Ocean: Thermohaline circulation II

ATM2106

1. Abyssal circulation

Not many observations, but..

- Localized sources provides water that must be come back to the surface to conserve the volume.

- The current is very slow satisfying geostrophic, hydrostatic and thermal wind balance.

The depth-integrated circulation

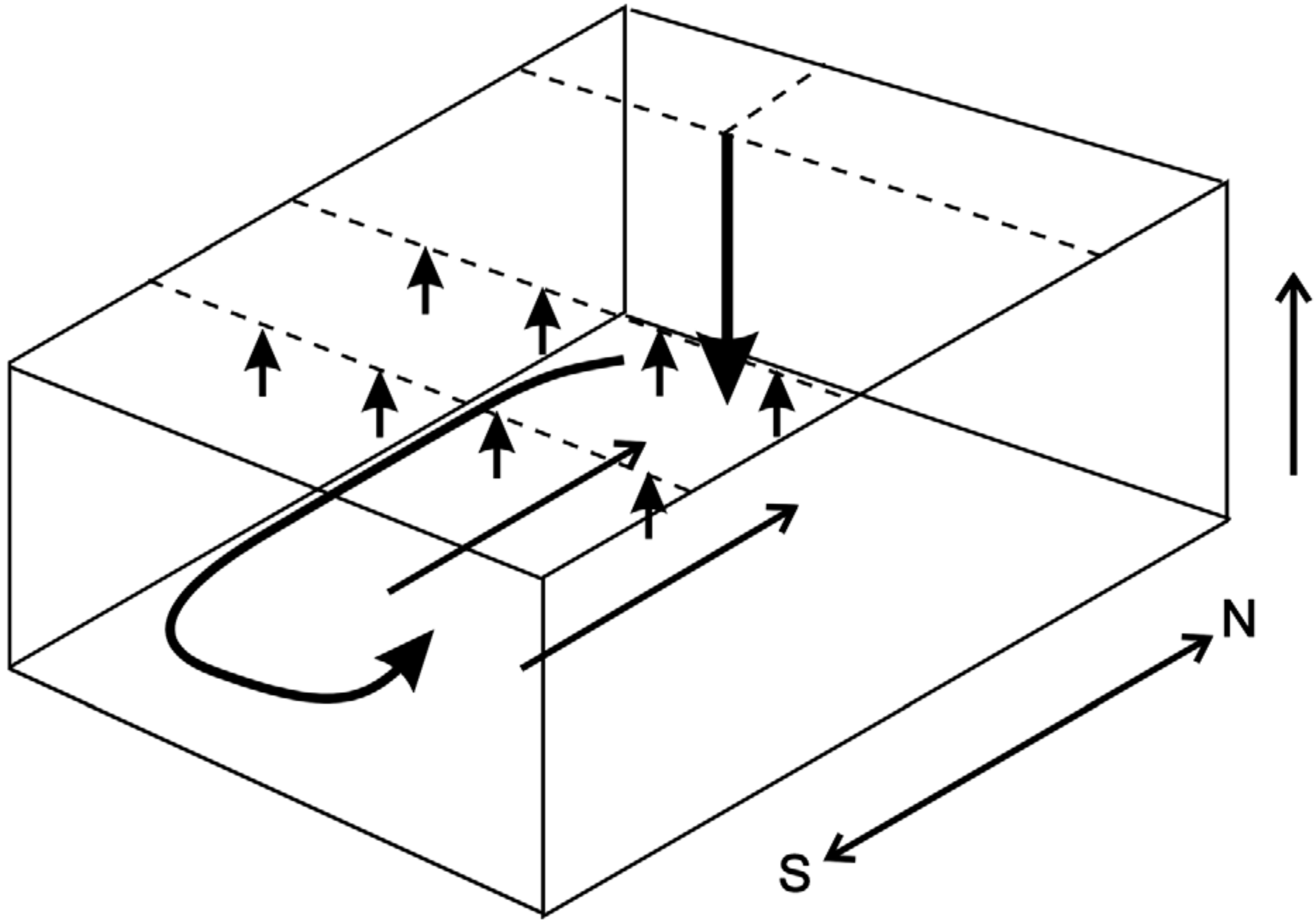
$$\beta v = f\frac{\partial w}{\partial z} + \frac{1}{\rho_{ref}}\frac{\partial}{\partial z}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)$$

$$\int_{-D}^{-d} \beta v dz = \int_{-D}^{-d} f\frac{\partial w}{\partial z} dz$$

$$w(-D) = 0$$

$$w(-d) > 0$$

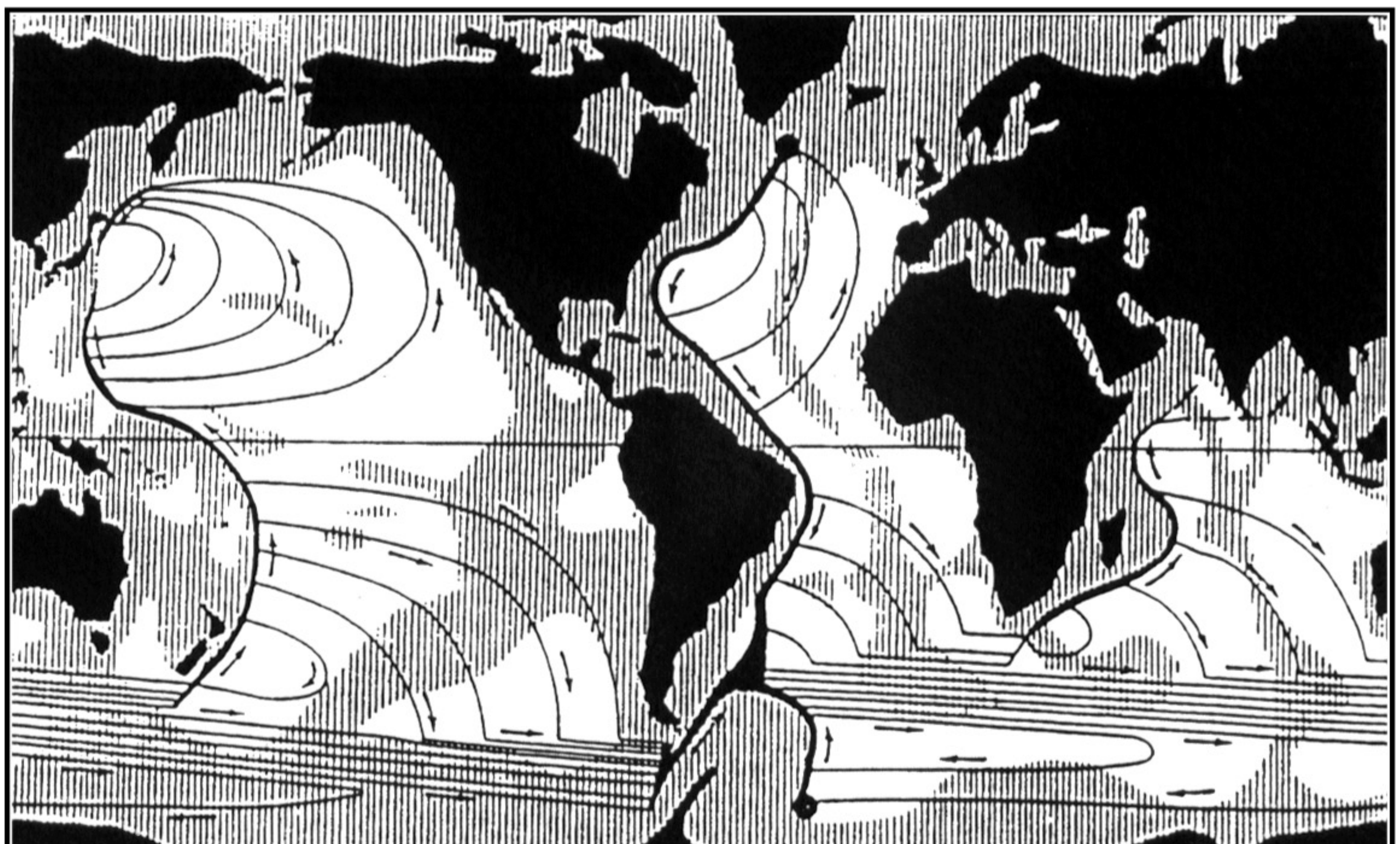
$$\int_{-D}^{-d} v dz = \frac{f}{\beta} w(-d) > 0$$



Water sinks in a localized region in polar latitudes with compensating upwelling distributed over the basin. Associated meridional flow is northward that is balanced by a western boundary current

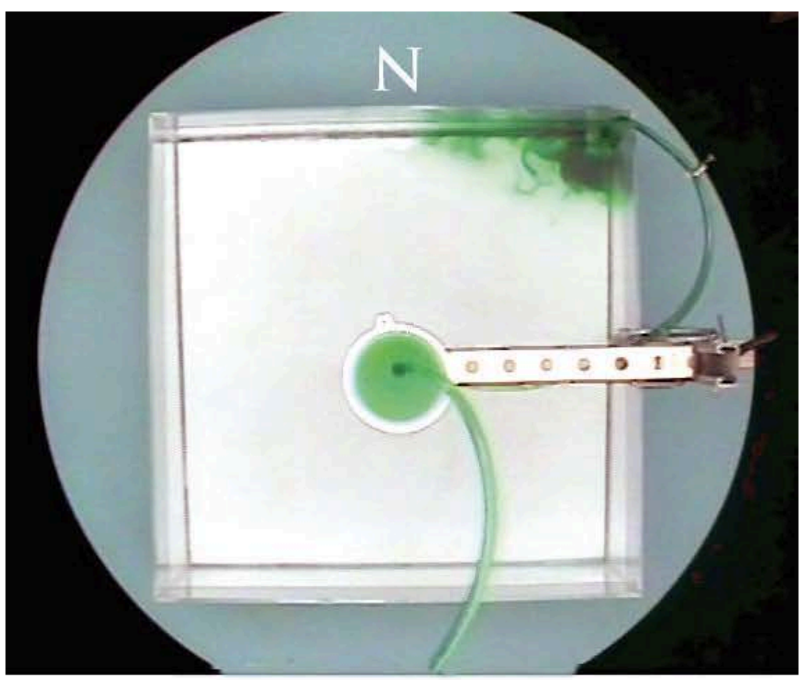
The equatorward western boundary current

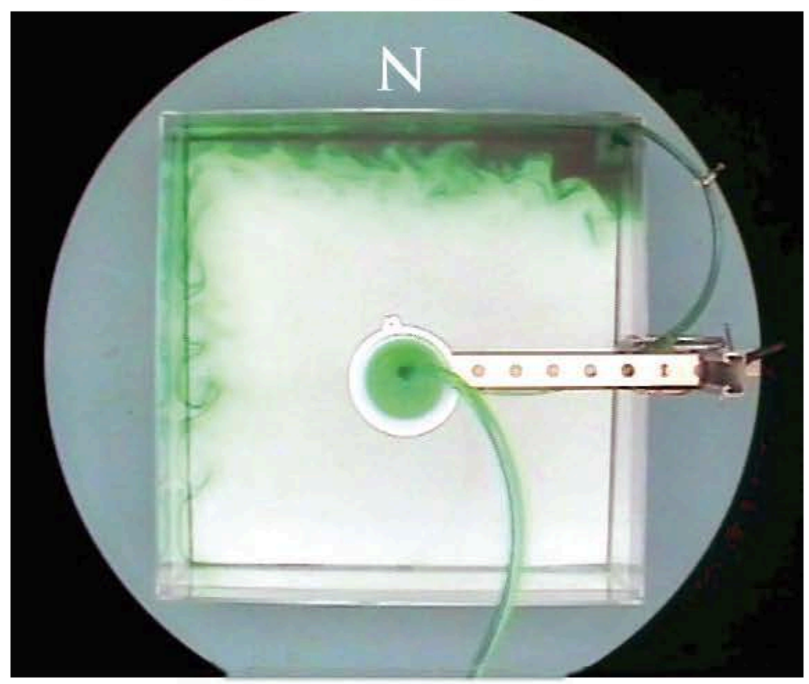
- The equatorward western boundary current beneath the poleward western boundary current!

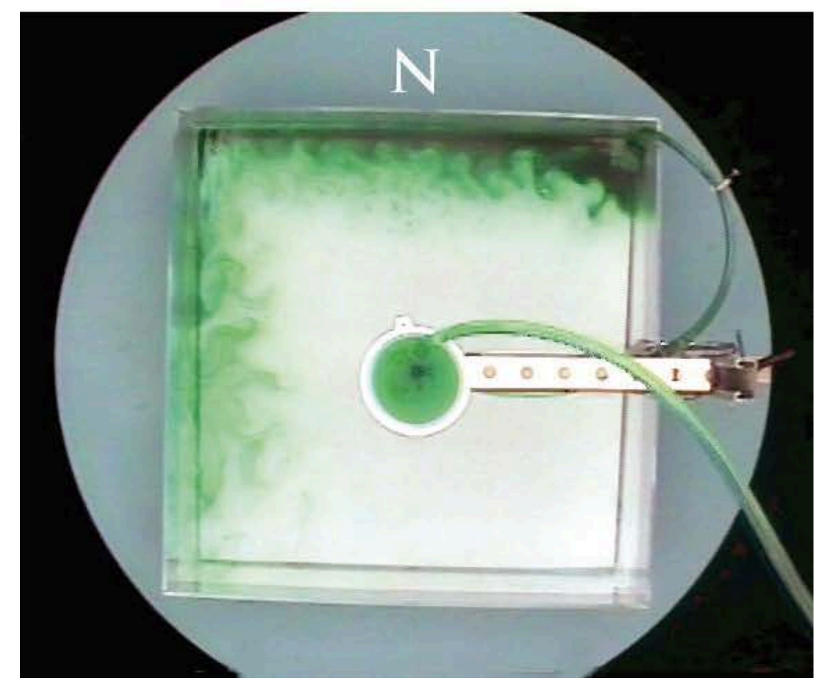


The equatorward western boundary current

- The equatorward western boundary current beneath the poleward western boundary current!







Why does it have to be on the west side?

$$\beta v = f\frac{\partial w}{\partial z} + \frac{1}{\rho_{ref}}\frac{\partial}{\partial z}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)$$

- This equation is valid when Coriolis force, pressure gradient force and wind stress are in balance.
- Near the boundary, the frictional force becomes important.
- Let’s follow the same procedure as we did for the equation above.

Why does it have to be on the west side?

$$-fv + \frac{1}{\rho_{ref}} \frac{\partial p}{\partial x} = -\epsilon u$$

$$fu + \frac{1}{\rho_{ref}} \frac{\partial p}{\partial y} = -\epsilon v$$

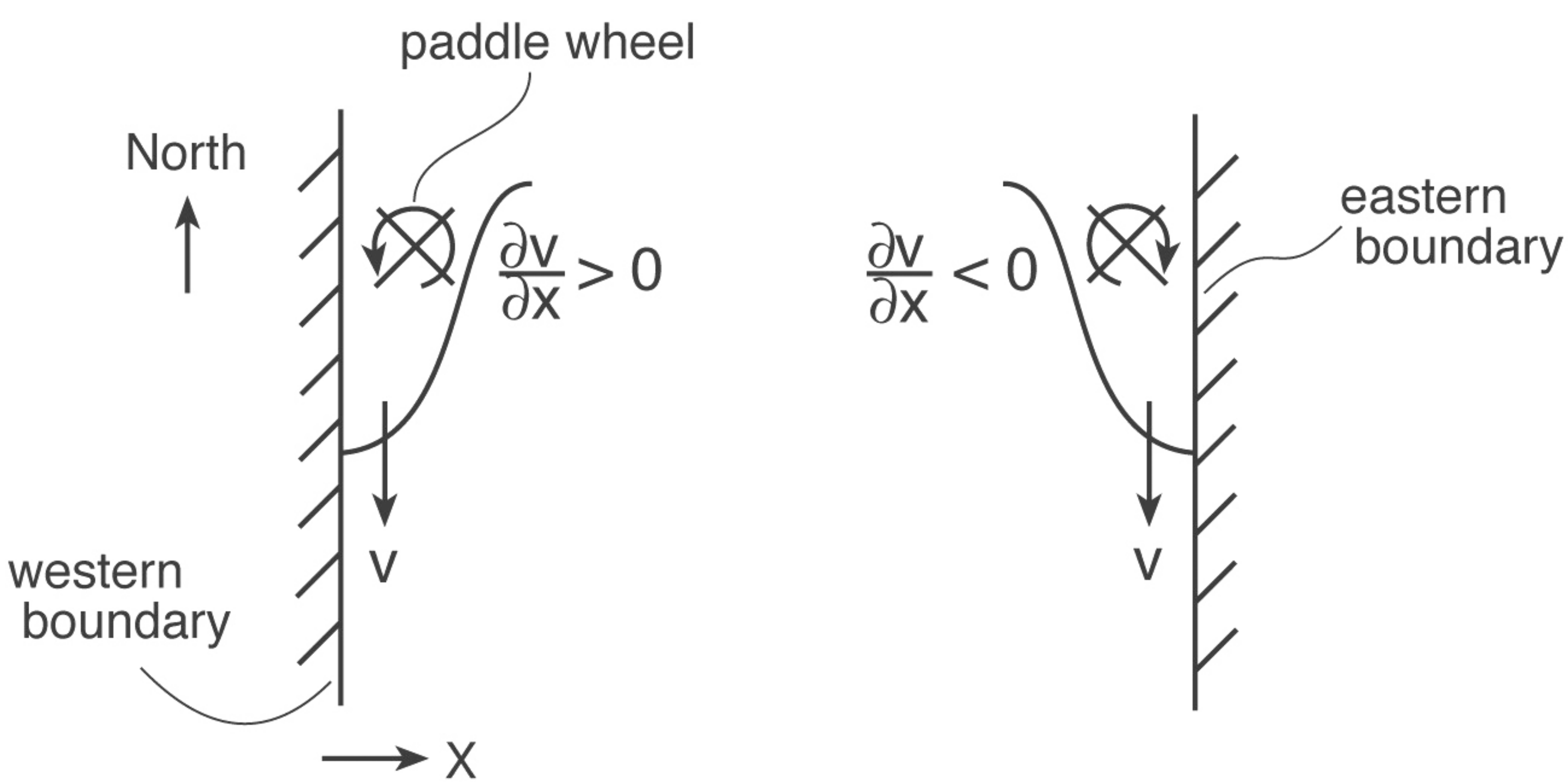
$$f \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = -\epsilon \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right)$$

$$f \left(\frac{\partial w}{\partial z} \right) = \epsilon \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right)$$

Why does it have to be on the west side?

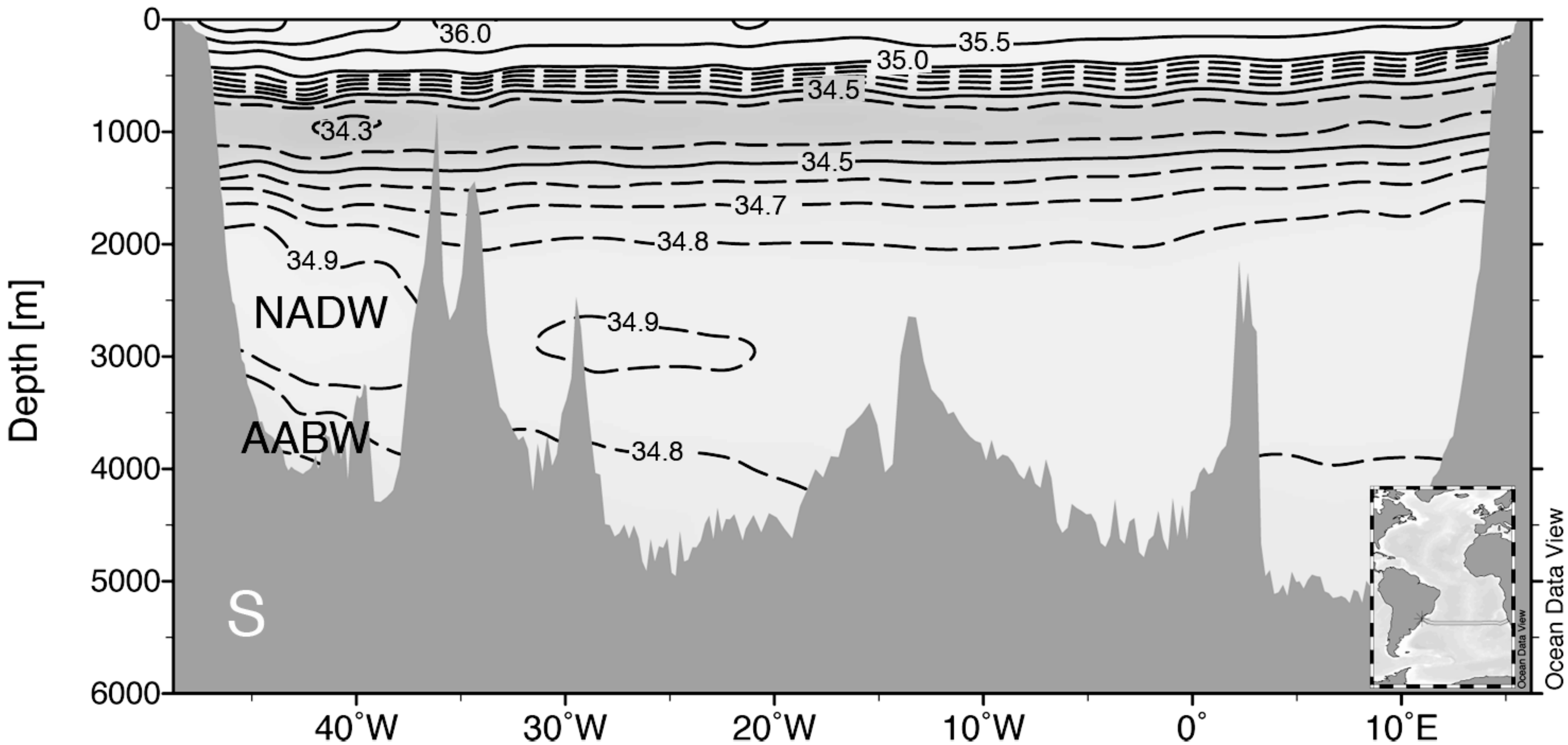
$$f\left(\frac{\partial w}{\partial z}\right) = \epsilon\left(\boxed{\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}}\right)$$

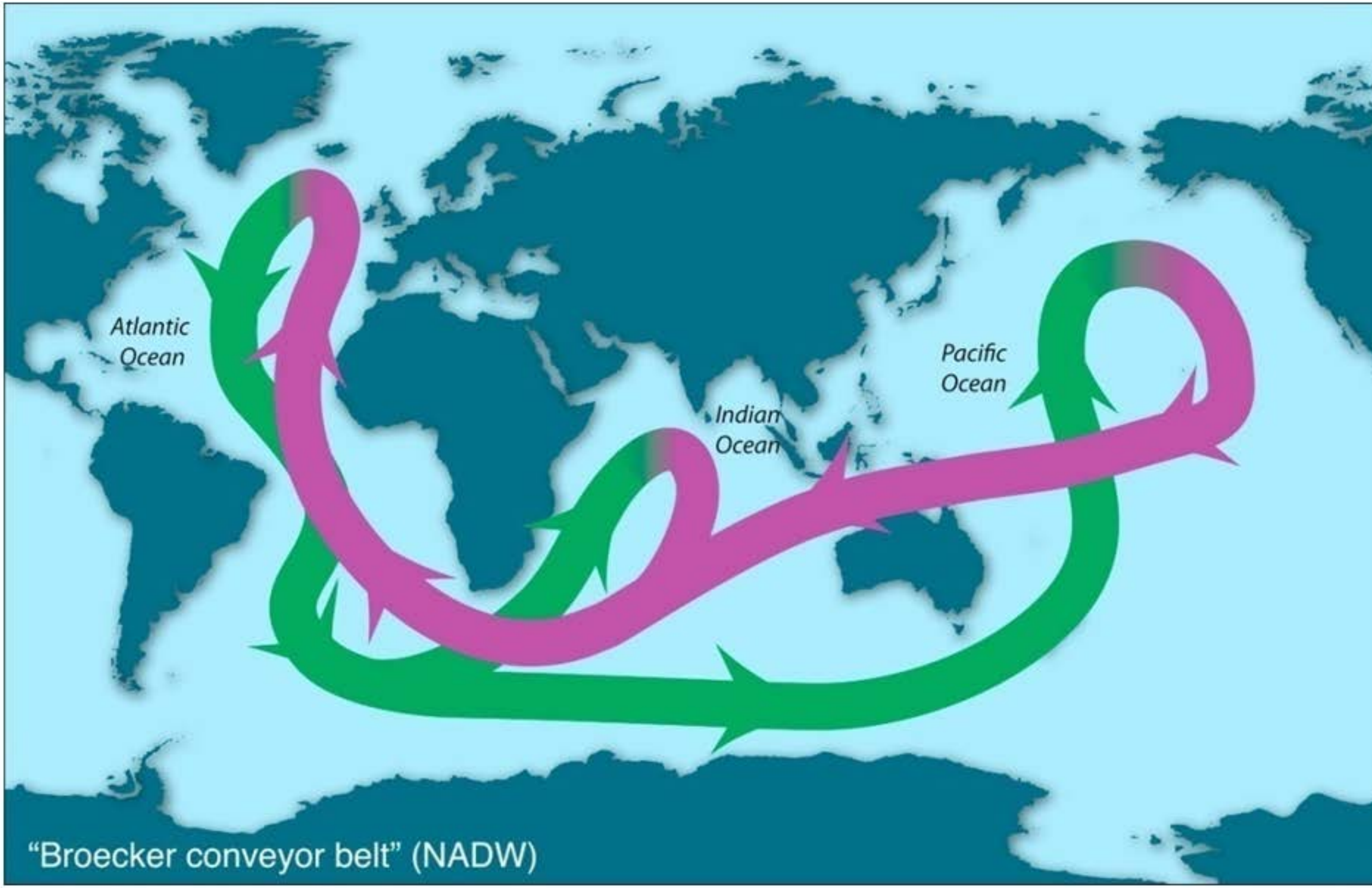
Vorticity



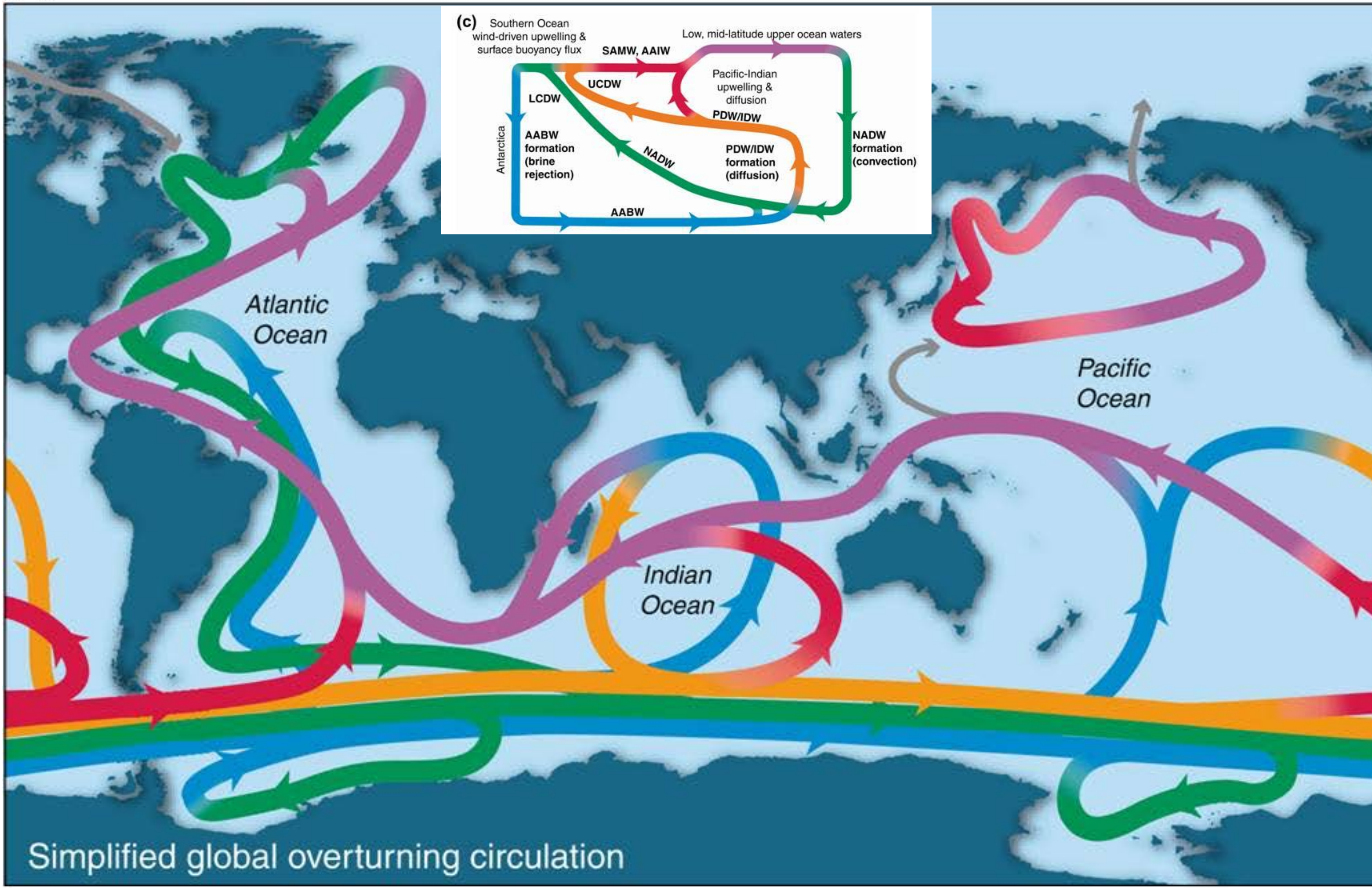
In order to satisfy *w(-d)>0*, the boundary current must exist on the west side of the basin.

Deep western boundary current in the observations

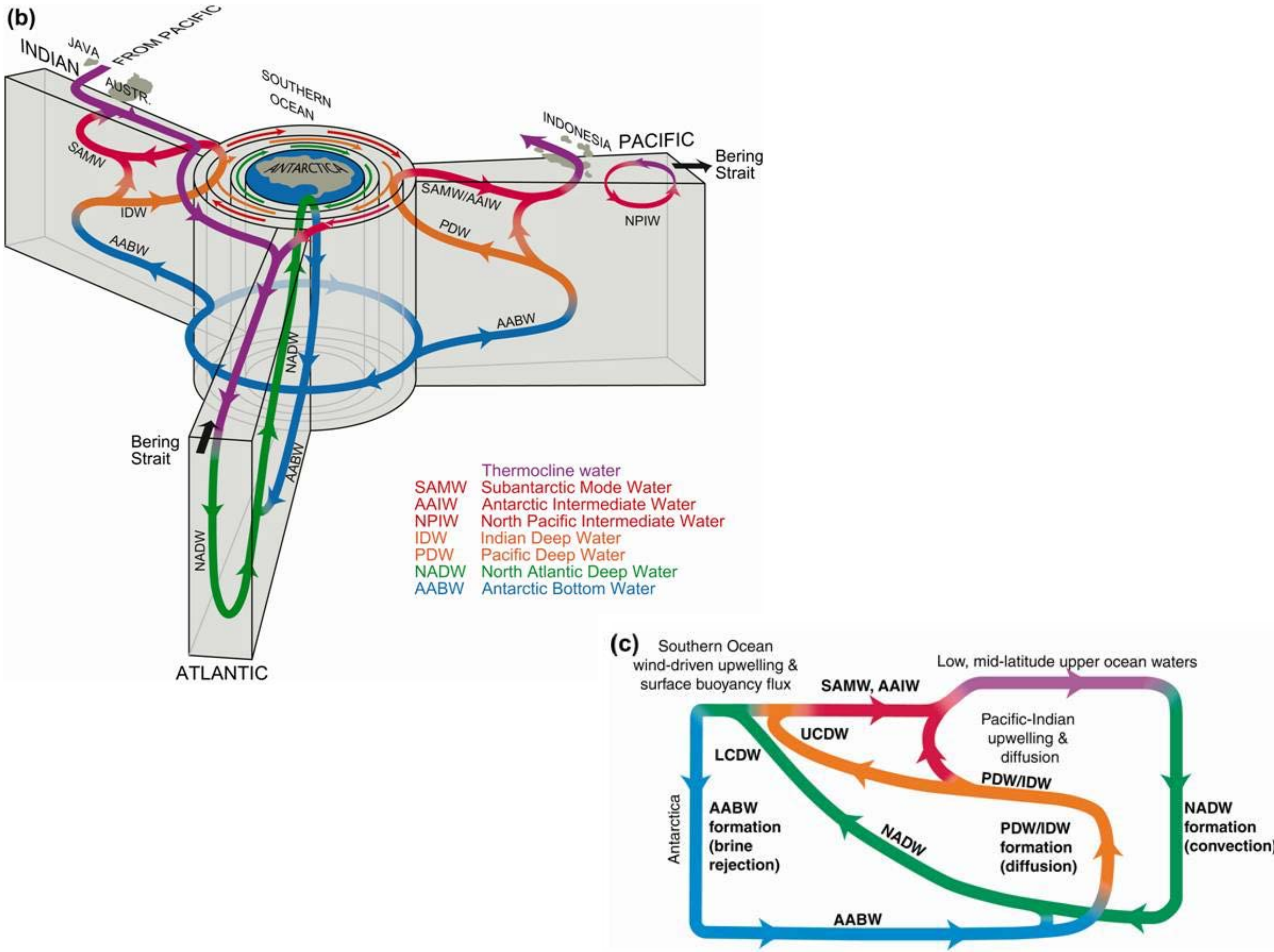




Simplified global NADW cell, which retains sinking only somewhere adjacent to the northern North Atlantic and upwelling only in the Indian and Pacific Oceans. See text for usefulness of, and also issues with, this popularization of the global circulation, which does not include any Southern Ocean processes. Source: After Broecker (1987).



Global overturning circulation schematics. (a) The NADW and AABW global cells and the NPIW cell. See Figure S14.1 on the textbook Web site for a complete set of diagrams. This figure can also be found in the color insert. Source: From Talley (2011).


(b)
INDIAN
JAVA
FROM PACIFIC
AUSTR.
SOUTHERN OCEAN
ANTARCTICA
INDONESIA
PACIFIC
Bering Strait
SAMW
IDW
AABW
SAMW/AAIW
PDW
NPIW
AABW
NADW
Bering Strait
AABW
NADW
ATLANTIC
Thermocline water
SAMW Subantarctic Mode Water
AAIW Antarctic Intermediate Water
NPIW North Pacific Intermediate Water
IDW Indian Deep Water
PDW Pacific Deep Water
NADW North Atlantic Deep Water
AABW Antarctic Bottom Water
(c)
Southern Ocean wind-driven upwelling & surface buoyancy flux
Low, mid-latitude upper ocean waters
SAMW, AAIW
UCDW
LCDW
Pacific-Indian upwelling & diffusion
PDW/IDW
Antarctica
AABW formation (brine rejection)
NADW
PDW/IDW formation (diffusion)
NADW formation (convection)
AABW

Ocean: Heat budget and transport

ATM2106

The heat budget for a column of ocean

$$\frac{\partial \mathrm{H}}{\partial t} = -Q_{net} - \left(\frac{\partial H_x}{\partial x} + \frac{\partial H_y}{\partial y} \right)$$

$$\mathrm{H} = \rho_{ref} c_w \int_{bottom}^{top} T dz$$

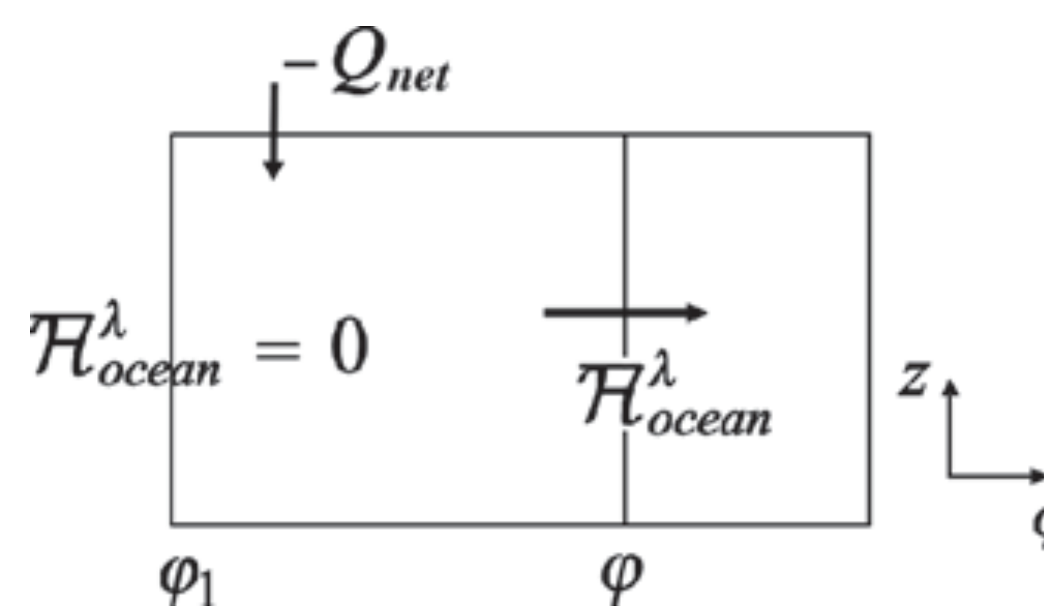
$$H_x = \rho_{ref} c_w \int_{bottom}^{top} uT dz$$

$$H_y = \rho_{ref} c_w \int_{bottom}^{top} vT dz$$

Changes in heat stored in a column of the ocean
= surface heat flux + horizontal heat flux by ocean currents

How to measure ocean heat transport?

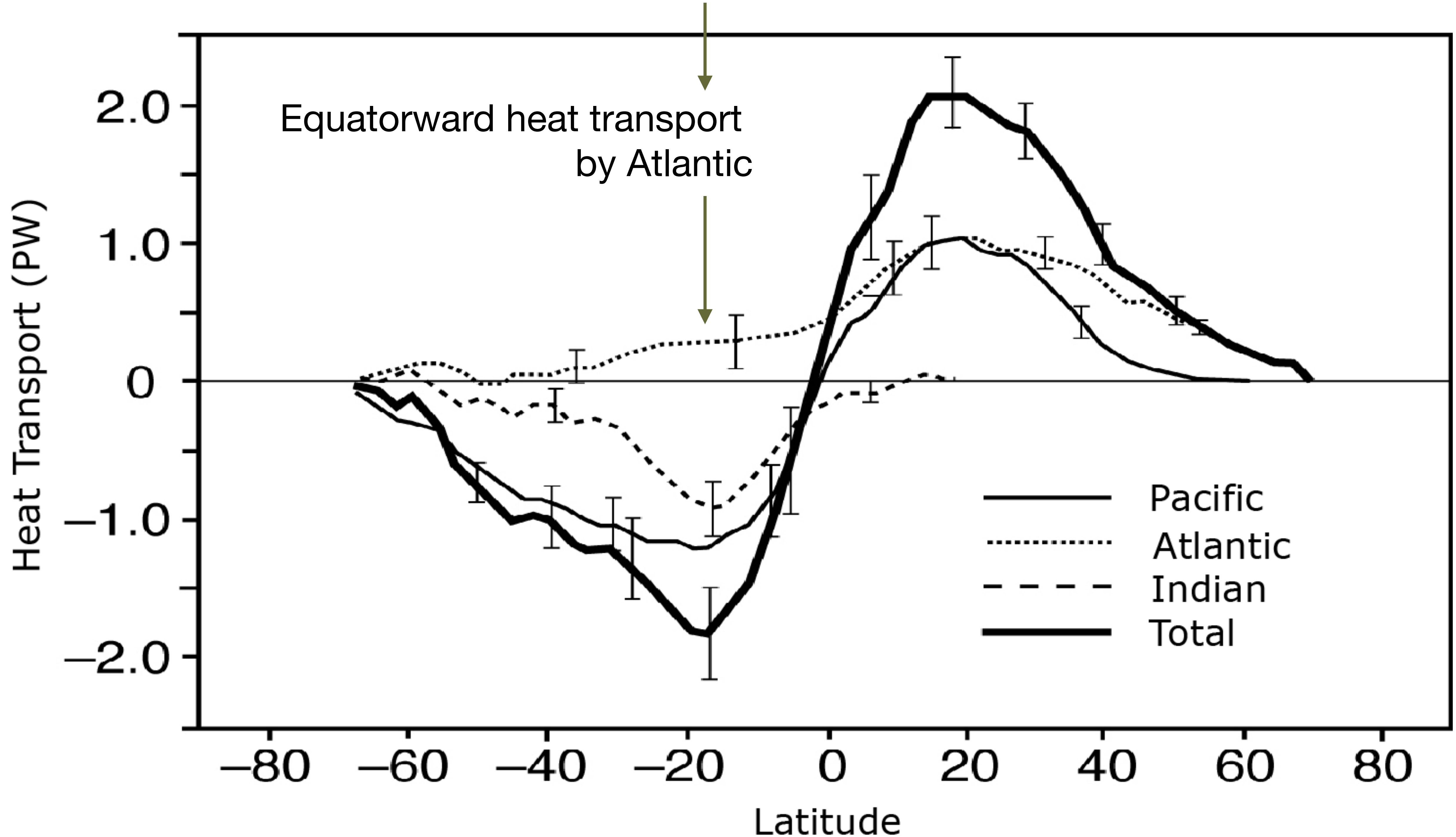
- By subtracting atmospheric heat transport from the total heat transport measured at the top of the atmosphere.
- By finding the heat transport that balances the surface heat flux under the assumption of steady state.
- By directly measuring the heat transport at a few locations.



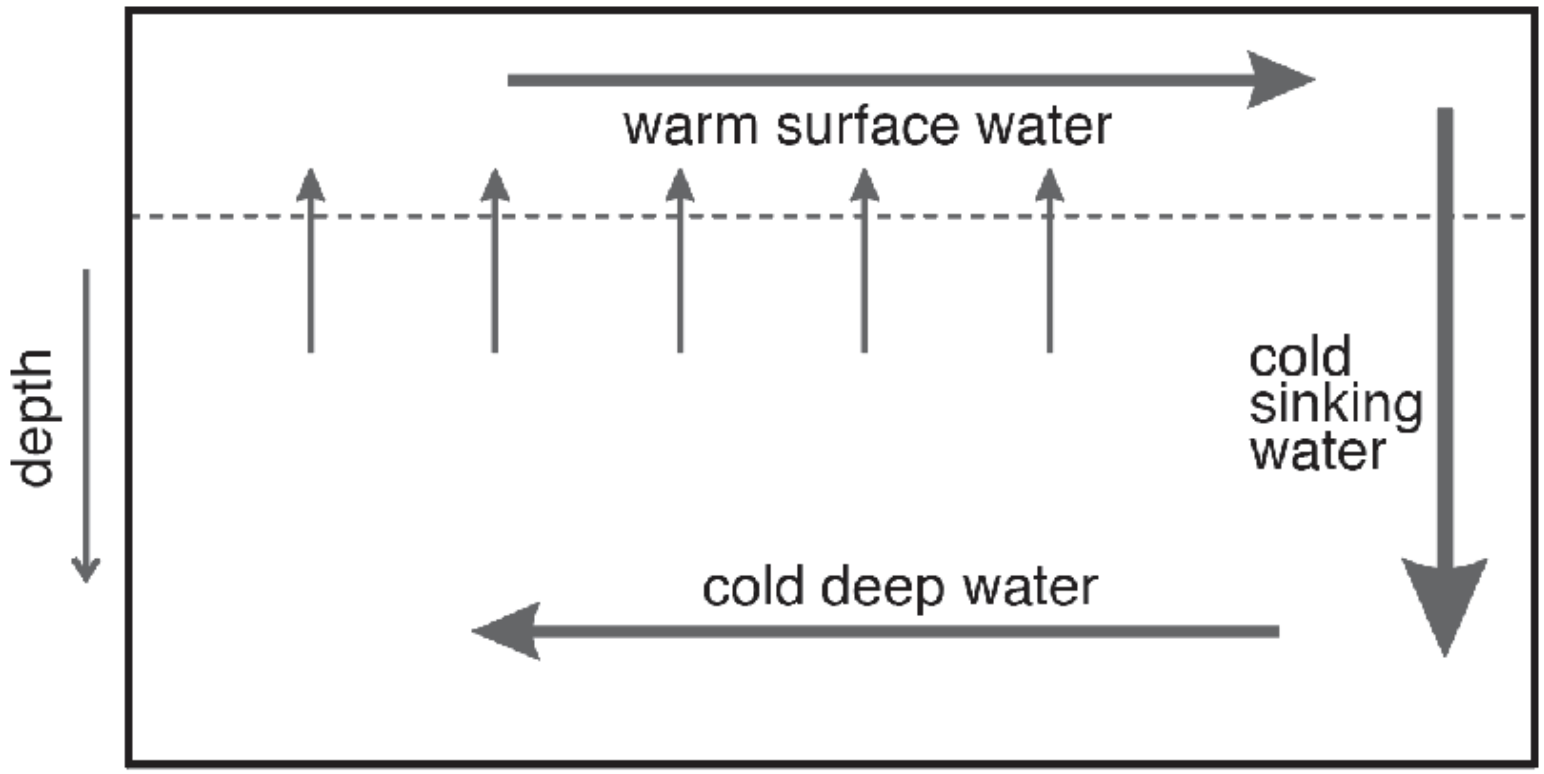
Northward heat transport in the world ocean

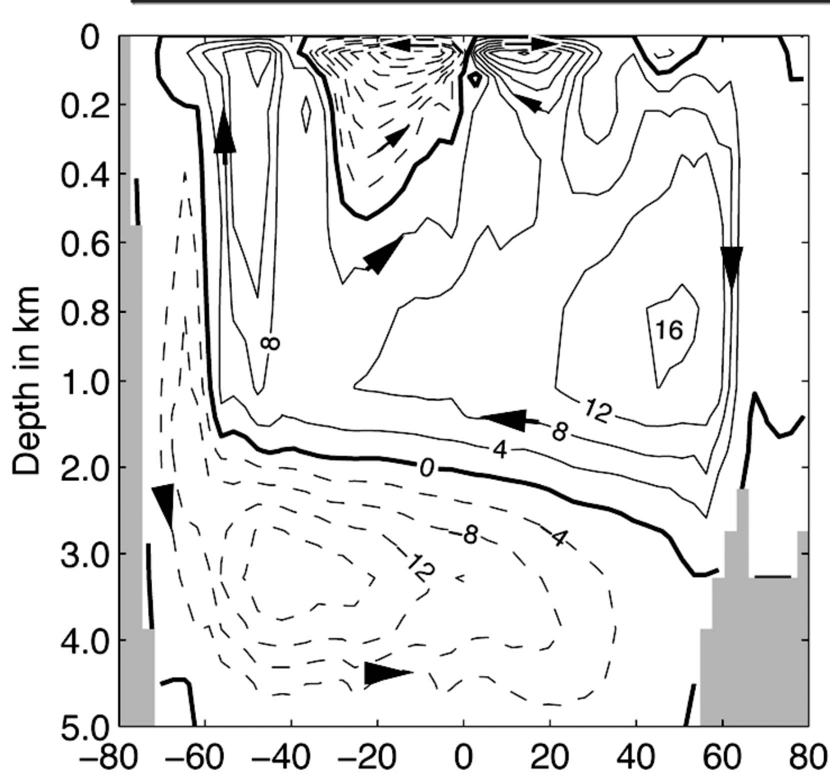
Atlantic Meridional Overturning Circulation

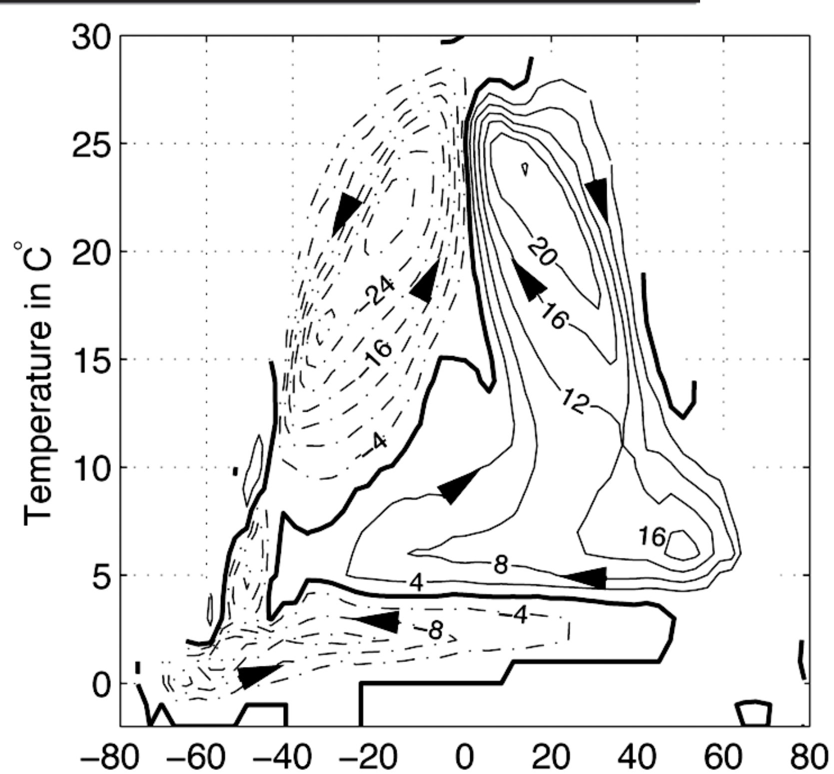
Peaks at 20°

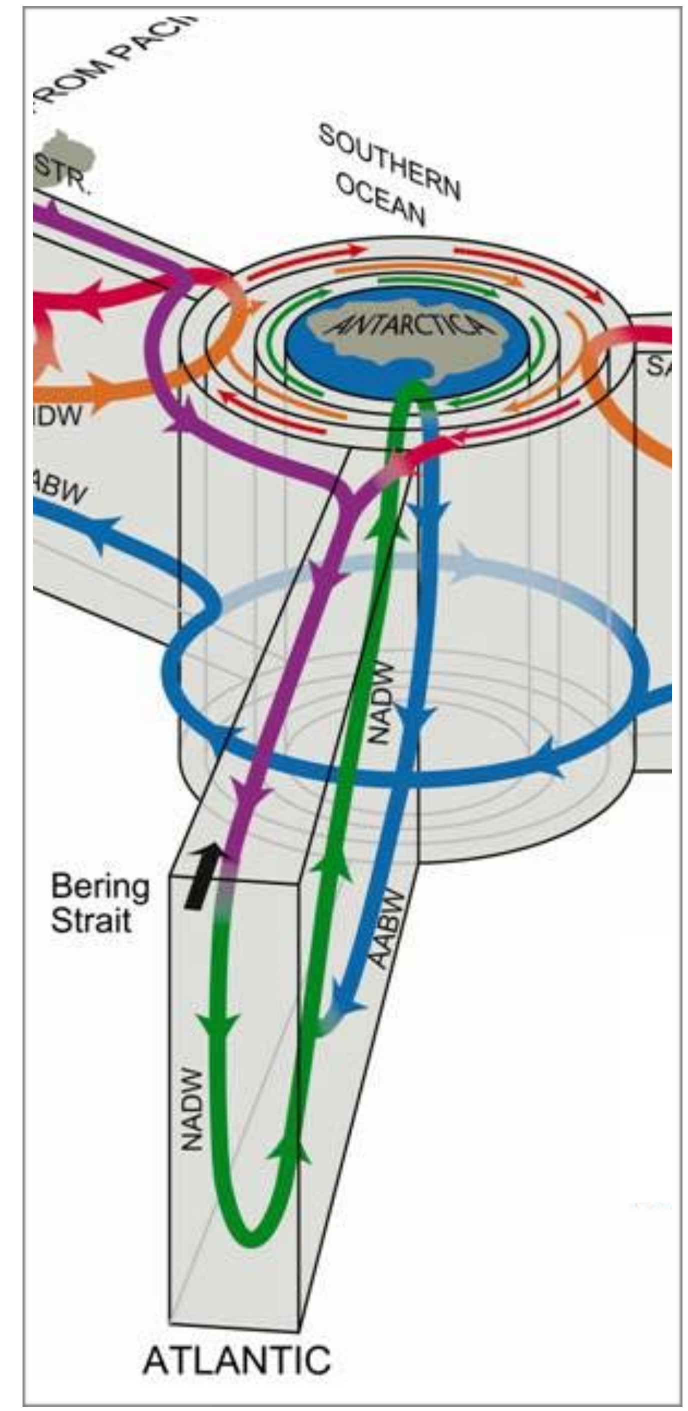


The ocean’s meridional circulation

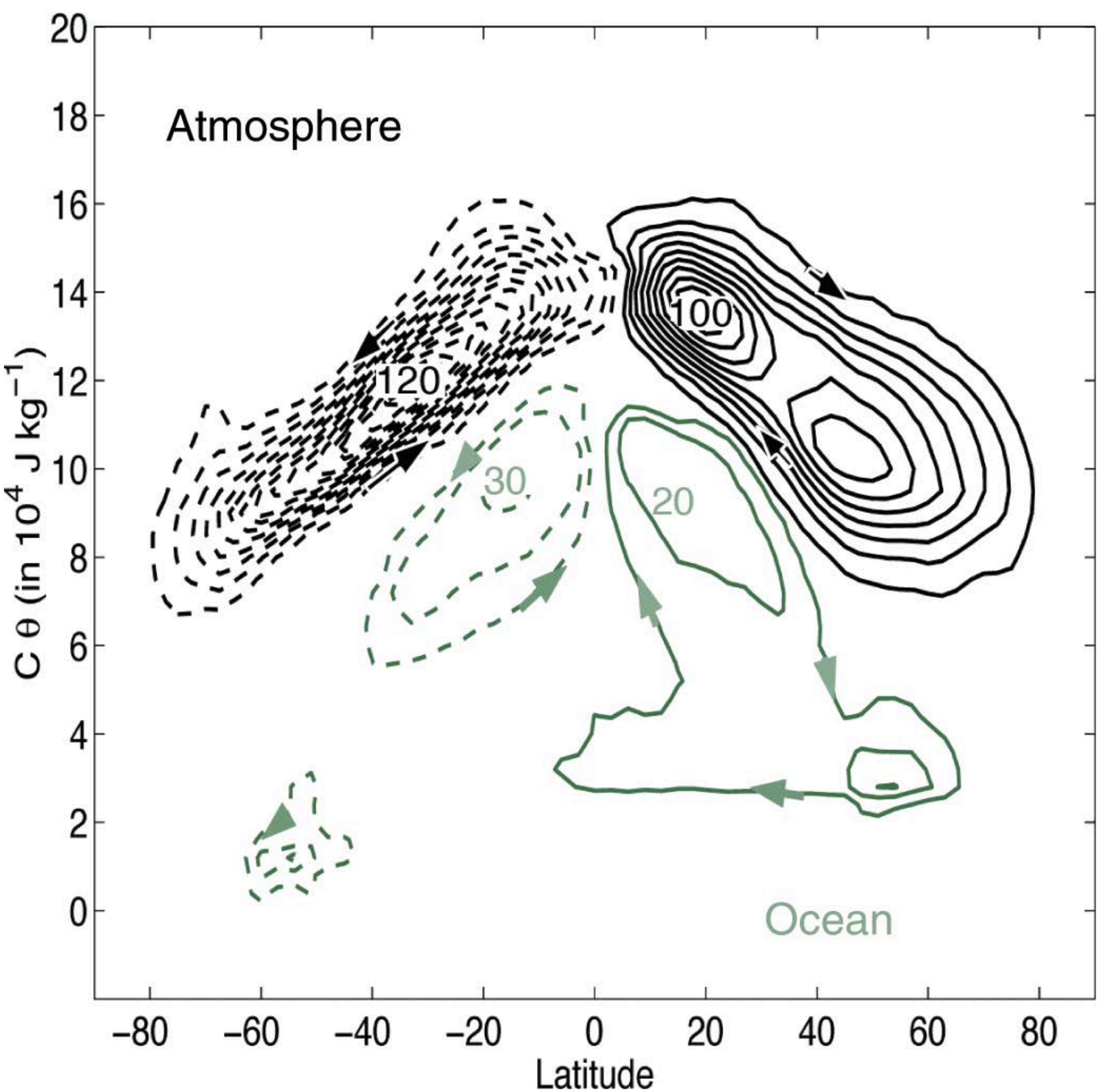








Annual mean mass streamfunction



- Atmospheric overturning circulation is much stronger than the one of then ocean.
- Similar “thickness” of the overturning circulation

higher heat capacity of the ocean v.s. greater temperature differences in the atmosphere