THE TURBULENT OCEAN

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ABSTRACT

The variability of the ocean over a wide range of scales, from the megameter to the millimeter, is examined in the light of turbulence theory.

The geophysical constraints which arise from the Earth's rotation and curvature and from the stratification are discussed with emphasis on the role they can play at different scales in inducing instabilities and a transfer of energy to other scales of motion.

INTRODUCTION

Turbulence in the ocean is still a very controversial subject. An innocent physicist, emboldened by a solid background in the theory of turbulence, who would approach the subject relying on his good understanding of, say, turbulent channel flow, might soon find himself confronted with a maze of conflicting experimental data and a nightmarish farrago of theories where he recognizes very little of what he regards as "turbulence".

Assuming that he is able to muster enough of his high school latin and greek to find his way among iso-halines, iso-pycnals and other proliferating peculiar surfaces which seem to fill the ocean with complexity, he may still require some time to adjust to new cabalistic concepts such as enstrophy, red energy cascade, eddies "which do not overturn", double diffusion or fossil turbulence.

Used to regard a turbulent flow as the superposition of a mean motion and turbulent fluctuations, he must face the fact that, if such things exist in the ocean, they are several thousands orders of magnitude apart in scales and apparently separated by a jungle of complicated movements which one refers to as the "variability of the ocean".

One of the most intriguing aspect of ocean variability is that, while the macroscale dynamics is claimed to be governed by a cascade of enstrophy to smaller scales, studies of mesoscale and microscale variability seem totally unconcerned with it as if it had disappeared somewhere on the way. (Planetary oceanographers talk about enstrophy "dissipation" but surely they mean something else like "annihilation"; even if one of them, answering a question at the Eleventh Liège Colloquium on Ocean Hydrodynamics, expressed the somewhat surprising view that "dissipated" enstrophy turns into heat).

For somebody trained in classical turbulence theory, it is not immediately obvious that the variability of the ocean is a form of turbulence.

If it is turbulence, it is clearly very different from the type of turbulence with which mechanical engineers, say, are familiar.

The reason why it should be so is found in the work of those oceanographers who, taking a quite opposite view, insist on describing ocean hydrodynamics in terms of non-turbulent theories such as linear wave propagation or molecular effects.

That they succeed in explaining some of the observations - after removing turbulence from the experimental signals in an operation which they gallantly call "decontamination" - is an indication of the mechanisms which are particular to geophysical flows and which are liable to modify, more or less drastically, geophysical turbulence.

The Coriolis force and the density stratification - which allow wave motions that do not exist in non-rotating non-stratified fluids -, the variations of density not only with temperature but also with salinity, - with different molecular diffusivities for momentum, heat and salt -, and their interference in other processes related to bottom or coastal topography and air-sea interactions, provide a great variety of mechanisms which eventually combine with mechanical effects to determine the stability or the instability of oceanic motions and the subsequent constraints on turbulence at different scales.

Including such geophysical constraints, it becomes possible to understand some essential characteristics of ocean variability and to build an image of what turbulence in the ocean may be.

TIME SCALES, LENGTH SCALES AND LINEAR WAVE THEORY

The equations of Geophysical Fluid Dynamics admit linear wave solutions of different kinds (e.g. Monin et al, 1977). Whether these waves can be observed in the ocean depends on a series of factors. Very small amplitude waves will not be affected by non-linear interactions but, on the other hand, they may be masked by stronger motions and remain unnoticed. Interactions between larger amplitude waves may create an intricated field of waves and wave packets of all scales, wave breaking and turbulence, from which the ideal individual wave of linear theory cannot be sorted out.

The theory of linear waves is however always a useful mathematical exercise as it helps to identify the dominant length scales (wave numbers) and time scales (frequencies) of motions.

From this point of view, it is convenient to divide ocean waves into three categories (Nihoul, 1979) :

A. Macroscale waves

These waves have frequencies (ω) in the range

$$10^{-8}s^{-1} \lesssim \omega \lesssim 10^{-5}s^{-1} \tag{1}$$

and operate over a range of horizontal wave-numbers (k)

$$10^{-6}m^{-1} \lesssim k \lesssim 10^{-3}m^{-1} \tag{2}$$

(Waves of larger periods could be considered but the limiting frequency 10^{-8} corresponding to periods larger than 10 years, they are presumably rather irrelevant to the present discussion). Macroscale waves are directly related to the spatial variations of the Coriolis parameter f ($f = 2\,\Omega \sin\phi$ where Ω is the angular velocity of the earth's rotation and ϕ the latitude) i.e. to the parameter

$$\beta = \|\underset{\sim}{\nabla} f\| \sim 10^{-11}m^{-1}s^{-1} \tag{3}$$

Macroscale waves include (e.g. Rhines, 1977 ; Nihoul, 1979)

(i) very slow baroclinic waves ($\omega k \ll \beta$) for which

$$\omega \sim \frac{N^2 \beta H^2}{f^2}\, k \tag{4}$$

where H is the depth of the ocean and N the Brunt-Väisälä frequency.

(The condition $\omega k \ll \beta$ yields $L \sim k^{-1} \gg R$ where $R = NHf^{-1} \sim 10^5$ m is the so-called "Rossby internal scale". These waves are thus very slow large scale small amplitude waves which, if excited, are likely to break rapidly under the effect of bottom slope and general baroclinic instability).

(ii) barotropic Rossby waves ($\omega k \sim \beta$) for which

$$\omega \sim \beta k^{-1} \sim f \gamma H^{-1} k^{-1} \qquad (5)$$

where γ is the non-dimensional mean bottom slope chosen here of the order $\gamma \sim 10^{-4}$ such that $f \gamma H^{-1} \sim \beta \sim 10^{-11}$.

(In a forced problem, with wind blowing accross the ocean surface, these waves constitute the most important mode at "weatherlike" time scales over horizontal scales greater than R).

(iii) fast baroclinic waves confined within a layer of thickness $f L N^{-1}$ above the sloping bottom, for which

$$\omega \sim \gamma N \qquad (6)$$

(If, in the dispersion relation of topographic Rossby waves $\omega \sim f \gamma H^{-1} k^{-1}$, one replaces the depth H by the "penetration height" $f L N^{-1}$, one obtains the dispersion relation (6). These waves may thus be regarded as topographic Rossby waves where density stratification provides a lid for vortex stretching).

B. <u>Mesoscale waves</u>

These waves have frequencies in the range

$$10^{-5} s^{-1} \lesssim \omega \lesssim 10^{-2} s^{-1} \qquad (7)$$

and include gyroscopic waves and inertial oscillations, tides, and internal gravity waves (e.g. Tolstoy, 1963 ; Monin et al, 1977).

The effect of the Earth's curvature becomes here negligible and the essential factors in the dispersion relations are the Coriolis parameter f and the Brunt-Väisälä frequencies N_{min} and N_{Max}.

C. <u>Microscale waves</u>

These waves have frequencies

$$10^{-2} s^{-1} \lesssim \omega \qquad (8)$$

They are essentially surface waves and acoustic waves. The forme only affect the upper layer of the ocean and may be regarded as an indispensable - unfortunately rather complicated - way of transferring momentum and energy directly from the wind to the sea ; the latter are marginally important in Ocean Hydrodynamics from which they are customarily excluded by the Boussinesq approximation.

MACROSCALE TURBULENCE

Macroscale motions in the ocean include large scale currents (gyres) and quasi-geostrophic or "synoptic" eddies which appear, from observational studies, to contain a large fraction of the ocean's kinetic energy.

The dynamics of the synoptic eddies is dominated by the earth's curvature - parameterized in terms of β - and their horizontal length scale is of the order of the Rossby internal scale R.

The spectral characteristics of the synoptic eddies, wave number κ_β, frequency ω_β and energy level $\kappa_\beta E(\kappa_\beta)$ where $E(\kappa)$ is the horizontal kinetic energy spectral density, can be estimated by turbulence similarity arguments. One finds

$$\kappa_\beta \sim R^{-1} \sim 10^{-5} m^{-1} \quad (9)$$

$$\omega_\beta \sim \beta\kappa_\beta^{-1} \sim 10^{-6} s^{-1} \quad (10)$$

$$\kappa_\beta E(\kappa_\beta) \sim \beta^2\kappa_\beta^{-4} \sim 10^{-2} m^2 s^{-2} \quad (11)$$

These estimates appear to be in good agreement with the observations (e.g. Koshlyakov and Monin, 1978).

The use of the term "synoptic" emphasizes the physical analogy between these eddies and the synoptic eddies of the atmosphere (cyclones and anticyclones, quasi-geostrophic motions at the Rossby scale).

The synoptic variability of the atmosphere, however, has time scales of the order of a week and is shaped by pressure lows and highs with characteristic horizontal scales of the order of the thousand of kilometers and one must exclude the hypothesis of the generation of synoptic ocean eddies by direct resonant interactions. Atmospheric disturbances - lows and highs - generate large-scale currents in the ocean and it is the barotropic and essentially baroclinic instability of these currents which provide the energy for the synoptic eddies (Kosklyakov and Monin, 1978).

(The kinetic - and approximately equal potential-energy of the synoptic eddies is essentially higher than the kinetic energy of the large scale currents and at the same time much smaller than the available potential energy of the latter. This constitutes strong experimental evidence of eddy generation through baroclinic instability of the large scale oceanic currents).

Large scale ocean experiments (Polygon, Mode, ...) give evidence of synoptic eddies of two kinds, "frontal eddies" produced by the cut-off of meanders from such frontal currents as the Gulf Stream and the Kuroshio, and much weaker "open-ocean eddies". The kinetic energy of the frontal eddies can be two orders of magnitude larger than the kinetic energy of the typical ocean eddies, the rotation velocity in the upper part of frontal eddies can reach meters per second. (e.g. Kosklyakov and Monin, 1978 ; Nihoul, 1979).

The vertical length scale of the synoptic eddies is of the order of the depth (e.g. Rhines, 1977 ; Woods, 1977 ; Nihoul, 1979) and it is very tempting to regard them as constituting a form of two-dimensional turbulence.

Potential vorticity would be conserved in such motion and, in the words of Gill and Turner (1979), "patches of marked particles would be teased out, into spindly shapes, leading to an enstrophy ('mean square vorticity') cascade to smaller scales".

Such a cascade predicted by the mathematical theory of homogeneous two-dimensional turbulence (e.g. Kraichnan, 1967 ; Batchelor, 1969) implies that the flow of kinetic energy is from smaller to larger scales (the "red cascade").

There must be however processes - presumably different at different levels and in different regions - which are not quasi-geostrophi and which limit the extension of potential vorticity contours.

Turbulent energy transfer to small scale may jump over an eventual synoptic valley via boundary turbulence, intermittent internal turbulence or non-local cascade into internal waves. Bottom roughness provides a permanent mechanism for the conversion from large to small scales. An initial cluster of eddies, surrounded by quiet fluid may cascade to longer scales but eventually the energetic patch will contain too few eddies to act as turbulence. Another obstacle to the 2D red cascade is the restoring force provided by the β-effect or its topographic equivalent. No matter how intense or how small the initial eddies, the red cascade carries the flow into the regime of linear waves. The red cascade is then not only blocked by Rossby wave propagation but it is reversed near western boundaries as fast long weak westward-propagating Rossby waves reflect at a western boundary into slow short strong eastward-propagating waves (Rhines, 1977).

Instabilities of fronts could play an important part. Such fronts can be formed by the same mechanisms which produce atmospheric fronts but according to Woods (1977, 1978), it is possible that they reach a limiting equilibrium form well before the larger scale velocity field which produced them has changed significantly.

The transfer of energy to internal waves has been strongly advocated. According to Müller (1976), for instance, internal waves could extract energy from synoptic eddies at about the same rate as that at which they gain energy from baroclinic instability of the wind - generated Sverdrup flow . Desquieting evidence against such a scheme was presented by Ruddick and Joyce (1979) from direct measurements of the vertical eddy momentum flux, due to internal waves, with moored current-meters and temperature sensors. They found no significant correlation with the mean shear and estimated an upper bound for the vertical eddy viscosity more than one order of magnitude smaller than Müller's suggestion (Garrett, 1979).

Following Panchev (1976), one can estimate the rate of energy transfer ε_β from the synoptic eddies to smaller turbulent oceanic scales as being of the order of

$$\varepsilon_\beta \sim 10^{-9}\ m^2 s^{-3} \tag{12}$$

i.e. of the same order as the atmospheric energy input into the largest oceanic scales (e.g. Ozmidov, 1965) but apparently one or two orders of magnitude smaller than the rate of energy transfer to larger scales through the red cascade (Panchev, 1976).

In the macroscale range, one expects - as a result of the two-dimensional turbulence enstrophy cascade - the energy spectral function $E(\kappa)$ to fall off as κ^{-3} i.e.

$$\kappa^3 E(\kappa) \sim \kappa_\beta^3\ E(\kappa_\beta) \sim 10^{-12} s^{-2} \tag{13}$$

Hence, at a scale of a few kilometers ($\kappa_w \sim 3\ 10^{-4} m^{-1}$, say) characteristic of eddies and intrusive layers which may emanate from fronts (Woods, 1978), the energy level would be, in the mean (all space and time intermittencies taken into account)

$$\kappa_w\ E(\kappa_w) \sim \kappa_\beta^3\ E(\kappa_\beta) \kappa_w^{-2} \sim 10^{-5} m^2 s^{-2} \tag{14}$$

Now, the rate of energy transfer from these scales, - which belong to the frontier districts between macro- and mesoscales -, into mesoscale turbulence can be estimated from turbulence similarity arguments. One finds

$$\varepsilon \sim \kappa_w \, E(\kappa_w)\omega_m \qquad (15)$$

where ω_m is the frequency of energy transfer from macroscale turbulence to mesoscale turbulence. Using eqs. (12) and (14), one gets

$$\omega_m \sim 10^{-4} \ s^{-1} \qquad (16)$$

This is precisely the characteristic frequency of the longuest mesoscale waves and one may speculate that the macroscale energy cascade is passed on to mesoscale turbulence through the instability of frontal eddies and intrusive layers transmuted by inertial, gyroscopic and tidal oscillations. The energy transferred into the mesoscales is found by eq. (12) to be only a small fraction of the energy which is apparently continuously recycled over the largest scales ($\kappa \sim 10^{-6}, 10^{-5} m^{-1}$) via transformations between kinetic and potential forms and exchanges between synoptic eddies, Rossby wave motions and gyres ; a very exclusive society, it would seem, which allows just enough bottom friction and limited leakage through physical space and Fourier space to maintain the energy balance with the atmospheric forcing.

MESOSCALE TURBULENCE

In the range of frequencies $10^{-5} \lesssim \omega \lesssim 10^{-2}$ one expects turbulent motions to be deeply intermingled with linear and non-linear waves related to tides, inertial oscillations and the general stratification of the ocean.

On the basis of turbulence, the ocean can be divided into three layers : (i) an upper mixed layer with a thickness $\sim 10^2$m which is continuously filled with turbulence generated by atmospheric factors working through the breaking of surface waves, drift currents and convection, (ii) an internal layer (practically the entire thickness of the ocean) in which only intermittent turbulence appears in the form of isolated patches or "blinis", (iii) a turbulent bottom layer with a thickness $\sim$ 10 m which is presumably similar to the atmospheric boundary layer.

Under stable stratification, turbulence loses energy in working against the buoyancy forces and turbulent mixing may become so difficult that, under natural conditions, it cannot extend to the whole water column and remains confined in distinct (well-mixed) "layers" separated by "sheets" where abrupt changes occur in temperature, velocity etc...

This is confirmed by experimental data and vertical profiles of temperature and other variables, almost everywhere in the ocean, show vertical step-like inhomogeneities generally referred to as the "fine structure" of the ocean.

As described by Woods (1977), the intermittent turbulence observed in the internal layers may be associated with trains of internal waves which, by locally increasing the vertical shear and reducing the Richarson number, allow turbulent patches to develop.

As a result of turbulent mixing, the water density becomes fairly homogeneous in a patch and in stably stratified surroundings, the density at the top becomes larger, and the density at the bottom smaller, than that of the ambient fluid. Under the action of buoyancy forces, the turbulent patch will then tend to flatten while it spreads aside by continuity forming a blini-shape intrusive layer contributing to the formation of the fine vertical structure.

Taking into account that there is an important input of energy in the ocean at the low frequency - low wave number end of the mesoscale range, through tidal and inertial oscillations, it is possible to conceive a coherent model of mesoscale motions based on the cohabitation of a chaotic field of internal waves and turbulent blinis.

Initially long waves will form large scale turbulent patches resulting in layers of great thickness. In such layers, internal waves of smaller periods and wave-lengths will develop forming turbulent patches of smaller dimensions and layers of smaller thickness and the process will continue, producing smaller and smaller scale motions down to the smallest waves and Kelvin-Helmoltz billows merging into three-dimensional turbulence.

The cascade "tidal-inertial waves → turbulent patches → fine structure layers → internal waves → turbulent patches → and so forth" is consistent with the observed spectra of ocean variability in the mesoscale range.

Another mechanism which could produce a fine vertical structure is double-diffusive convection.

The density of sea water being essentially a function of temperature and salinity, a given density distribution may mask important - but more or less compensating - variations of temperature and salinity.

Since the rates of diffusion of heat and salt are different, this situation allows potential energy to be released from the heavy component at the top. This type of "instability" can break smooth density gradients into a series of layers and interfaces with vertical transports accross the interfaces much larger than could be effected by classical diffusion down the mean gradients (e.g. Turner, 1973 a,b).

Double diffusive layering can be most effective near fronts when large anomalies of temperature and salinity may occur even with little net density differences.

According to Turner (e.g. Gill and Turner, 1979), quasi-vertical fluxes associated with double-diffusion processes can produce local density anomalies and so drive intrusive layers accross a front. Such layers could have vertical scales up to hundreds of meters and horizontal scales of several kilometers.

It is also possible that intrusions are produced by the dynamical instability of fronts and, unfortunately, experimental data are not yet sufficient to discriminate between the two mechanisms and assess their relative importance.

In any case, intrusive motions may be - as much as internal wave straining - an important cause of the observed fine vertical structure (Fedorov, 1978).

The role played by mesoscale fronts in ocean turbulence, has recently been emphasized by Woods (e.g. Woods, 1977 ; 1978) who suggested that frontogenesis is the inevitable outcome of the macroscale enstrophy cascade and that mesoscale fronts take over from the synopti eddies to transfer enstrophy to microscales.

Mesoscale fronts are formed by the synoptic scale deformation field ($\sim 10^5$m) acting on the gyre scale baroclinicity ($\sim 10^6$m). Hydrodynami instability of a mesoscale front produces meanders with wave-lengths λ ranging from a few tens of kilometers to a few kilometers (corresponding to typical wave-numbers $\kappa = 2\pi/\lambda$ of the order of $\kappa_w \sim 3\ 10^{-4} m^{-1}$) breaking into intrusive layers and eddies.

As mentioned before, such layers and eddies in the frontier districts between macroscale and mesoscale motions might provide the missing link between the synoptic eddies and the mesoscale blinis under direct influence of the stratification and the earth's rotation.

There is however an input of energy in the same range of scales and it must be taken into account.

According to Ozmidov (1965), this input corresponds to a rate of energy transfer of the order of

$$\varepsilon \sim 10^{-7} m^2 s^{-3}$$

How much of this energy affects the whole water column is still an open question.

One expects the effects of the tidal forcing to be felt at all depths but the amount of tidal energy which can be included in a turbulent cascade in the interior of the ocean appears rather uncertain (e.g. Garrett, 1979). An important fraction of the tidal energy is indeed dissipated on the continental shelf and in shallow coastal seas. Perhaps a gross estimate of $2\ 10^{-9}\ m^2 s^{-3}$ (Monin et al, 1977) may be retained for later comparisons.

(Bell (1975) has suggested a similar value for the energy flux into internal tides).

The effect of inertial oscillations is certainly not restricted to the upper ocean layers. Webster (1969) pointed out that inertial motions may ack like a kind of energy flywheel and explain the important and fairly permanent peak of kinetic energy about the inertial frequency.

Evidence of fairly energetic oscillations with frequencies close to the inertial frequency f, extending to the deep sea, have been given by several authors (e.g. Webster, 1968 ; Brekhovskikh et al, 1971 ; Perkins, 1972 ; Monin et al, 1977).

Knowing that this value is liable to be revised, one may perhaps write down an estimate

$$\varepsilon_f \sim 10^{-8}\ m^2 s^{-3} \tag{17}$$

for the rate of energy transfer through the mesoscales in the ocean interior (with a higher value of $10^{-7}\ m^2 s^{-3}$ in the upper layers).

Many experimental data seem to be reasonably well explained with values of ε of that order (e.g. Webster, 1969 ; Nihoul, 1979).

Frequency spectra of horizontal kinetic energy derived from long series of oceanic data show a marked narrow spectral peak at a frequency of the order of the inertial and tidal frequencies ($f \sim 10^{-4} s^{-1}$), followed by a gentle slope frequently fairly close to the classical - 5/3 line (e.g. Fofonoff, 1969 ; Webster, 1969 ; Nihoul, 1979).

If one considers, the largest scales on the gentle slope, at a frequency close to $f \sim 10^{-4}$ one may associate to them, by similarity arguments an energy level $\kappa_f E(\kappa_f)$ and a wave number κ_f of the order of

$$\kappa_f E(\kappa_f) \sim \varepsilon_f\ f^{-1} \sim 10^{-4}\ m^2 s^{-2} \tag{18}$$

$$\kappa_f \sim \varepsilon_f^{-1/2} f^{3/2} \sim 10^{-2}\ m^{-1} \tag{19}$$

for the value of ε given by eq. (17).

The corresponding eddies are not however the "energy containing eddies", corresponding to the spectral peak, which have about the same frequency but a level of energy about two orders of magnitude larger (e.g. Webster, 1969 ; Nihoul, 1979).

By similarity arguments, the energy containing eddies are characterized by a wave number κ_m given by

$$\kappa_m^3 \, E(\kappa_m) \sim \omega_m^2 \qquad (20)$$

i.e.

$$\kappa_m \sim \frac{\omega_m}{\left\{10^2 \kappa_f E(\kappa_f)\right\}^{1/2}} \sim 10^{-3} m^{-1} \qquad (21)$$

{With the higher value $\varepsilon \sim 10^{-7} m^2 s^{-3}$ (applicable to the upper layers), one finds, by the same calculation, $\kappa_m \sim 3\ 10^{-4} m^{-1} \sim \kappa_w$} .

In the range of frequencies between the Coriolis frequency and the maximum Brunt-Väisälä frequency ($10^{-4} \lesssim \omega \lesssim 10^{-2}$) turbulent eddies are intermingled with inertial-internal waves and although it should be possible to "see" fast internal waves passing through the turbulent eddies, internal waves are often so intense, interacting and diversified that it becomes difficult to separate the waves from the turbulence in a spectral analysis of current data. In fact, interactions and diversification of sources produce, in the internal wave field, an intricated collection of motions of various scales, a continuum of Fourier modes which appear as waves, wave packets or turbulent eddies depending on the distance they can propagate during their life-time (Nihoul, 1972) and which could very well be classed as turbulence if turbulence is defined as a "field of chaotic vorticity" (Saffman, 1968).

MICROSCALE TURBULENCE

In the range of frequencies $10^{-2} \lesssim \omega$, larger than the maximum Brunt-Väisälä frequency, the turbulence is no longer constrained by buoyancy and may be considered as three-dimensional.

Three-dimensional turbulence can be generated in the ocean by shear instability of local currents or by convection in layers with unstable density stratification.

The second mechanism would seem rather exceptional in the ocean where, unlike the atmosphere, the density stratification is always globally stable. However unstable layers may occasionally be produced by cooling of the ocean surface in the winter or by salt accumulation in the sub-surface waters during periods of intensive evaporation and, in deeper waters, by lateral instrusions (e.g. Fedorov, 1978).

One knows very little about convective turbulence in the ocean. It has been suggested that it might contribute something of the order of $10^{-8} m^2 s^{-3}$ to the rate of turbulent energy production ε, in some places.

Mechanical energy production is likely to be more widespread and to occur in all ocean currents whenever the local conditions of stability are not fulfilled. This process can be very important in boundary layers like the bottom boundary layer of the ocean where it may account for a rate of energy production ε larger than $10^{-8} m^2 s^{-3}$ (e.g. Nihoul, 1977), shallow continental seas where amplified tidal currents yield very high dissipation rates and in the upper mixed laye in association with drift currents generated by the wind.

In the upper mixed layer, however, the breaking of surface waves provides a generally more powerful mechanism for the generation of turbulence with rates of turbulent energy production of the order of $10^{-5} - 10^{-6} m^2 s^{-3}$.

In the deep ocean, local mechanical production of turbulence in macroscale and mesoscale currents is certainly less efficient and the largest estimate is a value of $\varepsilon \sim 10^{-7} m^2 s^{-3}$ for tidal and inertial non-stationary currents, with scales of the order of tens of kilometers (Lemmin et al, 1974).

Excluding boundary layers, the generation of three-dimensional turbulence in the ocean may thus be regarded as being largely the end product of the meso-scale cascade interpreted in terms of non-linear interactions of turbulent blinis or random, ultimately breaking, internal waves.

Microscale ocean turbulence will in general comprise an inertial range, where the energy spectral density $E(\kappa)$ is a function only of the wave number κ and of the energy ε transferred per unit time from one scale to the next, and a molecular range where molecular diffusivities play an essential role.

In the inertial range, similarity arguments predict the spectral law

$$E(\kappa) \sim \varepsilon^{2/3} \kappa^{-5/3} \tag{22}$$

and a similar $\kappa^{-5/3}$ dependence for the temperature fluctuations spectrum.

The inertial range exists for turbulent "frequencies"

$$\omega_\kappa \sim \kappa v_\kappa \sim \left(\kappa^3 E(\kappa)\right)^{1/2} \qquad (23)$$

smaller than the maximum Brunt-Väisälä frequency N_{Max} and larger than the viscous dissipation frequency

$$\omega_{\nu\kappa} \sim \nu\kappa^2 \qquad (24)$$

i.e.

$$N_{Max} << \varepsilon_3^{1/3}\kappa^{2/3} << \nu\kappa^2 \qquad (25)$$

The local rate of energy transfer - and ultimate dissipation rate - ε_3 in patches of microscale turbulence should be considerably larger than the mesoscale value but it is expected to average to the same order of magnitude once vertical intermittency is taken into account.

Taking a conservative value of 1 % for the intermittency factor, one estimates

$$\varepsilon_3 \sim 10^{-6} m^2 s^{-3} \qquad (26)$$

Then if $N_{Max} \sim 10^{-2} s^{-1}$ and $\nu \sim 10^{-6} m^2 s^{-1}$ eq.(25) is equivalent to

$$\kappa_N \sim 1\ m^{-1} << \kappa << \kappa_v \sim 10^3 m^{-1} \qquad (27)$$

so that the inertial range should be expected between typical scales of 1 metre and 1 millimetre.

Beyond κ_v, $E(\kappa)$ falls off rapidly as a result of viscous dissipation but the spectra of temperature and salinity fluctuations continue to higher wave numbers (with a κ^{-1} slope) before diffusion becomes significant and produces and exponential cut-off (Batchelor, 1959).

The predictions of the similarity analysis are confirmed by the observations in the ocean (e.g. Grant et al, 1968).

The persistence of temperature and salinity fluctuations of significant level beyond the viscous cut-off wave number κ_v is obviously a result of the smaller diffusivities of heat and salt.

The same persistence is likely to be observed in time, i.e., a patch of decaying turbulence will retain temperature and salinity heterogeneities of a given scale ℓ longer than velocity fluctuations of the same scale. The possibility of such "fossil turbulence" should not be overlooked when interpreting data based on temperature fluctuations.

THE CLIMATOLOGY OF OCEAN TURBULENCE

According to Munk (1966) a vertical eddy diffusivity of some $10^{-4} m^2 s^{-1}$ would be necessary to explain the vertical heat flux required in the ocean interior by global balances. Garrett (1979), reviewing the results of direct measurements, found that the experimental values were as a rule considerably smaller than Munk's theoretical estimate and suggested that perhaps isolated regions of very intense mixing existed which, somehow, the instruments had missed .

The discrepancy between the theoretical estimate and the experimental values cannot, according to Garrett (1979) be explained by the intermittency *in the vertical* (i.e. the unlikelihood of detecting a mixing event on a cast) but it might be related to the intermittency of mixing *in the horizontal* and *in time*.

The space and time intermittency of ocean turbulence is the inevitable consequence of the multiplicity of sources and physical processes acting over a wide range of scales. If one takes, for instance, the fine structure characteristic of the mesoscales, one finds that three different mechanisms at least may have generated the systems of layers and sheets. It is conceivable that all three operate in the ocean in different regions, or perhaps in the same region at different times, often under very different conditions. The question arises then whether it is possible to predict the mesoscale structure as a function of space and time or if one must regard the mesoscale turbulent blinis as appearing at haphazard with random sizes, locations and durations.

The same is true for energy dissipation or mixing intensity. Is it possible to survey the ocean, theoretically and experimentally and resolve the ocean dynamics, both in space and in time, with sufficient precision to associate numbers with given regions of the ocean and given periods of time ? Is this impossible and, then, is a statistical approach based on a limited number of indicative information the only feasible one ?

The answer may be yes to both questions.

To a certain extent, it should be possible to identify distinct ocean situations (semi permanent currents, intrusive plumes, down-and up-welling areas, regions of frontogenesis...), to ascertain the mechanisms that are actually operating and see that they are properly parameterized and to chart orders of magnitude showing the "climatology" of ocean turbulence.

To a large extent, however, this type of coarse climatology will not provide enough information on local instantaneous events to allow a deterministic approach. This means, to take a specific example, that it will not be possible to model the dispersion of a pollutant over scales going from 10^2m to 10^4m in the mesoscale range, taking into account the actual mechanisms which are responsible for the stirring, the mixing and the diffusion of the contaminant. Apart from measuring them simultaneously on the spot - in which case, it would be much simpler to measure the pollutant's concentrations directly - one can indeed have only statistical information of what they may be and what they may do. If these information are properly included into the parameterization, the model will be able to predict the extent of the contamination but not to reproduce the detailed dispersion pattern one might actually observe. Repeating the experiments many times in the same conditions and superposing the observations, however, a "mean" pattern would soon emerge which would look more and more like the model's predictions.

MODEL TURBULENCE FOR A MODELLED OCEAN

Modellers deal with averages taken over ensembles of identical oceans and, in doing so, they approach ocean variability with a resolute turbulence point of view.

Thinking in terms of turbulence, one likes to associate energy to each scale of motion and one tends to regard the transfer of energy from scale to scale as the essential mechanism of turbulence dynamics. The rate of downscale energy transfer ε - which may have very little to do with the rate of energy dissipation, at least its local value - is looked upon as the cogent factor which, associated with the wave number κ (the inverse of the length-scale), determines the strength of the turbulence and the efficiency of mixing.

This is of course nothing but the classical Kolmogorov theory of three-dimensional turbulence and many will undoubtedly object, to its application to the ocean, that the conditions of its validity are far from being satisfied in macroscale and mesoscale turbulence.

It is true that Kolmogorov's theory is essentially designed for mechanical turbulence in which large turbulent eddies are hydrodynamically unstable and disintegrate into smaller eddies transferring to them their kinetic energy (at a constant rate ε) until the eddies are small enough to be stabilized by viscous dissipation.

Ocean turbulence is evidently more complex but modellers will argue that the complicated processes which have been described as occuring in the macroscale and mesoscale ranges, simply explain, something which is self-evident in three-dimensional mechanical turbulence, i.e. why and how large eddies are unstable and finally transfer energy downscale to the dissipation range. The mechanisms of instabilities, subject to the geophysical constraints on the ocean, are obviously rather sophisticated. They involve conversion between kinetic and potential energy, energy transport by wave motions in physical space, trapped oscillations at the epidermis of turbulent blinis and, possibly, a form of "Echternacht procession" in the macroscales where energy goes from the large scale currents to the synoptic eddies and from them backwards - but not entirely - to the gyre circulation, via Rossby waves or more direct mechanisms.

But the detailed machinery is perhaps not important.

Macroscale quasi-twodimensional turbulence is overwhelmed by the enstrophy cascade to smaller scales, but the arresting result is the subsequent generation of fronts which generate large mesoscale eddies, - either directly or, indirectly, through the formation of intrusive layers -, of just the right scale to be grappled by inertial and tidal waves and initiate the cascade "layers - long waves - turbulent patches - layers - internal waves - turbulent patches ... microscale turbulence".

An individual turbulent blini, in the mesoscale range, does not "overturn", but turbulent blinis occur with random sizes, locations orientations and durations and an ensemble average is likely to reflect more the global randomness than any individual stiffness.

Internal waves and mesoscale turbulence are deeply intermingled, but the random field of interacting non-linear rotational internal waves and wave packets differs very little from turbulence and presumably less and less so as individual realizations of such fields are superimposed by ensemble averaging.

On the average, the ocean may turn out to be more turbulent than any particular oceanographic situation might suggest.

Admittedly, the values $\varepsilon_f \sim 10^{-8} m^2 s^{-3}$ and $\varepsilon_3 \sim 10^{-6} m^2 s^{-3}$ for the rates of kinetic energy transfer in the mesoscale and microscale ranges, respectively,are intentionally higher than current estimates trying to conciliate turbulent theories and observations. One must remember however that they are not intended to describe a particular, observed, situation but rather as ensemble averages appropriate to the parameterization of turbulence. The subjacent idea is that oceanic turbulence is highly *intermittent in time* and that the ocean cannot be adequately sampled in this respect ; the most intensive events being associated with severe weather conditions proscribing most of the oceanographic research.

Admittedly also, this is a wager.

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