A FORCED DISSIPATED PERSPECTIVE ON THE OCEAN MESOSCALE TURBULENCE

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Abstract By considering the ocean mesoscale turbulence as a forced dissipated dynamical system a natural question is what sets the mean state and the level of turbulence, given a steady large-scale forcing? The turbulence pops in spontaneously as a consequence of the instability of the mean state when pushed beyond its stability zone by the forcing. The turbulence is the process that relaxes the imbalance imposed by the forcing. Two dynamical systems are studied: a 3D ocean model integrating the primitive equations and a low dimensional Hamiltonian system. Long time numerical integrations are performed and statistical quantities are diagnosed. These results are used as guidelines and test points to derive a statistical theory.

QUESTION AND EXPERIMENTAL APPROACH

At scales smaller than about 400 km and down to 10 km the ocean dynamics is dominated by eddies, continuously generated, interacting, propagating and ultimately dissipated [2]. This is the so called mesoscale turbulence where the Earth rotation, the Earth curvature, the bathymetry and the stratification are major actors of the dynamics. This turbulence is sustained principally by the wind blowing at the surface of the ocean and in a less extent by the heat fluxes exchanged between the ocean and the atmosphere. However the wind is not directly stirring the ocean, the forcing is rather indirect: the large-scale zonal winds drive Ekman circulations that continuously tilt the stratification in the meridional direction and feed the potential energy on the large-scale. On the long term this tendency is balanced by the eddy fluxes [3]. The consequence is a mean state where the iso-density surfaces (isopcynals) are sloping. In return, the sloping is associated with large-scale currents via the thermal wind balance. There is thus a subtle interlinking between the forcing (a steady forcing is sufficient), the mean state (the general circulation) and the fluctuations (the eddies). The pivotal process that connects the mean state and the eddies is the baroclinic instability [4]. The baroclinic instability occurs when isopycnal slopes pass a critical value. The instability converts potential energy (PE) from the mean state to eddy kinetic energy (EKE). The interactions of eddies cause an inverse cascade of energy up to a scale where dissipation at the bottom of the ocean occurs. For the eddies to be continuously generated the mean state has to be kept unstable by the forcing. Our central question is what sets the mean state and the level of fluctuations for a given forcing? This is a formidable challenge. The question can be recast in a hierarchy of formulations from a semi-idealized continuous case to a low dimensional dynamical system:

Q1: for a given prescribed steady forcing, acting at the boundary of a zonally reentrant channel, and a given dissipation functional (e.g. linear drag), what sets the mean state (mean velocity, mean stratification at large-scale) and the level of turbulence (variance of velocity, variance of density at meso-scale)?

Q2: for a given low dimensional Hamiltonian system that admits a chaotic region for large enough energy level, forced by a steady forcing and given a certain dissipation functional, what sets the mean and the variance of the variables?

The common important ingredient of those two questions is that the turbulence (chaos) arises from a steady forcing that pushes the system beyond its stability region. Fluctuations are sustained by the instability and are the key mechanism by which the system tends to relax the forcing. They mediate the energy toward the dissipation.
Q1 is addressed using eddy resolving numerical simulations with an ocean model (ROMS) in reentrant channel on the beta plane [1]. Q2 is addressed using a six degrees of freedom Hamiltonian system with a degenerated Poisson bracket that aims to mimic a fluid Poisson bracket. We study the sensitivity of the statistical equilibrium state (means and variances) to the forcing. This is done by numerically integrating the experiments on very long times, discarding the spin-up phase and performing diagnostics on the statistical equilibrium regime.
In parallel of these numerical experiments we try to adapt existing theoretical tools from the statistical physics. The main hurdle is that the stochastic nature of the flow ultimately comes from the internal instability of the flow rather than from an external stochastic forcing.

RESULTS

One interesting feature of the problem Q1 is that because there is an inverse cascade of kinetic energy, there is no need to numerically resolve a wide range of scales to capture the essence of the equilibration. The key ingredient is to be eddy resolving, namely to allow the baroclinic instability to develop and to continuously generate eddies.

Our first diagnostics are integral ones: total volume transport, total EKE, total dissipation etc. We check how each variable scales with the wind. We then turn to the spatial structure of the flow and average quantities in the zonal direction. Understanding these patterns is central because the baroclinic instability completely depends on this pattern. We find a sensitivity of the isopycnal slopes to the wind. The weak forcing limit is interesting as the mean state is expected to be closer to its stability threshold. We diagnose the energy cycle of the system in terms of reservoirs (level of EKE) and of production (wind input, conversion PE to EKE, dissipation). We also try to develop a potential vorticity cycle of the flow. Finally we report on the temporal fluctuations of point-wise variables along with their probability density function (PDF). The horizontal velocity components and the density are Gaussian white noise. The vertical velocity is a white non Gaussian noise. A simple fit of its PDF is $P(w) = e^{-|w/w_0|^{\alpha}}$, with w_0 a scale parameter and α an exponent less than unity. Eddy fluxes such the meridional heat transport $v'\rho'$ or the energy conversion rate $w'\rho'$ are not Gaussian. However we can show that despite flatter tails and presence of extremes fluctuations, mainly weak fluctuations participate to the overall eddy flux.

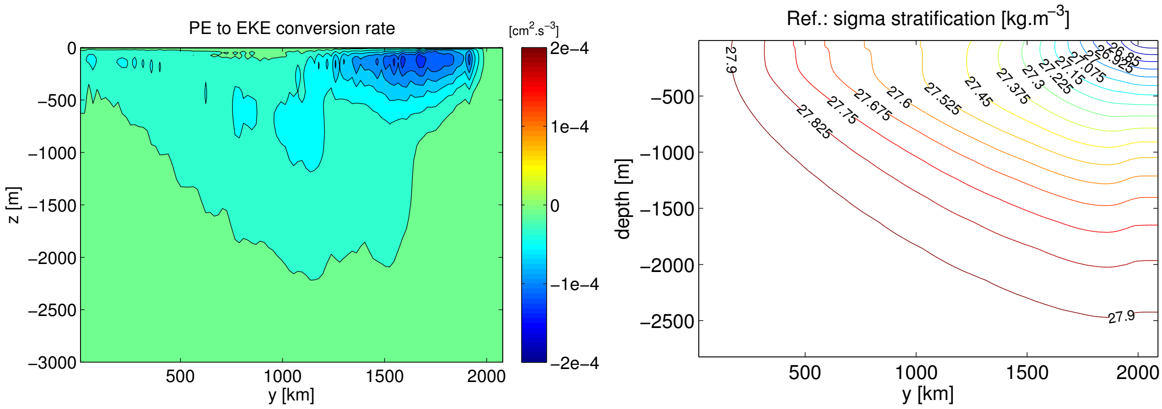

Figure 1. Mean conversion rate from potential to kinetic energy in the meridional plane $g\overline{w'\rho'}/\rho_0$ with $g = 9.8\,m.s^{-2}$ $\rho_0 = 10^3\,kg.m^{-3}$ (left). This is the footprint of baroclinic instability in action. Corresponding mean stratification $\sigma = \rho - \rho_0$ (right).

For the Q2 problem, we first show that the inviscid unforced system has a laminar to chaotic transition, that we believe is an important ingredient of the problem. We then apply the same methodology and report on the sensitivity to the forcing. Q2 is in spirit in the same line as the Lorenz's approach but differs by the fact that the system is basically forced-dissipated as opposed to purely inviscid. This model suffers from the arbitrariness of its equations.

Finally, at the time of writing of this abstract we are still trying to frame these results into a mathematical theory. Our hope is to manage to adapt a Fokker-Planck equation for this kind of problem. We expect to make real progress in the following months.

References

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