Outline

I General circulation

- ventilation
- subduction
- Large-scale Potential Vorticity
- transient tracers ^{3}He/Tritium transport

II Parameterization of geostrophic eddies

III geostrophic turbulence and coherent vortices

The general circulation of the oceans is related the distribution of potential vorticity (Rhines 1986, Pedlosky 1996)

$$Q = \frac{(2\vec{\Omega} + \vec{\zeta}) \cdot \nabla \rho}{\rho}$$

$\vec{\Omega}$ Earth rotation vector
$\vec{\zeta} = \nabla \times \vec{u}$ relative vorticity
ρ potential density

Ocean circulation is a blend of

- lateral gyrelike circulation driven by wind
- great convection cells that transport heat and salinity contrasts across the Earth (thermohaline component).

Planetary fluid dynamics is dominated by the conservation of Potential Vorticity (PV)

$$\frac{D}{Dt}Q = F - D$$

F external forcing
D dissipation (subgrid scale dynamics).
For planetary geostrophic flows, relative vorticity is negligible

$$Q = \frac{f}{\rho}\frac{\partial \rho}{\partial z}$$

where $f = 2\Omega\ sin\lambda$, with λ latitude.

For steady large-scale flows, PV conservation yields (for $\beta \equiv \frac{df}{dy}$),

$$\beta v = f\frac{\partial w}{\partial z}$$

which vertical integral leads to Sverdrup's (1947) classic prediction that, regardless of the structure of the density field and baroclinic circulation, the barotropic (depth-averaged) flow obeys

$$\beta V \equiv \beta \int_{-H}^{-H_{Ek}} v dz = f[w]|_{-H}^{-H_{Ek}} = f\, w_{Ek}$$

in the case where the flow vanishes at depth (or level sea-floor), with H_{Ek} and w_{Ek} are the depth and the vertical velocity at the base of the surface Ekman layer

$$w_{Ek} \; = \; -\nabla\cdot\vec{U_{Ek}} \; = \; \vec{k}\cdot(\nabla\times\frac{\vec{\tau}}{\rho_0 f}) \; = \; \frac{1}{f}\vec{k}\cdot(\nabla\times\frac{\vec{\tau}}{\rho_0}) \; + \; \frac{\beta}{f^2}\frac{\tau^x}{\rho_0}$$

with $\vec{\tau}$ wind-stress.

$$\Longrightarrow V = V_{Ek} + V_G$$

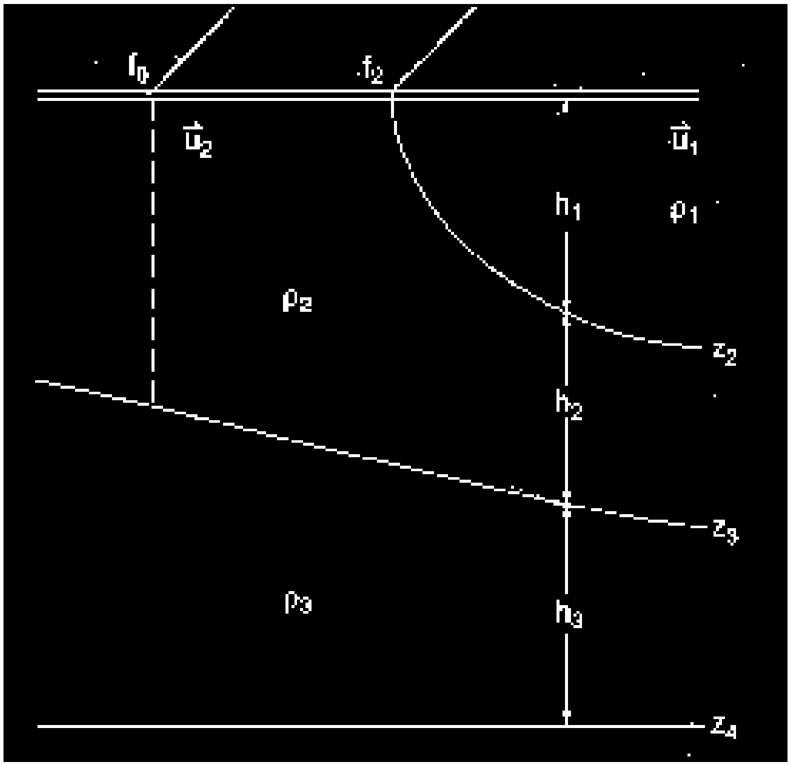
the total meridional transport is the sum of

V_{Ek} Ekman transport

V_g geostrophic transport

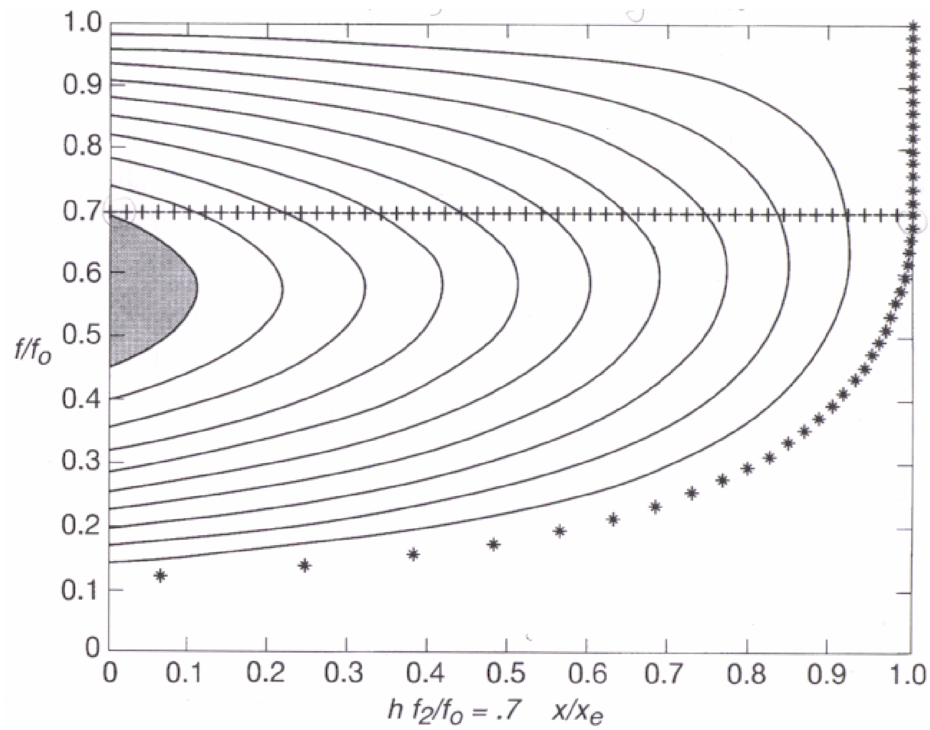
This relation encapsulates the way $\vec{\tau}$ at the sea-surface can force gyrelike circulations in the ocean.

Ventilated wind-driven gyre



Luyten, Pedlosky and Stommel (1983) (LPS model of ventilated thermocline

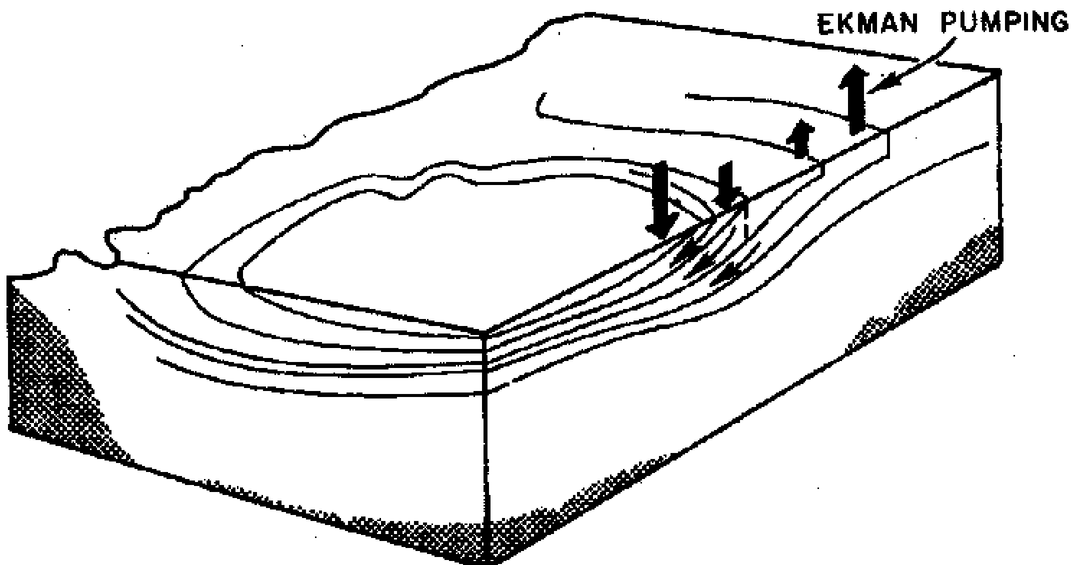
- 2 active layers overlying a third deep layer
- flow driven by wind and surface boundary conditions ("outcropped" density interfaces)
- calculate the flow as a downstream integration from the sea-surface forcing along characteristics
- domain bounded to the East by rigid N-S wall and Western Boundary Layer to the West (close mass budget)
- key assumptions:
  stationary flows
  Sverdrup dynamics
  Potential Vorticity and Bernouilli function conservation for layers that have subducted



LPS Height of the lower interface $D(x, y)$ for the subtropical gyre

- $D(x, y)$ indicates the flow in layer 2, which follows geostrophic contours south of $f = f_2$, where it is unforced.
- The region of southward-flowing circulation departs the Eastern boundary, which leaves a ”shadow zone” with no flow at all.
- The Western region that is not penetrated by characteristics from boundary conditions, isolates a pool of fluid that corresponds to a ”homogenized pool in potential vorticity” (Rhines and Young, 1982). Homogenization of a conservative advective scalar within a closed gyre occurs also in magnetohydrodynamics for instance.

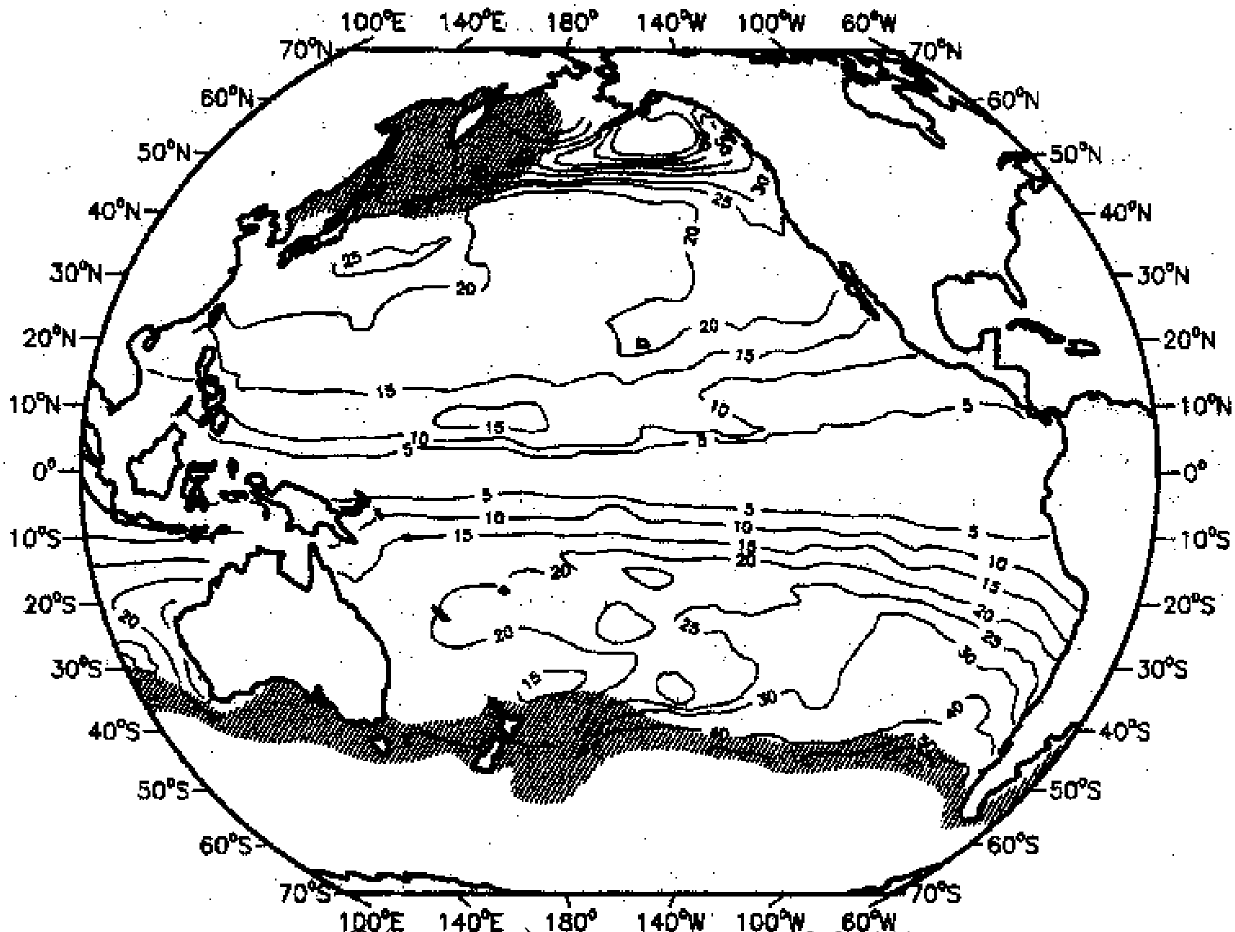
Summary



Schematic of surfaces of constant potential density, the Ekman vertical velocity, and the circulation of a subtropical gyre

- the fluid pumped from the upper boundary layer joins a greater stream recirculating from the western boundary current, or joining the equatorial region.
- the interior potential vorticity is determined
  - by the conditions where the fluid encounters the upper boundary layer,
  - by the dissipation balance integrated about the entire gyre

Potential Vorticity distribution



Potential vorticity $Q(x, y, \rho)$ on the $\sigma_\theta = 26.15$ potential density surface

$Q(x, y, \rho)$ two-dimensional surface at each density horizon is the principal field variable for the general circulation

$\sigma_\theta \equiv 1000(\rho - 1)$ potential density

The $\sigma_\theta = 26.15$ surface (typically 500 m-depth) is dominated by nearly uniform potential vorticity across the whole North Pacific, with inward advection of large q in the east.

The Atlantic ocean is markedly different with an intense buoyancy-driven ventilation (see below), as well as larger thermohaline component.

Flux form of the Potential vorticity equation

Marshall and Nurser (1992) have applied a formalism of generalized potential vorticity flux form to map the creation and transport of PV along isopycnic layers even in the presence of diabatic and mechanical forcing.

$$\frac{\partial}{\partial t}(\rho Q) + \nabla \cdot \vec{J} = 0$$
$$\vec{J} = \rho\, Q\, \vec{u} + \vec{N_Q}$$

$\vec{J}$ is a generalized flux of potential vorticity comprising the advective flux $\rho\, Q\, \vec{u}$ and the nonadvective flux $\vec{N_Q}$ defined by

$$\vec{N_Q} = \frac{1}{g} B\, \vec{\omega} + \vec{F} \times \nabla\sigma$$

$$\vec{\omega} = 2\vec{\Omega} + \nabla \times \vec{u}$$

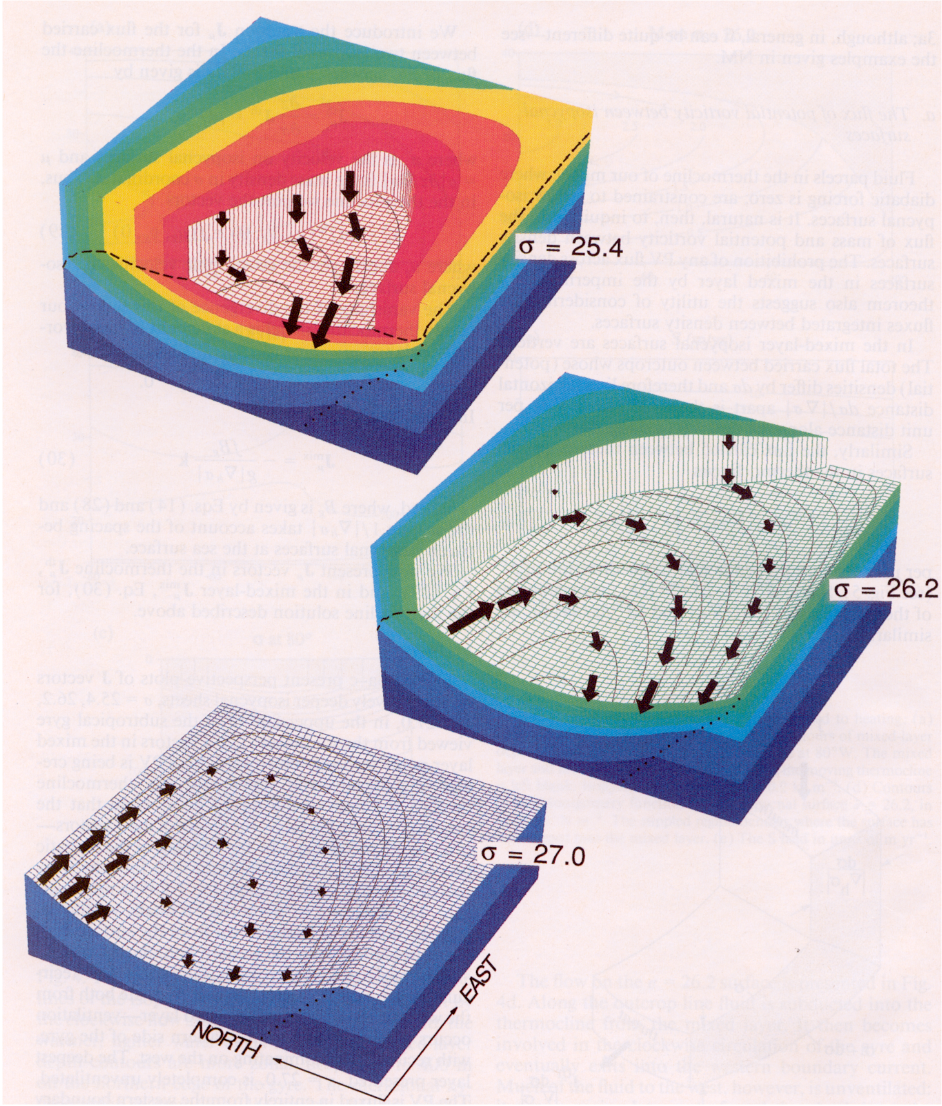
is the full 3D absolute vorticity, $\vec{\Omega}$ is the angular velocity of the Earth;

$B = -g\dfrac{d\sigma}{dt}$ buoyancy forcing, σ potential density

$\vec{F}$ nonconservative body force

$Q = -\dfrac{1}{\rho}\vec{\omega} \cdot \nabla\sigma$ potential vorticity

Haynes and McIntyre (1987) have shown that σ surfaces are impermeable to PV even in the presence of buoyancy forcing and frictional forces. This framework enables Marshall and Nurser to formulate the ventilation of the thermocline from the surface mixed layer and diagnose the potential vorticity flux at the ocean surface and into the thermocline.



- the $\vec{J}$-vectors in the mixed layer point vertically downward: PV is created diabatically and flows into the thermocline through the base of the ML. The shallow $\sigma = 25.4$ is totally ventilated with no recirculation from the western boundary.

- on the $\sigma = 26.2$ surface both ventilation and recirculation are evident. $\vec{J}$ vectors emanate both from the western boundary and the mixed layer (predominantly on the eastern side of the gyre), with recirculation dominating on the west.

- the deepest layer $\sigma = 27$ is completely unventilated. The PV is fluxed entirely from the western boundary layer to where it returns after its circuit around the gyre

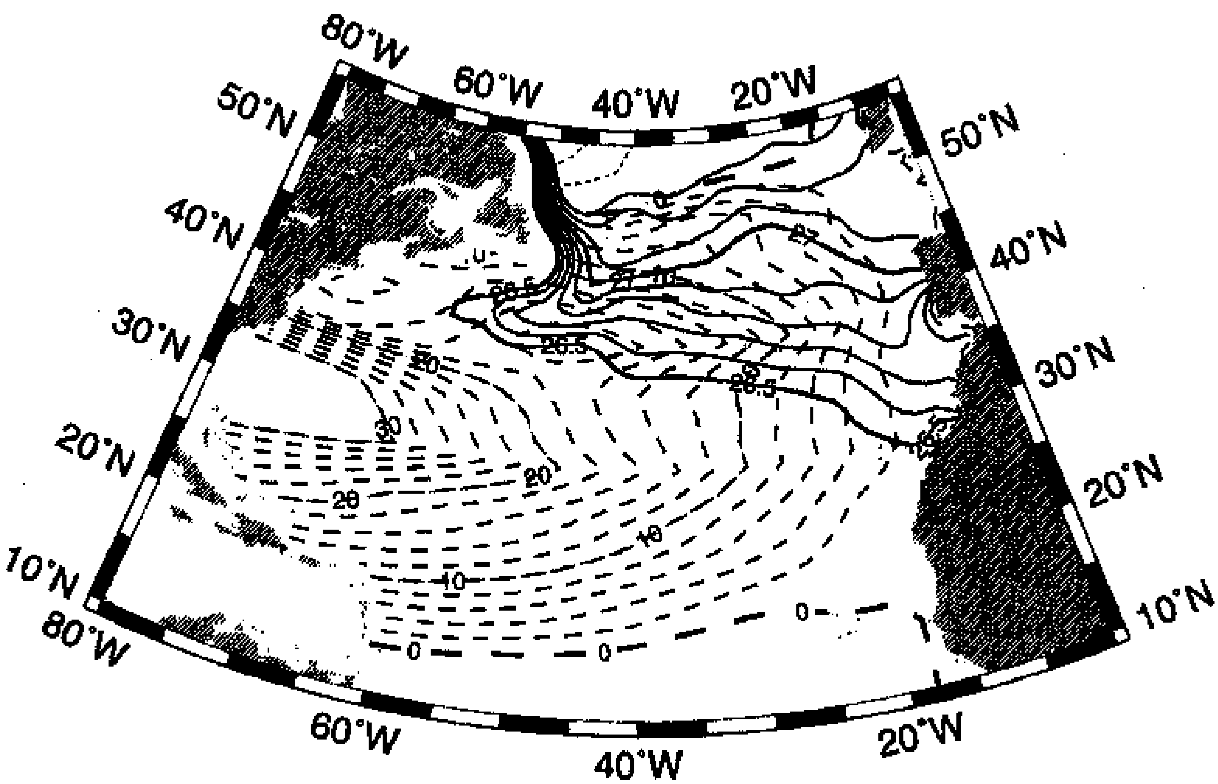
Summary

- the $\vec{J}$ vectors emphasize the role of buoyancy, as opposed to the sole mechanical forcing of the mixed layer in controlling subduction
- $\vec{J}$ is continuous at the base of the surface mixed layer, merely changing from a diapycnal nonadvective flux to an adiabatic, advective flux

Transient tracers ^{3}He/Tritium transport

Ventilation theory enables to rationalize the 3D "steady" largest scales of transport in the ocean that are compatible with the large-scale distribution of PV.

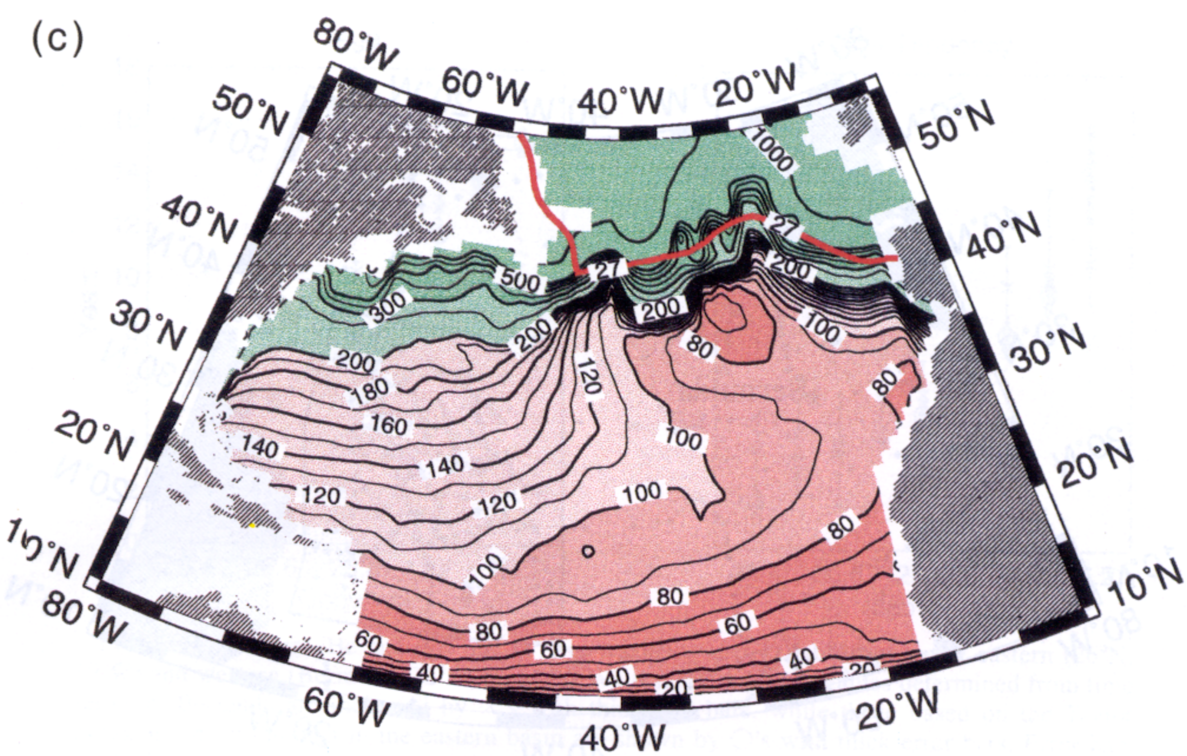
However, other tracer fields present evidence that they are not advected by the large-scale velocity. This is the case for instance of the distribution of transient tracers such as ^{3}He/Tritium distribution in the North Atlantic (Robbins et al., 2000).



Annual mean Sverdrup transport (dashed) based on Hellerman and Rosenstein (1983) wind stress and location of winter outcrop position (solid lines) of isopycnal surfaces ($\sigma_\theta = 26.5$ to 27.3) outcrop determined from Levitus et al. (1994)

http://stommel.tamu.edu/ baum/ocean_data.html (Levitus, wind, etc)

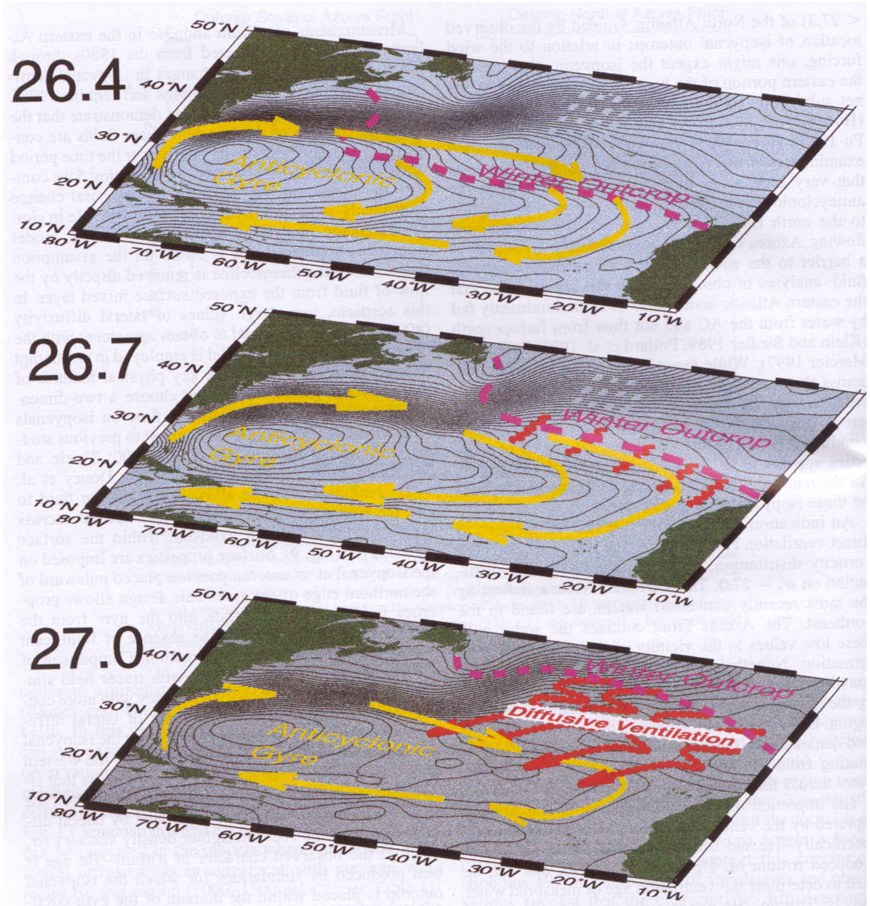
- From the isopycnal outcrop positions in relation to the wind forcing, one would expect that properties in the eastern basin to be set by the net subduction of fluid from the surface mixed layer.
- In contradiction, observed subsurface velocity fields show little net southward flow across the Azores front at 33° North that acts as a barrier to southward invasion of mass from the region of the isopycnal surface outcrops



PV distribution on $\sigma_\theta = 27.0$

PV distribution on $\sigma_\theta = 27.0$ present low values, indicating recently convected waters which emanate from the northeast and which are confined to the South by the Azores front.

- The radiodecay of bomb-produced tritium to ^{3}He provides a clock that can be used to determine the ventilation age of measured water parcels.
- Tritium-^{3}He age is only a precise estimate of true ventilation age if the subducted water parcels do not undergo mixing with surrounding fluids.
- Numerical simulations of the oceanic penetration of antrhropogenic tritium demonstrate that the presence of subsurface mixing distorts the measured tracer age field in a predictable manner. The amplitude of this distortion depends primarily on the strength of lateral mixing



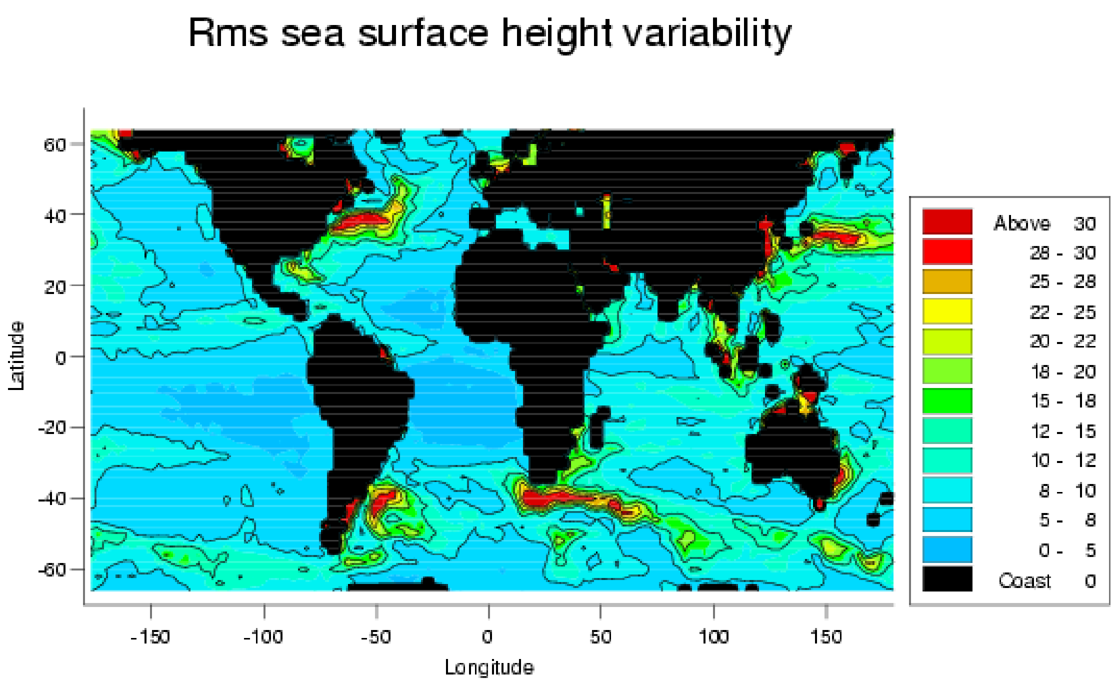
Robbins et al. 2000

- On $\sigma_\theta = 26.4$ the winter outcrop is within and to the south of the eastward flowing Azores Current (AC): properties are ventilated by direct advective flow originating in the winter mixed layer. This results in a Tritium-^{3}He age age field nearly steady with time.
- On the 2 deeper isopycnals $\sigma_\theta = 26.7$ and in particular on the $\sigma_\theta = 27$, surface outcrop is displaced northward of the AC and no streamlines of the anticyclonic gyre pass through the winter mixed layer. There are large observed temporal changes in the Tritium-^{3}He age field.
- Lateral mixing effects are vital to the evolution of basin-scale transient tracer fields on deeper isopycnals (”eddy-induced transport)

Parameterization of geostrophic eddies transport of tracers in climate models

Robbins et al.'s conclusions of the vital role of eddy lateral mixing of tracers fields are likely to hold for many regions in the ocean since geostrophic eddies are ubiquitous in the ocean.

Observations suggest that mesoscale variability is closely tied to the baroclinicity of the large-scale flow.

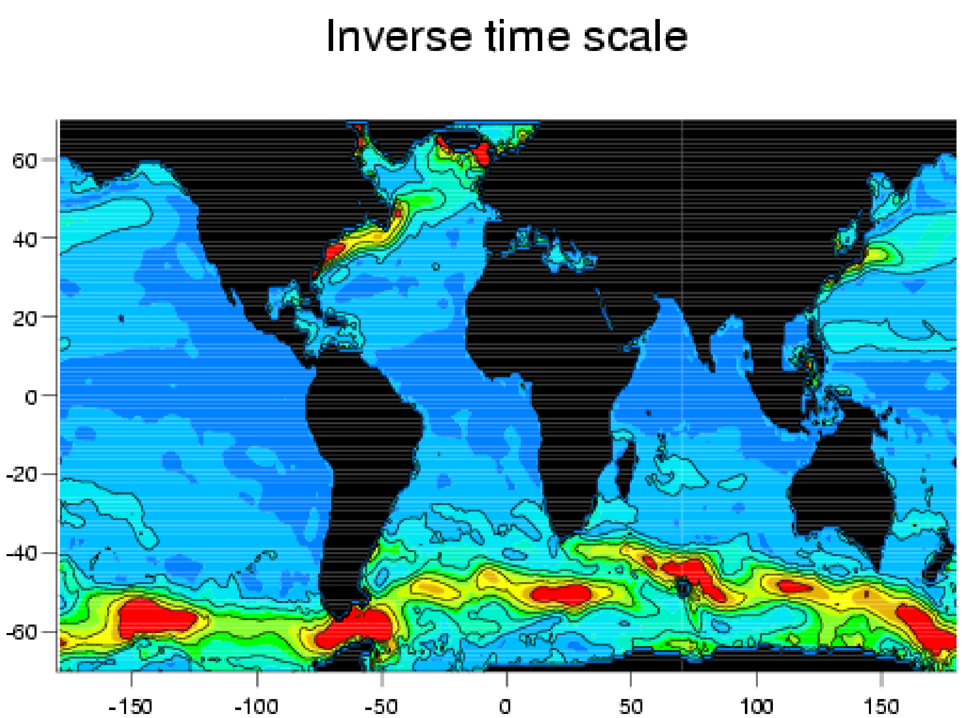


TOPEX rms sea-surface height variability averaged over 3 years (Tréguier et al. 1997)

On the other hand, one can also map the vertically-averaged growth rate of baroclinic waves using the classic Eady model scales as $f Ri^{-1/2}$ where f is the Coriolis parameter and

$$Ri = \frac{N^2}{(\partial_z U)^2 + (\partial_z V)^2}$$

is a Richardson number based on the large-scale mean flow U, V and the mean stratification $N^2 \equiv -\frac{g}{\rho_0}\partial_z \rho$. Held and Larichev (1996) argue that such time-scale averaged vertically is representative of the energy-containing eddies



Average time-scale of baroclinic energy containing eddies calculated from the Levitus density field (Tréguier et al. 1997)

The rough agreement between these two patterns supports the idea that baroclinic production is a major source of eddy activity and moreover that the relationship between baroclinicity and eddy statistics may be fairly local in the horizontal.

Such assumptions have been made by Gent & McWilliams (GM 1990, Gent et al. 1995, GM 1996) for proposing a parameterization of the important transfer properties of unresolved baroclinic eddies in coarse-resolution climate simulations. [Typical grid sizes in such models are 2-3° and cannot resolve typical eddy sizes which range from 5 to 200 kms.]

Gent & McWilliams approach is akin to that of studies in the middle atmosphere (Andrews et al., 1987) of chemical tracers distributions, which cannot be accounted for simply by sources/sinks and the Eulerian mean circulation. Instead the eddies transfer material by a process similar to Stokes drift, and this transport is comparable in magnitude and often counterbalances the Eulerian mean meridional circulation.

Andrews and McIntyre (1976) showed that one could add the wave-induced correction to the Eulerian circulation, which defines the the transformed Eulerian mean equations:

$$\begin{aligned} \overline{\boldsymbol{V}} &= \overline{\boldsymbol{v}} + \boldsymbol{v}^{\star} &&= \quad \overline{\boldsymbol{v}} + \frac{\partial \chi}{\partial z} \\ \overline{W} &= \overline{w} + w^{\star} &&= \overline{w} - \nabla_h \cdot \boldsymbol{\chi}, \end{aligned}$$

where $\boldsymbol{\chi}$ is a vector streamfunction proportional to the buoyancy flux given by

$$\boldsymbol{\chi} = -\frac{\overline{\boldsymbol{v}' b'}}{N^2}$$

$(\boldsymbol{v}^{\star}, w^{\star})$ are the ”residual mean” velocities and denote the wave-induced transport (formulated for 3D in GM90). Validity depends on assumptions of local growth and dissipation. These velocities are nondivergent and for an equilibrated eddy field, are the appropriate advecting velocity for scalars.

The large-scale tracer advection reads

$$\frac{\partial \overline{C}}{\partial t} + \overline{\boldsymbol{V}} \cdot \nabla_h \overline{C} + \overline{W} \frac{\partial \overline{C}}{\partial z} = \overline{S_0} - \overline{R}$$

here S_0 contains sources and sinks of tracer, R denotes an along-isopycnal mixing process, which can be represented in level models by a rotated diffusion tensor (Redi 1982).

However, in the special case $\overline{C}$ is the buoyancy, then the along-isopycnal mixing term R is identically zero. In that case, whenever $S_0 = 0$, then density surfaces are material surfaces and the large-scale flow can evolve in an adiabatic manner and the volume of fluid between isopycnal surfaces is conserved.

Gent and McWilliams (1990,1996) argue that eddy transfer in the ocean is just such an adiabatic process and choose for their residual mean velocities

$$
\begin{aligned}
\boldsymbol{v}^{\star} &= \frac{\partial}{\partial z}(\kappa\, \vec{\alpha_\rho}) \\
w^{\star} &= -\nabla_h \cdot (\kappa\, \vec{\alpha_\rho})
\end{aligned}
$$

where $\vec{\alpha_\rho}$ is the slope vector of the isopycnals.

Eddy-mixing tensor and eddy-induced advection

The eddy contribution to the tracer equation can be written quite generally as a 3 x 3 tensor operating on the gradient of the tracer, *when making an assumption of a local dependence between eddy fluxes and large-scale gradients*.

For illustrating purposes, let us consider the case where the tracer is temperature T and restrict ourselves to a 2D $y-z$ plane

$$\frac{\partial \overline{T}}{\partial t} + \frac{\partial}{\partial y}(\overline{v}\overline{T}) + \frac{\partial}{\partial z}(\overline{w}\overline{T})$$
$$+\frac{\partial}{\partial y}(\overline{v'T'}) + \frac{\partial}{\partial z}(\overline{w'T'}) = \text{Forcing + Dissipation}$$

$$\begin{bmatrix} \overline{v'T'} \\ \overline{w'T'} \end{bmatrix} = K \begin{bmatrix} \frac{\partial \overline{T}}{\partial y} \\ \frac{\partial \overline{T}}{\partial z} \end{bmatrix}$$

Eddy-mixing tensor

$$K = K_S + K_A$$

$$K_S = \begin{bmatrix} s_{yy} & s_{yz} \\ s_{yz} & s_{zz} \end{bmatrix}$$

$$K_A = \begin{bmatrix} 0 & -\chi \\ \chi & 0 \end{bmatrix}$$

The skew-symmetric part of the tensor amounts to an advection:

$$\begin{bmatrix} \overline{v'T'}_A \\ \overline{w'T'}_A \end{bmatrix} = \begin{bmatrix} 0 & -\chi \\ \chi & 0 \end{bmatrix} \begin{bmatrix} \frac{\partial \overline{T}}{\partial y} \\ \frac{\partial \overline{T}}{\partial z} \end{bmatrix}$$

$$\begin{aligned} \mathcal{D} &= \frac{\partial}{\partial y}(\overline{v'T'}_A) + \frac{\partial}{\partial z}(\overline{w'T'}_A) \\ &= - \frac{\partial \chi}{\partial y}\frac{\partial \overline{T}}{\partial z} + \frac{\partial \chi}{\partial z}\frac{\partial \overline{T}}{\partial y} \\ &= w^\star \frac{\partial \overline{T}}{\partial z} + v^\star \frac{\partial \overline{T}}{\partial y} \end{aligned}$$

In the adiabatic limit, K_S is oriented along isopycnals and the only eddy-effect on temperature is an advection represented by the meridional streamfunction χ

Parameterization of Eddy-induced velocity (summary)

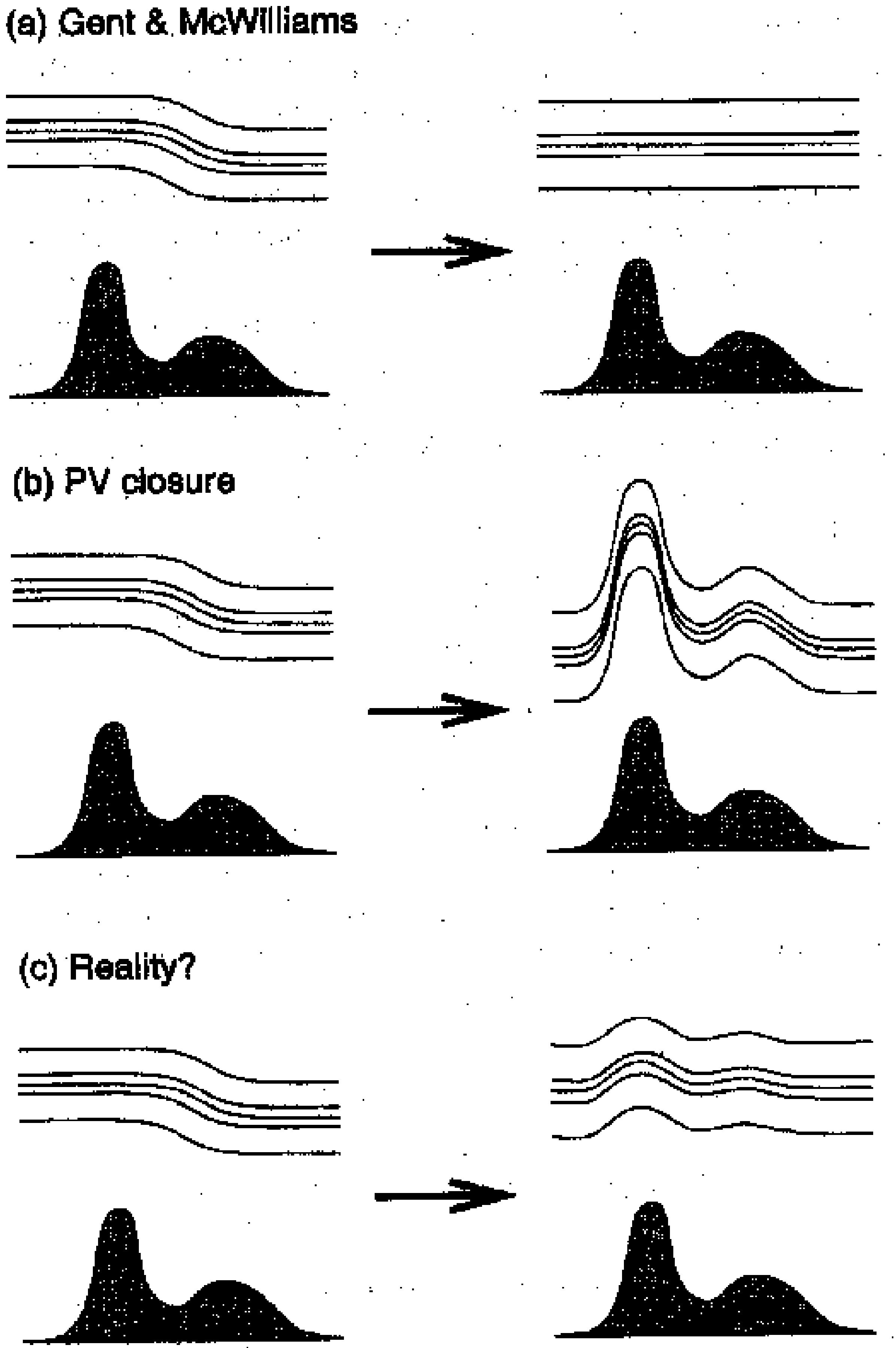
- Mixing of tracers by geostrophic eddies acts along potential density surfaces and advection of tracers is done by an effective transport velocity = large-scale velocity + eddy-induced transport velocity

- Gent et al. 1995 show that, for their choice of $(v^\star, w^\star)$, such a parameterization mimics baroclinic instability on average because it provides a negative definite sink in large-scale potential energy. An estimation of eddy heat transport using the Levitus density field yields a correct distribution with latitude of heat transport.

- Visbeck et al. 1997 advocate a parameterization of κ that varies in space and time which combines the ideas of GM90 and those of Green (1971) and Stone(1972) with a dependence of $\boldsymbol{\kappa}$ on the local Richardson number.

- Other studies advocate mixing of PV instead of mixing of thickness fluxes (Tréguier et al. 1997, Adcock and Marshall (1999) (see below).

- So far few studies have questioned the validity of the assumption of a local relation for the fluxes-gradient relation.



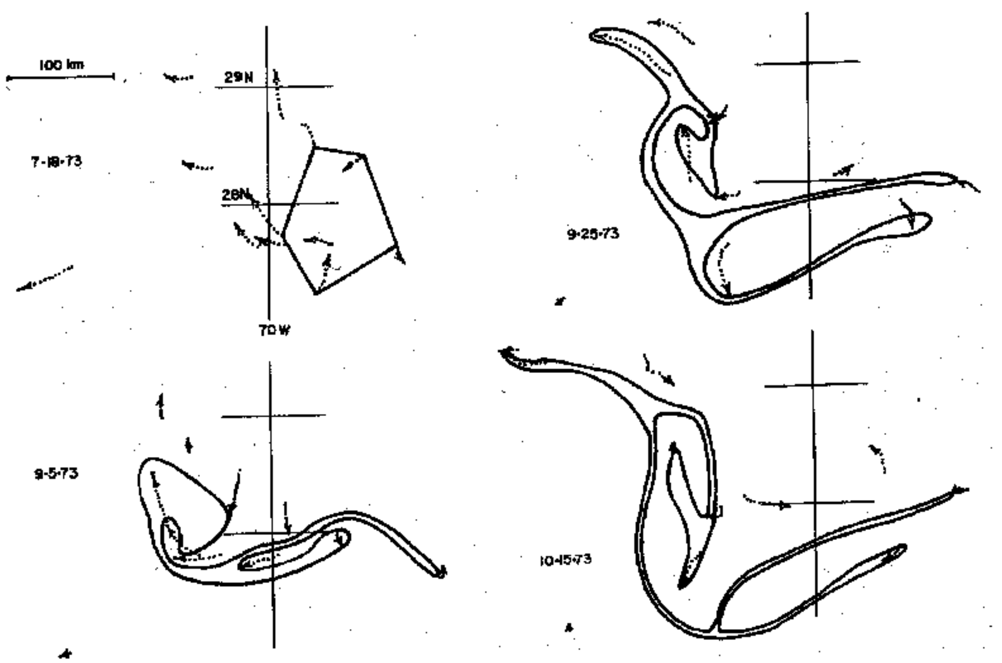
Adcock and Marshall, 1999

Mesoscale variability

Recent progress in float technology enabled to capture the many facets of the ubiquituous mesoscale variability in the ocean. Autonomous neutrally-buoyant devices, which passively follow the motion of water parcels and may be considered as ”Lagrangian”.

float technology site http://www.ifremer.fr/dtmsi/produits/Marvor/marvor.htm

A remarkable feature of geostrophic turbulence in presence of the β -effect is the co-existence of wavelike effects with patently turbulent straining fields. Freeland, Rhines and Rossby, 1975.



Distortion over a 3-month period of a polygon connecting 5 SOFAR floats.

Sketch of the smoothest figure that encloses a fixed area

subsurface WOCE float archive http://oceanic.cms.udel.edu/woce/FLOATS/index.htm)

A widely-used approach is to compute eulerian-averages of float statistics of absolute and relative dispersion (e.g. Davis, Griffa)

Geostrophic turbulence

A generic property of 2D turbulence is the spontaneous emergence of coherent structures (Basdevant et al. 1981, McWilliams 1984) and that property carries over to the stratified case of geostrophic turbulence as well (Hua and Haidvogel 1986, McWilliams et al. 1994, Dritschell 1998).

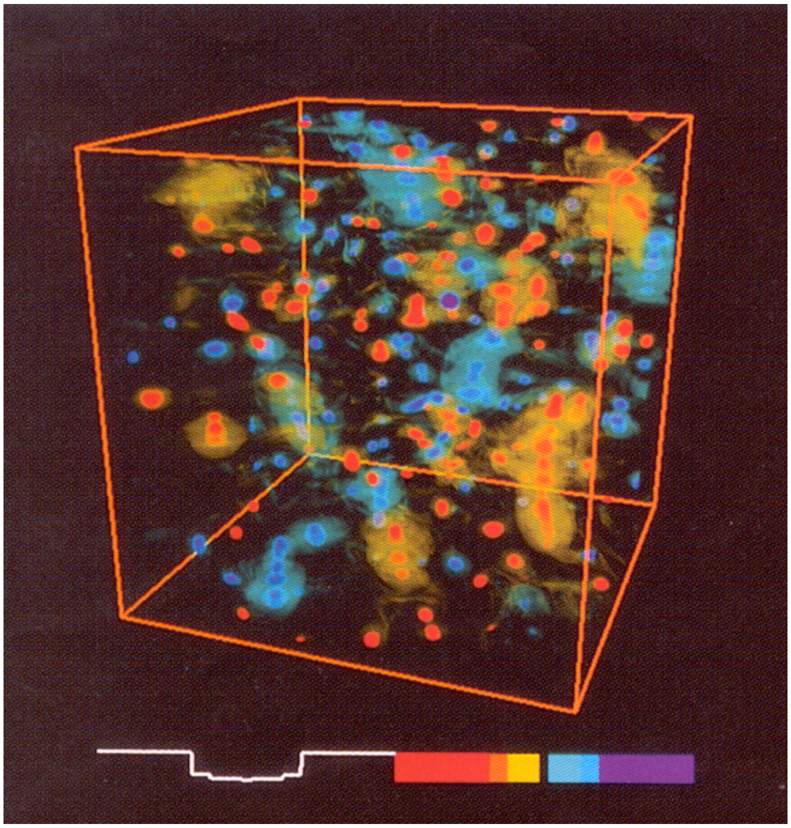
Such flows shoud obey the well-known quasi-geostrophic potential vorticity conservation

$$\frac{\partial q}{\partial t} + \frac{\partial \psi}{\partial x}\frac{\partial q}{\partial y} - \frac{\partial \psi}{\partial y}\frac{\partial q}{\partial x} = D$$

where D is the dissipation and q is the quasi-geostrophic potential vorticity

$$q = +\frac{\partial^2 \psi}{\partial x^2} + \frac{\partial^2 \psi}{\partial y^2} + \frac{f^2}{N^2}\frac{\partial^2 \psi}{\partial z^2}$$

Charney(1971) had originally predicted that geostrophic turbulence evolves to have a particular form of spatial isotropy of statistical average properties, when using an appropriate stretched vertical coordinate $z' = (N/f)z$ where N/f is typically of order 10-100 for the ocean.



Potential vorticity field at $t = 10.0$

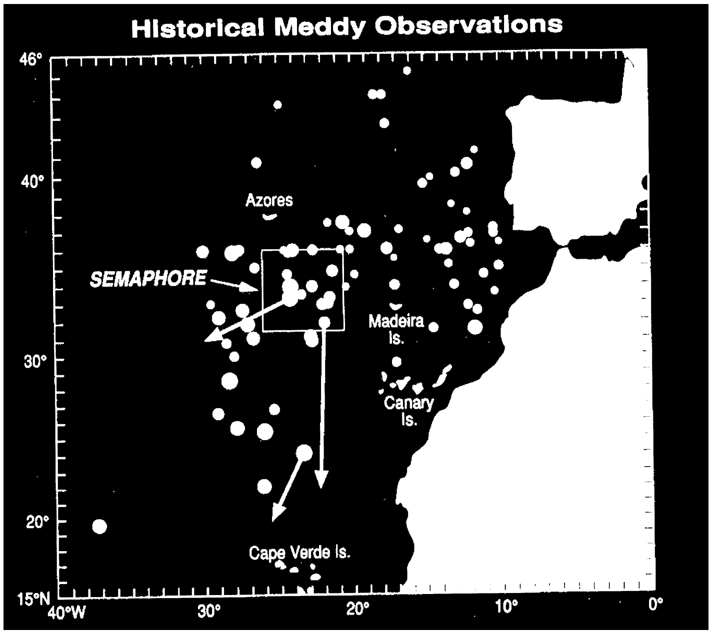
This was not confirmed by McWilliams et al. 1994, who found significant departures from isotropy, which they attributed to the large population of vortices in their $(320)^3$ multigrid simulation in a periodic cube. However their result has been a topic for debate: Dritschell 1998 found evidence for isotropization when using a contour surgery based code, but for different vertical boundary conditions.

The self-organization of geostrophic turbulence into long-lived coherent vortices has a bearing on oceanic observations of a very rich distribution of coherent vortices which horizontal size can vary from 5km to 200km in diameter, and which vertical size is such that they can either extend to the ocean bottom or be lens-like at intermediate depths.

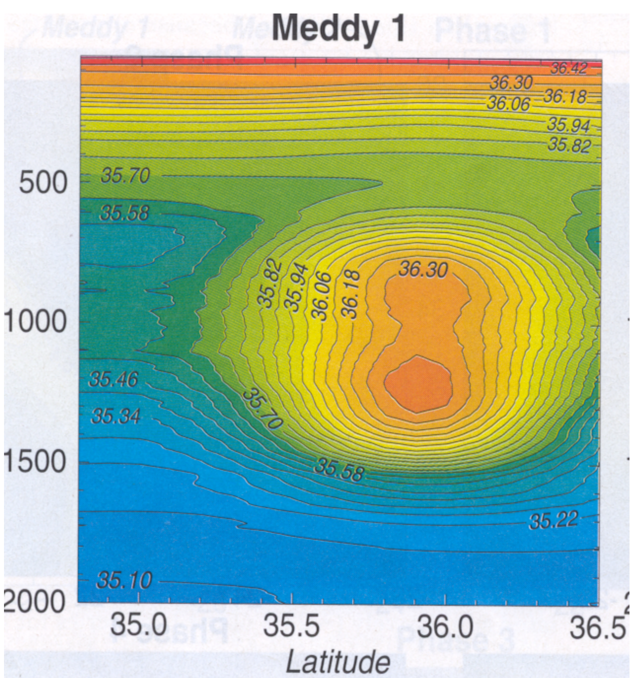
This has stimulated a wide body of litterature devoted to the hydrodynamics of coherent vortices and a recent review can be found in Carton (2001)

Meddies observations

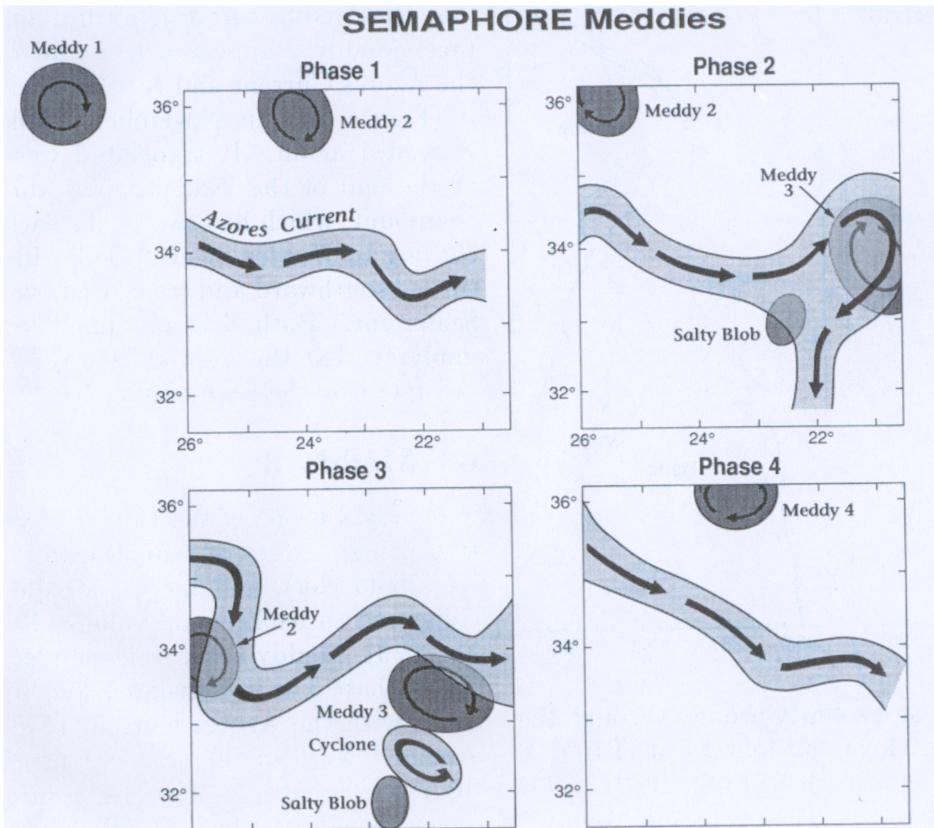
Meddies (Mediterranean water eddies) are coherent vortices which transport a significant amount of heat from the Mediterranean into the North Atlantic, at a mean depth of 1000m, and with a vertical extension of about 1000m. (Richardson and Tychensky 1998)



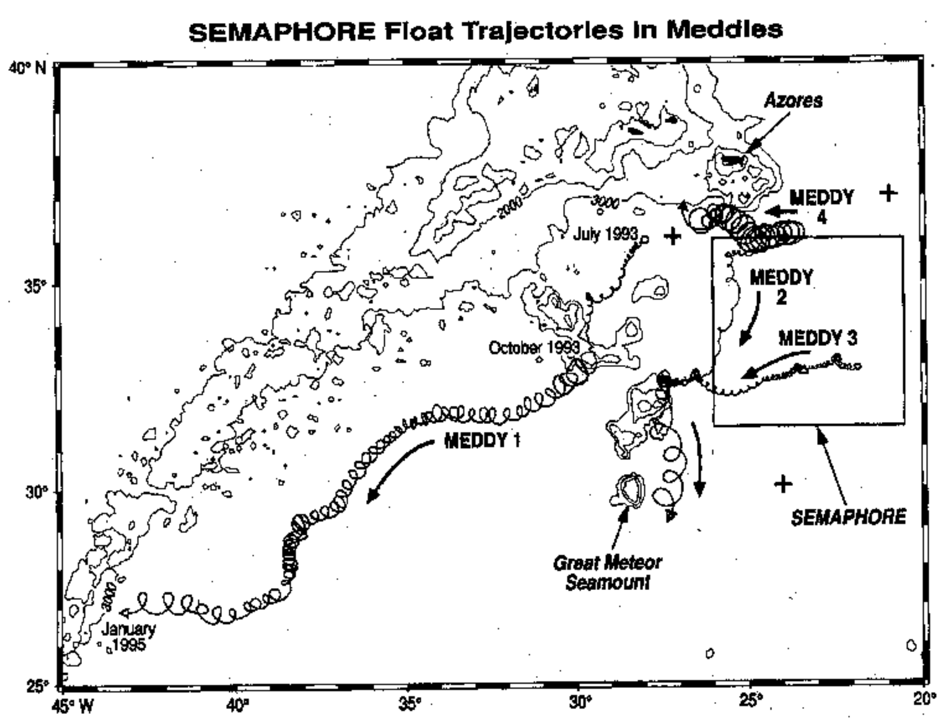
the diameter of dots indicates the magnitude of the salinity anomaly which ranges from 0.4 to 1.1 psu.



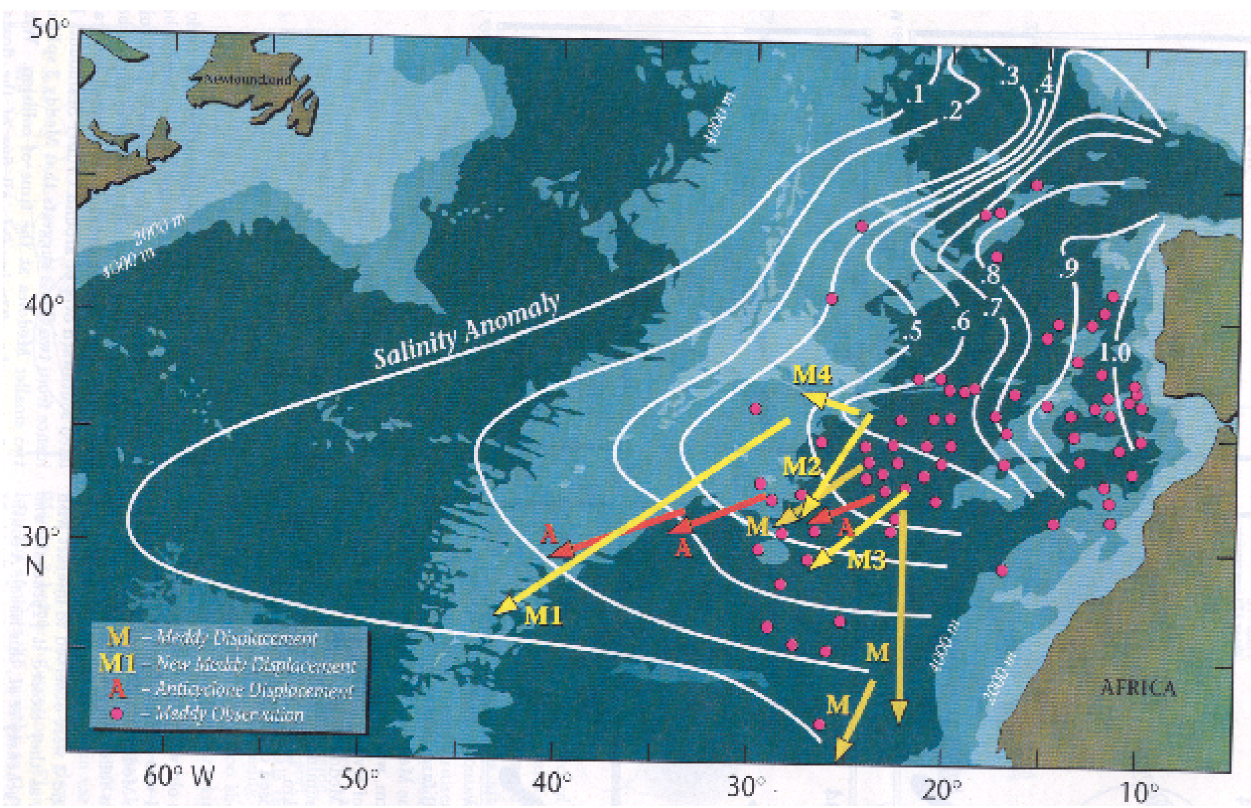
Meddies during the SEMAPHORE experiment were observed to interact with the Azores Current as they passed underneath or through it (Richardson and Tychensky 1998)



Three Meddies collided with tall seamounts, apparently fatally in two cases

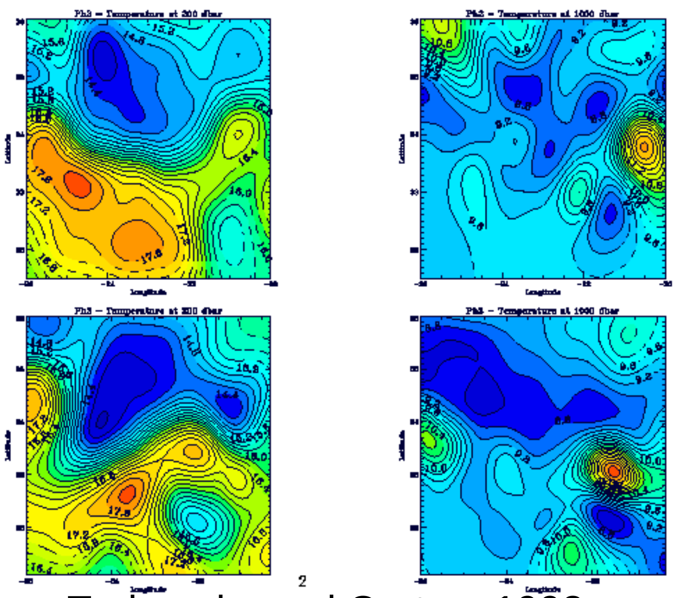


By blocking the Meddies and causing their early death, seamounts could have an important local influence on the transport of heat and salt and the structure of the Mediterranean salt tongue (Richardson and Tychensky 1998).

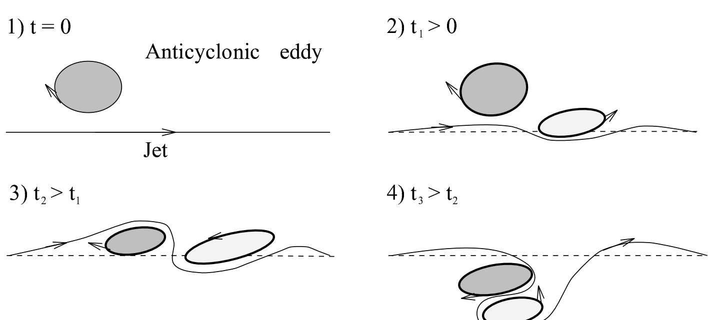


The study of the dynamics of these Meddies by Tychensky and Carton (1998) revealed a strong homogeneity of their core, a near-solid-body rotation in the core and a tripolar vertical structure in PV (an anticyclonic extremum at the center of the Meddy surrounded above and below by two positive PV anomalies).

Observations displayed quite a variety of coherent vortices dynamics, e.g. one of the Meddies (Meddy 3) crossed the Azores Current by forming a dipolar structure through the capture of a cyclonic meander of the AC



Tychensky and Carton, 1998



VanderMersch and Carton, 2001

Summary for coherent vortices

- Long-lived, abundant

- Transport heat and properties over very long distances

- Not well represented nor parameterized in present simulations of general circulation of the ocean

- Parity-bias: anticyclonic vortices more abundant than cyclonic ones in many parts of the water column (e.g. Lilly and Rhines (2001))

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