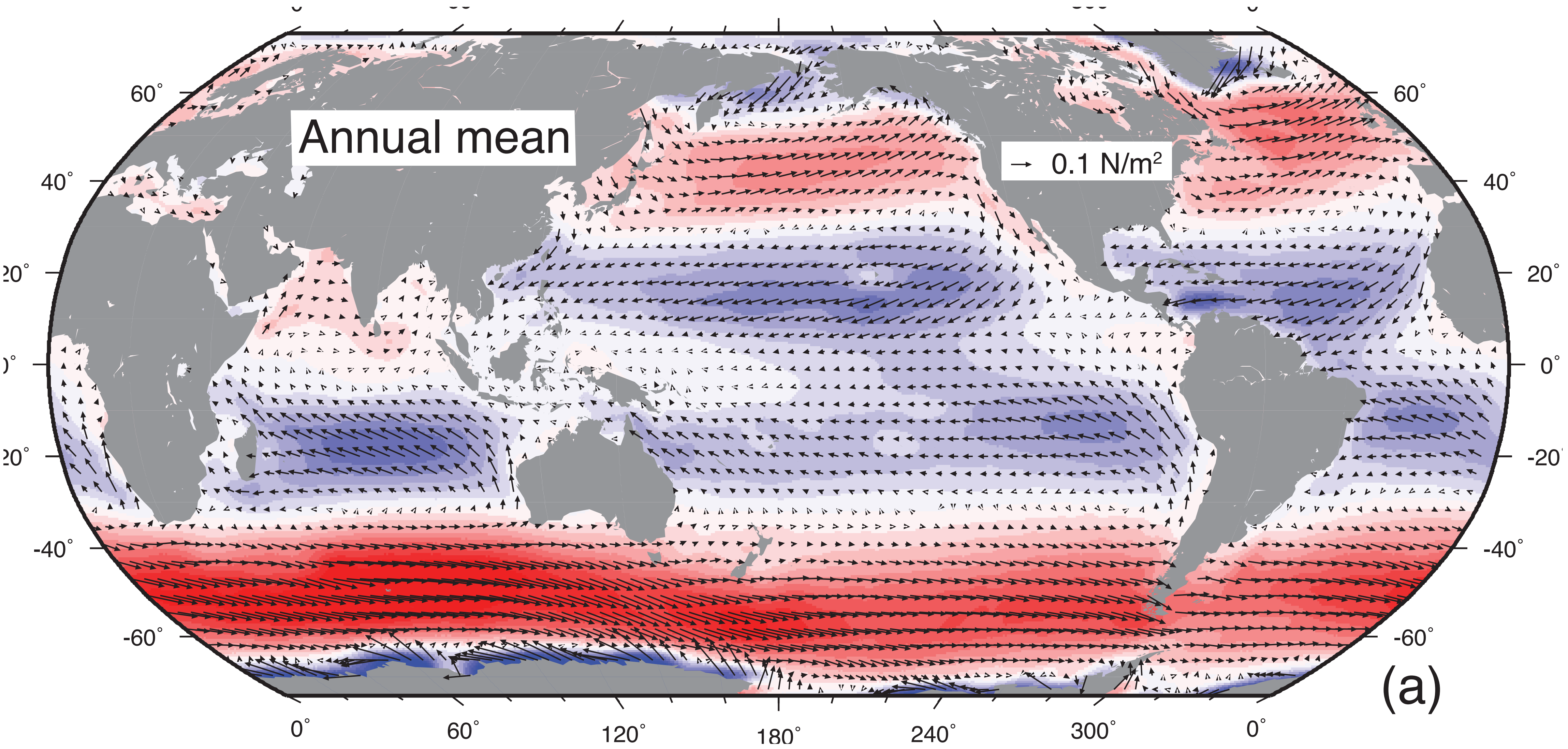
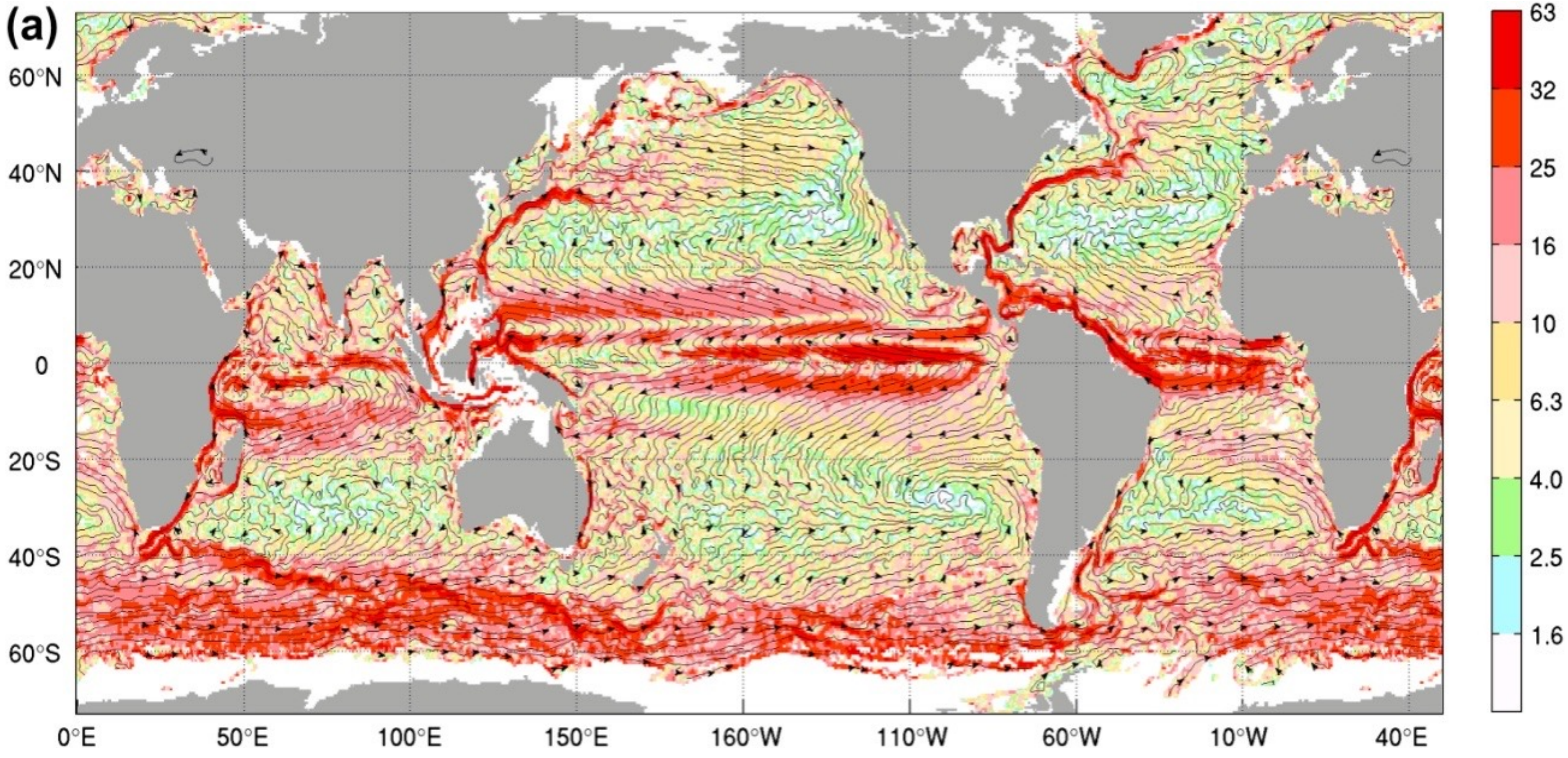
$\partial/\partial y(A_H \partial w/\partial y) + \partial/\partial z(A_V \partial w/\partial z)$ (7.11c)

Hydrostatic (doesn't figure in in Ekman)

Surface wind stress

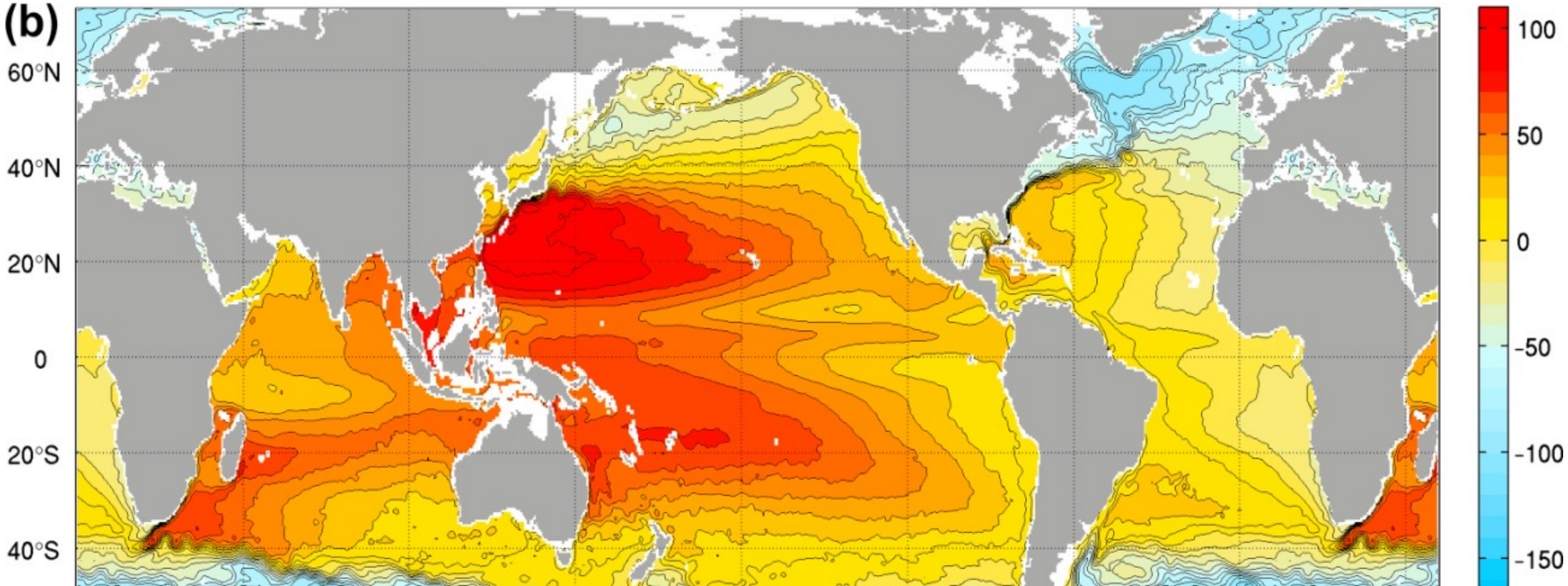


DPO Fig. 5.16a



(a) Surface velocity streamlines, **including both geostrophic and Ekman components;** color is the mean speed in cm/sec.

Calculated from surface drifters drogued at 15 m.

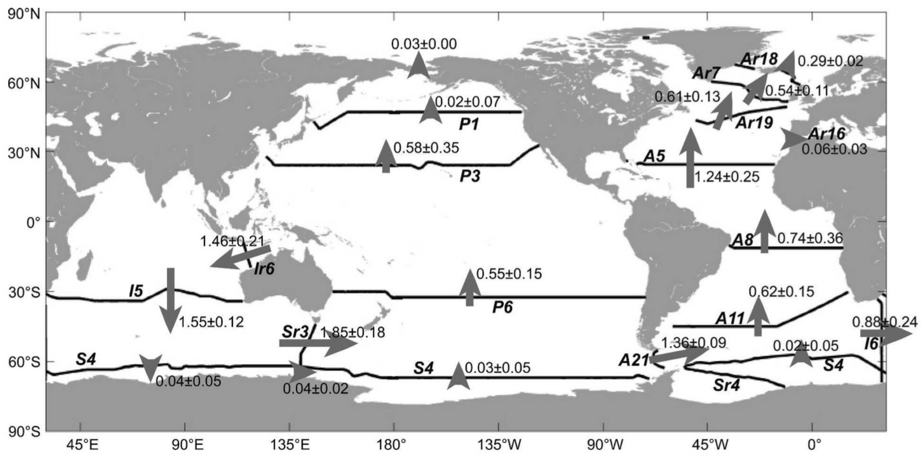


(a) Surface dynamic topography (dyn cm), with 10 cm contour intervals. Source: From **Maximenko et al. (2009).**

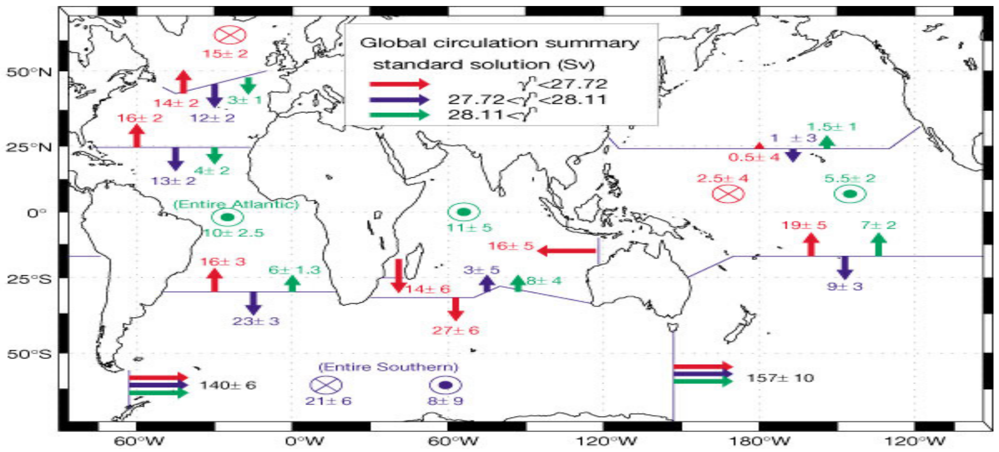
DPO FIGURE 14.2

Inverse models shown here, among others

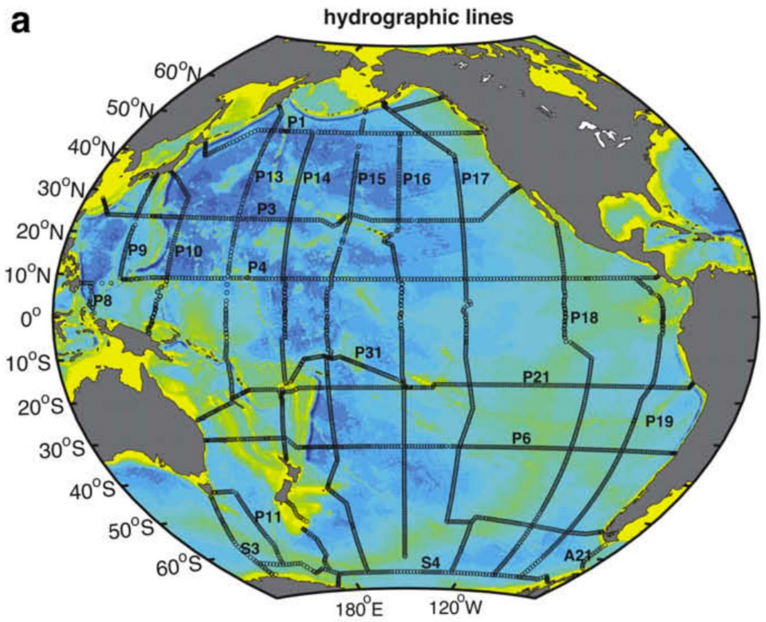
Lumpkin and Speer (2007) Global



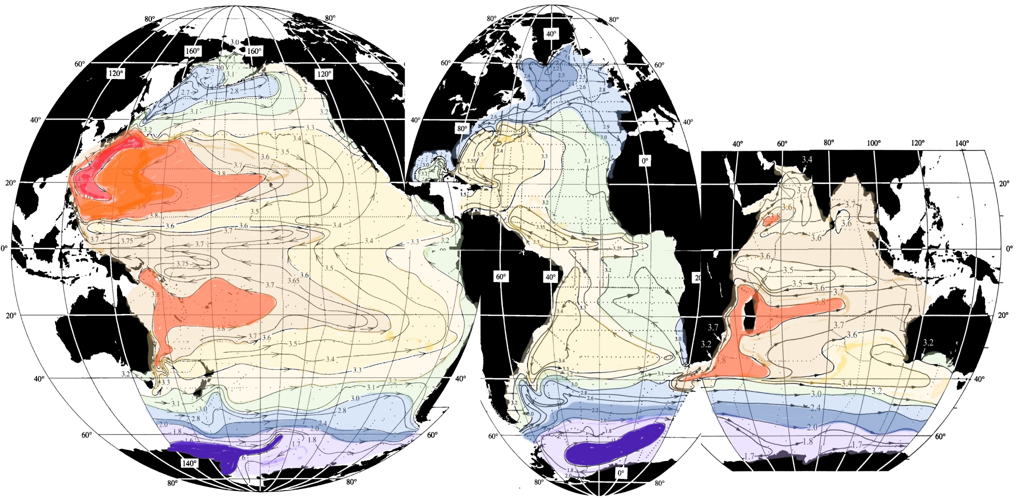
Ganachaud and Wunsch (2000) global



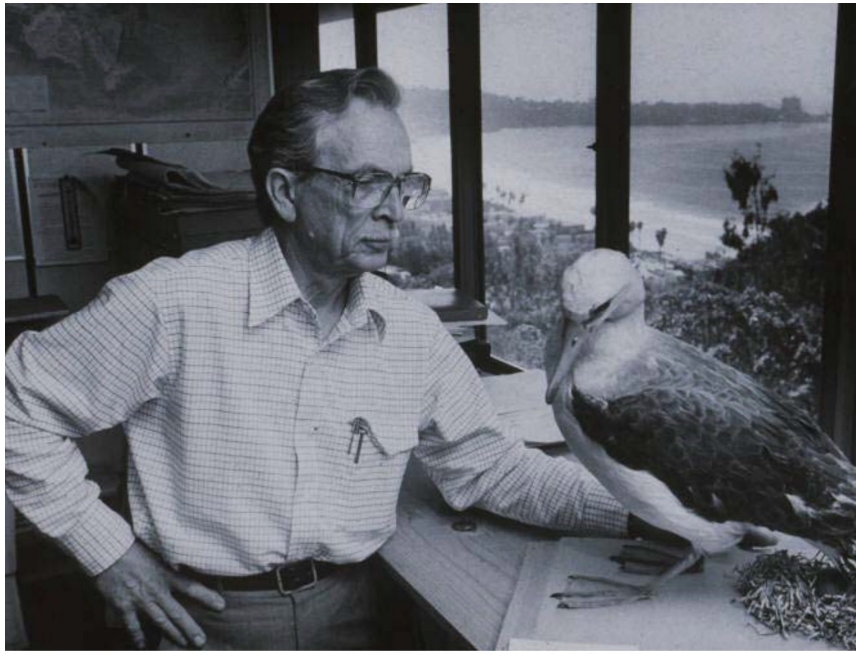
Macdonald et al. (2009) Pacific



Reid (1994, 1997, 2003) all basins



Global absolute geostrophic circulation mapping from Reid: A manual version of inverse modeling for hydrographic data



Joe Reid and the Albatross Award, 1988

1923-2015

On the total geostrophic circulation of the South Atlantic Ocean: Flow patterns, tracers, and transports

JOSEPH L. REID

Progress in Oceanography (1989)

On the total geostrophic circulation of the North Atlantic Ocean: Flow patterns, tracers, and transports

Progress in Oceanography (1994)

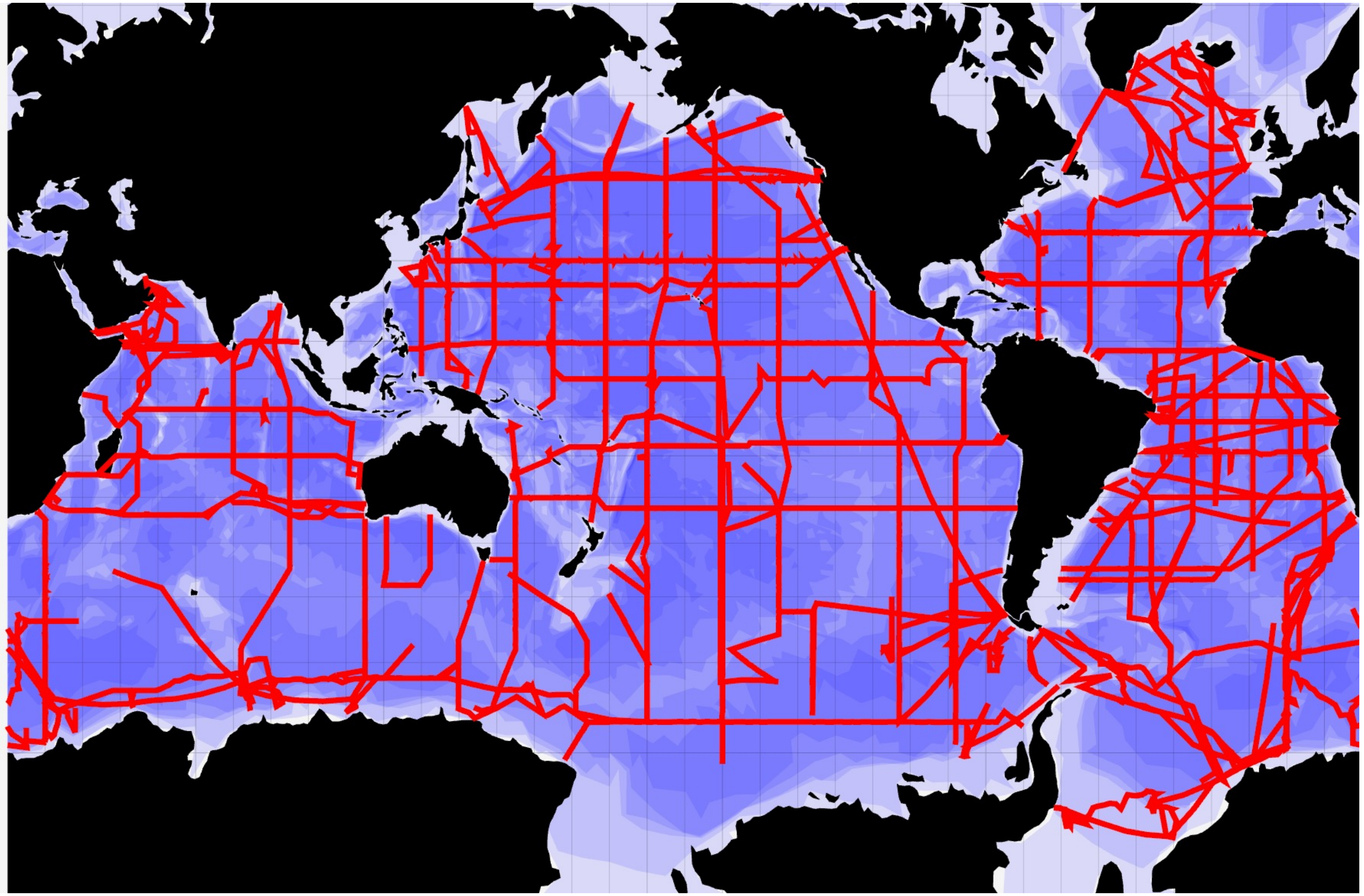
On the total geostrophic circulation of the pacific ocean: flow patterns, tracers, and transports

Progress in Oceanography (1997)

On the total geostrophic circulation of the Indian Ocean: flow patterns, tracers, and transports

Progress in Oceanography (2003)

Ship-based hydrographic observations with tracers



CCHDO WOCE One-Time survey

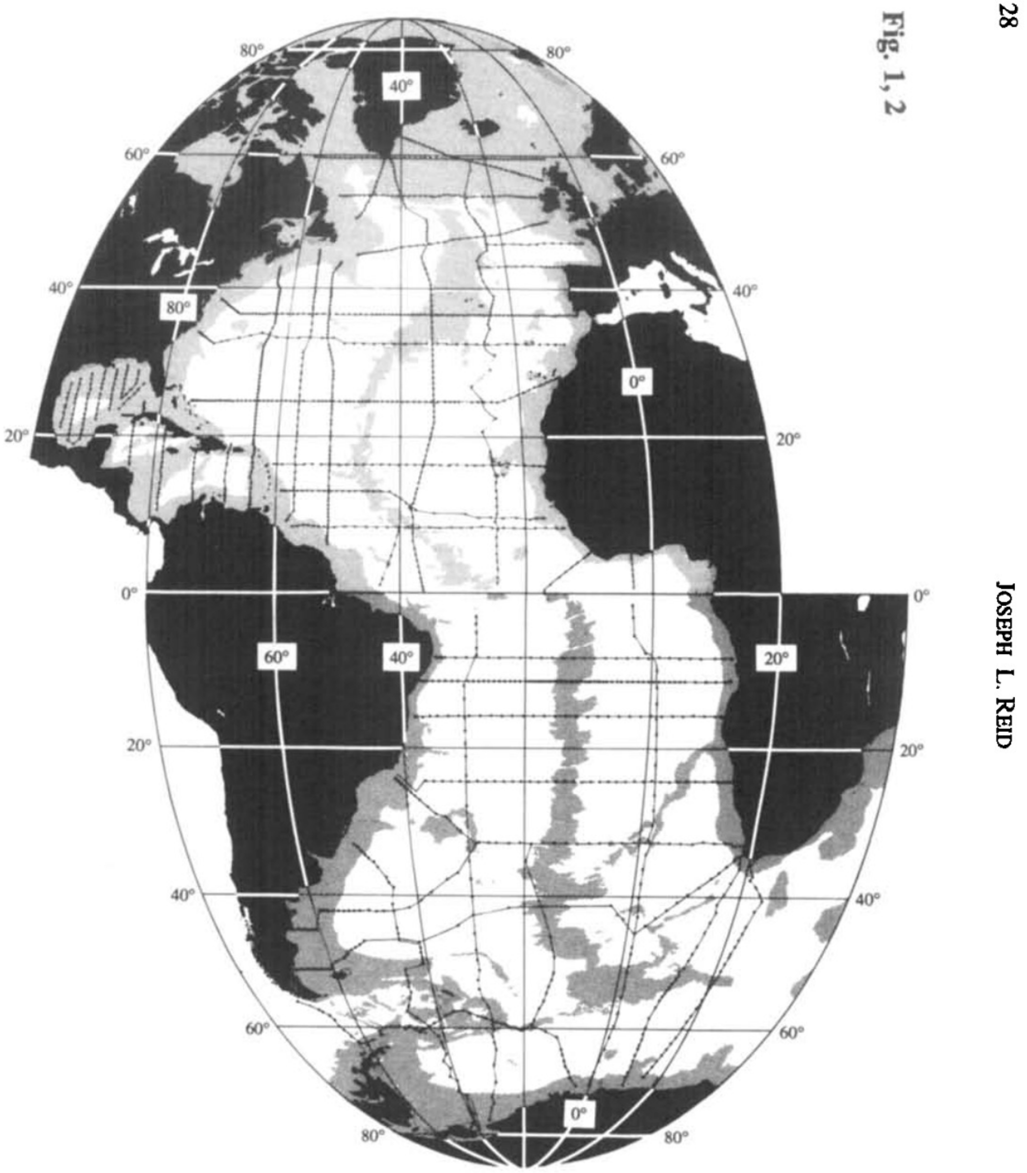
World Ocean Circulation Experiment, GO-SHIP and previous expeditions (IGY)

For steric height:
Temperature, salinity

For tracer patterns on isopycnals:
Oxygen
Nitrate, phosphate, silica

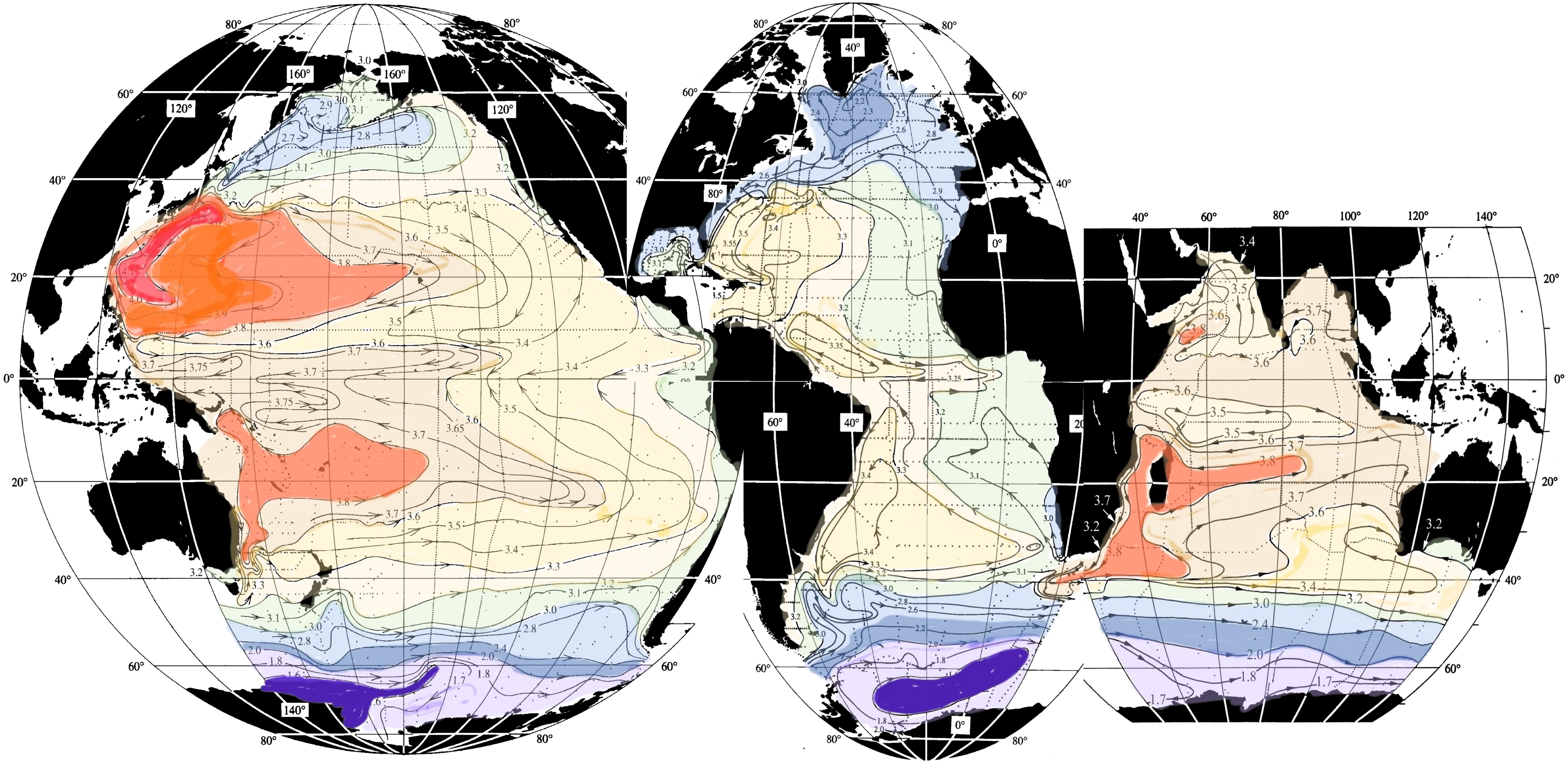
Reid circulation method

- Hydrographic profiles to ocean bottom (almost all)
- Assembled along lines
- Geostrophic flow relative to ocean bottom calculated for each station pair
- Flow compared with multiple tracer patterns on isopycnals; bottom reference velocity adjusted to be coherent with tracer patterns on potential density surfaces (0, 2000, 4000 dbar)
- Adjusted pressure gradients integrated horizontally from coast to obtain adjusted steric height
- Transport constraints on major currents applied



Example from Reid (1994)

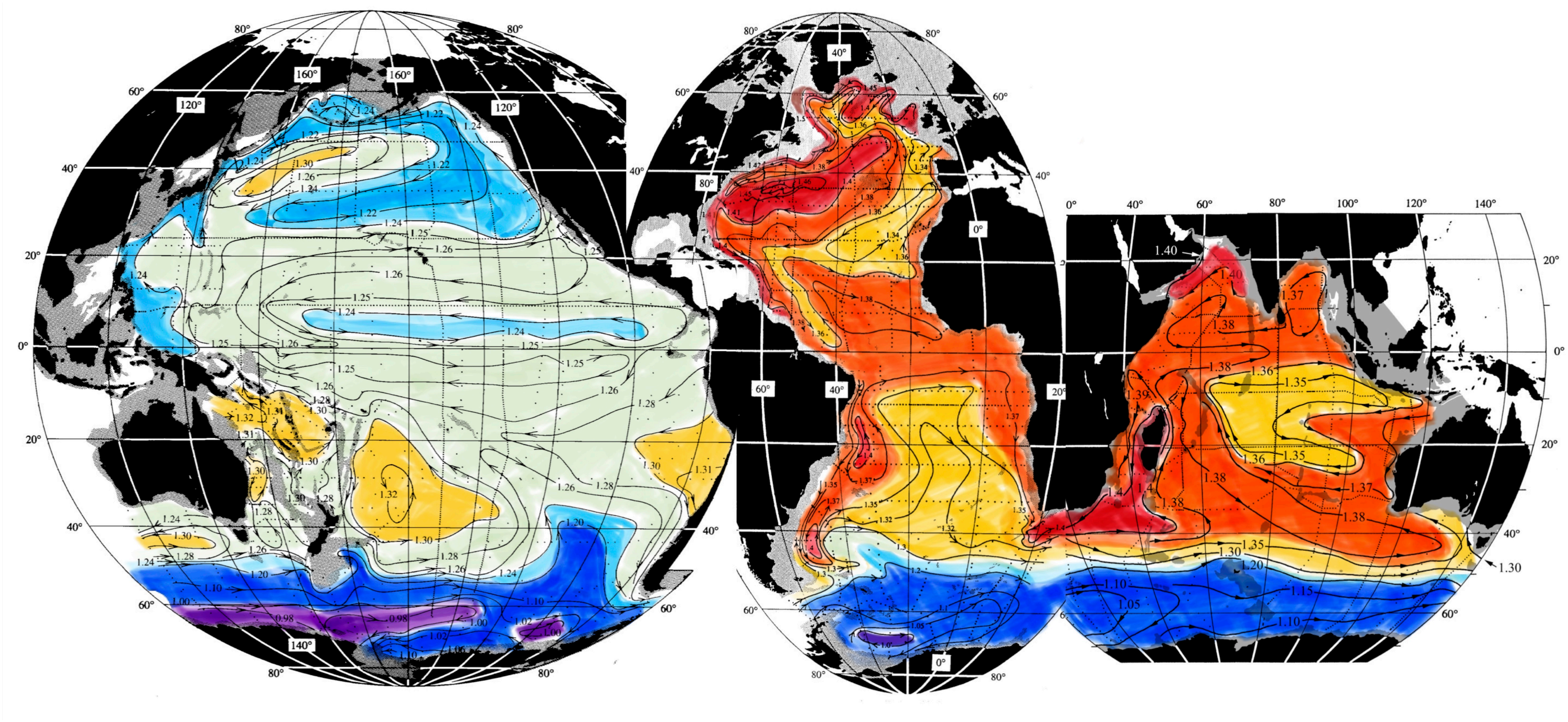
Surface circulation (absolute steric height from hydrography)



Similar to Maximenko et al on previous slide

(Reid, 1994, 1997, 2003)

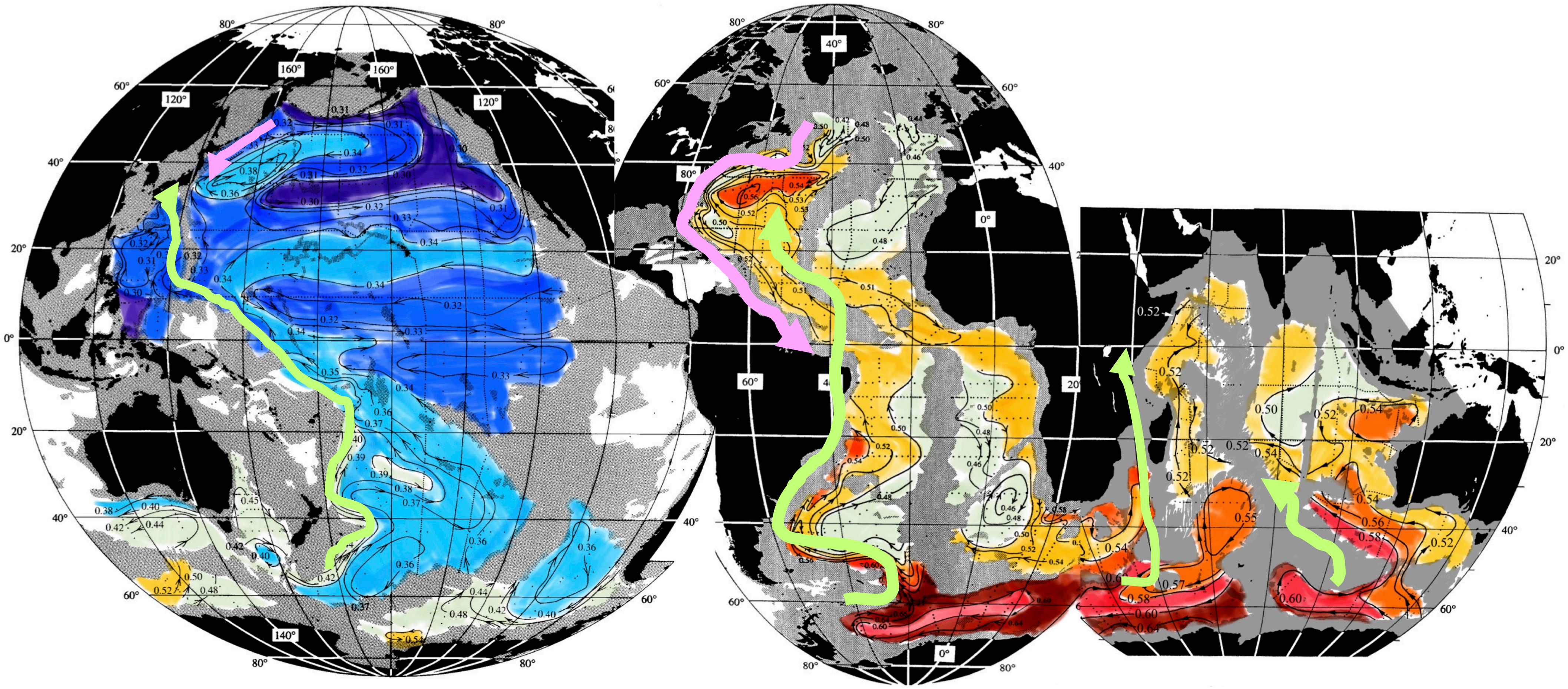
Circulation at 2000 dbar (adjusted steric height)



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

Water mass overview: 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

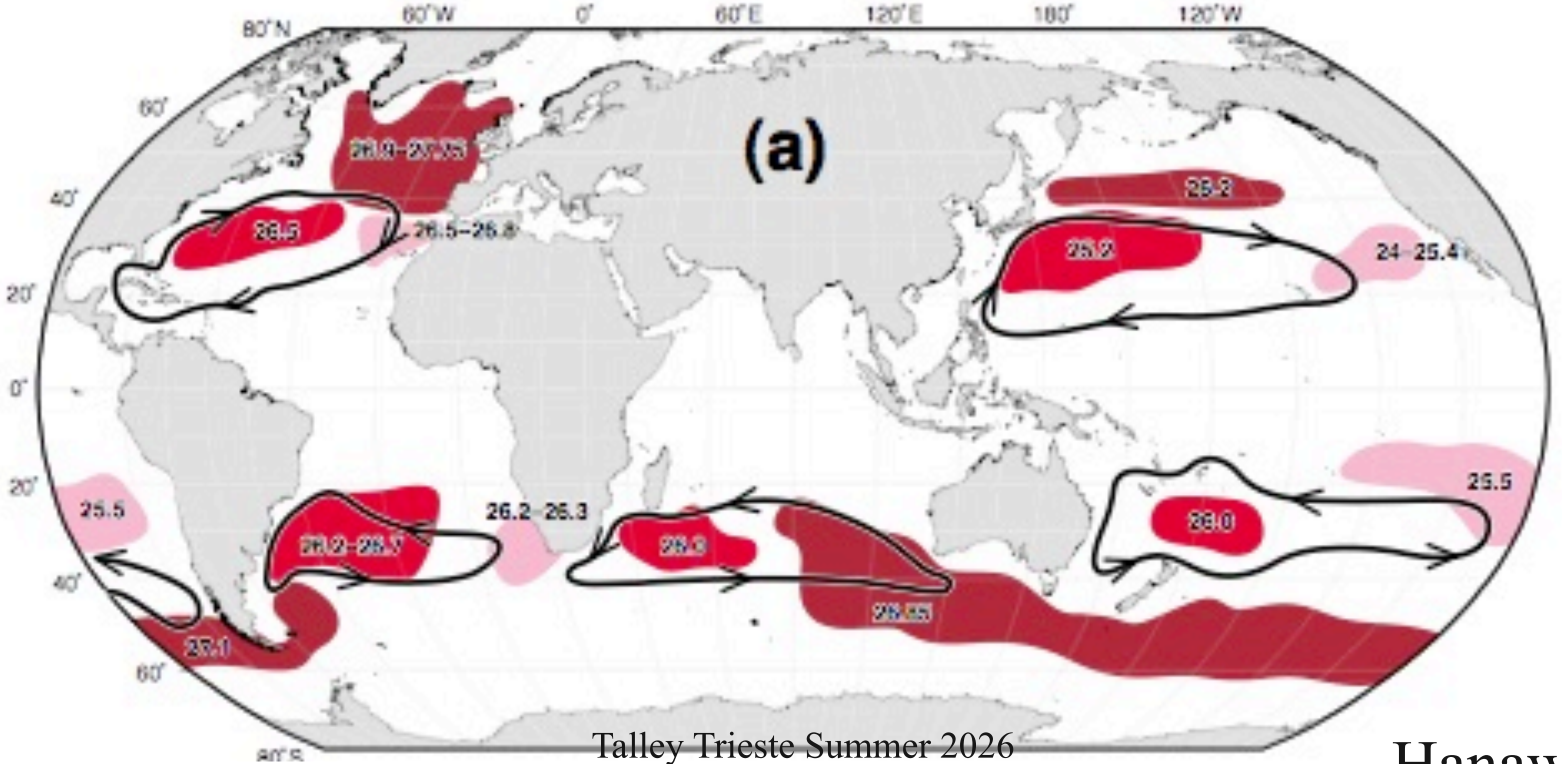
(4) Bottom layer: Antarctic Bottom Water (aka Lower Circumpolar Deep Water)

Remember these layer numbers!

(1) Upper ocean water masses

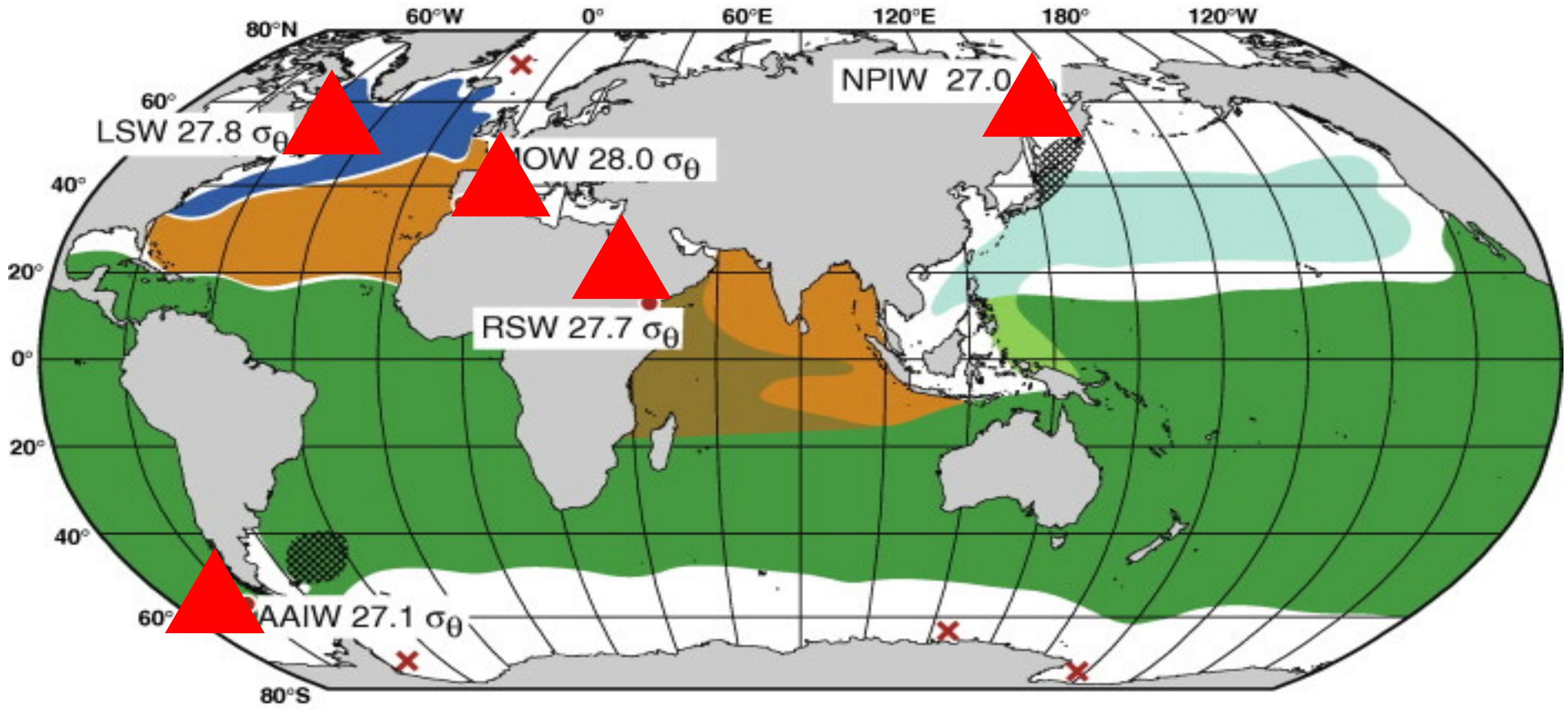
Central Waters (main subtropical thermocline, derived from broad subduction of subtropical surface waters) (not illustrated here)

- 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 waters

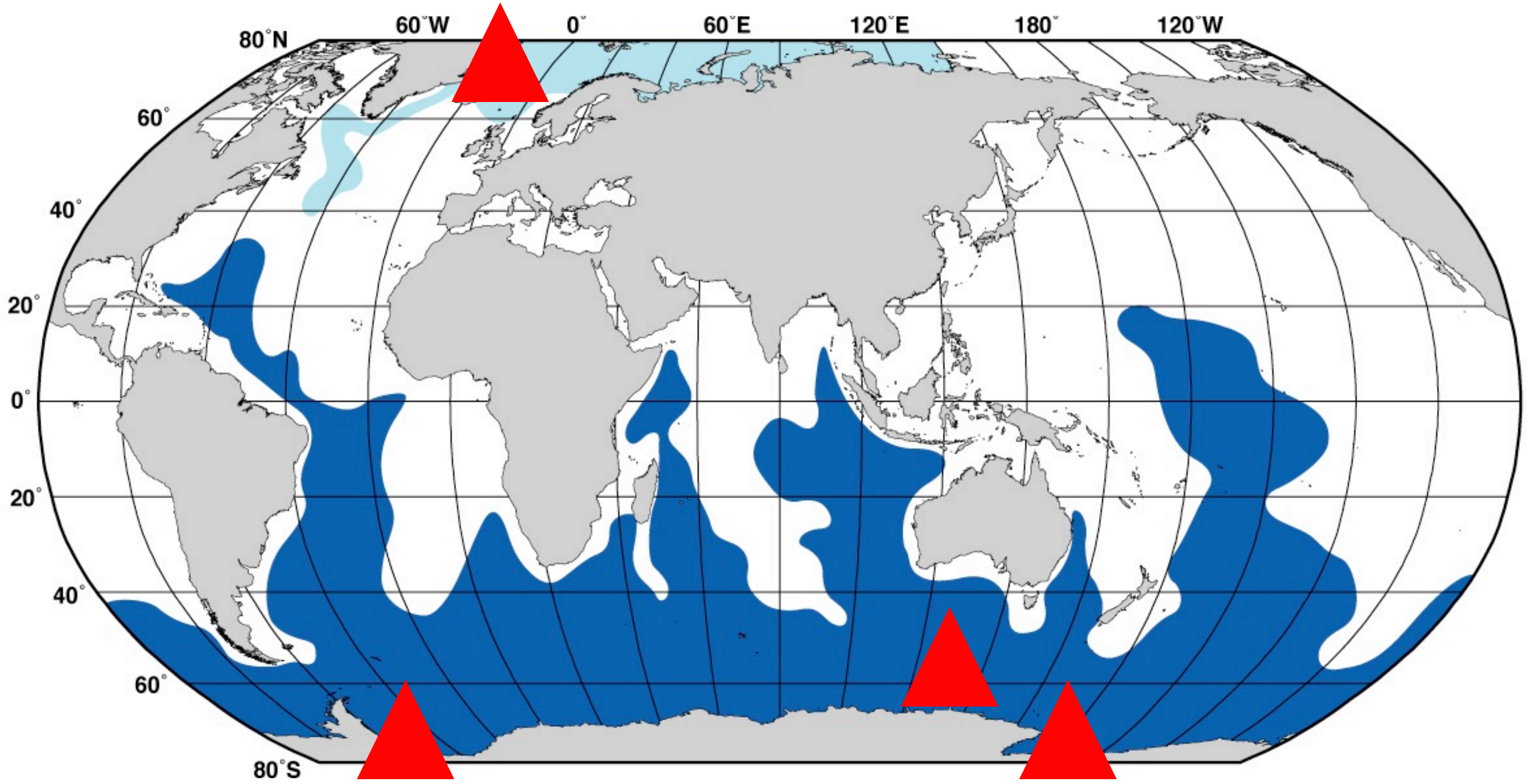


Low salinity: Labrador Sea Water, North Pacific Intermediate Water, Antarctic Intermediate Water

High salinity: Mediterranean Water, Red Sea Water

Talley (2008)

(3, 4) Deep waters

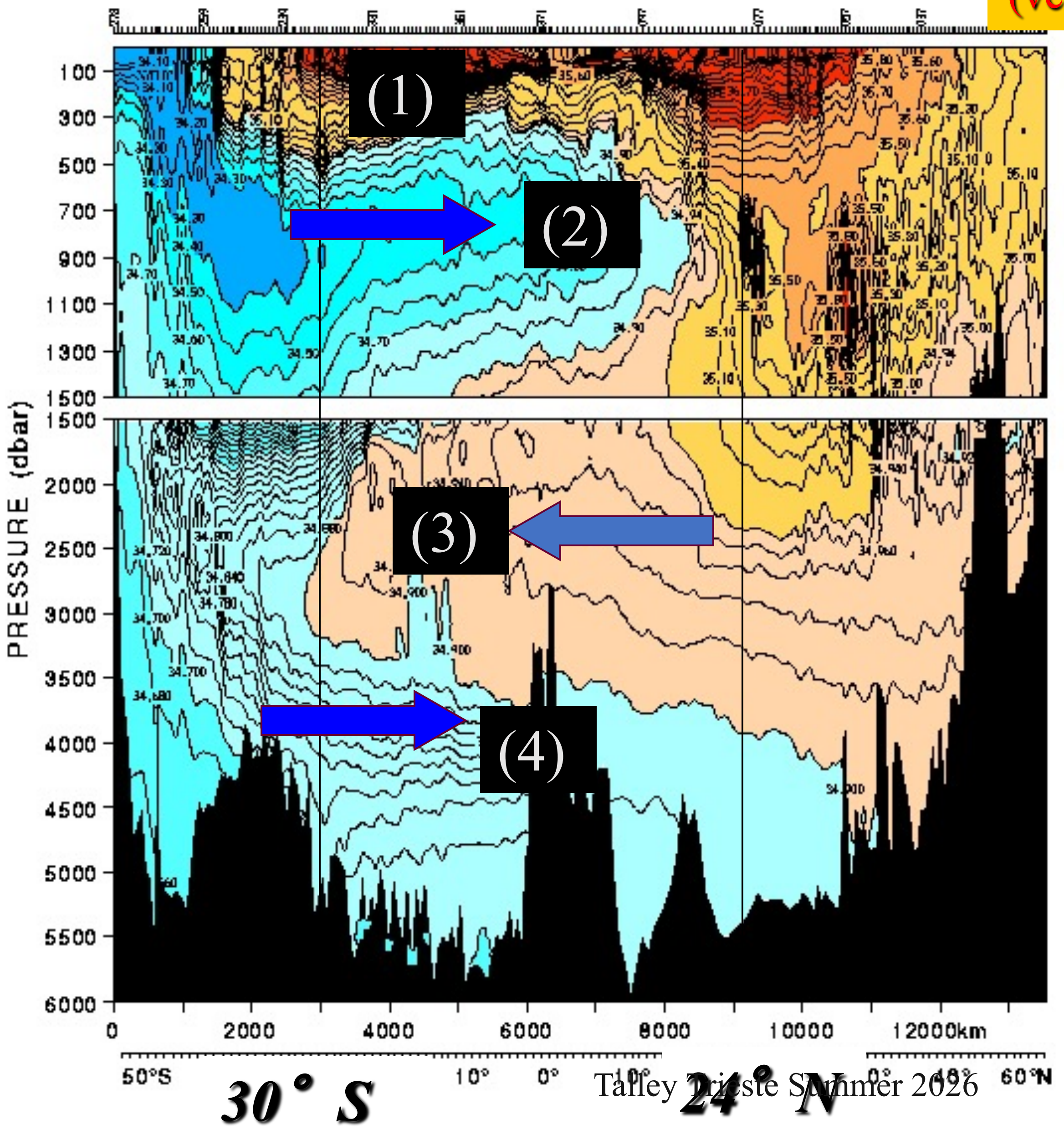


(3) Nordic Seas Overflow waters, contributing to NADW

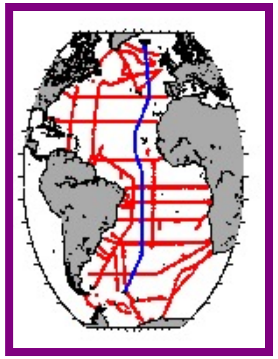
(4) Antarctic Bottom Water in Weddell, Ross Seas and Adelie Coast

Talley (1997)

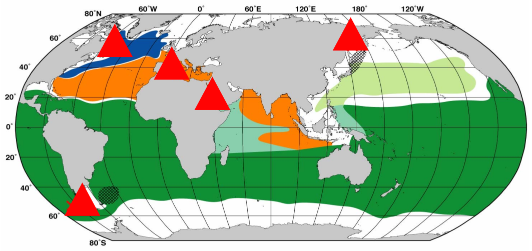
Salinity in the Atlantic at 25°W



(1) surface waters (ventilated thermocline)

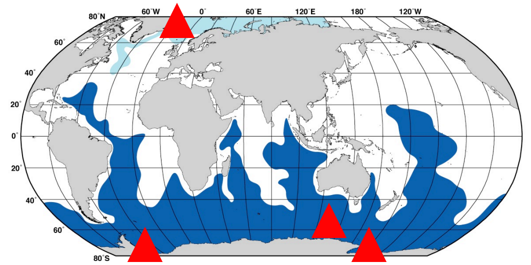


(2) Low salinity Antarctic intermediate water



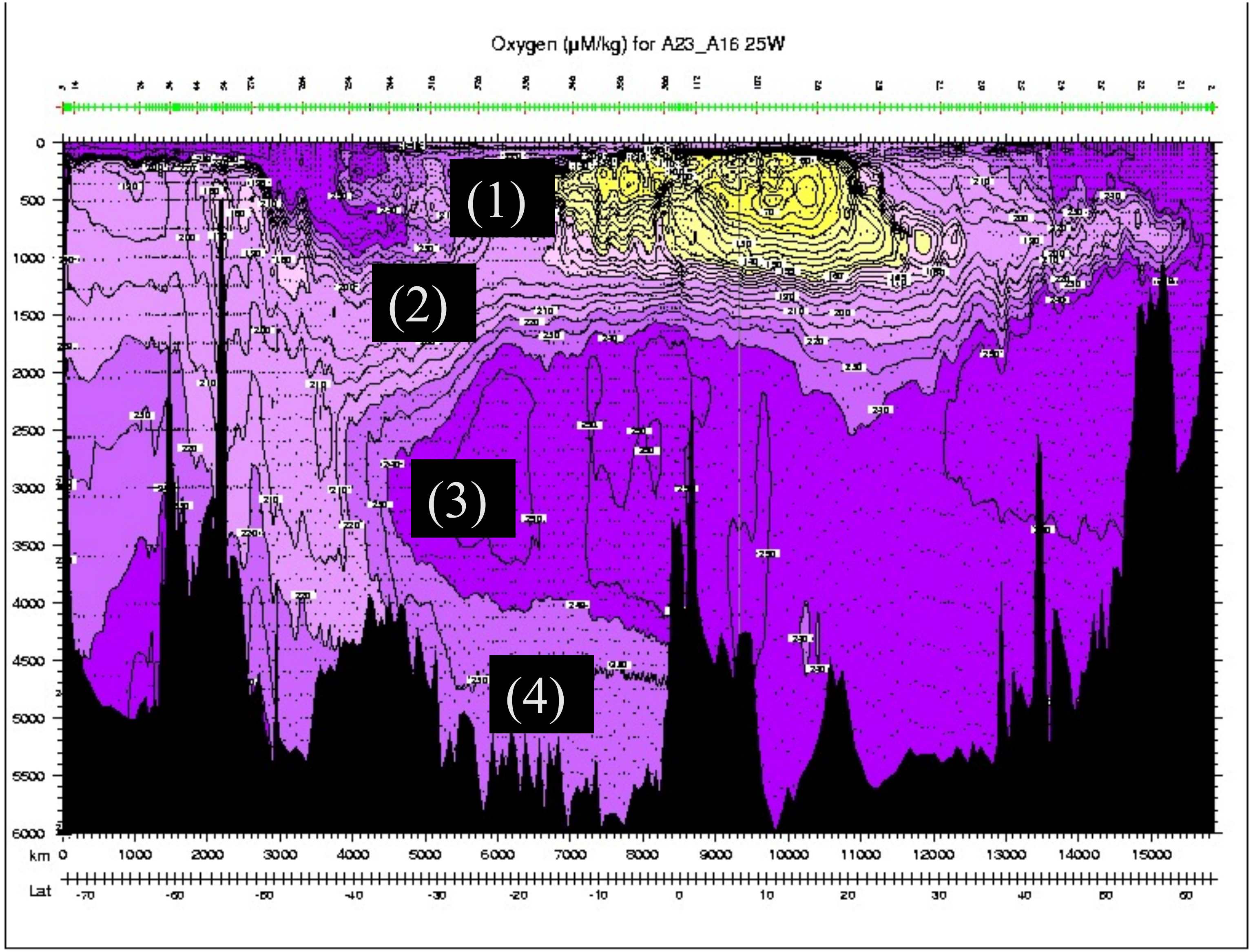
(3) High salinity

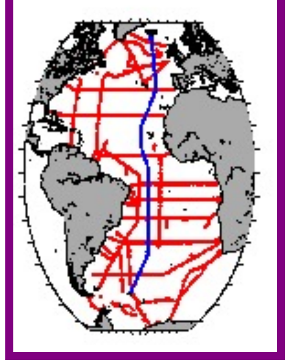
North Atlantic Deep Water



(4) Low salinity Antarctic bottom water

Oxygen in the Atlantic at 25W





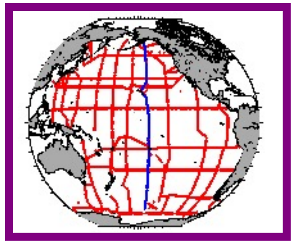
(1) Upper

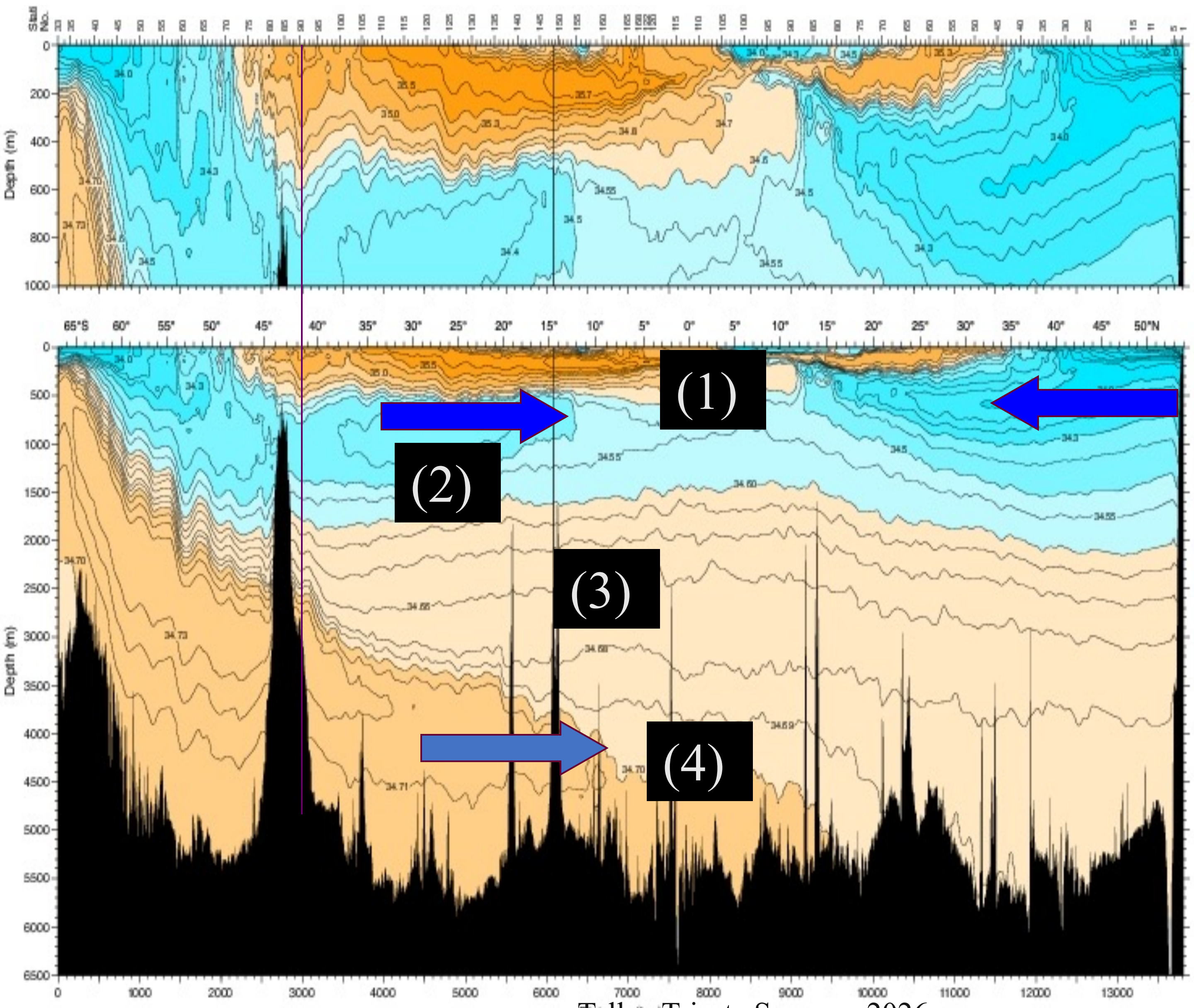
(2) AAIW and LSW

(3) NADW

(4) AABW

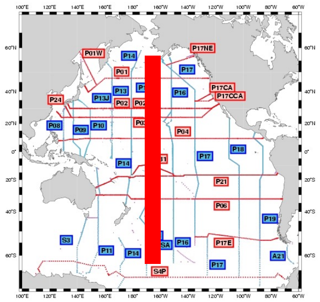
Salinity in the Pacific (150W)

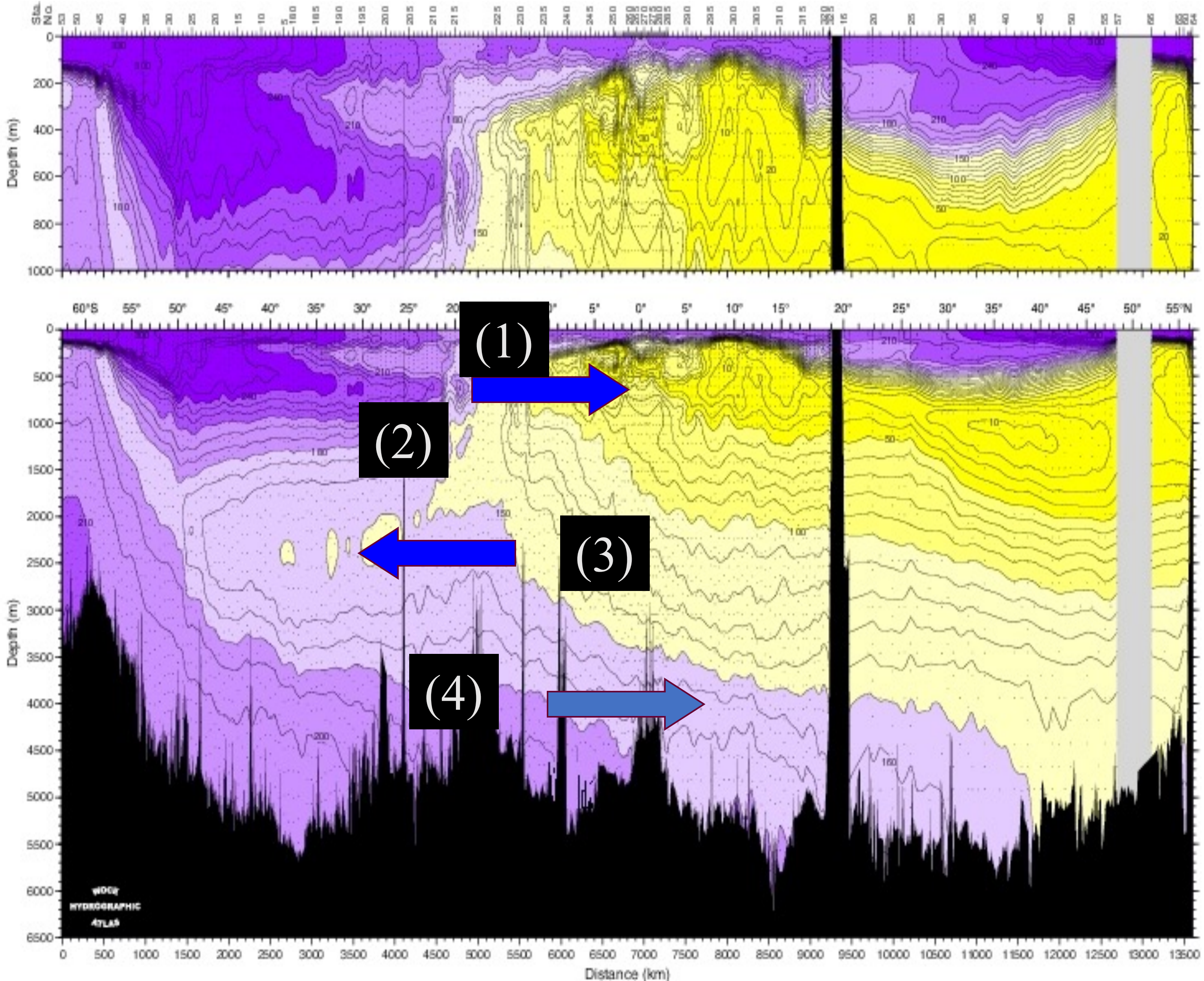




(1) Upper
(2) AAIW and NPIW
(3) PDW
(4) LCDW (AABW)

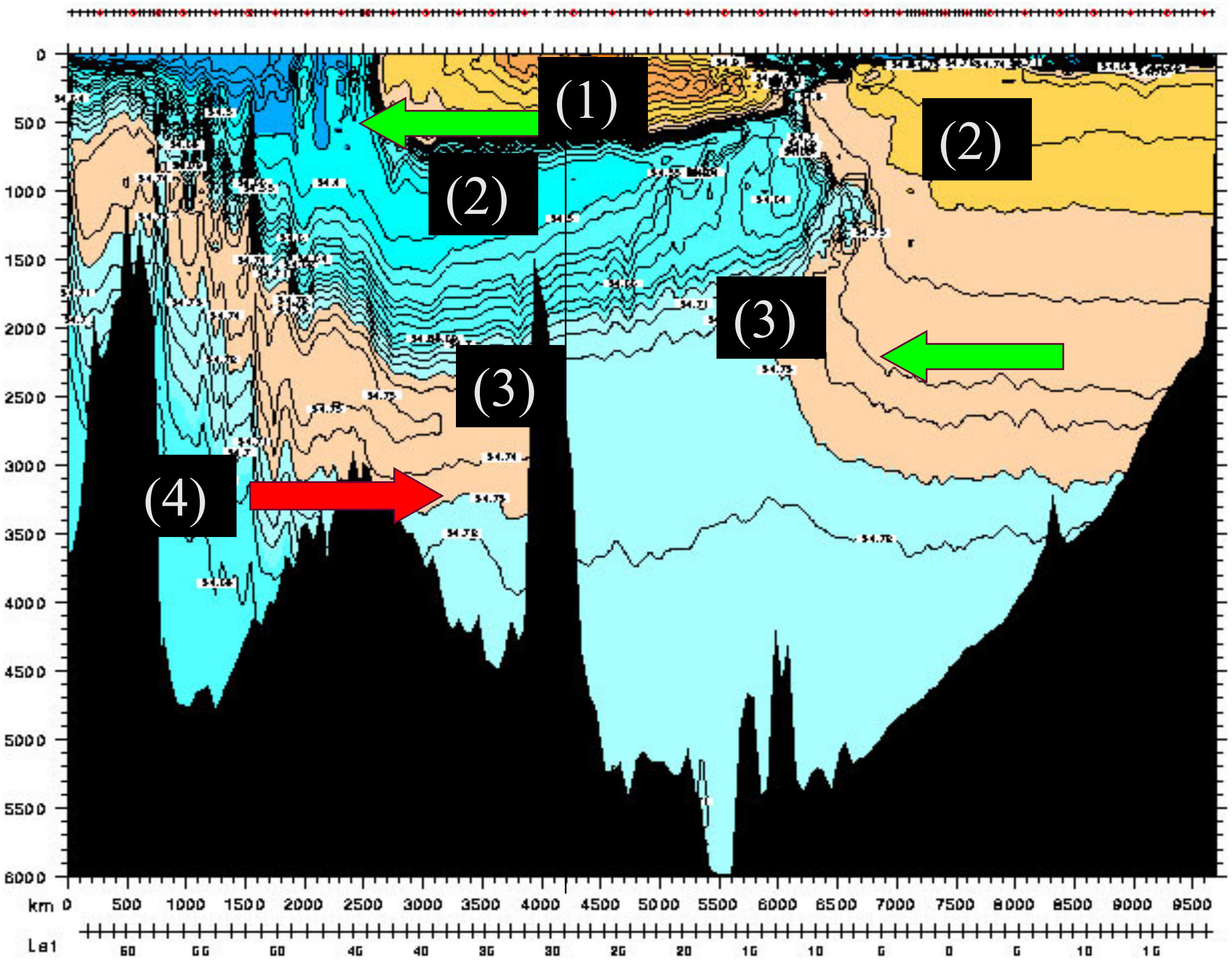
Oxygen in the Pacific (150W)

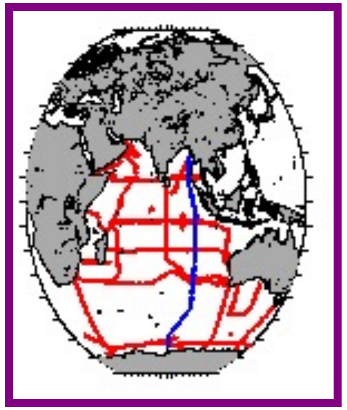




(1) Upper
(2) AAIW and NPIW
(3) PDW
(4) LCBW (AABW)

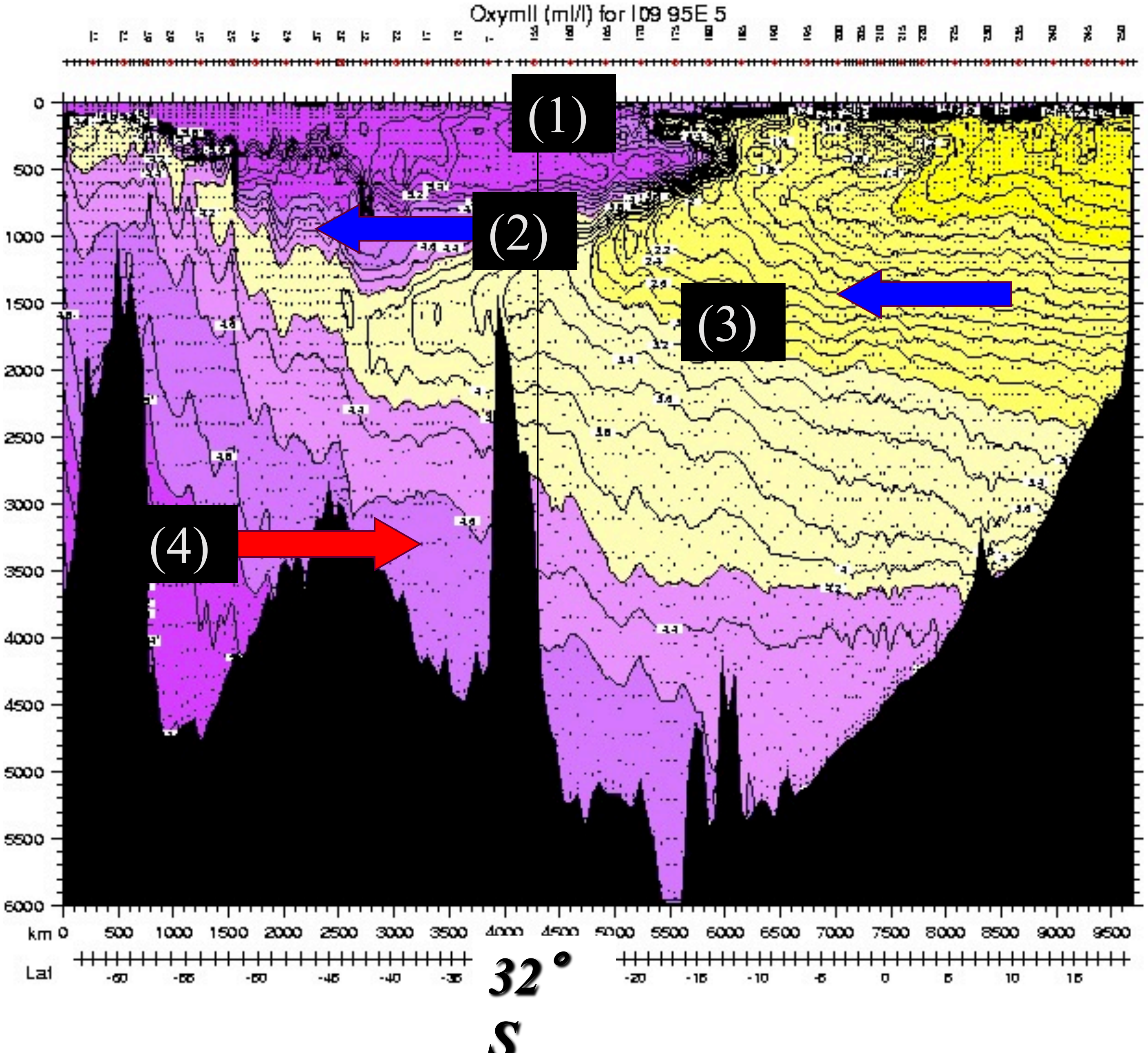
Salinity in the eastern Indian

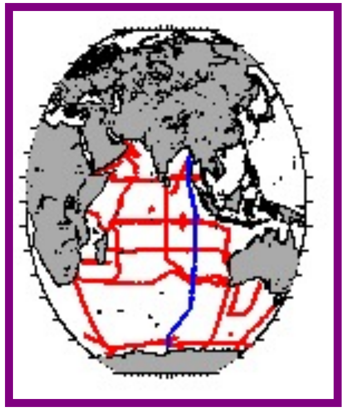




(1) Upper
(2) AAIW and RSW
(3) NADW and IDW
(4) LCDW (AABW)

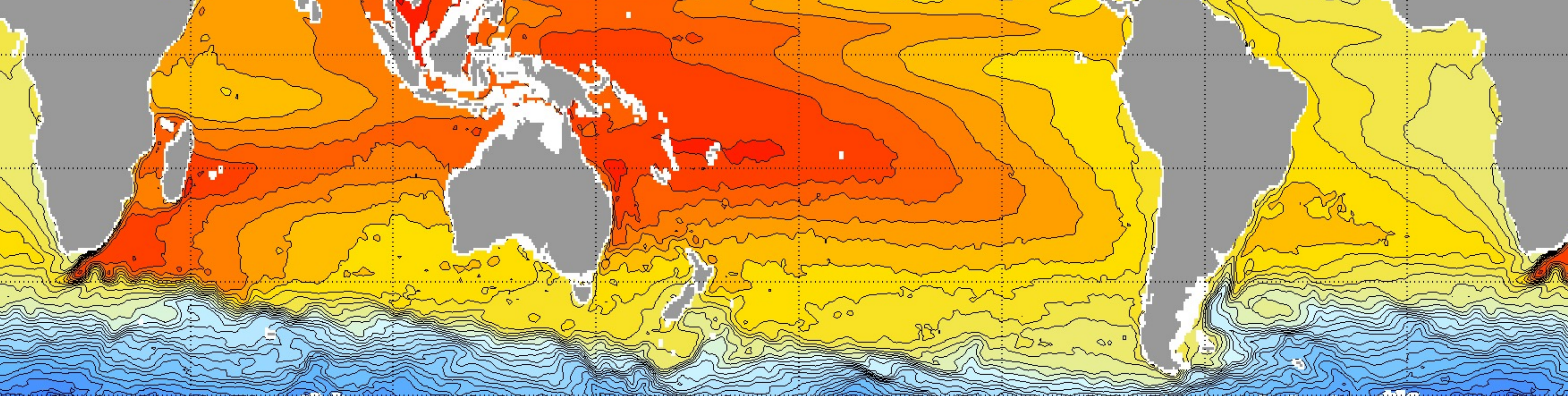
Oxygen in the eastern Indian





Lower oxygen: Red Sea Water and other northern Indian waters

Higher oxygen- Subantarctic Mode Water and Circumpolar Deep Water



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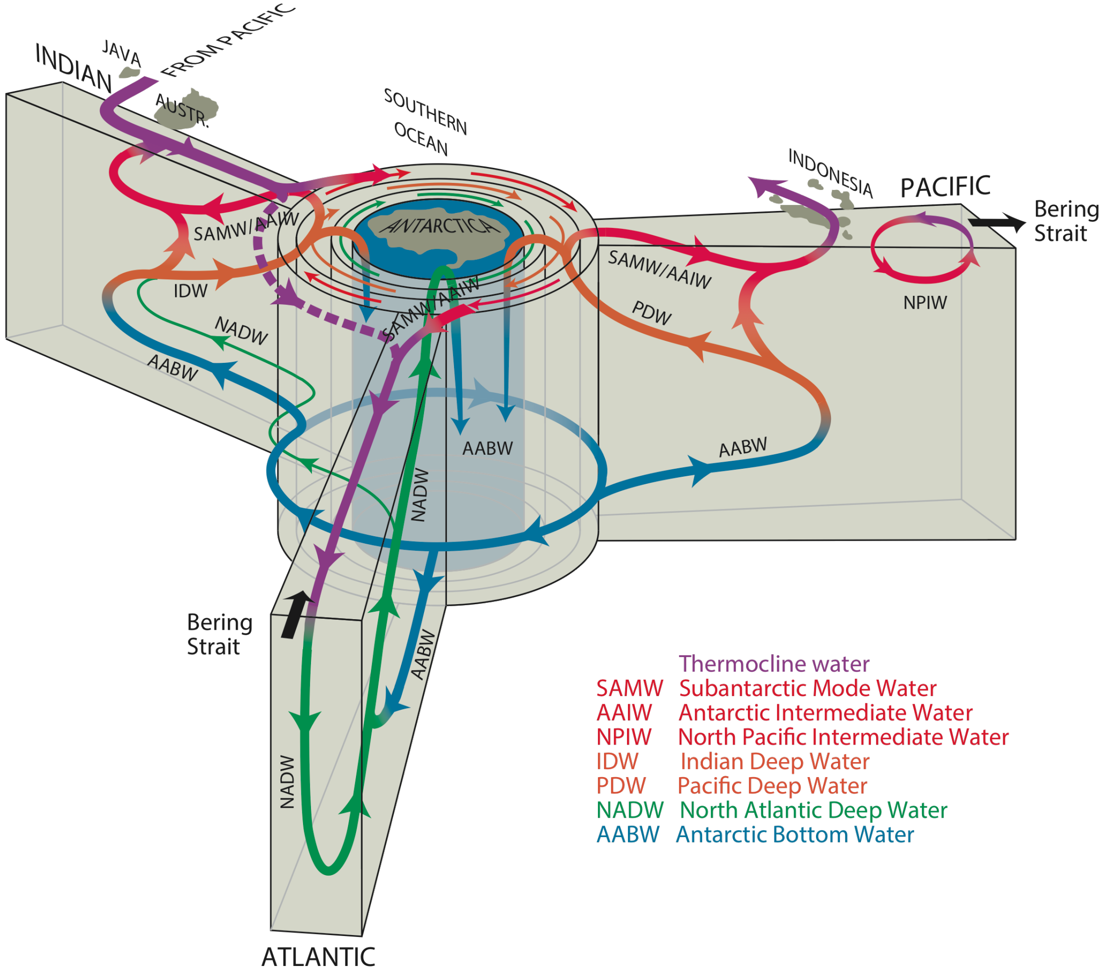
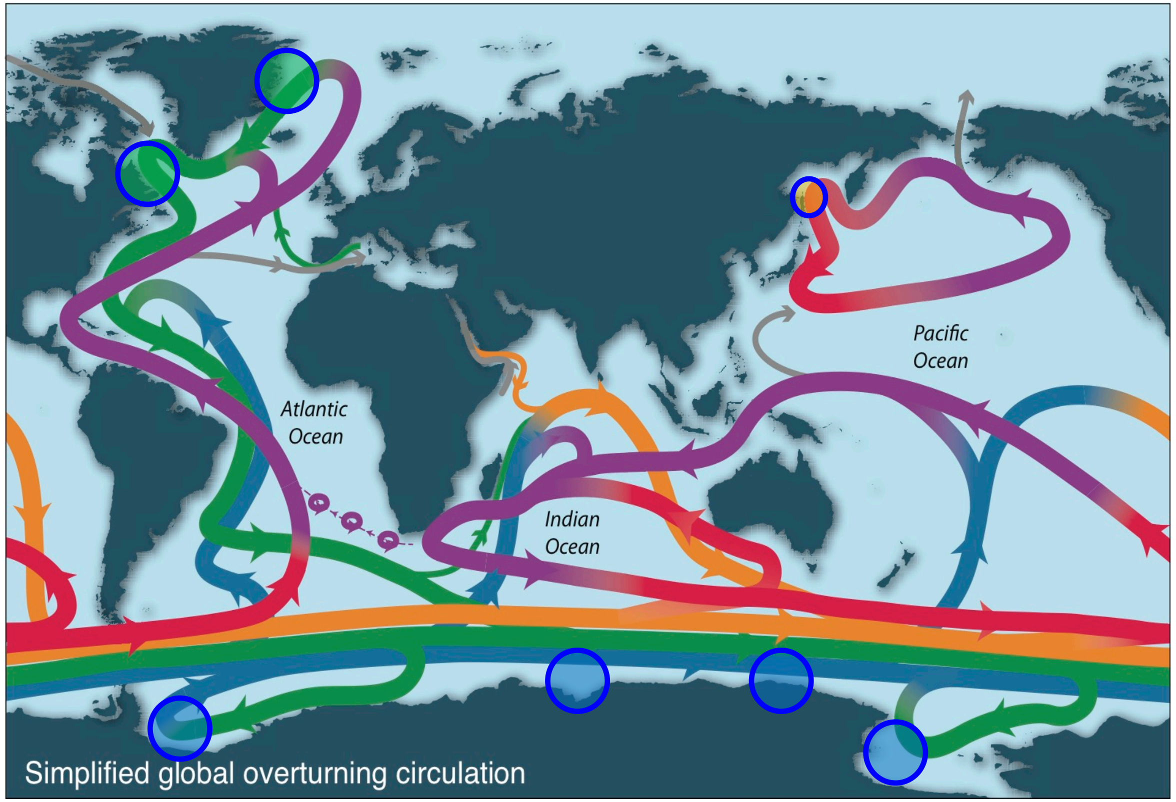
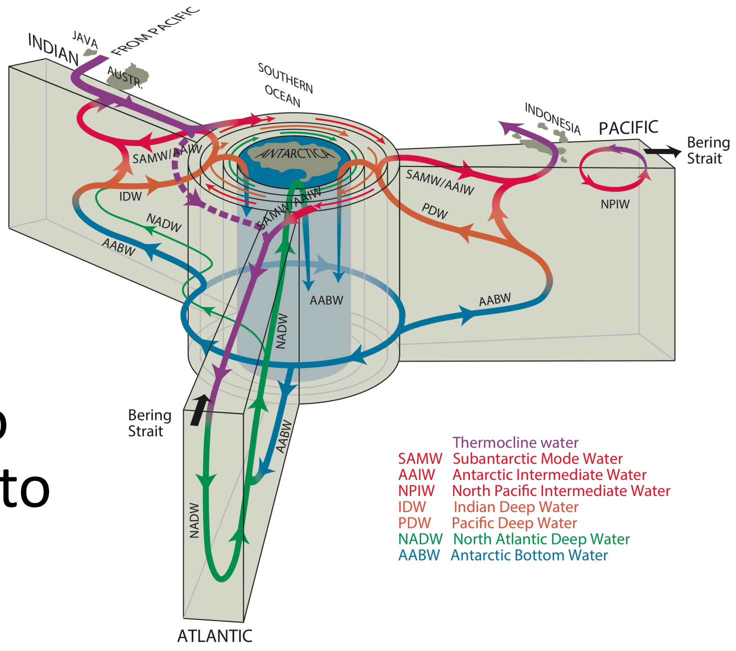
Global Overturning Circulation: build from 4-layer water masses and geostrophic transports





Talley (Oceanography, 2013)

Global overturning circulation: quantifying it



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

- Calculate NET MERIDIONAL TRANSPORT across latitudes.
- Use isopycnal layers (not depth layers), since flow is mostly along isopycnals.

- Also calculate heat and freshwater transports

Property transports

- Volume transport = $V = \sum v_i A_i = \iint v dA$ m³/sec
- Mass transport = $M = \sum \rho\, v_i A_i = \iint \rho v dA$ kg/sec
- Heat transport* = $H = \sum \rho\, c_p T v_i A_i = \iint \rho c_p T v dA$ J/sec=W
- Salt transport = $\mathscr{S} = \sum \rho\, S v_i A_i = \iint \rho S v dA$ kg/sec
- Freshwater transport = $F = \sum \rho\, (1\text{-}S)\, v_i A_i = \iint \rho (1 - S) v dA$ kg/sec
- Chemical tracers = $\mathscr{C} = \sum \rho\, C v_i A_i = \iint \rho C v dA$ moles/sec

* Heat: T is in Kelvin. Most useful when total mass transport $M = 0$ (flow in equals flow out), so convergence provides surface heat flux.

- Flux is these quantities per unit area
  e.g. volume flux is V/A, mass flux is M/A,
  heat flux is H/A, salt flux is $\mathscr{S}$/A, freshwater flux is F/A, chemical tracer flux is $\mathscr{C}$/A

Calculation of meridional overturn

DPO Section 14.2.2

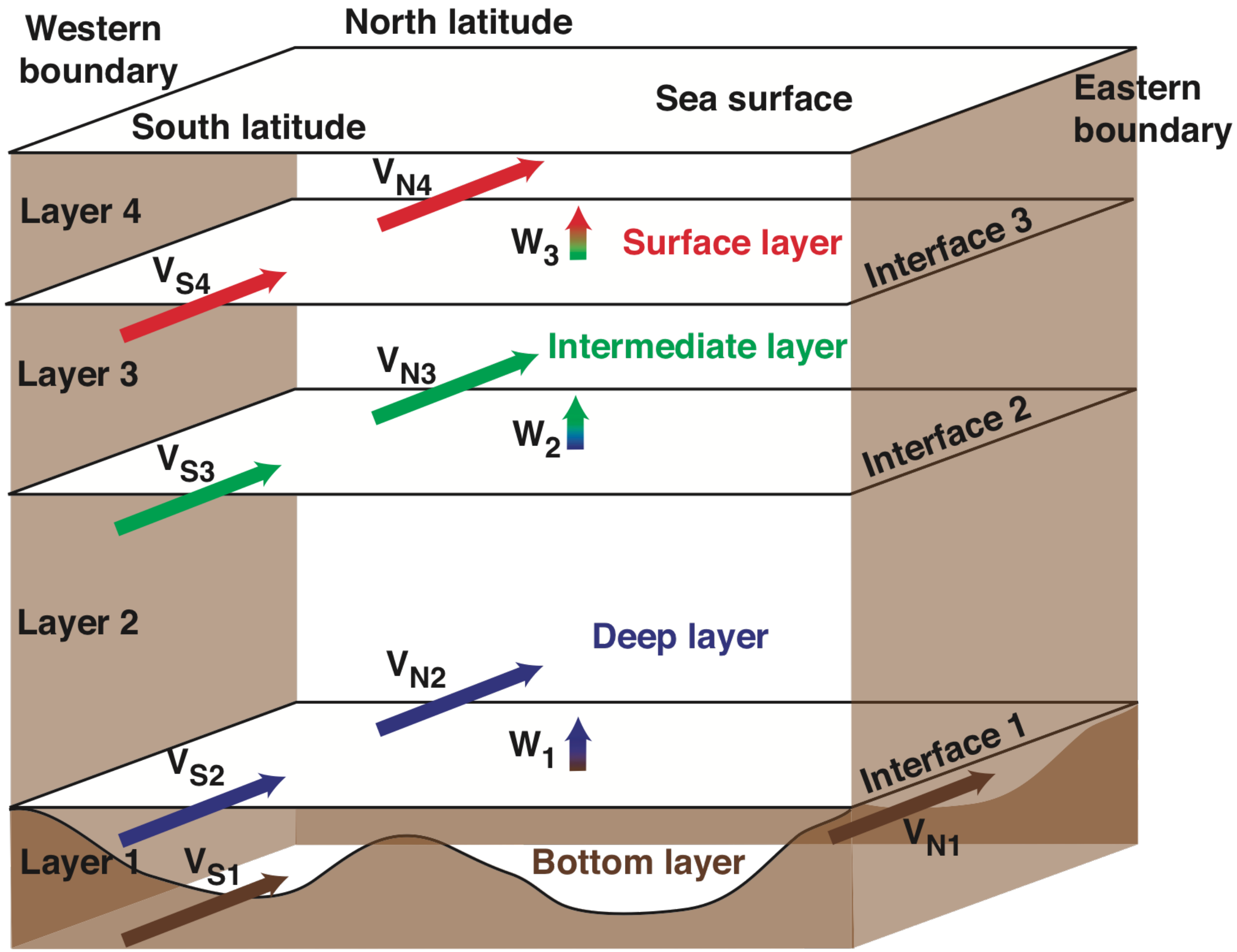
Use a zonal, coast-to-coast, top-to-bottom section

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

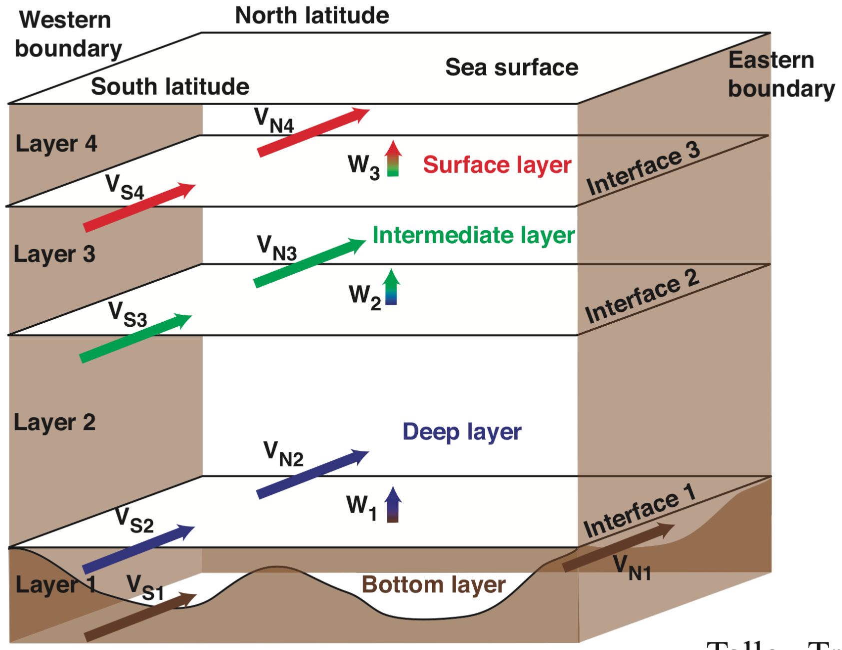
(3) Vertical velocity (average) = vertical transport divided by area of interface

Equation 14.1

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

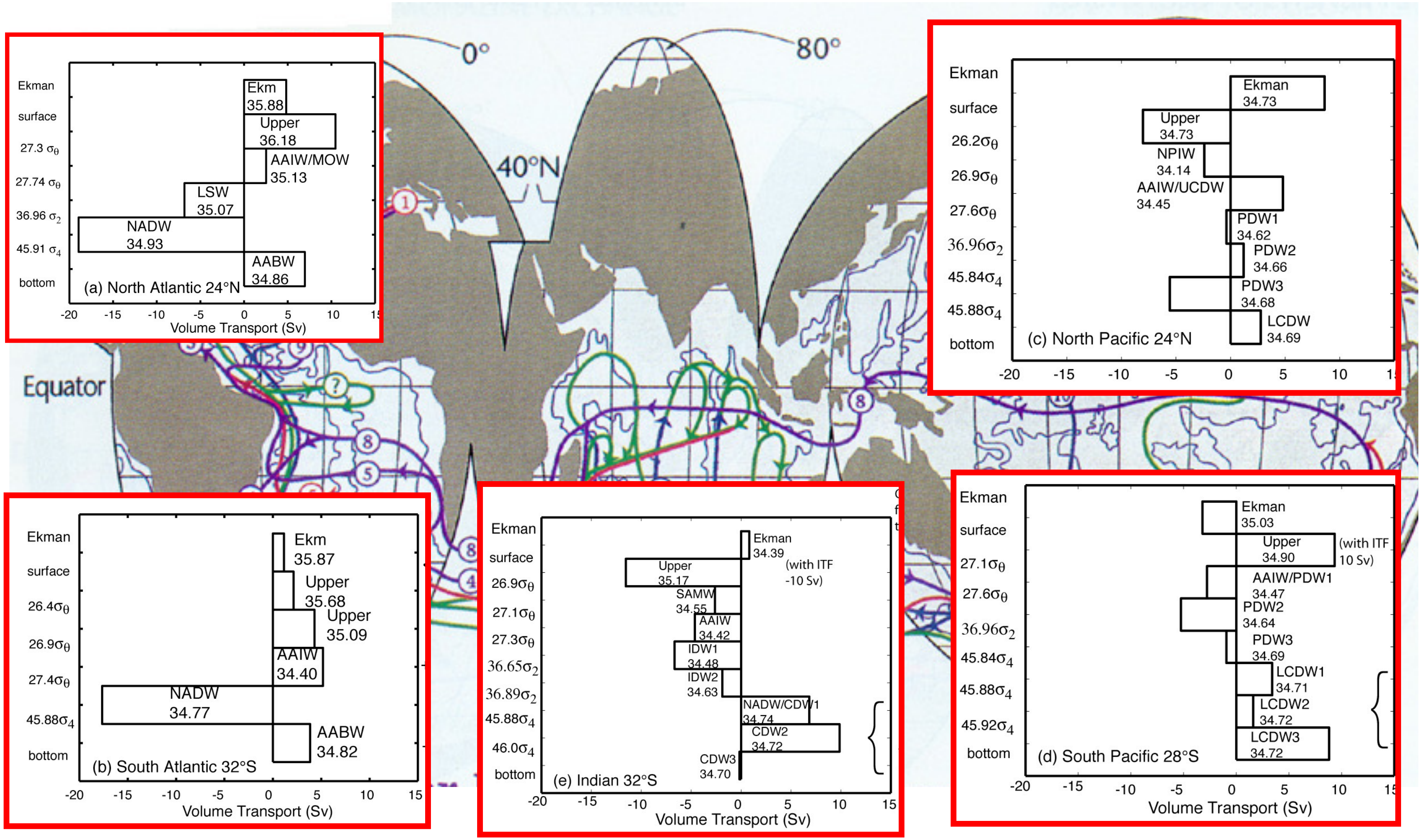
(2) $W_i = V_{Ni} - V_{Si} + W_{i-1}$

(3) $w_i = W_i/A_i$



DPO Fig. 14.5

Net meridional transports in isopycnal layers



Talley (Prog in Oceanogr. 2008)

Calculation of meridional overturning (vertical velocity)

Transport convergence 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.

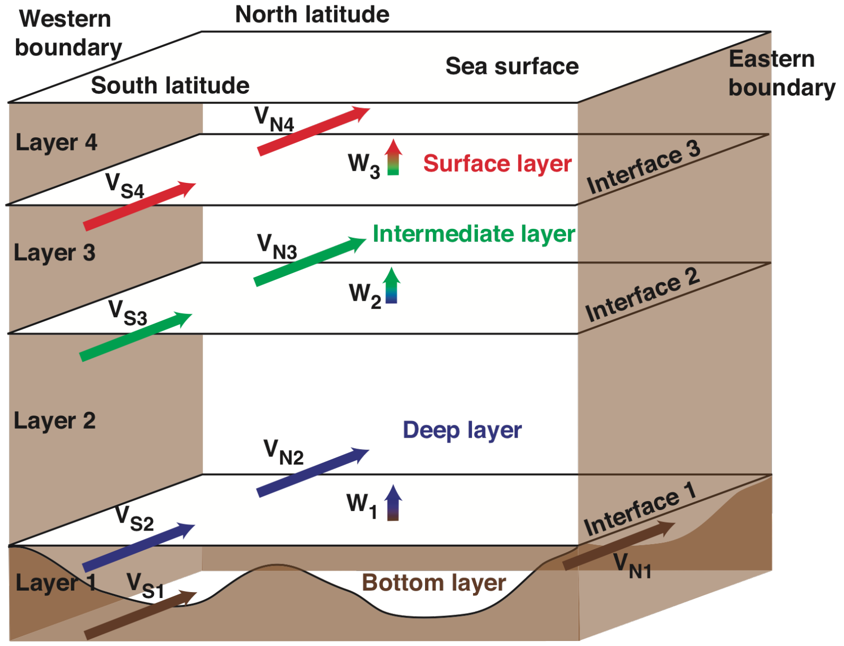
(3) If the area of the interface is 4,000 km x 1,000 km, the average vertical velocity is $w_1 = (1\text{x}10^6\ \text{m}^3/\text{sec})/(4 \text{ x } 10^{12}\ \text{m}^2) = 0.25\text{x}10^{-6}$ m/sec = $0.25\text{x}10^{-4}$ cm/sec

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) $w_i = W_i/A_i$

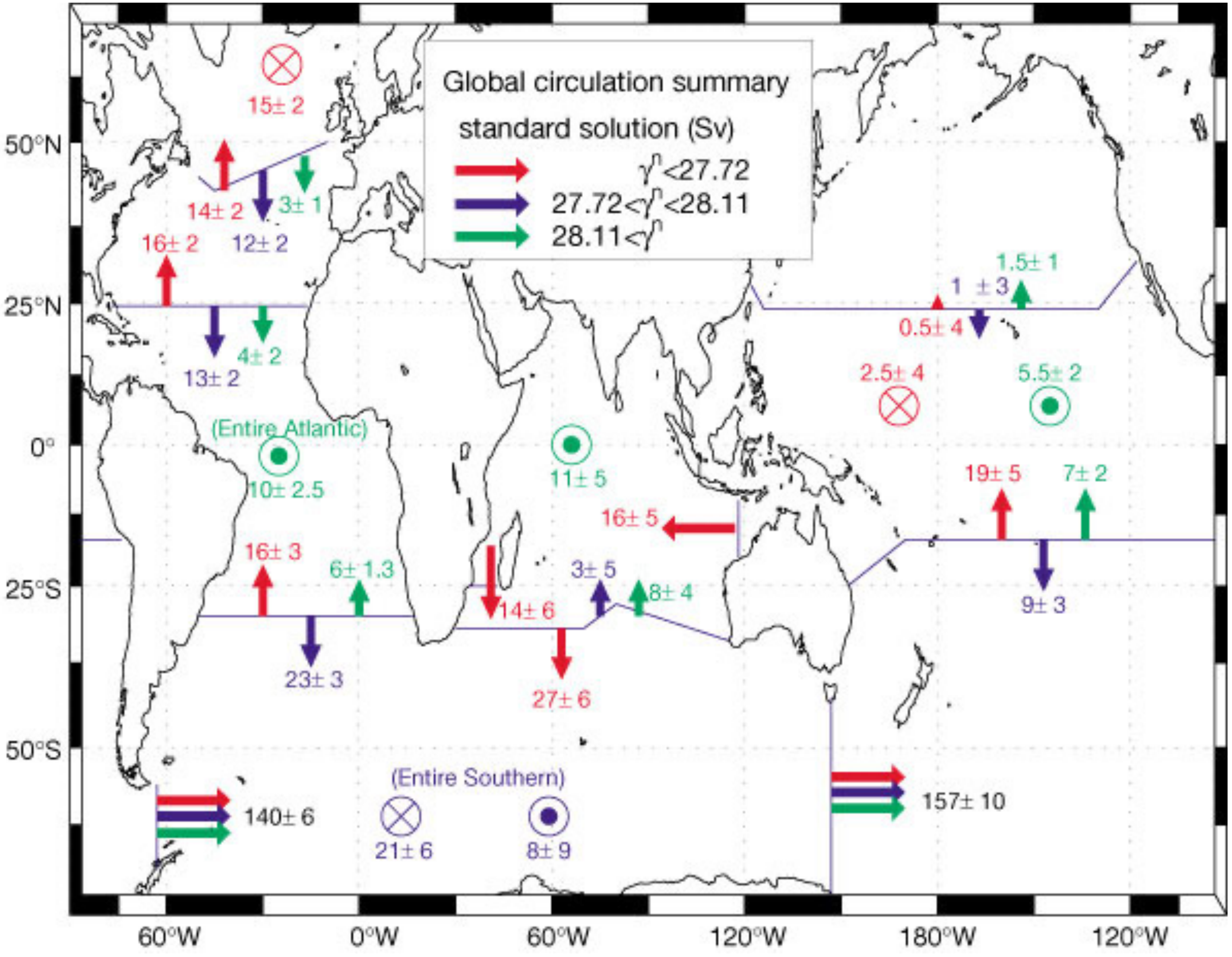


DPO Fig. 14.5

Global mass transport and vertical velocity

Inverse model

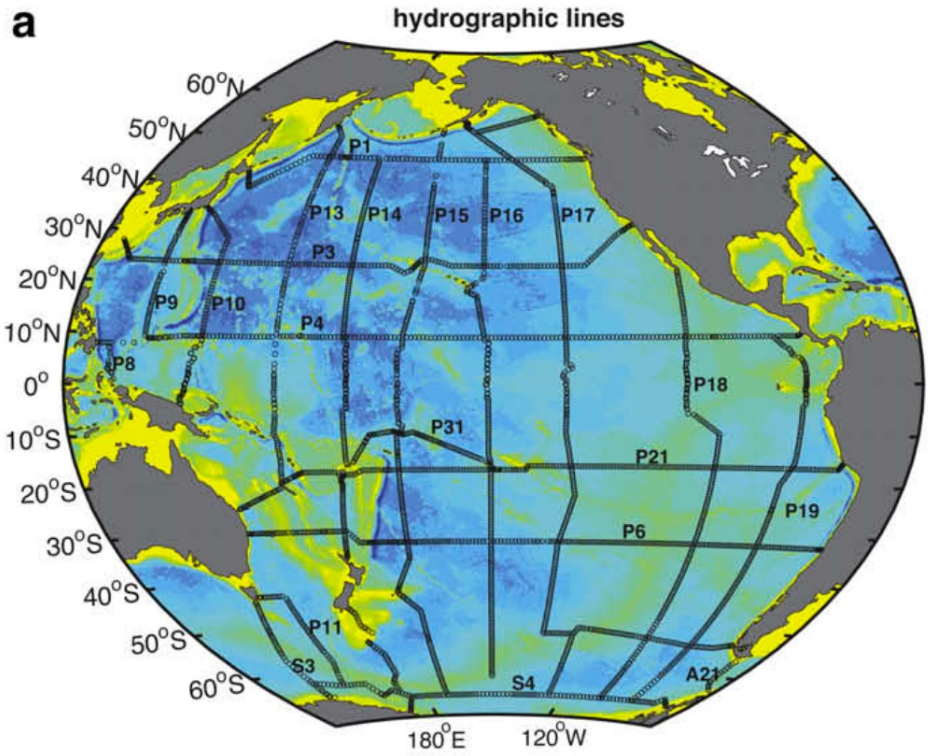
Ganachaud and Wunsch (2000)



Global mass transport and vertical velocity

Inverse model

Pacific WOCE



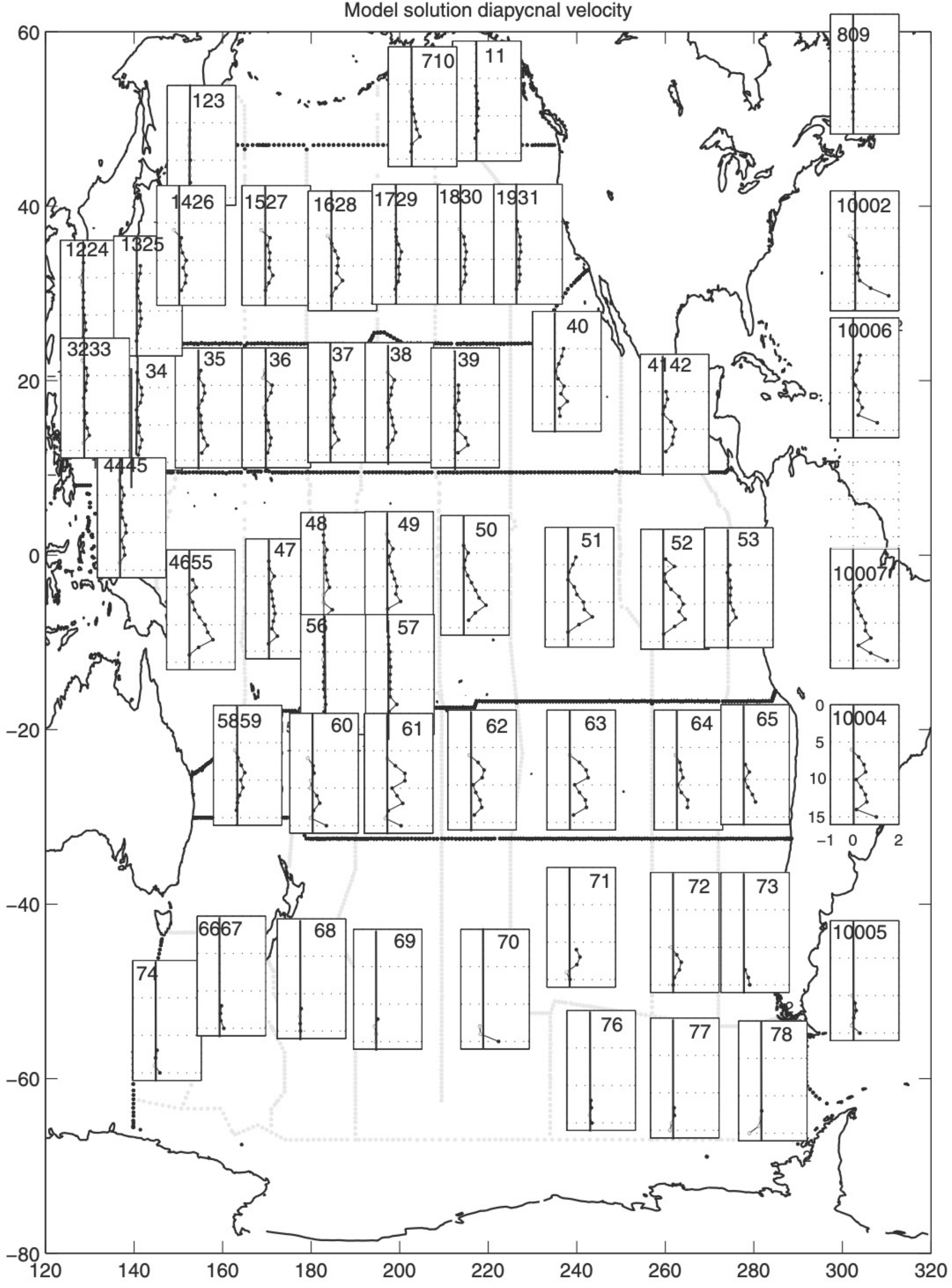


Fig. 9. Inverse estimates of diapycnal velocity in each of the small boxes (as labeled) and in each the large boxes (right-hand column). Axes are the same for all graphs, but labeled only on box 10004 (second up from the bottom on the right). Inverse solutions for **w** only exist for non-outcropping interfaces. Interface 0 is the surface, interface 15 is the bottom. Values are in units of 10^{-6} cm s^{-1}.

Macdonald et al. (PiO 2009)

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.

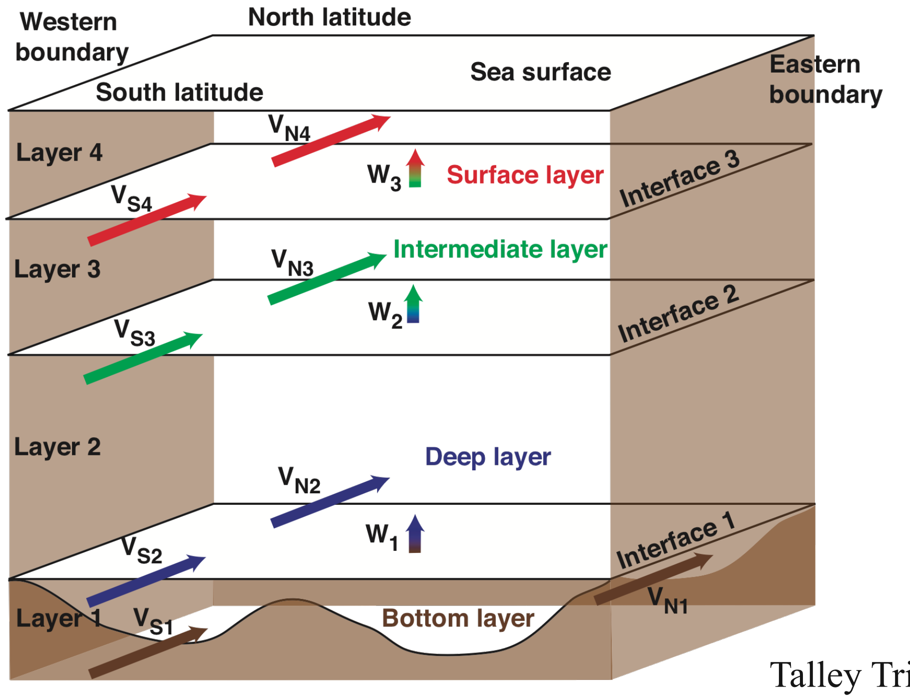
Useful to have transports at many latitudes (e.g. from models)

Equation 14.2

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

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

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



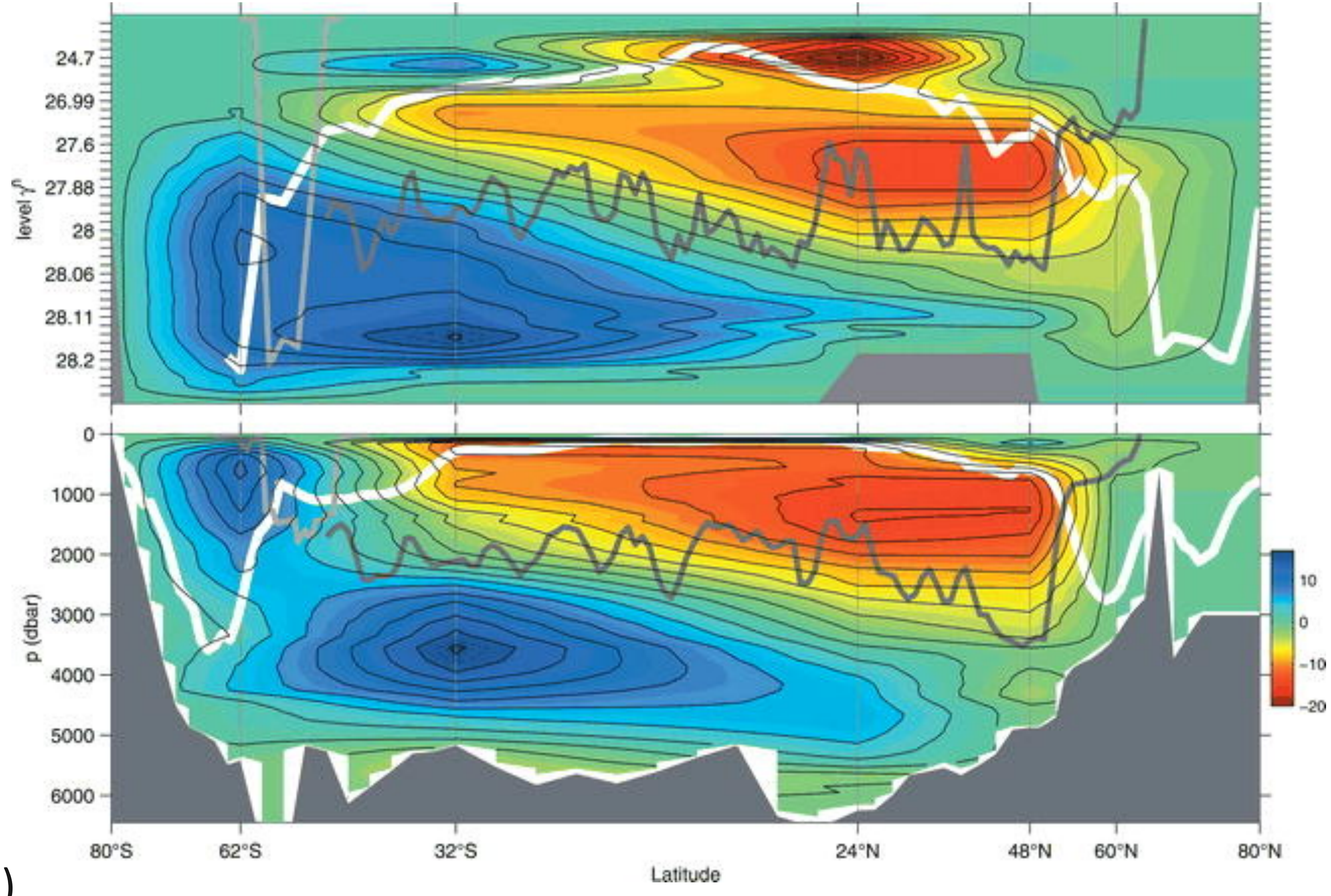
Meridional Overturning Streamfunction in isopycnal&depth /latitude coordinates

Inverse model

Isopycnal coordinates

Global

Pressure coordinates



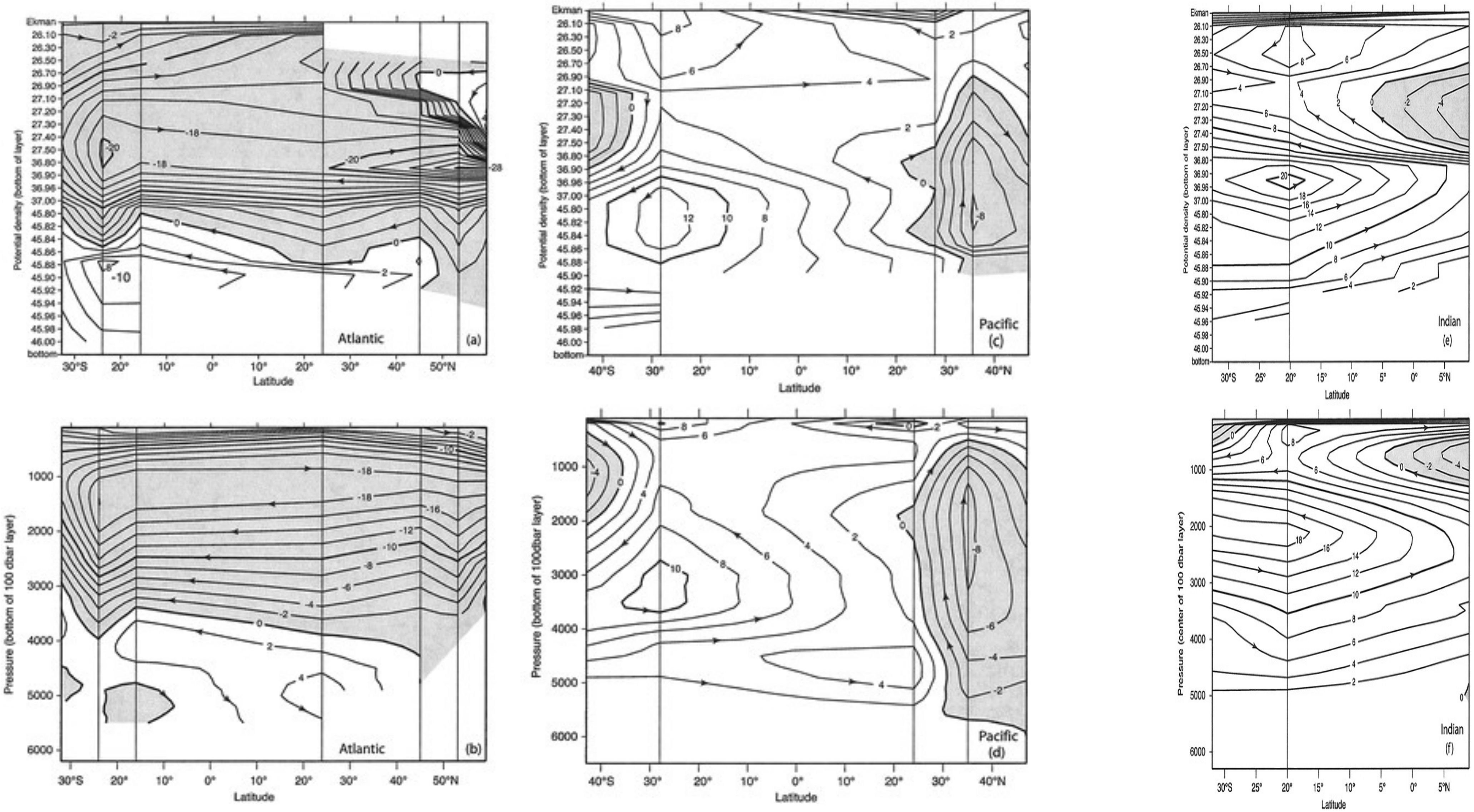
Lumpkin and Speer (2007)

Meridional Overturning Streamfunction in isopycnal/latitude coordinates

Atlantic | Pacific | Indian

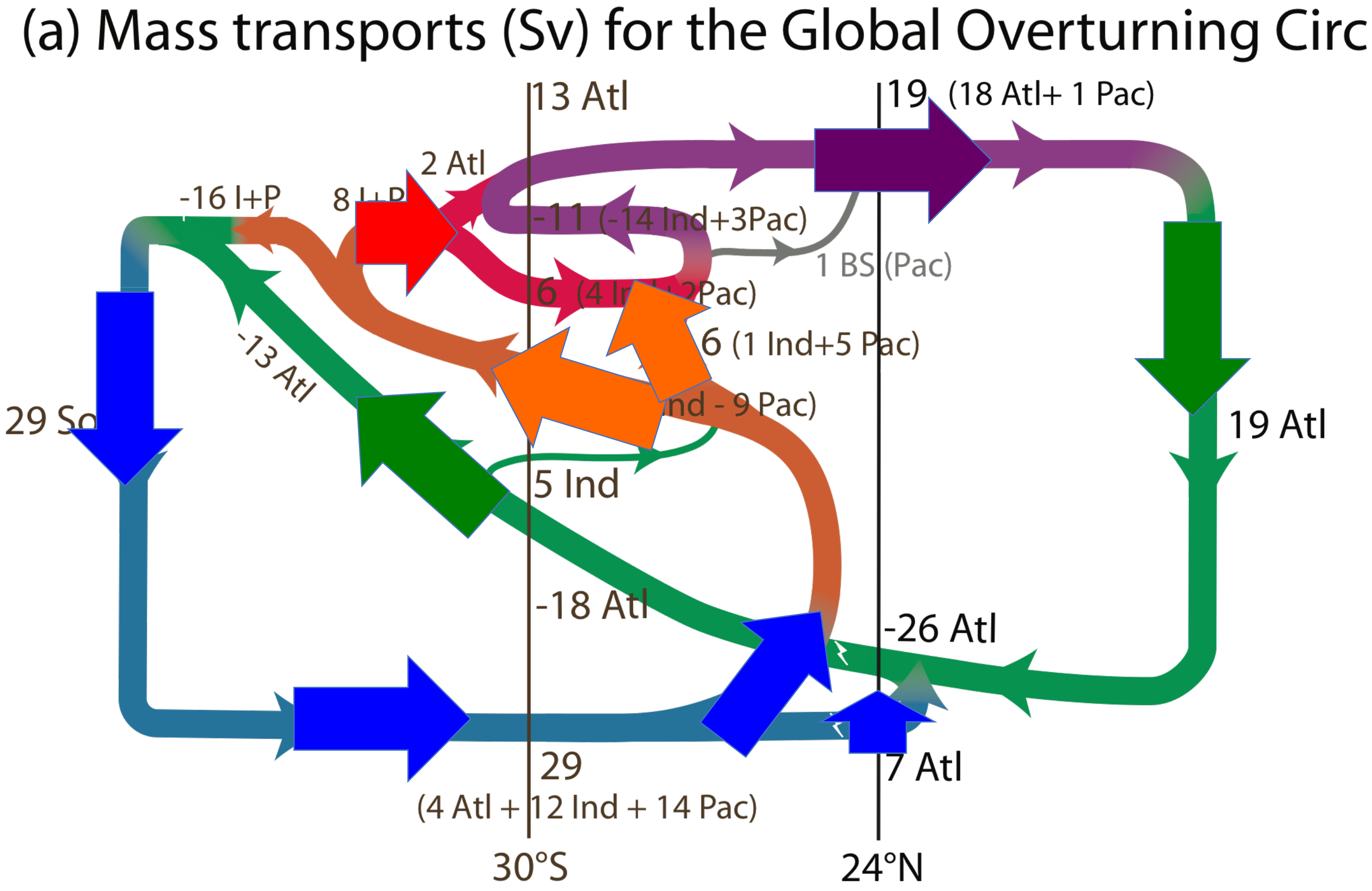
Isopycnal

Pressure



Talley et al. (2003) based on Reid velocities

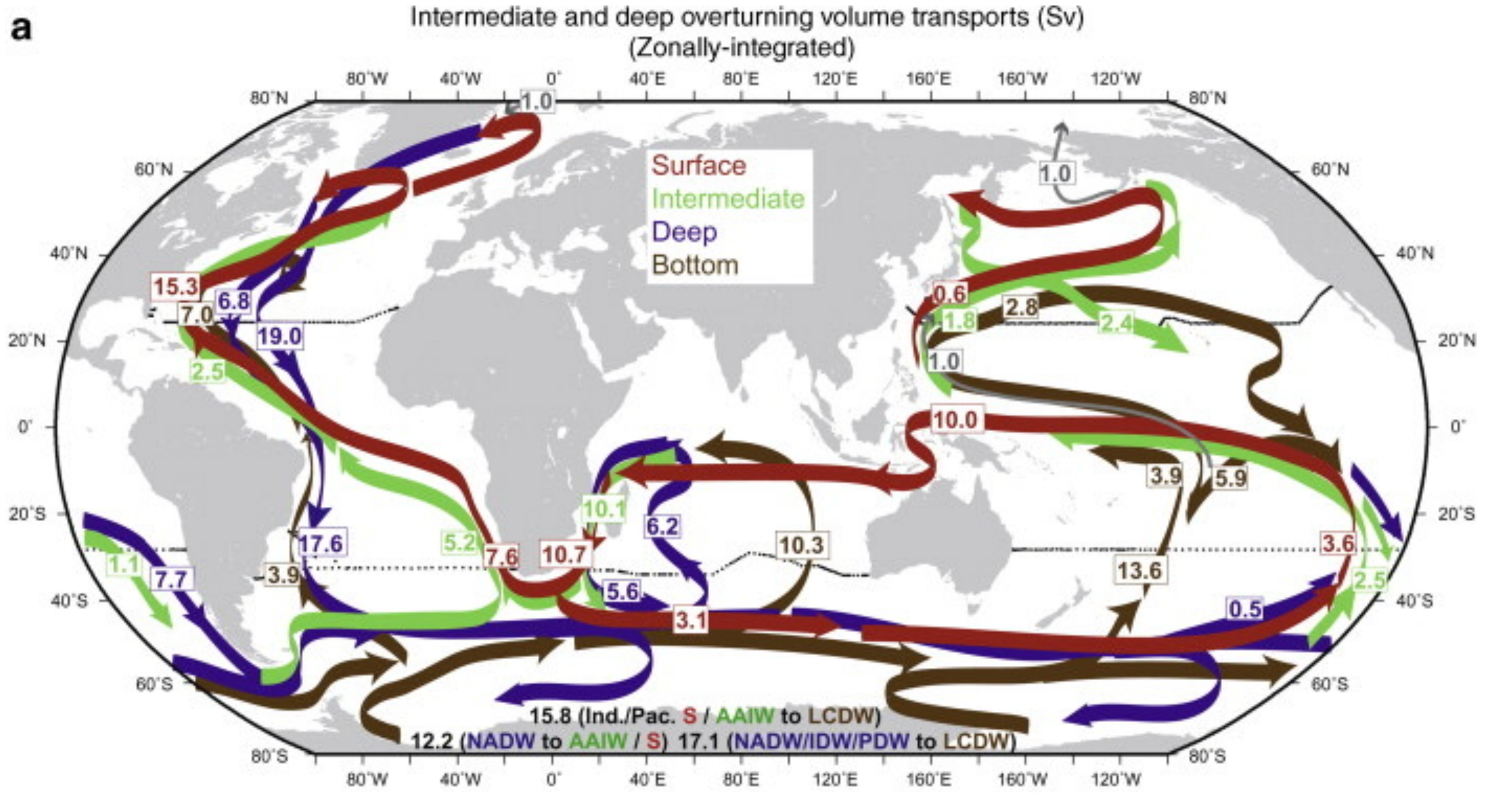
Mass transports for the Global Overturning Circulation



Talley (Oceanography, 2013)

What are the volume transport pathways through the oceans?

Talley (PiO, 2008) based on Reid (1994, 1997, 2003)



Overturning circulation

- NADW formation
- NPIW formation
- AABW formation

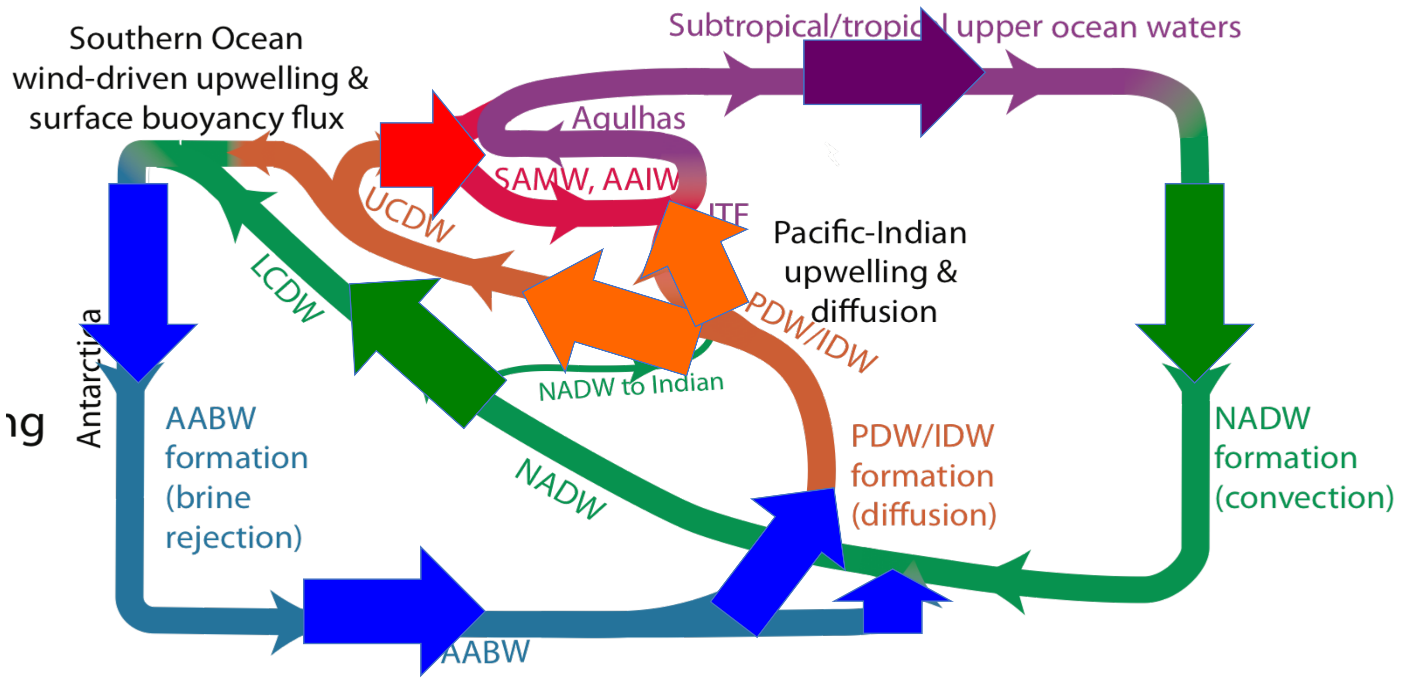
- Indian Ocean diapycnal upwelling
- Pacific Ocean diapycnal upwelling

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)

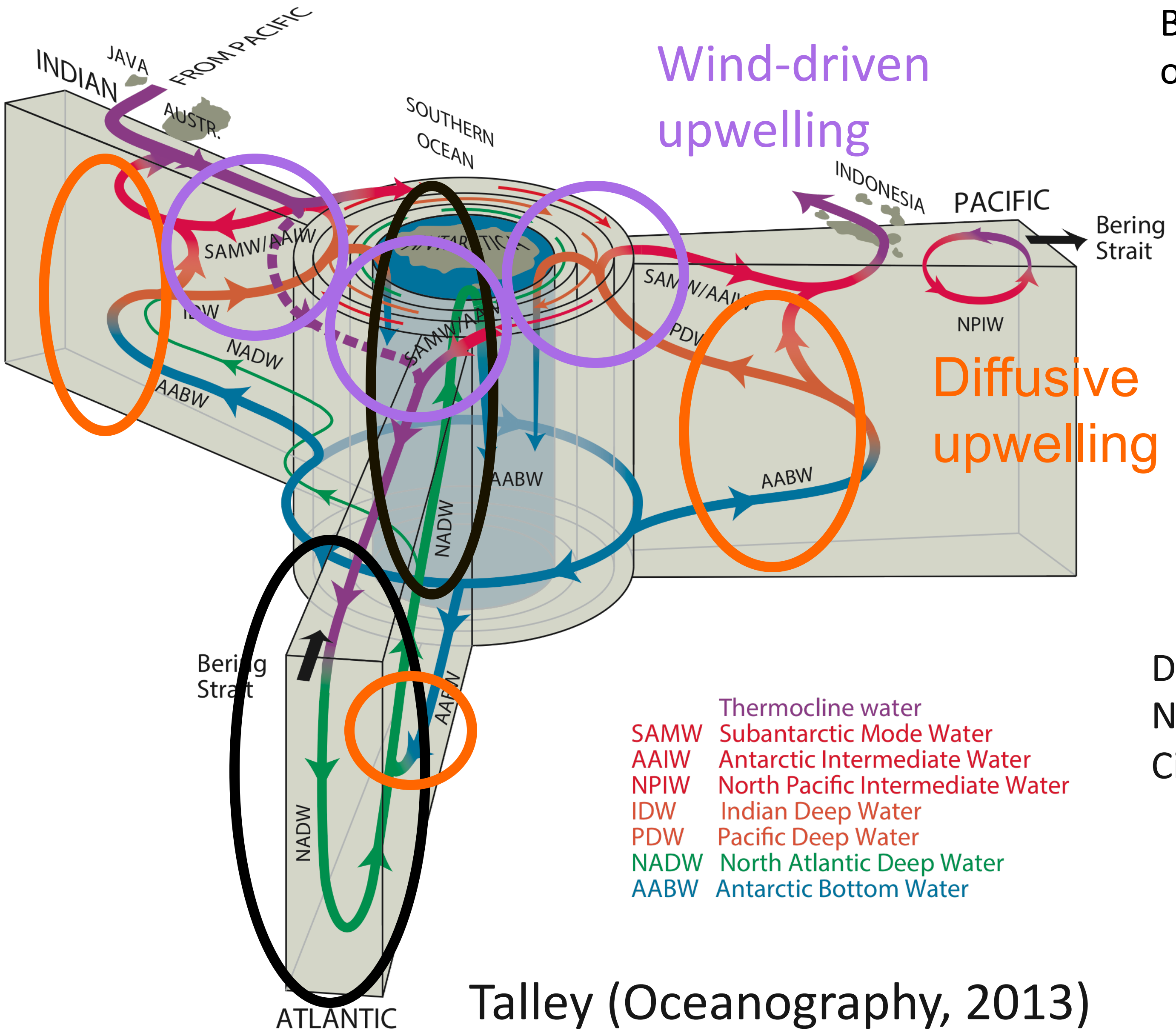


Talley (DPO, 2011; Oceanography, 2013)

DPO Fig. 14.11c

Global Overturning Circulation: more schematicized, with processes

Brine rejection: Southern Ocean overturning circulation (AABW)

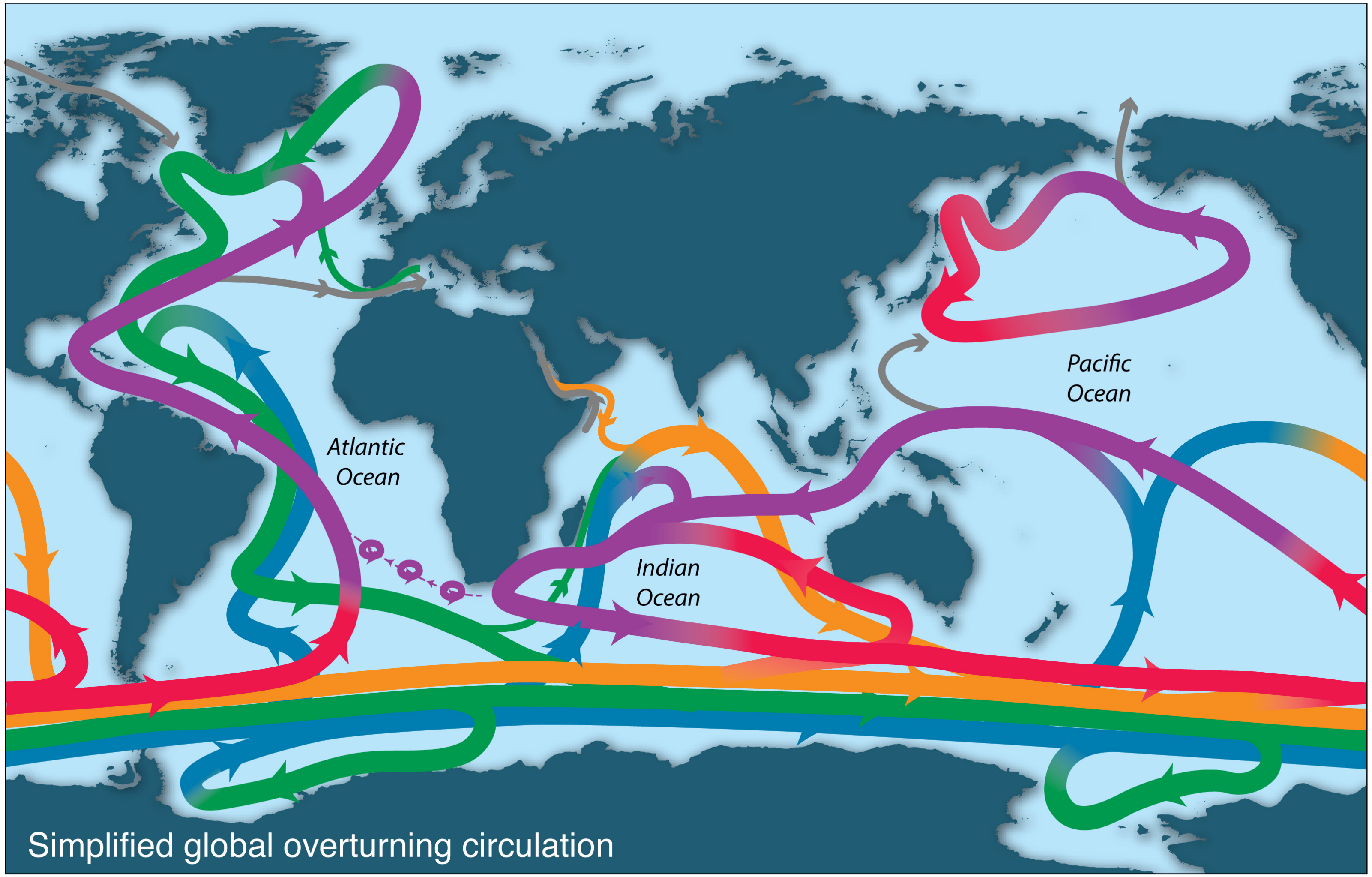


Deep convection

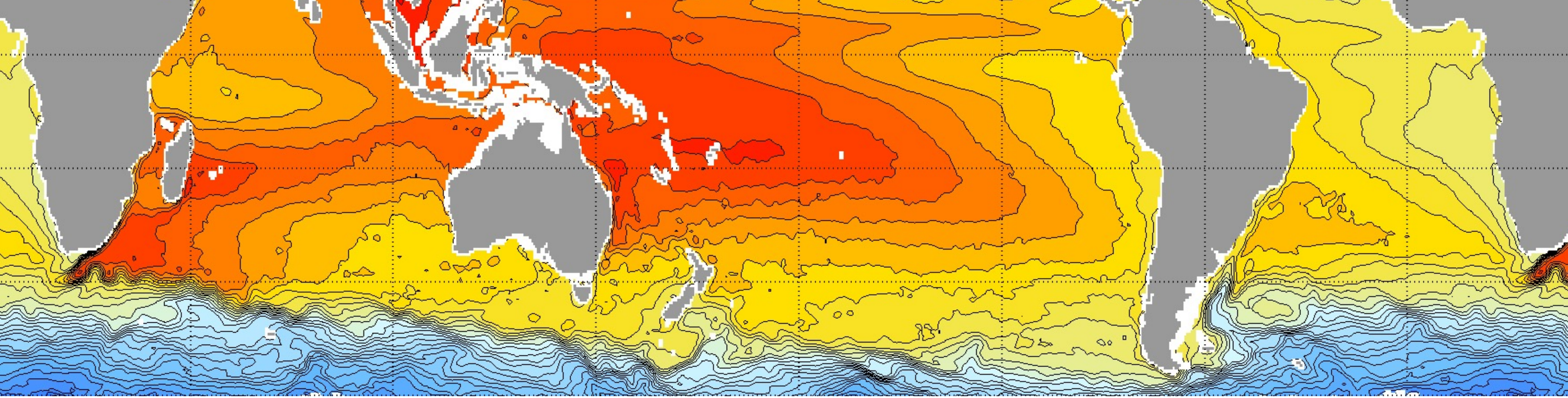
N. Atlantic Meridional Overturning Circulation (AMOC) (NADW)

Talley (Oceanography, 2013)

Global overturning schematic based on those transports and transformations



Talley (2013) based on DPO Fig. 14.11a



Dynamics: Geostrophy and Ekman flow, reference velocities
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Result: Meridional water mass distributions
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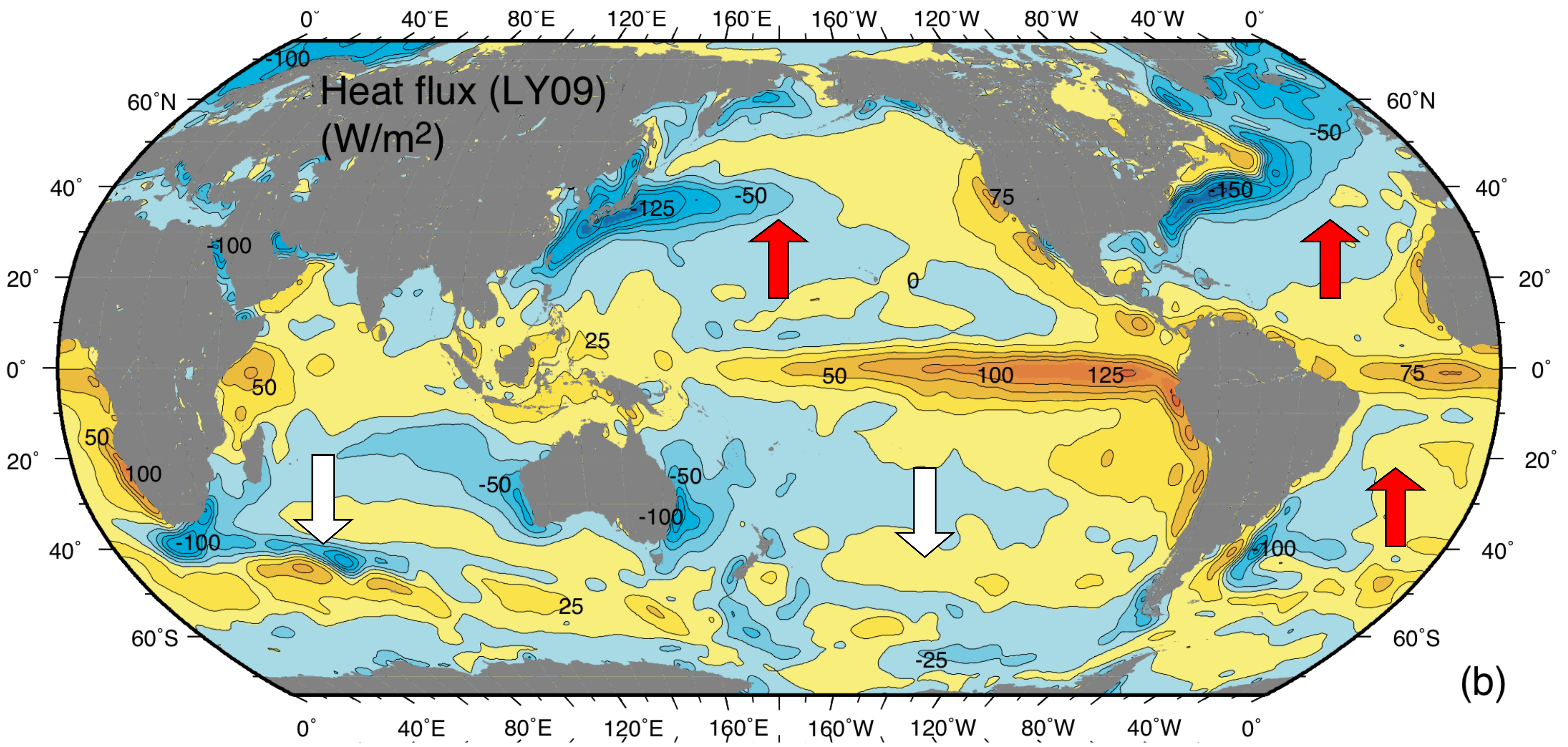
Result: Global overturning circulation mass transports
Result: Global heat, freshwater transports
Method: Budgets and implications for diffusivity
Result: Diapycnal diffusivity estimates
Dynamics: overturning circulation at large spatial scales
Result: Demonstration of large-scale force balance for GOC

Ocean heat transport: Air-sea heat flux



DPO Fig. 5.15

Yellow/orange - ocean gains heat. Blue - ocean loses heat.

Heat transport: red is northward, white is southward.

Two components:

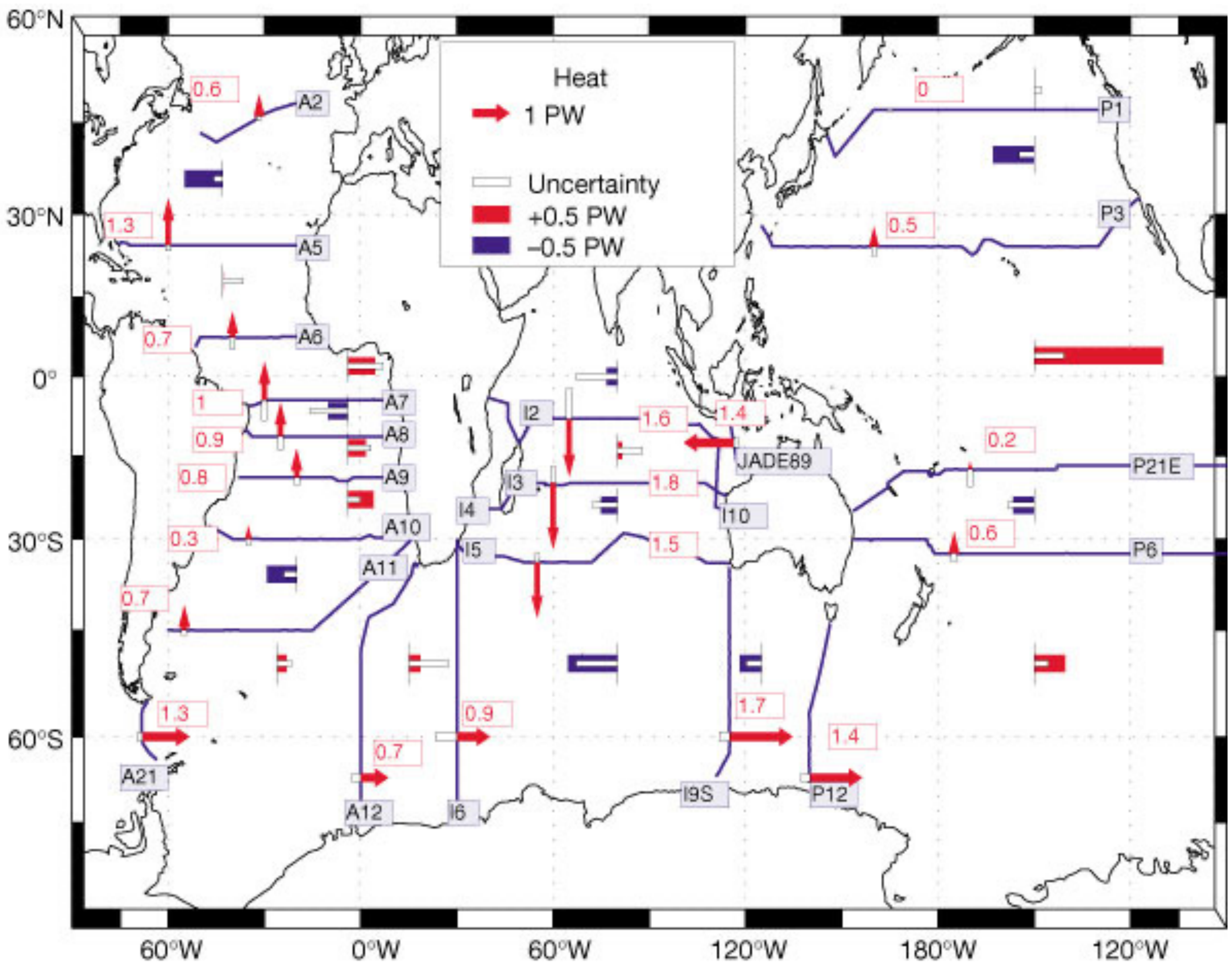
(1) poleward due to subtropical gyre circulation (warm western boundary current plus cooler subducted water).

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

Global heat transport

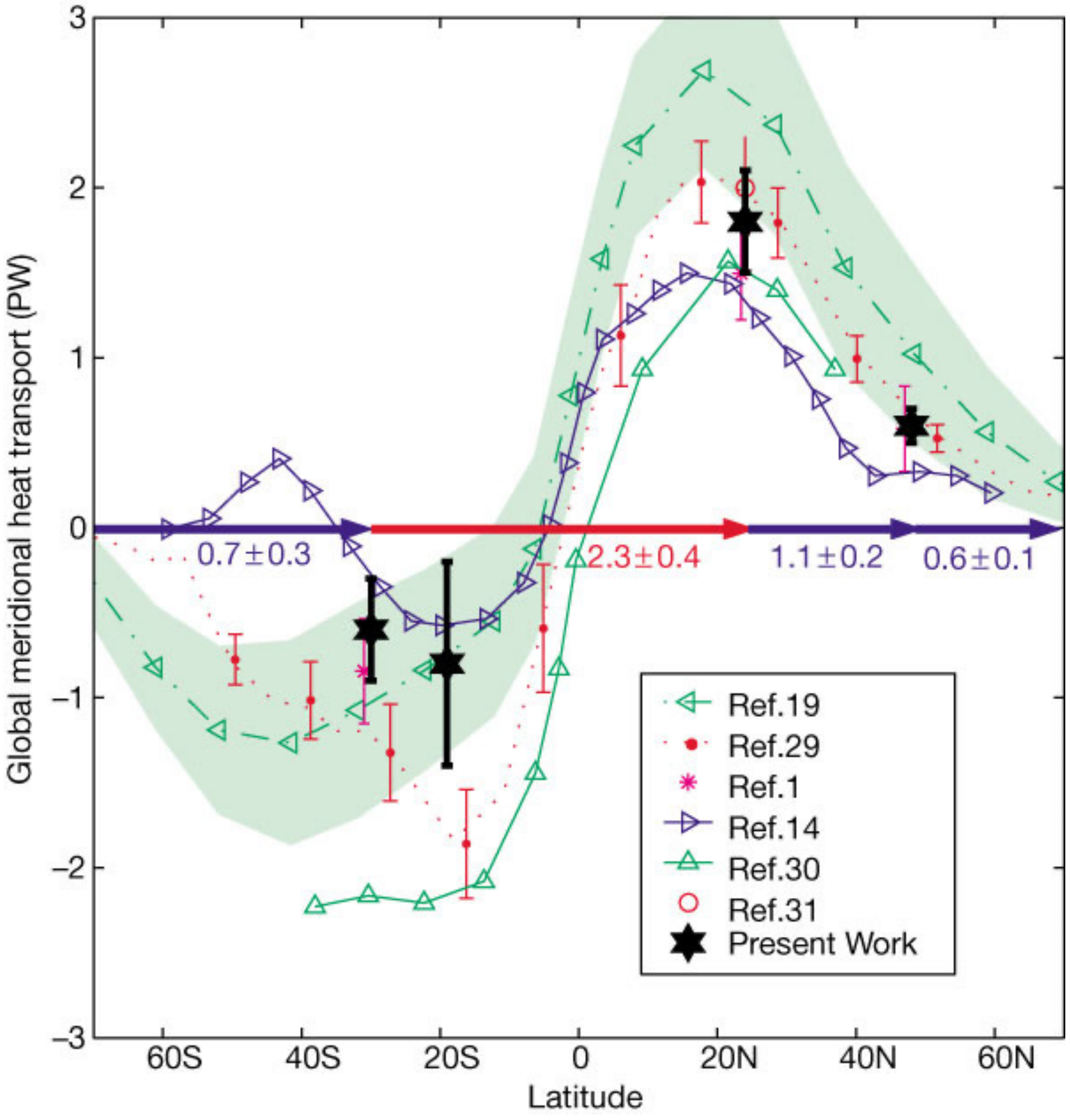
Inverse model

Ganachaud and Wunsch (2000)

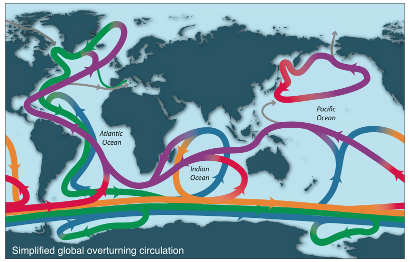


Global meridional heat transport

Ganachaud and Wunsch (2000)



Heat transports (mass balanced) for the Reid global overturning circulation



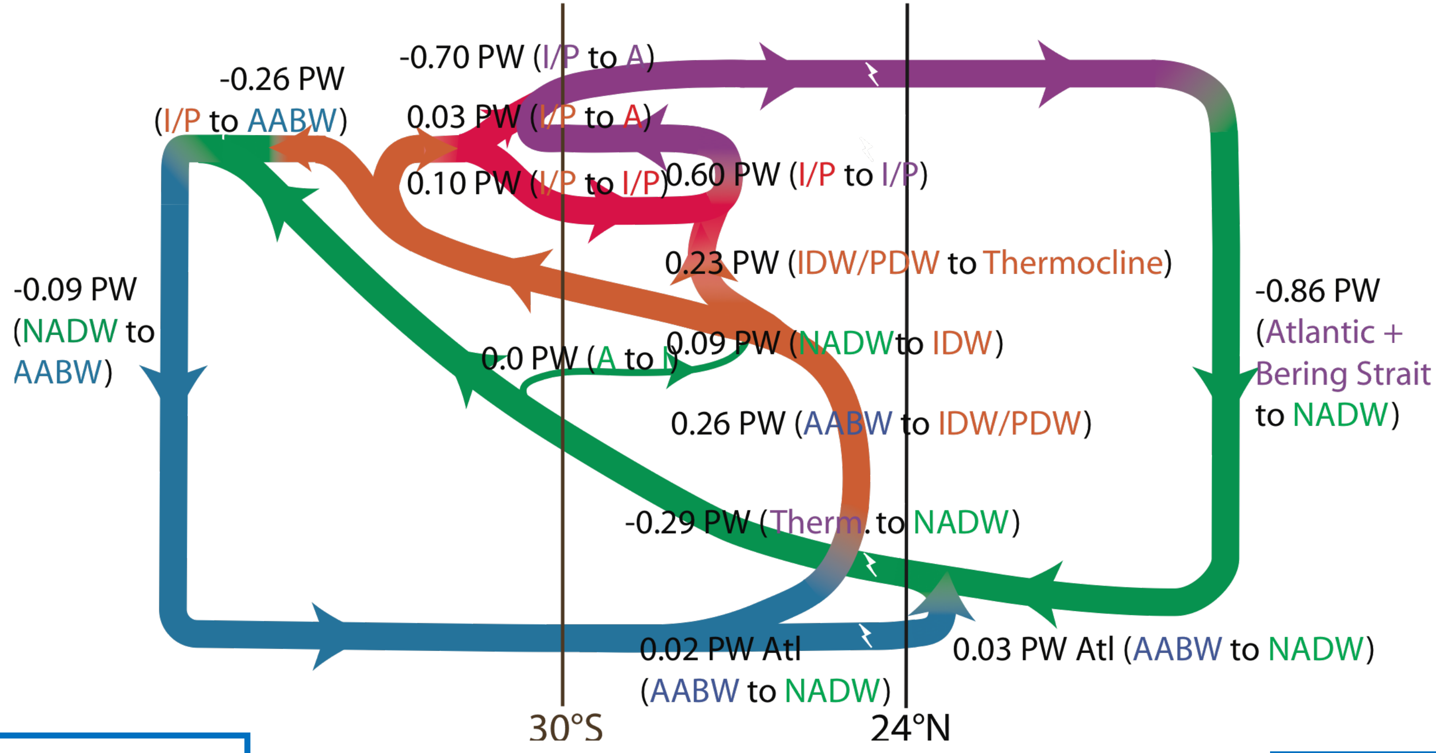
Overturning heat transports

Total gains and losses

-1.00 PW total heat loss

2.81 PW total heat gain

-1.81 PW total heat loss



Gains and losses due to GOC only

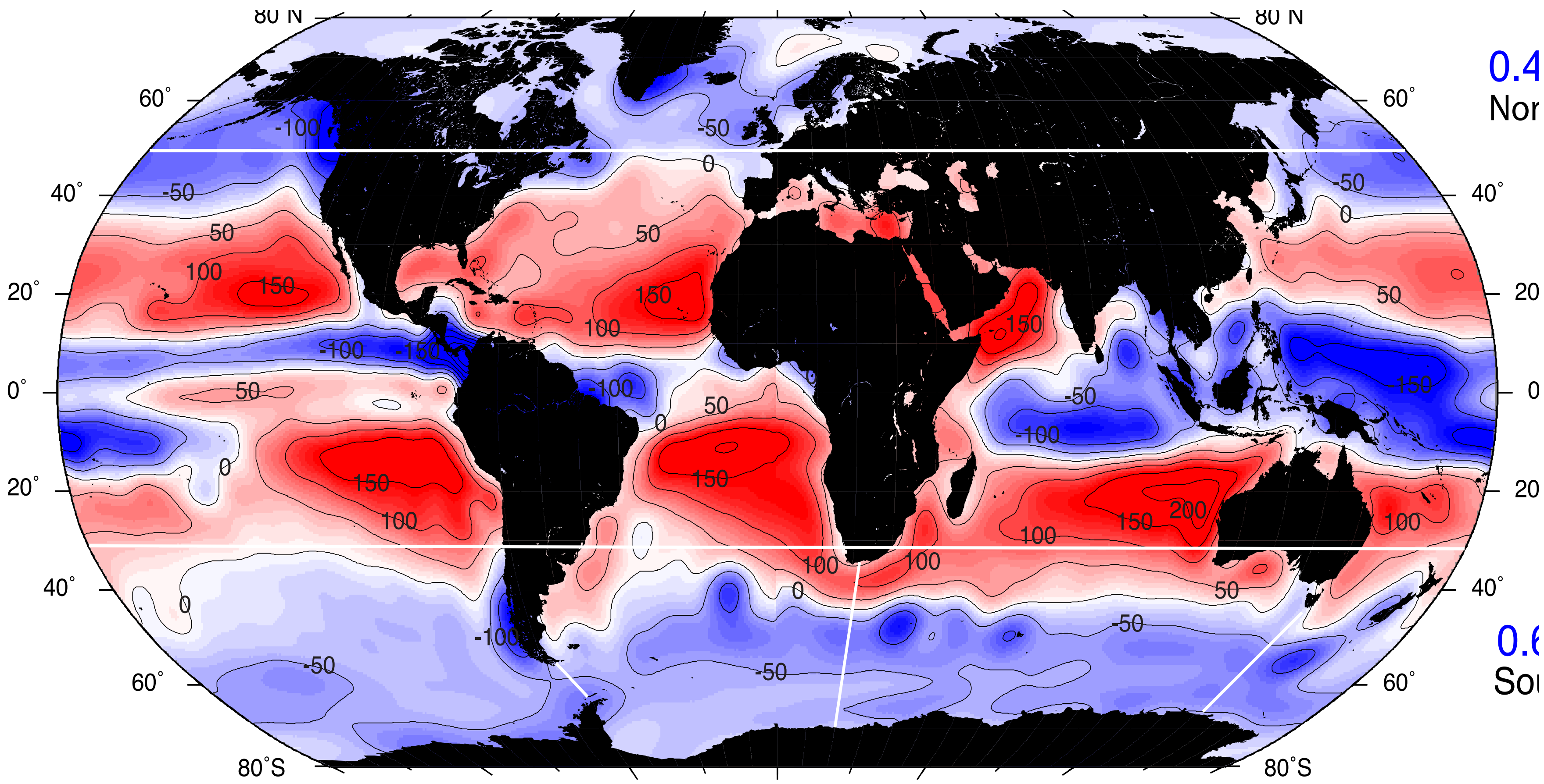
35% of SH heat loss goes into AABW production

42% of low latitude heat gain goes into the int./deep Indian and Pacific

48% of NH heat loss goes into NADW production

Talley (Oceanography 2013)

Ocean freshwater transport: Precipitation + runoff minus evaporation (cm/yr)



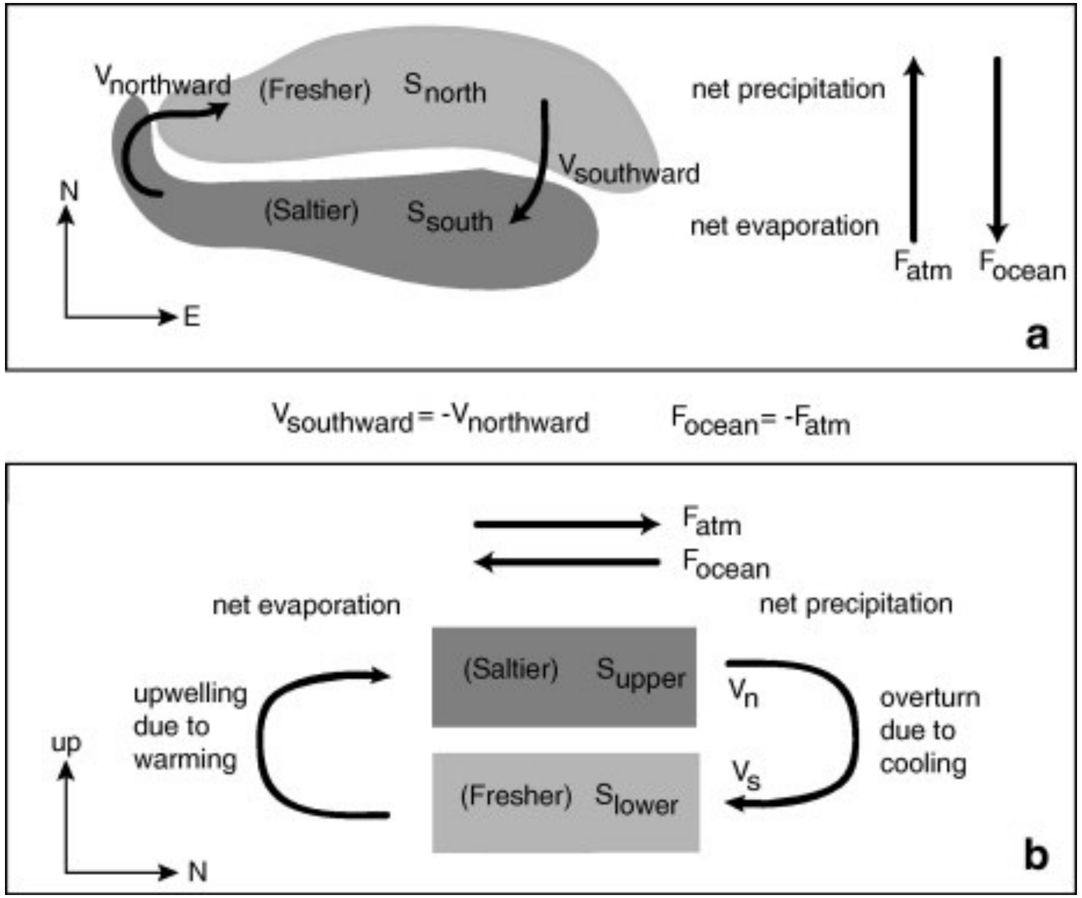
Blue: Net precipitation

Red: Net evaporation

DPO 5.4 NCEP climatology

Salinity is set by freshwater inputs and exports since the total amount of salt in the ocean is constant, except on the longest geological timescales

Define freshwater transport



Talley (2008)

- The atmosphere transports water vapor from evaporation to precipitation regions.
- The ocean transports freshwater from fresher to saltier regions.
- Atmosphere and ocean freshwater transports are equal and opposite
- Atmosphere F is set externally

- For larger atmospheric F transport, get larger ΔS
- For larger V transport exchange, get smaller ΔS

$F_{ocean} = V(S_{south} - S_{north})/S_o = V\Delta S/S_o$ where S_o is a reference salinity and V is the volume transport exchange between reservoirs

Therefore $\Delta S = FS_o / V$, so if F is set by the atmosphere and V is set externally by the flow, the salinity difference can be calculated. (Of course V can depend on ΔS, as in the Stommel thermohaline oscillator.)

Conservation of freshwater: the common practical approach

Mass: $F = -\rho_o V_o + \rho_i V_i = -(R + A_s P) + A_s E$ (positive is in to volume)

Salt: $\xi = -\rho_o V_o S_o + \rho_i V_i S_i = 0$

Salt divided by an arbitrary constant, choose to be about equal to mean salinity S_m:

$$\xi / S_m = \rho_o V_o S_o / S_m - \rho_i V_i S_i / S_m = 0$$

Subtract $F - \xi / S_m = F - 0$

$$F - \xi / S_m = \rho_i V_i (1 - S_i / S_m) - \rho_o V_o (1 - S_o / S_m)$$

Assume $\rho_i V_i \sim \rho_o V_o = \rho V$ given how small F really is, so

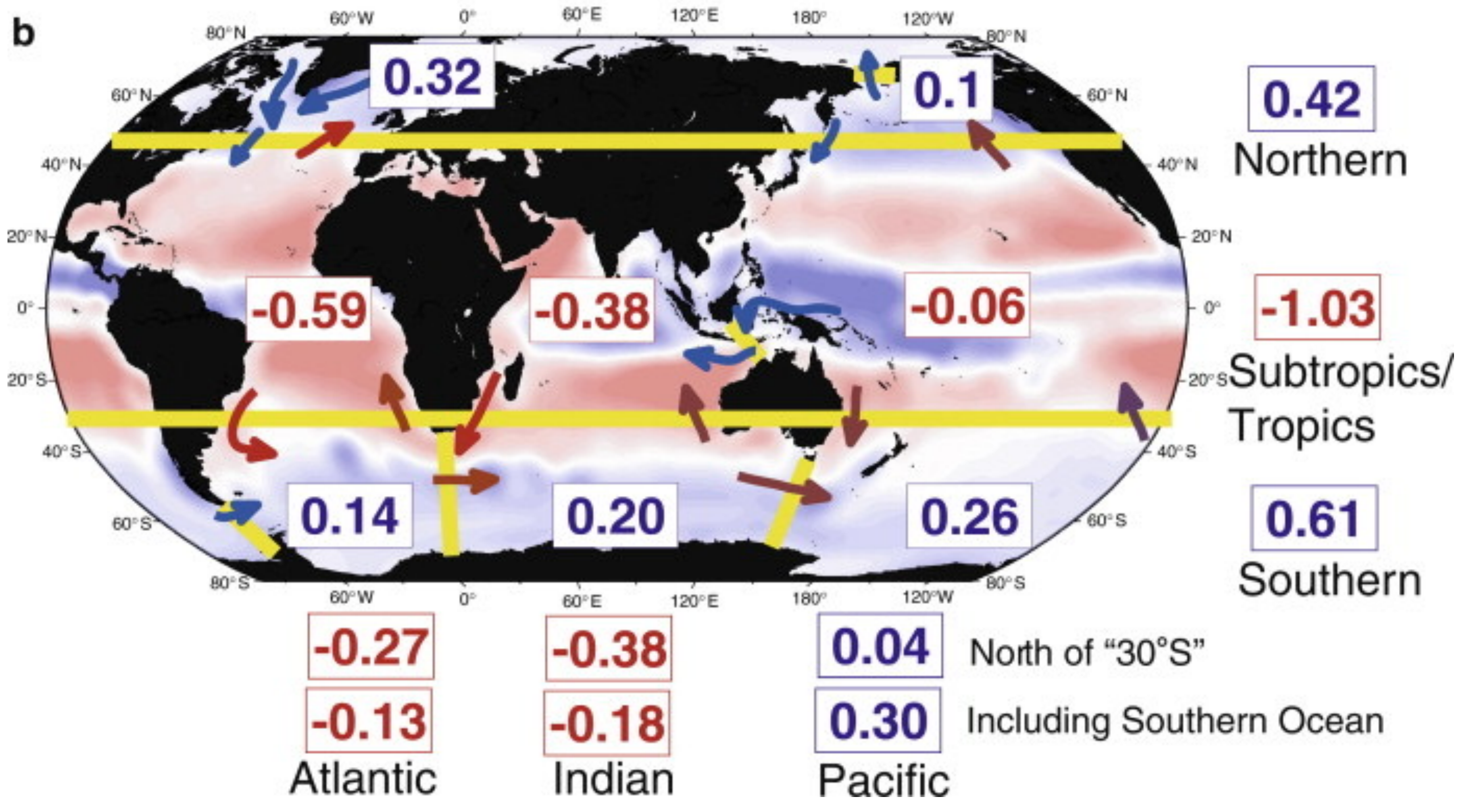
$$\boxed{F \sim \rho V (S_o / S_m - S_i / S_m) = \rho V (S_o - S_i) / S_m = -(R + A_s P) + A_s E}$$

Freshwater transport.

→ Freshwater input calculated from the difference in salinity between inflow and outflow equals the net precipitation, evaporation, runoff

Net freshwater divergence (Sv) from hydrographic section transports

- Calculated from geostrophic/Ekman velocity and salinity
- (Compare with air-sea flux E-P-R)



Meridional ocean transports of freshwater from high to low latitudes

"Zonal" ocean transports of freshwater from Pacific to Atlantic/Indian

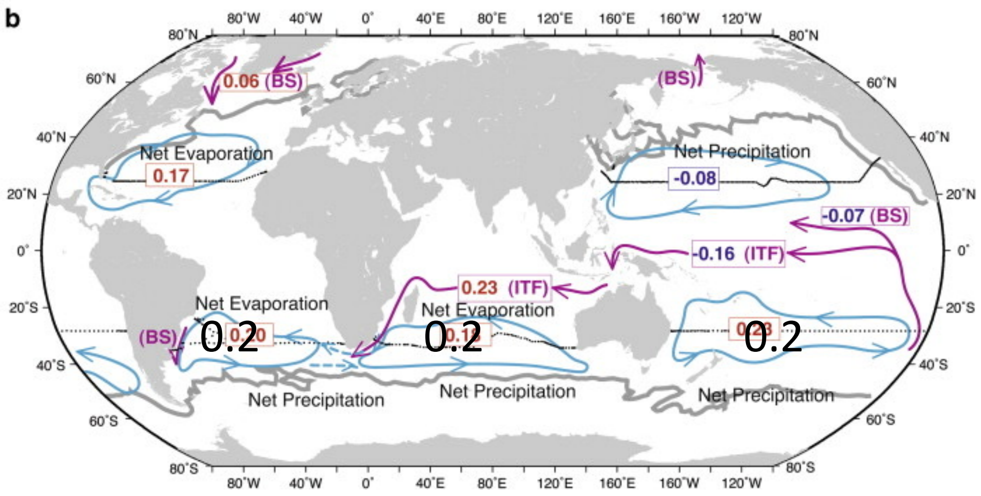
Talley (PiO 2008)

Northern and southern hemisphere FW transport mechanism asymmetry

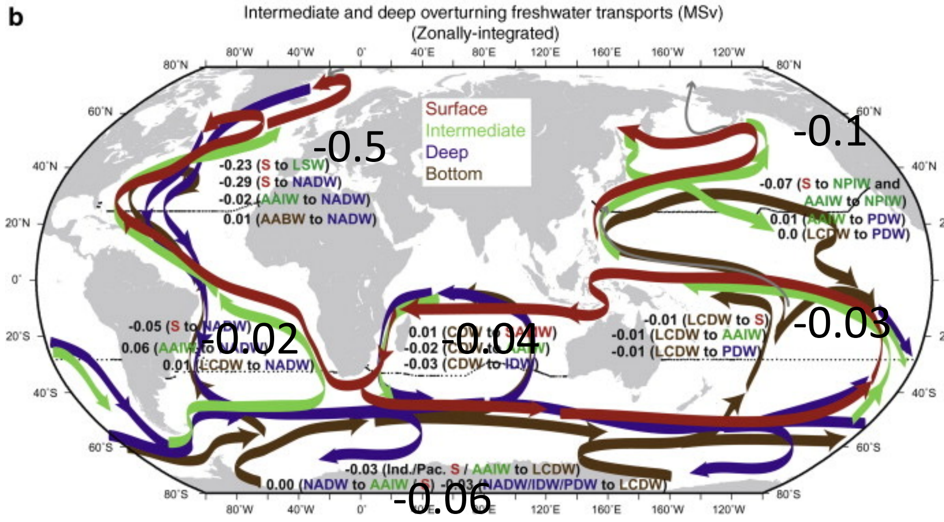
Talley (PiO 2008)

0.09 Sv

0.6 Sv



Upper ocean pathways



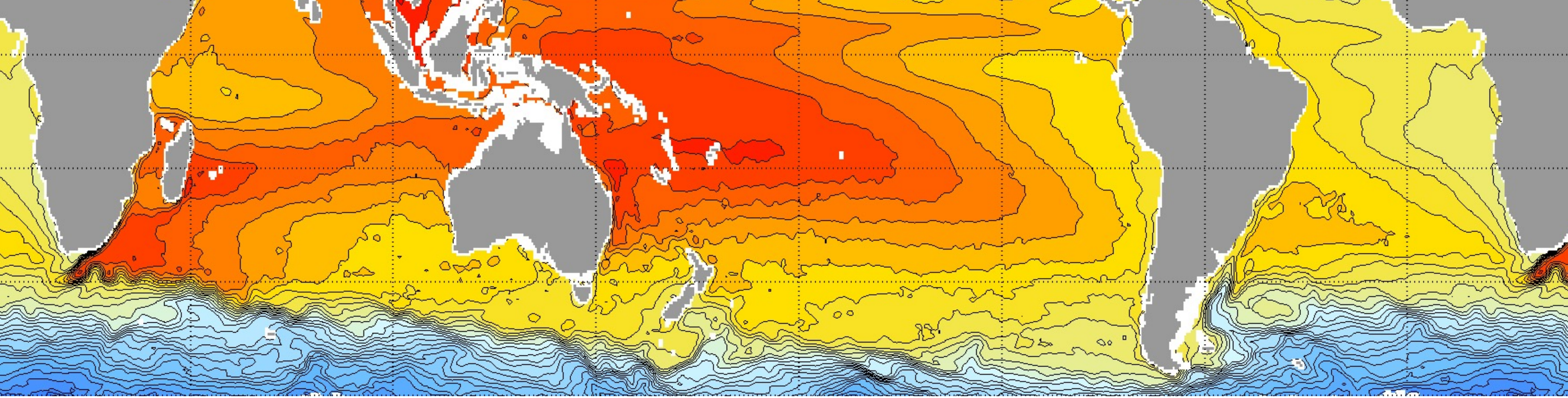
Intermediate and deep pathways

-0.6 Sv

-0.09 Sv

-0.06 Sv

- **Northern** hemisphere exports freshwater southward through **NADW and NPIW formation**
- **Southern** hemisphere exports freshwater northward through **upper ocean gyres**, (order of magnitude less through intermediate and deep water formation)
- Why: Salinity difference between inflow and outflow is larger if salty, warm subtropical surface water is transformed into fresh, cold deep/bottom water. **Drake Passage inhibits transport of surface waters to Antarctica** (Toggweiler and Samuels, 1995).
- Dumped freshwater on the Antarctic just floats in surface layer, can’t become dense enough to sink since temperature difference between surface and deep water is small.



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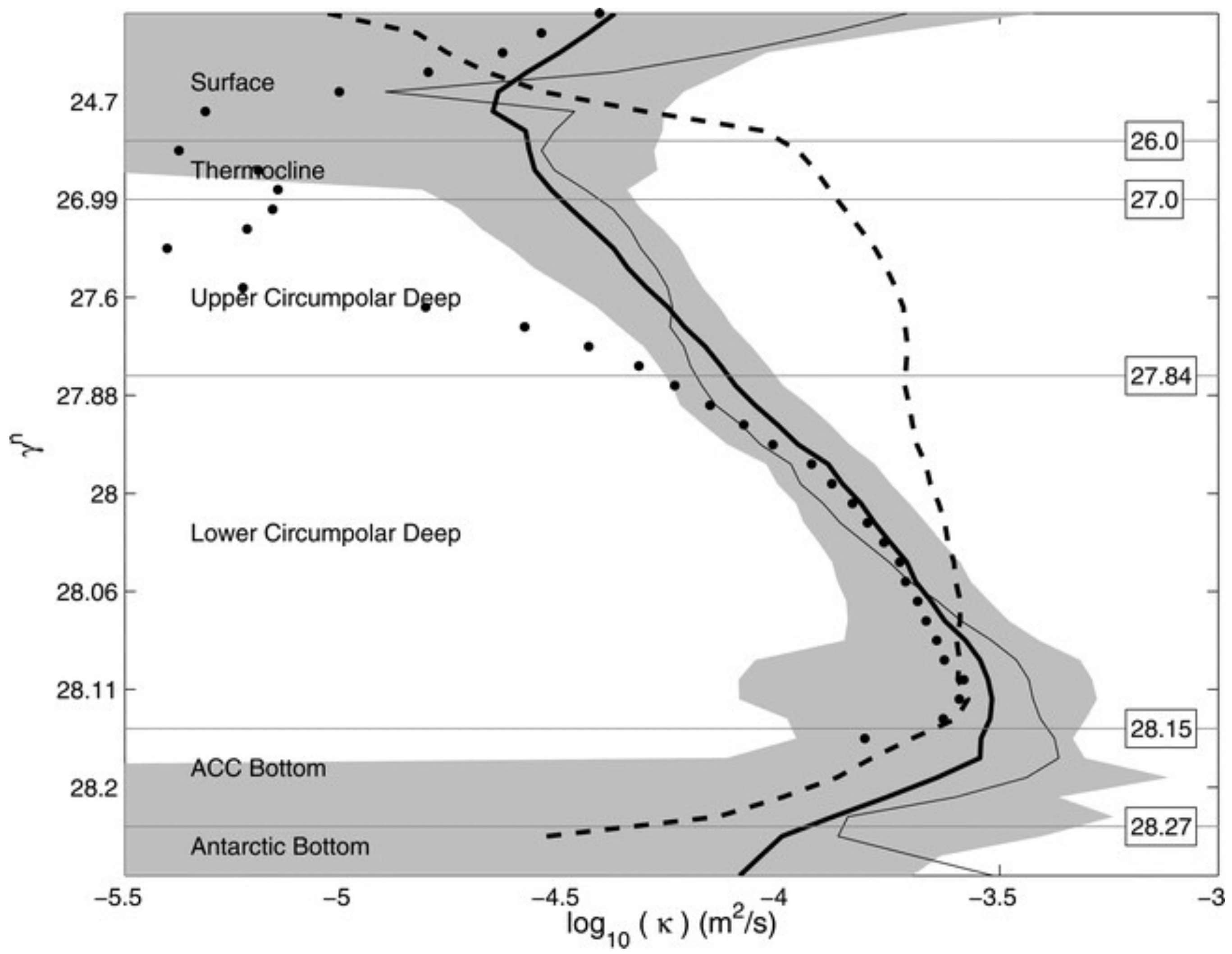
Dynamics: overturning circulation at large spatial scales
Result: Demonstration of large-scale force balance for GOC

Implications of calculated GOC for diapycnal diffusivity

Use vertical velocity w from inverse model
Vertical profiles of buoyancy
Estimate diapycnal diffusivity

Munk 1966

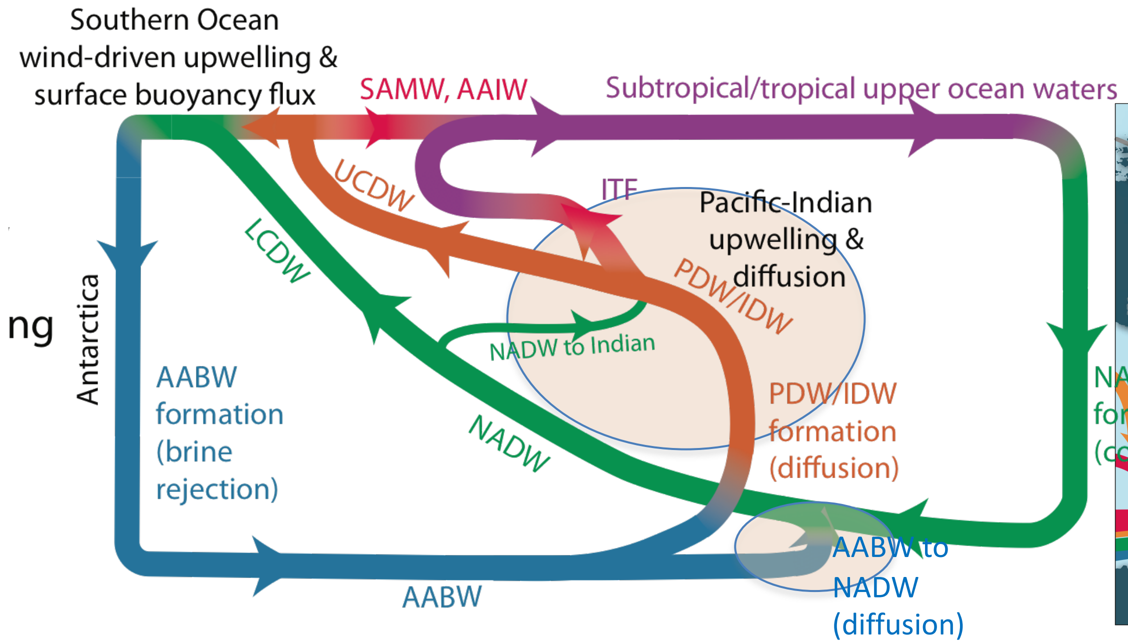
$$w\frac{\partial b}{\partial z} = \frac{\partial}{\partial z}\left(\kappa\frac{\partial b}{\partial z}\right)$$

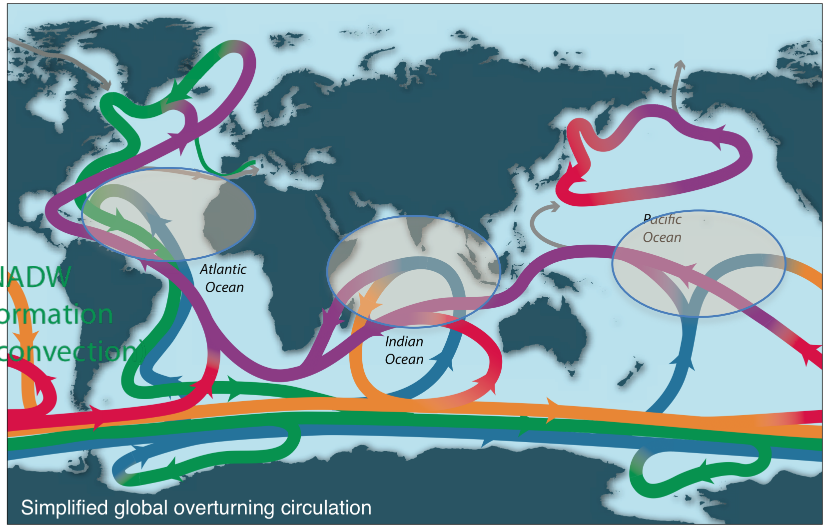


Lumpkin and Speer (2007) global hydrographic inverse model

Conclusions from Talley et al (2003) were very similar – need 10^{-4} m^2/sec

Implications of calculated GOC for diapycnal diffusivity





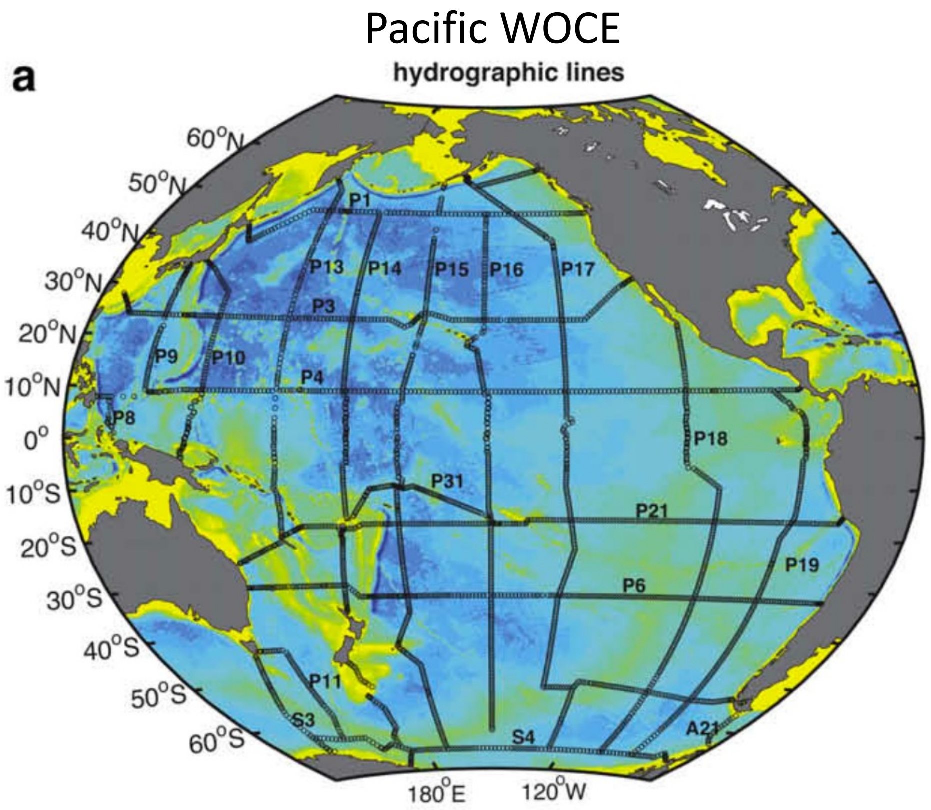
Low latitude vertical velocities and diffusivities diagnosed from this overturning:
Order 10^{-4} m^2/sec (Munk values)

(Talley, Reid, Robbins, J. Clim. 2003)

	w (cm/sec)	κ (m²/sec)	w (cm/sec)	κ (m²/sec)
	(if 10.5°S to 10.5°N)		(if all equatorial 2.5°S to 2.5°N)	
Indian	7×10^{-5}	1.7×10^{-4}	31×10^{-5}	6.8×10^{-4}
Pacific	4×10^{-5}	0.8×10^{-4}	16×10^{-5}	3.5×10^{-4}
Atlantic	7×10^{-5}	1.5×10^{-4}	25×10^{-5}	5.5×10^{-4}

Implications of calculated GOC for diapycnal diffusivity

Pacific WOCE



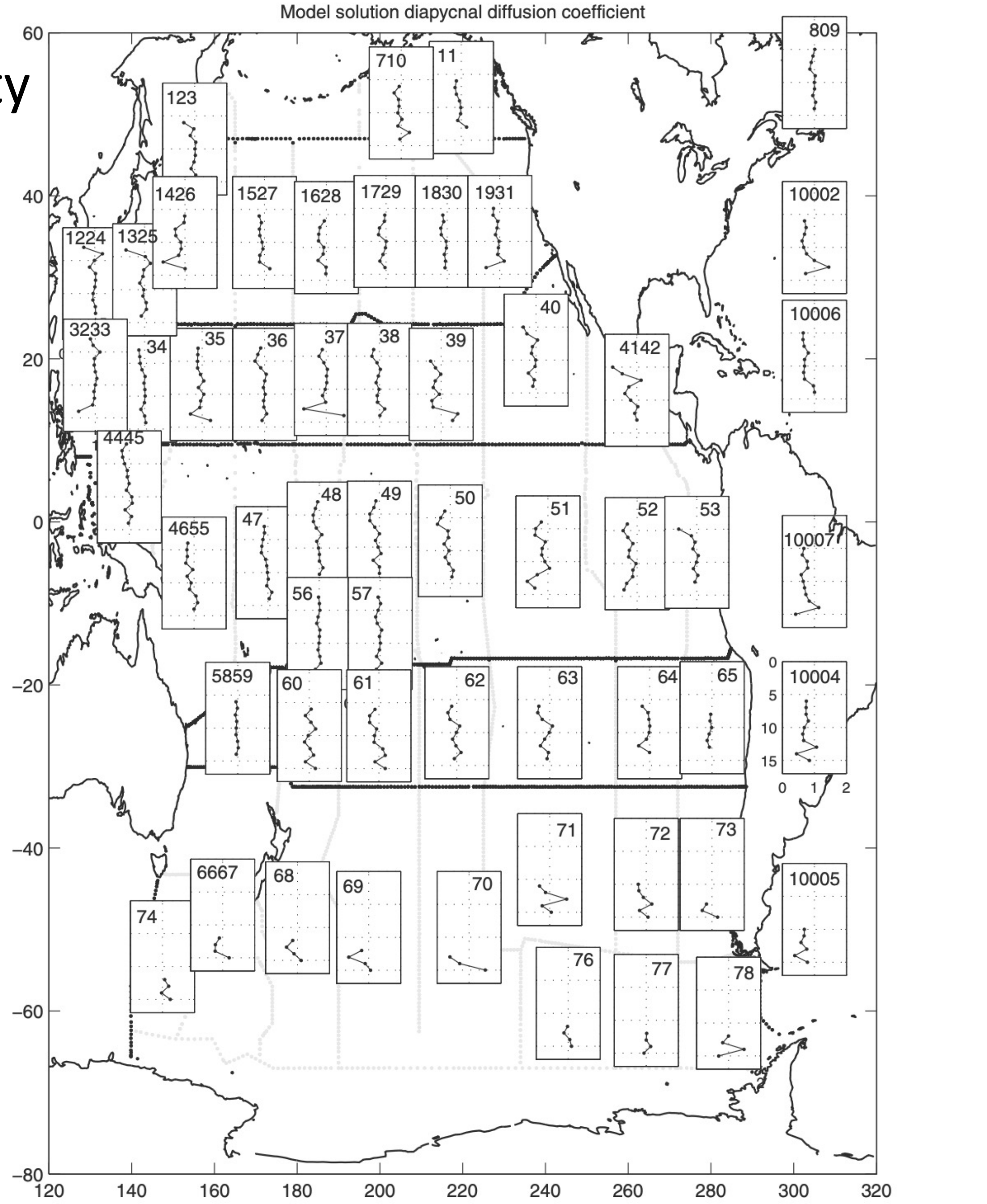


Fig. 11. Inverse estimates of diapycnal diffusivities in each of the small boxes (as labeled) and in each the large boxes (right-hand column). Axes are the same for a but labeled only on box 10004 (second up from the bottom on the right). Inverse solutions for $\mathbf{k}_z$ only exist for non-outcropping interfaces. Interface 0 is the surface. 16 is the bottom. Values are in units of $cm^2\ s^{-1}$.

Macdonald et al. (PiO 2009)

Overturning transports diagnosed from diapycnal diffusivity

Cimoli et al. (AGU Advances, 2023)

AGU ADVANCING EARTH AND SPACE SCIENCE

AGU Advances

10.1029/2022AV000800

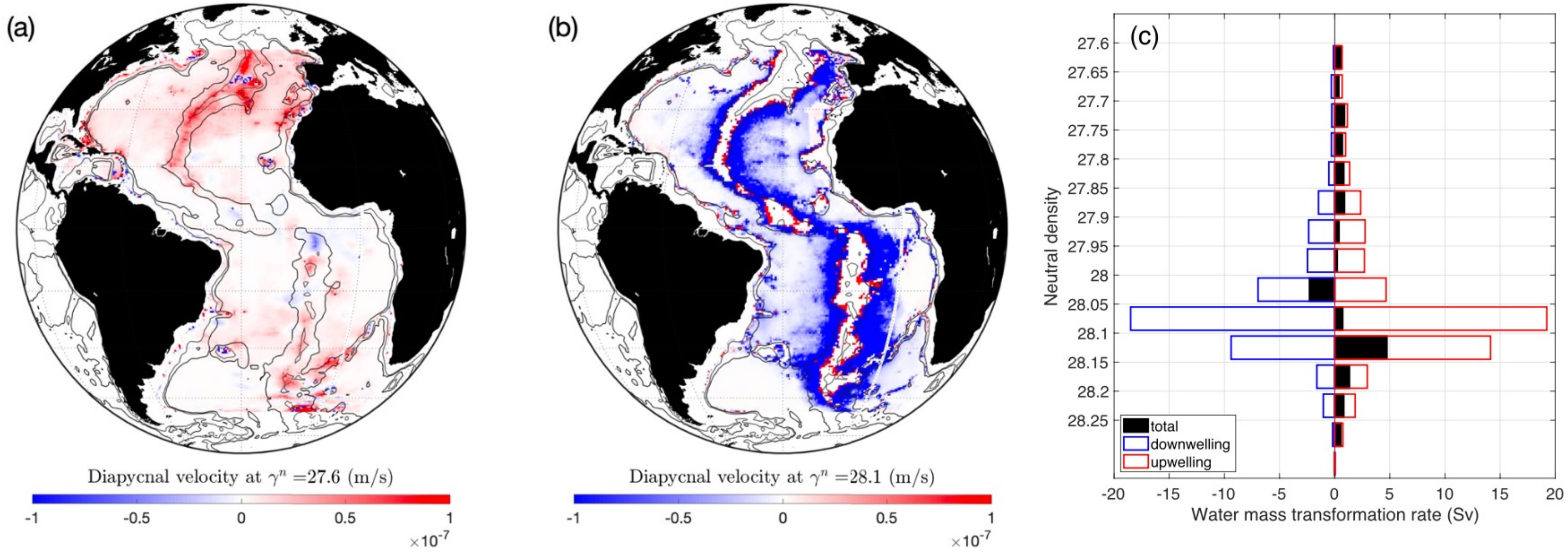
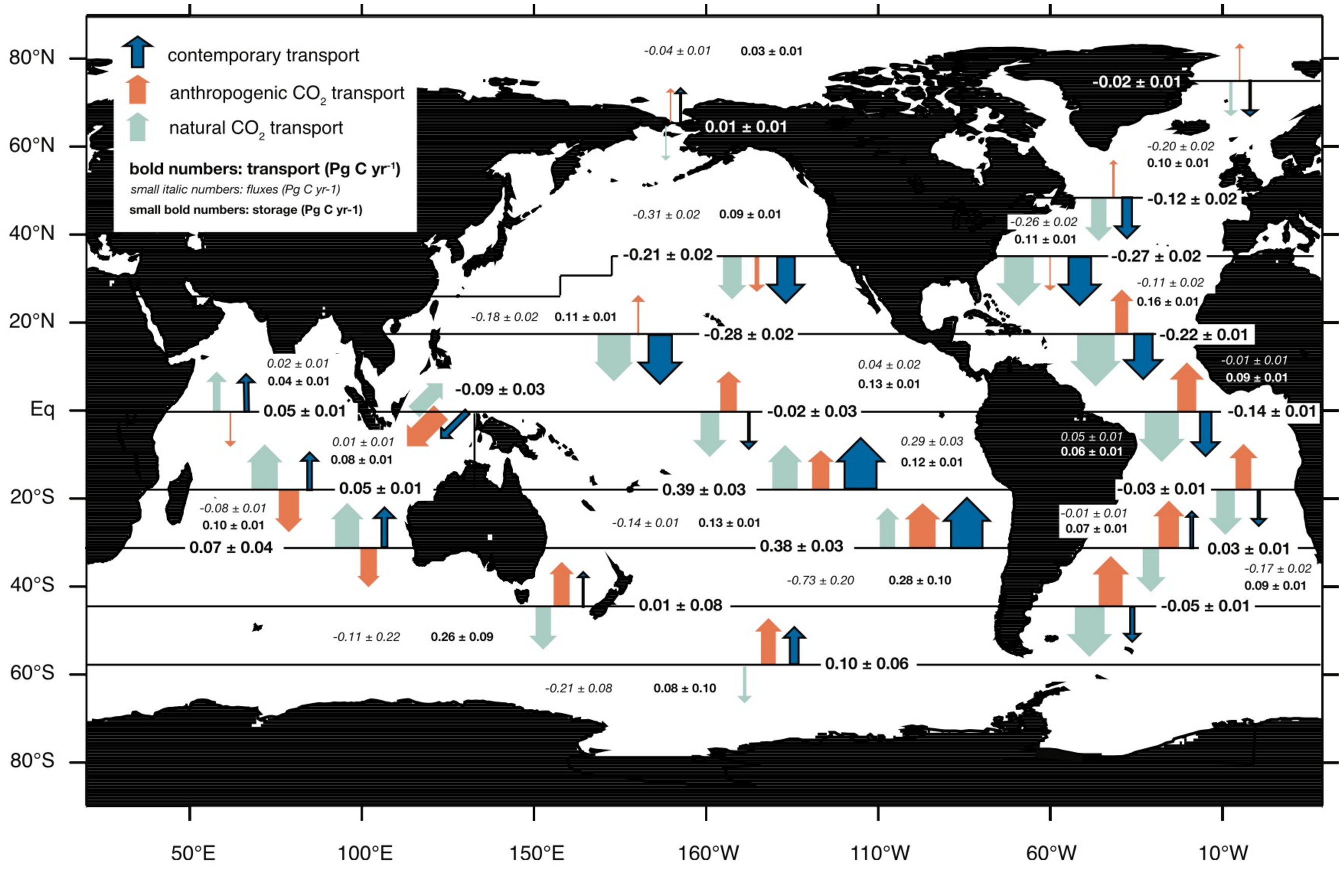
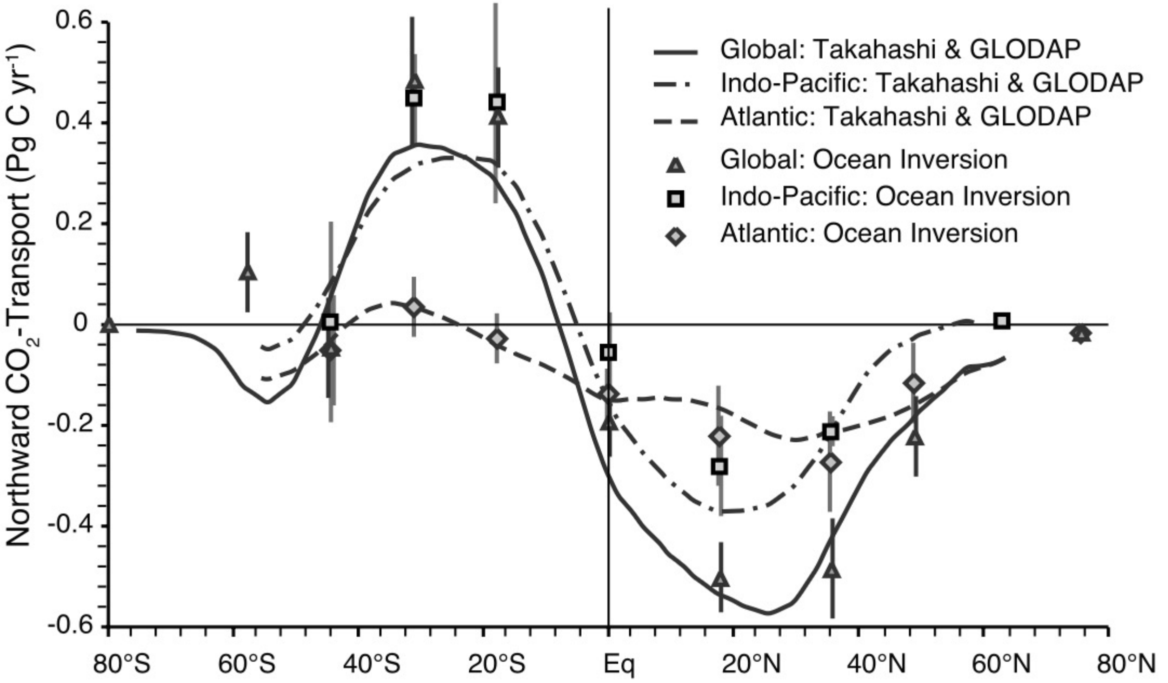


Figure 3. (a and b) Diapycnal velocity (Equation 2) calculated from the tidally driven mixing estimate on the neutral density surfaces $\gamma^n = 27.6$ (a) and $\gamma^n = 28.1$ (b). Positive values (red) indicate diapycnal upwelling, and negative values (blue) indicate diapycnal downwelling. The 3,000 and 4,000 m isobaths are also shown (thin black lines). (c) Water mass transformation rate (Equation 3) for the tidally driven mixing estimate across the density surfaces bounding the North Atlantic Deep Water flow ($\gamma^n = 27.6$–28.15). For each density surface, the total upwelling and downwelling are shown by the empty red and blue bars, respectively, while their residual is shown by the filled black bar. Positive transformation corresponds to a decrease in density.

Moving to other (GO-SHIP) properties: carbon

Hydrographic section inverse model



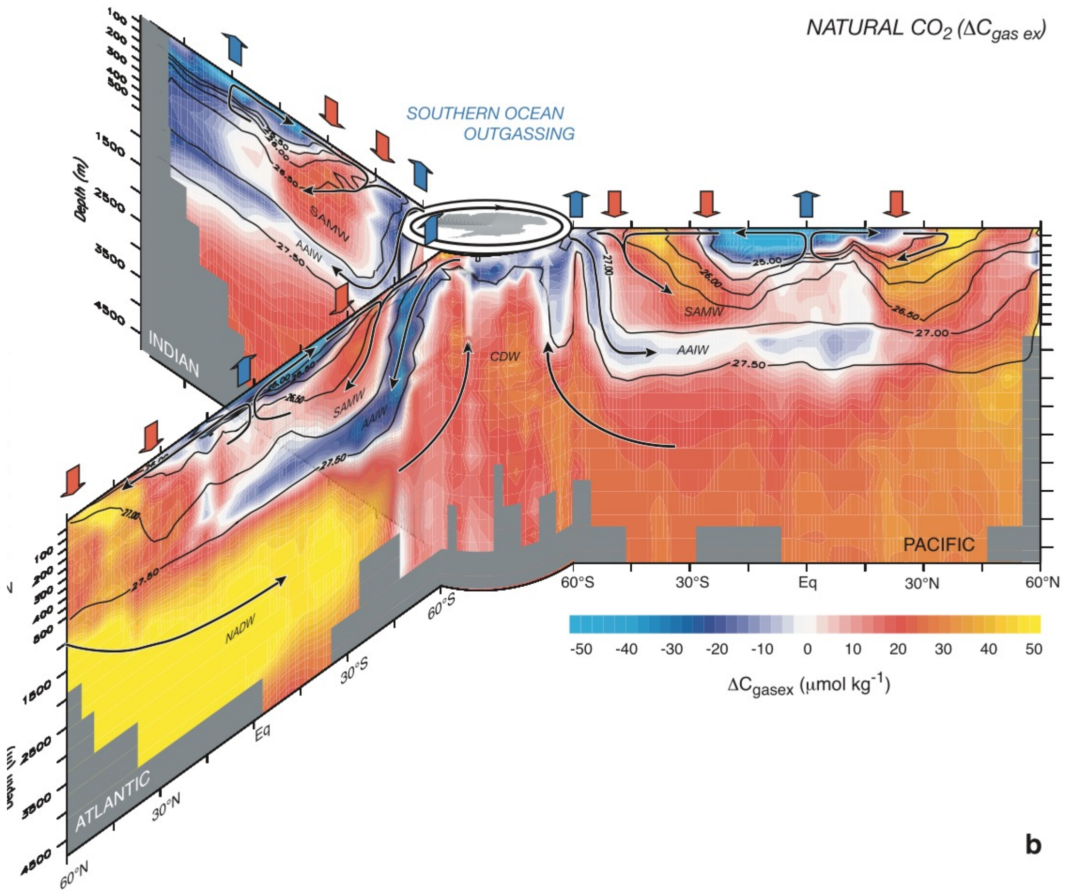


Gruber et al. (GBC 2009)

Moving to other (GO-SHIP) properties: carbon

Carbon: hydrographic inverse

Carbon: model

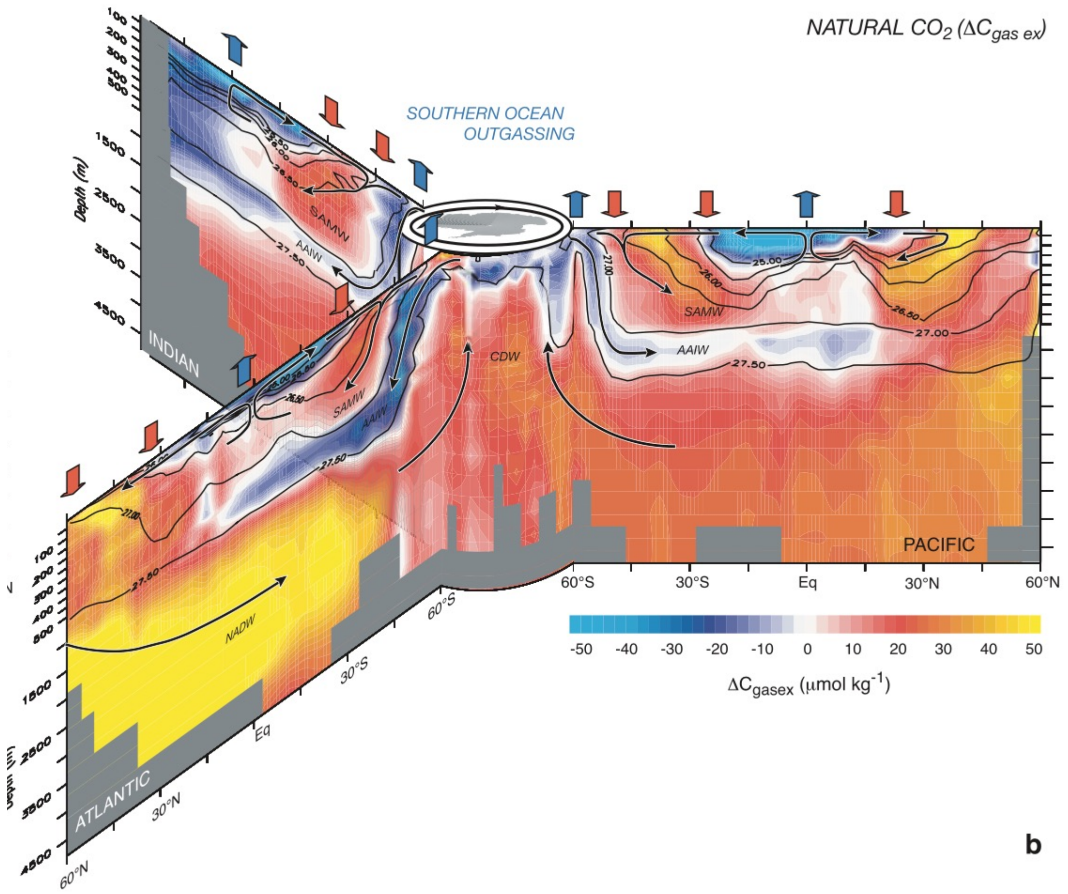


Gruber et al. (GBC 2009)

Aldama-Campino et al. (GBC 2020)

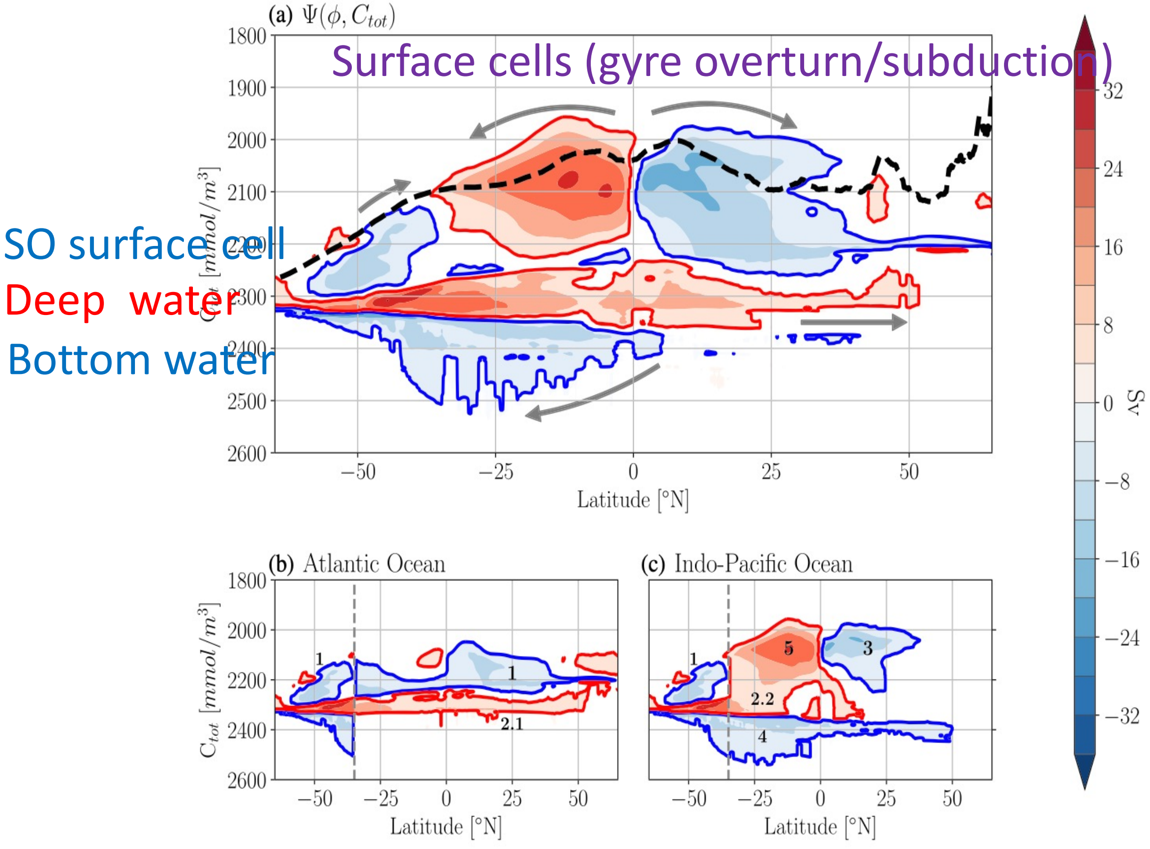
Moving to other (GO-SHIP) properties: carbon

Carbon: hydrographic inverse



Gruber et al. (GBC 2009)

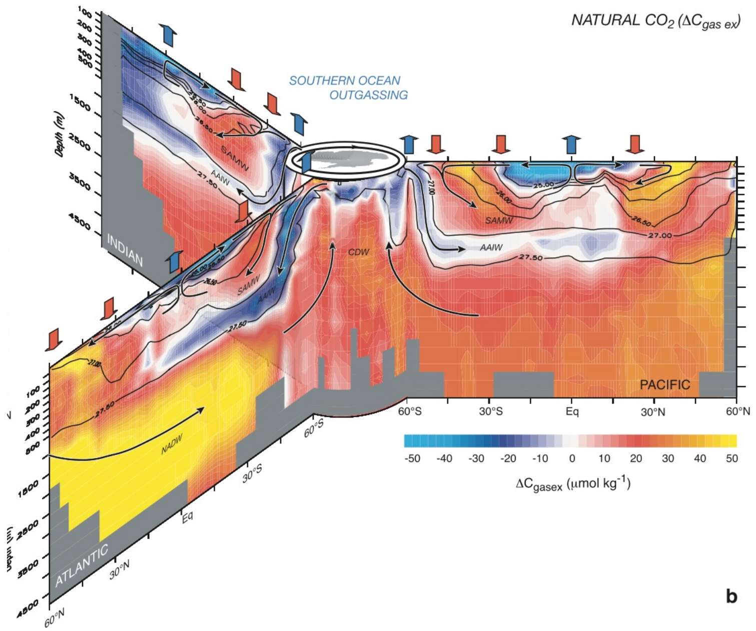
Carbon: model, focus on processes



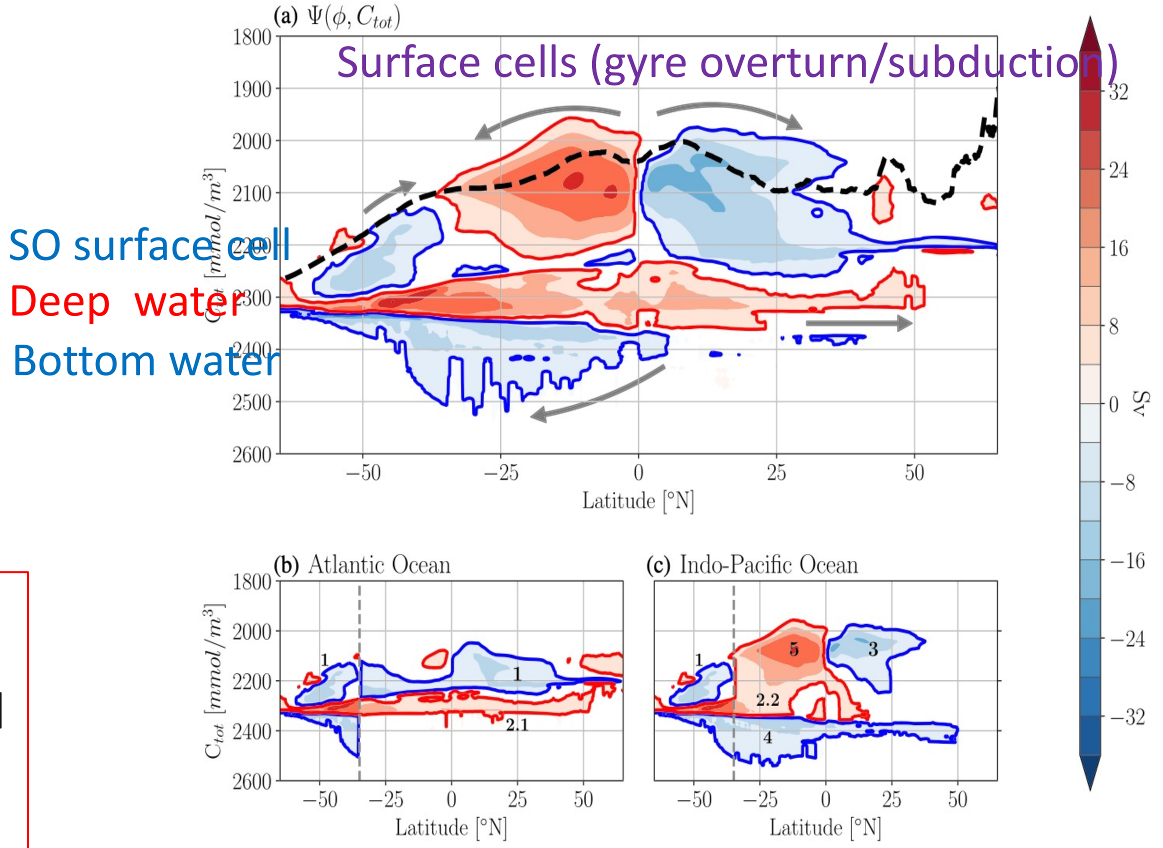
Aldama-Campino et al. (GBC 2020)

Moving to other (GO-SHIP) properties: carbon

Carbon: hydrographic inverse (Gruber et al., 2009)



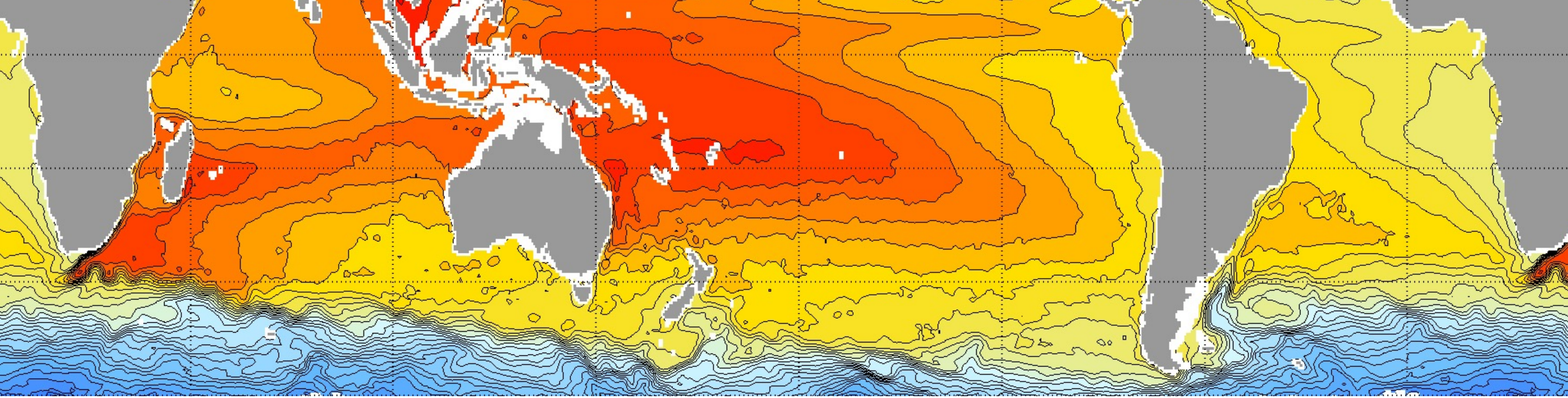
Carbon: model, focus on processes (Aldama-Campino et al., 2020)



Ideas to do:

Useful to revisit hydrographic inverse model results for processes identified in A-C.

Useful to use hydrographic inverses for decadal changes in carbon transports



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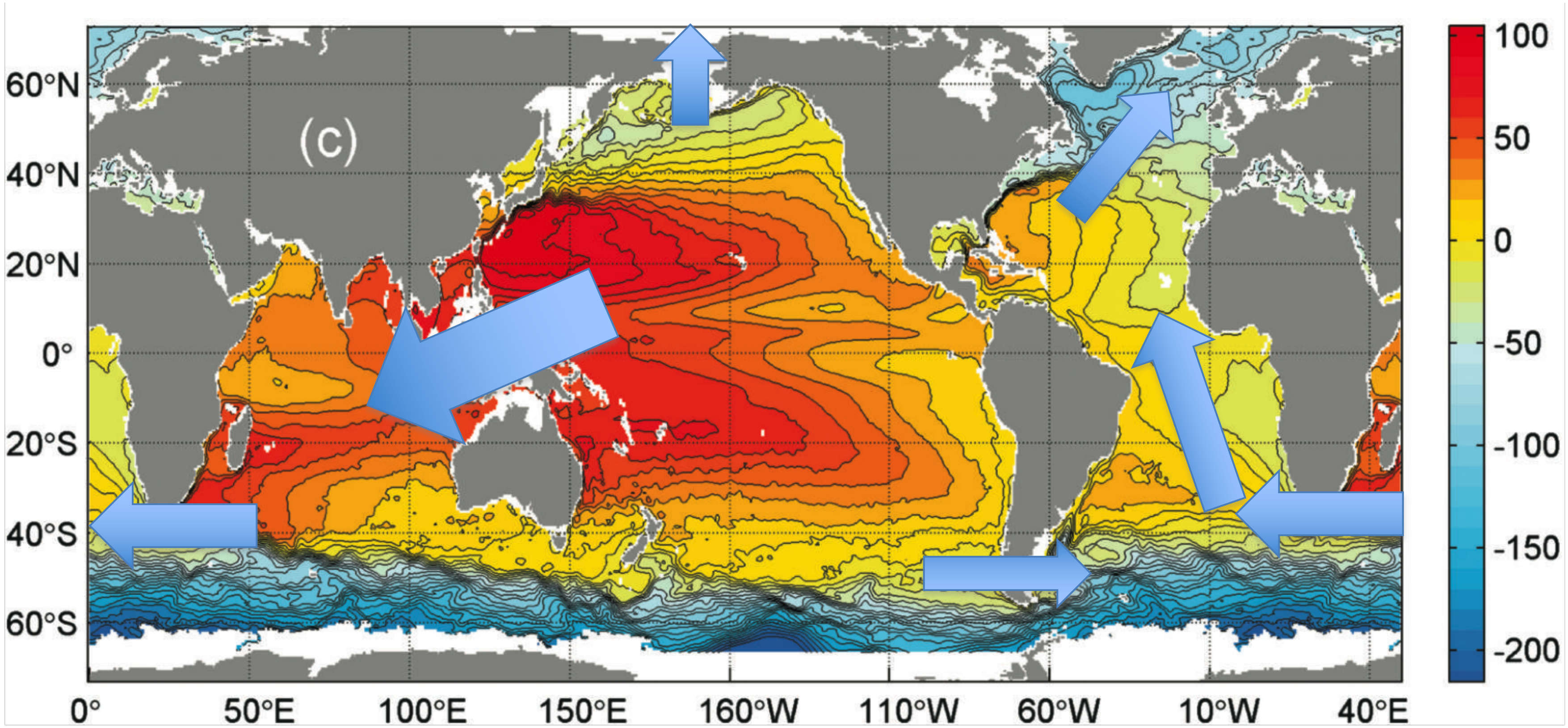
Result: Demonstration of large-scale force balance for GOC

Large-scale steric height distributions and the GOC

Sea surface height:

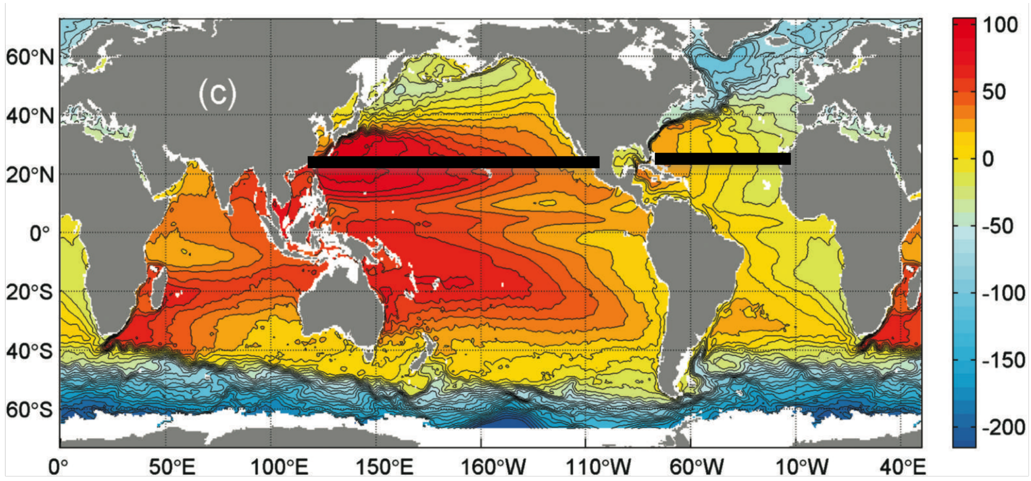
Pacific/Atlantic asymmetry

Superimpose Gordon conveyor belt plus 'cold water path' through Drake Psg. – remarkably similar to simply drawing vectors that go down the pressure gradient at the largest scale.

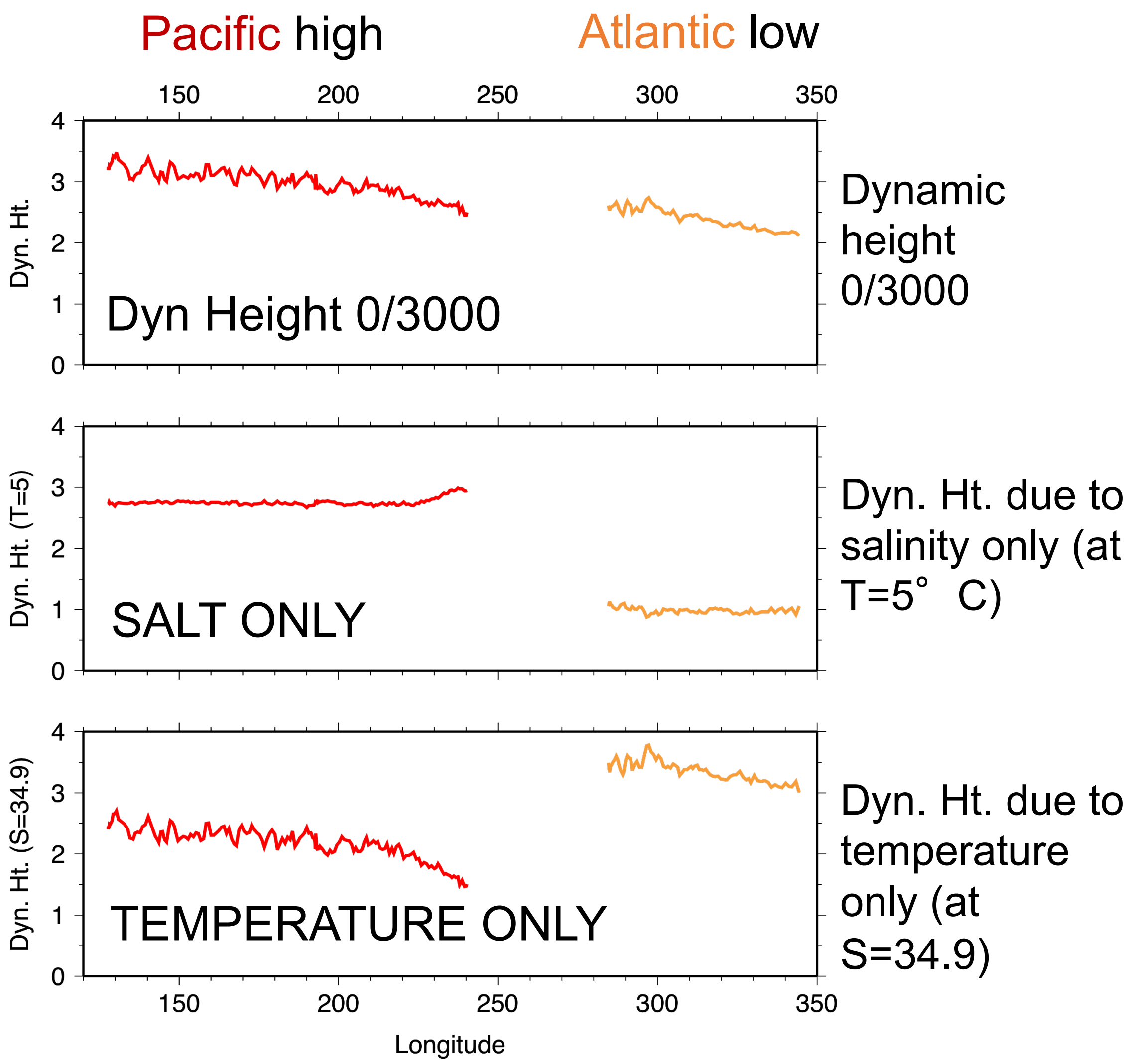


Global surface height based on surface drifter observations (Maximenko et al., 2009); similar to all modern surface height maps

Salinity difference causes the overall Pacific-Atlantic surface height difference



Surface dynamic height at 24°N



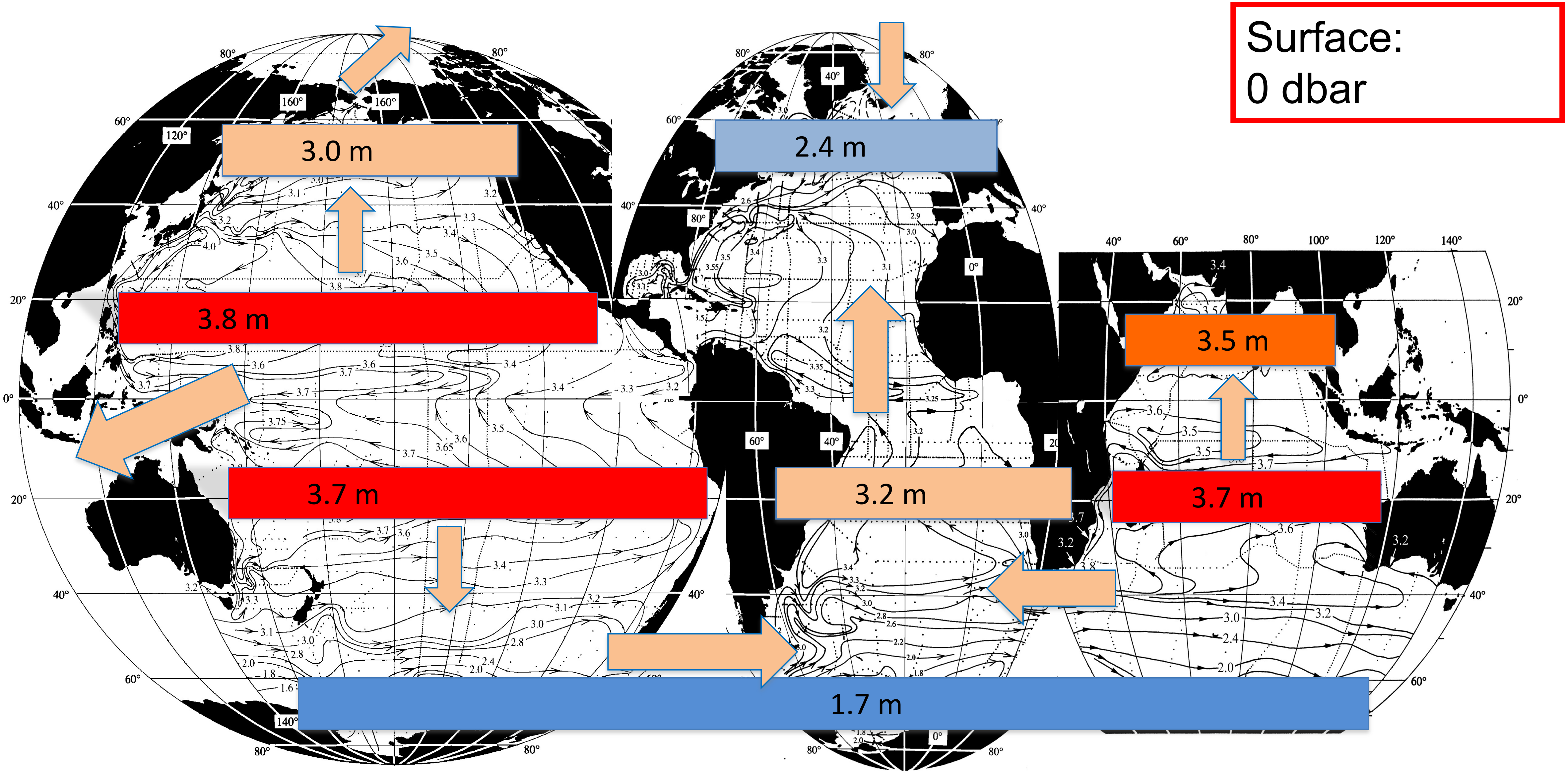
Fresher Pacific and saltier Atlantic leads to higher Pacific dynamic height

Colder Pacific partially offsets salinity difference.

Salinity difference drives the Global Overturning Circulation (dense N. Atlantic and light N. Pacific). (Stocker and Wright, 1991)

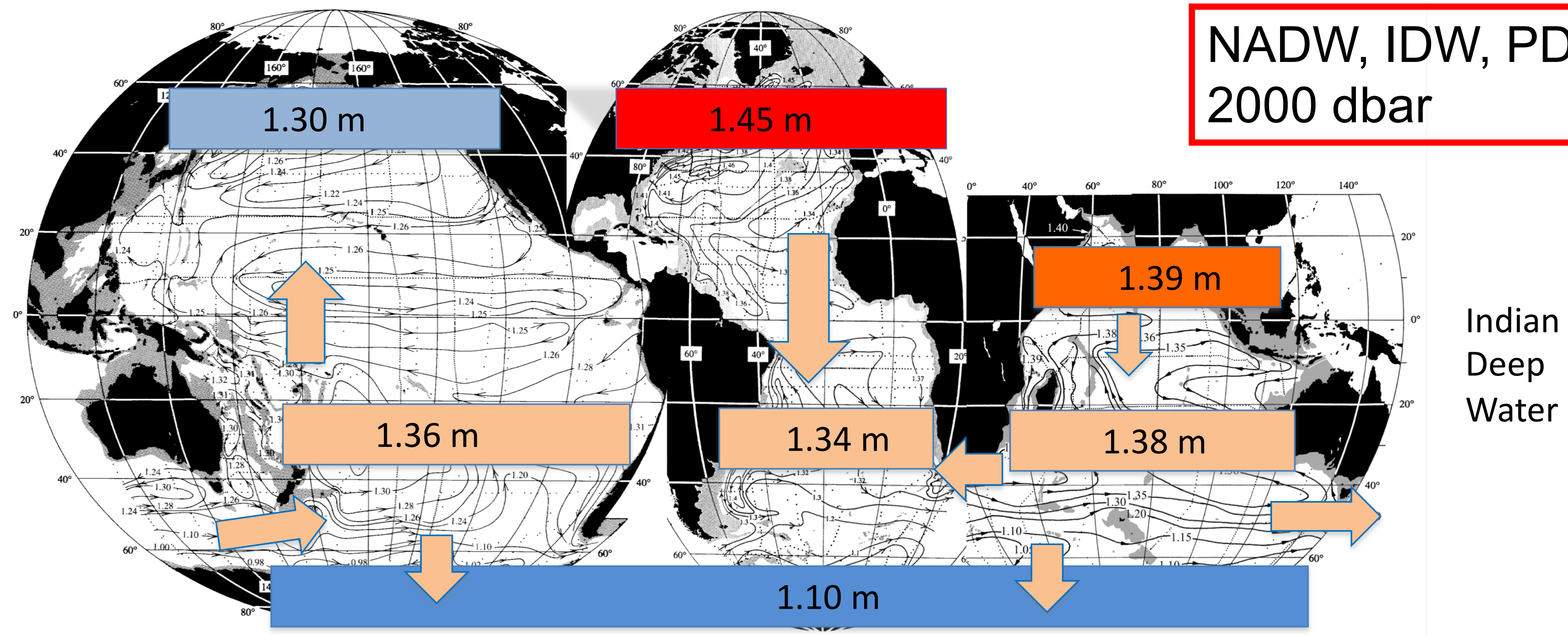
Salinity difference is due to atmospheric freshwater transport from Atlantic to Pacific (Zaucker and Broecker, 1992)

Adjusted steric height distributions from Reid (1994, 1997, 2003)



Large-scale steric height distributions and the GOC

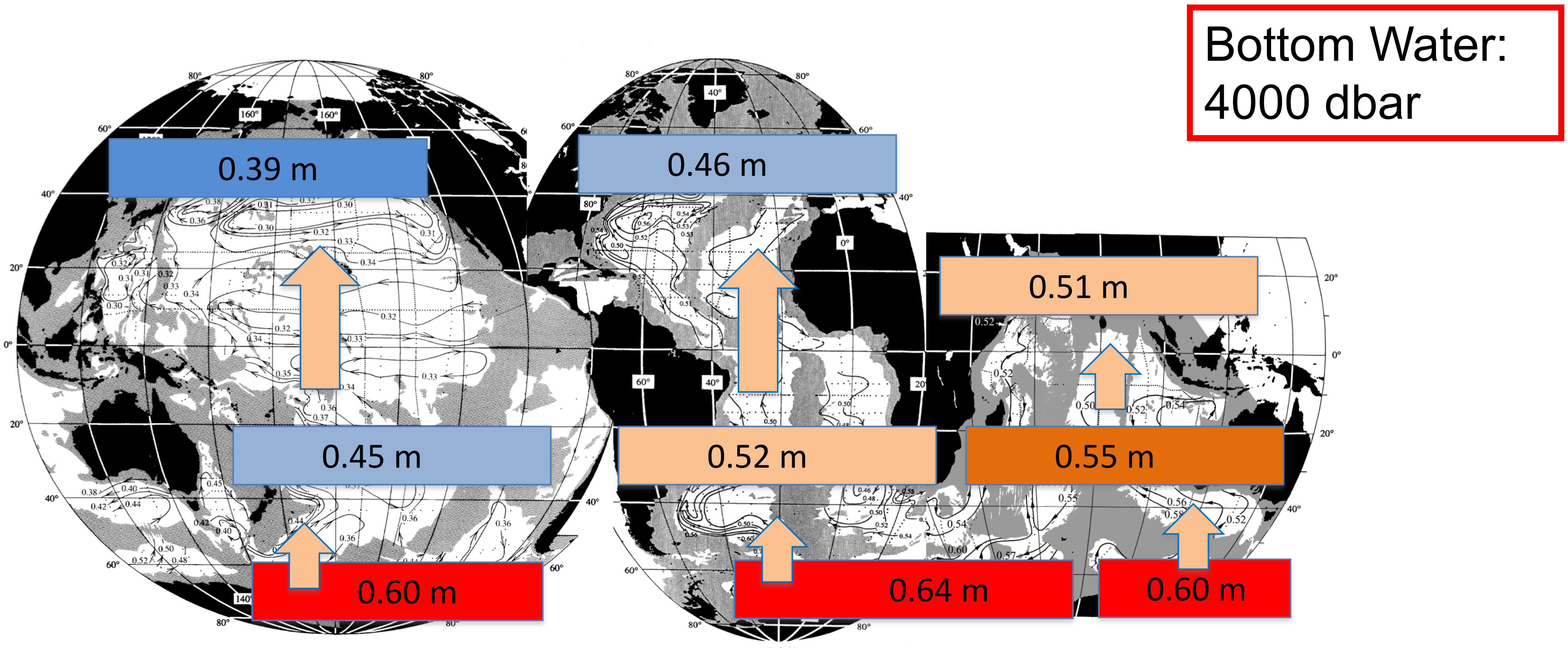
NADW, IDW, PDW:
2000 dbar



- Global adjusted steric height at 2000 dbar for the absolute geostrophic flow (Reid, 1994, 1997, 2003)
- Down-gradient from North Atlantic to S. Atlantic/S. Pacific to N. Pacific and to Antarctic: NADW formation and outflow. (Also net outflow of IDW to the south at this level)
-

W&S91 parameters (very rough):
Pacific $\varepsilon = 0.5$
Atlantic $\varepsilon = 0.4$
Indian $\varepsilon = 1$

Large-scale steric height distributions and the GOC



- Global adjusted steric height at 4000 dbar for the absolute geostrophic flow (Reid, 1994, 1997, 2003)
- Down gradient is from south to north in all basins – AABW filling in from the far south
- Pacific abyssal circulation is isolated from the Atlantic-Indian abyssal circulation
- Relevant theory and discussion: Marotzke et al. (1988), Wright and Stocker (1991), Stocker and Wright (1991). See R.-X. Huang (text, 2010), Vallis (text, 2006).

Summary for downgradient flow portion of talk

Hydrographic sections measured top to bottom and coast to coast provide the mass-balanced ocean transports needed for mapping the global circulation, overturning circulation, and estimated diffusivities required for the circulation.

Inverse models and 'manual' inverse (e.g. Reid) are the primary method for calculating the transports

(Inverse models allow estimate of uncertainty)

Forcing for GOC:

- Wind (Ekman)
- Heat
- Freshwater
- Turbulence (diffusivity)

Global overturning circulation is strongly controlled by ocean salinity distribution