Ocean Modeling II

Parameterized Physics

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MODEL EQUATIONS

Momentum equations:

$$\frac{\partial}{\partial t}u + \mathcal{L}(u) - (uv\tan\phi)/a - fv = -\frac{1}{\rho_0 a\cos\phi}\frac{\partial p}{\partial\lambda} + \mathcal{F}_{Hx}(u,v) + \mathcal{F}_V(u) \quad (2.1)$$

$$\frac{\partial}{\partial t}v + \mathcal{L}(v) + (u^2\tan\phi)/a + fu = -\frac{1}{\rho_0 a}\frac{\partial p}{\partial\phi} + \mathcal{F}_{Hy}(u,v) + \mathcal{F}_V(v) \quad (2.2)$$

$$\mathcal{L}(\alpha) = \frac{1}{a\cos\phi}\left[\frac{\partial}{\partial\lambda}(u\alpha) + \frac{\partial}{\partial\phi}(\cos\phi v\alpha)\right] + \frac{\partial}{\partial z}(w\alpha) \quad (2.3)$$

$$\mathcal{F}_{Hx}(u,v) = A_M\left\{\nabla^2 u + u(1-\tan^2\phi)/a^2 - \frac{2\sin\phi}{a^2\cos^2\phi}\frac{\partial v}{\partial\lambda}\right\} \quad (2.4)$$

$$\mathcal{F}_{Hy}(u,v) = A_M\left\{\nabla^2 v + v(1-\tan^2\phi)/a^2 + \frac{2\sin\phi}{a^2\cos^2\phi}\frac{\partial u}{\partial\lambda}\right\} \quad (2.5)$$

$$\nabla^2\alpha = \frac{1}{a^2\cos^2\phi}\frac{\partial^2\alpha}{\partial\lambda^2} + \frac{1}{a^2\cos\phi}\frac{\partial}{\partial\phi}\left(\cos\phi\frac{\partial\alpha}{\partial\phi}\right) \quad (2.6)$$

$$\mathcal{F}_V(\alpha) = \frac{\partial}{\partial z}\mu\frac{\partial}{\partial z}\alpha \quad (2.7)$$

Continuity equation:

$$\mathcal{L}(1) = 0 \quad (2.8)$$

Hydrostatic equation:

$$\frac{\partial p}{\partial z} = -\rho g \quad (2.9)$$

Equation of state:

$$\rho = \rho(\Theta, S, p) \rightarrow \rho(\Theta, S, z) \quad (2.10)$$

Tracer transport:

$$\frac{\partial}{\partial t}\varphi + \mathcal{L}(\varphi) = \mathcal{D}_H(\varphi) + \mathcal{D}_V(\varphi) \quad (2.11)$$

$$\mathcal{D}_H(\varphi) = A_H\nabla^2\varphi \quad (2.12)$$

$$\mathcal{D}_V(\varphi) = \frac{\partial}{\partial z}\kappa\frac{\partial}{\partial z}\varphi, \quad (2.13)$$

3-D primitive equations in spherical polar coordinates with vertical z-coordinate for a thin, stratified fluid using hydrostatic & Boussinesq approximations.

"The common parameterization hypothesis about turbulent processes is that they mix material properties, hence the most common operator form is *eddy diffusion* (e.g. by spatial Laplacians) with an *eddy diffusivity* as the free parameter."

McWilliams, 1998, "Oceanic General Circulation Models", Ocean Modelling and Parameterization, NATO Science Series.

PARAMETERIZATIONS IN CESM1 POP2

- Vertical mixing (momentum and tracers)
  - surface boundary layer
  - interior
- Horizontal viscosity (momentum)
- Lateral mixing: mesoscale eddies (tracers)
- Overflows
- Submesoscale eddies (tracers)
- Diurnal cycle for short-wave heat flux
- Solar absorption

VERTICAL MIXING SCHEME:
K-PROFILE PARAMETERIZATION (KPP)

OCEANIC VERTICAL MIXING: A REVIEW AND A MODEL WITH A NONLOCAL BOUNDARY LAYER PARAMETERIZATION

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S. C. Doney
National Center for Atmospheric Research
Boulder, Colorado

1994

Reviews of Geophysics, 32, 4 / *November 1994*
pages 363–403
Paper number 94RG01872

8755-1209/94/94RG-01872*$15.00*

- Unresolved turbulent vertical mixing due to small-scale overturning motions parameterized as a vertical diffusion.

- Guided by study and observations of atmospheric boundary layer

$$\partial_t X = -\partial_z \overline{w'X'} \quad \xrightarrow{\text{parameterize}} \quad \overline{w'X'} = -K_x \partial_z X$$

where K_x represents an "eddy diffusivity" or "eddy viscosity"
and X = { active/passive scalars or momentum }

VERTICAL MIXING SCHEME: K-PROFILE PARAMETERIZATION (KPP)

• KPP is not just a vertical diffusion scheme because the scalars (Temp and Salinity) have non-local or "countergradient" terms γ_x

$$\overline{w'X'} = -K_x(\partial_z X - \gamma_x)$$

• KPP involves three high-level steps:

1. Determination of the boundary layer (BL) depth: *d*
2. Calculation of interior diffusivities: υ_x
3. Evaluation of boundary layer (BL) diffusivities: K_x

• Diffusivity throughout the boundary layer depends on the surface forcing, the boundary layer depth, and the interior diffusivity.

• KPP produces quite large diffusivities below the boundary layer, which mixes temp and salinity quite deep in times of very strong surface wind stress, such as strong midlatitude atmosphere storms.

VERTICAL MIXING SCHEME: K-PROFILE PARAMETERIZATION (KPP)

1. BL depth *d* is minimum depth where the bulk Richardson # (Ri_b) referenced to the surface equals a critical Richardson # (Ri_{cr}=0.3).

$$Ri_b(d) = \frac{\overbrace{\left[B_r - B(d)\right]d}^{\text{Stabilizing buoyancy difference}}}{\underbrace{\left|\mathbf{V}_r - \mathbf{V}(d)\right|^2 + V_t^2(d)}_{\text{Destabilizing velocity shear}}}$$

unresolved shear

B_r : near-surface reference buoyancy

$\mathbf{V}_r$: near-surface reference horizontal velocity

$V_t(d)$: velocity scale of (unresolved) turbulent shear at depth *d*

Ri measures the stability of stratified shear flow. "Boundary layer eddies with mean velocity $\mathbf{V}_r$ and buoyancy B_r should be able to penetrate to the boundary layer depth, *d*, where they first become stable relative to the local buoyancy and velocity."

VERTICAL MIXING SCHEME: K-PROFILE PARAMETERIZATION (KPP)

2. Calculation of interior diffusivities

$$\upsilon_x(d) = \upsilon_x^s(d) + \upsilon_x^w(d) + \upsilon_x^d(d) + \upsilon_x^c(d) + \upsilon_x^t(d)$$

υ_x : interior diffusivity at depth *d* (below the boundary layer)

υ_x^s : (unresolved) shear instability

υ_x^w : internal wave breaking

υ_x^d : double diffusion

υ_x^c : local static instability (convection)

υ_x^t : tidal mixing

Superposition of processes sets interior vertical diffusivity, υ_x, below the surface boundary layer.

VERTICAL MIXING SCHEME: K-PROFILE PARAMETERIZATION (KPP)

Verification example at Ocean Weather Station Papa (50°N, 145°W):

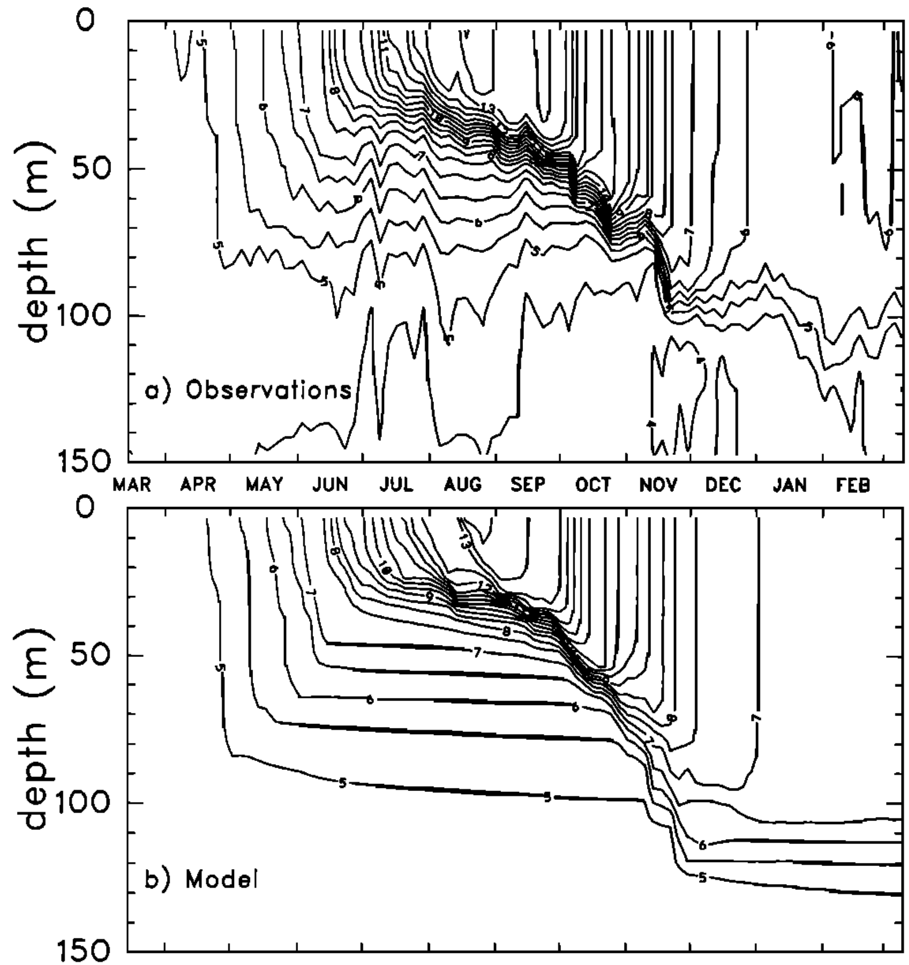

Figure 9. Time-depth sections of 4-day averages of observed temperatures in degrees Celsius (a) from ocean weather station (OWS) Papa during the ocean year March 15, 1961, to March 15, 1962 and (b) from the standard KPP simulation of OWS Papa.

Large et al (1994)

HORIZONTAL VISCOSITY

Spatially uniform, isotropic, Cartesian, Δ=250km grid for illustration

$$D(U) = A\ U_{xx} + A\ U_{yy}$$
$$D(V) = A\ V_{xx} + A\ V_{yy}$$

Grid Re (Diffuse Noise) $\rightarrow$ $A > 0.5\ V\,\Delta = 100{,}000\ m^2/s$

Resolve WBC (Munk Layers) $\rightarrow$ $A > \beta\ \Delta^3 = 80{,}000\ m^2/s$

Diffusive CFL $\rightarrow$ $A < 0.5\,\Delta^2 / \Delta t = 8000{,}000\ m^2/s$

Realism (EUC, WBC) $\rightarrow$ $A \sim$ physical $= 1{,}000\ m^2/s$

Smagorinsky $\rightarrow$ $A = C\,\Delta^2 \surd\ (\partial_x U)^2 + (\partial_y V)^2 + (\partial_x V + \partial_y U)^2$

ANISOTROPIC HORIZONTAL VISCOSITY

$$\partial_t u + \ldots = \partial_x (A \partial_x u) + \partial_y (B \partial_y u)$$

$$\partial_t v + \ldots = \partial_x (B \partial_x v) + \partial_y (A \partial_y v)$$

Grid Re (Diffuse Noise) $\rightarrow$ Live with the "noise"

Resolve WBC (Munk Layers) $\rightarrow$ $A = B = \beta \ \Delta^3$, only near WBC

elsewhere:

Realism (EUC, WBC) $\rightarrow$ $A = 300\ m^2/s$
$B = 300\ m^2/s$ in the tropics
$= 600\ m^2/s$ polewards of 30^o

Subject to diffusive CFL, but NO Smagorinsky

Mesoscale eddy mixing of tracers: Gent-McWilliams (GM) parameterization

150 JOURNAL OF PHYSICAL OCEANOGRAPHY VOLUME 20

1990

Isopycnal Mixing in Ocean Circulation Models†

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20 March 1989 and 14 August 1989

ABSTRACT

A subgrid-scale form for mesoscale eddy mixing on isopycnal surfaces is proposed for use in non-eddy-resolving ocean circulation models. The mixing is applied in isopycnal coordinates to isopycnal layer thickness, or inverse density gradient, as well as to passive scalars, temperature and salinity. The transformation of these mixing forms to physical coordinates is also presented.

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The Gent–McWilliams parameterization: 20/20 hindsight

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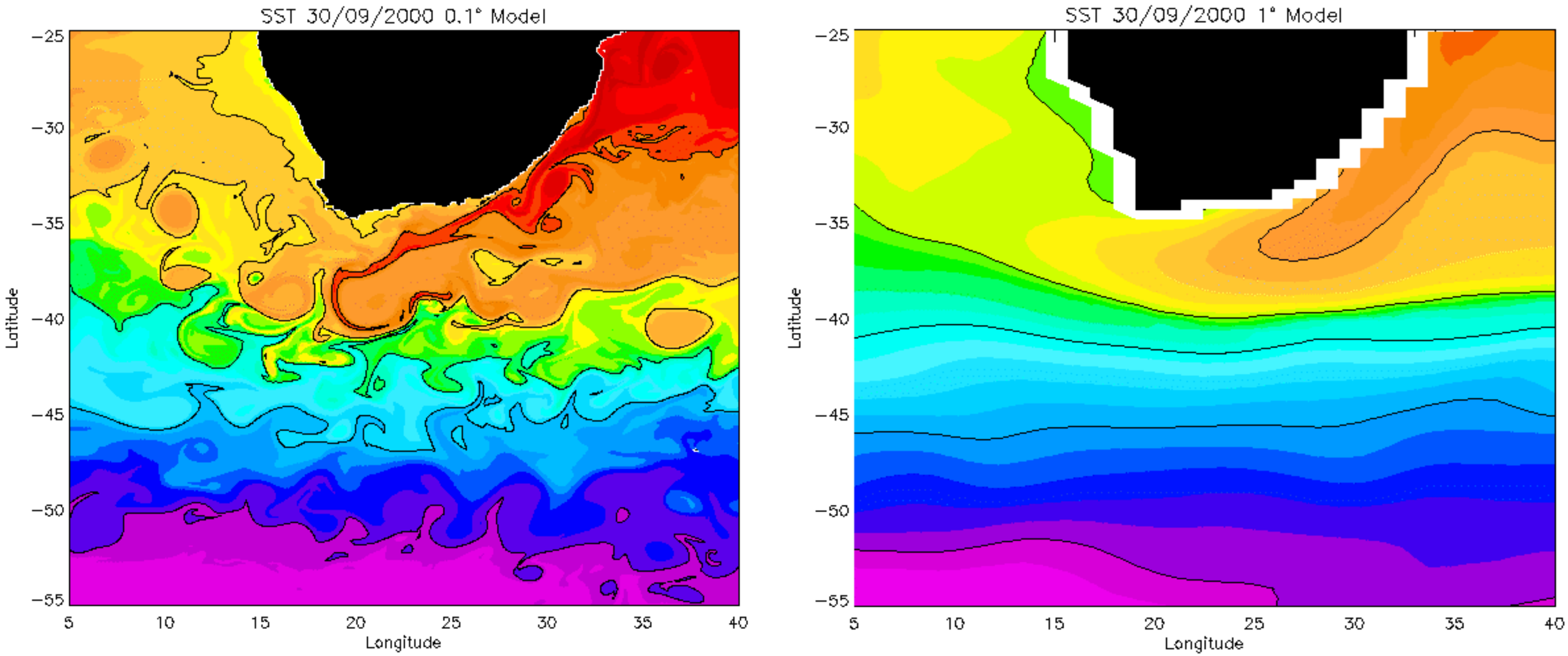
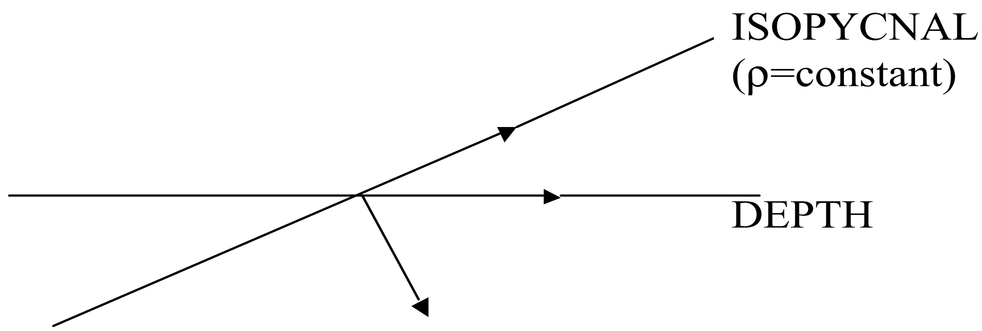
Why is GM needed?

Agulhas Retroflection



O(1°) models do not resolve the 1st baroclinic deformation radius away from the equatorial regions, and hence lack the mesoscale turbulence which mixes temperature, salinity and passive tracers in the real ocean.

Why is GM needed?



Isopycnal slopes are small $O(10^{-3})$ at most

Ocean Observations suggest mixing along isopycnals is $\sim 10^7$ times larger than across isopycnals.

- Early ocean models parameterized the stirring effects of (unresolved) mesoscale eddies by Laplacian *horizontal* diffusion with $K_H = O(10^3\ m^2/s)$, whereas the vertical mixing coefficient $K_v = O(10^{-4}\ m^2/s)$.

- Horizontal mixing results in excessive diapycnal mixing, which degrades the ocean solution: e.g. Veronis (1975) showed that it produces spurious upwelling in western boundary current regions which "short circuits" the N. Atlantic MOC.

- Thus, was a recognized need to orient tracer diffusion in z-coordinate models along isopycnal surfaces, to be consistent with observed ocean mixing rates.

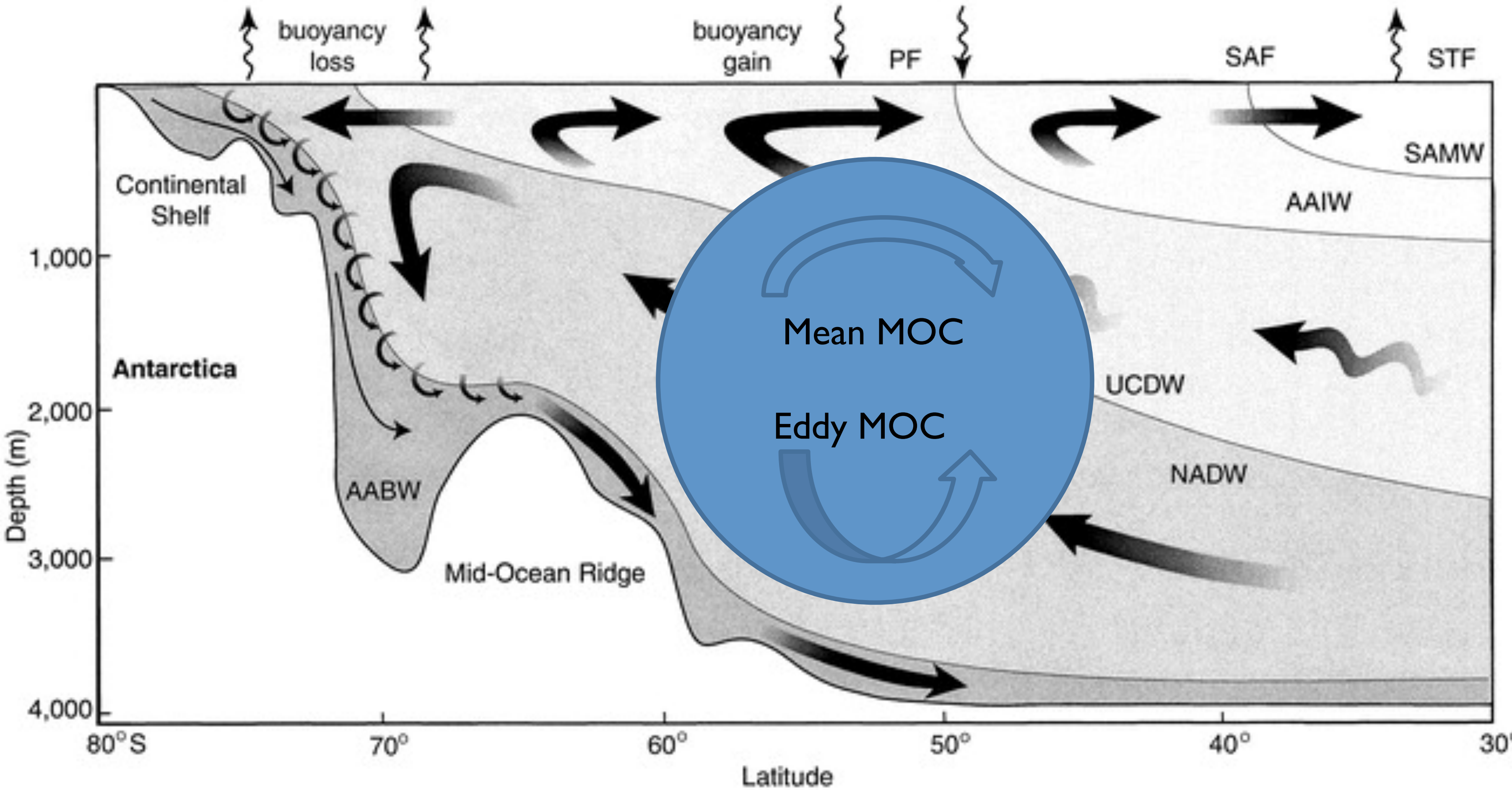
The GM Parameterization

$$\frac{\partial T}{\partial t} + (\underline{u} + \underline{u}^*).\underline{\nabla} T = \kappa \nabla_{\rho}^{2} T$$

$$w^* = -\underline{\nabla}.(\kappa \underline{\nabla} \rho / \rho_z), \underline{\nabla}.\underline{u}^* = 0.$$

GM (1990) proposed an eddy-induced velocity $\underline{u}^*$ in addition to diffusion along isopycnal surfaces.

Southern Hemisphere zonal wind jet



Baroclinic instability produces ACC eddies that try to flatten the isopycnals and produce a MOC that opposes the mean flow MOC.

GM impacts

Gent et al. (JPO, 1995): Eddy-induced velocity (v*,w*) acts to flatten isopycnals and minimize potential energy

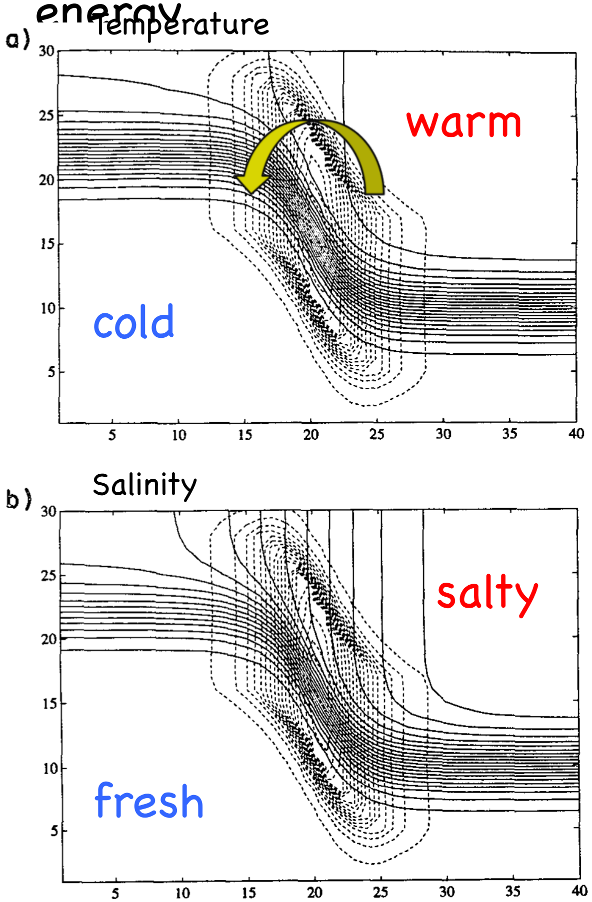


FIG. 3. Initial states of (a) temperature and (b) salt [contour interval one-quarter that of (a)]. Both panels also show the streamfunction $\kappa\rho_x/\rho_z$ for the parameterized eddy-induced transport velocity.

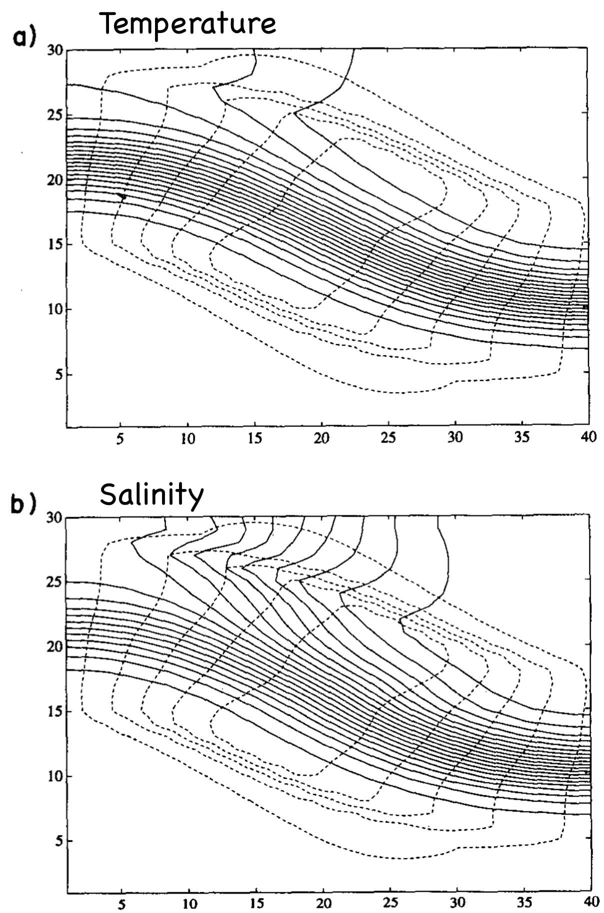


FIG. 4. Distributions of (a) temperature and (b) salt after an integration time of $20\Delta s^2/\kappa$. Both panels also show the streamfunction $\kappa\rho_x/\rho_z$ for the parameterized eddy-induced transport velocity. Contour intervals are the same as in Fig. 3.

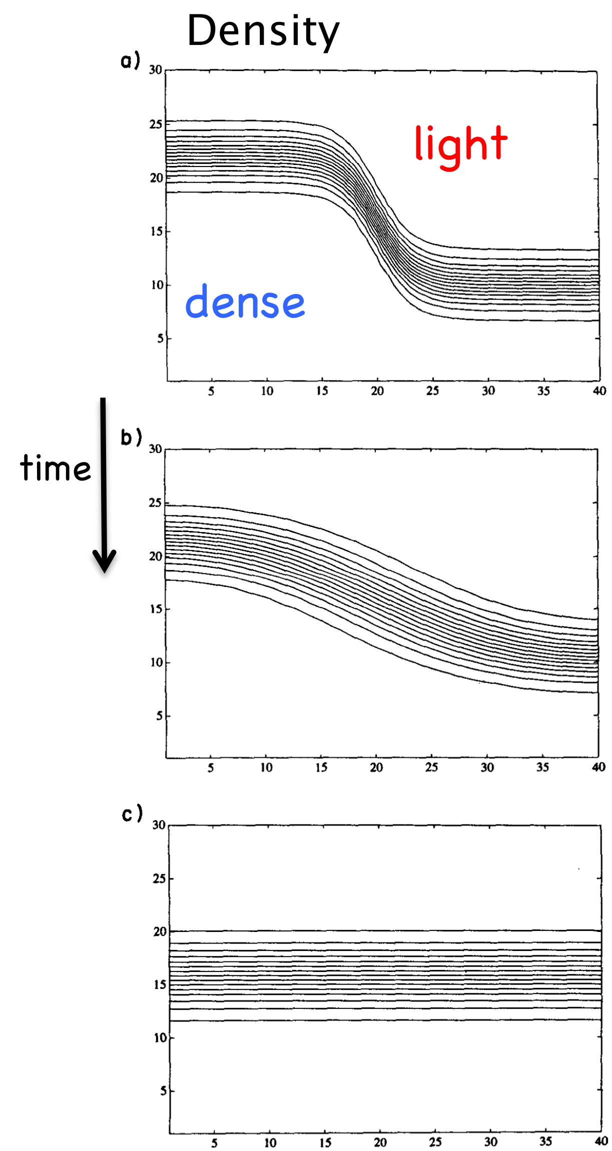
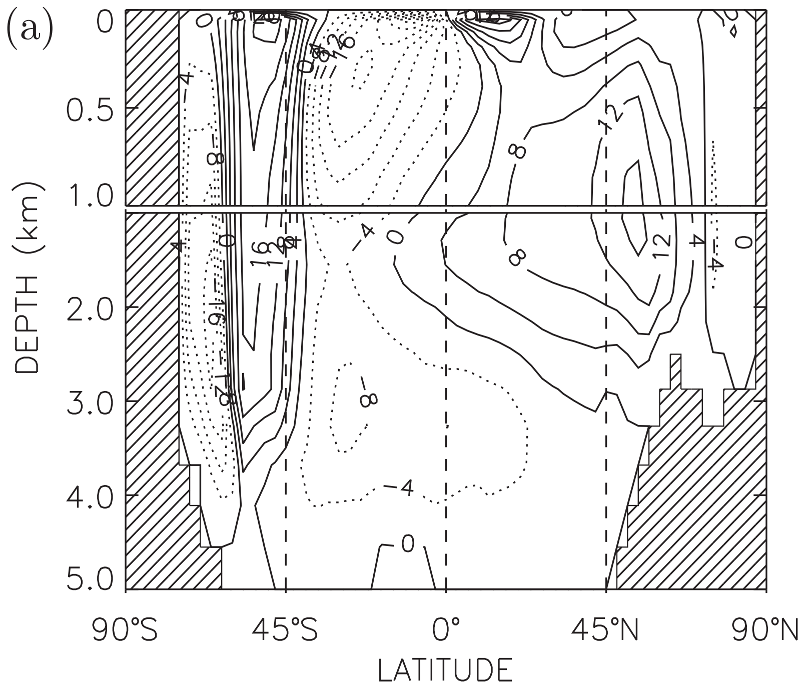


FIG. 5. Density distribution at various times of the integration. (a) Initial, (b) $20\Delta s^2/\kappa$, and (c) $1000\Delta s^2/\kappa$.

Impacts of GM



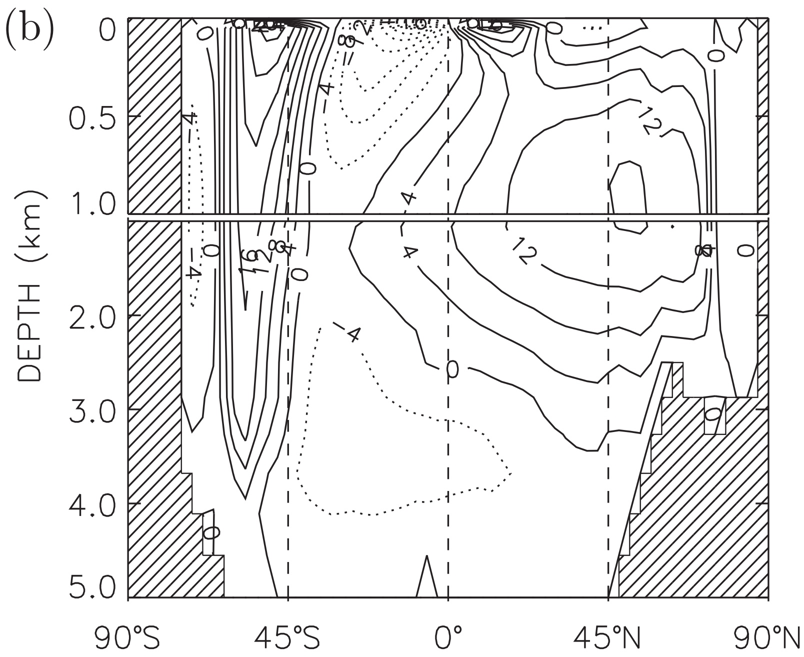
(a) Horizontal Diffusion, MOC (**u**)

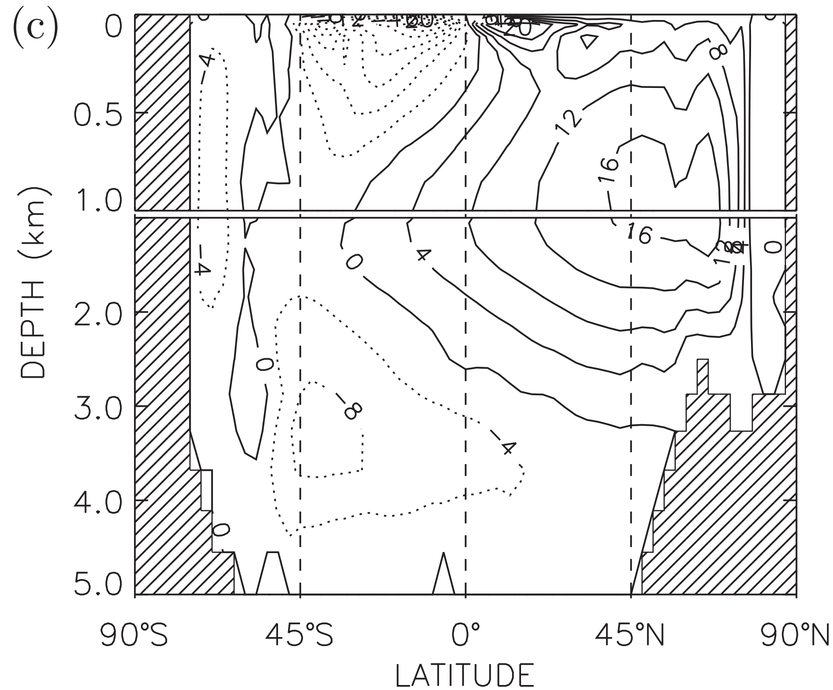
(b) GM, MOC (**u**)

(c) GM, MOC (**u+u***)

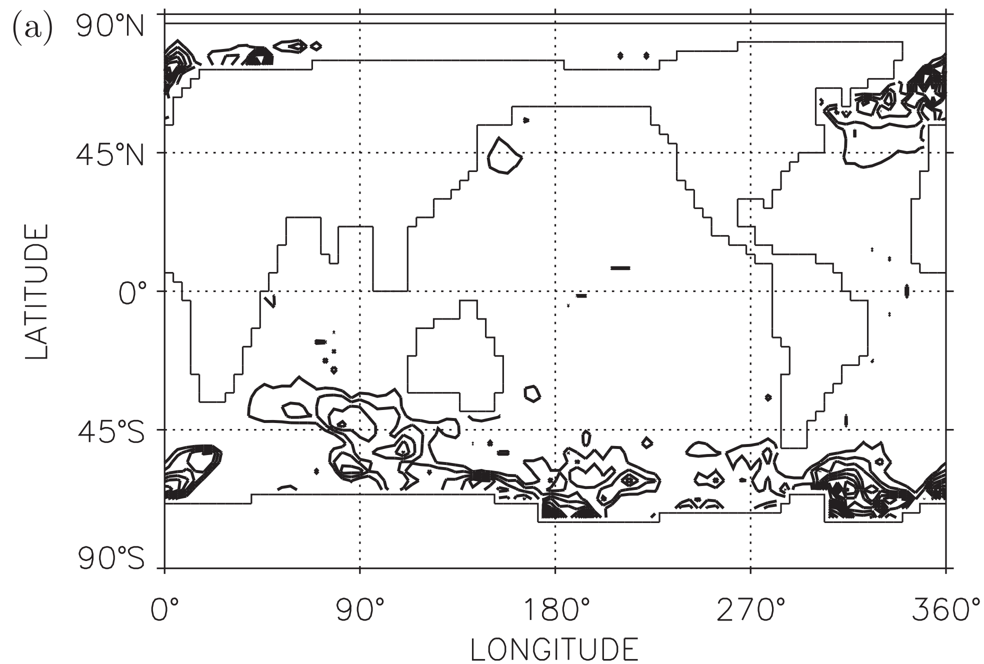
4° x 3° x 20L ocean model

Danabasoglu et al. (1994, Science)





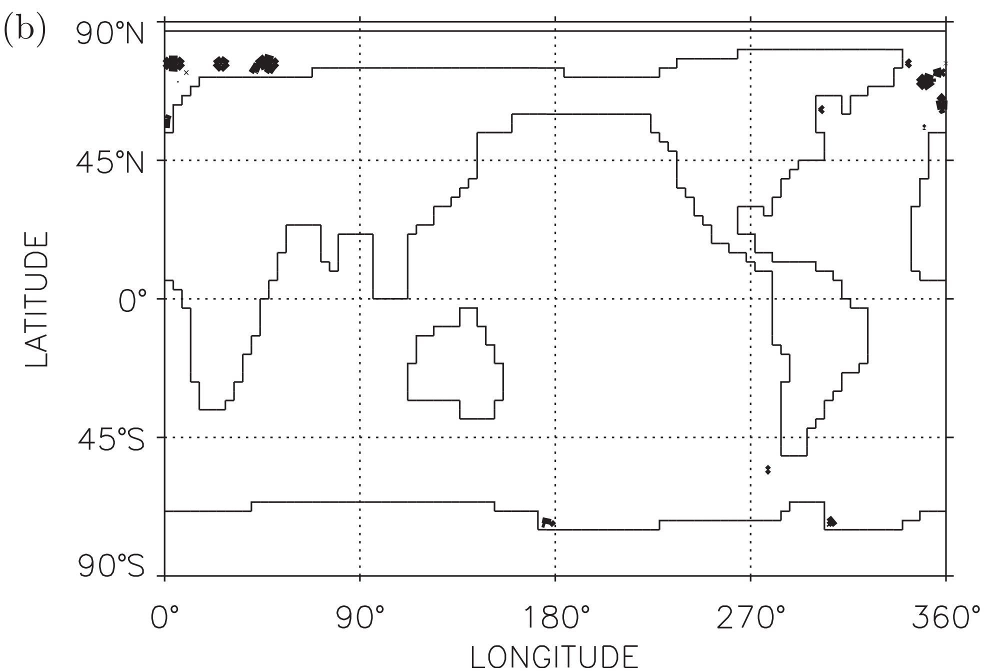
Impacts of GM



Deep Water Formation

(a) Horizontal Diffusion

(b) GM



In (b), deep water is formed only in the Greenland/ Iceland/Norwegian Sea, the Labrador Sea, the Weddell Sea and the Ross Sea.

$4^o \times 3^o \times 20L$ ocean model

Danabasoglu et al. (1994, Science)

GM: the near-surface eddy flux

GM90 is valid only in the nearly adiabatic ocean interior. Therefore, it needs to be modified near the surface as the isopycnals are nearly vertical in the mixed layer.

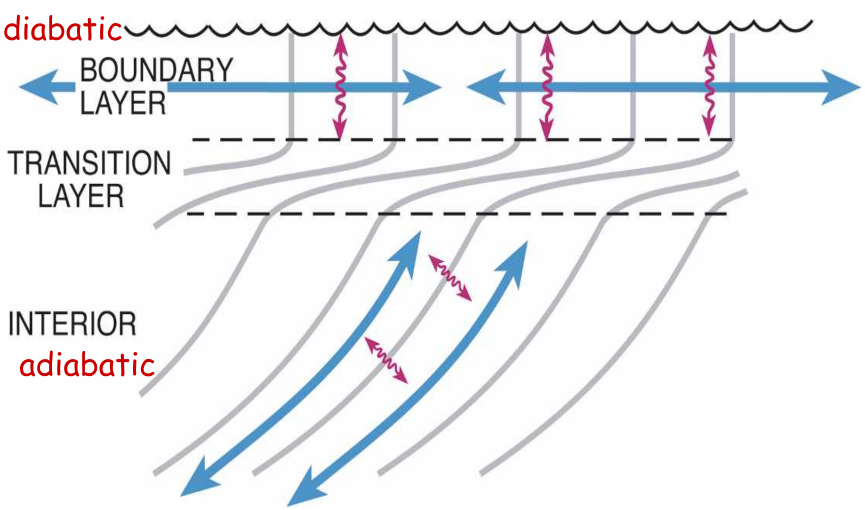
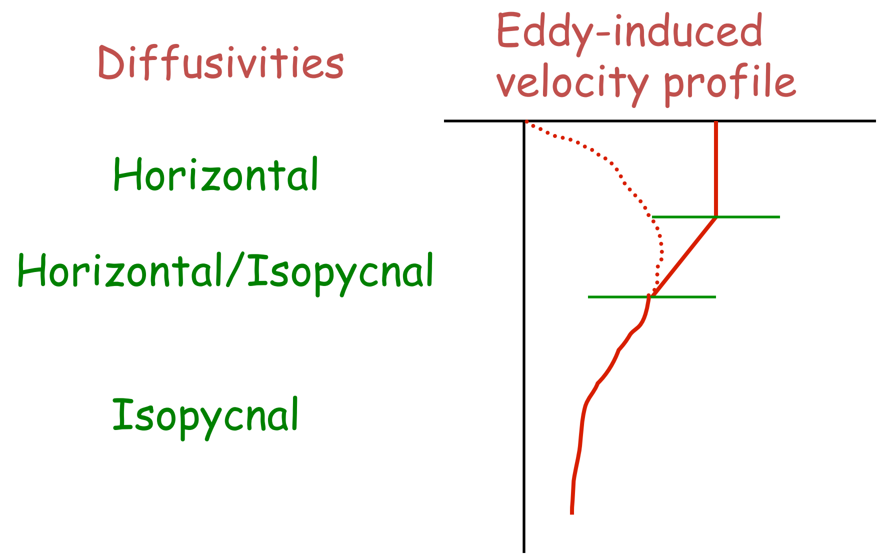


FIG. 2. A conceptual model of eddy fluxes in the upper ocean. Mesoscale eddy fluxes (blue arrows) act to both move isopycnal surfaces and stir materials along them in the oceanic *interior*, but the fluxes become parallel to the boundary and cross density surfaces within the *BL*. Microscale turbulent fluxes (red arrows) mix materials across isopycnal surfaces, weakly in the interior and strongly near the boundary. The interior and the BL regions are connected through a *transition layer* where the mesoscale fluxes rotate toward the boundary-parallel direction and develop a diabatic component.



The near-surface eddy flux scheme replaces the early approach of just tapering the GM coefficient to zero at the ocean surface.

Ferrari et al. (2008, J. Climate)

GM summary

Mimics effects of unresolved mesoscale eddies as the sum of

- diffusive mixing of tracers along isopycnals (Redi 1982),
- an additional advection of tracers by the eddy-induced velocity $\underline{u}^*$

Scheme is adiabatic and therefore valid for the ocean interior.

Acts to flatten isopycnals, thereby reducing potential energy.

Eliminates any need for horizontal diffusion in z-coordinate OGCMs

➔ eliminates Veronis effect.

Implementation of GM in ocean component was a major factor enabling stable coupled climate model simulations without "flux adjustments".