Ocean Mixing

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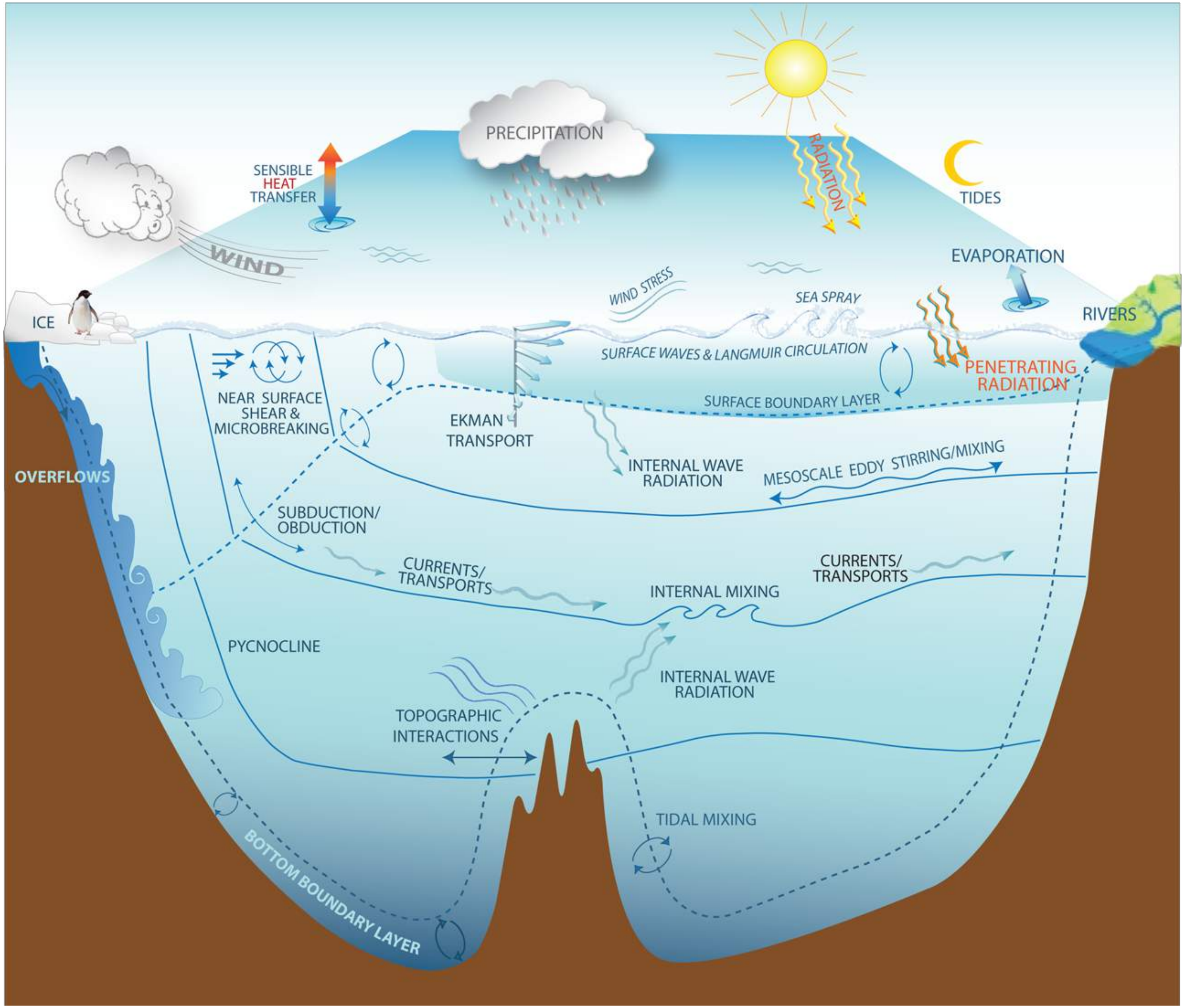
Outline

- Parameterizing subgridscale processes
- Ocean mixing processes and their parameterizations
  - Surface mixed layer: convection, wind-driven mixing, Langmuir turbulence
  - Diapycnal mixing in the ocean interior: shear-driven mixing, internal wave driven mixing – tidally driven internal waves, wind-driven internal waves, lee-waves.
  - Mesoscale eddies
  - Connecting mesoscale eddies and diapycnal mixing
- How has ECCO been used to:
  - Estimate values of parameters
  - Estimate sensitivity of circulation to parameters
- What more could be done with ECCO to improve understanding and parameterization of mixing?

Small-scale ocean processes



How does small-scale motion impact large-scale flow?

3D conserved tracer equation

$$\frac{\partial T}{\partial t} + \boldsymbol{u}.\nabla T = \kappa \nabla^2 T$$

Separate into large-scale and small-scale components: $\boldsymbol{u} = \boldsymbol{u}' + \overline{\boldsymbol{u}}$ $\quad T = T' + \overline{T}$

where $\overline{\boldsymbol{u}} = \frac{1}{V}\int_V \boldsymbol{u} dV$ and $\overline{T} = \frac{1}{V}\int_V T dV$ $\quad \frac{1}{V}\int_V \boldsymbol{u}' dV = \frac{1}{V}\int_V T' dV = 0$

$$\frac{\partial \overline{T}}{\partial t} + \overline{\boldsymbol{u}}.\nabla \overline{T} = \kappa \nabla^2 \overline{T} - \overline{\boldsymbol{u}'.\nabla T'}$$

Influence of small-scale motion (e.g. turbulence) on large-scale.

Eddy Diffusivity

Rewrite spatially averaged tracer equation, in Einstein notation:

$$\frac{\partial \bar{T}}{\partial t} + \bar{u}_j \frac{\partial \bar{T}}{\partial x_j} = \kappa \frac{\partial^2}{\partial x_j^2} \bar{T} - \overline{u_j' \frac{\partial T'}{\partial x_j}} = \frac{\partial}{\partial x_j}\left(\kappa \frac{\partial \bar{T}}{\partial x_j} - \overline{T' u_j'} \right)$$

(Using continuity equation)

$\overline{T' u_j'}$ = eddy flux of tracer

The evolution of the large-scale tracer depends on fluxes by small-scale motion.

By analogy with molecular viscosity/diffusivity, assume small-scale fluxes act down the large-scale gradients:

$$-\overline{u_j' T'} = \kappa_{Tj} \frac{\partial \bar{T}}{\partial x_j}$$

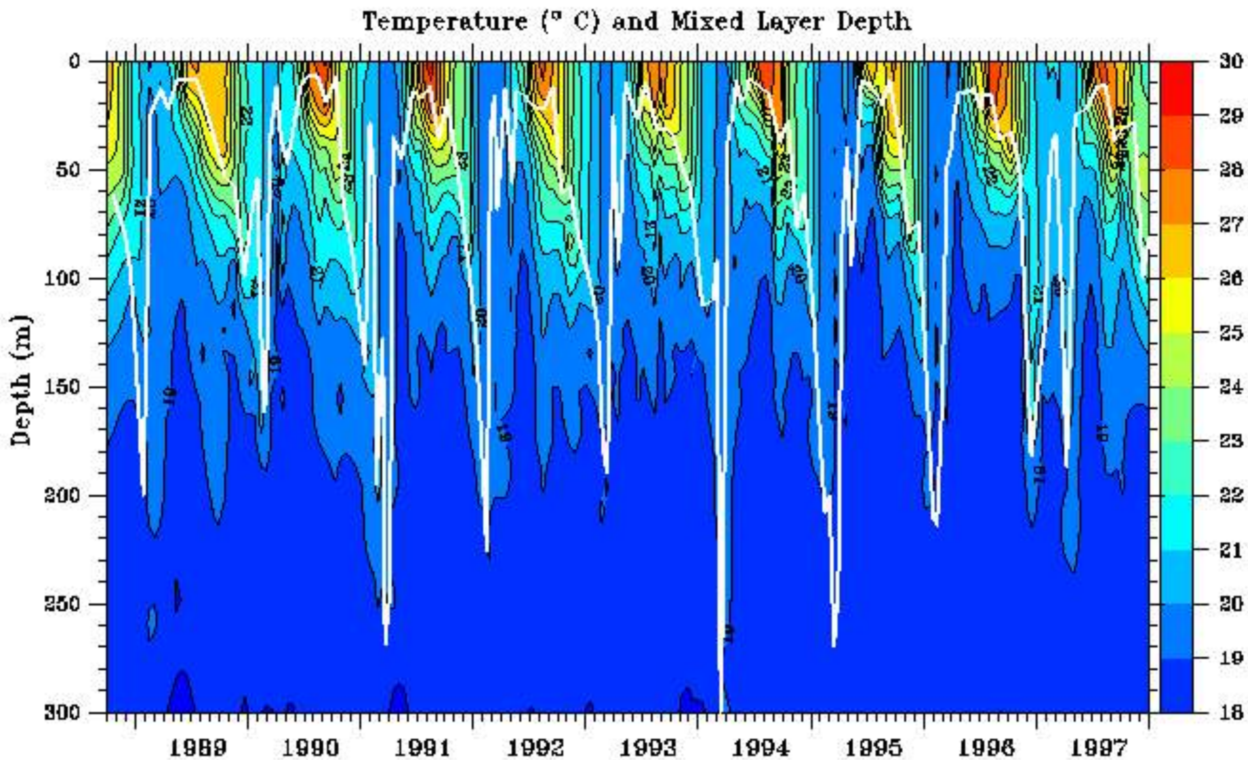
where κ_{Tj} is the eddy diffusivity in the j direction

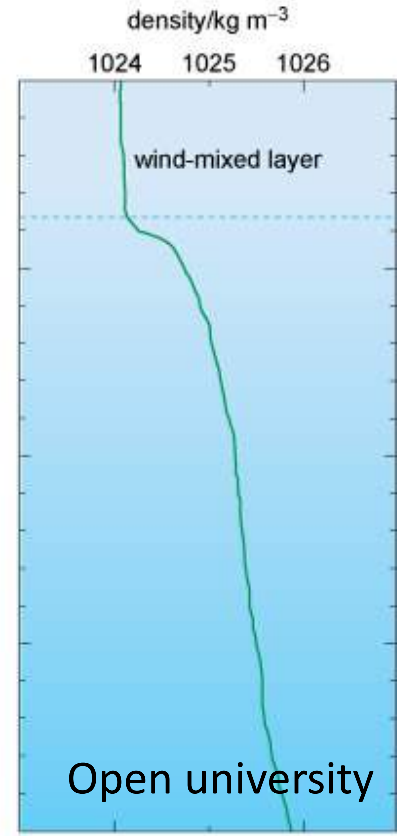
To parameterize the impact of small-scale motion on large-scale tracer distribution, we have to determine κ_{Tj} in terms of large-scale quantities.

In the stratified rotating ocean, κ_ρ, the diffusivity across isopycnal (diapycnal diffusivity) s, is very different from diffusivity along isopycnals.

Mixing at the ocean surface

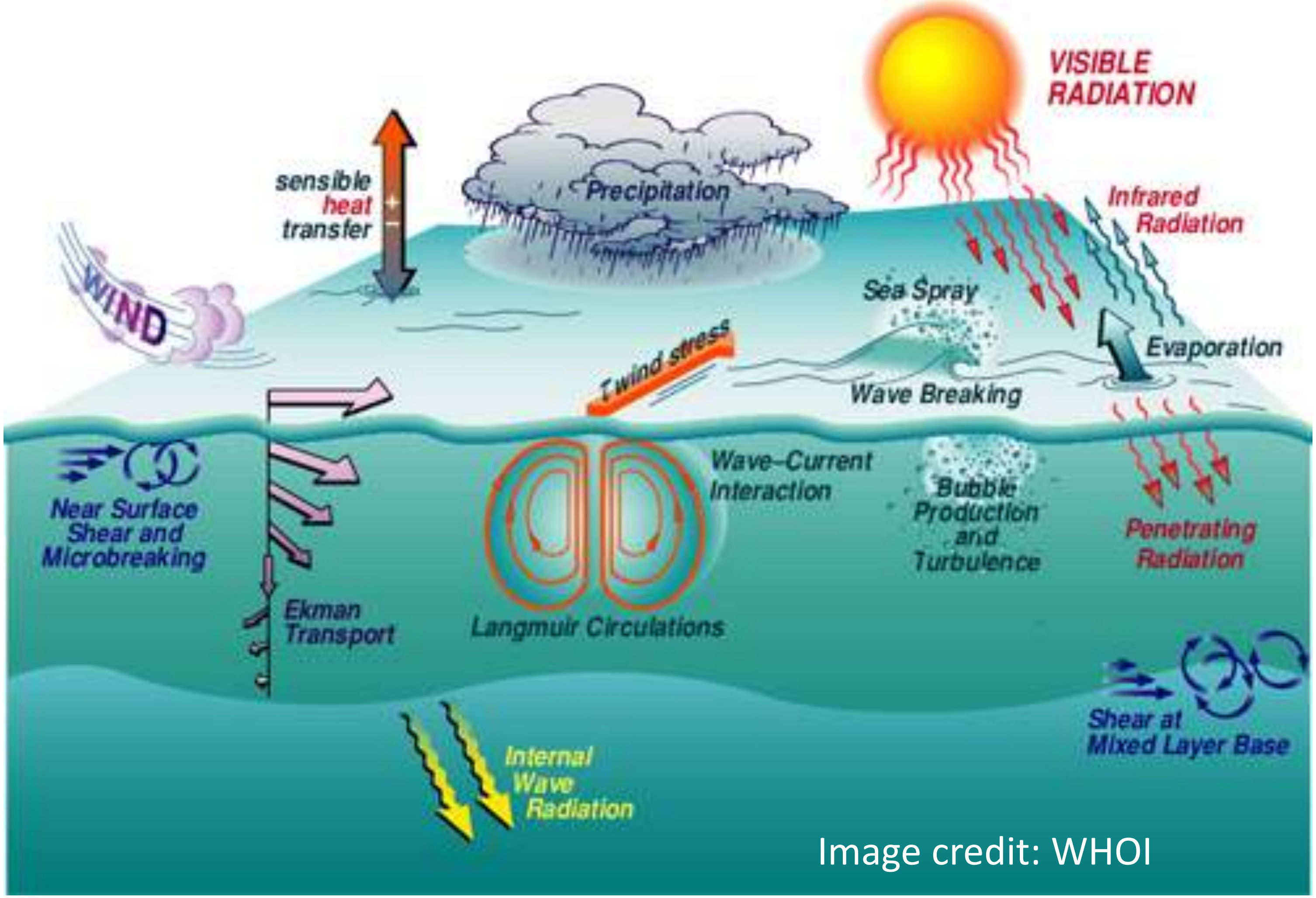
The Bermuda Atlantic Time-Series SITE (BATS)



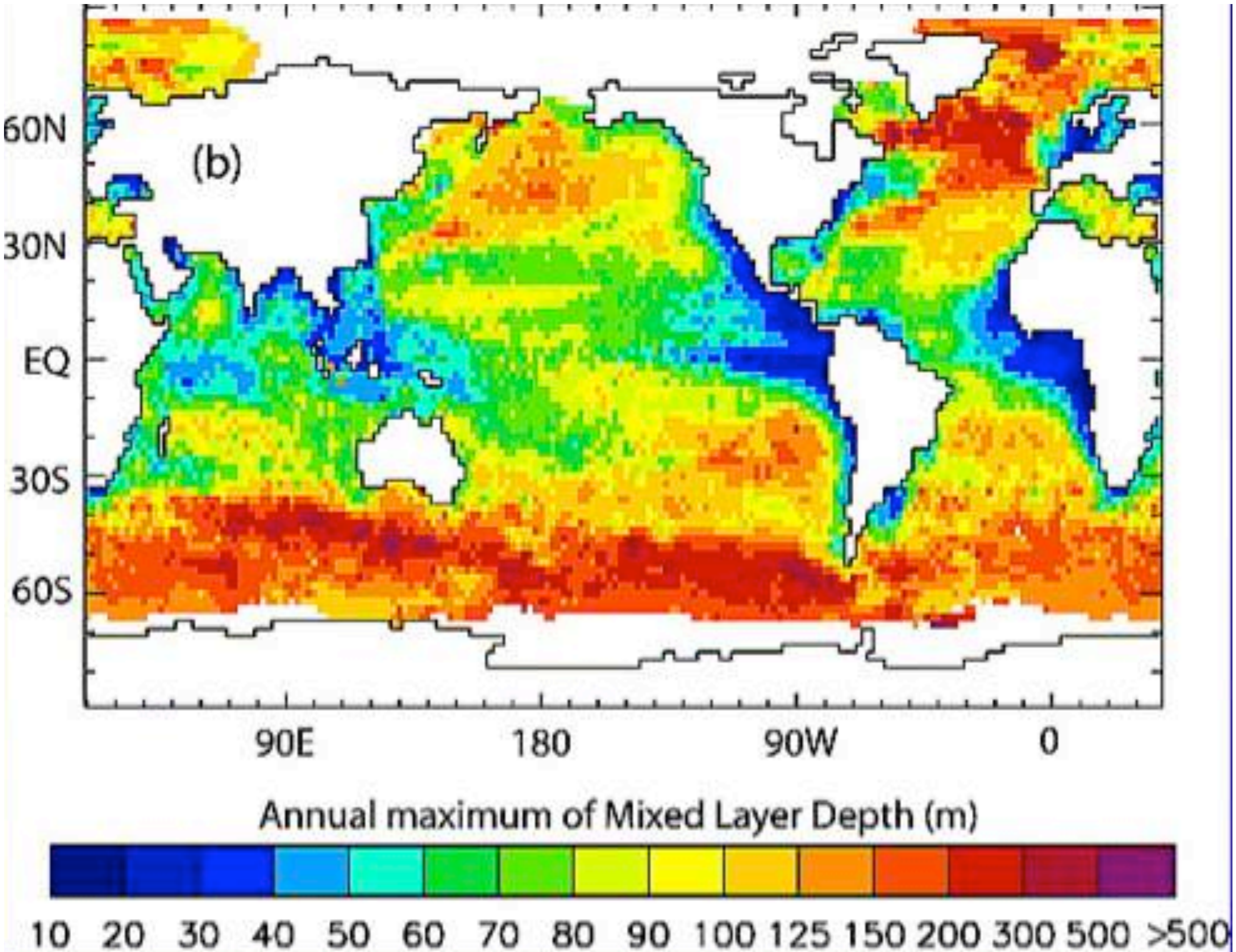


Mixing driven by surface forcing (wind/buoyancy loss) at the top of the ocean leads to a region of very weak density stratification: the surface mixed layer.

Surface mixed layer processes



Buoyancy loss and wind-driven mixing lead to deep mixed layers in subpolar latitudes



De Boyer Montegut et al. (2004)

Diapycnal mixing in the stratified ocean interior

Kelvin-Helmholtz billows in the laboratory

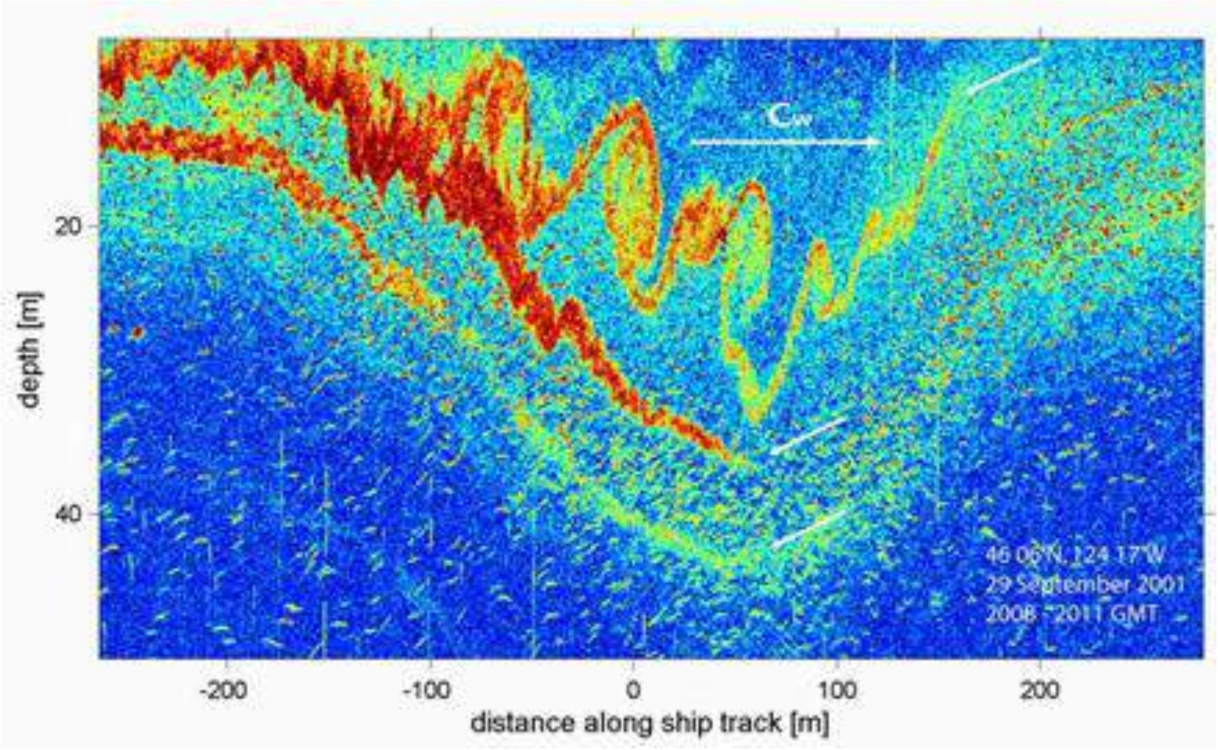
Shear instability can occur in stably stratified ocean if

$$Ri = \frac{N^2}{\left(\frac{\partial U}{\partial z}\right)^2} < \frac{1}{4}$$

(Miles-Howard criterion)

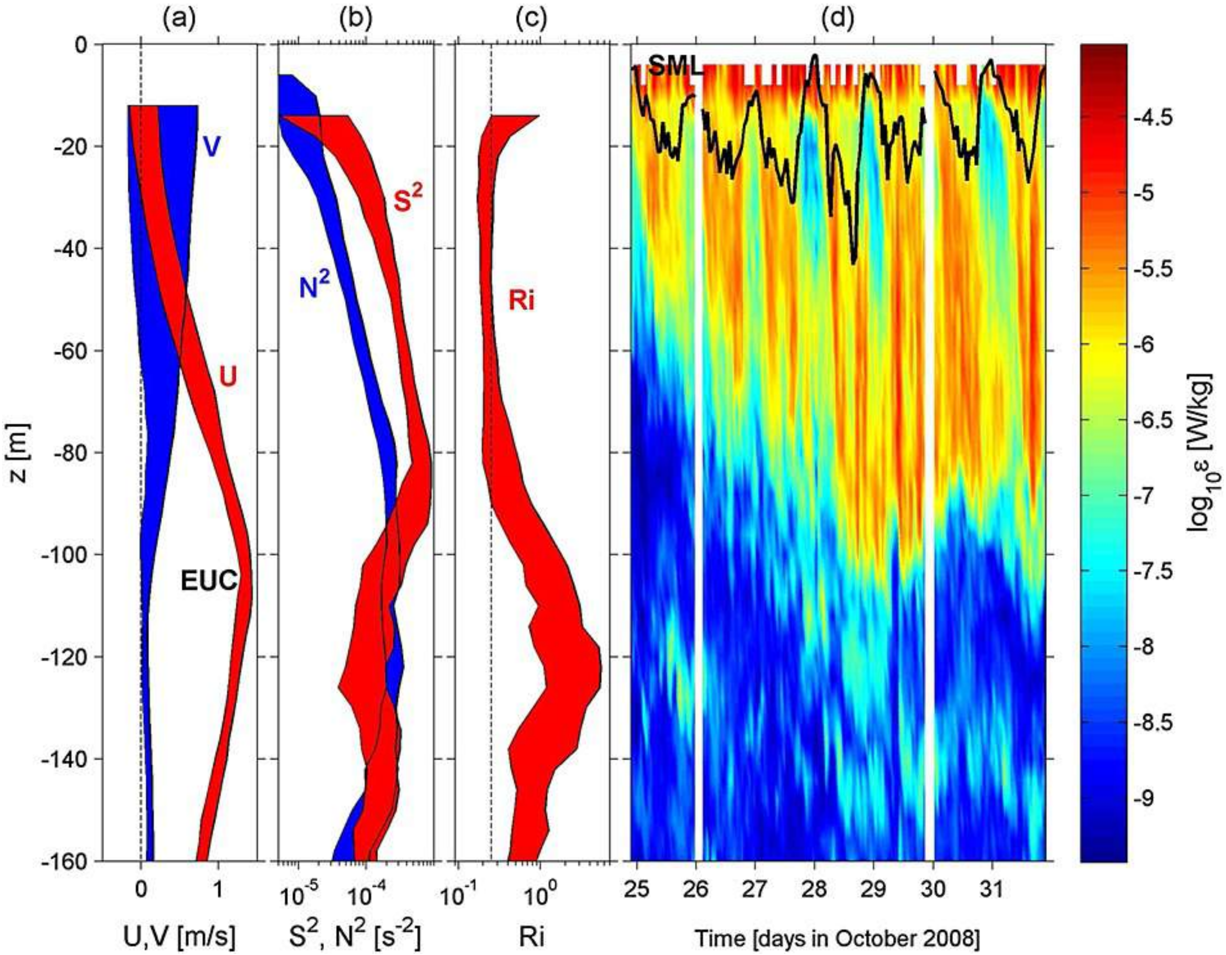
Shear-driven mixing in ocean interior

Acoustic backscatter image of internal solitary wave (Moum et al, 2003)



Whenever shears are large enough that Ri<O(1), (e.g. in density-driven currents, jets, internal waves) mixing can occur in the stably stratified ocean interior.

Shear-driven mixing driven by large-scale flow: Equatorial currents

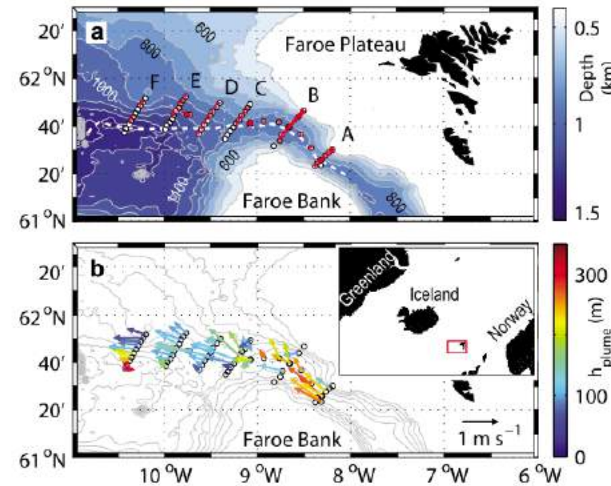


Vertically-sheared flow with $Ri \leq {}^{1}/_{4}$ is associated with turbulence below the diurnal mixed layer.

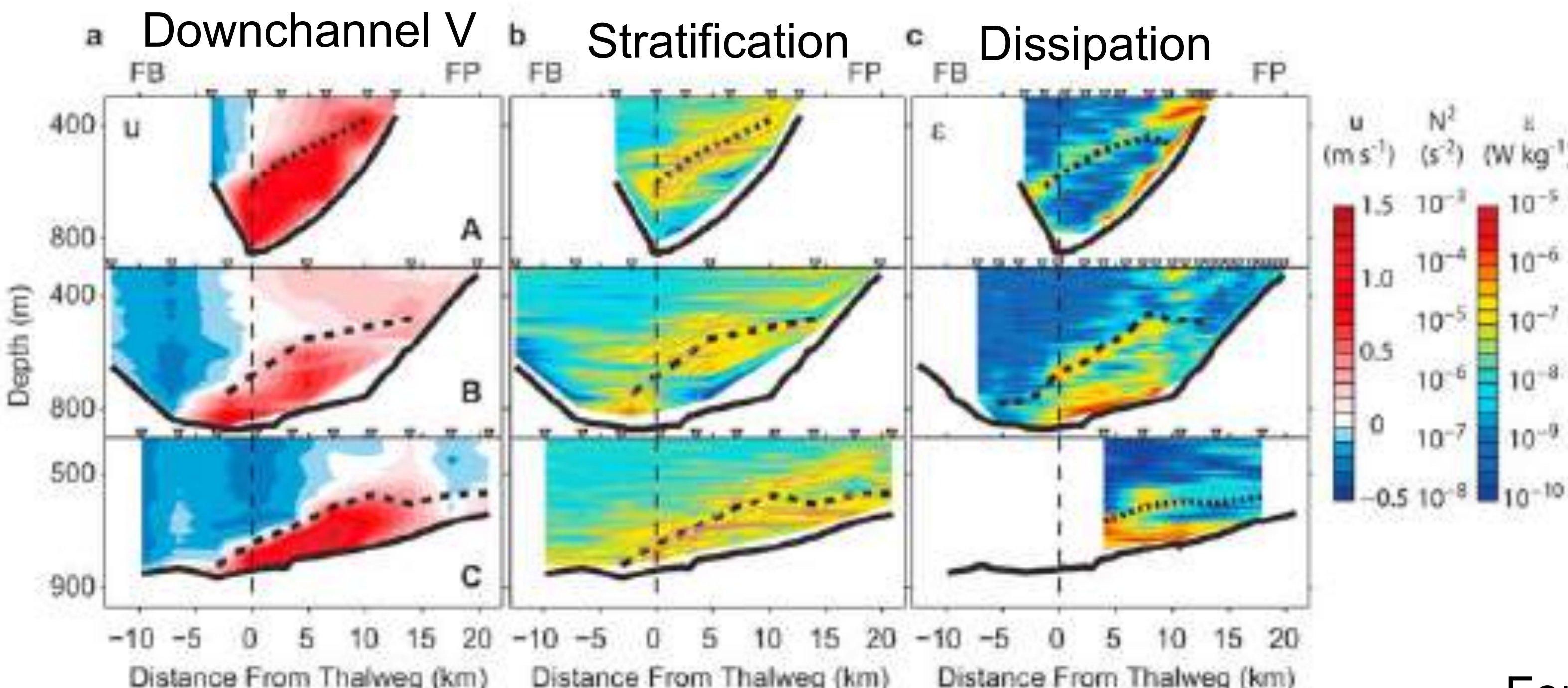
(Smyth and Moum, 2013)

Shear-driven mixing driven by large-scale flow: overflows

Faroe Bank Channel Overflow
(Fer et al, 2010)



Overflows are dense bottom currents, accelerating through topographic constrictions and down the continental slope. Large velocities confined to dense bottom layer can lead to shear instability.

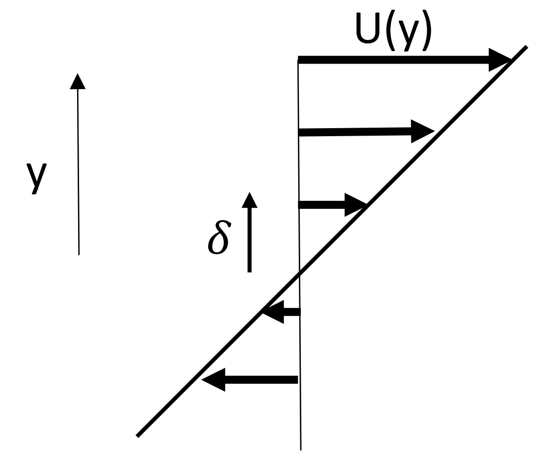


Fer et al, 2010

Parameterizing subgridscale mixing
Prandtl's mixing length model for the eddy viscosity

A parcel in a sheared parallel flow: $(\overline{U}(y), 0, 0)$

The average momentum flux due to parcel displacement δ:



$$-\overline{u'v'} = \overline{\delta v'}\frac{\partial \overline{U}}{\partial y} = \nu_T \frac{\partial \overline{U}}{\partial y}$$

$\overline{\delta v'} \sim \sqrt{(\overline{v'^2})}l$ where l is the mixing length

then the eddy viscosity is: $\nu_T = c\sqrt{(\overline{v'^2})}l$

and assuming isotropic flow: $\nu_T = c_\mu \sqrt{(\overline{q})}l$ where the turbulent kinetic energy density is $\overline{q}/2$.

$\kappa_T = \frac{\nu_T}{Pr_T}$ where Pr_T is the Turbulent Prandtl number

So if we know turbulent kinetic energy (TKE) and mixing length, we can estimate ν_T and κ_T

Turbulent kinetic energy equation

The TKE eqn:

$$\left(\frac{\partial}{\partial t} + \overline{u_j}\frac{\partial}{\partial x_j}\right)\frac{\bar{q}}{2} = \frac{\partial}{\partial x_j}\left(F_{i,j}\right) - \epsilon + P + B$$

$$F_{i,j} = \left(-\frac{1}{\rho_0}\overline{u_i' p'}\delta_{i,j} + \nu\frac{\partial}{\partial x_j}\frac{\bar{q}}{2} - \overline{u_j' u_i' u_i'}\right)$$

Transport

$$\epsilon = \nu\overline{\left(\frac{\partial u_i'}{\partial x_j}\right)^2}$$

Dissipation

$$P = -\overline{u_j' u_i'}\frac{\partial}{\partial x_j}\overline{u_i}$$

Shear production

$$\mathrm{B} = \overline{b' w'}$$

Buoyant production

This equation tells us how grid-scale averaged TKE changes with time, but contains subgridscale terms we don’t know….

TKE eqn models for eddy viscosity

TKE models for eddy viscosity use the prognostic eqn for TKE, with downgradient flux assumptions to parameterize terms on RHS, to then calculate $\nu_T = c_\mu \sqrt{\overline{q}} l$

$$\left(\frac{\partial}{\partial t} + \overline{u_j} \frac{\partial}{\partial x_j}\right) \frac{\overline{q}}{2} = \frac{\partial}{\partial x_j}\left(F_{i,j}\right) - \epsilon + P + B$$

$$B = -\kappa_T \frac{\partial \bar{b}}{\partial z} = -\frac{\nu_T}{Pr_T} \frac{\partial \bar{b}}{\partial z} = -\frac{c_\mu \sqrt{(\overline{q})} l}{Pr_T} \frac{\partial \bar{b}}{\partial z}$$

$$F_{i,j} = \frac{\nu_T}{\sigma_q} \frac{\partial}{\partial x_j} \frac{\overline{q}}{2}$$

$$\epsilon = \frac{c_\varepsilon \ q^{3/2}}{l}$$

(Kolmogorov)

$$\frac{-P}{\partial \overline{u}_i / \partial x_j} = \overline{u_j' u_i'} = -\nu_T \frac{1}{2}\left(\frac{\partial \overline{u}_j}{\partial x_i} + \frac{\partial \overline{u}_i}{\partial x_j}\right) = -c_\mu \sqrt{\overline{q}} l \frac{1}{2}\left(\frac{\partial \overline{u}_j}{\partial x_i} + \frac{\partial \overline{u}_i}{\partial x_j}\right)$$

$$\left(\frac{\partial}{\partial t} + \overline{u_j} \frac{\partial}{\partial x_j}\right) \frac{\overline{q}}{2} = \frac{\partial}{\partial x_j}\left(\frac{\nu_T}{\sigma_q} \frac{\partial}{\partial x_j} \frac{\overline{q}}{2}\right) - c_\varepsilon \frac{q^{\frac{3}{2}}}{l} + \nu_T \frac{1}{2}\left(\frac{\partial \overline{u}_j}{\partial x_i} + \frac{\partial \overline{u}_i}{\partial x_j}\right)^2 - \frac{\nu_T}{Pr_T} \frac{\partial \bar{b}}{\partial z}$$

To close this equation, and obtain ν_T, κ_T requires an assumption about lengthscale l

The Gaspar et al 1990 model for vertical mixing: one implementation of a TKE closure model.

The GGL (1990) model uses a TKE equation, assuming vertical fluxes only:

$$\left(\frac{\partial}{\partial t}\right)\frac{\bar{q}}{2} = \frac{\partial}{\partial z}\left(\frac{\nu_T}{\sigma_q}\frac{\partial}{\partial z}\frac{\bar{q}}{2}\right) - c_\varepsilon \frac{q^{\frac{3}{2}}}{l} + \nu_T\left(\frac{\partial \overline{u_h}}{\partial z}\right)^2 - \frac{\nu_T}{Pr_T}\frac{\partial \bar{b}}{\partial z} \qquad \nu_T = c_\mu \sqrt{(\bar{q})}\, l$$

Lengthscale $l = \sqrt{(l_u l_d)}$ where:

$$\frac{g}{\rho_0}\int_z^{z+l_u}[\rho(z)-\rho(z')]dz' = \bar{q} \qquad \frac{g}{\rho_0}\int_z^{z-l_d}[\rho(z)-\rho(z')]dz' = \bar{q}$$

= distances up and down a parcel moves to convert all TKE into PE.

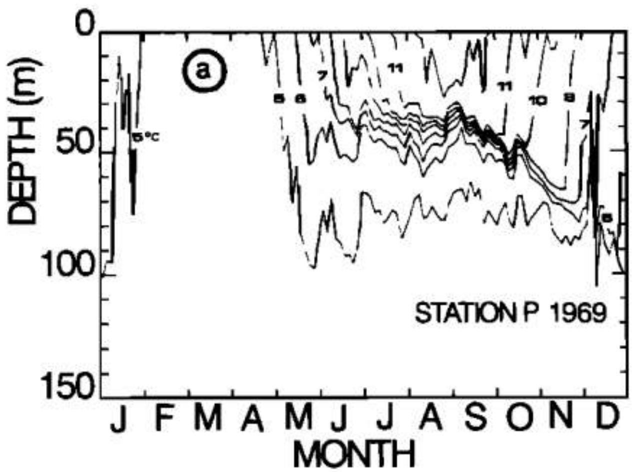
Coefficients determined by choice of mixing efficiency $\Gamma = 0.3$ where $\kappa_T = \frac{\nu_T}{Pr_T} = \Gamma\frac{\varepsilon}{N^2}$

(Osborne, 1980)

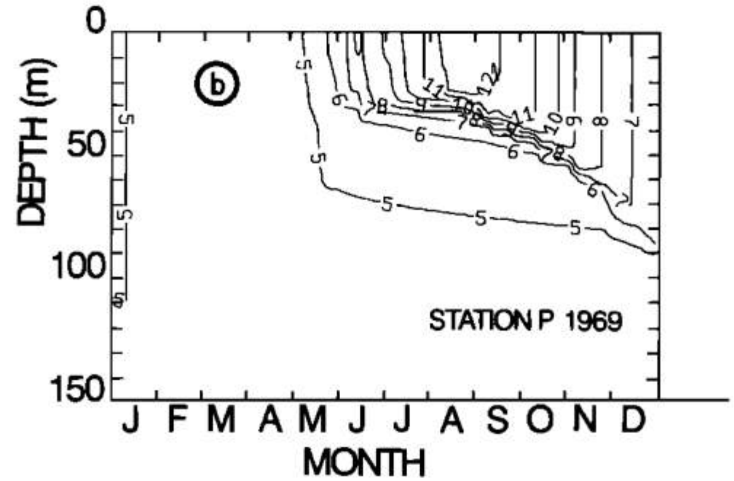
GGL1990 is the vertical mixed layer scheme in ECCOv4, parameterizing surface mixed layer (and also interior mixing driven by resolved shear)

GGL scheme reproduces observed mixed layer evolution

Isotherm depths at Ocean Station P



Observed

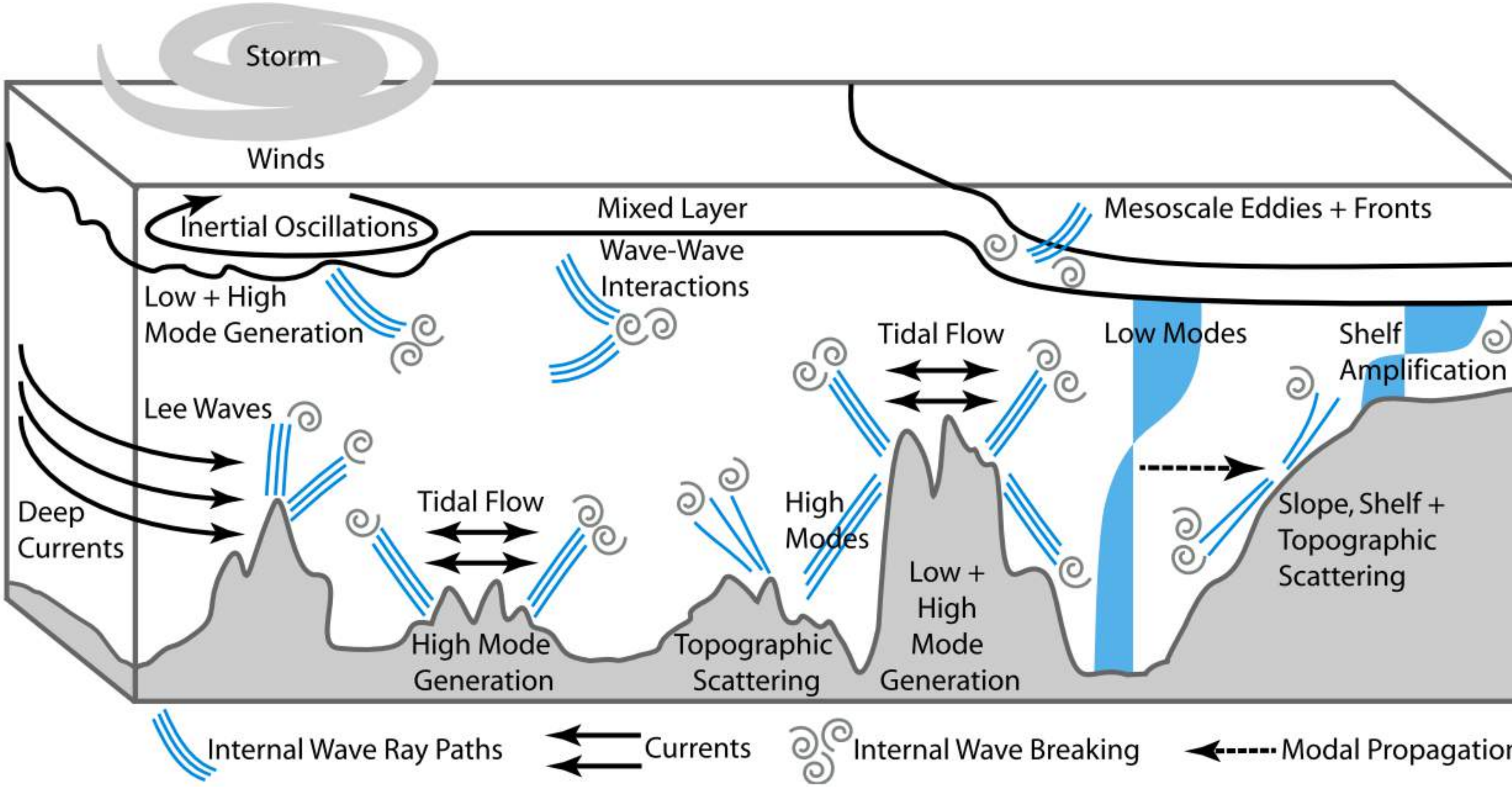


Modeled

GGL 1990 determines κ_T from resolved vertical shear and buoyancy gradient, and effectively parameterizes many mixed layer processes and mixing due to large-scale shear.

However, not all mixing is driven by the resolved shear/buoyancy gradient – e.g. internal waves provide shear on scales below GCM grid scale, leading to mixing.

Internal wave driven mixing: a nonlocal global problem



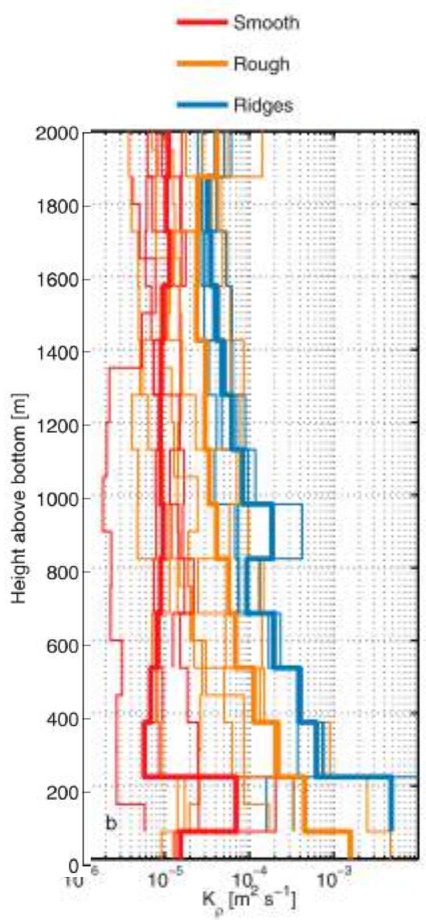
Winds, tides and subinertial flow generate internal waves, below global model grid-scale, which propagate, and eventually break. Some of the wave energy leads to diapycnal mixing, both near and far from the wave generation site.

MacKinnon et al, 2017, BAMS
Internal wave driven mixing Climate Process Team

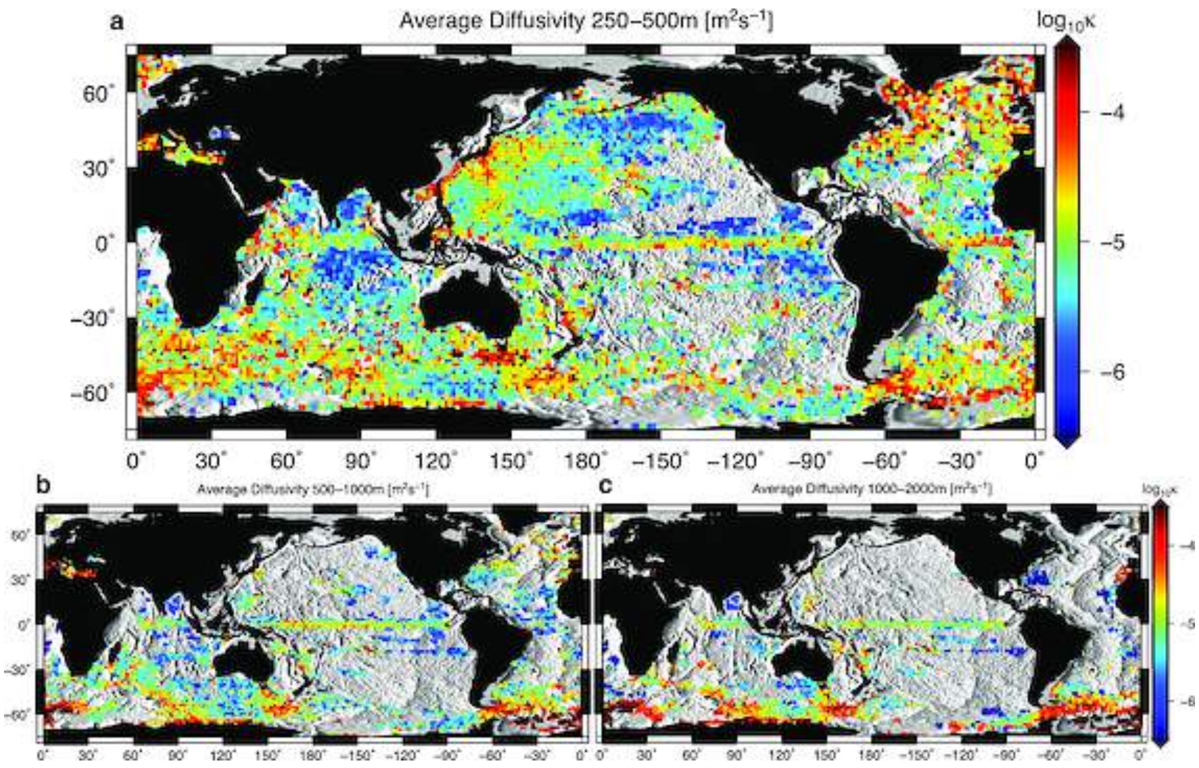
In ECCO this mixing due to unresolved waves is represented by spatially variable background diffusivity κ_d

Mixing is not uniform

Vertical distribution of observed mixing (Waterhouse et al, 2014)



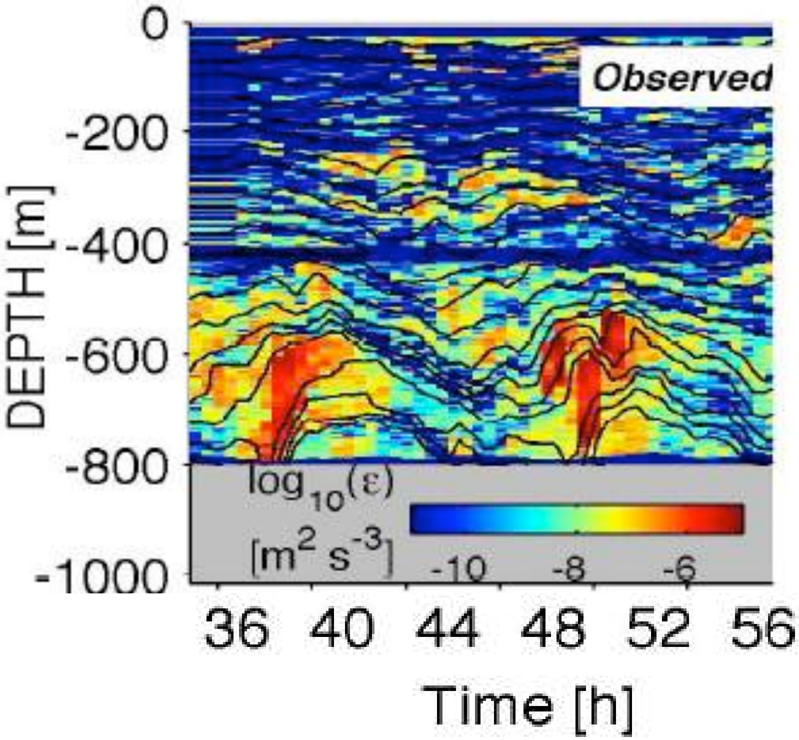
Horizontal distribution of observed mixing (Whalen et al, 2012) deduced from ARGO profiles



In much of the ocean, diffusivity and turbulent density flux increase with depth (e.g. due to breaking internal tides near the generation site).

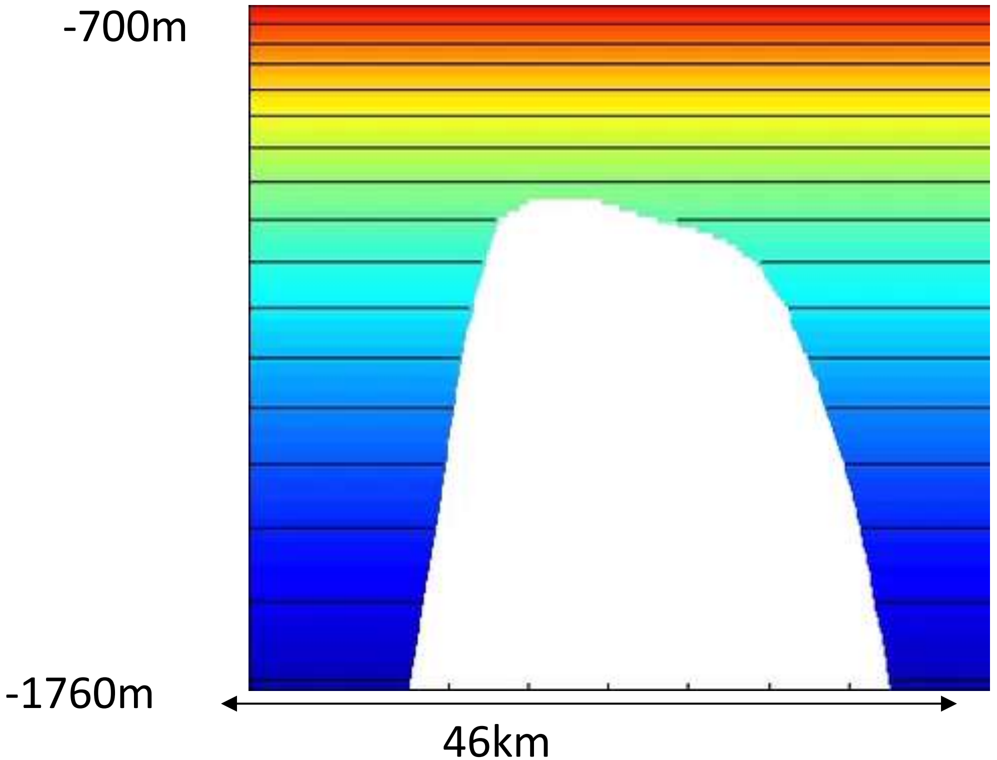
Mixing by tidally-driven internal waves at tall steep ridges

Observed dissipation at Hawaiian ridge



Klymak et al, 2008

Buoyancy field forced by M2 barotropic tide

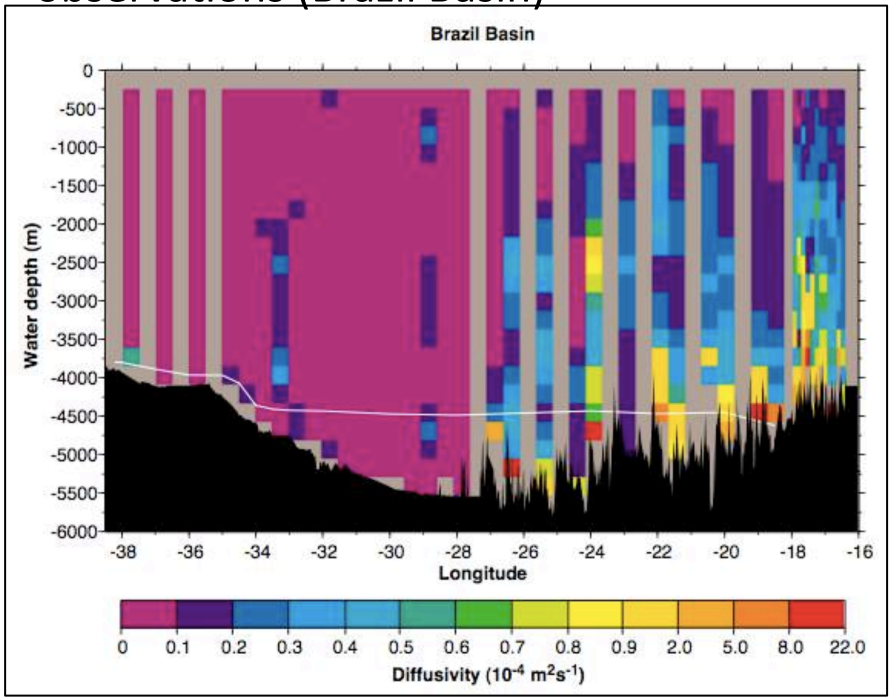


Legg and Klymak, 2008

At tall steep topography wave breaking occurs in transient internal hydraulic jumps

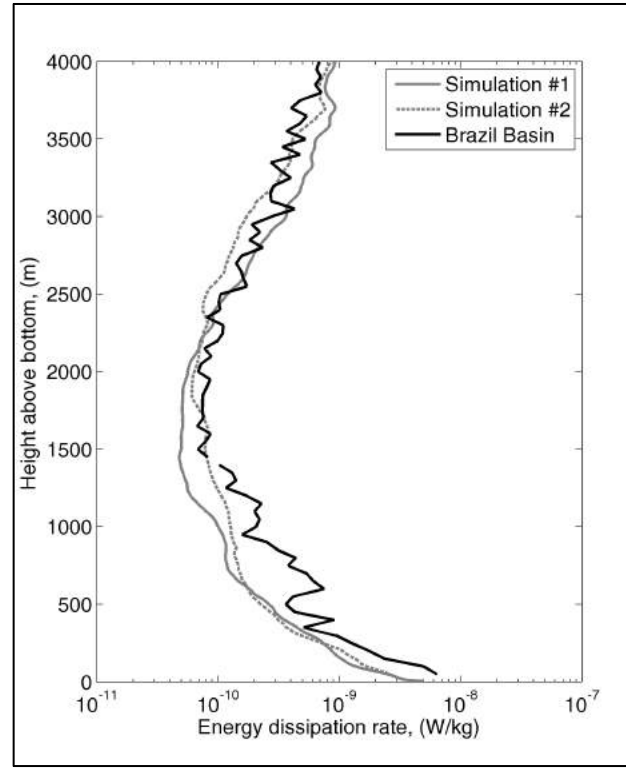
Mixing by tidally-driven internal waves at small amplitude rough topography

Diffusivity inferred from microstructure observations (Brazil Basin)



$$\kappa_\rho = \frac{\Gamma \varepsilon}{N^2}$$

(Polzin et al, 1997)



Comparison between observed and simulated dissipation for tidal flow over rough topography *(Nikurashin and Legg, 2011).*

Relative steepness is not large, topography is not tall: dissipation is not caused by transient internal hydraulic jumps

How do we currently parameterize mixing by tidally-generated internal waves?

In CESM and GFDL climate models, tidally-driven mixing is represented by a spatially and temporally variable eddy diffusivity κ.

Local dissipation

St Laurent et al, 2002

$$\varepsilon = \frac{1}{\rho} E(x,y) \cdot q \cdot F(z)$$

Rate of energy conversion from barotropic tide to baroclinic per unit area.

$$E(x,y) = \frac{1}{2}\rho_0 N_b k h^2 \langle u^2 \rangle$$

Fraction of local dissipation.
Set (arbitrarily) to 1/3 in current implementations

Vertical structure function
$\int_{-H}^{0} F(z)dz = 1$

Exponential decay with (arbitrary) constant vertical scale in most current implementations (alternatives exist: Melet et al, 2013)

$$\kappa = \varepsilon \frac{\Gamma}{N^2}$$

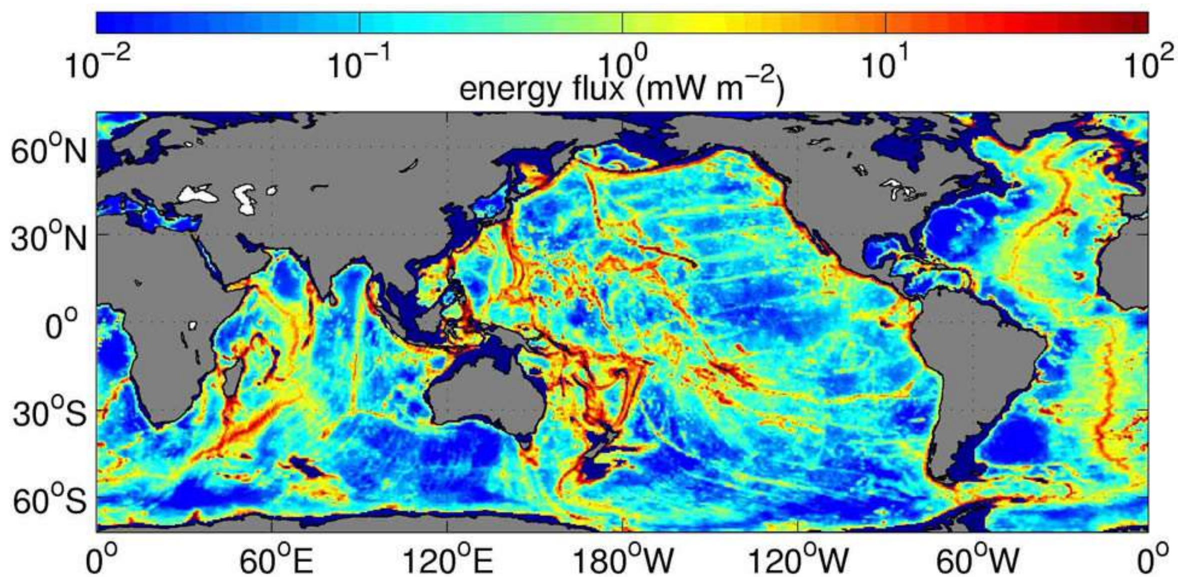
Mixing efficiency (~0.2)

Observational constraints on tidally-driven mixing: Barotropic to baroclinic energy conversion

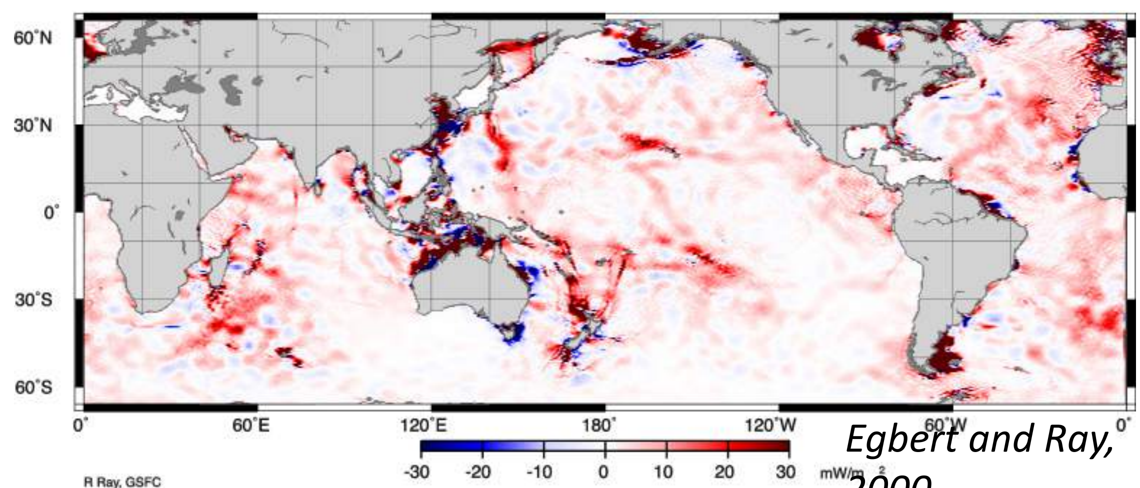
Energy conversion from barotropic to baroclinic tide, from parameterization.

$$E(x,y) = \frac{1}{2}\rho_0 N_b k h^2 \langle u^2 \rangle$$

Jayne and St Laurent, 2001



Energy loss from M2 tide, deduced from Topex-Poseidon SST: frictional dissipation in shallow seas, conversion to baroclinic tide in deep ocean.

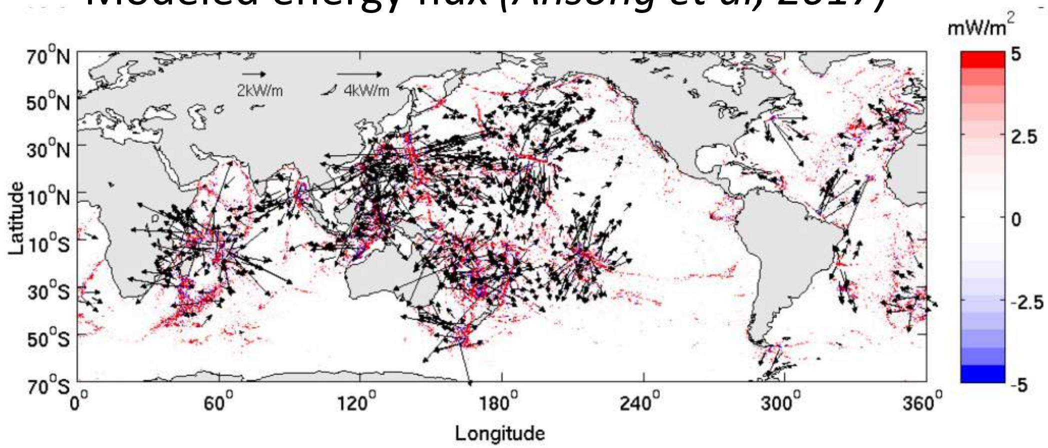


Egbert and Ray, 2000

Mixing by propagating tidally-driven internal waves: far-field tidal dissipation

A significant fraction of internal tide energy propagates away from generation site

Modeled energy flux *(Ansong et al, 2017)*



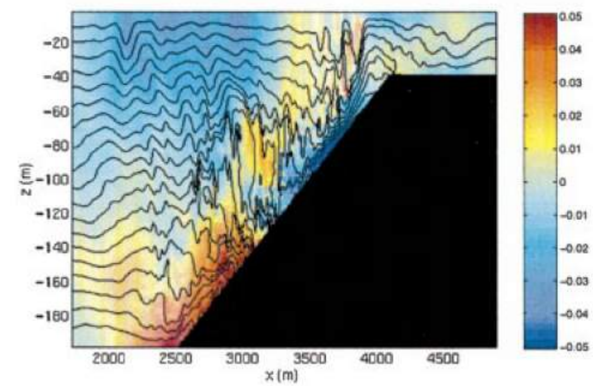
Current climate model parameterization: Constant or latitude dependent κ_b
Violates energy conservation!

For energy conservation by local and remote dissipation: globally require

$$\int \varepsilon dV = \frac{1}{\rho}\int E(x,y)dxdy$$

Breaking of propagating waves → farfield dissipation.

Processes leading to breaking include: wave-wave interactions, scattering from topography, interaction with subinertial flow.

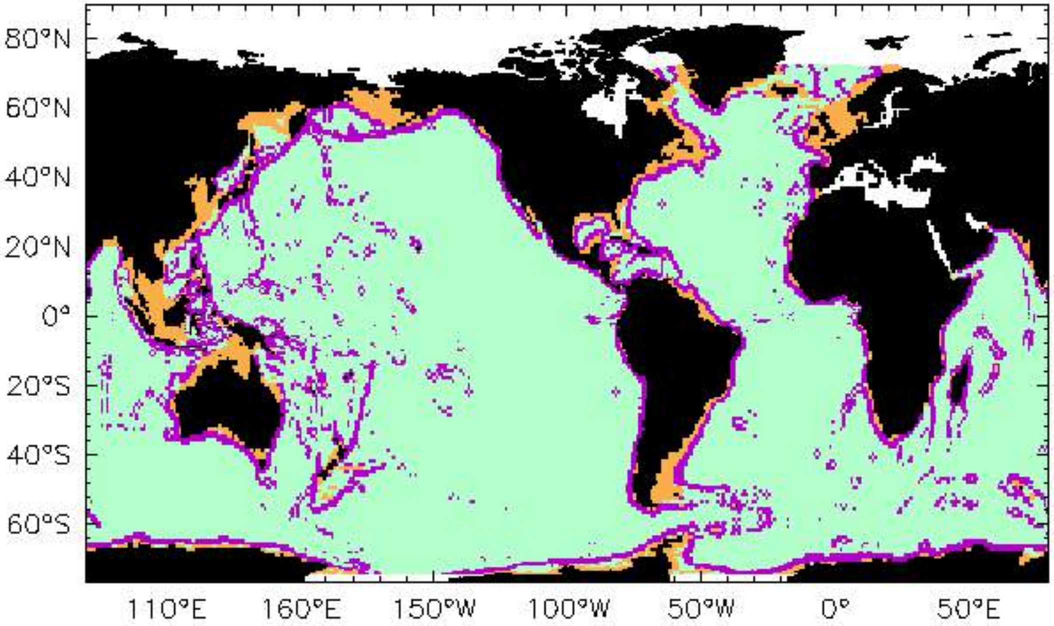


Internal wave breaking at continental slope
(Legg and Adcroft, 2003)

Does the spatial distribution of farfield internal tide-driven mixing matter for the ocean circulation?

Melet et al, 2016

Climate model thought experiment: examine impact of different idealized horizontal distributions of remote dissipation, dividing ocean into 3 zones



Slopes
Slope > 0.01.
Wave reflection/scattering

Basins
Depth >500m
Wave-wave interactions

Continental shelves
Wave shoaling

Remote dissipation parameterization:
Assign constant value of q_r for each zone

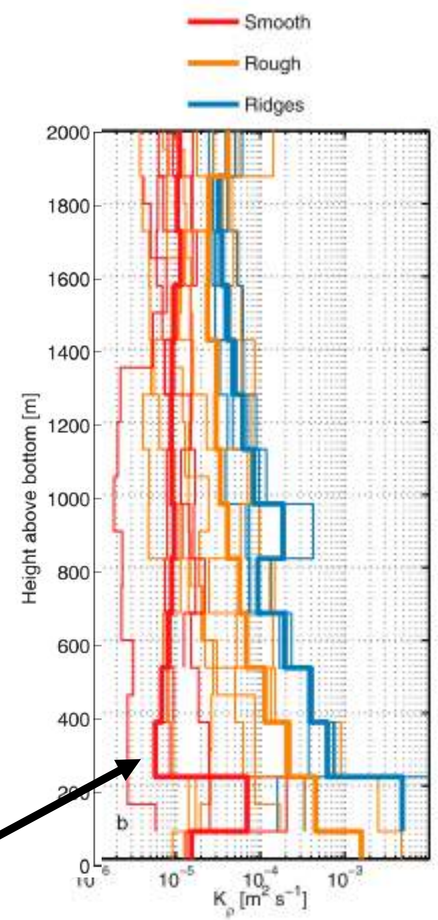
$$\varepsilon_r = \frac{1}{\rho} q_r \frac{E_{global}(t)}{Area_r} F(z)$$

- GFDL ESM2G 1000-year simulations with 1860 forcing include St Laurent et al (2002) representation of local tidal dissipation, with 20% dissipated locally.
- Remaining 80% dissipated in 1 of 3 zones.
- Reference experiment: 20% local, 80% dissipated via uniform $\kappa = 1.4\times10^{-5} m^2 s^{-1}$

Vertical structure of remote internal tide dissipation

Each experiment uses one of 3 vertical profiles for remote dissipation:

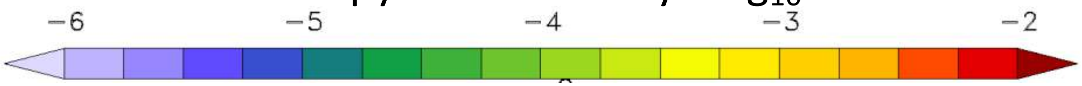
- $\varepsilon \sim e^{-\frac{z}{z_s}}$: Bottom intensified mixing as in St Laurent et al (2002), appropriate for topographic scattering
- $\varepsilon \sim E \sim N : \kappa \sim {}^{1}/_{N}$: Appropriate for direct breaking of large-scale waves
- $\varepsilon \sim E^2 \sim N^2$: κ constant with depth: agrees with nonlinear wave interaction theory and observations far from rough topography.

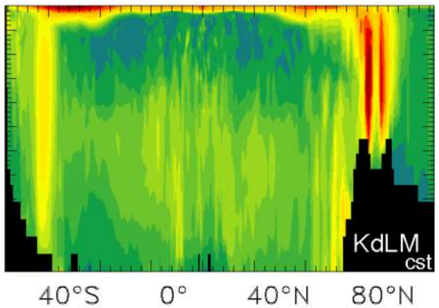


Synthesis of observed diffusivity: Waterhouse et al (2014)

How does net diffusivity depend on vertical and horizontal distribution of remote dissipation?

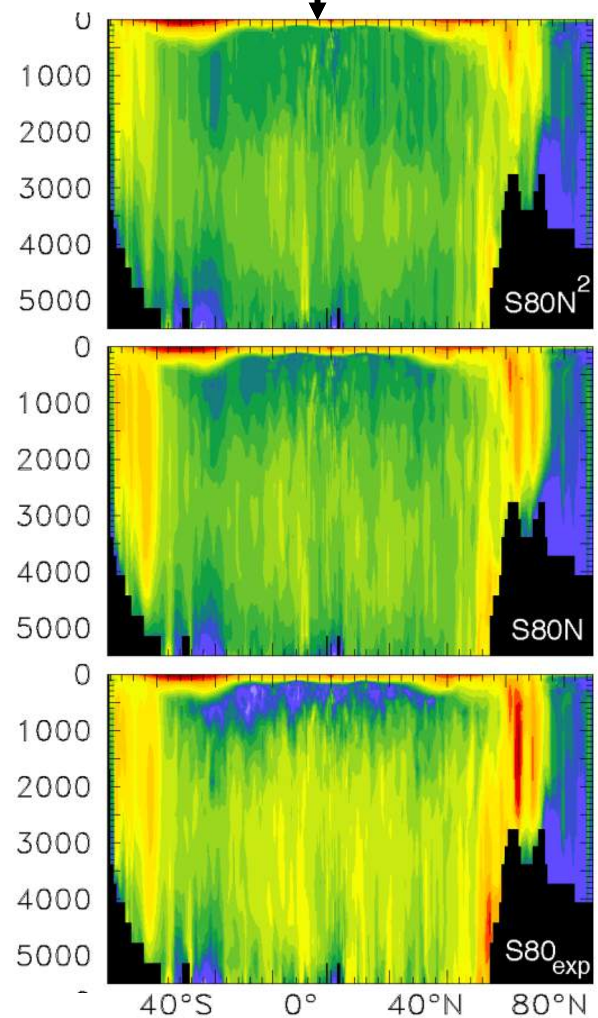
Zonal mean diapycnal diffusivity: $Log_{10} m^2 s^{-1}$





Reference: 20% local, 80% uniform diffusivity

20% local, 80% slopes, 3 different vertical structures for remote dissipation

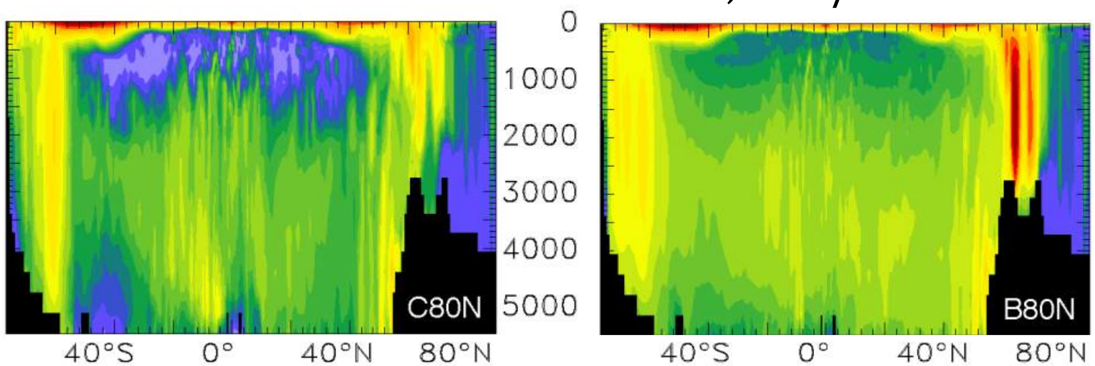


$\varepsilon \sim N^2$

$\varepsilon \sim N$

$\varepsilon \sim e^{-\frac{z}{z_s}}$

← 3 simulations with $\varepsilon \sim N, \kappa \sim 1/N.$ →



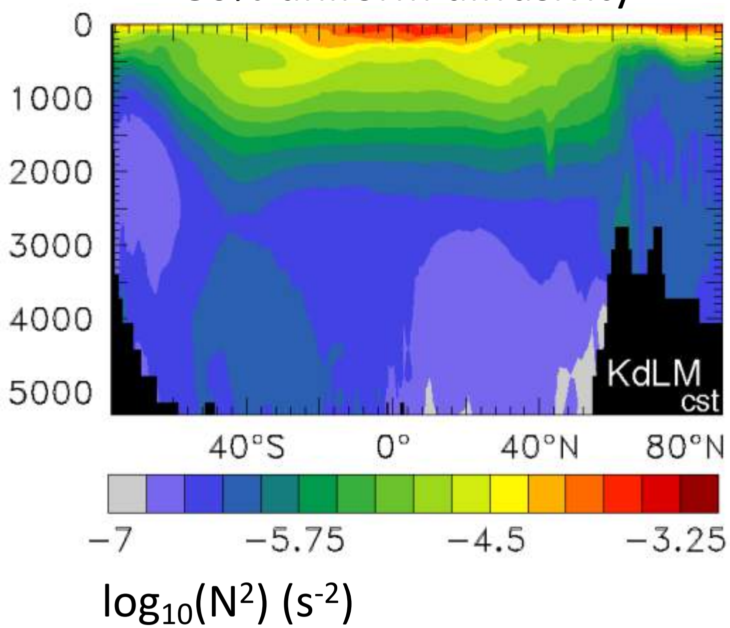
20% local, 80% coastal shelves

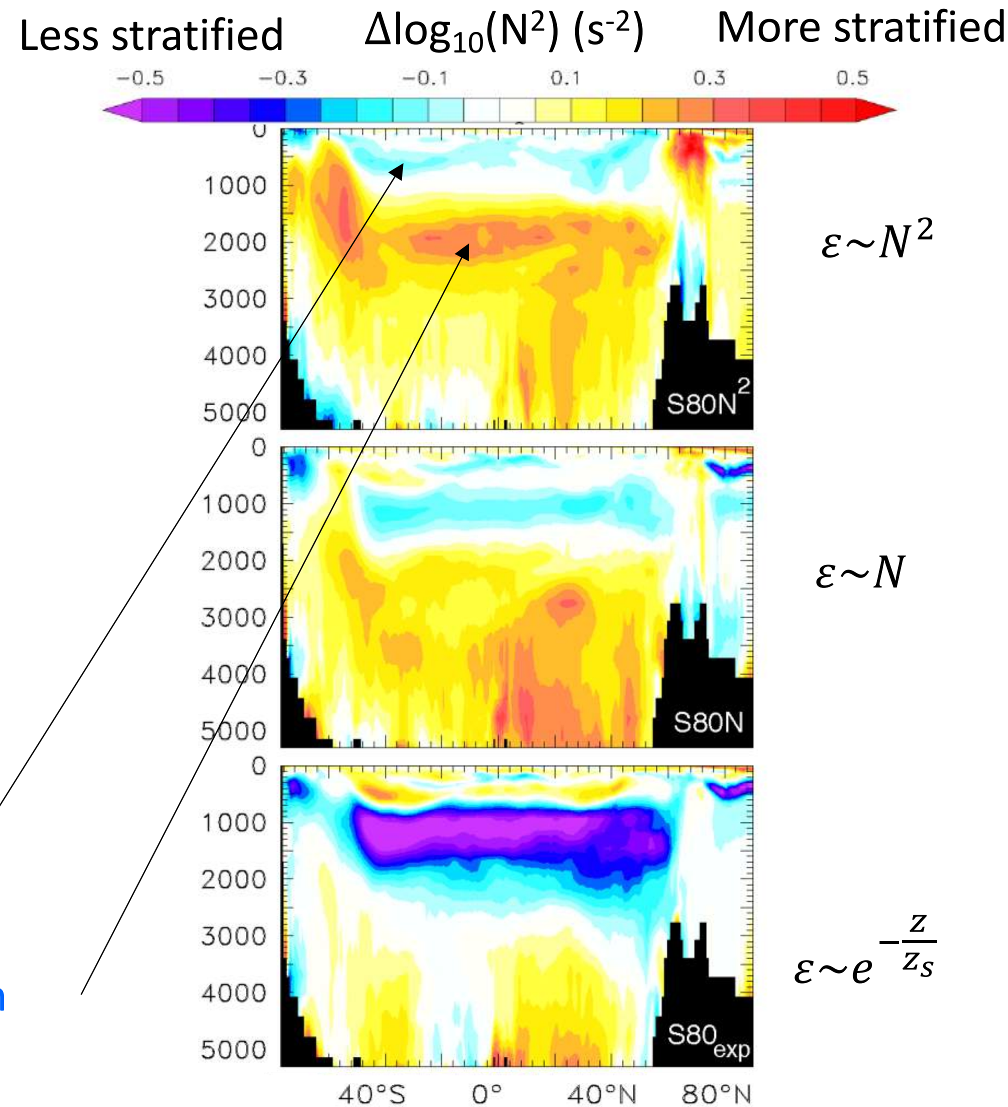
20% local, 80% basins

Vertical profile determines amount of mixing in thermocline relative to deep ocean.
Mixing on coastal shelves concentrates mixing at surface.

Influence of vertical distribution of remote dissipation on stratification

Reference: 20% local, 80% uniform diffusivity

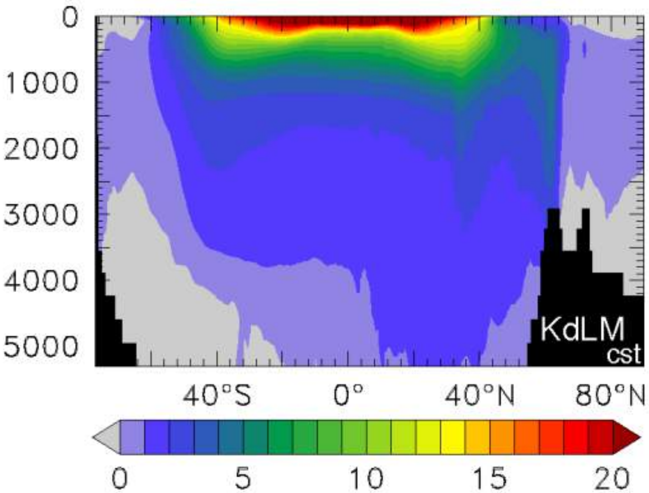




20% local, 80% slopes

Stronger near-surface diffusivity:→ reduced stratification above 1000m, increased stratification from 1000-2000m depth.

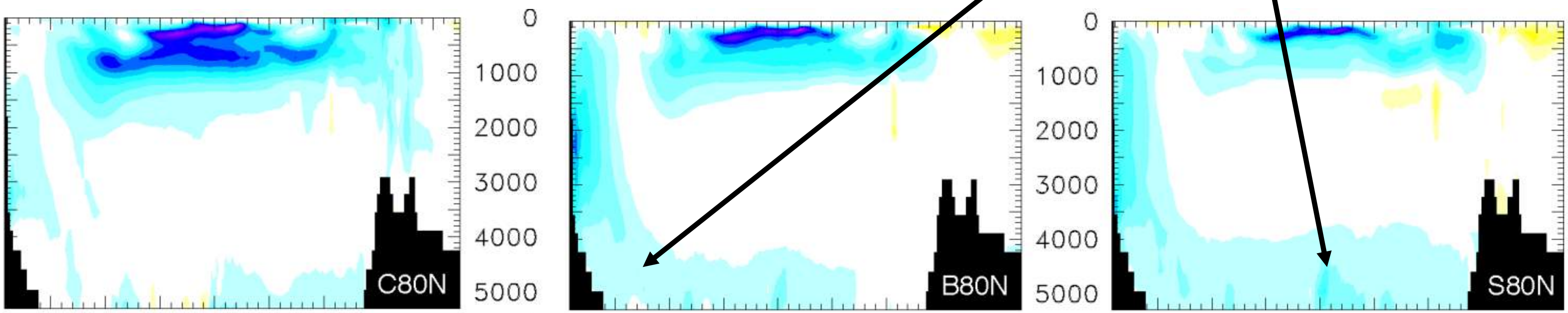
Horizontal location of mixing does matter if it influences bottom water properties



Zonal mean temperature (C)

Reference: 20% local, 80% uniform diffusivity

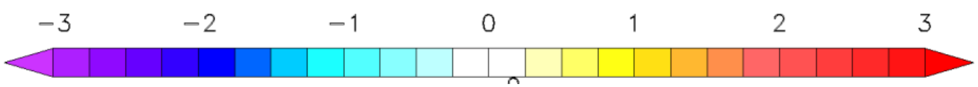
Less mixing on Antarctic shelves:→ colder AABW



20% local, 80% coasts

20% local, 80% basins

20% local, 80% slopes



colder

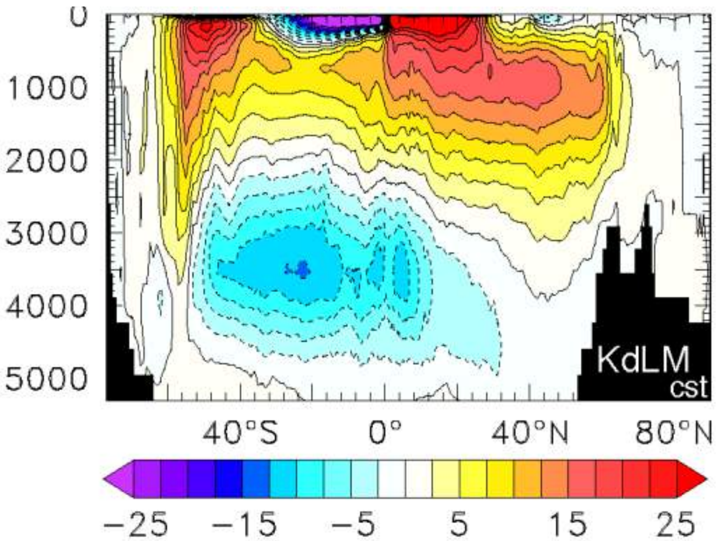
ΔT (C)

warmer

All have $\varepsilon \sim N$

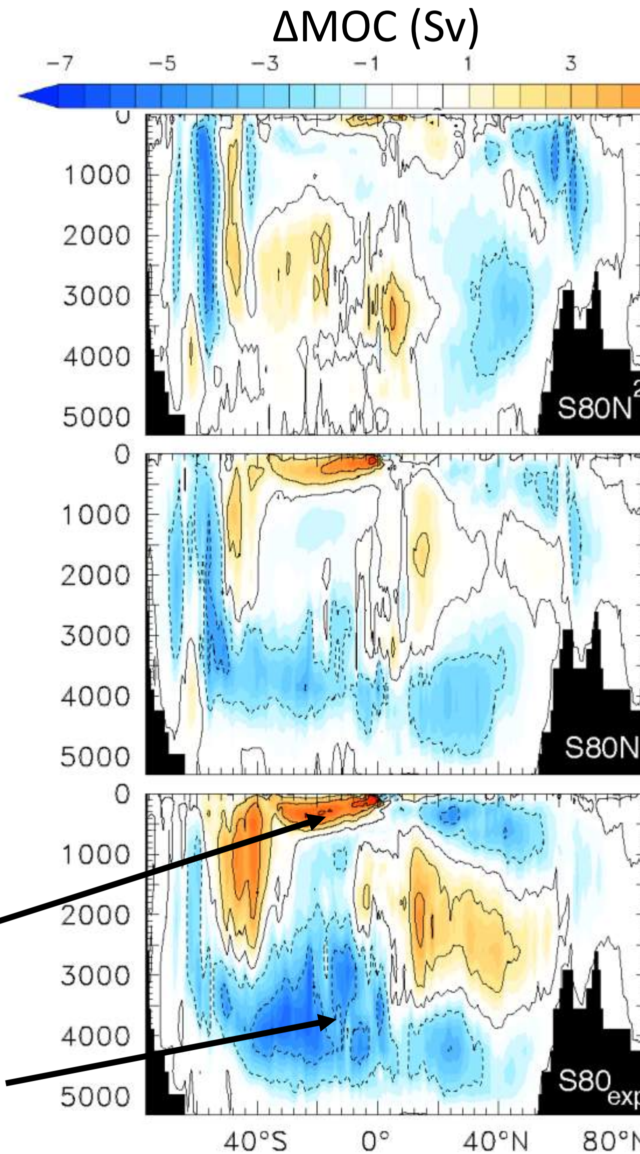
Vertical structure of mixing influences overturning strength

Reference: 20% local,
80% uniform diffusivity



Global meridional overturning circulation (Sv)

- Less near-surface mixing:→ weaker subtropical overturning
- More mixing at depth:→ stronger deep overturning



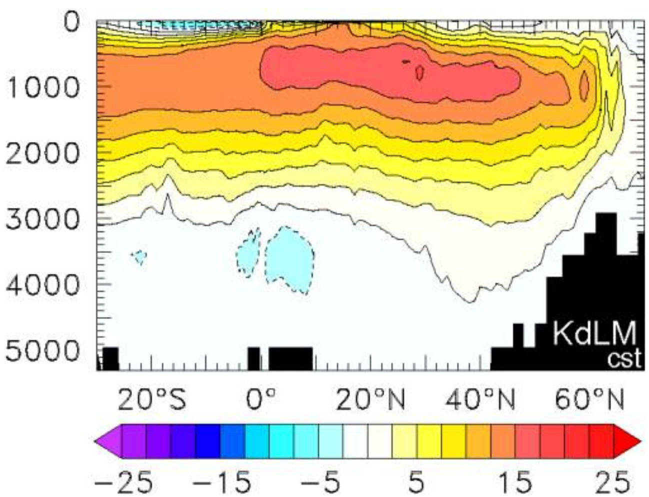
$\varepsilon \sim N^2$

$\varepsilon \sim N$

$\varepsilon \sim e^{-\frac{z}{z_s}}$

20% local, 80% slopes

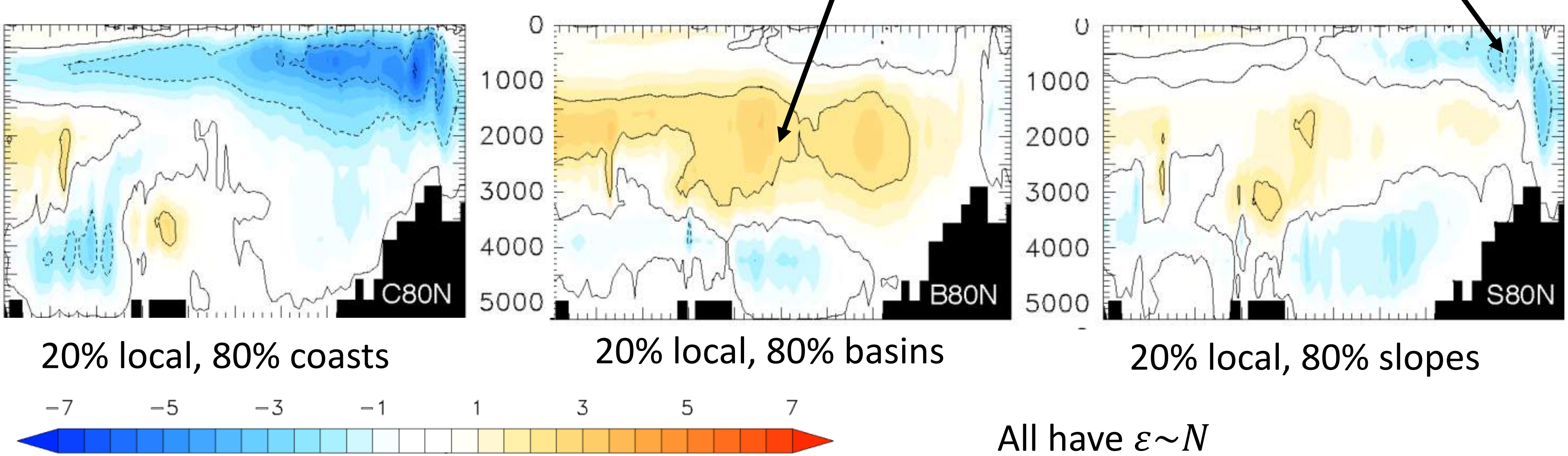
Horizontal location of mixing influences overturning if it dilutes dense overflows



Atlantic MOC (Sv)

Reference: 20% local, 80% uniform diffusivity

- Mixing on shelves/straits weakens AMOC
- Deep mixing strengthens/deepens AMOC



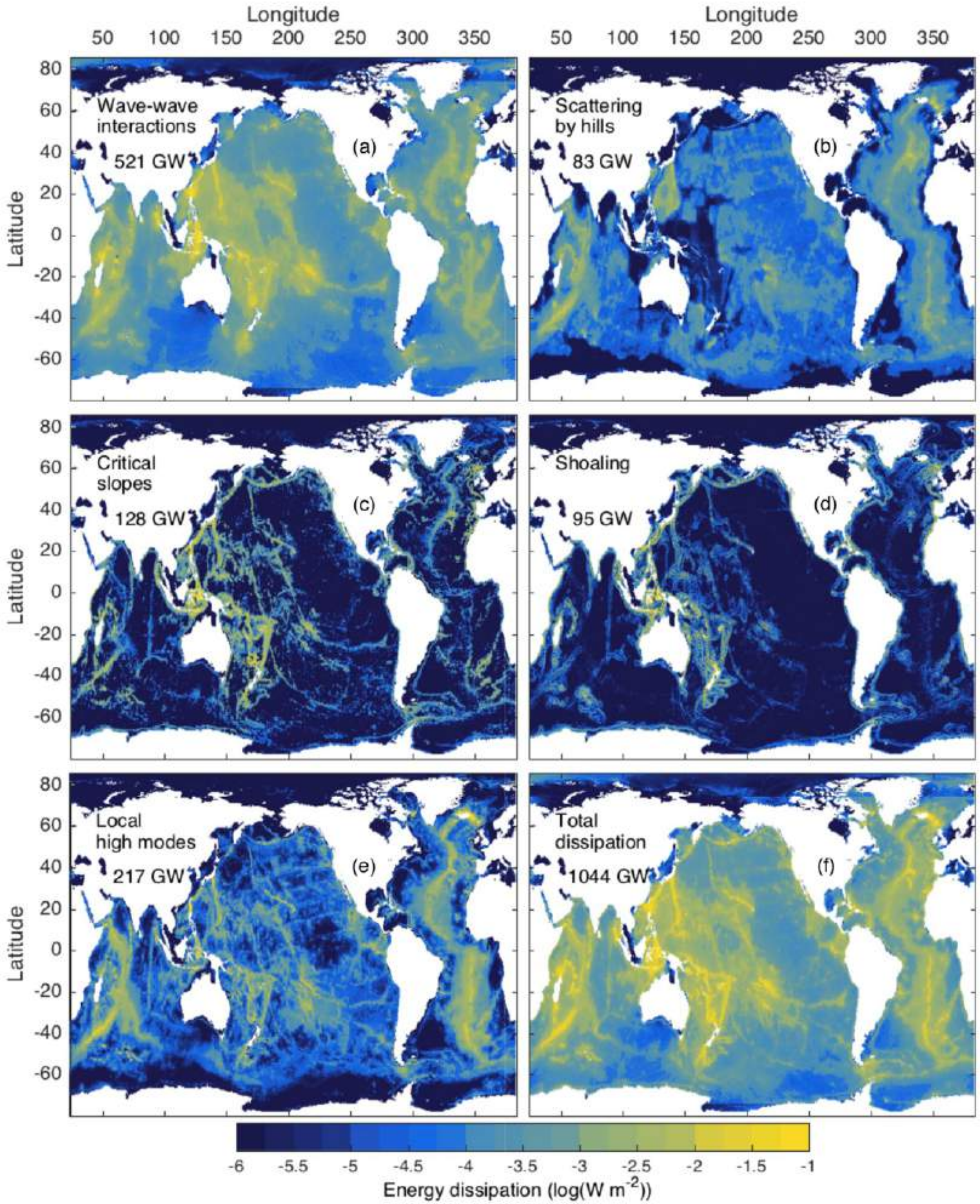
20% local, 80% coasts

20% local, 80% basins

20% local, 80% slopes

ΔAMOC (Sv)

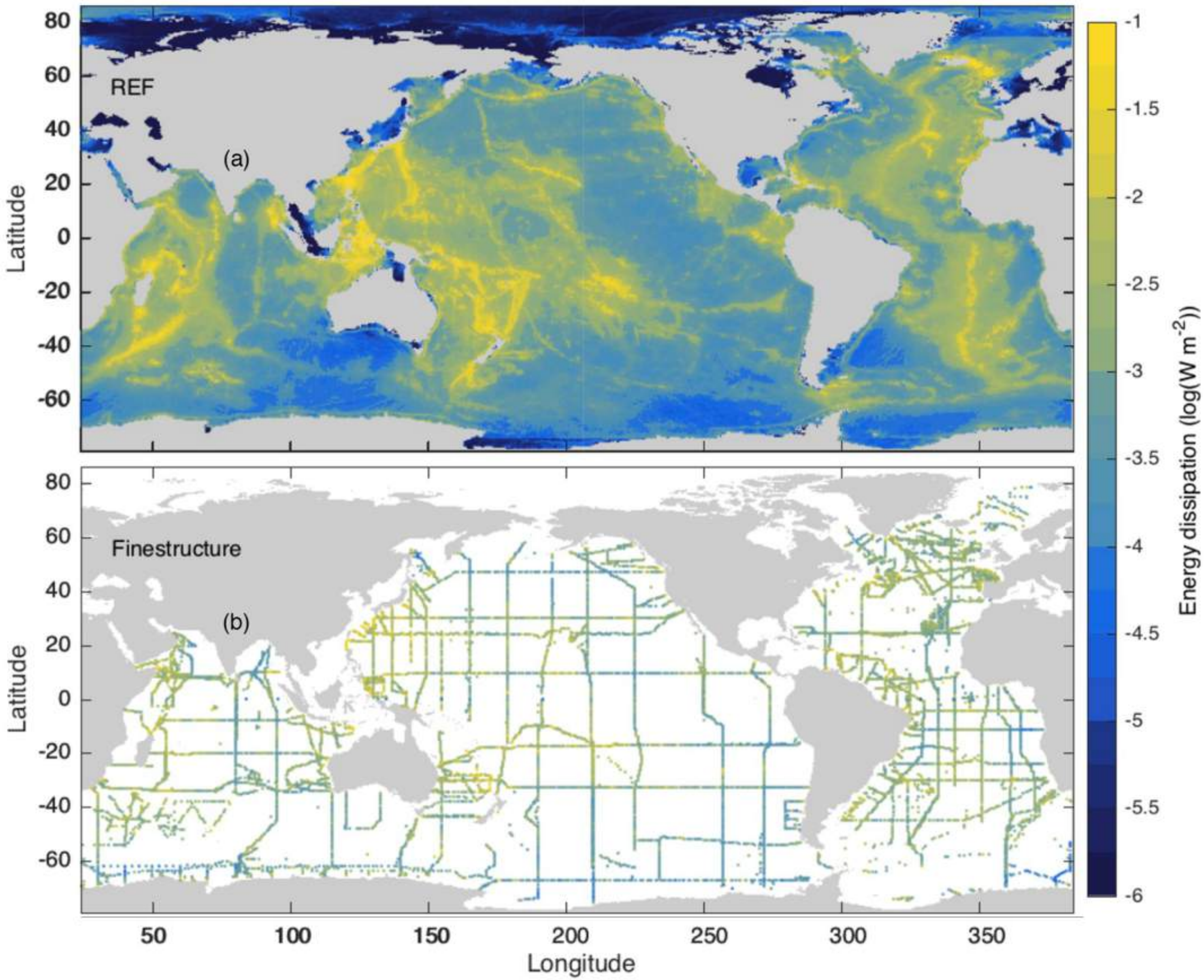
All have $\varepsilon \sim N$



The current best estimate of contributions of different processes to internal tide-driven mixing

(de Lavergne et al, 2019). Calculated using global 3D ray-tracing and WOCE climatology. Too expensive to calculate during run-time of a global model

Theoretical tidally-generated dissipation compares reasonably well with observational fine-structure estimates of dissipation

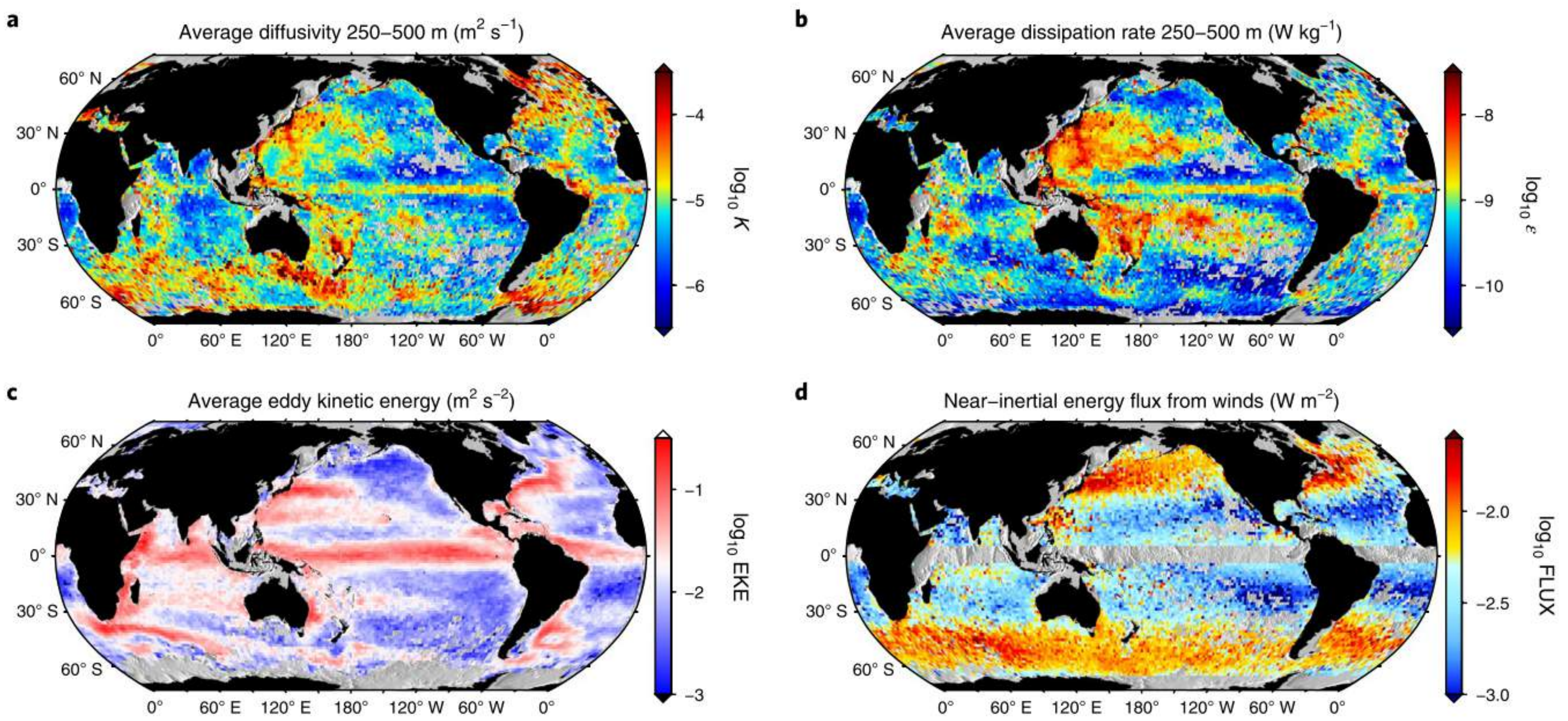


de Lavergne et al, 2019

Kunze 2017

Mixing in the stratified upper ocean due to winds

(Whalen et al, 2018)

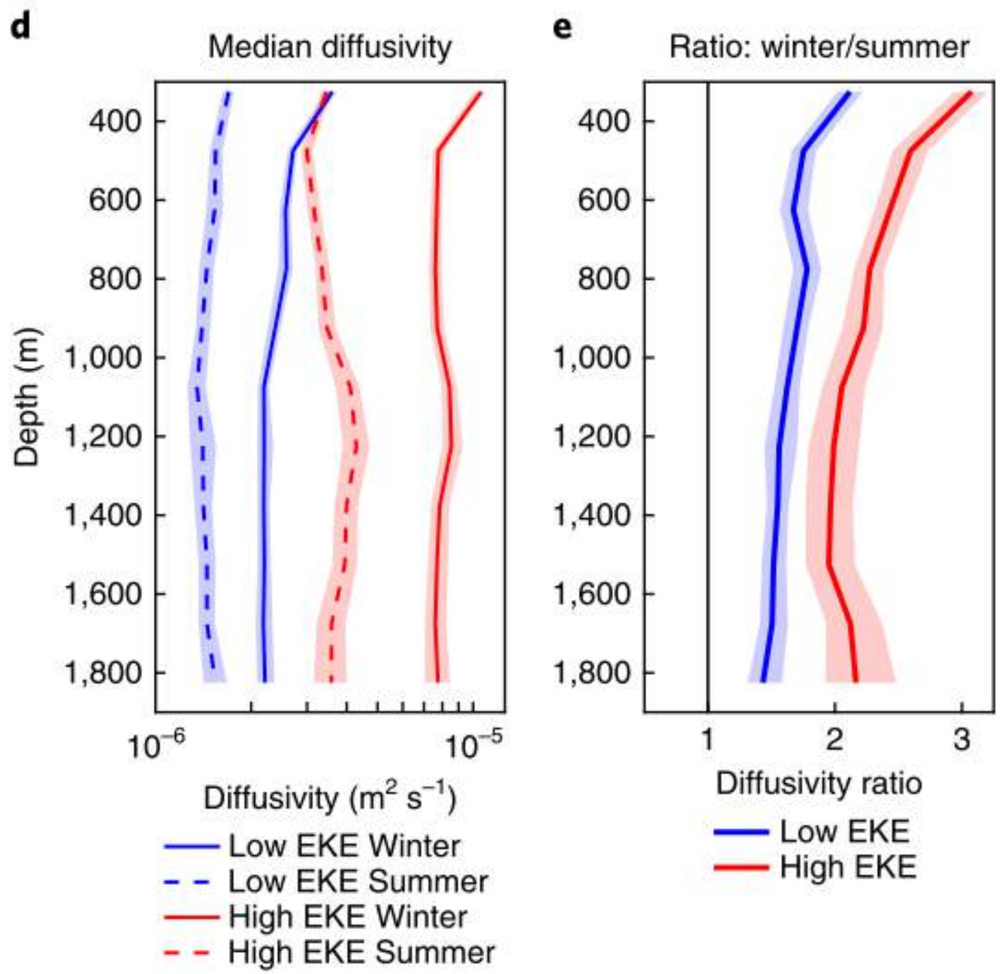


Upper ocean diffusivity/dissipation is enhanced by both (mesoscale) eddy kinetic energy and wind energy flux

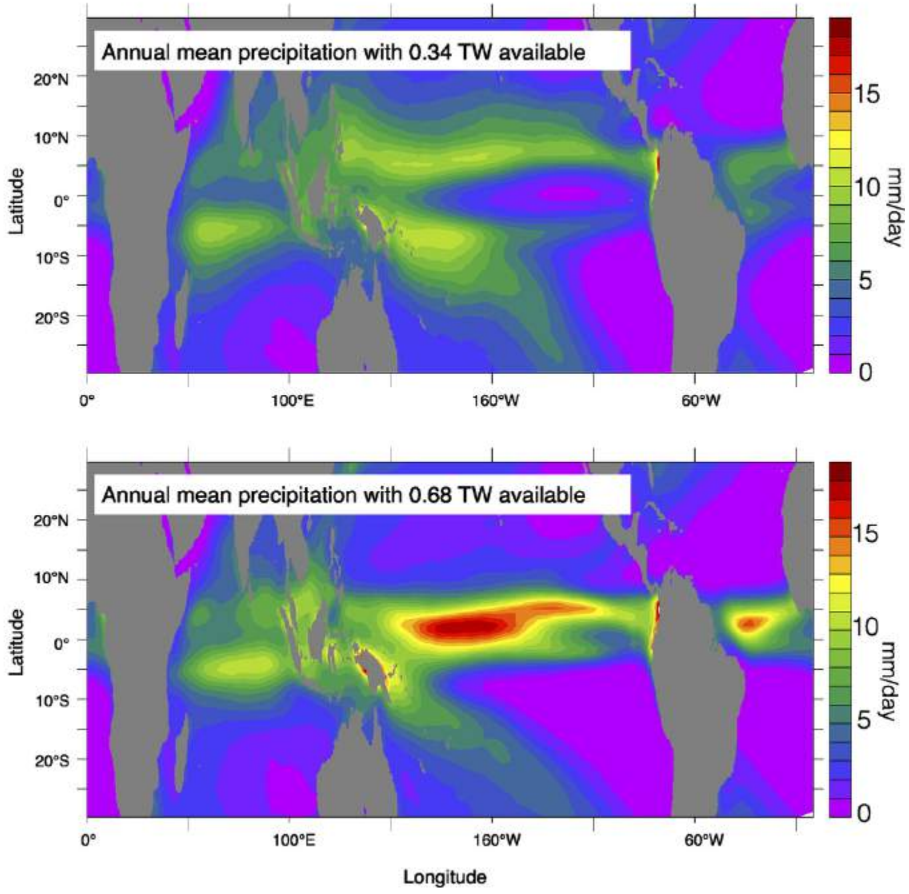
(Deduced from ARGO profiles and surface drifters)

Mixing in the stratified upper ocean due to winds

Diffusivities show a strong seasonal cycle, down to 2000m, enhancement in regions of high eddy kinetic energy, and stronger seasonal cycle in regions of high EKE (Whalen et al, 2018)



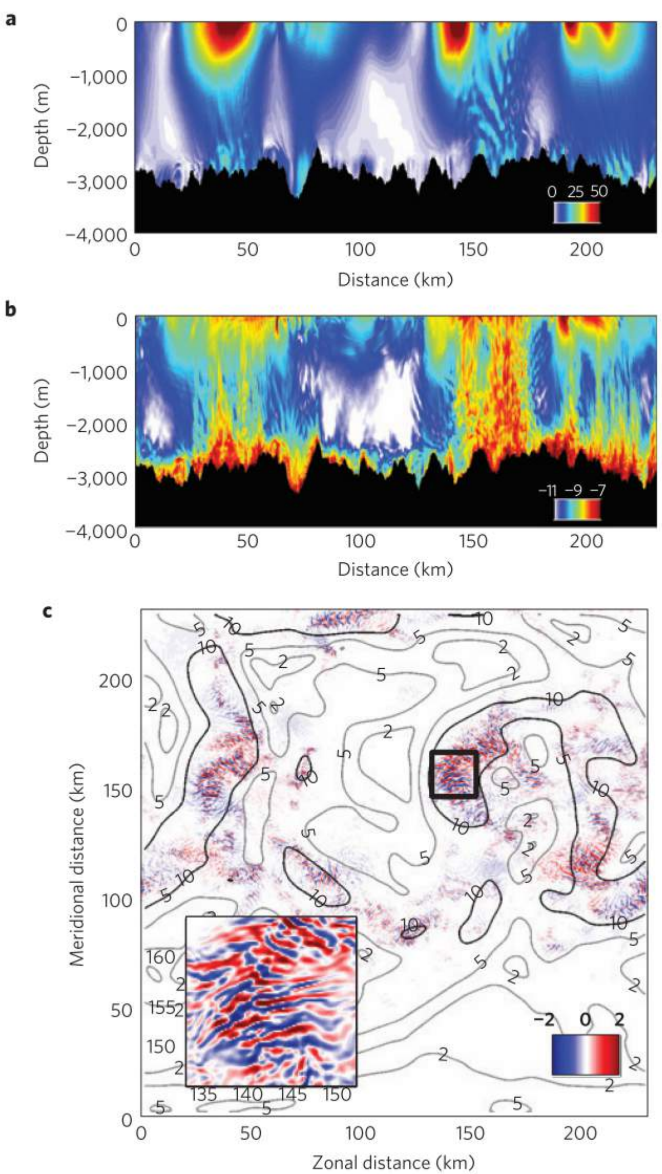
Impact of wind-generated diapycnal mixing in a global climate model



Jochum et al, 2013

NB: doesn't include interaction between wind and eddies

Mixing by oceanic lee waves



Speed (cm/s)

Energy dissipation (log10(W/kg))

Vertical velocity (cm/s) at 2km depth

Eddies flowing over rough bottom topography generate lee-waves (quasi-stationary internal waves) which lead to bottom-enhanced dissipation/mixing. *(MITgcm simulations, Nikurashin et al, 2012)*

Parameterizing mixing by oceanic lee waves

A first step is modeled on the internal tide parameterization

Local dissipation

St Laurent et al, 2002

$$\varepsilon = \frac{1}{\rho} E(x,y) \cdot q \cdot F(z)$$

Rate of energy conversion from subinertial flow to lee-waves per unit area.

Fraction of local dissipation.

Vertical structure function $\int_{-H}^{0} F(z)dz = 1$

(*Melet et al, 2014*)

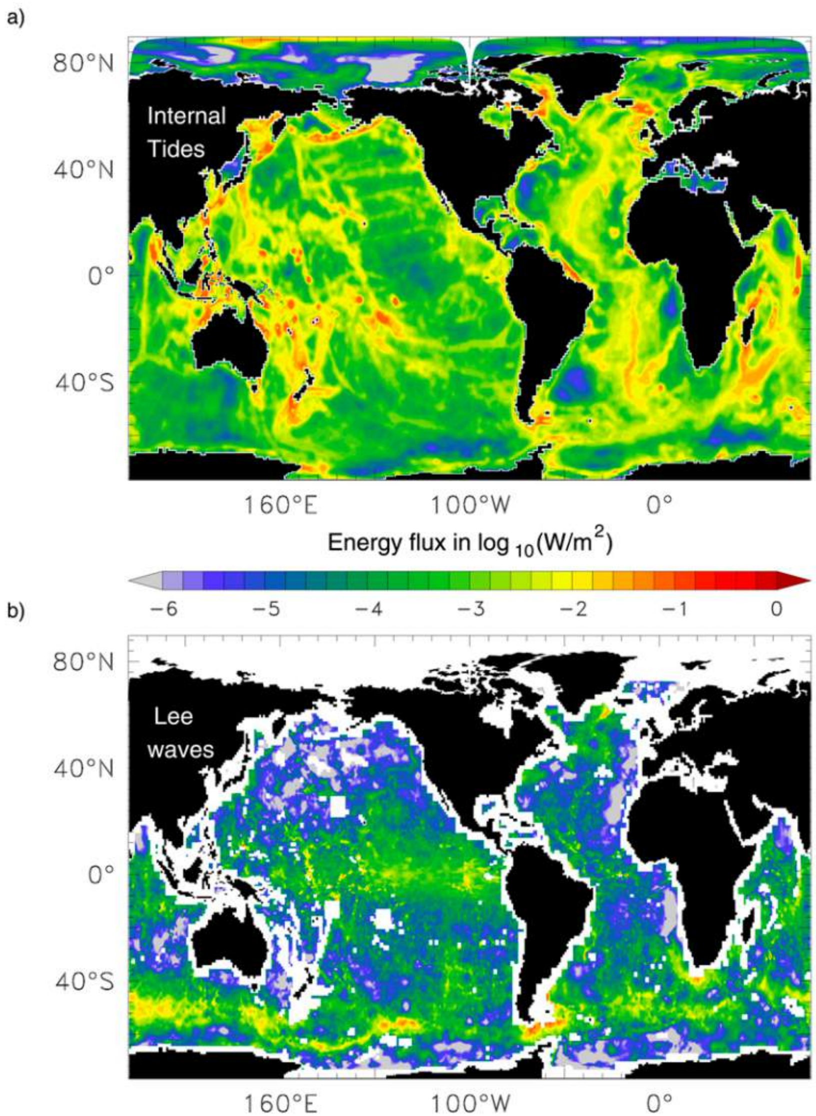
Energy conversion: linear prediction from Bell, 1975, given:
intrinsic frequency, $\omega = -Uk$, and wavenumber limits for lee-wave generation $\frac{f}{U} < k < \frac{N}{U}$.

$$E = \frac{\rho_0 |\mathbf{U}|}{2\pi} \int_{|f|/|\mathbf{U}|}^{N/|\mathbf{U}|} P_*(k) \sqrt{N^2 - |\mathbf{U}|^2 k^2} \sqrt{|\mathbf{U}|^2 k^2 - f^2} dk$$

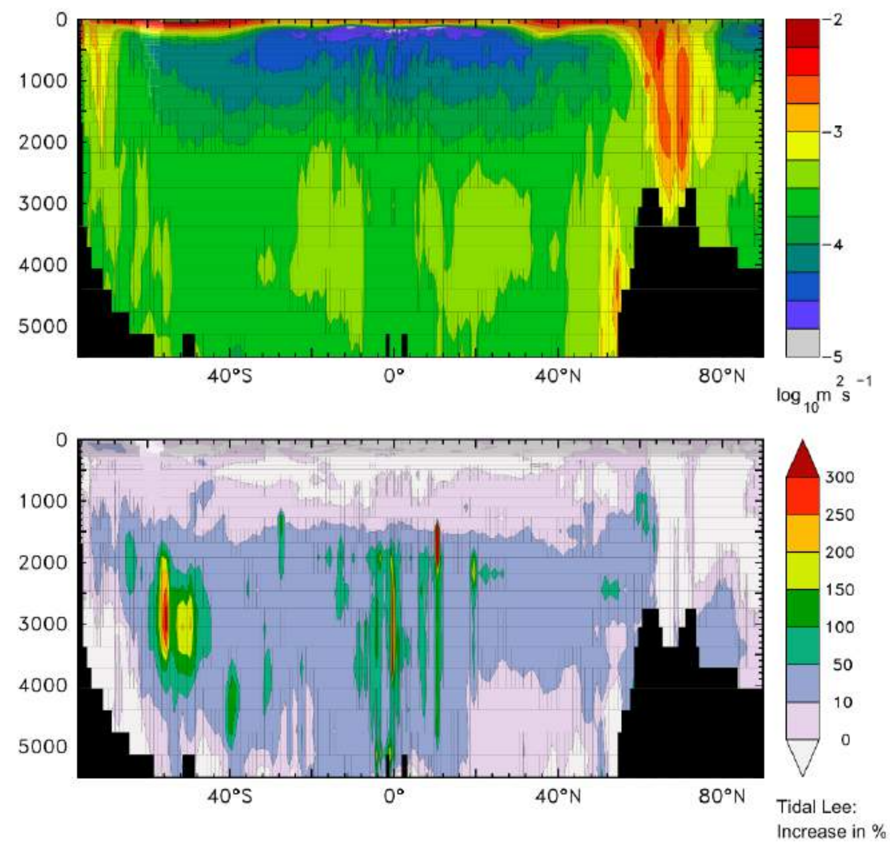
Nikurashin and Ferrari, 2011

Requires spectral representation of small-scale topography O(1km), and bottom velocities from mesoscale eddies, i.e. high-resolution global simulations or parameterization of mesoscale kinetic energy.

Impact of lee-wave parameterization



Diffusivity without lee-waves

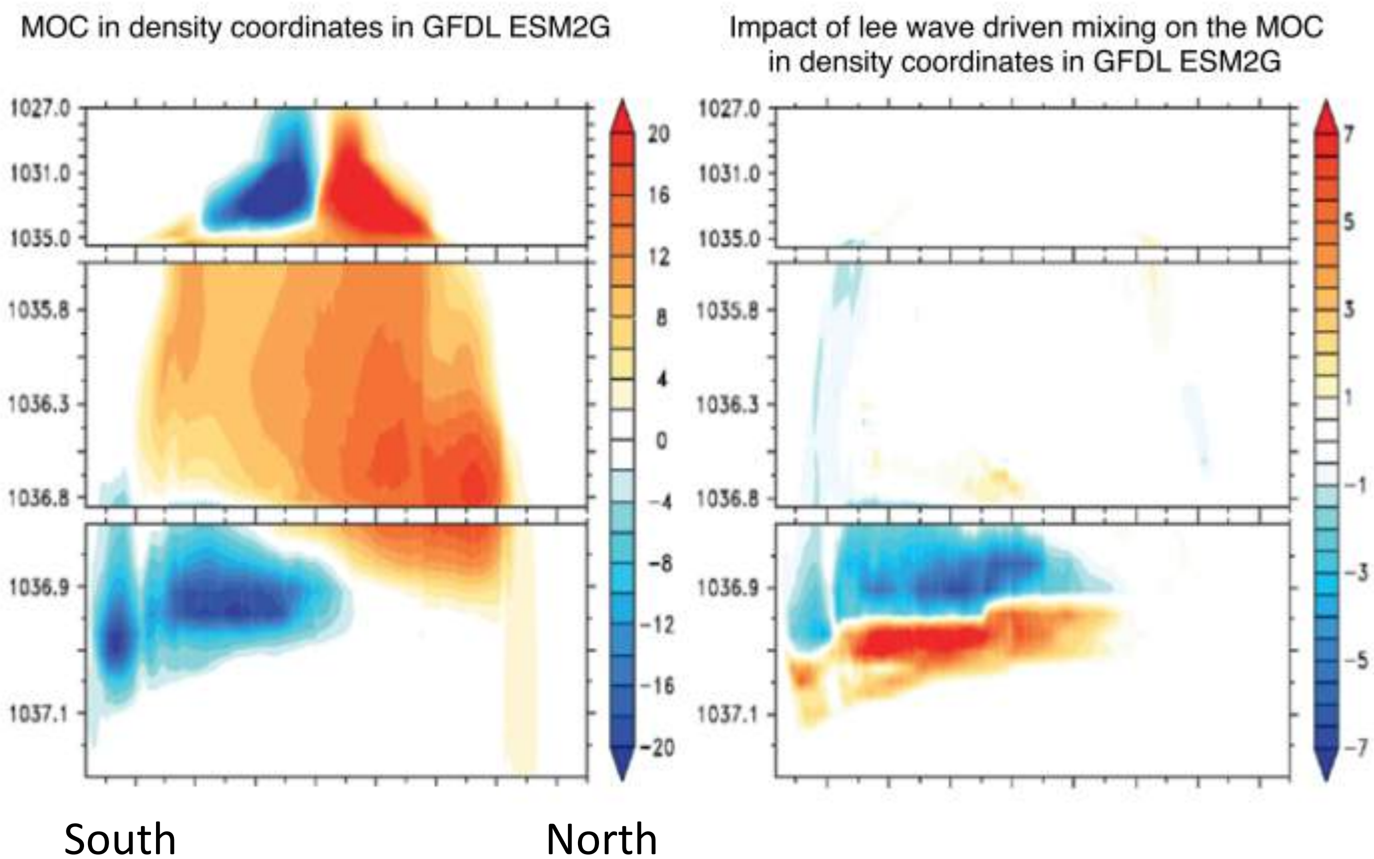


Additional diffusivity from lee-waves

(Melet et al, 2014)

Lee wave energy input is concentrated in Southern Ocean

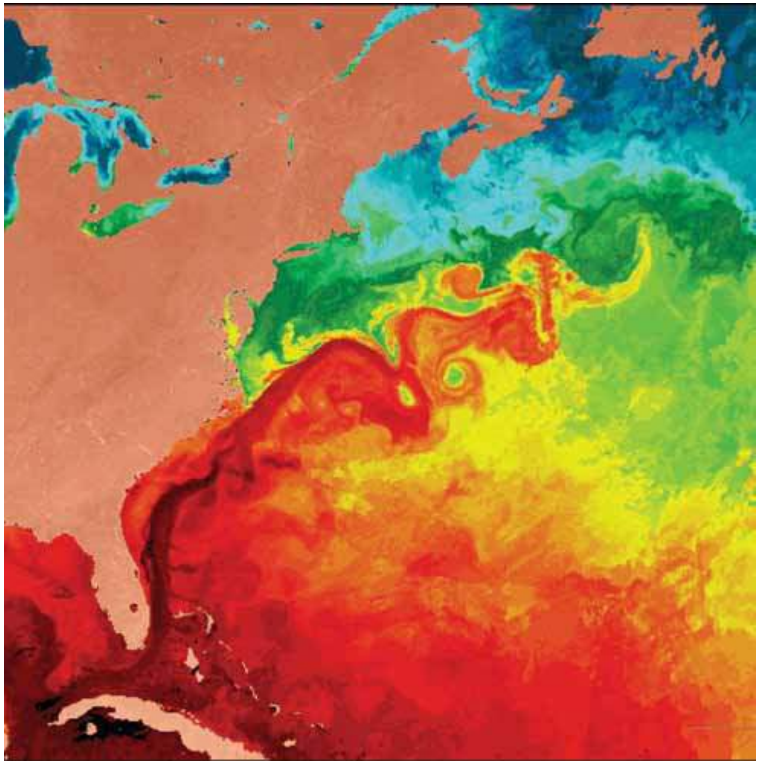
Impact of lee-wave mixing parameterization on overturning



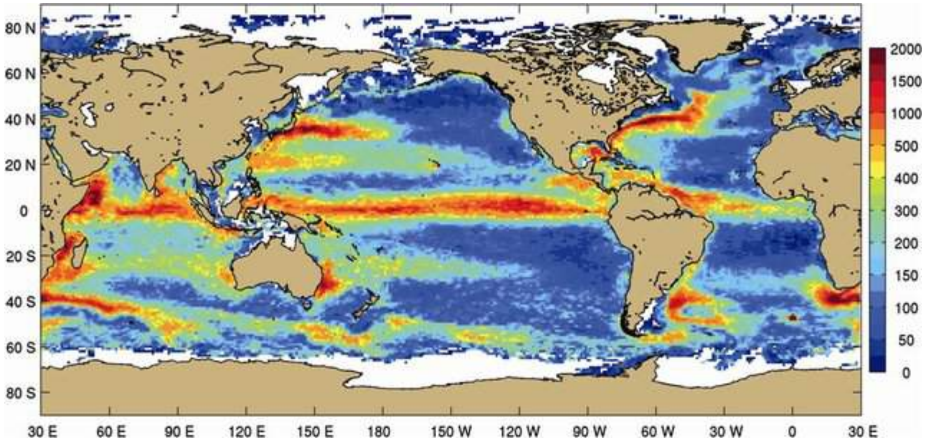
Melet et al, 2014

Additional mixing in the deep Southern Ocean leads to a lighter, stronger deep overturning cell.

Mesoscale eddies in the ocean



Sea Surface Temperature *(Talley, 2000)*



Eddy kinetic energy (cm^2/s^2) from surface drifters *(Maximenko et al, 2013)*

The ocean is full of geostrophic eddies generated by instability at density fronts. These deformation-radius scale O(5-50km) eddies are below the grid-scale of most global models, and their effects on larger scales need to be parameterized.

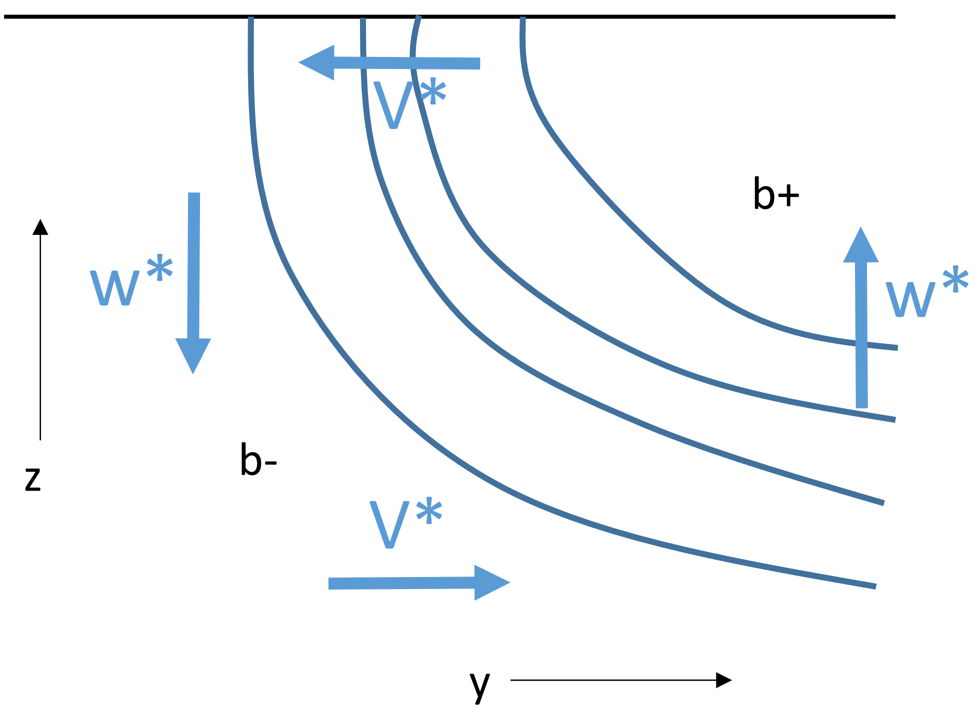
Parameterization of mesoscale eddies

Tracer equation $$\frac{\partial \varphi}{\partial t} + \nabla.(\boldsymbol{u}\varphi + \boldsymbol{u}^*\varphi) = \nabla.(\kappa_\sigma \nabla_\sigma \varphi) + \frac{\partial}{\partial z}\left(\kappa_z \frac{\partial \varphi}{\partial z}\right)$$

Eddy-induced velocity $\boldsymbol{u}^* = -\nabla \times \boldsymbol{\Psi}^*$

Along-isopycnal stirring

Diapycnal/vertical mixing



$$\boldsymbol{\Psi}^* = (\kappa_{GM} S_y, \kappa_{GM} S_{x,}, 0)$$

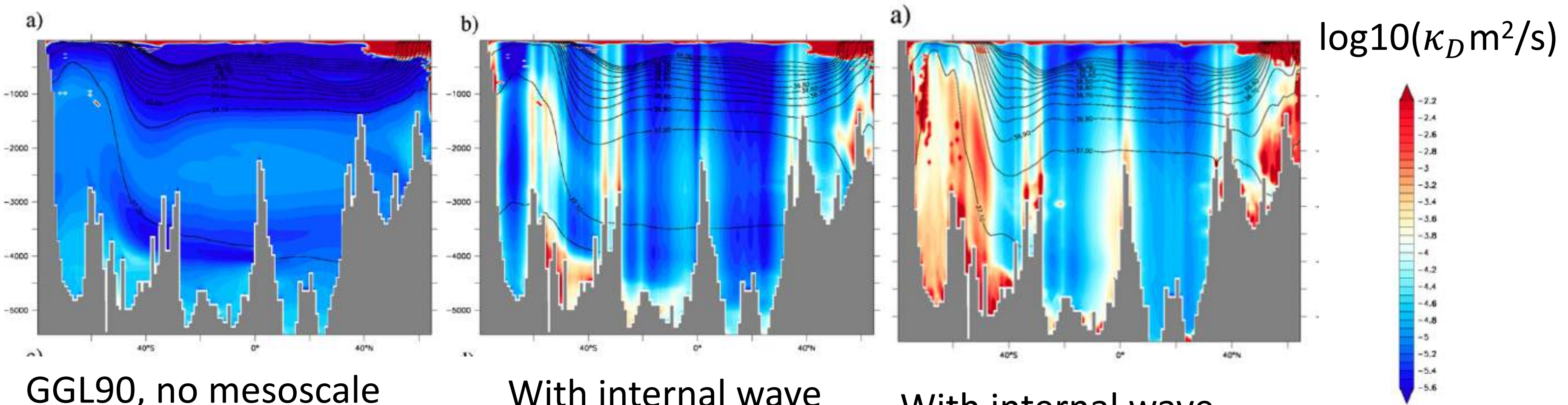
$$S_y = \frac{\partial b}{\partial y} \Big/ \frac{\partial b}{\partial z}$$

Eddy-induced velocities tend to adiabatically flatten isopycnals, reducing potential energy.
Along isopycnal stirring mixes tracers along isopycnals, without changing density.

Many different recipes for diffusivities:
κ_{GM}= Gent-McWilliams eddy diffusivity, κ_σ= Redi diffusivity.

Combining internal wave and mesoscale parameterizations in one energetically consistent framework

Mesoscale eddy kinetic energy contributes to internal wave-driven mixing, through interactions with surface winds and lee-wave generation.
Energetically consistent parameterizations should account for energy transfers between mesoscale eddies, internal waves, and small-scale turbulence.
Eden et al, 2014 is a first step to an energetically-consistent model.



GGL90, no mesoscale KE transfer, no internal waves

With internal wave parameterization, no mesoscale KE transfer

With internal wave parameterization and near-bottom mesoscale KE transfer

Use of ECCO for ocean mixing:
(a) estimating values of eddy diffusivities

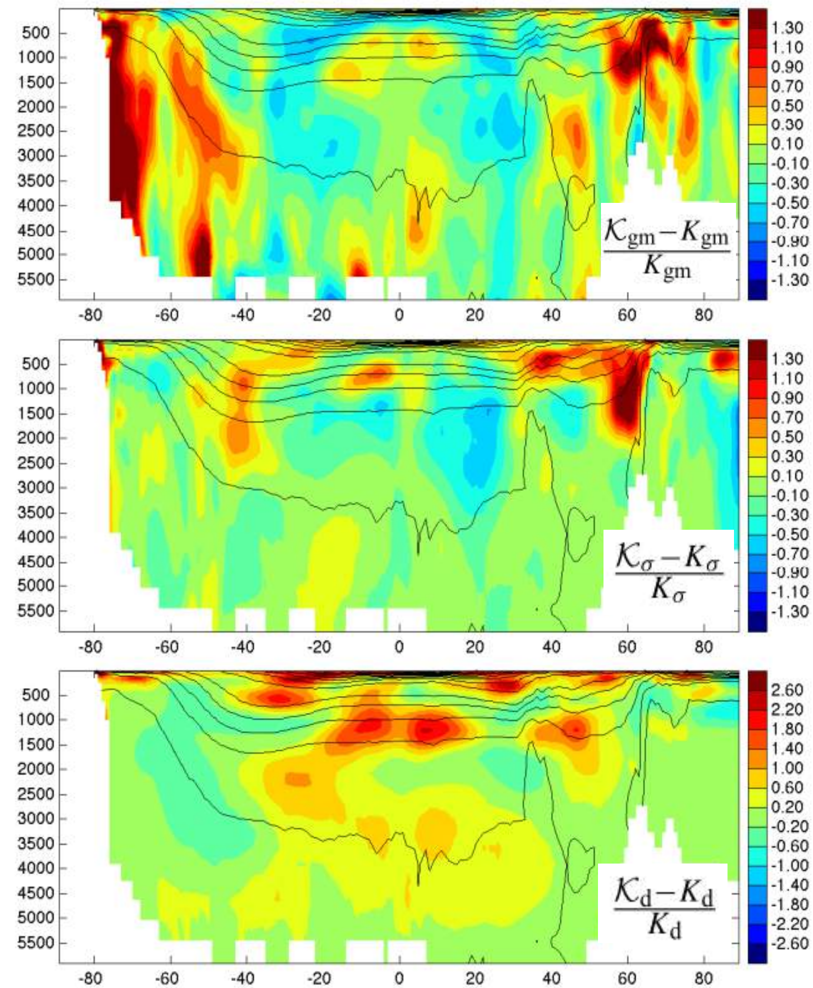
$$\frac{\partial \varphi}{\partial t} + \nabla.(\boldsymbol{u}\varphi + \boldsymbol{u}^{*}\varphi) = \nabla.(\kappa_{\sigma}\nabla_{\sigma}\varphi) + \frac{\partial}{\partial z}\left(\kappa_{z}\frac{\partial \varphi}{\partial z}\right)$$

$$\boldsymbol{u}^{*} = -\nabla \times \boldsymbol{\Psi}^{*}$$

$$\boldsymbol{\Psi}^{*} = (\kappa_{GM}S_{y}, \kappa_{GM}S_{x,}, 0)$$

$$\kappa_{z} = \kappa_{GGL} + \kappa_{d}$$

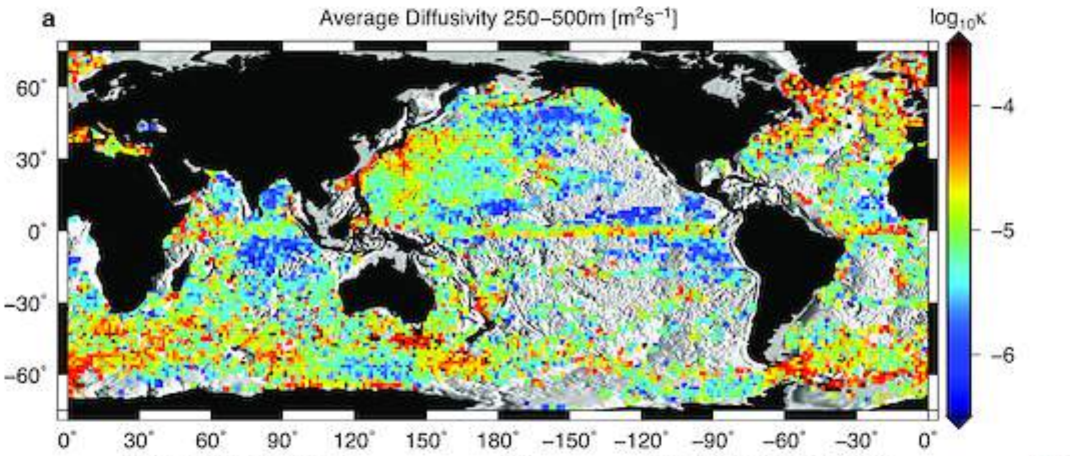
3D maps of κ_{GM}, κ_{σ}, and κ_{d} are estimated using a 20-year ECCO state estimate, constrained by ARGO data (Forget et al, 2015)



Zonal means of fractional change in eddy diffusivities (Forget et al, 2015), compared to initial uniform values.

Comparison of ECCO estimates with diapycnal diffusivities deduced directly from ARGO

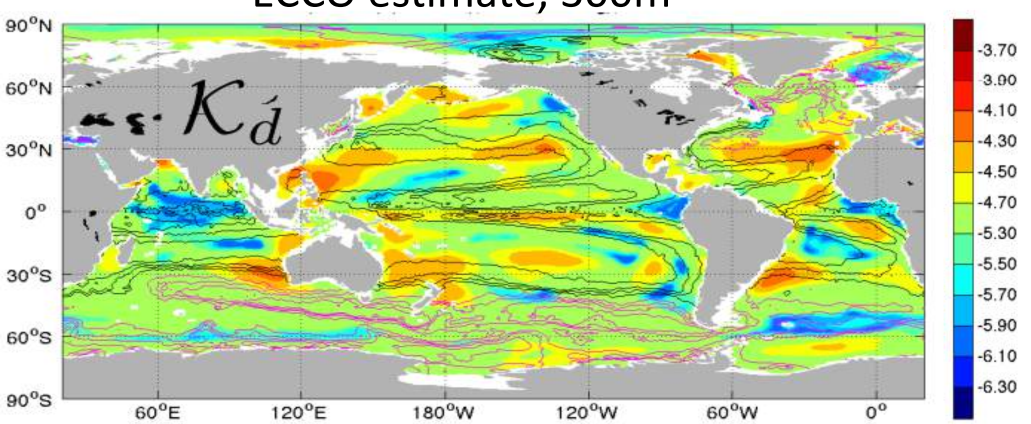
ARGO fine-structure estimate, 250-500m



(Whalen et al, 2012)

ECCO estimate, 300m

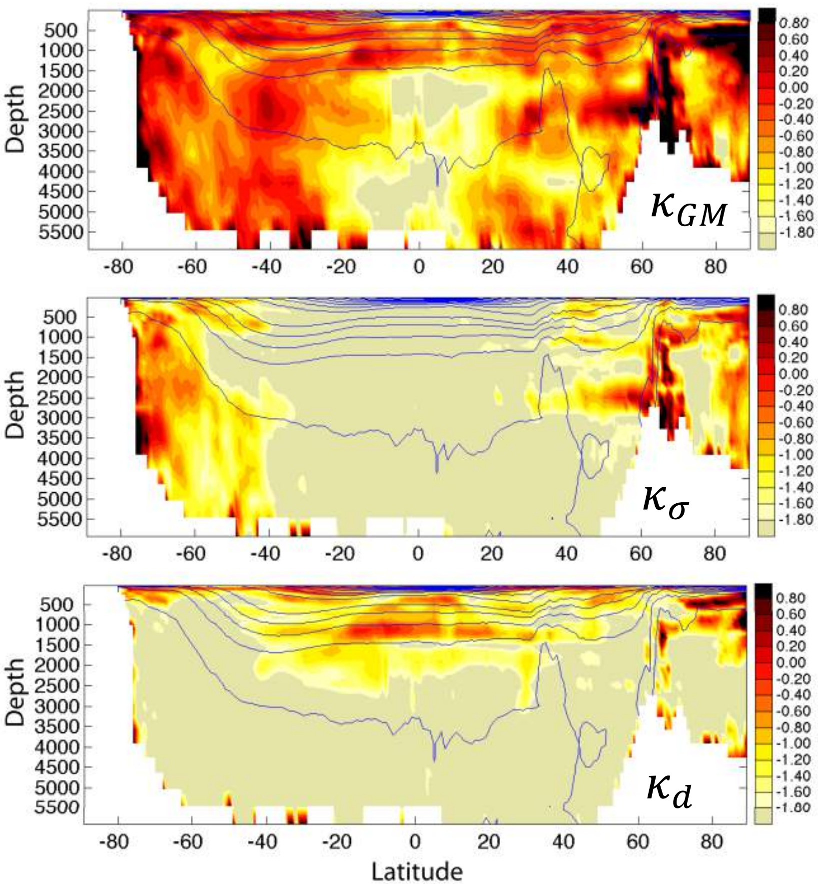
$\log 10(\kappa_d)$



(Forget et al, 2015)

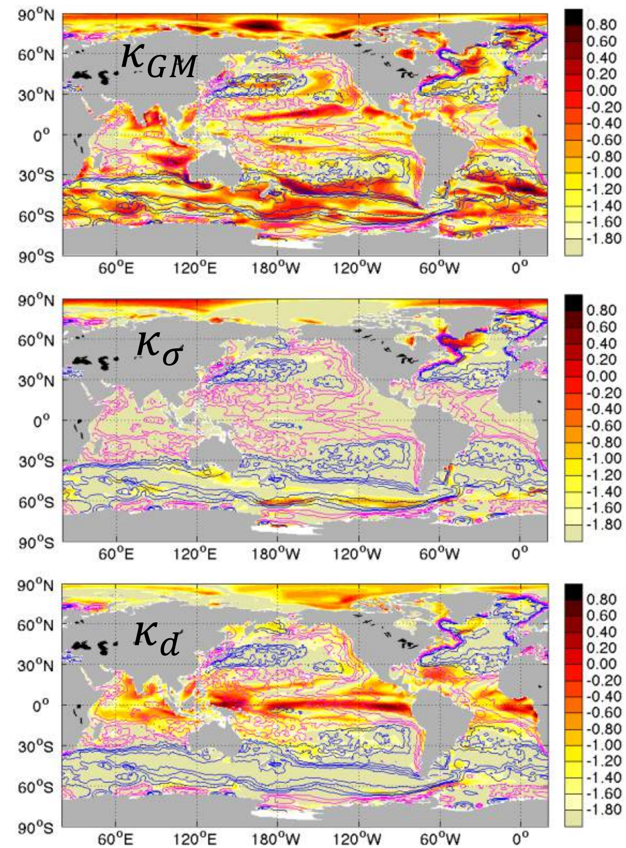
Use of ECCO for ocean mixing:
(b) sensitivity of solutions to eddy diffusivities

Sensitivity of ocean stratification



(log10(normalized mean square deviation))

Mixed layer depth sensitivity



Solutions are most sensitive to κ_{GM}.

What more could we do with ECCO to improve estimates of ocean mixing?

- 3D mixing maps from ECCO are static: mixing might change in future, as stratification changes
- Only physically-based parameterizations can predict how eddy diffusivities might evolve – can we use ECCO to better tune our physically-based parameterizations? (e.g. determine where baroclinic tidal energy is dissipated)
- Many sub-grid scale processes are missing in ECCO, e.g. submesoscale instabilities, mesoscale eddy backscatter, energy transfers from mesoscale to mixing, mixed layer fluxes due to Langmuir turbulence: can we include these?
- Can ECCO impose overall energetic constraints on parameterized mixing (as in Eden et al, 2014)?
- Can we constrain ECCO estimates with direct observational estimates of mixing (as in Kunze 2017 or Whalen et al, 2012)?

Difficulties in using ECCO to refine estimates of mixing

- Sometimes ECCO estimates of mixing may be reduced in order to compensate for unknown numerical mixing (e.g. in deep Southern Ocean).
- The best model for state estimates is not necessarily the best model for long-term climate simulations – how do we use one to improve the other?
- The data used to constrain solutions is limited in deep ocean – right where much of the ocean mixing occurs.