The Ventilated Ocean: Stratification and overturning in an ocean with adiabatic interior and Lagrangian ocean modeling

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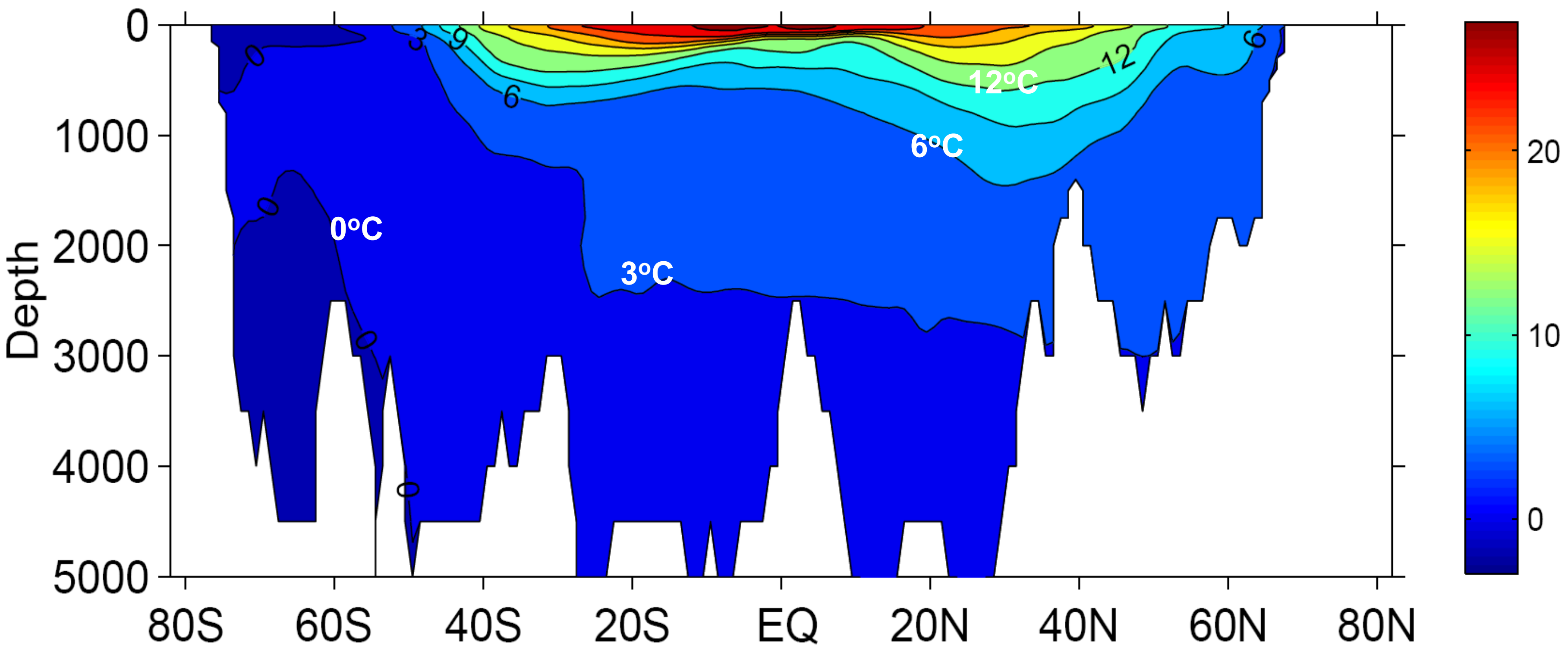
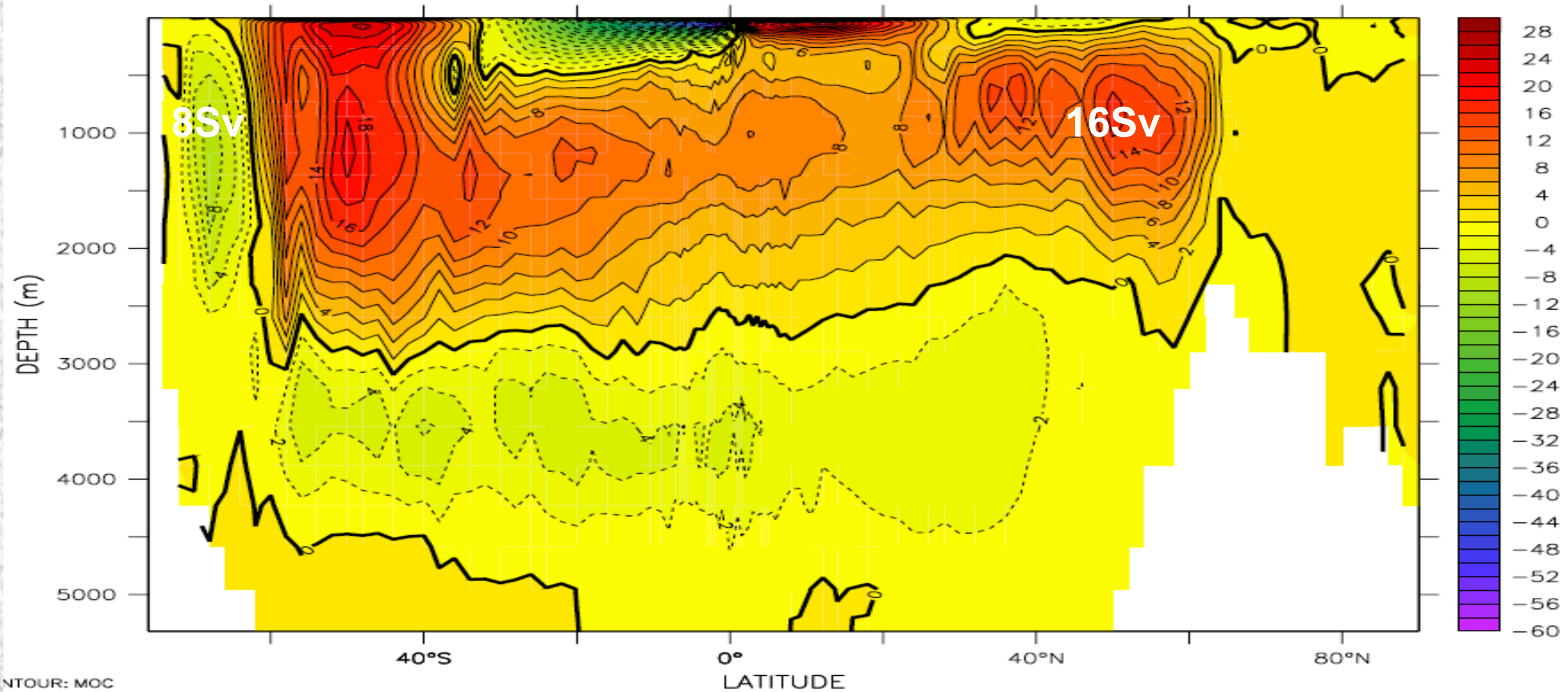
What controls ocean stratification and overturning circulation?

Atlantic ocean thermal structure

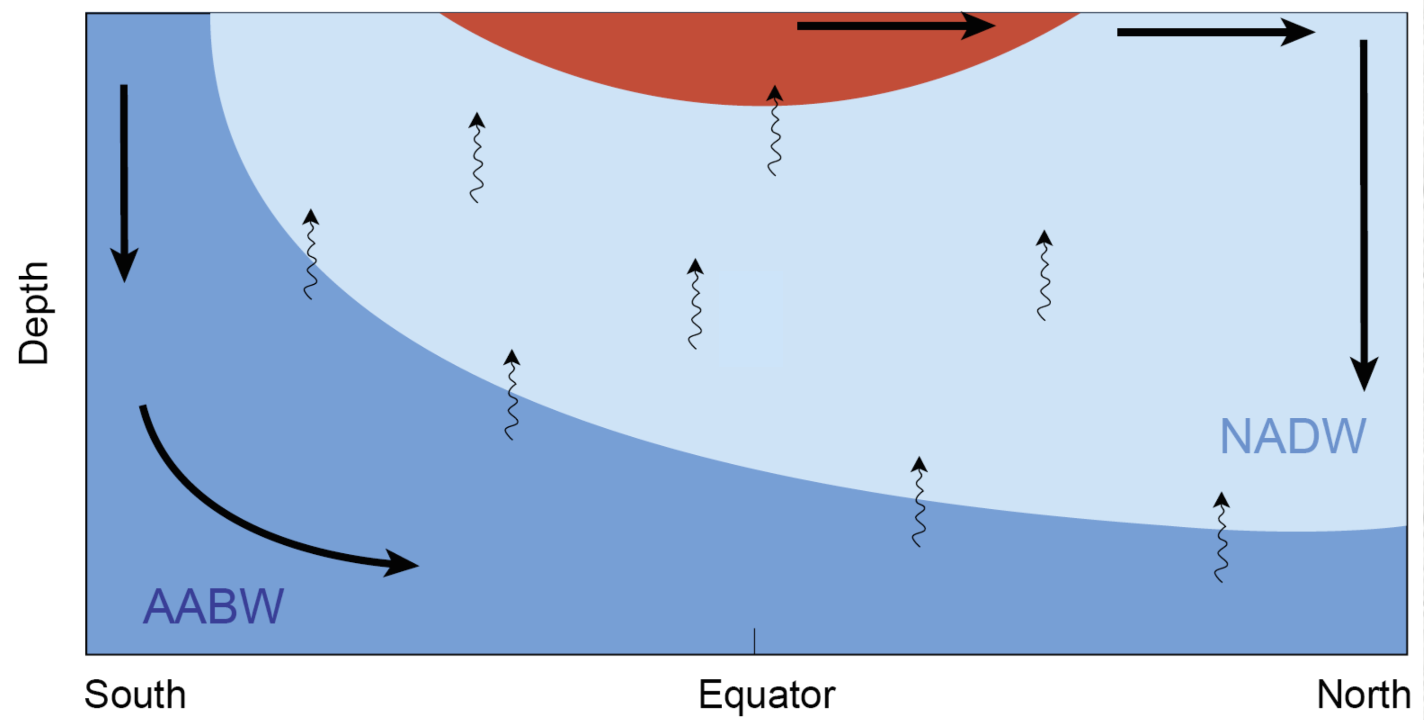


What controls ocean stratification and overturning circulation?

Meridional overturning circulation (MOC), GFDL model



Diffusive theories for the ocean thermal structure and deep ocean circulation

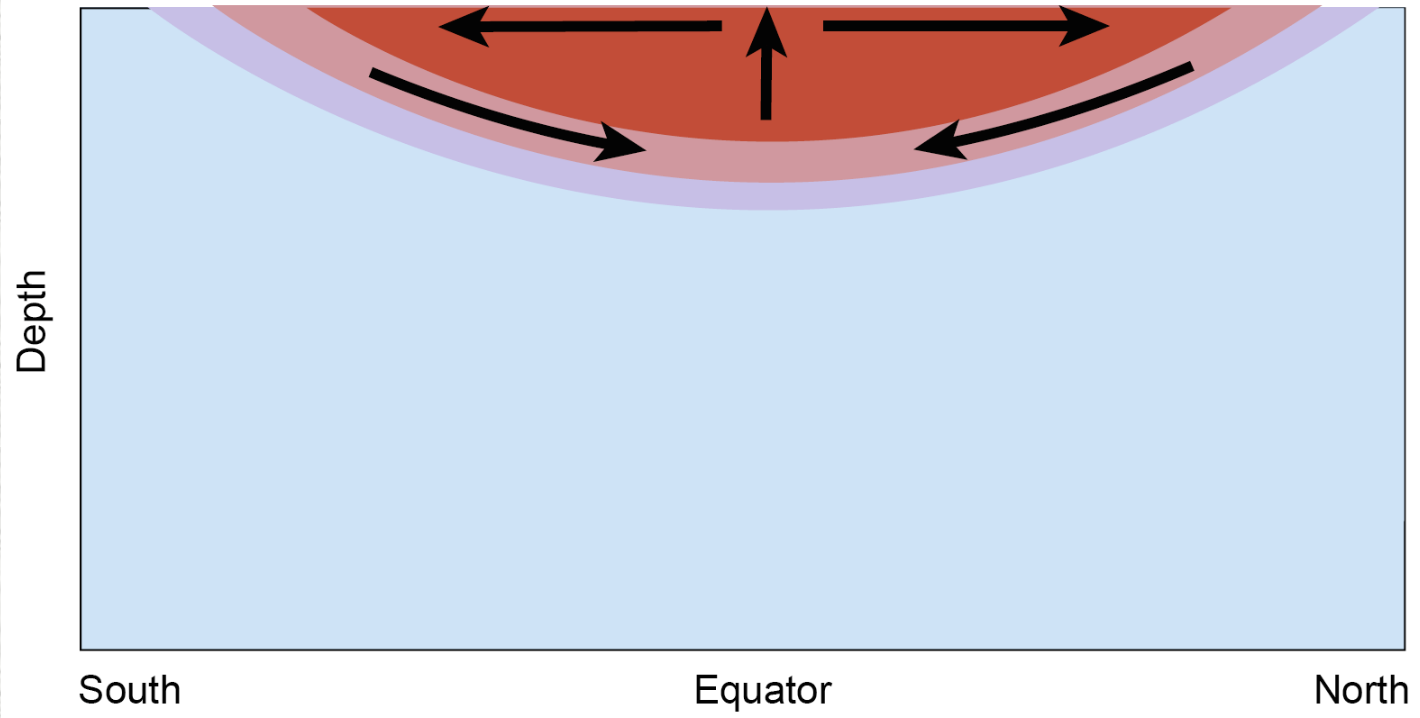


$$w\frac{\partial \rho}{\partial z} = k_d \frac{\partial^2 \rho}{\partial z^2}$$

k_d - diapycnal (vertical) diffusivity

Stommel 1958, Stommel and Arons 1960, Veronis 1976, others: ***Downward diffusion of heat is balanced by a broad upwelling of cold water***

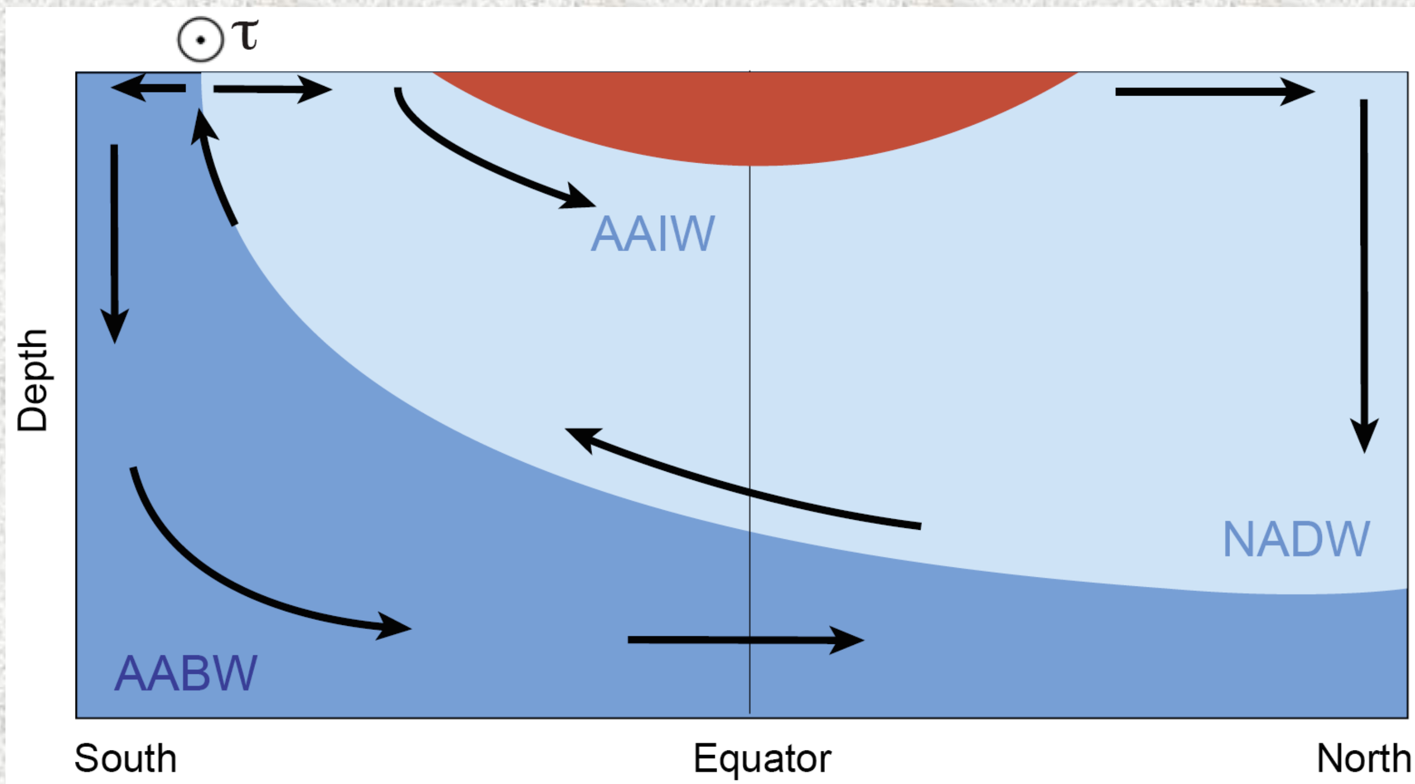
Adiabatic theories



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

Ekman pumping pushes water down along isopycnal surfaces.
Luyten, Pedlosky, and Stommel 1983: "The ventilated thermocline"; Iselin (1939), Welander (1959), Huang (1986)

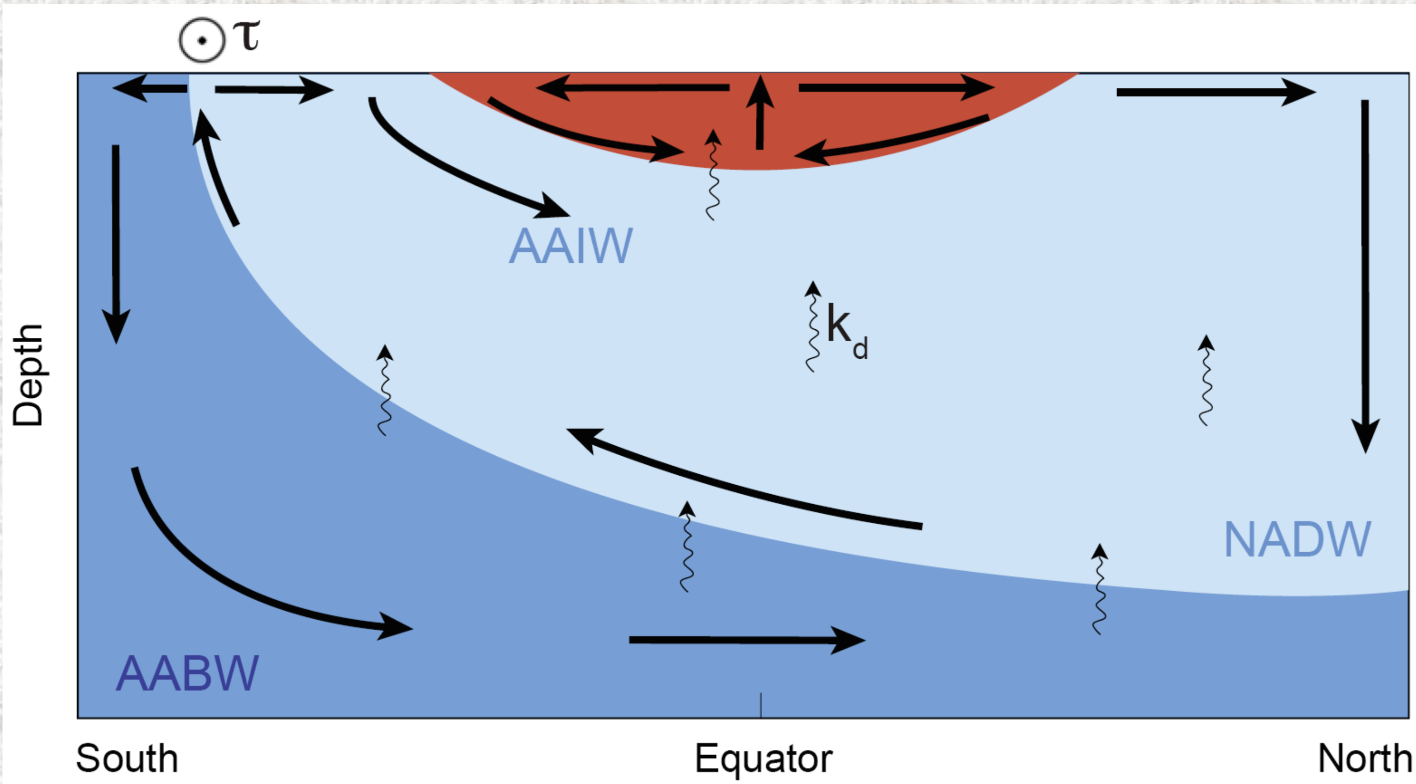
The role of the wind-driven upwelling in the Southern Ocean



τ – zonal wind stress in the Southern ocean

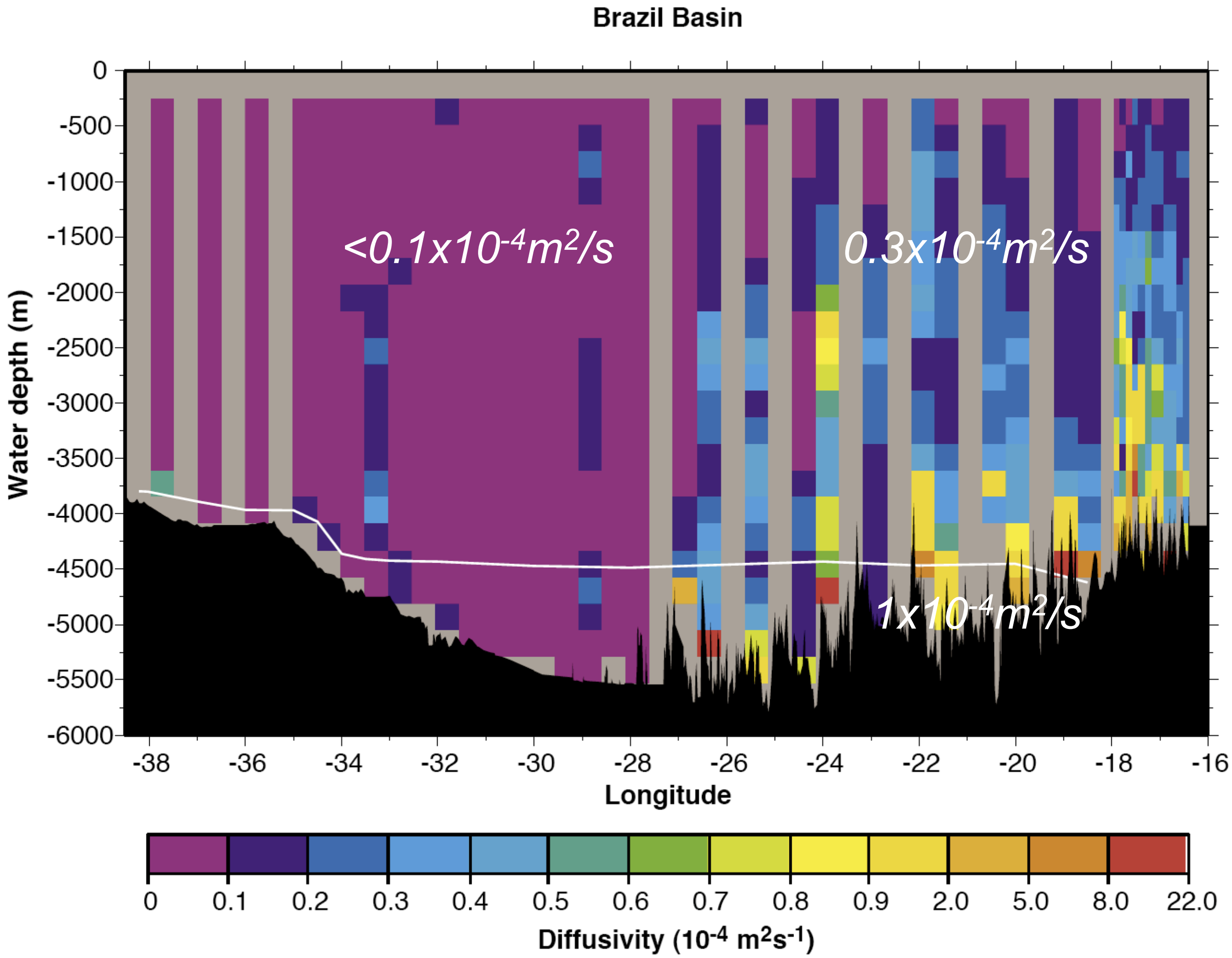
(e.g. Toggweiler and Samuels 1995, 1998, Gnanadesikan 1999)

A full schematic picture of the MOC



(e.g. Barreiro, Fedorov and co-authors 2008)

k_d - *diapycnal (vertical) diffusivity; very non-uniform*
$0.1 x 10^{-4} m^2/s$ in the upper and deep ocean to
$20 x 10^{-4} m^2/s$ in patches near bottom topography



Polzin et al 1997 (also Ledwell et al 1993)

Questions:

How important is diffusion (diabatic mixing) in the ocean interior for the ocean circulation and thermal structure?

What is the first-order solution for the ocean circulation and thermal structure? Is it diffusive or adiabatic?

What controls the ocean overturning circulation and stratification in an ocean with zero interior diffusion?

Approach:

Setting up a Lagrangian ocean model with a fully adiabatic interior flow

Navier-Stokes equations used in traditional GCMs

$$u_t = -\nabla \cdot (u\,\mathbf{u}) + v\left(f + \frac{u \tan\phi}{a}\right) - \left(\frac{1}{a\,\rho_\circ \cdot \cos\phi}\right) p_\lambda + (\kappa_m u_z)_z + F^u$$

$$v_t = -\nabla \cdot (v\,\mathbf{u}) - u\left(f + \frac{u \tan\phi}{a}\right) - \left(\frac{1}{a\,\rho_\circ}\right) p_\phi + (\kappa_m v_z)_z + F^v$$

$$w_z = -\nabla_h \cdot \mathbf{u}_h$$ *Vertical velocity*

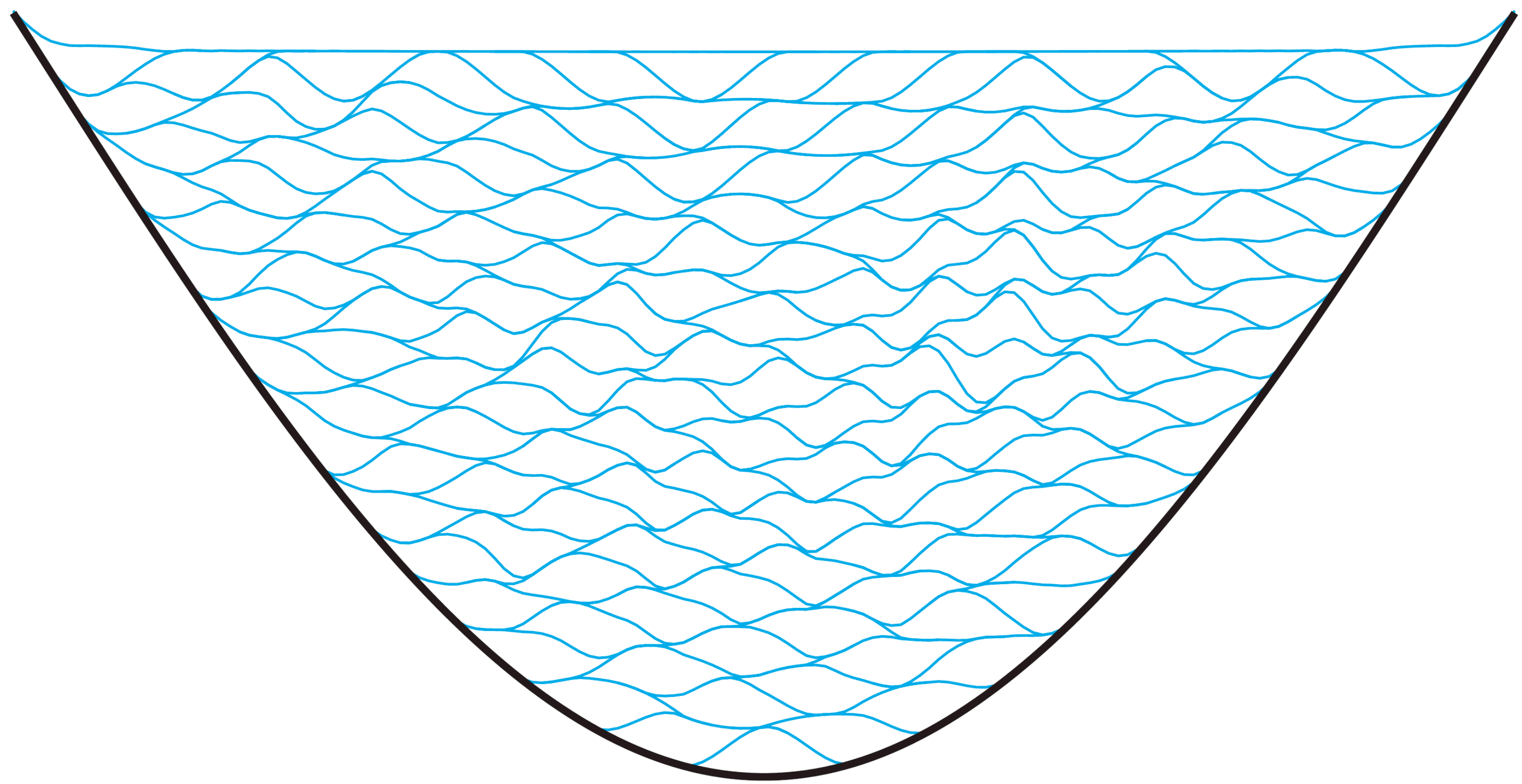
$$p_z = -\rho\, g$$ *Pressure*

$$\theta_t = -\nabla \cdot [\mathbf{u}\,\theta + \mathbf{F}(\theta)]$$ *Potential temperature*

$$s_t = -\nabla \cdot [\mathbf{u}\, s + \mathbf{F}(s)]$$ *Salinity*

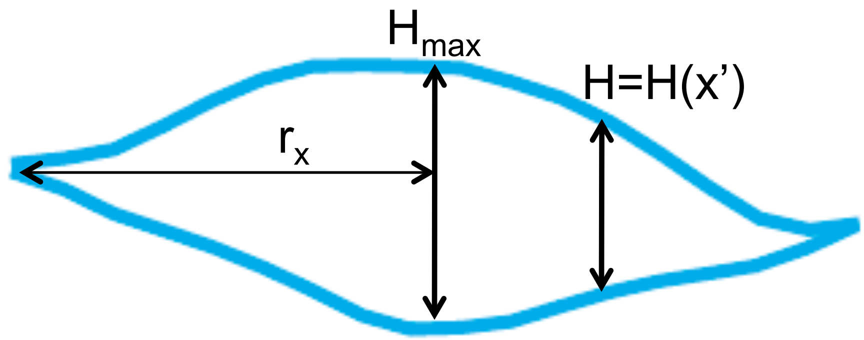
$$\rho = \rho(\theta, s, z).$$ *Density*

Lagrangian Ocean Model (Haertel et al 2010)



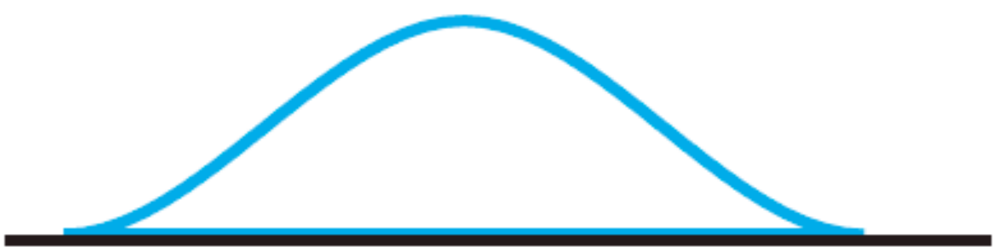
A stack of water parcels:
The stacking is done from the bottom up. Parcels conserve mass but change volume and shape. Pressure is hydrostatic. Dense parcels slide underneath less dense parcels.

Parcel's shape



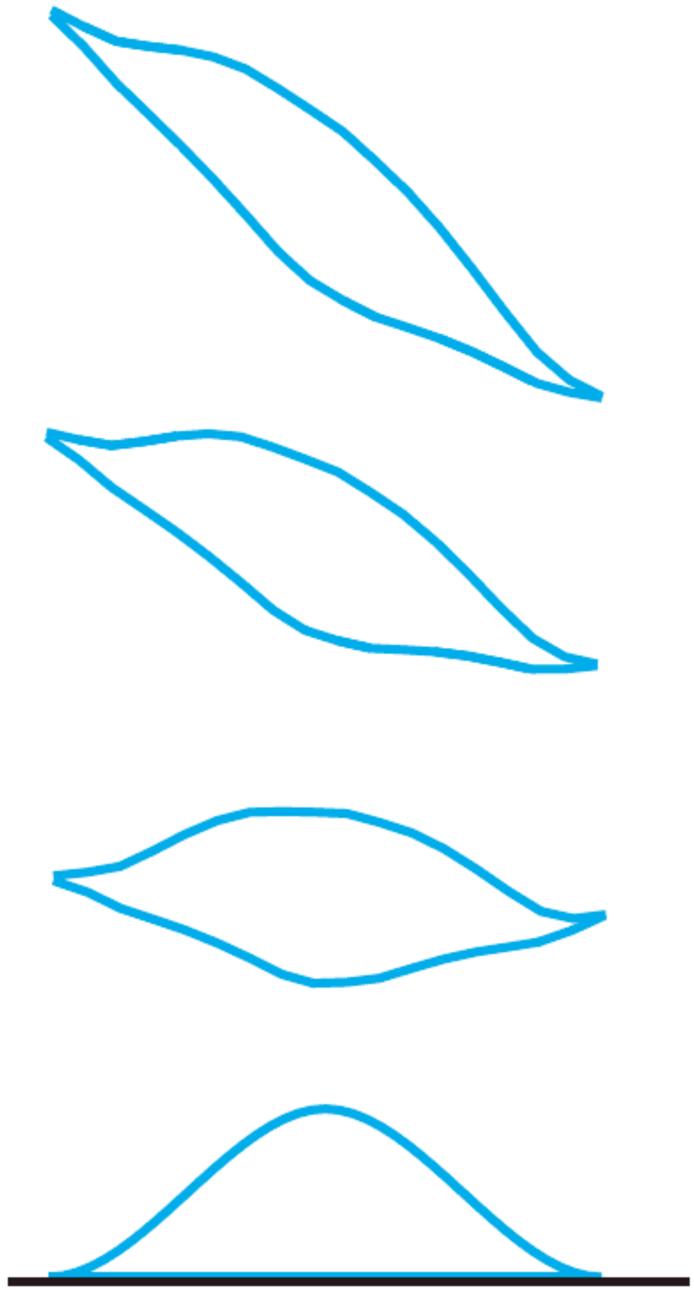
$$H(x', y') = H_{\max} p\left(\frac{|x'|}{r_x}\right) p\left(\frac{|y'|}{r_y}\right)$$ - parcel's thickness

p(x) - bell-shaped when set on a flat surface



$$p(x) = 1 + (2x - 3)x^2 \text{ for } x < 1$$

Parcel's shape



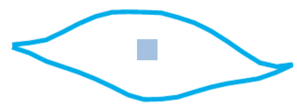
p(x) is fixed

$H_{max} = f(M,\rho)$

M – mass

ρ – density

Equations of motions



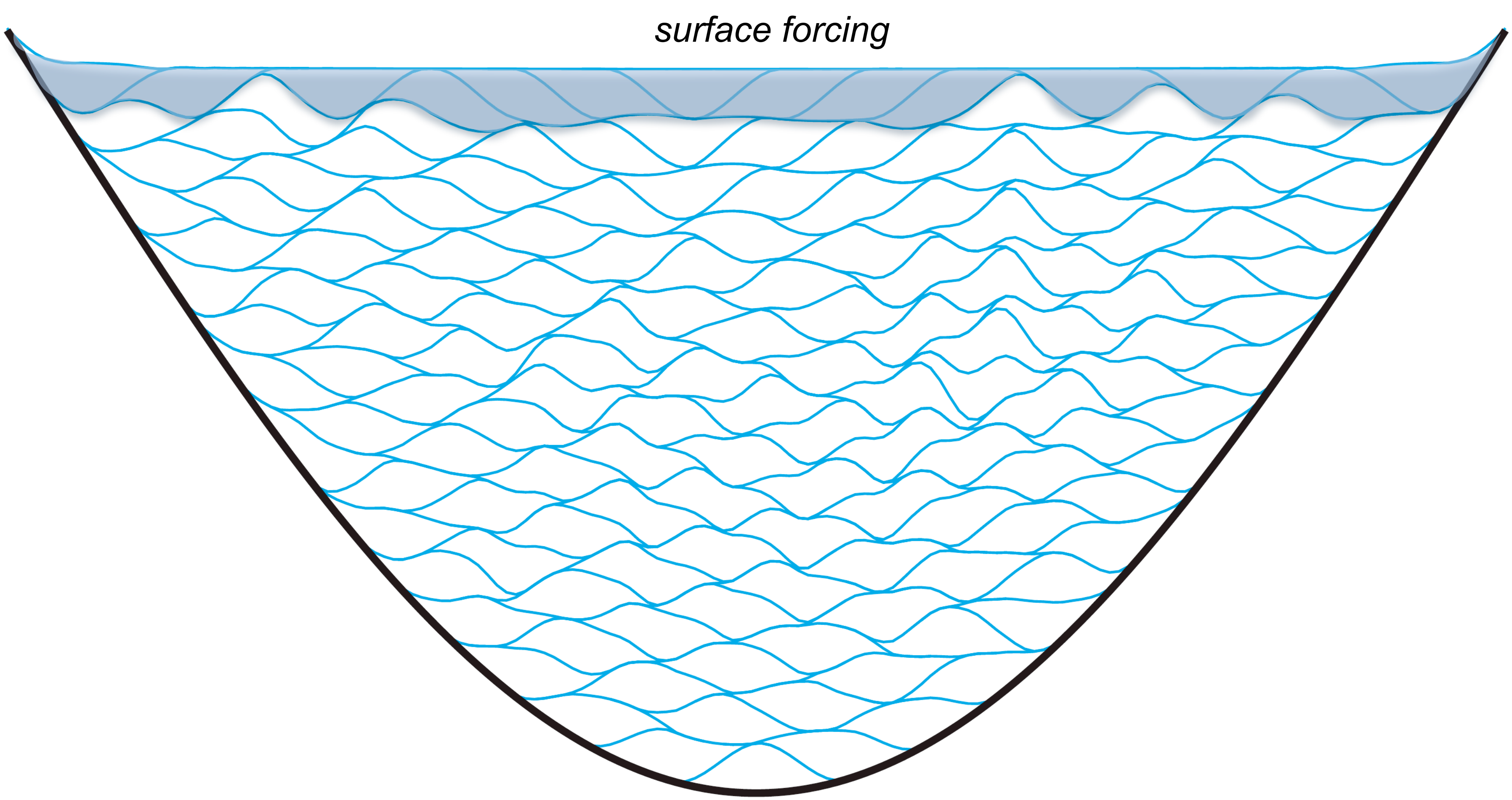
Newton's law :

$$\frac{d\boldsymbol{u}}{dt} = -f\,\boldsymbol{k} \times \boldsymbol{u} + \frac{1}{M}\int p\,ds + \nu\nabla^2\boldsymbol{u}$$

$$\frac{d\boldsymbol{x}}{dt} = \boldsymbol{u}$$

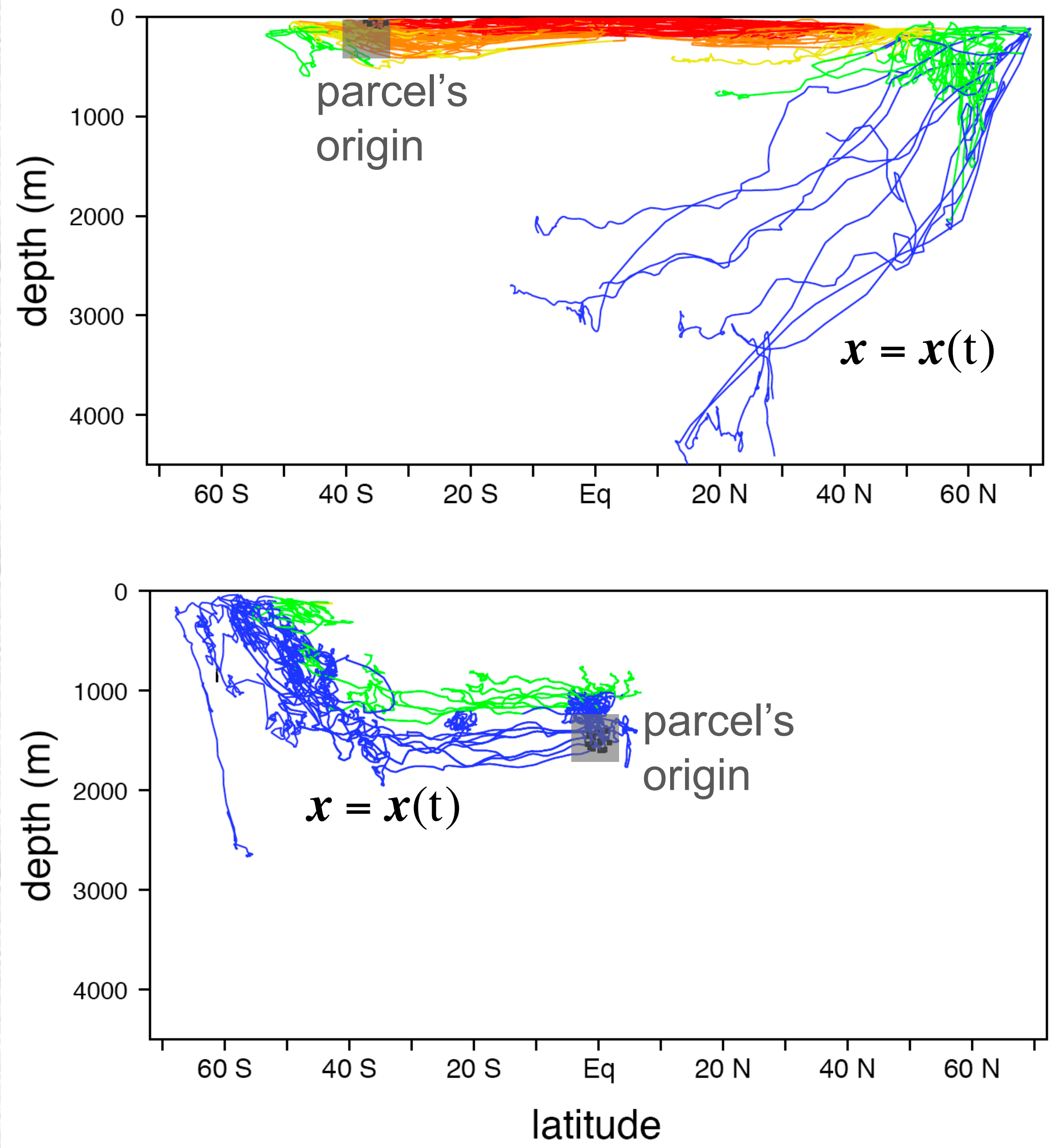
Density changes :

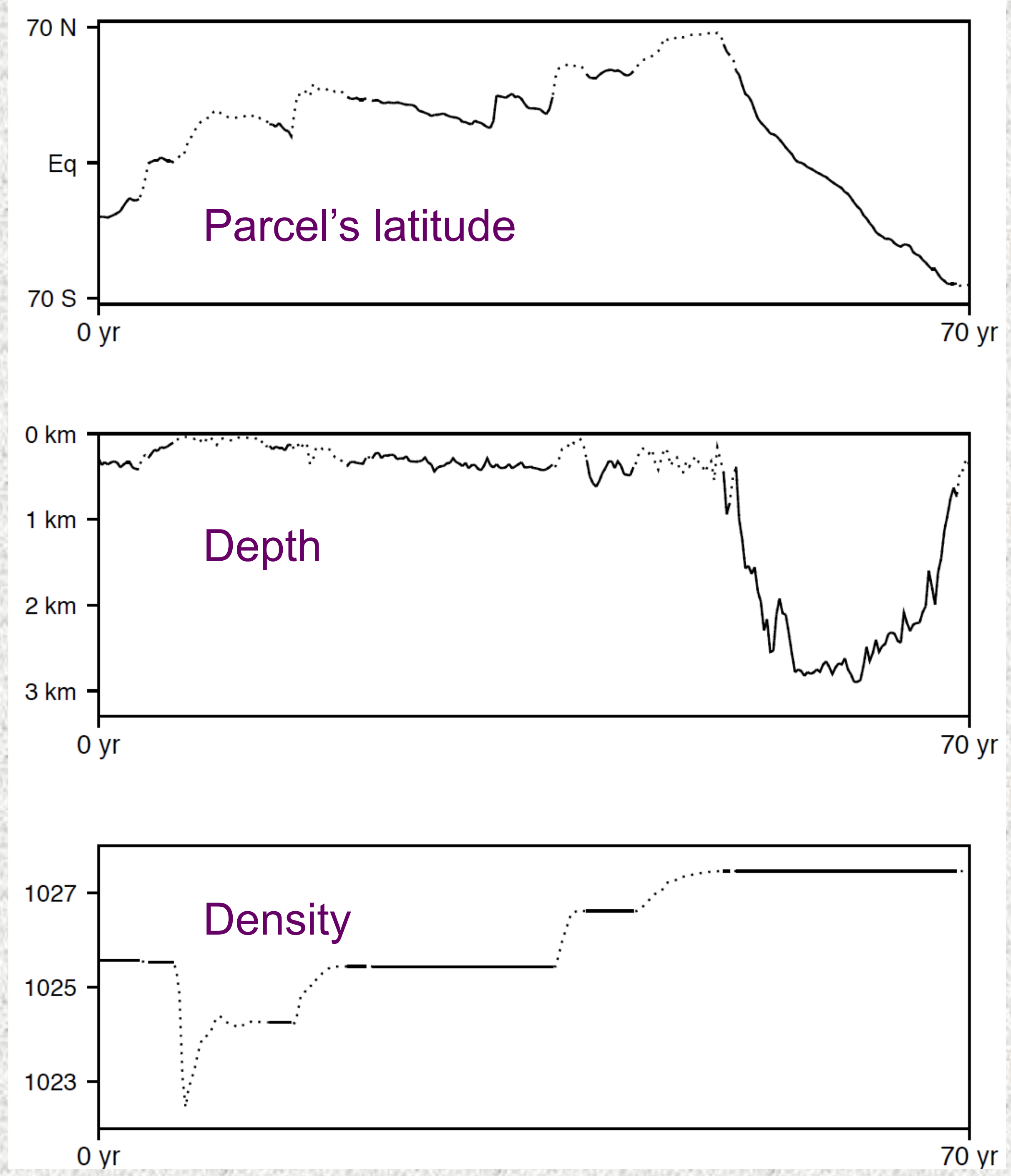
$$\frac{d\rho}{dt} = k_d\,\frac{\partial^2\rho}{\partial z^2}$$



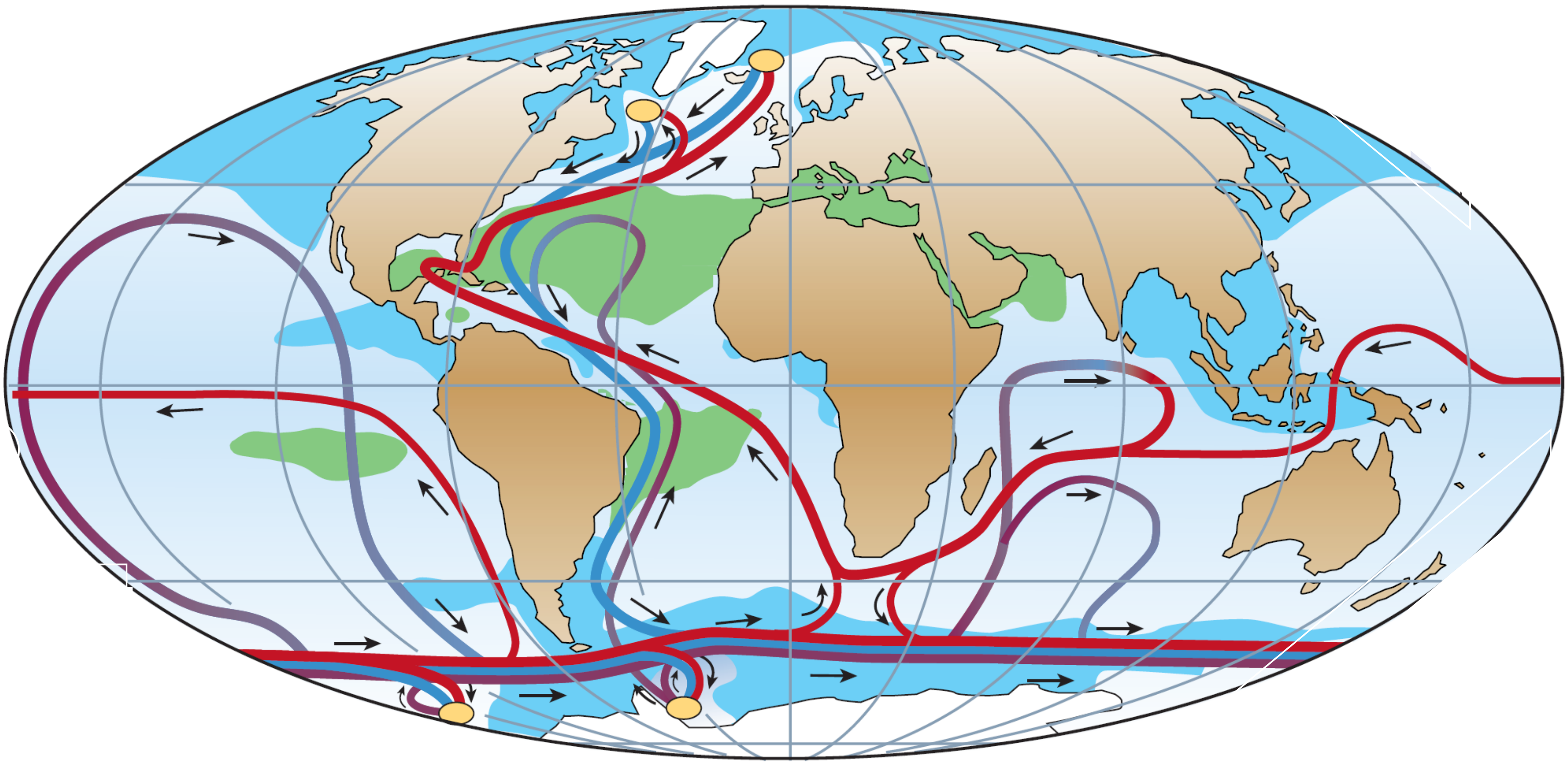
if $k_d = 0$, then below the surface layer $\frac{d\rho}{dt} = 0$

Examples of parcel's trajectories in the ocean



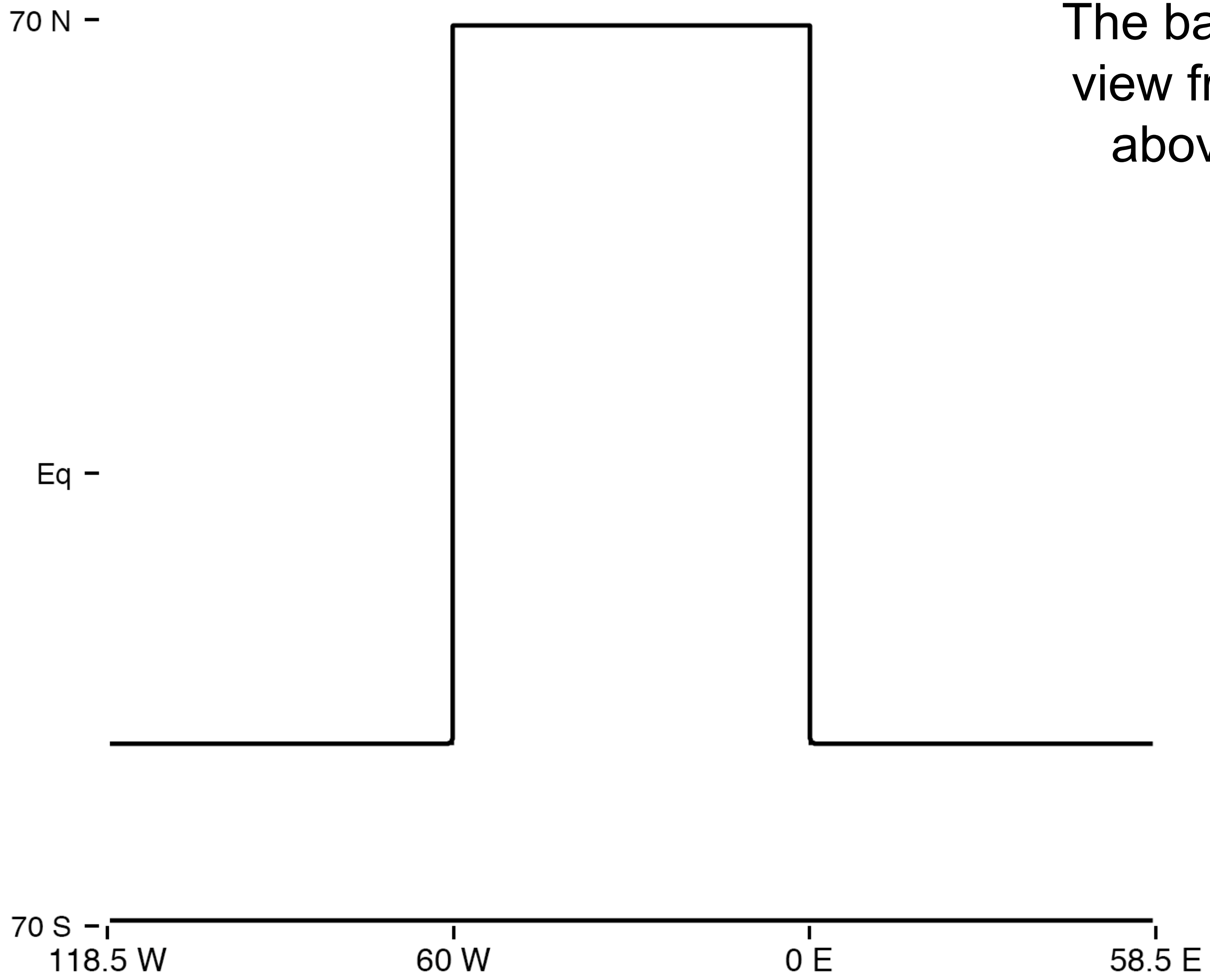


An example of a parcel's position and density with time

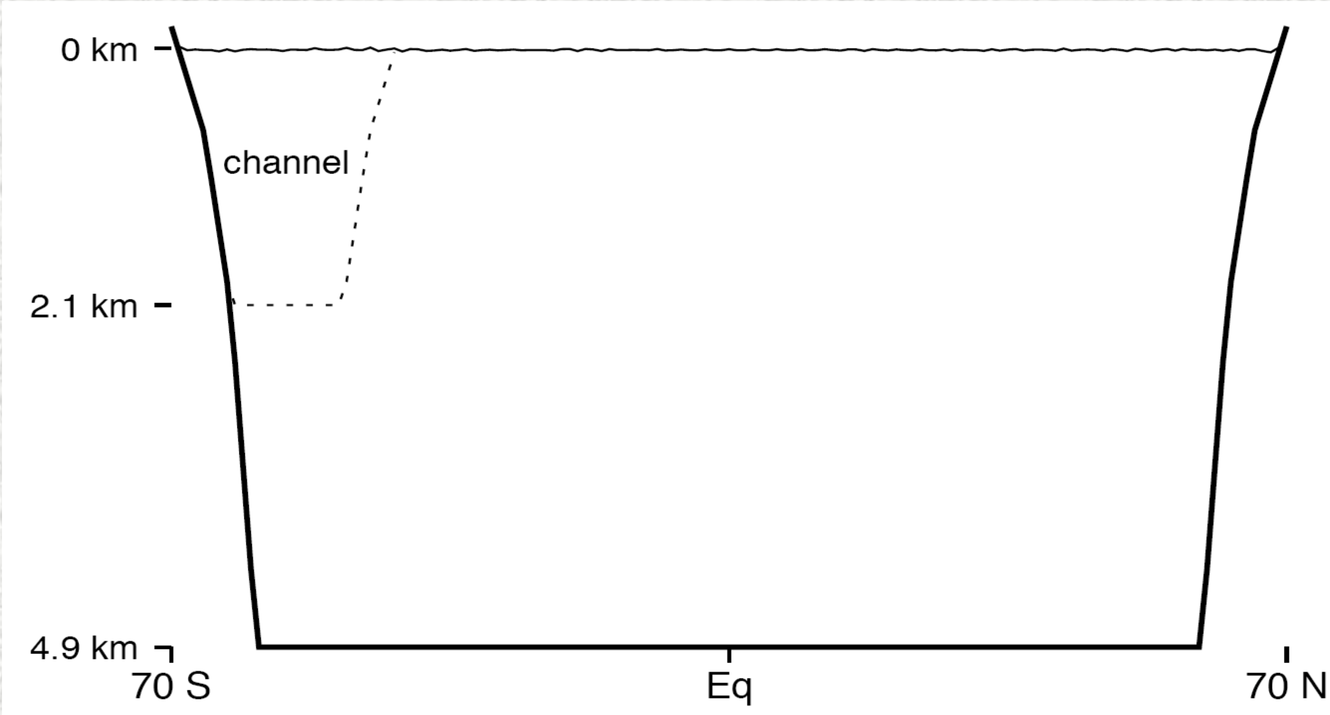


A cartoon of the global thermohaline circulation
after Rahmstorf 2003

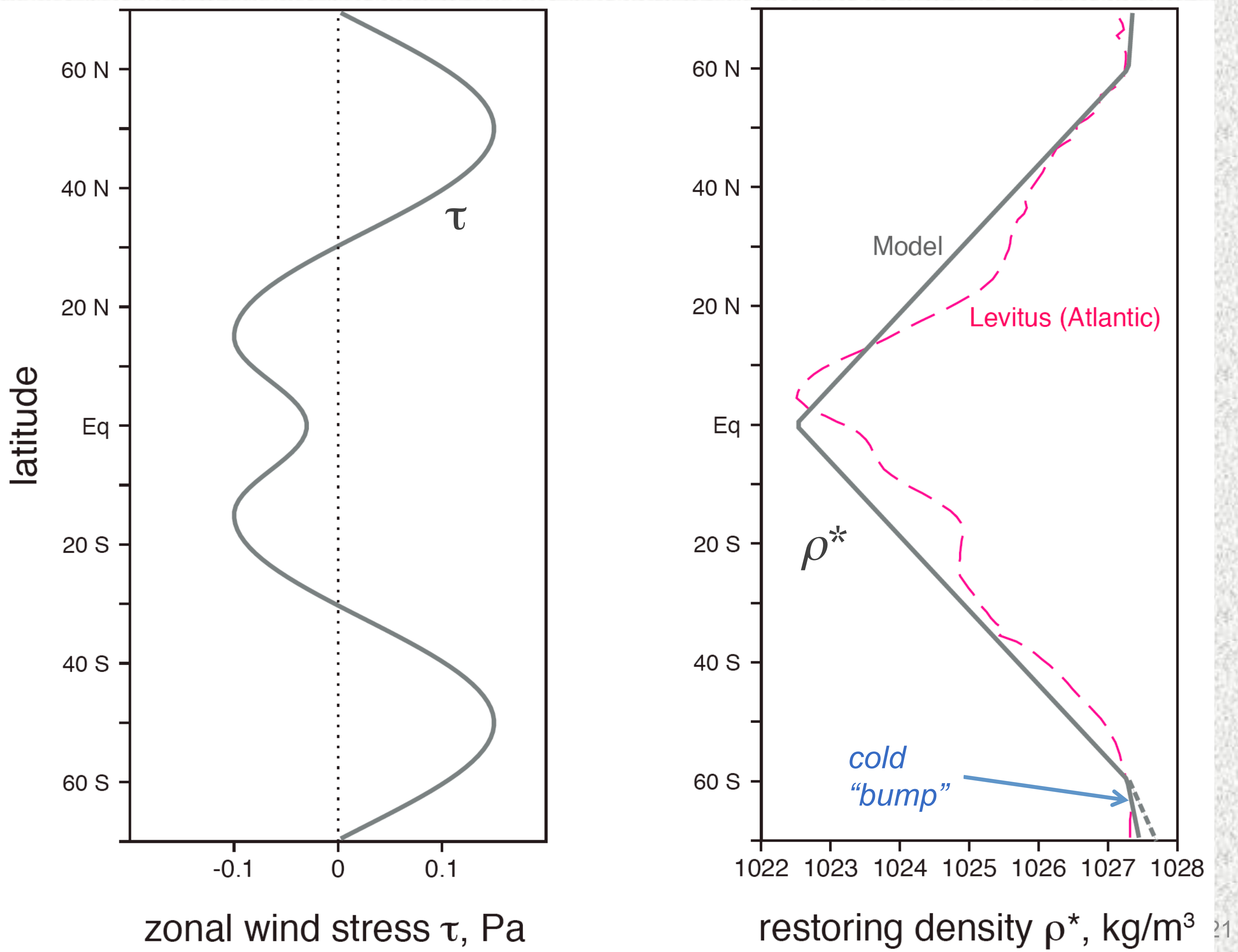
The basin, view from above

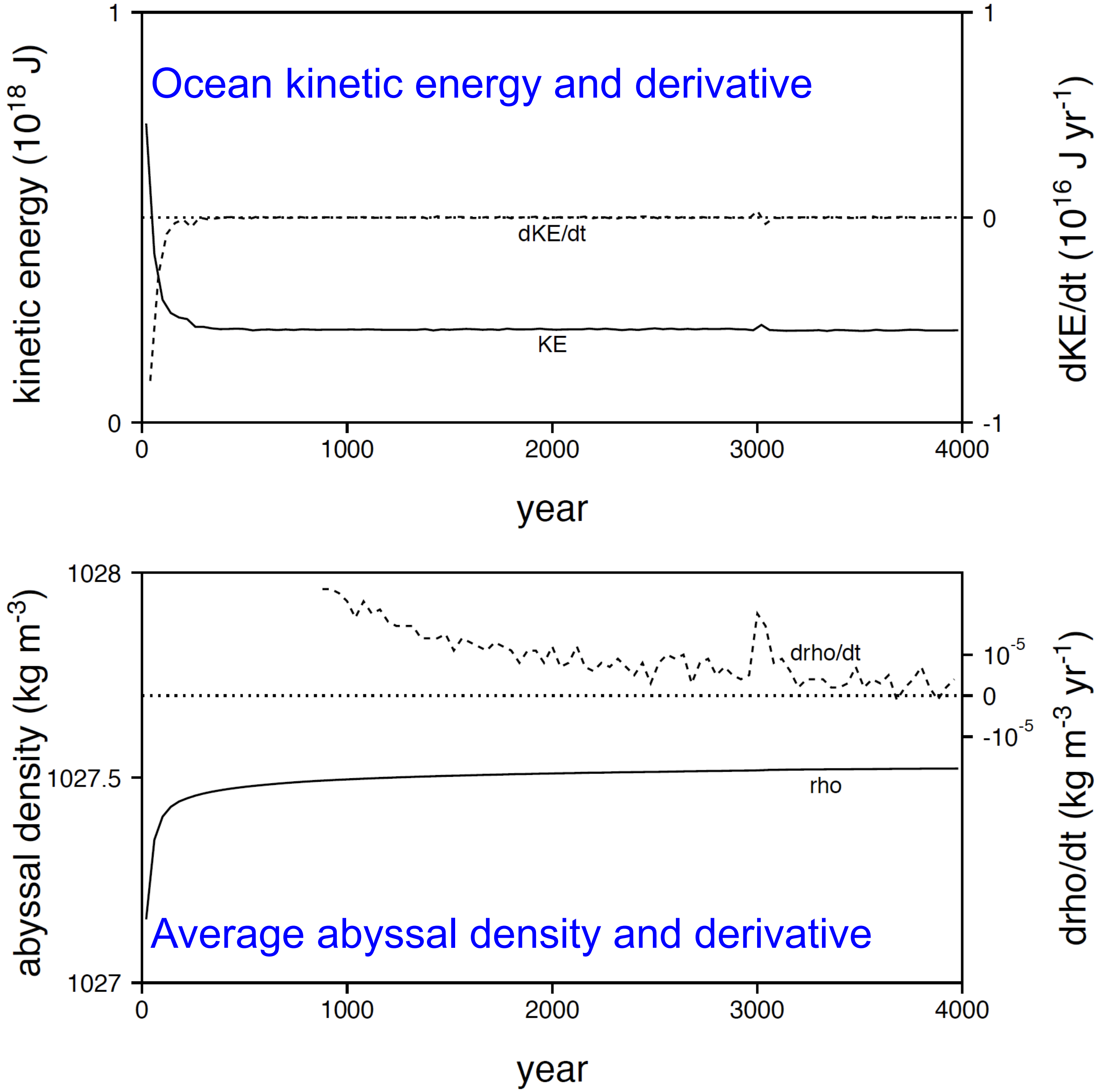


The basin, side view

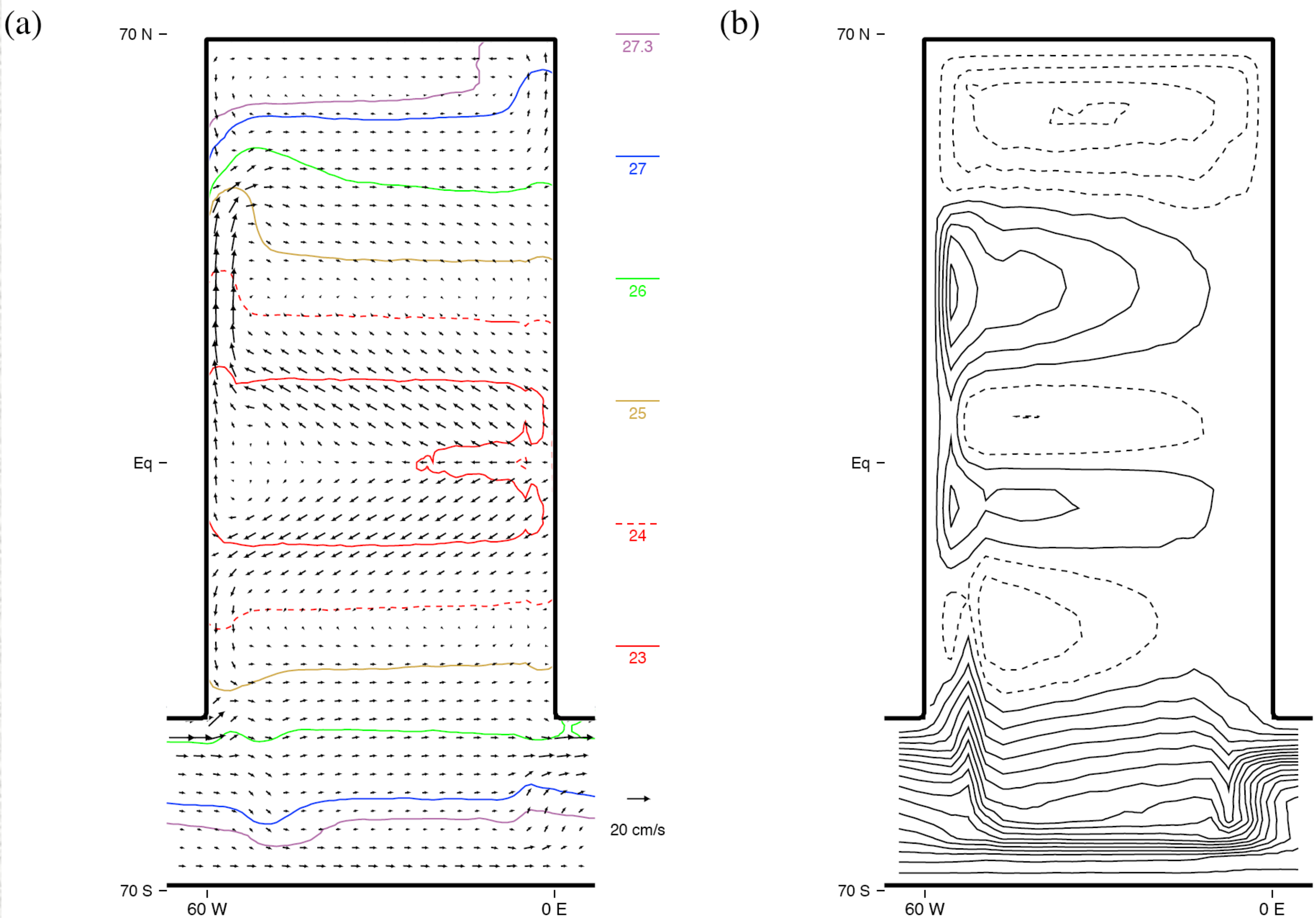


Surface forcing

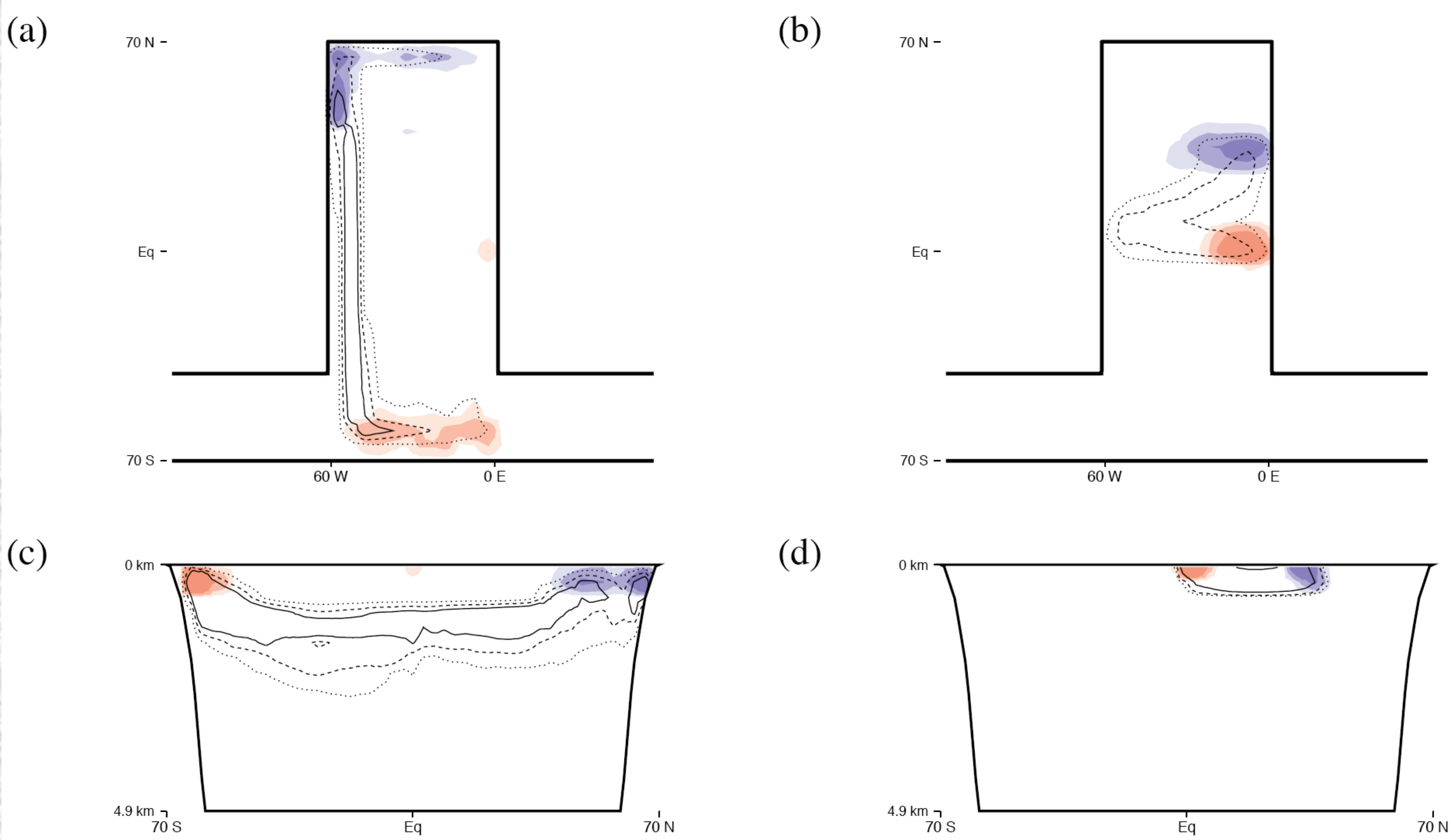



Ocean kinetic energy and derivative
kinetic energy (10^18 J)
dKE/dt (10^16 J yr^-1)
dKE/dt
KE
1
0
-1
0
1000
2000
3000
4000
year
Average abyssal density and derivative
abyssal density (kg m^-3)
drho/dt (kg m^-3 yr^-1)
drho/dt
rho
1028
1027.5
1027
10^-5
-10^-5
year

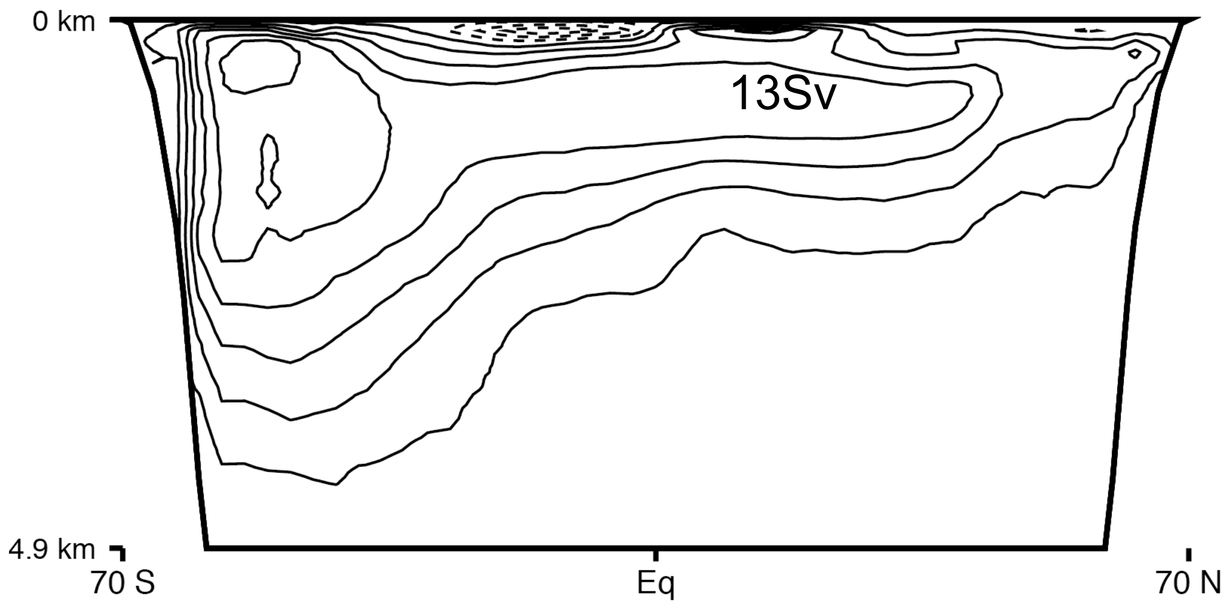
(a) Simulated surface density (kg/m³) and ocean currents
(b) Barotropic stream function, 7 Sv intervals

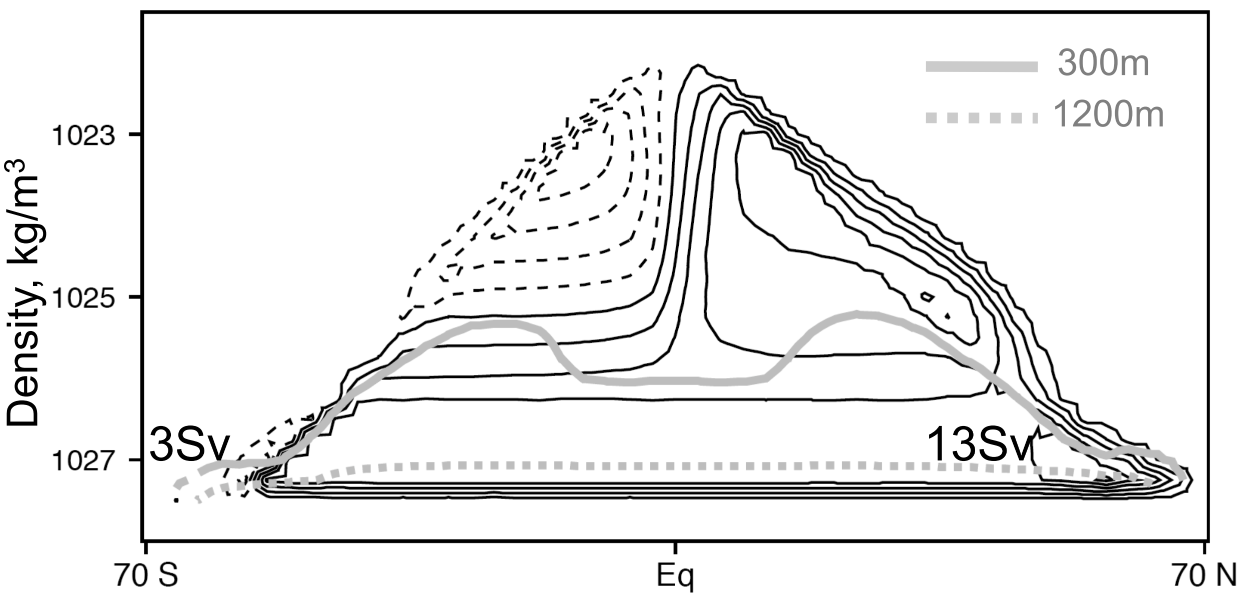


Examples of parcel routes : **Blue** – subduction, **red** – upwelling

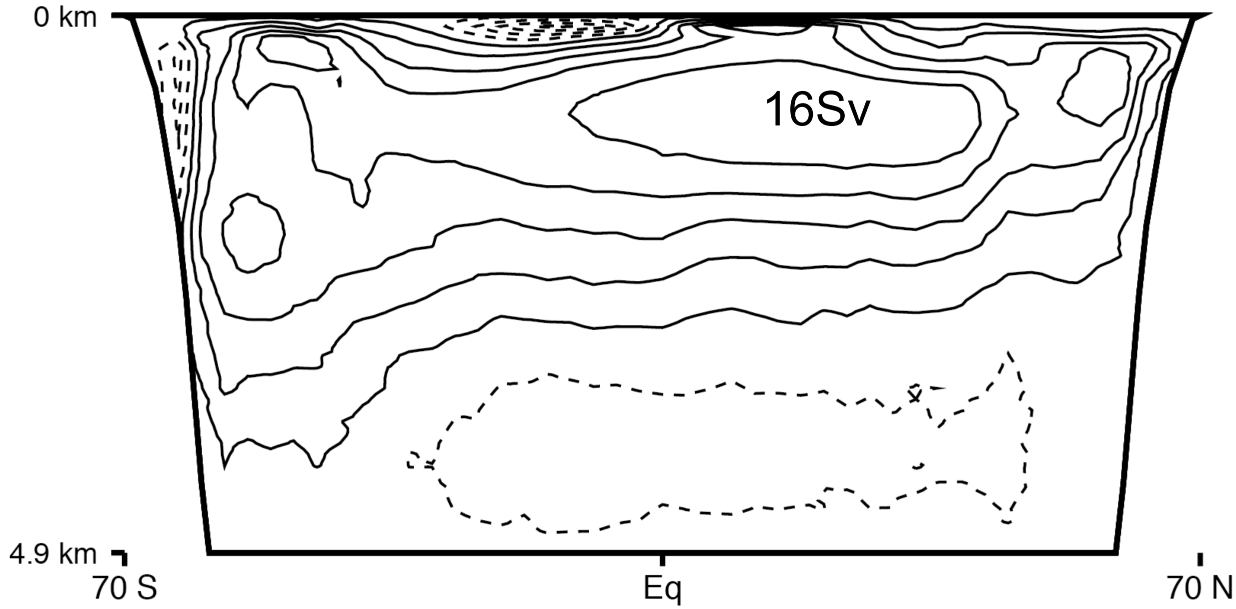


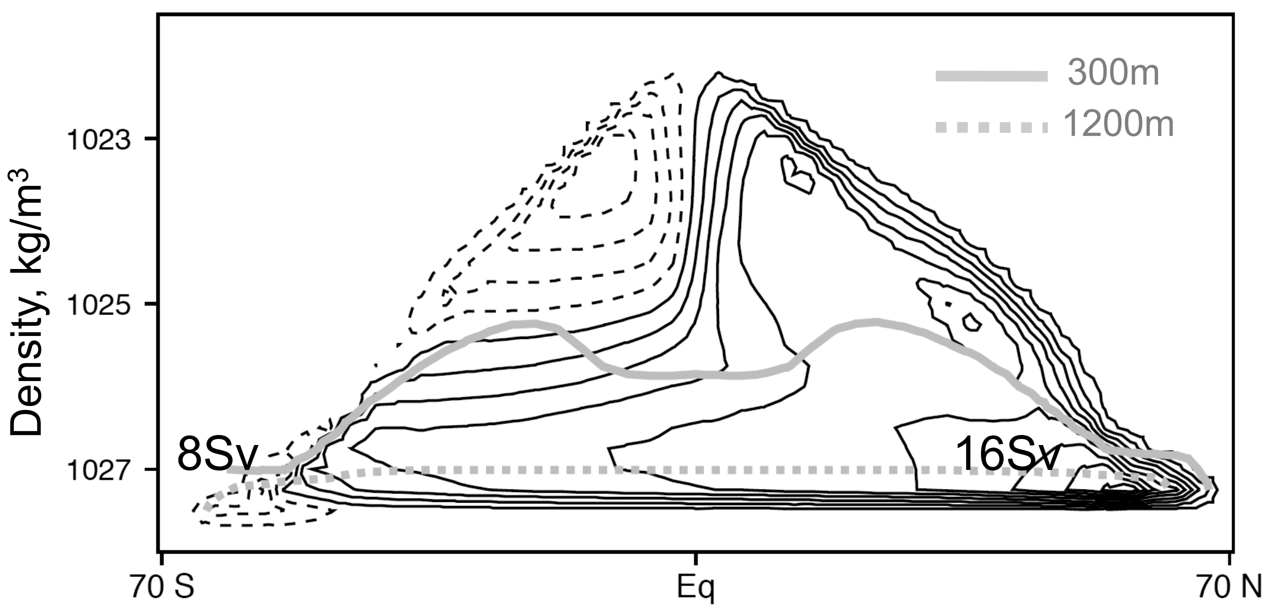
Overturning circulation: Non-diffusive

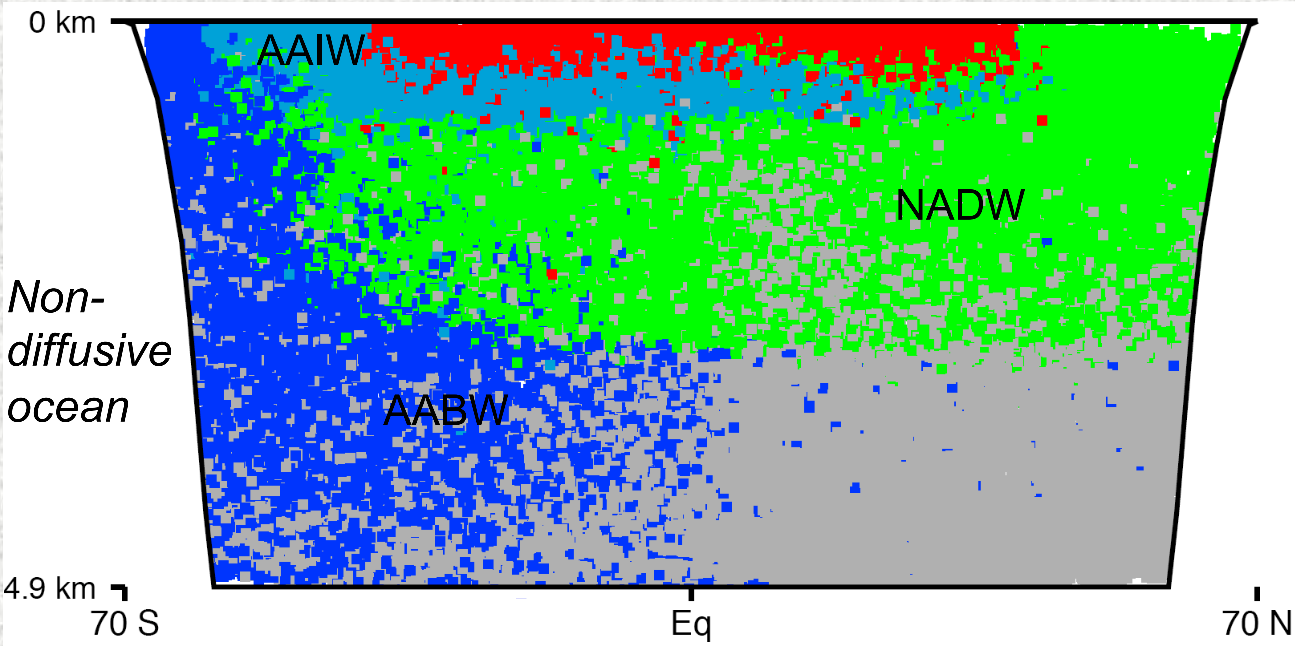




Overturning circulation: Diffusive ocean







Water Masses

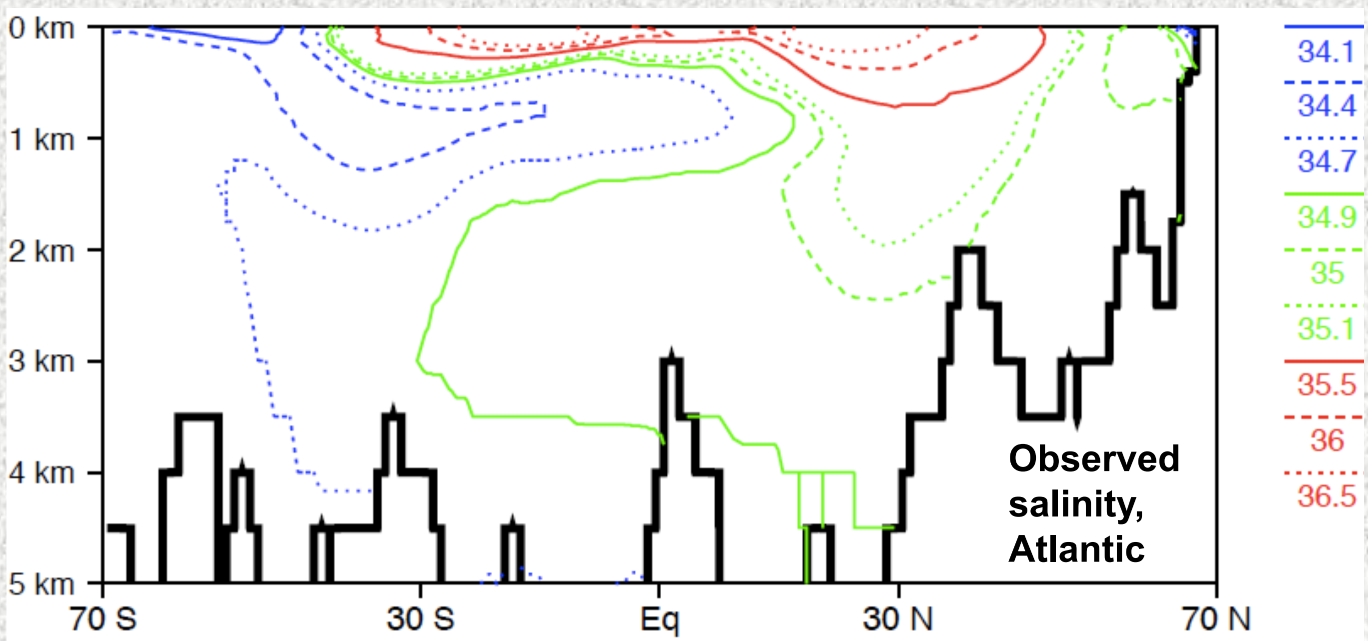
40°S - 40°N

60°S - 40°S

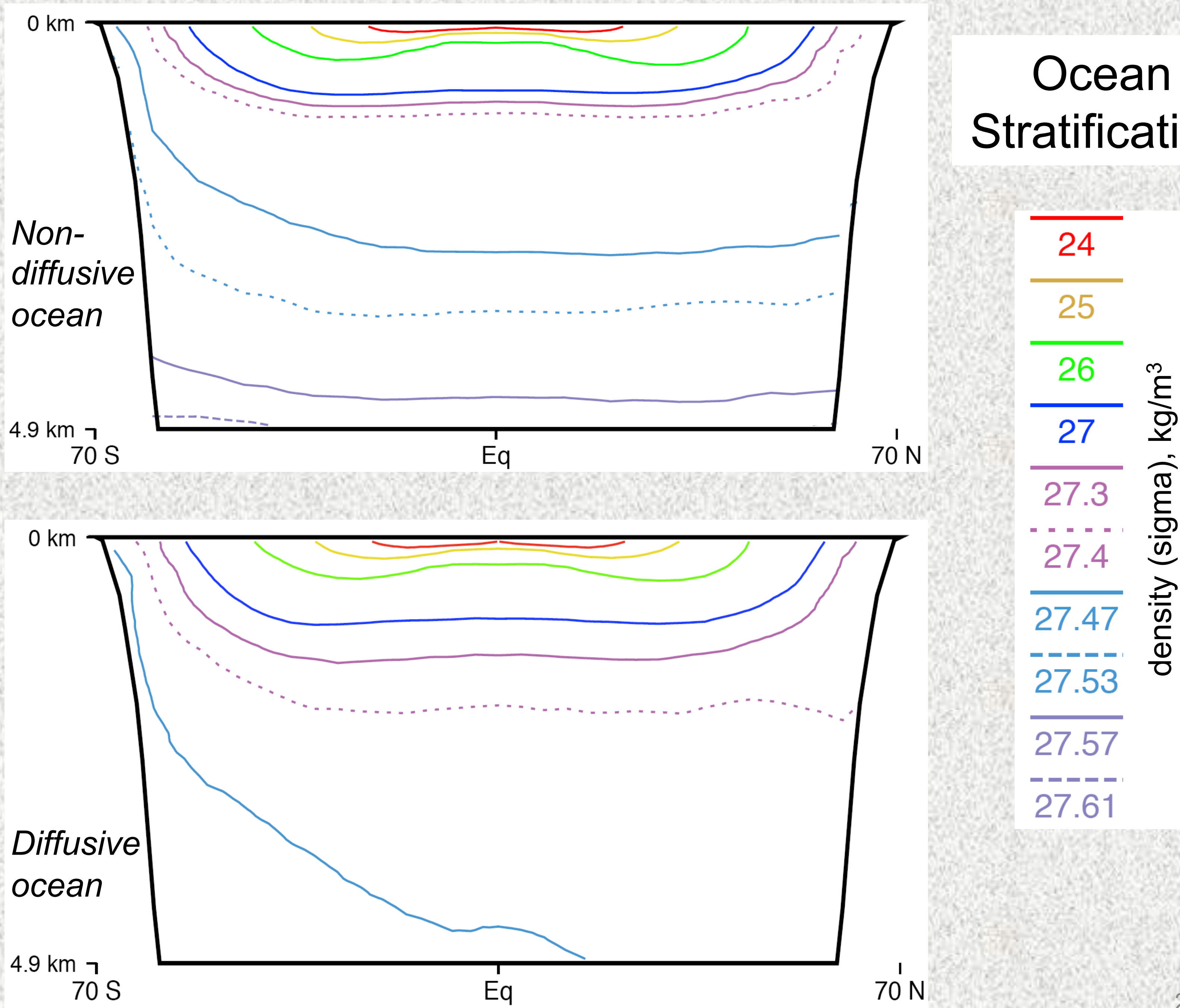
> 40°N

< 60°S

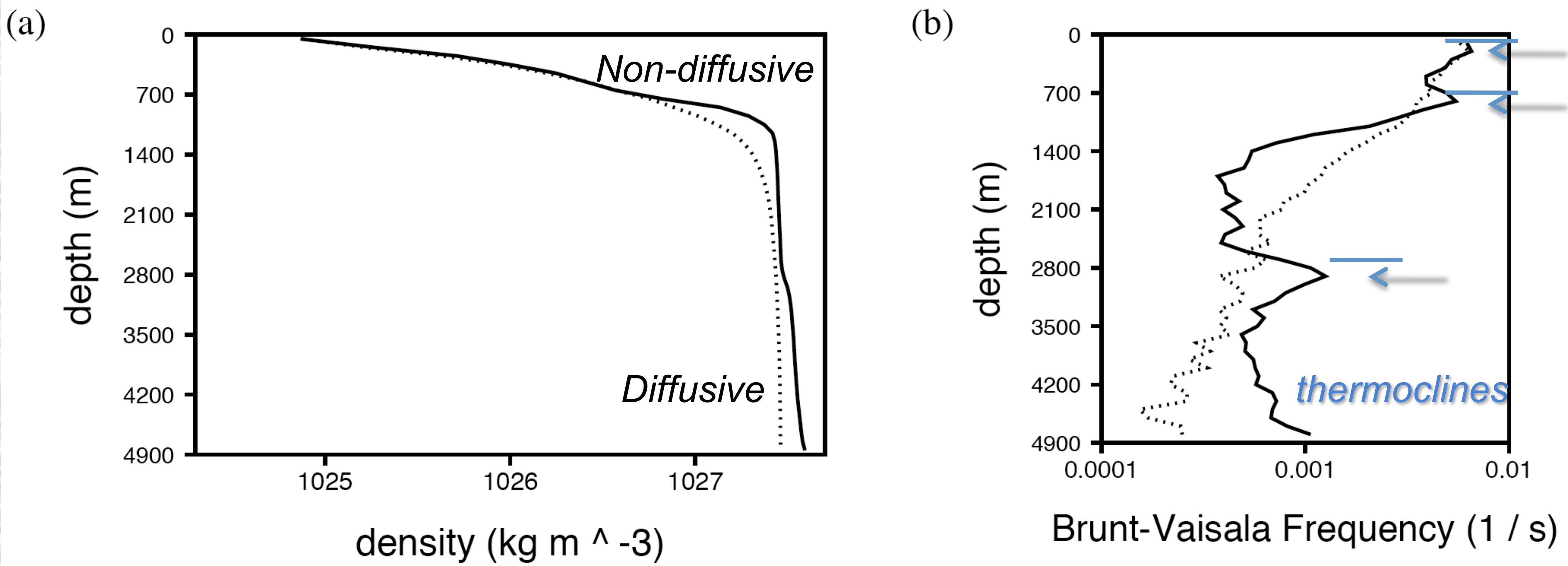
not ventilated after 700yrs



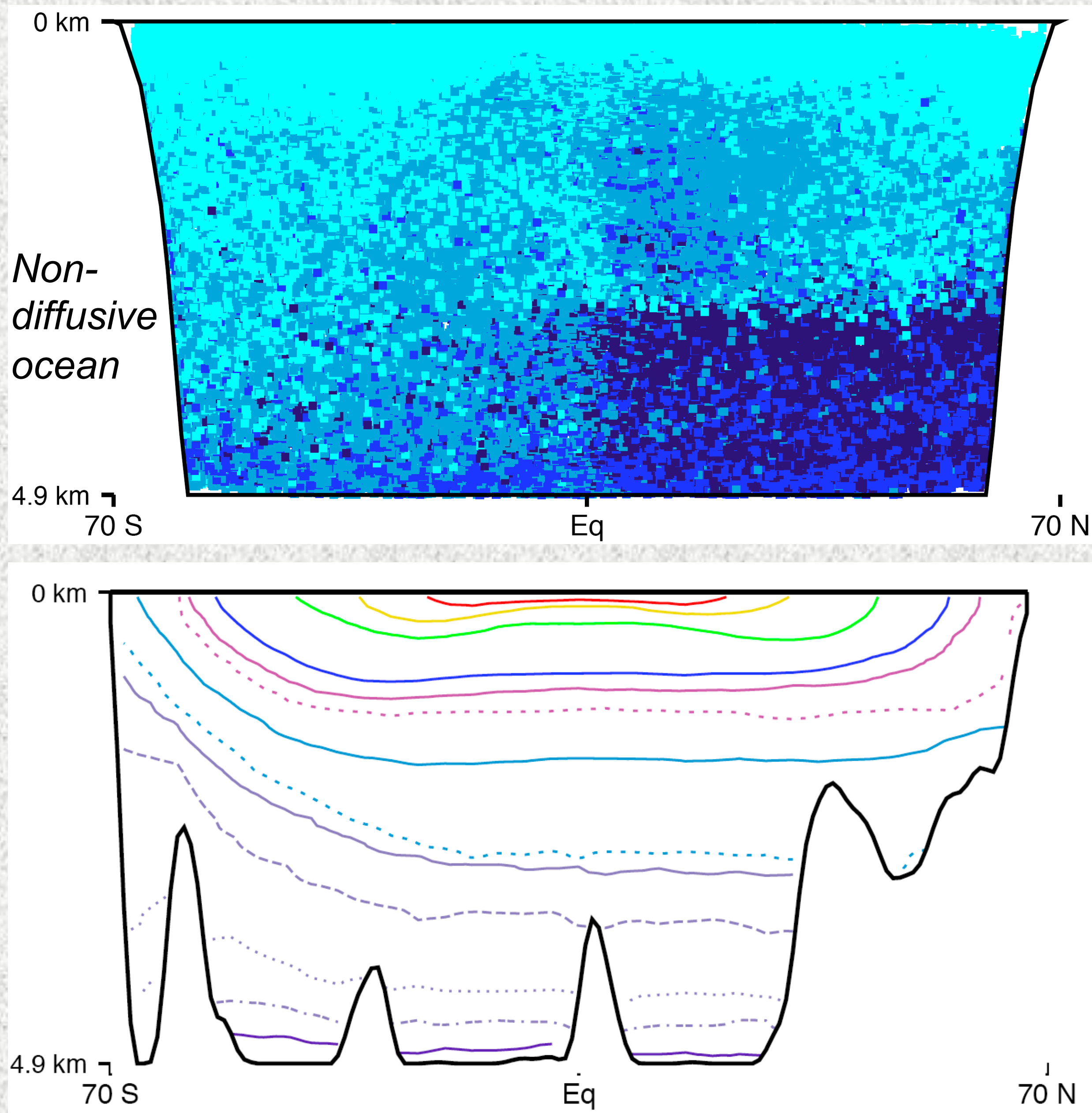
Ocean Stratification



Stratification and buoyancy frequency at 10^{o}W, 30^{o}N



Water Ventilation Age / Transit times



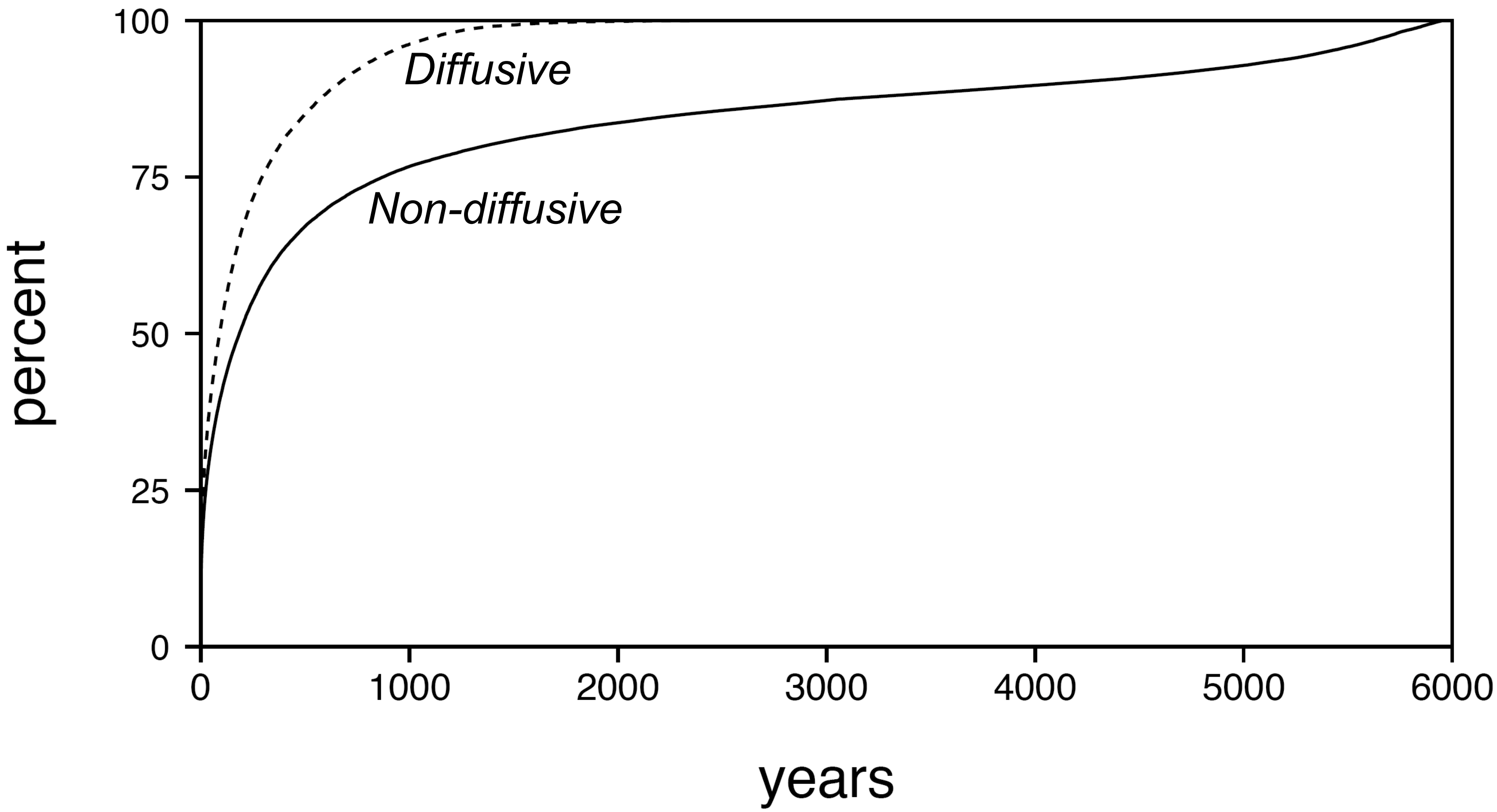
< 100 yrs

100-1000 yrs

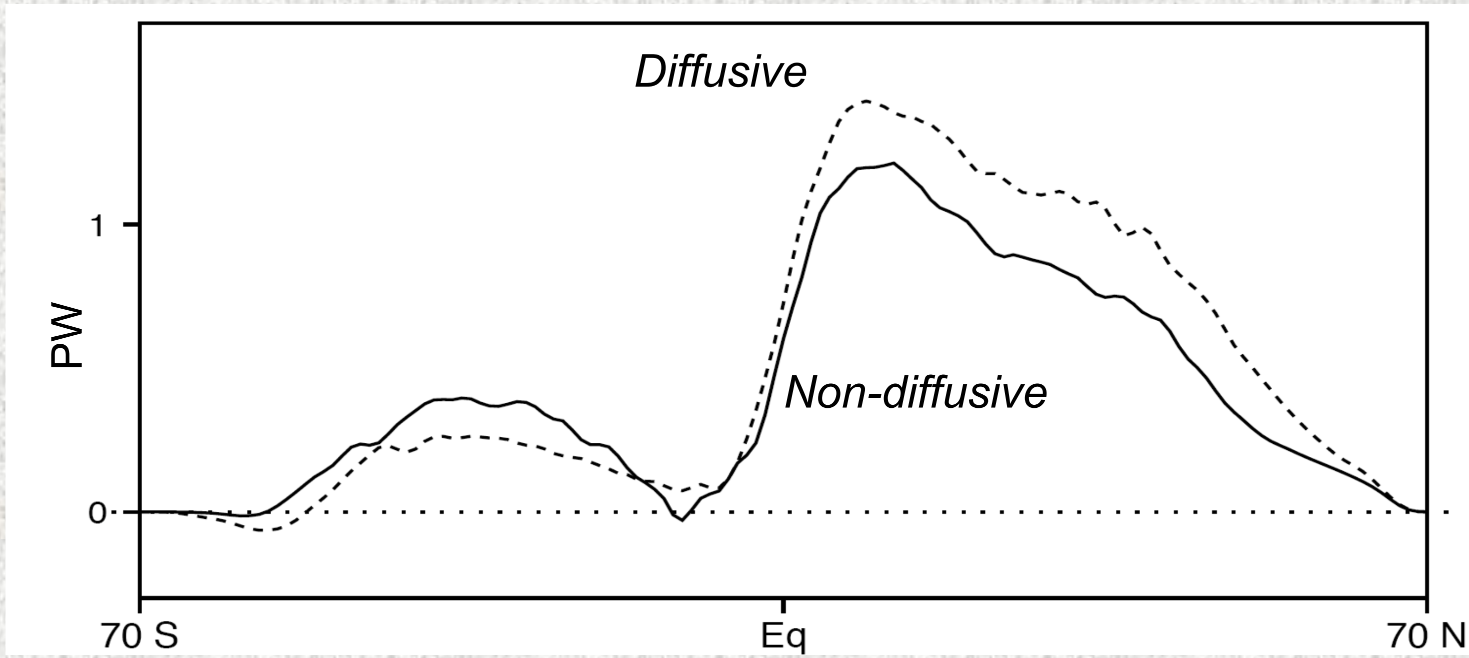
1000-3000 yrs

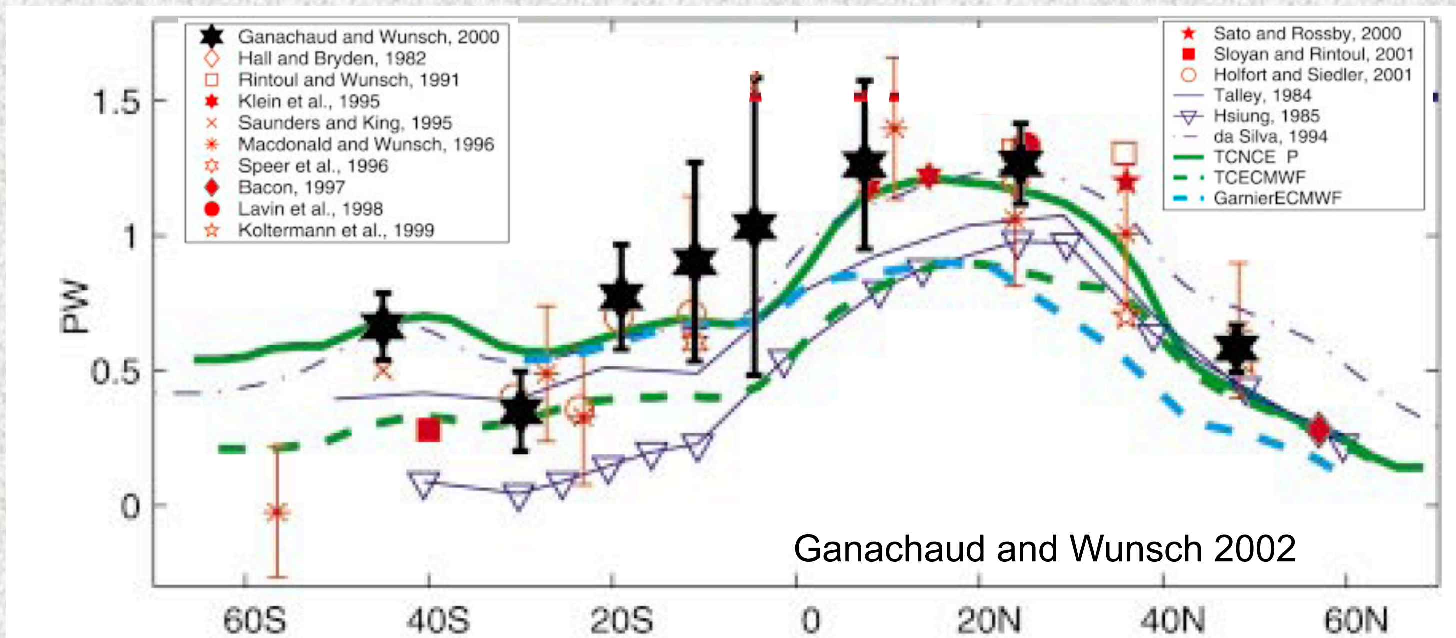
> 3000 yrs

Model cumulative distribution of water ventilation age (percentage of parcels younger than a certain “age”)

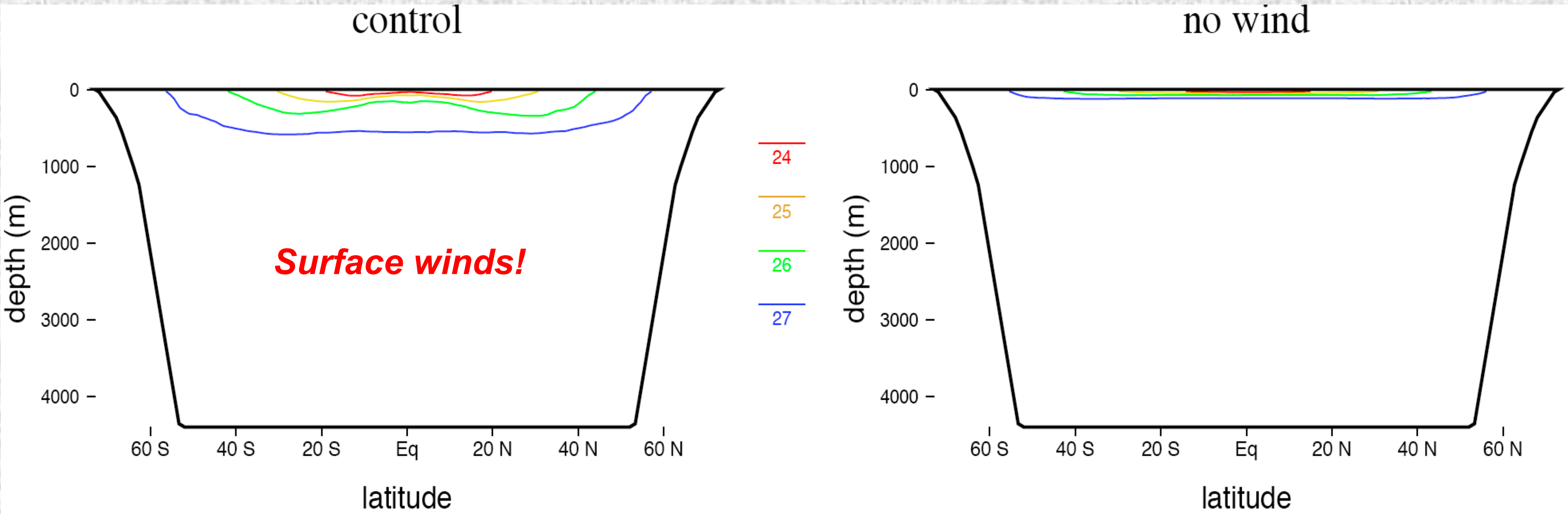


Northward heat transport in the Atlantic





Non-diffusive ocean: what controls stratification and the AMOC intensity?



Non-diffusive ocean:
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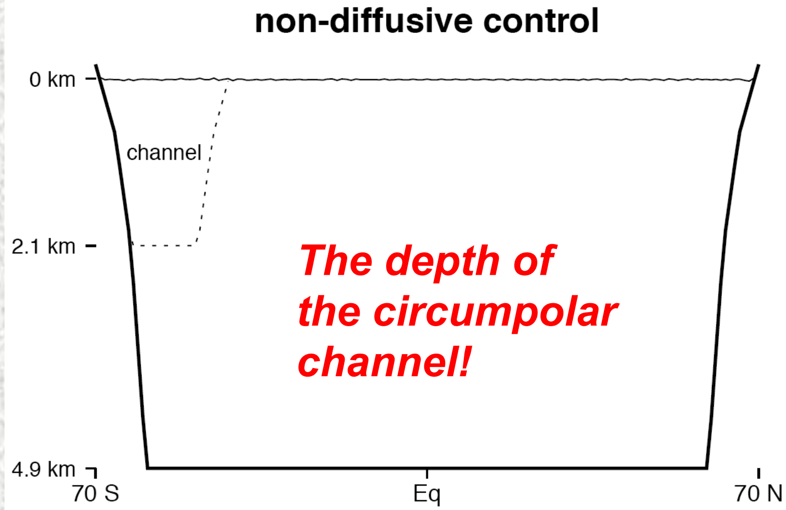
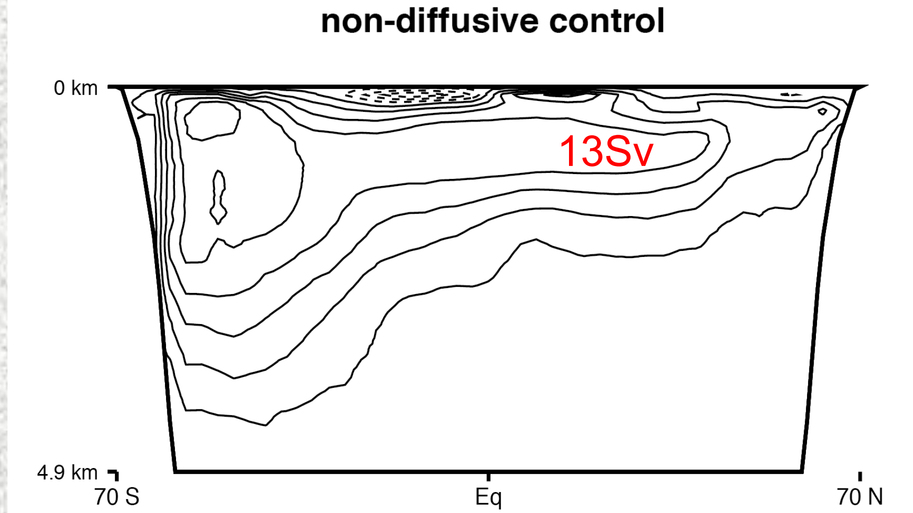
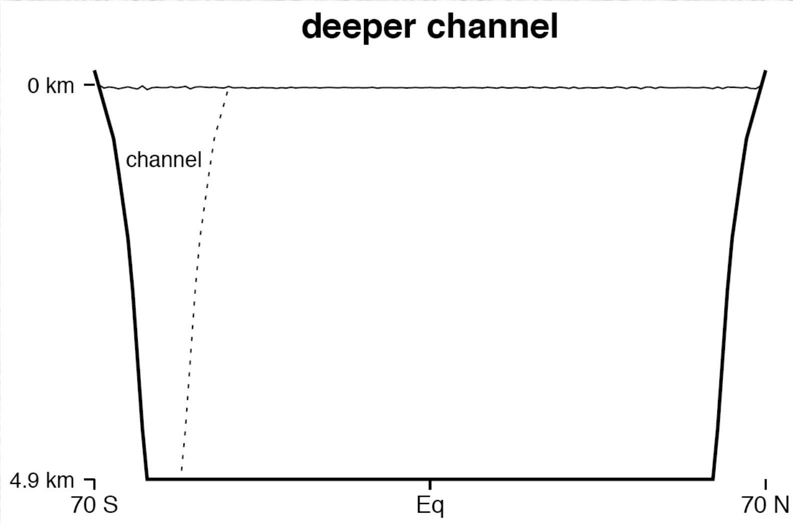
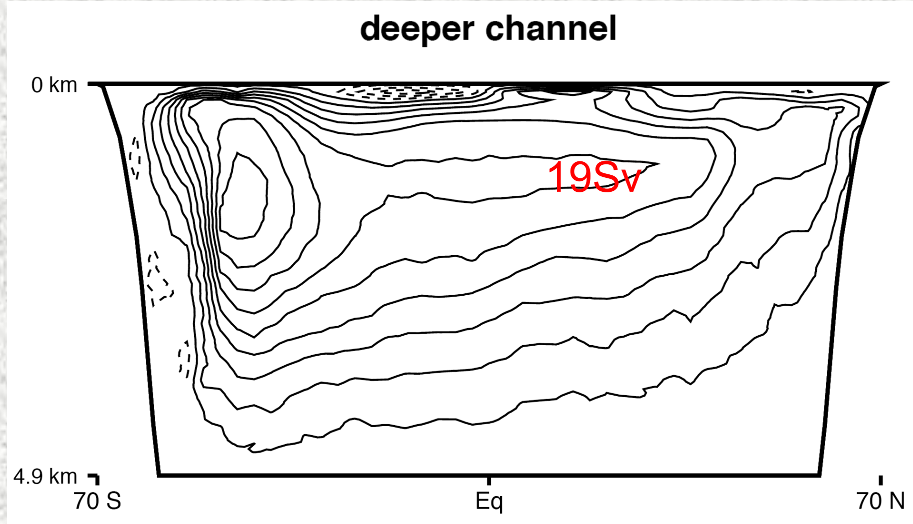








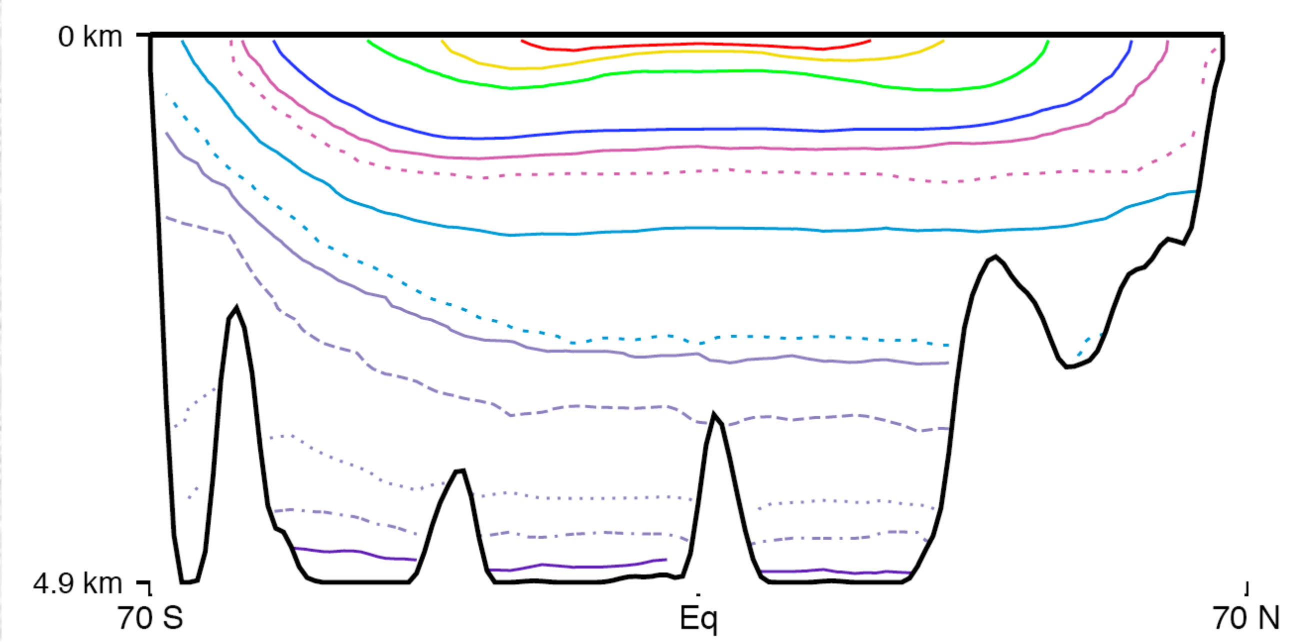
Summary:

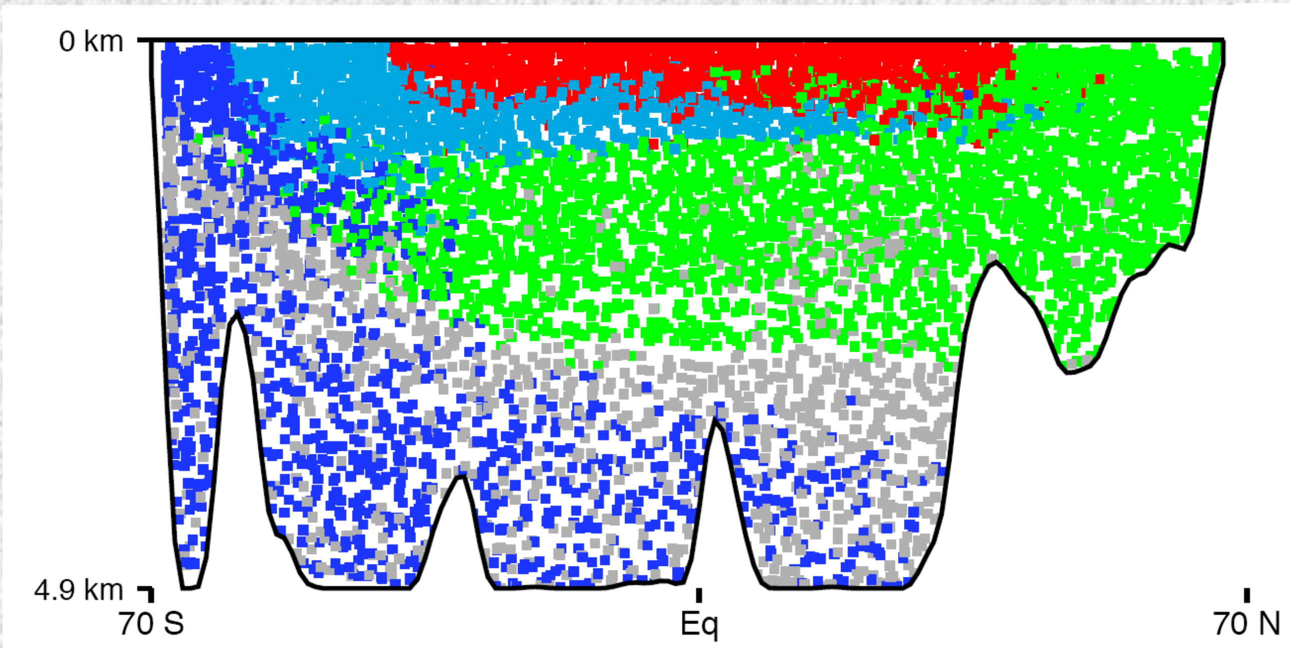
- *An ocean with a fully adiabatic interior gives the first-order solution for the ocean overturning and stratification problem, as long as the model includes zonal winds and a circumpolar channel.*

- *Deepening the channel or increasing the wind stress intensifies the overturning*

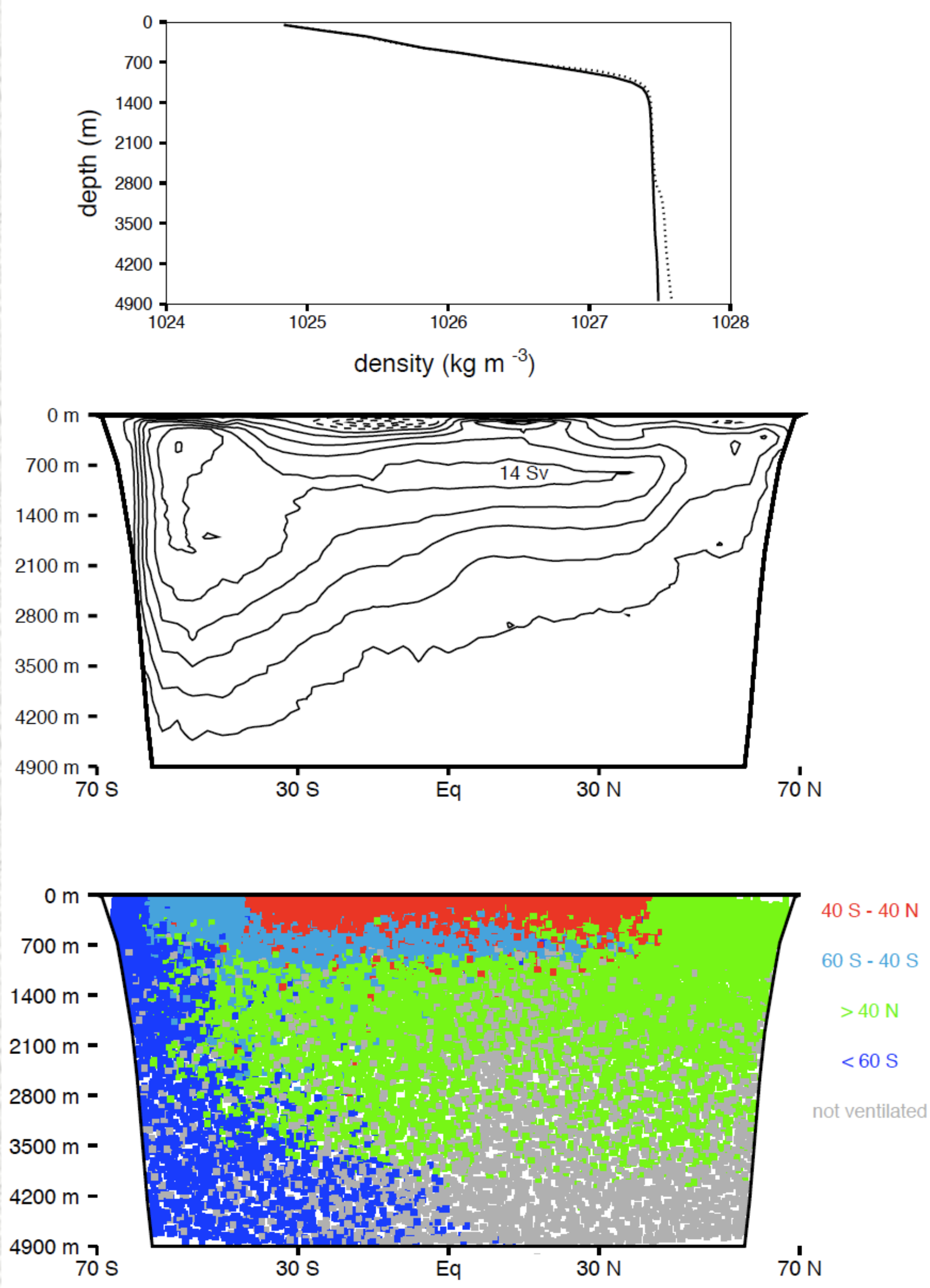
- *Diffusion provides the second-order perturbations, increasing stratification in the deep ocean (between 1000 and 3000m) and producing a more realistic bottom overturning cell*

The role of bottom



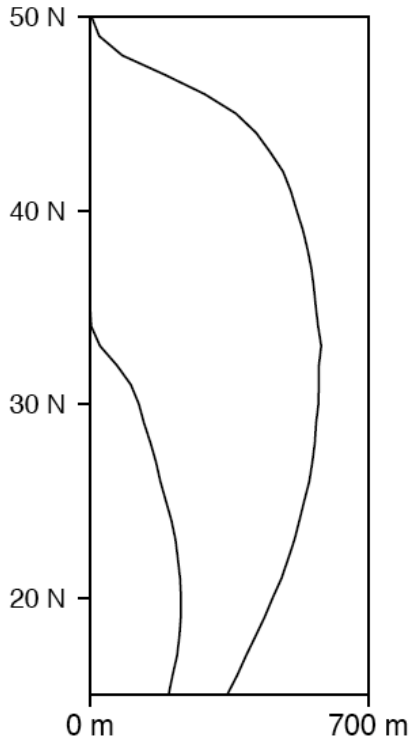


What controls stratification in the bottom ocean

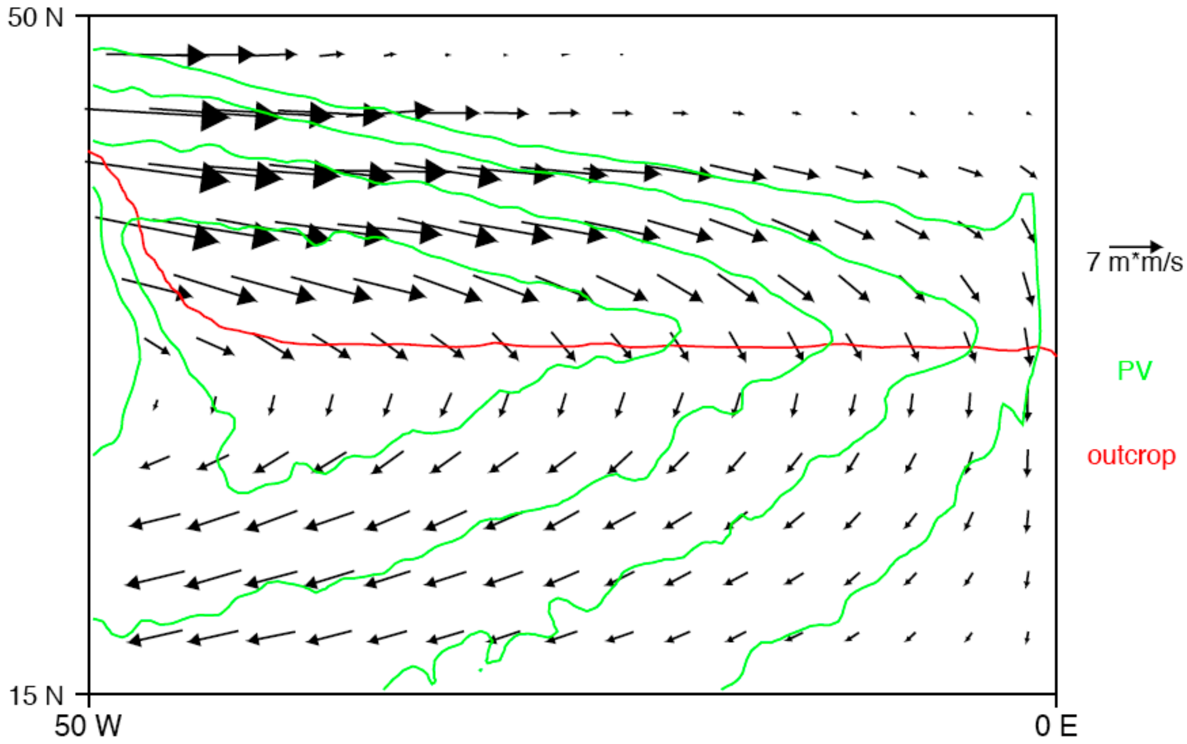


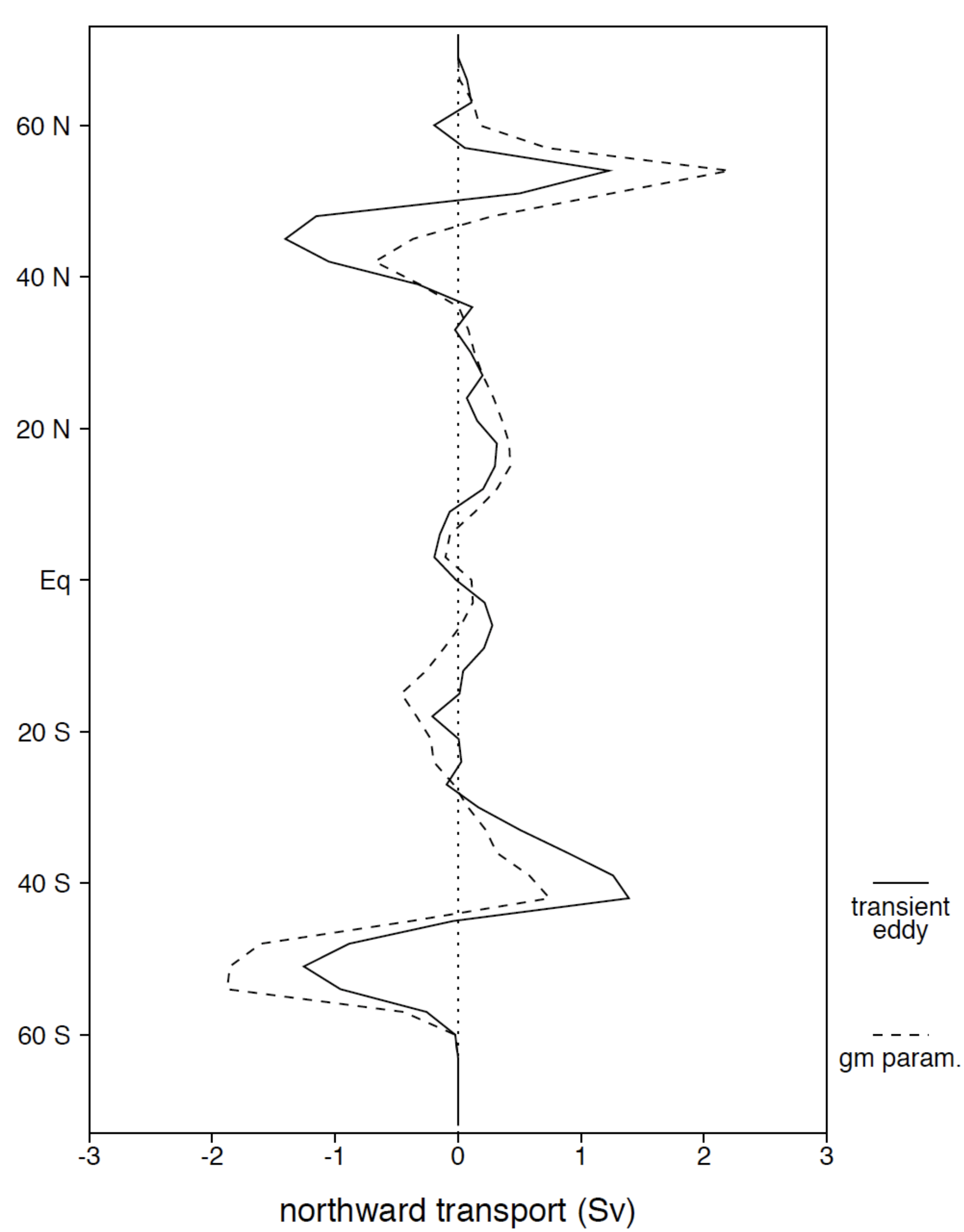
The ventilated thermocline

1025-26 kg/m³ density layer along 20°W



Currents and potential vorticity





Adiabatic horizontal eddy mixing in the model:

Northward bolus transport *of 1026-1027 kg m^{-3} layer thickness for 3^o model resolution in the non-diffusive run (solid line), ∫v'H'dx, and* ***bolus transport*** *in a conventional GM parameterization with a coefficient of 500 m^2s^{-1} (dotted line)*