Diapycnal Mixing Deductions from the Large-Scale, Time-Mean, World Ocean Temperature-Salinity Distribution

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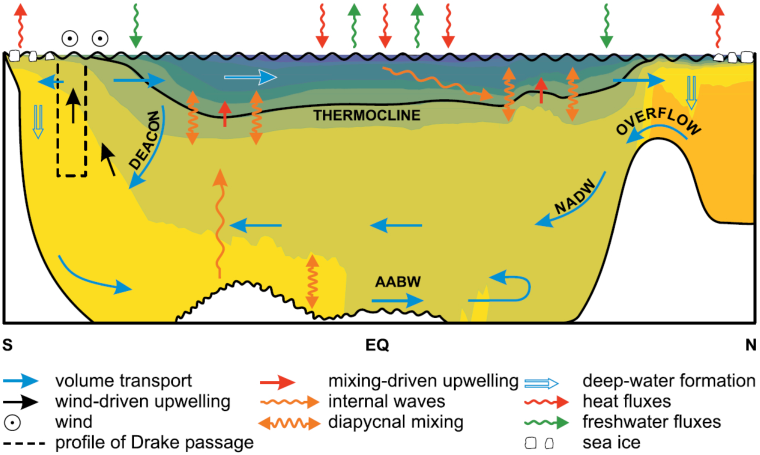
Idealized Atlantic MOC (from Kuhlbrodt)



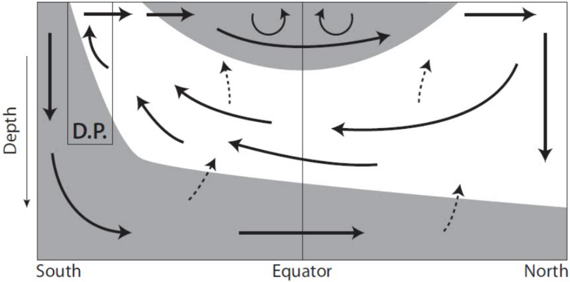
A brief history of the MOC vs diapycnal mixing

- 1950s: MOC supposed to be driven by buoyancy forcing and small scale mixing: thermocline and abyssal circulation theories (Robinson & Stommel, Stommel & Arons)
- 1960s: estimation of the deep mixing in the Pacific Ocean: Munk's "canonical" value 10^{-4} m^{2}s^{-1}.
- 1970s: Adiabatic ventilated thermocline (constrainted by Ekman pumping) above diffusive layer, MOC $\propto k_v^{2/3}$ under restoring BCs (Welander)
- 1980s: Measurements of the diffusivity (Ledwell, Gregg, Polzins, others) show that it is too small to drive the MOC ($\sim$ 30 Sv deep water production rate). The "missing mixing" problem !

A brief history of the MOC vs diapycnal mixing

- 1990s: Observations reveal a very different circulation than the classical one. Experiments with GCMs suggest a key role of the Southern Ocean in driving the mid-depth MOC: The "Drake-Passage effect" (Toggweiler & Samuels).
- 2000s: Strong mixing above rough topography (DIMES experiment, Naveira-Garabato).
- 2010s: A new paradigm emerged of a quasi-adiabatic mid-depth (1000-3000m) circulation with a key role of the ACC (Gnanadesikan, Henning & Vallis, Samelson, others).
- Very recently: Geographical distribution of diapycnal mixing in the upper 1000 m from a combination of observations and theories of small-scale turbulence. k_v highly non uniform (Waterhouse et al.)

A modern view of the interhemispheric MOC (from Vallis)



- deep cell (AABW) driven by a balance between small-scale mixing and combination of wind-driven upwelling and eddy-induced effects
- mid-depth cell (NABW) essentially adiabatic and results from the balance between wind-driven upwelling and deep water production in the north.

A brief review of methods used to infer diapycnal mixing

- Curve fitting of advection-diffusion balance provides w and k_v 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].
- Inverse modelling: strongly constrained problem based on large-scale (basin scale) mass budgets (Ganachaud; Wunsch). [Does not distinguish interior values from processes occurring at boundaries, sensitive to the choice of both the model and imposed constraints]
- 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]

A brief review of methods used to infer diapycnal mixing

- Indirect estimates from measurements of local buoyancy frequency and microscale shear: 1/ turbulent KE dissipation ϵ in isotropic turbulence, and 2/ parametrization of turbulent KE dissipation ϵ by internal waves breaking. [Accuracy of parametrizations ? fraction of ϵ available for mixing ($\propto$ *Ri* but usually taken as constant 0.2) ?]
- In-situ tracer release experiments: vertical dispersion (1D diffusion equation) of a tracer released on an isopycnal surface (Ledwell) [Provides a local value, difficult to implement]

This study: global distribution of k_v

- Mass budget at the grid point level (1° horizontal resolution) of an OGCM $\leftrightarrow$ Basin-scale mass budgets (Ganachaud)
- Signal over noise ratio ? Diapycnal velocity $w_c = O(1)$ m yr^{-1} (associated to a production rate of 30 Sv globally), while isopycnal vertical velocity $w = O(U\frac{\Delta z}{\Delta x}) = 300$ m yr^{-1} ($U = 1$ cm s^{-1}, slope of isopycnal$=10^{-3}$). So signal over noise ratio $\ll 1$.

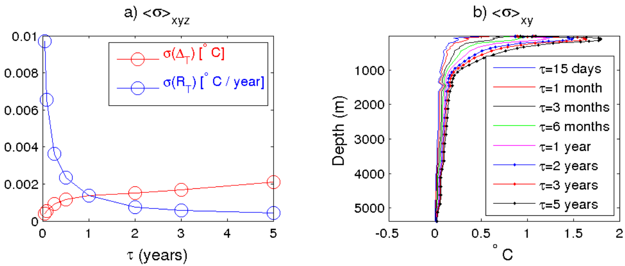
Can we estimate diapycnal mixing by calculating the oceanic circulation with an OGCM forced by the time-mean T ans S fields at $1^\circ \times 1^\circ$ resolution ?

Robust diagnostic method

- OGCM (MITgcm) brought to steady state (50 years spin-up) using restoring terms to observed time-mean climatological T and S distributions (WOA2009)
- Seasonal surface wind-stress (Large & Yeager)
- Strong restoring (2 months) in the upper 40 m to mimick observed heat exchanges ($\sim$ 30 W m^{-2} K^{-1}) in the surface mixed layer
- Weaker restoring (2 years) below (chosen to match the observed global ocean circulation - must be tuned).
- ETOPO-01 bathymetry dataset smoothed with a 100 km width Gaussian filter

Robust diagnostic method

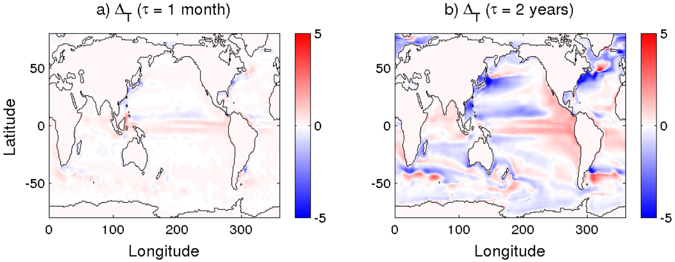
- No explicit diffusion terms in the tracer equations, but convection is retained in case of static instability (100 m^2s^{-1})
- Laplacian viscosity ($a_h = 5 \times 10^4 m^2s^{-1}$) chosen to solve barotropic Munk's boundary layer
- Vertical viscosity $a_v = 1$ cm^2s^{-1} (typical value)
- Quadratic bottom drag $C_D = 2 \times 10^{-3}$ (typical value)
- Model uncertainty estimated from the sensitivity of model solutions to the (deep) restoring timescale τ (1-3 years).



- Relatively weak sensitivity (Δ_T: model-data T difference and $R_T = \Delta_T/\tau$ restoring term) for $\tau > 1$ year and below 1000 m.
- Upper limit: $D\phi/Dt = R_\phi \to 0$ when $\tau \gg 1$ (adiabatic).

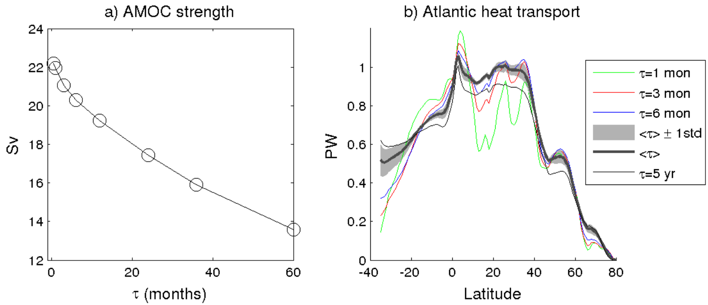
Restoring terms $\equiv$ convergence of eddy tracer fluxes

Calibration of τ: a) Thermodynamics



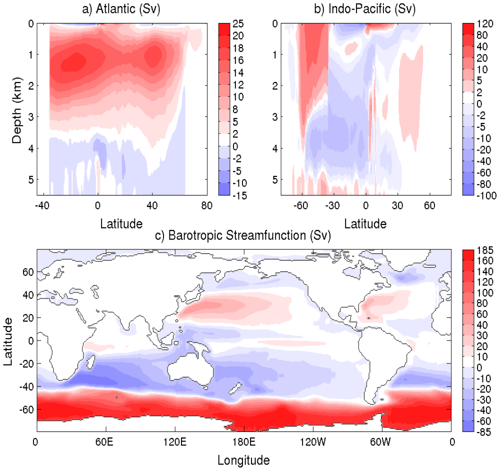
- Restoring terms (averaged in the upper 250 m) are typically nonzero in western boundary current regions and their eastward extension, fronts of the Southern Ocean, equatorial area.

Calibration of τ: b) Dynamics

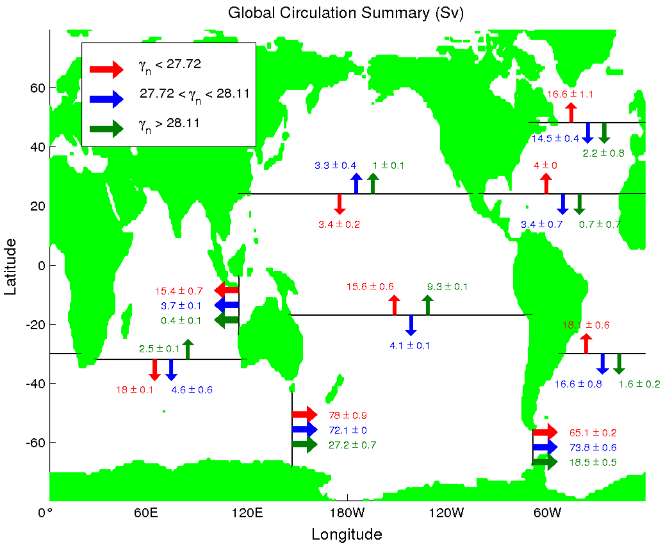


- Realistic Atlantic heat transport and circulation for $1 \leq \tau \leq 3$ years ($\sim$ range of values used by Sarmiento & Bryan; Lozier)

- Meridional circulation: Relatively weak mass transport in the 20-30N latitude band (Florida Strait), and absence of penetration of AABW in the Atlantic basin
- Barotropic circulation: Very good agreement with circulation inferred from ARGO floats displacements (in particular ACC, northern subpolar gyres, Ollitrault & Colin de Verdière) but too weak subtropical circulations (by a factor ∼2 in the northern hemisphere)



Mass transports (solution for $1 \leq \tau \leq 3$ years)



Global average $\overline{k_v}$: formulation

At equilibrium, equation governing the distribution of any tracer ϕ is

$$\nabla.(\mathbf{u}\phi) = R_\phi + C_\phi \tag{1}$$

Non-linear equation of state prevents us to compute divergence of density fluxes based on (1), but we are free to write

$$\nabla.(\mathbf{u}\sigma) = D_\sigma, \tag{2}$$

where σ is potential density (referenced to 0 db) and D_σ combines the diffusive and convective operators. Global horizontal average (overbar) yields

$$\frac{d\overline{w\sigma}}{dz} = \overline{D_\sigma} \tag{3}$$

Global average vertical mixing rates $\overline{k_v}$ obtained from advective diffusive balance, implying that

$$\overline{D_\sigma} = \frac{d}{dz}\left(\overline{k_v}\frac{d\overline{\sigma}}{dz}\right) \tag{4}$$

Global average $\overline{k_v}$: formulation

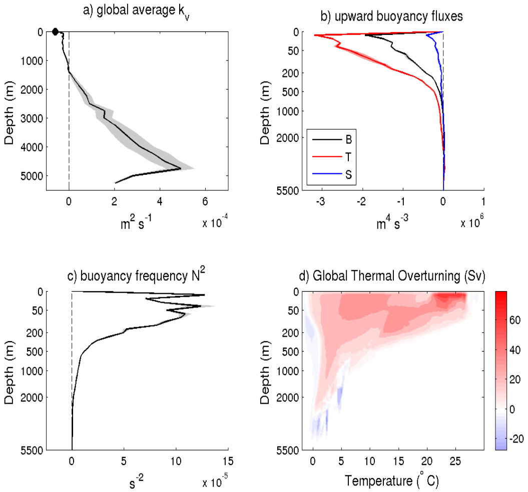
Integrating (3) from the bottom where buoyancy fluxes vanish, we get

$$\overline{k_v} = \frac{\overline{wb}}{\overline{N^2}}, \tag{5}$$

where $b = -g\sigma/\rho_0$ is the buoyancy and $N^2 = db/dz$ is the buoyancy frequency.

- This formulation provides *effective diffusivities* based on the knowledge of area-averaged density fluxes and buoyancy frequencies
- Result different from area-averaged mixing coefficients since co-variations of k_v and σ have been neglected in (4) ($\overline{k_v'\sigma'} \ll \overline{k_v}\overline{\sigma}$)
- The positiveness of $\overline{k_v}$ is not guaranteed by (5) (unlike inverse models, *a priori* bounded solutions)

- Upgradient mixing in the upper 1400 m (mean 2×10^{-5} m^{2}s^{-1}), downgradient mixing below (mean 2×10^{-4} m^{2}s^{-1})
- min value (-5×10^{-5} m^{2}s^{-1}) at 20 m depth (downward subtropical Ekman pumping)
- max value (5×10^{-4} m^{2}s^{-1}) at 4750 m depth
- upgradient mixing: volume of heavy (cold) waters that is transport upward is smaller than the volume of light (warm) water that is transported downward.



Global distribution $k_\rho(x, y, \gamma_n)$: formulation

We use the neutral density framework to estimate k_ρ: small isentropics displacements on γ_n do not produce restoring forces on the fluid parcel (as opposed to potential density surfaces). Neutral density surfaces defined as (McDougall)

$$\alpha \nabla_n \Theta - \beta \nabla_n S = 0, \tag{6}$$

where α and β are the thermal expansion and haline contraction coefficients computed from

$$\begin{aligned}
\rho &= \rho(S, \Theta, p, p_R) && \text{(7a)} \\
\alpha &= -\frac{1}{\rho} \frac{\partial \rho}{\partial \Theta}\bigg|_{S,p} && \text{(7b)} \\
\beta &= -\frac{1}{\rho} \frac{\partial \rho}{\partial S}\bigg|_{\Theta,p} && \text{(7c)}
\end{aligned}$$

Global distribution $k_\rho(x, y, \gamma_n)$: formulation

With those definitions, the buoyancy frequency simply becomes (McDougall 1987)

$$N^2 = g(\alpha\Theta_z - \beta S_z) \tag{8}$$

Tracer equations in γ_n coordinates are

$$\mathbf{u_h}.\nabla\Theta|_n + \Theta_z(w - \mathbf{u_h}.\nabla z|_n) = R_\Theta|_n + C_\Theta|_n \tag{9a}$$

$$\mathbf{u_h}.\nabla S|_n + S_z(w - \mathbf{u_h}.\nabla z|_n) = R_S|_n + C_S|_n \tag{9b}$$

Here w is the full vertical velocity and $\mathbf{u_h}.\nabla z|_n$ the iso-neutral contribution to the vertical velocity. Therefore we define the diapycnal velocity w_c as

$$w_c = w - \mathbf{u_h}.\nabla z|_n \tag{10}$$

Global distribution $k_\rho(x, y, \gamma_n)$: formulation

Now form the expression α times (9a) minus β times (9b) and use (6) and (8) to obtain

$$N^2 w_c = g[\alpha(R_\Theta|_n + C_\Theta|_n) - \beta(R_s|_n + C_s|_n)] \qquad (11)$$

Assuming that the diapycnal flux is equilibrated by a diffusive turbulent flux acting normal to the surface we get

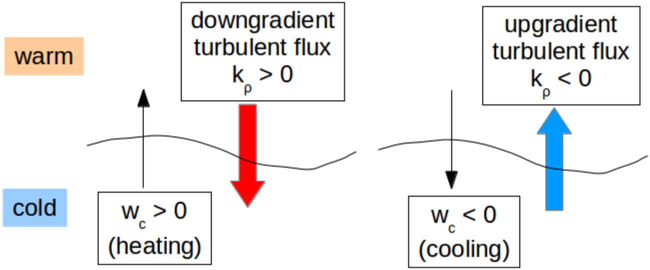
$$w_c = \partial_n(k_v \partial_z \gamma_n) \qquad (12)$$

where we have assumed a small slope approximation to assume the normal is in the z direction. Integrationg finally from the bottom upwards to an arbitrary neutral density level i using the no flux boundary condition at the bottom

$$k_v^i \partial_z \gamma_n^i = \int_{\gamma_n \geq \gamma_i} w_c d\gamma_n \qquad (13)$$

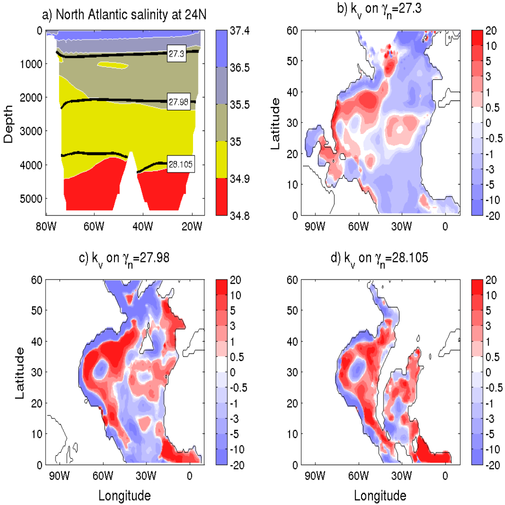
Global distribution $k_\rho(x, y, \gamma_n)$: remarks

- The positiveness of diapycnal mixing is not guaranteed by (13)
- Downgradient (upgradient) diapycnal mixing across a given neutral surface will occur wherever the vertically-integrated gain (loss) of buoyancy of the water column below that surface through heating (cooling) is larger (smaller) than the loss (gain) of buoyancy through salinification (freshening).



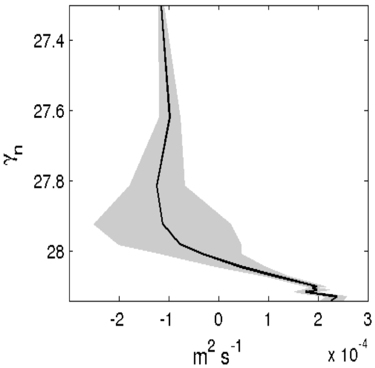
North Atlantic

- Downgradient diapycnal mixing enhanced in the Gulf Stream region and its eastward extension $O(10^{-3})$ m^2s^{-1}
- Relatively weak and upgradient diffusivities in the ocean interior in the upper ocean $O(10^{-4})$ m^2s^{-1}
- Strong downgradient mixing at bottom
- Signature of mid-Atlantic ridge clearly visible



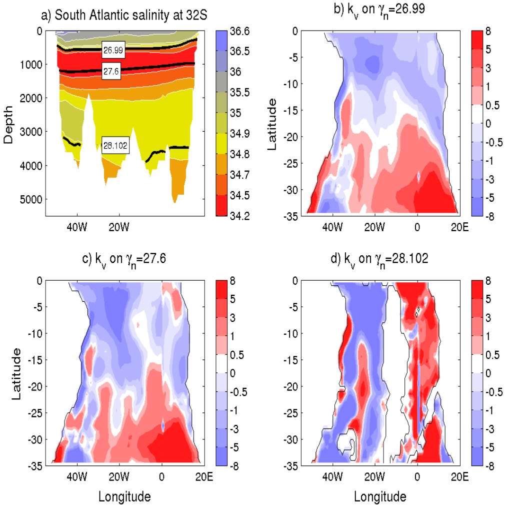
North Atlantic basin average

- Average mixing for the full water column is $0.6 \pm 0.4 \times 10^{-4}$ m^{2}s^{-1}
- Upgradient mixing found above 28.008 ($-0.9 \pm 0.7 \times 10^{-4}$ m^{2}s^{-1}) [might be biased by the misrepresentation of the Florida Strait current in the model]
- Downgradient mixing found below ($1.7 \pm 0.3 \times 10^{-4}$ m^{2}s^{-1})



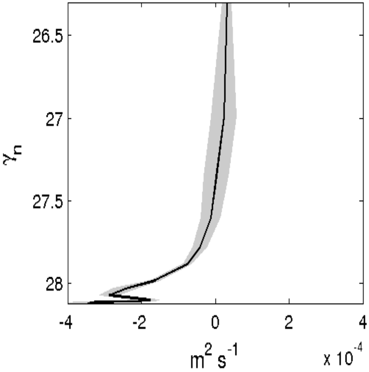
South Atlantic

- Large-scale tongue of downgradient mixing in the southwestern South Atlantic $O(10^{-3})$ m^{2}s^{-1} associated with the north-westward flow of AAIW into the basin
- Upgradient mixing found in the tropics
- Below the depth of NADW, strong upgradient (downgradient) mixing west (east) of the mid-Atlantic ridge [might be biased to to misrepresentation of AABW penetration into the basin]

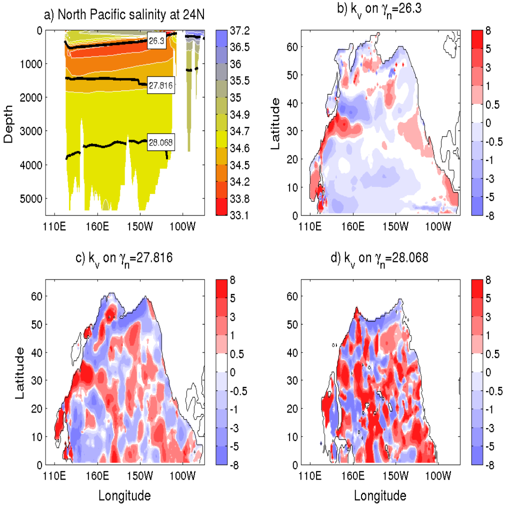


South Atlantic basin average

- Average mixing for the full water column is upgradient $1.4 \pm 0.3 \times 10^{-4}$ m^{2}s^{-1} (AABW bias)
- Peak value near bottom

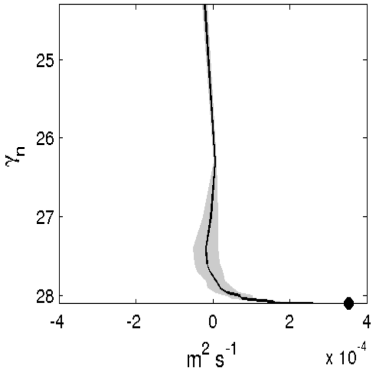


- Diapycnal mixing enhanced in the Kurushio region and its eastward extension $O(10^{-3})$ m^2s^{-1} (in agreement with indirect estimates from microstructure measurements)
- Patches of disorganized up- and downgradient diffusivities in the interior
- Strong downgradient mixing at bottom



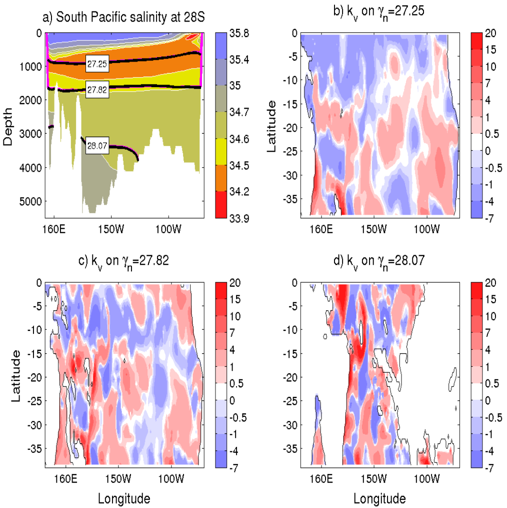
North Pacific basin average

- Average mixing for the full water column is $6 \pm 2 \times 10^{-5}$ $m^2 s^{-1}$
- Weak (not significant) upgradient mixing found above 27.7 ($1.4 \pm 2.4 \times 10^{-5}$ $m^2 s^{-1}$)
- Downgradient mixing found below ($1.1 \pm 0.2 \times 10^{-4}$ $m^2 s^{-1}$)
- Peak value on the deepest neutral surface (3.5×10^{-4} $m^2 s^{-1}$ on $\gamma_n = 28.11$)



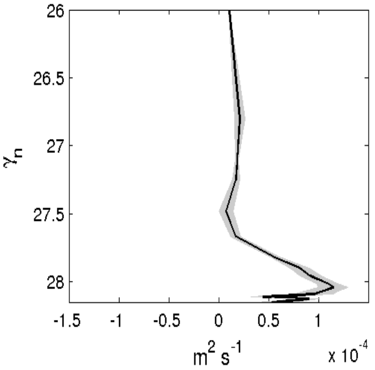
South Pacific

- Diapycnal mixing enhanced along western boundaries (East Australian Current, off New Zealand, and deeper) $O(10^{-3})$ m^{2}s^{-1}
- Patches of disorganized up- and downgradient diffusivities in the interior
- Downgradient mixing largely dominates at bottom

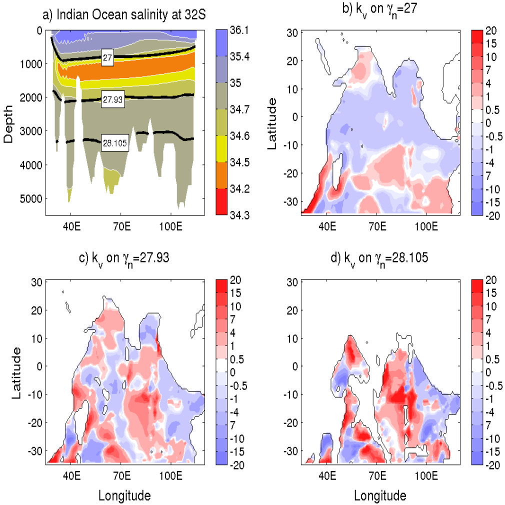


South Pacific basin average

- Average mixing for the full water column is $5.8 \pm 1 \times 10^{-5}$ m^2s^{-1}
- Average in thermocline waters and upper AAIW is $1.7 \pm 0.3 \times 10^{-5}$ m^2s^{-1}
- Downgradient mixing found below ($1.1 \pm 0.2 \times 10^{-4}$ m^2s^{-1})
- Peak value 1.1×10^{-4} m^2s^{-1} on $\gamma_n = 28.04$

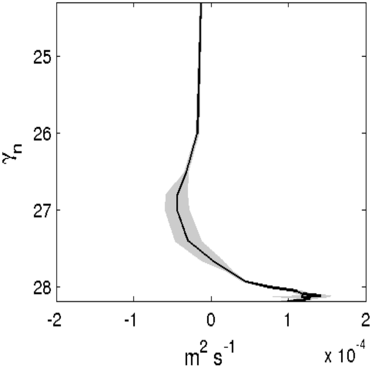


- Enhanced mixing along western boundary (Mozambique-Agulhas)
- Upgradient diffusivities in the tropics
- Signature of topograhic ridge at intermediaite depths
- Mixing of LCDW with CDW results in downgradient mixing at depth



Indian basin average

- Average mixing for the full water column is $4.2 \pm 1 \times 10^{-5}$ m^2s^{-1}
- Relatively weak upgradient mixing above 27.4 ($3 \pm 0.8 \times 10^{-5}$ m^2s^{-1})
- Relatively strong downgradient mixing below ($8.6 \pm 1 \times 10^{-5}$ m^2s^{-1})
- Peak value in the abyss (1.4×10^{-4} m^2s^{-1}) on $\gamma_n = 28.105$



Summary

- ... ongoing work: results are being written down ...
- A new method has been developped to infer diapycnal mixing rates at the global level
- This method makes use of a robust diagnostic simulation with restoring towards to observed distribution of temperature and salinity fields
- On the global average, weak (upgradient) mixing is found in the upper ocean (consistent with in-situ measurements and numerical simulations that indicates that heat is being pumped downward in the upper ocean), while strong (downgradient) diffusivities are found in the abyss.
- Strong spatial heterogeneity of diapycnal mixing: western boundary currents and fronts vs ocean interior

Summary

- Signature of major topographic features (mid-Atlantic ridge) clearly visible on diapycnal mixing
- Basin-scale average mixing usually indicates weak (upgradient) mixing in the upper $\sim$2000 m (baroclinic instability $\overline{w'b'} > 0$?) but strong (downgradient) mixing in the abyss, consistent with recent measurements (DIMES) and the current understanding of the MOC.