velocity profile usually exists in the lower few percent of the Ekman height and the slope of this profile can be used to infer the bottom stress. It has only recently become feasible to make detailed vertical velocity profiles and these reveal two logarithmic regions. The inner layer is controlled by the very local characteristics of the bottom and its slope gives the stress experienced by particulates on the bottom. The outer layer reveals the large addition of form drag due to long horizontal-scale bottom features and this drag provides the boundary condition for the flow well above the bottom. That is, circulation models should use a drag coefficient consistent with the friction velocity derived from the outer layer. The outer layer may be important to sedimentation after the onset of suspension.

See also

Ekman Transport and Pumping.

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TURBULENCE SENSORS

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Introduction

This article describes sensors and techniques used to measure turbulent kinetic energy dissipation in the ocean. Dissipation may be thought of simply as the rate at which turbulent mechanical energy is converted into heat by viscous friction at small scales. This is a complicated indirect measurement requiring mathematical models to allow us to envisage and understand turbulent fields. It will require using this theory to understand how sensors might be developed using basic principles of physics to measure properties of a turbulent field to centimeter scales. Instruments must be used to carry these sensors into the ocean so that the researcher can measure its turbulent characteristics in space and time. It is also this sensor–instrument combination that converts the sensor output into a quantity, normally a voltage varying in time, that is used by the experimenter to calculate turbulent intensity. Thus, both the characteristics of sensors and the way in which the sensor–instrument combination samples the environment must be understood and will be discussed below.

Understanding Turbulence in the Ocean

There is no universally accepted definition of turbulence. Suppose that one stirs a bowl of clear water and injects some colored dye into it. One sees that filaments of dye become stretched, twisted and contorted into smaller and smaller eddies and eventually the bowl becomes a uniform color. This experiment leads to one definition of turbulence. It includes the concept that eddies in the water are distributed randomly everywhere in space and time, that energy is transferred from larger to smaller eddies, and that over time the mean separation of the dyed particles increases. In contrast, the ocean is typically stratified through a density that is determined by the temperature and salt in the water as well as the pressure. In this environment, a vertical shear in the velocity in the water column can be large enough to overcome the stability. Energy from the mean flow is converted into large-scale eddies determined by flow boundary conditions that characterize turbulent kinetic energy at its maximum scales. Further vortex stretching creates smaller and smaller eddies resulting in a turbulent cascade of energy (velocity fluctuations) to smaller scales until viscous forces begin to dominate where the energy is eventually dissipated as heat. This article focuses on sensors to measure this dissipation process directly by measuring the effect of viscosity on the turbulent cascade.

The irregular and aperiodic velocity fluctuations in space and time characteristic of turbulence, accompanied by energy transfer between scales and associated fluid mixing, may be described mathematically through nonlinear terms in the Navier–Stokes equation. Nevertheless, it is difficult to solve numerically in oceanographic applications. At the dissipation scales, typically a few meters and smaller, we normally assume that the turbulent field is homogenous and that it has definable statistical averages in all parts of the field. We further assume that direction is unimportant (isotropy) and statistical distributions depend only on separation distances between points. With the turbulence controlled only by internal parameters, we assume the nature of the nonlinear cascade of energy from large to small scales generates a universal velocity spectrum. An example of this spectrum is shown schematically in **Figure 1A**. At low wavenumbers, k, no energy is taken out by viscous dissipation, so the energy flux, ε, across each wave number, or down the cascade, is constant. Through dimensional arguments, the three-dimensional turbulent energy spectrum, $E(k)$ in this region (called the inertial subrange) as a function of wave-number k is given by

$$E(k) = \alpha\varepsilon^{2/3}k^{-5/3} \qquad [1]$$

where α is a constant determined experimentally to be approximately 1.5. In practice the three-dimensional spectrum given in eqn [1] cannot easily be measured and one must use the one-dimensional analogy where k is replaced by a component k_i.

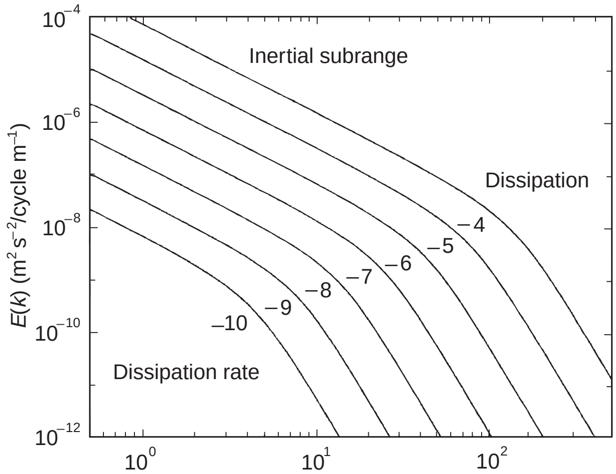
At higher wavenumbers or smaller scales the velocity gradient spectrum (obtained by multiplying the spectrum in eqn [1] by the square of the wave number k^2) shows more clearly where dissipation occurs. **Figure 1B** shows the spectra of velocity shear for velocity fluctuations for one component of k for values of ε most typically found in the ocean. In this case, the spectra of fluctuations transverse to the measurement direction are shown but the picture for along-axis fluctuations would look almost identical. At the highest wavenumbers (smallest scales), viscous dissipation reduces the energy per unit wavenumber to zero. At small scale, it is assumed that turbulent motion is determined only by kinematic viscosity, $v(=1.3\times10^{-6}\,\mathrm{m^2\,s^{-1}}$ at 10°C), and the rate, ε, at which energy passed down from larger eddies, must be dissipated. By dimensional arguments the length scale at which viscous forces equal inertial forces, and viscosity dissipates the turbulent energy as heat, is given by viscous cutoff scale

$$L_v = 2\pi(v^3/\varepsilon)^{1/4} \qquad [2]$$

The factor 2π gives a length scale from the radian wave number. This is an important scale for the design of instruments and sensors because it defines the smallest diameter eddies that must be measured.

The dissipation ε is given by integrating the spectrum shown in **Figure 1B**.

$$\varepsilon = 15v\int_0^\infty k_1^2E_1(k_1)\mathrm{d}k_1 = 7.5v\int_0^\infty k_1^2E_2(k_1)\mathrm{d}k_1 \qquad [3]$$



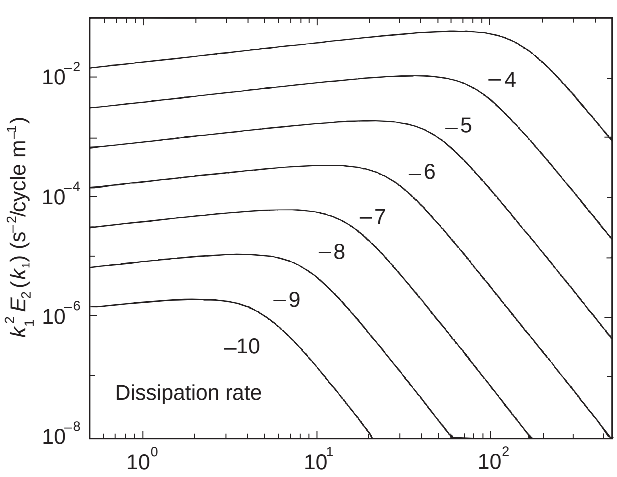


Figure 1 (A) The universal, velocity spectra for dissipation rates that typically occur in the ocean. Power density in velocity is plotted as a function of wavenumber. The shape of the spectrum remains the same but, as the energy in the turbulent field increases, the spectrum moves to higher wavenumbers and to higher intensities. (B) The equivalent universal, velocity shear spectra. (A) and (B) both show the inertial subrange and dissipation region, but in (B), the dissipation portion is more strongly emphasized.

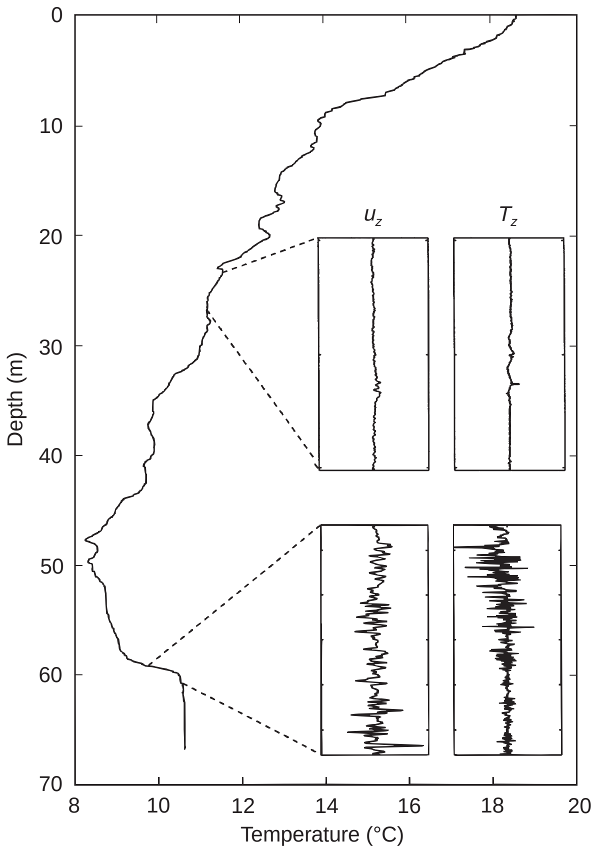


Figure 2 A representative vertical profile of temperature is shown from the surface to bottom obtained with a vertical falling instrument. In panels at the right are shown expanded portions of the velocity shear (u_z) and the gradients in temperature (T_z). The panels represent small sections of the vertical record that are treated as time-series to calculate spectra similar to those in **Figure 1(B)**. The upper panel of u_z at mid-depth is a region of low dissipation and the one below represents higher dissipation.

$E_1(k_1)$ is the one-dimensional wavenumber spectrum of longitudinal velocity, and $E_2(k_1)$ is the one-dimensional spectrum of transverse velocity and one assumes isotropy to estimate the factors 15 and 7.5, respectively. In practice, the upper integration limit may be replaced with the viscous cutoff scale. For the transverse turbulent velocity u, the shear variance in the z direction, $\overline{(\mathrm{d}u/\mathrm{d}z)^2}$ is equivalent to the integral of equation [3] and ε is given by

$$\varepsilon = \frac{15}{2}\nu\overline{\left(\frac{\mathrm{d}u}{\mathrm{d}z}\right)^2} \qquad [4]$$

These assumptions are important to the way in which sensors are designed. A common way to observe turbulent fields is by making measurements of velocity and other mixing quantities along a trajectory through a turbulent field assuming that it is frozen in space and time. Measurement along a line, recorded as a time-series (**Figure 2**), is interpreted as spatial variability by assuming stationarity and using the known sensor velocity to convert into distance. Standard Fourier transform techniques allow one to generate spectra similar to those in **Figure 1** from which dissipation, ε, may be estimated.

If there is a temperature gradient in the water column when turbulence is generated, the velocity field strains the temperature field, creating strongly interleaved temperature filaments over the vertical range of the overturn. The temperature microstructure intensity depends not only on the mean gradient but also on the energy in the turbulent field, in particular dissipation, ε. Temperature fluctuations recorded as a time-series (**Figure 2**) can be represented by spectra similar to those shown in **Figure 1**. As with the velocity fluctuations, there is a subrange where diffusive and viscous effects are unimportant where temperature fluctuations are transferred towards higher wave numbers. Temperature spectra persist to length scales smaller than the viscous cutoff scale. In this range, not only kinematic viscosity, ν, and dissipation, ε, are important but also, thermal diffusion, $\kappa_T(\approx 1.4\times 10^{-7}\,\mathrm{m^s\,s^{-1}})$. The cutoff wavelength for temperature fluctuations is given by

$$L_T = 2\pi(\nu k_T^2/\varepsilon)^{1/4} \qquad [5]$$

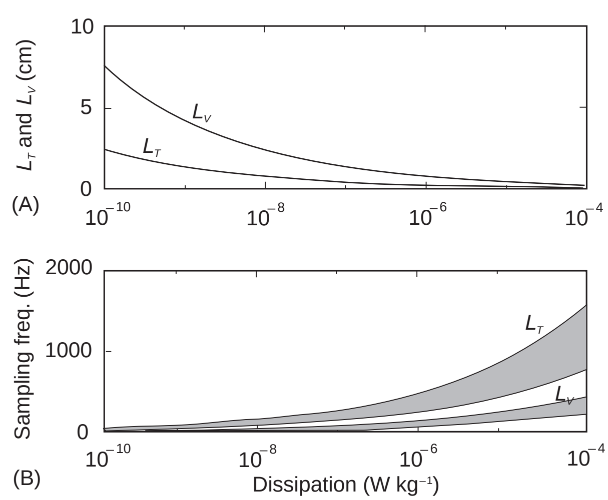


Figure 3 (A) The decrease in the cutoff scale with increasing dissipation for viscous dissipation, L_ν (eqn [2]) and thermal dissipation, L_T (eqn [5]). (B) The sampling frequency that is required for a particular ε for viscous dissipation, L_ν and for thermal cutoff, L_T. The upper and lower boundaries of the shaded bands correspond to measurement at flow speeds of 1.0 and 0.5 m s^{-1}, respectively.

Under restricted circumstances, the temperature gradients or temperature microstructure can be measured in the ocean to this scale. Under these circumstances one can determine L_T and hence estimate dissipation ε.

The sensors used most commonly in oceanography to measure dissipation make use of the above ideas. Velocity fluctuations may be used to determine dissipation ε directly using eqns [3] or [4]. Measuring temperature fluctuations allows dissipation to be calculated indirectly from eqn [5]. The units used to express dissipation in the ocean are $\mathrm{W\,kg^{-1}}$ (watts of mechanical energy converted into heat per kilogram of sea water). Typical values range from $10^{-10}\,\mathrm{W\,kg^{-1}}$ in the deep ocean to $10^{-4}\,\mathrm{W\,kg^{-1}}$ in active boundary layers on continental shelves. (To put these numbers into a simple perspective, energy dissipated in the ocean may range from the almost insignificant rate of $100\,\mathrm{W\,km^{-3}}$ to the very large rate of $100\,\mathrm{MW\,km^{-3}}$.) Present sensors and instruments are capable of measuring over this range of dissipation. Regions of higher dissipation such as river outflows and tidal channels are not normally measurable with sensors and instruments described here.

Measuring Dissipation in the Ocean

The most common technique of estimating dissipation in the ocean involves measuring small-scale velocity and temperature fluctuations. This may be accomplished by dropping a profiler vertically, towing one horizontally or setting it at a fixed position and measuring the fluctuations in velocity and temperature as the water moves past the sensors. This allows a time-series of turbulent velocity fluctuations to be recorded. A typical platform used to measure dissipation in the ocean is a vertical profiler that falls typically at a speed of 0.5–$1.0\,\mathrm{m\,s^{-1}}$. There have been many such instruments built and each one typically carries a number of sensors to measure some components of the turbulent velocity as well as temperature microstructure. A sample time-series for a vertical profiler is shown in **Figure 2**. Assuming that the turbulent field is isotropic, homogeneous and stationary one can use the mean flow velocity to determine the wavenumber scale and calculate the one-dimensional turbulence spectrum, $E_1(k_1)$ or $E_2(k_1)$, as defined above and from this determine the dissipation, ε, using eqns [3] and [4].

As the turbulent dissipation gets larger, the wavenumber at cut-off gets larger. Alternatively, L_v and L_T become smaller as shown in **Figure 3(A)**. As one tries to measure higher dissipation one must have a sensor with better spatial resolution and higher frequency response. We convert from wavenumber, k (cycles $\mathrm{m^{-1}}$), to frequency using the relationship $f = kV$ (Hz) where V ($\mathrm{m\,s^{-1}}$) is the flow speed past the sensor. The cutoff frequencies corresponding to L_v and L_T are given by $f_{ci} = V/L_i$. In practice, one does not have to measure the microstructure variance to the cutoff frequency because of the universal characteristic of the dissipation curves. A usual compromise is to consider that if 90% of the dissipation curve is measured then a satisfactory measure of dissipation can be achieved. This is summarized in **Figure 3B** which shows the sampling frequency that must be achieved to resolve a particular dissipation. (It must be remembered that to resolve the energy at any frequency one must sample at least twice that frequency.)

Turbulence Dissipation Sensors

Airfoil Probes

One of the most commonly used turbulence sensors to measure turbulent velocity fluctuations is called an airfoil probe. This sensor is an axially symmetrical airfoil made of flexible rubber surrounding a sensitive piezoelectric crystal. The sensitive tip of the probe approximates a parabola of revolution, several millimeters in diameter and about 1 cm long. The crystal generates a voltage proportional to the magnitude of a force applied perpendicular to its axis. The crystal is rigid in one transverse direction so responds to a cross force only in one direction. Thus, two sensors are required to measure the two transverse components of turbulent velocity fluctuations. The sensor is placed on the leading end of an instrument that is moving relative to the water at a mean speed V. In a mean flow along the axis of the shear probe, no lift will be generated and no force applied to the crystal. If there is an off-axis turbulent velocity, a lift will be generated which will apply a force to the piezoelectric crystal through the flexible rubber tip. Thus, the sensor will provide a voltage that is linearly proportional to the turbulent velocity. The effective resolution of the sensor is of order 1 cm, the smallest scale of turbulence that can be effectively measured by this sensor. From **Figure 3**, it can be seen that for values above $10^{-5}\,\mathrm{W\,kg^{-1}}$ this type of sensor will begin to underestimate dissipation. Normally the signal from the sensor is differentiated to emphasize the high frequency part of the turbulence spectrum. This gives the velocity shear, and analysis of this signal allows direct generation of spectra similar to the theoretical ones shown in **Figure 1B**. For this reason, airfoil probes are often called shear probes. These sensors measure the component of turbulence perpendicular to the drop direction of the instrument. As such, it is eqn [4] which is most relevant to calculating dissipation, ε. **Figure 3B** shows that to measure dissipation to $10^{-5}\,\mathrm{W\,kg^{-1}}$ in a flow speed of $1\,\mathrm{m\,s^{-1}}$ (along the axis of the sensor) the output must be sampled to at least as rapidly as 200 Hz.

Of the many instruments that use this sensor to measure dissipation, the most common are vertical profilers. Those used near the surface are often called tethered free-fall profilers because they have a light, loose line attached to the instrument for quick recovery and redeployment. The line is usually a data link to the ship where data are recorded on computers for analysis. Because of the intermittent nature of turbulence, it is important to have many profiles (or independent samples) in measuring dissipation to be able to obtain a statistically robust average value. For deeper measurements of dissipation, free-fall profilers are used that have no tether line. They are deployed to a predetermined depth in the ocean where their buoyancy is changed to allow them to return to the surface. These instruments record internally and can be inherently quieter than tethered free-fall instruments but are slower to recover and redeploy. In practice, both types of profilers can measure dissipation as low as $10^{-10}\,\mathrm{W\,kg^{-1}}$. In shallow regions of high dissipation such as the bottom boundary layer of tidally generated flow over banks, in bottom river channels or in active regions such as the Mediterranean outflow tethered free-fall instruments have been most successful. Where the dissipation exceeds $10^{-5}\,\mathrm{W\,kg^{-1}}$, these profilers and the shear probe sensor give limited results.

The airfoil probe has also been used successfully to obtain dissipation measurements horizontally. It has been used as a sensor on a towed fish pulled horizontally at speeds of order $1\,\mathrm{m\,s^{-1}}$. The results look similar to those in **Figure 2** where the depth axis is replaced by a horizontal axis and similar techniques to those described above are used to extract dissipation. Because of towline vibration, a towed instrument is generally noisier than a free-fall profiler. If the vibration noise of the platform is transferred to the airfoil sensor, it will generate velocity signals relative to the sensor indistinguishable from turbulence in the water with the sensor not vibrating. Generally, a measurable dissipation lower limit for these instruments of $10^{-9}\,\mathrm{W\,kg^{-1}}$ would be considered good. These shear probes have also been mounted on submarines for horizontal measurements. Nevertheless, this platform has had only limited use because of vehicle noise and expensive operating costs. More recently, unmanned submarines called autonomous underwater vehicles have been used as suitable platforms for turbulent kinetic energy measurements. They are expected to have similar noise characteristics to towed instruments. Another interesting way of obtaining horizontal measurements is to place shear probes on a moored instrument. The turbulence in the water is measured as it flows past the sensor at a speed $V\,\mathrm{m\,s^{-1}}$. In this case, the water velocity must typically be faster than $0.1\,\mathrm{m\,s^{-1}}$ for the measurements to be within the sensor capabilities and mooring vibrations generate similar problems to towed instruments.

Thin Film Sensors

One of the original sensors used to measure turbulence and dissipation is called a hot film sensor. In these sensors, a platinum or nickel film is deposited on the surface near the conical tip of a glass rod of order 1 mm diameter and covered with a thin film of quartz to insulate it from the water. The film is heated to several degrees centigrade above the ambient temperature and special electronics are used to maintain a constant thin film temperature. Water flowing across the probe cools the platinum. Fluctuations in the current, required to keep the sensor at a constant temperature, are a measure of the turbulent velocity fluctuations along the axis of the sensor. This sensor measures the $E_1(k_1)$ component of the turbulent field as opposed to the $E_2(k_1)$ component measured by the shear probe. Therefore, the first part of the eqn [3] is relevant to estimating dissipation. The primary advantage of this sensor over the shear probe is that it has much smaller spatial resolution and a much higher frequency response. As one can see from **Figure 3**, this allows one to measure to higher dissipation rates. The disadvantages of this probe are that the electronics to run it are much more complicated than for shear probes and the sensors are more difficult to fabricate and quite expensive. They also require a lot of power to heat since they are very low in resistance (of order 5–10 Ω). Because the quartz insulation must be extremely thin to provide good heat transfer, thin films are also very fragile and easily damaged by impact with particles in the water. These probes do not provide an output voltage that is linear with turbulent velocity fluctuations. They also tend to be noisy and subject to fouling. They are seldom used today in ocean measurements.

Pitot Tubes

Another recently developed sensor used to measure dissipation makes use of a Pitot tube. If a Pitot tube is placed in water flowing at a speed W along its axis, the pressure generated by the flow is proportional to W^2. This technique has been applied to turbulence measurements by carefully designing an axisymmetric port a few millimeters in diameter on the tip of a sensor of order 1 cm in diameter. By connecting the port to a very sensitive differential pressure sensor, fluctuations in pressure along the axis of the probe can be measured. Using suitable electronic circuits, a signal is produced that is linearly proportional to along-axis fluctuations in turbulent velocity. In this sense, it is similar to the heated-film sensor and different from the shear probe which measures fluctuations perpendicular to the mean flow. This sensor has been used in conjunction with a pair of shear probes to simultaneously measure all three components of turbulent velocity fluctuations.

Temperature Microstructure Sensors

As outlined above, if there is turbulent mixing occurring in a region where there is a temperature gradient, the turbulent velocity will cause the temperature to be mixed. If, for example, warmer fluid overlays colder fluid, turbulence will move parcels of warm fluid down and cold fluid up. A temperature sensor that traverses a patch of fluid such as this will measure fluctuations in temperature as shown in **Figure 2**. The spectrum of these fluctuations can be used to determine the dissipation using eqn [5]. Because the molecular diffusivity of heat for water is much smaller than the molecular viscosity, the scale at which temperature fluctuations cease is about a factor of three smaller than the scale at which velocity fluctuations cease. This is shown clearly in **Figure 3(A)** that compares L_T and L_ν. These facts place a severe restriction on the speed and size of a temperature sensor compared to a shear probe, or alternatively limits the speed that an instrument may fall. For the same fall speed, a temperature sensor must be sampled at a much higher rate than a shear probe. The simplest temperature sensor with the precision and noise level to measure temperature microstructure in the ocean is the thermistor. The smallest thermistors that are used in sea water are a fraction of millimeter in diameter and have a frequency response of order 10 ms. At a flow speed of $1\,\mathrm{m\,s^{-1}}$, one is able to delimit the spectrum of temperature for dissipations up to about $10^{-7}\,\mathrm{W\,kg^{-1}}$. Some success has been obtained by using very slow moving profilers that fall or rise at about $0.1\,\mathrm{m\,s^{-1}}$. An alternative to the thermistor is a thin film thermometer. It is similar to the hot film velocity sensor described above and is constructed identically. Used as a thermometer, the change in the resistance of this sensor is a measure of change of temperature. Thin film sensors are faster than thermistors, typically with a time-constant of 2 ms which means that for any sensor velocity the temperature fluctuations may be measured to a higher wave number. These sensors are nevertheless at least an order of magnitude noisier than thermistors, which means that they are suitable for measuring microstructure only in regions where there are strong mean gradients. Using thermometry to measure dissipation is subject to large errors because, as indicated in eqn [5], dissipation is proportional to $(L_T)^4$, and this requires accuracy in determining L_T that is seldom achieved.

Some success has been made using sensors that measure conductivity as a proxy for temperature. These sensors make use of the fact that the conductivity of sea water is determined by both salt and temperature and in most cases, the temperature causes most of the fluctuations. The techniques used are similar to those described above for temperature. Some of the sensors are smaller and faster than thermistors and less noisy than thin film thermometers. They are still limited to the same constraints as thermometers in that they must fully resolve the spectrum in order to estimate L_T and utilize eqn [5].

Acoustic Current Meters

Acoustic techniques have also been used to measure water velocity in the ocean and indirectly to infer dissipation rates. One such technique utilizes an acoustic Doppler current meter optimized to measure vertical velocity fluctuations in the water column. In these instruments, a sound pulse is transmitted into the water and the sound scattered back to a sound receiver. The back-scattered pulse contains information about the water velocity because of the Doppler shift in the sound frequency. This technique is unable to measure to dissipation scales but instead, measures vertical velocities in the $k^{-5/3}$ wavenumber range defined by eqn [1]. By suitably defining a turbulent timescale, dissipation is estimated from the intensity in the fluctuations in the vertical velocity. This technique is very useful in studying turbulence in regions of intense mixing such as tidally driven flows.

In another technique, an array of small acoustic transmitters and receivers is configured such that the transit time of a pulse of sound can be measured over a short distance of around 10–20 cm. Velocity fluctuations in the water change the transit time and allow water velocity fluctuations to be inferred. This configuration of sensors is generally mounted as a fixed array on a platform on the bottom and has been used to measure turbulent mixing in many places on continental shelves. This technique has the advantage over profiling dissipation sensors of measuring three components of velocity fluctuations over long periods of time at a single place. Dissipation is estimated from the $k^{-5/3}$ wavenumber range.

Conclusions

The measurement of mixing rates in the ocean is important to our understanding of the distributions of temperature, salinity, and nutrients in the ocean. We need to understand this to include them correctly in climate and biological ocean models. The way in which energy is converted from sources at large scale and dissipated at small scales has required the development of a variety of ocean sensors. Some of these are described briefly above. It is hoped that enough of the key words and ideas have been put forward for the reader to understand some of the principles involved in turbulence measurement and at least some of the sensors and techniques used.

Symbols used:

α	An experimentally determined spectral constant
$E(k)$	Energy spectral density
$E_1(k_1)$	One-dimensional energy wavenumber spectrum – fluctuations along the axis of measurement
$E_2(k_1)$	One-dimensional energy wavenumber spectrum – fluctuations perpendicular to the axis of measurement
ε	Dissipation of turbulent kinetic energy
f	measurement or sampling frequency
κ_T	molecular diffusivity of heat
L_ν	viscous cutoff scale
L_T	temperature cutoff scale
u	horizontal velocity fluctuation
ν	kinematic viscosity
V	flow velocity along axis of sensor
W	drop velocity
z	distance coordinate (normally vertical)

See also

Dispersion and Diffusion in the Deep Ocean. Dispersion in Shallow Seas. Fossil Turbulence. Internal Tidal Mixing. Intrusions. Island Wakes. Langmuir Circulation and Instability. Meddies and Sub-surface Eddies. Mesoscale Eddies. Profiling Current Meters. Three-dimensional (3D) Turbulence. Topographic Eddies. Turbulence in the Benthic Boundary Layer. Under-ice Boundary Layer.

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