Baroclinic instability and turbulence in the ocean

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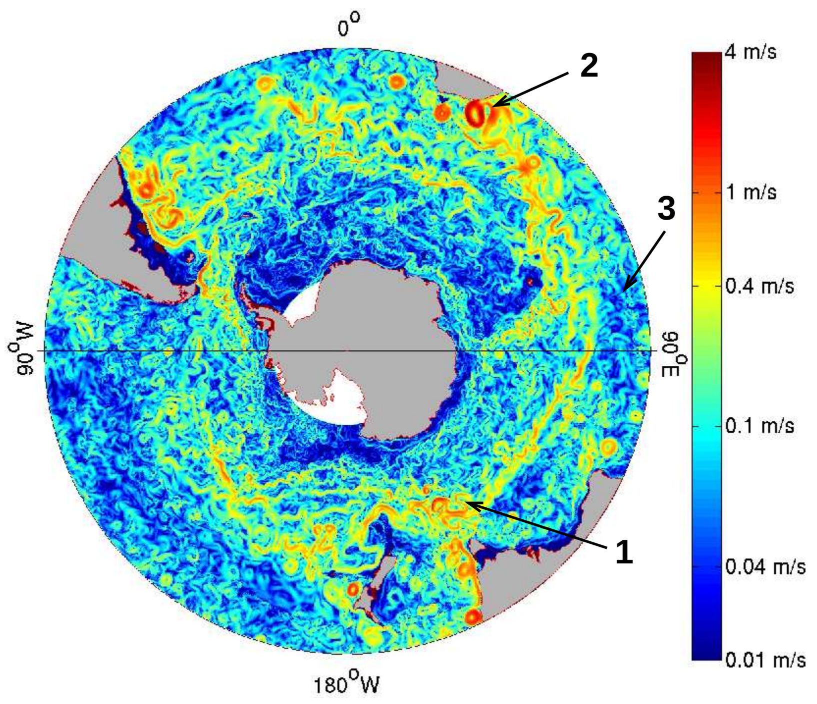
The ocean is a sea of eddies

- What determines their horizontal scales and their magnitude ?
- What is their horizontal organization (rings, jets,...)?
- What is their vertical partition (barotropic, baroclinic,...)?
- What is their geographic repartition (western boundary currents, ...)?
- What is their generation mechanism (baroclinic instabilities,...)?



Method 1: diagnostics from a comprehensive primitive equations ocean model

As a first step before future analysis of data from observations, we have performed diagnostics of the $1/6°$ resolution simulations of the MESO project Hallberg, Gnanadesikan JPO 2006. Their are simulations of an isopycnal hemispheric ocean model with realistic geometry and idealized forcing See above one snapshot of surface velocity modulus.
Projections on **Baroclinic modes** $\phi_m(z)$ (**Rossby radius** R_m):

$$\frac{d}{dz}\left(\frac{f^2}{N^2}\frac{d\phi_m}{dz}\right) = -\frac{\phi_m}{{R_m}^2} \text{ with } \left.\frac{d\phi_m}{dz}\right|_{z=0} = \left.\frac{d\phi_m}{dz}\right|_{z=-H} = 0.$$

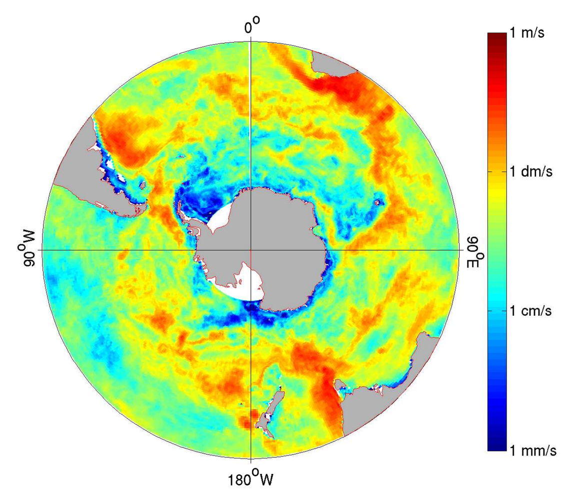
Method 2: quasi-geostrophic (QG) dynamics forced by an imposed vertical shear

- From MESO, at a given point (lat, lon) and after one year averaging: $\mathbf{U} = (U(z), V(z))$ and $\partial_x Q = \partial_z\left(S^2\partial_z V\right)$, $\partial_y Q = \beta - \partial_y\left(S^2\partial_z U\right)$, with $S = f/N(z)$.
- QG model in doubly periodic domain ($1000\ km \times 1000\ km$), effective horizontal resolution 256×256, quadratic bottom drag ($1/Cd \sim 400\ km$):

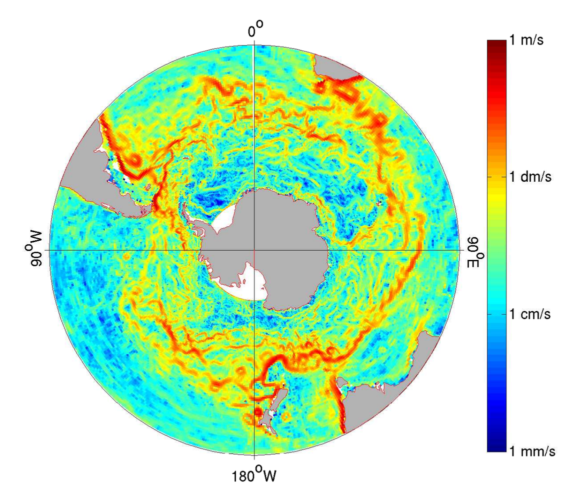
$$\partial_t q + \mathbf{u}\nabla q = -\mathbf{U}\nabla q - \mathbf{u}\nabla Q + \mathcal{D}_{lateral} + \mathcal{D}_{bottom}$$
$$q = \Delta\psi + \partial_z\left(S^2\partial_z\psi\right) \quad \text{with} \quad \mathbf{u} = \mathbf{k}\times\nabla\psi$$

Diagnostics from MESO

Eddy and Mean Kinetic Energy (EKE-MKE) (one year average)



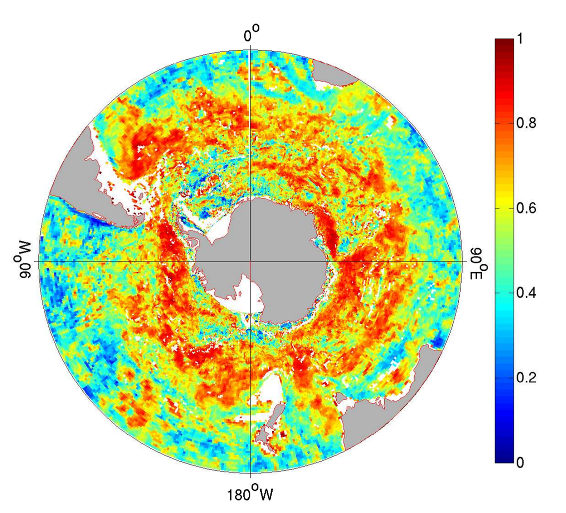
Verticaly integrated rms velocity ($\sqrt{2EKE}$).



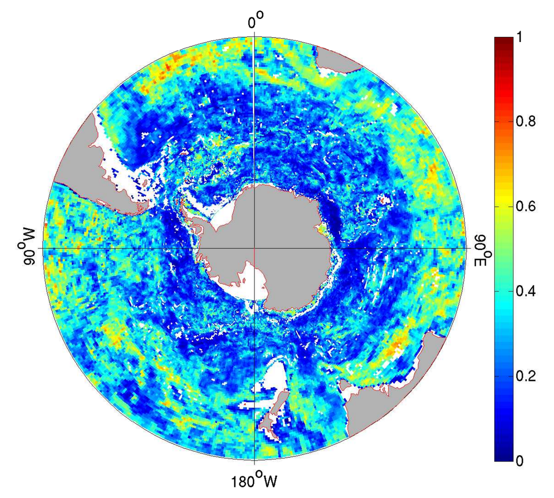
Verticaly integrated mean velocity modulus ($\sqrt{2MKE}$).

Vertical partition of the EKE on barotropic and 1st baroclinic modes

As in idealized simulations Smith and Vallis JPO 2001 and observations Wunsh JPO 1997.



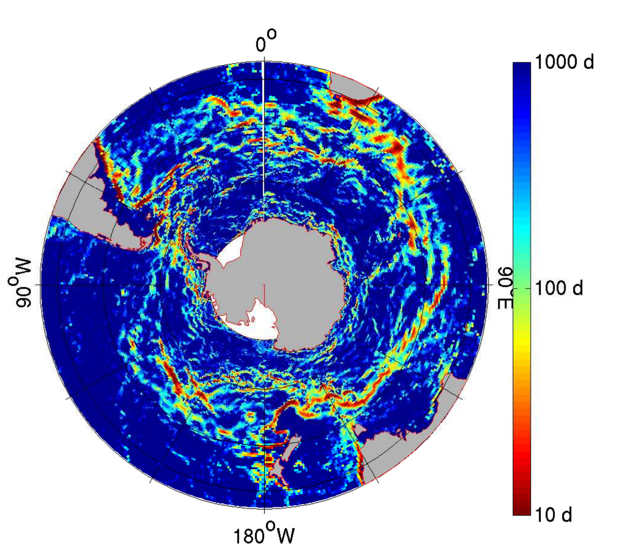
ratio of the EKE of the **barotropic mode over the total EKE**



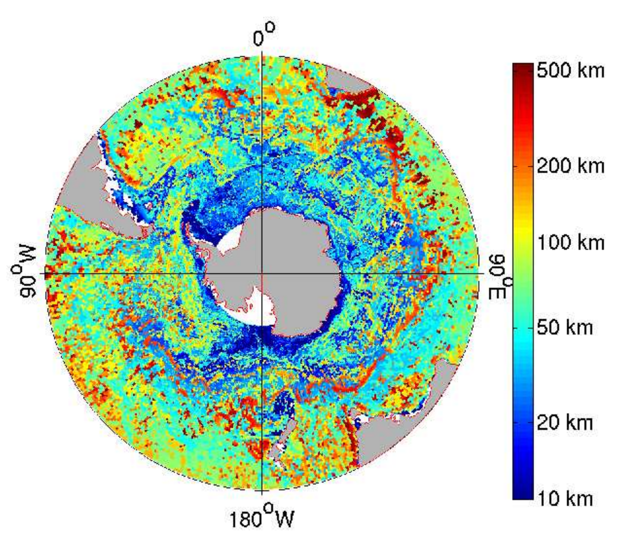
ratio of the EKE of the **1st baroclinic mode over the total EKE**

Linear instability analysis

Regions of fast growthrate are localised in space and associated to large wavelength L_{inst}, see also Tulloch *et al*, subm. to JPO



τ_{inst}: time scale of most unstable mode

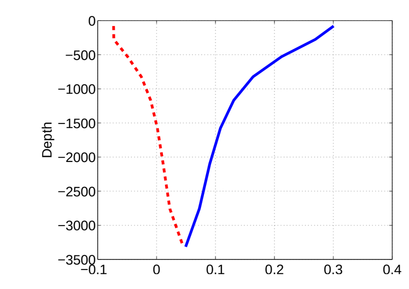


L_{inst}: wavelength of the most unstable mode

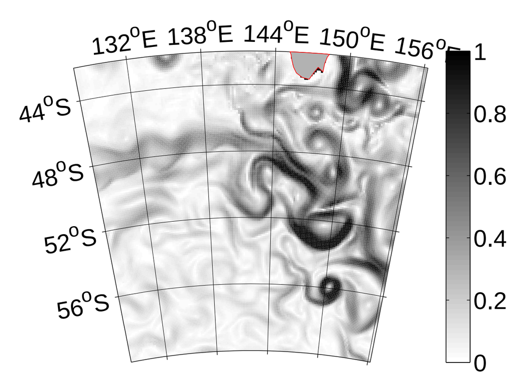
Diagnostics from QG simulations: three examples

Point 1 (142e51s): small rings.

Results presented below are consistent with similar recent studies Smith and Marshall JPO 09 and in reasonnable agreement with observations.

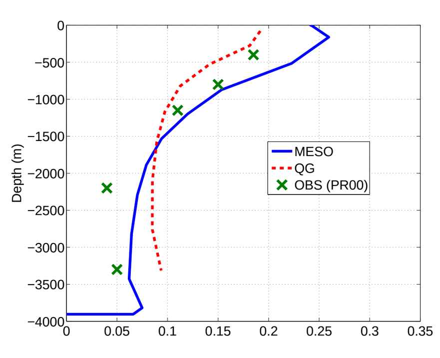


$U(z)$ plain blue, $V(z)$ dashed red ($m.s^{-1}$).

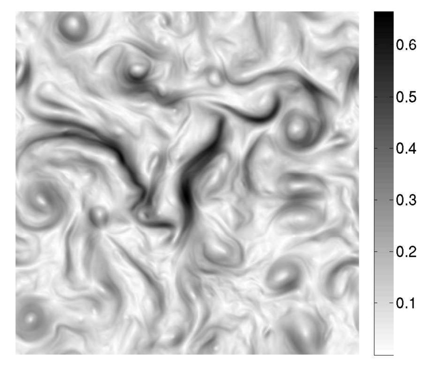


QG simulations

Domain size $L = 1000\ km$
Rossby Radius $R = 14\ km$
Instability wavelength $L_{inst} = 161\ km$
Barotropic E centroid: $L_{Et} = 288\ km$
1st baroclinic E centroid: $L_{Ec} = 182\ km$
$54\ \%$ of KE is barotropic
$24\ \%$ of KE is 1st baroclinic



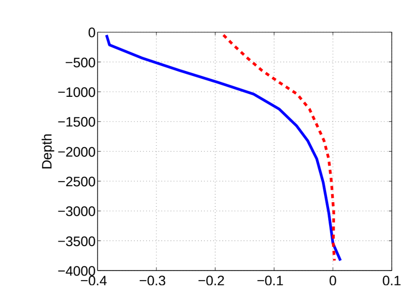
rms velocity ($m.s^{-1}$)



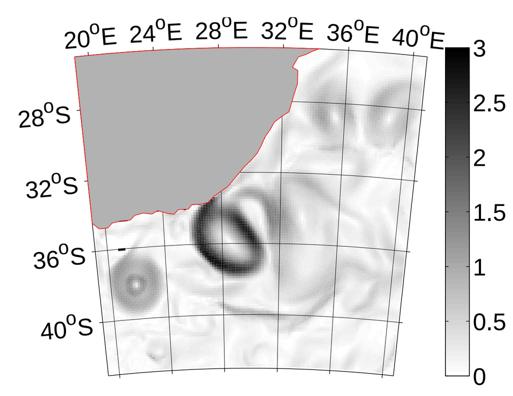
$|\mathbf{u}_{surface}|$ ($m.s^{-1}$) (top: MESO, $L \sim 2000\ km$; bottom: QG $L \sim 1000\ km$

Point 2 (30e34s): vortex lattice.

Taking $V = 0$ does not change qualitatively the results presented below. Both β and bottom friction have to be no zero to observe the vortex lattice, as in the two layers simulations of Arbic Flierl JPO 04.

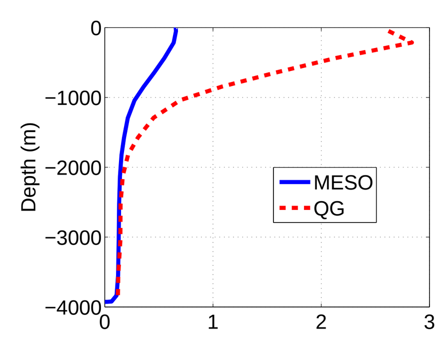


$U(z)$ plain blue, $V(z)$ dashed red ($m.s^{-1}$).

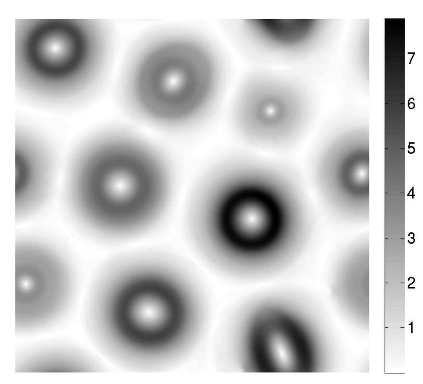


QG simulations

Domain size $L = 1000\ km$
Rossby Radius $R = 36\ km$
Instability wavelength $L_{inst} = 385\ km$
Barotropic E centroid: $L_{Et} = 396\ km$
1st baroclinic E centroid: $L_{Ec} = 365\ km$
$35\ \%$ of KE is barotropic
$49\ \%$ of KE is 1st baroclinic



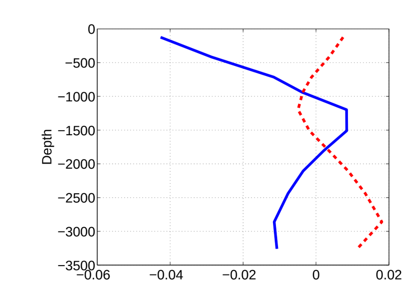
rms velocity ($m.s^{-1}$)



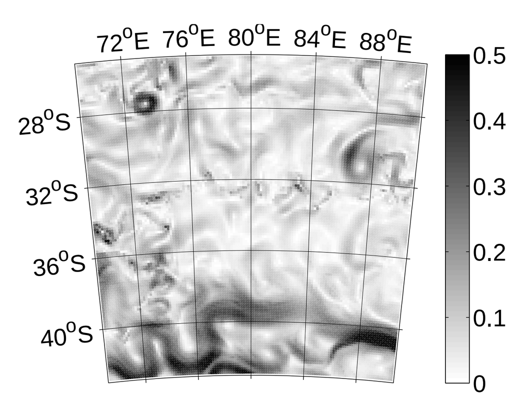
$|\mathbf{u}_{surface}|$ ($m.s^{-1}$) (top: MESO, $L \sim 2000\ km$, bottom: QG, $L \sim 1000\ km$

Point 3 (80e34s): eastward jet.

Taking $U = 0$ does not change qualitatively the results presented below: importance of the combined effect of β and of a meridional shear, Spall JMR 00 and Smith JPO 07.

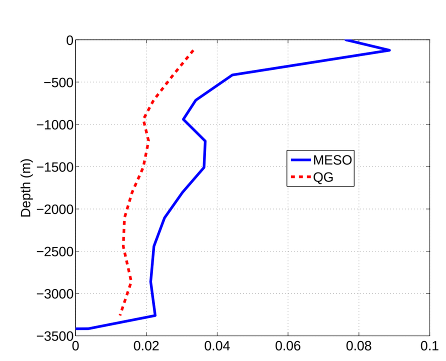


$U(z)$ plain blue, $V(z)$ dashed red ($m.s^{-1}$).

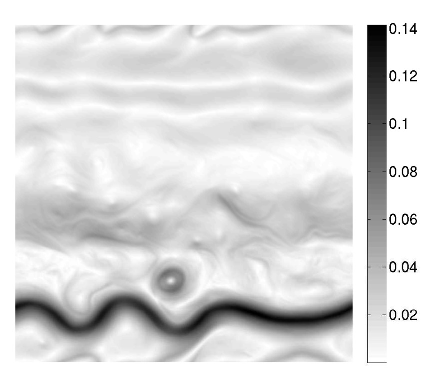


QG simulations

Domain size $L = 1000\ km$
Rossby Radius $R = 38\ km$
Instability wavelength $L_{inst} = 125\ km$
Barotropic E centroid: $L_{Et} = 338\ km$
1st baroclinic E centroid: $L_{Ec} = 260\ km$
$34\ \%$ of KE is barotropic
$27\ \%$ of KE is 1st baroclinic



rms velocity ($m.s^{-1}$)



$|\mathbf{u}_{surface}|$ ($m.s^{-1}$) (top: MESO, $L \sim 2000\ km$, bottom: QG, $L \sim 1000\ km$

Take home messages

Vertical structure of oceanic eddies projects well on the barotropic mode and the first baroclinic mode (QG and MESO)

Oceanic turbulence is self-organized into rings or quasi-stationary intense zonal jets

Strikingly, this formation of such cohorent structures are observed even without taking into account other mechanisms such as bottom topography or barotropic instabilities. We need a theory for this. **Statistical mechanics** might be a good candidate Venaille and Bouchet, *in prep.* for JPO, Bouchet and Venaille, *Lecture notes* World Scientific 2010.

Limits of the approach

i) Artificial separation between mean and eddying flow
ii) Locality assumption (Regions characterized by strong energy levels are localized in space, and eddies can propagate away from the generation region)
iii) The important role of bottom friction