Determination of diapycnal mixing rates in the World Ocean: implications for the Atlantic circulation

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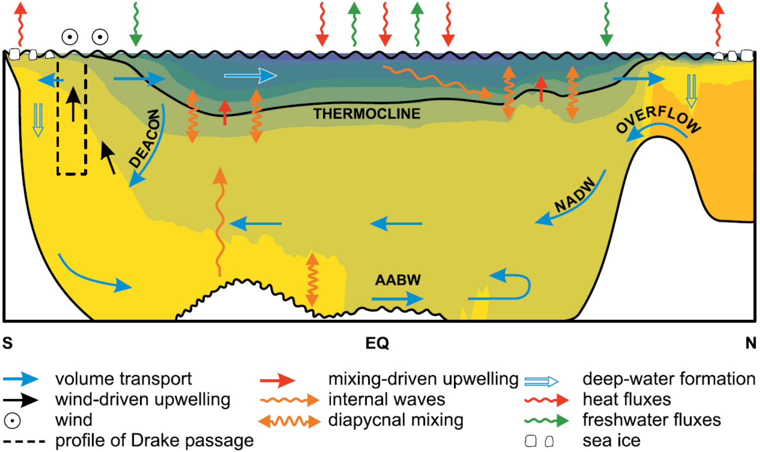
3-5th May 2017, AMOC meeting, IUEM

Outline

Idealized Atlantic MOC (from Kuhlbrodt et al. 2007)



scaling arguments show that $\Psi \propto D^{1/2} - D^{2/3}$

A brief review of methods used to infer diapycnal mixing

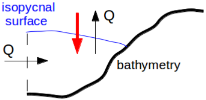
- **Curve fitting of advection-diffusion balance** provides e (diapycnal velocity) and D (diapycnal diffusivity) from the knowledge of $C(z)$ (vertical distribution of tracer) and Q (the production rate) (Munk; Munk & Wunsch) [works reasonably well in the ocean interior (away from boundaries) where isopycnals are mostly horizontal].

$$\overline{e}\frac{\partial \overline{C}}{\partial z} = \overline{D}\frac{\partial^2 \overline{C}}{\partial z^2} \tag{1}$$

With $\overline{e} = Q/A \sim 10^{-7}$ m s^{-1} we obtain $\overline{D} \sim 10^{-4}$ m^2 s^{-1} for a vertical scale height $(\overline{D}/\overline{e})$ of ~ 1 km. The overbar denotes an average over a broad region.

A brief review of methods used to infer diapycnal mixing

- **Large-scale mass budgets** (Ganachaud; Wunsch; Lumpkin and Speer) between a subset of hydrographic sections (WOCE). [Does not distinguish interior values from processes occurring at boundaries, sensitive to the choice of both the model and imposed constraints + unknown reference level + time-dependent observations].
- **Direct methods**: mass budget in a topographically blocked flow (Vema channel, Hogg et al.) [Provides accurate measure of diapycnal mixing, but only in very specific regions].



- **Adjoint techniques** (Forget 2015) provide 3D maps of eddy mixing coefficients. [Simulations typically integrated over 50-100 yr $\ll$ adjustment timescale of the deep ocean $\Rightarrow$ D at depth $\sim$ first guess value (low)].

A brief review of methods used to infer diapycnal mixing

- **Indirect estimates from measurements of local buoyancy frequency and microscale shear**: 1/ turbulent dissipation rate ϵ in isotropic turbulence ($7.5\nu(\partial u/\partial z)^2$), and 2/ "fine-scale" parametrizations of turbulent dissipation rate ϵ by internal waves breaking. [Accuracy of parametrizations ? fraction of ϵ available for mixing ($\propto R_f$ but usually taken as constant $\Gamma = 0.2$; $DN^2 = \Gamma\epsilon$, Osborn 1980) ?].
- **In-situ tracer release experiments**: diapycnal dispersion (diffusion equation) of a tracer released on an isopycnal surface (Ledwell) [Provides a local value, difficult to implement].
  $D_{Ledwell} \ll D_{Munk} \Rrightarrow$ "Missing mixing problem".

This study: global distribution of D

- Mass budget at the grid point level (1° horizontal resolution) of an OGCM $\leftrightarrow$ Basin-scale mass budgets (Ganachaud)

The central question of this study

Can we estimate diapycnal mixing by calculating the oceanic circulation with an OGCM forced by the time-mean Θ and S fields at $1^\circ \times 1^\circ$ resolution ?

Theory

At equilibrium, the generic tracer ϕ (either Θ or S) will obey:

$$\nabla.(\mathbf{u}\phi) = \frac{\phi_{CLIM} - \phi}{\tau_\phi} + C_\phi - \nabla.(\overline{\mathbf{u}'\phi'})$$

- $R_\phi = (\phi_{CLIM} - \phi)/\tau_\phi$: restoring to climatology
- C_ϕ : convection
- $\overline{\mathbf{u}'\phi'}$: **adiabatic** turbulent fluxes (isopycnal mixing + adiabatic version of baroclinic instability (GM))

Equations of conservation of Θ and S in γ_n coordinates

$$\begin{aligned} \mathbf{u_h^r}.\nabla_n\Theta + e\partial_z\Theta &= R_{\Theta,n} + C_{\Theta,n} + \nabla.(K_S\nabla\Theta)_n \quad (2a) \\ \mathbf{u_h^r}.\nabla_n S + e\partial_z S &= R_{S,n} + C_{S,n} + \nabla.(K_S\nabla S)_n \quad (2b) \end{aligned}$$

where e is the diapycnal velocity and the indice r states for the residual mean circulation.

$$e = w^r - \mathbf{u_h^r}.\nabla_n z \quad \text{under the small slope approximation.}$$

Forming the expression α times (2a) with minus β times (2b) and use the definition of neutral surfaces $\alpha\nabla_n\Theta - \beta\nabla_n S = 0$ and the buoyancy frequency $N^2 = g(\alpha\Theta_z - \beta S_z)$, we obtain

$$eN^2 = g[\alpha(R_\Theta + C_\Theta + I_\Theta) - \beta(R_s + C_s + I_S)]_n,$$

where I denotes the isopycnal mixing term.

Diapycnal mixing rates can finally be estimated using a simple adective-diffusive balance in the diapycnal direction

$$e = \partial_n (D \partial_z \gamma_n) \tag{3}$$

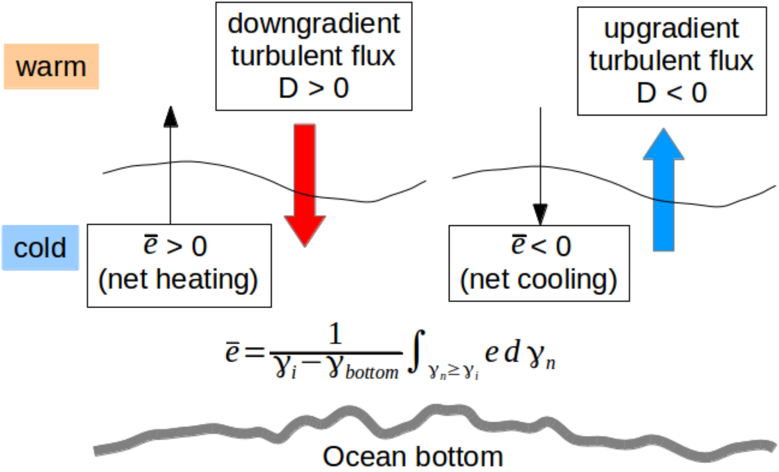
where we have neglected nonlinearities due to the fact that $N_z^2 \neq g(\alpha\Theta_{zz} - \beta S_{zz})$

Diapycnal mixing on the specific surface γ_n^i is thus calculated as

$$D(\gamma_n^i) = \frac{\int_{\gamma_n \geq \gamma_i} e d\gamma_n}{\partial_z \gamma_n|_{\gamma_n^i}}, \tag{4}$$

where a zero flux boundary condition is used at the bottom ($D\partial_z\gamma_n = 0$ at $z = -H(x, y)$)

The positiveness of D is not guaranteed by (4)



Model configuration

- MITgcm at 1° horizontal resolution, 44 vertical levels ($\delta z : 10m \rightarrow 250m$).
- Near global : extends from 80S to 80N
- Model brought to steady state (50 years spin-up) using 3D restoring to observed time-mean (annual mean) climatological Θ and S distributions (World Ocean Atlas 2009)
- Seasonal surface wind-stress (Large & Yeager)
- ETOPO-01 bathymetry dataset smoothed with a 100 km width Gaussian filter

- GM parameterization ($D_A = 500$ m^{2}s^{-1})
- Isopycnal mixing ($D_S = 1000$ m^{2}s^{-1})
- **No explicit vertical mixing** in the tracer equations ($D = 0$, since this is what we want to infer).

- Restoring timescale increases linearly with depth

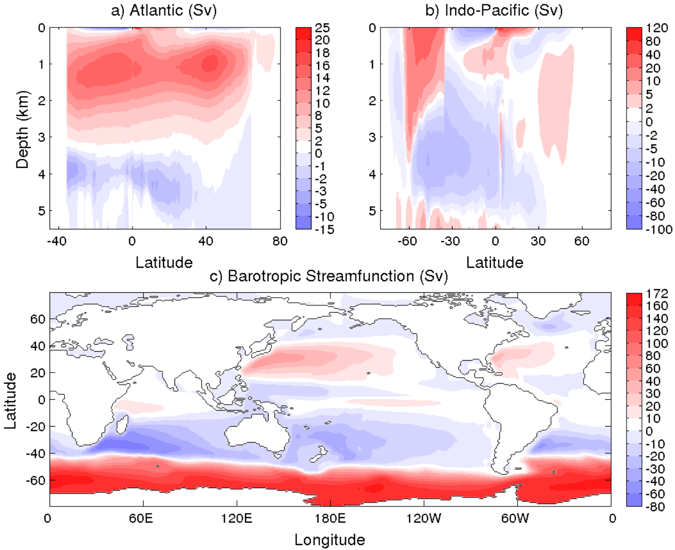
$$
\begin{aligned}
z < 40\ \mathrm{m}\ :\ \tau_\phi &= 2 \text{ months for } \Theta \text{ and } 1 \text{ year for } S \\
z > 40\ \mathrm{m}\ :\ \tau_\phi &= \tau_u + (\tau_b - \tau_u)\frac{z - z_u}{z_b - z_u}
\end{aligned}
$$

- Restoring timescales in the "surface mixed layer" tuned to match observed heat and freshwater fluxes ($\tau_\theta \neq \tau_s$)
- Below the surface mixed layer, turbulent eddies mix heat and salt in much the same way ($\tau_\theta = \tau_s$).
- Model uncertainty estimated from the sensitivity of model solutions to the bottom restoring timescale $\tau_b = (5 - 10)$ years.

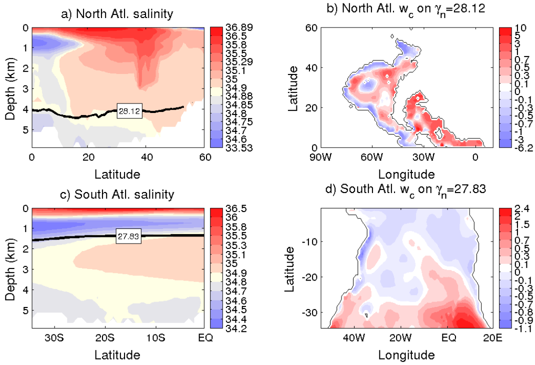
Link between τ_ϕ and diapycnal mixing

If $\tau_\phi \to \infty$ then $\mathbf{u}^{\mathbf{r}}.\nabla\phi \to 0$: adiabatic circulation

Structure of the circulation

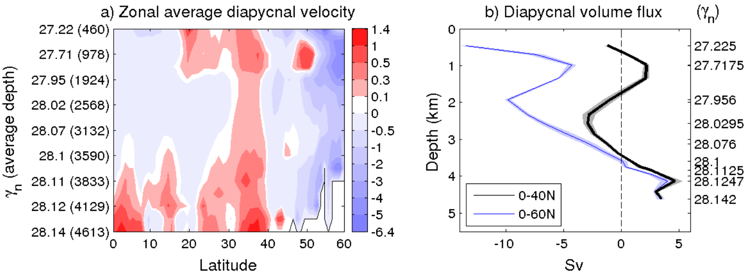


Diapycnal velocity on specific neutral surfaces



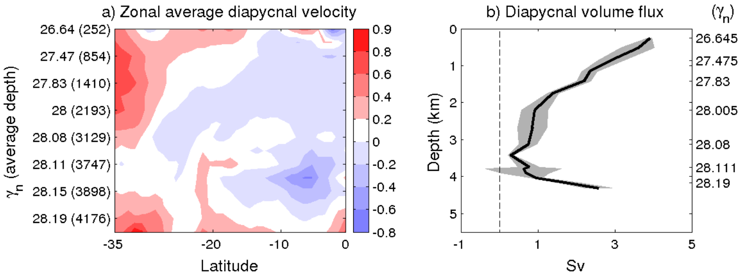
- Deep upwelling ($\times 10^{-6}$ m s^{-1}) found East of the North Atlantic Ridge where the AABW flows northward and is capped by topography in the North.
- Upwelling collocated with the northward flowing (cold and fresh) AAIW ($e \sim gN^{-2}(\alpha R_\Theta - \beta R_S)_n > 0$)

Zonal and isoneutral average diapycnal velocities ($\times 10^{-6}$ m s^{-1}) in the North Atlantic



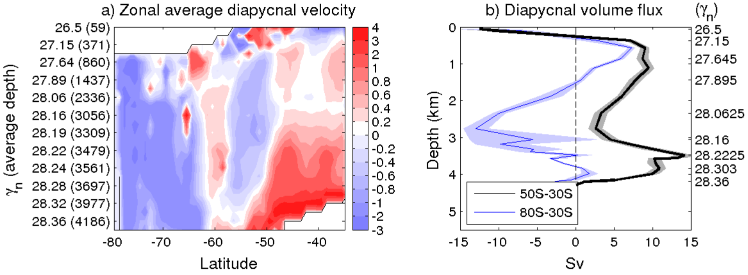
- Water mass conversion from light to heavy waters north of 40N ($e < 0$). Upwelling at nearly all latitudes below 3500 m ($\sim$ 5 Sv). Upward diapycnal mass transport throughout the entire water column at 30N-40N.

Zonal and isoneutral average diapycnal velocities ($\times 10^{-6}$ m s^{-1}) in the South Atlantic



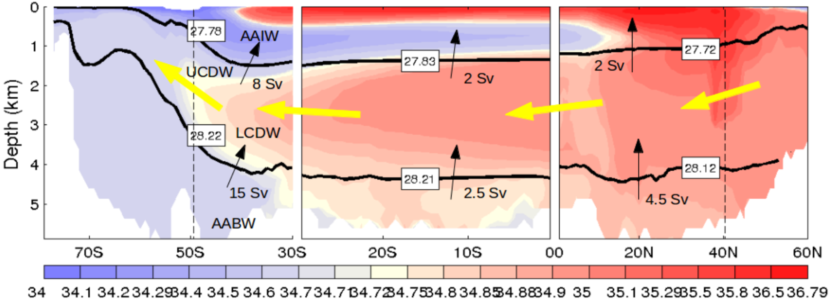
- AAIW upwelling with $e \sim 10^{-6}$ m s^{-1}. Upwelling of AABW below 3500 m between 35S and 10S. Downwelling elsewhere. Overall upwelling at all depths $\sim$ 1-3 Sv. [7 Sv AABW upwelling for the Atlantic basin (35S-60N) in agreement with Talley 2013]

Zonal and isoneutral average diapycnal velocities ($\times 10^{-6}$ m s^{-1}) in the Southern Ocean



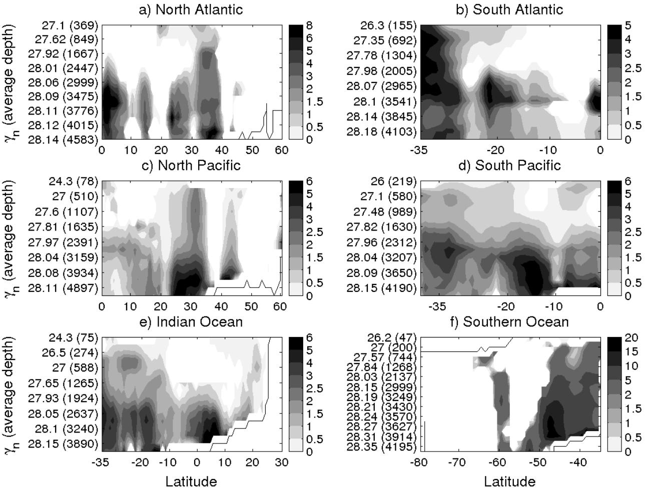
- Creation of heavy waters south of 50S with $e \sim -3 \times 10^{-6}$ m s^{-1}. Strong upwelling ($e \sim 4 \times 10^{-6}$ m s^{-1}) between 50S and 30S (15 Sv) and between LCDW and AABW. A second maximum occurs between UCDW and AAIW (8 Sv).

Summary of Atlantic circulation

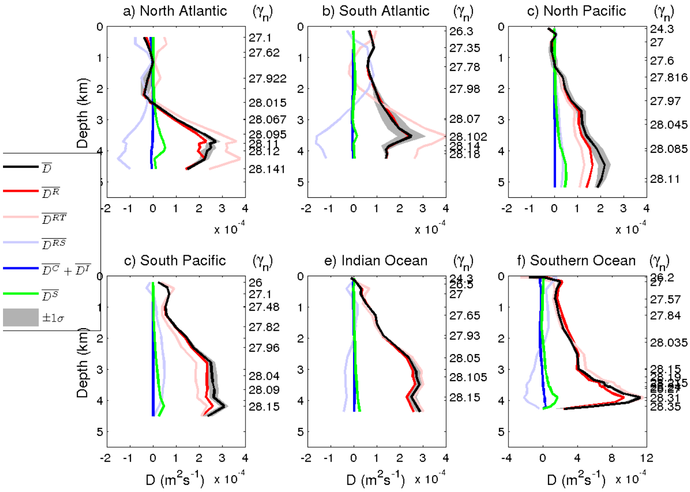


Diapycnal fluxes are large in the abyss and upper ocean and much weaker in the interior supporting the quasi-adiabatic character of the mid-depth overturning

Zonal and isoneutral average diapycnal diffusivivities ($\times 10^{-4} m^2 s^{-1}$) [negative values are white shaded]



Ensemble mean isoneutral average diapycnal diffusivities



- A new method has been proposed to infer (3D maps of) diapycnal velocities and diffusivities: tracer fields of an OGCM restored to a 3D time-mean climatology [Arzel and Colin de Verdière (2016), JPO, 46, 3751-3775]
- Advantage of the method: very easy to implement and very low computational cost (as compared to adjoint techniques for instance)
- Caveat: the restoring term contains contribution from (diabatic) turbulent eddy fluxes and model error
- Source of uncertainty: adiabatic diffusivities (K and K_{GM}) and (bottom) restoring timescale τ.

- Diapycnal diffusivities are low $O(10^{-5})$ m^{2}s^{-1} in the thermocline and increase below to Munk's values $O(2 \times 10^{-4})$ m^{2}s^{-1} in all ocean basins but in the Southern Ocean $O(12 \times 10^{-4})$ m^{2}s^{-1}.
- Values in good agreement with inverse models (Ganachaud, Lumpkin and Speer) and in the range of values inferred from local measurements (Waterhouse et al, 2014).

- Diapycnal mass transport in the Atlantic strong in the abyss (between AABW and NADW) and near the base of the thermocline, and weaker in the interior supports the quasi-adiabatic character of the mid-depth Atlantic overturning (theories from Vallis, Cessi)

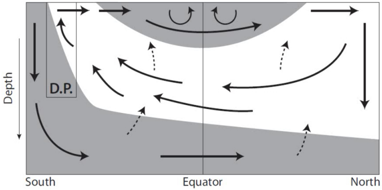


Figure: (figure from Vallis) Mid-depth cell (NABW) essentially adiabatic and results from the balance between wind-driven upwelling and deep water production in the north.