Deep Ocean Mixing

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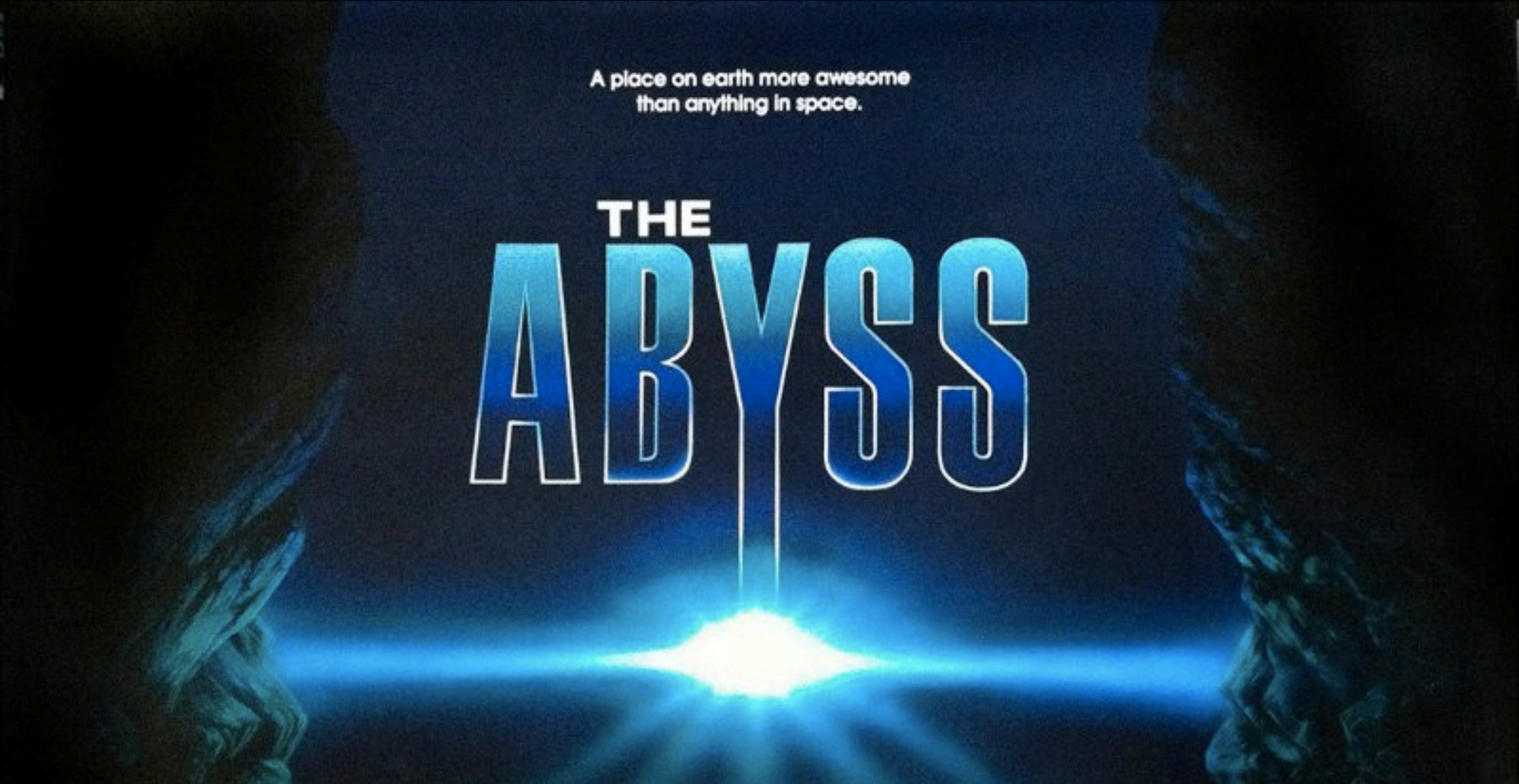
Les Houches, August 2017



Outline

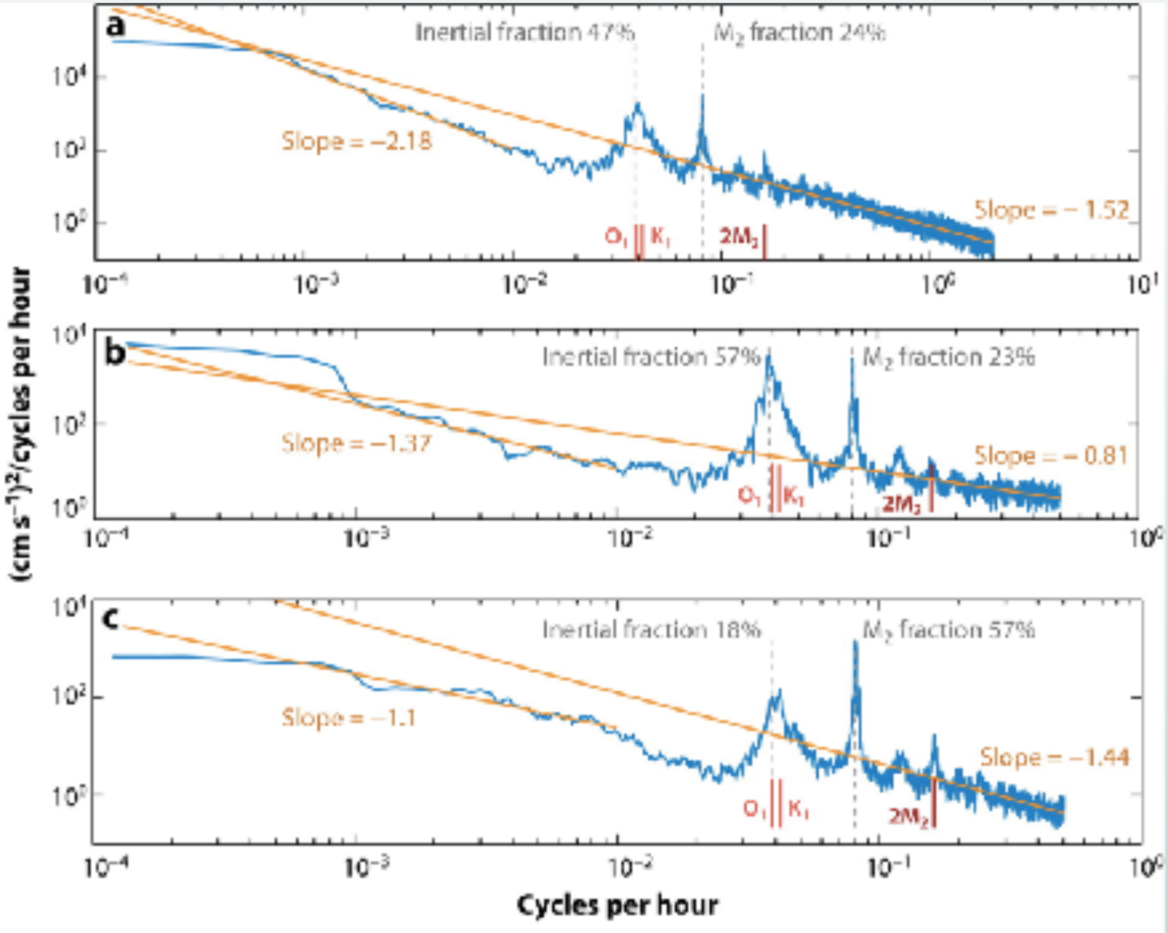
- Turbulence and diapycnal mixing

- Methods used to estimate ocean diapycnal mixing
  - temperature variance budget
  - turbulent kinetic energy budget
  - tracer release experiments

- Direct estimates of ocean diapycnal mixing
  - North Atlantic (shallow)
  - Brazil Basin (deep)
  - Drake Passage (mid-depth)

Turbulence and mixing

Turbulence in ocean interior

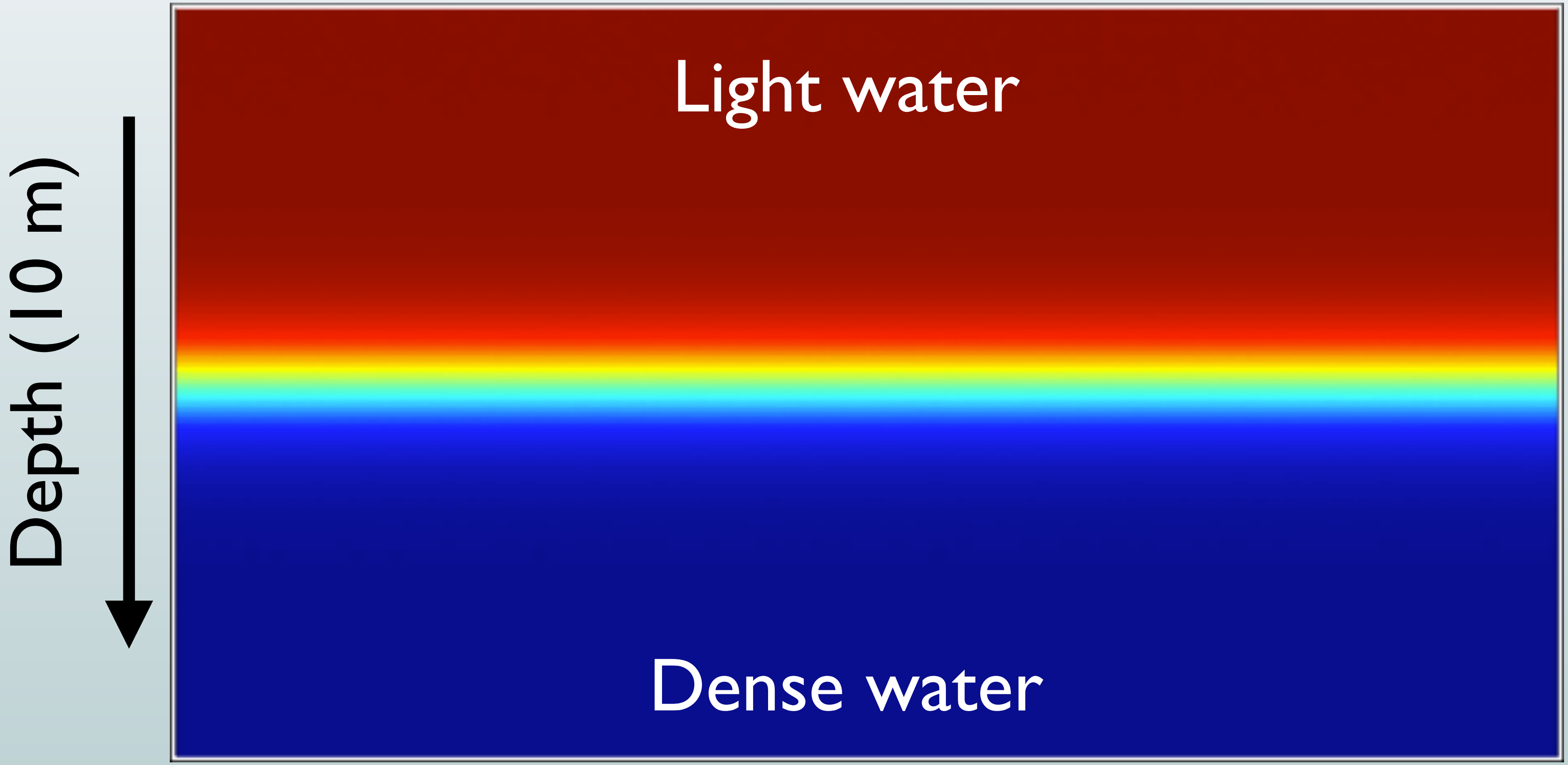


Ferrari R, Wunsch C. 2009.
Annu. Rev. Fluid Mech. 41:253–82

Subinertial motions have large Ri and small Ro

Superinertial motions can develop small Ri and large Ro

Breaking internal waves



Turbulent buoyancy fluxes

Boussinesq Equations

$$\partial_t \mathbf{u} + (\mathbf{u}\cdot\nabla)\mathbf{u} + f\hat{\mathbf{z}}\times\mathbf{u} = -{\rho_0}^{-1}\nabla p + b\mathbf{z} + \nabla\cdot(\nu\nabla\mathbf{u}),$$

$$\nabla\cdot\mathbf{u} = 0,$$

$$\partial_t\theta + (\mathbf{u}\cdot\nabla)\theta = \nabla\cdot(\kappa_\theta\nabla\theta),$$

$$\partial_t S + (\mathbf{u}\cdot\nabla)S = \nabla\cdot(\kappa_S\nabla S),$$

$$b = g\alpha(\theta-\theta_0) - g\beta(S-S_0), \quad \rho = \rho_0\left(1 - g^{-1}b\right)$$

$$\implies \quad \partial_t b + (\mathbf{u}\cdot\nabla)b = \nabla\cdot(\kappa_\theta g\alpha\nabla\theta - \kappa_S g\beta\nabla S)$$

Reynolds decomposition

Buoyancy budget

$$\partial_t b + (\mathbf{u} \cdot \nabla) b = \nabla \cdot (\kappa_\theta g \alpha \nabla \theta - \kappa_S g \beta \nabla S)$$

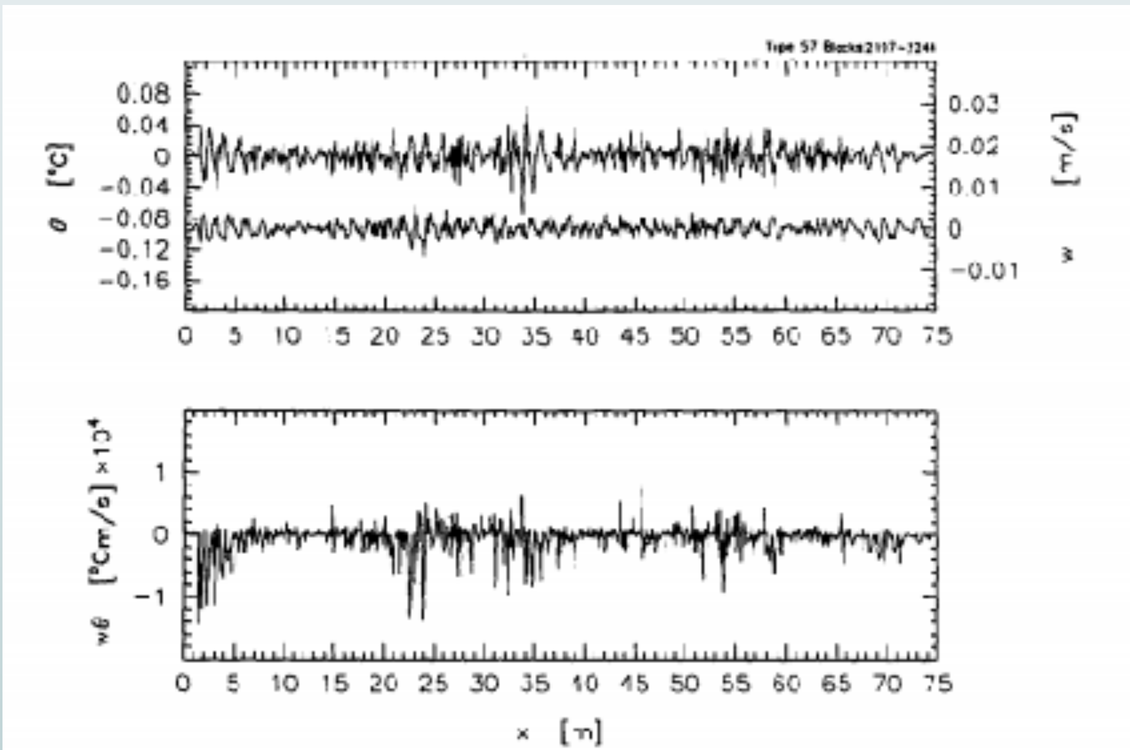
Subinertial buoyancy budget

$$\begin{aligned} \partial_t \bar{b} + (\bar{\mathbf{u}} \cdot \nabla) \bar{b} &= -\nabla \cdot (\overline{\mathbf{u}' b'} - \kappa_\theta g \alpha \nabla \bar{\theta} + \kappa_S g \beta \nabla \bar{S}) \\ &\simeq -\nabla \cdot \overline{\mathbf{u}' b'} \\ &\simeq -\partial_z \overline{w' b'} \end{aligned}$$

Measurements of turbulent buoyancy fluxes

I. Direct eddy correlations

- Compute w and b (θ and S)
- High pass signals to extract super-inertial signals w' and b'
- Compute correlations by averaging over time $\overline{w'b'}$



Fleury and Lueck, 1994

- Challenge I: measure w and b (θ and S) at same location from stable platform
- Challenge II : average over long enough time

II. Temperature variance budget

- Potential temperature budget

$$\partial_t \theta + (\mathbf{u} \cdot \nabla)\theta = \nabla \cdot (\kappa_\theta \nabla \theta)$$

- Mean and eddy potential temperature budget

$$\partial_t \bar{\theta} + (\bar{\mathbf{u}} \cdot \nabla)\bar{\theta} = \nabla \cdot (-\overline{\mathbf{u}'\theta'} + \kappa_\theta \nabla \bar{\theta})$$

$$\partial_t \theta' + ((\bar{\mathbf{u}} + \mathbf{u}') \cdot \nabla)\, \theta' + \mathbf{u}' \cdot \nabla \bar{\theta} = \nabla \cdot (\overline{\mathbf{u}'\theta'} + \kappa_\theta \nabla \theta')$$

- Potential temperature variance budget

$$\frac{1}{2}\partial_t \overline{\theta'^2} + \frac{1}{2}\nabla \cdot \left[\overline{(\bar{\mathbf{u}} + \mathbf{u}')\theta'^2} - \kappa_\theta \nabla \overline{\theta'^2}\right] = -\overline{\mathbf{u}'\theta'} \cdot \nabla \bar{\theta} - \kappa_\theta \overline{|\nabla \theta'|^2}$$

II. Temperature variance budget

- Assuming that turbulence in stationary, homogeneous and isotropic:

$$\cancel{\frac{1}{2}\partial_t \overline{\theta'^2}} + \frac{1}{2}\nabla \cdot \left[\cancel{\overline{(\bar{\mathbf{u}} + \mathbf{u}')\theta'^2} - \kappa_\theta \nabla \overline{\theta'^2}}\right] = -\overline{\mathbf{u}'\theta'} \cdot \nabla\bar{\theta} - \kappa_\theta \overline{|\nabla\theta'|^2}$$

$$-3\,\kappa_\theta \overline{(\partial_z \theta')^2}$$

stationarity homogeneity isotropy

- Assuming that $\nabla\bar{\theta} \simeq \partial_z \bar{\theta}\ \hat{\mathbf{z}}$

$$\overline{w'\theta'} \simeq -\frac{3\kappa_\theta \overline{(\partial_z \theta')^2}}{\partial_z \bar{\theta}} \equiv -\frac{1}{2}\frac{\chi}{\partial_z \bar{\theta}}$$

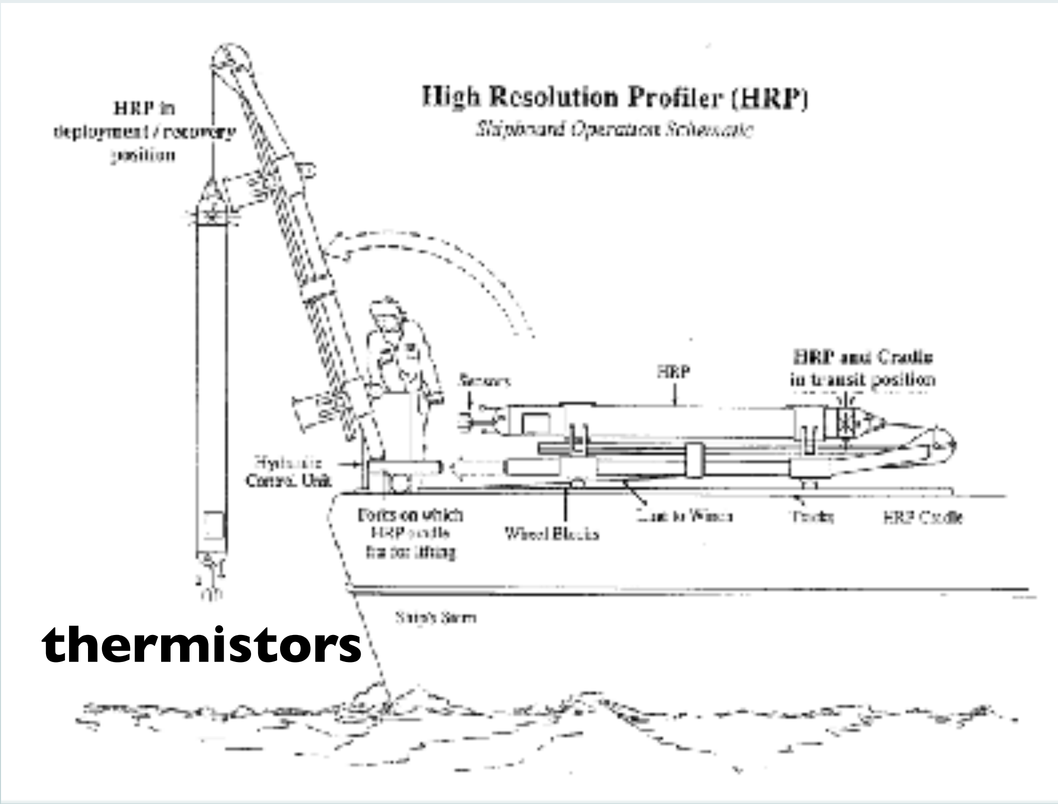
- Assuming that temperature dominates buoyancy gradients

$$\overline{w'b'} \simeq g\,\alpha\,\overline{w'\theta'} \simeq -\frac{g\alpha}{2}\frac{\chi}{\partial_z \bar{\theta}}$$

(*Osborn-Cox, 1972*)

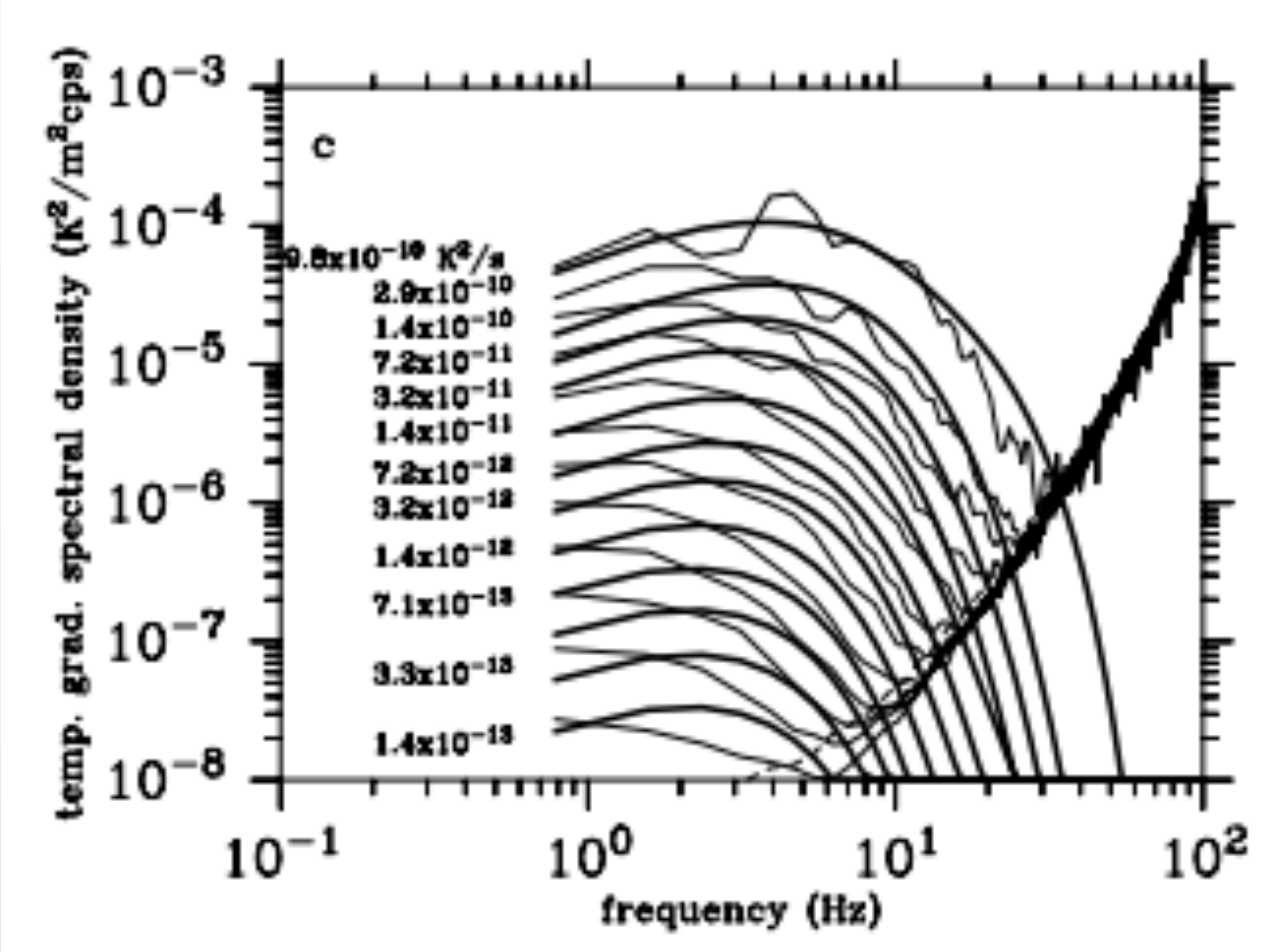
Microstructure thermistors

High Resolution Profiler



Microstructure thermistors

Temperature gradient spectra (χ)



III. Kinetic energy budget

- Full turbulent kinetic energy budget

$$\frac{1}{2}\partial_t\overline{|\mathbf{u}'|^2}+\frac{1}{2}\nabla\cdot\left[\overline{(\bar{\mathbf{u}}+\mathbf{u}')|\mathbf{u}'|^2}-\nu\nabla\overline{|\mathbf{u}'|^2}+\rho_0^{-1}\overline{p'w'}\right]=$$
$$=-\overline{\mathbf{u}'\mathbf{u}'}\cdot\nabla\bar{\mathbf{u}}+\overline{w'b'}-\nu\overline{|\nabla\mathbf{u}'|^2}$$

- Turbulent kinetic energy budget for stationary, homogeneous, isotropic turbulence

$$-\overline{\mathbf{u}'\mathbf{u}'}\cdot\nabla\bar{\mathbf{u}}+\overline{w'b'}\simeq\frac{15}{2}\nu\overline{(\partial_z u')^2}$$

- Assuming that $\nabla\bar{\mathbf{u}}\simeq\partial_z\bar{\mathbf{u}}_h\ \hat{\mathbf{z}}$

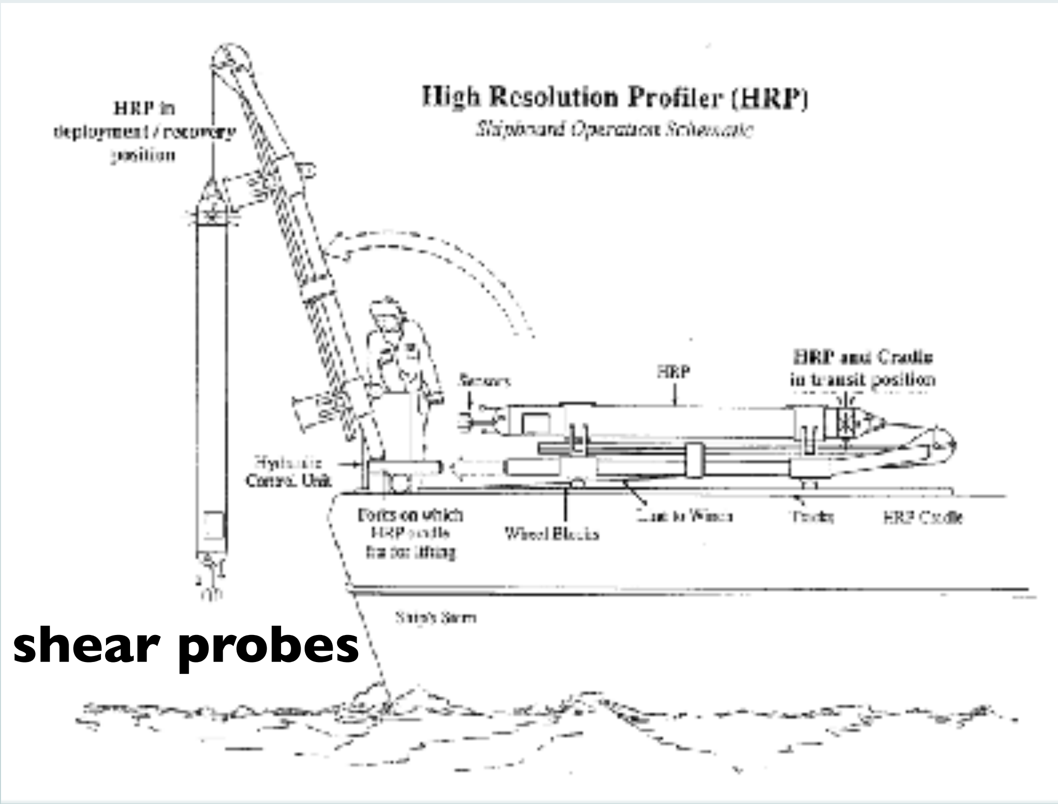
$$-\overline{\mathbf{u}'_h w'}\cdot\partial_z\bar{\mathbf{u}}_h+\overline{w'b'}\simeq\frac{15}{2}\nu\overline{(\partial_z u')^2}$$

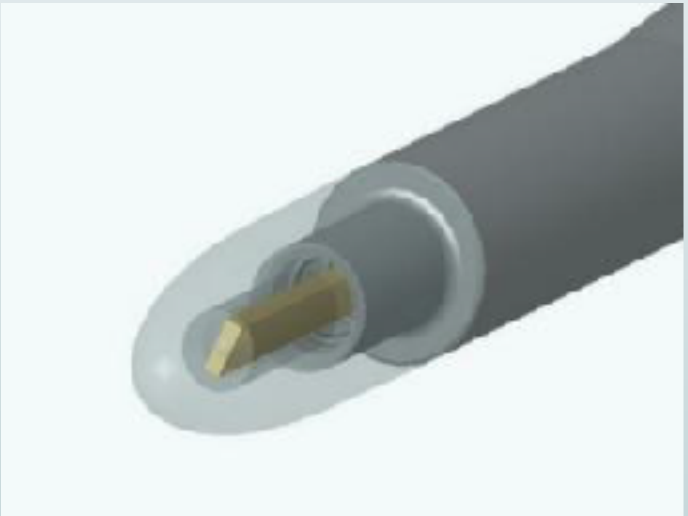
- Introducing the flux Richardson number (*Osborn, 1981*)

$$Ri_f\equiv\frac{\overline{w'b'}}{\overline{\mathbf{u}'_h w'}\cdot\partial_z\bar{\mathbf{u}}_h}\qquad\Longrightarrow\qquad\overline{w'b'}\simeq-\frac{Ri_f}{1-Ri_f}\frac{15}{2}\overline{(\partial_z\mathbf{u}')^2}\equiv-\Gamma\epsilon$$

Microstructure shear probes

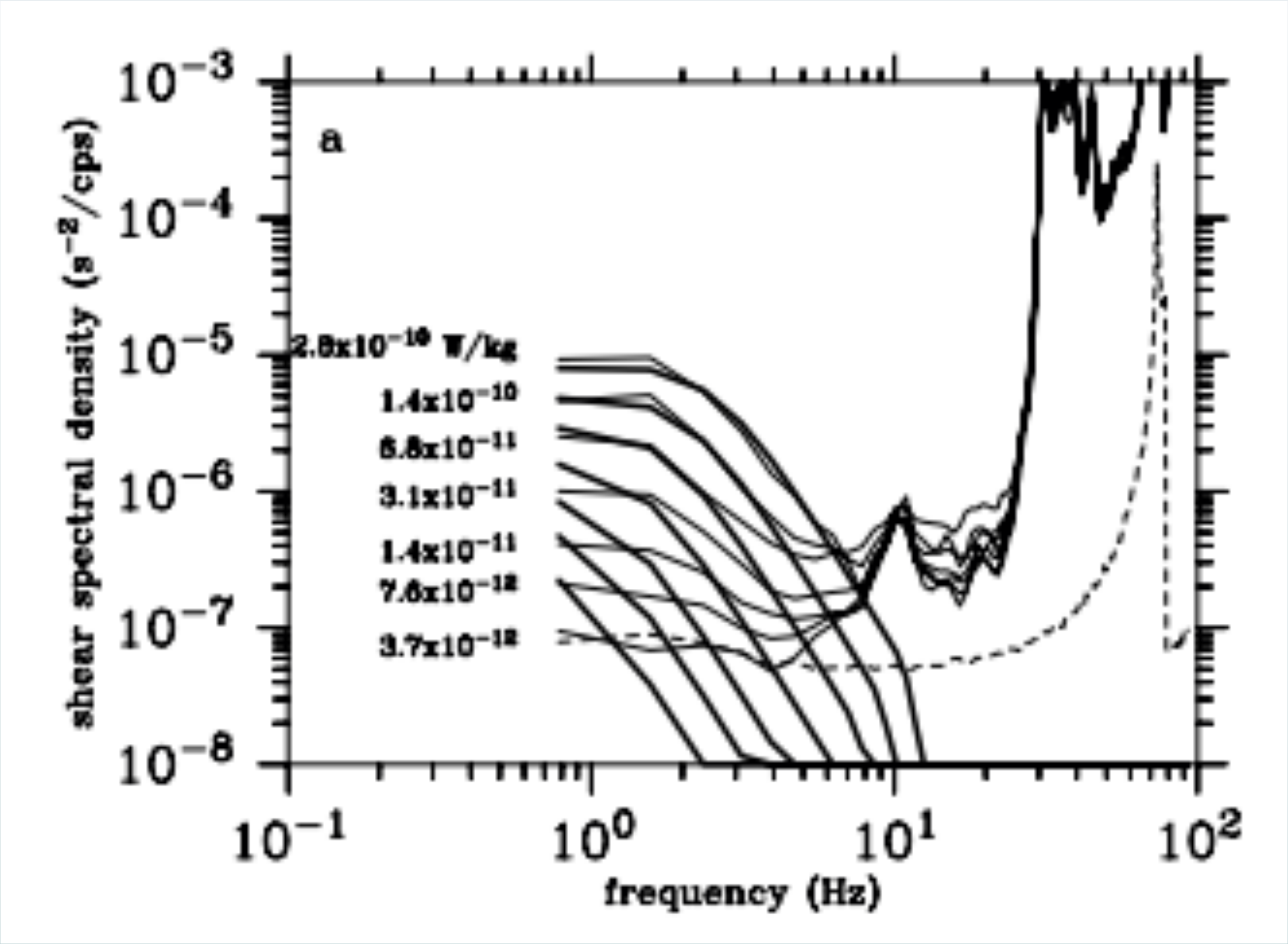
High Resolution Profiler





Microstructure shear probes

Turbulent kinetic energy dissipation (ϵ)



Mixing coefficient

Temperature variance budget: $\overline{w'b'} = -\frac{g\alpha}{2}\frac{\chi}{\partial_z\bar{\theta}}$

Turbulent kinetic energy budget: $\overline{w'b'} = -\Gamma\epsilon$

$$\implies \Gamma\epsilon = \frac{g\alpha}{2}\frac{\chi}{\partial_z\bar{\theta}} \qquad \implies \Gamma = \frac{g\alpha}{2\partial_z\bar{\theta}}\frac{\chi}{\epsilon} = 0.1 - 0.3$$

Table 4 $\gamma_{\chi\epsilon}$ **in pycnoclines from direct** χ_T **and** ϵ **measurements, mostly by profiling**

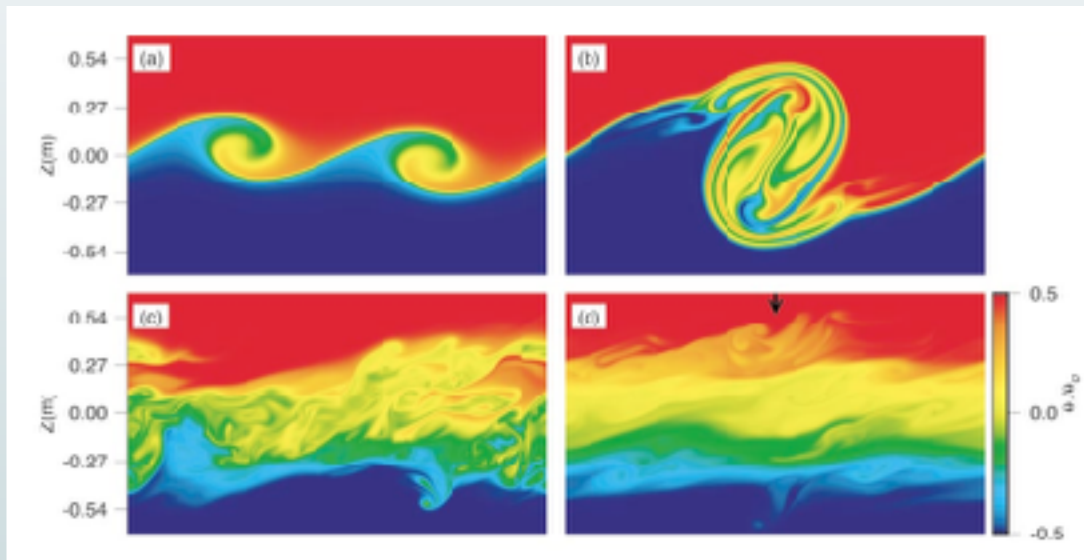
Location	R_ρ	Re_b	Γ	Reference
Rockall Trough	>7	300–3,000	0.05–0.32	Oakey 1982, 1985[a]
California Current	Variable		0.18[b]	Gregg et al. 1986
Equator, 140°W[c]	Variable	1–10^6	0.12[b]	Peters & Gregg 1988
Equator, 140°W[c]	—	—	0.12–0.48[d]	Moum et al. 1989
California Current	—	—	0.05[e]	Yamazaki & Osborn 1993
~~Admiralty Inlet~~	~~0.05~~	~~$\approx 2\times10^4$~~	~~0.58[f]~~	~~Seim & Gregg 1994~~
Tidal channel	—	—	0.25, 0.23[g]	Gargett & Moum 1995
Equator, 140°W[h]	—	20–10^5	0.14[h]	Peters et al. 1995
Northeast Pacific			≈ 0.3–0.4	Moum 1996
Northeast Atlantic	−0.8 to 5	30–4,000	0.14, 0.21[i]	Ruddick et al. 1997
Northeast Atlantic[j]	−100 to −1	—	0.16 ± 0.04	St. Laurent & Schmitt 1999
	>1	—	0.2–0.3	
~~Monterey shelf[k]~~			~~0.0022~~	~~Gregg & Horne 2009~~
~~Equator, 80.5°E[k]~~			~~<0.02~~	~~Fujiwara et al. 2015~~

Gregg et al., 2018

Eddy diffusivities

Diffusive closure: $\overline{w'b'} = -K_T \partial_z \bar{b}$

$\partial_z \bar{b}$



$\overline{w'b'}$

Temperature variance budget

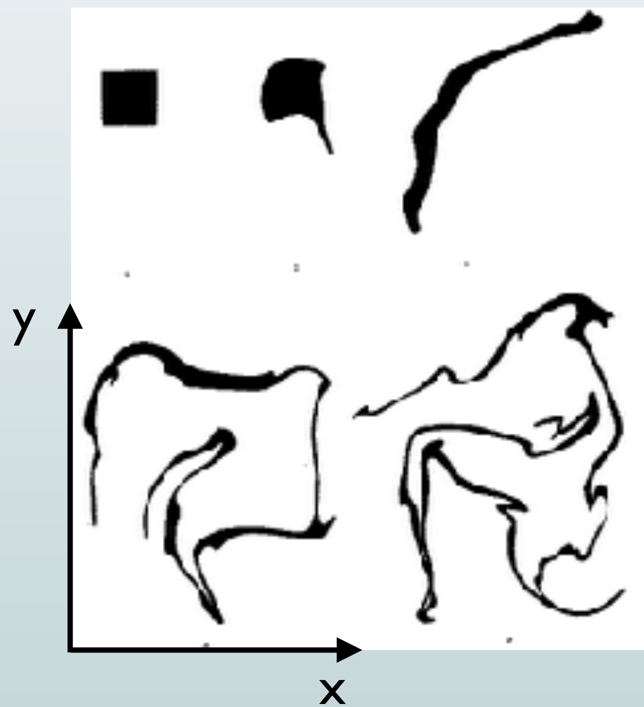
$$\overline{w'b'} = -\frac{g\alpha}{2}\frac{\chi}{\partial_z \bar{\theta}} \quad \Longrightarrow \quad K_T^{\chi} = \frac{1}{2}\frac{\chi}{(\partial_z \bar{\theta})^2}$$

Turbulent kinetic energy budget

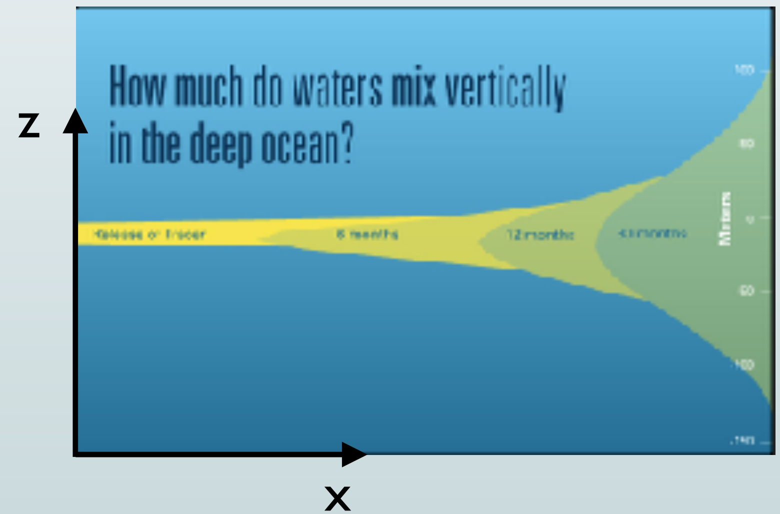
$$\overline{w'b'} = -\Gamma \epsilon \quad \Longrightarrow \quad K_T^{\epsilon} = \Gamma \frac{\epsilon}{\partial_z \bar{b}}$$

IV. Tracer release experiments

Lateral stirring by geostrophic motions



Vertical mixing by breaking waves



$$\partial_t \bar{c} + \bar{\mathbf{u}} \cdot \nabla \bar{c} = -\nabla \cdot \overline{\mathbf{u}' c'}$$
$$= \nabla \cdot K_T \nabla \bar{c}$$

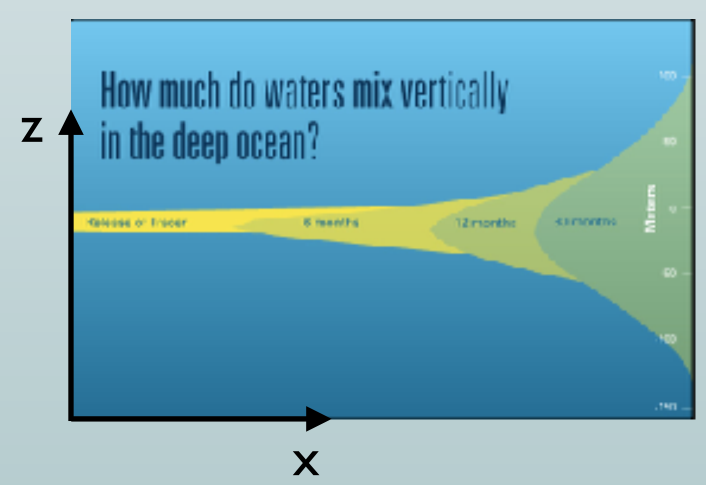
IV. Tracer release experiments

- The eddy diffusivity K_T is inferred from the evolution of the tracer z-moments

$$Z_1 \equiv \frac{\langle zc \rangle}{\langle c \rangle}, \qquad Z_2 \equiv \frac{\langle (z - Z_1)^2 c \rangle}{\langle c \rangle}, \qquad \langle \cdot \rangle = \iiint \cdot \, \mathrm{d}V$$

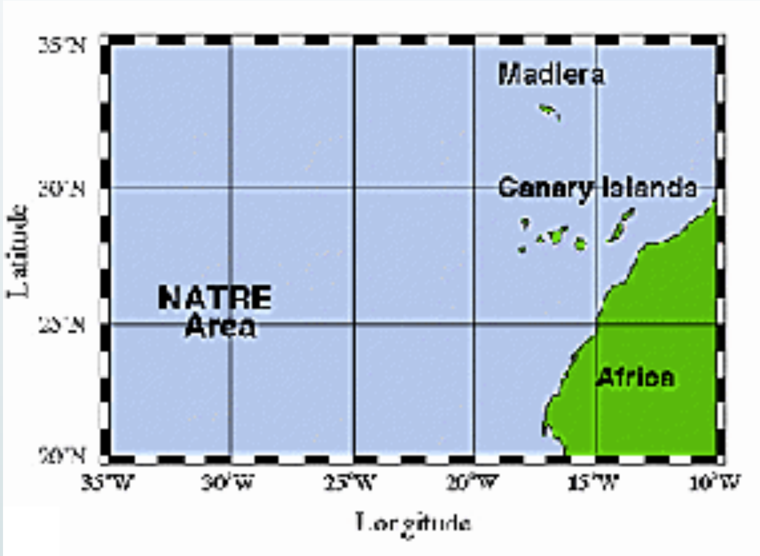
- Assuming that
  - isopycnals are flat (can be relaxed using isopycnal coordinates)
  - the stirring velocity $\overline{\boldsymbol{u}}$ is along isopycnals ($\overline{w}=0$)
  - K_T is independent of depth z

$$\frac{dZ_1}{dt} = 0$$
$$\frac{dZ_2}{dt} = 2\frac{\langle K_T c \rangle}{\langle c \rangle}$$

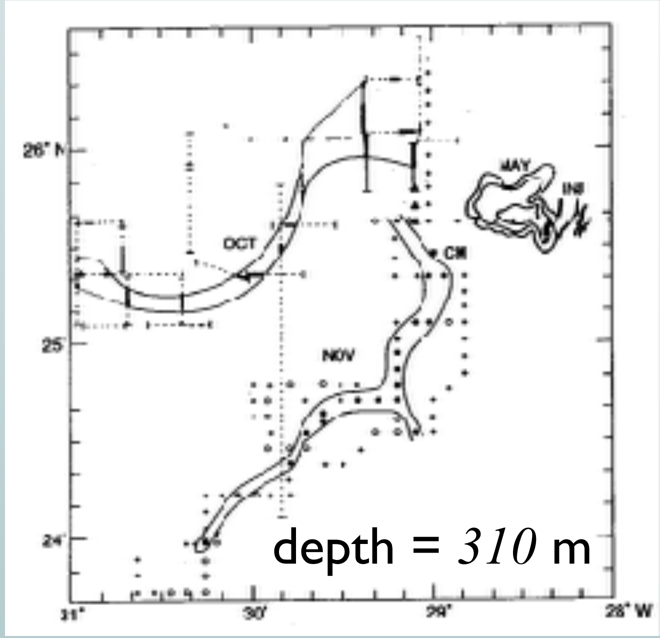


North Atlantic Tracer Release Experiment (NATRE)

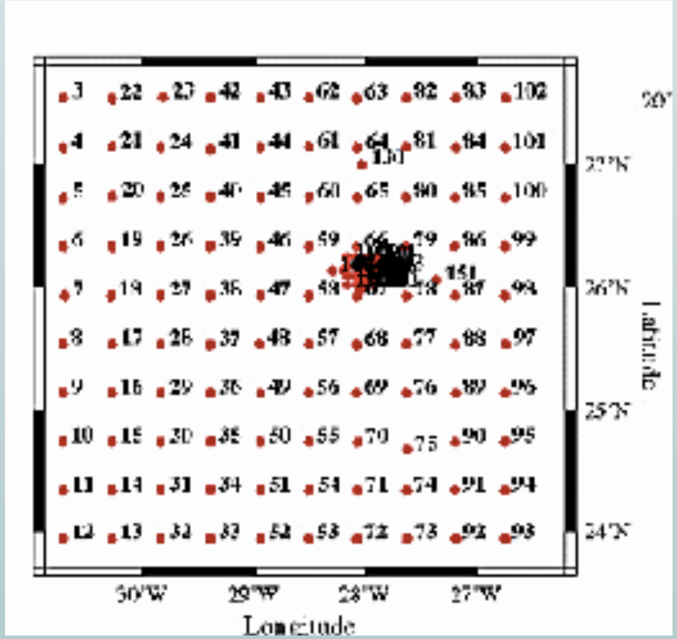
NATRE



Tracer release (*Ledwell et al., 1993*)



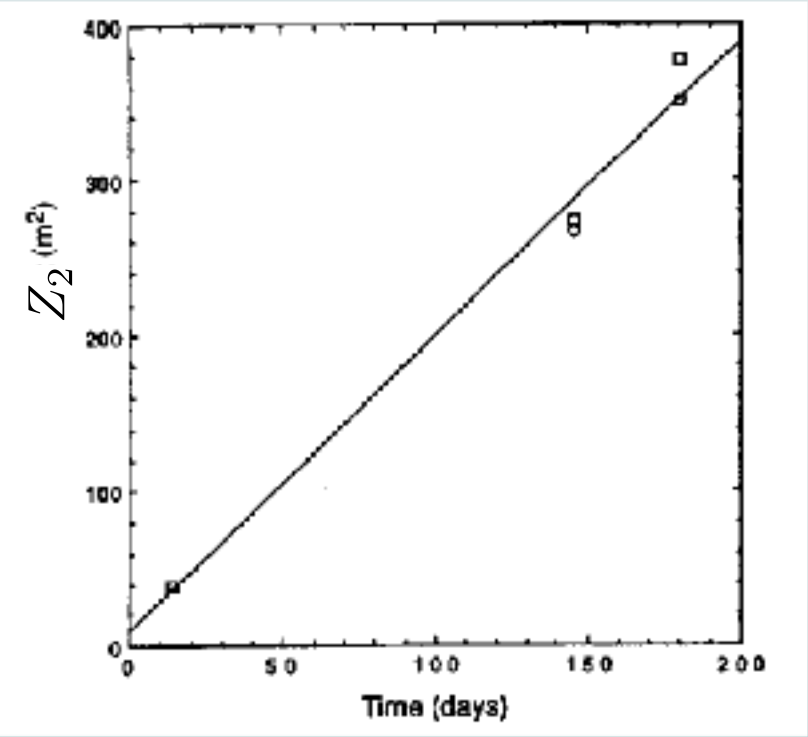
HRP profiles (*Toole et al., 1994*)



Three methods are consistent

Tracer release (*Ledwell et al., 1993*)

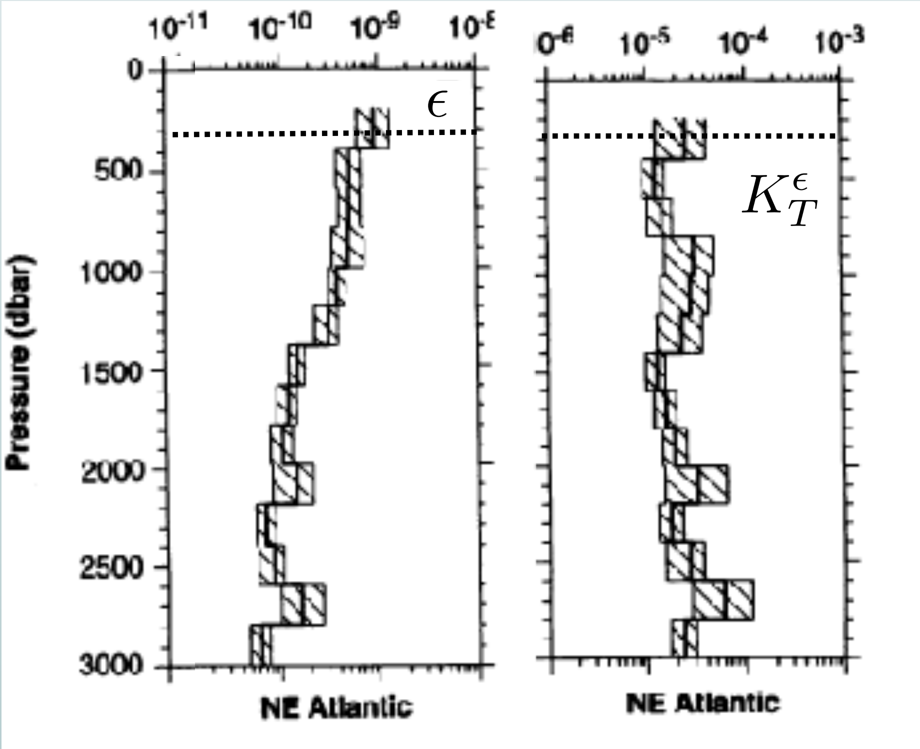
$$K_T^{tracer} = (1.1 \pm 0.2) \times 10^{-5}\mathrm{m}^2\ \mathrm{s}^{-1}$$



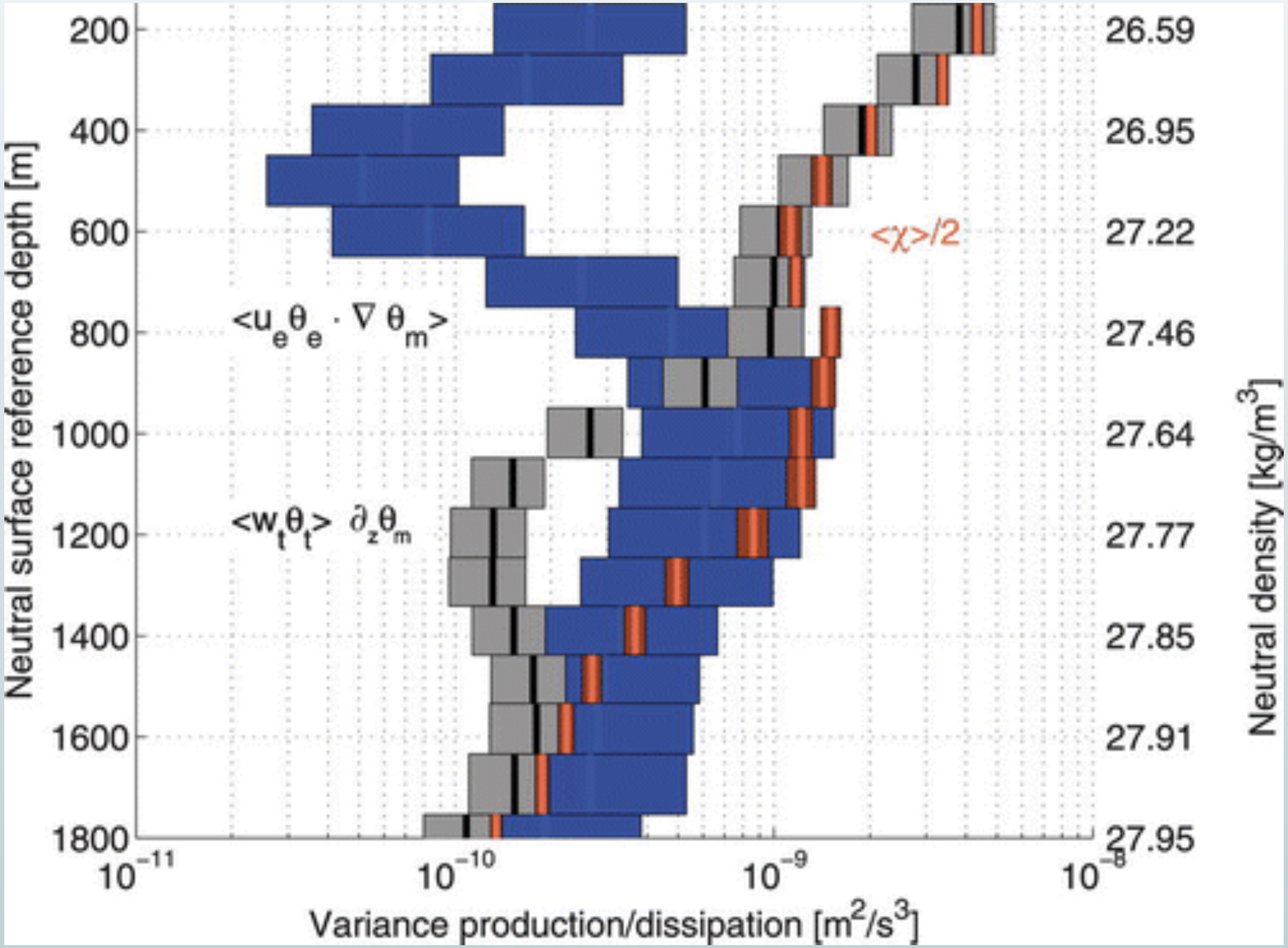
HRP profiles (*Toole et al., 1994*)

$$K_T^{\chi} = \frac{1}{2}\frac{\chi}{(\partial_z\bar{\theta})^2} = (1.0 \pm 0.2) \times 10^{-5}\mathrm{m}^2\mathrm{s}^{-1}$$

$$K_T^{\epsilon} = 0.25\frac{\epsilon}{\partial_z\bar{b}} = (1.2 \pm 0.2) \times 10^{-5}\mathrm{m}^2\mathrm{s}^{-1}$$



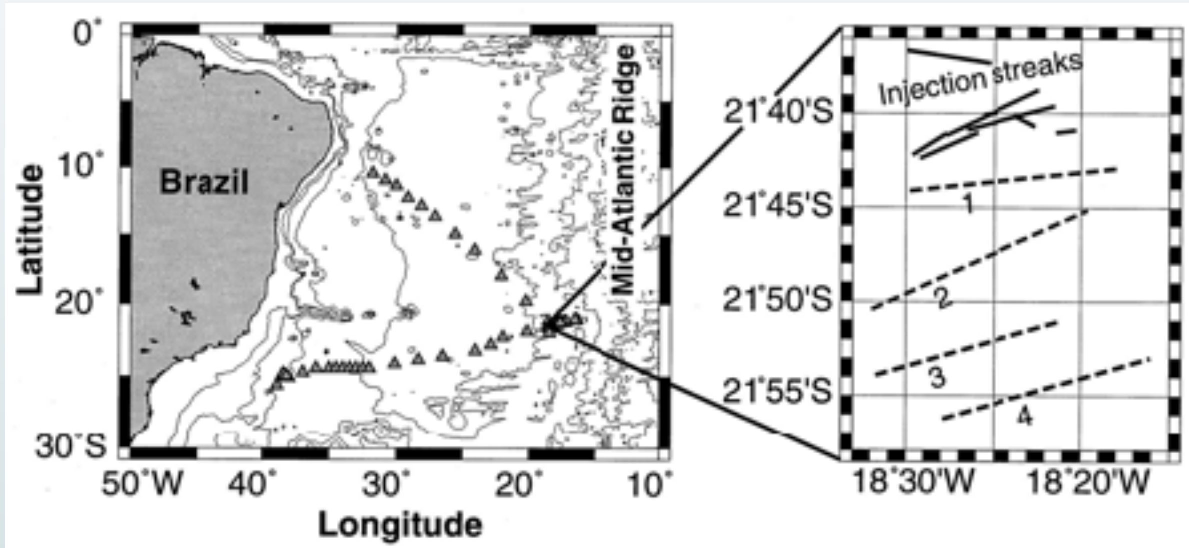
Contamination of χ



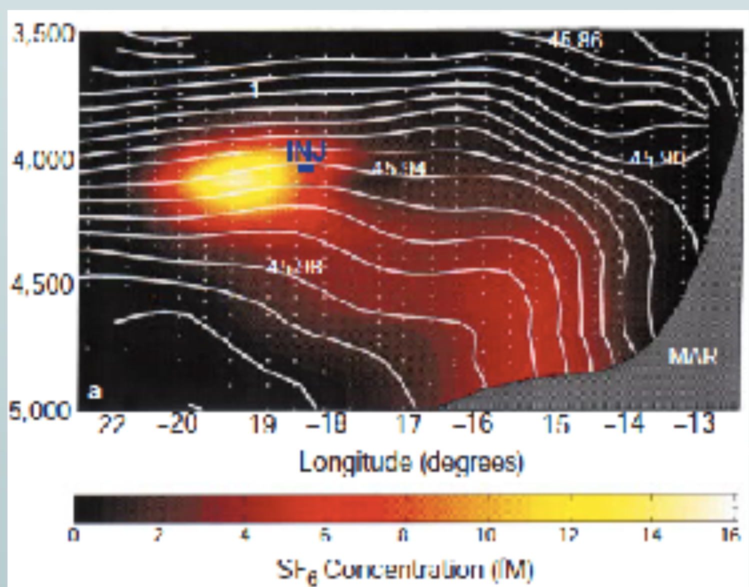
Ferrari and Polzin, 2005

Brazil Basin Tracer Release Experiment (BBTRE)

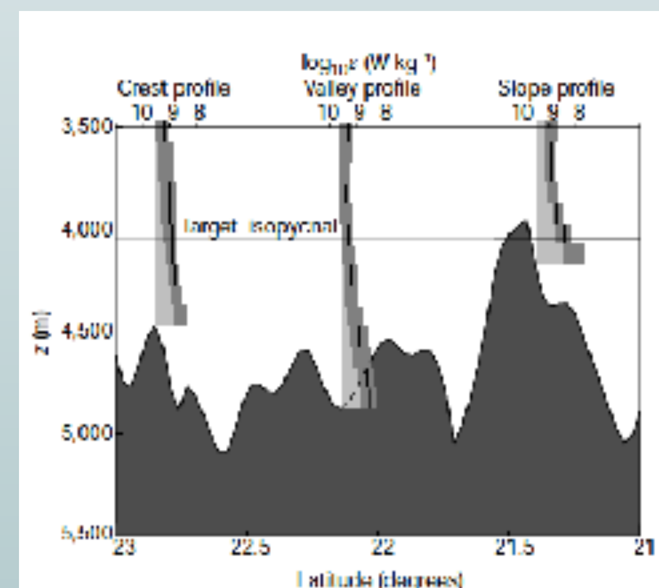
BBTRE



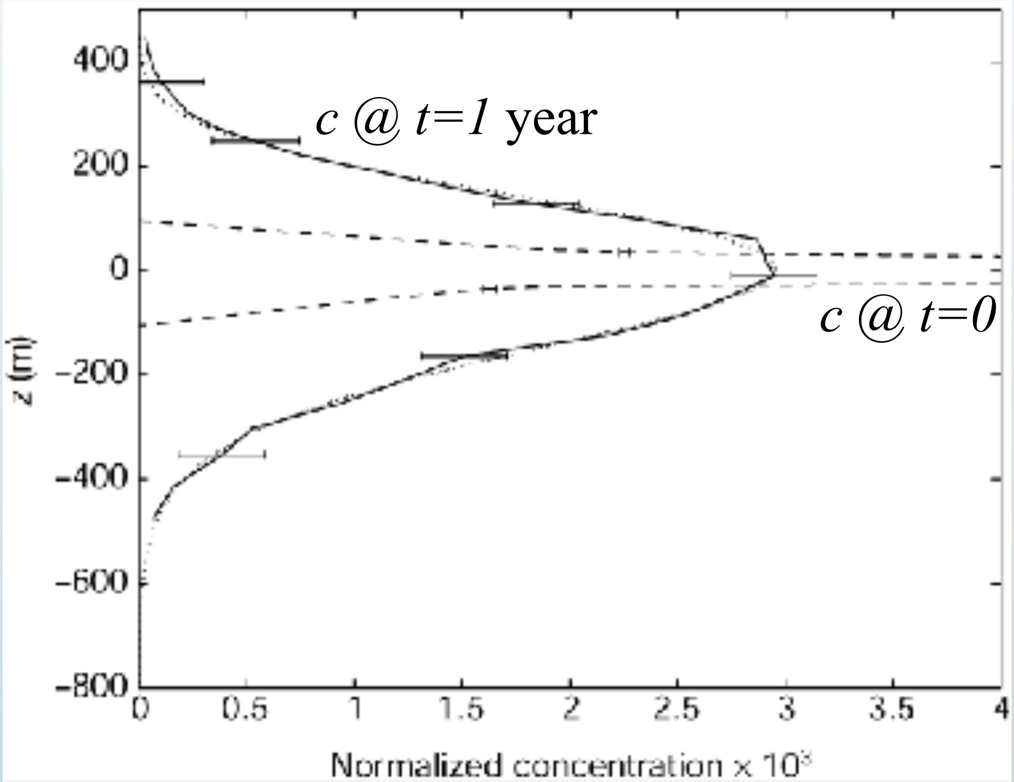
Tracer release (*Ledwell et al., 2000*)



HRP profiles (*Polzin et al., 1997*)



K_T is bottom enhanced

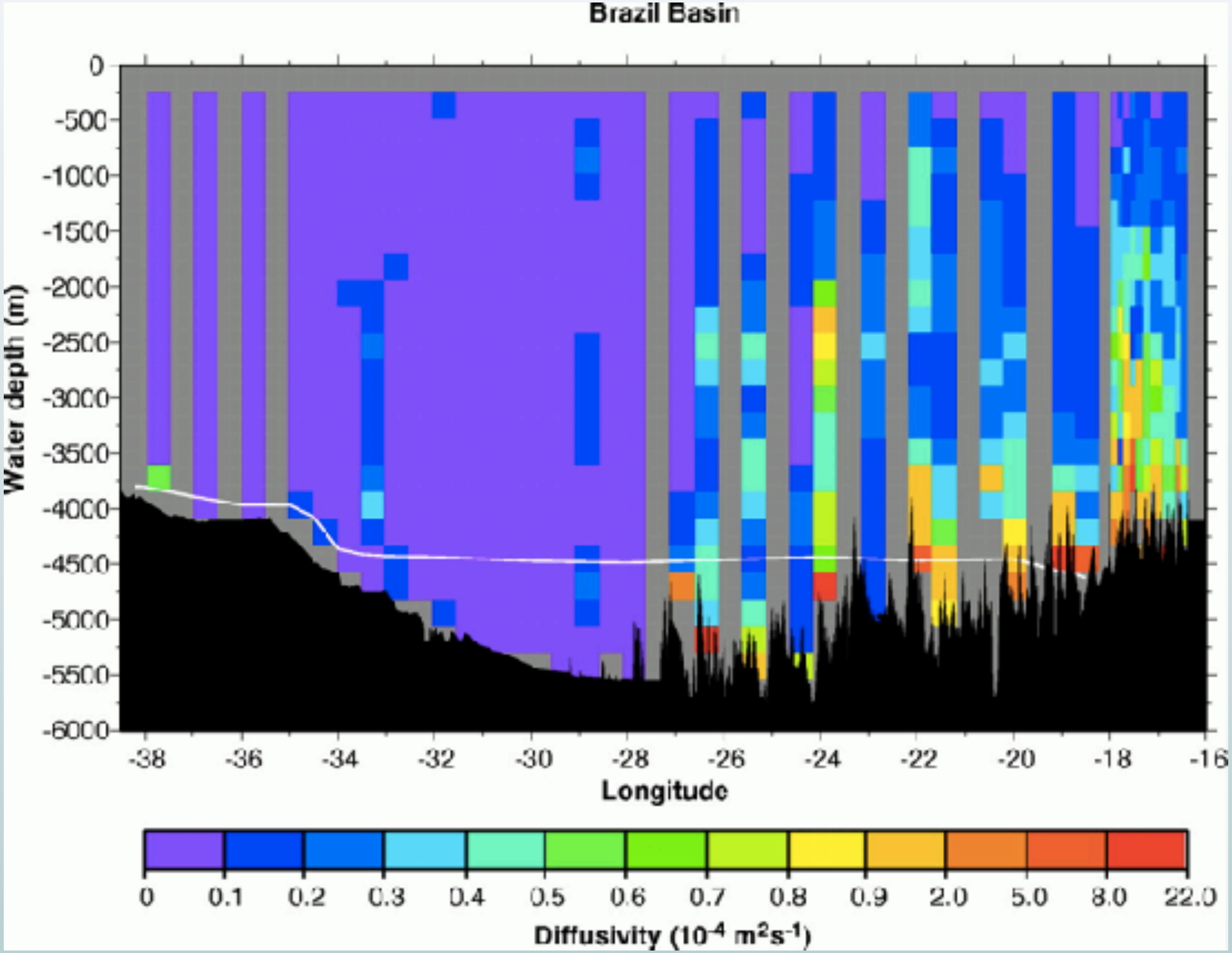


$$K_T^{tracer} = \frac{1}{2}\frac{Z_2|_{t=1\,\text{year}} - Z_2|_{t=0}}{1\,\text{year}} = (3-8)\times 10^{-4}\text{m}^2\text{s}^{-1}$$

$$K_T^{\epsilon} = (2 \pm 1)\times 10^{-4}\text{m}^2\text{s}^{-1}$$

Ledwell et al., 2000

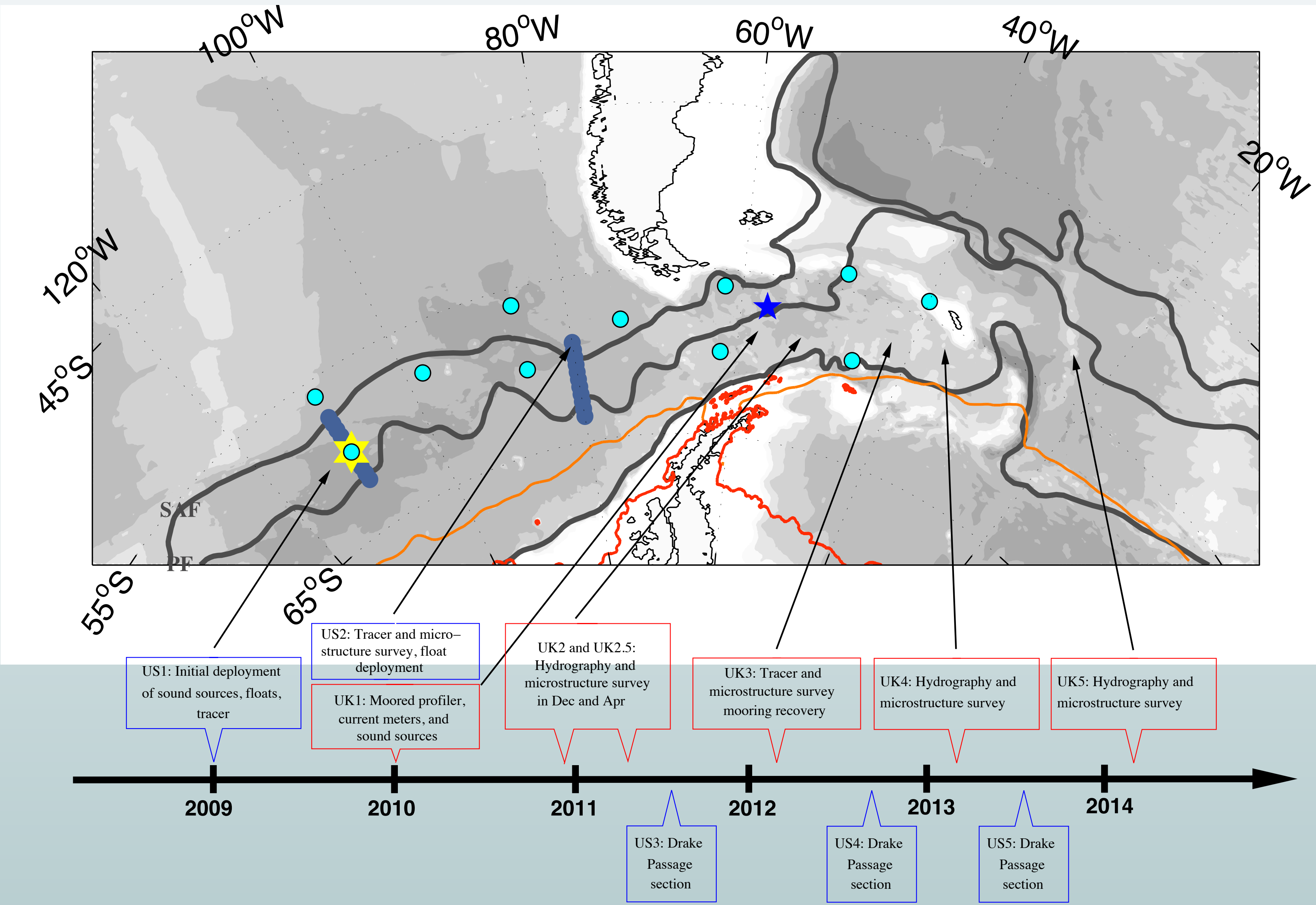
K_T is bottom enhanced



Polzin et al., 1997

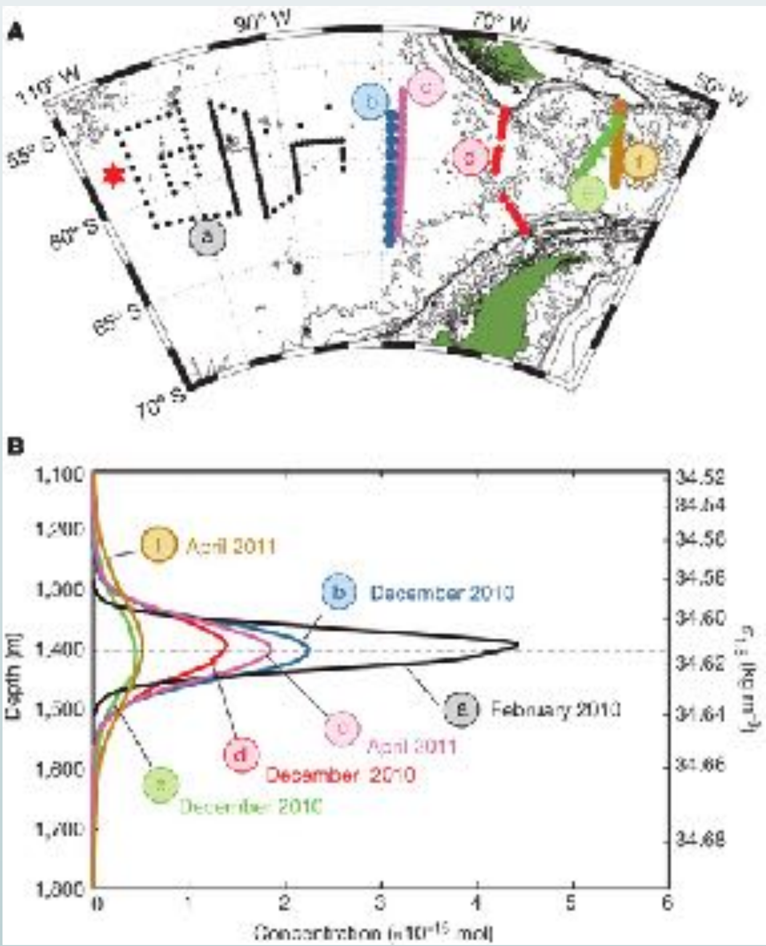
Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean (DIMES)

DIMES


100°W
80°W
60°W
40°W
20°W
120°W
45°S
55°S
65°S
SAF
PF
US1: Initial deployment of sound sources, floats, tracer
US2: Tracer and micro–structure survey, float deployment
UK1: Moored profiler, current meters, and sound sources
UK2 and UK2.5: Hydrography and microstructure survey in Dec and Apr
UK3: Tracer and microstructure survey mooring recovery
UK4: Hydrography and microstructure survey
UK5: Hydrography and microstructure survey
2009
2010
2011
2012
2013
2014
US3: Drake Passage section
US4: Drake Passage section
US5: Drake Passage section

Tracer estimate of κ

- Passive tracer (CF_3SF_5) was released at 1500m depth, 2000m above seafloor



Upstream of Drake Passage

between transects (a) and (d)

$$K_T^{tracer} = \frac{1}{2}\frac{Z_2|_b - Z_2|_\star}{2\,\mathrm{years}} \simeq 2\times 10^{-5}\mathrm{m}^2\mathrm{s}^{-1}$$

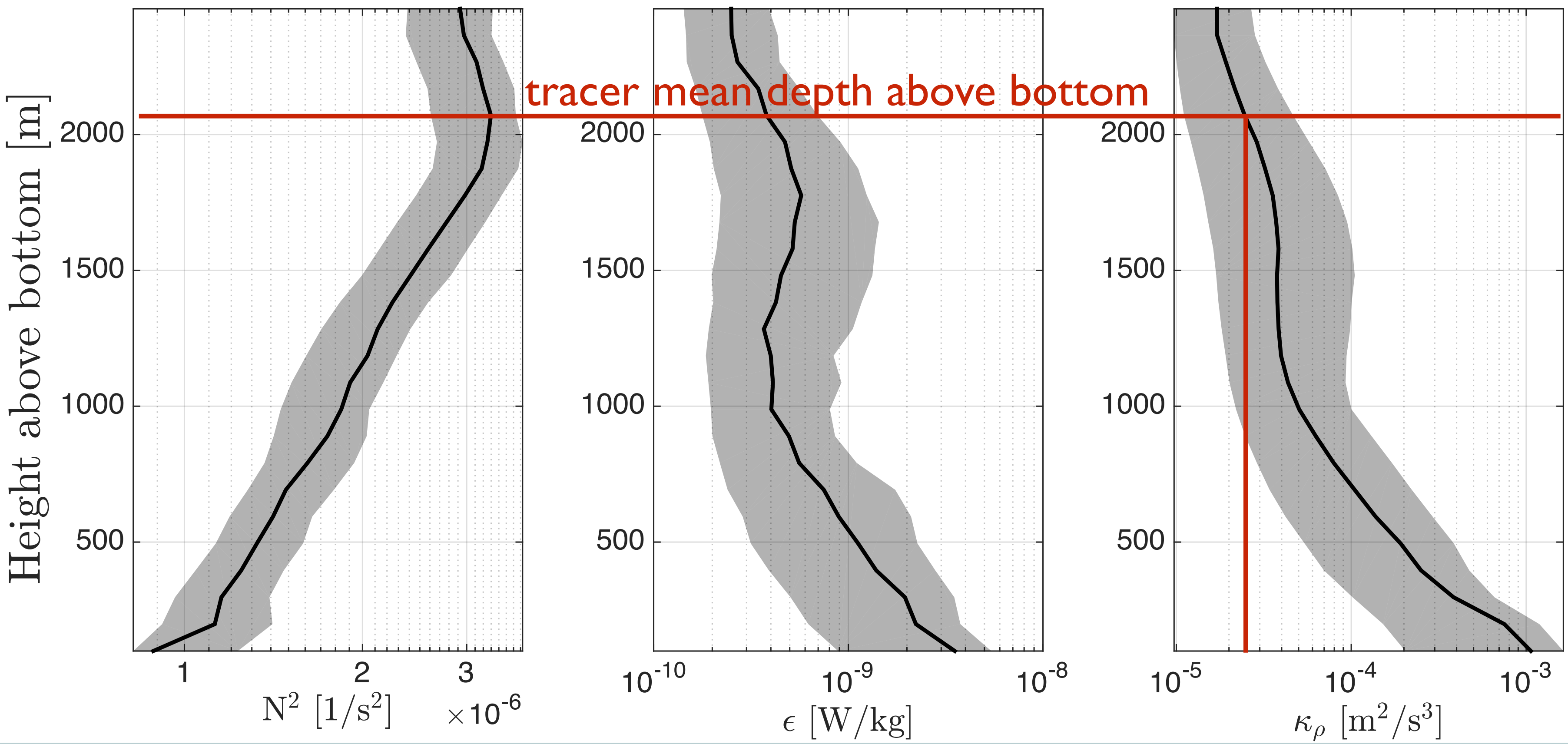
Downstream of Drake Passage

between transects (d) and (f)

$$K_T^{tracer} = \frac{1}{2}\frac{Z_2|_f - Z_2|_d}{3\ \mathrm{months}} \simeq 3\times 10^{-4}\mathrm{m}^2\mathrm{s}^{-1}$$

Microstructure estimate of K_T

- Upstream of Drake Passage K_T^{tracer} and K_T^{ϵ} are similar
- Downstream of Drake Passage K_T^{ϵ} at mean depth of tracer is an order of magnitude smaller that K_T^{tracer}

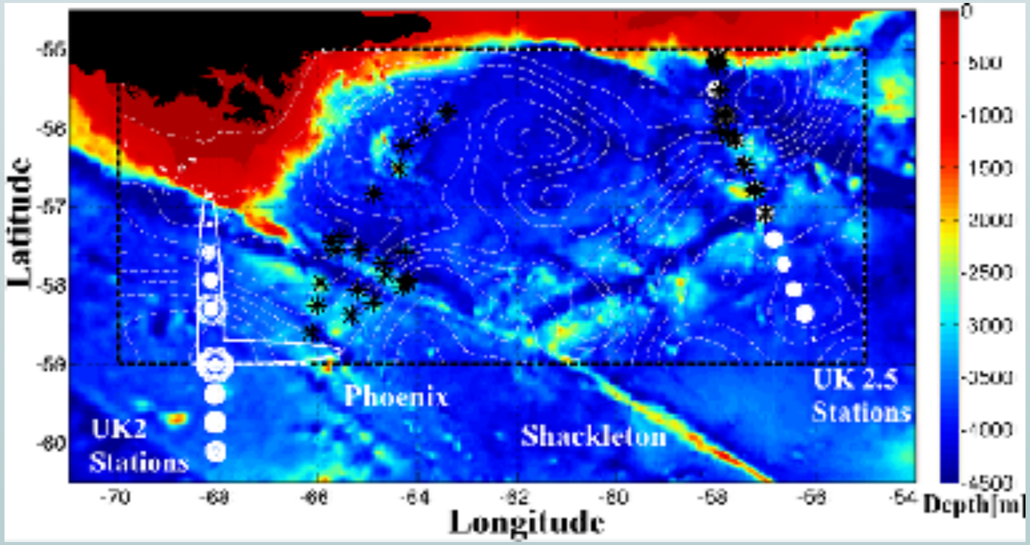


Mashayek, Ferrari et al., 2017

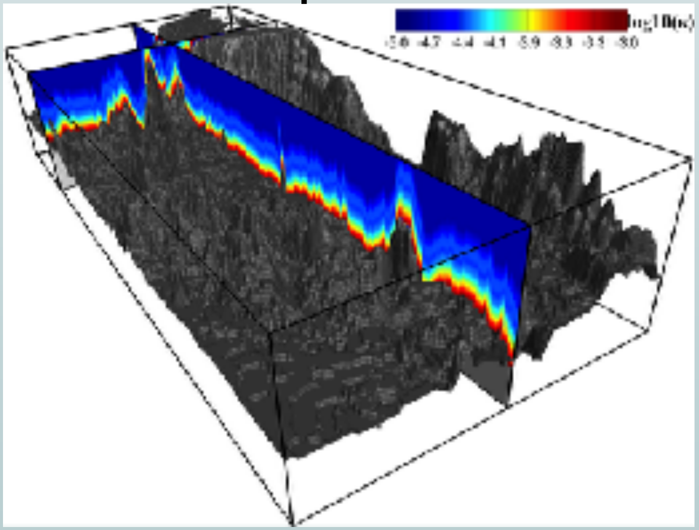
Modeling approach

- MIT general circulation model of a 1400x400 km patch of Scotia Sea
- Resolution of 500-600 m, 100 vertical levels (30 m resolution at tracer depth)
- Model forced at boundaries with a coarser resolution patch of ACC constrained to observations (Tulloch, Ferrari et al., 2014)
- Smith and Sandwell topography at one minute (1/60th of degree)
- Vertical profile of K_T from microstructure profile imposed everywhere as a function of height above the bottom

Model domain



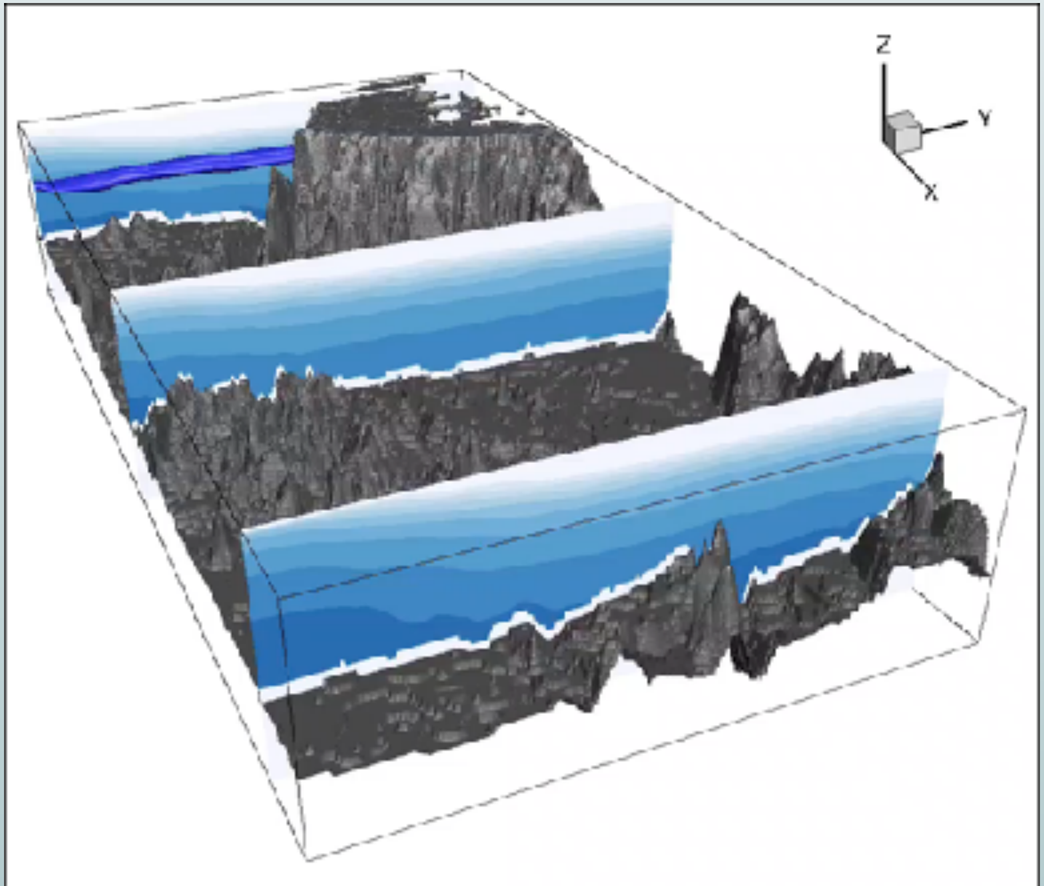
κ_m profiles



Mashayek, Ferrari et al., 2017

Tracer evolution

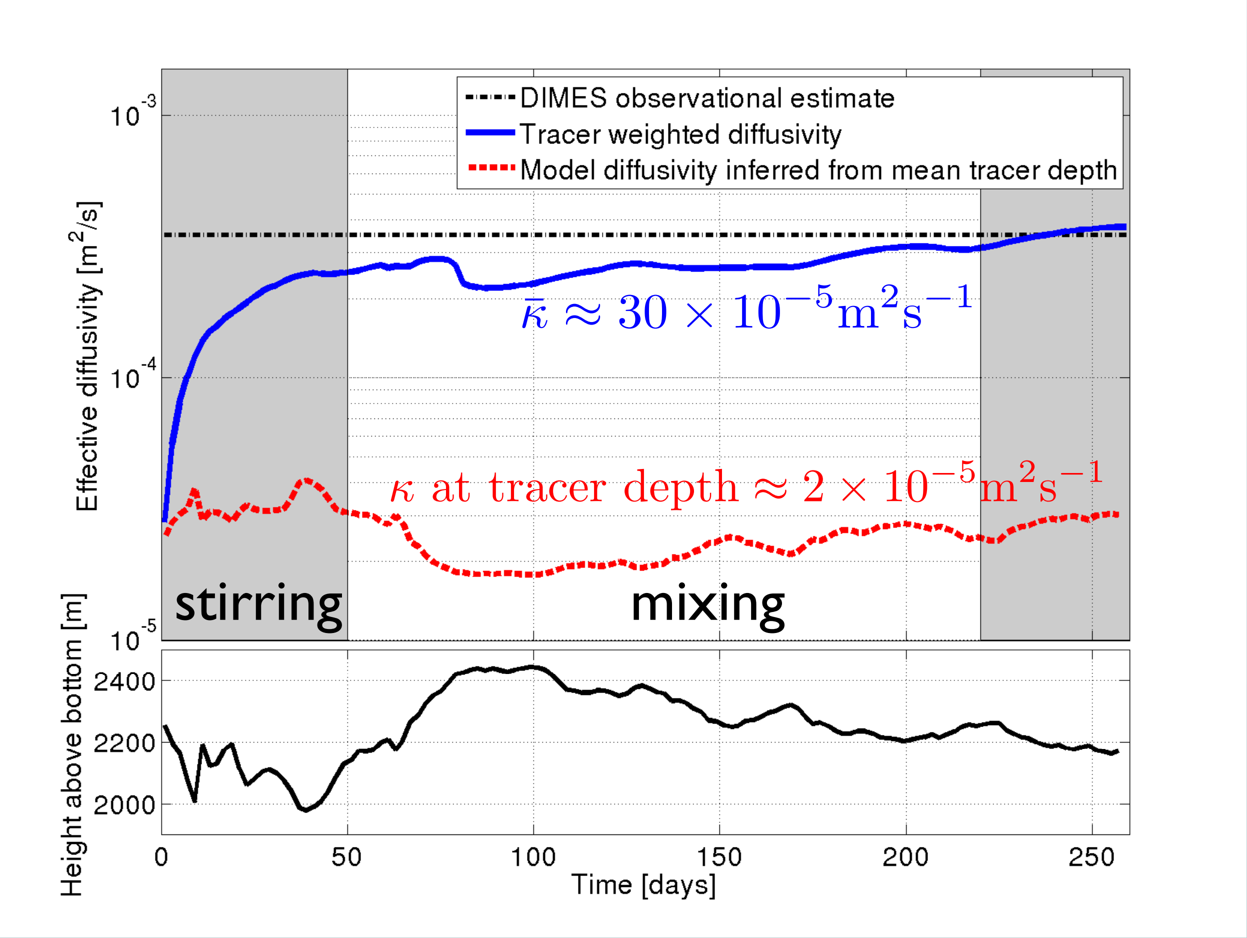
- Tracer is released along western boundary of domain based on sampling of real tracer along that line
- Movie shows that tracer is
  - stirred laterally by geostrophic eddies toward seamounts/ridges
  - mixed vertically by enhanced K_T next to seamounts/ridges



Numerical estimate of K_T^{tracer}

- We can estimate the diffusivity experienced by the tracer as

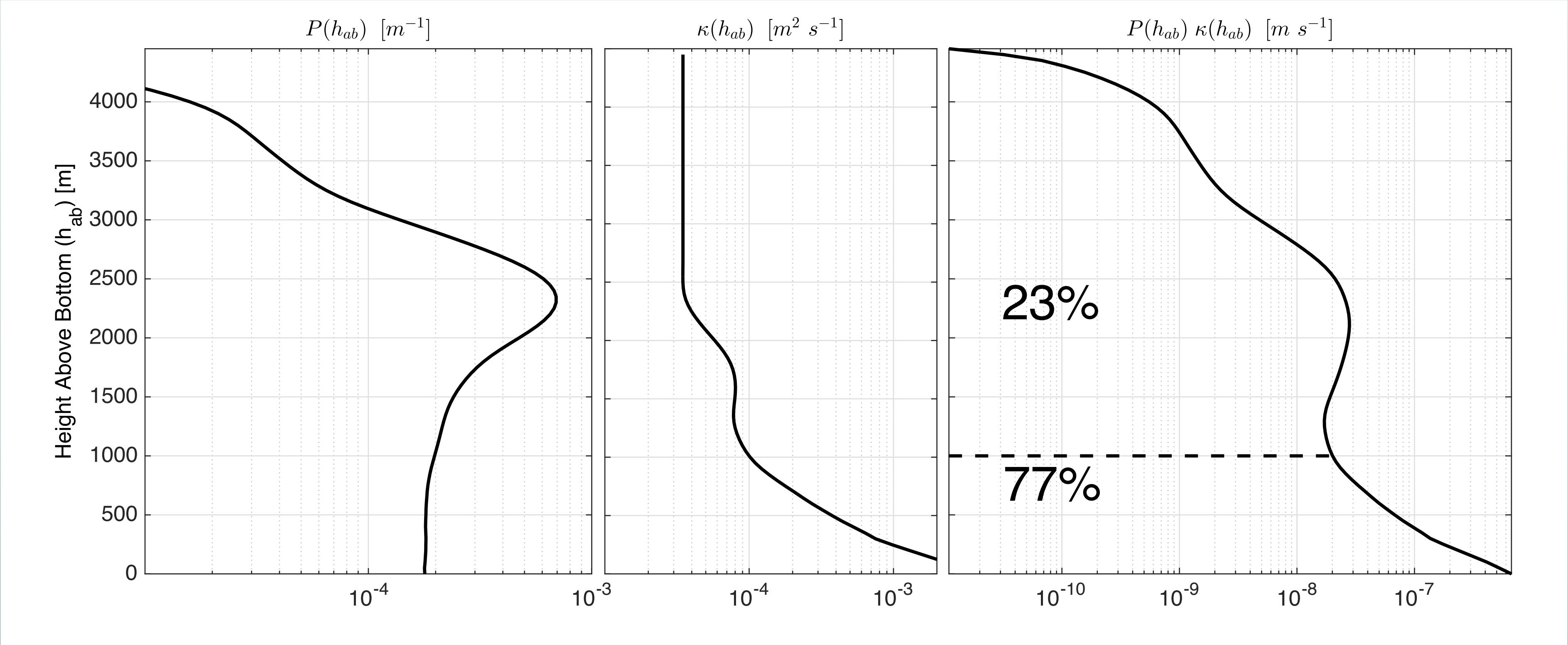
$$\bar{K} \equiv \frac{\iiint K_T c \, \mathrm{d}V}{\iiint c \, \mathrm{d}V}$$



Mashayek, Ferrari et al., 2017

Numerical estimate of K_T^{tracer}

- Panel 1: distribution of tracer peaks ~2300 m above seafloor and decays below
- Panel 2: profile of K_T increases toward seafloor
- Panel 3: increase of K_T is so large that estimate of $\overline{K}$ is dominated by integral over bottom 1000 m
- Mixing occurs when tracer is within 1000 m of seafloor

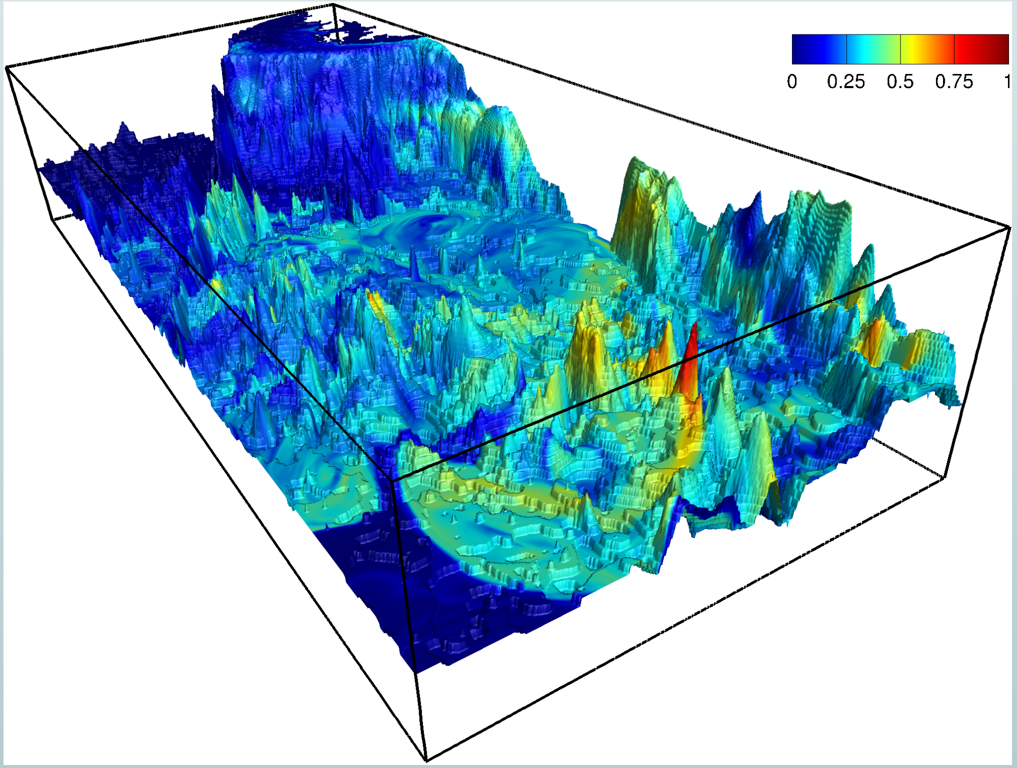


Mashayek, Ferrari et al., 2017

Numerical estimate of K_T^{tracer}

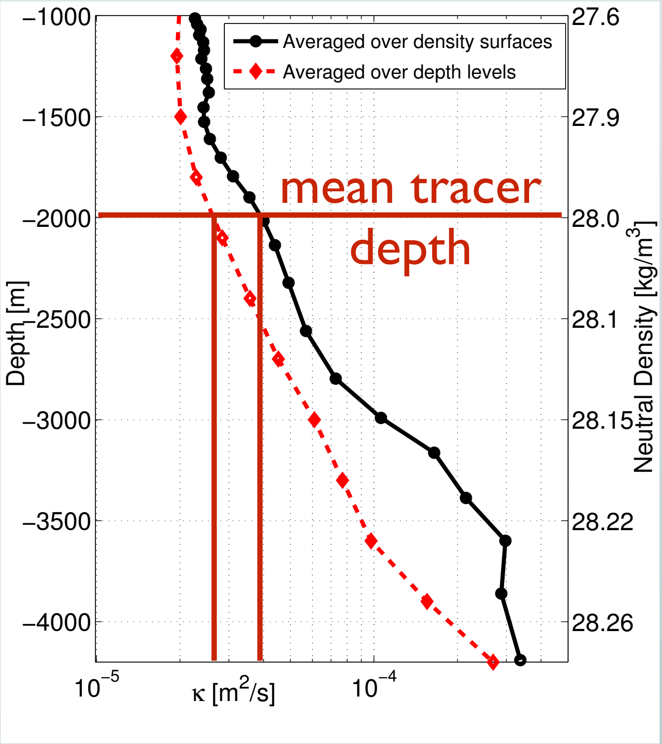
- Tracer tends to accumulate over remounts and ridges where mixing is strong
  - velocities are weaker close to seafloor
  - bottom enhanced mixing drives tracers toward the seafloor
  - other reasons?

Vertically integrated tracer concentration



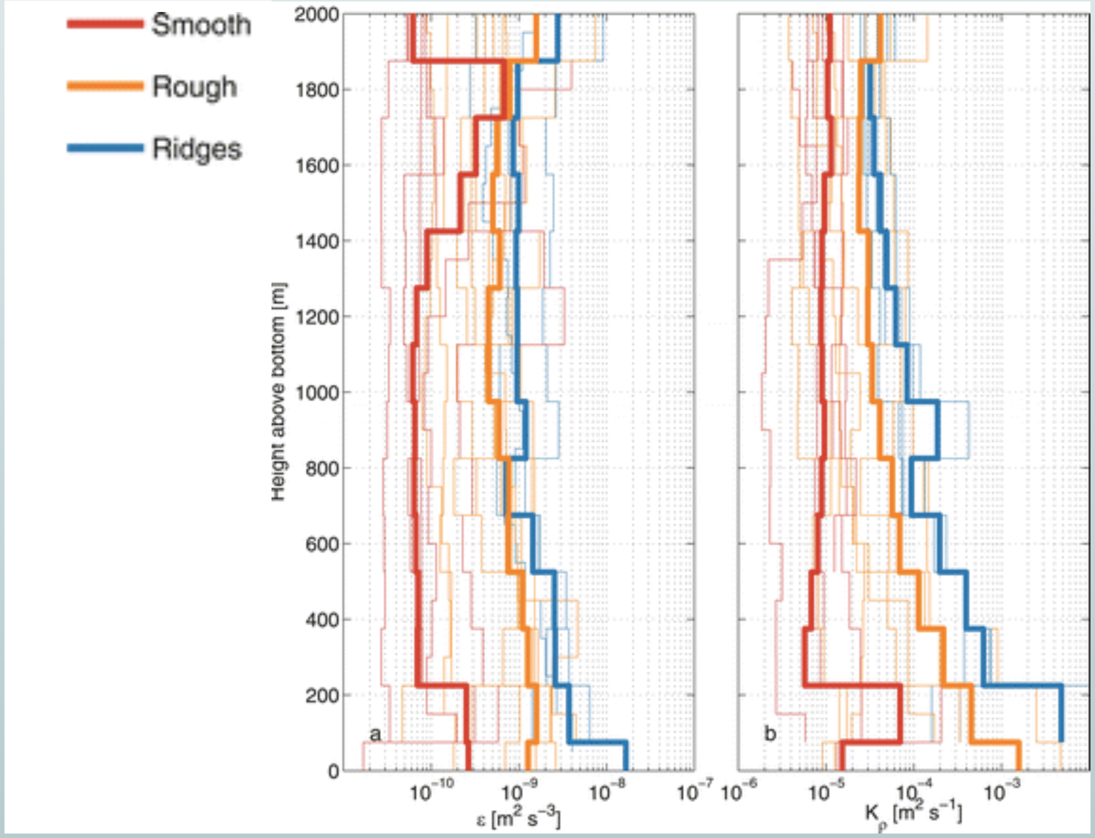
Numerical estimate of K_T^{tracer}

- Simply averaging along density surfaces gives little enhancement (mixing hotspots are too rare)
- Accumulation of tracer near topography is crucial to explain large K_T^{tracer} in deep ocean (hotspots trap tracer)



Conclusions

- $K_T^\epsilon = \Gamma\epsilon/\partial_z\bar{b}$ increases toward rough seafloor
- $\overline{w'b'} = -\Gamma\epsilon$ increases in magnitude toward rough seafloor
- mixing is confined within a few hundred meters of seafloor



Waterhouse et al., 2014

Conundrum

- $\overline{w'b'} = -\Gamma\epsilon$ increases in magnitude toward rough seafloor
- mixing drives sinking, not upwelling of ocean waters