Bubble clouds and the dynamics of the upper ocean

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SUMMARY

This is a review of turbulence in the upper ocean close to the sea surface, particularly of the information that has been obtained from sonar observations of bubble clouds produced by breaking wind waves. These clouds provide tracers of the turbulent motions and are important, especially at high wind speeds, in the process of air–sea gas transfer. The observations of bubble clouds are here related to other measurements of turbulence, particularly to direct measurements of currents and temperatures in lakes or at sea, and to laboratory studies. Some novel observations of bubble clouds and breaking waves, their frequency and relation to Langmuir circulation, are presented.

There is now emerging a pattern of clues that point to the dominance of breaking surface gravity waves as a source of turbulence to a depth below the surface of 0.04 to 0.2 times the wavelength of the dominant breaking waves, although the effect of swell has yet to be clarified. The relative depth appears to increase with increasing values of \( W_0/c \), where \( W_0 \) is the wind speed and \( c \) the phase speed of the dominant waves. Below this region the generation of turbulence may be dominated by shear-stress or convection. Here, turbulence is generally similar to that in the atmospheric boundary layer. There are, however, coherent motions on the scale of the mixing layer that persist for periods of a few minutes to an hour or so, to which the transport of a large part of the momentum and heat fluxes can be attributed, and which strongly affect the dispersion of buoyant particles or bubbles. These motions deserve special study to establish their contribution to heat and momentum transport, and hence to determine if, or when, they should be specifically represented in models of the upper ocean devised, for example, to describe the dispersion of passive and non-passive tracers or the air–sea transfer of gases.

1. INTRODUCTION

An understanding of the upper-ocean boundary layer, the region from the ocean surface to the main thermocline, that part of the ocean most directly affected by, and responsive to, the atmosphere, is incomplete and at worst totally inadequate given the importance of the layer in models seeking to estimate the oceans’ response to atmospheric forcing or to estimate the exchange of heat, momentum or gases between the oceans and the atmosphere. The layer is a buffer zone between the atmosphere and the ocean, reacting slowly and with some resistance to those signals that reach and manage to penetrate the highly stable interface, the ocean surface, where density increases sharply; water is some 800 times more dense than air. Surface currents are typically far less than surface winds but may carry far greater momentum per unit volume. Resistance is exemplified by gas diffusion. The diffusion of gases in air is typically \( 10^4 \) times greater than in water so that, generally, air–sea transfers are controlled by the liquid phase. Recall the facts, so clearly articulated in Gill’s (1982) book, that 2.5 m of the ocean has the same heat capacity per unit area as has the whole depth of the atmosphere, and that the ocean, unlike the atmosphere, transmits light poorly so that, in consequence, typically 80% of the energy at wavelengths of solar radiation that enters the surface of the ocean is absorbed in the top 10 m. The dynamics of the upper-ocean boundary layer determine
how the heat entering the ocean is distributed and hence, for example, the temperature of the sea surface, a vitally important lower-boundary condition for models of atmospheric circulation. Lack of knowledge of the layer is identified by the Intergovernmental Panel on Climate Change (1990) as contributing significantly to the uncertainty of climate prediction.

Recognizing the central role of the upper-ocean boundary layer in atmosphere-ocean dynamics, it is surprising how little is known for certain of the processes that create and drive mixing and turbulence within it. Can we think of it and treat it simply as a sort of upside-down atmospheric boundary layer?

The fairly extensive and thorough studies of the atmospheric boundary layer have recently been described by Donelan (1990) in a review of air-sea interaction. He concludes, however: 'The widest gaps in our knowledge of air-sea interaction are in the water boundary layer and its intimate relationship with surface waves'. Very few reliable measurements of turbulence have been made in the upper ocean in comparison with those in the atmospheric boundary layer. Accurate measurement is difficult, not only because of the general absence of stable platforms from which to make observations, but also because wave-induced motions often exceed turbulent motions and mean currents, and it is hard to separate the various components. The technology now exists to address one particular area, and here more information may soon be forthcoming. Donelan remarks: 'The study of wave breaking and near-surface turbulence is probably the most exciting and rapidly expanding aspect of air-sea interaction'. This is at the heart of this review.

It is fortunate that, in the process of breaking, waves introduce into the sea a tracer—or cloud of tracers, bubbles—which may help provide clues to establish the relationship between the waves themselves and turbulence. Bubbles are not, however, a passive tracer; they rise and dissolve, and their introduction to the ocean adds, of course, to the complexity of the multiphase medium, already containing particles and live organisms which may, in some circumstances, be so dense that they absorb significant amounts of short-wave solar radiation and control the surface temperature (Sathyendranath et al. 1991). However, as I hope to show, the benefits of bubbles as tracers that aid observation outweigh the disadvantages they introduce by adding to the complexity of the physics. Although the ocean is relatively opaque to light, it transmits sound very effectively, and sonar can be used to detect and track bubble clouds. This offers a powerful diagnostic research technique to study the connections of waves and turbulence, the kinetics and dynamics of the upper ocean, without interference with the flow field.

My focus is on what, pertinent to the investigation of the dynamics of upper ocean, can be learnt from bubble clouds. Although the volume of this information is relatively small in comparison with the vast amount of data obtained about winds, convection, lee waves, turbulence, fronts, cyclones, and so on, from clouds of droplets in the atmosphere (how fortunate are meteorologists that clouds are visible!), it has helped considerably to clarify and identify processes in an important region where information was previously lacking. Some comparison of the dynamics of the upper-ocean boundary layer with those of the atmospheric boundary layer is useful and profitable (see section 4(a)), but this does not form the major thrust of this review. Nor shall I say much about the importance of bubbles in scouring particles and, for example, bacteria from the ocean (Blanchard and Syzdek 1982) in producing jet drops and film drops (Resch et al. 1986) or aerosols (Bortkovskii 1983) as they burst, of their effect on albedo (Jerlov 1976), of the production of sound as bubbles are generated (Pumphrey and Crum 1988) or resonate as clouds (Prosperetti 1988; Lu et al. 1990; Yoon et al. 1991), of the sound caused by bubble oscillations after their formation by rain (Pumphrey et al. 1989; Laville et al. 1991) or of
the muffling of sound from the sea surface caused by a layer of bubbles (Farmer and Lemon 1984). Their importance in these areas is considerable and self-evident, but I leave review to others who are better qualified to write it.

2. THE FORMATION OF Bubbles

Blanchard and Woodcock (1957) identified three main sources of bubbles—breaking waves, precipitation nucleation, and the formation of bubbles on particles in supersaturated conditions. Little appears to be known of the latter, but it seems that in the absence of precipitation and in wind speeds exceeding about 3 m s\(^{-1}\), wave breaking generally provides the dominant source of bubbles.

Wave breaking and the formation and nature of bubble clouds are subjects in which knowledge is still wanting. There are several distinct types of breaking waves, though until recently these were largely unquantified either in their frequency or their efficiency in producing bubbles or turbulence. Spilling waves, in which air is entrained near the crest, subsequently producing an air–water mixture that advances down the forward face of the wave, and plunging breakers, in which a jet produced at the crest falls on to water on the forward face of the wave (see later), both occur in deep water (Cokelet 1977; Longuet-Higgins 1988) and contribute to producing bubble clouds. These breakers appear to be most commonly associated with waves near the peak of the wave-frequency spectrum and with wave groups, which impose a deterministic pattern to breaking (Donelan et al. 1972; Thorpe and Humphries 1980); successive waves break with a periodicity approximately twice that of the dominant wave period. This pattern of breaking has been identified in sonographs of sound scattered from bubble clouds (Thorpe and Hall 1983) and in the spectrum of sound produced by breaking waves (Farmer and Vagle 1988).

More subtle, but perhaps less violent, forms of air injection take place via the steepening and breaking of short waves on the crests of longer waves (Longuet-Higgins 1987) or by the generation of capillary waves on the forward faces of short gravity waves, where they may steepen and break with bubble formation in their troughs (Longuet-Higgins 1963; Crapper 1970). Evidence for the coexistence of two distinct populations of breaking, perhaps reflecting these different mechanisms, comes from measurements by Xu et al. (1986) of the height and period of breaking gravity waves in a laboratory wind-wave channel (see Longuet-Higgins 1988). The relative importance of these different breaker types in the generation of turbulence or bubbles is unknown, but probably depends on the relation between the local sea state (a function of fetch and time-history of the wave field) and the wind speed (a factor which contributes a measure of the inherent unsteadiness of the wave field contributing to wave breaking (Yuen and Lake 1980)). Winds and wave fields that may be regarded as steady are rarely found and, in the absence of a better indicator, a parameter, the inverse ‘wave age’ \(W_{10}/c\), is often used to characterize the wind–wave relationship rather than referring it to the wind speed alone. \(W_{10}\) is the wind speed at 10 m above the mean sea surface level and \(c\) the phase speed of the dominant waves. The friction velocity in the air is often used in the parameter in place of \(W_{10}\), but \(W_{10}\) is preferred since it can be measured more easily. The mean sea surface slope (the ratio of the significant wave height to a wavelength determined from the zero-crossing period) derived from the JONSWAP and Pierson–Moskowitz spectra increases with \(W_{10}/c\), favouring more frequent breaking as the ratio becomes larger.

The spatial and temporal variability of breaking waves is not well known. Measurements of the number of wave crests breaking at a fixed point per wave period have been
reported by Longuet-Higgins and Smith (1983), who used a ‘surface jump meter’ and a criterion based on the time derivative of wave height to detect the breaking waves; and by Weissmann et al. (1984), who determined ‘breaking’ from the higher-frequency component of surface elevation. Given the uncertainty in the choice of the subjective criteria for identifying breaking, and differences in wave and fetch conditions, it is not surprising to find some discrepancy in the estimates; Longuet-Higgins and Smith found about one breaking or very steep wave per 100 wave periods, and Weissman et al. found 8.6 waves breaking per 100. Thorpe and Humphries (1980) estimated the equivalent spatial scale, the number of breaking waves per wavelength, to be in the range 0.01 to 0.1. (These relatively simple, but fundamental, observations form, as we shall see later, the basis of other key estimates, and should be further extended to a variety of wind and fetch conditions; see section 4(d).) Other than the wave-group-related periodicity referred to earlier, no spatial or temporal pattern of breaking waves in deep water has been noticed, except where atmospheric effects such as rainfall and gustiness are prevalent, or where internal waves or oceanic fronts play a significant role in modifying the energy density of the wave field and create local regions or bands of breaking waves by providing an underlying field of current convergence and divergence or perhaps affecting the stability of the atmospheric boundary layer.

Several mechanisms are involved in the entrainment of bubbles into the ocean by a breaking wave. The jet produced at the crest of a plunging breaker falls and strikes downward on the forward face of the wave, partly rebounds producing droplets and spray, and again falls to strike the surface (Longuet-Higgins 1988). This rebounding process may be repeated several times. The part of the jet penetrating the surface carries a downward component of horizontal momentum, imparting an asymmetry in momentum transport at the air–sea interface, and entrains with it air which immediately breaks up into bubbles. The falling spray further contributes to bubble entrainment. There is, moreover, a tube of air trapped between the forward face of the wave and the falling jet, and the trapped air that does not escape laterally along the tube breaks rapidly into a cloud of bubbles.

A similar process of air entrainment, but on a diminished scale, accompanies the initiation of a spilling breaker. The subsequent process of aeration is discussed by Longuet-Higgins and Turner (1974) in what appears to be the only dynamical model of a spilling breaker. Here the overrunning of a layer of air by the advancing plume of air and water incorporates more air into the whitecap, leading to a void fraction that, in Longuet-Higgins and Turner’s study, must exceed 8% for a steady flow to be sustained. (A further process of aeration through the surface of the plume may occur if it becomes vigorously turbulent as in spillways.)

Production of intense turbulence and vorticity accompanies the creation of bubbles. Conservation of vorticity in the water, so far as it is ensured by Kelvin’s circulation theorem, ceases when a multiply connected region, violating the conditions of the theorem, is created by the overturning wave, and the smaller bubbles are rapidly diffused by the rotational turbulent flow or carried by a local circulation generated by the breaking. Within the turbulent region there is evidence from the bubbles’ motion of downward extending vortices with near-vertical axes (Su et al. 1984), perhaps reflecting the along-crest variability often visible in breaking waves, and a consequential strong variability in vorticity.

This description of breaking and the generation of bubbles and turbulence has so far been largely qualitative. An important quantitative study of breaking waves has been made by Rapp and Melville (1990) who have created carefully reproducible, almost two-dimensional, breakers in a wave group in a laboratory fresh-water tank. (Important
differences are found in the size distribution of bubbles in fresh water and salt water. Coalescence occurs rapidly in fresh water leading to larger-sized bubbles that quickly rise to the surface (Scott 1975.) Their most significant results relate to the loss of excess momentum flux carried by the wave groups in their experiments. This ranges from 10% in spilling waves to up to 25% in plunging breakers. Given the, albeit, rough estimates of breaking frequency which are available (see above), Melville and Rapp (1985) conclude that essentially all the momentum flux from the wind may pass through the wave field before being transferred by breaking to the mean current in the water column. This alone places wave breaking in a pivotal position in the understanding of air–sea interaction. Although a study of bubble dispersion does not play a central role in Rapp and Melville's (1990) work, they did examine the effect of wave breaking on the dispersion of a patch of dye, which proved also to be indicative of how wave-generated turbulence spreads following breaking. Dye, and turbulence produced by the breaking waves, is mixed down to depths of about 0.1\* in two wave periods after breaking, where \* is the wavelength. Subsequent deepening, due to diffusion in the (spreading but decaying) turbulence left by the passing breaking wave, occurs at a rate proportional to \( t^{-0.25} \), where \( t \) is the time following the start of breaking, causing the dye to reach about 0.15\* at a time of ten wave periods after the onset of a spilling breaker, or 0.2\* after ten wave periods for a plunging breaker. The dye is also diffused horizontally to a total length of about 0.8\*—slightly larger values of length being associated with large plunging breakers, lower values with spilling waves. The rate of dissipation of turbulent kinetic energy falls as \( t^{-3.5} \) following breaking, more than 90% of the energy lost from the waves being dissipated in four wave periods.

Surface currents resulting from the momentum transferred to the flow by the breaking wave produce a circulation in a region about one wavelength long. Their decay is observed to be slow. Mean surface currents of 0.02c-0.03c, where c is the phase speed of the waves, are generated and take some 60 wave periods to decay to 0.005c. This decay time is comparable with the expected breaking repetition period, so that much of the drift in the upper layers of the ocean may be a direct consequence of momentum transfer by breaking waves. Loewen and Melville (1991) have further shown that the sound generated by breaking waves, scales with the amount of energy lost from the waves during breaking.

Phillips (1988) has devised a mechanistic argument relating wave spectra to turbulence that, in particular, offers a firm justification for relating the rate of dissipation of turbulent kinetic energy, \( \varepsilon \), to \( u_\ast^3 \), where \( u_\ast \) is the friction velocity in the air. This theory provides predictions for the frequency of occurrence of breaking and the length of a whitecap front which are yet to be confirmed by observation.

In summary we notice that although the importance of wave breaking is firmly established, its frequency and effects, particularly on turbulence or on bubble production in the ocean, are rather poorly quantified. We may anticipate that, provided bubbles can survive and be dispersed like dye, they may spread horizontally downwind to scales of about one wavelength and vertically to 0.2\* within some ten wave periods after generation.

To what extent can bubbles serve as tracers?

3. The Vertical Distribution of Bubbles in the Sea

A complete discussion of the properties and distributions of bubbles in the upper ocean is beyond the scope of this review*, but certain facts are pertinent. There have

* For more details of the properties, rise speed, dissolution rate, etc. of bubbles, see Thorpe (1982).
been few direct observations at sea of the size distributions of subsurface bubbles. Kolovayev (1976), Johnson and Cooke (1979), Mulhearn (1981) and Walsh and Mulhearn (1987) used photographic techniques, whilst Medwin (1970, 1977), McDaniel (1988) and Medwin and Breitz (1987) made acoustic measurements. Although the results differ considerably, none of the techniques being ideal, there is a general consensus that there is, on average, a rapid fall in the total number of bubbles as depth increases, the exponential decay-scale being 1–2 m, the number of bubbles increases rapidly with wind speed and, for bubbles of radius greater than about 50 μm, the number of bubbles of a given radius falls rapidly as that radius increases. This final fact is hardly surprising since the larger bubbles rise most rapidly to the surface. (A bubble of radius 50 μm rises at about 0.5 cm s⁻¹ whilst a 500 μm bubble rises at 7 cm s⁻¹.) Small bubbles dissolve rapidly, a 5 μm radius bubble in less than a minute. Bubbles smaller than 50 μm are very difficult to detect, and it is as yet uncertain at what radius the bubble-size distribution is 'peaked'.

Acoustic measurements with upward-pointing 250 kHz sonars moored to the sea bed (Thorpe 1982, 1986a) or with a 119 kHz sonar mounted on a submarine (Crawford and Farmer 1987) show that a stratus layer of bubbles persists at the surface when the wind speed, $W_{10}$, exceeds about 7 m s⁻¹, and the clouds below appear as columnar plumes. The mean cloud depth, $d$, increases with wind speed, reaching 10 m at winds of about 12 m s⁻¹. The ratio $d/λ$ increases from about 0.04 when $W_{10}/c = 0.6$ to about 0.2 when $W_{10}/c = 2$, where $λ$ and $c$ are the wavelength and phase speed of the dominant waves. Figure 1 shows the variation of cloud depth as measured by an automatically

![Figure 1](image.png)

**Figure 1.** The variation of bubble clouds measured at sea as a function of time in different wind speeds (see Thorpe 1986a). The logarithm of the acoustic scattering cross-section (in m⁻³) is contoured in increments of 0.5 from a threshold value of −4.22. In (a) $W_{10} = 6.9$ m s⁻¹, $W_{10}/c = 0.63$; (b) $W_{10} = 10.1$ m s⁻¹, $W_{10}/c = 0.86$; (c) $W_{10} = 13.3$ m s⁻¹, $W_{10}/c = 1.22$. The depth is relative to the surface level shown by the upper line. The depth is also shown as a function of wavelength, $λ$ (on the left), and of significant wave height, $H_s$ (on the right).
Figure 1. Continued.
recording sonar (Thorpe 1986a). For low values of \( W_{10}/c \) the deepest clouds reach to about 0.16\( \lambda \) or about five times \( H_s \), the significant wave height. Increase in \( W_{10}/c \) results in deeper-going clouds, their depths sometimes exceeding 0.25\( \lambda \), or 5.2\( H_s \), at \( W_{10}/c = 1.22 \). These values are consistent with those observed for spilling waves (more typical of low values of \( W_{10}/c \)) and plunging breakers (more common at high \( W_{10}/c \)) in Rapp and Melville's laboratory experiments. The mean down-wind length of bubble clouds, \( L \), measured at 10.3 m below the surface and made non-dimensional with \( \lambda \), also increases with \( W_{10}/c \); \( L/\lambda \sim 0.1 \) at \( W_{10}/c = 0.6 \) and \( L/\lambda \sim 1.2 \) at \( W_{10}/c = 2 \). The larger values of \( d/\lambda \) and \( L/\lambda \) at \( W_{10}/c = 2 \), when breaking is more prevalent and, in the range of wind speeds studied, bubbles frequently reached 10.3 m, are in accordance with the depth or length to which dye is mixed in Rapp and Melville's (1990) laboratory study. When breaking is weaker and less frequent (see section 4(d)) at low values of \( W_{10}/c \), bubble clouds are more intermittent and the mean depth of bubble clouds is therefore less.

It is, however, presumptuous to imply immediately a relationship between the distributions of turbulence, or dye, in the laboratory study and the bubbles observed by sonar in the ocean. The extent to which bubbles modify turbulence in the time-dependent evolution of a wave-produced bubble cloud is not known except in a general way in the highly aerated zone of spilling breakers (see section 2) where bubble volume (void) fractions are high. Close to the mean water level, Lamarre and Melville (1991) found void fractions that exceeded 0.2 at times up to half a wave period after breaking. These subsequently fall rapidly with an e-folding time of about 0.25 wave periods; large bubbles rising rapidly to the surface and perhaps, by vortex shedding, contributing to the turbulence. Beyond this near-surface zone the average void fractions appear to be small, typically \( 10^{-5} - 10^{-7} \) or less at depths exceeding 0.7 m and in winds of about 12 m s\(^{-1} \), as may be inferred from the measured mean size distributions. Although there is insufficient information about the more pertinent maximum void fractions and their subsequent development, it is likely that the dynamical effect of bubbles on turbulence in this zone is negligible.

The radius of a bubble that is resonant, near the sea surface, and therefore particularly effective in scattering sound at the frequency of the 250 kHz sonar, is 17 \( \mu \)m. This 'resonant' radius increases as the square root of pressure as depth increases. Such bubbles rise slowly (at about 1 mm s\(^{-1} \)) and may be effectively neutral both in their effect on the flow and in acting as tracers, like dye. Consider for example waves of 6 s period. These have a wavelength of 56 m and the rise of a 17 \( \mu \)m radius bubble in ten wave periods is about 6 cm. This is negligible in comparison with the depth, about 0.2\( \lambda \) = 11.2 m, to which a dye layer is spread in the same time according to Rapp and Melville's results. The effects of dissolution and of pressure in changing the radius of bubbles during the time of ten wave periods that accompanies their downward transport by the turbulence derived from the breaking waves are generally negligible (see Thorpe 1982, Fig. 18); dissolution is not significant at times less than 100 s, and will have little effect for wind-waves of period less than 10 s. Dissolution may be an important factor over the time-scale (5–10 minutes) for which bubble clouds persist (Thorpe and Hall 1983), but not during the relatively short period of the injection of turbulence by breaking waves. There are other concerns. Although the scattering cross-section of resonant bubbles is large, the total scattering from a cloud may not be dominated by bubbles of near-resonant size. Moreover, the ability of the sonar to 'see' bubbles depends upon the detection threshold of the instrument. Nevertheless, the small rise speeds of resonant
bubbles and their consequent tendency to follow fluid motions suggest (but do not prove) that the acoustically detected bubbles may be quasi-passive tracers. The point is germane to understanding the relation between the turbulent dispersion of solutes and of buoyant particles, a field where general rules are still lacking. The generally favourable agreement between the field observations and the laboratory studies indicates that the vertical diffusion of bubbles by turbulence from sources other than the breaking waves may be relatively small. This conjective deserves further study because, if true, it provides a powerful tool for studying breaking-wave-produced turbulence at sea.

Analytical and numerical models have been developed to describe the vertical distribution of bubbles and to infer information about the turbulence and gas transfer from a balance between turbulent diffusion and upward bubble rise (including a flux in radius space to account for dissolution and compression of the bubbles (Thorpe 1982, 1984a, 1986b)), and some aspects of the models have been tested in laboratory experiments (Bowyer, personal communication). It is possible from these models to reproduce the form of the bubble-size distributions that have been observed and to examine, for example, the sensitivity of these distributions to changes in water temperature, gas saturation levels and particle numbers (Thorpe et al. 1992). The models have also been used to predict the acoustic scattering cross-section of the bubbles and to compare its mean vertical distribution with observation (Thorpe 1986a). Turbulence has been represented in the models by a diffusivity of the form \( K_v = kw_\ast z \) appropriate to a constant stress layer, where \( k \) is von Kármán's constant, \( z \) is the depth and \( w_\ast \) is a friction velocity, to be determined by comparison between the model predictions and the observations at a depth of 0.1\( \lambda \). An independent estimate of the friction velocity in the water, \( w_\ast \), may be found by assuming that the wind stress, \( \rho_\ast C_D W_{10}^2 \), is transmitted through the surface to the water (Donelan 1978; Mitsuyasu 1985), where \( \rho_\ast \) is the density of air and \( C_D \) is a drag coefficient, about \( (1-3) \times 10^{-3} \) (see, for example, Large and Pond 1981), so that

\[
\rho u_\ast^2 = \rho_\ast C_D W_{10}^2
\]  

(1)

where \( \rho \) is the water density. Except when \( W_{10}/c \) is large, the ratio \( w_\ast / u_\ast \) is found to be greater than unity, reaching about 3 (error \( \pm 1 \)) at \( W_{10}/c = 0.6 \) and implying that the effective diffusivity of bubbles is higher than can be accounted for by the assumption of a constant stress layer and the conventional law of the wall. It appears that the mean bubble distribution (and recall that clouds persist for periods far exceeding the period of their generation and initial dispersion by the wave-produced turbulence) reflects a level of turbulence that is more intense than the expected background levels.

4. Turbulence in the upper ocean

I shall confine my discussion of turbulence to that part of the turbulent spectrum, and to processes with time and length scales, that are directly related to bubble clouds. Bubbles are produced abruptly, as we have seen, by breaking waves. The bubbles carried downwards to even a 1 m depth are very small, although clouds extend to many metres. They persist typically for periods of 5–10 minutes (Thorpe and Hall 1983). The relevant turbulent motion is that which is effective in producing diffusion on these scales, scales of time from perhaps a second up to an hour, and of space from the Kolmogorov scale, typically of the order of 1 cm, to the depth of the so-called mixed layer and of horizontal dimension at least of the order of the length of the largest waves.
(a) Similarities between the upper-ocean boundary layer and the atmospheric boundary layer

Many of the direct measurements of turbulence, or of associated fluid properties, in the near-surface ocean point to there being a general similarity to the atmospheric boundary layer.

Although Donelan (1990) has questioned whether the friction velocity is the only available length-scale in the water, pointing out that the horizontal velocity at the sea surface is effectively unconstrained unlike the atmospheric boundary layer over land, the velocity spectra appear to scale with the friction velocity in the water and the depth, z (Jones and Kenney 1977). In neutral conditions, measurements of fine-scale temperature variations in a lake were used by Dillon et al. (1981) to infer that $\varepsilon$ scales with $z^{-1}$ in a manner consistent with the law of the wall, $\varepsilon = u_\\ast^3/kz$. Oakey and Elliot (1982) found a vertically averaged rate of dissipation of turbulent kinetic energy, $\varepsilon$, that, although subject to considerable scatter, shows a scaling with $W_{30}^3$ (see also the reference to Phillips (1988) in section 2). Soloviev et al. (1988) used an upward-rising turbulence profiler in the Atlantic and found values of $ekz/u_\\ast^2$ close to, but slightly greater than, unity in the range of depths, $z$, from within a few centimetres of the surface to some 5 m in winds of 1.9 to 6.5 m s$^{-1}$ and waves of 3–4 s period and 0.5 to 1 m in height. Churchill and Csanady (1983) observed a roughly logarithmic variation of drift currents with depth, consistent with the existence of a constant stress layer. Logarithmic profiles were also found in laboratory experiments by Cheung and Street (1988) but, in the presence of waves, the gradients of the Eulerian mean flows were lower than predicted from the law of the wall (i.e. corresponding to larger-than-expected values of von Kármán's constant). The separation of wave-induced motions and turbulence is a non-trivial problem, and care has to be taken to assess the Stokes drift contribution to Lagrangian measurements. Toba (1988) also argues for a similar boundary-layer structure in the air above and in the water below waves, but expresses the need for care in comparing laboratory-generated and sea wind-waves. The parameter $W_{10}/c$ is suggested as an appropriate comparator. In conditions when the wave field was strongly affected by swell, the values of the friction velocity inferred by Churchill and Csanady were higher than could be accounted for by use of Eq. (1). This is another clue that wave-generated turbulence may be important near the surface.

In convective conditions, Shay and Gregg (1984, 1986) have drawn attention to the similarity in the scaling of $\varepsilon$ and of the effective buoyancy flux, $J_0$, due to heat leaving the ocean surface; $\varepsilon$ is uniform throughout much of the convective layer and its ratio to $J_0$ is close to that found in the atmosphere by Caughey and Palmer (1979; see, however, Moum et al. 1989). Relatively higher values of $\varepsilon/J_0$ were found as the surface was approached in the ocean, but those higher values of $\varepsilon$ are consistent with estimates derived from the buoyancy flux and the law of the wall (Lombardo and Gregg 1989). The atmospheric observations of Caughey and Palmer also show close agreement with Brubaker's (1987) measurements of the rate of dissipation of temperature variance in a lake. This has a power-law scaling which follows that of local free-convective similarity.

There is also evidence of some structural similarity between the atmospheric and oceanic boundary layers as measured by the presence and tilt of temperature ramps and by the magnitude and sign of the skewness of the horizontal temperature gradient. The skewness is mainly caused by the presence of abrupt, vertically coherent, temperature changes; the temperature ramps are apparently generated by the straining of the mean temperature field by eddies with nearly horizontal axes aligned roughly across wind in the direction of the vorticity of the mean flow. The skewness changes sign when the heat flux is reversed (Antonia et al. 1979; Thorpe 1985; Soloviev 1990; Thorpe et al. 1991).
The temperature ramps in the oceanic boundary layer have an associated ramp-like structure in the acoustic scattering from bubbles (Thorpe and Hall 1987), pointing again to the interrelationship between the bubbles and the dynamics of the layer.

(b) Near-surface turbulence

Having now sought and found some similarity with the atmospheric boundary layer, I return to consider the region very close to the surface. Evidence from Rapp and Melville (1990), as well as from the observations of bubbles, suggests that turbulence derived directly from the breaking waves may be limited to a relatively narrow zone within a distance of about 0.2λ from the sea surface. I shall henceforth disregard the effects of buoyancy flux through the surface. My remarks therefore apply at depths much less than the magnitude of the Monin–Obukov length scale.

Measurements made from a fixed tower in Lake Ontario (Kitaigorodski et al. 1983) suggest that the effect of breaking waves dominates and enhances the turbulence to a depth of about ten times the root-mean-square wave amplitude, which generally lies between 0.04 and 0.16 times λ (see discussion in Thorpe 1986a), consistent with the mean depth of the bubble-cloud distribution (section 3, see also Fig. 1). Below this depth the law of the wall appears to apply. Gregg (1987) and Anis and Moun (1992) using vertical profiling probes have also presented evidence of exceptionally high values of ε near the surface in strongly convective conditions, values higher than can be explained by shear stress and buoyancy.

An experiment has been made by a group led by Dr T. Osborn to study breaking waves, turbulence and bubbles in this near-surface region from the US submarine Dolphin. Measurements made using turbulence probes and sonars mounted on the submarine operating close to the surface in mean winds of 5–9 m s⁻¹ and waves of 4 s period (0.86 < W₁₀/c < 1.4) and weakly convective conditions, show a distinct correspondence between enhanced values of turbulent dissipation and acoustically detected bubble clouds. Values of the mean dissipation exceeded the values expected from the law of the wall and buoyancy flux by up to ten times, anomalously high dissipation being in water that contained bubbles and the largest values being close to the surface. The dissipation rate, ε, near the surface has a horizontal patchiness associated with the horizontal variability of the bubble clouds. Below the depth of about 5 m (0.2λ) to which clouds extend there was no apparent correspondence between ε and the position of the overlying clouds. At 10 m depth, still within the apparently ‘mixed layer’ that extended to 60 m below the surface, the turbulence was still ‘patchy’, but patches were typically some 100 m apart horizontally, a far greater separation than the bubble clouds. Regrettably, owing to a major failure of the submarine’s generators, few data were collected. Evidence was, however, found of another major process of turbulent dispersion in the upper ocean, Langmuir circulation.

(c) Langmuir circulation

Much effort has been given to explaining the cause of the linear, wind-aligned bands of convergence and divergence on the sea surface, and the underlying circulation, with downward-going flow beneath convergence and upwelling below divergence. This pattern was described by Langmuir (1938) as one of ‘helical vortices’, but more accurately, perhaps, is one of parallel vortices of alternating sign and with axes aligned down-wind, or of helical fluid-particle paths constrained within the pattern of the circulation cells. Knowledge of the circulation has been comprehensively reviewed by Leibovich (1983). Theory and laboratory experiments point to a combination of waves and drift current as being essential ingredients, at least in providing a vortex force to initiate the longitudinal
vortices. Given, however, that Langmuir's paper reporting persuasive, if not convincing, evidence for his claim that circulation patterns 'apparently contribute the essential mechanism by which the epilimnion' (the near-surface 'mixed' layer) is produced' was published as long ago as 1938, and recognizing the vivid visual clues presented in the form of 'wind-rows' of floating organisms, debris or foam on the surface of the sea and lakes, it is astonishing that the importance, or otherwise, of the circulation as a process of turbulent dispersion and transfer in the upper ocean has not yet been fully established.

The most extensive and thorough measurements at sea are those of Weller and Price (1988) from the floating instrument platform, FLIP. They found downward vertical and downwind horizontal velocity components sometimes exceeding 20 cm s⁻¹ below the surface convergence and above mid-depth in the mixed layer. The speed was not simply proportional to wind speed. It appears that Langmuir circulation can be an important direct mixing mechanism in the upper one third to one half of the 'mixed' layer, rapidly mixing away shallow near-surface stratification associated with diurnal heating. There was, however, no evidence that the circulation plays a direct role in mixing near the base of the 40–60 m deep mixed layer present during the experiment. It was, unfortunately, not possible to calculate the Reynolds stress associated with the circulation accurately, but the results support the early conclusions of Gordon (1970) and Pollard (1977) that the stress may sometimes be an order of magnitude larger than that produced by the wind stress. An appreciable component of the vertical heat flux may also be carried by the circulation (Thorpe 1984b). The mean shears created by the vertical and horizontal variations in the mean current probably contribute to the shear-stress maintenance of turbulence, but this is yet to be quantified.

Whilst there may still remain uncertainty about the role of Langmuir circulation in transporting heat or momentum, there can be little doubt as to its effect in ordering the dispersion of floating and buoyant objects or bubbles. With the exception of the sargassum weed which Langmuir noted as being drawn into long lines down-wind, there are rarely markers in the sea or, perhaps more pertinent, suitable vantage points, to make clear the wind-row patterns so often visible on the surface of lakes. The submarine observations using side-scan sonar have, however, demonstrated that linear subsurface bands of bubbles aligned in the wind direction can be detected even in conditions in which there is no evidence of wind-rows visible on the surface. These bubble bands have been reported before (Thorpe and Hall 1983; Thorpe et al. 1985; Zedel and Farmer 1991). They are caused by bubbles from breaking waves being carried by the convergent flow into the downwelling regions of the Langmuir circulation where, in rising against the downward flow, bubbles are partly trapped and where they accumulate and sustain the linear subsurface bands so long as the adjacent convergence pattern is maintained, typically 20 min or more (Thorpe 1984b). Although the submarine data are insufficient to support robust statistical tests, no significant difference was apparent between vertical extent of bubble clouds in wind-rows and those left by recently breaking waves. It was noticed in section 3 that the vertical extent of bubble clouds observed at sea is consistent with that observed in the quiescent conditions of a laboratory. The 'background turbulence', including Langmuir circulation, appears to play a relatively minor role in the vertical diffusion of bubbles, but a major role in their horizontal dispersion.

(d) Recent observations using sonar

Sonar measurements in a freshwater lake, Loch Ness, by Mr M. Curé (private communication) have vividly revealed the formation of the bands of bubbles and their
disintegration, and have shown that they are detectable whenever the wind speed is sufficient to cause breaking waves and bubbles ($W_{10} > 2.5 \text{ m s}^{-1}$). As an example of the information that is now available, Fig. 2 shows the appearance of bands detected by a 90 kHz sonar located on the side of the loch at a depth of 35 m and pointing directly across the loch. The loch (1 km wide), orientated south-west to north-east, has a mean depth of 150 m and a length of 34 km. The fetch at the site in south-westerly winds is 20 km. The acoustic returns have been digitized, the effect of slant range has been corrected and the images processed to reduce variation in signal level caused by attenuation. Figure 2 shows the evolution of structure during a period of south-westerly winds. The pattern migrates away from the sonar to the south-east, to the right of the down-wind direction. This movement is typical, and has been previously noticed in wind-rows in lakes (Thorpe and Hall 1982). Amalgamation of the bands is a feature commonly noticed in the sonographs. This process leads to the lateral (or cross-wind) dispersal of buoyant material which might, except for small-scale, inter-cell turbulent exchange, be trapped indefinitely within the circulation of a single cell (Csanyi 1973; Faller and Auer 1987; Thorpe 1992). Figure 2 also shows that there is a hierarchy of bubble bands, with smaller scales converging into the larger.

Higher-frequency sonars, having greater resolution, though poorer range, can detect breaking waves (Thorpe and Hall 1983). Figure 3 is a sonograph from a 250 kHz sonar pointing in the same direction as the 90 kHz, showing occasional sharp-fronted bands aligned roughly in the range direction that are associated with breaking wave crests and clearly initiate sonar echoes from ‘plumes’ of bubbles. These remain visible for a short time dependent on the down-wind length and lifetime of the bubble plumes and the mean current through the sonar beam. Dr M. White (private communication) has identified the breaking waves by applying a threshold criterion to the time derivative of the signal at constant range (see Fig. 4). As in other methods of identifying breaking waves (see section 2) the criterion is somewhat subjective, but this technique can locate and determine the number and scale of breaking waves that produce detectable bubble clouds (a potentially quantifiable measure of ‘breaking’), and suggests the use of a fully calibrated sonar to quantify the variation of the acoustic target strength of bubble clouds produced by waves and to study its relation to ambient sound generation.

Figure 4 and similar sets of data in different wind speeds have been used to determine the frequency of wave breaking at a fixed position as a function of wind speed, as shown in Fig. 5(a). Also shown in Fig. 5 are data obtained by Longuet-Higgins and Smith (1983) in the North Sea where the fetch was about 100 km, and by Weissman et al. (1984) in a lake with fetch of 7 km, referred to in section 2. There is considerable scatter. This is greatly reduced when the number of breaking waves per wave is plotted as a function of the parameter $W_{10}/c$, as shown in Fig. 5(b), in spite of the different techniques and criteria used to detect breaking. The proportion of waves which break at a fixed position increases with $W_{10}/c$. We have also examined the location of breaking relative to the bubble bands. Figure 6 shows that although there is some evidence that slightly more frequent breaking occurs near, or in, the bands, this is not a significantly greater frequency than elsewhere. The observations support a hypothesis that it is primarily a redistribution of randomly injected bubbles by the Langmuir circulation that leads to the formation of the bubble bands, rather than locally enhanced breaking. The time taken for bubbles to be carried into the bands is typically 2–3 min, less than the lifetime of the bands themselves (15–30 min). The sonograph, Fig. 2, thus represents the space–time variation of convergence and divergence on the water surface.
Figure 2. Scattering from subsurface bubble clouds detected from below using a 90 kHz side-scan sonar directed across-wind at an acute angle to the surface. The target strength is not calibrated but the colour-scale, dark brown to white, shows the relative intensity of scattering. Bubble clouds (dark) lie in bands some 15 m apart which have been intensified in the image and which, as time advances, migrate away from the sonar. The wind speed, $W_{10} = 15 \pm 2$ m s$^{-1}$. Viewed at a low angle from the left it can be seen that smaller bands migrate to join the larger and that some of the large-scale bands amalgamate (e.g. at about 20 min, 100 m range).

Figure 3. Sonograph using a 250 kHz side-scan sonar aligned as for the 90 kHz sonar in Fig. 2; range (m) versus time (min). The near-vertical lines are due to sound scattered from breaking waves. The crest lines are slightly inclined to the sonar direction. As the waves advance across the beam the position of the breaking-wave crests consequently approach the sonar at a speed much greater than the wave phase speed. The 'trails' to the right of these lines are bubble clouds left by the breaking waves. $W_{10} = 6$ m s$^{-1}$, $W_{10}/c = 1.8$. 
Figure 4. An image showing the breaking-wave crests derived from Fig. 3 by identifying high time-derivatives of the sonar signal strength at fixed range: range (m) versus time (min). The criterion for selection is subjective, but is consistent with targets that lead to identifiable bubble clouds.

Figure 5. (a) The variation of the number of breaking waves as a function of wind speed, $W_{10}$. The triangle shows a measurement by Longuet-Higgins and Smith (1983) and the circle one by Weissman et al. (1984). (b) The same data, but now plotted to show the proportion of breaking waves as a function of $W_{10}/c$, both on log-scales.
Figure 6. The frequency of wave breaking as a function of position from one bubble band (at 0) to its neighbour (at 10). $W_{d0} = 11.5 \text{ m s}^{-1}$. Measurements were made over a period of 48 min with 3100 breaking waves included in the analysis. The mean band separation was about 20 m. There is little evidence to support the hypothesis that the bands are strongly supported by bubble input from waves breaking predominantly in their vicinity.

5. Bubbles and gas flux

The surface area of bubbles below the sea surface per unit area is, on average, very small, so they do not add appreciably to the total area of contact between the atmosphere and the ocean. Nevertheless, at sufficiently high wind speeds, bubbles may contribute significantly to the net air-sea transfer of some gases, notably those with low solubility in sea water. The pressure within bubbles is the sum of the hydrostatic pressure, which at a depth of 10 m is approximately twice that at the sea surface, and an additional pressure due to surface tension. The latter is inversely proportional to the radius of the bubble and is equal to the atmospheric pressure at the sea surface for bubbles of radii about 1 μm. The additional pressure to which gases in bubbles are subjected below the surface will help to promote supersaturated conditions in the water. Gases are effectively pumped into the water via bubbles and a supersaturated equilibrium is maintained in the water by the gases being diffused back into the atmosphere via evasion through the sea surface.

Studies of bubble-mediated gas transfer have been made by Thorpe (1982, 1984c), Merlivat and Memery (1983), and Memery and Merlivat (1985). Supersaturations of 1–2% in nitrogen and of about 0.5–1% in oxygen and argon (all gases with low solubilities in sea water) will be supported by the flux associated with bubbles at wind speeds of 10 m s$^{-1}$. Carbon dioxide is relatively soluble. Bubbles appear to play only a minor role in its transport and will not contribute measurably to its supersaturation (Woolf and Thorpe 1991). Accurate prediction using numerical models is hindered by the lack of knowledge of the precise form of the bubble-size distribution (see section 3) and its variation with environmental factors such as wind speed, waves, gas saturation levels and particulates present in sea water (see Thorpe et al. 1992), of the precise values of gas evasion rates and their variation with wind speed (Liss and Merlivat 1986; Watson et al. 1991), and also by the uncertainty in the appropriate representation of upper-ocean dynamics and its relationship to bubbles (section 4). The effect of Langmuir circulation
on the air–sea flux of low-solubility gases may be appreciable (Thorpe 1984b). In view of the results presented in section 3, and Fig. 5, it appears timely to look more closely at the sensitivity of low-solubility-gas transfer rates to variations in wave conditions.

6. DISCUSSION

Much of the experimental evidence that I have assembled in this review appears contradictory, particularly that which describes the nature of turbulence in the upper layers of the ocean within distances less than about 0.2A and much less than the magnitude of the Monin–Obukov length scale from the surface. There is, for example, on the one hand evidence that the rate of dissipation of turbulent kinetic energy near the surface follows the law of the wall (Dillon et al. 1981; Soloviev et al. 1988). On the other hand there is evidence of much higher values within distances of a few wave heights from the surface (Kitaigorodski et al. 1983; Gregg 1987; Anis and Moun 1992). Measurements are difficult to obtain and, at best, fragmentary. Oceanic turbulence is generally highly intermittent, requiring extensive data sets to establish reliable estimates of the mean values. Many vertical profiles are needed to obtain an accurate average of the near-surface wave-induced turbulence; turbulent energy decays rapidly following breaking, most of the energy being lost within four wave periods (Rapp and Melville 1990; section 2), and the frequency of wave breaking is low (section 4(d)).

More information is needed about the effect of transient conditions, especially variable winds. Some progress in the study of the development of wind–wave spectra in changing winds has been made by Toba et al. (1988). The time-scale of adjustment of the equilibrium range wave spectrum following a change in wind is remarkably small (of the order of 10 min) in comparison with the time needed to develop the equilibrium spectrum. It is, however, comparable with the time-scale for the adjustment of windrows in changing winds (1–30 min, Leibovich 1983) although it is hard to see how deeply-penetrating circulation patterns could be established so rapidly. It is particularly important to establish the stress carried by the circulation. It is generally assumed in estimating \( \varepsilon kz/\nu^3 \) that the stress in the water is given by Eq. 1, or that the stress at the sea surface is continuous. Melville and Rapp’s (1985) experiments indicate the importance of the waves in momentum transport. Gordon’s (1970) and Pollard’s (1977) conclusions (section 4(c)) suggest that there may often be occasions when a high proportion of the flux of the horizontal momentum from air to water is passing into or out of the waves, when the wave field is gaining or losing a significant amount of momentum.

If, as Melville and Rapp (1985) argue, wave breaking is important in the transfer of momentum from the air to the sea, then the turbulence near the surface is likely to be dependent on the waves as well as the wind. A simple explanation of the variety of near-surface dissipation profiles referred to at the beginning of this section (but one not entirely consistent with observations) appears to be that there is a dependence on a parameter linked closely to \( W_{10}/c \) (‘linked’, because it might be appropriate to use the friction velocity in the air rather than \( W_{10} \), and the definition of \( c \) is less than precise). At large values of \( W_{10}/c \) (low wave age) when breaking occurs frequently (Fig. 5(b)) and is more vigorous, driving relatively deeper bubble clouds (section 3), the wind stress is very effectively transferred to the water and, although a little momentum is being transferred to the wave field, a friction velocity in the water may be accurately derived from Eq. 1. However, at low values of \( W_{10}/c \), breaking is much less intense, so that the linkage between the air and the water via the waves is weaker. The rate of loss of momentum from the wave field, partly used in driving and sustaining Langmuir circulation, contributes significantly to the effective friction velocity and hence dissipation.
rates in the sea, which are greater than those derived from the wind stress above. The effective friction velocity in the water will exceed that derived from Eq. 1.

Soloviev et al.'s (1988) experiments at values of $W_{10}/c < 1$ appear to contradict this simple explanation. Perhaps other effects, including buoyancy flux, need careful attention. Whilst Rapp and Melville's valuable studies of breaking waves show rapid dissipation to occur, they exclude the effects of winds and wind-generated currents. Few of the near-surface bubble clouds found in the submarine's observations to contain the high-dissipation rates (section 4(c)) could have had an age of less than the four wave periods in which turbulence is observed to be lost in the laboratory experiments. This suggests that turbulence might be supported and sustained for longer periods by shear-stress production when waves break in a shear flow (e.g. a wind-drift current or flow due to Langmuir circulation). It is also possible that swell may play a role. Dr J. Moum (private communication) tells me that he is examining this question, having obtained several data sets which show levels of $\varepsilon$ that are enhanced above the law of the wall, but also several which do not, one being in a fjord where there was no swell. The equipment presently available to measure turbulence and detect the generation and presence of bubble clouds and Langmuir cells provides the essential technology needed to mount a definitive study of dispersion, diffusion and turbulence in the upper ocean.

It is a sign of our lack of knowledge that there is generally no specific representation of the effects of surface waves or Langmuir circulation in numerical models that describe the diffusion of heat and momentum or the spread of pollutants in the upper-ocean boundary layer. We are beginning to learn how the state of the wave field may affect the wind-drag coefficient, $C_D$, which appears in Eq. 1 (Donelan 1990). It is time to unravel the complex relationship of waves and turbulence and of the subsurface bubbles that demonstrate so vividly the short time-scale connections between the atmosphere and the ocean and which may have important long-term consequences for heat and momentum exchange.

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