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INTERNAL WAVES, FINESTRUCTURE,
MICROSTRUCTURE, AND MIXING IN THE OCEAN

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M. C. Gregg
and
Melbourne G. Briscoe

November 1980

TECHNICAL REPORT

*Prepared for the Office of Naval Research
under Contract N00014-76-C-0197; NR 083-004
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INTERNAL WAVES, FINESTRUCTURE, MICROSTRUCTURE,
AND MIXING IN THE OCEAN

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Abstract. Progress in measuring, interpreting, and understanding oceanic internal gravity waves and fine and microstructure is reviewed; we emphasize the quadrennium 1975-1978. The context is how these subjects contribute to oceanic mixing. The overlap between the areas is examined, as is the relevance of the subjects to other aspects of oceanography.

Present trends and suggestions for future work are included, and we offer some speculation on possible progress during the next quadrennium, which may be substantial especially for finestructure understanding.

Introduction

This is the first combined IUGG (International Union of Geodesy and Geophysics) review on internal waves and fine and microstructure.

Our task is made considerably simpler by very recent reviews on internal waves (Garrett and Munk, 1979), on mixing in the ocean (Garrett, 1979b), and on internal waves and small-scale processes (Munk, 1979), that contain much of the information we could have included here; this present report is therefore written to be complementary to those three reviews. They were broad and covered many decades of work; we emphasize the last few years and have tried to include sufficient detail about the major papers to suggest their contribution within the broader context. We emphasize fine and microstructure in this review because they are the subjects most in need of a framework for understanding.

The Bibliography is not complete, even for the quadrennium. In particular, it makes no great attempt to cover non-English language material, nor to describe the gray literature of company reports and newsletters. It principally contains refereed articles from the major journals with a few extra items included if they were particularly interesting or indicated a class of work not otherwise covered. Not all the listed items are discussed in the text.

In its essence, the study of internal waves is concerned with a major source of variability and a probable contributor to oceanic mixing. Guided by a theoretical framework, the subject has matured with the experimental work becoming more complicated and the ideas and analyses more sophisticated. The recent theoretical benchmarks are Garrett and Munk (1975) (called GM75 herein), Thorpe (1975), Müller and Olbers (1975), and McComas (1977).

There is no equivalent theoretical framework for mixing, particularly in stratified fluids. Mixing is a very general term that means the combining of things so that the resulting substance is more uniform in composition. If the ocean is considered on a global scale, mixing is accomplished by a hierarchy of processes ranging from the wind-driven gyres through mesoscale eddies to viscosity and diffusion, which are the ultimate mixing processes and refer to intermingling on a molecular scale. These processes act over a 10^{10} range of scales (up to 10⁴ km for ocean basins to about 1 mm for the diffusive cut-off of salinity fluctuations). If one is concerned with only a restricted scale range in the middle of this hierarchy, then the processes acting over larger scales will usually tend to increase the variability of interest while all of those at smaller scales will be "mixing". Thus, in Eckart's (1948) terminology, one person's stirring may be another's mixing. Confusion can only be avoided by being precise about the scales and measures of inhomogeneity being considered.

Microstructure is the smallest scale variability in the ocean, traditionally 1 m to about 1 mm where the molecular processes of viscosity and diffusion become important. It has always been hoped and sometimes asserted that classical ideas of isotropic three-dimensional turbulence, perhaps modified slightly by buoyancy, are appropriate. It is still not clear to what extent this is true, and the varying analyses

and interpretations share this confusion.

Finestructure refers to fluctuations that are small, typically 1 m to 100 m in the vertical, in comparison to the scale depth of the thermocline, but are too large to directly involve molecular processes. Since the corresponding horizontal scales of finestructure are about 10 km or less, finestructure and microstructure together span a 10^7 range of scales; the processes producing finestructure and microstructure encompass a large fraction of the spectrum of ocean physics.

The present excitement about mixing work stems from the ability to make observations of the signatures of at least some of the mixing processes. Observationally based flux estimates provide consistency tests of the more traditional flux values, which are based on large-scale budgets and are usually parameterized by eddy coefficients. The canonical presentation of the eddy coefficient approach in the vertical is Munk (1966). Since the dissipation of internal wave energy by shear and convective (overturning) instabilities is one of the processes provoking this mixing, the two subjects are closely coupled.

This coupling is shown schematically in Figure 1. The widespread confirmation of GM75 (e.g., Wunsch, 1976b, Wunsch and Webb, 1979) has demonstrated that there is at least a zeroth order steady state level for the internal wave field, suggested by the size of the "bucket" in the Figure. Theoretical analyses show an abundance of mechanisms for feeding energy into the internal wave field (c.f., Thorpe, 1975) but little is known about how much transfer actually does take place. If the flow of energy into the field is large (shown by the open top to the bucket) then a large amount must also splash out to preserve the universal steady state. On the other hand, if the production mechanisms are effectively weak (only a small hole exists in the bucket lid), then the rate of loss is also low.

In the thermocline the leakage, represented by ϵ (the rate of dissipation of turbulent kinetic energy by viscosity) drives the turbulent fluxes of heat and salt. The microstructure measurements to date are consistent with weak turbulent gradient fluxes. However, knowledge of the mixing efficiency is so uncertain that the weak fluxes could be consistent with either of the extreme internal wave states.

In the following, the recent work on internal waves and mixing processes are reviewed separately since the direct linkage in the literature is not yet very strong. For internal waves much of the literature can be categorized as either phenomenology or dynamics, and there is a growing interest in the effects of internal waves on acoustics and sea surface signatures. A recommended reading list is offered below for those with limited time, or interest, for fully exploring the literature. The mixing contributions divide nicely between thermocline processes and those occurring in various boundary regions. Finally we conclude with a discussion of both areas and their intersection.

The outline of the review is therefore:

Introduction

Internal Waves

Phenomenology

Universality.
Upper-Ocean.
Acoustical Fluctuations.
Remote Sensing from Satellites.
Solitons, etc.

Dynamics

Large-Scale Energy Sources.
Surface-Internal Wave Interactions.
Generation by Wind.

Recommended Reading

Mixing

Thermocline Processes

Finestructure - from Moorings.
Finestructure - from Profiles.
Microstructure - Fluxes and Dissipation
Levels.
Microstructure - Detailed Characteristics.
Discussion.

Boundary Processes

Mixed Layers and the Upper Ocean.
Interleaving.
Jets.
Bottom Boundary Layers.
Side Walls.
Hydrothermal Activity.

Discussion

Acknowledgements.

Bibliography

Internal Waves

Three main ideas have been pervasive in the internal wave work of the quadrennium: the unifying concept of a universal deep-ocean spectrum (Garrett and Munk, 1975); the utilitarian view of internal waves as a viscosity that drains energy from the mesoscale motions and, incidentally, energizes the internal wave field (Müller, 1976); and the separatists' view of the upper-ocean internal waves as being quite different in nature from the deep-ocean internal waves (e.g., Brekhovskikh, et al., 1975; Pinkel, 1975; and Käse and Clarke, 1978). All three ideas are elegant, but how correct are they? Some evidence is discussed below.

Other major areas of work have been on surface and internal wave interactions, solitons and other non-linear behavior both statistical and non-statistical, acoustical fluctuation induced by internal waves, remote sensing from satellites, and generation by wind fields.

Phenomenology

Universality. Wunsch (1976b) and Wunsch and Webb (1979) have tried to find instances of internal wave spectra that do not resemble the canonical, universal form, i.e., a spectrum like Garrett and Munk (1975),* called GM75.

* Since at least Cairns and Williams (1976), Desaubies (1976b), Bell (1976), Müller, et al. (1978), and Katz and Briscoe (1979) have all contributed to the description of the hypothesized universal spectrum, to call it solely a Garrett-Munk spectrum is no longer accurate.

Their results are intriguing. Within a kilometer above Muir Seamount or 10km beside it, and within submarine canyons, the energy levels are higher than the universal form. Under the Gulf Stream and at the equator there are subtler disparities due to decreased coherences. Roth, et al. (1979) have similarly examined near-surface measurements only and also find the universal spectrum to be a good descriptor of the "base state"; that is, observed deviations (in terms of energy and coherence) were consistent with the presence of internal waves in addition to the universal form, not instead of it. Near the island of St. Croix, and underneath strong atmospheric forcing, they found more energy than in the canonical form; in the data sets they examined they could find no instances of significantly less energy. Some oddities of coherence, especially for vertical separations, were ascribed primarily to buoyancy-profile effects and finestructure contamination.

The tentative picture, then, is of a canonical, universal form for the internal wave spectrum - see footnote above - that represents a lower bound to the energies present in the upper ocean as well as the deep ocean. Near strong sources more energy is present. Coherences may either increase or decrease depending on the coherence of the additional energy, and there is a suggestion that the distribution of energy among the many modes present may also be modified (Wunsch and Webb, 1979) even if the total energy is invariant (within a factor of 2 or 3; by "universal" we do not mean exactly the same energy everywhere).

Inertial waves and internal tides are not clearly part of the universal spectrum; the latter are not even present in the Garrett-Munk models, and both are more deterministic than random. The inertial motions are clearly wind related (e.g., Kundu, 1976; and Yao, et al., 1977) while the internal tides are directional (e.g., Hendry, 1977), vertically asymmetric (e.g., Müller, et al., 1978; so are the inertial motions), and probably geospecific (Wunsch, 1975b). The internal tides constitute perhaps one-third of the total displacement field due to internal waves (e.g., Briscoe, 1975a) even in the deep, open ocean, while the inertial motions account for half or more of the current field (e.g., Sanford, 1975); incorrect modeling of these motions is a severe error. Although there is an inertial cusp in the canonical form (c.f.,

Garrett and Munk, 1975), the concept of an average spectrum ignores the consistent observation of inertial motions as occurring in bursts (e.g., Webster, 1968), sometimes correlated with the wind.

Garrett and Munk's (1972a) suggestion that the deep ocean spectrum might be universal was surprising; in fact it now seems that a universal spectral form may be the underlying state in the upper ocean as well as the deep. Attention will be increasingly directed toward studies of the anomalous cases, as Wunsch (1975a) suggested. The central problem is unchanged: why does the universal spectrum exist? Some answers are available in McComas (1977) where calculations of the mechanism and rate of decay of a perturbation to the universal form are given. The predicted disappearance of a perturbation into the background unfortunately occurred so quickly as to cast doubt on the very validity of the theory used, but the consequences qualitatively were clear: the spectral slopes and vertical symmetries of the universal spectrum are due to non-linear interactions within the wave field. Still open is why the particular spectral levels are so invariant, although the original suggestion of Garrett and Munk (1973) about a saturation mechanism based on shear instability is still a contender.

Upper-Ocean. Brekhovskikh, et al. (1975) inferred from their "line sensor" temperature observations that high-frequency internal waves (near the local buoyancy frequency) occurred in groups of low-mode, nearly sinusoidal oscillations, and that sometimes the waves seemed to be standing, rather than progressive (in the horizontal). Sabinin and Serikov (1976) used data-adaptive spectral analysis to look further at the short-term properties of the data set and hypothesized that the standing waves were caused by billow turbulence in the weakly stratified mixed layer; the billows were due to a Richardson-number instability when the shear from inertial motions constructively added to the large-scale mean shear present. The waves seemed to be standing because the observation site saw similar waves coming at it from billows on both sides of it. (The test for standing vs. progressive looked only at oppositely directed wave vectors, and not at the phases of the waves.)

The data-taking and data-analysis techniques used by the Soviet teams tend to force the internal wave field to appear low-mode and non-stationary. The line sensors act as low-pass filters in vertical mode space (Roth, et al., 1979), while the data-adaptive methods take a momentary realization of the random process and emphasize it. It is crucial to distinguish between random fluctuations of a vector Gaussian process, and actual non-stationarity/intermittency that may be due to forcing functions or perhaps to non-linear behavior. Briscoe (1977b) tried this approach with deep-ocean measurements (and concluded that only the inertial motions looked particularly non-Gaussian), but the upper-ocean remains untested. We suspect it will seem like the deep ocean, i.e. basically Gaussian with its fluctuations in energy, coherence, and isotropy being

statistically expected, except during detectable generation events when the energy will increase, coherence will probably rise, and isotropy will probably diminish.

Pinkel (1975), working from a large and carefully taken data set based on an array of temperature profilers, argued that the wave-number bandwidths and energy distributions of the Garrett and Munk (1975) universal form were quite inappropriate in the last couple of octaves before the local buoyancy frequency cutoff. He points out that this is the range of internal wave frequencies that are trapped in the pycnocline, and thus may not be subject to the same processes that tend to make the lower-frequency spectrum universal. On the other hand, Desaubies (1975) offered an explanation for Pinkel's "anomalous" coherences based on the vertical modes that are trapped in the waveguide: away from the local buoyancy frequency the modes interfere destructively; near the buoyancy frequency the interference tends to be constructive, thus increasing the vertical coherence.

Käse and Clarke (1978) looked further at the question of energy and coherence peaks due to the shape of the buoyancy profile and forcing by surface stress. They concluded that the high-frequency energy and coherence anomalies (with respect to a universal spectrum) in their GATE data could be due to the rather narrow pycnocline overlying a thick stable layer: surface-generated internal waves, which must be of sufficiently low mode to penetrate through the mixed layer to the pycnocline, but not of large enough vertical scale to pass through the pycnocline, accumulated (it was hypothesized) in that narrow waveguide and seemed super-energetic. Presumably if one waited long enough, that extra energy would interact non-linearly (c.f., McComas and Bretherton, 1977) with the underlying universal spectrum and thus appear at lower frequencies through the parametric sub-harmonic instability mechanism. It would also become vertically symmetric through the elastic scattering mechanism, and effectively be absorbed into the full-depth universal spectrum to become part of that background internal wave field.

We feel that the Käse and Clarke (1978) view is likely to be the correct one: the seemingly special aspects of upper ocean internal waves are probably traceable to peculiarities of the buoyancy profiles and to propinquity to a strong forcing function, i.e. the wind stress. (Whether the wind stress acts directly or through surface wave generation as an intermediary is still an open question; probably both are relevant.)

Still relatively untested is the Roth et al. (1979) hypothesis of a universal spectrum as the background state in the upper ocean, but frankly we can't see how this could be incorrect: the interaction processes that shape the spectrum seem to be so strong, and the dissipation processes so weak, that expectations are low to find a sub-energetic spectrum.

An interesting speculation derives from the possible need for a certain kind of buoyancy profile (e.g., Käse and Clarke, 1978) to accumulate newly-generated internal wave energy:

perhaps a reason the Wunsch (1976b) and Wunsch and Webb (1979) searches for sources have been relatively unsuccessful is that the buoyancy profiles at their tested measurement points had no waveguide-like accumulation regions, hence any generated energy was immediately interacted away into the entire bulk of the fluid where it appeared as only a perturbation on the background universal spectrum.

Acoustical Fluctuations. Numerous analyses of internal-wave induced fluctuations of acoustic propagation (e.g., Flatte, et al., 1977; Desaubies, 1976a; Munk and Zachariasen, 1976; and Dyson, et al., 1976) have made substantial progress by applying the universal internal wave spectrum to the problem. The phase fluctuations (arrival times) are in good agreement with predictions based on the internal wave spectrum but the amplitude fluctuations are not. Ewart, et al. (1978) and Desaubies (1978) suggest the irreversible finestructure discussed below may be the significant process affecting the amplitude of the fluctuations.

Attention has also turned to the inverse problem: what can acoustic fluctuations tell us about the internal wave field? The pessimist's answer is that the open internal-wave questions are probably subtler than the resolution and accuracy of inverse methods.

Remote Sensing from Satellites. Present technology, whether in the visible or infrared spectrum and even considering synthetic aperture radar systems, so far permits only a glimpse of the surface effects of internal waves. Although these effects are sometimes substantial (see section below on Surface-Internal Wave Interactions), they are not detectable at all times and places even though the internal waves themselves might be present. In fact, with the possible exceptions of Apel, et al., (1975a) and Proni, et al., (1978), all reported satellite observations have been in relatively shallow water, i.e. continental shelf depths or less, and when pycnoclines were strong and shallow.

It is still not definite exactly what the satellite sees at the surface, although the preponderance of opinion (e.g., Phillips, 1977a) is that a mechanism like that of Hughes (1978) is operating: namely, the small-scale wind waves are affected by the patterns of surface currents due to the internal waves, and the resulting pattern of root-mean-square surface slopes causes a reflectivity pattern that is visible when the lighting conditions are right. Fedorov (1976) disagrees, feeling that the satellite is seeing alternate bands of turbid and clear water; this suggestion is unsubstantiated by *in situ* measurements (e.g., Apel, et al., 1975b).

Surely the most exciting use of a satellite in this regard would be in a series of measurements coordinated with those from a ship and/or moorings, but the few attempts so far (Apel, et al., 1975b; Haury, et al., 1979) have not been very informative. It would seem, however, that it is the coastal internal waves, probably the wave-packet cum solitary wave systems described below, that will continue to be the focus of satellite observations.

Solitons, etc. A surprising amount of work

has gone into theoretical study of solitary waves, or "solitons", although no proof exists yet that the ocean actually contains such things. The soliton is a mathematically-defined propagating disturbance that retains its form because of a fragile balance between dispersive effects trying to spread the disturbance out, and non-linear effects trying to steepen and close it up. The phrase "soliton-like" has become popular, and evidently means something that may be a soliton but can't be so proved. (R. Pinkel, personal communication, 1978, has a manuscript that may provide the proof, but it is still in negotiation.)

The Soviet workers have proposed the concept (e.g., Miropol'skiy and Sabinin, 1977) of a "soliton gas", which is a random ensemble of solitons propagating around in the upper ocean, passing through each other from time to time and responsible for the upper ocean being fundamentally different from the deep ocean; the difference arises from the small vertical length scale associated with the distance to the surface. Pelinovskiy and co-workers are leading proponents of this view, and have also presented (Pelinovskiy and Rayevskiy, 1977) the alternate case of interacting weakly non-linear waves.

The early observations of Perry and Schimke (1965) have been supplemented by Osborne, et al. (1977); their measurements of non-linear internal wave packets in the Andaman Sea show the most extreme and intense version of this phenomenon that has yet been seen, although a possibly stronger case may exist in the Sulu Sea northeast of Borneo (Apel, personal communication, 1978). Similar phenomena have been seen in Massachusetts Bay, most recently by Haury, et al. (1979), the Straits of Gibraltar (e.g., Boyce, 1975), in the fjords on the west coast of Canada (e.g., Farmer and Smith, 1978), probably in the Caspian Sea (Ivanov and Konyayev, 1976), in a few other locations, and in numerous aerial and satellite photographs (e.g., Sawyer and Apel, 1976; Elachi and Apel, 1976; and Fedorov, 1976). The existence seems usually due to a fortuitous combination of tidal currents, hydrography, and bathymetry (Maxworthy, 1979).

Whether these packets of internal waves are actually solitons or not is interesting but seems a point more of concern to fluid dynamicists than to oceanographers; they clearly are large, intense, momentary disturbances of the water column that contribute to vertical mixing (Farmer and Smith, 1978) and biological variability (Haury, et al., 1979).

It so far has not proven necessary to describe the oceanic internal wave field as being composed of a collection of solitons; weakly interacting non-linear waves (c.f. McComas and Bretherton, 1977) has been a sufficient description. Possibly some middle-ground between the two will prove fruitful, albeit complicated.

Dynamics

Large-Scale Energy Sources. Müller and Olbers (1975) and Müller (1976) proposed that the large-scale vertical shear due, for example, to eddies would strain the internal wave field in such a way that during its relaxation to its

original state it would drain energy from the shear, and thus keep the waves energized. The internal wave field acts as an eddy viscosity of about $0.4 \text{ m}^2 \text{ s}^{-1}$ against which the large-scale motions must work. Garrett (1979b) reviews the energy balances implicit in the argument. Consequences of the idea are intriguing because the mechanism allows the internal wave field to stay energized, the eddy field to be dissipated, and vertical mixing of at least $3 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ to occur (via dissipation of internal wave energy by breaking, and assuming a mixing efficiency of $1/4$; c.f., Olbers, 1976).

Unfortunately, the only real test of Müller's model has failed to support it. Ruddick and Joyce (1979) examined moored current meter data from the North Atlantic in an attempt to find the required correlations between large-scale shears and internal wave energies (they actually used vertically-acting Reynolds' stresses), and at best could find an eddy viscosity about 20 times smaller than Müller's estimate. Their explanation for the failure of the model involved Müller's assumption about the relaxation characteristics of the sheared internal wave field: Müller supposed the equilibrium state of the field to be maintained even if the mean flow changed, while Ruddick and Joyce proposed that the field has a new, different equilibrium state for a new, different mean flow. The former case requires the field to work against the mean flow to keep its basic state, whence the apparent eddy viscosity, whereas the latter case of a more flexible internal wave field produces no back effect on the mean flow.

Thus, progress has in a sense been backwards during the quadrennium. The summary energy balance in Briscoe (1975a), based on Müller and Olbers (1975), would no longer seem to be very convincing, and we are left with the suspicion that the internal wave field is eclectic about its energy sources: it will draw from winds (e.g., Goodman and Levine, 1977), mixed layer motions (e.g., Bell, 1978), surface wave interactions (e.g., Olbers and Herterich, 1979, and Watson, et al., 1976), flows over rough bottoms (Bell, 1975), whatever, as needed. The progress in the next quadrennium should be substantial in verifying this suspicion: JASIN (Pollard, 1978) could provide some of the clues.

Surface-Internal Wave Interactions. Watson, et al. (1976) and Olbers and Herterich (1979) have worked out the spectral growth of the internal wave field by an energy flux due to the interaction of two surface waves and one internal wave. Their conclusions are quite different: Watson, et al., argue that more than enough energy is readily available from the interaction to provide all the internal wave energy needed to maintain a dissipation rate yielding an eddy coefficient for momentum of $10^{-1} \text{ m}^2 \text{ s}^{-1}$. Olbers and Herterich, on the other hand, feel that while high-frequency internal waves in the seasonal thermocline under a not thick mixed-layer might be so generated, the flux to internal waves from surface waves is negligible compared to other source processes.

The different results are due to different assumptions about the degree of randomness of

the triad interactions: Watson, et al., do not perform any averaging across a randomly-phased ensemble of interacting waves, as do Olbers and Herterich. This is equivalent to assuming a fully deterministic system versus a fully random system, and the small fluxes obtained by Olbers and Herterich reflect the inefficiency of the latter system. Although waves in the real ocean may sometimes lean toward determinism, the random view would seem to be more reasonable most of the time.

No published experimental verification of surface-wave generation of internal waves exists, although the anecdotal literature supports the concept. The point is hard to prove because the surface waves needed for the interaction are usually associated with wind, which is also a direct source of internal waves (e.g., Käse, 1979, and Käse and Tang, 1976), and the theoretical result that the generated internal waves should propagate at right angles to the surface waves is difficult to substantiate because it is not easy to get a directional measure of internal waves.

The other piece of the interaction, namely internal wave effects on surface waves, has received attention theoretically (e.g., Rizk and Ko, 1978; Thomson and West, 1975; and especially Hughes, 1978), but the only oceanic experimental verification (finally!) is Hughes and Grant (1978); the satellite (and aircraft) observations depend greatly on this interaction in order to have something to look at, and the "soliton workers" (above) often report a visible surface effect.

Generation by Wind. Many analyses now exist of the generation of inertial and higher frequency internal waves by wind forcing; Thorpe (1975) reviewed the subject, and Goodman and Levine (1977), Krauss (1976a, b, 1978a, b), Käse and Tang (1976), Käse (1979), Miropol'skiy (1975a), and numerous others have added to the picture.

Johnson, et al. (1978) found a significant increase in the vertical displacement spectra when the winds were strong, and found the vertical profile to be step-like although it relaxed to a smoother profile two days later when the winds dropped. This presumably supports the idea of winds as a source of internal wave energy.

The missing feature of all analyses has been a description of the wind field: the problem is described in wavenumber-frequency space and depends upon the overlap of the time-space scales in the wind and in the internal wave field. The Garrett-Munk spectra and their modifications give some idea of the background internal wave field and Willebrand (1978) describes the large-scale, low-frequency space-time wind field, but not one wind description exists (that is, based on over-water measurements) that is really appropriate to the internal wave generation problem.

Until this deficiency is remedied, this area will not progress.

Recommended Reading

Garrett and Munk (1979) is an excellent review of internal waves, and is recommended

Fig. 1. Pictorial flow of energy through the internal wave field into fine and microstructure, and mixing. See text for discussion. Note especially the two kinds of internal wave fields: one that draws much energy from many sources and consequently returns much energy via dissipation in order to maintain its own energy level, versus one that needs little and delivers little.

reading to new and oldcomers to the subject. It does trail off in hopeful speculation on why the universal spectrum may exist, but the idea they present (based on Cox and Johnson, 1979) is simple and fascinating: if the internal waves do not dissipate quickly (Figure 1, bottom), then they can propagate large distances and diffuse (by wave-wave interactions) to even greater distances. This would spread throughout the ocean a "hot spot" due to strong sources, and hence make the energies everywhere seem similar, i.e. universal. Their estimates may be a little high (the 100-day estimate for low-mode interaction times from McComas, 1977, perhaps should be only 10 days, depending on the kind of interaction involved), but the idea is still applicable and is certainly attractive.

Thorpe (1975) is still a comprehensive review of generation, dissipation, and interaction, but needs to be updated by McComas (1977), Bell (1978), Olbers and Herterich (1979), and this present review.

Müller (1976) is important for the strength of its central concept of the large-scale shear as an internal wave energy source, and Ruddick and Joyce (1979) is equally important for showing that the concept may not be applicable.

Käse and Clarke (1978) contains the view we think will prevail for the upper ocean, and Leaman's (1976) use of rotary polarization needs exploitation, especially into the frequency domain, to test Käse and Clarke's ideas and to extend them down into the bulk of the ocean.

The analytical techniques of "consistency relations" in Müller and Siedler (1976) and of vector correlations in Calman (1978a,b) are powerful schemes to sort out the dynamics of the measured motions.

The Müller, et al. (1978) inverse analysis of IWEX (c.f. Briscoe, 1975b) virtually closes out

a decade of effort along a certain line of inquiry, namely moored observations of deep ocean internal wave fields. One suspects that it will require major advances in instrumentation or major changes in ideas before the subject will be reopened.

Internal tides (Wunsch, 1975b) are usually the largest vertical displacement signal in the spectrum, but the connection with coastal generation and low-mode, horizontally directional structure is still uncertain.

Mixing

The vast increase in small scale observations has made possible the identification of many mixing processes and has provided some information about where they may be found. Most of the major processes known are shown schematically in Figure 2.

To understand the significance of the mixing work it must be viewed in the context of the larger scale setting. A major division in settings can be made between the thermocline of the open ocean and the boundary regions, which include water mass transitions as well as the air-sea and solid boundaries.

Thermocline Processes

The classical thermocline problem sought to explain the smooth, nearly exponential decrease in temperature below 1 km by a vertical balance between a slow upwelling of cold bottom water and a downward turbulent heat flux. Contrary to some of the earlier work it is now apparent that the shape of the profile can be satisfied equally well by a wide range of turbulent fluxes (Welander, 1971). For fluxes corresponding to an eddy coefficient (see below), K_H , less

Fig. 2. Sources of fine and microstructure (c.f. Thorpe, 1975, Figure 5). Mixing results from some, but not all, of the processes.

than $10^{-5} \text{ m}^2 \text{ s}^{-1}$, Welander maintains that the thermocline will behave essentially as though no diffusion occurred, i.e. as an ideal fluid. Many of the models have assumed $K_H \sim 10^{-4}$, a value corresponding to turbulent fluxes 1000 times greater than molecular diffusion.

As shown schematically in Figure 1 different values of K_H will be accompanied by varying intensities of the accompanying microstructure signatures. The justification for obtaining fluxes from the microstructure is the same as used in the atmospheric case: the local production of temperature fluctuations by overturning motions is assumed to be balanced by local diffusive smoothing or viscous dissipation. A similar argument is used for turbulent momentum transports. By assuming that the other terms in the respective turbulent balance equations are negligible, the fluxes can be parameterized as turbulent exchange or 'eddy' coefficients. The fluxes can then be obtained from measurements of the mean profiles and the variances of the appropriate gradients; the dominant contributions to these variances come from the scales at which viscosity or diffusion smooths the fluctuations. These arguments, which were first applied to oceanic temperature data by Osborn and Cox (1972), can be expressed for the heat flux and horizontal momentum flux as:

$$F_H = K_T^H \frac{\partial \bar{T}}{\partial z} \quad \text{where} \quad K_T^H = \frac{\langle (\nabla T')^2 \rangle}{\langle \partial T / \partial z \rangle^2} k \quad (1)$$

$$F_M = K_Z^M \frac{\partial \bar{U}'}{\partial z} \quad \text{where} \quad K_Z^M = \frac{\epsilon}{\langle \partial U' / \partial z \rangle^2} \quad (2)$$

k is the molecular heat diffusivity and ϵ is the rate of dissipation of turbulent kinetic energy by molecular viscosity. The Cox number, defined by

$$C = \frac{\langle (\nabla T')^2 \rangle}{\langle \partial T / \partial z \rangle^2} \quad (3)$$

thus represents the ratio of the variance of the temperature gradients to the square of the mean gradient. For a "diffusive" thermocline with $K_H = 10^{-4} \text{ m}^2 \text{ s}^{-1}$ the Cox number would be 1000.

It should be noted that the use of eddy coefficients in the atmosphere has been most successful in the heated surface layer where the turbulence is strong and steady over periods long in relation to the turbulent time scales. Much less success has been experienced in stable surface layers or in free turbulent shear layers. By comparison with the atmosphere, the thermocline is very strongly stratified and the turbulence much weaker. Thus the above solutions should be regarded as hypotheses rather than as proven relationships. However, they also serve as good consistency checks; if the assumed turbulent gradient fluxes are present, then there must be observable signatures in the microstructure with correspondingly intense Cox numbers.

Another process of potential significance in the flux balance of the thermocline is salt fingering. The major question is whether salt fingering actually does occur throughout those large regions in the thermocline where it is possible. Theoretical arguments indicate that it should (Huppert, 1971), but the direct observations have been very difficult and the indirect signatures of step-like finestructures are largely absent. Enhanced downward fluxes of both heat and salt are produced by fingering; however, the density flux due to the salt is greater than that due to the heat. This net downward mass flux is in contrast to the turbulent gradient flux, which mixes light water downward. Thus, a thermocline balance between salt fingering and rising deep water is not possible. One alternative is that in the upper ocean, i.e. depths less than 1 km, salt

fingering balances the downward flux of light water due to Ekman pumping while the classical thermocline balance is maintained at greater depths (which is the region actually considered by the thermocline models).

Away from coastal boundaries or open ocean frontal regimes, the formation of salt-stabilized temperature inversions occurs infrequently. The finescale temperature profile thus decreases monotonically, if irregularly, with depth over large areas of the thermocline in mid-gyre. There is evidence for irregularities produced by vertical mixing of the kind assumed to produce the turbulent gradient flux (these are irreversible fine-structure) and by the transient strain of the smooth mean profile by internal waves (reversible finestructure; c.f., Lazier, 1973). A large part of the effort in finestructure has been directed toward separation of these effects. One approach has been largely indirect, the inference of finestructure properties required to explain the "noise" in moored observations, while another has emphasized the direct measurement of profiles.

Finestructure - from Moorings. Joyce and Desaubies (1977) used a Gaussian internal wave field and a Poisson model for the irreversible finestructure to explain time series of temperature and temperature difference obtained from the IWEX mooring (Briscoe, 1975b). Internal wave parameters were obtained by fits to the low frequency part of the moored temperature spectrum. The underlying assumption of the Poisson model is that the finestructure is due to a step-like sequence of completely mixed regions separated by thin regions of high temperature gradients. The temperature difference, ΔT , across a given interval, ΔZ , is assumed to be due to a random number of very thin discrete steps. The Poisson parameters were obtained from first-differenced CTD records taken near the IWEX mooring and used to form a structure function and then a vertical wavenumber spectrum. This representation of the finestructure accounted for approximately half of the variance under the observed vertical wavenumber spectra in the range of 0.03 to 1 cpm, and about 10% under the observed frequency spectra in the internal wave range. By acting as white noise, the finestructure also accounted for a decrease in the moored vertical coherence, MVC, between temperature sensors separated by more than the rms internal wave displacement. Time series of temperature difference, measured over 1.7 m in the vertical, were very sensitive to the finestructure and the modeled finestructure accounted for most, but not all, of the temperature difference variance for frequencies less than N , but gave an overestimate for frequencies greater than N .

A similar need for temperature finestructure to explain moored observations was reported by Müller, et al. (1978), and Eriksen (1978). The entire IWEX data were subjected to cross-spectral analysis by Müller, et al., to obtain internal wave parameters by least square fitting. In examining the coherences, they found that all slant or (nearly) vertical coherences showed rapid decreases for vertical

separations less than a few meters, compared to the slow and uniform decrease of horizontal coherence. They attributed this rapid initial decrease to temperature finestructure with vertical coherence scales of 2 m or less. Similarly to Joyce and Desaubies, they found a 10% finestructure contamination of the vertical displacement spectrum in the internal wave frequency band. Eriksen analyzed data from an H-shaped microscale array that had a string of temperature sensors and triaxial clusters of current meters. He found that all of the Poisson models gave a -2 spectral slope for the finestructure spectrum whereas a -2.5 slope gave a much better fit to his measurements.

The necessity for a current finestructure to explain various "noises" in moored velocity data was also a general result of these models. Briscoe (1975a) suggested such a possibility, and then (Briscoe, 1977a) analyzed temperature and current meter data from a string of instruments that had vertical separations of 2, 4, and 6 m. For a given vertical separation, the coherence between the temperature sensors remained nearly unity for frequencies less than N , and then decreased rapidly for higher frequencies. However, the current coherences began decreasing at very low internal wave frequencies. The "noise" causing the decrease in coherence had some vertical coherence itself and therefore was evidently not of instrumental origin. Briscoe attributed the loss of coherence to current finestructure with a vertical scale of approximately 1 m. Müller, et al., and Eriksen also found a need for current finestructure.

In the results discussed above, finestructure, both in temperature and velocity, appears as passive fields with vertical scales of 0(1 m) that are advected vertically past the moored sensors by low frequency internal waves. Although Joyce and Desaubies used a model that assumed irreversible finestructure, this assumption was not substantiated and, as pointed out by Eriksen, the models do not distinguish between the possible origins. The range of possible parameters for the finestructure has been narrowed, however.

Eriksen's observations also yielded a time series of Richardson number that was computed from the first difference shear and temperature observations made with a vertical separation of about 7 m. Using density that was estimated from temperature by assuming a mean TS relationship, Eriksen reported a sharp cut-off in the Richardson number at 0.25, which appeared to be due to increases in the shear rather than to decreases in the stratification. The time series of arctan Richardson number appeared to be spectrally white; hence, breaking did not occur preferentially at any frequency in the internal wave band (see footnote below). The low Richardson number values tended to be associated with temperature inversions across the array, suggesting the presence of shear instabilities in the density profile; however, due to the lack of salinity data the presence of thermohaline intrusions could not be ruled out. Eriksen's interpretation was that the added shear, due to current finestructure, caused the breaking of large scale internal waves* rather

than that the breaking represented the shear instabilities in smaller scale waves. The occurrence of these breaking events as well as the microstructure observations indicates that at least some of the modeled finestructure is formed irreversibly. The key to this is the high wavenumber behavior of low-frequency internal waves, because most of the internal wave-caused shear comes from that region of Fourier space.

* There is a semantic problem here. In a random superposition of internal waves, which produces a random field of shear, a Richardson number can locally decrease to less than one-quarter in a way unassociated with any particular internal wave Fourier component. There is no identifiable "breaking internal wave"; there is only a breakdown due to internal waves. The process is local in physical space, hence broadband in Fourier space.

Finestructure - from Profiles. GM75, together with the modification by Cairns and Williams (1976), assumed that the internal wave spectrum followed a -2 to -2.5 vertical wavenumber dependence to a sharp cut-off. This cut-off was assigned (by GM) to 2 cpm at the top of the thermocline, and was assumed to decrease linearly with N . From this Munk (1979) deduced that shear instabilities and convective overturning were equally likely to happen, but that convective overturning would cause more mixing and finestructure. The rate at which instabilities would occur goes like N or $N^{1/2}$, depending on the assumptions; either way, most of the action is in the thermocline.

Gregg (1977a) presented evidence, based on least-square fits to ensemble averaged spectra from CTD casts at mid-latitudes in both the Atlantic and Pacific, for a change in the slope of the vertical-profile temperature spectrum at a scale of approximately 10 m. The spectral slope was nearly -2 for less than 0.1 cpm, and about -3 for 0.1 to 2.0 cpm. Hayes (1978) reported additional evidence from low latitudes in the Pacific for the same spectral form. Johnson, et al. (1978) also reported a spectral break at 0.1 cpm. In addition, Hayes found, from a set of yo-yo profiles, that the dropped lag coherence fell off more rapidly than predicted by the GM model for vertical scales less than 25 m. This was attributed to non-internal wave effects that could be due to the semidiurnal tide or to finestructure. However, advection of the structures and ship drift could easily contaminate these measurements so that horizontal coherence was actually being measured at high wavenumbers.

Gregg's interpretation of the break-in-slope was that it was due either to a change in the dynamics of the internal waves at that scale or to finestructure effects. Clearly, if some of the structure at scales of 0.5 to 10 m is irreversible finestructure, then the internal wave spectrum must roll off to a slope steeper than -3 at 0.1 cpm in order to keep the total spectrum below the observed levels. As indicated by Garrett (1979b), this leads to a very

different picture of internal wave breaking than given by GM75: larger scale convective overturning becomes more likely.

Direct observations by velocity profiling of this cut-off of the internal wave spectrum have been elusive due to incomplete bandwidth resolution of the instruments. Sanford's Electromagnetic Velocity Profiler (EMVP) (c.f., Sanford, et al., 1978) has yielded information on the shear from the largest scales down to somewhat greater than 10 m. Other devices measure velocity fluctuations over scales of 1 m or less relative to free-fall bodies. Simpson (1975), using a small deflection vane mounted on the front of a 5 m long tube, found sharp shear structures over scales of several meters, as did Van Leer and Rooth (1975), who dropped dye pellets from a research submarine and estimated the shear from subsequent distortion of the dye streaks. Unfortunately, uncertainties in the response of the free-fall vehicles (such as Simpson's) to the larger scale shear make reconstruction of shears over scales larger than the body quite difficult.

A very different technique for distinguishing reversible from irreversible finestructure is to follow particular small scale gradient features laterally with a series of closely spaced profiles, or in time, with repeated casts at one site. The latter approach has so far proved ambiguous in the ocean due to the larger scale shear, but has worked in lakes. Lazier and Sandstrom (1978) accounted for the vertical migration of gradient features in terms of the forced modes of oscillation in a lake by including a weak turbulent viscosity, $K_M = 10^{-5} \text{ m}^2 \text{ s}^{-1}$, in the theoretical response function of the lake. This produces a gradual change of phase across the nodes of the displacement profile, and hence the vertical migration. The implication for the ocean is that this may also be true for progressive waves.

Johnson, et al. (1978) traced gradient features laterally by yo-yo'ing a CTD from a slowly moving ship (a technique now called a tow-yo). When the winds were relatively high, 20 m s^{-1} , they found a striking series of step-like structures between 200 and 400 m depth that were 20-30 m thick and extended laterally for at least 6 km. When the winds dropped, these structures disappeared, and the profile returned to a more typical appearance with irregular fluctuations that had little coherence between records that were separated by 0.6-2.0 km.

From the similarity in the apparent displacement profiles in temperature and salinity, Johnson et al. deduced that internal wave deformation was responsible for most of the structures larger than 5 m, which was the estimated resolution of their technique. Their analysis shows that the structures are consistent with an internal wave origin, but in our opinion is not a unique proof; structures formed by vertical mixing would have the same signature. However, they argue quite plausibly that the geostrophic adjustment of mixed structures would limit their horizontal extent to less than the observed coherence. At least some of the finestructure must be formed irreversibly, although this may represent a relatively small

fraction of the total fluctuations in typical profiles. Several structures 5 to 10 m thick with mean gradients of about 10^{-4} °C m^{-1} and sharp boundaries were found in the Sargasso Sea by Gregg and Sanford (1979). The strain required to produce such low gradients over relatively thick regions is very high and quite unlikely.

Microstructure - Fluxes and Dissipation Levels. Gregg (1977b) reported the results from three cruises, each of about two weeks duration, to the same site in the subtropical gyre of the North Pacific, and one cruise to an equivalent location in the South Pacific. The microstructure measurements were mostly from depths between 100 m and 1 km, but some records were taken as deep as 3900 m. When all of the profiles (each 125 m in extent) were averaged, those from October had a mean Cox number of 2, June a Cox number of 10, and February-March a Cox number of 59. These values include no allowance for possible isotropy in the temperature fluctuations which could increase the above values by, at most, a factor of 3. Contrary to the GM prediction of a rapid decrease of K_T with depth, the lowest Cox numbers were found in the main thermocline, with higher values in the seasonal thermocline and below the main thermocline. A much more limited set of observations was reported by Gregg and Sanford (1979) from an equivalent location in the Sargasso Sea. These measurements found individual 50 m thick sections in the Sargasso Sea that had Cox numbers from 4 to 500. The data were too few to form a representative mean value, but are consistent in level with those found in the Pacific.

Elliott and Oakey (1979) reported an extensive series of measurements taken between 30 and 130 m during the GATE experiment in the Equatorial Atlantic at 9° N and 23° W. The profiles were routinely made at least twice a day and at 10 minute intervals for several intensive periods. The Cox number throughout this time remained near or below 10. Also, an extensive series of measurements was reported by Williams (1976) from a drifting depth cycling capsule used 1000 km off the California Coast. In sections that did not contain finescale inversions (presumably due to thermohaline intrusions), the Cox numbers were roughly 20.

A series of velocity microstructure measurements was reported by Osborn (1978), who measured the small scale shear in the upper 500 m at several sites in the vicinity of the Azores Islands. He found that 80% of the column contained measurable shears over scales of approximately 0.1 m that corresponded in level to dissipation rates less than 10^{-8} W kg^{-1} . (By comparison, a global balance of dissipation and the mean rate of working by the tide and by the wind on the ocean corresponds to a dissipation rate of 10^{-9} W kg^{-1} .) Osborn's measurements assumed that the velocity structures were isotropic. The lack of such isotropy would reduce the values by possibly as much as a factor of 7.5.

Microstructure - Detailed Characteristics. The major contributions to the microstructure variances occur in relatively small scale

patches. No definitive statistics have been done on these patches, either in the vertical or the horizontal, but investigators usually report typical patch scales one to two times the length of the intervals which they use to average the raw data. Gregg (1977b) showed that some of the major contributions come from sections which have obvious density instabilities over scales of about 1 m. The difficulty in doing the patch statistics on the temperature microstructure is that the effect of vertical turbulent motions on the temperature field is modulated by the fine-structure. Elliott and Oakey (1979) used 10 m vertical averaging lengths in analyzing their very rapidly repeated profiles of temperature microstructure, and found significant changes on about a two-hour scale, which they suggest may be the lifetime of a typical microstructure patch. However, the difficulty of separating advection and ship drift from temporal evolution leave this an open question.

Spectral characteristics of the microstructure appear to be essentially the same as those outlined by Gregg (1975). Most investigators report a microstructure spectral range from 2-20 cpm in places where the Cox number is above background. Gregg (1977b) reported a range of variation from +1 to -1 of the slope gradient spectrum at these scales. The higher slopes are associated with the larger Cox numbers. The +1 slope is the same as the Batchelor (1953) prediction for passive scalars in a turbulent flow, and apparently reflects a patch with active turbulence. The -1 slope is the characteristic of the stratified fine-structure, and appears to reflect the absence of any mixing events. Elliott and Oakey (1976), investigating the detailed spectral characteristics for wavenumbers greater than the spectral maximum, found that the roll-off is faster than the form of the exponential decay predicted by Batchelor. They argued that due to the stratification of the profile, there is no turbulent cascade similar to that assumed by Batchelor, and that the microstructure is created over essentially the same scale range as it is observed; i.e. scales of 10 to 20 cpm. The observed spectral slope was approximated by assuming that initially sharp gradient features were smoothed by diffusion. Using this technique, they estimated the ages of various gradient structures.

Two approaches continue to be used to examine the three-dimensional temperature fluctuations. One uses a small thermistor mounted on the tip of one of the rotating wings of the free-fall vehicles, cutting through the water at a very shallow angle to the horizontal. Comparison of the spectra from this probe with those from a simultaneous vertical profile give evidence for the isotropy, or lack of it, in the profile. Gregg (1977b) found many cases with complete stratification, i.e., no indication of isotropy, out to scales of approximately 80 cpm; however, minor departures were found in some cases. Elliott and Oakey (1976) and Marmorino and Caldwell (1978) used another technique, that of comparing simultaneous observations from different temperature sensors mounted at the same level on free-falling vehicles. By examining the persistence of gradient structures

across their arrays, Elliott and Oakey found that features as small as 0.10 m were coherent across their array of four parallel thermistors. Marmorino and Caldwell used arrays of thermocouples as thermopiles, and compared the results with those of the thermistor mounted at the center. They too found that the horizontal coherence dropped at a scale of 10 cm, indicating that at least structures smaller than that were not sheets, although little can be said about the isotropy.

Discussion. The work described above appears to be consistent with an internal wave field that, in an equilibrium state at least, has a vertical wavenumber dependence that changes from a -2 to -2.5 slope to more than -3 at a scale of about 10 m. Internal wave deformation is the dominant process producing fluctuations in the profiles at larger scales. The case seems strong for some irreversible density fine-structure with vertical scales of a few meters but the extent to which it dominates the profile at 1-10 m scales is still unresolved. The nature of the current finestructure is also uncertain, and there are several possible origins: small-scale shear induced by larger scale internal waves shearing across the irreversible finestructure, sheet waves on the density finestructure (recall Woods', 1968, movies), and motions due to the lateral spreading of the partially mixed regions.

The microstructure measurements to date indicate the very weakly diffusive or ideal fluid type of thermocline model. Cox numbers high enough to support the $10^{-4} \text{ m}^2 \text{ s}^{-1}$ diffusivity obtained in models such as "Abyssal Recipes" (Munk, 1966) have not been found in the mid-gyre oceanic regimes. From the comparison of microstructure and finestructure the evidence seems strong that mixing in the thermocline does not often go to completion, i.e., few midwater mixed layers are found; rather, most of the profile has a very irregular sequence of high and low gradient regions.

One explanation for the low microstructure levels has been proposed in a paper by Schmitt and Evans (1978). They note that in the main thermocline the mean gradients necessary for salt fingers are far exceeded, and that the rapid growth rates predicted for fingering indicate that they should occur on time scales comparable to those of internal waves. They propose that the salt fingers become active on high gradient regions which are produced by the deformation of a smooth mean profile by internal wave straining and that, in turn, the buoyancy flux due to the salt fingers produces fine-structure with an alternation of high and low gradient regions. The salt fingering is predicted to be intermittent, but the net effect could result in very significant heat and salt fluxes over large areas of the main thermocline. The estimates of salt fluxes correspond to an eddy coefficient of $1-6 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$. They maintain that the microscale signature of the fingers has not been observed due to the vertical alignment of the fingers and the limitation of the resolution of the wing tip sensors to scales greater than 0.1 m.

Observations of salt fingering in the main thermocline in mid-gyres have not been obtained; the only observations are those from high gradient regions between sequences of nearly homogeneous layers that form in a few special locations: the original observations under the Mediterranean outflow, and more recent observations in the Tyrrhenian Sea (Molcard and Williams, 1975) and the Caribbean Sea (Lambert and Sturges, 1977). Direct observations were reported by Williams (1975), who used an optical profiler, and Magnell (1976), who towed a high frequency conductivity probe. The predicted signal due to the salt fingers in terms of the temperature and salinity signal is much weaker in the main thermocline and, if present, has so far escaped detection. A prediction of the spectral signature of a field of salt fingers made by Huppert and Linden (1976) and modified by Schmitt (1979) provide a reference for detection.

Boundary Processes

Mixed Layers and the Upper Ocean. A number of observations were reported during this period that tried to elucidate the mechanisms responsible for mixing at shallow depths in the ocean. The differences in these observations and in the conditions and processes occurring when they were made shows the very complex nature of the dynamics of the upper ocean.

Gregg (1976) made some observations of a late winter storm at moderate latitudes in the North Pacific. Initially, there was very strong lateral variability, apparently the residue of previous mixing events. When a mild storm went by, there was surface cooling with convection that formed deep adiabatic mixed layers 100 m thick. Spectra of the temperature profiles were consistent with those expected for fully developed homogeneous turbulence when the depth of the mixed layer was less than the Monin-Obukov length, L_b ; later in the storm, when the winds had relaxed and the depth of the mixed layer was much greater than L_b , the spectra were no longer consistent with the "universal" forms.

In spite of the strong surface-driven vertical mixing, the change in the heat content of the upper 200 m in the study area was dominated by lateral effects; the net heat flux through the surface was only one-eighth as large as the lateral flux due to several intrusions. During the storm a 50 m thick lens, which had no apparent connection to the surface, came into the area at a depth between 150 and 200 m. Later, when the winds relaxed, lateral spreading of light surface water from some other locations overrode the upper half of the local mixed layer at one station. This, together with the earlier deep intrusion, produced a triple-step structure of 50 m thick layers in the density profile.

A beautiful series of measurements was reported from Loch Ness in a series of papers by Thorpe and co-workers. These observations utilized a fixed, and at times, a yo-yo'ing array mounted on the bottom that gave time series observations of the detailed temperature and shear structure through shallow mixing events. These observations, reported by Thorpe,

et al. (1977), Thorpe (1977a), Thorpe and Hall (1977), and Thorpe (1978a), generally showed that many of the mixing events were consistent with Kelvin-Helmholtz instabilities. In one case, they observed thermal structures (at depths comparable to, or greater than, L_b) that appeared to be due to eddies, were aligned across the wind direction, and had scales comparable to the height of the sub-surface mixing layer. The mixing layer was found to be statically stable in the mean, but with a low Richardson number. The overturning scale of individual events within the mixing layer was about 0.15 of the size of the active patch. For scales less than the rms overturning scale, the temperature structure tended to be isotropic. In another series of observations, the Reynolds number of the mixing events appeared to be about 4000, a rather low value in comparison to what one would expect for fully developed turbulence. The billows were found to be overhanging only 20 to 30% of the time.

Preliminary results from the MILE experiment reported by Dillon and Caldwell (1979a) show sharp, overhanging events at the base of the mixed layer during a deepening phase of a warm, heating storm in the North Pacific. Their observations are quite similar to the structures reported by Thorpe.

Osborn (1978) noted in his observations of dissipation near the Azores that regions of very high dissipation up to 45 m thick were present at the base of the mixed layer at stations close to the islands following a very strong wind event. This was consistent with the mixing expected for a slab response of the mixed layer to the winds. A quite different case was a 150 m deep turbulent mixed layer observed in the Sargasso Sea by simultaneous observations of three profilers; one measured the large scale shear, another the velocity microstructure, and the third temperature microstructure. As described by Gargett, et al. (1979), and by Gregg and Sanford (1979), a mean shear of 10^{-2} s^{-1} across the base of the mixed layer produced turbulent spectra in both the micro-scale shear and temperature spectra. Estimates of ϵ , based on an integration of the shear spectrum and an assumption of isotropy, agreed within a factor of 4. Since surface winds were light during these observations, the turbulent stirring of the mixed layer resulted from a larger scale shear, possibly due to a mesoscale eddy or to convective cooling.

Unlike the work discussed above, which was based on vertical profile observations, Dillon and Powell (1976) estimated dissipation rates in the mixed layer of Lake Tahoe assuming that their observations of velocity fluctuations from moored current meters could be interpreted as three-dimensional turbulence. As D'Asaro (1978) showed, in the unstratified mixed layer those velocity fluctuations having typical internal wave time scales are probably the exponential tails protruding into the mixed layer from the free internal waves in the thermocline underneath. True, the internal wave currents decay exponentially into the mixed layer, but the rate of decay is scaled by hundreds or thousands of meters, hence are nearly undiminished even at the surface. Only the vertical motions due to

internal waves really disappear in the mixed layer.

A very powerful new technique was demonstrated by Hauray, et al. (1979) who obtained simultaneous observations of a 15 m high overturning event in Massachusetts Bay with a CTD and a towed high-frequency acoustic echosounder. The ability to detect such events from a device that can be towed at speeds of 4 m s^{-1} or more offers exciting possibilities for future work.

Interleaving. During this quadrennium, a great many additional examples have been reported of salt-stabilized temperature inversions formed by lateral advection and a survey of much of the earlier work was published by Fedorov (1978). Most of the inversions have been found in regions with larger scale horizontal gradients, usually connected with fronts of some sort. One group of these papers is concerned primarily with those features that generally have vertical scales from a few meters to 10 or 20 meters, and are found most commonly in coastal regions. These features have been reported south of New England by Voorhis, et al. (1976), on the Oregon Coast by Marmorino and Caldwell (1978), in Bristol Bay, Alaska by Coachman and Charnell (1977), off Nova Scotia by Horne (1978), in the New York Bight by Posmentier and Houghton (1978), and off the Washington-British Columbia Coast by Gargett (1976b, 1978). Most of these interleavings are believed to have been formed by cross-frontal mixing.

Other examples of interleaving are found on the boundaries of continental slopes. The largest example was recorded along the continental margin of Antarctica (Carmack and Killworth, 1978): water formed at the surface and flowed along the shelf until it stabilized at an appropriate density level, where it was observed as a 200 m thick lens with horizontal dimensions of 100 km x 800 km. Much smaller examples, reported by Armi and Millard (1976) and by Armi (1978) in the North Atlantic, are believed to have been formed by current-induced mixing over irregularities on the sea floor with a subsequent flow out into the ocean interior where they were found as lenses of water above the bottom.

Fronts, and associated interleaving, in the open ocean were recorded by Georgi (1978) in the Antarctic Polar Front, by Williams (1976) at the outer edge of the California Current, and by Gregg and Sanford (1979) in the Gulf Stream.

The microstructure observations that were made through intrusive features (Gargett, 1976b, 1978; Williams, 1976; and Gregg and Sanford, 1979) continue to show evidence for intense microstructure on the upper and lower boundaries of the salt-stabilized temperature inversions. As pointed out by Gargett, the fact that the variances are high in these regions is indirect evidence of double diffusive activity since the gradients are in the correct sense. The existence of prominent step structures found in many of these cases by the free-fall observations is more direct evidence. Particularly noteworthy are the very pronounced step structures above and below the inversions found in the Gulf Stream, indicating that the very strong

shears in the Stream do not completely inhibit the double diffusivity processes.

A lens of intrusive water was tagged with an acoustic float by Voorhis, et al. (1976), and followed for a period of several days. CTD casts, using the float as a reference, showed the layer decreasing in thickness from 15 m to 3 m. The thinning was attributed to vertical shear while changes in temperature and salinity, possibly due to double diffusive activity, appeared to give a time scale for the lenses of one to three days.

Both Georgi (1978) and Horne (1978) observed that the temperature features seemed most coherent parallel to fronts, leading to the suggestion by Horne that the layers originate as sheets and not as tongues, i.e. the structures appear as thin "ribbons" lying parallel to the front rather than perpendicular to it. The intrusions appear to have been stretched along the frontal direction, as suggested by Voorhis, et al., (1976).

Some papers report density instabilities on the boundaries of the thermohaline intrusions: an example taken with a Plessey STD had 10 m scale instabilities while another, using a Brown CTD, had density inversions over a 1 m scale. These are at variance with the microstructure observations and we believe that they are instrumental artifacts due to mismatch in time constants for the conductivity and temperature sensors.

Georgi (1978) explored the possibility that intrusions near the Antarctic circumpolar front were due to horizontal internal wave displacements in the presence of the strong mean horizontal gradient associated with the front. However, displacements based on the GM75 model yielded predicted "reversible inversions" much weaker than the observed finestructure.

Based on these observations of the ubiquitous nature of intrusions and their often intimate association with double diffusivity processes, laboratory experiments are being done to simulate the formation of intrusions in doubly diffusive regimes. Turner (1978) reported a series of experiments in which an intrusive lens of water was released into a profile at its boundaries; the intruding fluid moved across isopycnals in a way that was related to the mean gradients in the environment and their relation to the source fluids. Shadowgrams showed pronounced salt-fingering and diffusive-convective layering in the appropriate places on the intrusions. He also showed that similar features could arise from opposing mean gradients if there are lateral gradients present in the water. In this case, it is not necessary for the intrusion to be initially formed by a density current; it can form by a small perturbation on the diffusively unstable profile. Observations of large scale intrusions at the Polar Front in the Drake Passage by Joyce, et al. (1978) show the type of behavior predicted by Turner's results, as do Horne's (1978) contour maps.

The implication of these results is that strong horizontal and vertical fluxes are associated with these intrusive features. Joyce (1977) has proposed that the horizontal fluxes associated with intrusive motions can be

estimated by using a triple-scale decomposition of motion near a front; the largest scale of variability is the front itself, the intermediate scale is the intrusion, and the smallest scale is that of the microstructure on the intrusions. With this proposal, he obtained estimates of the horizontal fluxes based on measurements of horizontal and vertical fine-structure and microstructure gradients. The suggestion that the fluxes can be obtained so simply is appealing but Turner's experiments suggest that the processes are much more complex. Before such an approach is used more must be learned about the temporal and spatial scales of the intrusions and the processes forming them.

Jets. During the quadrennium, the first microstructure measurements have been reported from intense jets at the Equatorial Undercurrent and in the Gulf Stream. The equatorial observations revealed mixing regions much more intense than those found in the main thermocline. Gregg (1976) reported observations from the undercurrent at 155° W in the Central Pacific, where a 30 m thick region midway between the surface and the velocity maximum showed intense mixing with a Cox number greater than 3300. Although the section had an overall stable stratification, there were sharp 0(1 m) overturning events throughout the mixing layer. The microstructure levels at the velocity maximum were much lower, with a Cox number of about 40. A somewhat less intense region of microstructure was found in the weakly stratified thermocline between 300 and 400 m deep in which the mean Cox number was about 100. The pattern and level of these measurements, if the gradient flux relationship and equation (1) are used, are in quite good agreement with the model of the Equatorial Undercurrent put forward by Robinson (1966).

A series of papers by Osborn and co-workers (Crawford and Osborn, 1979a, 1979b; Osborn and Bilodeau, 1979) reported on observations made during the GATE experiment in the Equatorial Undercurrent of the Atlantic. The temperature structures found were similar to those reported from the Pacific: a region above the velocity maximum with a mean Cox number of 4100; a relative minimum between 10 and 45 in the Cox number at the velocity maximum; and another mixing region below the maximum. The most intense velocity microstructure was found above the core where the mean dissipation rate was found to be $3 \times 10^{-7} \text{ W kg}^{-1}$. Using the dissipation method, (2), the value of K_M ranged from 10^{-2} to $8 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ in the high dissipation regions, to less than 10^{-5} near the velocity maximum. They also applied their measurements to a discussion of the overall energetics of the undercurrent.

Two sets of observations were reported from the Gulf Stream. Oakey and Elliott (1977) made two broad transects of microstructure profiles beginning in the Sargasso Sea and crossing the Stream into slope water. The data, analyzed over 100 m thick sections, showed values of the Cox number that were primarily functions of the water mass, irrespective of the depth or north-south position of the section along the

transect. The values of the Cox number obtained ranged from 20 to 520. They suggested that the mechanism of formation of the microstructure might be the dissipation of internal waves on finestructure features.

Gregg and Sanford (1979) reported several simultaneous profiles of velocity shear and temperature microstructure made in the Stream. They found Cox numbers very similar to those reported by Oakey and Elliott (1977) with a maximum value of 500, averaged over 50 m. The detailed signatures of the microstructure showed the major activity to be due to double diffusivity signatures on the upper and lower boundaries of the numerous salt-stabilized temperature inversions. The Cox numbers from regions that were monotonic on the fine scale ranged from about 30 to 50. Thus, from these observations, it is apparent that the level of small scale mixing in the Gulf Stream is much less intense than that found in Equatorial Undercurrents. While there may be some small regions of intense mixing, there were not the thick sections of overturning events and very high Cox numbers found on the Equator. Thus, the Cox numbers indicate vertical mixing considerably less than the values of up to $10^{-3} \text{ m}^2 \text{ s}^{-1}$ that have been used in some of the numerical models of the Gulf Stream. Lateral intrusions across the Stream may therefore be the dominant process in the flux balances of temperature and salinity.

Bottom Boundary Layers. The use of Brown CTD profilers close to the sea floor has shown well-mixed boundary layers that are very similar in appearance to surface-mixed layers. Armi and Millard (1976) presented numerous examples of 10 - 100 m thick layers whose occurrence was related to the velocity obtained from nearby moorings. Armi (1978) later reported that some layers appear to advect away from the site of formation on the bottom and form mid-water mixed layers at sites where the water depth is greater. He argued that all of the necessary vertical flux for global balances similar to the thermocline models could be supplied by this mechanism; Garrett (1979a) and Armi (1979) discuss the energetics further. The similarity of these intrusions, resulting from the mixing over the bottom, to the intrusions occurring at the surface is most striking.

Microstructure measurements were made in a mixed layer on the Oregon Shelf by Caldwell (1976). He reported a mixed layer 10 m deep with a super-adiabatic gradient in the bottom 5 m. Spectra of the temperature gradients had some of the characteristics of those expected for homogeneous fully turbulent fluids. The number of observations, however, were too limited to make a detailed estimate of the dynamics taking place, but the measurements were not inconsistent with a logarithmic velocity profile beneath a turbulent Ekman layer.

Side Walls. Hogg, et al. (1978) reported the presence of several regions of intense temperature finestructure activity immediately adjacent to Bermuda. The regions of intense activity appeared where mesoscale eddies or rings were forcing flow parallel to the bottom

contours. It was also found that the shear levels measured by EMVP over scales of 20 to 50 m were more energetic at stations adjacent to the island than at the off-shore sites. However, no relationship was found between variations of the shear intensity and the temperature finestructure. Microstructure measurements by Gregg and Sanford (1979) showed intense activity in the same regions as the strong finestructure. Some of these regions had Cox numbers (averaged over 50 m or more) as high as 750, values much greater than those reported from the open ocean thermocline in monotonic profiles. There was a close relationship between these regions of intense microstructure and pronounced step-like finestructure of high gradient regions separating nearly adiabatic layers from 3 to 8 m thick. Spectra of the temperature fluctuations in some of these nearly adiabatic regions showed rising gradient spectra, which suggested active mixing. However, the well-mixed nature of the finestructure indicated that the regions had been mixing for some time before the observations were made. These measurements of well-mixed regions yielded a much smaller vertical scale than those which had previously been found in the vicinity of Bermuda. Nevertheless, the correlation of the finestructure and microstructure observations showed that very strong mixing activity can take place on the slopes of islands.

Hydrothermal Activity. Several examples have been found of hydrothermal activity. Rona, et al. (1975) discuss one example from the hydrothermal field on the Mid-Atlantic Ridge. Over one wall of a rift valley, 20 m above the bottom, a horizontally towed thermistor array found what appeared to be a plume with a 350 m wide temperature anomaly of 0.1°C . Over another wall of the valley, a super-adiabatic temperature gradient was found in the bottom 500 m of the profile. Rona, et al., hypothesized that this was due to larger scale convective transport from the bottom. Too little is known about these features to decide whether they are significant mixing agents or just interesting curiosities.

Discussion

The Garrett-Munk framework for internal waves has been an asset to the field for it has helped bring together disparate observations into a common pattern, and has helped guide additional experimental work. There has been, however, a tendency for purely descriptive work on internal waves to be inhibited. Good, careful descriptive contributions should always be welcomed, for they are part of the necessary feedback between theory and experiment.

Fine and microstructure work has, on the other hand, no similar theoretical framework, and purely descriptive contributions are common. If, however, they are not carefully done, they will not be helpful and can even hurt the growth of the field. There has also been a tendency to overstate the importance and generality of a particular process and its effects on the ocean.

Many, although undoubtedly not all of the pieces, of the mixing puzzle have been found. It is entertaining, and can be instructive, to imagine the entire jigsaw puzzle from the available segments, and the picture we see in the thermocline is the following. Much evidence for finestructure with vertical scales greater than the rms internal wave displacement ($7 - 20$ m in the thermocline, varying in depth as $N^{-1/2}$) is consistent with formation by internal wave strain. Except for the thermohaline intrusions there is not much suggestion of other processes creating finestructure at those scales. The evidence for scales less than the rms internal wave displacement points to dual roles of internal waves and irreversible mixing. Whether one is dominant is not clear; there are direct observations of overturning events on these scales and this is also the range of thicknesses found in the microstructure patches.

The relatively low intensity of the microstructure and the fact that little of the mixing produces midwater well-mixed layers are consistent with a weak turbulent heat flux. If Welander's (1971) criterion of K_H^2 less than $10^{-5} \text{ m}^2 \text{ s}^{-1}$ is used, the thermocline appears to be much closer to the ideal fluid case than to a diffusive steady state. The few dissipation measurements, ϵ , reported from the thermocline seem low but uncertainties in the mixing efficiency make it difficult to relate this to the heat fluxes. Hence, evidence favors a weakly dissipative internal wave field but the question is still very open. To help close the question, we need simultaneous and direct measurements of kinetic energy losses from internal waves and potential energy gains in the finestructure field.

The specific mechanisms by which the turbulent mixing takes place in the thermocline are largely undetermined. The various mixing length arguments deduced from the GM internal wave parameterization are strongly dependent on the high wavenumber behavior of the internal wave field, which is poorly known, and which may even be a poorly-posed question. However, a strong case has been made for the importance of the Kelvin-Helmholtz instability at shallow depths, at least in some special situations. The microstructure signatures contradict the requirements for fully developed homogeneous turbulence but a maximum $+1$ slope of the temperature gradient spectrum in active patches seems to hold. However, even though that Batchelor spectrum was derived without assuming a "fully developed" turbulence, there are definite departures from the predicted form in the cut-off range.

An exciting development in the field has been the progress in our understanding of double-diffusivity processes in the ocean. This has developed by a coupling of laboratory and theoretical work with observations (the same kind of coupling has greatly aided internal wave understanding) and has been particularly fruitful with thermohaline intrusions. Much more work is required to understand the dynamics of intrusions, both diffusively stable and unstable ones, and to be able to make realistic estimates of the associated fluxes. This is essential for frontal and continental shelf

dynamics, but, the broader implications are unknown. The various types of intrusions appear to be major factors in the vertical as well as the horizontal flux budgets of heat and salt in the ocean. If the thermohaline intrusions, and possibly salt fingering in the thermocline, are dominant in the budgets of the upper ocean, then there must be a general downwelling of light surface water, presumably driven by Ekman pumping.

The developments in mixed layer work, particularly when the work in progress on MILE and JASIN is considered, offer the prospect of a major advance in understanding during the next quadrennium. Strong turbulence, with forms close to the "universal" spectra, has been found within mixed layers when the surface forcing is sufficiently high. It is clear that meteorological measurements sufficient to make good estimates of the Monin-Obukhov length are required as well as at least some information on the full three-dimensional response of the upper several hundred meters.

Most of the problems in mixing and internal waves require observational answers before enough pieces can be assembled to permit comprehensive theoretical treatment. This will require simultaneous three-dimensional time histories of the major parameters. The work discussed above represents steady progress in that direction.

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