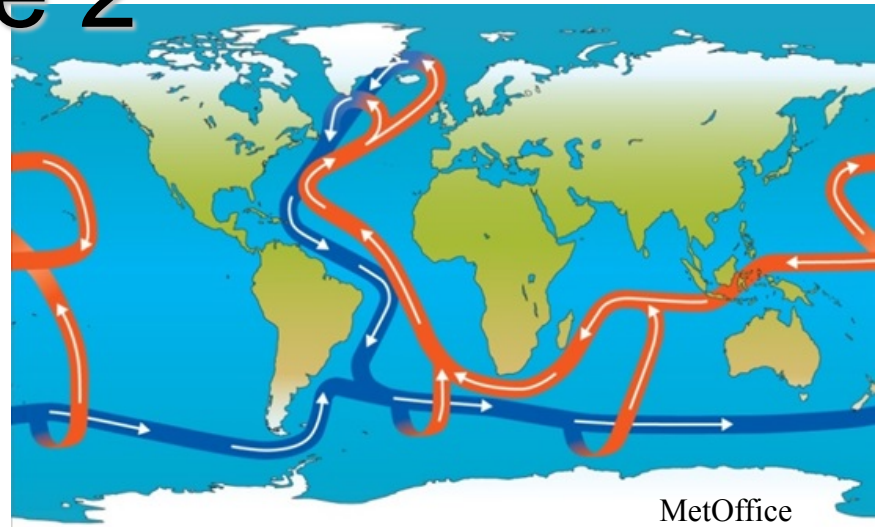


The Role of small scale turbulence for Ocean Circulation

Advanced Methods in Oceanography

Lecture 2



For the end of the the lectures you should be able to...

- Describe the AMOC
- Describe the two paradigms of the AMOC
- For the traditional paradigm, derive the main balance of the AMOC
- For the alternative paradigm, derive the SO dynamics relative to the AMOC

The Role of small scale turbulence for Ocean Circulation

- Turbulence
 - What is small scale turbulence: Mixing and Stirring
- Why does mixing matter for the large-scale ocean circulation?
 - The two ocean circulation paradigms
 - Theory of the AMOC

Turbulence in the Ocean

Why does “turbulence” matter?

Exercise

$$\frac{\partial \rho}{\partial t} + u \frac{\partial \rho}{\partial x} + v \frac{\partial \rho}{\partial y} + w \frac{\partial \rho}{\partial z} = 0$$

$$\rho = \bar{\rho} + \rho', \quad (u, v, w) = (\bar{u}, \bar{v}, \bar{w}) + (u', v', w'),$$

Here the “bar” can denote a spatial or a time average.

1. Decompose the equation in mean and anomalous terms.
2. Find the equation for the mean.

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$$\frac{\partial \bar{\rho}}{\partial t} + \bar{u} \frac{\partial \bar{\rho}}{\partial x} + \bar{v} \frac{\partial \bar{\rho}}{\partial y} + \bar{w} \frac{\partial \bar{\rho}}{\partial z} = -\overline{u' \frac{\partial \rho'}{\partial x}} - \overline{v' \frac{\partial \rho'}{\partial y}} - \overline{w' \frac{\partial \rho'}{\partial z}}$$

Turbulence in the Ocean

$$\frac{\partial \bar{\rho}}{\partial t} + \bar{u} \frac{\partial \bar{\rho}}{\partial x} + \bar{v} \frac{\partial \bar{\rho}}{\partial y} + \bar{w} \frac{\partial \bar{\rho}}{\partial z} = -\overline{u' \frac{\partial \rho'}{\partial x}} - \overline{v' \frac{\partial \rho'}{\partial y}} - \overline{w' \frac{\partial \rho'}{\partial z}}$$

Turbulent terms:

- How the large/long scale is affected by the smaller scale process?
- How could we represent the smaller/shorter scale effect?

“Subgrid” process need to be parameterized, *i.e.* the right handside operator has to be represented as a function of large/long scale.

In general one can split an operator in two terms, through:

- 1.its rotation effect and
- 2.its deformation effect.

Turbulence in the Ocean

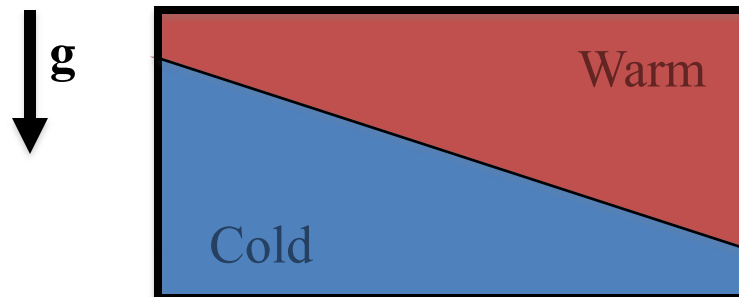
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Exercise:

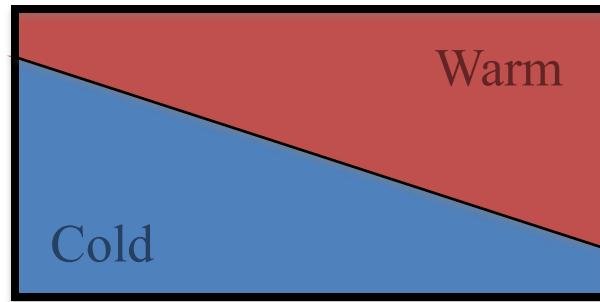
➤ Determine the effect of 1. and 2. on a dual-density profile flow



Turbulence in the Ocean

Hence:

1. Rotation conserved the properties -> Stirring/Transport process
2. Deformation destroy gradient -> Mixing



Effect of Rotation



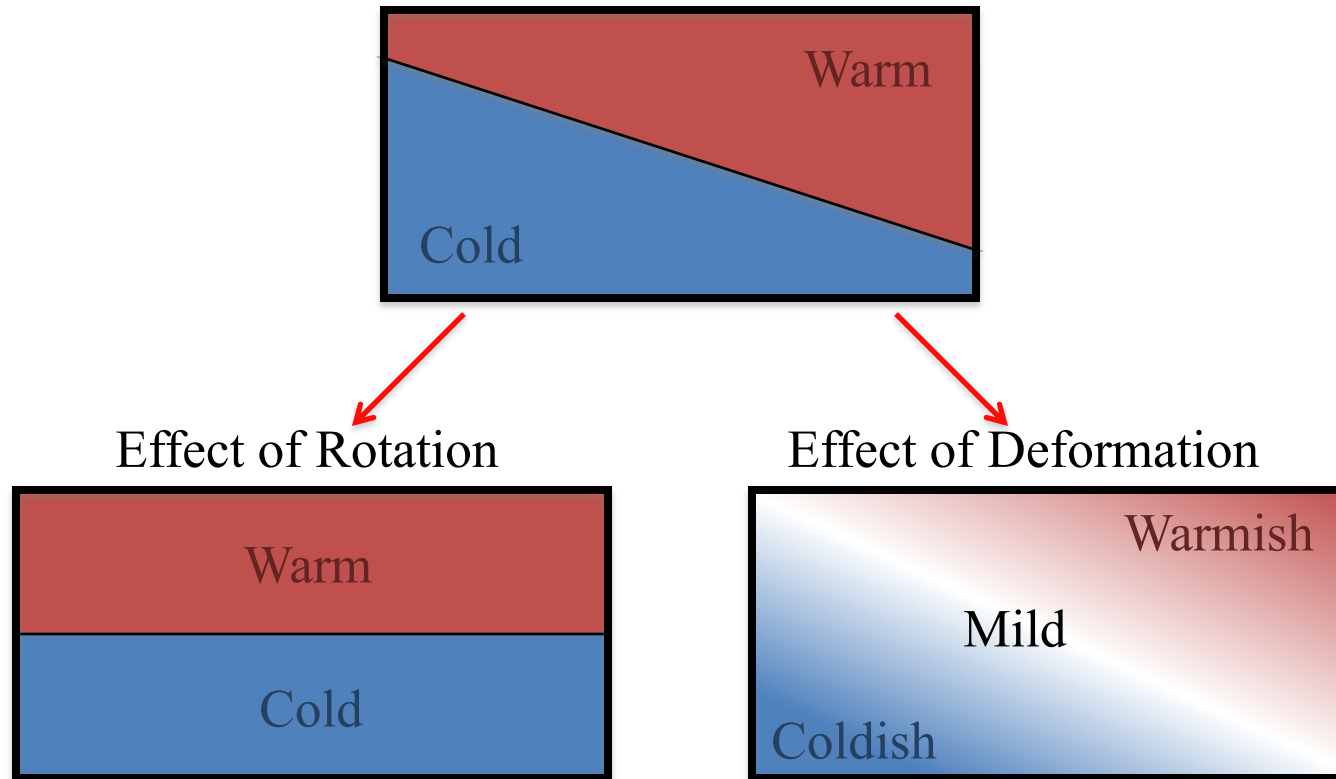
Effect of Deformation



Turbulence in the Ocean

Hence:

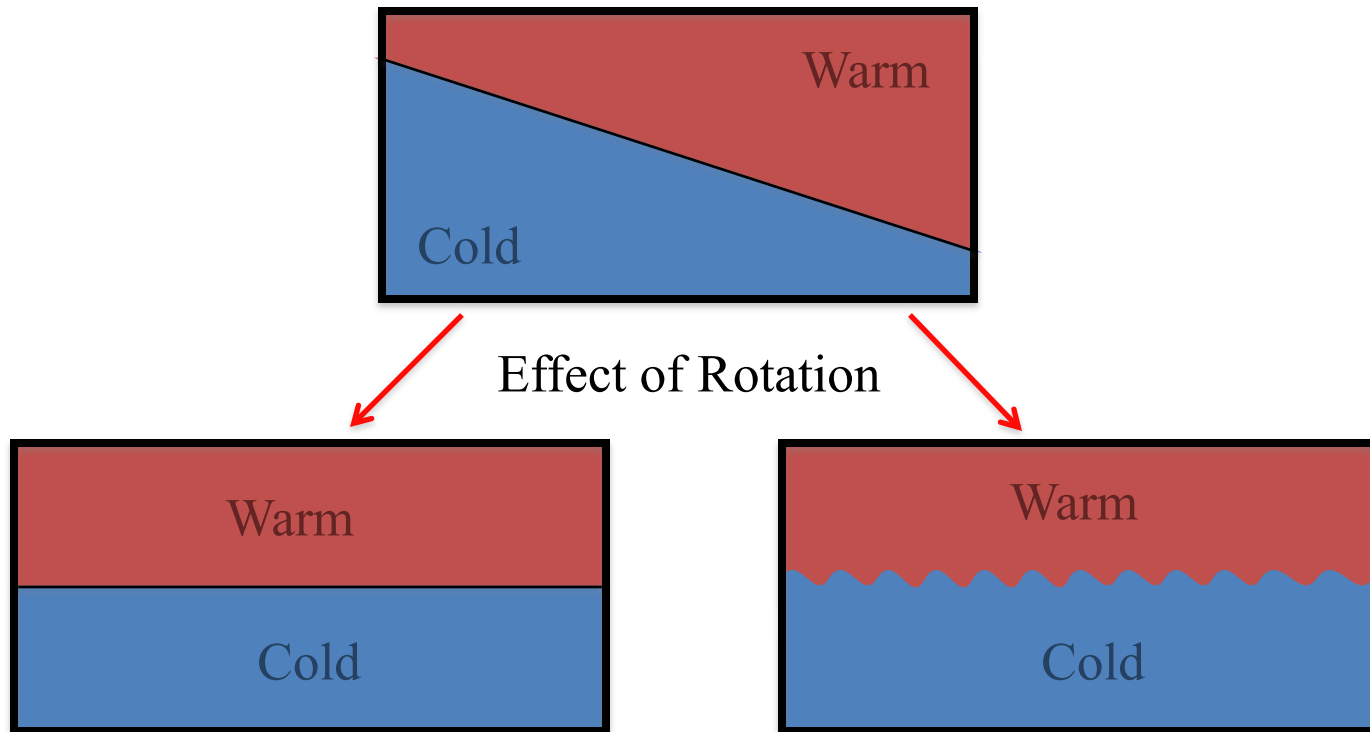
1. Rotation conserved the properties -> Stirring/Transport process
2. Deformation destroy gradient -> Mixing



Turbulence in the Ocean

Side comment on stirring:

It technically can also increase small scale features, hence leading to instability, through the creation of smaller scale but sharper gradients.

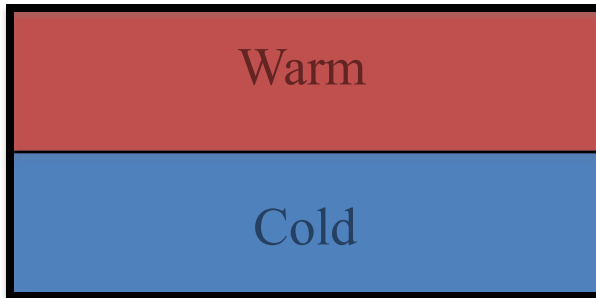


Turbulence in the Ocean

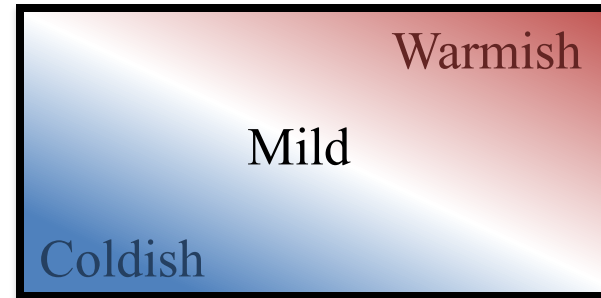
Hence:

1. Rotation conserved the properties -> Stirring/Transport process
2. Deformation destroy gradient -> Mixing

Effect of Rotation



Effect of Deformation



Possible parameterization:

Following Gent McWilliams:

$$\overline{u' \frac{\partial \rho'}{\partial x}} \approx \tilde{u} \frac{\partial \bar{\rho}}{\partial x}$$

Where $\tilde{u}$ is proportional to the large/long scale slope (to mimic Baroclinic instability).

Following Fick's law:

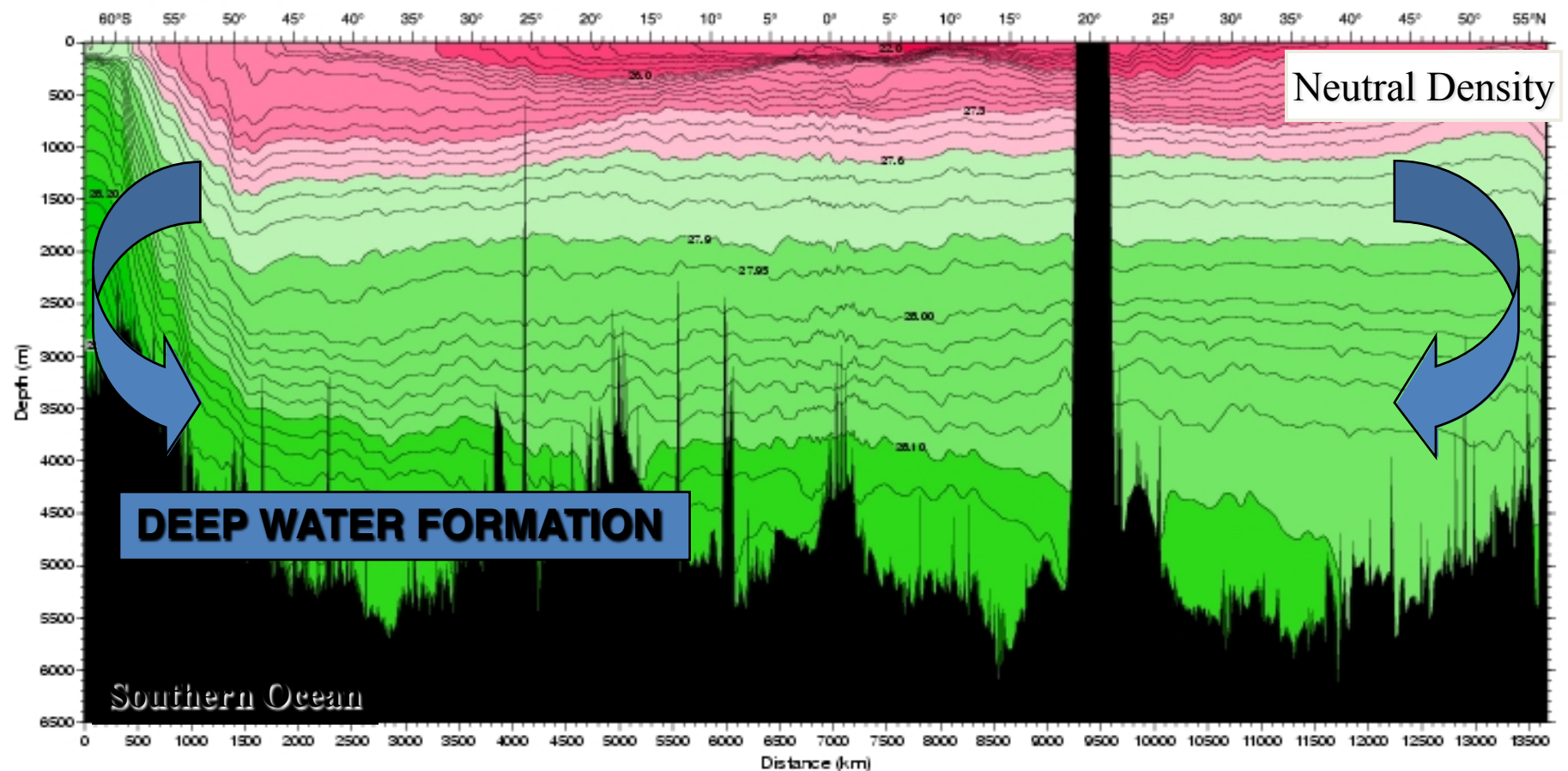
$$\overline{u' \rho'} \approx -K \frac{\partial \bar{\rho}}{\partial x}$$

The flow is proportional to the large/long gradient (to induce a downgradient flow).

Why does turbulence matter for the large-scale ocean circulation?

(1) The ocean has a nearly ubiquitous stable stratification. How is it maintained?

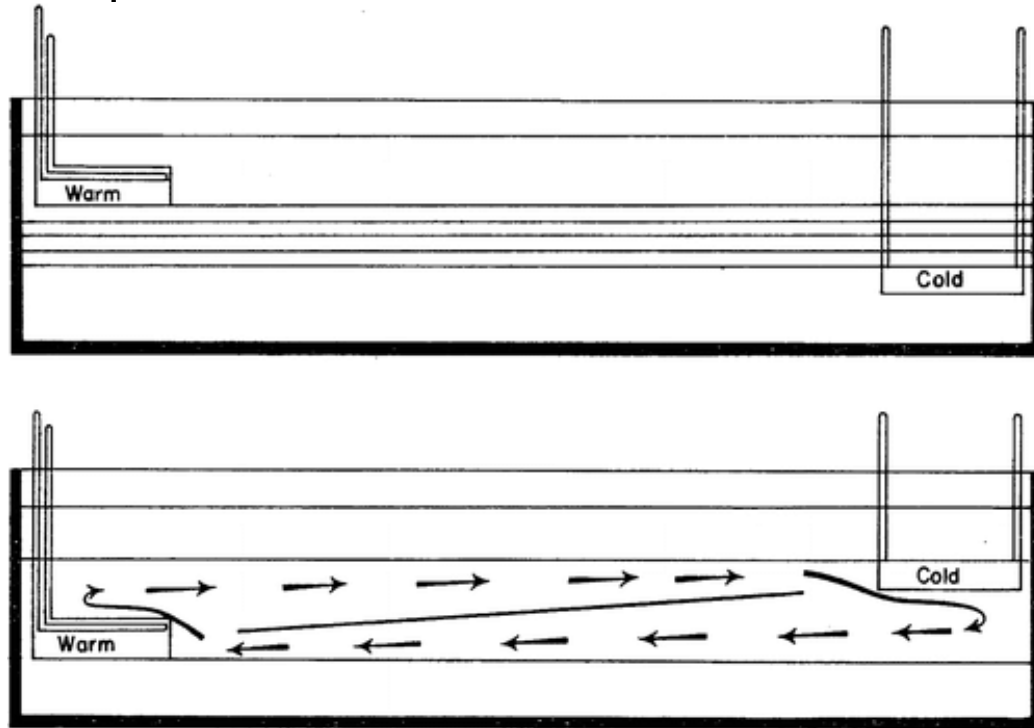
Simple conceptual model: stratification arises from the competition between the removal of buoyancy from the deep ocean by dense water formation and the addition of buoyancy to the deep ocean by downward fluxes of light upper ocean waters, presumed to be warmed directly by the sun.



Why does turbulence matter for the large-scale ocean circulation?

(2) Simple energetic arguments suggest that maintaining the ocean stratification equates to driving the ocean's meridional overturning circulation (MOC), and that surface buoyancy forcing is not a significant driver...

Sandström's theorem: 'In the limit of no diffusion, heating must be at a higher pressure than cooling, in order that a steady circulation can be maintained against the dissipative effects of friction'.



2 configurations of convective systems (from Defant 1961) illustrating the arguments of Sandström's theorem.

1. Top fluids are stably stratified, with the cold source fluid filling the abyss of the container, with almost **no-circulation**.

2. Bottom, convection gives rise to fluid instabilities analogous to those in the Rayleigh-Bénard case, with **circulation**.

Why does turbulence matter for the large-scale ocean circulation?

(2) **Sandström's theorem:** 'In the limit of no diffusion, heating must be at a higher pressure than cooling, in order that a steady circulation can be maintained against the dissipative effects of friction'.

Since there is a steady circulation in the ocean, *i.e.* the AMOC exists, for consistency with Sandström's theorem either:

1. The Sandström theorem does not apply.

2. There is a heat source below the heat sink... Or a way to put warm water below cold one!

Why does turbulence matter for the large-scale ocean circulation?

(2) **Sandström's theorem:** 'In the limit of no diffusion, heating must be at a higher pressure than cooling, in order that a steady circulation can be maintained against the dissipative effects of friction'.

Since there is a steady circulation in the ocean, *i.e.* the AMOC exists, for consistency with Sandström's theorem either:

1. The Sandström theorem does not apply.

-> There is vertical mixing in the ocean.

Small scale mixing!

2. There is a heat source below the heat sink... Or a way to put warm water below cold one!

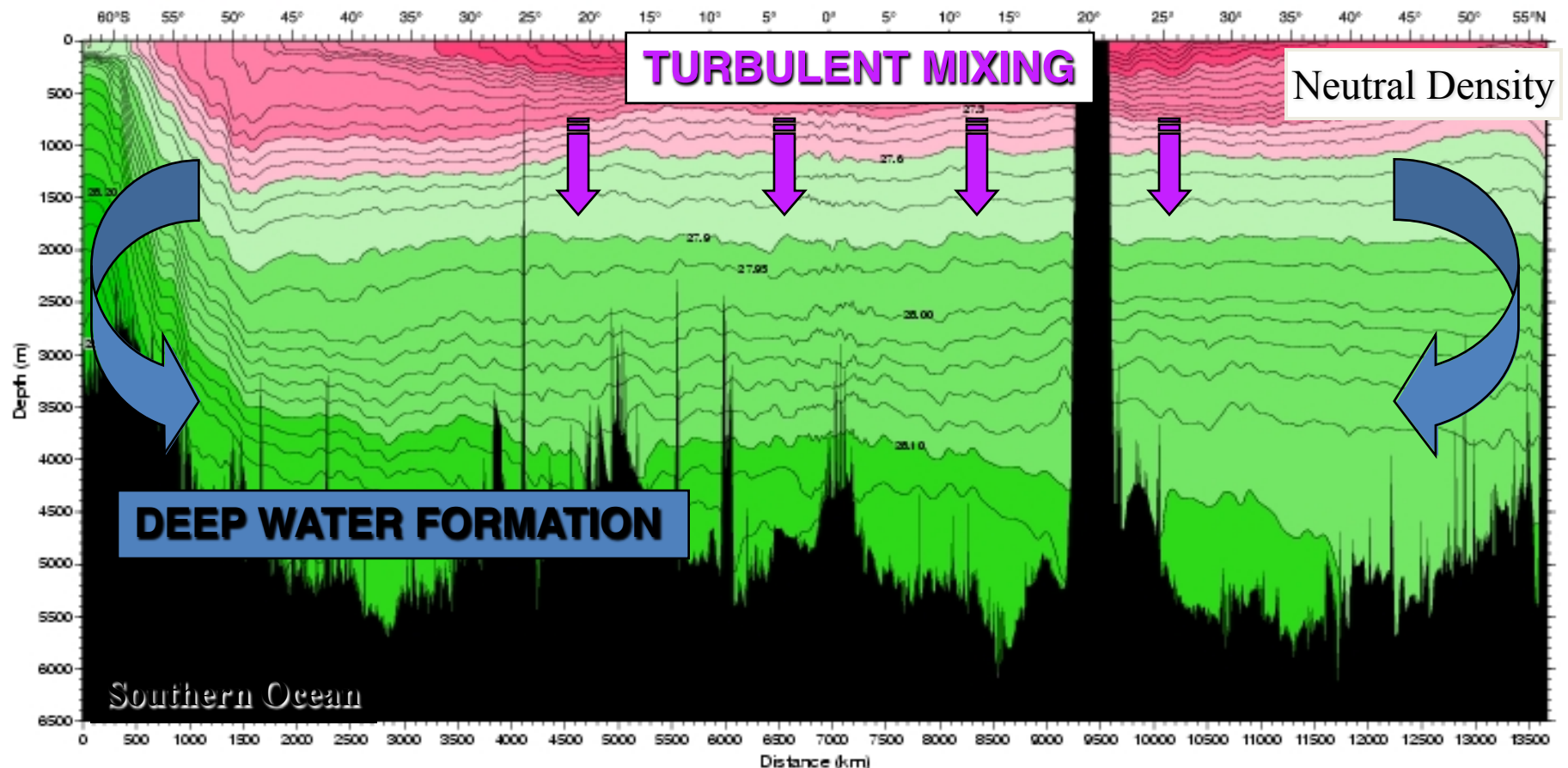
-> Mechanical downward fluxes of warm water through eddies process.

Small scale stirring!

This summarizes the two main paradigms of the AMOC, both evolving small scale turbulent processes.

The traditional paradigm of ocean circulation: the abyssal ocean

Munk (1966) and Munk and Wunsch (1998) consider the density budget of the abyssal ocean (below ~ 1000 m). For a deep water formation rate of ~ 25 Sv and the observed stratification, they show that a global-mean turbulent diapycnal (*i.e.* directed across density surfaces) diffusivity of $K_\rho \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ is required to sustain the stratification.



How is the conclusion that $K_\rho \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ reached?

The calculation of the global-mean turbulent diapycnal diffusivity by Munk and Wunsch is based on an advection - diffusion balance (of density):

$$w \frac{\partial \rho}{\partial z} = \frac{\partial}{\partial z} (K_\rho \frac{\partial \rho}{\partial z})$$

or **VERTICAL ADVECTION = TURBULENT VERTICAL DIFFUSION**
(where w = vertical velocity, ρ = density, z = height relative to the sea surface).

Solving for K_ρ we obtain:

$$K_\rho = \frac{1}{\partial \rho / \partial z} \int w \frac{\partial \rho}{\partial z} dz$$

If we further assume that K_ρ varies slowly relative to the vertical gradient of density, we have:

$$K_\rho = (w \partial \rho / \partial z) / (\partial^2 \rho / \partial z^2)$$

How is the conclusion that $K_\rho \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ reached?

In order to understand where the $K_\rho \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ result comes from, let us try some vertical velocity and stratification values characteristic of the ocean in

$$K_\rho = (w \partial\rho/\partial z) / (\partial^2\rho/\partial z^2)$$

Munk and Wunsch assumed an input of dense water to the deep ocean (nominally) deeper than 1000 m of $T_{\text{deep}} \sim 25 \text{ Sv}$ at high latitudes. With a global ocean area of $A \sim 2.5 \times 10^{14} \text{ m}^2$, this translates into a global-mean vertical velocity $w = T_{\text{deep}} / A \sim 10^{-7} \text{ m s}^{-1}$ (or $\sim 4 \text{ m yr}^{-1}$).

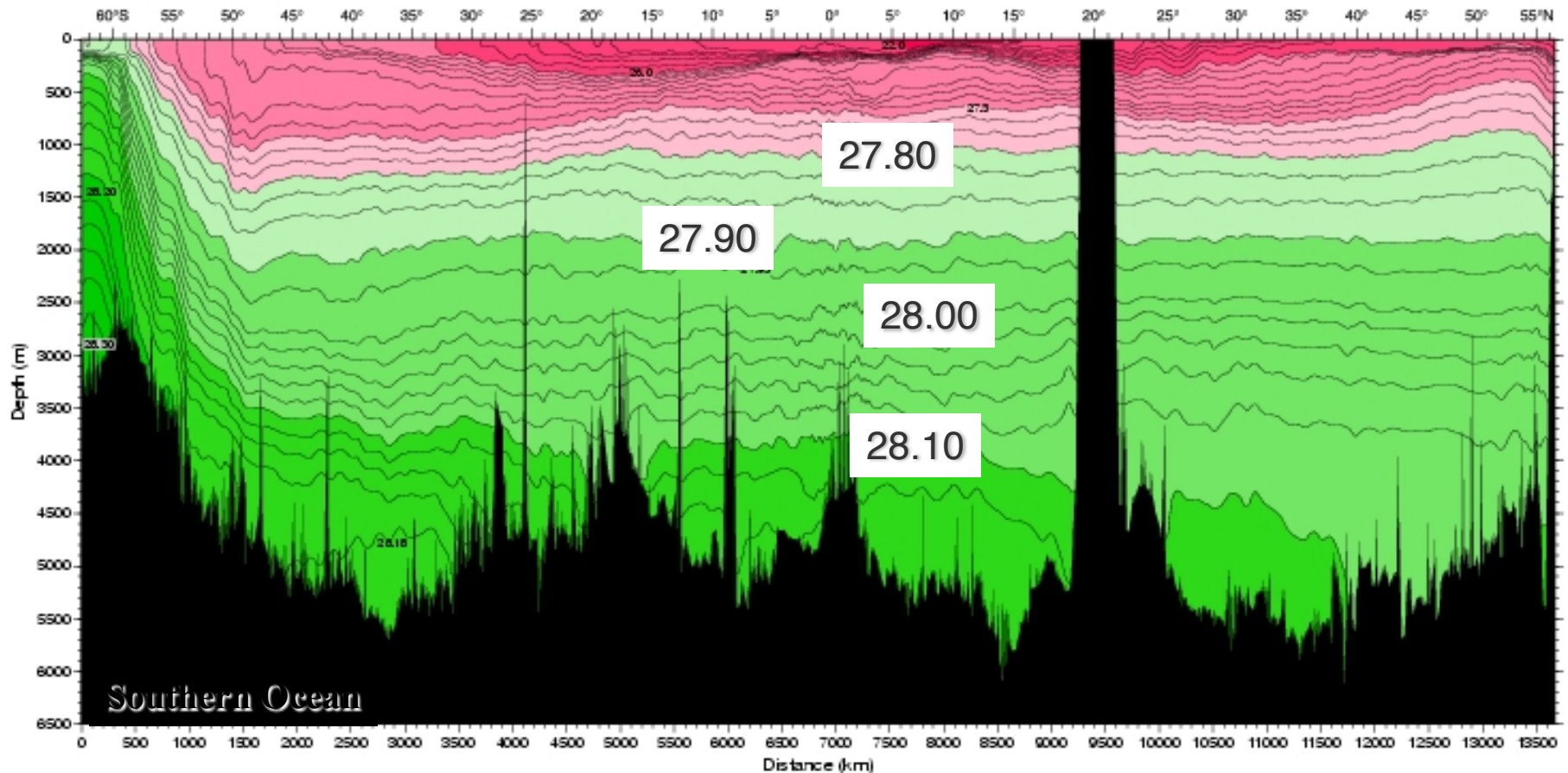
Returning to the meridional section of density across the ocean, we can quantify $\partial\rho/\partial z$ and $\partial^2\rho/\partial z^2$ in the abyssal ocean volume of interest...

How is the conclusion that $K_\rho \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ reached?

Density typically increases by $\sim 0.3 \text{ kg m}^{-3}$ between 1000 and 4000 m, yielding $\partial\rho/\partial z \sim -10^{-4} \text{ kg m}^{-4}$.

Density typically increases by $\sim 0.1 \text{ kg m}^{-3}$ between 1000 and 2000 m, and by $\sim 0.04 \text{ kg m}^{-3}$ between 3000 and 4000 m. This implies that $\partial^2\rho/\partial z^2 \sim -3 \times 10^{-8} \text{ kg m}^{-5}$.

Plugging these values into the advection - diffusion balance...



How is the conclusion that $K_\rho \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ reached?

Plugging in the values that we have just obtained for w , $\partial\rho/\partial z$ and $\partial^2\rho/\partial z^2$ into the advection - diffusion balance equation

$$K_\rho = (w \partial\rho/\partial z) / (\partial^2\rho/\partial z^2)$$

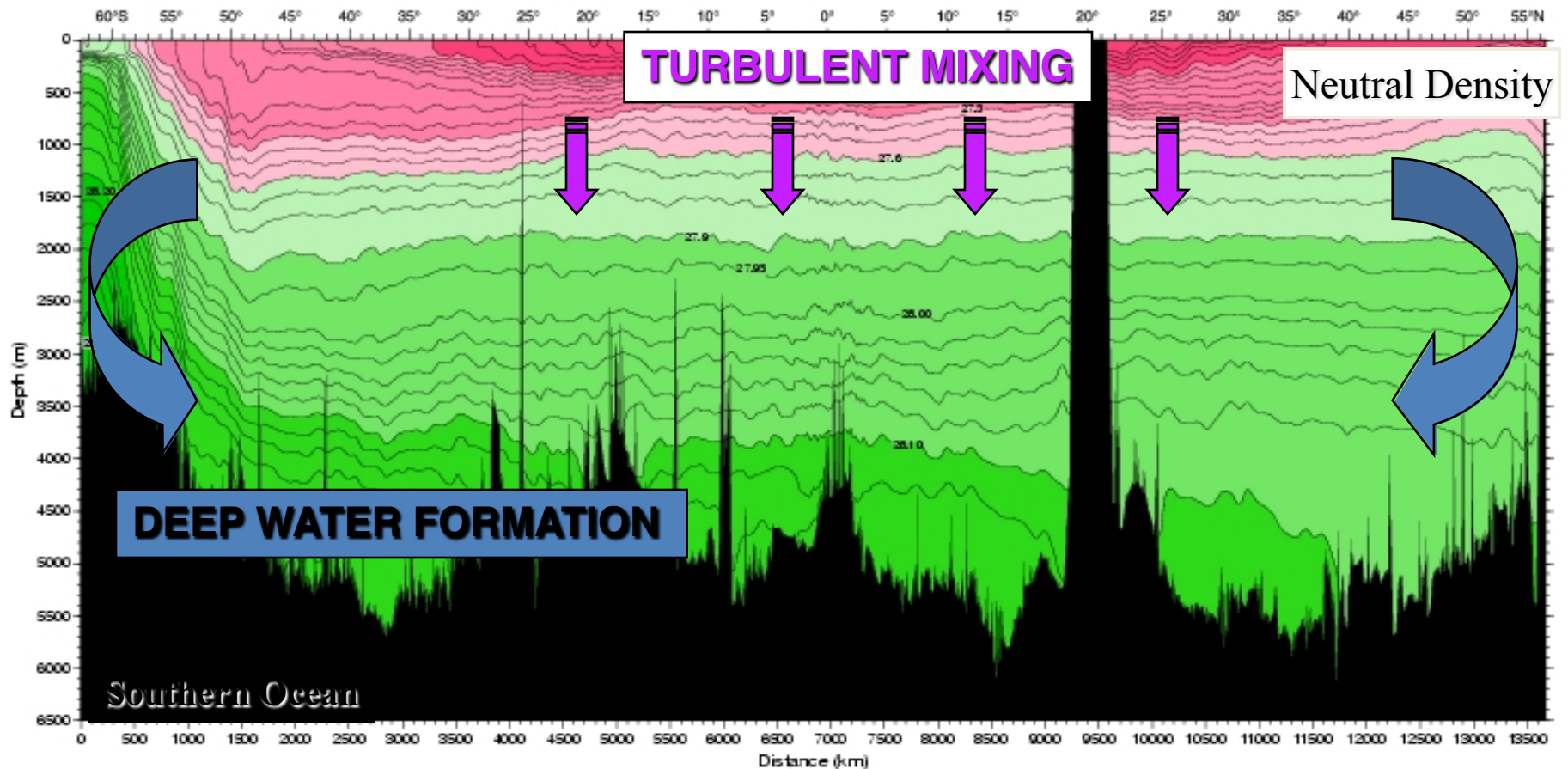
we obtain

$$K_\rho \sim (10^{-7} \text{ m s}^{-1}) \times (-10^{-4} \text{ kg m}^{-4}) / (-3 \times 10^{-8} \text{ kg m}^{-5}) \sim 3 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$$

which agrees approximately with Munk and Wunsch's estimate.

What does this imply for the ocean's energy budget?

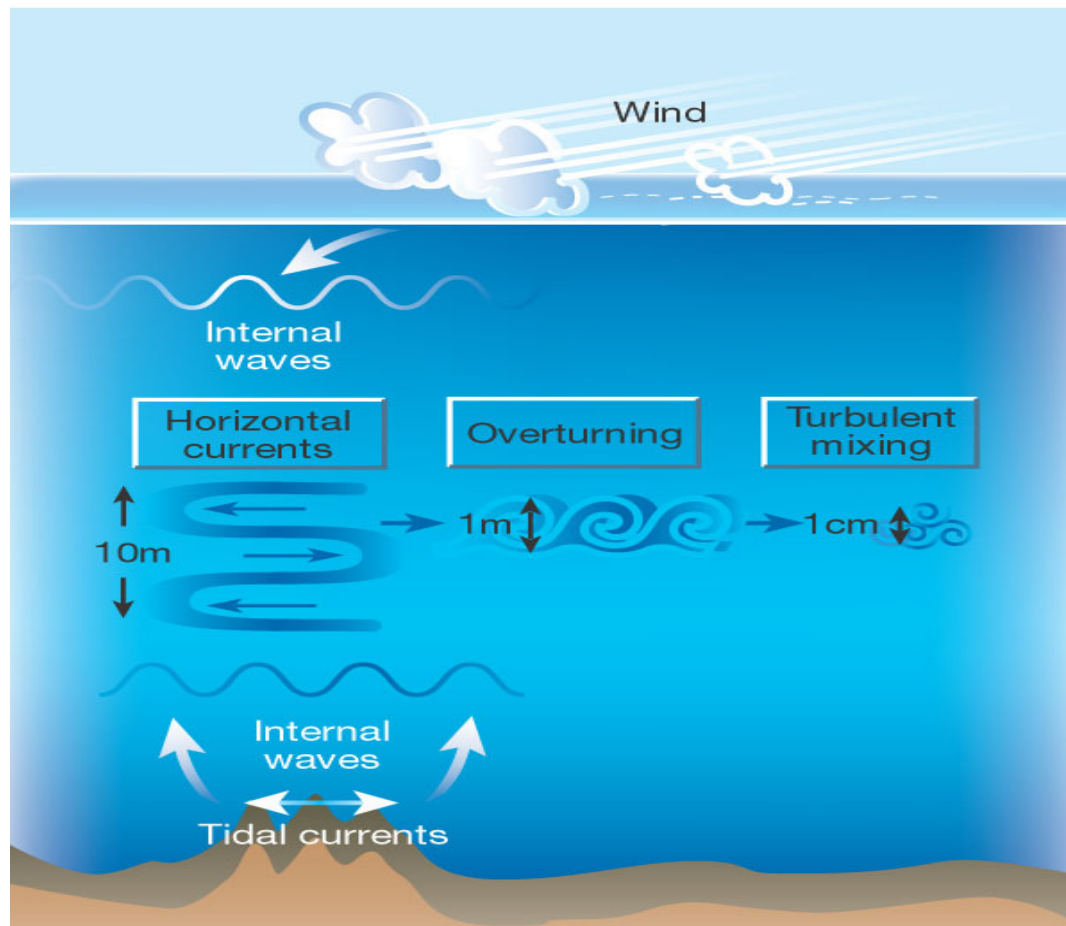
Using $K_p = \gamma \varepsilon / N^2$, where γ is the mixing efficiency (0.2) and ε is the turbulent kinetic energy dissipation rate, and integrating over the global abyssal ocean, Munk and Wunsch show that to sustain the ocean stratification and drive the MOC via turbulent mixing, winds and tides must supply kinetic and potential energy to the deep ocean at a rate of 2 - 3 TW (with 1.5 - 2.5 TW being dissipated by turbulence and ~ 0.5 TW being expended in raising the ocean's centre of mass).



What powers the ocean circulation?

(Where does all the energy come from?)

The energy from winds and tides is imagined to reach the ocean interior via the propagation of internal waves generated as the mixed layer is deformed by the wind or as tide- or wind-generated flows impinge on rough topography. Internal waves would then lead to turbulent mixing upon breaking.



Gregg (2003)

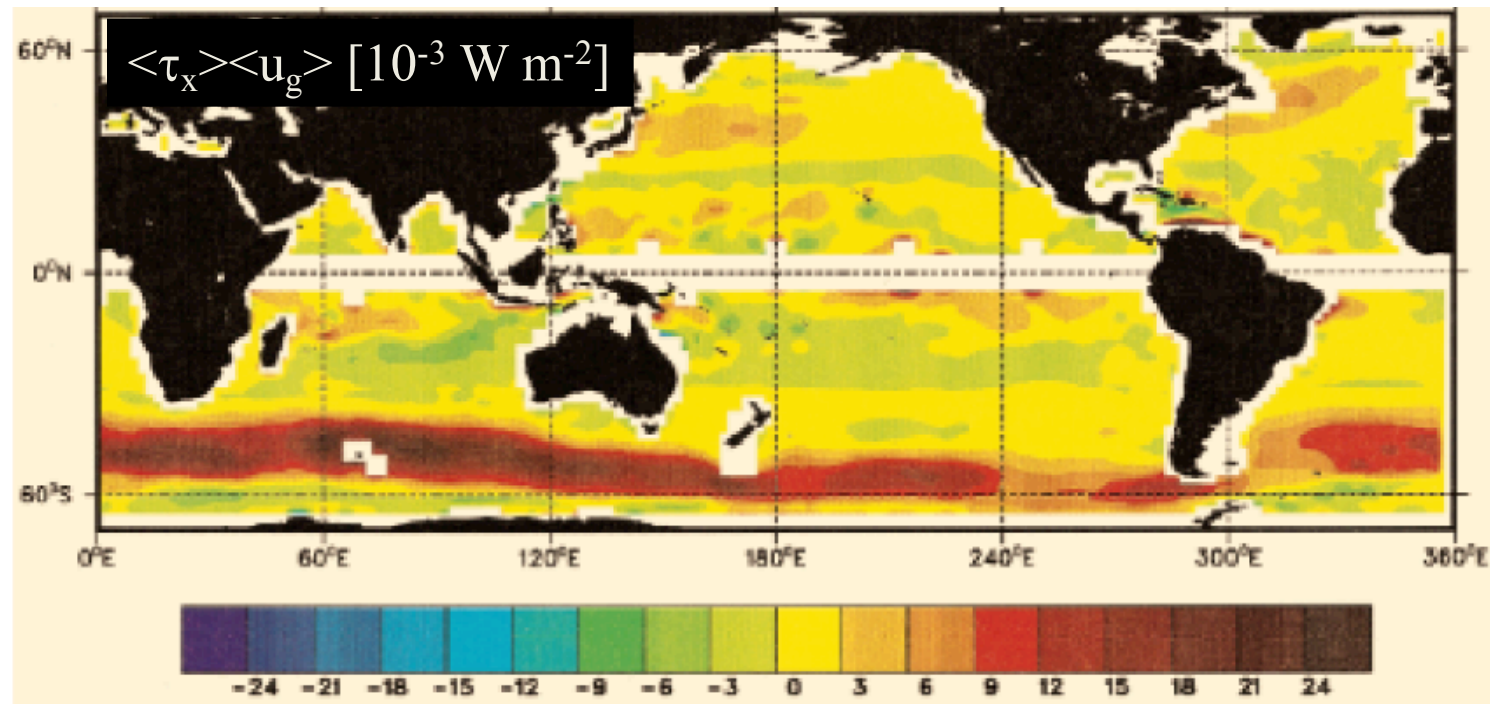
What powers the ocean circulation?

(Where does all the energy come from?)

The plausibility of the traditional ocean mixing paradigm is suggested by the broad correspondence between the power required to support it (2 - 3 TW) and estimates of the rate at which energy is input to the deep ocean by tides and winds, which contribute ~ 1 TW each.

The wind work (= rate of energy input) can be split into two components: wind work on the ocean's background geostrophic flow; and wind work on upper-ocean near-inertial motions (remember that internal waves must have frequencies between f , the inertial frequency, and N , the buoyancy frequency). The former likely dominates the total wind work. Next, we examine some state-of-the-art estimates of these two components.

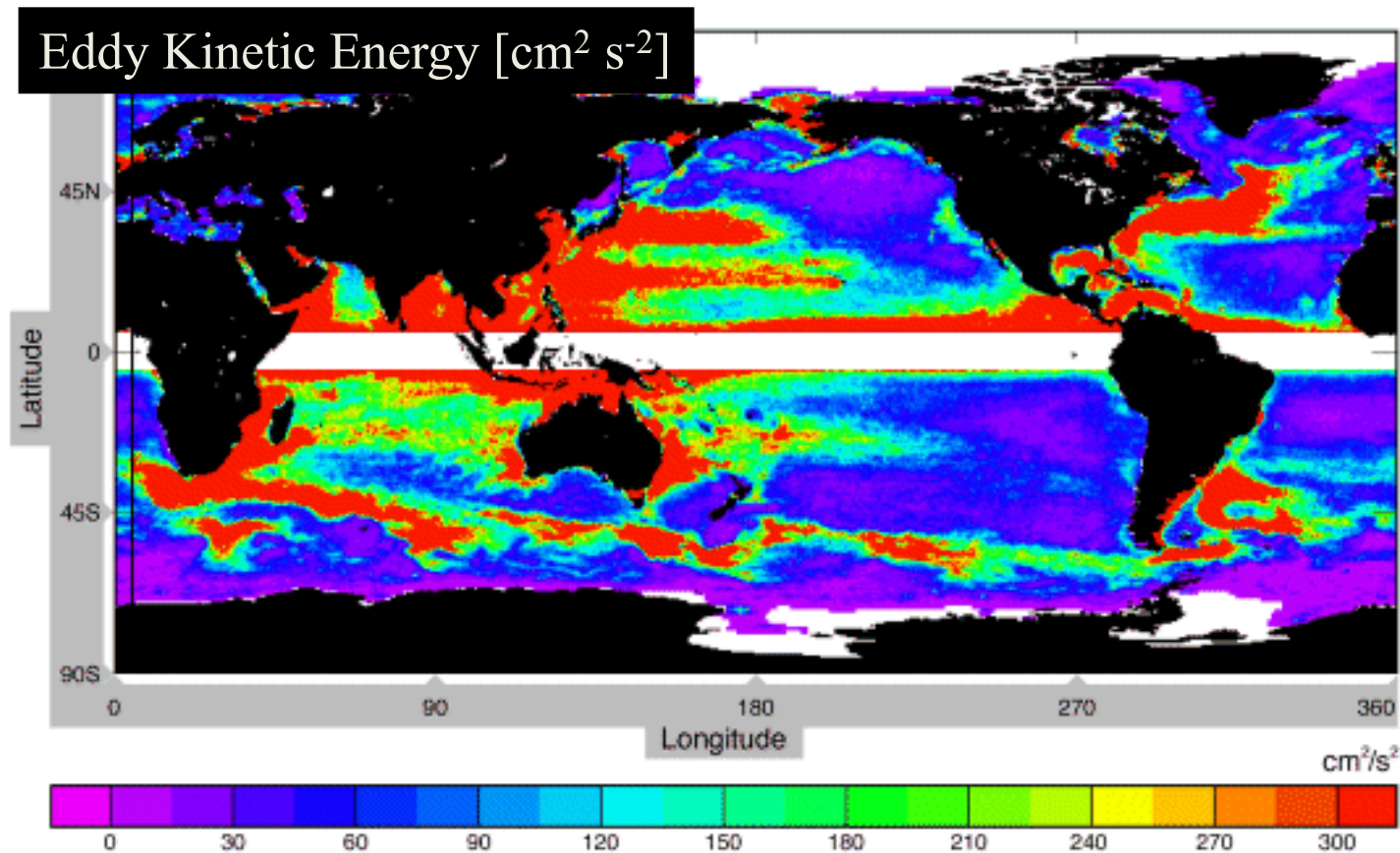
Wind work on the geostrophic circulation of the ocean



Wunsch (1998)

Global wind work $\sim 0.8 \text{ TW}$ ($\sim 80\%$ in the ACC) cf. $\leq 2 - 3 \text{ TW}$ required to drive MOC.

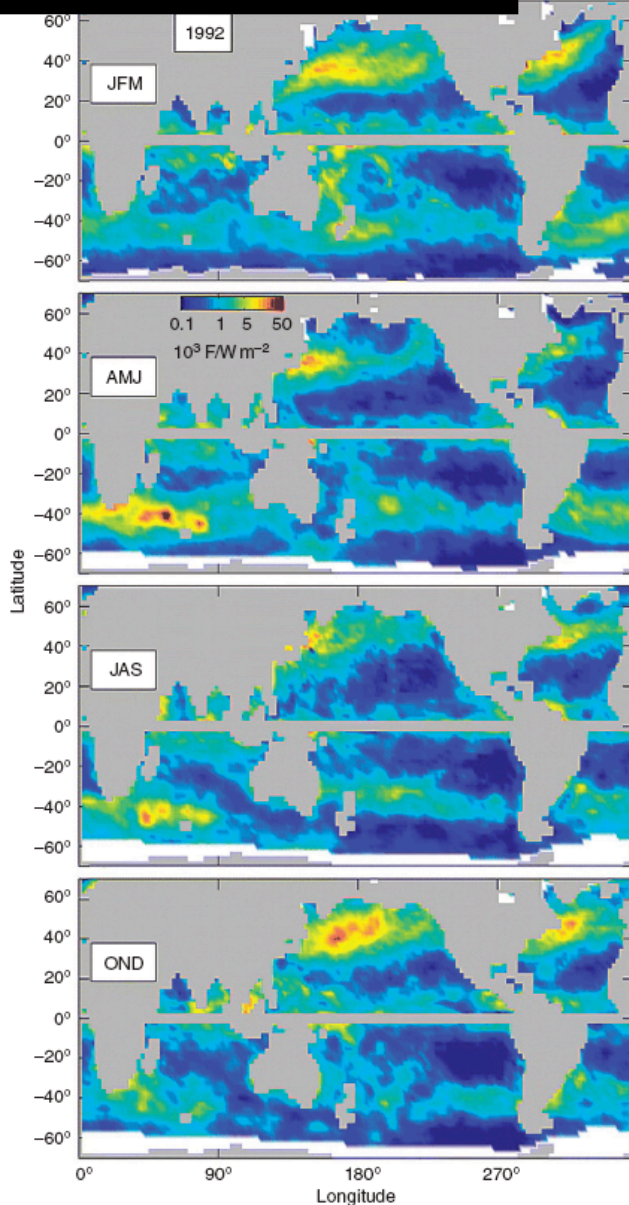
Fate of the wind work on the geostrophic circulation of the ocean



The bulk of this energy is transferred to the ocean's mesoscale eddy field via baroclinic instability. The subsequent pathway toward dissipation is uncertain. The traditional paradigm of ocean mixing assumes that the energy eventually finds its way to the internal wave field via several poorly understood processes.

Wind work on upper-ocean inertial motions

$$\langle \tau_x \rangle \langle u_i \rangle [10^{-3} \text{ W m}^{-2}]$$

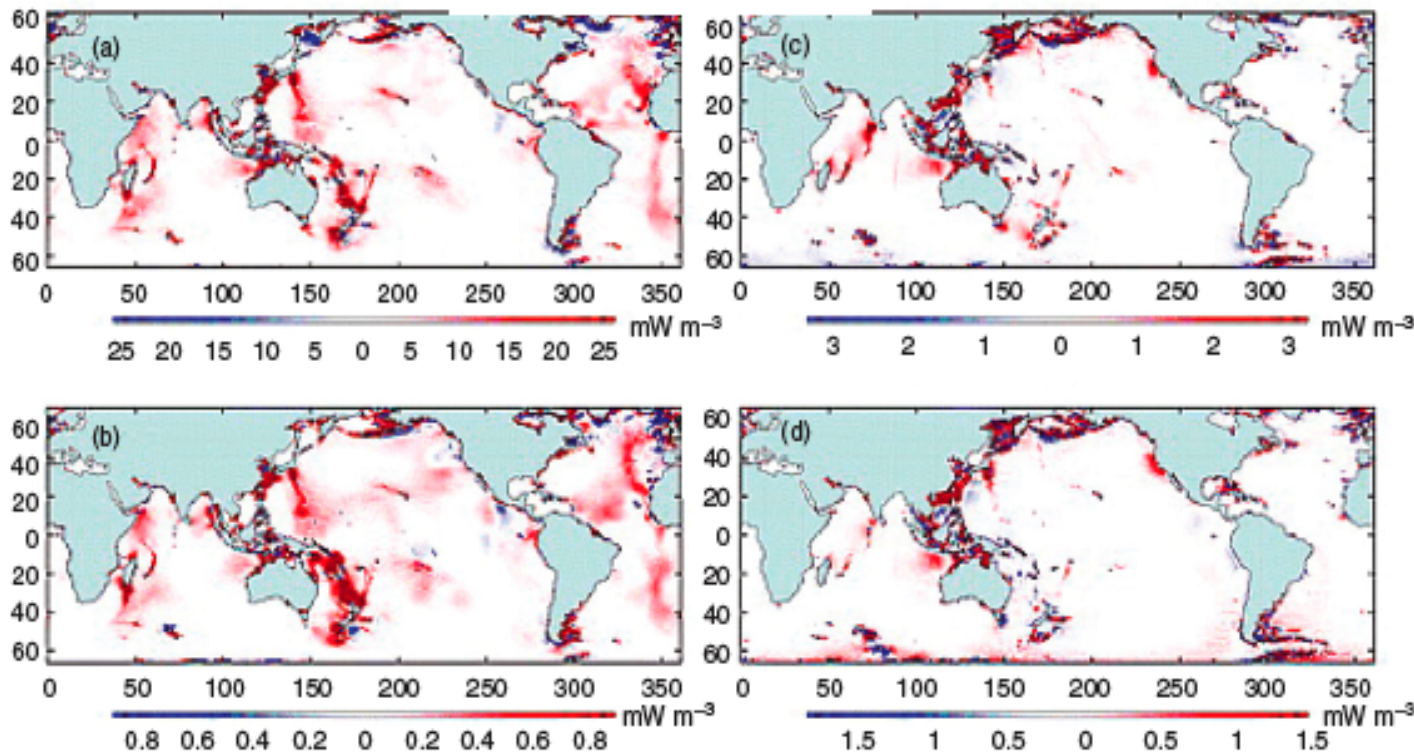


Global wind work on mixed layer $\sim 0.5 \text{ TW}$ (focused on wintertime, mid-latitude storm tracks), but only a minor fraction (20 - 30%) may enter deep ocean, with the rest being expended in the upper-ocean mixed layer.

This energy is put directly into frequencies that can propagate as internal waves. Near-inertial internal waves are present widely across the top 0.5-2 km of the ocean, although they carry relatively little energy and so the associated turbulent dissipation and mixing are likely small.

Tidal work on the ocean

The rate at which the sun and the moon work on the ocean has been estimated as ~ 3.5 TW, but much (~ 2.6 TW) of this energy is dissipated in shallow shelf seas and does not access the deep ocean. The remaining ~ 0.9 TW enters the deep ocean as a barotropic tide that forces flow over rough and steep topography. In doing so, it generates internal tides and boundary layer turbulence. Internal tide generation is focused over mid-ocean ridges, continental slopes and island arcs.



Maps of dissipation of the four major tidal constituents, estimated from altimetry: (a) M_2 ; (b) N_2 ; (c) K_1 ; (d) O_1 .

Egbert and Ray (2003)

Motivation for observing vertical ocean Mixing

- The traditional paradigm of ocean mixing proposes that turbulent mixing by breaking internal waves (at a rate of $K_p \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$) sustains oceanic stratification and overturning circulation. The energy for turbulent mixing is thought to be supplied by winds and tides at a total rate of 2 - 3 TW.

Observations of vertical turbulent mixing in the deep ocean

While the traditional paradigm of ocean mixing is plausible in terms of the available energy sources, observations of turbulent mixing in the ocean appear to challenge its validity above the deepest layers of the ocean...

Evidence for slow mixing across the pycnocline from an open-ocean tracer-release experiment

James R. Ledwell*, Andrew J. Watson†
& Clifford S. Law†

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Woods Hole Oceanographic Institution, Woods Hole,
Massachusetts 02543, USA

† Plymouth Marine Laboratory, Prospect Place, West Hoe,
Plymouth PL1 3DH, UK

THE distributions of heat, salt and trace substances in the ocean thermocline depend on mixing along and across surfaces of equal density (isopycnal and diapycnal mixing, respectively). Measurements of the invasion of anthropogenic tracers, such as bomb tritium and ^3He (see, for example, refs 1 and 2), have indicated that isopycnal processes dominate diapycnal mixing, and turbulence measurements have suggested that diapycnal mixing is small^{3,4}, but it has not been possible to measure accurately the diapycnal diffusivity. Here we report such a measurement, obtained from the vertical dispersal of a patch of the inert compound SF_6 released in the open ocean. The diapycnal diffusivity, averaged over hundreds of kilometres and five months, was $0.11 \pm 0.02 \text{ cm}^2 \text{ s}^{-1}$, confirming previous estimates¹⁻⁴. Such a low diffusivity can support only a rather small diapycnal flux of nitrate into the euphotic zone; it justifies the neglect of diapycnal mixing in dynamic models of the thermocline²⁵⁻²⁷, and implies that heat, salt and tracers must penetrate the thermocline mostly by transport along, rather than across, density surfaces.

Diapycnal spreading of a man-made tracer released in the NE Atlantic thermocline

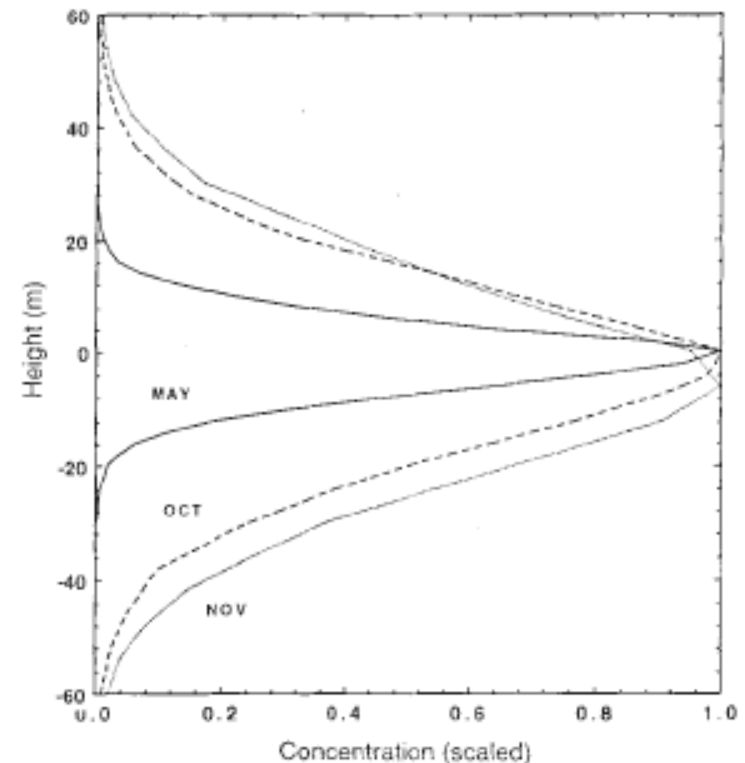


FIG. 2 Evolution of the vertical distribution of the tracer. The mean profiles have been scaled so that the widths can be compared.

Observations of vertical turbulent mixing in the deep ocean

While the traditional paradigm of ocean mixing is plausible in terms of the available energy sources, observations of turbulent mixing in the ocean appear to challenge its validity above the deepest layers of the ocean...

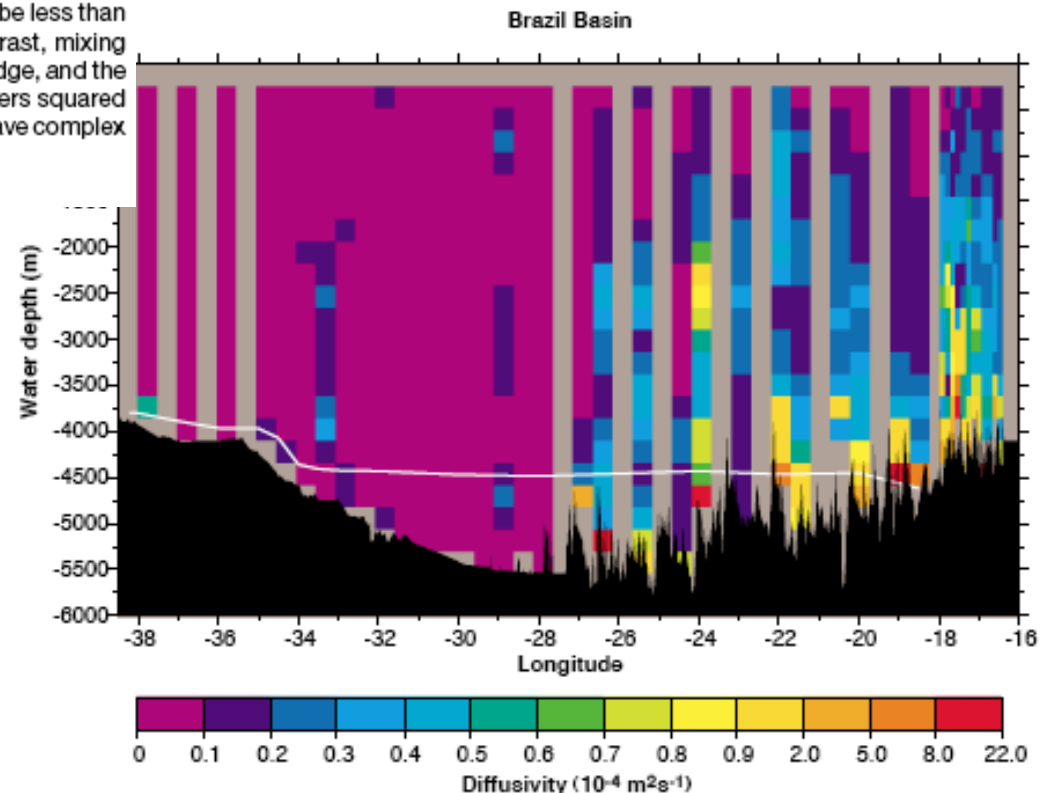
Spatial Variability of Turbulent Mixing in the Abyssal Ocean

K. L. Polzin,* J. M. Toole, J. R. Ledwell, R. W. Schmitt

Ocean microstructure data show that turbulent mixing in the deep Brazil Basin of the South Atlantic Ocean is weak at all depths above smooth abyssal plains and the South American Continental Rise. The diapycnal diffusivity there was estimated to be less than or approximately equal to 0.1×10^{-4} meters squared per second. In contrast, mixing rates are large throughout the water column above the rough Mid-Atlantic Ridge, and the diffusivity deduced for the bottom-most 150 meters exceeds 5×10^{-4} meters squared per second. Such patterns in vertical mixing imply that abyssal circulations have complex spatial structures that are linked to the underlying bathymetry.



Turbulent diapycnal diffusivity across the Brazil Basin of the South Atlantic measured with a microstructure profiler

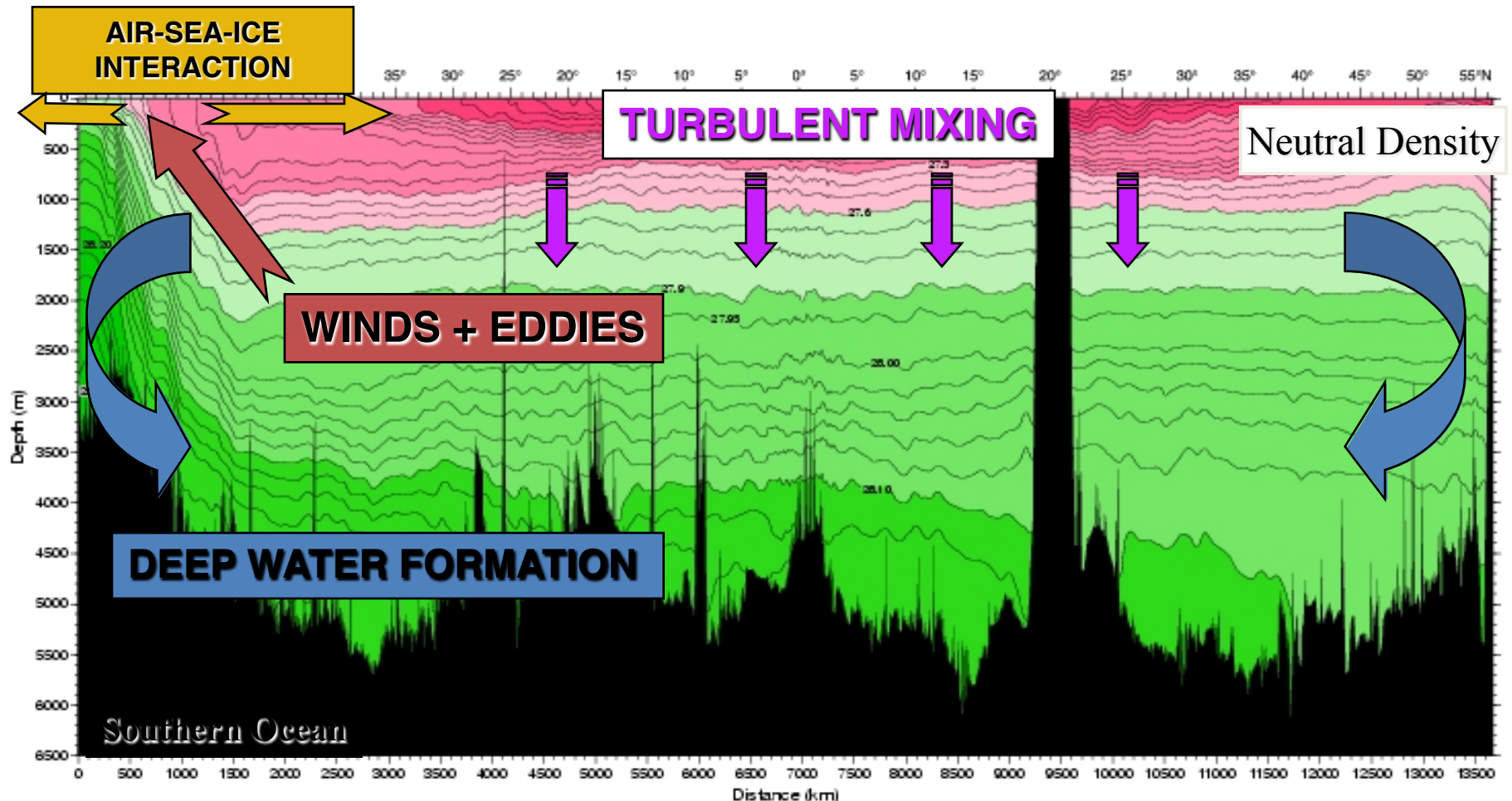


Summary - The Role of vertical mixing in Ocean Circulation

- The oceanic stratification and overturning circulation owe their existence to mechanical (rather than buoyancy) forcing.
- The essential energetics consists of the conversion of 2 - 3 TW of wind and tidal work into oceanic potential and kinetic energy through the generation of the large-scale ocean circulation.
- The traditional paradigm of ocean mixing proposes that small-scale turbulence (primarily due to internal wave breaking / mixing at sills?) is the critical process that links energy input with energy dissipation, thereby sustaining the large-scale ocean circulation.
- Direct observations of turbulent mixing in the ocean challenge the validity of the traditional paradigm above the ocean's deepest layers, and thus motivate an alternative paradigm.

An alternative paradigm of ocean circulation: the permanent pycnocline

An alternative paradigm, consistent with observations of weak turbulent diapycnal mixing across the permanent pycnocline, suggests that the requirement that $K_p \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ is relaxed if much of the deep water upwells along isopycnals, primarily in the Southern Ocean. The energy input required to sustain the ocean stratification and drive the MOC in this way is $\sim 0.6 \text{ TW} \ll 2 - 3 \text{ TW}$.



An alternative paradigm of ocean circulation: the permanent pycnocline

These ideas stem from a family of theoretical models describing the renewal of water masses in the permanent pycnocline, and showing that a realistic circulation may be obtained in the uppermost ~ 1 km of the ocean without any need for diapycnal mixing.

The Ventilated Thermocline¹

J. R. LUYTEN, J. PEDLOSKY AND H. STOMMEL

Woods Hole Oceanographic Institution, Woods Hole, MA 02543

(Manuscript received 3 May 1982, in final form 24 August 1982)

ABSTRACT

A simple theoretical model for the oceanic thermocline and the associated field of currents is presented. The model consists of a finite but arbitrarily large number of inviscid, homogeneous fluid layers each with a different density. The dynamical balances everywhere are Sverdrupian. In regions where the Ekman pumping is negative (downward) the surface density is specified, i.e., the position of the outcrop of density interfaces is specified. This outcropping of density layers allows deep motion to be excited by the ventilation provided by Ekman pumping even in latitudes far south of the outcrop where the layer is shielded from direct influence of the wind. Analytical solutions are presented in the case where the density-outcrop lines are coincident with latitude circles. The solutions are not self-similar and important sub-domains of the solution are defined by critical potential vorticity trajectories which separate the ventilated from the unventilated regions in the lower thermocline. These critical trajectories also separate regions of strong variations in potential vorticity from regions of fairly weak variation in potential vorticity although the small variations in potential vorticity in the latter are crucial to the dynamics.

Comparison is made between the predictions of the model and data from the Atlantic with encouraging results.

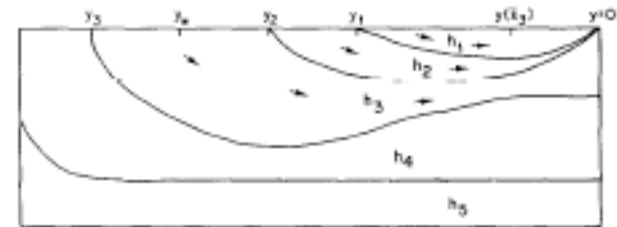
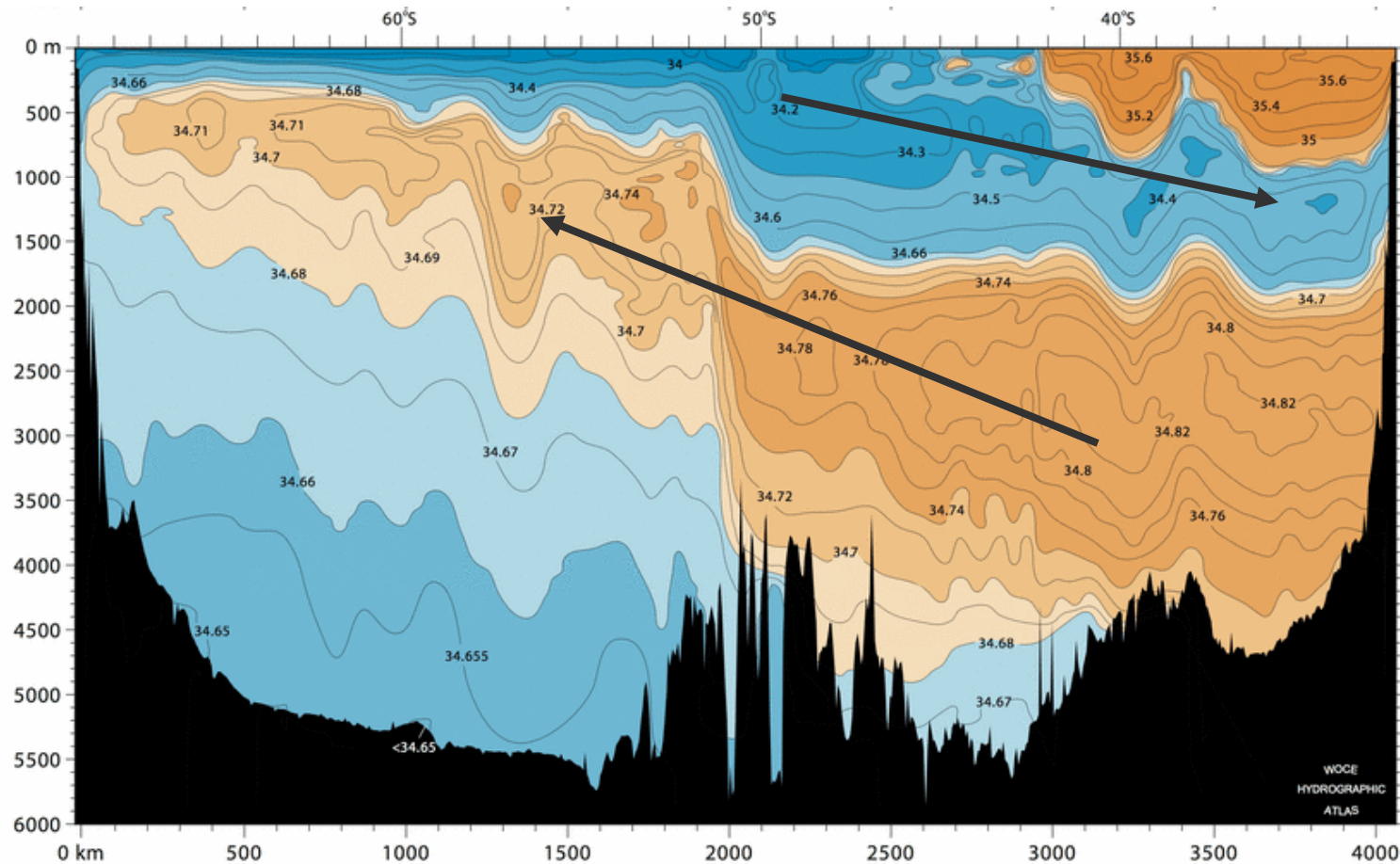


FIG. 1. A schematic meridional cross section of the model. The latitudes of the density outcrops are y_1 , y_2 and y_3 as described in the text. The Ekman pumping changes sign at $y_4(x)$. South of $y(x_3)$ layer 3 is motionless.

These models invoke a three-way interaction between wind-forced (Ekman) vertical motion, isopycnal transport of water masses by mesoscale eddies, and air-sea buoyancy fluxes, to explain how the subduction of relatively unmodified upper ocean waters into the ocean interior comes about.

How much deep water upwells in the Southern Ocean?

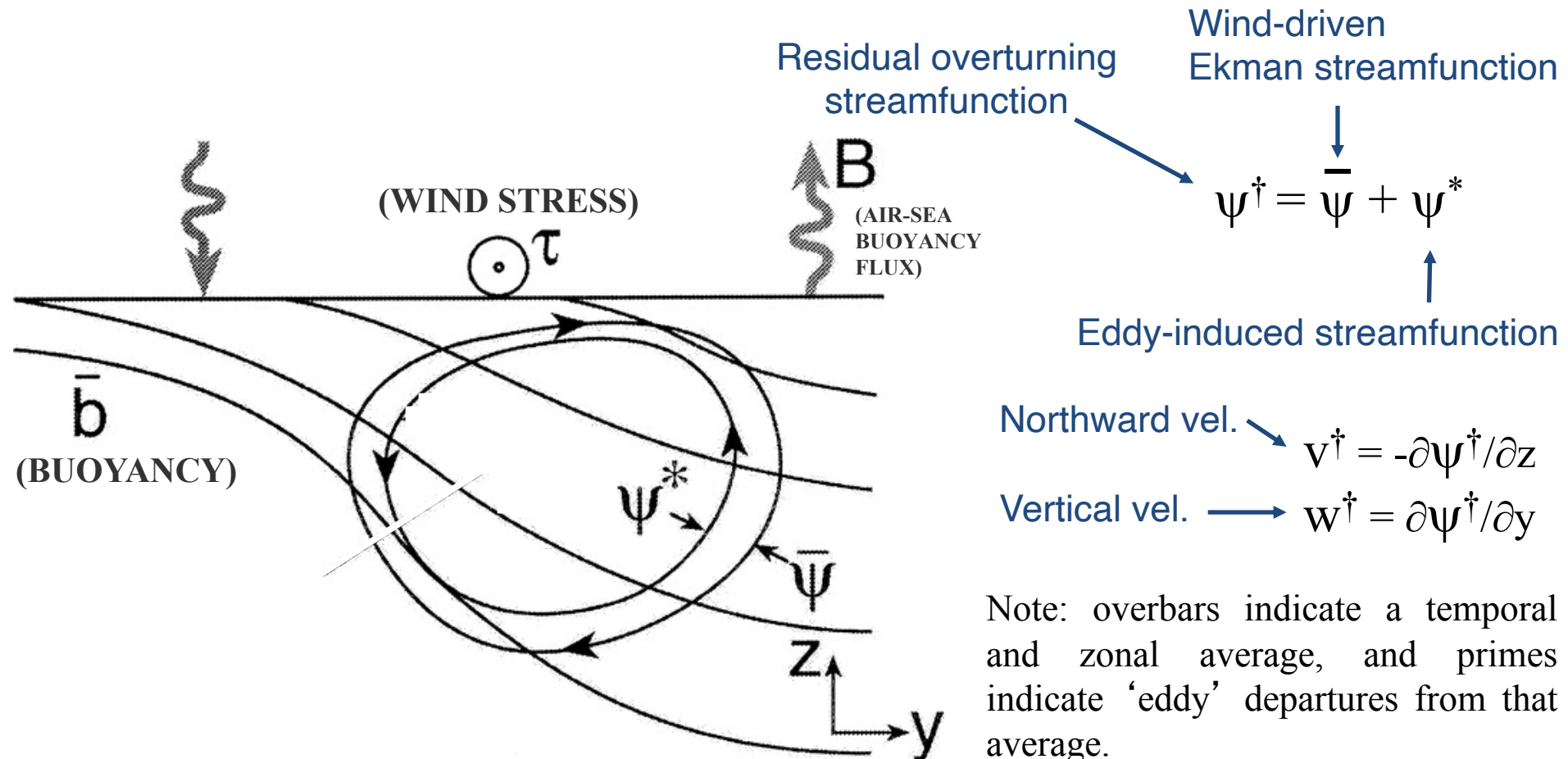
The extent to which the upwelling of deep water in the Southern Ocean may relax the requirement that $K_p \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ depends on how much water upwells as a function of density. We know from water mass properties that at least some deep water does upwell in the Southern Ocean...



Salinity section across the Southern Ocean south of Africa

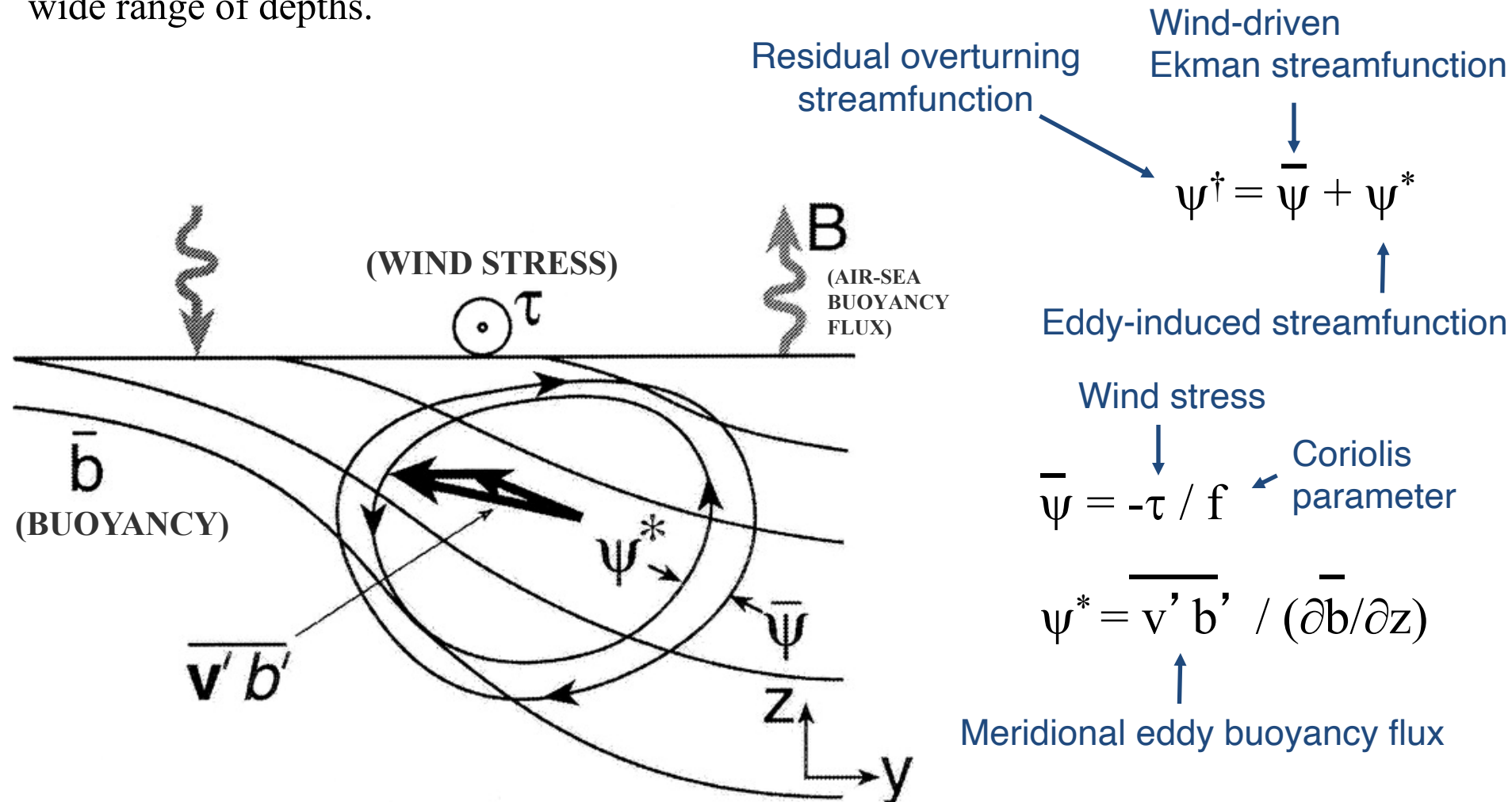
How much deep water upwells in the Southern Ocean?

The extent to which the upwelling of deep water in the Southern Ocean may relax the requirement that $K_p \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ depends on how much water upwells as a function of density. Our current understanding of this subject owes much to a family of *residual mean* analyses (e.g., Johnson & Bryden, 1989; Karsten & Marshall, 2002; Marshall & Radko, 2003...), which describe the meridional overturning circulation of the Southern Ocean as resulting from two largely opposing contributions:



How much deep water upwells in the Southern Ocean?

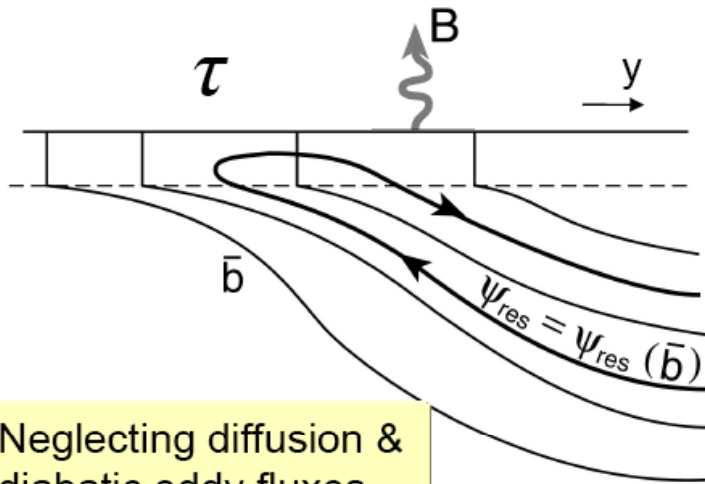
The Ekman overturning consists of a thin (~ 100 m) northward Ekman flow at the surface, and a much thicker southward geostrophic flow below the crests of topographic features, where time-mean zonal pressure gradients can be sustained. It can be quantified from the wind stress. The eddy-induced overturning results from eddies flattening isopycnals, over a wide range of depths.



How much deep water upwells in the Southern Ocean?

To estimate the rate of overturning, consider the buoyancy ($b = -g \rho / \rho_0$) budget of the residual overturning circulation....

At the sea surface, the meridional advection of buoyancy by the residual circulation has to match the air-sea buoyancy flux for a steady state to be maintained. In the limit of no diapycnal mixing in the ocean interior, the residual overturning streamfunction value set at the surface outcrop of an isopycnal will stay constant as that isopycnal extends down into the ocean interior.



Thus, at each isopycnal, we can estimate the residual overturning streamfunction $\psi^\dagger$ (i.e. the rate of upwelling and downwelling) from the air-sea buoyancy flux at the surface outcrop of that isopycnal:

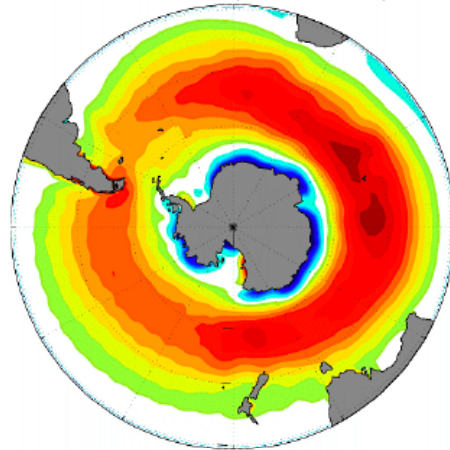
$$\psi^\dagger (\partial \bar{b} / \partial y) = B$$

How much deep water upwells in the Southern Ocean?

We can now infer the residual overturning circulation of the Southern Ocean from observations of the air-sea buoyancy flux, and show that (because of the eddy contribution to the overturning) it is very different from the wind-driven Ekman contribution (as estimated from wind data).

Wind Stress

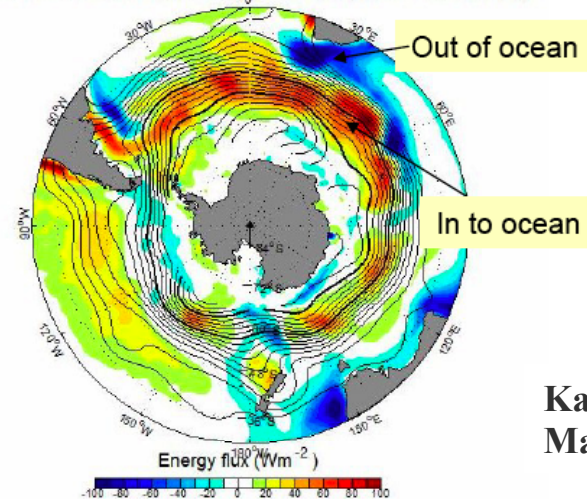
Annual mean TAU_X (NCEP, 1980–2000)



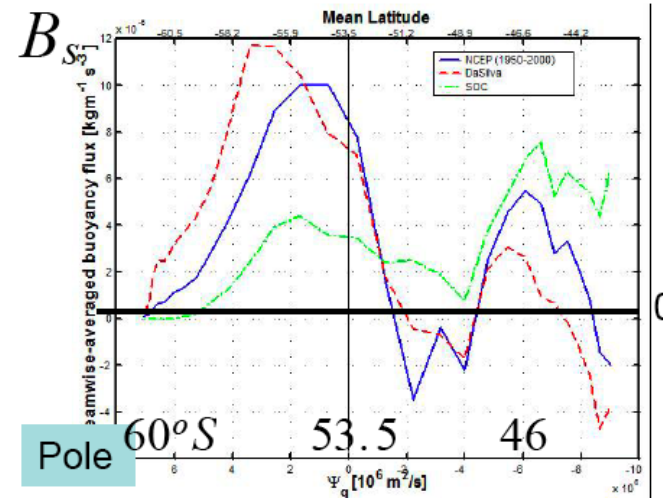
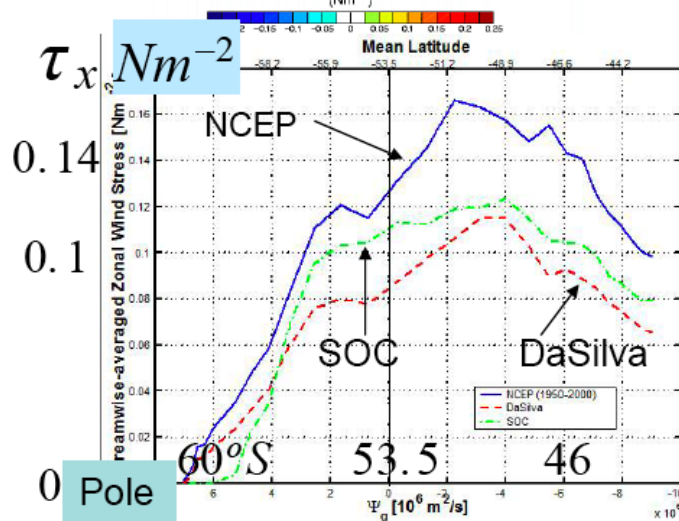
Air-sea buoyancy flux

Annual mean net surface heating of ocean (NCEP 1950–2000)

NCEP



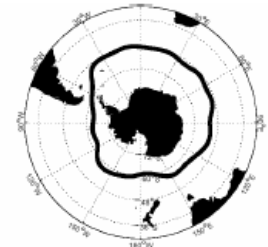
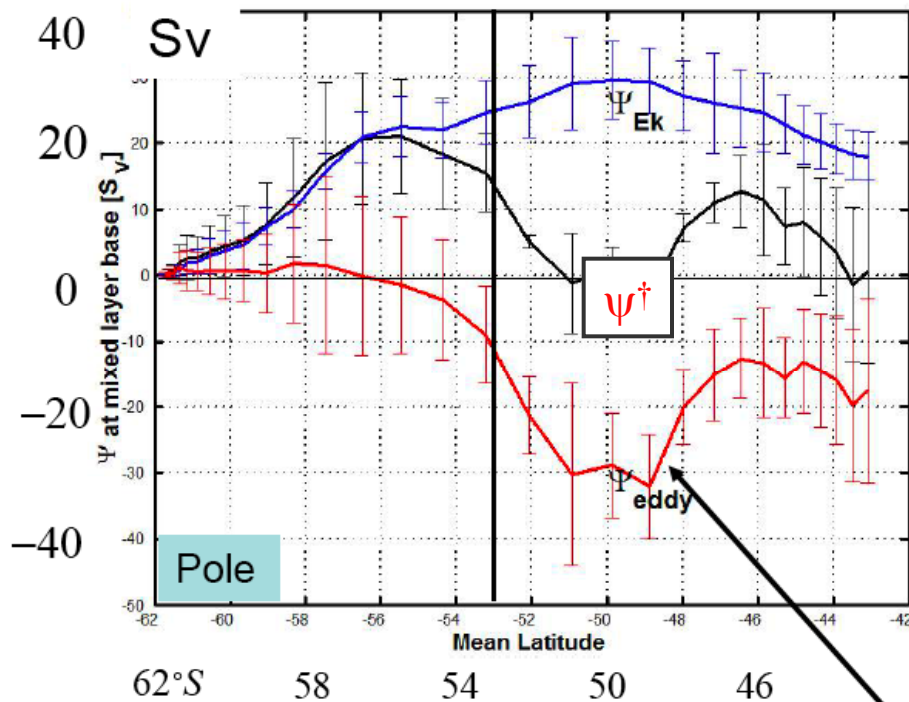
Karsten & Marshall (2002)



How much deep water upwells in the Southern Ocean?

The residual overturning is significantly weaker than the Ekman overturning, and there is an important question as to whether eddies are able to explain the difference. Large uncertainties exist, but the suggestion is that ~20 Sv of deep water is upwelled along isopycnals in the Southern Ocean → **Implications for the traditional paradigm of ocean mixing.**

Cross-Stream transports (Up- or downwelling)



$$\overline{\Psi} = -\frac{\tau}{f}$$

Ekman

$$\Psi^\dagger = \frac{B_s}{\frac{\partial \bar{b}}{\partial y}}$$

Residual overturning

$$\Psi^\dagger - \overline{\Psi}$$

'Inferred' eddy transport

Can eddies achieve such a large transport?

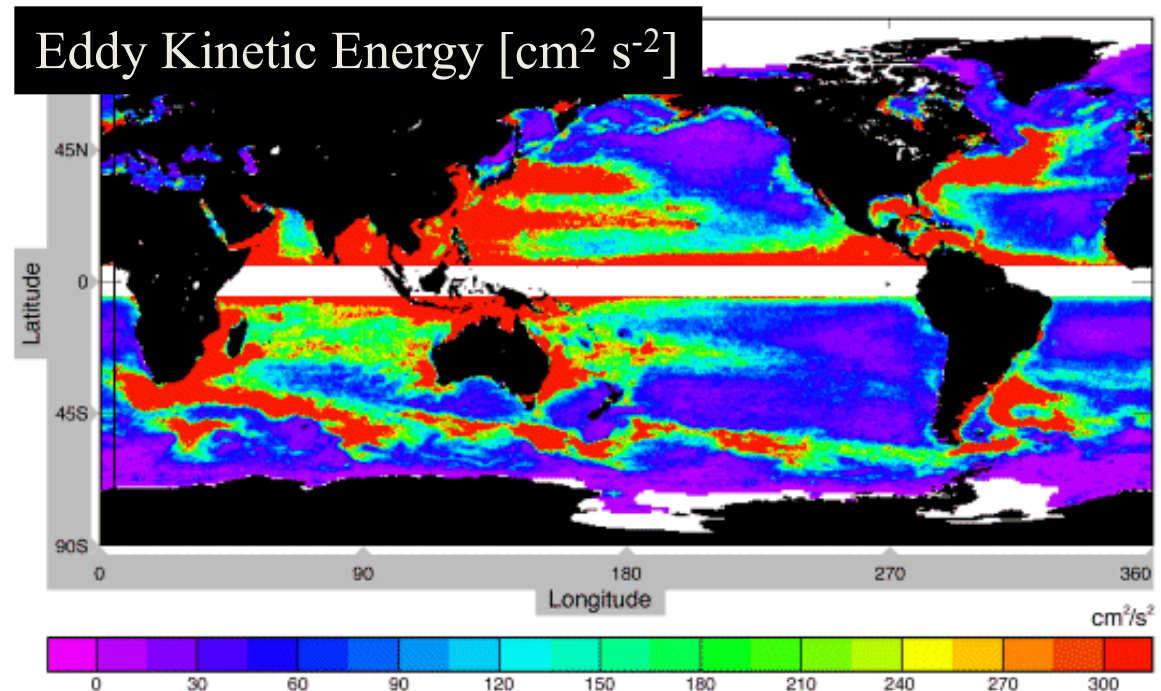
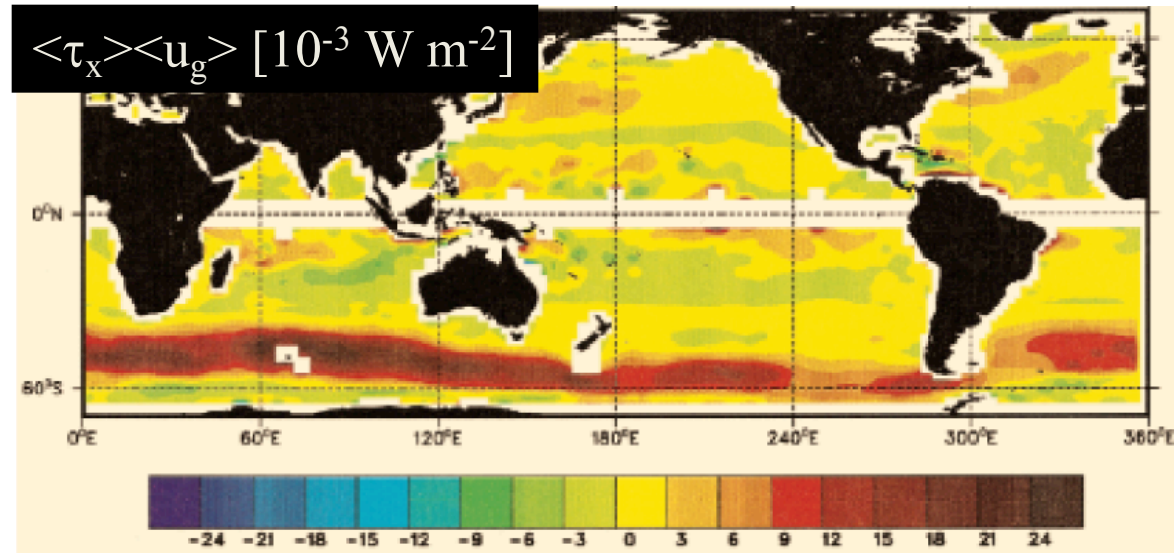
Issues with the alternative ocean circulation paradigm

While isopycnal upwelling in the Southern Ocean relieves the need to find some ‘missing turbulent mixing’ (as would appear from the traditional ocean mixing paradigm), significant open questions remain. Specifically, the energy budget of the circulation is far from closed, and Southern Ocean eddies seem to have rather central (and possibly conflicting) roles in both ocean mixing paradigms...

- In the traditional paradigm, Southern Ocean eddies channel a large fraction (~ 0.6 TW or $\sim 30\%$ of the total) of the net energy input by wind and tides to the ocean toward dissipation scales, presumably leading to eventual turbulent mixing.

Fate of the wind work on the geostrophic circulation of the ocean

The bulk of the wind work is transferred to the ocean's mesoscale eddy field via baroclinic instability. The subsequent pathway toward dissipation is uncertain. The traditional paradigm of ocean mixing assumes that the energy eventually finds its way to the internal wave field via several poorly understood processes.



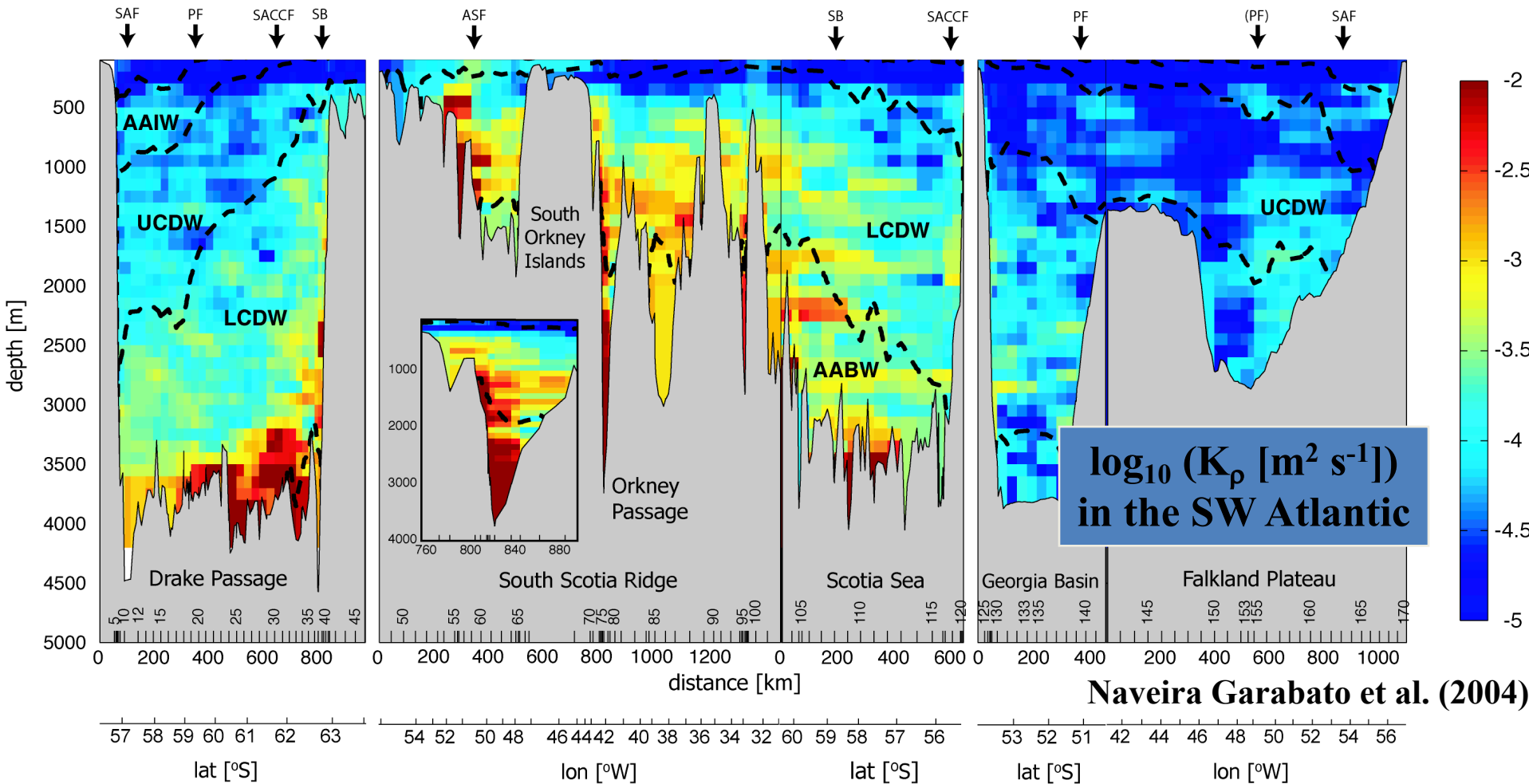
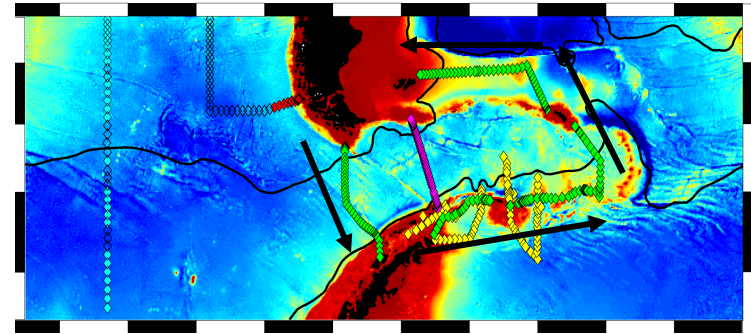
Issues with the alternative ocean mixing paradigm

While isopycnal upwelling in the Southern Ocean relieves the need to find some ‘missing turbulent mixing’ (as would appear from the traditional ocean mixing paradigm), significant open questions remain. Specifically, the energy budget of the circulation is far from closed, and Southern Ocean eddies seem to have rather central (and possibly conflicting) roles in both ocean mixing paradigms...

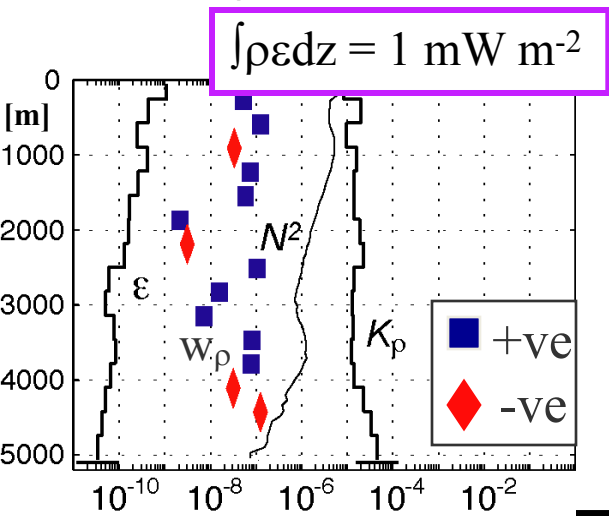
- In the traditional paradigm, Southern Ocean eddies channel a large fraction (~ 0.6 TW or $\sim 30\%$ of the total) of the net energy input by wind and tides to the ocean toward dissipation scales, presumably leading to eventual turbulent mixing.
- In the alternative paradigm, Southern Ocean eddies are also responsible for much of the isopycnal upwelling of deep water. Eddies are implicitly assumed to be dissipated in a viscous fashion, without leading to any turbulent mixing. But how do eddies actually dissipate in the ocean?

Intense turbulent mixing in the Southern Ocean associated with eddy flow - topography interaction

But the very same Southern Ocean wind work and eddy field at the core of the alternative paradigm of ocean mixing leads to intense turbulent mixing...



K_ρ , stratification, ε and w_ρ in the SE Pacific / SW Atlantic

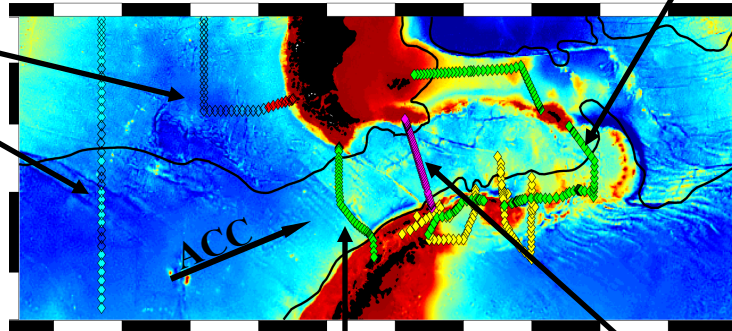
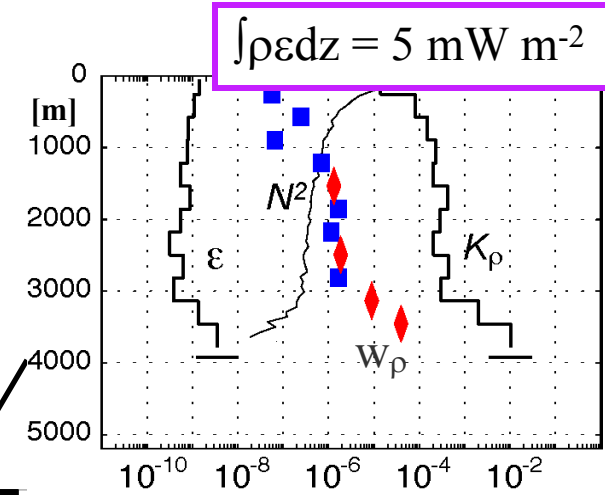


$$K_\rho [\text{m}^2 \text{ s}^{-1}]$$

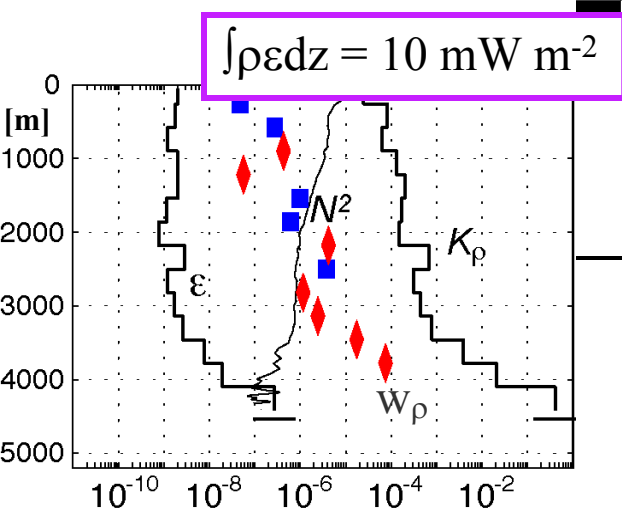
$$N^2 [\text{s}^{-2}]$$

$$\varepsilon [\text{W kg}^{-1}]$$

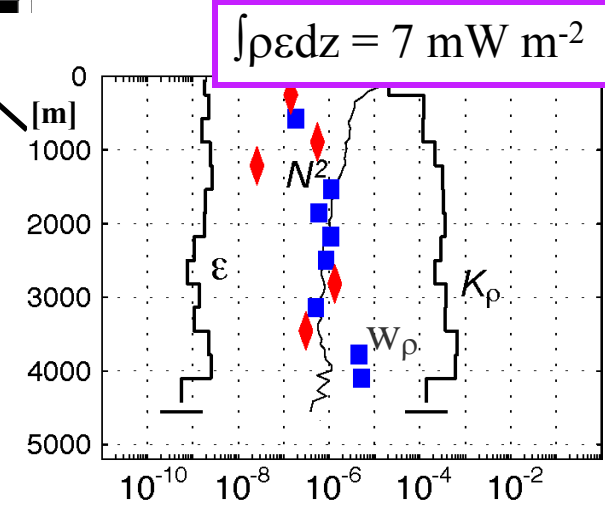
$$w_\rho = N^{-2} (K_\rho N^2)_z [\text{m s}^{-1}]$$



Naveira Garabato et al. (2004)



cf. wind work $\sim 10 \text{ mW m}^{-2}$



Coupling of the traditional and alternative paradigms

Taken together, evidence to date suggests that the two paradigms of the impact of small scale ocean turbulence on large scale ocean circulation may be coupled. The emerging picture of ocean mixing is best described as the combination of two spatially and physically intertwined ‘pycnocline isopycnal’ and ‘abyssal diapycnal’ regimes.

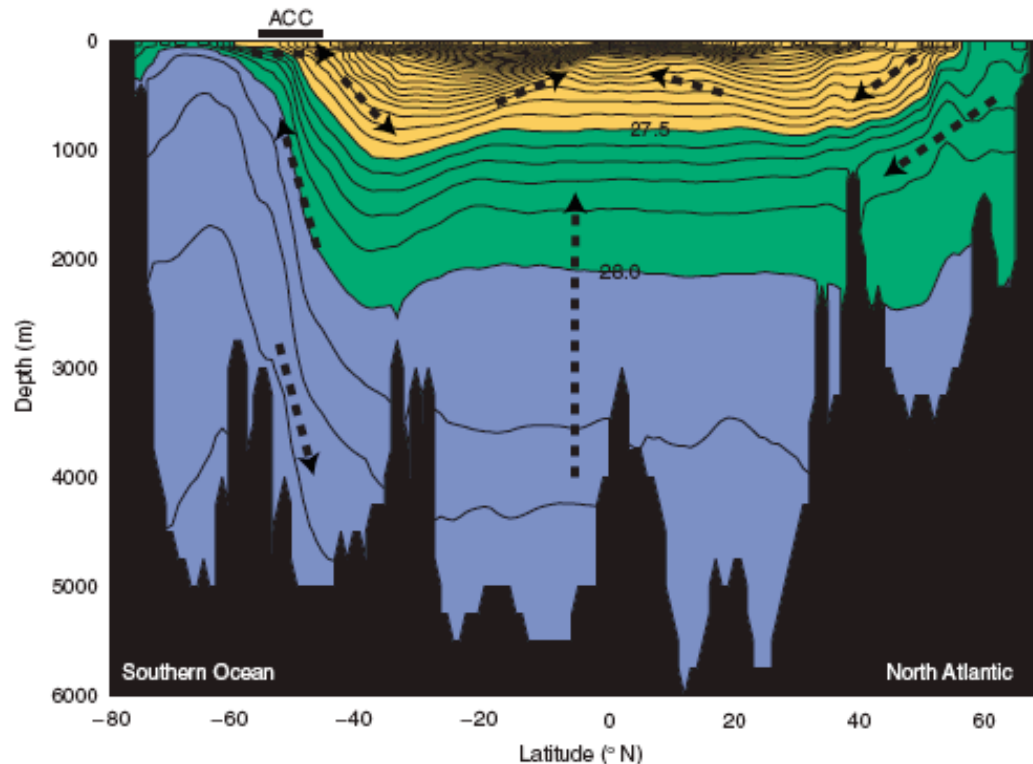


Figure 1 Vertical distribution of neutral density in the Atlantic Ocean along 30° W. Contours denote isopycnal surfaces with values between 23 and 28.4 kg m^{-3} at intervals of 0.1 kg m^{-3} . Colors indicate three density classes (separated by the 27.5 and 28.0 kg m^{-3} isopycnals) that are subject to different mixing regimes. The stratification in the abyssal ocean (purple) arises primarily from the balance between upwelling of dense water produced in the high-latitude Southern Ocean and turbulent diapycnal mixing elsewhere. In contrast, the stratification within and above the permanent pycnocline (orange) is mainly shaped by the wind- and eddy-driven subduction of near-surface water masses along isopycnals. Finally, the thick layer at the base of the permanent pycnocline, which outcrops into the upper ocean mixed layer within the Antarctic Circumpolar Current (ACC), represents a transition between the pycnocline and abyssal mixing regimes and is subject to a combination of both. The dashed arrows indicate the broad sense of the global ocean overturning driven by this set of mixing processes.

Coupling of the traditional and alternative paradigms

Recent theoretical studies (e.g. Nikurashin and Vallis, 2012) try to describe the ocean circulation in conjunction with the two paradigms.

Observational arguments:

3 regions = 3 dynamics

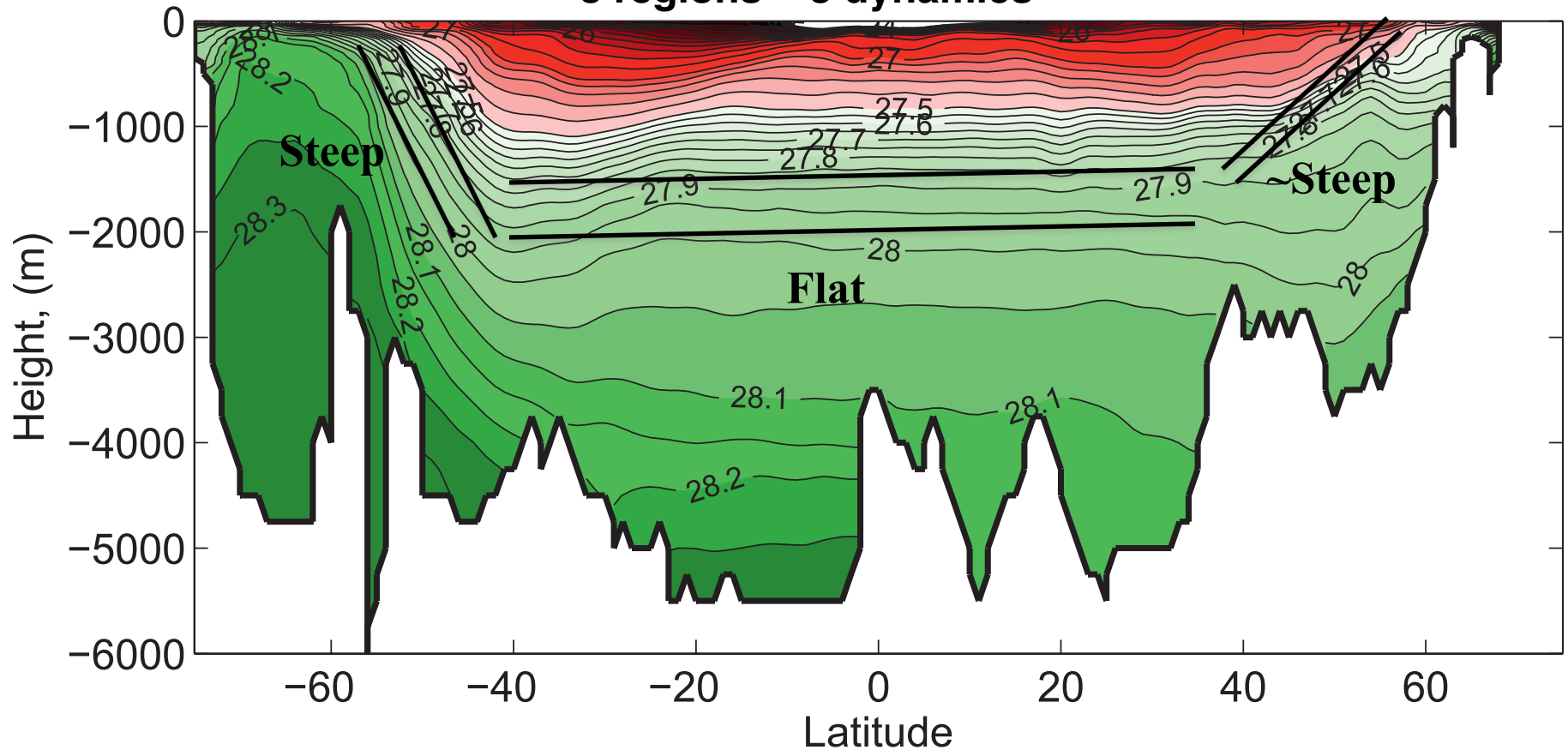


FIG. 1. Meridional section of neutral density (kg m^{-3}) in the Atlantic Ocean at 25°W from the World Ocean Circulation Experiment (WOCE). A weakly stratified water mass at middepth, roughly between the 28 and 28.1 kg m^{-3} isopycnals, corresponds to a thermostat associated with the inflow of the NADW in the Atlantic Ocean. Note that the contour intervals are 0.1 and 0.05 kg m^{-3} for isopycnals greater and lower than 27.5 kg m^{-3} , respectively.

Coupling of the traditional and alternative paradigms

Recent theoretical studies (e.g. Nikurashin and Vallis, 2012) try to describe the ocean circulation in conjunction with the two paradigms.

Dynamical arguments: 3 regions = 3 dynamics

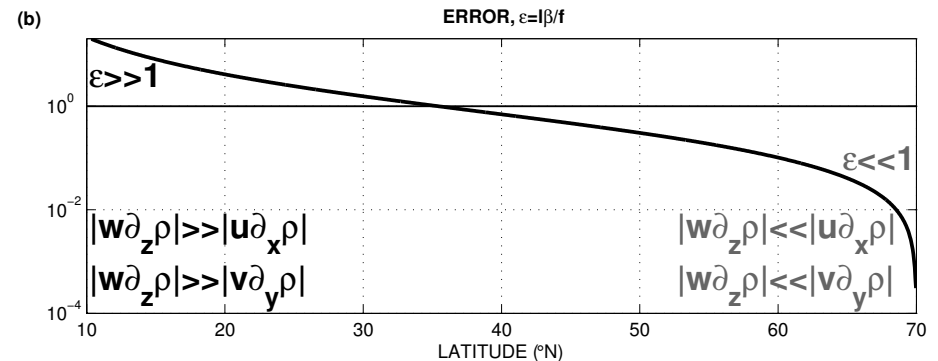
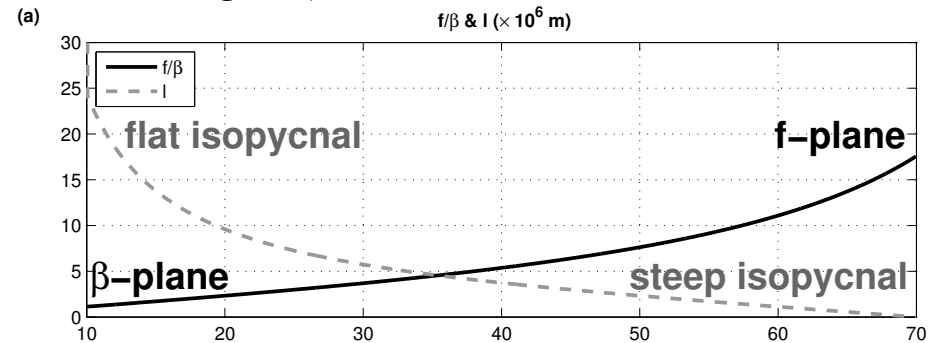
1. The first region is the Southern Ocean where zonal Pressure gradient vanishes after zonal averaging (i.e. zonally periodic region).

$$-f \frac{\partial v}{\partial z} = \frac{g}{\rho_0} \frac{\partial \rho}{\partial x}$$

$$+f \frac{\partial u}{\partial z} = \frac{g}{\rho_0} \frac{\partial \rho}{\partial y}$$

$$u \frac{\partial \rho}{\partial x} + v \frac{\partial \rho}{\partial y} + w \frac{\partial \rho}{\partial z} = 0$$

2. South of 35°N vertical dynamics
(Rossby waves propagation)
3. North of 35°N horizontal dynamics
(Baroclinic instability)



$$\varepsilon = l(\beta / f), l = -h \frac{\partial \rho}{\partial y} / \frac{\partial \rho}{\partial z}$$

Coupling of the traditional and alternative paradigms

Recent theoretical studies (e.g. Nikurashin and Vallis, 2012) try to describe the ocean circulation in conjunction with the two paradigms.

The equations:

3 regions = 3 dynamics

1. The Southern Ocean zonally periodic

$$J(\psi, \rho) \approx 0, \psi = \bar{\psi} + \psi^*$$

2. The subpolar region is conducive to Rossby waves propagation:

-> Vertical dynamics

$$w \frac{\partial \rho}{\partial z} = \frac{\partial}{\partial z} (K_z \frac{\partial \rho}{\partial z})$$

3. Northern North Atlantic is conducive to baroclinic instability:

-> Horizontal dynamics

$$v \approx u \propto \frac{\partial P}{\partial y}$$

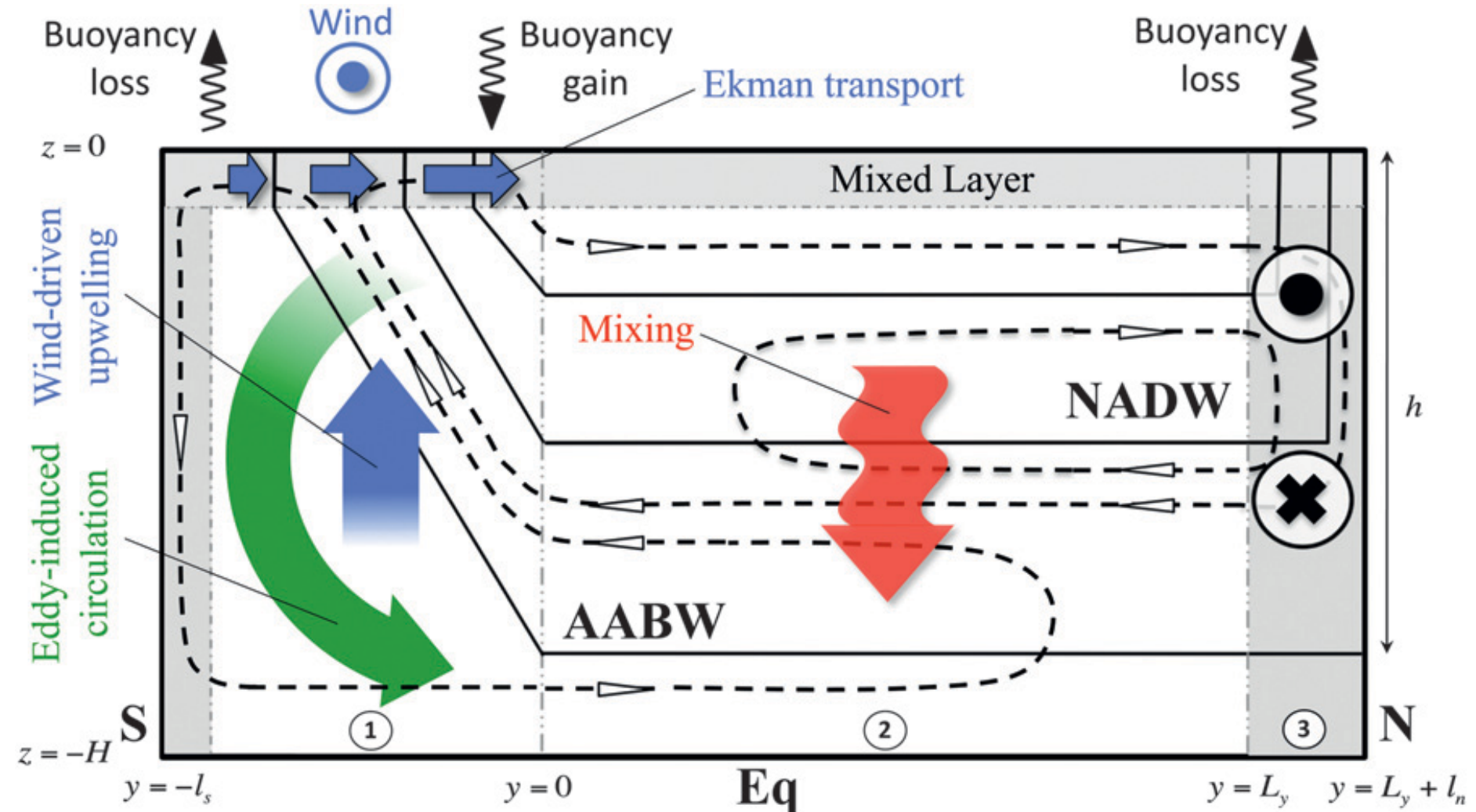
Here, each terms are zonally averaged

Coupling of the traditional and alternative paradigms

Recent theoretical studies (e.g. Nikurashin and Vallis, 2012) try to describe the ocean circulation in conjunction with the two paradigms.

The model:

3 regions = 3 dynamics



Coupling of the traditional and alternative paradigms

1. The Southern Ocean sets the stratification, 3. The northern North Atlantic produces North Atlantic Deep Water accordingly to this stratification, 2. The subtropical links the two regions, **Solutions:**

$$\begin{aligned}
 &1. \\
 &J(\psi_1, \rho) \approx 0, \psi_1 = \bar{\psi} + \psi^* \\
 &\bar{\psi} = W \frac{\tau}{f \rho_0} \frac{z+H}{H} \\
 &\psi^* = W \kappa s = -W \kappa \left(\frac{\partial \rho}{\partial y} / \frac{\partial \rho}{\partial z} \right) \\
 &\bar{\psi} = -\psi^* \Leftrightarrow s = -\frac{\tau}{\rho_0 \kappa f} \frac{z+H}{H}
 \end{aligned}$$

$$\begin{aligned}
 &2. \\
 &w \frac{\partial \rho}{\partial z} = \frac{\partial}{\partial z} \left(K \frac{\partial \rho}{\partial z} \right), \quad \frac{\partial \psi_2}{\partial y} = W w \\
 &\psi_2(y, z) = \psi_2(y_{2/3}, z) - W \int_y^{y_{2/3}} \left[\frac{\partial}{\partial z} \left(K \frac{\partial \rho}{\partial z} \right) / \frac{\partial \rho}{\partial z} \right] dy \\
 &\psi_2(y, z) = \psi_2(y_{2/3}, z) + (y - y_{2/3}) W \left[\frac{\partial}{\partial z} \left(K \frac{\partial \rho}{\partial z} \right) / \frac{\partial \rho}{\partial z} \right]
 \end{aligned}$$

$$\begin{aligned}
 &3. \\
 &v \propto u, -f \frac{\partial v}{\partial z} = \frac{g}{\rho_0} \frac{\partial \rho}{\partial x}, \quad \frac{\partial \psi_3}{\partial z} = -W v \\
 &v = -\frac{\tau}{f H \rho_0} + \frac{1}{f \rho_0} \frac{\partial (P - \langle P \rangle)}{\partial y} \\
 &\psi_3(y, z) = W \frac{\tau}{f \rho_0} \frac{z+H}{H} - \frac{W}{f \rho_0} \int_{-H}^z \frac{\partial (P - \langle P \rangle)}{\partial y} dz
 \end{aligned}$$

Coupling of the traditional and alternative paradigms

Summary:

1. The Southern Ocean sets the stratification:

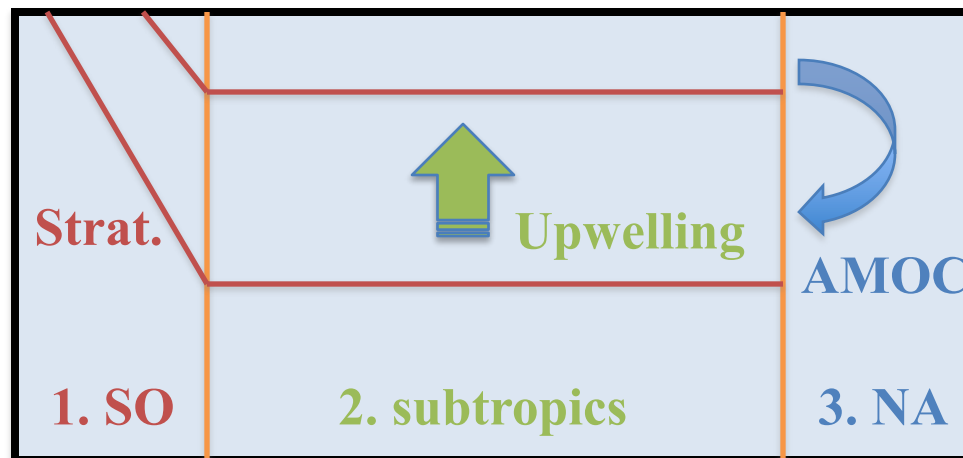
Through the competition of surface wind stress and eddies (small scale turbulence) the warm water is pushed at depth (i.e. virtually corresponding source of warm water at depth).

2. The subtropical linked the two regions:

A global upwelling “consume” water bringing dense deep water upward through vertical small scale mixing.

3. The northern North Atlantic produces North Atlantic Deep Water accordingly to this stratification:

The North Atlantic set the intensity of the AMOC in response to the stratification, i.e. in absence of stratification set in the SO (no SO winds) the AMOC vanishes.

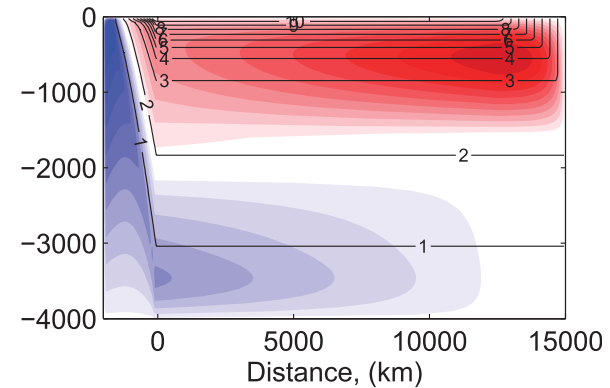
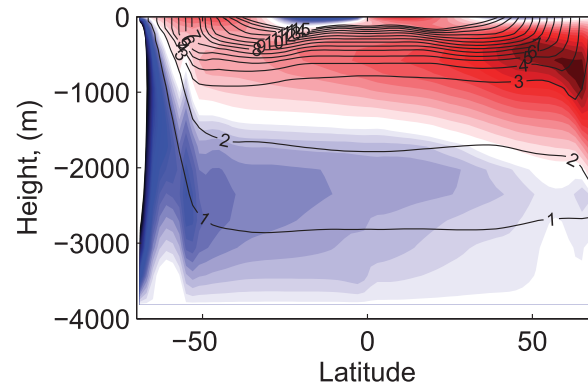


Coupling of the traditional and alternative paradigms

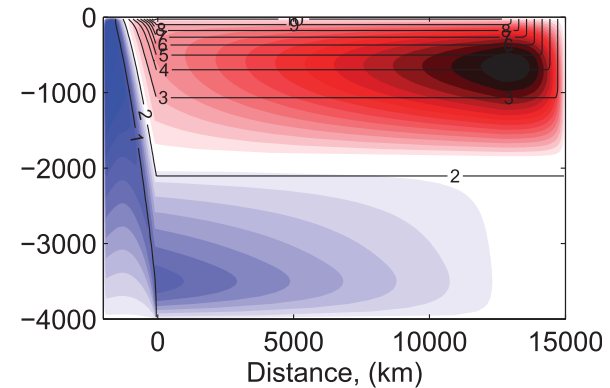
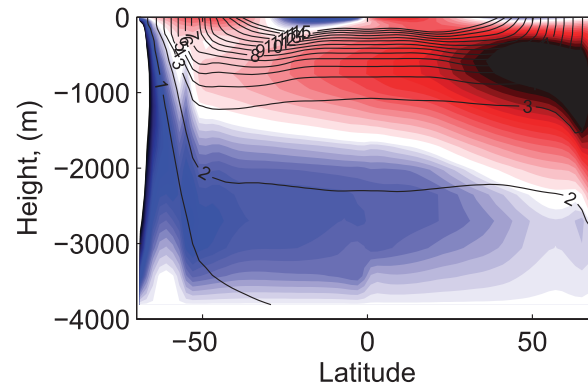
Application: GCM in idealized Configuration

Theoretical Solution

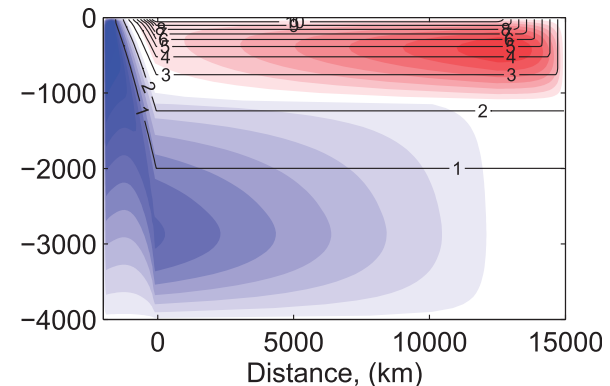
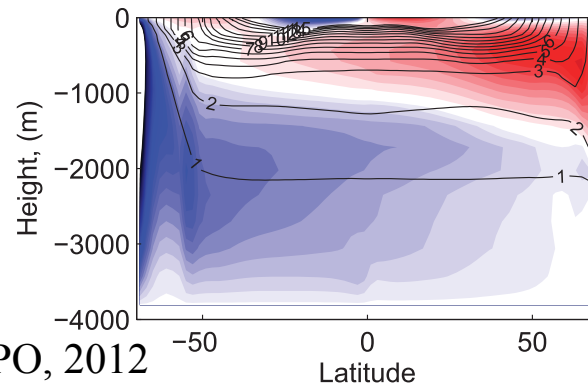
$$K_z = 2 \times 10^{-5} \text{ m s}^{-2}$$
$$\tau = 0.2 \text{ N m}^{-2}$$



$$K_z = 1 \times 10^{-5} \text{ m s}^{-2}$$
$$\tau = 5 \text{ N m}^{-2}$$



$$K_z = 2 \times 10^{-5} \text{ m s}^{-2}$$
$$\tau = 0.1 \text{ N m}^{-2}$$



Conclusions

- The ocean's stratification and MOC owe their existence to mechanical (rather than buoyancy) forcing
- Essential energetics: 2 - 3 TW are input by winds and tides and dissipated by internal wave breaking and near-boundary turbulence
- There are two physically and spatially coupled (through wind work / Southern Ocean eddies) ocean mixing regimes: an 'abyssal diapycnal' regime, and a 'pycnocline isopycnal' regime
- Significant open questions remain regarding most aspects of how turbulence drives the global ocean circulation