models and measurements suggest that typical energy fluxes at shelf breaks are of order $100\,\mathrm{W\,m^{-1}}$, leading to a global total of order 15 GW. This is perhaps an underestimate, because it may not fully account for shelf canyons and other three-dimensional features, but the order of magnitude seems reliable. Internal tide generation by deep-ocean topography, however, may be far more important. Recent research based on global tide models as well as on empirical estimates of tidal dissipation deduced from satellite altimetry suggests that generation of internal tides by deep-sea ridges and seamounts could account for 1 TW of tidal power. Refining such estimates, and understanding the role that internal tides play in generation of the background internal wave continuum, in vertical mixing, and in maintenance of the abyssal stratification, are some of the outstanding issues of current research.

See also

Acoustics in Marine Sediments. Internal Tidal Mixing. Internal Waves. Satellite Altimetry. Tides.

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INTERNAL WAVES

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Introduction

Waves at the sea surface are a matter of common experience. Surface tension is the dominant restoring force for waves with a wavelength less than 17 mm or so; longer waves are more affected by gravity. They have periods up to about 20 s and amplitudes that may be many meters.

Given the stable density stratification of the ocean, it is not surprising that there are also 'internal gravity waves,' with a water parcel displaced vertically feeding a gravitational restoring force. The wave periods depend on the degree of stratification but may be as short as several minutes and can be long enough that the Coriolis force plays a major role in the dynamics. Vertical displacements are typically of the order of ten meters or so, with horizontal excursions of several hundred meters. The associated horizontal currents are typically several tens of millimeters per second. An interesting difference from the surface wave field is that internal waves always seem to be present, without the intense storms or periods of calm that exist at the surface.

The existence of internal waves complicates the mapping of average currents and depths of particular density surfaces. They have also been the objective of intensive military-funded research because of the possibility that wakes of internal waves generated by submarines might be detectable by remote sensing, thus betraying the submarine's location. More conventional acoustic means of submarine detection are complicated by the deflection of acoustic rays by the rather random variations in sound speed induced by internal waves. In civilian activities, the currents and buoyancy changes associated with internal waves are a matter of concern in offshore oil drilling.

Most importantly, perhaps, the current shear of internal waves, including those of tidal frequency, can lead to instability and turbulence, and so the waves are the main agent for vertical mixing in the ocean interior. This mixing plays a major role in determining the strength of ocean circulation, and hence the poleward heat flux and climate. The mixing, along with the associated circulation, also provides nutrient fluxes into the sunlit upper ocean where primary biological production occurs. Understanding internal waves is thus of vital importance, particularly since they occur at too small a scale to be treated explicitly in computer models of the ocean. Their effects must be 'parametrized,' or represented by formulas that involve only the quantities that are carried in the model. In this respect, internal waves in the ocean are somewhat akin to clouds in the atmosphere – they play a vital, perhaps even controlling, role in global-scale problems. (The atmosphere also has internal waves, of course, which are known to play a major role in redistributing momentum.)

This short article will first describe the waves that can occur at sharp density interfaces in clear analogy to waves at the sea surface. This will be followed by a description of the waves that can propagate through a continuously stratified ocean, and a discussion of their generation, evolution, and relationship to ocean mixing.

Interfacial Waves

If the ocean consists of an upper layer of density $\rho - \Delta\rho$ and thickness h_1 above a layer of density ρ and thickness h_2, then waves that have a wavelength much greater than both h_1 and h_2 travel at a speed $[g'h_1h_2/(h_1 + h_2)]^{1/2}$ independent of wavelength, where $g' = g\Delta\rho/\rho$ is known as the 'reduced gravity'. If $h_2 \gg h_1$, this becomes $(g'h_1)^{1/2}$, in clear analogy to the speed $(gh)^{1/2}$ for surface waves that are long compared with the water depth h. This formula for the speed of interfacial waves also holds for $h_2 \gg h_1$ even if the wavelength is not long compared with h_2.

The theory behind this requires that the amplitude of the waves is much less than the thickness of the layers. Many observed interfacial waves (**Figure 1**) violate this assumption and also the requirement that their wavelength is long compared with the

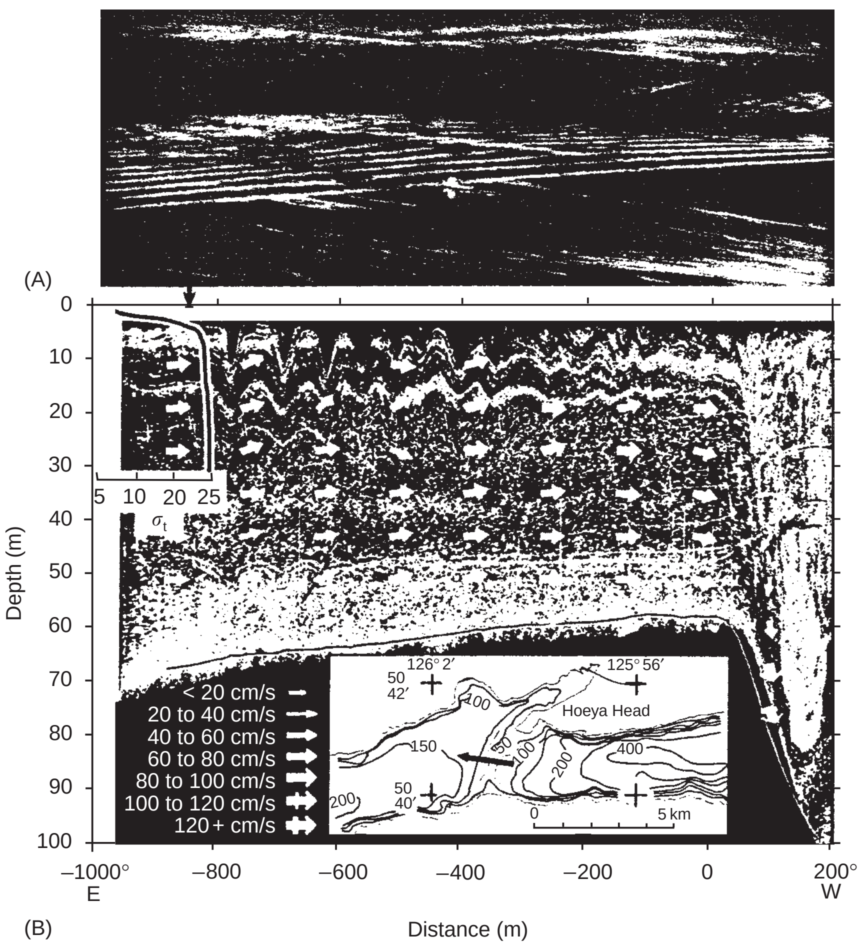

Figure 1 A group of interfacial solitary waves generated by tidal flow over the sill at Knight Inlet, British Columbia. (A) The banded surface manifestation. (B) An echo-sounder image of the same waves. Current vectors are also shown. The location and ship track are shown in the bottom inset. The upper left inset of the density (σ_t) profile shows a strongly stratified thin upper layer above a much thicker more homogeneous layer, rather than an ideal two-layer situation. (From Farmer D and Armi L (1999) The generation and trapping of solitary waves over topography. *Science* 283: 188–190; courtesy of D. Farmer.)

layer thicknesses. Finite amplitude is associated with a tendency for waves to steepen, much as in the development of a tidal bore at the sea surface. On the other hand, a horizontal scale that is not very long, compared with at least the thickness of the thinner layer, leads to dispersion, the break-up of a disturbance into waves of different wavelengths traveling at different speeds. Interestingly, these effects can cancel, leading to the possibility of 'internal solitary waves', waves of finite amplitude that can be spatially localized and travel without change of shape. They can occur singly, or in groups as in **Figure 1**. If in a group, the crests pointing away from the thinner layer are sharper than the troughs. Even if they occur at a density interface many meters, or tens of meters, below the surface, they are often visible if the upper layer is turbid, so that the crests appear from above as more opaque tubes than the surrounding water. More frequently they are seen because the associated currents cause visible variations in surface roughness (**Figure 1A**). (Whether the water is rougher above the crests or troughs of the interfacial waves depends on the relative directions of propagation of the surface waves and the interfacial waves, as can readily be established by considering the interaction in a frame of reference moving with the interfacial waves.)

The generation of these packets of internal solitary waves, or trains of waves with similar properties, often occurs when tidal flow over a sill, or off the edge of the continental shelf, leads to a leeward depression in the interface (as in **Figure 1**). As the tidal current reverses, this depression propagates back over the sill, or onto the continental shelf, and breaks up into large amplitude interfacial waves. (In the situation shown in **Figure 1**, interfacial waves have actually formed before the current reversal.)

The internal solitary waves, or solitons, typically have periods of tens of minutes. This would appear to be too short for the Earth's rotation to be a factor, but it does seem that the break-up of an internal tide into internal solitons may be inhibited by rotational effects.

While it is generally only the shape of these interfacial waves that propagates, with little net water movement, they can be sufficiently large that they do, in fact, carry water along with them. Remarkable behavior also occurs as the waves approach shore: although they have sharp downward crests offshore where the lower layer is thicker, they must switch to having sharp upward crests as the lower layer becomes thinner than the upper layer! They do this only with considerable loss of energy into smaller-amplitude dispersive waves and turbulence.

The waves described above have a vertical motion that is maximum at the interface and tends to zero at top and bottom boundaries. The horizontal currents in the two layers are in opposite directions. A sharp interface is not necessary; large-amplitude internal motions can persist even if the density jump is smeared out in the vertical. In the case of small-amplitude waves, the motion can then be thought of as the first vertical mode, made up of propagating waves that reflect off the sea surface and seafloor. We therefore turn next to a discussion of these building blocks.

Internal Waves

The Basic Physics

A particle displaced vertically in a continuously stratified fluid experiences a restoring buoyancy force. If a whole vertical fluid column is displaced, the vertical uniformity of the motion means that there is no change in the hydrostatic vertical pressure gradient and the restoring force on each particle is just gravity g times the density perturbation, which is minus the vertical displacement times the vertical density gradient $\mathrm{d}\rho/\mathrm{d}z$. This leads to a simple harmonic oscillator equation for the motion of the column, with the frequency of oscillation given by the 'buoyancy frequency' N, where $N^2 = -(g/\rho)(\mathrm{d}\rho/\mathrm{d}z)$. This frequency is independent of the horizontal scale of the fluid columns, suggesting that the frequency of motions that are wavelike in the horizontal is N, independently of scale.

If the fluid columns are now allowed to oscillate obliquely, at an angle θ to the vertical, the vertical restoring force is reduced by a factor $\cos\theta$, as is the component of this force parallel to the motion. The extra factor $\cos^2\theta$ in the oscillator equation then means that the frequency is reduced to $N\cos\theta$, again independently of the lateral scale.

Regarding these motions as waves, it is clear that the motion is transverse, as required for an incompressible fluid with $\nabla\cdot\mathbf{u} = 0$. In a rotating world the motion is acted upon by the Coriolis force, so that fluid oscillations in inclined sheets now develop a transverse motion, within the sheet but orthogonal to the motion with no rotation. The relationship between the frequency ω of the oscillations now involves the Earth's rotational frequency. Provided that N is sufficiently greater than the Coriolis frequency f, which is twice the vertical component of rotation, the connection between frequency ω and orientation θ becomes eqn [1].

$$\omega^2 = N^2\cos^2\theta + f^2\sin^2\theta \qquad [1]$$

Equation [2] is an alternative expression, in terms of the wavenumber $\mathbf{k} = (k, l, m)$.

$$\omega^2 = \frac{N^2(k^2 + l^2) + f^2 m^2}{k^2 + l^2 + m^2} \qquad [2]$$

While the frequency can be as high as N when the particle motion is vertical, it cannot be lower than the Coriolis frequency f. In this limit, the particle motion is horizontal in 'inertial' circles, expressing the tendency for steady rectilinear motion with respect to a nonrotating reference frame. Any frequency of motion between these limiting frequencies is possible, depending on θ, or, equivalently, the ratio of vertical to horizontal wavenumber. Moreover, at any frequency, any wavelength is possible.

The group velocity (the velocity with which a wave packet, or energy, propagates) is given by $(\partial\omega/\partial k, \partial\omega/\partial l, \partial\omega/\partial m)$. For internal waves this is easily shown from (2) to be at right angles to the wavenumber vector $\mathbf{k}$. In other words, energy propagates parallel to the wave crests, rather than at right angles as for surface waves! This remarkable feature can be demonstrated in a laboratory experiment (**Figure 2**). In terms of vertical propagation, waves with downwards phase propagation have upwards energy flux, and vice versa.

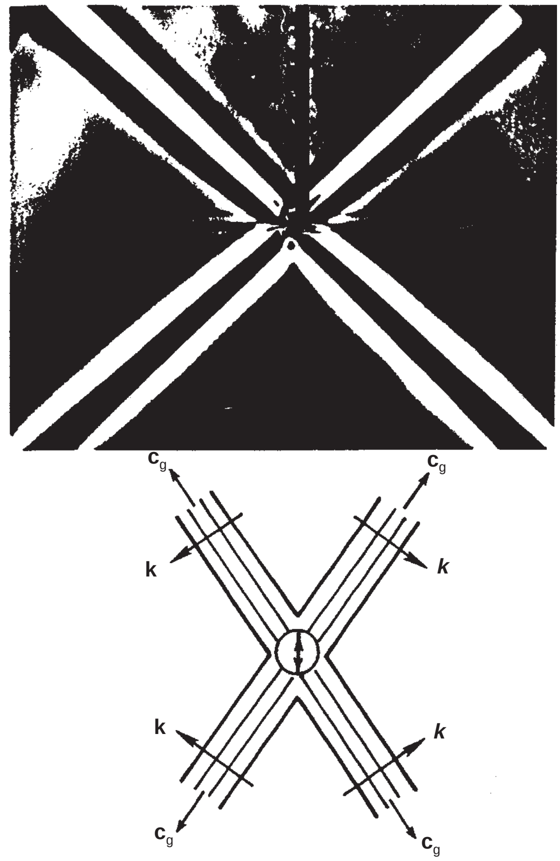

Figure 2 Internal waves of a fixed frequency less than N are generated by vertical oscillations of a wavemaker in a tank of stratified fluid. In the photograph the light and dark radial lines are contours of constant density perturbation and so are lines of constant wave phase. The schematic diagram shows the directions of wave phase propagation (**k**) and group velocity ($\mathbf{c_g}$).

Observations

Measurements of the frequency spectra of internal waves can be obtained from analysis of time-series of measurements, at a fixed point, by current meters or by temperature sensors that show changes associated with vertical motion of the temperature-stratified water. Such measurements do show a block of energy at frequencies between f and N, falling off above and below these frequencies and with an energy distribution in between that seems close to ω^{-2}. For currents there is typically an extra peak (an 'inertial cusp') near f, but this is suppressed in temperature data as the near-inertial motions are largely horizontal.

Measurements at a single fixed point do not, however, provide information on the wavenumber content, or spatial scales, of the energy at any frequency. For this one needs information from many locations (such as from many current meters on a mooring) or the continuous vertical profile obtainable from an acoustic Doppler current profiler (ADCP). Current meter arrays are, of course, limited by the cost and logistics of deploying large numbers of instruments, and moored ADCPs with good resolution have a range much less than the depth of the ocean.

Invaluable high-resolution vertical profiles of horizontal currents over the whole ocean depth have been obtained from dropped or lowered profiling current meters that measure the tiny electric potentials generated by movement of conducting sea water in the Earth's magnetic field, and also by lowered ADCPs. These techniques do not, however, provide much information on the frequency content of the motions. Further information has also come from horizontal tows of sensors, or arrays of sensors, hence mapping horizontal scales though, again, not providing frequency information.

Various syntheses of the information from these types of experiments have shown a tendency for energy to be distributed in vertical wavenumber and frequency somewhat as shown in **Figure 3**: the frequency dependence is roughly like ω^{-2}, and at each frequency there is a tendency for there to be more energy at small vertical wavenumbers (large scales), with a roll off to high wavenumbers with a power

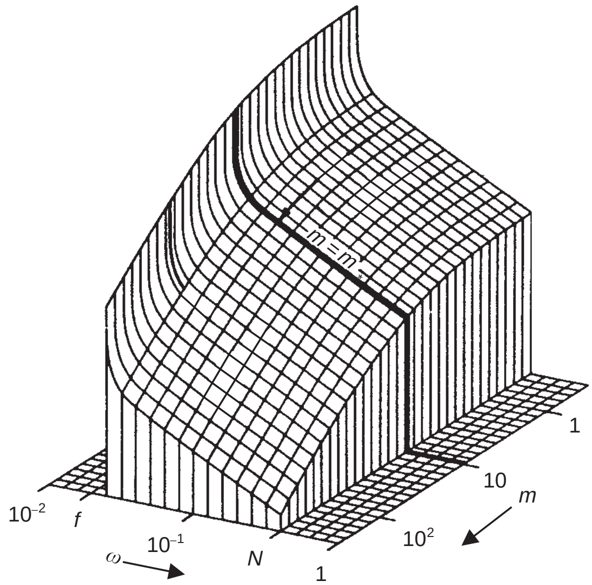

Figure 3 The so-called 'Garrett–Munk' spectrum, giving the distribution of energy in a space defined by frequency ω and vertical wavenumber m. The spectrum has a peak at the inertial frequency f and falls off like ω^{-2} at higher frequencies up to N. In vertical wavenumber the spectrum is fairly flat at small wavenumbers (large scale), then falls off rapidly to high wavenumber. The scales are as multiples of f for ω and in terms of an equivalent vertical mode number (number of half-wavelengths in the ocean depth) for m.

law like m^{-2} or $m^{-5/2}$, though the exponent here is certainly not well-established or universal. While a given frequency and vertical wavenumber magnitude are associated with a given magnitude of horizontal wavenumber via [2], the direction of the wavenumber is not specified, either up versus down or in the 360° available horizontally. It is generally assumed that the energy is horizontally 'isotropic,' or distributed evenly among all possible directions of propagation. It is also assumed that there is as much energy propagating up as down, though there is evidence for preferential downward transmission for waves with frequency within about 10% of f at mid-latitude.

The inertial peak, in fact, deserves special consideration, given its dominance. One interesting aspect is that the current vectors spiral with depth, with a connection between the direction of rotation of the spiral and the direction of energy propagation: in the Northern Hemisphere currents rotate clockwise with time, so that a current vector profile that shows increasing clockwise rotation with increasing depth below the sea surface must have phase propagation upward and hence group, and energy, propagation downward.

The model spectrum shown in **Figure 3** is only a very rough approximation. Observed spectra typically have considerable additional energy at tidal frequencies, with this energy also distributed over various vertical scales.

Generation

It seems that the inertial peak may be generated by fast-moving storms that set up currents in the upper ocean with the corresponding Coriolis forces unmatched by pressure gradients. Near-inertial motions result, with the current vectors rotating with a frequency close to the local f. The large horizontal scale of these motions means, however, that they experience different f at different latitudes. The current vectors at different latitudes then rotate at different rates, increasing the latitudinal gradients in current vectors and decreasing the horizontal scale. The resulting convergence and vertical motion means that the waves can no longer be purely inertial; they retain their frequency but propagate equatorward to a region where f is smaller. At the same time they develop an increasing vertical group velocity and propagate downward into the ocean. The evolution of this important near-inertial part of the internal wave spectrum is also affected by wave interactions with lower-frequency eddies.

Higher-frequency waves may be generated by storms that move more slowly, by turbulence in the surface mixed layer, by subtle interactions between surface waves, or as part of the decay process of ocean eddies. They may also arise from interactions between preexisting internal waves, as will be discussed shortly.

The tides are another important source of energy for internal waves observed throughout the ocean, as already discussed for interfacial waves near the sea surface. The barotropic, depth-independent, tidal currents associated with tidal changes in sea level move density-stratified water over topographic features on the seafloor, setting up internal oscillations much as if the topographic features were oscillating wavemakers in an otherwise still ocean. These 'internal tides' are mainly at the tidal frequencies, though there may also be energy at multiples of these.

Lower-frequency currents in the deep ocean are generally much weaker than tidal currents, but in areas, such as the Southern Ocean, where they are significant, they may set up quasi-steady 'lee waves,' or standing internal waves behind topographic features (as occurs in the atmosphere). These may propagate upward into the ocean.

The relative importance of these different mechanisms on a global basis has not been established, but the current view is that wind and tides are the dominant sources and are of comparable importance.

Evolution

A number of processes can contribute to the filling in of the continuous spectrum typically observed. One seems to be resonant wave–wave interactions: the nonlinear terms, involving $\mathbf{u} \cdot \nabla$, in the governing fluid dynamical equation vanish identically for a single wave, but produce interaction terms if two waves are present. These terms, in the momentum and density equations, may be regarded as forcing terms with frequencies and wavenumbers given by the sum and difference frequencies and wavenumbers. For some pairs the sum (or difference) frequency is exactly what would be expected for a free wave with the sum (or difference) wavenumber, so this wave is now resonantly excited, acquiring energy from the original two waves. Detailed calculations for this theory, and using a different approach when the assumptions of weak interaction break down, do not actually make it clear how a typical spectrum arises, but suggest that, once it is resent, there is a cascade of energy to waves with shorter vertical scales. As will be discussed later, these shorter waves are more likely to become unstable, break down into turbulence, and cause mixing.

The direction in which energy flows in frequency is less clear, though one interaction mechanism, akin to the excitation of a simple pendulum by oscillation of its point of support with twice the natural frequency of the pendulum, can produce small-scale waves with half the frequency of a large-scale parent wave (provided that this half-frequency is still greater than f).

Bottom Reflection and Scattering

Internal waves have a frequency less than the local value of the buoyancy frequency N. This typically decreases with increasing depth below the sea surface, so that some downward-propagating waves must undergo internal reflection at a level where their frequency matches the local N. Waves with a frequency less than N at the seafloor (or just above some well-mixed bottom boundary layer) will be scattered and reflected there.

The reflection process is unlike that for, say, sound waves, in that for internal waves to conserve their frequency on reflection, they must, by eqn [1], conserve their angle to the vertical, not their angle to the normal. As a consequence, waves reflected

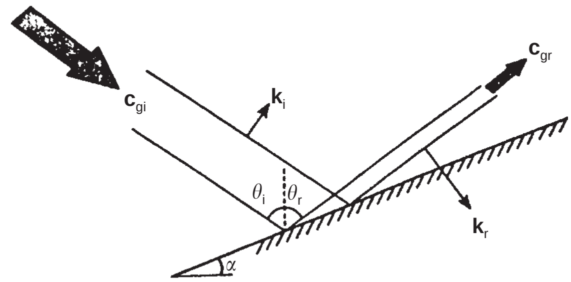

Figure 4 Internal wave rays are reflected at the seafloor at an equal angle to the normal to a bottom slope. Subscripts i and r indicate incident and reflected waves.

upslope will have a shorter wavelength and narrower ray tube (**Figure 4**). The latter effect, combined with a reduction in group velocity, causes an increase in wave amplitude, particularly near the 'critical frequency' for which the wave rays are parallel to the slope. The waves may be amplified less, or even reduced, for other azimuthal angles of incidence, but it turns out that overall amplification is expected for an isotropic incident spectrum, as has been well documented for internal waves near the steeply sloping sides of Fieberling Guyot in the Pacific Ocean.

This analysis certainly applies if the length scale of the slope is large compared with the wavelength. For smaller-scale topography, some energy may be backscattered without as much amplification, but, in general, internal wave interaction with bottom topography tends to redistribute energy toward shorter wavelengths. This may be just as important as wave–wave interactions in shaping the wavenumber part of the internal wave spectrum, though bottom interactions do not affect the frequency distribution.

Energetics and Mixing

The general picture that has emerged for internal waves in the ocean is that there is a cascade of energy to smaller scales as a consequence of wave–wave interactions and bottom scattering. This leads to a tendency for shear instability of the horizontal currents, with an expected vertical scale of the order of 1 m for a typical spectrum. It is generally assumed that a fraction of about 15–20% of the energy lost in the breaking leads to an increase in the potential energy of the water column (with the rest of the energy being dissipated and ultimately appearing as a negligible internal heating rate). The associated vertical mixing rate, or 'eddy diffusivity,' is of the order of $10^{-5}\,\mathrm{m^2\,s^{-1}}$, again for typical spectral energy levels in the main thermocline. This agrees rather well with estimates based on measurements of turbulent microstructure, and with even more direct estimates based on observations of the vertical spread of an artificial tracer.

The mixing may be considerably more intense throughout the water column in regions, such as the Southern Ocean, where internal wave energy levels seem higher, perhaps as a consequence of seafloor generation of the waves by strong mean currents over rough topography. Stronger mixing is also observed in general within a few hundred meters of the bottom in areas of rough bottom topography, though it is not clear whether this results directly from the increased shear of the reflected and scattered waves, or via stronger wave–wave interactions at increased internal wave energies. The relative importance of wind-generated internal waves and internal tides in these regions is also still unsettled, though a topic of active research.

Rather weak mixing in the main thermocline of the ocean, together with much weaker stratification in abyssal areas of strong mixing, means that the overall energy loss from the internal wave field is small enough that it would take many tens of days to drain the observed energy levels. This may tie in with the remarkable feature, mentioned earlier, that observed internal wave energy levels in the ocean seem to be rather uniform in space and time, at least much more so than for surface gravity waves; there is no such thing as an 'internal calm.' The interpretation is that the decay time of internal waves is, unlike the situation for surface waves, considerably longer than the interval between generation events. There is still some seasonal modulation of the internal wave energy levels, but less than that in the wind and also in accord with a decay time of tens of days.

There are exceptions to this picture, of course, with, for example, much lower internal wave energy levels and mixing in the Arctic Ocean (except near some topographic features), perhaps as a consequence of less wind generation, because of the protective ice cover, as well as rather weak tidal currents.

The overall dynamical balance of the internal wave field in the ocean is qualitatively summarized in **Figure 5**.

Internal Waves on the Continental Shelf

The above discussion of internal waves has been focused on the deep-sea situation. There are some similarities on the much shallower continental shelves, though with a considerable fraction of observed internal wave energy often being associated with the rather large interfacial waves, or their equivalent in a smoothly stratified fluid, discussed at the beginning of this article. It is not clear how much of the rest of the internal wave field on shelves is locally generated and how much propagates there from the deep ocean.

Other Aspects

In the atmosphere, internal waves are crucial in establishing the general circulation by transporting momentum from one location to another and then depositing it when they break. One reason for this breaking is that as internal waves propagate vertically into thinner air they must increase their amplitudes in order to conserve their energy flux, and so become more prone to instability. This is not a factor in the oceans, where the density change is very minor. The ratio of mean flow speeds to wave speeds is also less in the ocean than in the

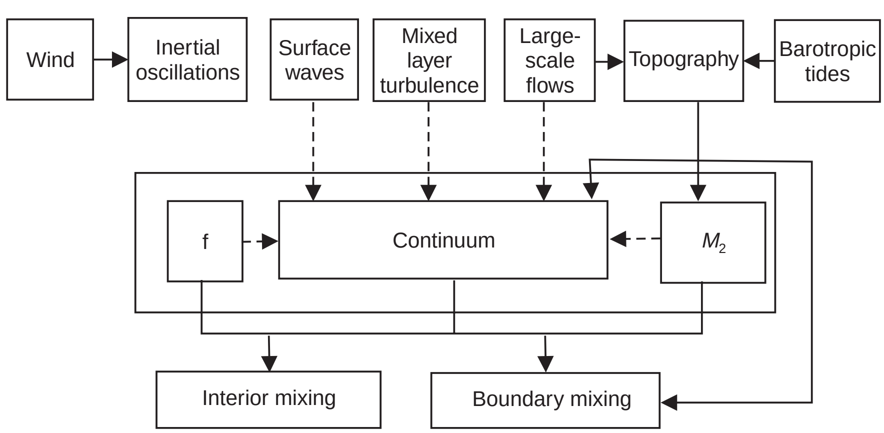

Figure 5 A summary of internal wave generation, evolution, and eventual dissipation causing mixing in the ocean interior or near boundaries. The dashed lines indicate conjectured energy pathways. (From Müller and Briscoe (2000).)

atmosphere, making interactions between waves and currents less important in the ocean. None the less, it does seem likely that there are some locations in the ocean where internal wave breaking should drive mean flows. One possible location is the continental slope; internal waves generated as lee waves at one location may propagate shoreward, break on the slope, and drive an along-slope current, much as ocean swell incident at an angle to a beach may drive longshore currents inside the breaker zone.

The role of internal waves in other situations may also have been somewhat unrecognized so far. One is their effect on surface mixed layer deepening. The waves alternately shallow and deepen the layer, making the shear across the mixed layer base more destablizing during the shallow phase and enhancing the overall mixing. This is an effect that has been excluded from models of the surface layer, and may be a partial reason why these models sometimes need to include ad hoc extra mixing just below the base of the layer.

Conclusions

Internal waves are both an unavoidable nuisance and a key ingredient of the behavior of the ocean; perhaps they are like the clouds in the atmosphere. The analogy is certainly a good one when one thinks of modeling the large-scale circulation of the two media for applications such as climate prediction. Numerical models fail by several orders of magnitude to have sufficient resolution to treat them explicitly, so their effects must be parameterized. This requires not just an understanding of their role in present conditions, but also a submodel that will predict their characteristics and effects in a changing mean state. A model for internal waves will need to account for the whole awkward mix of generation, propagation, wave–wave interactions, interactions with the mean state, and reflection and scattering from the rough seafloor. We have a partial understanding of many of the pieces but are a long way from putting them all together.

See also

Breaking Waves and Near-surface Turbulence. Internal Tidal Mixing. Internal Tides. Surface, Gravity and Capillary Waves. Wave Generation by Wind.

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INTERNATIONAL ORGANIZATIONS

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Introduction

This article is limited to a description of the most important organizations related to Ocean Sciences, rather than exhaustively cataloging perhaps hundreds of entities which may cross one or more national borders. I have classified the major international organizations into governmental (or more accurately intergovernmental) and nongovernmental organizations, and within those groupings, global and regional organizations. With the proliferation of the world wide web, much information can be readily accessed from each organization's web page, some of which I have edited and utilized.

Intergovernmental Organizations (Global)

The overarching global intergovernmental organization is the United Nations (UN). Within it, the Intergovernmental Oceanographic Commission (IOC) has the main responsibility for the coastal and deep oceans. Organizationally, the IOC is a component commission of the United Nations Educational, Scientific and Cultural Organization (UNESCO).