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INTERNAL TIDAL MIXING

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Introduction

Any meaningful attempt towards understanding how the ocean works has to include an allowance for mixing processes. A great deal is known about the transport of heat and solutes by molecular processes in laboratory-scale experiments. Quite naturally, at the dawn of ocean science, these concepts were borrowed to speculate about the oceans. On that basis it was suggested by Zoeppritz that the temperature structure T(z) at 1000 m depth could be interpreted in terms of the time history T(t) at the surface some 10 million years earlier. It would indeed be nice if past climate could be inferred in this simple manner.

Once it was realized that molecular diffusion failed miserably to account for the transports of heat and salt, generations of oceanographers attempted to patch up the situation by replacing the molecular coefficients of conductivity and diffusivity by enormously larger eddy coefficients, but leaving the governing laws (equations) unchanged. By an arbitrary choice of the magnitude of the coefficients it was generally possible to achieve a satisfactory (to the author) agreement between theory and observation, a result aided by the uncertainty of the measurements. But the clear danger signal was there; each experiment, each process required a different set of values.

The present generation of oceanographers has come to terms with the need for understanding the mixing processes, just as they had come to terms some years ago with the need for understanding how ocean currents, ocean waves etc. are generated. Wind mixing and tidal mixing are very different processes. And the widespread use of parametrization of mixing processes will not succeed unless there is an underlying understanding of what is being parametrized. Here we have come a long way and have a long way to go.

Stirring and Mixing

In a Newtonian fluid the down-gradient flux of a quantity θ is given by

$$F_\theta = -\kappa\, d\theta/dx \qquad [1]$$

It is the very smallness of the molecular diffusivity κ which requires large gradients $d\theta/dx$ in order to attain significant fluxes F_θ. The fundamental distinction between stirring and mixing was first made in 1948. Stirring produces the gradients whereas molecular mixing reduces the gradients. For the purpose of this article stirring and mixing are both included in the discussion of the contribution of internal tides to mixing processes.

For an ocean in steady-state there needs to be an overall balance between the generation and dissipation of mean-square gradients. Nearly all of ocean dynamics deals with processes that generate gradients. This takes place over a wide variety of scales, all the way up to the scales of ocean basins. Dissipation takes place on the 'microscale,' i.e., millimeters to centimeters. This is the scale which includes the dominant contributions to the gradient spectrum. A further increase in the spatial resolution of the measurements does not lead to a significant increase in the measured mean-square gradients.

The Battle for Spatial Resolution

It is very difficult to attain a quantitative measure of mixing in the turbulent ocean interior. The problem is the need for very high spatial resolution. An eddy coefficient can be estimated as follows:

$$\kappa_{\text{molecular}} \times \text{rms}(d\theta/dx) = \kappa_{\text{eddy}} \times \text{mean}(d\theta/dx) \qquad [2]$$

(The subscript 'molecular' is introduced here to emphasize the distinction.) The ratio: mean square gradient/square mean gradient (the 'Cox number') has been used to estimate the ratio of the eddy coefficient to the molecular coefficient. A typical value away from boundaries is $\kappa_{\text{eddy}} = 10^{-5}\,\text{m}^2\,\text{s}^{-1}$, two orders of magnitude in excess of the molecular coefficient.

Achieving the required resolution has been a major accomplishment; but there are many problems with the measurements, and even more problems with the interpretation of the measurements along the lines of eqn [2]. It was only with the confirmation by a tracer release experiment that the community has come to accept the value $\kappa_{\text{pelagic}} = 10^{-5}\,\text{m}^2\,\text{s}^{-1}$ for the eddy diffusivity in the interior ocean, away from rough topography. There is of course considerable variability from place to place, but the surprising finding is not how large this variability is but how small it is.

Maintaining the Stratification

A quite different estimate of eddy diffusivity associated with pelagic mixing can be made from the following considerations. Bottom water is formed in the winter by convective overturning in just a few places: the Greenland Sea, the Labrador Sea and along the Antarctic continent. The formation is estimated at $Q = 25$ Sverdrups ($25 \times 10^6\,\text{m}^3\,\text{s}^{-1}$). This would fill the oceans with cold water in a few thousand years. The reason this does not happen is that turbulent diffusion downward from the warm surface balances the upwelling of cold water. With reasonable assumptions this leads to an estimate of eddy diffusivity $\kappa_{\text{stratification}} = 10^{-4}\,\text{m}^2\,\text{s}^{-1}$, ten times the measured pelagic value.

Measurements near topography do indeed give high diffusivities, orders of magnitude above the pelagic value. One simple interpretation is that there are concentrated areas of mixing (just as there are concentrated areas of bottom water formation) from which the water masses (but not the turbulence) are exported into the interior ocean. We can ask the question whether the global stratification can be maintained by vertical mixing in 10% (say) of the ocean volume with an average diffusivity 100 times the pelagic value?

The work done against buoyancy by turbulent mixing in a stratified fluid can be written

$$\varepsilon_b = \kappa(g/\rho)(-d\rho/dz) = \kappa N^2\,\text{W kg}^{-1} \qquad [3]$$

where N is the buoyancy frequency. Only a fraction γ (called the 'mixing efficiency') of the work goes into increasing potential energy (the rest goes into joule heat). A typical value is $\gamma = 0.2$. The total work per unit area is $\varepsilon_{\text{total}} = \varepsilon_b/\gamma$. For the world ocean of area A, the total work done is

$$D = A\int \rho\varepsilon_{\text{total}}\,dz = g\gamma - 1\kappa A\Delta\rho\ \text{W} \qquad [4]$$

where $\Delta\rho = 1\,\mathrm{kg\,m^{-3}}$ is taken as the difference between surface and bottom density. Then for $A = 3.6 \times 10^{14}\,\mathrm{m^2}$ and $\kappa = \kappa_{\mathrm{stratification}} = 10^{-4}\mathrm{m^2\,s^{-1}}$, one has $D = 2\,\mathrm{TW}$ (1 terawatt $= 10^{12}\,\mathrm{W}$) for the power required to maintain the global stratification in the face of 25 Sverdrups of bottom water formation. To maintain the pelagic turbulence requires only 0.2 TW.

Tidal Dissipation: The Astronomic Evidence

It is interesting to compare these numbers with the dissipation of tidal energy. We know this number with remarkable accuracy to be $2.5 \pm 0.1\,\mathrm{TW}$ for the principal lunar tide (M_2); it is obtained from the measured rate of $3.82 \pm 0.07\,\mathrm{cm\,y^{-1}}$ at which the Moon is moving away from the Earth. For all solar and lunar tides the dissipation is 3.7 TW, but with lesser certainty. We note that the tidal dissipation is of the same magnitude as the 2 TW required for maintaining the ocean stratification. Is this an accident?

The astronomic evidence tells us nothing about how and where the dissipation takes place. Allowing for dissipation in the solid Earth and atmosphere leaves 3.4 TW to be dissipated somewhere somehow in the ocean. Ever since it was estimated in 1919 that the dissipation in the Irish Sea is at 0.060 TW, the traditional sink has been in the turbulent bottom boundary layers of marginal seas, about 60 Irish Seas for the world. And before the astronomic estimates settled down to their present value, the ocean estimates kept rising and falling with the astronomic estimates.

Boundary Layer Dissipation Versus Scatter

When we speak of tides we usually refer to surface (barotropic) tides which have a nearly uniform current velocity from top to bottom, and a maximum vertical displacement at the surface. However, there is also a class of internal (baroclinic) tides with velocities that vary with depth and with maximum displacements in the interior.

A surface (barotropic) tide has essentially no shear in the interior ocean, There is shear near the bottom boundary, but the barotropic tidal velocities are so low in the deep ocean that the dissipation is negligible. In shallow seas the barotropic tidal currents are amplified, and the dissipation (proportional to the current cubed) is greatly amplified. This is the basis on which the global tidal dissipation has been attributed to the marginal seas.

Internal tides are part of a larger class of internal waves with frequencies other than tidal frequencies. A possible mechanism of tidal dissipation is the scattering of surface tides into internal tides, with subsequent transfer of energy into the broad spectrum of internal waves, and finally into turbulent dissipation: surface tides $\rightarrow$ internal tides $\rightarrow$ internal waves $\rightarrow$ turbulence. What is required at the second stage is some nonlinear frequency splitting which converts the low-frequency tidal line spectrum into a closely packed high-frequency continuum that resembles the observed internal wave spectrum. The final step is associated with the fact that the internal wave spectrum is at or near instability in the Richardson sense: the means-square shear is roughly $4N^2$ (N is the buoyancy or Brunt-Väisälä frequency).

Scattering of surface tides into internal tides can take place along wavy bottoms. A second possibility is scattering along submarine ridges. An acoustic tomography experiment north at Hawaii detected internal waves of tidal frequency radiating northward.

Satellite Altimetry to The Rescue

A subsequent analysis of satellite altimetry clearly showed internal tides emanating from the Hawaii submarine ridge. This is shown in **Figure 1**. The radiated power was estimated at 0.015 TW. So 14 Hawaiian chains will radiate 0.2 TW, enough to power the pelagic mixing associated with $\kappa_{\mathrm{pelagic}} = 10^{-5}\mathrm{m^2\,s^{-1}}$. The discovery of internal tide signatures by means of satellite altimetry was an altogether unexpected dividend from a technology that has revolutionized tidal analysis. The global sampling of surface elevation has introduced a new element into a subject that had gone to bed (in the opinion of some) with the work of Victorian mathematicians. With the Laplace tide equation as a guide for the tidal response of a nondissipative ocean, the assimilation of TOPEX/POSEIDON altimetry can lead to estimates of where one needs to introduce dissipation for agreement with the satellite data.

The most recent result allocates roughly 1 TW to the open ocean, mostly over rough terrain. The tentative conclusion is that tidal dissipation is a significant factor in open ocean turbulent mixing.

Supporting evidence comes from the measurements of tracer dispersion and microstructure in the Brazil Basin. Diffusivities of $\kappa = 2 \times 10^{-4}$–$4 \times 10^{-4}\mathrm{m^2\,s^{-1}}$ at an elevation of 500 m above

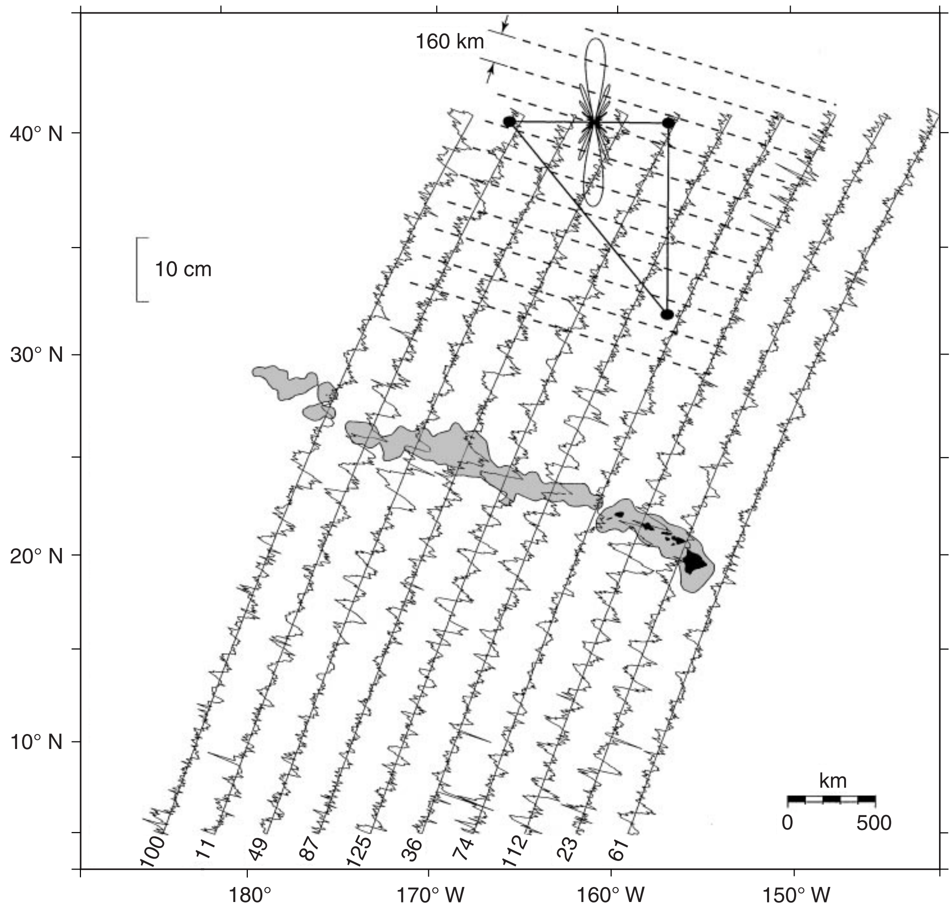

Figure 1 Surface manifestation of internal M_2 tides emanating from the Hawaii'an Island Chain. (Reproduced from Ray and Mitchum 1997.) The wiggly curves show the amplitudes along the ascending orbits of TOPEX/POSEIDON, with positive elevations on the north side. The dashed lines are the inferred crests of the mode 1 component of internal tides. Background shading corresponds to bathymetry, with darker areas denoting shallower water. The triangle to the north east shows the position of the tomographic array. Adapted from Ray & Mitchum (1997).

the abyssal hills of the Mid-Atlantic Ridge, increasing to $10 \times 10^{-4}\,m^2\,s^{-1}$ near the bottom have been obtained. Perhaps the most important result is that over a period of a month the diffusivities vary by a factor of two, with the large values occurring at spring tide and the small values at neap tide.

Discussion

There is more than enough tidal dissipation to feed the measured pelagic turbulence associated with $\kappa_{\text{pelagic}} = 10^{-5}\,m^2\,s^{-1}$. With regard to the larger value $\kappa_{\text{stratification}} = 10^{-4}\,m^2\,s^{-1}$, the present best estimates would suggest that tidal dissipation could power half the turbulence needed to account for the observed ocean stratification. **Figure 2** attempts an allocation of tidal energy flux, but there are many uncertainties, some by factors of two or more. The assumed one-dimensional balance between upward advection and downward diffusion as a measure of $\kappa_{\text{stratification}}$ is itself somewhat uncertain. However, the present conclusion is that tidal dissipation is a significant, possibly dominant factor driving mixing in the ocean interior.

The combination of detailed *in situ* measurements of turbulent mixing subject to a global lid on available tidal energy has led to giant strides towards a meaningful parametrization of ocean mixing, whether or not tidally induced. The conversion of wind energy to turbulent mixing plays a major role, particularly in the upper oceans. In **Figure 2**, equal weight has arbitrarily been assigned to tides and winds. We shall have to await the outcome of this competition.

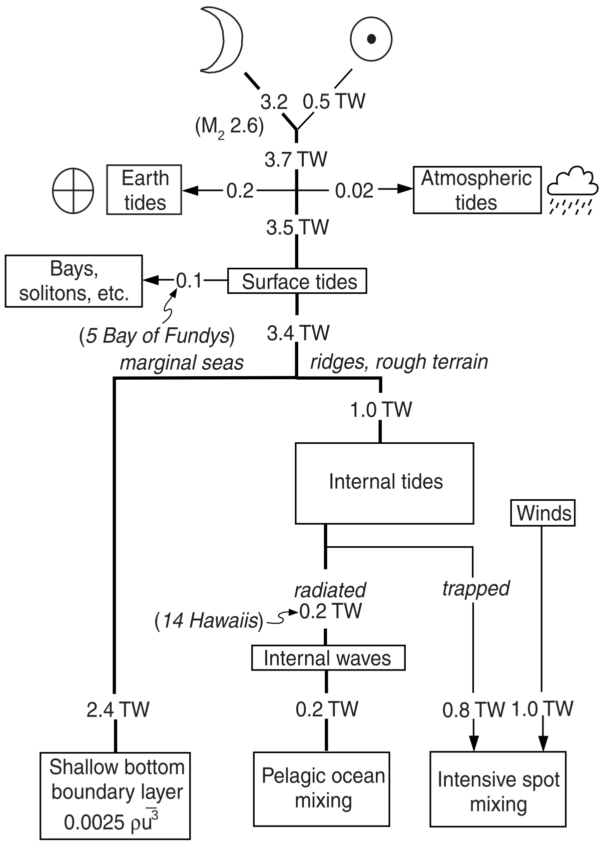

Figure 2 Sketch of proposed flux of tidal energy (modified from Munk and Wunsch, 1997). The traditional sink is in the turbulent boundary layer of marginal seas. Scattering into internal tides over ocean ridges (by the equivalent of 14 Hawaii's) and subsequent degradation into the internal wave continuum feeds the pelagic turbulence at a level consistent with $\kappa_{\text{pelagic}} = 10^{-5}\,\text{m}^2\,\text{s}^{-1}$. Most of the ocean mixing is associated with a few concentrated areas of surface to internal mode convergence over regions of extreme bottom roughness and with severe wind events. Light lines represent speculation with no observational support.

See also

Dispersion and Diffusion in the Deep Ocean. Internal Tides. Internal Waves. Tides. Turbulence in the Benthic Boundary Layer.

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INTERNAL TIDES

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Introduction

Oceanic internal tides are internal waves with tidal periodicities. They are ubiquitous throughout the ocean, although generally more pronounced near large bathymetric features such as mid-ocean ridges and continental slopes. The internal vertical displacements associated with these waves can be extraordinarily large. Near some shelf breaks where the surface tides are strong, internal displacements (e.g., of an isothermal surface) can exceed 200 m. Displacements of 10 m in the open ocean are not uncommon. The associated current velocities are usually comparable to or larger than the currents of the surface tide. Internal tides can occasionally generate packets of internal solitons which are detectable in remote sensing imagery. Other common nonlinear features are generation of higher