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COURSE ON CLIMATE VARIABILITY STUDIES IN THE OCEAN
"Tracing & Modelling the Ocean Variability"
16 - 27 June 2003

301/1507-14

Tracing and Modelling the Ocean Variability - I

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Tracing and modeling ocean variability

ICTP – June 2003

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$$\text{variability} \begin{cases} \text{Ocean's 'weather'} \\ \text{Geostrophic eddies} \end{cases}$$

Discuss:

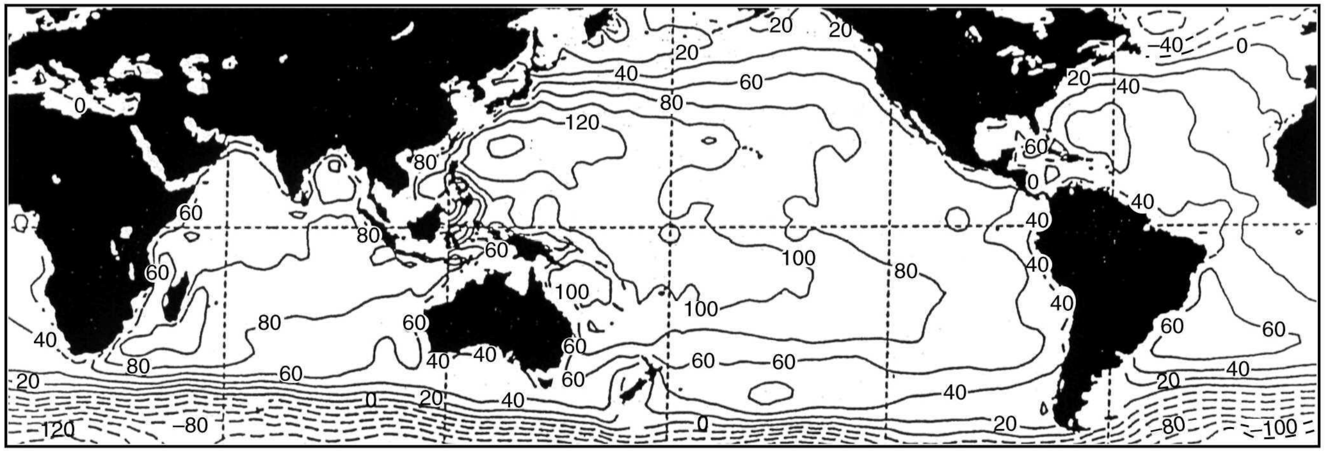
1. Eddies: are they important for climate?

2. Eddies and their role in setting structure of thermocline, mixing….

Sea surface height

mean

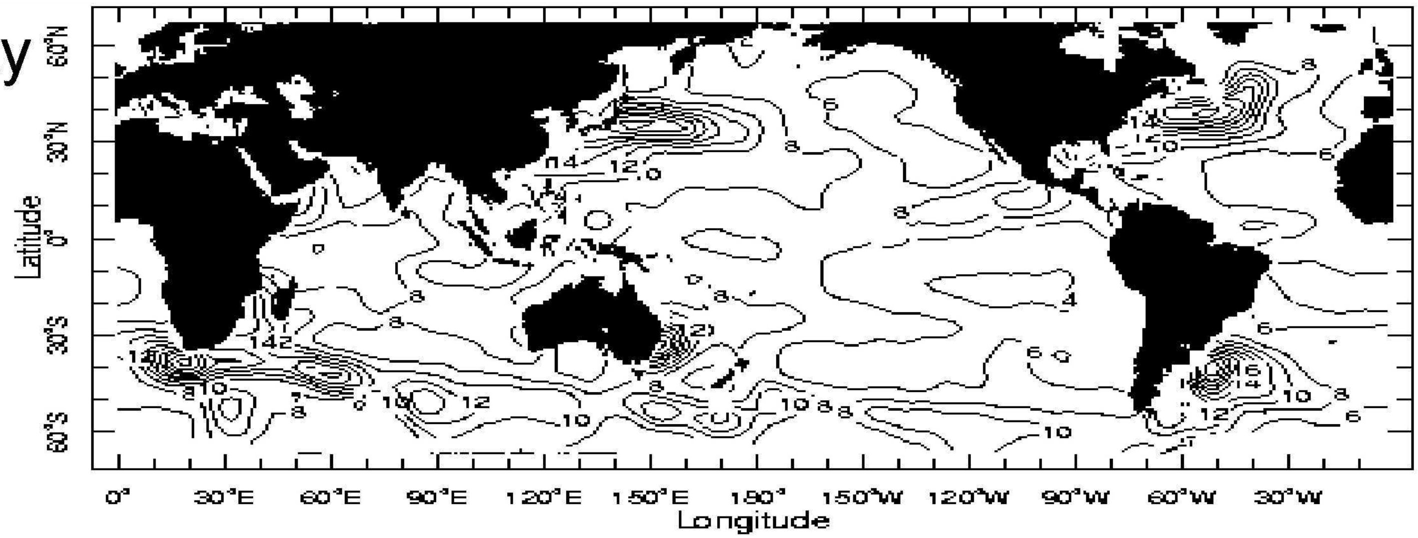
$\overline{h}$



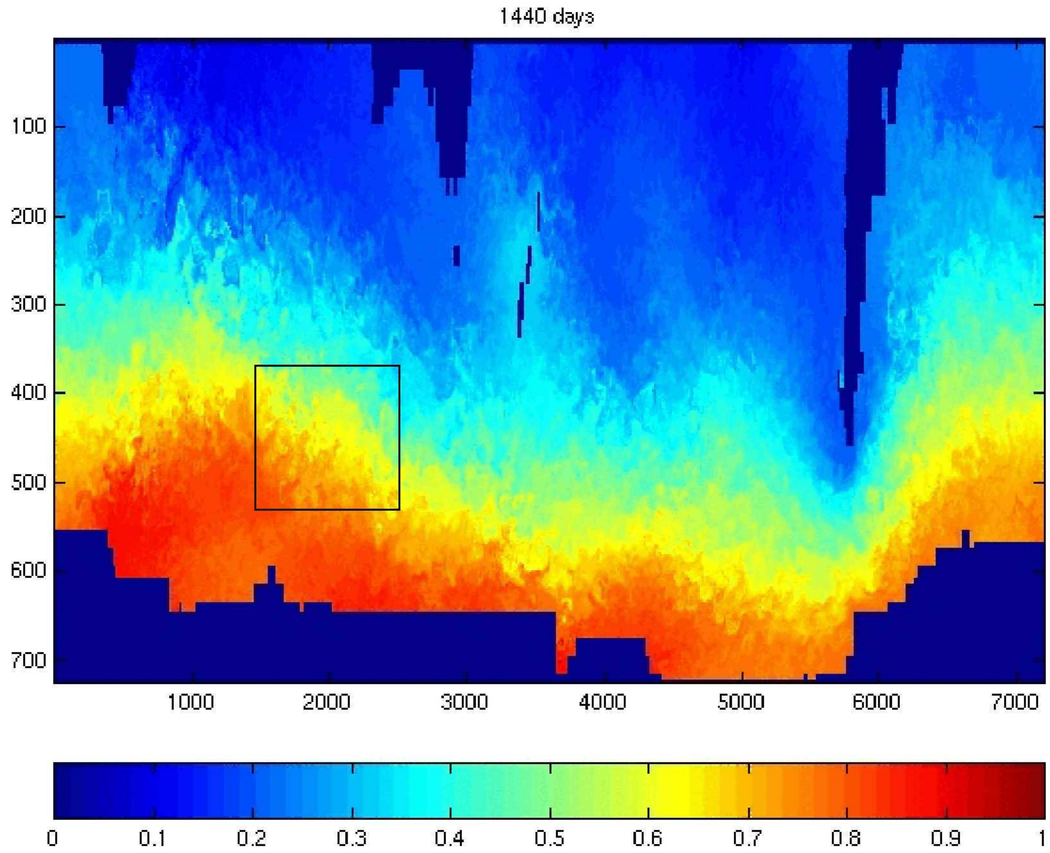
cm

variability

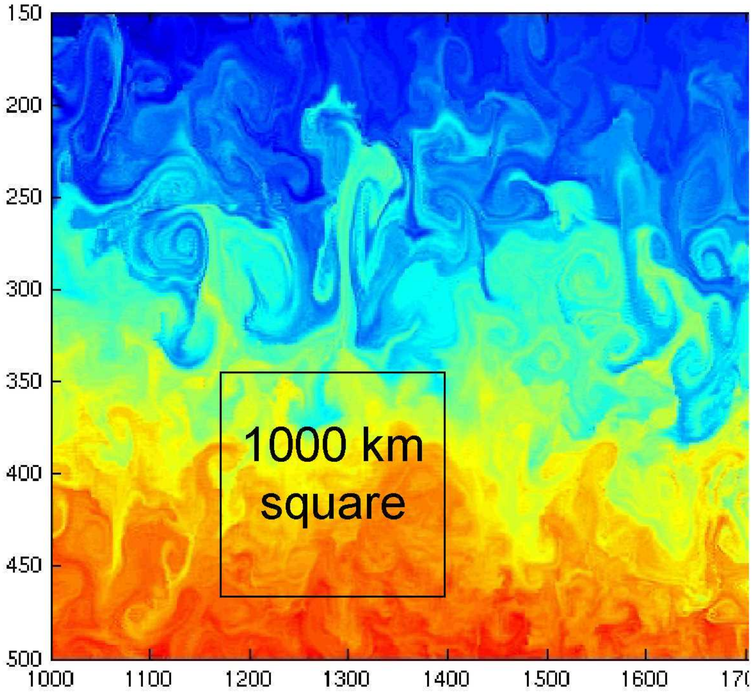
$\overline{h'^2}^{\frac{1}{2}}$



Evolution of passive tracer



Idealized tracer driven by surface currents in ACC



Ocean eddies are small and energetic

Difficult to observe and model

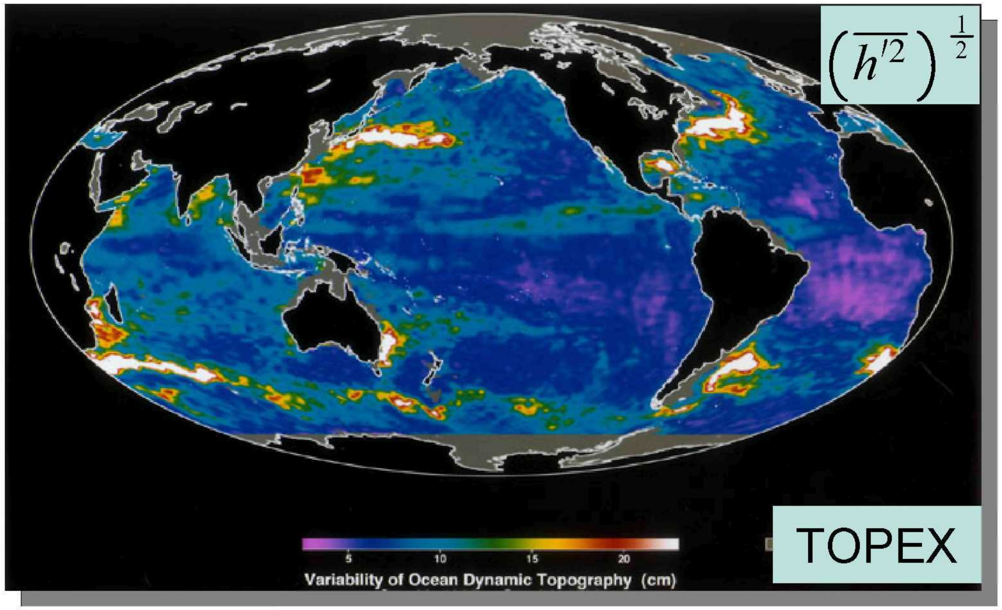
Are eddies important for climate?

Models suggest eddies play major role in tracer, momentum and vorticity budgets

e.g. vanishing of ‘Deacon Cell’

Doos and Webb

Danabasoglu, McWilliams

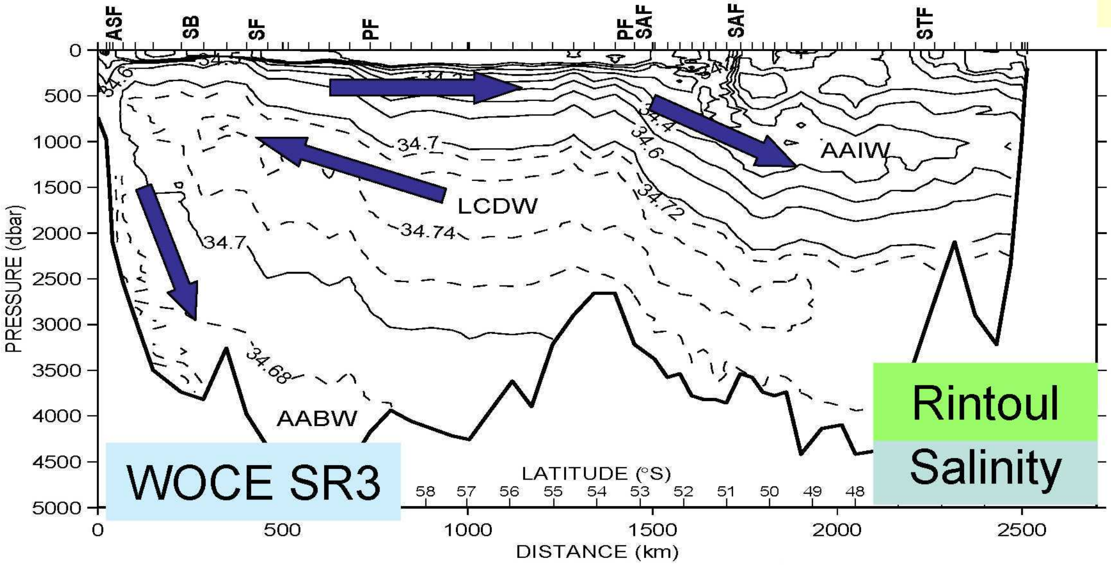


But

What can we infer from observations?

Tracer distributions

Circulation inferred from tracers is the ‘residual flow’



WOCE SR3

Rintoul

Salinity

$$v_R = \overline{v} + v^*$$

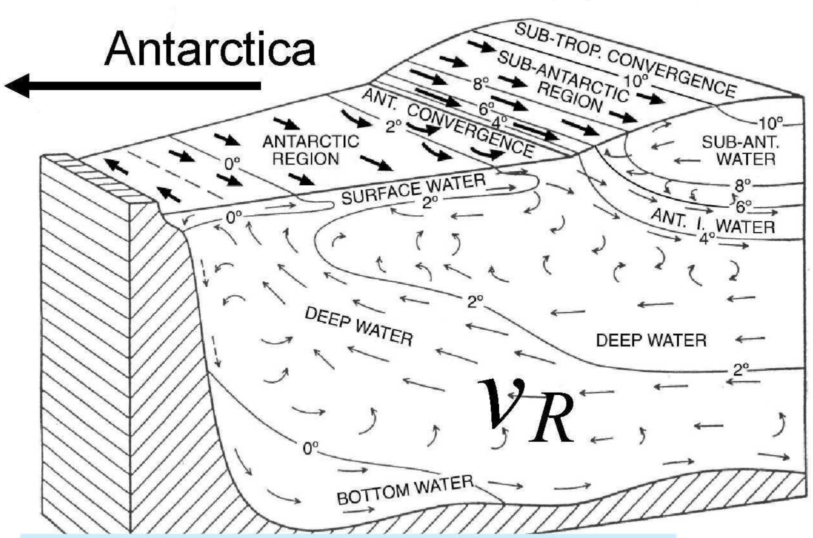
Residual mean

Eulerian mean

Bolus transport

$$\frac{\overline{vh}}{\overline{h}} \qquad \frac{\overline{v'h'}}{\overline{h}}$$

Rhines

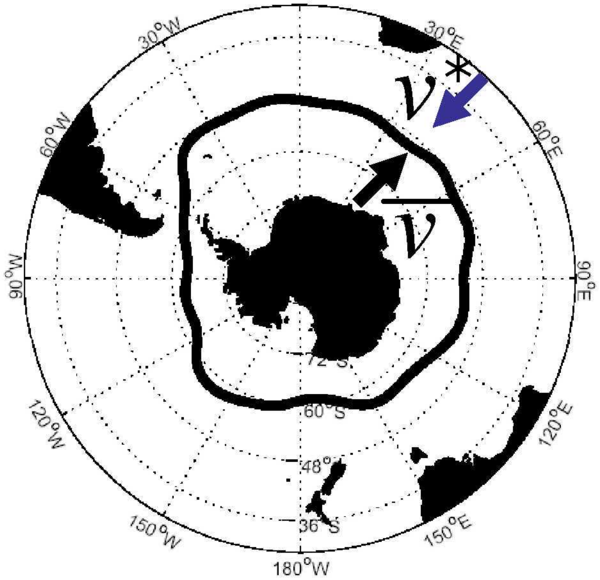


Schematic: Sverdrup

What is the role of eddy transfer in setting up v_R ?

Infer Eddy transport from observations

$$v_R = \overline{v} + v^*$$



1 Buoyancy transport

Diabatic
Deacon Cell

Surface buoyancy fluxes $\rightarrow v_R$

Winds $\rightarrow \overline{v}$

$$\Longrightarrow v^*$$

Speer, Rintoul &
Sloyan

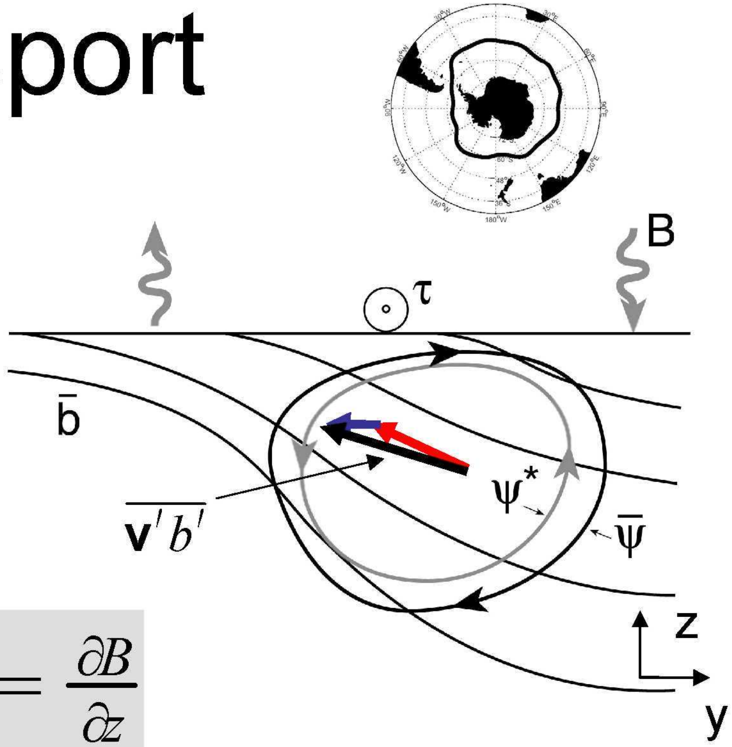
2 Eddy diffusivity, K

How can we estimate a K from observations?

3 Momentum balance

Compare eddy stress to wind stress

1 Buoyancy Transport



(i) Stream-wise average

$$2\text{d} \Longrightarrow \Psi$$

(ii) Rewrite Eulerian buoyancy budget:

$$\overline{v}\frac{\partial\overline{b}}{\partial y}+\overline{w}\frac{\partial\overline{b}}{\partial z}+\frac{\partial}{\partial y}\left(\overline{v'b'}\right)+\frac{\partial}{\partial z}\left(\overline{w'b'}\right)=\frac{\partial B}{\partial z}$$

Decompose eddy flux

$$(\overline{v'b'},\overline{w'b'}) = (\overline{w'b'}/s_\rho,\overline{w'b'}) + (\overline{v'b'}-\overline{w'b'}/s_\rho,0)$$

isopycnal horizontal

where $s_\rho = -\frac{\overline{b}_y}{\overline{b}_z}$ is the isopycnal slope

Residual-mean buoyancy budget

Held and Schneider, 1999

isopycnal

$$\nabla \cdot (\overline{w'b'}/s_\rho, \overline{w'b'}) = \overline{v}^{*}\,\overline{b}_y + \overline{w}^{*}\,\overline{b}_z$$

where $\Psi^* = -\frac{\overline{w'b'}}{\overline{b}_y}$

leftover

$$\nabla \cdot (\overline{v'b'} - \overline{w'b'}/s_\rho, 0) = \frac{\partial}{\partial y}\left[(1-\mu)\overline{v'b'}\right]$$

$$(\overline{v} + v^*)\frac{\partial \overline{b}}{\partial y} + (\overline{w} + w^*)\frac{\partial \overline{b}}{\partial z} = \frac{\partial \overline{B}}{\partial z}$$

+ dia

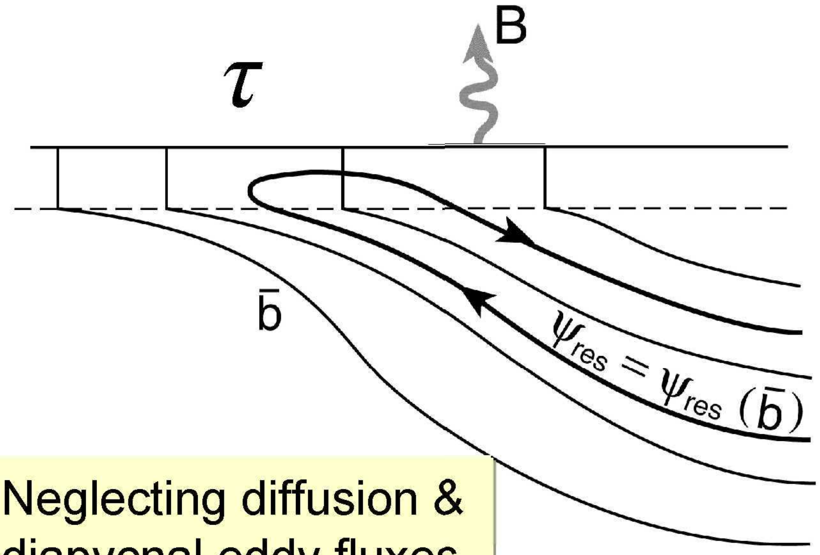
where $$\Psi^* = -\frac{\overline{w'b'}}{\overline{b}_y} = \frac{\overline{v'b'}}{\overline{b}_z}$$

If eddies are ‘adiabatic’

Residual mean theory

Andrews and McIntyre
Held & Schneider
Gent and McWilliams

Cross-stream transport inferred from air-sea fluxes



Neglecting diffusion & diapycnal eddy fluxes

$$\Psi_R \frac{\partial \overline{b}}{\partial y} = B_s$$

from air-sea buoyancy fluxes

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

from winds

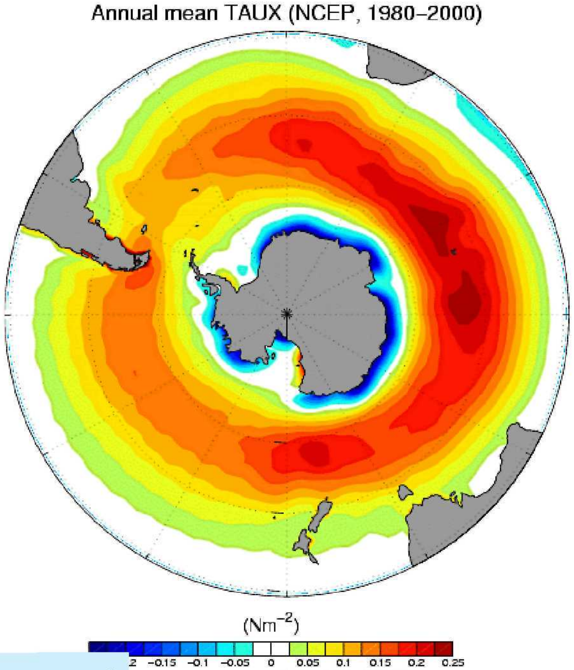
Cross-stream transport implied by the wind is not the same as that implied by air-sea buoyancy fluxes

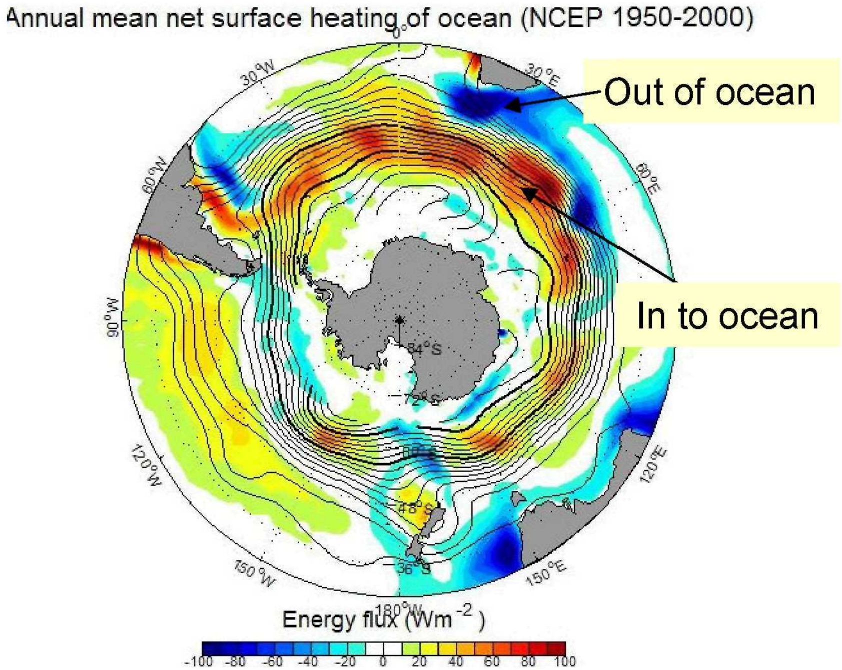
David Marshall
Speer, Rintoul & Sloyan

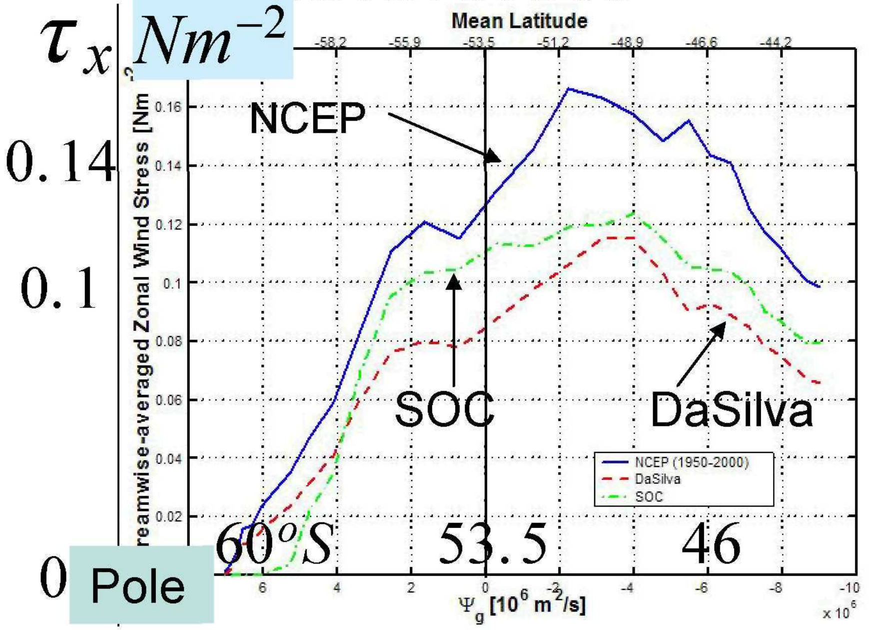
Wind Stress

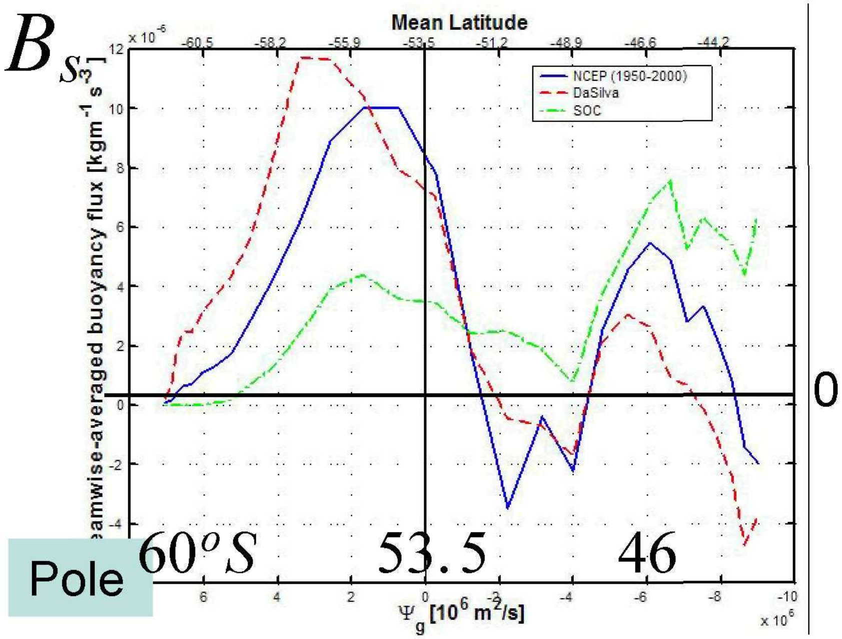
Air-sea buoyancy flux

NCEP

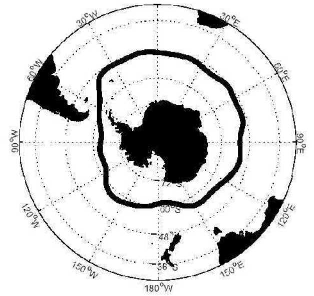

Annual mean TAUX (NCEP, 1980–2000)

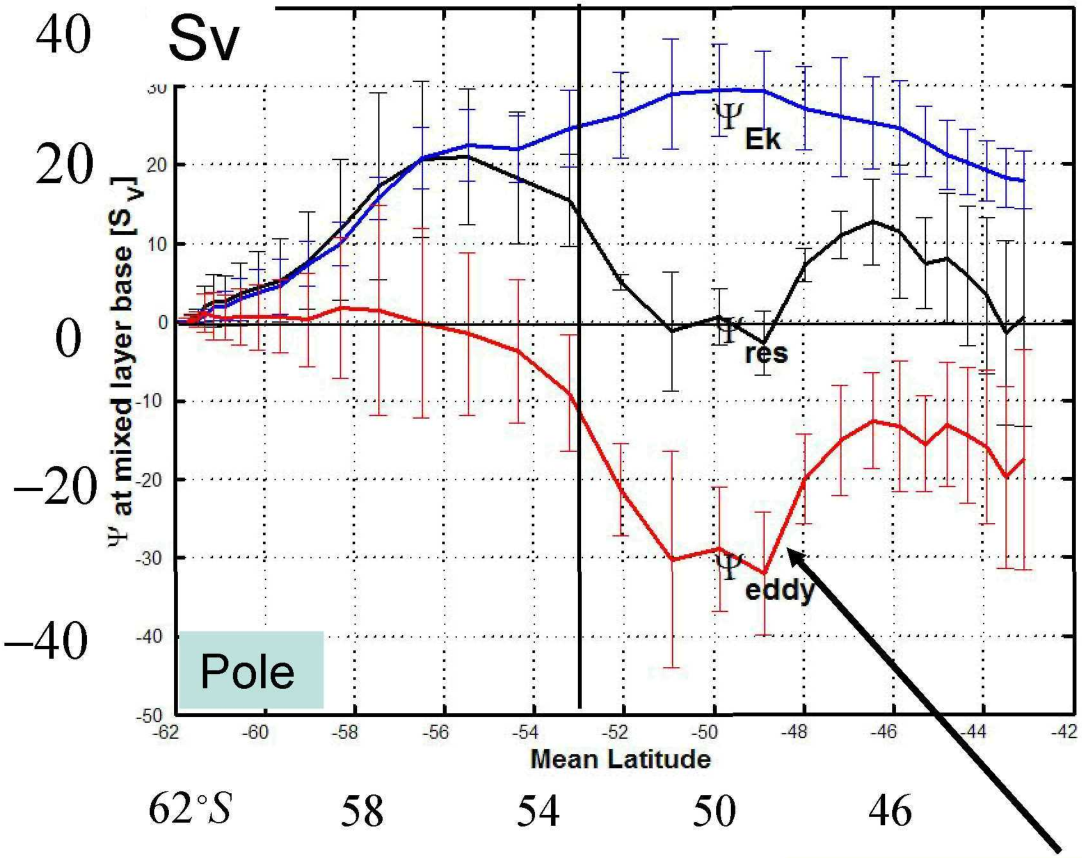

Annual mean net surface heating of ocean (NCEP 1950-2000)





Cross-Stream transports





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

$$\Psi_R = \frac{B_s}{\frac{\partial \bar{b}}{\partial y}}$$ Buoyancy fluxes

$$\Psi_R - \overline{\Psi}$$ Inferred 'eddy induced' transport

Taka Ito

Can eddies achieve such a large transport?

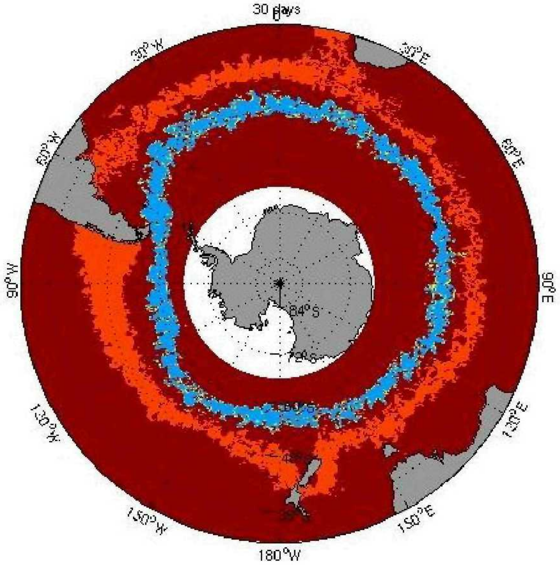
WOCE SAC climatology

Gouretski and Jancke

2 Eddy transfer

Can we characterize eddies with a diffusivity, K?

If so, how large is the K?



Use ‘idealized tracer’ driven by observed surface geostrophic flow to yield K

altimetry

‘idealized tracer’ ⇝ Device to yield diffusivity from observations

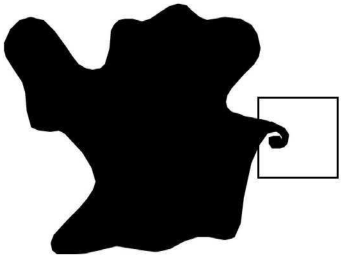
Advection-diffusion in 2-d

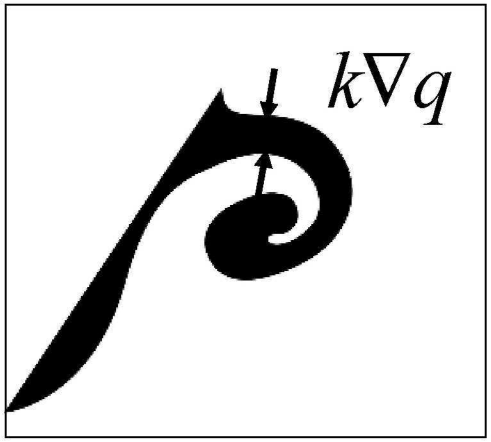
Nakamura
Haynes and Shuckburgh

$$\boxed{\frac{\partial q}{\partial t} + \mathbf{v} \cdot \nabla q = k\nabla^2 q}$$

'small-scale' diffusion

Large-scale **v** drives small scale mixing $k\nabla q$





In 'area' coordinates problem can be rephrased as:

$$\frac{\partial q}{\partial t} = \frac{\partial}{\partial A}\left(K\frac{\partial q}{\partial A}\right)$$

where $$K = \frac{\chi}{\left(\frac{\partial q}{\partial A}\right)^2}$$

$$\chi = k\frac{\partial}{\partial A}\int_{A(q,t)} |\nabla q|^2 dA$$

'dissipation rate' of q

Theory 'intersects' with observations very nicely

Drive tracer with altimetric data

$$\frac{\partial q}{\partial t} + \mathbf{v} \cdot \nabla q = k \nabla^2 q$$

Topex data
every 10 days

$$k = 100 m^2 s^{-1}$$

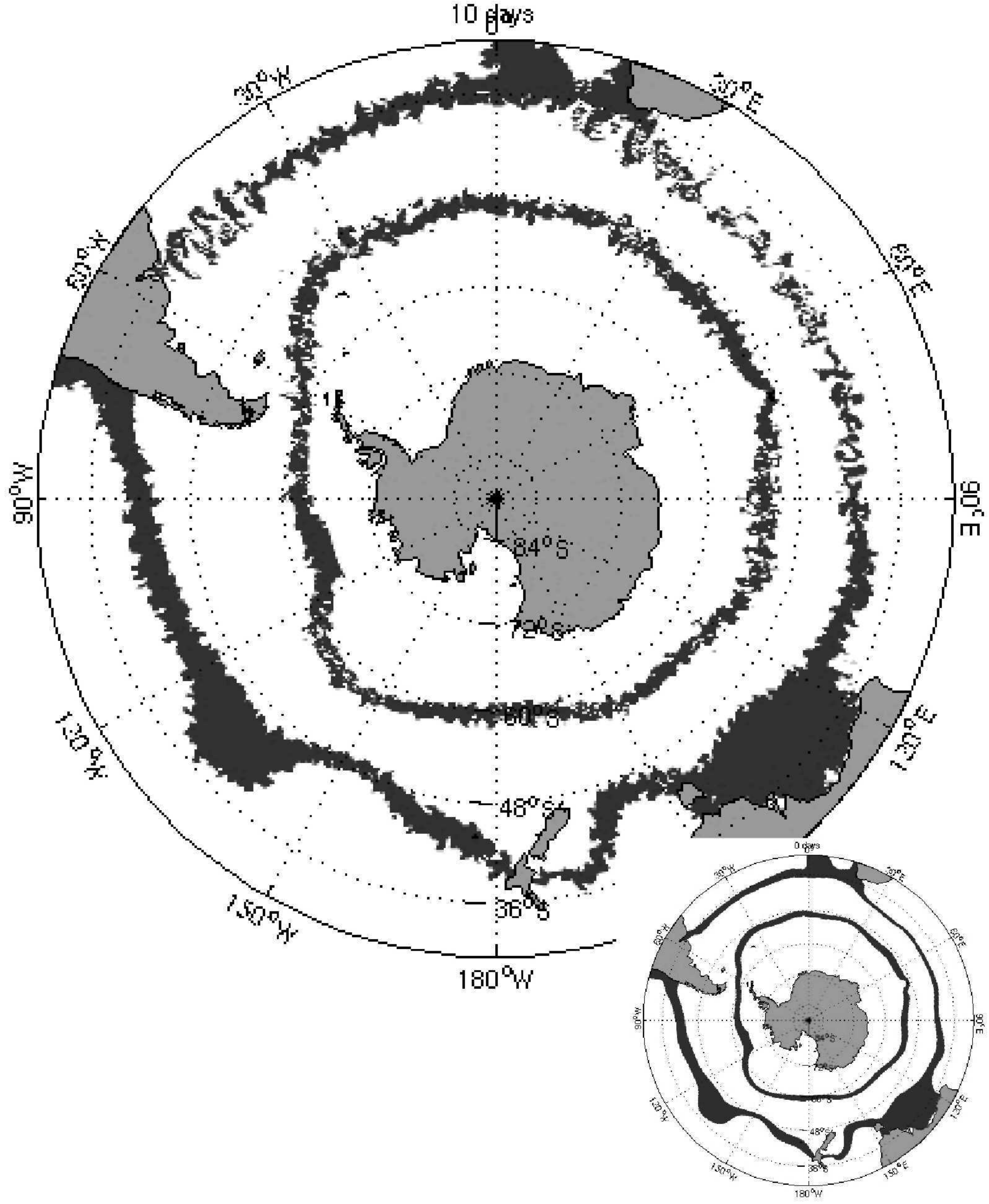
Resolution $\frac{1}{6}^{\circ} \times \frac{1}{6}^{\circ}$

Assumption: large-scale K is independent of small-scale k

Planning to repeat at

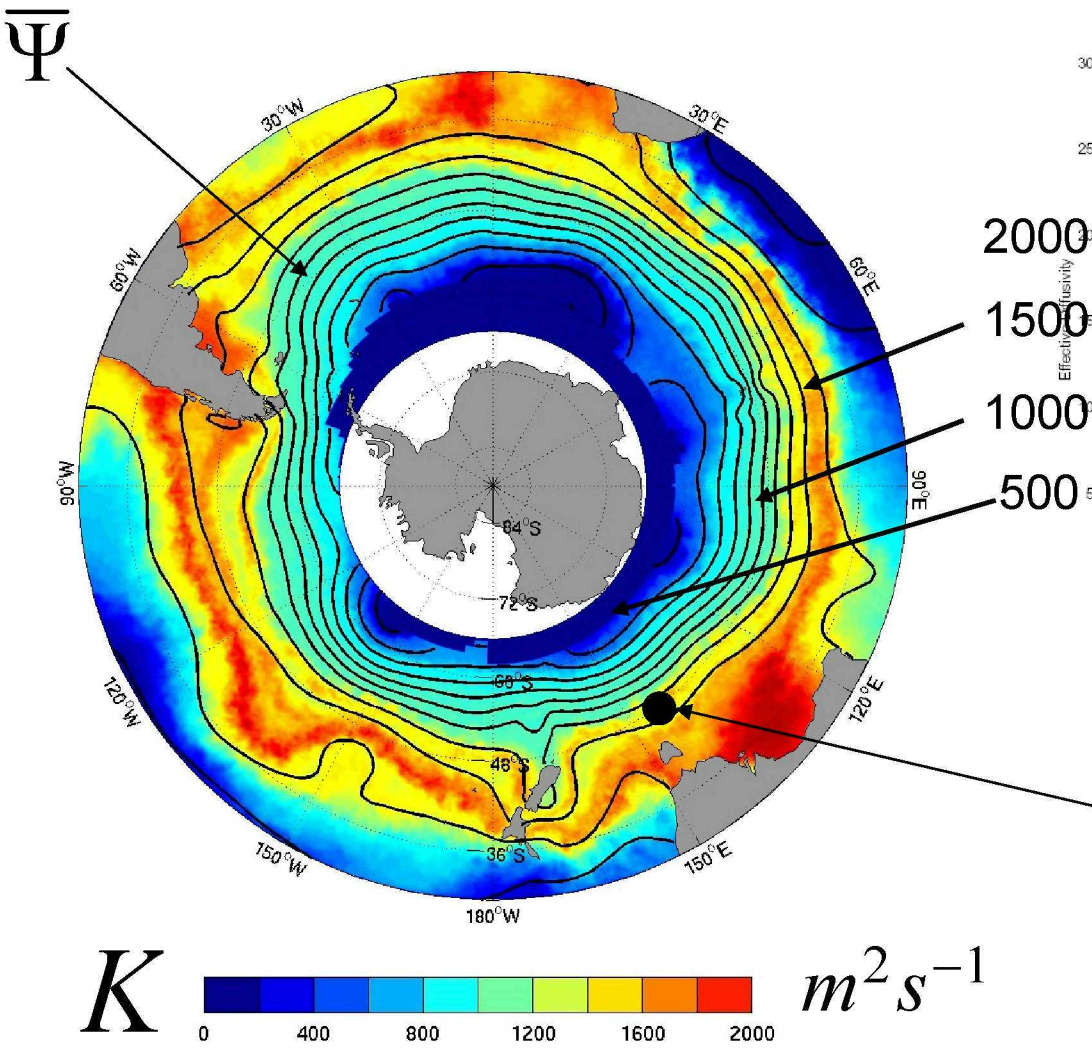
$\frac{1}{30}^{\circ} \times \frac{1}{30}^{\circ}$ with $k = 10 m^2 s^{-1}$

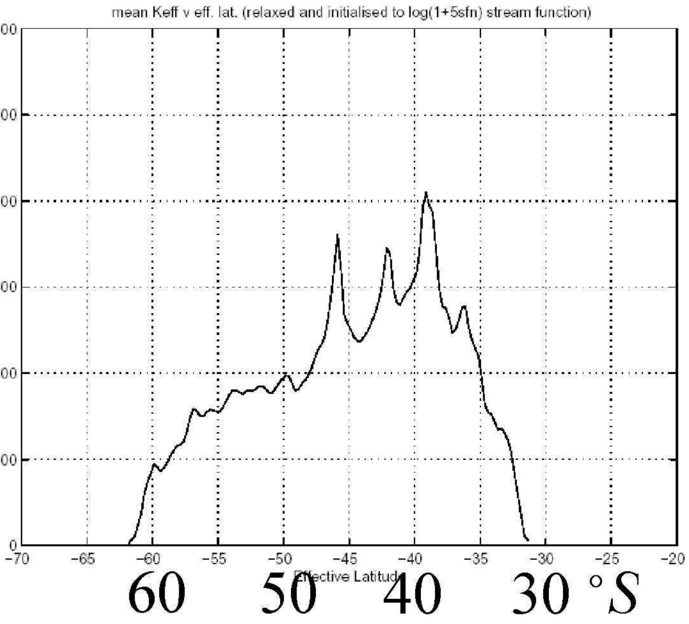
Emily Shuckburgh (DAMPT)
Helen Jones (MIT)

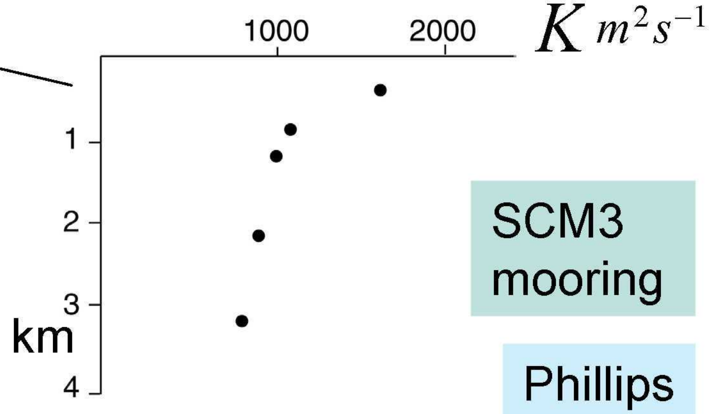


Eddy diffusivity

Near surface K





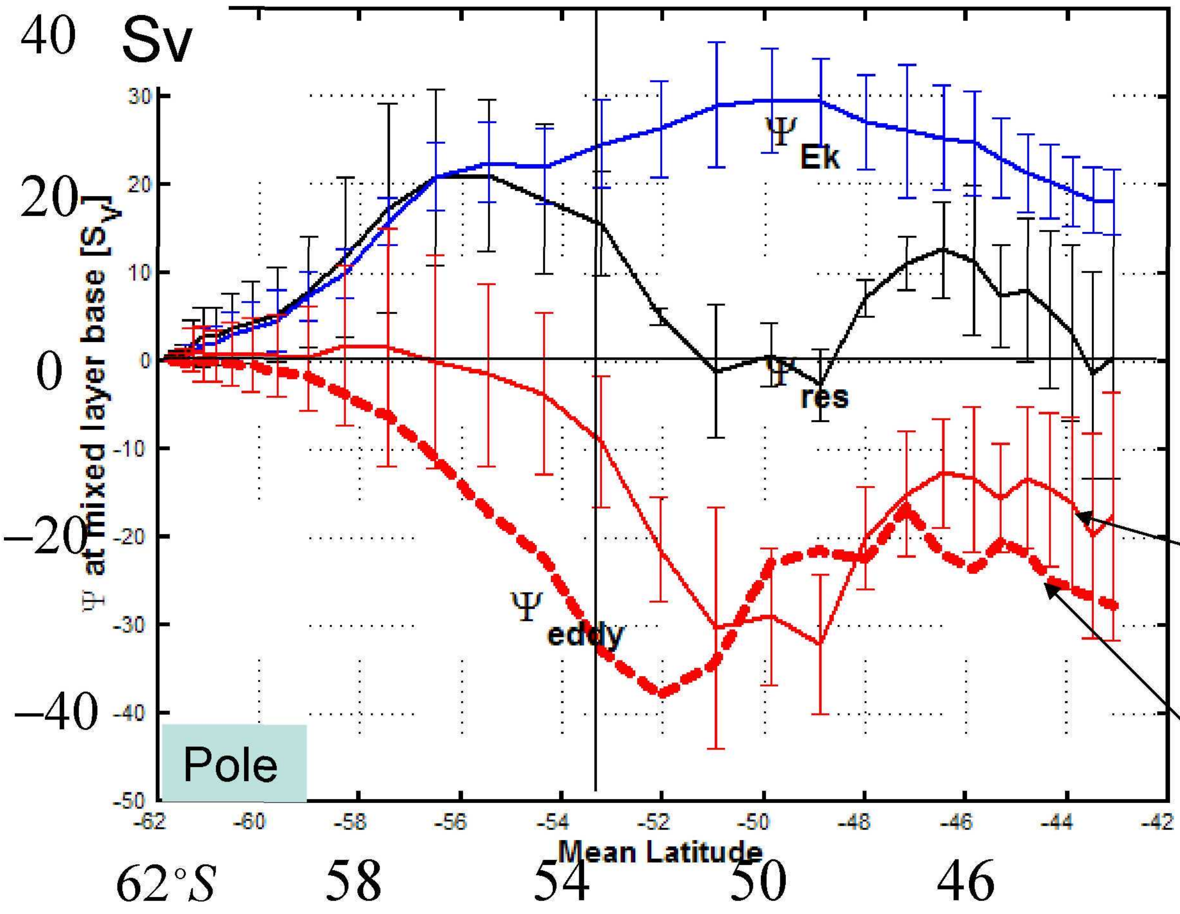


SCM3 mooring

Phillips

Implications for parameterization

Deducing Ψ^* from K



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

$$\Psi_R = \frac{B_s}{\frac{\partial \overline{b}}{\partial y}}$$

$$\Psi_R - \overline{\Psi}$$

$$\Psi^* = \frac{\overline{v'b'}}{\overline{b}_z}$$

$$= K s_\rho$$

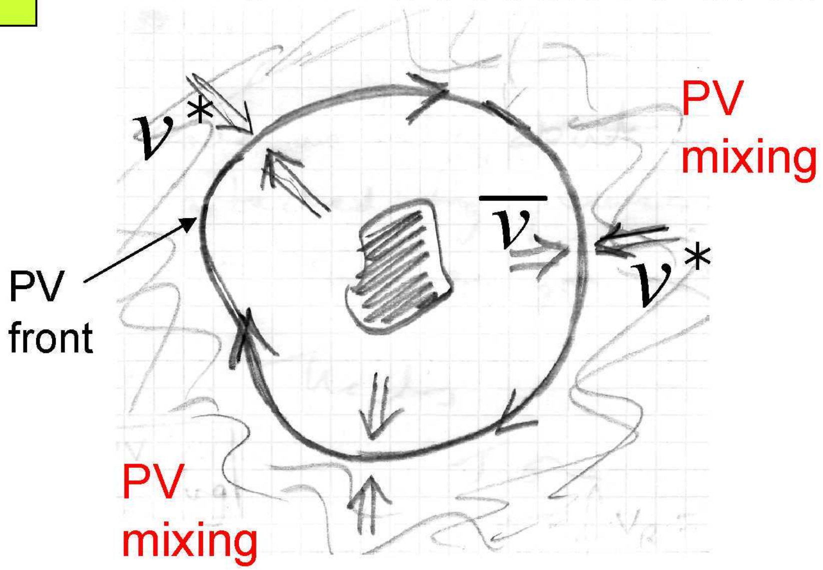
isopycnal slope at mixed layer base

$$s_\rho = -\frac{\overline{b}_y}{\overline{b}_z}$$

WOCE SAC climatology

Gouretski and Jancke

3 Momentum balance



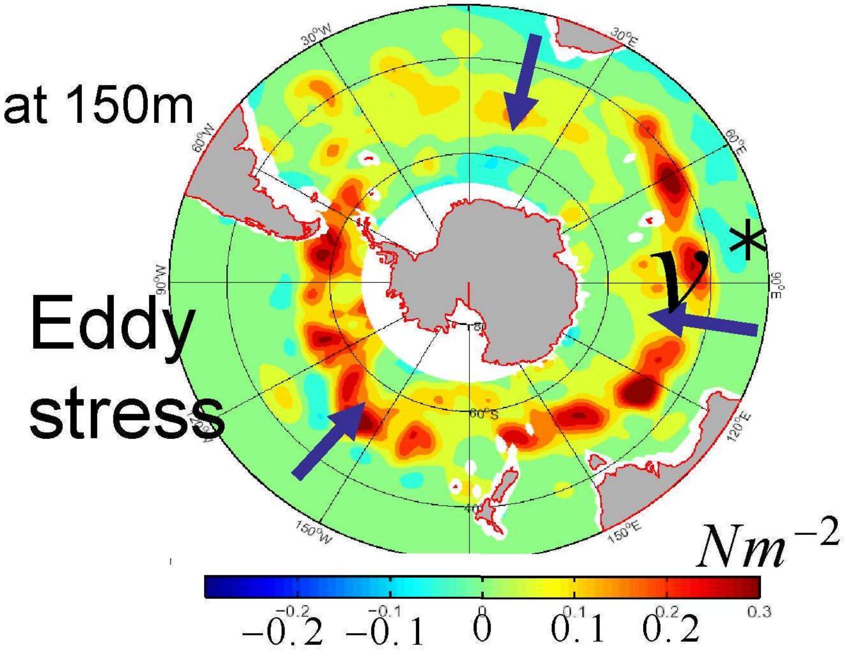
$$-fv_R = F + \overline{v'q'}$$

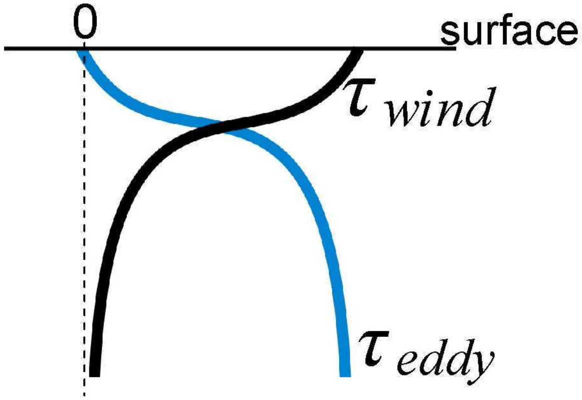
$$\frac{\partial \tau_{wind}}{\partial z}$$

Eddy PV flux

$$\frac{\partial \tau_{Eddy}}{\partial z}$$

$$\tau_{Eddy} = f\Psi^* = fKs_\rho$$





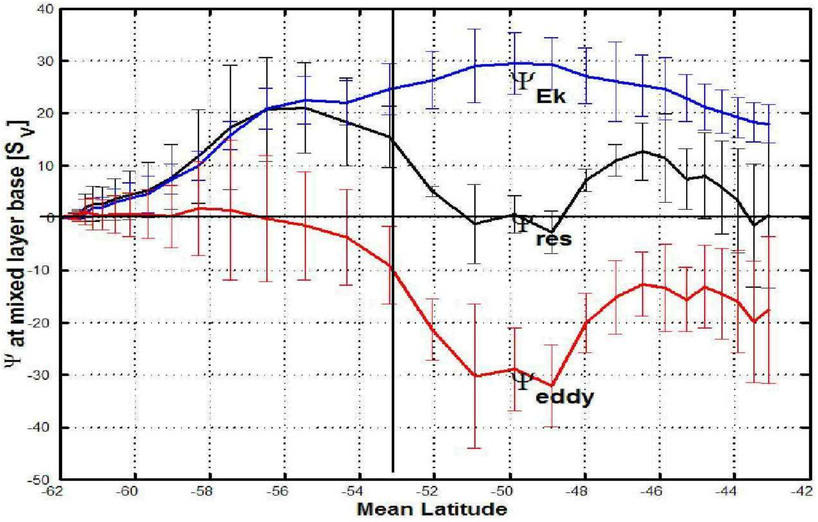
if $\Psi_R = 0$

$\tau_{eddy} = \tau_{wind}$

Johnson and Bryden

Mapping down to depth

Karsten, Marshall

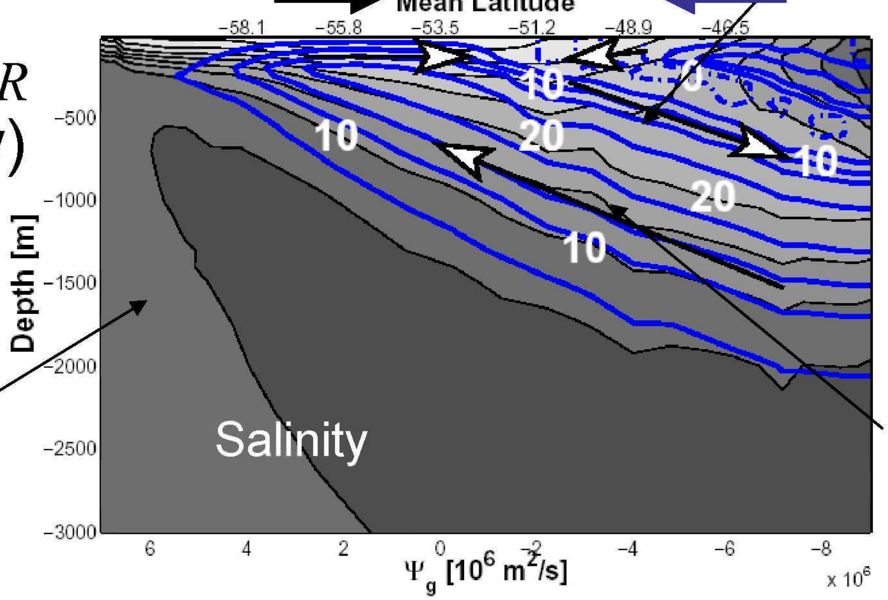


$\overline{v}$

v^*

Intermediate Water

Ψ_R (Sv)



Upper Circumpolar Deep Water

No bottom water

Conclusions

1 Transport implied by τ and B_s are very different

$\Longrightarrow$ Ψ^* must be large and oppose $\overline{\Psi}$

30 Sv

2 Estimated K from altimetry

2-d mixing of idealized tracer

yields near-surface $K \sim$ 500 to 2000 $m^2 s^{-1}$

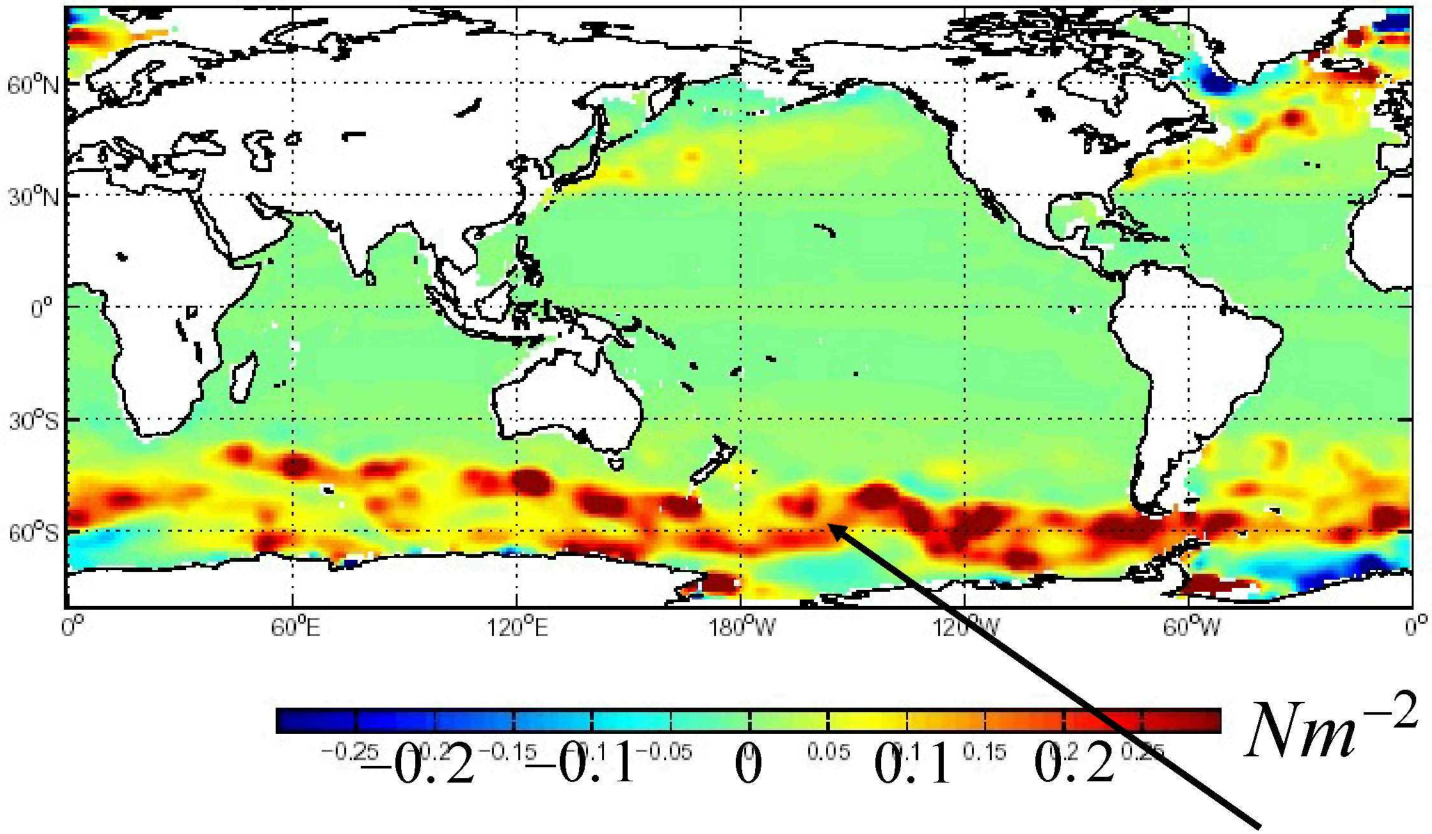
and consistent pattern of Ψ^*

3 Eddy stresses are as large as wind stress

4 Eddies are hugely important in ACC and elsewhere…..

Estimate of global eddy stress

At 300m



Parameterized in our global climate models!

assuming K = 1000$m^2 s^{-1}$

using WOCE SAC climatology

Observations

Bryden,1979
Gille,1997
Phillips and Rintoul, 2000
Speer, Rintoul and Sloyan, 2000
Karsten and Marshall, 2002
Gille, 2002

Modeling

McWilliams, Holland and Chow, 1978
Semtner, Chervin, 1992
Döös and Webb 1994
Danabasoglu, McWilliams 1994
Ivchenko, Richards, Stevens 1996

Theory

Munk and Palmén, 1951
deSzoeke and Levine, 1981
Marshall, 1982
Johnson and Bryden, 1989
Gent and McWilliams, 1990
Marshall, Olbers, et al, 1993
Nakamura, 1996
David Marshall, 1997
McDougall,1998
Haynes and Shuckburgh, 2001
Olbers and Ivchenko, 2002
Marshall and Radko, 2002

Reviews

Nowlin and Klink, 1986
Rintoul, Hughes and Olbers, 200