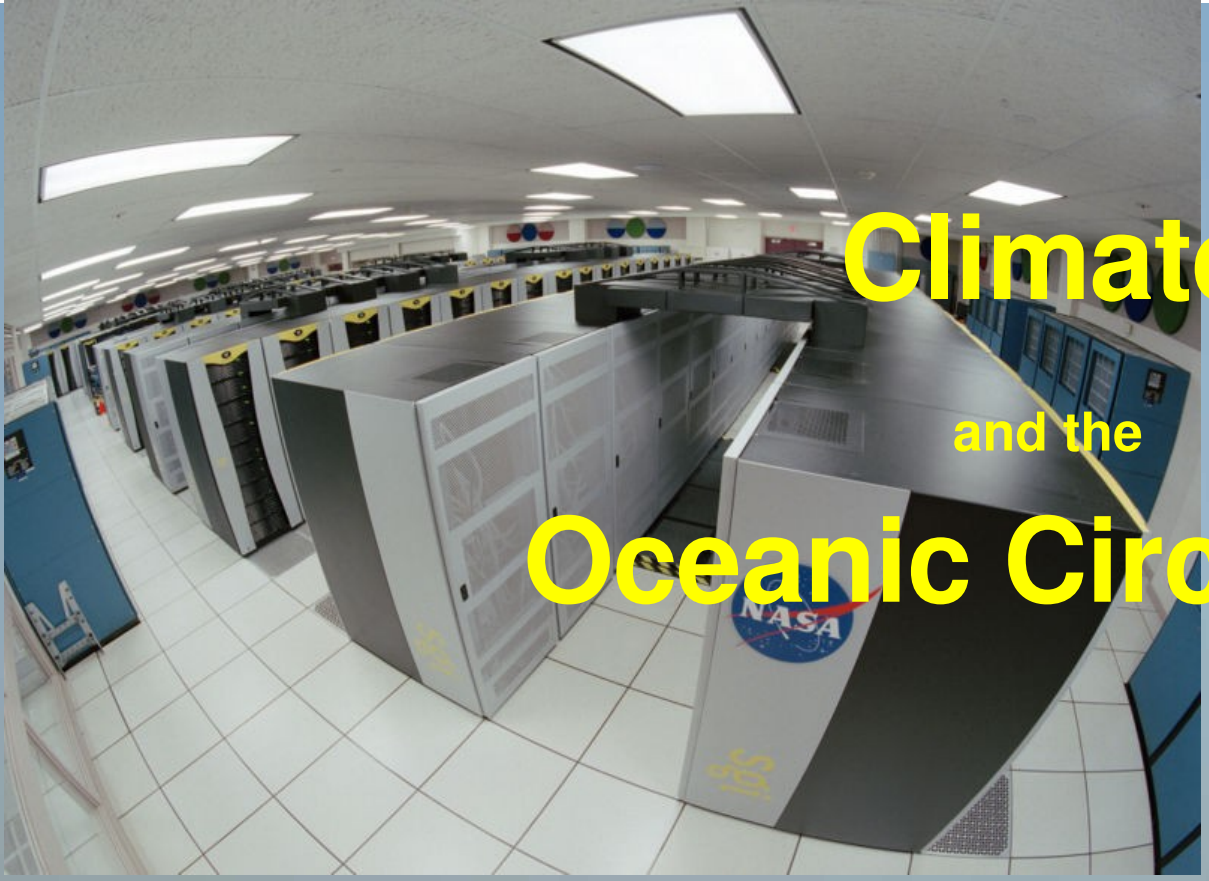


Climate

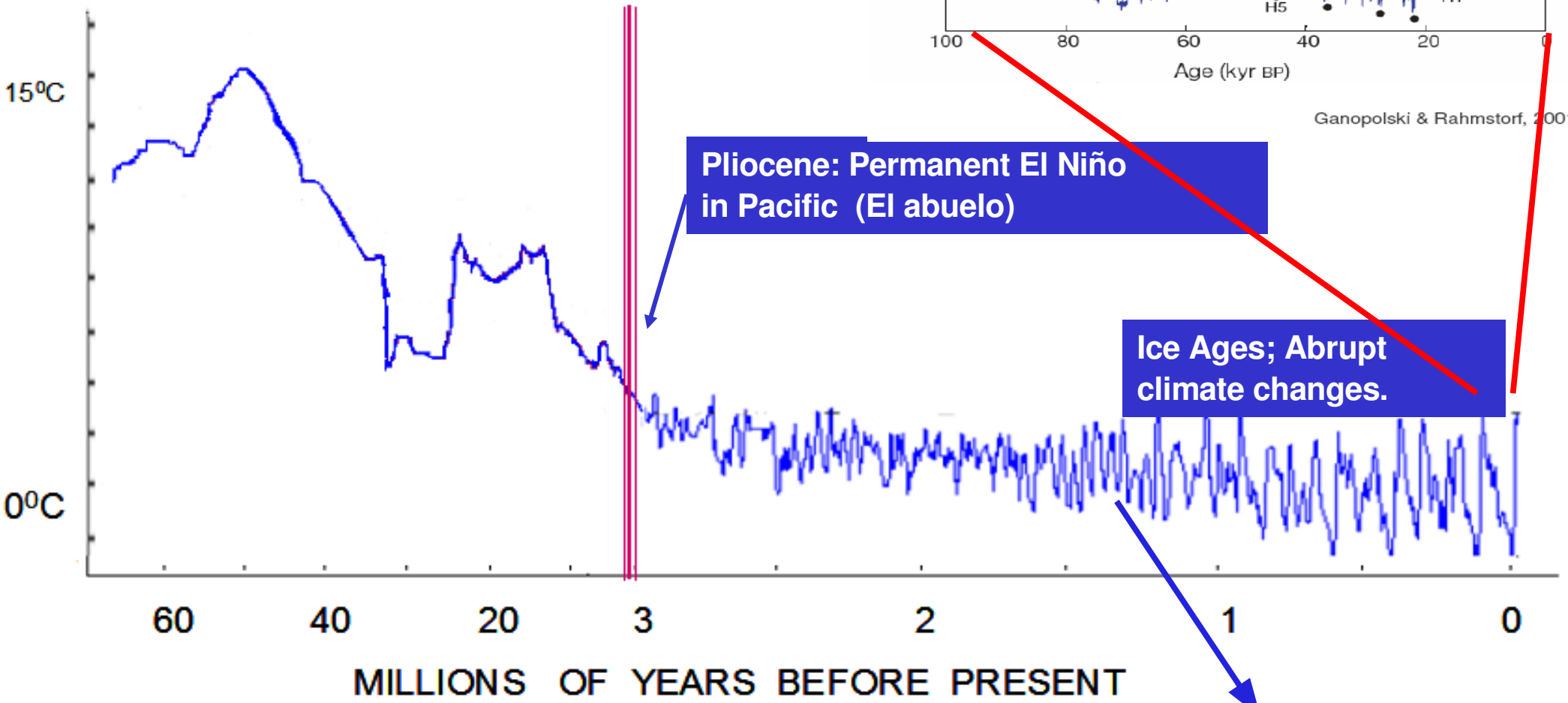
and the

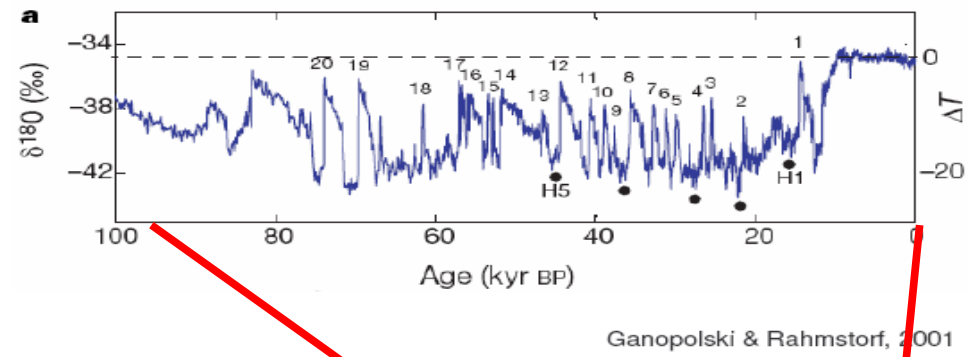
Oceanic Circulation

Marcelo Barreiro
Universidad de la República
Uruguay

GLOBAL COOLING OVER THE PAST 60 MILLION YEARS.

The response to MILANKOVITCH forcing AMPLIFIED ENORMOUSLY over the PAST 1 Ma





Ganopolski & Rahmstorf, 2001

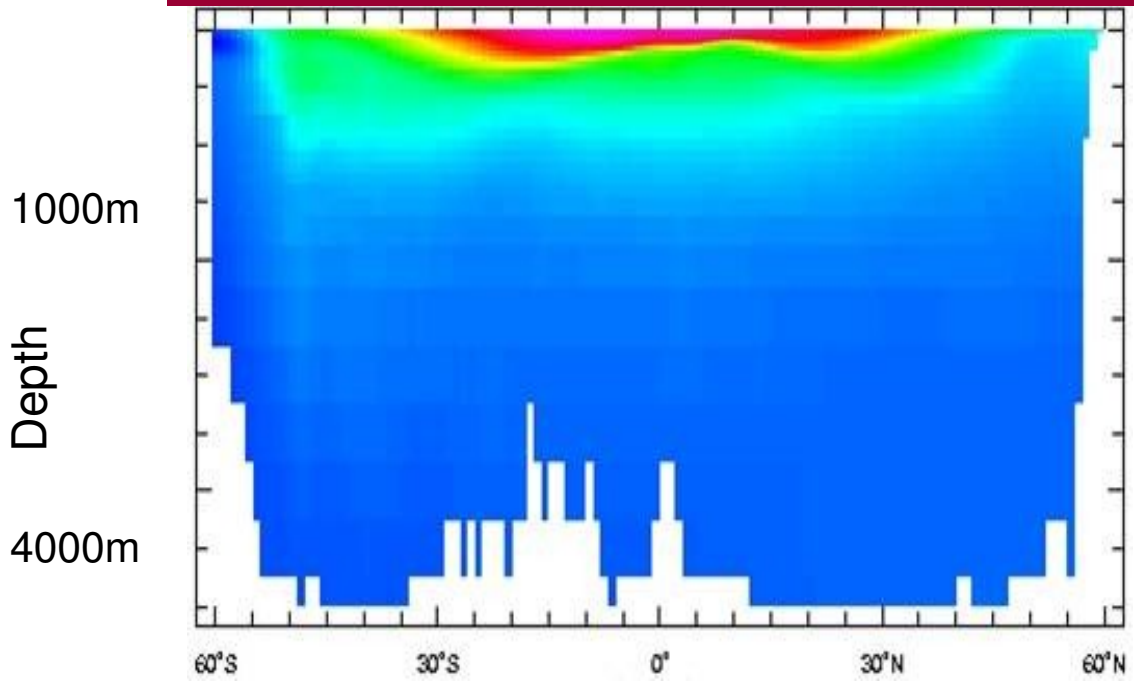
Pliocene: Permanent El Niño in Pacific (El abuelo)

Ice Ages; Abrupt climate changes.

Equatorial Pacific varies with obliquity, but following high latitude insolation, not local.

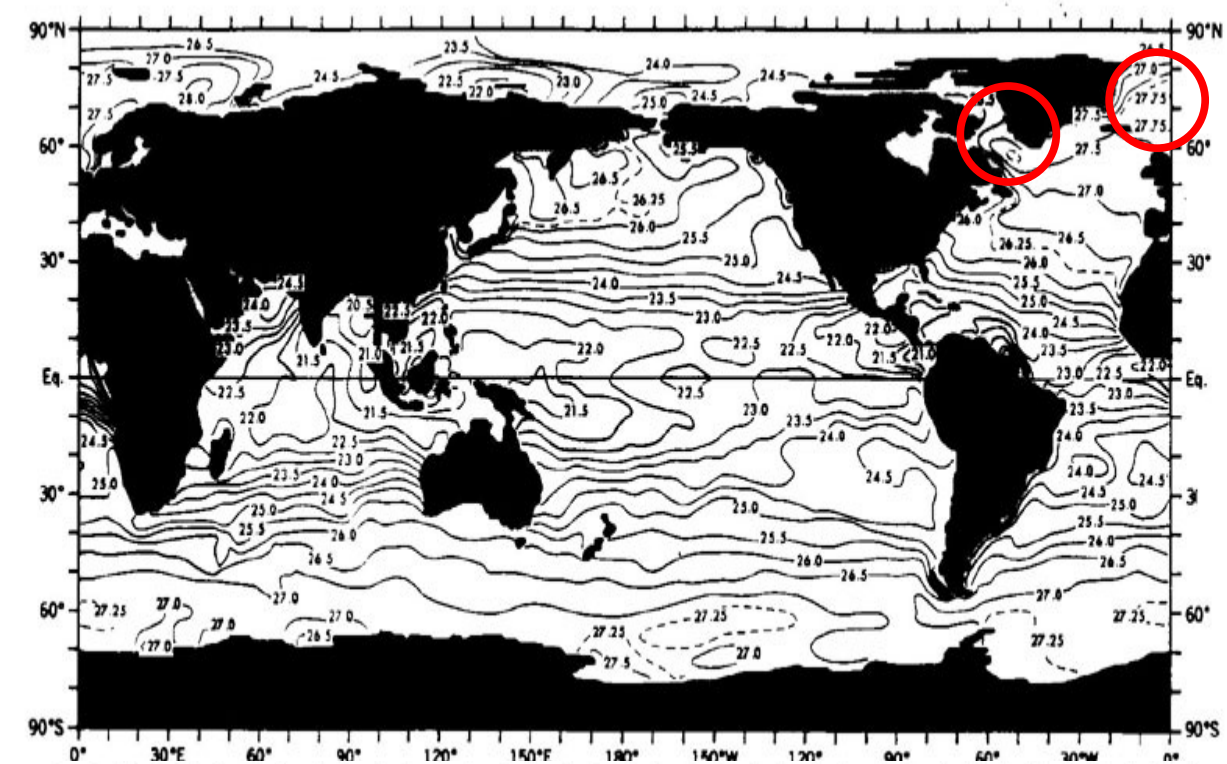
NOTE: Time Scale changes at 3 Million Years

WHY IS THE OCEAN SO COLD?



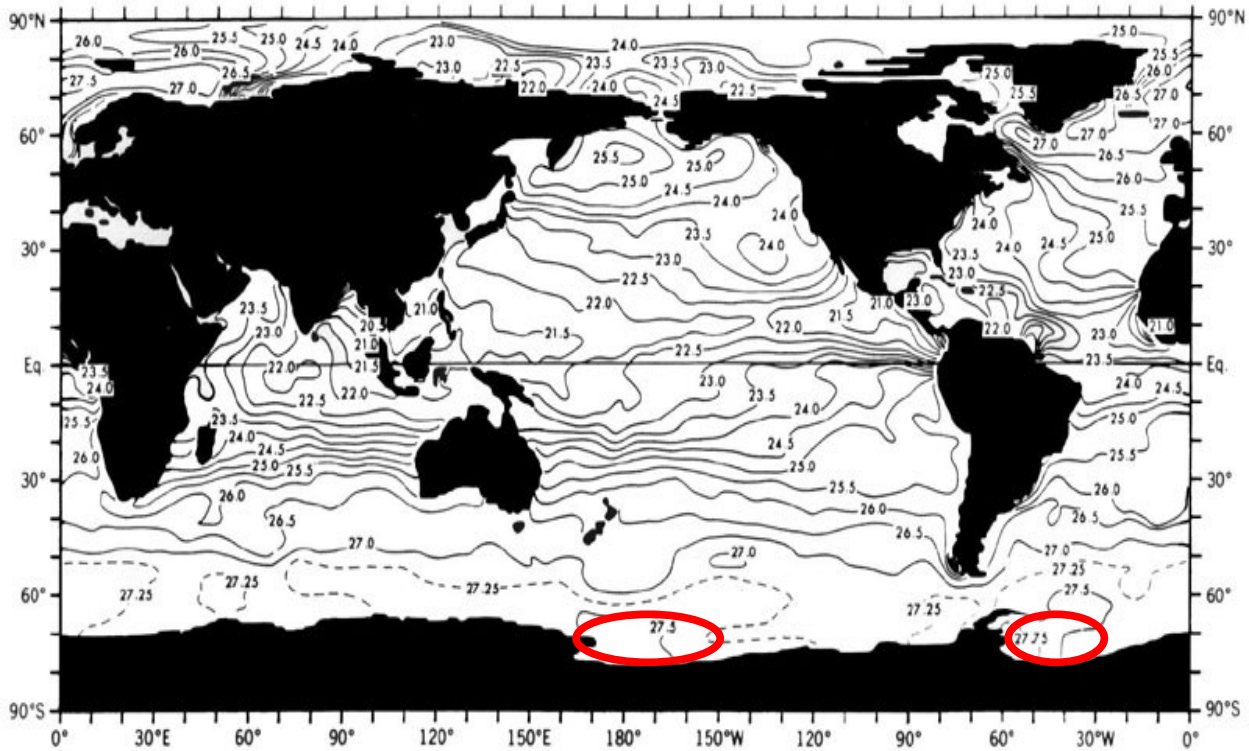
Temperature along a section in the mid-Pacific (152W)

The thermohaline circulation



NH winter distribution of surface density.

Labrador and Greenland Seas sites of deep water production



SH winter distribution of surface density.

Weddell and Ross Seas

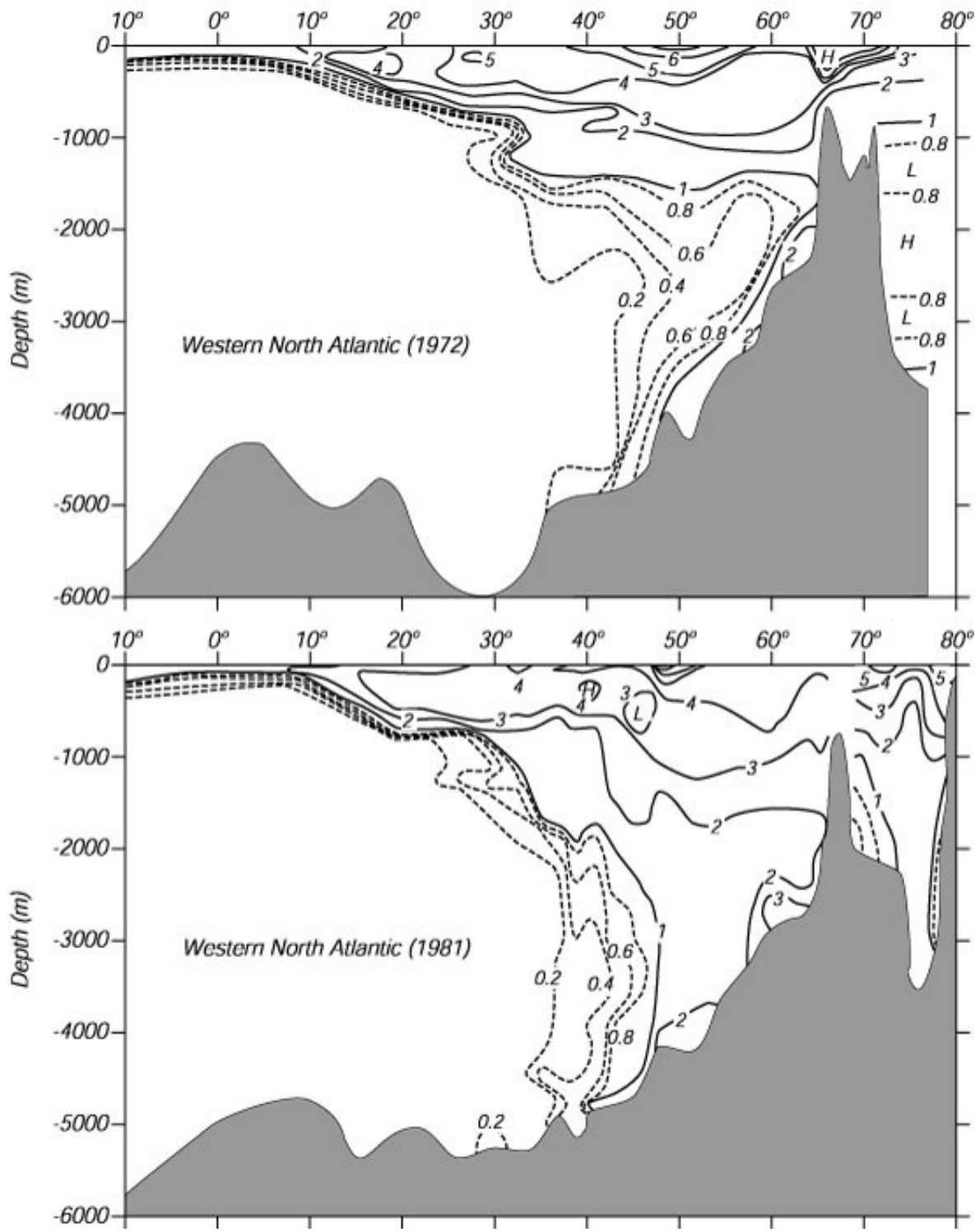
Evidence of a deep circulation:

Distribution of tritium in the North Atlantic.

The tritium entered the ocean thanks to atomic bomb testing from 1950 to 1970.

Currents are very slow here, about 1.6 mm/s.

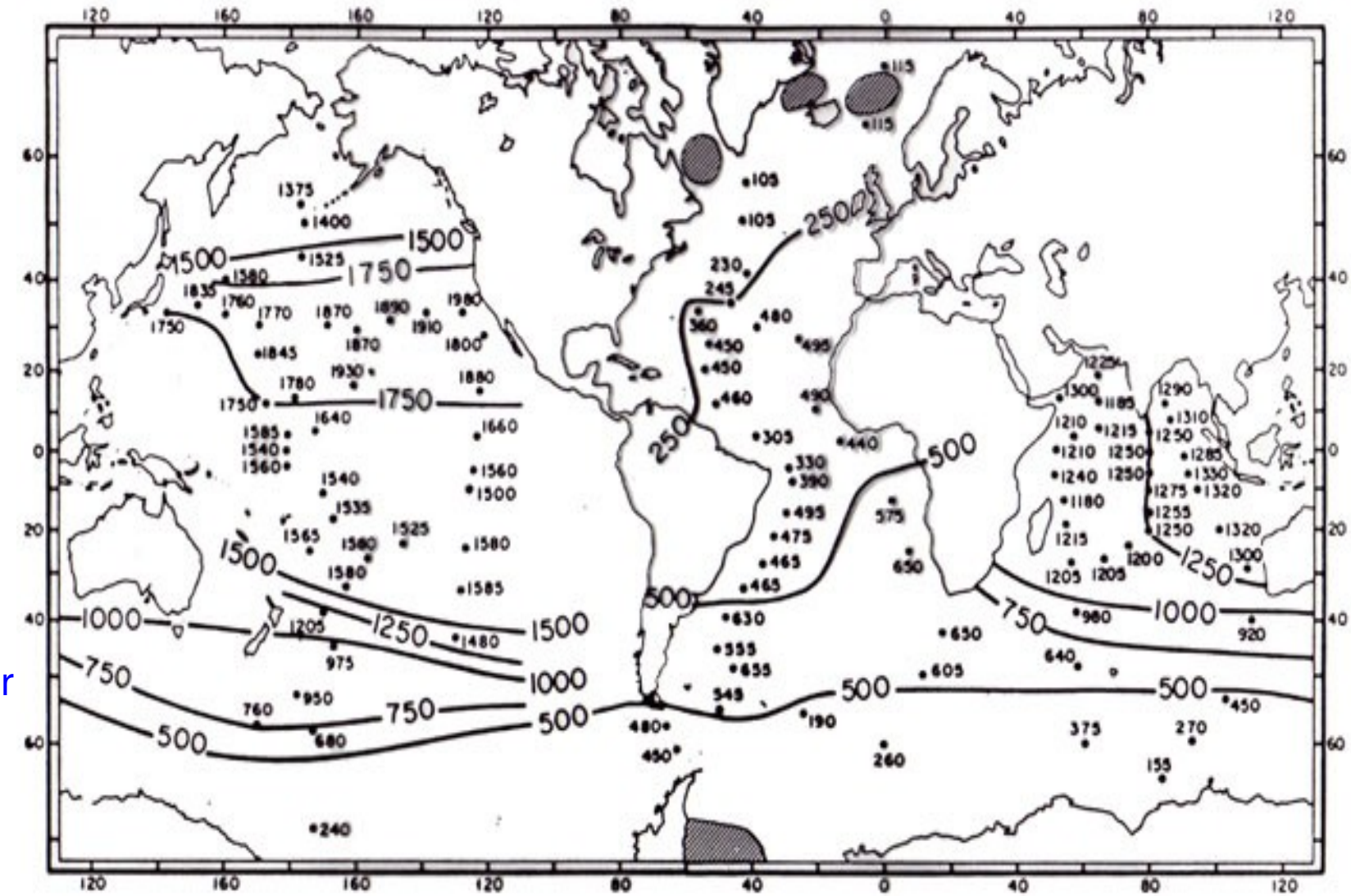
Units are tritium units, where one tritium unit corrected to the activity levels that would have been observed on 1 January 1981 (Toggweiler 1994)

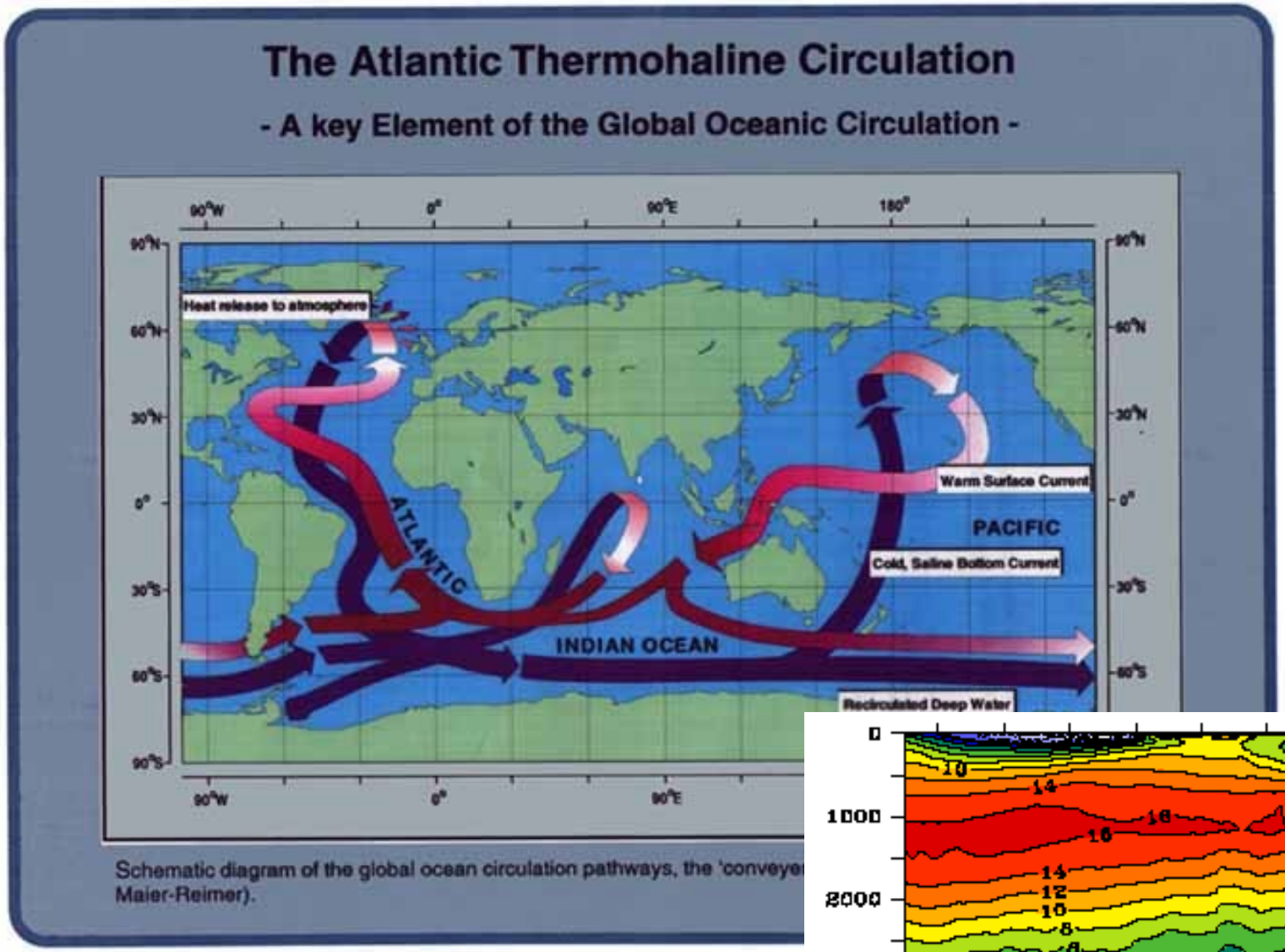


More evidence: Distribution of radiocarbon ages for deep sea water (3km). Dots indicate station locations, shaded areas sites of deep water formation (Broe

Radiocarbon is created in the upper atmosphere due to cosmic rays. It enters the ocean through absorption of CO_2, and once below the surface it is isolated from the source and starts to decay.

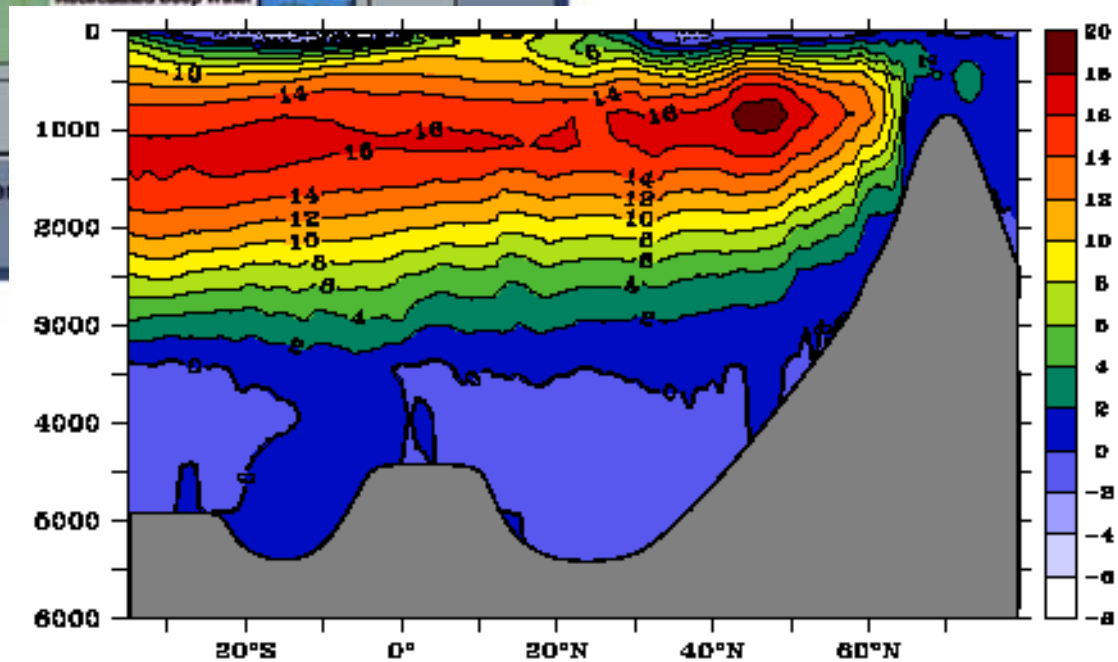
Sites of deep water formation shows the younger radiocarbon ages.





From CLIVAR website

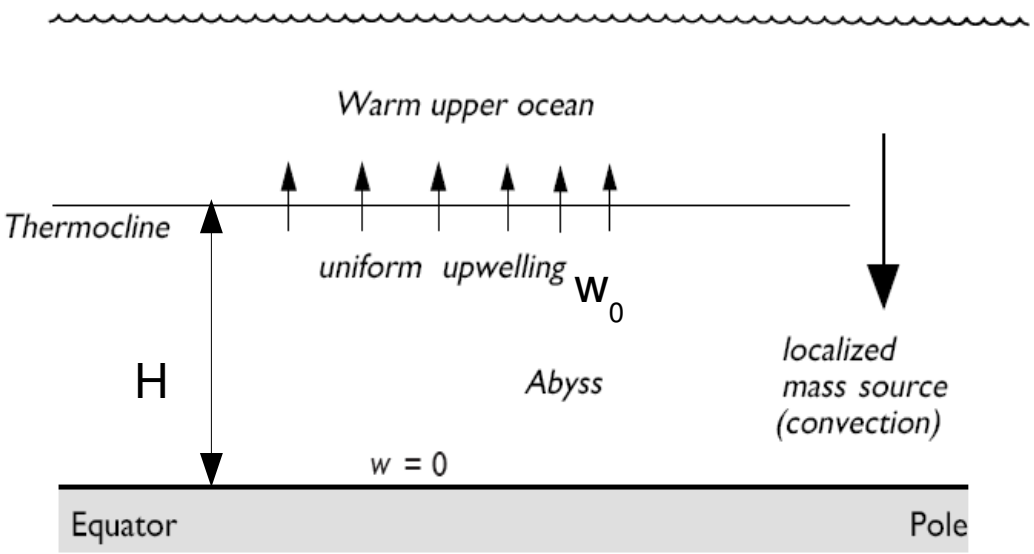
Atlantic Meridional Volume Streamfunction



Sommel-Arons model for the abyssal circulation (Vallis 2006)

Assumptions:

1. Cold, deep water is supplied by deep convection at high-latitude locations.
2. Mixing in the ocean brings the cold, deep water back to the surface uniformly.
3. Flat bottom ocean.
4. The abyssal circulation is strictly geostrophic in the interior of the ocean, and thus potential vorticity is conserved.



$$f u = \frac{-\partial \phi}{\partial y}, \quad f v = \frac{\partial \phi}{\partial x} \qquad f = f_0 + \beta y$$

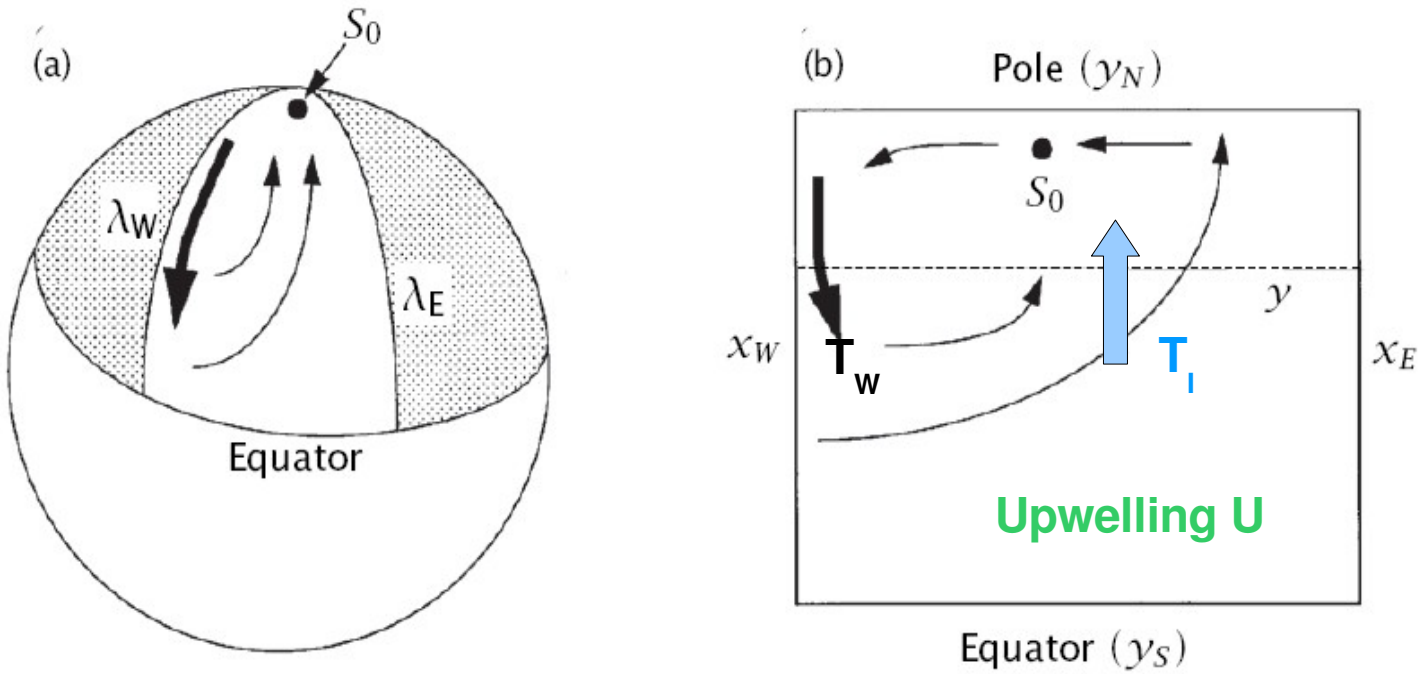
$$\nabla . u = \frac{-\partial w}{\partial z}$$

$$(curl + continuity) \Rightarrow \quad \beta v = f \frac{\partial w}{\partial z} \quad \Rightarrow \quad v = \frac{f}{\beta} \frac{w_0}{H}$$

The interior flow is polewards everywhere.

$$u = \frac{-1}{f} \frac{\partial \phi}{\partial y} = \frac{-1}{f} \frac{\partial}{\partial y} \int_x^{x_E} \frac{-f^2 w_0}{\beta H} dx' = \frac{2}{H} w_0 (x_E - x)$$

The interior flow is eastward and is independent of y.



Since the interior flow $\mathbf{T_I}$ is northwards, there must be a boundary flow moving southward $\mathbf{T_W}$. By conservation of mass in the region north of y

$$S_0 + T_I(y) = T_W(y) + U(y), \qquad S_0 = w_0(x_E - x_W)(y_N - y_S)$$

$$T_I(y) = \int_{x_W}^{x_E} vH\,dx = \frac{\int_{x_W}^{x_E} f w_0}{\beta}\,dx = \frac{f}{\beta} w_0 (x_E - x_W)$$

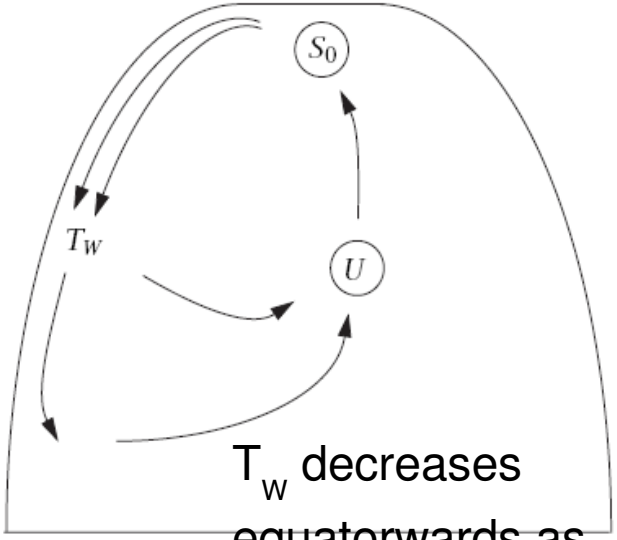
$$U = \int_{x_W}^{x_E} \int_{y}^{y_N} w\,dy\,dx = w_0(x_E - x_W)(y_N - y)$$

$$\Rightarrow \quad T_W(y) = S_0 + T_I - U = S_0 + \frac{f}{\beta} w_0 (x_E - x_W) - w_0(x_E - x_W)(y_N - y)$$

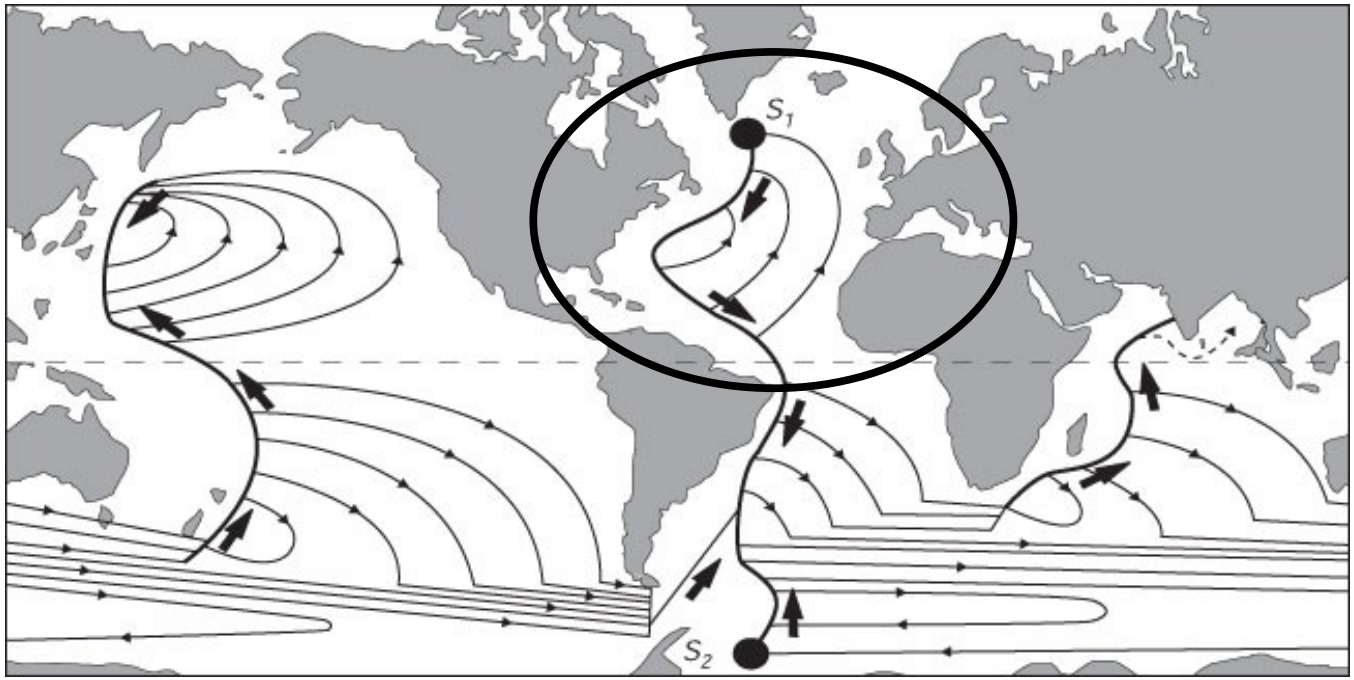
$$\Rightarrow T_W(y) = 2 S_0 \frac{y}{y_N}, \qquad (y_S = 0, f_0 = 0)$$

Properties of the solution:

1. the western boundary current is equatorward everywhere
2. the northward mass flux at the northern boundary is $T_I(y_N)=S_0$
3. the transport of the western boundary current at the northern boundary is twice the strength of the source $T_W(y_S)=2S_0$ => recirculation



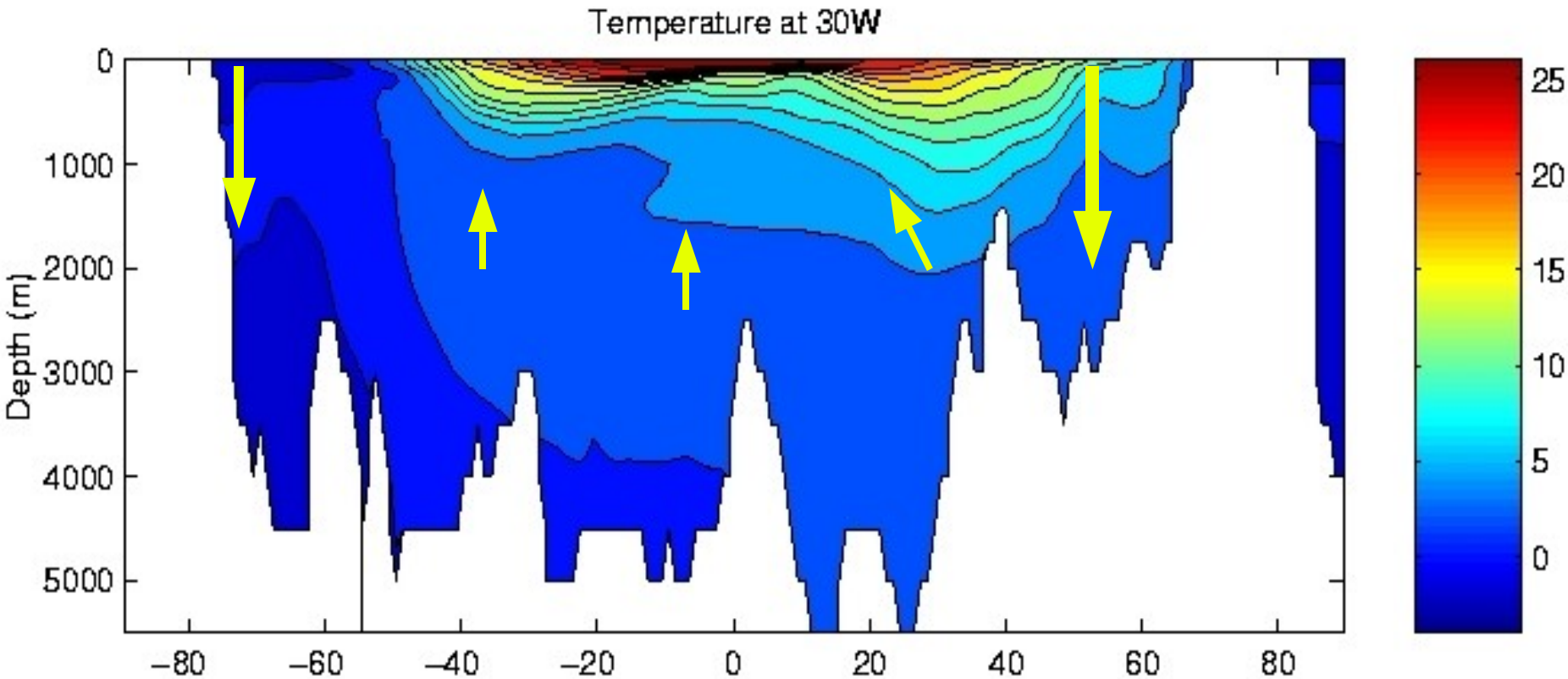
T_W decreases equatorwards as it loses mass to T_I and then upwelling.



Stommel 1958

Deep convection sets the deep ocean water properties and guarantees it is cold even when the surface is very warm. As heat is constantly diffusing downwards from the surface, the deep waters would eventually warm up if this diffusion was not countered by deep sinking injecting cold water to the bottom. Thus the thermohaline circulation is partly responsible for maintaining the stable stratification of the world's oceans.

The main vertical balance is advection-diffusion: $w\frac{\partial T}{\partial z}=\kappa\frac{\partial T^2}{\partial z^2}$



Scaling of the deep circulation

1) Thermal wind
(geostrophy + hydrostatic)

$$f\frac{\partial u}{\partial z}=-g\gamma\frac{\partial T}{\partial y} \quad \Rightarrow \quad f\frac{U}{D}=g\gamma\frac{\Delta T}{L}$$

2) Geostrophic vorticity equation
(geostrophy + incompressible).

$$\beta v=f\frac{\partial w}{\partial z} \quad \Rightarrow \quad \beta U=f\frac{W}{D}$$

3) Thermodynamics: The slow diffusion of heat is balanced by the slow transport of cold waters from the poles.

$$w\frac{\partial T}{\partial z}=\kappa\frac{\partial T^2}{\partial z^2} \quad \Rightarrow \quad W=\frac{\kappa}{D}$$

From 1) and 2)

$$W=\beta\frac{D^2}{f^2}g\gamma\frac{\Delta T}{L}$$

Using 3) gives a scaling for the vertical velocity W and the depth of the thermocline

$$W=\kappa^{2/3}\left[\frac{\beta g\gamma\Delta T}{f^2 L}\right]^{1/3}$$

$$D=\kappa^{1/3}\left[\frac{f^2 L}{g\beta\gamma\Delta T}\right]^{1/3}$$

Both, the upwelling strength and thermocline depth depend on the diapycnal turbulent diffusion.

Issues

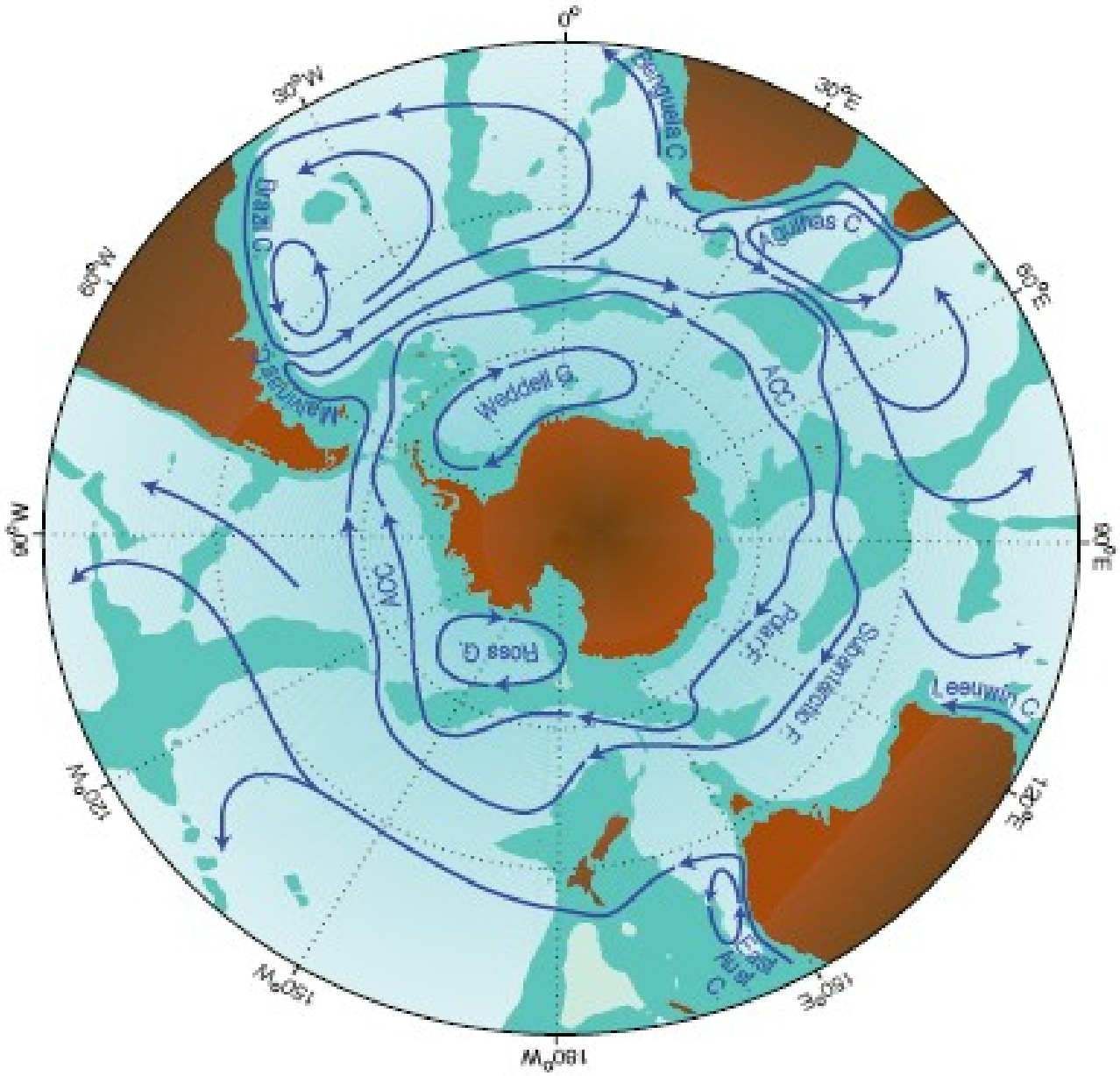
- Deep convection alone leads to a pool of cold stagnant deep water (Munk and Wunsh 1998). Thus, deep convection does NOT drive the circulation, mixing does. Diapycnal mixing is necessary to pump cold water upward through the thermocline and drive the circulation. The power of mixing comes from winds and tides.

- **Missing mixing problem**: the measured mixing in the main thermocline is small (~0.1 cm/s^2). It is large near ocean ridges and rough terrain (> 1 cm/s^2), but can not account for the observed circulation.

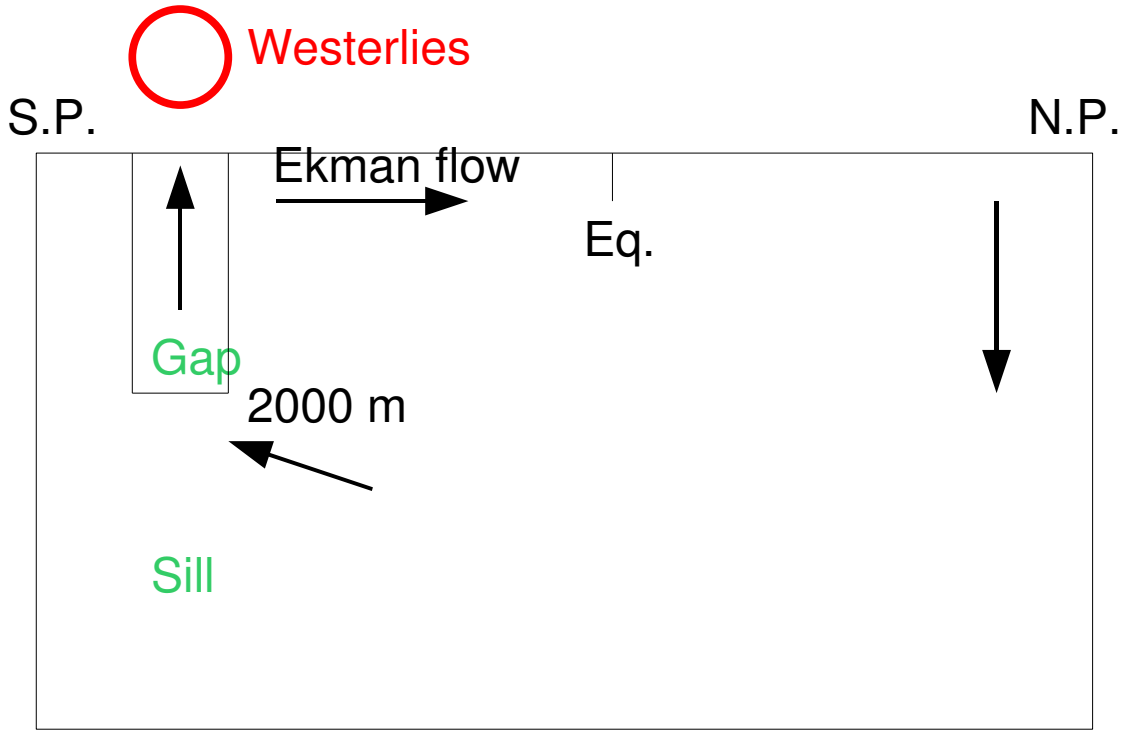
- Water does not upwell uniformly across the thermocline in mid and low latitudes.

An alternative viewpoint is that the deep circulation is directly driven by the winds over the Southern Ocean.

The Drake Passage allows the existence of the Antarctic Circumpolar Current



The Drake passage effect (Toggweiler and Samuels 1995)

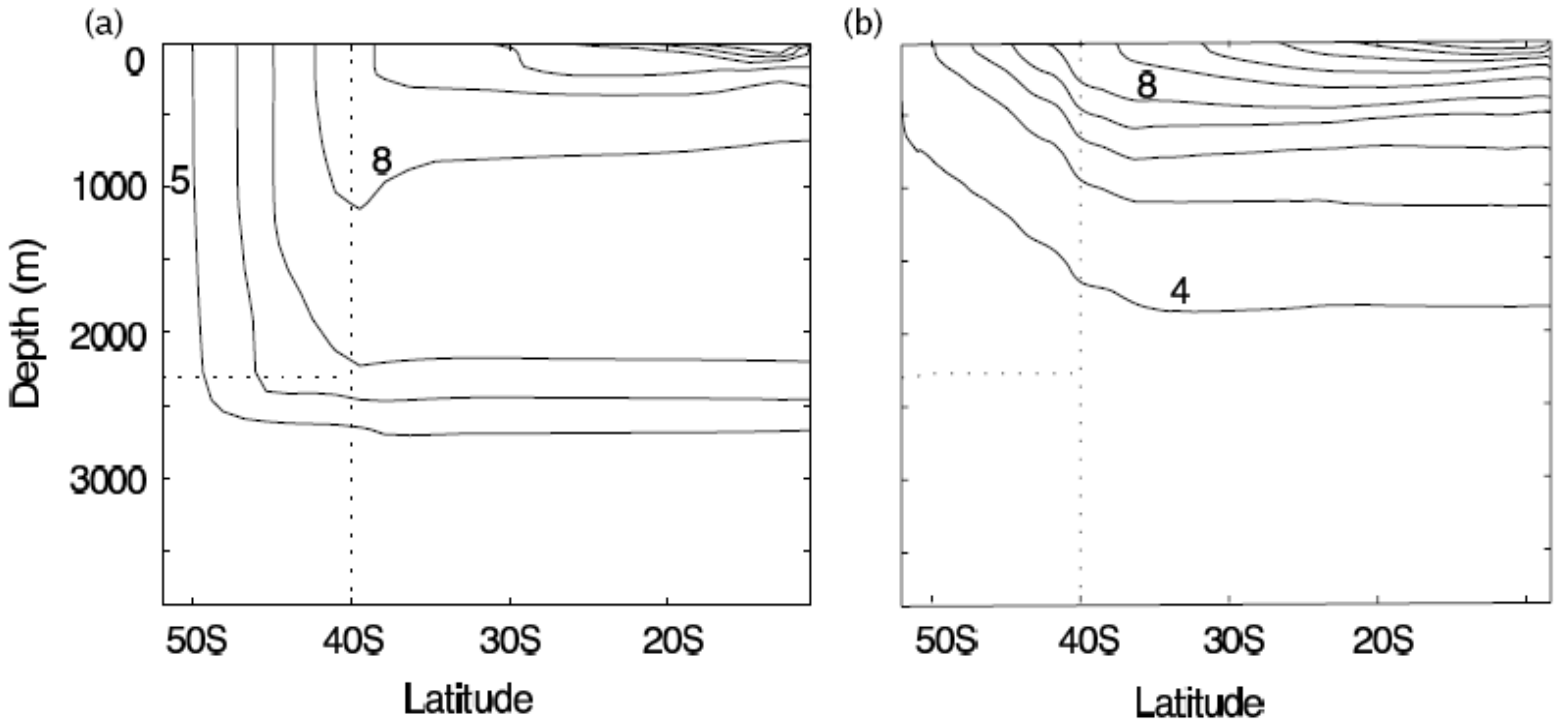


• Westerlies over the Southern Ocean lead to divergence and wind-driven upwelling.

• The northward waters pushed by the wind come only from deep waters due to the lack of topographic barriers at the latitude band of the Drake Passage. Only below 2000 m the topography can sustain a pressure gradient that drives a southward flow.

• It is argued that the downward branch that connects the deep southward flow with the surface northward flow can only occur in the North Atlantic due to stable stratification elsewhere. Thus the energy input for the THC comes from the Southern Ocean winds, and the THC results from a continuity requirement. This argument ensures the existence of a deep circulation even in the limit of very small mixing.

The role of the eddies in the ACC



Simulations of the temperature structure in the ACC region in a diffusive model with no baroclinic eddies. The channel creates steep isotherms and the flow is baroclinically unstable (large APE).

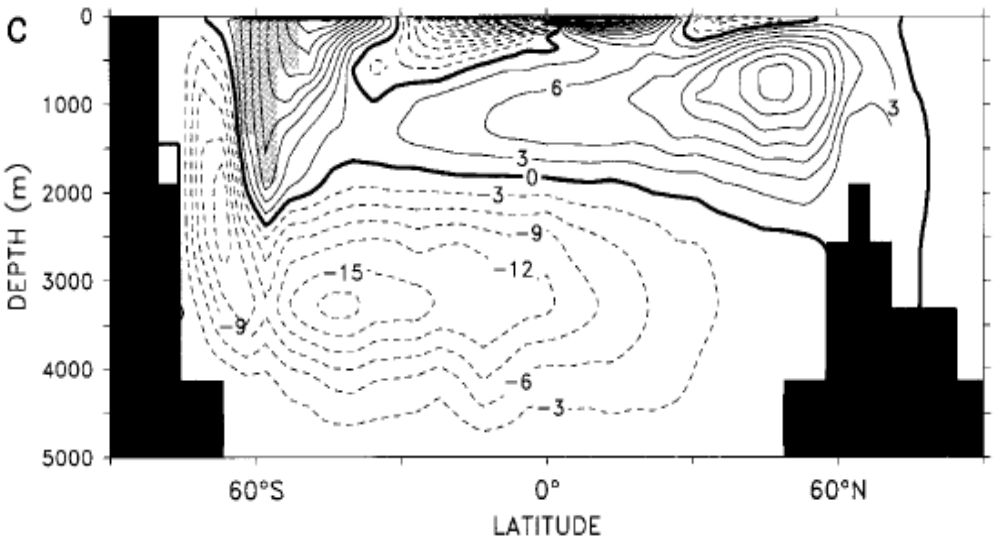
Simulations of the temperature structure in a high-resolution model that allows baroclinic eddies to form. The mesoscale eddies release the APE and isotherms become less vertical. Thus eddies control ACC dynamics.

Vallis 2006

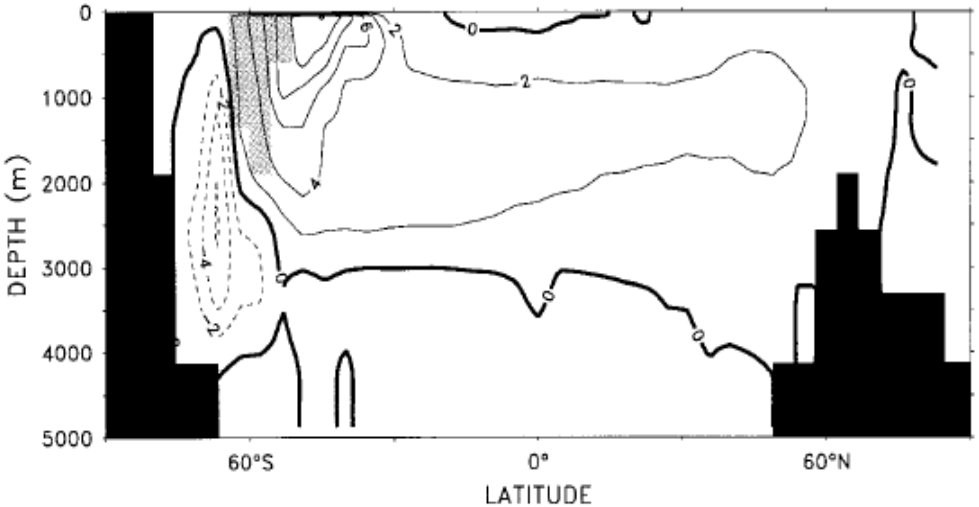
The Deacon Cell

In ocean models, part of the Ekman drift recirculates locally. This so-called Deacon Cell reflects the obstacle of the Drake Passage gap (note that the three other Ekman cells are closed within the upper 1000m).

Ocean volume streamfunction

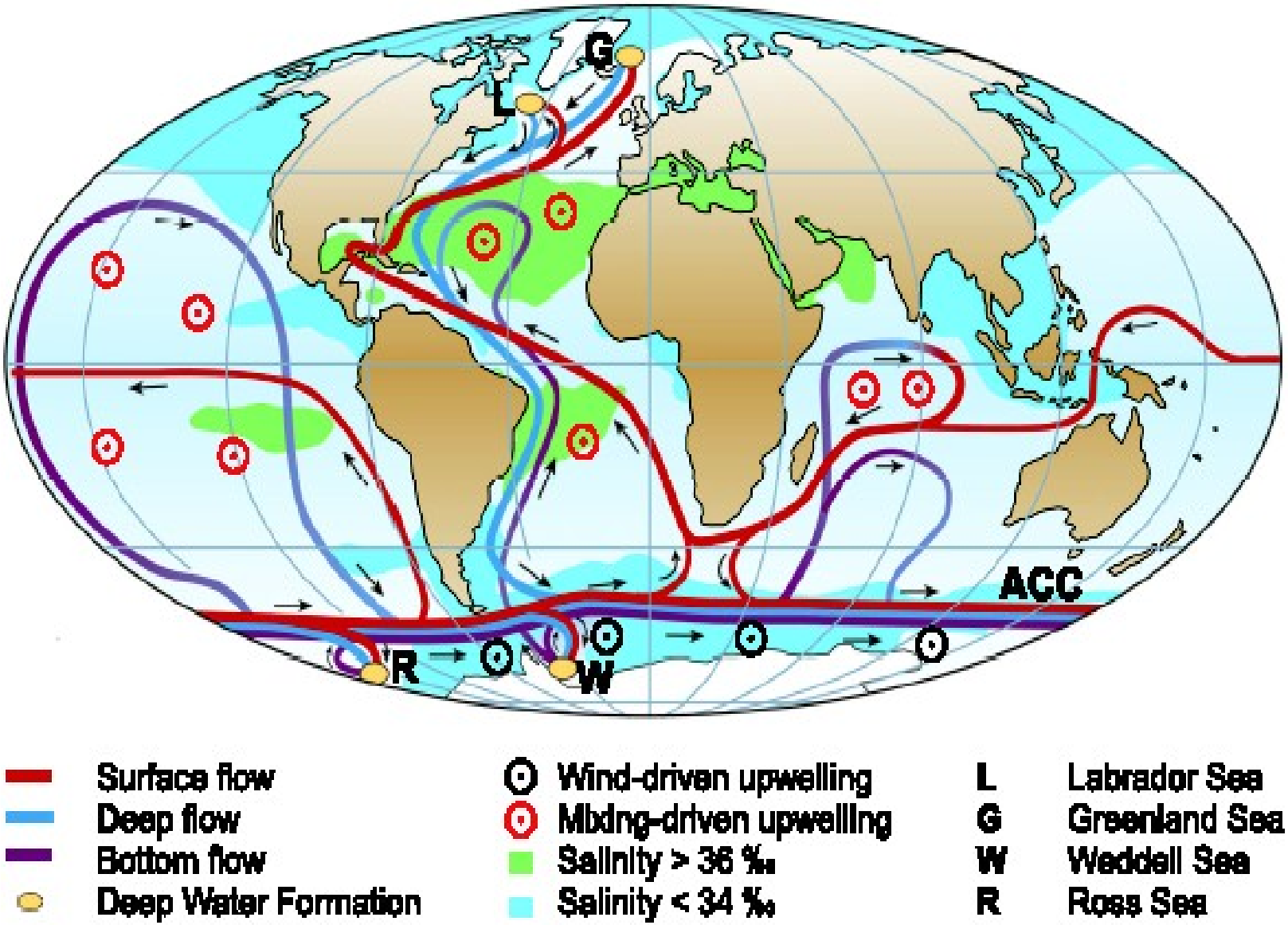


A 40% increase in Southern Ocean zonal wind stress increases the Deacon Cell by 12 Sv, but only 3 Sv sink in the North Atlantic as NADW. The Deacon Cell associated sinking is along isopycnals, and is consistent with density stratification noted by Toggweiler and Samuels (1995)



Rahmstorf and England 1997

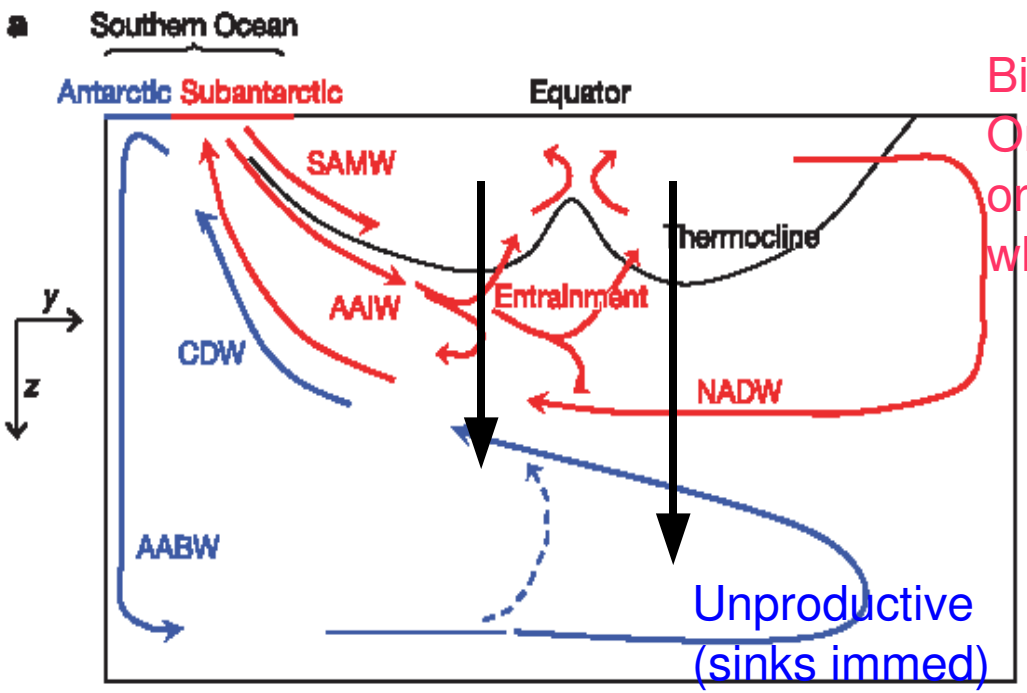
A better description of the thermohaline circulation: turbulent mixing + S.O. wind-driven upwelling



Rahmstorf (2006)

Deep circulation and CO_2 impacts (Toggweiler et al 2006)

Westerlies



Marinov et al (2005)

Biologically productive (travels a lot). Organisms strip the nutrients, and produce organic particles that fall into the deep ocean, where they are respired elevating pCO2

Unproductive (sinks immed) high pCO2.

If the red circulation remains active, but the blue is weak, the escape of CO2 back to the atmosphere is restricted. CO2 will accumulate in the deep ocean as the CO2 content in atmosphere and upper ocean fall.

CDW- Circumpolar Deep Water
AABW-Antarctic Bottom Water

NADW-North Atlantic Deep Water
SAMW-Subantarctic Mode Water
AAIW-Anatarctic Intermediate Water

Positive feedback:

↓ Poleward shifted westerlies

↓ Stronger upwelling in S.O.

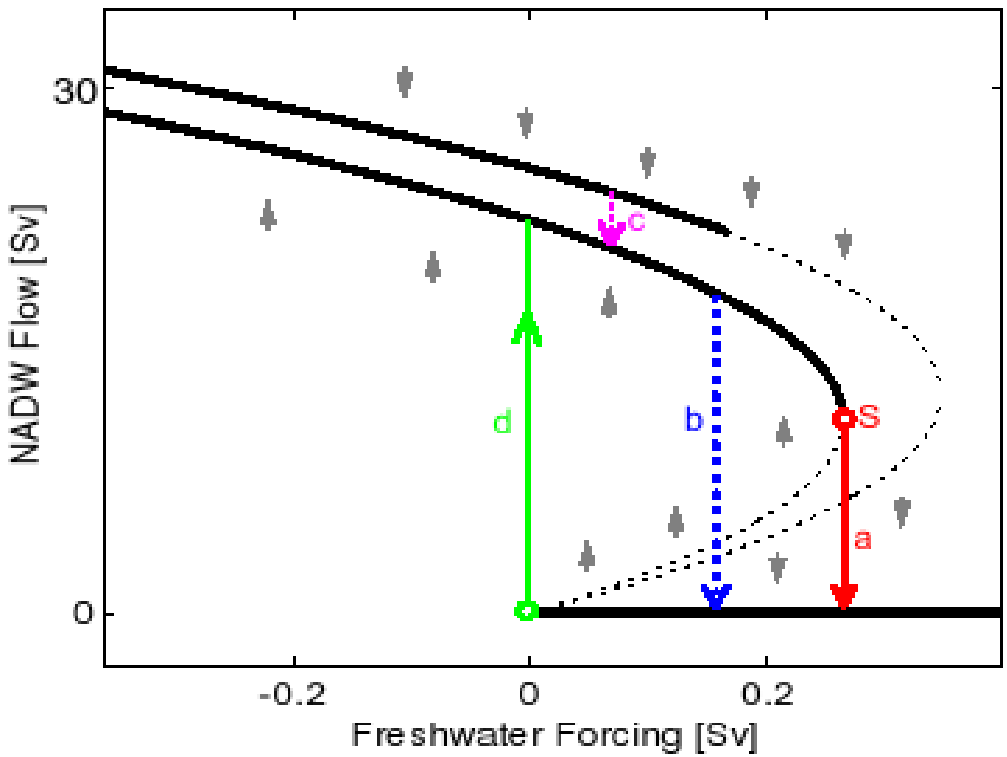
↓ Higher atmospheric CO2

Tropospheric warming ↑

Stability of the Thermohaline Circulation

• The dependence on salinity of the THC creates a positive feedback (Stommel 1961) : higher salinity in the area of deep water formation enhances the circulation, which, in turn, brings more salt to the area.
• This feedback results in a nonlinear dependence of the strength of the THC on freshwater and the system presents bi-stability.

Stability diagram



Hysteresis in coupled models

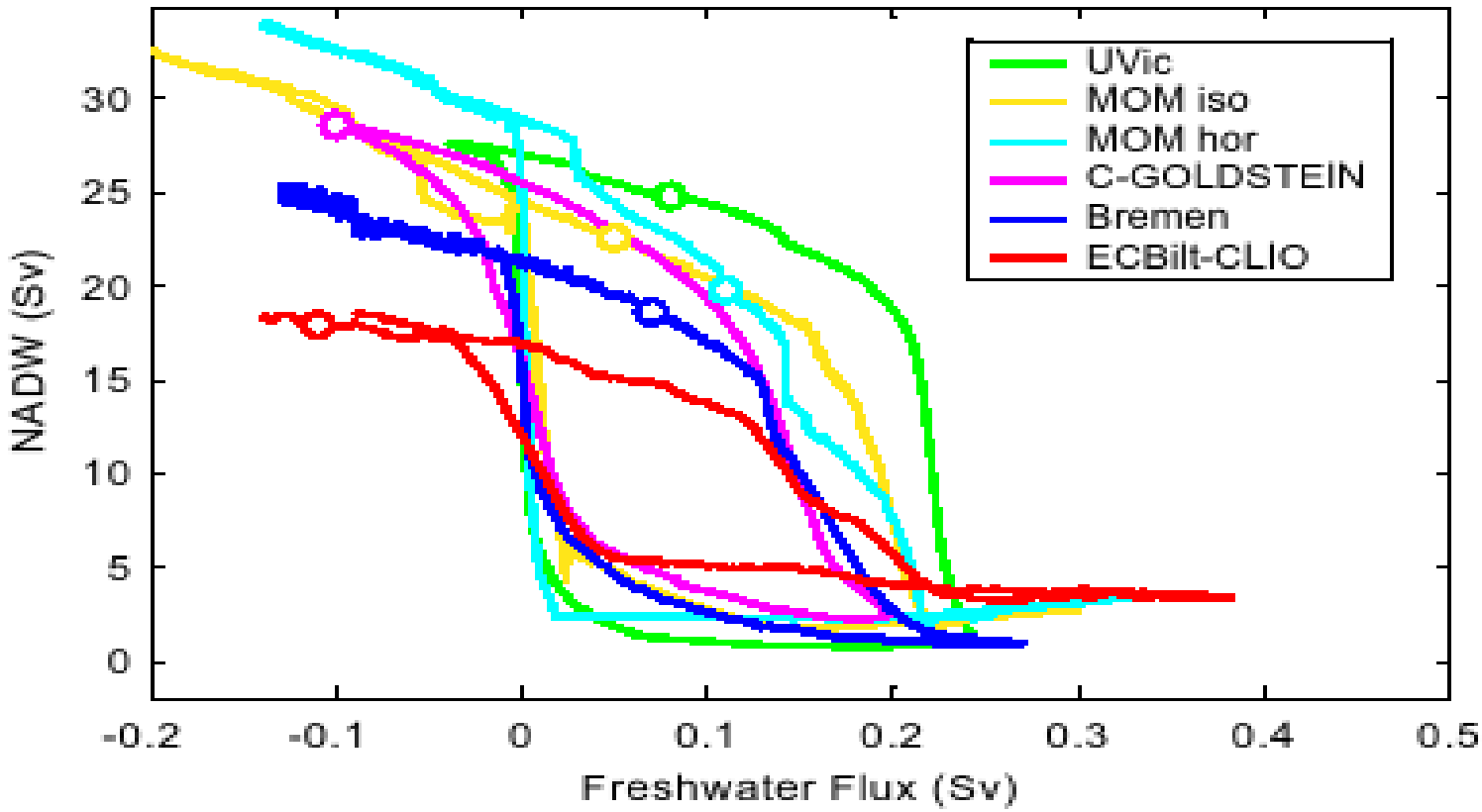


Fig. 10 Hysteresis curves from a range of different intermediate-complexity climate models including 3-dimensional ocean models. The curves are derived by slowly increasing the freshwater influx to the north Atlantic, then decreasing it again, from a present-day initial state of each model marked by an open circle. From [Rahmstorf, et al., 2005]

Summary of Thermohaline Circulation

- Deep water formation occurs in the Greenland-Norwegian Sea, the Labrador Sea, Weddell and Ross Seas.
- Deep waters spread mainly as deep western boundary currents (e.g. North Atlantic Deep Water).
- An estimate of ~0.4 TW is necessary to drive the THC (Munk and Wunsch 1998). This energy power comes from the turbulent mixing in the ocean's interior, which, in turn, is due to winds and tides. The Drake Passage effect may also help in driving the THC.
- The THC is sensitive to buoyancy perturbations in the sites of deep water formation because decreases the meridional density gradient. $W \sim \kappa^{2/3} \Delta T^{1/3}$

$$D \sim \kappa^{1/3} \Delta T^{-1/3}$$

Issues:

- The thermohaline circulation can be described with a coarse ocean model.

- The equator is not a special region, the cold tongues are weak, and the ocean absorbs heat more uniformly than observed.

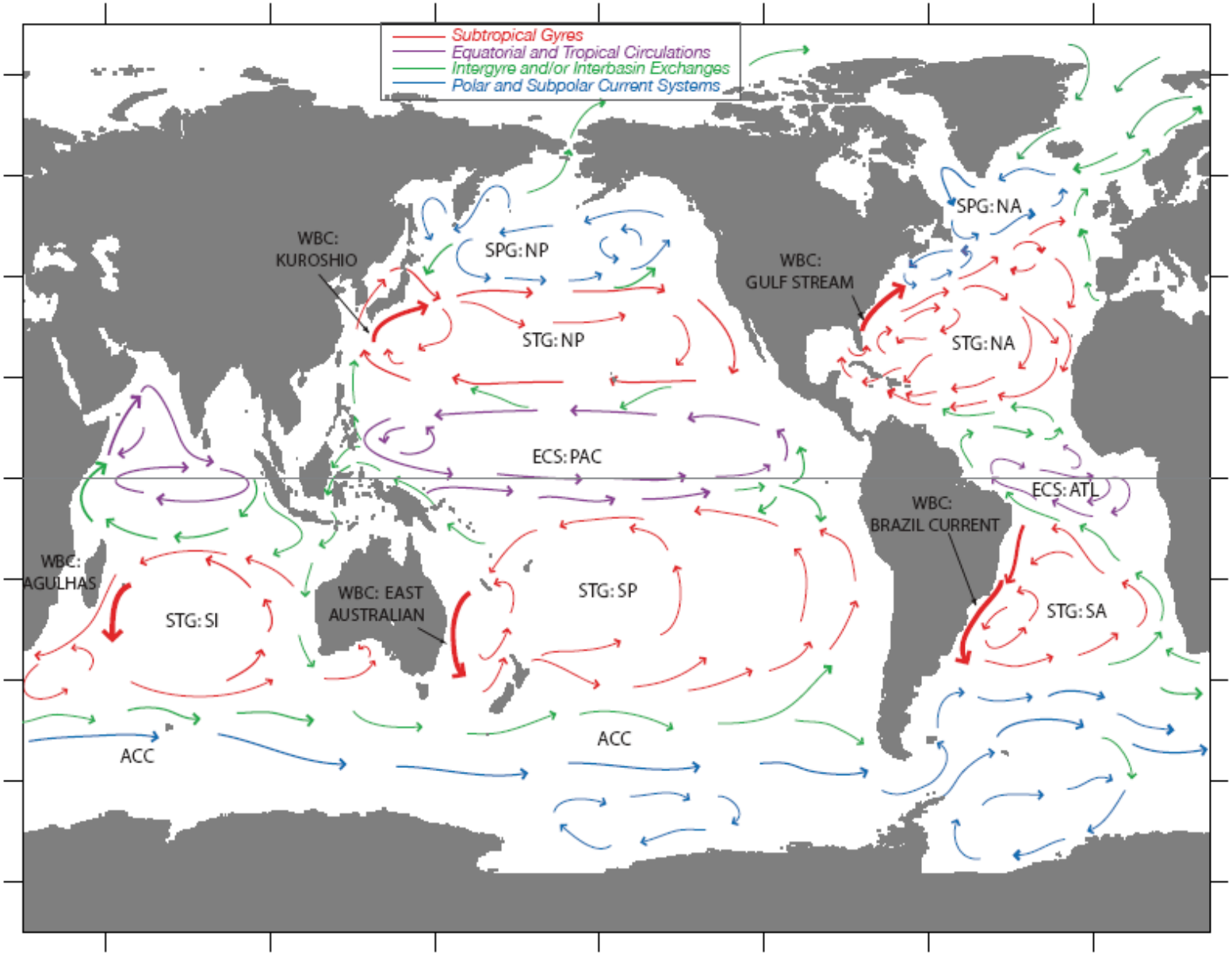
- The good representation of this large scale circulation does not ensure a good representation of the tropical circulation.

- The thermocline structure associated with this circulation is more diffuse than the observed.

- Except in the Southern Ocean, the winds do not have a direct influence in the circulation thus far described. Nevertheless, in the subtropical gyres wind converges and downwells, and therefore influences the structure of the thermocline.

The oceanic wind-driven shallow circulation

Surface ocean Currents



Vallis 2006

Ekman dynamics

Consider a boundary layer where the flow is given by the wind stress, and the balance is

$$\vec{f} \wedge \vec{u}_E = \frac{1}{\rho_0} \frac{\partial \vec{\tau}}{\partial z} \qquad \vec{f} = f\,\vec{k}$$

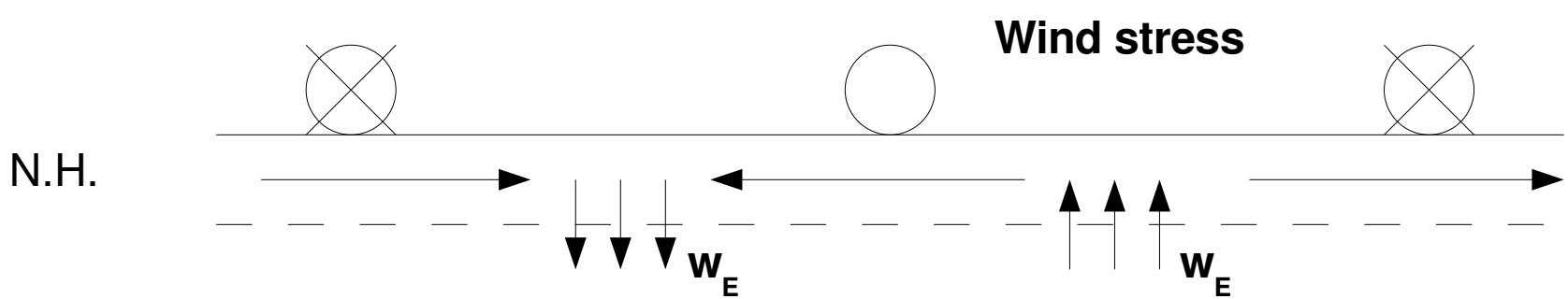
$$(integrate\ on\ the\ thickness\ of\ the\ layer) \Rightarrow \quad \vec{M}_E = \int_{Ek} \rho_0 \vec{u}_E\, dz = \frac{-1}{f} k \wedge \vec{\tau}_T$$

and the transport is always at right angles to the stress. Using the continuity equation

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

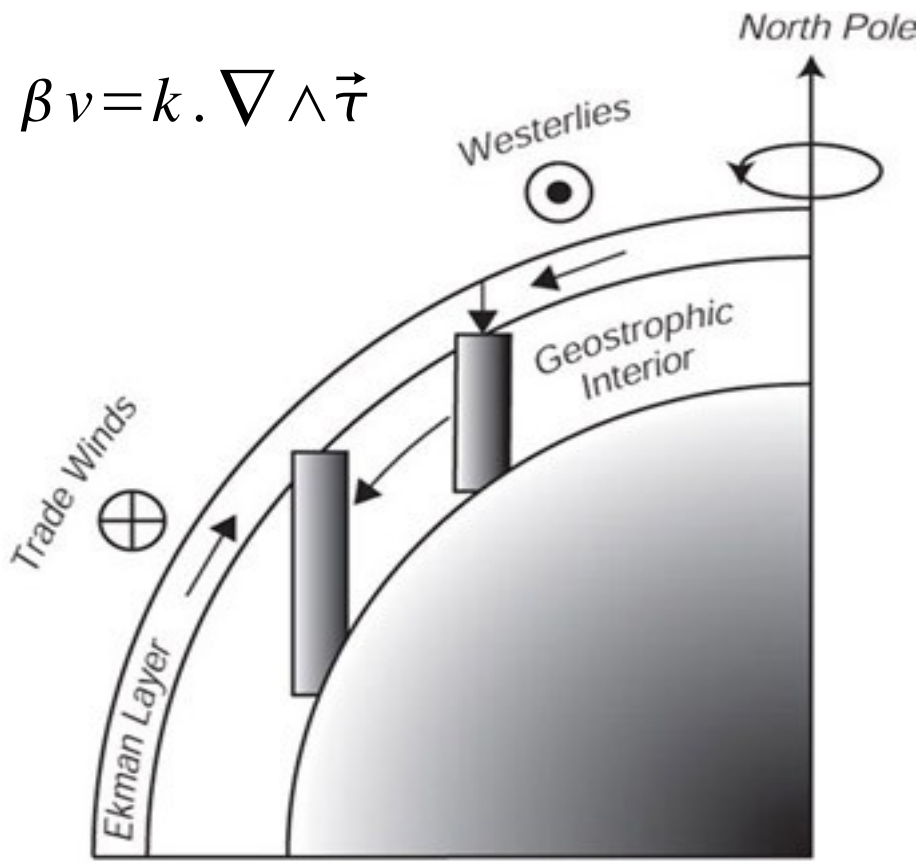
$$(integrate\ over\ depth\ layer) \Rightarrow \quad \frac{1}{\rho_0} \nabla . \vec{M}_E = -w_E \qquad (neglect\ geostrophy)$$

$$\Rightarrow \quad w_E = \frac{1}{\rho_0} k . \nabla \wedge \vec{\tau}_T$$



Sverdrup Dynamics

$$\beta v = k \,.\, \nabla \wedge \vec{\tau}$$



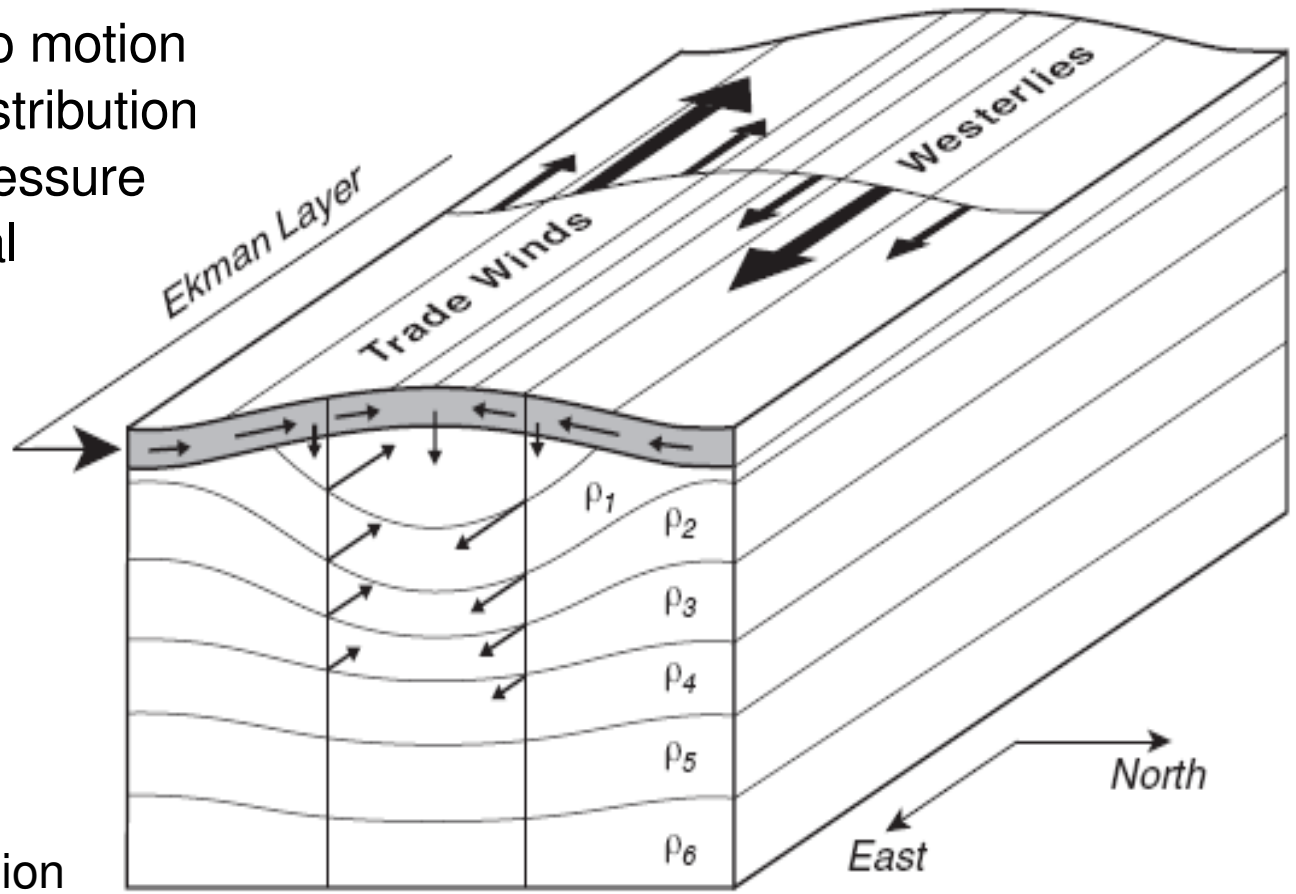
Convergence of Ekman flow in the frictional layer induces downward motion to the ocean's interior, squashing the water column.

Due to conservation of potential vorticity f/H, the water column needs to move southward (Sverdrup flow).

How does Ekman pumping drive geostrophic flow in midlatitudes?

The converging Ekman drift produces a pool of warm water that domes up if there is a level of no motion at depth. This density distribution produces north-south pressure gradients that drive zonal gesotrophic flows.

$$u = \frac{-1}{\rho f} \frac{\partial p}{\partial y}$$



Level of no motion

Surface height calculated from winds

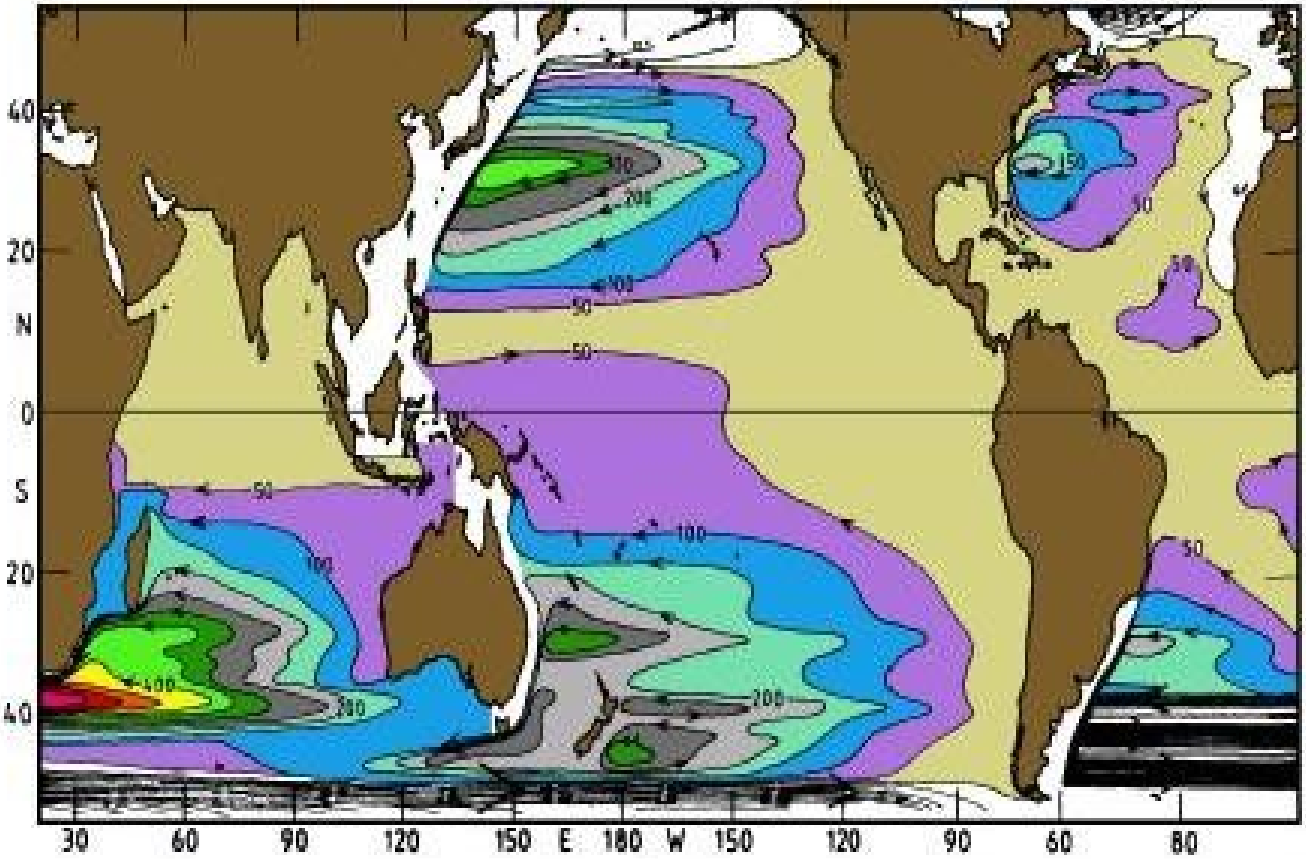
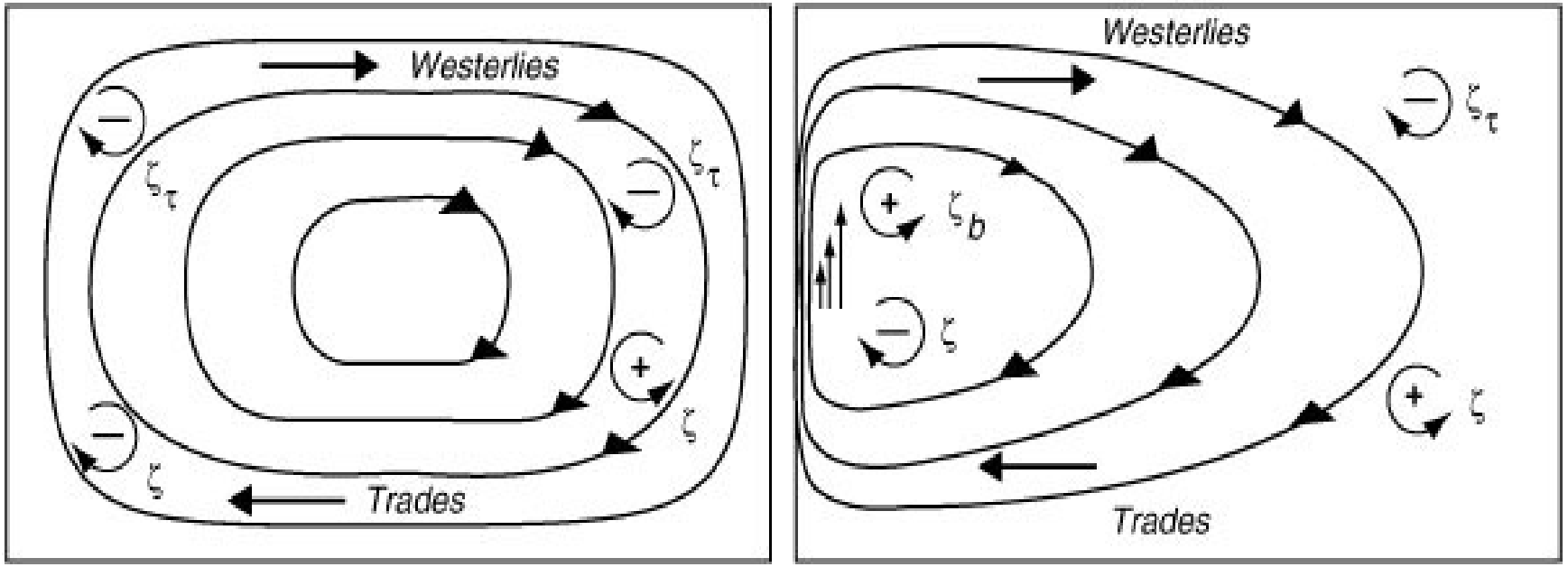


Fig. 4.4. Depth-integrated steric height P, calculated from the right-hand side of the Sverdrup relation (eqn (4.5)), using the data from Hellerman and Rosenstein (1983). Units are 10^1 m^2. For details of the integration procedure see Godfrey (1989).

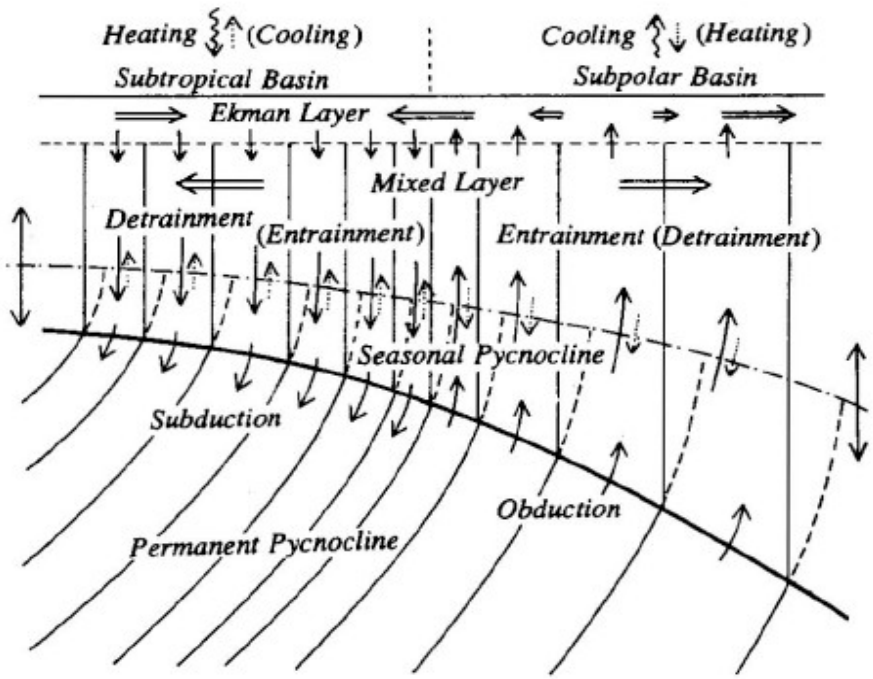
Western Intensification

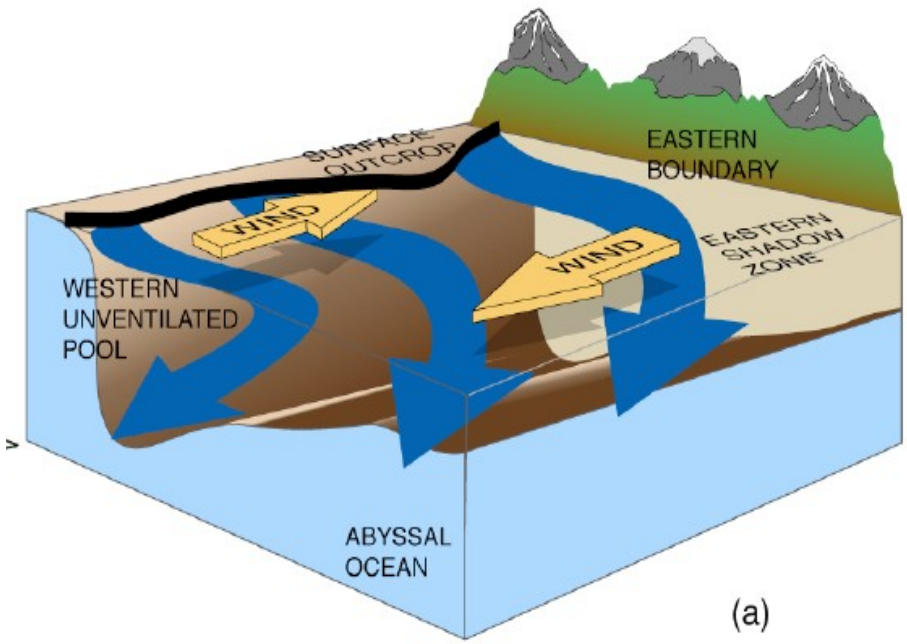


The wind through friction puts "negative" vorticity into the subtropical ocean. For a steady state this vorticity must be removed somewhere. It is removed through friction in a narrow western boundary current.

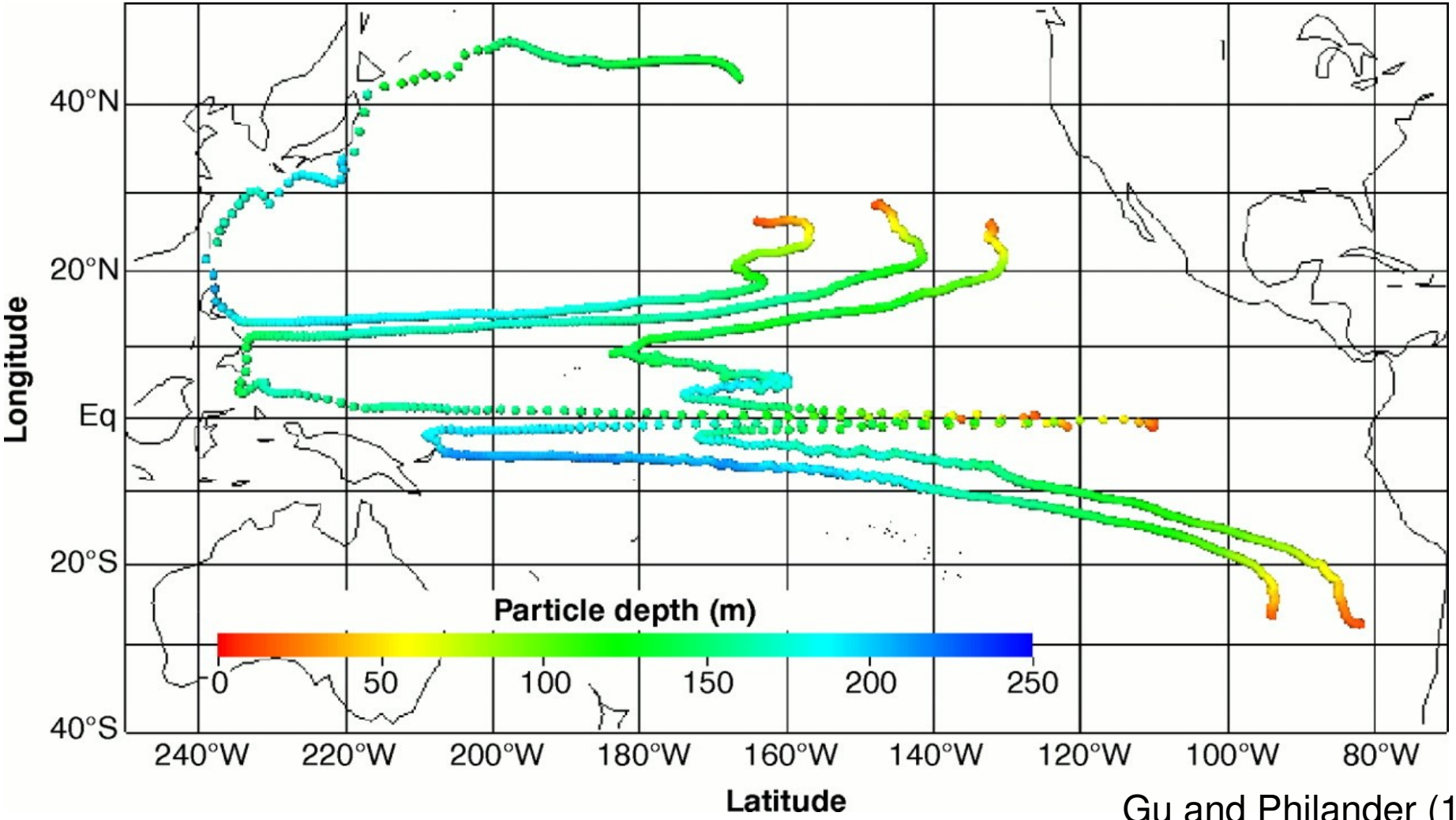
3-dimensional structure of the wind-driven circulation

In the subtropical gyres wind converges and downwells, and therefore influences the structure of the thermocline. What is the fate of the subducted water?





The shallow wind-driven circulation of the ventilated thermocline



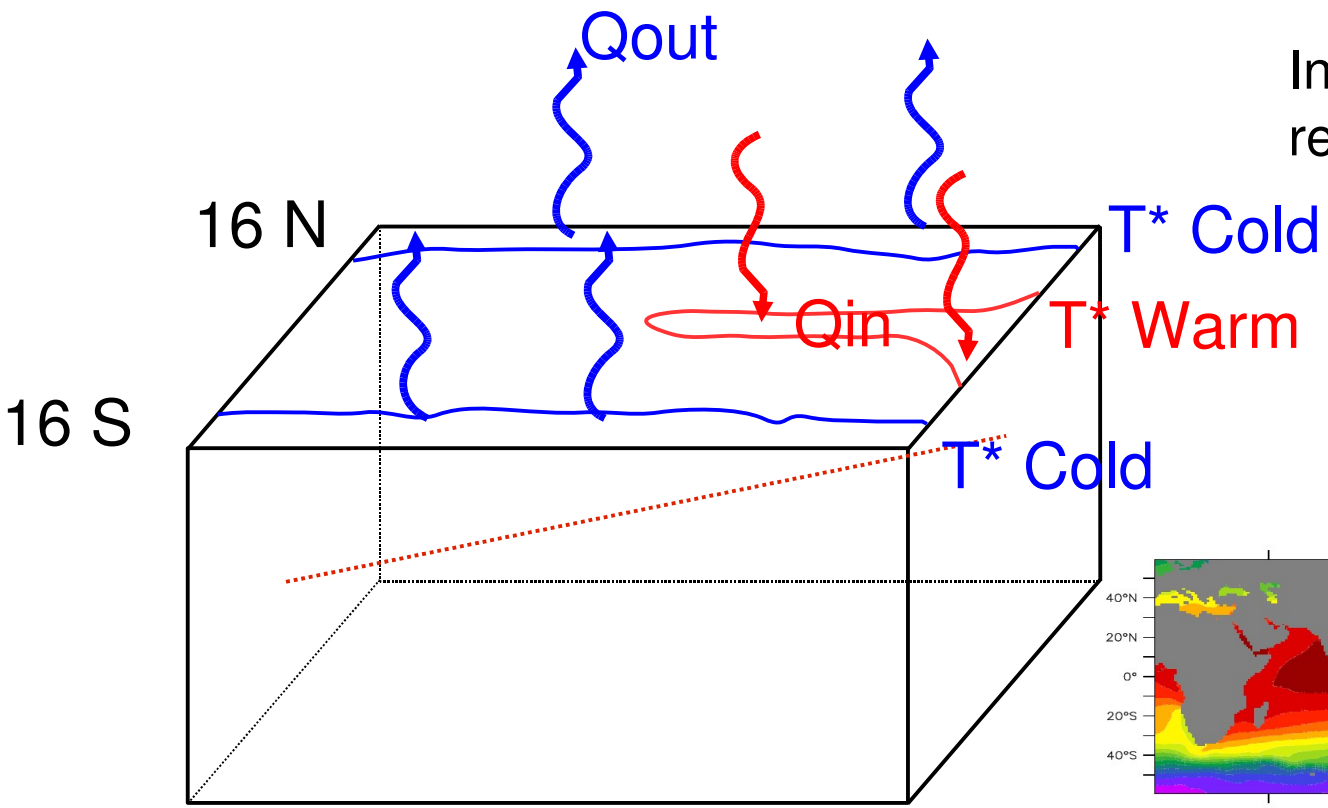
Gu and Philander (1997)

Evolution of water parcels over a 16-year period. Water subducts in the eastern subtropical basins, propagate westward, mostly adiabatically, converge in the equatorial thermocline where they upwell, and move poleward as Ekman drift.

What do we need to represent this circulation?

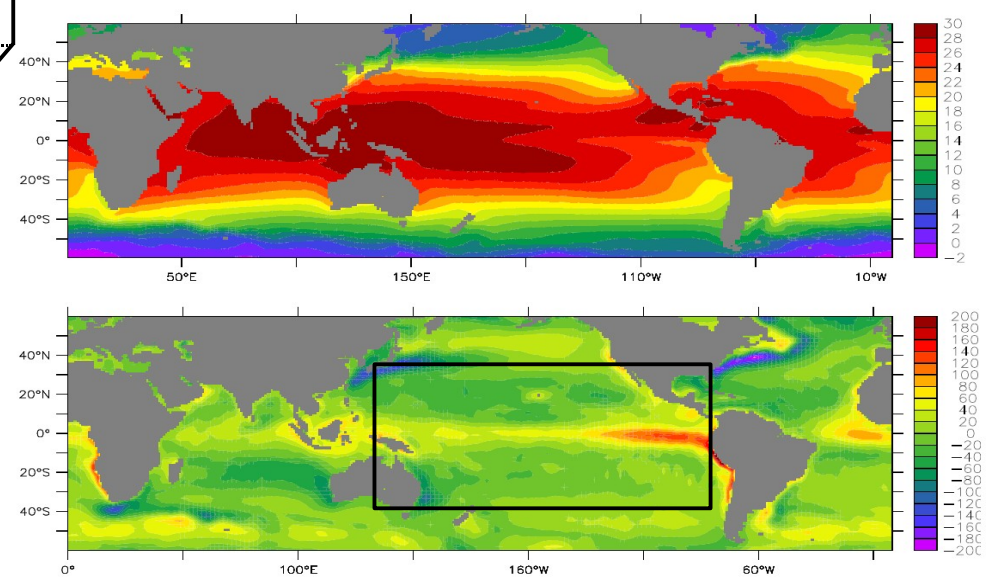
Consider a one-basin ocean GCM forced with:

- easterly wind stress
- meridionally varying air temperature T*, so that the surface heat flux is Q=a(T*-T).



In this small basin the THC is represented by diffusion.

MOM4, 0.5x0.5, 32 levels



Mean state of the ocean model

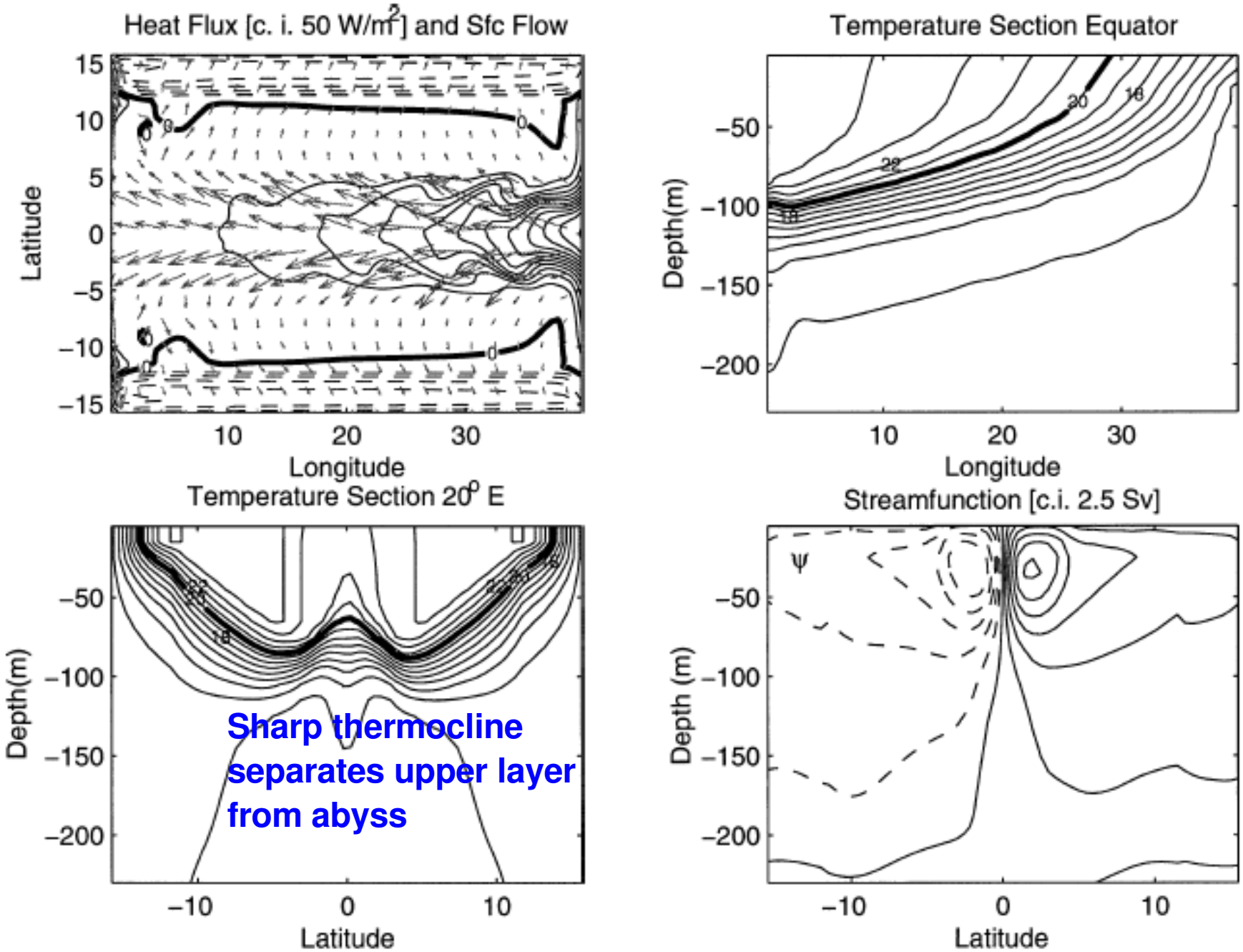
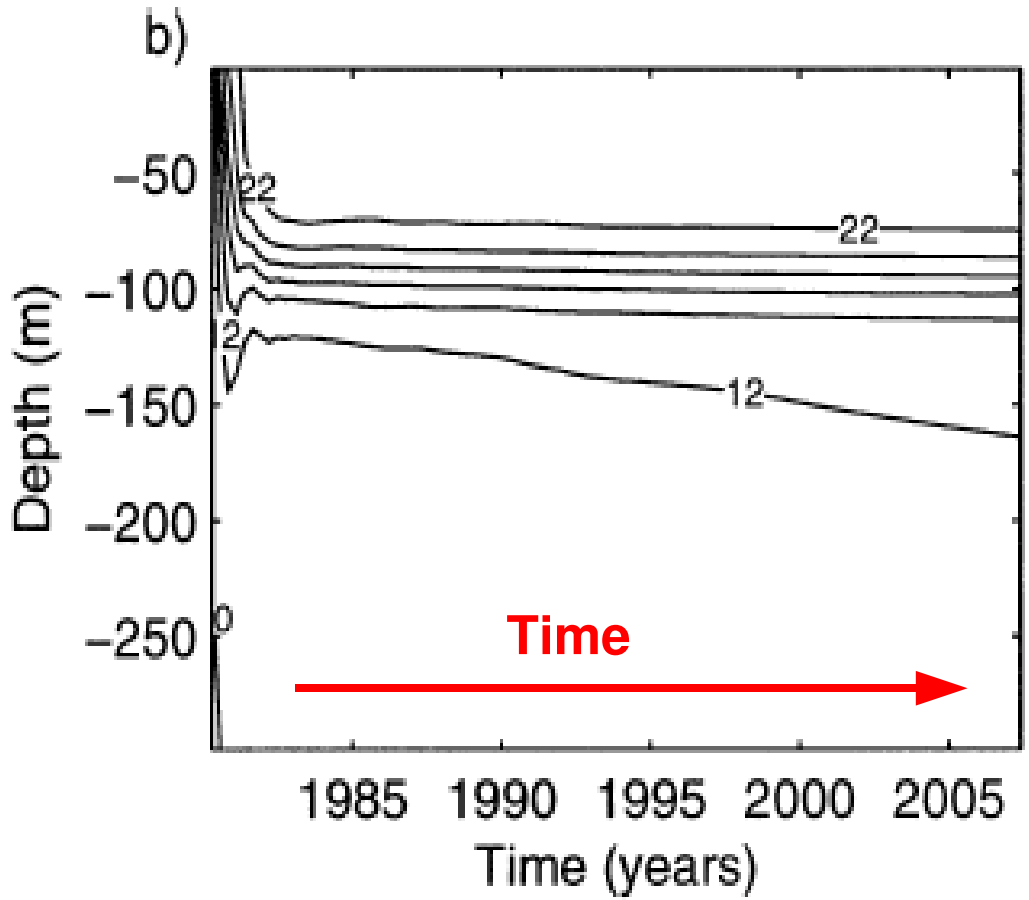


FIG. 3. Mean structure of the steady-state solution for the default case.

The ventilated thermocline sets up very fast in a matter of years, and does not significantly change afterwards.

The deeper ocean warms up much slower until a balance between diffusion of heat and cold upwelling sets up just below the ventilated thermocline.

There is clear separation of time scales between the set up of the ventilated thermocline and the adjustment of the deep ocean.



Evolution of the temperature on the equator in the middle of the basin

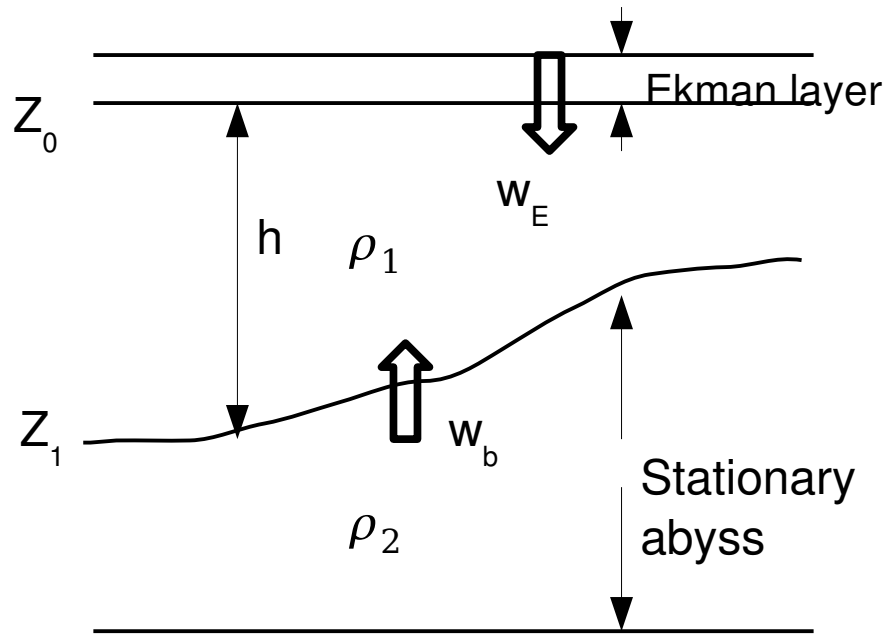
A 1 layer reduced-gravity model of the ventilated thermocline

Assuming geostrophic, steady flow

$$f \wedge \vec{u} = -g' \nabla h, \qquad g' = g \frac{(\rho_2 - \rho_1)}{\rho_0}$$

Taking the curl and using continuity

$$\nabla . \vec{u} = \frac{-\partial w}{\partial z}$$



gives the geostrophic vorticity equation, which integrated over the depth of the active layer

$$h \beta v = f (w_E - w_b)$$

The velocity at the base of the layer w_b under steady conditions $w_b = u . \nabla h = 0$

Thus, $$\frac{\partial (h^2/2)}{\partial x} = \frac{f^2 w_E}{g' \beta}$$ $\Rightarrow$ (integrating from eastern boundary)

$$h^2 = -2\frac{f^2 w_E}{g'\beta}(x_e - x) + h_e^2$$

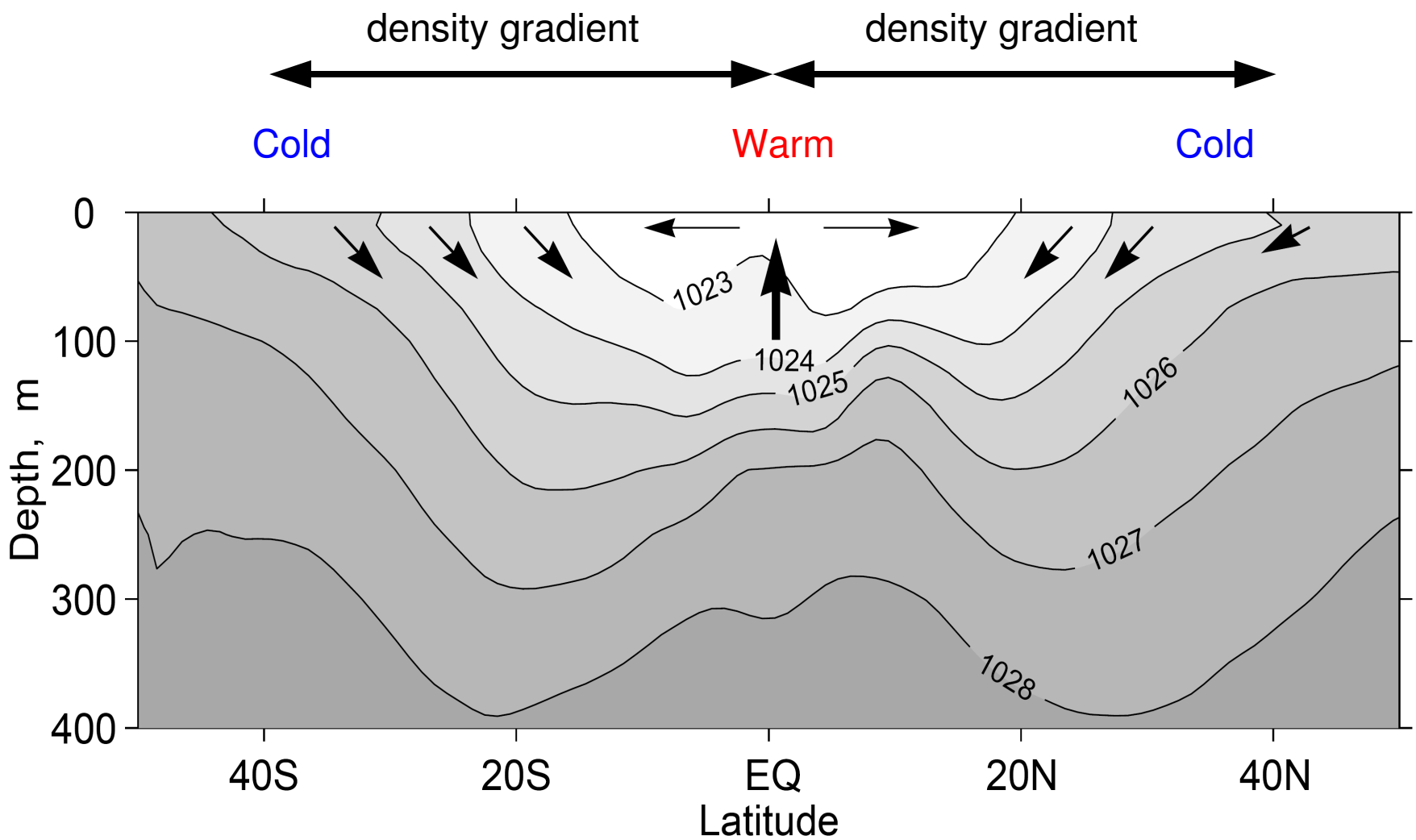
- Note that this equation gives a scaling for the thermocline depth in the context of the theory of the ventilated thermocline

$$g' = g\Delta\rho/\rho_0 = g\gamma\Delta T$$
$$\beta = f/L$$

$$D = \left[\frac{W_E f L^2}{g\gamma\Delta T}\right]^{1/2}$$

Thus, the depth of the thermocline increases with the magnitude of the wind strength, and decreases with the meridional temperature (density) gradient.

Mean state of tropical Pacific shows the structure of the ventilated thermocline.



Summary of circulation of the ventilated thermocline

- Meridional circulation: water parcel subducts in the subtropics, propagates adiabatically towards the equator, joins the Equatorial Undercurrent in the western side of the basin, upwells in the eastern equatorial region, and moves poleward due to Ekman drift.
- The time scale of the circulation is of the order of decades.
- The depth of the thermocline increases with wind strength, and with decreasing surface temperature (density) gradient. $D \sim W_E^{1/2} \Delta T^{-1/2}$