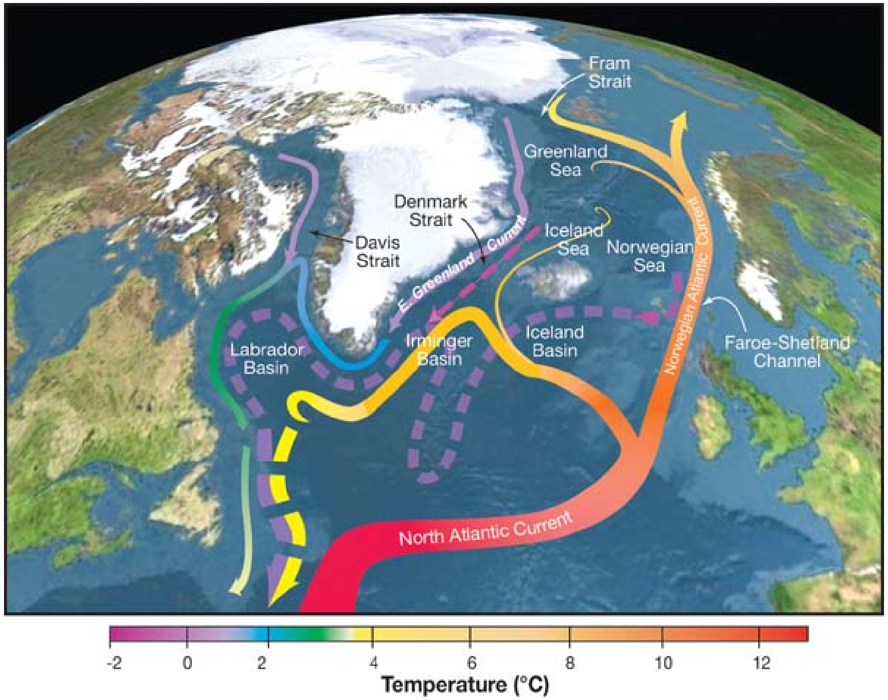
# The Sinking branch of the Atlantic Meridional Overturning Circulation

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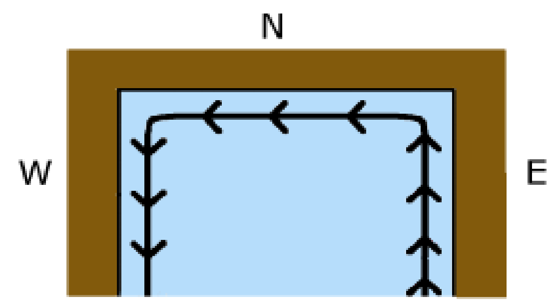
# There is sinking near a boundary when there is buoyancy forcing.

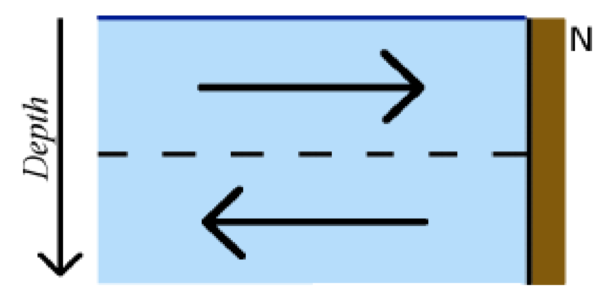
Consider the thermodynamics balances within a mixed layer subject to a cooling $Q$.

2 main assumptions :

- Without cooling - Flow along the boundary




- With cooling - No integrated net flow towards the boundary layer. Slight deviation.




Spall and Pickart: Where does dense water sink? A subpolar gyre example JPO 31, 810-826, 2004

Steady state, no horizontal diffusion. Density conservation :

$$u\rho_x + v\rho_y = Q$$

Polar form components :

$$u = V\cos\theta \quad , \quad v = V\sin\theta$$

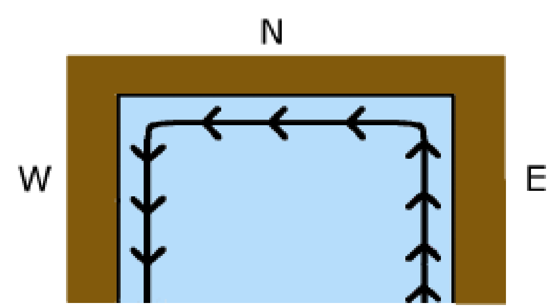
Assumption: adiabatic flow along boundary: $\theta = 0$
diabatic flow much weaker: $\theta$ is small

So:

$$u = V$$

$$v = V\theta$$

$$V\rho_x = Q$$



Geostrophic & hydrostatic balances :

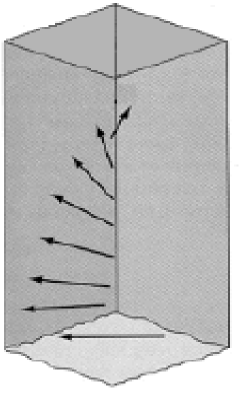
$$\rho_x = -\frac{f\rho_0}{g} v_z \qquad ,$$

Density conservation:

$$-\frac{Vf\rho_0}{g} V\theta_z = Q$$

Which implies that the velocity vector rotates with depth:

$$\theta_z = -\frac{gQ}{V^2 f\rho_0}$$



Cooling forces the velocity vector
to spiral anti-clockwise

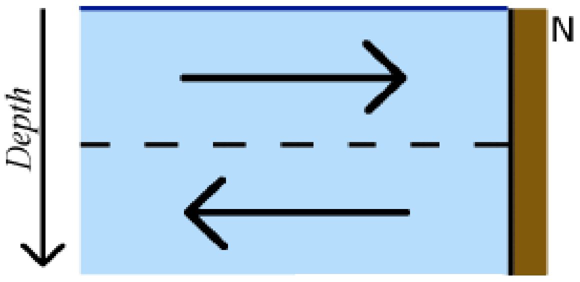
The direction of the flow becomes:

$$\boxed{\theta(z) = \theta_0 - \frac{gQz}{f\rho_0 V^2}}$$

1st order : Flow along the boundary. Then : $v \sim \theta v$

$$v(z) \sim \frac{ghQ}{2\rho_0 f V}\left(1 - \frac{2z}{h}\right)$$

The meridional flow is northward in the upper mixed layer and southward in the lower mixed layer



The meridional transport per unit depth :

$$\int_0^{L_x} v\,dx = \frac{gh}{2\rho_0 f}\left(1 - \frac{2z}{h}\right)\int_0^{L_x} \frac{Q}{V}\,dx$$

$$Q \sim V\rho_x$$

The total meridional overturning forced by boundary convection

$$M_B = \int_0^{\frac{h}{2}} \int_0^{L_x} v\,dx\,dz = \frac{g\Delta\rho_B h^2}{8\rho_0 f}$$

The net sinking does not depend on the surface density flux, length of the boundary, magnitude of the advection, or on any dynamics and thermodynamics away from the boundary, although all these factors may indirectly influence $\Delta\rho_B$

# What about eddies and transience?

Eddies: Q becomes Q*, with

$$Q^* = Q - \nabla . (\overline{u' \rho'})$$

This changes the velocity profile, but not the sinking flux relation. Only the mixed layer depth *h* is no longer the relevant depth-scale but *H*; the depth over which the eddies transport heat.

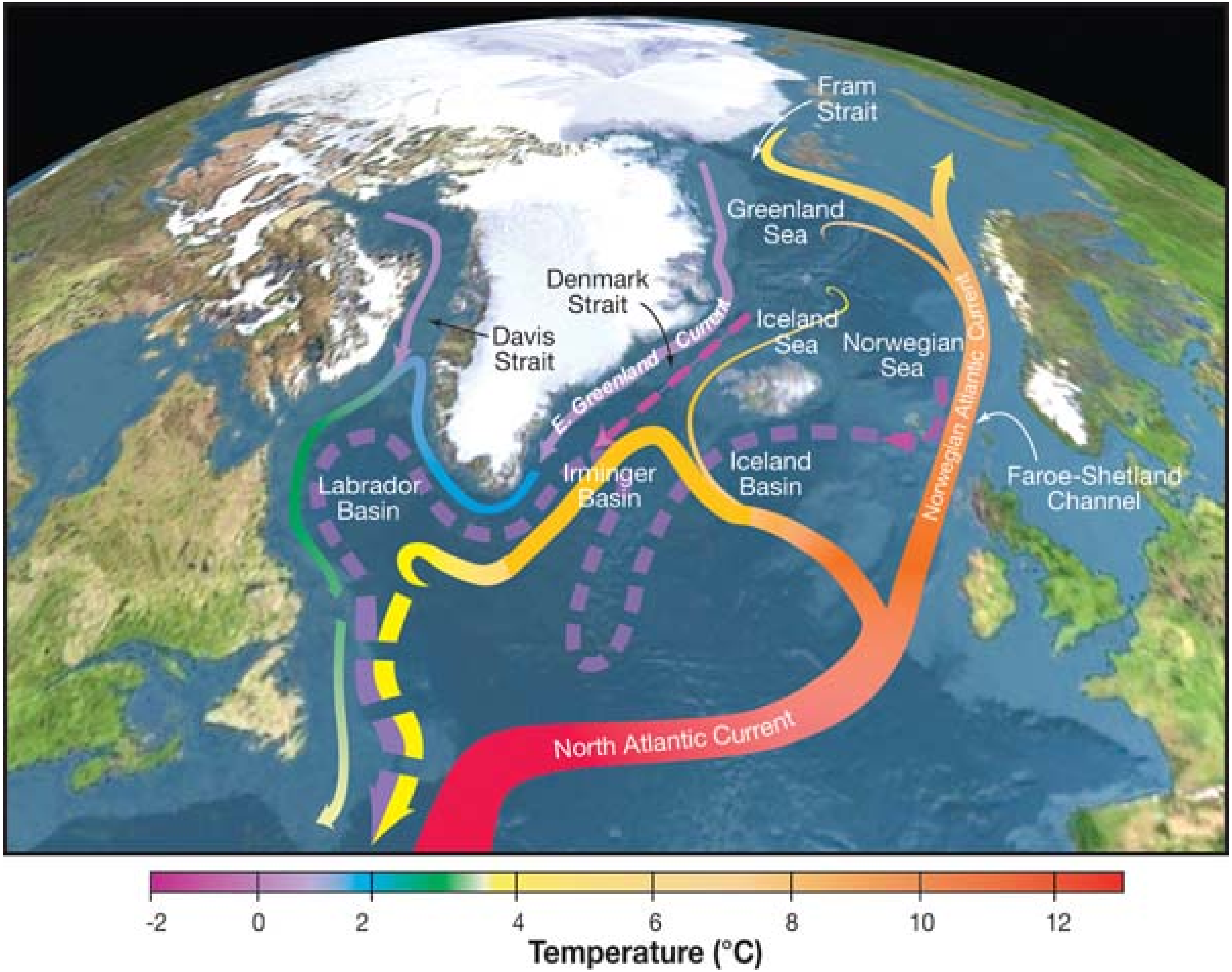
Transience: we can add to the density equation

$$Q_{an} = \rho_t' + c\rho_x'$$

This does not change anything, apart from the eddy flux divergence. So, the relation

$$M_B \sim \Delta\rho_B$$

is robust.


Fram Strait
Greenland Sea
Denmark Strait
Davis Strait
E. Greenland Current
Iceland Sea
Norwegian Sea
Norwegian Atlantic Current
Faroe-Shetland Channel
Labrador Basin
Irminger Basin
Iceland Basin
North Atlantic Current
-2
0
2
4
6
8
10
12
Temperature (°C)

Boundary sinking versus sinking in the interior:

Planetary geostrophic dynamics :

$$v = \frac{fW}{\beta h}$$

Total downward mass flux :

$$M_I = \frac{\beta h L_x L_y v}{f}$$

Total northward mass flux :

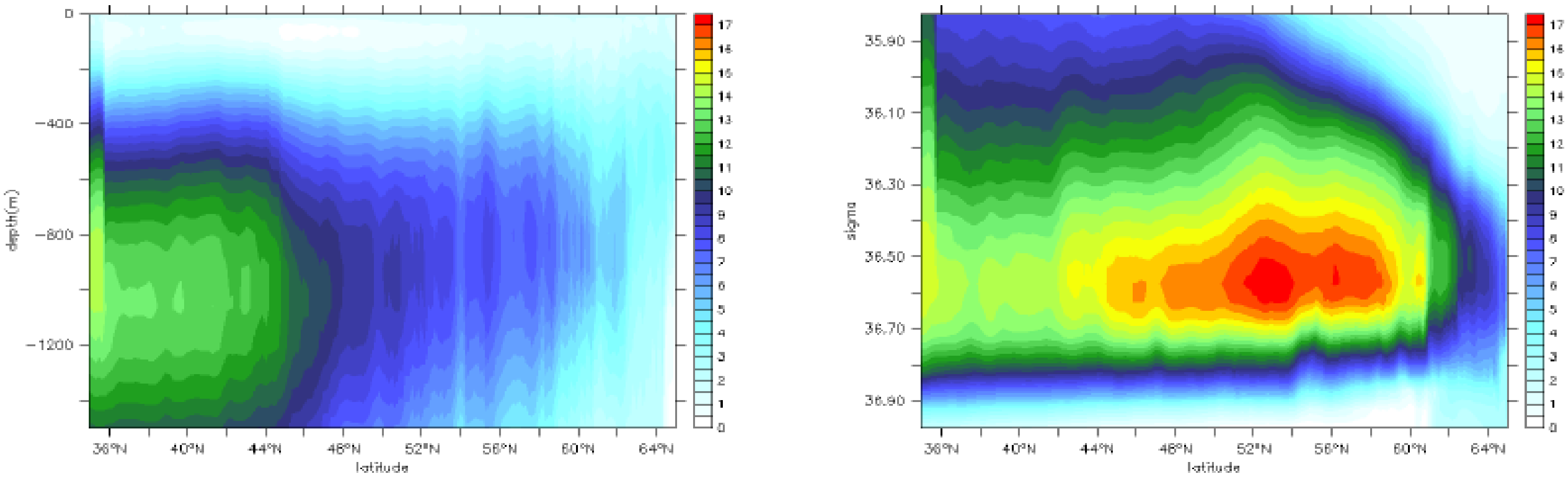
$$M = vhL_x$$

Ratio of the sinking rate to the recirculation strength :

$$\boxed{\frac{M_I}{M} = \frac{\beta L_y}{f}}$$

This ratio is O(0.1). To obtain sufficient sinking the geostrophic constraint must be broken and this only occurs near lateral boundaries.

If sinking does not coincide with convection/densification, the overturning streamfunction in z and density coordinates must be different.



In a coarse-resolution model the interior is no longer governed by planetary geostrophics; friction becomes important;

$$\beta v = f w_z - A\nabla_H^2 \zeta$$

We scale v and $\zeta$ with $V$ and $V/L_y$ respectively

The vorticity balance for the meridional flow then becomes:

$$V = \frac{fW}{\beta H} - \frac{AV}{\beta L_y^3}$$

The total downward mass flux then becomes:

$$M_I = L_x L_y W = \frac{\beta H L_x L_y V}{f}\left(1 + \frac{A}{\beta L_y^3}\right)$$

$$R_F = \frac{A}{\beta L_y^3}$$ is a nondimensional number that measures the importance of numerical friction in driving the interior downwelling branch of the AMOC: $R_F = 0.5$ for coarse resolution models, 5 – 10 times smaller for eddy-resolving models

Again, the total northward mass flux driven by buoyancy loss is given by:

$$M = vHL_x$$

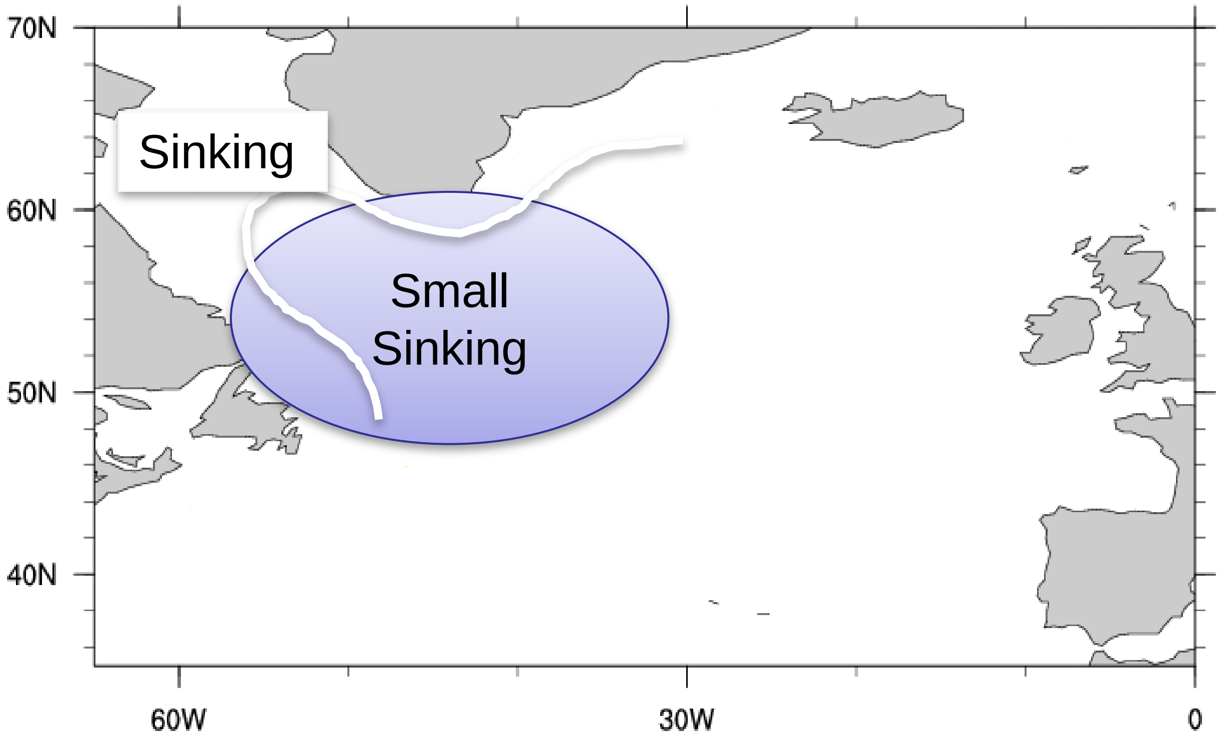
The ratio of the interior to total sinking then becomes:

$$\frac{M_I}{M} = \frac{\beta L_S}{f}\left(1 + \frac{A}{\beta L_y^3}\right) = r_g + r_f$$

with $r_g = \frac{\beta L_S}{f}$ and $r_f = \frac{A L_S}{f L_y^3}$

$$r_g \approx 0.1 \quad , \quad r_f = (0.005 \sim 0.05) \times G_f$$

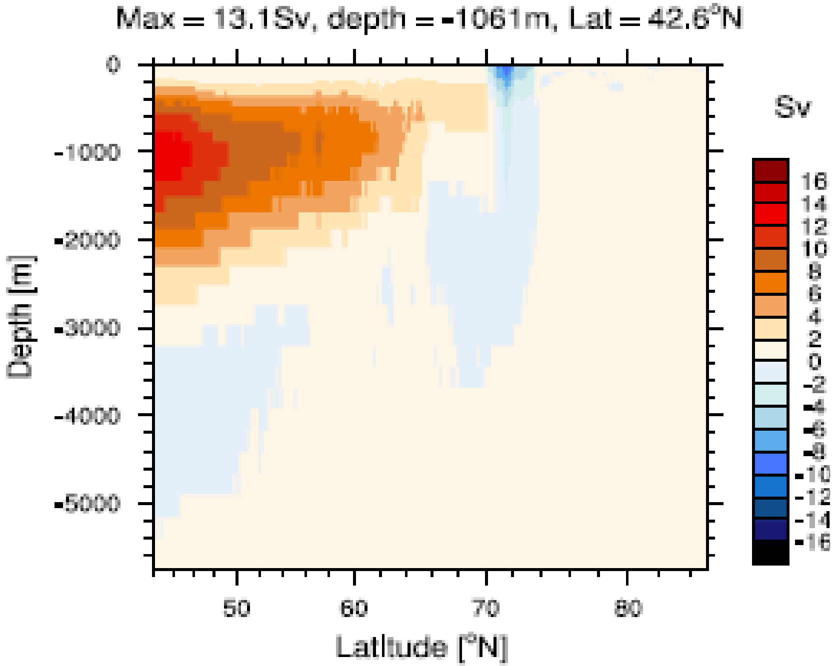
# The theoretical prediction



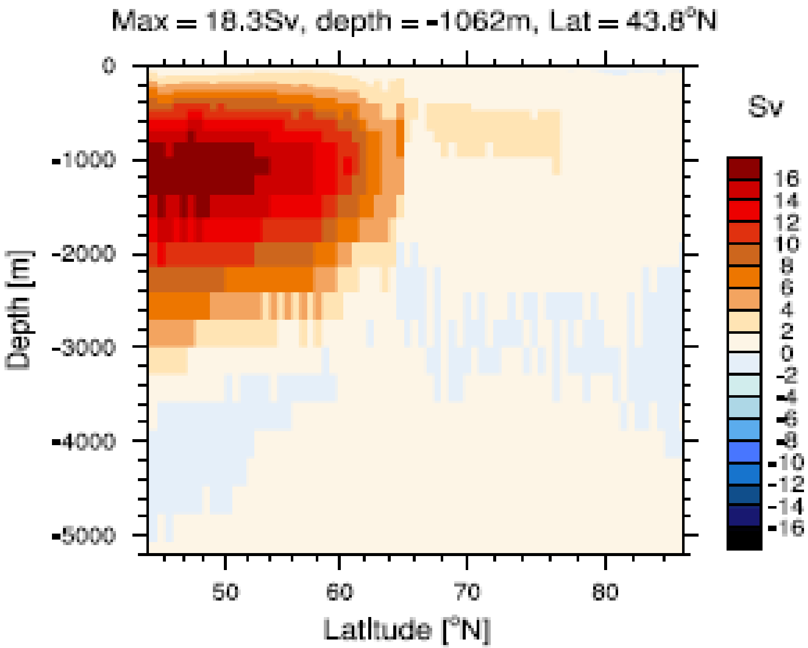

Net sinking in the interior should be small

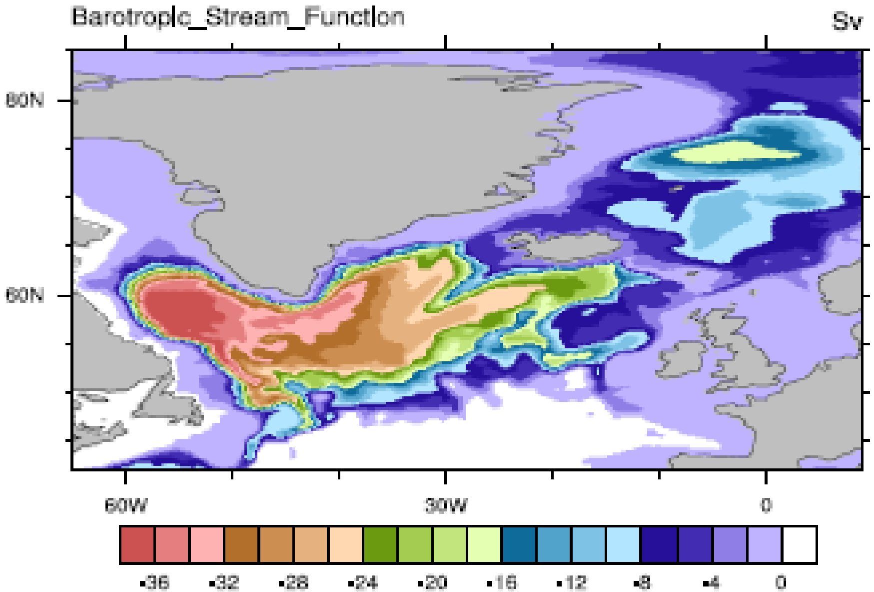
Sinking should occur near the boundaries

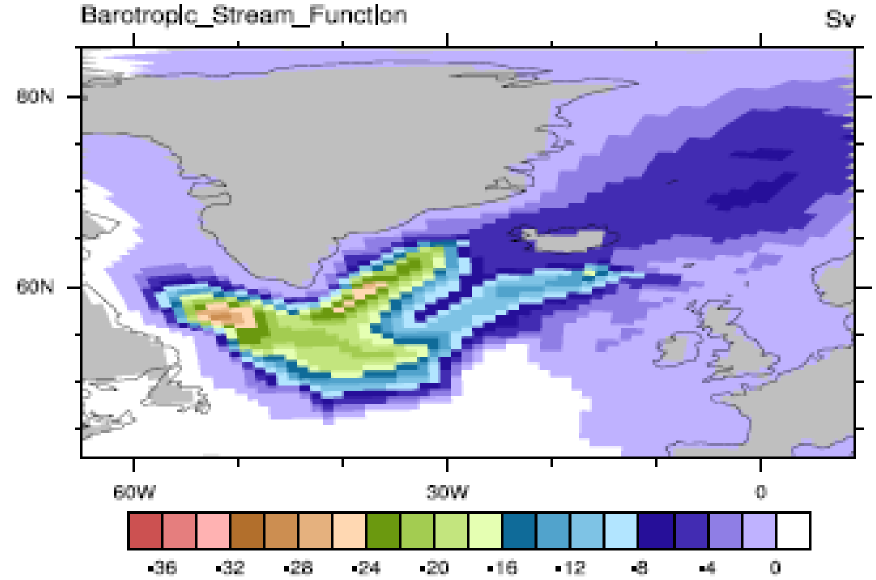
Orca-025




Orca-1




Barotropic_Stream_Function
Sv
80N
60N
60W
30W
0
-36 -32 -28 -24 -20 -16 -12 -8 -4 0


Barotropic_Stream_Function
Sv
80N
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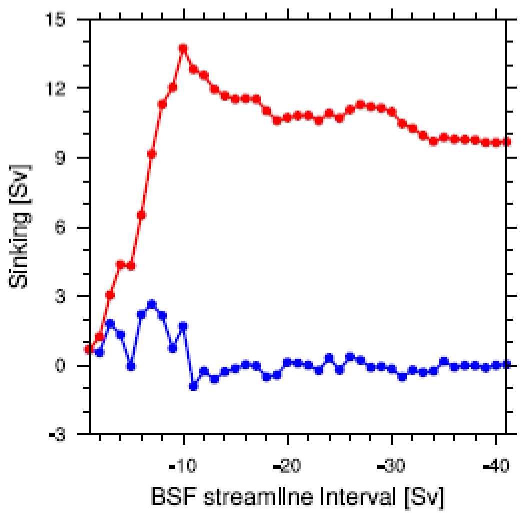
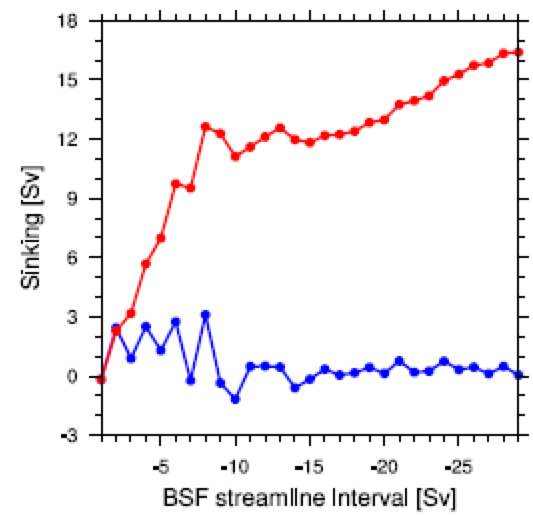
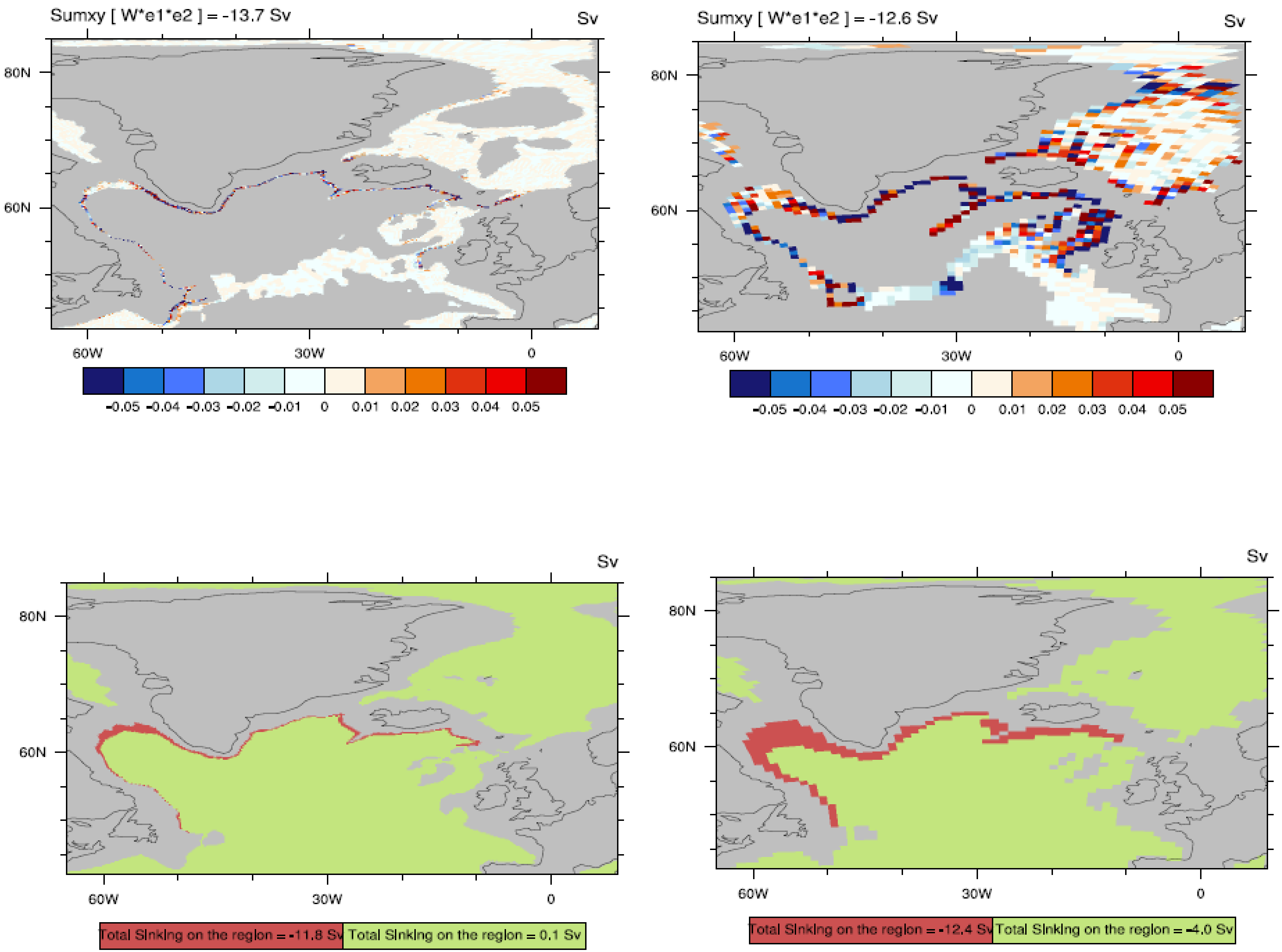
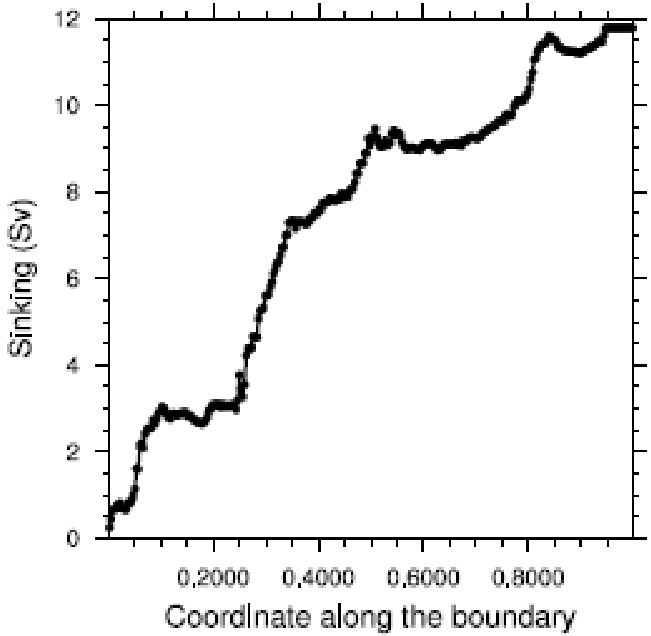
0
-36 -32 -28 -24 -20 -16 -12 -8 -4 0


Sinking [Sv]
BSF streamline interval [Sv]
15 12 9 6 3 0 -3
-10 -20 -30 -40

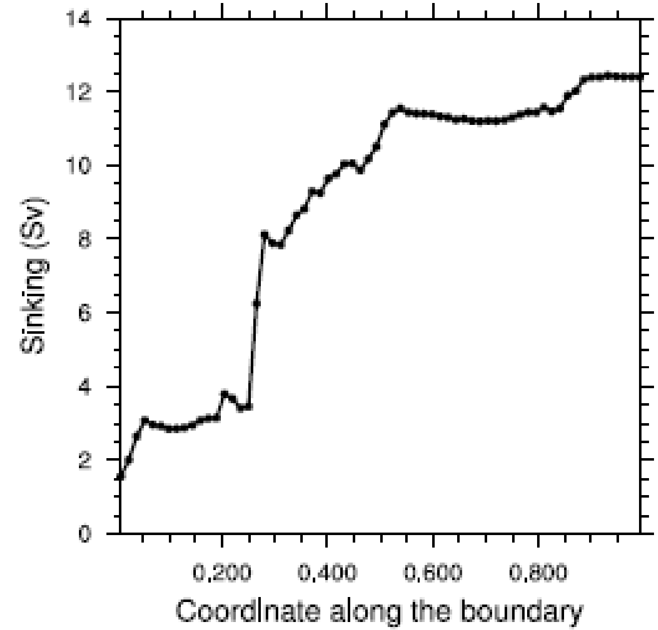

Sinking [Sv]
BSF streamline interval [Sv]
18 15 12 9 6 3 0 -3
-5 -10 -15 -20 -25



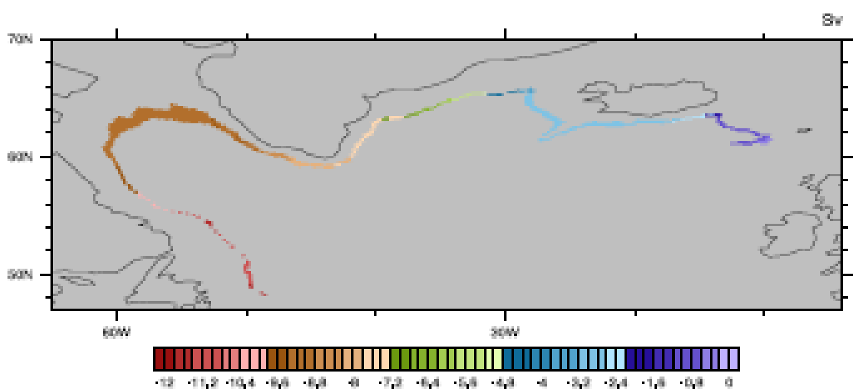
Sumxy [ W*e1*e2 ] = -13.7 Sv
Sv
80N
60N
60W
30W
0
-0.05 -0.04 -0.03 -0.02 -0.01 0 0.01 0.02 0.03 0.04 0.05
Sumxy [ W*e1*e2 ] = -12.6 Sv
Sv
80N
60N
60W
30W
0
-0.05 -0.04 -0.03 -0.02 -0.01 0 0.01 0.02 0.03 0.04 0.05
Sv
80N
60N
60W
30W
0
Total Sinking on the region = -11.8 Sv
Total Sinking on the region = 0.1 Sv
Sv
80N
60N
60W
30W
0
Total Sinking on the region = -12.4 Sv
Total Sinking on the region = -4.0 Sv

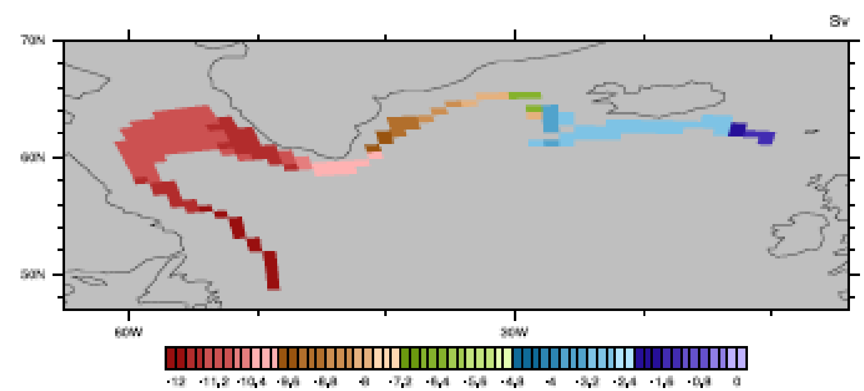

Sinking (Sv)
Coordinate along the boundary
0,2000
0,4000
0,6000
0,8000

(a)


Sinking (Sv)
Coordinate along the boundary
0,200
0,400
0,600
0,800

(a)


Sv
70N
60N
50N
60W
30W


Sv
70N
60N
50N
60W
30W

# Basic assumptions

**Does the Spall scaling work in realistic models?**

Steady state, no horizontal diffusion. Density conservation :

$$u\rho_x + v\rho_y = Q$$

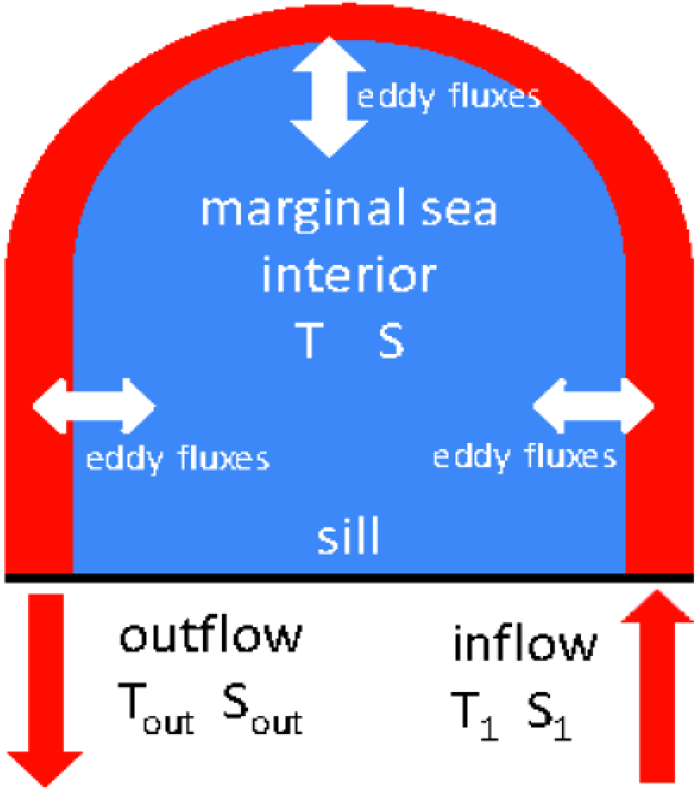
$$Q \sim V\rho_x$$

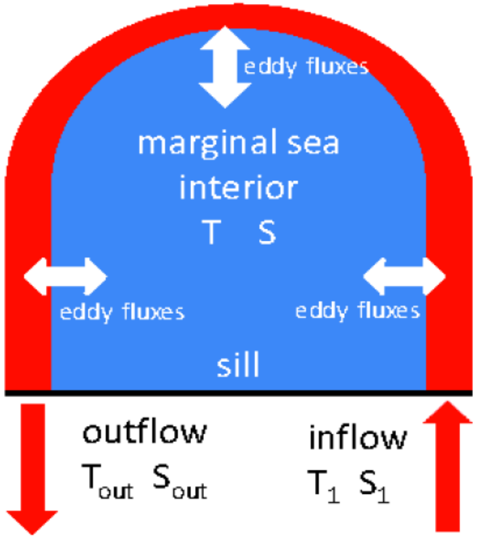
The total meridional overturning forced by boundary convection

$$M_B = \int_0^{\frac{h}{2}} \int_0^{L_x} v \, dx \, dz = \frac{g\Delta\rho_B h^2}{8\rho_0 f}$$

How are changes in **atmospheric forcing and oceanic advection** related to changes in **sinking**?

Spall: Influences of precipitation on water mass transformation and deep convection, JPO, in press.






Heat budget for the Marginal sea:

$$(T_1 - T_{out})VHL = \frac{A\Gamma}{\rho_0 C_p}(T - T_A + PL/A(T_1 - T_A))$$

Salt budget for the marginal sea:

$$(S_1 - S_{out})VHL = -EAS_0$$

After connecting the boundary current to a convective interior variations in sinking can be related to the forcing:

$$W = W^*/\Psi = \frac{1}{2\Delta\rho}(2\mu(1 - \Delta T + PL/A) + \gamma/4).$$

Which can be approximated by:

$$W = 0.5\epsilon(\Delta T - \Delta S).$$

with: $\Delta T = (T_1 - T)/T^*$ and $\Delta S = (S_1 - S)\alpha_S/\alpha_T T^*$

and $T^* = T_1 - T_A$

# Objectives

- Obtain a complete 2D picture of the sinking branch of the AMOC from a global eddy-permitting ocean model
- Contrast different depths to elucidate role overflows and GIN-Sea versus Labrador Sea
- Contrast 2D picture at constant depth with picture at constant rho
- Assess the typical spatial & temporal characteristics of the sinking branch in relation to variations in the atmospheric forcing fields and possible other dynamical controls
- Connect theoretical and modelling results to observational programs