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INTERNAL WAVES IN THE OCEAN: A REVIEW

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Introduction

This review documents the advances in our knowledge of the oceanic internal wave field during the past quadrennium. Emphasis is placed on studies that deal most directly with the measurement and modeling of internal waves as they exist in the ocean. Progress has come by realizing that specific physical processes might behave differently when embedded in the complex, omnipresent sea of internal waves. To understand fully the dynamics of the internal wave field requires knowledge of the simultaneous interactions of the internal waves with other oceanic phenomena as well as with themselves.

This report is not meant to be a comprehensive overview of internal waves. The focus is on topics that have been discussed most actively in the literature; subjects that may be important, but have not received recent attention, are omitted. Often only a recent reference is given; the earlier studies upon which the work is based may not be cited. Excellent reviews of internal waves are provided by Garrett and Munk (1979), Munk (1981) and Olbers (1982). Hendershott (1981) gives a useful summary of internal tides; interesting discussions by many authors on the nonlinear properties of internal waves can be found in West (1981).

The reference list is intended to be a comprehensive compilation of papers published since 1979; not all works are mentioned in the text. The list is generally confined to papers written or translated into English. A number of references from disciplines where an understanding of the internal wave field is important, such as oceanic acoustics, is also included.

Random Sea

Internal waves in the ocean exist over a sufficiently wide frequency(ω)-wavenumber(κ) band so as to obscure the identification of individual plane waves. With the rapid increase of observations during the late 1960's and early 1970's it was realized that in order to describe and model the wave field one must resort to statistical methods. Much of the research over the last decade has been directed toward measuring the statistics of time-space fluctuations of velocity and temperature and determining to what extent these observations are consistent with the fundamental properties of internal waves, such as those dictated by the dispersion relation. A major advance in describing the internal wave field has been the empirical model of Garrett and Munk (Garrett and Munk, 1972, 1975; hereafter referred to as GM). The GM model organizes many diverse observations from the deep ocean into a fairly consistent statistical representation by assuming that the internal wave field is composed of a sum of weakly interacting waves of random phase.

Once a kinematic description of the wave field is determined, it may be possible to assess quantitatively the dynamical processes responsible for maintaining the observed internal waves. The concept of a weakly interacting system has been exploited in formulating the energy balance of the internal wave field (e.g., Olbers, 1976) in a manner analogous with the treatment of surface waves (Hasselmann, 1967). Recently, other sophisticated techniques for studying the dynamics of the random sea have been adapted from other branches of physics, such as statistical mechanics, quantum mechanics and plasma physics. The major advances have come in understanding the nonlinear interaction of internal waves with themselves. The dominant generators and dissipators of internal wave energy and momentum are still unknown, and their identification remains a challenge for theoreticians and experimentalists.

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Identification: observations and verification

Deep ocean. The Garrett-Munk model is designed for the deep ocean, where the buoyancy frequency, N(z), varies slowly enough with depth that the WKB approximation can be used for describing the dispersive characteristics of the internal wave field. The surprising aspect is that the energy levels and coherence structure of the GM model have been found to be remarkably constant throughout the world's oceans (Wunsch, 1976; Wunsch and Webb, 1979). A host of experiments, such as IWEX (Internal Wave Experiment), has provided much evidence which basically confirms the tenets of the GM model (Briscoe, 1975; Müller et al., 1978). Researchers now emphasize differences from the GM model in their observations. Finding significant departures from the GM universal model may lead to the identification of the most important processes that control the dynamic balance of the internal wave field (Wunsch, 1975a).

Agreement with the GM model was found to extend to low latitude (5°-10° N) in measurements of vertical wavenumber spectra (Hayes and Powell, 1980). However, very near the equator the observed spectral levels increased to ten times the GM level. Direct application of the GM model near the equator is dangerous because of the failure of the f-plane approximation. Eriksen (1980) has constructed a model of waves on an equatorial β-plane in a manner analogous to that used in formulating the GM model. Although based on limited data, the model provides a first attempt at organizing equatorial observations into a consistent framework.

At mid-latitude horizontal wavenumber spectra from towed sensors in the N. Atlantic thermocline show good agreement with the GM spectrum (Katz and Briscoe, 1979). However, there are anomalously high peaks in vertical coherence at 0.7-2 cpkm that may be due to relatively high-frequency waves that "tunnel" through a region of low N where they cannot exist as free waves. The deviation from the GM model near N is discussed in the next section.

Advances in instrumentation have made possible the measurement of small-scale vertical shear of horizontal current on vertical scales from 100 m to 1 cm, spanning the region from internal waves to turbulence (Gargett et al., 1981). The observed vertical shear spectrum is characterized as flat from 0.01 to 0.1 cpm in vertical wavenumber (β). Above 0.1 cpm the spectrum falls off as β^{-1} out to a buoyancy wavenumber $\beta_b = (N^3/\varepsilon)^{1/2}$, where there is a spectral minimum before an increase due to the dissipation of turbulent kinetic energy (ε). The level of the flat portion of the spectrum follows the WKB scaling for internal waves. The break in slope at 0.1 cpm appears to indicate the transition from internal wave motion to the increasing influence of turbulence. This is consistent with the observations and conjecture originally made by Gregg (1977).

Further effort has been made to modify, expand and explain certain consequences of the GM model especially near f (inertial frequency) and N where the WKB scaling breaks down. Munk (1980) compares peaks in the velocity spectrum near f and N that arise from using the energy distribution of the GM model with the exact solutions of the vertical wavefunctions in the neighborhood of both the horizontal and vertical turning points. The modification of the GM spectrum by a vertical turning point, where the wave frequency equals the local value of N, was first discussed by Desaubies (1973). Predicted peaks in spectra and coherences at frequencies near N are consistent with deep internal wave observations (Desaubies, 1975). In the upper ocean, however, there appears to be proportionally more energy and less wavenumber bandwidth at high frequency than in the GM energy distribution (e.g., Pinkel, 1975). The structure of the inertial spectral peak was modeled and compared with current meter data from the N. Atlantic by Fu (1981). With the effect of the turning latitude included the GM model provides a good description of the level and shape of the inertial peak in the deep ocean over smooth topography. However, at stations located in the upper ocean or in the deep ocean over rough topography, the local generation of inertial motion at the top and bottom of the ocean was found to be significant.

A deviation from the isotropic GM spectrum near topography is not necessarily an indication of a source or sink region but may be a linear distortion of the deep-ocean internal wave field. Eriksen (1982) investigated the effect of a sloping bottom on internal waves and found an intensification of velocity and temperature spectra over a frequency band centered at the local critical frequency—the frequency at which an incident ray is reflected parallel to the bottom. This linear theory predicts that these waves will be polarized with the major axis of the current ellipse oriented up-slope; there are some near-bottom observations on the continental slope that follow this behavior (Eriksen, 1982).

Another study of the effect of topography on the internal wave field was based on vertical temperature and velocity profiles near Bermuda (Johnson and Sanford, 1980). The polarization and horizontal anisotropy of the near-island current profiles are consistent with energy propagation from the ocean bottom and away from the island. These observations are in contrast to deep-ocean stations where the data indicate near-surface generation.

Upper ocean. In the upper ocean many of the assumptions of the GM model should not be valid for several reasons: 1) N varies rapidly with depth, invalidating the WKB scaling, 2) waves may be strongly forced and coupled with each other due to the proximity of atmospheric forcing, and 3) the wave field may not be stationary or homogeneous because of the large time and space variability in the upper layers. Still, the GM spectrum provides a description of the deep-ocean wave field that is useful for comparing with upper-ocean measurements.

With the increase in upper-ocean observations from experiments such as GATE (GARP Atlantic Tropical Experiment), MILE (Mixed Layer Experiment) and JASIN (Joint Air-Sea Interaction), generalizations about the upper-ocean internal wave field can begin to be made. The greatest deviations from the GM frequency spectrum are found near f, in the tidal band, and in a high-frequency band preceding the spectral roll-off. The remaining spectrum (which may not be much) sometimes follows WKB scaling (Levine et al., 1983a), but there are significant exceptions in regions where N varies rapidly (Käse and Siedler, 1980). In a summary of upper-ocean measurements Roth et al. (1981) find that spectral levels are usually equal to or above the GM spectrum. The upper-ocean wave field is hypothesized to be composed of a "base state," such as that described by GM, plus waves generated locally in the energetic upper layers.

Pinkel (1981b) discusses some of the consequences when the GM model is extended to the upper ocean. The waves of large vertical wavelength are attenuated as a result of two effects: they approach the surface boundary and the vertical waveguide where they propagate as free waves becomes thinner. This reduction of energy in the low-mode waves causes the red GM vertical wavenumber spectrum to become whiter and results in lower vertical coherence. Vertical velocity measurements from FLIP below the high-frequency band, with the semidiurnal tide and its harmonics removed, are consistent with this model (Pinkel, 1981b).

In a review of upper-ocean internal waves in the ice-covered Arctic Ocean, Morison (1983) finds frequency spectral levels to be significantly lower than those found in temperate latitudes. Although the field evidence is sparse, the spectral shapes are similar to those found in other upper-ocean experiments. The unique Arctic environment suggests many possible reasons for the low energy levels: the ice cover eliminates surface forcing by gravity waves, the tides are generally weak in the Arctic basin, and the ice sheet enhances dissipation at the surface. Additional experiments are necessary to substantiate these preliminary observations.

The deviation from the GM model at high frequency, just below the spectral roll-off, is usually evident in vertical displacement and horizontal velocity measurements as a spectral peak or shoulder accompanied by high vertical and horizontal coherence. These features appear to be due to fewer, relatively more energetic modes in this frequency band than the GM model proposes.

From a horizontal array deployed during GATE the high-frequency energy centered at 3 cph was found to be dominated by horizontally anisotropic waves of mode one, propagating to the west against a strong eastward flowing mean current (Käse and Siedler, 1980). Further analysis by Peters (1982a) indicates that the observed anisotropy cannot be explained primarily by a kinematic modification of the wave field by the mean shear flow. The implication is that the anisotropy is related to dynamic processes. The high-frequency statistics from MILE were modeled by a few low modes and incoherent noise (Levine et al., 1983a). The best fit to vertical displacement observations indicates that mode one is dominant with signal-to-noise ratios of order one. However, many features of the observed horizontal velocity spectra did not agree well with the model, such as the model-predicted 180° phase shift across the node of mode one. Some of these discrepancies are attributed to the contaminated response of a VACM (vector-averaging current meter) on a surface mooring (Halpern et al., 1981). Velocity observations in JASIN from VMCMs (vector-measuring current meter) attached to a surface mooring did show high coherence and a 180° phase shift in the high-frequency band, which is indicative of modal structure (Levine et al., 1983b). In JASIN the high-frequency band occurred over lower frequencies (0.5-3 cph) than in GATE and MILE (2-5 cph), which apparently is related to differences in the N profiles. A three-mode model was found adequate to reproduce many of the observed spectral features of both moored and towed data in JASIN (Levine et al., 1983b).

Finestructure. It is often difficult to distinguish from observations between irreversible finestructure, the signature of mixing, and reversible finestructure, the distortion due to small-scale internal waves. This problem does not solely result from our experimental inadequacies to sample densely or accurately enough but arises also from the ambiguity in defining wave-like and non-wave-like motion at small scales. In studying dynamic balances from observations it is necessary to be able to distinguish between processes that can transport mass and heat (irreversible) from those that only transport energy and momentum (reversible). This problem is usually tackled by formulating a statistical model and comparing it with observations--disagreement with internal wave statistics is blamed on irreversible finestructure (Desaubies and Gregg, 1981; Levine and Irish, 1981). Methods have also been developed to identify irreversible finestructure by attempting to remove the effect of internal wave strain on individual profiles of temperature and salinity. This is generally accomplished by removing density fluctuations (e.g., Johnson et al., 1978). However, these techniques are not unique and can lead to obscuring important processes, such as double diffusion (McDougall and Ruddick, 1982).

Indirect detection: Acoustics, Radar, Magnetometry. There is interest among acousticians to understand better the effects of the internal wave field on sound transmission through the ocean. These studies are usually undertaken from the perspective of the acoustician trying to interpret acoustic observations in terms of the sound-speed and velocity fluctuations caused by internal waves. When sound-speed fluctuations are modeled by the passive advection of a mean sound-speed profile by internal waves, the GM representation leads to a reasonable description of observed acoustic travel-time fluctuations over a fixed distance (Munk and Zachariasen, 1976; Desaubies, 1976). However, the corresponding fluctuations of the acoustic amplitude are not well predicted. Several explanations have been offered to account for this discrepancy: the effect of multiple scattering by the internal waves (Uscinski, 1980), and the strong scattering by finestructure, both reversible and irreversible (e.g., Ewart, 1980; Flatté et al., 1980; Unni and Kaufman, 1981). Whether acoustic measurements will be a useful tool for monitoring aspects of the internal wave field, such as vertical momentum flux (Munk et al., 1981a), or helping to distinguish between wave-like and non-wave-like finestructure remains unanswered.

Satellite observations have the desirable property of covering vast areas in a short time. The obvious difficulty with using this tool for the study of internal waves is that only the surface manifestations of the wave can be seen. With SAR (synthetic aperture radar) surface patterns indicative of internal waves can be detected (Vesecky and Stewart, 1982). The radar is sensitive to short surface waves ($\simeq$ 30 cm) that are believed to be modulated by internal waves through the convergence of oil films or surface wave-internal wave interactions. Internal waves are usually observed with SAR near coastal areas occurring in separate groups 10-60 km apart with crests 10-100 km long (Fu and Holt, 1982). These packets can often be associated with the interaction of tidal currents with topography and appear to exhibit characteristics of nonlinear dispersion (Apel, 1981). An attempt has also been made to use measured phase speeds of these wave packets to estimate heat content in the upper ocean (Mollo-Christensen and Mascarenhas, 1979). With increasing coverage and sampling-rate the satellite may prove useful for studying the coastal internal wave field, especially at tidal frequencies.

Internal waves propagating in the earth's magnetic field induce a secondary magnetic field. The magnetic induction spectrum for the GM model has been calculated, and magnetometry has been suggested as a potential oceanographic tool (Petersen and Poehls, 1982).

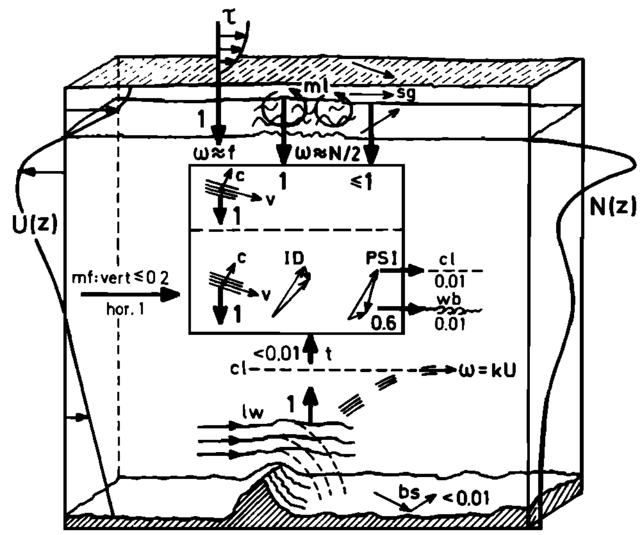

Figure 1. Sketch of interaction processes affecting the internal wave field in the upper and the deep ocean. Energy fluxes are in units of 10^{-3} W/m^2. Abbreviations and references for flux estimates are as follows: τ, wind stress (Käse, 1979; Käse and Olbers, 1980); ml, mixed layer turbulence (Bell, 1978); sg, surface gravity waves (Olbers and Herterich, 1979); cv, near-inertial waves (Käse and Olbers, 1980; Müller et al., 1978); mf, large scale mean flow (Ruddick and Joyce, 1979; Brown and Owens, 1981); t, baroclinic tides (Olbers and Pomphrey, 1981); lw, lee waves (Bell, 1975); bs, bottom scattering (Bell, 1975); cl, critical layers (Ruddick, 1980); wb, wave breaking; ID, induced diffusion; PSI, parametric subharmonic instability (Pomphrey et al., 1980). (From Olbers, 1982).

Dynamics

Great advances have been made in understanding the dynamics of the internal wave field. More realistic theoretical calculations have resulted by treating the wave field stochastically. By describing the wave field as a random process, one can more easily account for the multitude of interactions that are possible in a random sea.

The goal is to understand the flow of energy through the internal wave field: the sources, the sinks and the processes that transfer energy among the internal waves themselves. Despite the fact that the spectrum of the internal waves is nearly universal, the primary sources and sinks have still not been identified. This may not be surprising. Since the wave field is so insensitive to geographical variation, atmospheric forcing, etc., it follows that the waves spend much time as free waves and only slowly receive energy to replace the small amount lost to dissipation (e.g., Garrett and Munk, 1979; Olbers, 1982); hence, the direct link between the forcing or dissipation and the wave field itself is difficult to make. A schematic diagram from Olbers (1982) indicating some of the physical processes that may affect the internal wave field is presented in Fig. 1. Many processes, such as scattering at fronts and generation by atmospheric pressure fluctuations, were not included because their effect on the wave spectrum could not be estimated or turned out to be small (see Olbers (1982) for details).

Transfer. Much of the theoretical work has concentrated on determining the redistribution of energy among the internal waves. This calculation is the most straightforward since it depends only upon the energy distribution of the internal wave field in frequency and wavenumber—no knowledge of sources and sinks is necessary. By calculating the rate of energy transfer among the waves, the location and magnitude of the sources and sinks in frequency-wavenumber space can be inferred by assuming a steady-state internal wave field.

Most studies of nonlinear transfer in the internal wave field have used the framework of weak-interaction theory. An essential assumption of the theory is that the time scale of the nonlinear transfer of energy be slow compared with the period of the wave; the wave must maintain its integrity for at least a period or wavelength before being significantly modified. This permits the use of a two time-scale perturbation analysis, and only resonant triad interactions contribute significantly to the energy transfer. The application of weak-interaction theory to the internal wave field was pioneered by Olbers (1976) and McComas and Bretherton (1977). In these studies the internal wave field is governed by a radiative transport equation that describes the changes of the wave action density through wavenumber space. A significant contribution by McComas and Bretherton (1977) was to identify three classes of triad interactions that dominate in the internal wave field: Induced Diffusion (ID), Parametric Subharmonic Instability (PSI) and Elastic Scattering (ES). The categorizing of these interactions has facilitated the mathematical treatment as well as discussions of energy transfer. The ID mechanism represents the interaction of a high-frequency, high-wavenumber wave with a wave of low-frequency, low-wavenumber to produce another high-frequency, high-wavenumber wave. This process acts as a diffusion of wave action in vertical wavenumber space. In ES a high-frequency wave is scattered into another wave with nearly opposite vertical wavenumber (with similar horizontal wavenumber and frequency) by a low-frequency wave of twice the vertical wavenumber. This mechanism is effective in removing vertical asymmetry in the energy flux, except at frequencies near f. The PSI interaction transfers energy from a low vertical wavenumber wave into two waves of high vertical wavenumber at half the frequency. This mechanism is most efficient at near-inertial frequencies.

The most recent and complete description of the energy balance is presented by McComas and Müller (1981b) (Fig. 2). Using the GM model and an analytical simplification of the radiative transfer equation, energy is found to be generated at low vertical wavenumbers $\beta < \beta_*$ and dissipated at high wavenumbers $\beta > \beta_c$. Between β_* and β_c there is a region of constant energy flux through which the ID mechanism at high frequency ($N > \omega > 4f$) and the PSI mechanism at low frequency (2f–4f) transfers energy to high wavenumber. The constant flux region has a wavenumber spectrum with a β^{-2}

dependence, corresponding to a flat vertical shear spectrum as given by the GM model and measured directly by Gargett et al. (1981). Above $\beta_c \approx 0.1$ cpm there is a break in the slope, and the spectrum falls off more rapidly, also consistent with observations.

Questions have been raised as to the validity of applying weak-interaction theory to the internal wave field. It is claimed that the nonlinear transfer time is, in fact, not slow compared to the wave period for much of the high-frequency, high-wavenumber spectrum of internal waves (Holloway, 1980, 1982; Pomphrey et al., 1980). This led to recalculation of the energy transfer rates by the ID and ES mechanisms without using the assumption of weak interaction and thereby avoiding the need for time-scale separation. For the GM model the ES interactions were found to be adequately described by weak-interaction theory (Watson, 1981). However, with regard to the ID mechanism the transfer times were significantly reduced from those calculated by McComas and Müller (1981b) (Meiss and Watson, 1982). Despite this modification the scenario of the internal wave energy balance given by McComas and Müller (1981b) remains qualitatively unchanged (Pomphrey, 1982).

Recently, the ID interaction has been reexamined using eikonal theory (Henyey and Pomphrey, 1982). The high-frequency, high-wavenumber waves are modeled as a superposition of wave packets that move through a large-scale, low-frequency flow. The numerical experiments, which involve following wave packets propagating through the GM internal wave field, suggest different results from those of Meiss and Watson (1982). It is argued that ID actually behaves as diffusion in frequency-depth space rather than wavenumber-time. Further research is needed to determine the consequences on the energy balance proposed by McComas and Müller (1981b).

Another view of the energy balance is presented by Orlanski and Cerasoli (1981). Numerical experiments of a two-dimensional random internal wave field suggest strong interactions may be important in the transfer of energy. When low wavenumber energy is added to a spectrum saturated at the dissipation scales, the vertical gradient is increased in localized areas causing overturning. Hence, dissipation is increased without first transferring the energy to high wavenumber. Quantitative application of these ideas to ocean spectra would require further experimentation.

Sources. A variety of sources may be responsible for maintaining the observed internal wave field (e.g., Thorpe, 1975). Few of these have been evaluated accurately enough to determine which are the most important.

Calculations by Müller (1976) indicated that the mean shear in quasi-geostrophic flow could provide a high rate of energy input into the internal wave field. The calculated wave-induced vertical and horizontal eddy viscosities calculated from the GM model are 0.4 and 7.0 m^2/s respectively. This implies the energy gain from the vertical shear alone ($\approx 10^{-3}$ W/m^2) would be enough to maintain the internal wave field while the energy transfer rate from the horizontal shear is insignificant. However, field evidence

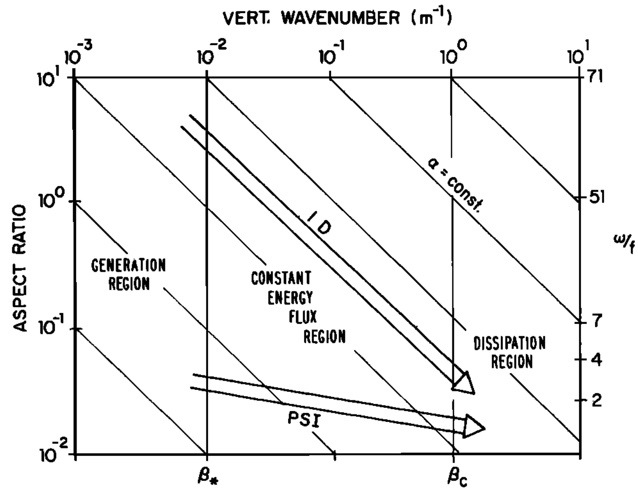

Figure 2. Schematic view of energy balance of the internal wave field. Energy is generated at low vertical wavenumbers $\beta < \beta_*$. It is transferred at high frequencies by the ID mechanism and at low frequencies by the PSI mechanism to high wavenumber $\beta > \beta_c$. Here dissipation is dominant over the weak nonlinear transfer (from McComas and Müller, 1981).

indicates that this estimate of vertical viscosity is too high by at least a factor of 100 (Frankignoul and Joyce, 1979; Ruddick and Joyce, 1979). In contrast, results from the Local Dynamics Experiment of Polymode indicate a correlation between the horizontal shear and internal wave variance resulting in a wave-induced horizontal viscosity of 10^2 m^2/s, much larger than predicted by Müller (Brown and Owens, 1981). Further theoretical effort and more conclusive experimental evidence are needed to resolve the discrepancies.

Recent calculations by Olbers and Herterich (1979) showed that the nonlinear transfer of energy from surface waves to internal waves is relatively insignificant with the possible exception of internal waves at high frequency. The importance of this mechanism on the internal wave field is much less than originally estimated by Watson et al. (1976), where the interacting wave components were modeled by a fixed-phase relationship rather than a more reasonable random-phase representation.

The atmosphere may generate internal waves directly by fluctuating wind stress, surface pressure, and buoyancy flux. It has been established that near-inertial motion in the surface layers can be forced by the wind stress; however, the effects of the wind on the internal waves below the mixed layer are not well known. Käse (1979) estimates with a theoretical argument that the wind stress forcing is potentially strong enough at near-inertial frequencies to maintain the entire internal wave field. Measurements from GATE indicate a downward propagation of near-inertial energy into the thermocline with marginally significant correlation between the wind stress and the wave field (Käse and Olbers, 1980). These measurements are consistent with a model of Ekman suction at the base of the mixed layer providing direct linear coupling of the wind stress to the wave field in the thermocline.

The response of the ocean to an intense storm, such as a hurricane, has been studied by Price (1982 a,b) with a numerical model supported by limited field observations. The forced, near-inertial internal waves are found to transfer energy directly through the mixed layer to the thermocline more effectively than free internal waves. These waves can drain the energy in the mixed layer by 1/e in just five wave periods. These results are expected also to hold qualitatively for less severe storms.

Evidence of correlation between the wind and the high-frequency internal wave field is minimal. No significant correlation was found between the high-frequency variance in the internal wave field and the wind stress during MILE or JASIN from measurements made within 20 m of the mixed layer (Levine et al., 1983a; de Witt et al., 1982)

Using data from JASIN, Briscoe (1983) has examined the variations of the kinetic energy in the internal wave band between 0.1-2 cph away from the tidal and inertial frequencies. For a single five-day period there is a relationship between the energy flux into the atmospheric boundary layer and the changes in the vertically integrated internal wave energy. A 1.5 day lag is observed before the wave field responds to the wind stress perhaps indicating that the interaction with internal waves is accomplished indirectly through the surface wave field.

The internal tide generally contains a significant fraction of the total energy in the internal wave field. The ability of the internal tide to act as an energy source for the internal wave continuum (e.g., Bell, 1975) has been investigated theoretically using weak-interaction theory (Olbers and Pomphrey, 1981). Assuming an internal tide superimposed on the GM spectrum, the rate of energy transfer from the tide to the continuum is too low to be significant.

Sinks. The most important dissipative processes for internal waves in the ocean interior are probably shear and gravitational instabilities, or some amorphous combination of the two (Holloway, 1981). In order to describe the complete energy balance McComas and Müller (1981b) devised a simple parametrization of internal wave dissipation. By assuming that a dissipation event is localized in space and time and that the mixing in each event can be described by a vertical eddy viscosity, the time rate of change of the spectral density of the internal wave field is proportional to the spectrum of vertical shear. This is a useful result since the shear spectrum can be measured. According to this model the dissipation in the GM spectrum occurs primarily at high vertical wavenumber, where the shear is large.

A mixing model based on the random occurrences of shear instability in a stochastic representation of the internal wave field was developed by Desaubies and Smith (1982). Assuming that complete vertical mixing takes place over the depth range where the Richardson number (Ri) drops below 0.25, estimates of eddy diffusivity are comparable to those inferred from microstructure measurements (Garrett, 1979). The probability density function of Ri is found to depend only on the rms strain of the internal wave field, where strain is ∂(vertical displacement)/∂z, and to be very sensitive to its value. The strain is proportional to the spectral energy level and high vertical-wavenumber cutoff; an increase in either of these quantities results in a dramatic increase in the probability of instability. The model indicates that shear instabilities caused by the stretching and straining of the internal waves themselves are sufficient to result in significant internal wave dissipation.

Another possibility for removing internal wave energy and momentum is by loss to the mean flow through the mechanism of critical-layer absorption. The critical level is the location where the phase velocity of an internal wave propagating in a mean shear flow becomes equal to the mean velocity. Booker and Bretherton (1967) have demonstrated theoretically that most of the momentum of the internal wave is absorbed by the mean flow near a critical level when Ri is much greater than one. The phase speeds of oceanic internal waves of high vertical wavenumber (high mode) and low frequency can be comparable to the speeds of mean flows, thus making this mechanism a potentially important one. Ruddick (1980) has examined some of the effects on the GM spectrum, such as anisotropy and momentum loss, that could be caused by critical-layer absorption. This model, which assumes complete absorption of the wave momentum by the mean flow at a critical level, predicts a maximum wave-induced vertical eddy viscosity in a 400 m thick mean-shear layer of ~200 cm^2/s.

Extensive investigations into the details of critical-layer phenomena have been made theoretically with analytical and numerical studies (e.g., Brown and Stewartson, 1980, 1982a,b; Hirt, 1981) and experimentally in the laboratory (e.g., Koop, 1981; Thorpe, 1981). The application of these studies to the oceanic internal wave field has yet to be made.

Top and bottom boundaries are potential sites for internal wave dissipation as well as generation. The bottom boundary has been suggested as a sink for near-inertial internal waves to explain the often observed downward energy flux (e.g., Leaman, 1976). However, estimates of the absorption of near-inertial internal waves in the benthic boundary layer have indicated that it is probably a relatively unimportant process over smooth topography (Fu, 1981; D'Asaro, 1982b). However, Eriksen (1982) argues that near-inertial waves may be intensified after critical reflection off relatively flat slopes of 1%. The resulting strong inertial current and shear are suggested as energy sources which aid in the formation of bottom mixed layers. There is evidence that boundary layer dissipation may also be important where large amplitude internal waves encounter steep topography, such as in a submarine canyon (Hotchkiss and Wunsch, 1982).

Deterministic Waves and Specific Processes

While much of the geophysical internal wave field can only be described effectively using statistical measures, some phenomena lend themselves to a more deterministic description. This is usually only possible with oscillations at a dominant frequency or wavelength that are strong enough to be distinguished from the background. A deterministic description is also useful in theoretical models and laboratory experiments that are designed to study a specific process. This allows one to analyze a particular phenomenon in detail without the additional complications of the random sea.

Internal tides

The internal tide is an internal wave at tidal frequency that usually distinguishes itself from the background internal wave field by its large amplitude. Unlike most internal waves the source of the internal tide is known; it is generated from the barotropic tide probably by interaction with topography (see review by Hendershott, 1981).

The continental slope is a potential site of significant generation of internal tide. Theory indicates that the internal tide propagates from the slope area in beams of energy following characteristics (Rattray et al., 1969; Baines, 1982). The tide may not be observed as predicted since it can be modified as it propagates through the ocean by the spatial and temporal variations of N and mean shear (Mooers, 1975; Chuang and Wang, 1981). Current meter data from the Oregon slope and shelf were consistent with an 80 km wide beam-like structure of semidiurnal internal tide emanating from the slope where the topography was steeper than the characteristic (474-1050 m) (Torgrimson and Hickey, 1979). The beam could be traced at least 50 km from the generation area. Evidence of the beam-like nature of the tide was also found in the Rockall Trough in the N.E. Atlantic during JASIN (de Witt et al., 1982). During a one-week period, a strong tidal signal was observed, and its generation traced to Rockall Bank about 100 km away. Although the beam-like nature of the tide was observed on large space scales, locally the oscillations could be described by a dominant mode-three plane wave propagating horizontally with a wavelength of 36 km. On the W. Florida shelf a strong diurnal internal tide did not show beam-like behavior; the vertical structure was primarily composed of first and second modes (Leaman, 1980). The temporal variation of the tidal energy appeared to be related to low-frequency vertical shear. Since the diurnal tide is near the critical frequency at this location, the structure of the tide is very sensitive to changes in the slope of the characteristic caused by variation of the shear.

Far from the generator the modes are expected to become uncorrelated, the beam degenerating into uncorrelated vertical modes. The higher modes will be damped more effectively, and therefore only the lowest modes will be observed, probably with mode one dominant (Hendershott, 1981). However, recent observations in the deep ocean have indicated that higher modes, from 3-5, may dominate (Simpson and Paulson, 1979; Lyshenko and Sabinin, 1980).

Solitary waves

In addition to internal tides, flows over large ridges and sills can produce a variety of nonlinear phenomena, such as hydraulic jumps and solitary waves. Some of these phenomena that have been studied both theoretically and in the laboratory have been identified in geophysical flows. From laboratory studies Maxworthy (1979) describes a scenario of the generation of solitary waves from tidal flow: the ebb flow produces lee waves behind the sill; as flow slackens, these waves propagate upstream over the sill and develop into a series of solitary waves. Phenomenon similar to this have been seen in Massachusetts Bay (Haury et al., 1979). When the tides turned, lee waves were observed to steepen as they propagated over the sill and formed up to three wave packets, composed of 8-10 minute period waves, modulated at a period of 90 minutes. In the Andaman Sea packets of internal solitary waves that are probably generated by tidal currents have also been observed (Osborne and Burch, 1980). Interaction of the solitary waves with surface waves is demonstrated to be responsible for observed regions of short, choppy, breaking surface waves.

Internal waves generated at the sill of a fjord propagate into the fluid and when dissipated at boundaries may be responsible for vertical mixing (Stigebrandt, 1979). Extensive observations in Knight Inlet, British Columbia, indicate a large variety of responses resulting from tidally driven flow over a sill depending on the strength of the stratification and tidal currents (Farmer and Smith, 1980). The densiometric Froude number is found to be a useful quantity for characterizing whether the response will result in an internal hydraulic jump or a train of lee waves. Evidence of vertical mixing is often observed within energetic bores traveling away from the sill--this mechanism for mixing differs from Stigebrandt's model where the mixing occurs at boundaries.

Strongly nonlinear internal motion has also been observed in the thermocline of the open ocean far from the coast (100 km) over deep water (4 km) (Pinkel, 1979). Using Doppler sonar, features of 20 km horizontal scales were detected with phase velocities of 40 cm/s and water speeds >20 cm/s. Although comparison with theory is not conclusive, these features have similarities with solitary wave solutions of mode two.

It is also possible to generate solitary waves by the collapse of a localized mixed region that may be caused by turbulent overturning (e.g., Maxworthy, 1980; Kao and Pao, 1980). The geophysical significance of this mechanism is unknown.

Horizontal inhomogeneity. If the scales of the internal waves are small compared to the scales of the mean velocity and buoyancy fields, the propagation of an internal wave can be described using the WKB approximation. Olbers (1981b) applied WKB theory to internal wave propagation through a mean geostrophic flow with horizontal as well as vertical shear. Critical layers are found that exhibit a valve effect--incident waves from one side can penetrate the layer while waves from the other side are absorbed. The conditions for reflection and absorption are presented in detail for geostrophic flow with constantly sloping isopycnals.

A general theory of internal wave interaction with inhomogeneities without using WKB scaling has been developed by Olbers (1981a). The approach is to cast the equations of motion of internal waves into the form of the Schrödinger equation and adapt scattering theory from quantum mechanics. The theory is general enough to apply to scattering by topography (e.g., seamounts and continental shelves) as well as by variations in mean density and velocity fields. The theory is applied to an oceanic front and indicates that near the front the internal wave field will be anisotropic with more energy along the axis of the front than perpendicular to it.

Into the Future

The study of oceanic internal waves is at a crossroad. The extensive observations made during the past decade have been reconciled with the unforced equations of motion, and a kinematically consistent description of a "universal" internal wave field has emerged. Yet, the physical processes responsible for maintaining the pervasive oceanic internal wave field remain unidentified. While the description of the internal waves rests on a firm base of observations, the theories explaining the flow of energy through the wave spectrum are largely unsubstantiated. The path of research that will best provide the key answers is uncertain; progress in understanding the forces that drive the wave field will not be achieved easily.

Much of the recent effort by theoreticians has been directed toward describing the nonlinear energy transfer among the internal waves. Although some consensus appears to be emerging from the mathematically complicated calculations, the increasing activity by theoretical physicists from other disciplines is sure to lead to new insights and stimulate debate. The application of mathematical techniques from other branches of physics will undoubtedly be implemented in the ongoing search for the significant mechanisms of wave generation and dissipation. One of the severe theoretical limitations is the inability to treat strong nonlinear interactions. When the interaction of waves can no longer be described by weak-interaction theory, the transition to turbulence begins, and the notion of a wave itself becomes fuzzy.

Where does this leave the experimentalist? In order to make advances in understanding the dynamic balance of the internal wave field, the observationalist must be guided by theories that can be tested by measurements. As yet theoreticians have not been able to indicate such straightforward tests. The task is inherently complicated by a random sea that prevents the association of a wave of a certain frequency with its wavenumber. Indeed it is not easy even to conceive of a definitive experiment that would reveal the nature of the energy balance.

Although the prospects for breakthroughs in determining the energy balance are somewhat discouraging, there are still many topics that would benefit from further exploration. The gathering of better quality data in a variety of oceanic environments will permit the continuing refinement of the empirical description of the internal wave field and provide substance to stimulate and guide theoretical ideas. Even though the concept of a "universal" internal wave field is well founded, a closer examination of the deviations is needed. How do variations in topography, wind stress, eddy fields and fronts affect the internal wave climate? Exploratory experiments designed to investigate systematically the temporal and spatial variability of spectral level and coherence structure in diverse oceanic environments may provide useful clues.

Most recent theoretical and experimental research has been guided by the premise that internal waves exist at a continuous spectrum of frequency and wavenumber. However, most of the internal wave energy does not fit into a power law description of a spectral continuum but is, in fact, concentrated in narrow frequency bands around the inertial, tidal and buoyancy frequencies. As these are the most energetic internal waves, uncovering the details of their interactions may provide information relevant to the dynamics of the continuum. Further theoretical studies treating the interactions of these narrow-band waves with the broad-band continuum and with each other would be useful.

Internal waves occupy an intermediate time and space scale of oceanic phenomena between turbulent and geostrophic flows. Because they are ubiquitous, internal waves are familiar to investigators of microscale and mesoscale phenomena alike. Internal waves are often considered a nuisance because their complicated temporal and spatial fluctuations are blamed for obscuring and contaminating experimental observations. Consideration of the internal wave field is often necessary, if only to demonstrate their relative unimportance. However, recent work has indicated that internal waves may play a significant role in the energy and momentum balances of processes at other scales, such as an effective viscosity for mesoscale motions or as a source of shear for turbulent production at the microscale. Because internal waves can be significant participants in many physical processes in a stratified ocean, an improved understanding of their behavior is important for the specialist and for the physical oceanographer in general.

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ADVANCES IN SATELLITE OCEANOGRAPHY

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Abstract. Progress has been made in the past four years by U.S. scientists in the development and application of active and passive satellite remote sensing techniques to the study of oceanic processes. This report summarizes technical advances and recent applications. Major advances have been made in developing and applying quantitative measurements from active and passive satellite based sensor systems launched in the late 1970's and that proven methodologies now exist to observe sea surface temperature, ocean elevation, ocean color, surface wind stress and waves, and to locate free drifting buoy data collection platforms. Many of the advances in technique and application have occurred using sensors which were experimental, i.e., not part of an operational satellite observing system. Consequently future geophysical application and development of advanced techniques to enhance our

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