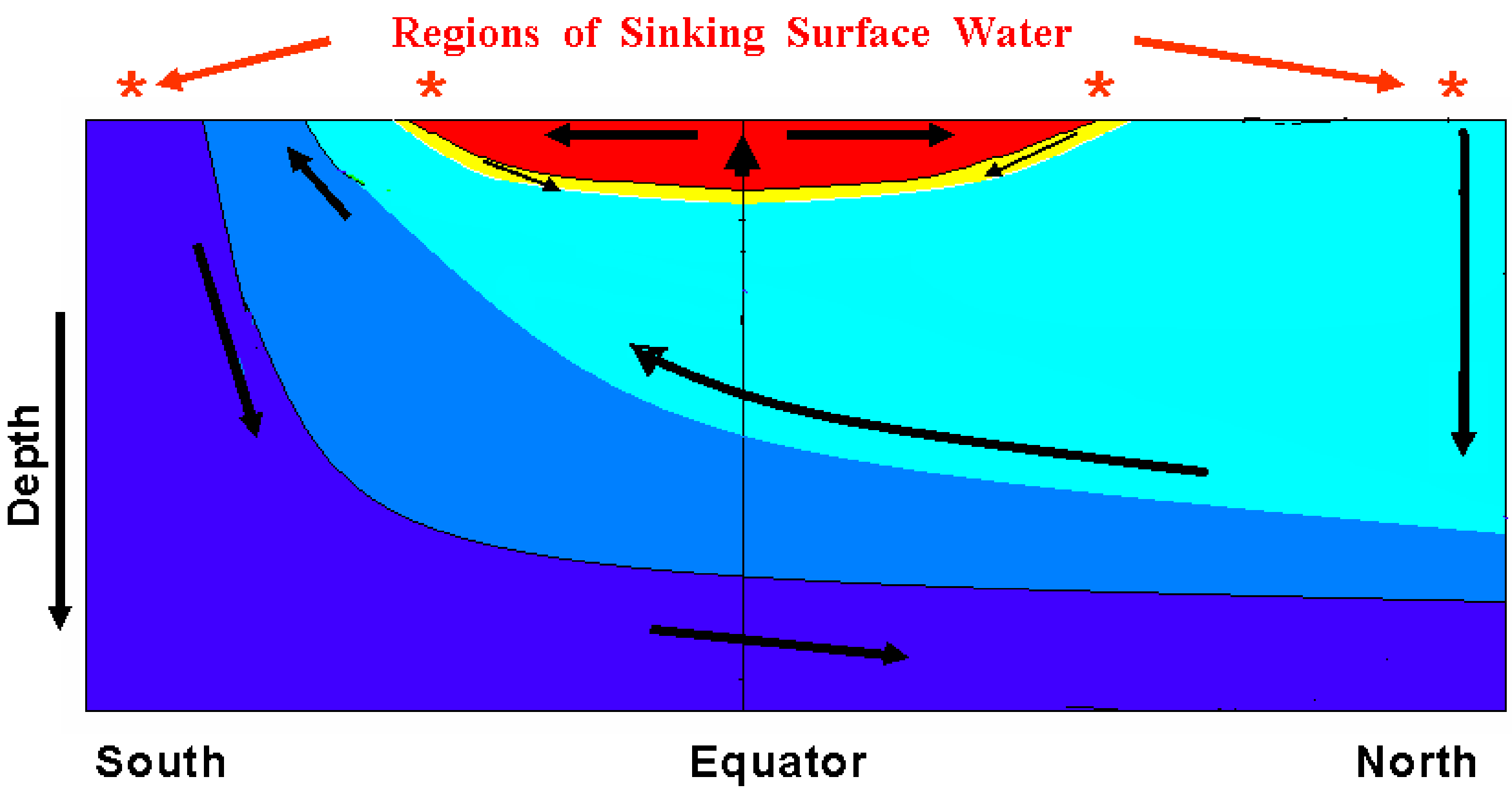
The thermal structure of the upper ocean

Boccaletti et al (2004)

Both Components of the Oceanic Circulation are Conveyor Belts because both involve meridional overturning.

(a) Shallow, rapid wind-driven circ. $uT_x + vT_y + wT_z = 0$

(b) Deep, slow, thermohaline circ. $wT_z = kT_{zz}$



Both circulations transport heat polewards, maintain the surface heat balance, and can affect climate.

Deep thermohaline circulation

1) Thermal wind (geostrophy + hydrostatic)

$$f\frac{\partial u}{\partial z} = -g\gamma\frac{\partial T}{\partial y}$$

$$f\frac{\partial v}{\partial z} = g\gamma\frac{\partial T}{\partial x}$$

2) Geostrophic vorticity equation (geostrophy + incompressible).

$$\beta v = f\frac{\partial w}{\partial z}$$

3) Thermodynamics: vertical advection-diffusion

$$w\frac{\partial T}{\partial z} = \kappa\frac{\partial T^2}{\partial z^2}$$

Depth of the thermocline (applicable to a high diffusive ocean)
-increases with diffusion
-decreases with increased vertical stability

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

Shallow wind-driven circulation of ventilated thermocline (below Ekman layer)

1) Thermal wind (geostrophy + hydrostatic)

$$f\frac{\partial u}{\partial z} = -g\gamma\frac{\partial T}{\partial y}$$

$$f\frac{\partial v}{\partial z} = g\gamma\frac{\partial T}{\partial x}$$

2) Geostrophic vorticity equation (geostrophy + incompressible).

$$\beta v = f\frac{\partial w}{\partial z}$$

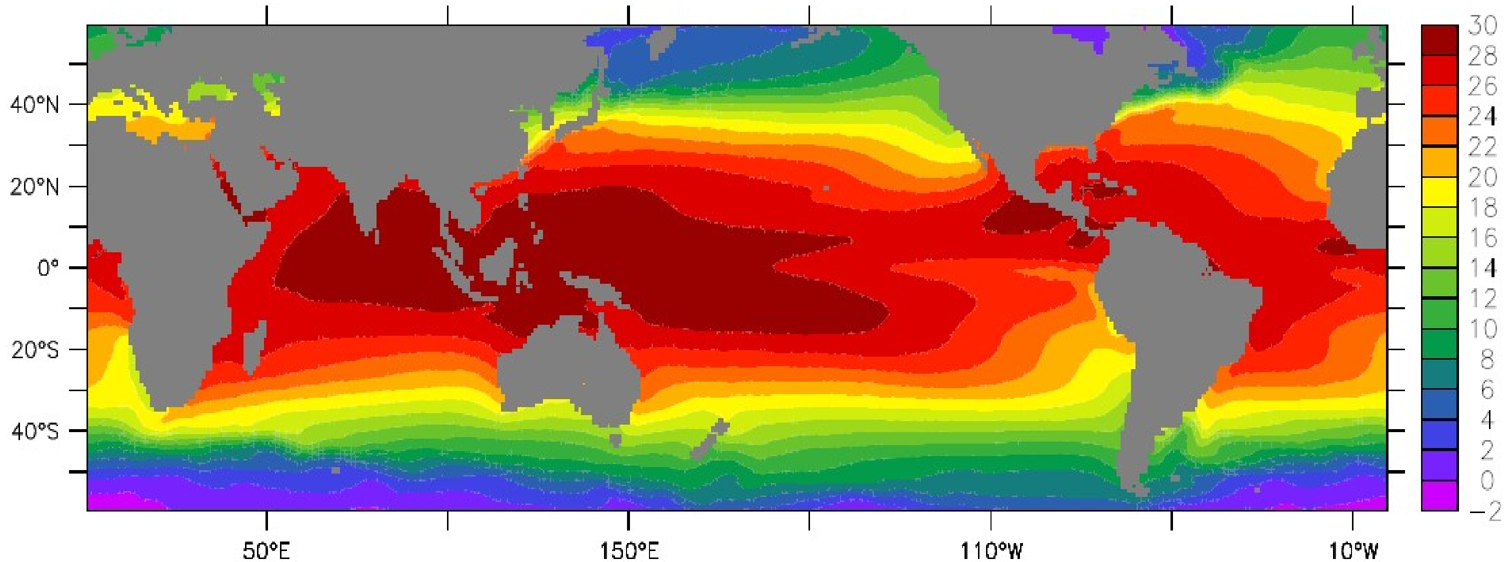
3) Thermodynamics: conservative flow

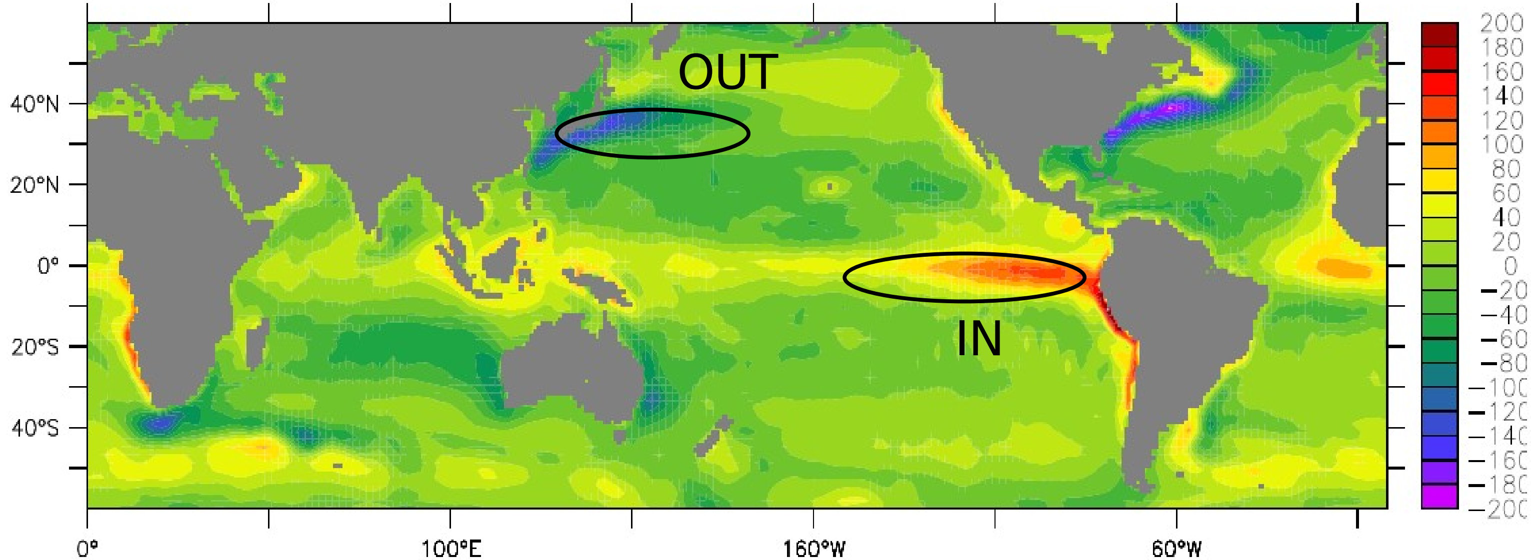
$$\vec{V}.\nabla T = 0$$

For a two-layer ocean the depth of the thermocline:
-depends on the depth at the eastern boundary
-increases with wind strength
-decreases with increased vertical stability of the ocean

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

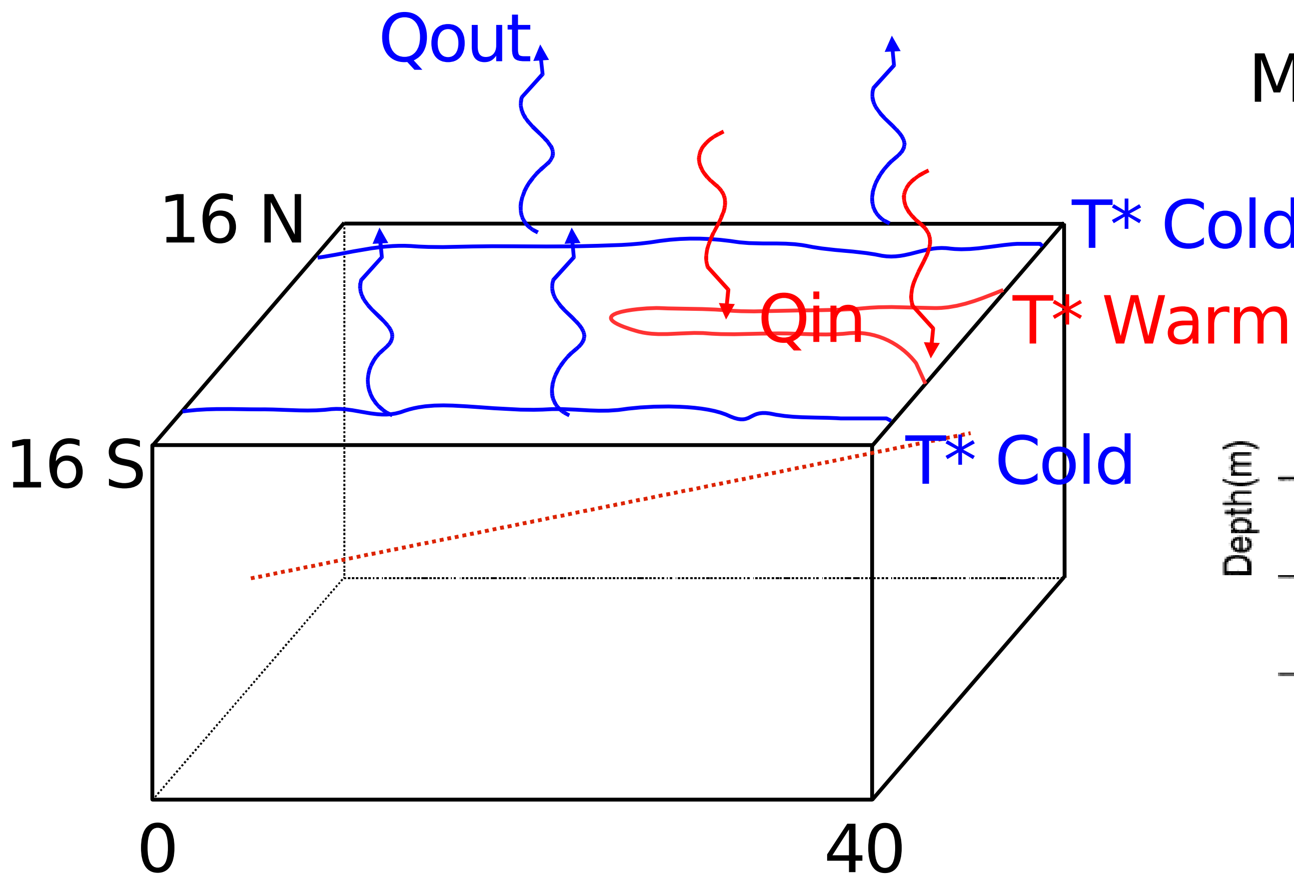
SST and net heat flux into the ocean

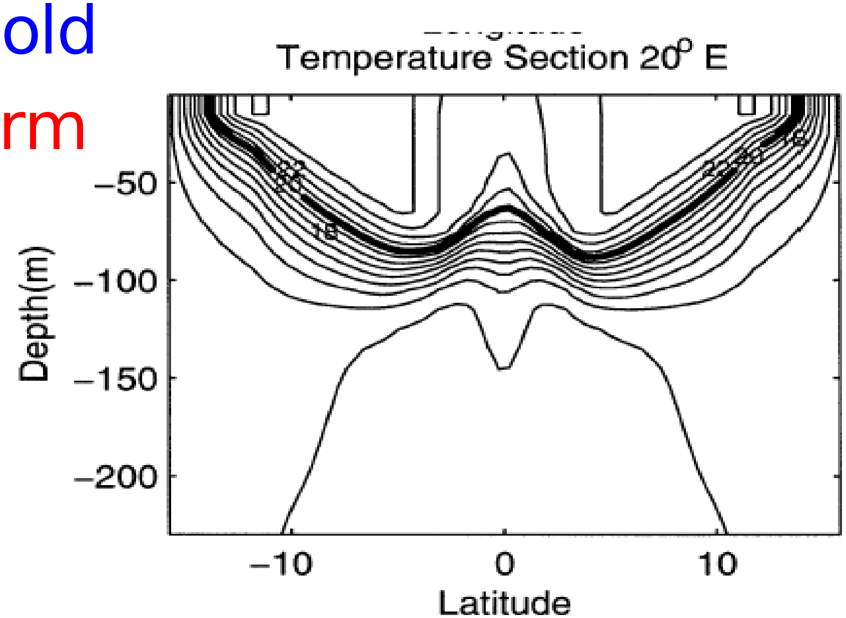




Consider a one-basin ocean GCM forced with constant wind stress and meridionally varying air temperature T*, so that the surface heatflux is Q=a(T*-T).

MOM4, 0.5x0.5, 32 levels

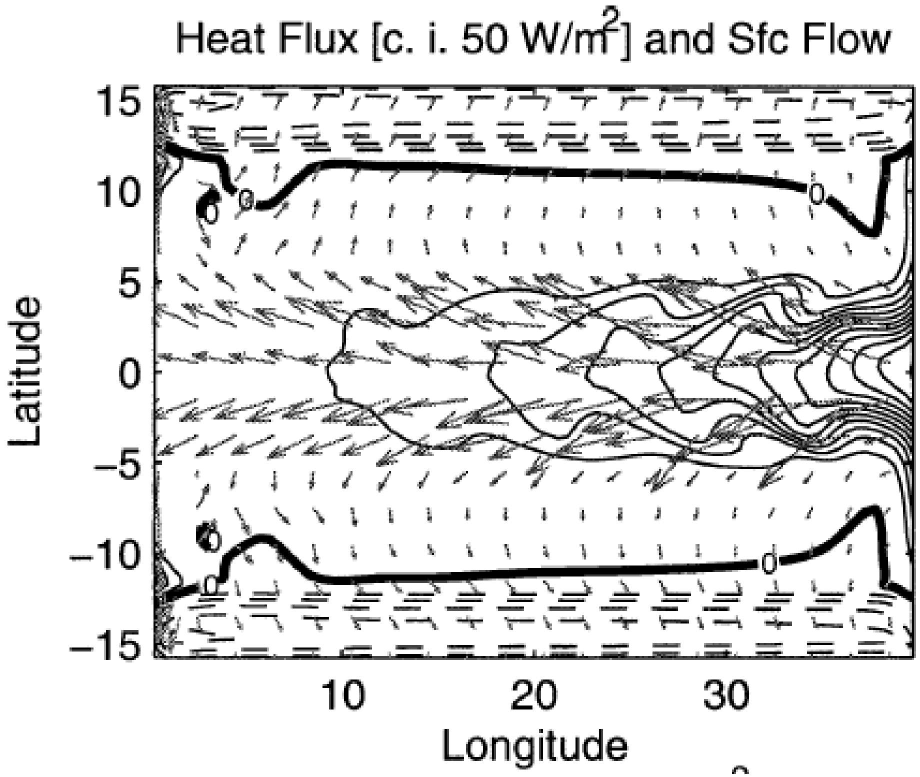


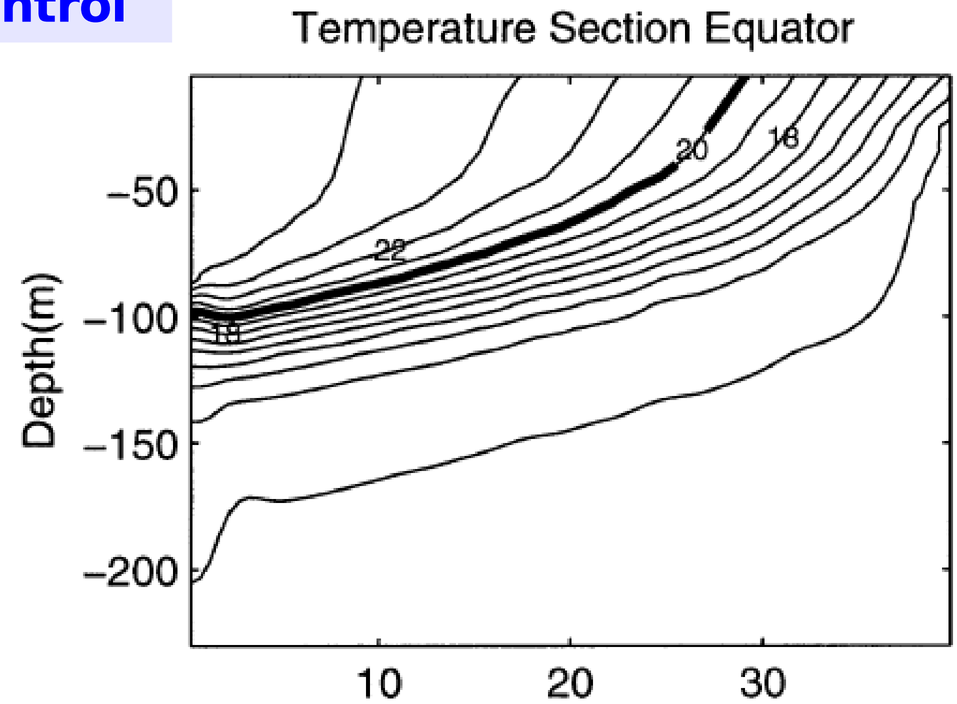


Under the constraint of a balanced heat budget Qout=Qin

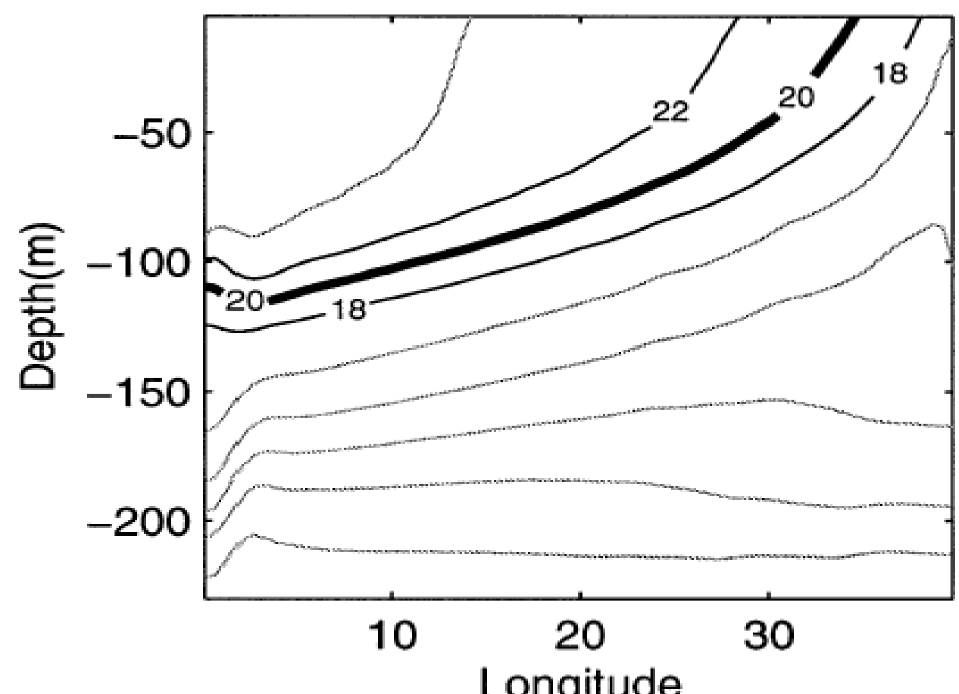
What happens if the high latitude T* changes??

Control





Warmer “high latitudes”



Even though the momentum forcing and diffusivity is the same the thermocline depth is different.

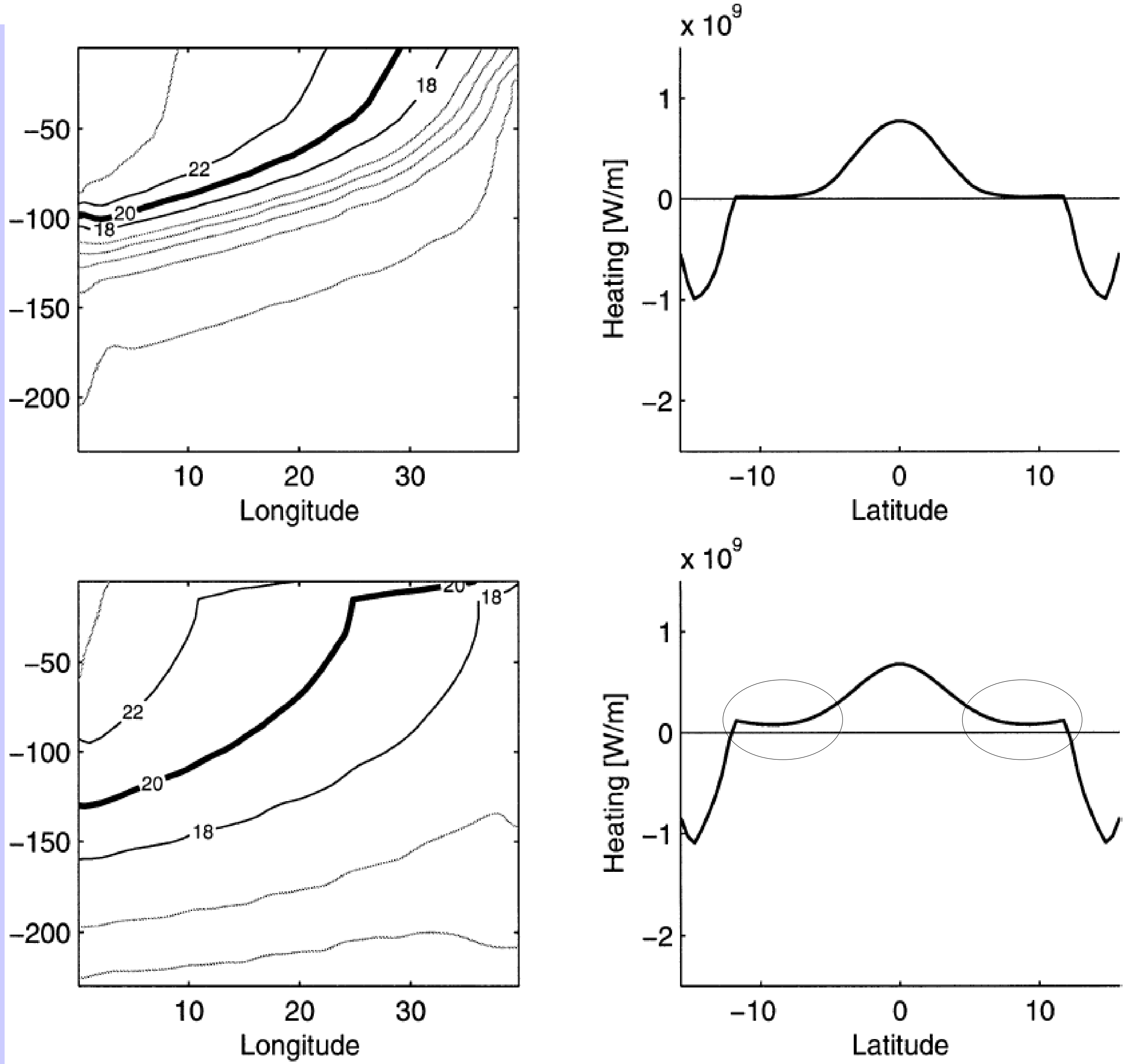
As the high latitude T* gets warmer the ocean loses less heat. To decrease the heat uptake and balance the heat budget, the equatorial thermocline deepens and exposes less cold water

The role of diffusion

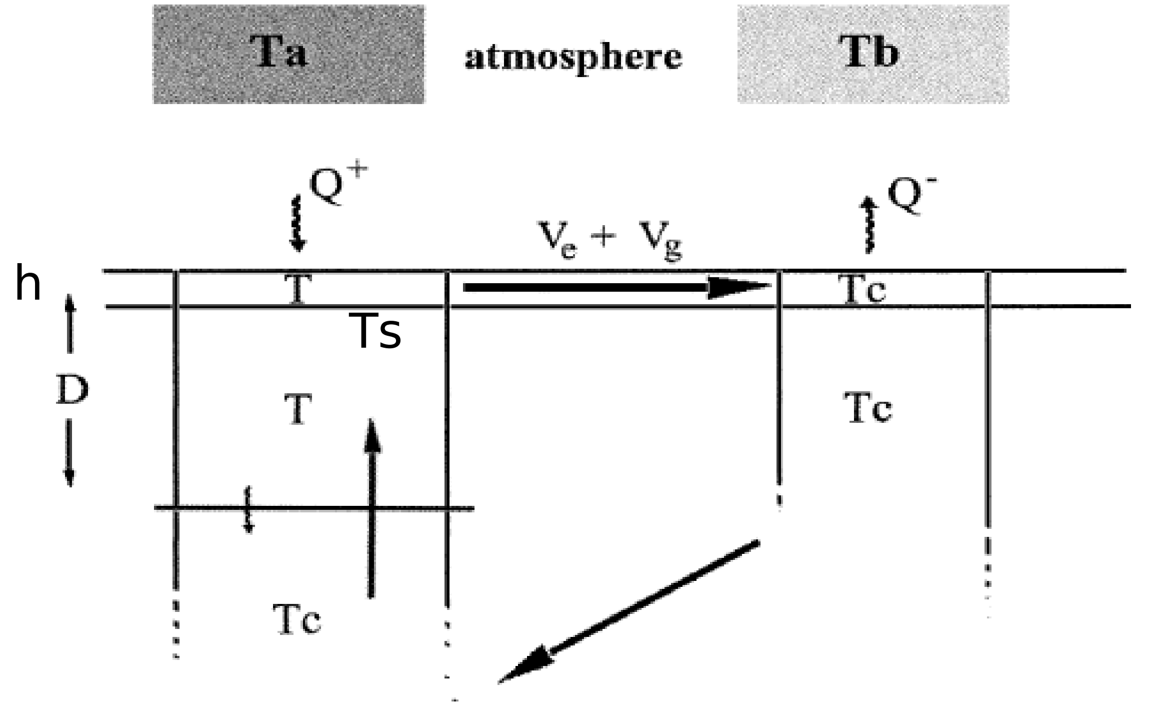
For the same T*, an increase in diffusion “diffuses” the thermocline and reduces the static stability.

The ocean is now able to gain heat over a large area, not restricted to the cold tongue. Thus, diffusion also controls the depth of the thermocline and the strength of the cold tongue.

The partition of heat transport between shallow and deep circulations depends on diffusion. Note that in this model the THC is represented by mixing.



$$Q_P = \alpha(T_a - T)$$
$$Q_N = \alpha(T_b - T_c)$$



Constraints:

$$Q_P + Q_N = 0$$
$$\Delta T = (T - T_c) = (T_a - T_b) - 1/\alpha(Q_P - Q_N)$$

diffusion upwelling

$$Q_P = L^2\left(\kappa\frac{\Delta T}{D} + w_E(T - T_S)\right) = L^2\left(\kappa\frac{\Delta T}{D} + w_E h\frac{\Delta T}{D}\right)$$

$$Q_N = -v_E \Delta T(hL) - v_G \Delta T(DL) = -v_E \Delta T(hL) - g\gamma\frac{D^2(\Delta T)^2}{f}$$

Ekman drift

Boundary geostrophic flow

Theory Results eqns for D and ΔT:

$$v_E \Delta T h + g\gamma \frac{D^2 (\Delta T)^2}{f L} = \frac{L}{D}(\kappa + w_E h)\Delta T$$

$$\Delta T = (T_a - T_b) - \frac{2\Delta T}{D\alpha}(\kappa + w_E h) L^2$$

Note that the first equation transforms to a statement of mass conservation by removing the "common" ΔT.

If only diffusion is important in transporting heat $v_E = w_E \Delta T = 0$ and recover diffusive scaling of thermocline depth

geostrophic flow ~ diffusive upwelling => $D = \kappa^{1/3} [\frac{f^2 L}{g \beta \gamma \Delta T}]^{1/3}$

In the limit of no mixing (K=0), the equation is third order in D. For large thermocline depth D:

$$D = (\frac{f L^2 w_E h}{g \gamma \Delta T})^{1/3}$$

where D depends only on winds and mixed layer.

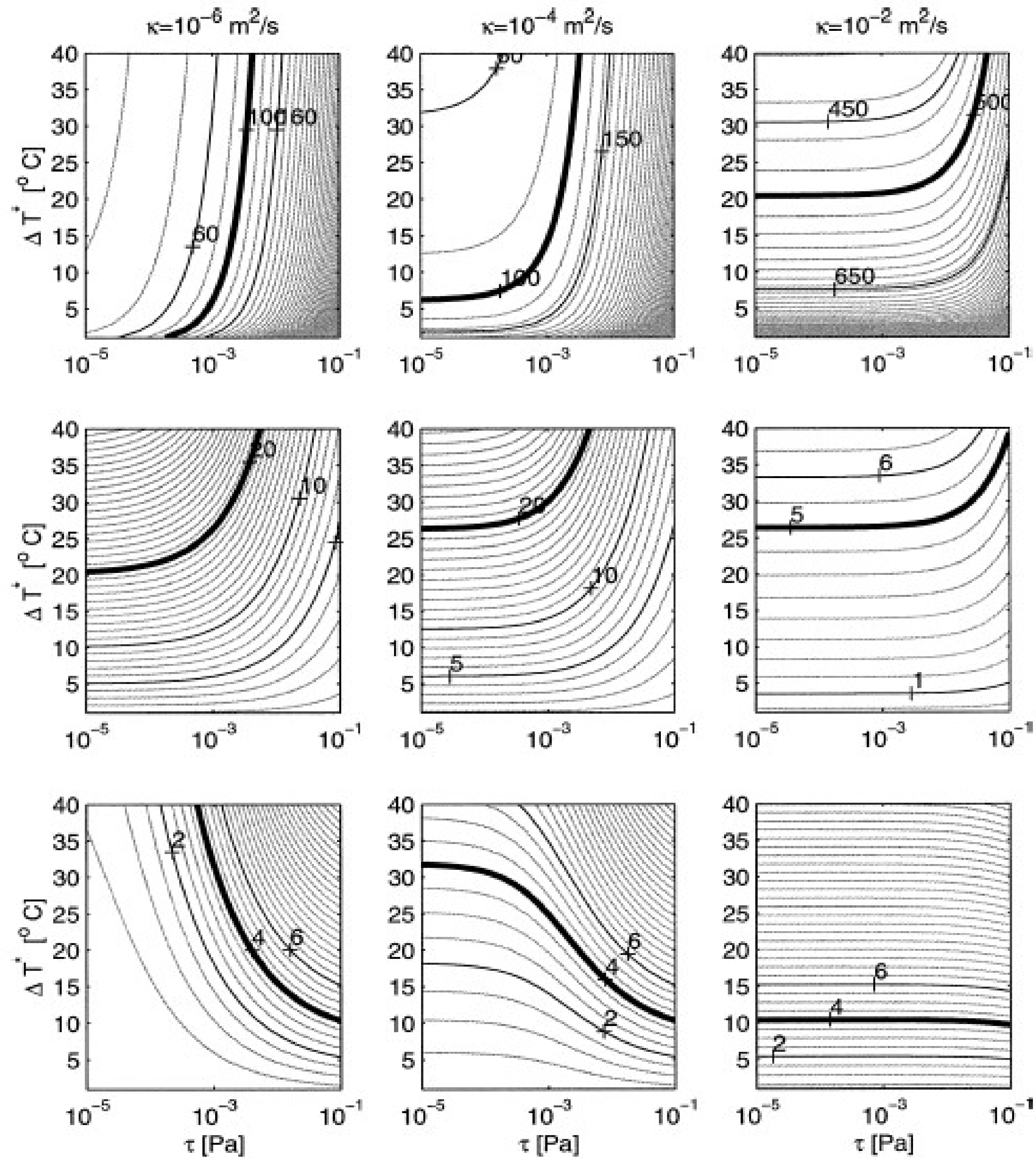
The larger the diffusivity the sensitivity to wind forcing decreases and thermohaline effects dominate.

Thermocline Depth

Static stability

Ocean heat transport

$$H = \rho_0 C L^2 (\kappa + w_E h) \frac{\Delta T}{D}$$

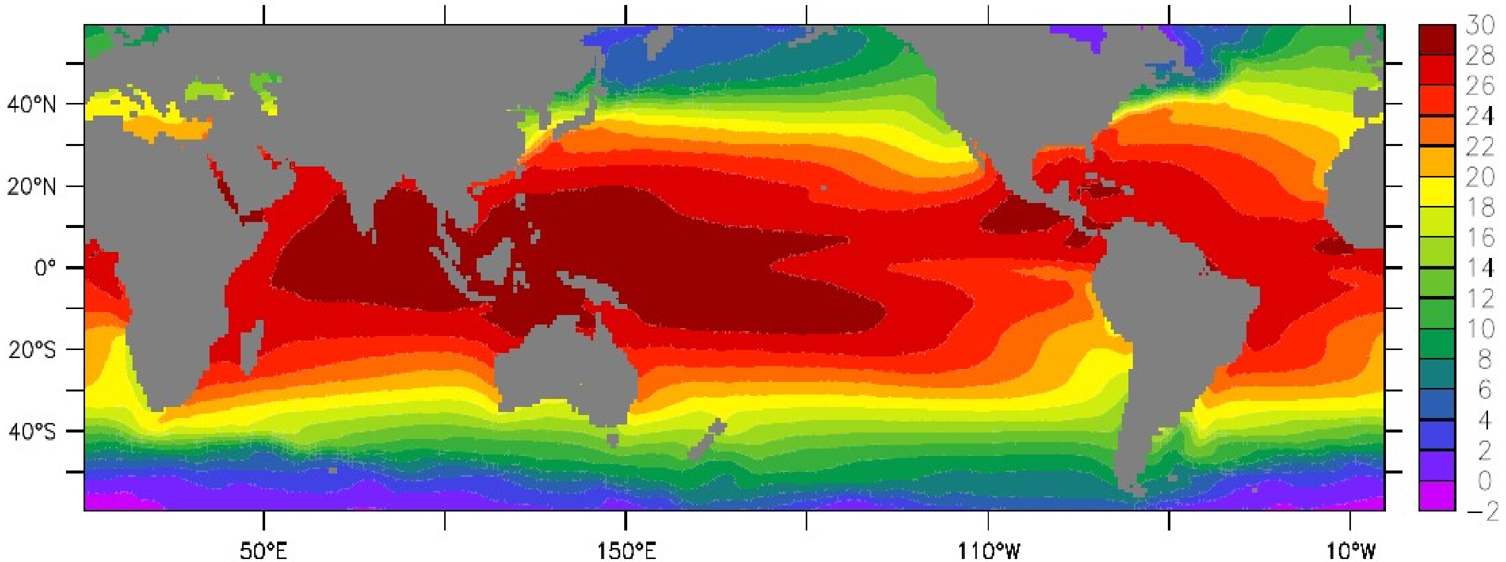


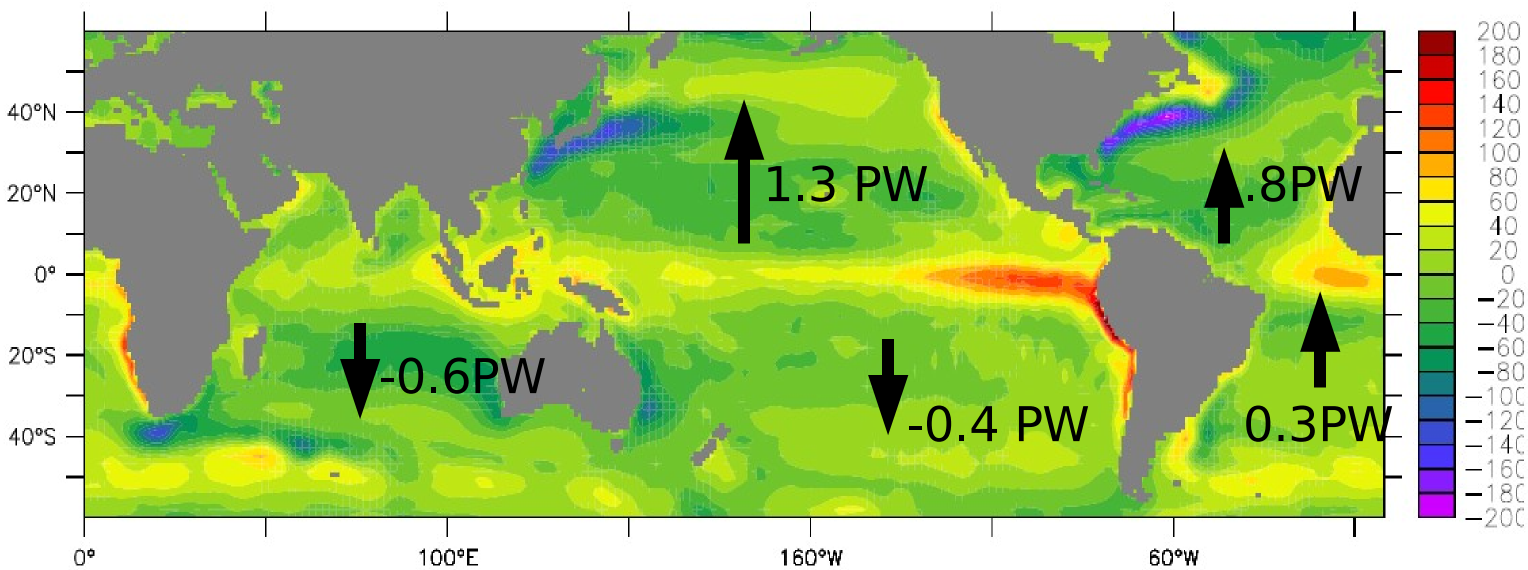
Summary

- The constraint of a balanced heat budget strongly constraints the thermal structure of the upper tropical oceans.

- In a small closed basin this constraint implies that as the extratropics lose less heat, the equatorial thermocline has to deepen in order to expose less cold water and gain less heat.

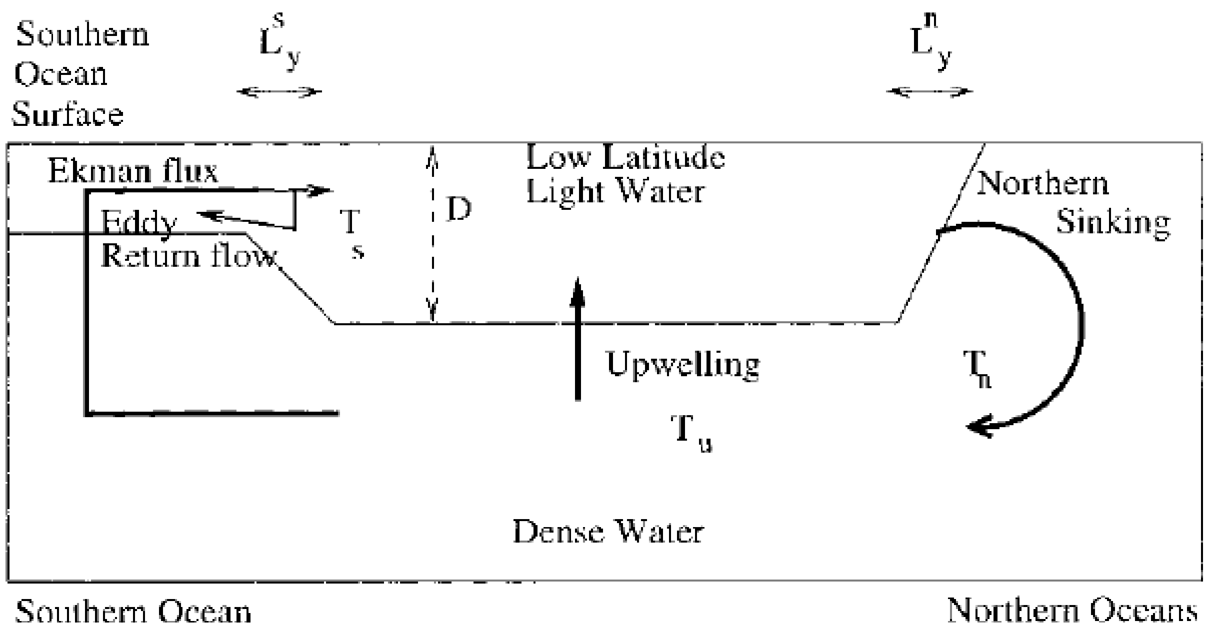
Problem: Ocean basins are not independent, and are connected through ocean and atmospheric bridges

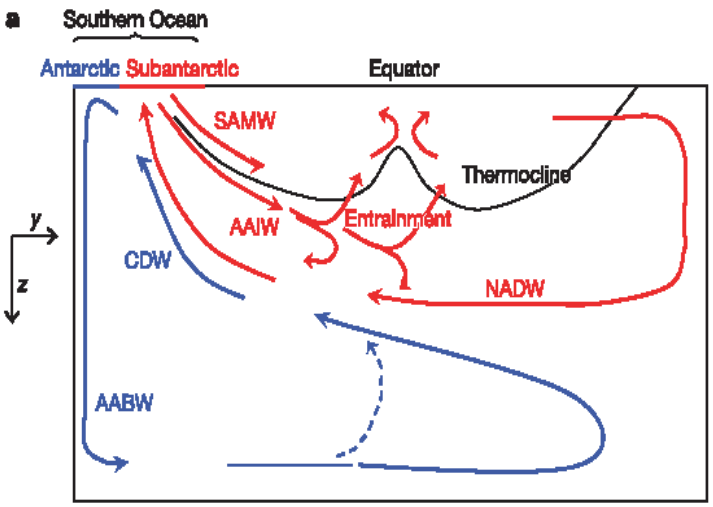




Talley 2003

Gnanadesikan (1999) model of pycnocline





Marinov et al 2006

Mass budget: $T_U = T_N - T_S$

$$\frac{K_V A}{D} = \frac{C g' D^2}{\beta L_y^N} - \left(\frac{\tau}{\rho f} L_X - \frac{A_I D}{L_y^S} L_X\right)$$

Tropical diffusive Upwelling

Northern Flow (geostrophic boundary current)

Southern Ocean: Ekman - eddies