
VMP 2020 Training

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Instructors: Evan Cervelli
Justine McMillan

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Prepared by: Justine McMillan
Prepared on: 2020-04-30
Authors: Rolf Lueck
Emma Murowinski
Evan Cervelli
Justine McMillan
Cynthia Bluteau

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info@rocklandscientific.com
tel: +1 250 370 1688

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1 Introduction to Oceanic Turbulence

1.1 What is Turbulence?

Turbulent flows are ubiquitous in both natural systems (e.g. atmospheric and oceanic boundary layers) and engineering applications (e.g. pipe flow, wake generation). Despite its universality, turbulence is difficult to define. Instead, it is described as a *syndrome* characterized by a set of symptoms, which together, distinguish it from non-turbulent fluid motions (Stewart, 1969). In this sense, all turbulent flows can be described as (Tennekes and Lumley, 1989):

- *irregular*: Turbulent flows are disordered, unsteady, and chaotic. The kinetic energy is concentrated into bursts that are separated by comparatively quiescent regions.
- *rotational*: Turbulent flows possess three-dimensional vorticity appearing as identifiable structures (e.g., eddies) that exist over a broad range of length scales.
- *diffusive*: Turbulent flows cause rapid mixing. Transfer rates of momentum, heat and other scalar properties are much higher than if only molecular diffusion processes are involved.
- *dissipative*: Turbulent flows irreversibly lose kinetic energy to heat through the action of viscosity on the smallest scales.
- *a continuum*: The size of the smallest eddies in a turbulent flow is still many of factors of 10 larger than any molecular length scale.

An example of a weak turbulent flow depicting these characteristics is shown in [Figure 1](#).



Figure 1: An iced-latte. An example of weak turbulence with a background gradient of two scalars – temperature and cream-concentration.

Turbulent flows are quantitatively characterized by a high Reynolds number, which is the ratio of inertial and viscous forces acting within a fluid. Mathematically, the Reynolds number is defined as

$$\text{Re} = \frac{\mathcal{U}\mathcal{L}}{\nu}, \quad (1)$$

where $\mathcal{U}$ is a velocity scale, $\mathcal{L}$ is a length scale and ν is the kinematic viscosity of the fluid.

Turbulence is a characteristic of a *flow* and not of a *fluid*. There are non-turbulent flows that have some but not all of the above-listed characteristics. For example, surface waves can be highly irregular, but the water motion induced by surface gravity waves is irrotational and, hence, non-turbulent. In fact, surface waves are highly linear until they grow to the verge of breaking. After the waves break, they can induce considerable turbulence into near-surface water.

1.2 Navier–Stokes equation

The starting point for all analytic treatments of turbulence are the Navier-Stokes equations,

$$\frac{\partial \tilde{u}_i}{\partial t} + \tilde{u}_j \frac{\partial \tilde{u}_i}{\partial x_j} = -\frac{1}{\rho_0} \frac{\partial \tilde{P}}{\partial x_i} - \frac{g}{\rho_0} (\tilde{\rho} - \rho_0) \delta_{i3} + \nu \frac{\partial^2 \tilde{u}_i}{\partial x_j \partial x_j}, \quad (2)$$

$$\frac{\partial \tilde{u}_i}{\partial x_i} = 0, \quad (3)$$

$$\frac{\tilde{\rho}}{\rho_0} = 1 - \alpha \tilde{T} + \beta \tilde{S}, \quad (4)$$

$$\frac{\partial \tilde{T}}{\partial t} + \tilde{u}_j \frac{\partial \tilde{T}}{\partial x_j} = \kappa_T \frac{\partial^2 \tilde{T}}{\partial x_j \partial x_j}, \quad (5)$$

$$\frac{\partial \tilde{S}}{\partial t} + \tilde{u}_j \frac{\partial \tilde{S}}{\partial x_j} = \kappa_S \frac{\partial^2 \tilde{S}}{\partial x_j \partial x_j}, \quad (6)$$

where the instantaneous quantities are identified by a tilde ($\sim$), u is the velocity, P the pressure, ρ the fluid density, which is determined by the temperature T and the salinity S , ρ_0 is an arbitrary reference density, ν the kinematic molecular velocity, and κ_T and κ_S are the molecular diffusivity of heat and salt, respectively. The subscripts i and j are the standard Einstein notation for the three components of direction. Coriolis acceleration has been ignored.

1.3 The Reynolds Decomposition – mean properties

If the flow is statistically stationary, velocity and scalar quantities can be decomposed into the sum of a mean value and a zero-mean fluctuation – the Reynolds decomposition –

$$\begin{aligned} \tilde{u}_i &= U_i + u_i, \\ \tilde{P} &= P + p, \\ \tilde{\rho} &= \rho + \rho', \\ \tilde{T} &= T + T', \\ \tilde{S} &= S + S'. \end{aligned} \quad (7)$$

where

$$\overline{u_i} = \overline{p} = \overline{\rho'} = \overline{T'} = \overline{S'} = 0, \quad (8)$$

and the over-line denotes a spatial average. Incompressibility holds for the mean and the fluctuation velocity

$$\frac{\partial U_i}{\partial x_i} = \frac{\partial u_i}{\partial x_i} = 0 . \quad (9)$$

Mean momentum equation

Substituting eq. (7) into eq. (2) and using eq. (9) we can derive the equation for the mean momentum. Each of the terms can be simplified as follows:

- mean rate of change:

$$\overline{\frac{\partial}{\partial t} (U_i + u_i)} = \frac{\partial U_i}{\partial t} + \frac{\partial \overline{u_i}}{\partial t} = \frac{\partial U_i}{\partial t} + \frac{\partial \overline{u_i}}{\partial t} = \frac{\partial U_i}{\partial t} , \quad (10)$$

- mean advection:

$$\begin{aligned} \overline{(U_j + u_j) \frac{\partial}{\partial x_j} (U_i + u_i)} &= U_j \frac{\partial U_i}{\partial x_j} + U_j \frac{\partial \overline{u_i}}{\partial x_j} + \overline{u_j} \frac{\partial U_i}{\partial x_j} + \overline{u_j \frac{\partial u_i}{\partial x_j}} \\ &= U_j \frac{\partial U_i}{\partial x_j} + \frac{\partial}{\partial x_j} \overline{u_i u_j} , \end{aligned} \quad (11)$$

- mean pressure gradient:

$$\overline{\frac{\partial}{\partial x_i} (P + p)} = \frac{\partial P}{\partial x_i} , \quad (12)$$

- mean mass (and similarly for the other scalars of T and S):

$$\overline{(\rho + \rho' - \rho_0) \delta_{i3}} = (\rho - \rho_0) \delta_{i3} , \quad (13)$$

- mean viscous drag:

$$\overline{\nu \frac{\partial^2}{\partial x_j \partial x_j} (U_i + u_i)} = \nu \frac{\partial^2 U_i}{\partial x_j \partial x_j} . \quad (14)$$

Combining the terms gives the mean momentum equation

$$\frac{\partial U_i}{\partial t} + U_j \frac{\partial U_i}{\partial x_j} + \frac{\partial}{\partial x_j} \overline{u_i u_j} = -\frac{1}{\rho_0} \frac{\partial P}{\partial x_i} - \frac{g}{\rho_0} (\rho - \rho_0) \delta_{i3} + \nu \frac{\partial^2 U_i}{\partial x_j \partial x_j} , \quad (15)$$

which can be rewritten as

$$\frac{DU_i}{Dt} = -\frac{1}{\rho_0} \frac{\partial P}{\partial x_i} - \frac{g}{\rho_0} (\rho - \rho_0) \delta_{i3} + \nu \frac{\partial^2 U_i}{\partial x_j \partial x_j} - \frac{\partial}{\partial x_j} \overline{u_i u_j} , \quad (16)$$

where D/Dt is the material derivative and represents the total rate of change of a fluid parcel. This equation closely resembles the form of eq. (2), except for the last term on the right hand side of eq. (16). Importantly, the covariance of the velocity fluctuations, $\overline{u_i u_j}$ and its gradient are generally non-zero even though $\overline{u_i} = 0$, so this term cannot be ignored.

Reynolds stress

The covariance $-\overline{u_i u_j}$ in eq. (16) is the kinematic Reynolds stress¹ and has some similarity to the viscous term in the mean momentum equation. For example, we can combine these two terms, which are the last two on the right hand side of eq. (16), to read

$$\frac{\partial}{\partial x_j} \left(\nu \frac{\partial U_i}{\partial x_j} - \overline{u_i u_j} \right). \quad (17)$$

The first term is the stress due to molecular motions and is a property of the *fluid*. The second term is also a stress but it is a property of the *flow*, and is usually much larger than the viscous term.

The Reynolds stress, $-\overline{u_i u_j}$, is a 9-component symmetric tensor with 6 possibly unique terms. If the turbulence is isotropic (independent of direction), the off-diagonal terms ($i \neq j$) are zero, and the diagonal terms ($i = j$) are equal.

It is easy to see why the Reynolds stress is positive in a sheared flow (Figure 2). If a fluid parcel has a mean position of y , then a positive v perturbation brings it into an environment of faster u -velocity, making the instantaneous product $uv < 0$. Likewise, if the parcel moves down, $v < 0$, it will have a positive anomaly with respect to its environment, $u > 0$, and the instantaneous product is also negative. Moreover, a parcel that has temporarily gone up, will return to its equilibrium position y , with more momentum than it had before going up and, similarly, a parcel that has gone down will return with less momentum. The average result is that $\overline{uv} < 0$ and momentum has been transferred downward and, hence, down the gradient of mean velocity.

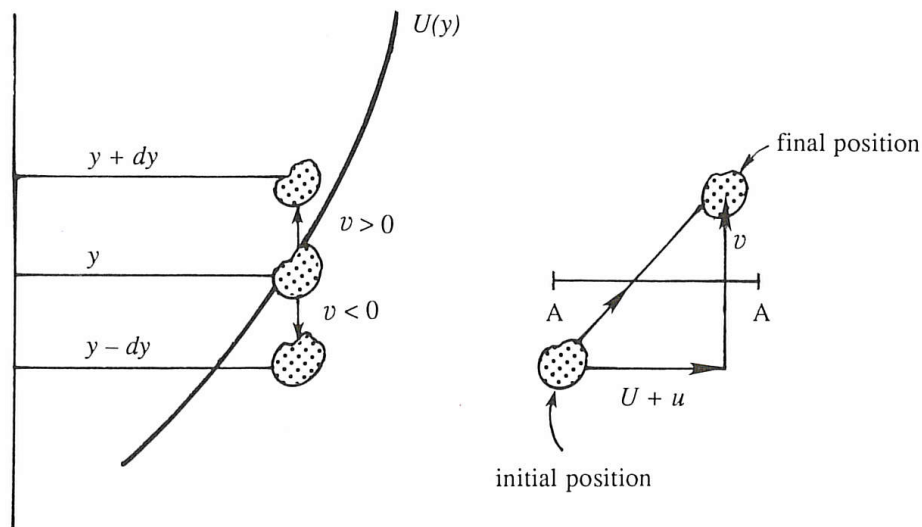


Figure 2: A schematic of macroscopic fluid perturbations in a turbulence vertically sheared flow, after Kundu (1990).

The Reynolds stress is the main “friction” in a turbulent flow and it is produced through *macroscopic* parcels of fluid and not by *microscopic* molecules. The exchange of momentum is far more effective than it would be through molecular friction. In a turbulent tidal channel, $\sqrt{-\overline{uv}}/U \approx 0.2$.

¹the Reynolds stress per unit mass.

Mean scalar equation

The mean equation for either scalar (T or S) can be derived following the procedure used for momentum. For example, using eq. (5) and eq. (7), the mean heat equation is

$$\begin{aligned} \frac{\partial T}{\partial t} + U_j \frac{\partial T}{\partial x_j} + \frac{\partial}{\partial x_j} \overline{u_j T'} &= \kappa_T \frac{\partial^2 T}{\partial x_j \partial x_j} , \\ \frac{DT}{Dt} &= \frac{\partial}{\partial x_j} \left(\kappa_T \frac{\partial T}{\partial x_j} - \overline{u_j T'} \right) , \end{aligned} \quad (18)$$

where the turbulent heat flux $-\overline{u_j T'}$ is generally non-zero and directed down-gradient, if there is a mean background gradient of temperature, T . This can be visualized by replacing the background profile of velocity $U(y)$ in Figure 2, with a profile of $T(y)$ subject to fluctuations of velocity v . The molecular heat flux (the first term in the bracket on the right-hand side of eq. (18)) is usually very small compared to the turbulent heat flux. The mean salinity has a similar equation except that the molecular diffusivity, κ_S , is about 100 times smaller than κ_T .

There is a fundamental difference between the momentum and scalar equations – the scalar equation is linear. In addition, if we view $U(y)$ in Figure 2 as a temperature profile, $T(y)$, then the exchange process of heat is different from that for momentum. Momentum can be transferred to a displaced parcel of water by inertial- and pressure-forces and by molecular (viscous) diffusion. However, heat can only be transferred to a displaced parcel by molecular (thermal) diffusion. This has implications for the smallest scales that are possible for momentum and scalar fluctuations.

There is no equation for the ‘conservation of density’. The fluid density is determined by its temperature and salinity, and the diffusion and turbulent transport of these scalars determines the behaviour of the fluid density.

The closure problem – Something gained, something lost

The Reynolds decomposition has converted the equations of motion into ones that look like simple viscous flow. However, it has also generated a new term – the Reynolds stress – which is a function of the flow and is, therefore, a new unknown variable that must be related to the velocity field in order to “close” the problem. Some functional form must be hypothesized before we can solve the equations for the mean flow. The validity of the solution depends directly on the validity of the hypothesis. This is the crux of the problem regardless of whether we try to solve the equations analytically or numerically.

One simple but almost always inappropriate approach, and one that is often used in global circulation models, is to treat the Reynolds stress as a viscosity – an eddy viscosity – in analogy to molecular friction. For example,

$$-\overline{uw} = K_e \frac{\partial U(z)}{\partial z} , \quad (19)$$

then,

$$\tau = \frac{\partial}{\partial x_j} \left(\nu \frac{\partial U_i}{\partial x_j} + K_e \frac{\partial U_i}{\partial x_j} \right) = \frac{\partial}{\partial x_j} \left((\nu + K_e) \frac{\partial U_i}{\partial x_j} \right) = (\nu + K_e) \frac{\partial^2 U_i}{\partial x_j \partial x_j} , \quad (20)$$

for a unidirectional and vertically sheared flow. The eddy viscosity, K_e , is seldom a constant. In a bottom boundary layer, the Reynolds stress is constant and independent of height in the lower

few metres (where the vertical shear is large) and then decreases with increasing height above the bottom. The classic Ekman layer spiral is analytically derived by hypothesizing a constant eddy viscosity. However, no one has ever seen a surface flow that is 45° to the wind direction. Typically the surface flow is 10° to 20° off the direction of the surface wind.

Thus, a major occupation of turbulence research is to derive closure hypotheses that are reasonably in agreement with observations.

1.4 The Reynolds Decomposition – second-order properties

Kinetic energy of the mean flow

Multiplying eq. (16) by U_i and summing over i , gives us the equation for the kinetic energy of the mean flow, namely

$$\begin{aligned} \frac{D}{Dt} \left(\frac{1}{2} U_i^2 \right) &= \frac{\partial}{\partial x_j} \left(-\frac{1}{\rho_0} P U_j + \overline{u_i u_j} U_i + \nu \frac{\partial \frac{1}{2} U_i^2}{\partial x_j} \right) \\ &\quad + \overline{u_i u_j} \frac{\partial U_i}{\partial x_j} - \nu \left(\frac{\partial U_i}{\partial x_j} \right) \left(\frac{\partial U_i}{\partial x_j} \right) - \frac{g}{\rho_0} (\rho - \rho_0) U_3 . \end{aligned} \quad (21)$$

The term on the left side of the equality is the total rate of change of the kinetic energy of the mean flow, following a water parcel. In a non-accelerating and steady flow, this term is zero. The first term on the right side of the equality is the transport (or divergence) of kinetic energy. It is the rate at which energy enters due to internal forces directed along the flow. The second term is the loss due to turbulence because the Reynolds stress, $-\overline{u_i u_j}$, is usually positive. It is a stress working against the mean shear. The third term is the loss of kinetic energy due to molecular friction of the mean flow and is always small. The last term is a loss to gravitational potential energy by the mean vertical velocity, which is also negligible in most geophysical flows².

In a steady geophysical flow, the transport term must compensate for the loss of mean KE given by

$$\overline{u_i u_j} \frac{\partial U_i}{\partial x_j} , \quad (22)$$

Typically, the pressure-velocity term is the main compensator. Think of a simple tidal channel of constant cross-section, flowing in the x -direction. The pressure-velocity term is

$$\frac{U}{\rho_0} \frac{\partial P}{\partial x} = g U \frac{\partial \eta}{\partial x} , \quad (23)$$

where η is the surface elevation. Water is flowing “down hill”. The kinetic energy is generated from a lowering of the potential energy, and lost to the Reynolds stress term in eq. (21).

Kinetic energy of the turbulence

We can derive the equation for the kinetic energy of the turbulence using the following steps. First we take the Navier-Stokes equation (eq. (2)) for the total velocity and subtract from it the equation for the mean flow given by eq. (16), to derive an equation for the fluctuation of velocity, u_i . We then take this equation and multiply it by u_i and take its average. The process

²This term would not be negligible in a flow such as that in a water fountain

is tedious and is well described in Kundu (1990). The result is

$$\begin{aligned} \frac{D}{Dt} \left(\frac{1}{2} \overline{u_i^2} \right) = & - \frac{\partial}{\partial x_j} \left(\frac{1}{\rho_0} \overline{p u_j} + \frac{1}{2} \overline{u_i^2 u_j} + \frac{\nu}{2} \frac{\partial \overline{u_i^2}}{\partial x_j} \right) \\ & - \overline{u_i u_j} \frac{\partial U_i}{\partial x_j} - \nu \overline{\left(\frac{\partial u_i}{\partial x_j} \right) \left(\frac{\partial u_i}{\partial x_j} \right)} - \frac{g}{\rho_0} \overline{\rho' u_3} . \end{aligned} \quad (24)$$

The term on the left side of the equality is the total rate of change of turbulence kinetic energy. The first term on the right side is the transport (or divergence) of turbulence kinetic energy. It is frequently small or simply ignored in many common situations. The second term on the right involving the Reynolds stress is now familiar. It is the negative of the corresponding term for the mean kinetic energy found in eq. (21). It is the transfer of kinetic energy from the mean flow to the turbulence – the rate of production of turbulence kinetic energy – and is usually the only source of energy that sustains the turbulence. The third term is the rate of dissipation of kinetic energy by molecular friction and is usually identified by the symbol ϵ . It is always a loss of, or sink for, kinetic energy. It is huge compared to the viscous term in the equation for the mean kinetic energy (eq. (21)) because the gradients of the mean velocity are never very large and the loss of mean kinetic energy is small. However, the gradients of the velocity fluctuations can be enormous because the eddies can have very small linear dimensions, $\sim 1 \times 10^{-2}$ m. The last term is the loss of turbulence kinetic energy due to the vertical flux of buoyancy. In a stably stratified flow, the buoyancy flux is a sink for kinetic energy because $\overline{\rho' u_3}$ is positive. It raises the potential energy of the fluid by transporting mass upwards. Over a heated boundary, such as the atmosphere on a sunny day, this term is a source of energy for the turbulence. It lowers the potential energy of the fluid by moving mass downwards.

In a steady and non-divergent flow, only the last three terms are significant and eq. (24) reduces to

$$0 = \mathcal{P} + \mathcal{B} - \epsilon, \quad (25)$$

where

$$\mathcal{P} = \text{Shear Production} = -\overline{u_i u_j} \frac{\partial U_i}{\partial x_j}, \quad (26)$$

$$\mathcal{B} = \text{Buoyancy Production} = -\frac{g}{\rho_0} \overline{\rho' u_3}, \quad (27)$$

$$\epsilon = \text{Viscous Dissipation} = \nu \overline{\left(\frac{\partial u_i}{\partial x_j} \right)^2}. \quad (28)$$

The flow of TKE

A simple example of the flow of kinetic energy is that of a steady, vertically sheared, and stably stratified flow over a level bottom, such as might be found in a long tidal channel of uniform cross-section. In such a case, the mean velocity has only one component, $U(z)$, all cross-channel and down-stream gradients of mean quantities vanish, $\partial/\partial x = 0$ and $\partial/\partial y = 0$. The only exception is $\partial P/\partial x$, because there has to be a gradient of pressure to drive the flow, usually from a sloping free surface. The mean kinetic energy equation (eq. (21)) reduces to

$$\frac{\partial}{\partial x} (PU) = \overline{uw} \frac{\partial U}{\partial z}. \quad (29)$$

The production of kinetic energy by flowing “down hill” is transferred to the turbulence by the shear working against the Reynolds stress. The equation for the turbulence kinetic energy (eq. (24)) reduces to

$$\frac{\partial}{\partial z} \left(\frac{1}{\rho_0} \overline{pw} + \frac{\nu}{2} \frac{\partial}{\partial z} \overline{u_i^2} \right) + \overline{uw} \frac{\partial U}{\partial z} + \epsilon + \frac{g}{\rho_0} \overline{\rho'w} = 0. \quad (30)$$

The vertical divergence – i.e. the first term – is usually taken to be small, which leaves a balance of production, buoyancy flux, and dissipation. Thus, the kinetic energy produced by the mean flow is lost to viscous dissipation and some goes into producing potential energy.

To see the interaction between individual velocity components of the turbulence, and how energy flows from one component to the other, we use the following procedure. We take the equation for the instantaneous flow (eq. (2)) and subtract from it the equation for the mean flow (eq. (16)), to produce a set of three equations, $i = 1 - 3$, for the velocity fluctuations. For each case of i we multiply the equation by u_i and take the average of this product. The difference between the instantaneous and mean equations is

$$u_j \frac{\partial U_i}{\partial x_j} + u_j \frac{\partial u_i}{\partial x_j} - \frac{\partial}{\partial x_j} \overline{u_i u_j} = -\frac{1}{\rho_0} \frac{\partial p}{\partial x_i} - \frac{g}{\rho_0} \rho' \delta_{i3} + \nu \frac{\partial^2 u_i}{\partial x_j \partial x_j}, \quad (31)$$

where the total derivative $D/Dt = 0$. Multiplying the above by u for the case of $i = 1$ gives

$$-\overline{uw} \frac{\partial U}{\partial z} - \frac{1}{2} \left(\frac{\partial}{\partial x} \overline{uu^2} + \frac{\partial}{\partial y} \overline{vu^2} + \frac{\partial}{\partial z} \overline{wu^2} \right) = \frac{1}{\rho_0} \frac{\partial}{\partial x} \overline{pu} - \frac{\nu}{2} \nabla^2 \overline{u^2} + \nu \overline{\left(\frac{\partial u}{\partial x_j} \right)^2}, \quad (32)$$

and multiplying by v for the case $i = 2$ gives

$$-\frac{1}{2} \left(\frac{\partial}{\partial x} \overline{uv^2} + \frac{\partial}{\partial y} \overline{vv^2} + \frac{\partial}{\partial z} \overline{wv^2} \right) = \frac{1}{\rho_0} \frac{\partial}{\partial y} \overline{pv} - \frac{\nu}{2} \nabla^2 \overline{v^2} + \nu \overline{\left(\frac{\partial v}{\partial x_j} \right)^2}, \quad (33)$$

and multiplying by w for the case $i = 3$ gives

$$-\frac{1}{2} \left(\frac{\partial}{\partial x} \overline{uw^2} + \frac{\partial}{\partial y} \overline{vw^2} + \frac{\partial}{\partial z} \overline{ww^2} \right) = \frac{1}{\rho_0} \frac{\partial}{\partial z} \overline{pw} - \frac{\nu}{2} \nabla^2 \overline{w^2} + \frac{g}{\rho_0} \overline{\rho'w} + \nu \overline{\left(\frac{\partial w}{\partial x_j} \right)^2}, \quad (34)$$

where

$$\nabla^2 = \frac{\partial}{\partial x^2} + \frac{\partial}{\partial y^2} + \frac{\partial}{\partial z^2}, \quad (35)$$

and we have used the identity

$$u \frac{\partial^2 u}{\partial x_j \partial x_j} = \frac{1}{2} \frac{\partial^2 u^2}{\partial x_j \partial x_j} - \frac{\partial u}{\partial x_j} \frac{\partial u}{\partial x_j}, \quad (36)$$

which holds for all of the three velocity components, u , v , and w . Only the equation for the downstream velocity fluctuations (eq. (32)) has a source of energy – the Reynolds stress working against the mean shear – and this energy comes from the mean flow. However, all three components of velocity fluctuations have a sink to viscous dissipation (the last term in each equation). Therefore, the energy in the v - and w -fluctuations does not come directly from the mean flow. Instead, it comes from a re-distribution of the kinetic energy in the u -component. This redistribution is accomplished by pressure–velocity correlations and by inertial forces (the triple velocity correlations). The vertical velocity fluctuations have an additional loss due to the

vertical buoyancy flux, the second last term in eq. (34). If the density is completely homogeneous (a well mixed channel), then the buoyancy flux is zero. If the bottom is a source of heat, such as is the case for the atmospheric boundary layer on a sunny day, then this term is a source of turbulence KE , which gets redistributed to the other velocity components.

The kinetic energy is transferred to the turbulence at the largest scales of the flow and, so, we can expect the largest scales to be anisotropic because the energy enters only into the u -fluctuations. The energy will be transferred to ever smaller scales by the redistribution process, and this will tend to make the velocity isotropic at smaller scales. It is usually assumed that at the very smallest scales – where the kinetic energy is irreversibly lost to friction – that the velocity fluctuations are completely isotropic.

Scalar fluctuation variance

We can develop the equation for the conservation of scalar variance by taking the equation for the total temperature ((5)) and subtract from it the equation for the mean temperature, T , to derive the equation for the fluctuation temperature, T' . That is,

$$\frac{\partial T'}{\partial t} + u_j \frac{\partial T}{\partial x_j} + u_j \frac{\partial T'}{\partial x_j} + U_j \frac{\partial T'}{\partial x_j} - \frac{\partial}{\partial x_j} \overline{u_j T'} = \kappa_T \frac{\partial^2 T'}{\partial x_j \partial x_j} . \quad (37)$$

We now multiply this equation by T' and average it to get

$$\begin{aligned} \frac{\partial}{\partial t} \overline{T'^2} + 2\overline{u_j T'} \frac{\partial T}{\partial x_j} + U_j \frac{\partial}{\partial x_j} \overline{T'^2} + \frac{\partial}{\partial x_j} \overline{u_j T' T'} &= 2\kappa_T \overline{T' \frac{\partial^2 T'}{\partial x_j \partial x_j}} \\ &= \kappa_T \nabla^2 \overline{T'^2} - 2\kappa_T \overline{\left(\frac{\partial T'}{\partial x_j} \right) \left(\frac{\partial T'}{\partial x_j} \right)} , \end{aligned} \quad (38)$$

which can be reduced to

$$\frac{D}{Dt} \overline{T'^2} + 2\overline{u_j T'} \frac{\partial T}{\partial x_j} + \frac{\partial}{\partial x_j} \overline{u_j T' T'} = \kappa_T \nabla^2 \overline{T'^2} - 2\kappa_T \overline{\left(\frac{\partial T'}{\partial x_j} \right) \left(\frac{\partial T'}{\partial x_j} \right)} . \quad (39)$$

The last term on the right side of the equality is always a sink for the variance of temperature fluctuations and it must be large because it is proportional to the square of the gradient of temperature fluctuations. The first term on the right must be small because it is proportional to the curvature of the scalar *variance*. In a steady state, where the total rate of change is negligible, the only term that can produce scalar variance is the second term on the left side of the equality. In this case, the flux of heat, $-\overline{u_j T'}$, down the gradient of mean temperature. So this is the source of scalar-fluctuation variance. The third term on the left side is a turbulent transport of temperature fluctuations and is usually small.

Importantly, the scalar fluctuations are “passive” – they do not drive anything – and are a consequence of *velocity* turbulence in the presence of a mean gradient of a scalar. Thus, they are an indicator of turbulence, subject to the presence of a background scalar gradient. However, because the scalar fluctuations are driven by the velocity field, the structure of the fluctuations of a scalar does have a relationship to the structure of the velocity fluctuations. This property is sometimes exploited when a velocity measurement is not possible.

1.5 The Energy Cascade and Isotropy

The production of turbulence occurs at large scales where large eddies extract energy from the mean flow or the buoyancy flux. Spatial gradients within these eddies are weak and, hence, the direct effect of viscosity is negligibly small. Richardson (1922) proposed that there exists an energy cascade in which the TKE is transferred from these large eddies to smaller and smaller scales, until eventually, the smallest eddies are dissipated as heat by viscosity. He described this phenomenon poetically as,

*Big whorls have little whorls
That feed on their velocity,
And little whorls have lesser whorls
And so on to viscosity.*

At the large scale end of this cascade, the flow is spatially inhomogeneous and is directionally dependent on the gradients in the mean flow – that is, the flow is *anisotropic*. However, as the eddies interact with each other and pass their energy to the smaller scales (through overturns and vortex stretching), the directional biases are lost and the motions become locally *isotropic*: i.e.,

$$\overline{u_1^2} = \overline{u_2^2} = \overline{u_3^2}. \quad (40)$$

At these small scales, the velocity fluctuations have essentially lost their memory of the largest scales and are no longer influenced by the mean flow.

1.6 Dissipation Rates

The rate of viscous dissipation of turbulent kinetic energy³ (eq. (28)) contains 9 terms, but not all of them are independent. If the turbulence is isotropic, then ϵ can be reduced to

$$\epsilon = \frac{15}{2} \nu \overline{\left(\frac{\partial u}{\partial z}\right)^2} = 15\nu \overline{\left(\frac{\partial u}{\partial x}\right)^2}, \quad (41)$$

where $\partial u/\partial z$ can be any of the 6 components of the shear of the fluctuations, and $\partial u/\partial x$ can be any of the 3 rates of strain.

A similar quantity – called the rate of dissipation of scalar variance (χ_T) – is defined from the gradient of the temperature fluctuation

$$\chi_T = 2\kappa_T \overline{\left(\frac{\partial T'}{\partial x_j}\right)^2} = 2\kappa_T \left[\overline{\left(\frac{\partial T'}{\partial x}\right)^2} + \overline{\left(\frac{\partial T'}{\partial y}\right)^2} + \overline{\left(\frac{\partial T'}{\partial z}\right)^2} \right], \quad (42)$$

where $\kappa_T \approx 1.7 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$ is the molecular diffusivity of heat. If the scalar fluctuations are isotropic, then

$$\chi_T = 6\kappa_T \overline{\left(\frac{\partial T'}{\partial x}\right)^2}, \quad (43)$$

where $\partial T'/\partial x$ can be any of the three components of the gradient of the scalar fluctuations.

³This is a mouthful and is usually just called “the dissipation” or “the rate of dissipation”, unless it needs to be distinguished from the rate of dissipation of scalar variance.

1.7 Turbulent Length Scales

Turbulence is primarily a *spatial* phenomenon because its sources and sinks depend on the *gradients* of velocity, pressure and scalar properties. Time rates of change apply primarily to the *variance* of the properties of the flow and are, therefore, small and often negligible. This is somewhat counterintuitive, because at a fixed position within a turbulent flow, the properties will fluctuate rapidly. For example, in a 2 m s^{-1} tidal flow, a turbulence eddy of size 1 m will typically advect past the position in 0.5 seconds. Over this short time, the properties of the eddy itself will hardly change because the TKE will grow and diminish with the mean tidal flow, which varies on time scales of approximately 2 hours. Similarly, in a stratified ocean, a characteristic time scale of eddy evolution might be several hours as it would scale with the inverse of the buoyancy frequency. Because the properties of the turbulence vary more significantly in space, it is more useful to characterize turbulent flows using length scales as opposed to time scales.

Most scales are derived by dimensional analysis along with a consideration for the forces that could be significant. The rate of dissipation of kinetic energy (ϵ) is frequently a parameter that determines scales because this rate will quickly adjust itself to match the rate of production at the largest scales and to the flux of buoyancy, if there is any.

The length scale of the *energy containing eddies*, in an unstratified or weakly stratified flow (e.g. tidal channel) will be comparable to the dimensions of the flow itself, usually the depth, because there is nothing else to constrain the size of eddies. If the flow is stratified, the scale will be limited by the buoyancy frequency, N , given by

$$\text{Buoyancy frequency} \quad N^2 = -\frac{g}{\rho_0} \frac{\partial \rho}{\partial z} - \left(\frac{g}{c}\right)^2 \approx \frac{g}{\rho_0} \frac{\partial \sigma_\theta}{\partial z}, \quad (44)$$

where c is the speed of sound and σ_θ is the potential density. The scale of the largest dissipating eddies – i.e. the over-turning scale – is therefore a function of both N and ϵ and, by dimensional analysis, is given by

$$\text{Ozmidov scale,} \quad L_O = \left(\frac{\epsilon}{N^3}\right)^{1/2}. \quad (45)$$

At larger scales, the production of kinetic energy is not sufficient to maintain the flow of TKE into gravitational potential energy.

The length scale at which viscous dissipation damps out the velocity fluctuations of the *smallest eddies* is

$$\text{Kolmogorov scale,} \quad L_K = \left(\frac{\nu^3}{\epsilon}\right)^{1/4}. \quad (46)$$

In the range of $L_O \rightarrow L_K$ we expect that both viscous forces and buoyancy forces to be unimportant. This range is often called the “equilibrium subrange” and is discussed in [Section 1.10](#). There is also a length scale for the smallest scalar (e.g. temperature) fluctuations, which are erased by molecular diffusion. This scale is,

$$\text{Batchelor scale,} \quad L_B = \left(\frac{\nu \kappa^2}{\epsilon}\right)^{1/4} = \left(\frac{\nu^3}{\epsilon}\right)^{1/4} \left(\frac{\kappa}{\nu}\right)^{1/2} = L_K P_R^{-1/2} \quad (47)$$

where κ is the molecular diffusivity of the scalar (e.g. κ_T for temperature) and $P_R = \nu/\kappa$ is the Prandtl number. For temperature fluctuations, $P_R \sim 7$; therefore, the fluctuations of temperature have scales smaller than those of velocity. For other scalars, such as salinity, which has a diffusivity, κ_S , 100 times smaller than that of heat, the scales of the fluctuations can be

even smaller. There are many other scales but the three given in eqs. (45)–(47) are the ones commonly used for geophysical scale turbulence.

Scales have little value without some numeric estimation. Some estimates of buoyancy frequency and dissipation rates are used to derive typical Ozmidov and Kolmogorov length scales for a range of natural waters (Table 1). Importantly, these are scale estimates. The actual scales of motions will equal these within a factor of order unity. That is, the actual scales can be up to 3 times larger to 3 times smaller than the scale estimates.

Table 1: Typical values for turbulence length scales for a range of natural waters. Batchelor scales are computed using a Prandtl number of 7 that is typical for temperature fluctuations.

Environment	N^2 s^{-2}	ϵ W kg^{-1}	L_O m	L_K m	L_B m
Surface mixing layer	0	3×10^{-7}	∞	1×10^{-3}	0.4×10^{-3}
Seasonal thermocline	100×10^{-6}	3×10^{-8}	0.2	2×10^{-3}	0.8×10^{-3}
Permanent thermocline	10×10^{-6}	3×10^{-9}	0.3	3×10^{-3}	1.1×10^{-3}
Abyssal ocean	2×10^{-6}	5×10^{-10}	0.4	4×10^{-3}	1.5×10^{-3}
Swift tidal channel	0	1×10^{-4}	∞	0.3×10^{-3}	0.1×10^{-3}

Clearly, buoyancy strongly limits the size of turbulence eddies in the ocean away from boundaries. The Ozmidov scale (L_O) seldomly exceeds 1 m. The Kolmogorov scale (L_K) is usually smaller than one centimetre and does not range widely because it depends on the fourth-root of the rate of dissipation. Therefore, the smallest scales of motion are at least 2 factors of 10 smaller than the overturning scale, L_O . So, there is a range of motions, spanning at least 2 decades, in which neither buoyancy nor viscosity are important – roughly speaking, from one metre to 1 centimetre.

1.8 Osborn and Cox model of Turbulence

One of the earliest simple models of oceanic turbulence is that of Osborn and Cox (1972) in which they asserted that the primary background gradient of temperature is the vertical component, that the flux of heat is entirely vertical, and that, in a steady or slowly evolving state, the production of temperature fluctuation variance is balanced by its rate of dissipation. Therefore, using eq. (39),

$$-\overline{wT'}\frac{\partial T}{\partial z} = 3\kappa_T \overline{\left(\frac{\partial T'}{\partial z}\right)^2} = \frac{1}{2}\chi_T, \quad (48)$$

where both T and T' are measurable from a vertical profiler, in principle. The rate of dissipation of scalar variance gives a measure of the vertical flux of heat due to turbulence. If we hypothesize that the flux is given by an eddy diffusivity (in analogy with molecular diffusivity), then

$$-\overline{wT'} = K_T \frac{\partial T}{\partial z}, \quad (49)$$

and the eddy diffusivity is

$$K_T = 3\kappa_T \overline{\left(\frac{\partial T'}{\partial z}\right)^2} / \left(\frac{\partial T}{\partial z}\right)^2, \quad (50)$$

The ratio of the variance of the gradient of the temperature fluctuations to the background gradient squared – the ratio on the left side of eq. (50) – is called the Cox number. It gives a

measure of the strength of the vertical flux due to turbulence relative to the flux by molecular diffusion. There is considerable ambiguity on whether the Cox number is the ratio or 3 times the ratio. When the Cox number is small (~ 1) turbulence is insignificant. When the water is turbulent, the Cox number readily exceeds 100. So, measuring the gradient of a scalar fluctuations provides one means of assessing the eddy diffusivity of an environment. However, it is an indirect estimate and, because momentum can be transferred by inertial and pressure forces, the eddy diffusivity may be different from the eddy viscosity. In addition, technical limitations of the scalar sensors make it difficult to accurately measure the scalar variance. See [Section 3.2](#) and Sommer et al. (2013a) for further discussion about the sensor limitations.

1.9 Osborn Model Of Turbulence

Another simple model of oceanic turbulence is that of Osborn (1980) in which he assumed that for steady flow there is a local balance between the production of turbulence kinetic energy, its rate of dissipation, and the vertical flux of buoyancy. This reduces [eq. \(24\)](#) to $0 = \mathcal{P} + \mathcal{B} - \epsilon$. Laboratory observations indicate that for a uni-directional shear flow, the buoyancy flux (i.e. $-\mathcal{B}$) is approximately 1/6 of the rate of shear production ($\mathcal{P}$), implying that

$$\mathcal{B} = -\Gamma\epsilon, \quad (51)$$

where Γ , often called the mixing efficiency, is approximately 1/5. In the ocean, there are very few estimates of Γ and they range from 0.05 to 0.3, for example Oakey (1982b) and Peterson and Fer (2014). The appropriate value of Γ to use for oceanographic studies is an ongoing debate in the research community.

Despite its simplicity and inherent assumptions, the model of Osborn provides a method for estimating the eddy diffusivity (K_ρ) using the buoyancy frequency (N) and the dissipation rate, which are both measurable. By assuming that the buoyancy flux can be written as a product of the eddy diffusivity and the density gradient (analogous to molecular diffusivity), $\mathcal{B}$ can be expressed in terms of K_ρ by

$$\mathcal{B} = -\frac{g}{\rho_0} \overline{\rho'w} = K_\rho \frac{g}{\rho_0} \frac{\partial \rho}{\partial z} = -K_\rho N^2. \quad (52)$$

Then, equating [eqs. \(51\) and \(52\)](#) yields

$$K_\rho = \frac{\Gamma\epsilon}{N^2}. \quad (53)$$

This is the most frequently used method of estimating the eddy diffusivity in the ocean, however, it makes the implicit assumption that the eddy diffusivity is the same for heat as it is for density.

1.10 Spectra of Velocity

The wavenumber spectrum of velocity fluctuations was first explored by Kolmogorov (1941) where he proposed that for steady turbulence, and for scales smaller than those of the largest energy containing eddies, l , the spectrum of velocity fluctuations, S , should depend only on the rate of dissipation, viscosity, and the wavenumber itself, namely

$$S = S(k, \epsilon, \nu), \quad (54)$$

where k is the wavenumber in units of radians per metre. The total variance of velocity is

$$\overline{u^2} = \int_0^\infty S(k) dk . \quad (55)$$

The wavenumber range of $k \gg l^{-1}$ is usually called the *equilibrium range* (Figure 3). In this range, the velocity fluctuations should not depend on the details of the energy containing scales, but only on the rate of flow of kinetic energy from the energy containing scales towards the dissipating scales. In addition, for eddies significantly larger than the Kolmogorov scale – i.e. for wavenumbers $k \ll L_K^{-1}$ – the velocity fluctuations are not dampened by viscosity. This range of $l^{-1} \ll k \ll L_K^{-1}$ is known as the *inertial subrange*. In this range, the spectrum does not depend on the viscosity and is only a function of the rate of dissipation (which equals the rate of through-flow of kinetic energy) and the wavenumber itself.

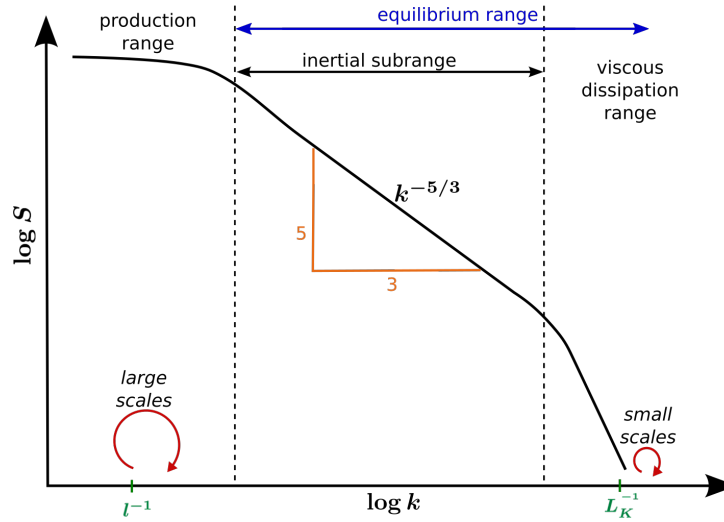


Figure 3: Schematic of a turbulence velocity spectrum.

On dimensional grounds, the three-dimensional spectrum of velocity in the inertial subrange must be

$$S = A_3 \epsilon^{2/3} |\mathbf{k}|^{-5/3} , \quad (56)$$

which is depicted schematically in Figure 3. The constant $A_3 \approx 1.5$ is of order unity and is now known as the three-dimensional Kolmogorov constant, and $\mathbf{k} = (k_1, k_2, k_3)$ is the three-dimensional wavenumber.

One-dimensional spectra

We cannot simultaneously measure all three of the components of the wavenumber. Usually, we measure only one component and this component is in the direction of profiling. In isotropic turbulence, the one-dimensional wavenumber spectrum can be derived from the three-dimensional spectrum by integrating over the entire range of the other two components of the wavenumber. Examples of this procedure can be found in Hinze (1959), Sections 3–4.

The level of the one-dimensional spectrum in the inertial subrange depends on the direction of the wavenumber component relative to the direction of the velocity component. There are two possibilities: (1) the wavenumber and the velocity component are parallel, or (2) the wavenumber and velocity component are orthogonal. For example, if we measure u as a function of x , then this is the parallel case, and we denote the spectrum by the symbol Φ_{11} . If we measure v or w as

a function of x , then this is the orthogonal case, and we denote the spectrum by the symbol Φ_{22} . For both cases, the one-dimensional spectrum retains its $k^{-5/3}$ dependence, where k is now the wavenumber in the direction of profiling. That is, the one-dimensional wavenumber spectrum is

$$\Phi_{ii} = A\epsilon^{2/3}k^{-5/3}, \quad (57)$$

where i is either 1 or 2 and summation is not implied. The value of the Kolmogorov constant, A in eq. (57), is summarized in Table 2.

Direction	A	A
k -units	[rad m ⁻¹]	[cpm]
Φ_{11}	1/2	3/20
Φ_{22}	2/3	1/5

Table 2: The one-dimensional Kolmogorov constant, A , for the cases of a velocity component that is parallel, Φ_{11} , and orthogonal, Φ_{22} , to the direction of profiling, and for the two different units of wavenumber. Measurements of A differ by only a few percent from the fractional values given here, and the accuracy of measurements does not justify using a decimal number.

There is endless, and needless, confusion over the units of wavenumber. In all theoretical derivations, the unit of wavenumber is rad m⁻¹ and is loosely taken as the inverse of a length scale. The measure of an angle is dimensionless and sloppy writers frequently abbreviate the units of wavenumber to m⁻¹. This is unacceptable because in data processing it is more natural to use a wavenumber expressed in units of cycles per metre, or simply cpm. These two measures of an angle differ by a factor of 2π , which is not small, and has resulted in significant errors (not to mention embarrassment) when the units of wavenumber are not rigorously provided on the annotation of graphs.

Validation of the Kolmogorov spectrum with ocean measurements

The final proof of the existence of the Kolmogorov spectrum did not occur until the early 1960s – almost 20 years after Kolmogorov’s paper – because of the difficulty of taking measurements in high Reynolds number turbulence. The critical evidence was obtained by towing a large torpedo-shaped vehicle through Discovery Passage located on the coast of British Columbia, Canada (Grant et al., 1962). The channel depth is ~ 300 m and speeds exceed 5 m s^{-1} , giving a Reynolds number of

$$\text{Re} = \frac{UD}{\nu} = 1.5 \times 10^9. \quad (58)$$

These measurements of the horizontal velocity fluctuation (in the direction of profiling) not only revealed an extensive inertial subrange, but they also reached into the range of wavenumbers affected by viscosity (i.e. the *viscous dissipation range*), where the spectrum rolls off more rapidly than $k^{-5/3}$ (Figure 4).

About one decade later, Nasmyth (1970) collected a very large set of velocity measurements in Discovery Passage, that spanned a wide range of rates of dissipation. He unified his measurements into a single non-dimensional spectrum of velocity fluctuations in the direction of profiling. Unfortunately, the information was presented in a PhD thesis that was never published in a scholarly journal. In fact, he did not even tabulate the empirical non-dimensional spectrum, and showed only a figure. The spectral values were later tabulated by Oakey (1982b) for general reference.

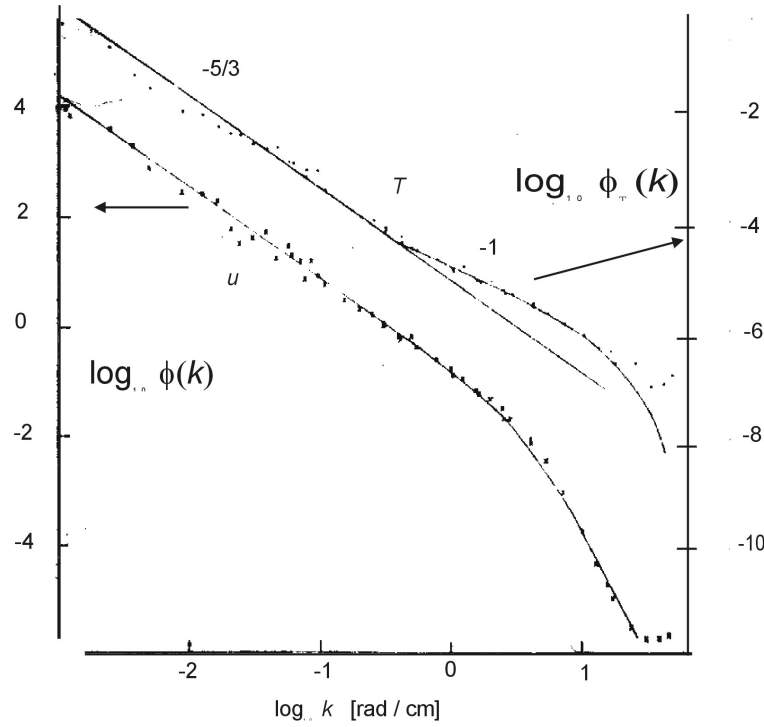


Figure 4: The one-dimensional wavenumber spectrum of velocity fluctuations u (lower curve) and temperature fluctuations T' (upper curve) at $Re \approx 6 \times 10^8$, after Grant et al. (1962).

1.11 Spectra of Velocity Gradients (Nasmyth Spectrum)

Whenever we have sensors with enough spatial resolution to measure into the viscous dissipation range of wavenumbers, we usually measure (or derive) the gradient of the fluctuations in order to achieve a better signal-to-noise ratio. This is necessary because high-wavenumbers are heavily attenuated by the $k^{-5/3}$ behaviour of velocity and scalar fluctuations. If a signal has a spectrum, say $\Phi_{11}(k)$, then its gradient in the direction of the wavenumber k is simply $k^2\Phi_{11}(k)$, provided that the wavenumber is in units of rad m^{-1} . Thus, spectra of the rate of strain, such as $\partial u/\partial x$, and spectra of shear, such as $\partial u/\partial z$, have a $k^{1/3}$ form in the inertial subrange.

The non-dimensional shear spectrum derived from Nasmyth (1970) (Figures 5 and 6) shows that the effect of viscosity becomes significant at a non-dimensional wavenumber of $\sim 3 \times 10^{-2}$, where the spectrum has its peak. This seems to be a rather small wavenumber. However, if the wavenumber is expressed in units of rad m^{-1} , then the peak of the spectrum occurs at a non-dimensional wavenumber of 0.2 which is consistent with the notion that viscosity becomes important at scales comparable to, and smaller than, the Kolmogorov length scale, L_K .

An empirical equation for the non-dimensional shear spectrum of Nasmyth – devoid of any physics – is

$$\Psi = \frac{8.05\hat{k}^{1/3}}{1 + (20.6\hat{k})^{3.715}}, \quad (59)$$

where $\hat{k} = kL_K$, L_K is given by eq. (46), and k is in units of cpm . The dimensional Nasmyth spectrum (Figure 7), shifts upwards with increasing ϵ and its peak shifts to higher wavenumbers.

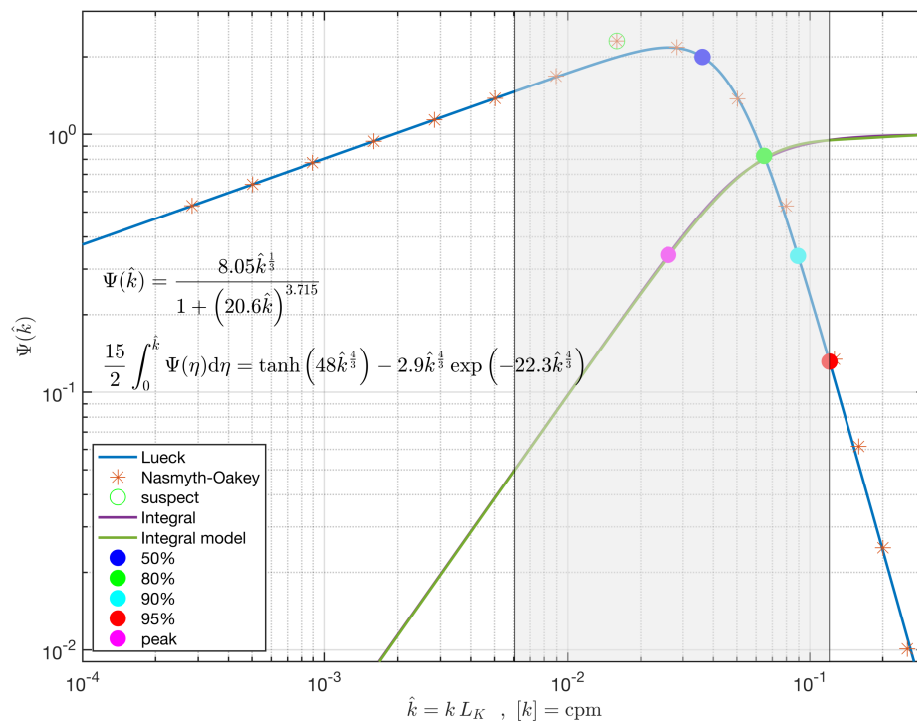


Figure 5: The non-dimensional wavenumber spectrum of velocity shear as tabulated in Oakey (1982b), where the wavenumber is expressed in units of cpm (orange asterisks). The blue line is an empirical mathematical fit, Ψ . The integral of the tabulated values (purple) is almost completely obscured by its empirical model (green). Important points are highlighted by coloured disks. The microstructure range is shaded in grey.

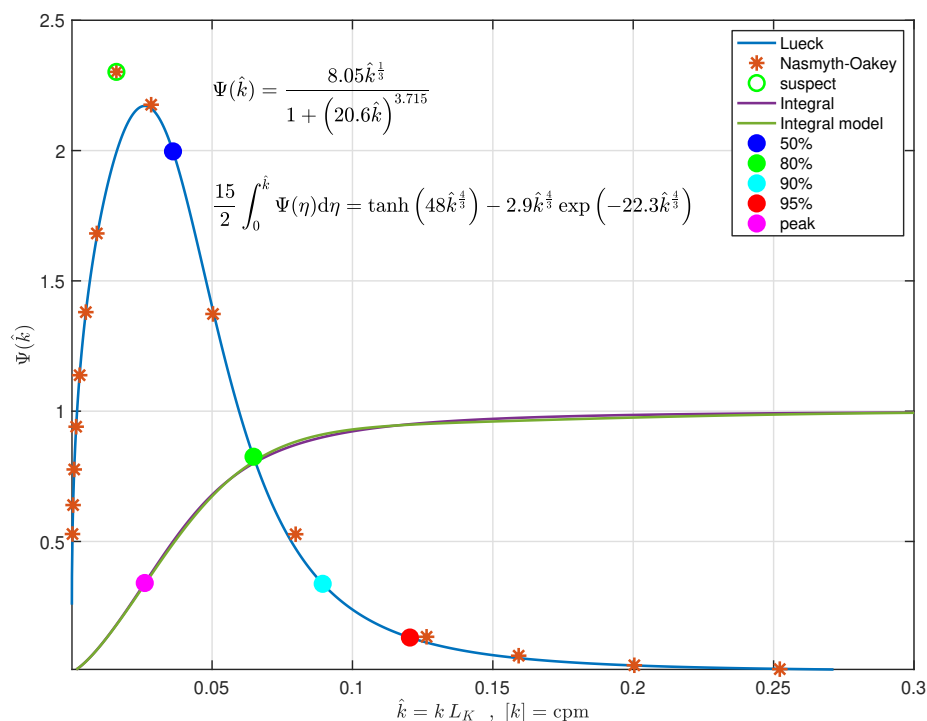


Figure 6: Same as Figure 5, but in linear space.

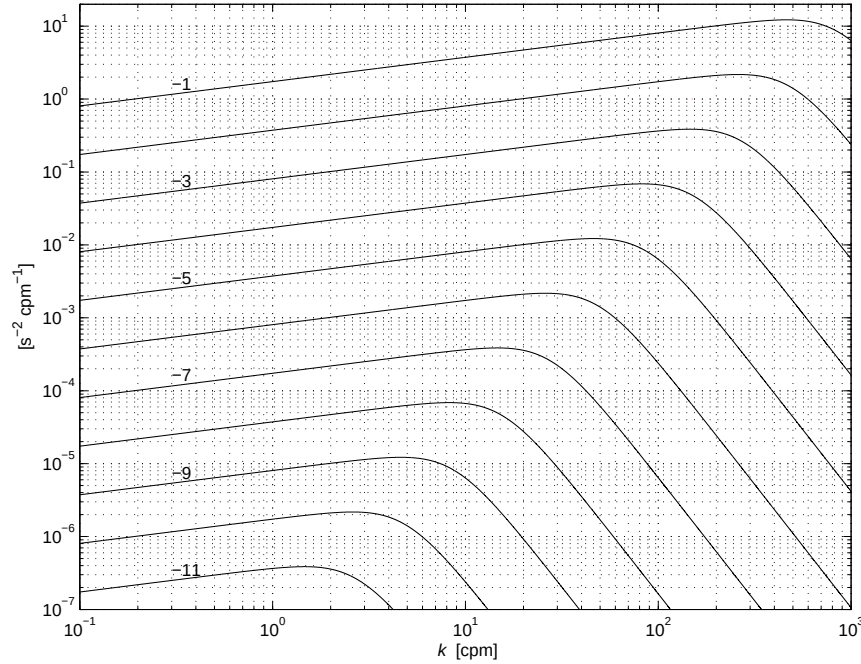


Figure 7: The dimensional Nasmyth spectrum for dissipation rates $10^{-11} < \epsilon < 10^{-1} \text{ W kg}^{-1}$.

Microstructure Range

The integral of the Nasmyth empirical spectrum (Figure 5, green line) shows that 90 % of all dissipation occurs in the wavenumber range from

$$6.1 \times 10^{-3} \leq \left(\frac{\nu^3}{\epsilon} \right)^{1/4} k \leq 0.12 , \quad (60)$$

which is shaded in grey in Figure 5. This is the *microstructure range* of velocity fluctuations. In the open ocean, the range is approximately 0.2 to 10 cpm for quiescent regions, such as the abyssal ocean, and it is 2 to 100 cpm in active regions such as the surface mixing layer (Table 3). In coastal regions, particularly in tidal channels, the rate of dissipation can be a thousand times larger yet, giving a microstructure range of 10 to 600 cpm.

Environment	ϵ [W kg ⁻¹]	range [cpm]
Abyssal Ocean	1×10^{-10}	0.2 – 10
Permanent Thermocline	1×10^{-8}	0.6 – 30
Seasonal Thermocline	1×10^{-7}	1 – 60
Surface Mixing Layer	1×10^{-6}	2 – 100
Tidal Channel	1×10^{-3}	10 – 600

Table 3: The microstructure wavenumber range for various rates of dissipation of kinetic energy, ϵ .

Spectral Estimates of the Dissipation Rate

The Nasmyth spectrum is a valuable tool for estimating the rate of dissipation because we can use a spectral measurement, $\Psi(k)$, to estimate ϵ by

$$\epsilon = \frac{15}{2} \nu \int_{k_0}^{k_1} \Psi(k) dk, \quad (61)$$

where k_0 and k_1 are the limits of spectral integration. Deriving the variance of the velocity shear (or its rate of strain) by spectral integration offers several advantages over calculating the variance in the spatial domain.

- The measured spectrum can be compared to the Nasmyth spectrum for quality assessments.
- Measurement noise, particularly at high wavenumbers, can be eliminated by a judicious choice of k_1 .
- Signal contamination by vibrations – an ever present problem – may be eliminated if it is spectrally narrow, or if it is coherent with accelerometer signals.
- k_1 can be non-dimensionalized to estimate the fraction of the variance that is resolved, by using its value as the coordinate in the integral of the Nasmyth spectrum (cyan and magenta curves in Figures 5 and 6).

Nearly all oceanic estimates of the rate of dissipation are made using some form of spectral integration to derive the variance of shear.

On a few occasions, the rate of dissipation has been estimated using space-domain processing. When the rate of dissipation is high and vibrational contamination resides at high wavenumbers (above $\approx 1/2$ of the Kolmogorov wavenumber), then it is possible to derive a “clean” signal by low-pass filtering the shear to retain only the wavenumbers that contribute significantly to viscous dissipation. This low-passed shear signal can then be squared and multiplied by $15\nu/2$ to produce an “instantaneous” dissipation rate. This space series becomes meaningful after it is averaged over some distance. Yamazaki and Lueck (1990) used this technique to explore the expected log-normal distribution of dissipation rates from shear data collected with a large submarine. Johnson et al. (1994) used this method to get dissipation rates right to the bottom in the boundary layer of the Mediterranean outflow. The data were collected with an expendable profiler and the dissipation rates were used to derive estimates of the bottom stress.

1.12 Spectra of Scalars

The general form of the scalar turbulent spectrum is illustrated in Figure 8 using the arguments summarized in Tennekes and Lumley (1972, section 8.6). In high Reynolds flows, when the largest turbulent overturns L are significantly larger than the Kolmogorov length scale, L_K , as defined in eq. (46), an inertial-convective subrange exists in the scalar spectrum over similar wavenumbers as the inertial subrange of the velocity spectrum. At large Prandtl number, such as those associated with heat and salt in water, viscous-convective and viscous-diffusive subranges develop, which extend to larger wavenumbers than the viscous subrange of velocity (Figure 8). The wavenumbers associated with the viscous subranges of the scalar spectrum are dictated by the Batchelor wavenumber, $k_B = L_B^{-1}$, where L_B is defined in eq. (47). This wavenumber depends on the dissipation of turbulent kinetic energy, ϵ , and molecular diffusion κ of the scalar, such that k_B for salt will be approximately 10x larger than the wavenumber for heat at a given

ϵ . It is thus theoretically more difficult to resolve the viscous-diffusive subranges for salinity than heat.

Spectra of temperature spatial gradients $\Phi_{\partial T/\partial z}$ for a range of ϵ are presented in Figure 9. In this form, a spectral peak appears at wavenumbers $k \approx 0.2k_B$ before the roll-off in the viscous-diffusive subrange. An increase in ϵ will shift the spectra along a -1 slope in log-log space, while increasing χ shifts the spectra upwards (see Figure 9). The viscous subranges, and thus the spectral peak becomes progressively more difficult to resolve when ϵ increases. Batchelor fitting exploits the presence of this spectral peak and its roll-off to estimate concurrently ϵ and χ (e.g., Luketina and Imberger, 2001; Merrifield et al., 2016), while Ruddick et al. (2000) integrated the viscous subranges of the scalar spectra to estimate χ prior to fitting Batchelor's theoretical model to obtain ϵ . Spectral fitting over the viscous subranges is impeded because the theoretical model, Kraichnan vs Batchelor, is debated at high wavenumbers, in addition to the universality of the constants used in these models (see Bogucki et al., 2012, for a review). Using controlled laboratory experiments, these authors demonstrated that the Kraichnan spectral form with a “universal” constant $q = 5.25$ agreed better with temperature observations than Batchelor's form. The use of the temperature gradient spectra to determine ϵ is also limited by the frequency response of the sensors which is discussed further in Section 3.2.

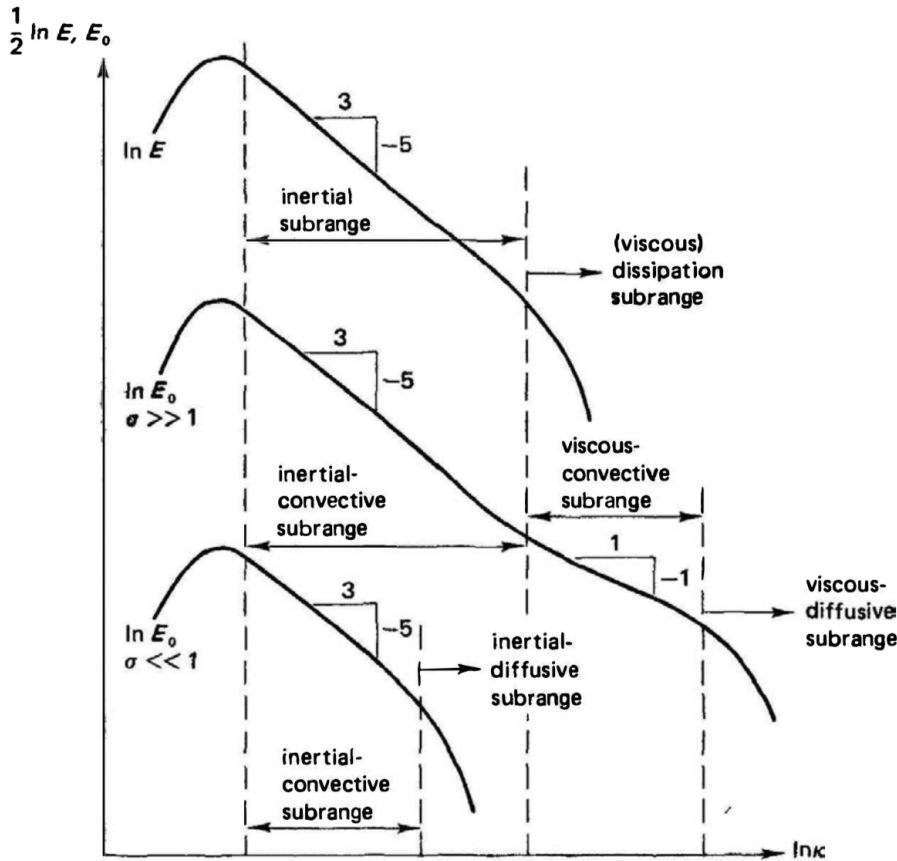


Figure 8: Spectral shapes for scalars E_o and velocity E according to Prandtl number $\sigma = \frac{\nu}{\kappa}$ where κ is molecular diffusivity of the scalar. Reproduced from Tennekes and Lumley (1972, Fig 8.11). Heat and salt in water both have large Prandtl number ($\sigma \approx 7$ for heat).

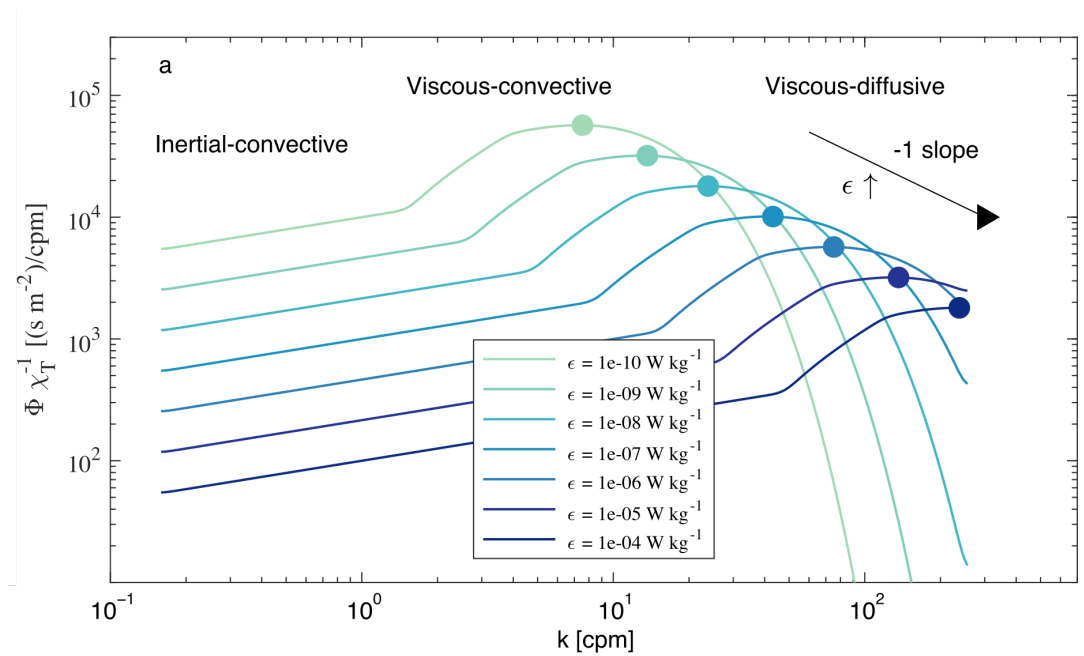


Figure 9: Kraichnan's model spectra for the temperature gradient $\partial T / \partial x_3$ spectra Φ for varying ϵ with the universal constant set to $q = 5.25$ (Bogucki et al., 2012). The circles represent the wavenumber where 25% of the variance (i.e., χ) is resolved. Note: Figure adapted from Bluteau et al. (2017).

2 Ocean Turbulence Measurement Techniques

2.1 Taylor Frozen Field Hypothesis

Turbulence measurements are typically made by mounting sensors either (1) on a profiler that moves through the water (e.g. free-fall instrument, glider, float), or (2) on a fixed platform where the mean current acts to pass fluid past the instrument. In both of these cases, temporal measurements are made, whereas turbulence properties are described in terms of spatial scales (Sections 1.5 and 1.7). Because turbulent eddies evolve slowly in time (Section 1.7), the properties of the flow are essentially “frozen” as the profiler passes through them (or the flow is advected past the instrumentation), provided that the measurement interval is short compared to the time-scale of evolution of the eddies⁴. This is the notion of “Taylor Frozen Field Hypothesis” which allows the time series of any turbulence quantity to be converted into a spatial sequence using

$$g(t) \longrightarrow g(z = Wt) , \quad (62)$$

where W and z are the speed and direction, respectively, of profiling (or of the mean current for fixed instrumentation). Any time rate-of-change of a turbulence quantity can then be converted into an along-path gradient using

$$\frac{\partial}{\partial t} = W \frac{\partial}{\partial z} . \quad (63)$$

The concept of Taylor’s Frozen Field Hypothesis is identical to the label “synoptic” that is frequently used in oceanography, where we assume that the field obtained with a series of stations is identical to the one that would have been obtained by an instantaneous measure of the field (e.g. a satellite image).

The necessity of using Taylor’s Frozen Field Hypothesis places a lower limit on the profiling speed of an instrument (or the speed of the mean flow for a fixed platform). The time-scale for the evolution of an eddy of size l by viscosity is $(l^2/\epsilon)^{\frac{1}{3}}$ (see Lueck et al. (2002) for an in depth discussion). A length scale of $O(1\text{ m})$ is sufficient for an estimate of the rate of dissipation. For typical dissipation rates in the ocean (Table 1), the time scale of the evolution of the eddies over a 1 m interval ranges from approximately 2000 s to 20 s for ϵ values of 10^{-10} and $10^{-4} \text{ W kg}^{-1}$, respectively. Therefore, for profiling speeds larger than $\sim 0.1 \text{ m s}^{-1}$ the turbulence is essential frozen.

There are a few instruments that inherently produce a space-series. For example, an acoustic Doppler current profiler produces velocity estimates that are uniformly spaced along its acoustic beams.

2.2 What can we measure?

There is a natural tendency to try to measure the Reynolds stress, $-\overline{uw}$, because it is the only turbulence term in the Reynolds equation for the mean flow. Knowing the Reynolds stress “closes” the problem and should lead us to a solution – whether analytic or numerical. Similarly, we would like to measure the vertical flux of scalars such as heat, $-\overline{wT'}$ and close the equation for the mean scalar. Unfortunately, measuring these fluxes is not practical in all but a few limited regions of the ocean.

The problems of directly measuring, say, the vertical flux of heat are numerous.

⁴More specifically, over the time interval of an estimate of the statistical properties of the turbulence, such as the rate of dissipation of turbulent kinetic energy, ϵ .

- We need a stable platform. This is only possible near a solid boundary, such as the ocean bottom, or under an ice sheet. In all other places the platform (or vehicle) carrying our sensors will be in motion and contribute to the fluctuations of u , w , and T' . Separating the sensor motions from the oceanic signal is challenging, to put it mildly.
- Most velocity measuring instruments – particularly those that use acoustic methods – tend to have a high noise level compared to the oceanic signals away from boundaries.
- Deriving a statistically reliable value takes a long time of sampling, because the flux containing eddies tend to be large (or comparable to the energy containing eddies).
- Internal wave motions can be large, yet they do not contribute to the turbulence flux. It can be extremely difficult to separate the wave signal from the turbulence.

Let us briefly look at the last item – contamination from internal waves. If we could hold rigidly a vertical velocity and temperature sensor in the ocean thermocline, then this pair of sensors would see a signal due to internal waves. Vertical undulations of the isotherms with an amplitude ζ will generate a temperature fluctuation of

$$T' \sim \zeta \frac{\partial T}{\partial z} . \quad (64)$$

The vertical velocity of this undulation is

$$w \sim \zeta N , \quad (65)$$

where the symbol $\sim$ indicates the same order of magnitude. Naturally, w and T' are exactly 90° out of phase and their co-product averages to zero. However, this quadrature signal is extremely large compared to the in-phase signal expected for turbulence. Let us take $\zeta = 10$ m for the internal wave amplitude, and $N = 3 \times 10^{-3} \text{ rad s}^{-1}$ for its frequency. In addition, we now know that the eddy diffusivity for the thermocline is $K_T \approx 1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$, and even the “Holy Grail” of Munk’s Abyssal Recipe (Munk, 1966) is only 10 times larger. Therefore, the false eddy diffusivity signal, $\hat{K}_T$, produced by internal waves is

$$\hat{K}_T = -\overline{wT'} / \frac{\partial T}{\partial z} \sim \zeta^2 N = 0.3 \text{ m}^2 \text{ s}^{-1} . \quad (66)$$

This quadrature signal is bigger than the expected in-phase signal by a factor $\sim 10^4$. Even minute errors in the *timing* or *phase* of the measurement of either signal will lead to an erroneous estimate of the vertical flux of heat! The phase shift required to turn the quadrature signal $\hat{K}_T$ into a co-signal K_T , is their ratio, $K_T/\hat{K}_T$. The timing error that could produce such a phase shift is

$$\delta t \sim \frac{1}{N} \frac{K_T}{\hat{K}_T} \sim 0.01 \text{ to } 0.1 \text{ s} , \quad (67)$$

depending on the choice of K_T . Platform motions will only add to this already hopeless situation.

In order to measure the Reynolds stress in a tidal channel, Lu and Lueck (1999a) and Lu and Lueck (1999b) found that one has to average the uw co-product over about 100 eddies to get a statistically reliable estimate. If the largest eddies are of order 10 m and the current is 1 m s^{-1} , then the estimates require $1 \times 10^3 \text{ s}$ – a considerable length of time. In the thermocline one would probably have to average over several buoyancy periods in order to achieve some measure of reliability, and this duration is also about 10^3 s .

If, on the other hand, we measure the velocity (or the scalar) field into the dissipation range, then the duration of measurements required for statistical reliability is much shorter than it is for measurements in the flux-containing range. The dissipation scale is almost always shorter than 0.01 m and a few hundred dissipating eddies are sampled in only 1 m of profiling, or within 1 to 10 s, depending on the speed of profiling. That is the good news.

The bad news is that measurements in the dissipation range make far greater demands on the characteristics of the sensors. The sensor must be able to resolve fluctuations with scales smaller than 0.01 m *and* be fast enough to detect them as the sensor profiles through the water. In addition, turbulence signals are spectrally red – they have a $k^{-5/3}$ form (which is also an $f^{-5/3}$, form due to profiling). In order to detect the dissipating eddies, the sensor must have very low noise compared to one that is used in the flux-containing range. These three challenges – spatial and temporal resolution and noise – have been surmounted by the shear probe and, consequently, oceanic turbulence measurements are predominately taken in the dissipation range.

2.3 Basic Constraints for Dissipation Measurements

Following Lueck et al. (2002) there are three elements required to take oceanic profiles of the dissipation rate:

- A sensor or probe that detects the physical parameter of interest, excluding (as much as possible) all other parameters;
- Electronic circuitry that amplifies and filters the signal produced by a probe, and records this signal for later analysis;
- A platform that moves the sensor smoothly through the ocean to produce a space series of the parameter of interest, i.e. a profile.

A wide variety of analytic and numerical models indicate that the average eddy diffusivity of the ocean is $K_T \sim 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, while four decades of observations indicate that the typical diffusivity is close to $\sim 1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ in regions away from boundaries. An absolute lower limit to the eddy diffusivity would be 10 times the molecular diffusivity, i.e. $10\kappa_T \sim 1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$, because turbulence cannot be of any importance if it does not substantially increase the diffusivity above this value. Using the model of Osborn (eq. (52)), we have

$$\epsilon = \frac{K_T N^2}{\Gamma} . \quad (68)$$

The buoyancy frequency squared ranges from $N^2 \sim 10^{-6} \text{ s}^{-2}$ in the deep abyssal ocean to $\sim 10^{-4} \text{ s}^{-2}$ in the seasonal thermocline (Table 1). Using a mixing efficiency of $O(0.1)$, and the rather pessimistic value of the $K_T = 10\kappa_T$, the minimum rate of dissipation that must be detected is then

$$\epsilon \sim \frac{10^{-6} 10^{-6}}{0.1} = 1 \times 10^{-11} \text{ W kg}^{-1} , \quad (69)$$

So, to detect weak turbulence present in the deep ocean, the velocity sensor must be able to provide a noise level below $10^{-10} \text{ W kg}^{-1}$. Using eq. (41), the root-mean-square (rms) velocity shear (or rate of strain) at the lowest rates of dissipation is

$$\left(\frac{\partial u}{\partial z} \right)_{\text{rms}} \lesssim \left(\frac{2}{15} \frac{\epsilon}{\nu} \right)^{1/2} = 3 \times 10^{-3} \text{ s}^{-1} . \quad (70)$$

The smallest eddy size, L_K (eq. (46)), will be ~ 0.01 m and, so, the rms velocity is $u_{\text{rms}} \sim 3 \times 10^{-5} \text{ m s}^{-1}$. There is no acoustic-type sensor that comes even remotely close to this level of noise.

The wavenumber range of microstructure (Table 3) sets explicit and implicit requirements for profiling:

1. The probes must have the spatial resolution to span the microstructure range. Basically, to resolve wavenumbers as large as $k = 100$ cpm, the sensor must be smaller than 0.01 m to provide measurements with little or no attenuation due to averaging of the signal over the surface of the sensor.
2. The platform carrying the sensor must move smoothly through the water so that the samples, which are invariably sampled regularly in time, can be converted unambiguously into a regularly sampled spatial series using Taylor's Frozen Field Hypothesis (Section 2.1). A platform which does not satisfy this requirement is a cable lowered CTD because its direction of travel will frequently reverse when waves get large.
3. Because all practical velocity sensors measure velocity *relative* to their platform, the platform must not vibrate or have other spurious motions in the microstructure range. Such motions would be added to the environmental signal measured by the probe and yield overestimates of the turbulence quantities.

The speed of profiling cannot be chosen arbitrarily. The *upper limit* is set by the frequency response of the probes. Environmental signals with a wavenumber of k (in units of cpm) will appear to the sensor at a frequency of $f = Uk$, where U is the speed of the platform relative to the ambient water. Sometimes this is different from the speed of a profiler in an earth-fixed reference frame (e.g. within large upwelling/downwelling currents). The speed of profiling must be chosen so that the apparent frequency of the signal falls within the frequency response of the probe. This requirement is often not achieved with temperature sensors. The *lower limit* for the profiling speed is set by the time-evolution of the turbulence itself (Section 2.1). The time scale of dissipating eddies is $(\nu/\epsilon)^{1/2}$ and the profiler must traverse over the largest scale of the microstructure range, $70(\nu^3/\epsilon)^{1/4}$, during an interval that is short compared to the time scale of the dissipating eddies. Thus, the lower limit of the speed of profiling is

$$U \gg \frac{70(\nu^3/\epsilon)^{1/4}}{(\nu/\epsilon)^{1/2}} = 70(\nu\epsilon)^{1/4} . \quad (71)$$

For the open ocean, where $\epsilon \sim 10^{-6} \text{ W kg}^{-1}$ is possible, the right hand side of eq. (71) is 0.07 m s^{-1} and most profilers satisfy the minimum-speed requirement by profiling faster than 0.3 m s^{-1} .

2.4 How do we get those gradients?

This is not a trivial question. There are at least two methods that we can use to convert a time series of a signal, say $u(t)$, into a gradient along the direction of profiling, say $\partial u(z)/\partial z$. The simplest approach is to take the *sampled* signal $u(n)$, taken at time $t = n\Delta t$, where $n = 1, 2, \dots, N$ is the index to the samples, and apply a scaled first-difference operation, such as

$$\frac{\partial u}{\partial z} = \frac{u(n) - u(n-1)}{W\Delta t} , \quad (72)$$

where W is the speed of profiling. This approach is rarely satisfactory because of unavoidable sampling noise. To understand why it is unsatisfactory, we have to take a close look at the act of sampling.

Data quantization

We are usually rather vague about the value of a sample and implicitly assume that it is identical to the value of the signal at the moment of sampling. This is unrealistic. A physical signal (temperature, pressure, velocity, etc.) is continuous in both time *and* value. For example, if the temperature is decreasing from 4 °C to 3 °C, then it will at some moment during that cooling have a temperature of π °C.

The sampled data must be represented by a finite number of digits or bits. This means that it is quantized into quantum levels. For example, if a system measures a signal, $s(t)$, over a range from s_l up to s_u , and the values are quantified by a B -bit number, then there are 2^B quantum levels for the samples, and the step size of the quantum levels is

$$\delta_s = \frac{s_u - s_l}{2^B - 1} \approx \frac{s_u - s_l}{2^B} . \quad (73)$$

The difference, $s_u - s_l$, is usually called the full-scale range of a measurement system. The number of steps in a staircase is one less than the number of levels and, if $B > 10$, the difference between these two numbers is inconsequential.

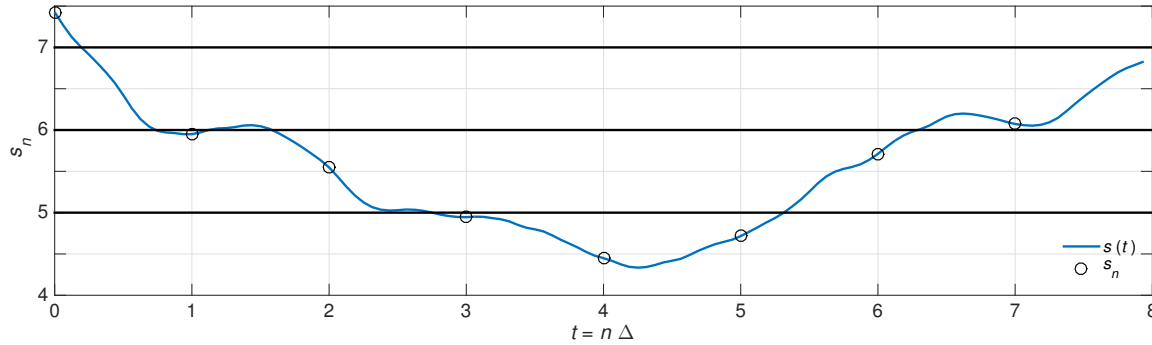


Figure 10: An example of the quantization of the samples, s_n , of a continuous signal $s(t)$. The ideal sampler assigns the nearest quantum level (black horizontal lines) to the samples.

The ideal sampler will assign the measured value to the nearest quantum level which would produce the data [7, 6, 6, 5, 4, 5, 6 and 6] for the example in Figure 10.

It is now quite common for the sampler (which is also called an analog-to-digital converter, or ADC) to provide 16-bit values. The step size (eq. (73)) is then about 1.5×10^{-5} of the full-scale range of the instrument. Some samplers provide up to 24-bit values.

Because the data are quantized, every sample is wrong but, ideally, by no more than $\pm\delta_s/2$. The samples are, effectively, the true value of the signal plus random noise. If the signal spans many quantum levels, then the error of the samples taken with an ideal sampler must be uniformly distributed over the range of $\pm\delta_s/2$. The probability density function of such a uniformly distributed random variable is

$$\begin{aligned} p(x) &= \frac{1}{\delta_s} \quad |x| \leq \delta_s/2 \\ &= 0 \quad |x| > \delta_s/2 . \end{aligned} \quad (74)$$

The mean error is the first moment of the probability density function

$$\mu = \int_{-\infty}^{\infty} \frac{1}{\delta_s} x dx = \frac{1}{\delta_s} \int_{-\delta_s/2}^{\delta_s/2} x dx = 0 , \quad (75)$$

and so the ideal sampler does not bias the data. However, the variance of the error

$$\sigma^2 = \int_{-\infty}^{\infty} \frac{1}{\delta_s} x^2 dx = \frac{1}{\delta_s} \int_{-\delta_s/2}^{\delta_s/2} x^2 dx = \frac{2}{\delta_s} \frac{1}{3} \left(\frac{\delta_s}{2} \right)^3 = \frac{\delta_s^2}{12}, \quad (76)$$

is not zero. Your data are the actual values of the signal *plus* uniformly distributed noise with a standard deviation of $\delta_s/\sqrt{12}$. This additional noise is called the *sampling noise*, or the *quantization noise*. Good samplers get within a factor of 2 of the ideal quantization. The sampling noise cannot have any particular frequency and, therefore, it must be white and uniformly distributed over the Nyquist band, which means that the spectrum of this noise is $\phi_{\delta_s} = \delta_s^2/(12f_N)$, where $f_N = f_s/2$ is the Nyquist frequency.

Your estimate of the spectrum of a signal is never smaller than the sampling noise. For example, the vibration spectra of Figure 11 flatten out to the sampling noise of $7 \times 10^{-4} \text{ Hz}^{-1}$, near the Nyquist frequency, even though the continuous domain signals are attenuated to a lower level by the anti-aliasing filters. For the example of Figure 11, the ideal sampling noise is

$$\phi_{\delta_s} = \frac{\delta_s^2}{12f_N} = \frac{1}{12 \times 256} = 3.3 \times 10^{-4} \text{ Hz}^{-1}, \quad (77)$$

which is only two times smaller than the actual sampling noise.

The sampling noise variance is independent of the rate of sampling, and the noise spectrum decreases inversely with increasing sampling rate. This is the main benefit of over sampling a signal.

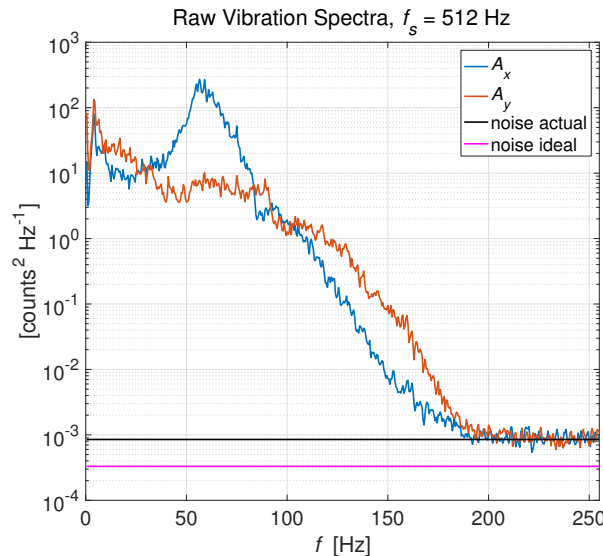


Figure 11: The spectra from two vibration sensors, A_x and A_y , that have not been converted into physical units. The signals were sampled at 512 s^{-1} . The anti-aliasing filter was a cascade of two 4-pole, low-pass, Butterworth filters with an effective cut-off frequency of 98 Hz.

The continuous-domain signal $u(t)$ itself has noise and it will add to the sampling noise. However, for many sensors and signals, the continuous-domain noise is many factors of 10 smaller than the sampling noise, even for 16-bit analog-to-digital converters.

Now, we return to the estimation of a gradient. Clearly, the first-difference approach (eq. (72))

is better represented by

$$\frac{\partial u}{\partial z} = \frac{u(n) - u(n-1)}{W\Delta t} + \frac{e(n) - e(n-1)}{W\Delta t}, \quad (78)$$

where e is the sampling noise. The standard deviation of the additional noise due to sampling is $\sqrt{2}\delta_s/(W\Delta t)$. The added noise can be very large when Δt is small, and this error is particularly acute when the signal is “red” – i.e., if it has most of its variance at low frequency and very little at high frequency. Turbulence, because of its $k^{-5/3}$ characteristic, is a perfect example of a red signal. The high-wavenumber (high-frequency) parts of the signal are readily lost because of sampling noise. Fortunately, there is a solution that does not require increasing the number of bits, B , of the data sampler.

Signal pre-emphasis

The alternative approach is to use *pre-emphasis* which is discussed fully in Mudge and Lueck (1994) and Rockland Technical Note 002. It is quite possible with analog (continuous-domain) electronics to form the time-derivative of a signal, or to add the time derivative of a signal to the signal itself, to form a new signal before it is sampled. For example

$$\begin{aligned} y(t) &= G_D \frac{d}{dt} x(t), \text{ or} \\ y(t) &= x(t) + G_D \frac{d}{dt} x(t), \end{aligned} \quad (79)$$

for any time-continuous signal x , where G_D is the gain of the differentiator and has units of s^{-1} . That is, one can obtain the derivative multiplied by any scalar value.

Turbulence fluctuations (and a great many other geophysical signals), have a red spectrum – one that decreases rapidly with respect to frequency. The act of time-differentiation boosts the spectrum of a signal in proportion to f^2 and effectively whitens its spectrum – a process frequently called pre-emphasis. This process raises the continuous-domain signal and its noise *above* the sampling noise, before the signal is sampled, so that it is not degraded by the sampling process itself. It is the sensor and its analog electronics that set the lowest signal that can be resolved and not the sampling noise of an analog-to-digital converter. Another way of looking at this is that pre-emphasis maximizes the environmental information that can be obtained from a sampling of a signal.

For the case of the shear probe, we usually take the first form of eq. (79) and use a gain of $G_D \approx 1$ s. The sampled signal is then proportional to the time derivative of the original signal, for example $\partial u/\partial t$, and, after dividing the sampled signal by the speed of profiling, we obtain a signal proportional to $\partial u/\partial z$, when profiling in the z -direction.

We are rarely interested in the original shear-probe measured velocity signal because the probe is an AC-sensor – the velocity is not measured for frequencies smaller than 0.1 Hz. However, for a scalar signal, such as temperature, we are interested in all frequencies, including the mean. For such signals we use the second form of eq. (79). To recover the original signal, x , we exploit the Fourier transform property of differentiation. If a signal $x(t)$ has a Fourier transform $X(f)$, then its derivative has the transform $2\pi j f X(f)$, where $j^2 = -1$. Thus, the Fourier-space representation of eq. (79) is

$$Y(f) = X(f) + G_D 2\pi j f X(f) = X(f) (1 + 2\pi j f G_D) \quad . \quad (80)$$

The transfer function of the pre-emphasis operation is then

$$\frac{Y(f)}{X(f)} = 1 + 2\pi j f G_D , \quad (81)$$

and we can recover a very high-resolution rendition of the original signal $x(t)$ by low-pass filtering the sampled data $y(n\Delta t)$ using a filter that has a transfer function of

$$H(f) = \frac{1}{1 + 2\pi j f G_D} , \quad (82)$$

which is a first-order low-pass Butterworth filter with a half-power response at $f_c = 1/(2\pi G_D)$. It is important to note that the pre-emphasis [eq. \(81\)](#) is conducted in the continuous domain, whereas the undoing of the pre-emphasis is achieved in the discrete domain. The characteristics of discrete-domain filters are very close to those of their counterparts in the continuous domain, but they are not identical.

The time-derivative can be obtained from this low-pass filtered signal using the first-difference technique without any contamination from sampling noise. Alternatively, the derivative can be obtained by high-pass filtering the sampled signal, to remove the original signal and leave only its derivative. Again, the cut-off frequency must be $(2\pi G_D)^{-1}$. This technique improves signal resolution by typically the equivalent of 7-bits, that is, by a factor of 100. Spectral noise, which is proportional to the square of the extra resolution, is reduced by a factor of 10^4 (Mudge and Lueck, 1994). An example of the application of pre-emphasis is shown in [Figure 12](#).

2.5 Goodman coherent noise removal technique

Vibrations from the instrument are picked up by the shear probes and introduce noise to the data. We use a methodology described in (Goodman et al., 2006) to remove vehicular contamination of shear probe measurements. We assume that shear probe signals are linearly related to real environmental turbulence, plus a contribution measured by the accelerometers:

$$\mathbf{s} = \hat{\mathbf{s}} + B_{ik} * a_k , \quad (83)$$

where $\mathbf{s}$ is the time series of measured shear at each shear probe, $\hat{\mathbf{s}}$ is the uncontaminated real environmental signal at each shear probe, a_k is the signal reported by the k^{th} accelerometer, and B_{ij} is the multivariate weighting function that represents the contribution from each acceleration signal into each shear-probe signal. The convolved acceleration represents the portion of the acceleration signal that is coherent with the shear signal. It is important to remember that the actual acceleration at each shear probes is not (and cannot be) measured. The process described by [eq. \(83\)](#) only assumes that the measured accelerations are linearly related to the actual accelerations of the probes. At any given frequency, the vibrations of the probes are proportional to the measured accelerations and these may be phase shifted. The proportionality factors and the phase shifts can be frequency dependent, but these frequency dependences must be time-invariant. The time indices in [eq. \(83\)](#) are implicit and the added noise of the measured shears has been ignored for simplicity.

By taking the Fourier transform of each signal, we can describe [eq. \(83\)](#) as

$$\phi_i = \hat{\phi}_i + \beta_{ik} \alpha_k , \quad (84)$$

where ϕ_i , $\hat{\phi}_i$, β_{ik} , and α_k are the Fourier transforms of s_i , $\hat{s}_i$, B_{ik} , and a_k , respectively. The frequency indices are implicit. If we multiply each side of [eq. \(83\)](#) by its complex conjugate

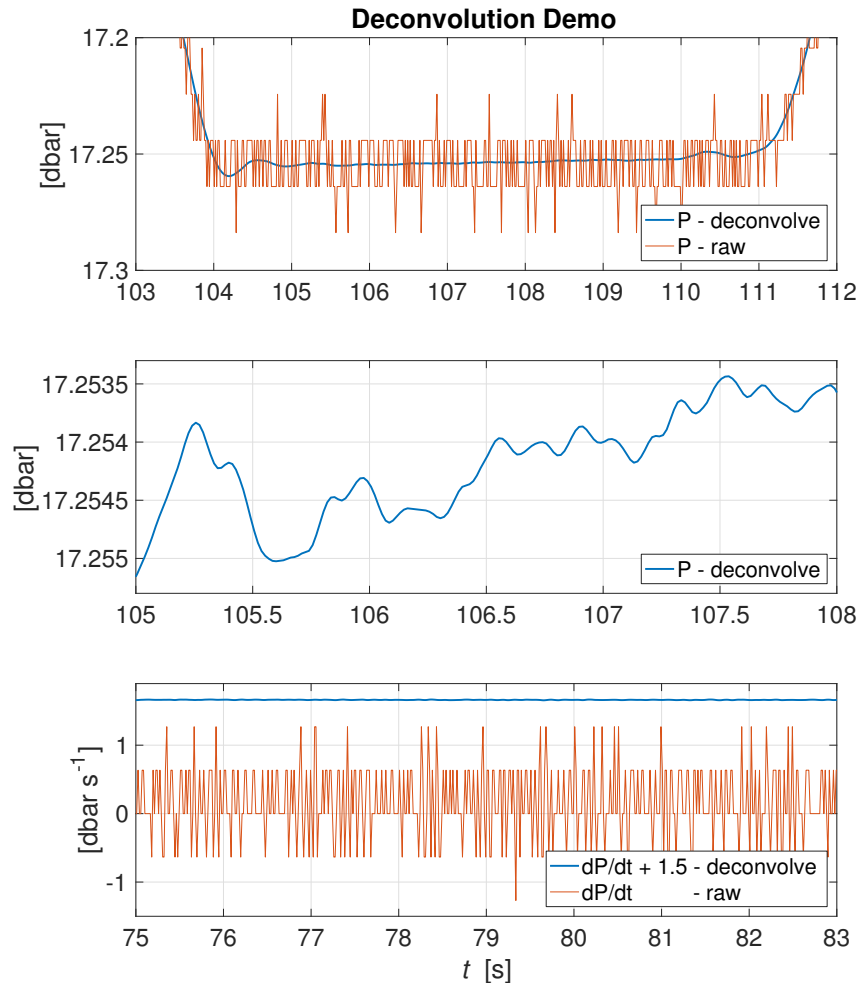


Figure 12: Upper panel: Pressure data from a profiler that has just impacted the soft bottom of a lake. The transducer has a full-scale range of 500 dbar. The red line is a standard pressure signal, recorded without the benefit of pre-emphasis. The blue line is the pressure signal recovered from the pre-emphasized signal using $G_D = 20$ s. No smoothing has been applied to either signal. Middle panel: Same as above but with a highly zoomed-in view. Bottom panel: The rate of change of pressure during free-fall, which is used to determine the speed of profiling. No smoothing has been applied to either signal. However, the enhanced rate-of-change of pressure signal has been shifted up by 1.5 dbar for visual clarity.

and take the average of each term, then each term is proportional to its spectrum, and we can correct the measured shear spectra using

$$\hat{\Phi}_{ij} = \Phi_{ij} - \chi_{ik} \Gamma_{kl}^{-1} \chi_{lj}^* , \quad (85)$$

where $\hat{\Phi}_{ij}$ is the corrected cross-spectrum of $\hat{s}_i$, Φ_{ij} is the cross-spectrum of the original contaminated signal, χ_{ij} is the cross-spectrum of the contaminated shear signal and accelerometer output, and Γ_{kl} is the cross spectrum of acceleration. In addition, we must assume that the accelerations are independent of environmental turbulence, such that $\overline{\hat{s}_j a_j} = 0$. Turbulent eddies with length scales longer than the length of the profiler will accelerate the profiler and violate this assumption. Eddies with length scales shorter than the profiler will, through averaging their effect over the length of the profiler, contribute progressively less to the measured accelerations as their length scale decreases. Vibrations induced by drag on the profiler (such as vortex shedding), and other self-generated accelerations, are usually unrelated to the environmental turbulence and meet the assumption that $\overline{\hat{s}_j a_j} = 0$.

The terminology of “cross-spectra of shear” may seem strange if you are not familiar with multivariate signal processing. The cross-spectrum of the measured shear, $\Phi_{ij}(f)$, is a three-dimensional matrix with a size of $N_P \times N_P \times N_f$, where N_P is the number of shear probes, and N_f is the number of frequency indices in the spectra. If you have two shear probes then, for each frequency index, Φ_{ij} is two-dimensional, where Φ_{11} and Φ_{22} are the real auto-spectra of probes 1 and 2, respectively, Φ_{12} is the complex cross-spectrum of probes 1 and 2, and Φ_{21} is the cross-spectrum of probes 2 and 1 and equal to the complex conjugate of Φ_{12} .

The Goodman multivariate coherent vibration removal algorithm is computationally efficient because it uses all signals simultaneously and, thereby, avoids redundant Fourier transform calculations. It also takes into account that partial coherency between the acceleration signals themselves. In weakly turbulent environments, where the relative contamination may be large, the correction can be significant.

A simple univariate example

Let the shear in the ocean be represented by $\hat{s}(t)$, the measured shear by $s(t)$, and the added noise of the measured shear by $e_s(t)$. Then, the measured shear can be represented by

$$s(t) = \hat{s}(t) + h(t) * a(t) + e_s(t) \quad (86)$$

where $a(t)$ is a simultaneous measure of the vibration that is linearly related to the vibration of the shear probe in the direction of its sensitivity, and $h(t)$ is a time-invariant weighting-function that characterizes how the vibration, $a(t)$, are related to the measured signal $s(t)$.

The frequency domain equivalent of eq. (86) is

$$S(f) = \hat{S}(f) + H(f) A(f) + E_s(f) , \quad (87)$$

where the upper-case variables are the Fourier transforms of their lower-case counterparts. If we multiply the left- and right-sides of eq. (87) by their complex conjugates we have

$$\begin{aligned}
SS^* &= \left(\hat{S} + HA + E_s \right) \left(\hat{S} + HA + E_s \right)^* , \\
&= \hat{S}\hat{S}^* + HH^* AA^* + E_s E_s^* \\
&\quad + \hat{S}H^* A^* + \left(\hat{S}H^* A^* \right)^* \\
&\quad + \hat{S}E_s^* + \left(\hat{S}E_s^* \right)^* \\
&\quad + E_s^* HA + (E_s^* HA)^* ,
\end{aligned} \tag{88}$$

where the frequency dependence is not explicitly shown for clarity. We can be quite certain about the expectations of the terms involving the noise in the measurement of shear, namely

$$\begin{aligned}
\hat{S}^* E_s &\rightarrow 0 \\
E_s^* HA &\rightarrow 0
\end{aligned} \tag{89}$$

because the noise in the electronic circuitry supporting the shear probe should have no relationship to the oceanic shear nor to the vibration of the instrument. It is also likely that the vibration of the instrument, a , is not related to the oceanic shear, $\hat{s}$, for length-scales of the shear that are short compared to the length of the vehicle that carries the shear probes (see the discussion above). Thus, for the range of frequency over which the oceanic shear does not induce vibrations, we have the expectation that

$$\hat{S}H^* A^* \rightarrow 0 . \tag{90}$$

Thus, only the first three terms on the right hand side of [eq. \(88\)](#) are non-zero. The expectation of the measured shear can then be expressed by the spectrum, C , of each term, namely

$$C_{ss}(f) = C_{\hat{s}\hat{s}}(f) + |H|^2 C_{aa}(f) + C_{e_s e_s}(f) , \tag{91}$$

where $C_{ss}(f)$, $C_{\hat{s}\hat{s}}(f)$, $C_{aa}(f)$, and $C_{e_s e_s}(f)$ are the auto-spectra of the measured shear, the oceanic shear, the measured vibrations, and the added noise, respectively. An estimate of the hypothesized, but unknown, transfer function, $H(f)$, is obtained using

$$\hat{H}(f) = \frac{C_{sa}(f)}{C_{aa}(f)} . \tag{92}$$

$\hat{H}(f)$ is an estimate of the transfer function and, like all estimates of a transfer function, it is limited by the caveats of this technique. First, there can be no added noise to the vibration signal. Its auto-spectrum is in the denominator of [eq. \(92\)](#) and the transfer function is biased low by the added noise of a vibration measurement. Zero-noise measurements are impossible. This is one reason that Rockland uses piezo-accelerometers to measure vibrations. They have an extremely high signal-to-noise ratio ($\sim 10^6$ and higher) and provide an estimate with very little bias. The shear and vibration signal are ‘statistical’ and sufficient data must be used to get a statistically reliable estimate of the transfer function.

The best estimate of the oceanic shear spectrum, one without vibration-coherent contamination, is obtained by using [eq. \(91\)](#). That is,

$$C_{\hat{s}\hat{s}}(f) = C_{ss}(f) - \left| \hat{H}(f) \right|^2 C_{aa}(f) - C_{e_s e_s}(f) . \tag{93}$$

The noise of the measured shear, $C_{e_s e_s}(f)$, is usually small but it too can be estimated by collecting data with test (dummy) probes installed and scaling the resulting spectrum to convert it into physical units, with the identical scaling factor that is used to convert the measured shear into physical units.

Equation (93) is the univariate version of eq. (85) with the addition of the noise in the measured shear, which was ignored in the multivariate equation.

3 Ocean Turbulence Sensors

3.1 Shear Probes

The most commonly used and best-suited sensor for measuring microstructure velocity fluctuations is the airfoil (shear) probe. This probe was initially conceived by H.S. Ribner and T. Siddon at the University of Toronto and was developed for atmospheric and wind-tunnel measurements (Siddon, 1965) and (Siddon and Ribner, 1965). The probe was subsequently modified for use in water (Siddon, 1971) and, in 1972, Tom Osborn made the first successful deployment of the shear probe on a microstructure profiler in a fjord on the west coast of Canada (Osborn, 1974).

Osborn and Crawford (1980) describe the theory of operation and calibration of the shear probe. The sensing element is a piezo-ceramic bimorph (two-layer) beam which generates an electric charge in response to bending forces (Figure 13).

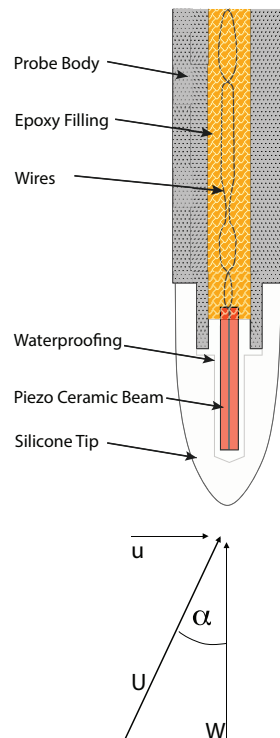


Figure 13: A sketch of the shear probes. The speed of vertical profiling is W , the horizontal velocity fluctuation is u , the total velocity is U , and the angle of attack is α . The length of the sensing tip is 8.5 mm.

For the shear probe made by Rockland, the bimorph is about 13 mm long, 1.6 mm wide and 0.5 mm thick. About one-half of the beam is solidly anchored into its supporting stainless steel sting, and the other half is cantilevered out of the sting and will generate a charge if it is bent. The bending under a load is in the direction normal to the wide surface of the beam – much like a diving board at the side of a swimming pool bends in the vertical direction but not in the horizontal direction. A water blocking layer is placed over the surface of the ceramic, and the entire tip is covered with rubber to form an air-foil surface of revolution – much like the shape of a bullet.

A piezo-ceramic is inherently an AC-sensor. It responds only to fluctuations and produces no mean signal. The resistance of the beam must exceed $1 \times 10^{10} \Omega$ in order for the shear probe to respond properly at the lowest apparent frequencies of the microstructure range (0.1 Hz). Any leakage of moisture on to the surface electrodes of the ceramic beam degrades the low-frequency response. More than 10 years were spent developing a long-term water-proofing method, without which, it would not be possible to use the shear probe for long-term deployments on autonomous vehicles, such as gliders, and on Apex floats. The longest test conducted so far is 3 months at a pressure of 1000 dbar. All probes maintained a resistance greater than $5 \times 10^{10} \Omega$.

Why is water-proofing so important? The capacitance, C , of a bimorph is typically 1 nF. If there is a leakage resistance of R ohms across the bimorph, then the signal produced by the bimorph relaxes (disappears) with a time-constant $\tau = RC$. The resistance and capacitance form a high-pass filter that attenuates signals with frequencies smaller than $1/(2\pi\tau)$. We want an unattenuated response down to ~ 0.1 Hz which is equivalent to $\tau \gtrsim 1$ s. Therefore, we must have $R > 1 \times 10^9 \Omega$.

Several versions of the shear probe are in use by scientists around the world. They differ mainly in the length of the tip measured relative to the point of cantilever and the diameter at the point of cantilever. The original probe by Osborn is 9 mm long and 5 mm wide. The probe developed and used by N. Oakey at the Bedford Institute of Oceanography, and also used by M. Gregg at the University of Washington, is 14 mm long and 5 mm wide. The probe developed by J. Moum at Oregon State University is 25 mm long and 6 mm wide (see Moum et al. (1995), their figures A2 and B2, for more details). The shear probe is now commercially available from Rockland Scientific. Its length and width are 8.5 mm and 4.7 mm, respectively (Figures 14 and 15).



Figure 14: The FP07 thermistor (top), the shear probe (centre), and the SBE7 micro-conductivity sensor (bottom). The copper disk is a Canadian penny and has a diameter of 19 mm.

Calibration procedure and probe sensitivity

The shear probe is calibrated using a vertical jet of water after the method of Osborn and Crawford (1980). Think of the vertical jet as being represented by the vector U shown in Figure 13. The angle of attack, α , is then varied from -10° to $+10^\circ$ in steps of 2° . Because the shear probe has no steady output, it is rotated around its longitudinal axis at a rate of 1 revolution per

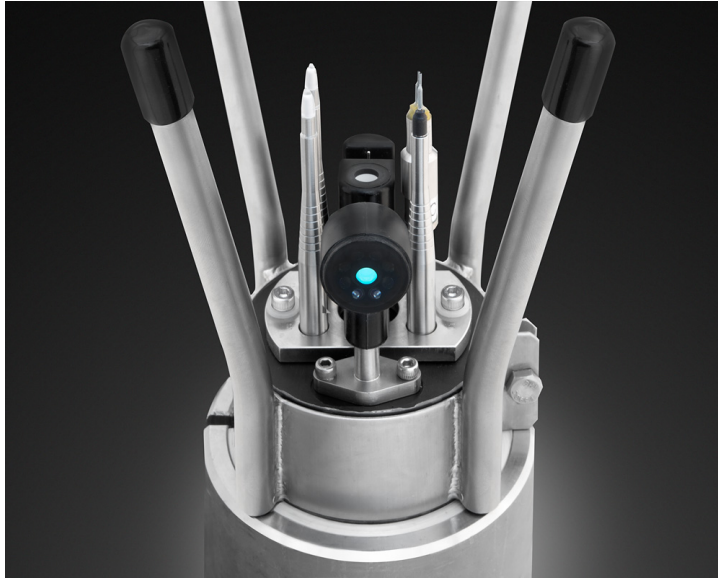


Figure 15: A microCTD with two shear probes, one FP07 thermistor, a micro-conductivity sensor, a reference CT and a fluorometer.

second to produce a sinusoidal signal with a frequency of 1 Hz and an amplitude corresponding to a cross-stream velocity $u = U \sin \alpha$. The velocity of the jet along the longitudinal axis of the shear probe is $W = U \cos \alpha$. By the theory of aerodynamic lift over slender bodies of axial symmetry, the lift force is given by

$$F \propto \frac{1}{2} \rho U^2 \sin 2\alpha = \rho (U \cos \alpha) (U \sin \alpha) = \rho W u . \quad (94)$$

The charge produced by the piezo-ceramic element in the shear probe is proportional to the force of [eq. \(94\)](#). The electronics transfer this charge to a capacitor, to produce a voltage. In the calibration procedure, we measure the rms voltage (rather than the peak voltage), and fit it linearly to $U^2 \sin 2\alpha$, to derive the probe sensitivity, S , so that

$$E_{rms} = S U^2 \sin 2\alpha . \quad (95)$$

The calibration is done at room temperature in fresh water, $\rho \approx 1000 \text{ kg m}^{-3}$, and typical values of the sensitivity are 0.05 to $0.15 \text{ V m}^{-2} \text{ s}^{-2}$. Nobody makes adjustments for the higher density of seawater because the calibration is only good to about 5%. Notice that the instantaneous voltage (the voltage that you would use to derive the instantaneous cross-stream velocity) is given by

$$E_P = \sqrt{2} S U^2 \sin 2\alpha = 2\sqrt{2} S U \cos \alpha U \sin \alpha = 2\sqrt{2} S W u . \quad (96)$$

Thus, the shear probe output is linear with respect to the cross-stream velocity u and its output increases with increasing speed of profiling W – the faster you profile, the larger the signal. The factor of $\sqrt{2}$ is an artifact of using the rms voltage for calibration and the factor of 2 comes from the $\sin 2\alpha$ identity. For any direction of profiling, there are always two orthogonal, or cross-stream, directions. A single shear probe can only detect one of these cross-stream velocity components. However, two shear probes with their longitudinal axes in parallel can detect both components when one probe is rotated by 90° around its axis. In isotropic turbulence, the two cross-stream velocity components are statistically independent. Therefore, it is common to mount two shear probes on a profiler, for redundancy and for increased statistical confidence.

Frequency and spatial response of the probes

The shear probe has a frequency response that is determined by the mechanical stiffness of the probe tip and its mass. Vibration tests indicate that the frequency response is several kilo-hertz and, therefore, not an issue at the usual speeds of profiling. However, its finite size does induce spatial averaging which limits the wavenumber response of the shear probe. Macoun and Lueck (2004) indicate that the response is in the form of a first-order low-pass filter with a half-power wavenumber of 48 cpm. That is, the measured spectrum is reduced by the factor

$$H(k) = \frac{1}{1 + (k/48)^2} . \quad (97)$$

Temperature effects on the shear probe

The shear probe responds to temperature in two ways: (1) its sensitivity, S , changes with temperature, and (2) it has a “pyro-electric” response.

The response of the shear probe to cross-stream velocity fluctuations depends on temperature because the amount of charge released (or absorbed) by the piezo-ceramic beam, per unit of strain, is slightly temperature dependent. The so-called g_{31} coefficient relates the amount of charge produced per unit of strain and its value usually decreases by $\sim 10^{-3} \text{ }^\circ\text{C}^{-1}$. However, the mechanical components surrounding the ceramic beam also have temperature-dependent properties. The silicon rubber, the Teflon sheath, and the epoxy all become less stiff with increasing temperature. The net result is that the sensitivity of the shear probe increases with increasing temperature. At least, that was the commonly held view until recently.

Tests conducted at Rockland in 2015 on eight shear probes indicate that there is a spread of temperature coefficients. Most samples had a positive temperature coefficient of about $0.5 \% \text{ }^\circ\text{C}^{-1}$, some had a negligible temperature dependence, and one had a *negative* coefficient of $-0.5 \% \text{ }^\circ\text{C}^{-1}$. Further testing is needed. The shear probes are usually calibrated at room temperature ($\sim 20^\circ\text{C}$) and if they are used in an environment of much different temperature, for example in arctic waters, then it may be wise to calibrate them at the target operating temperature.

The pyro-electric effect is a response to a sudden change of temperature of a ceramic. A high rate-of-change of temperature will cause a charge to be liberated (or absorbed) by the ceramic, even when there is no strain applied to the beam. The ceramic is fairly well insulated from the environment by the inner layer of epoxy, the Teflon shroud and by the silicon rubber. It does not experience rapid temperature changes in most oceanographic environments. Because of the insulation surrounding the beam, the pyro-electric effect manifests itself as a very low frequency ($\ll 0.1 \text{ Hz}$) response. The front-end charge-transfer amplifier of the shear probe is usually set to high-pass signals above 0.1 Hz and, consequently, the pyro-electric effect is not noticeable in almost all application. However, when a researcher (Wolk and Lueck, 2001) attempted to extend the low-frequency response of the charge-transfer amplifier to 0.01 Hz , the pyro-electric effect became noticeable.

3.2 FP07 Thermistors

A thermistor is a metal-oxide resistor that has a large temperature-coefficient of resistance. The coefficient is usually negative. The most commonly used microstructure temperature sensor is the FP07⁵ thermistor (Figures 14 – 16). It consists of a metal-oxide bead of 200 μm diameter, covered with glass to a thickness of 25 μm . This bead is fused on to a larger glass bulb of 2 mm diameter which forms a stem of 10 mm length. The end of the stem is epoxied into a stainless-steel sting.

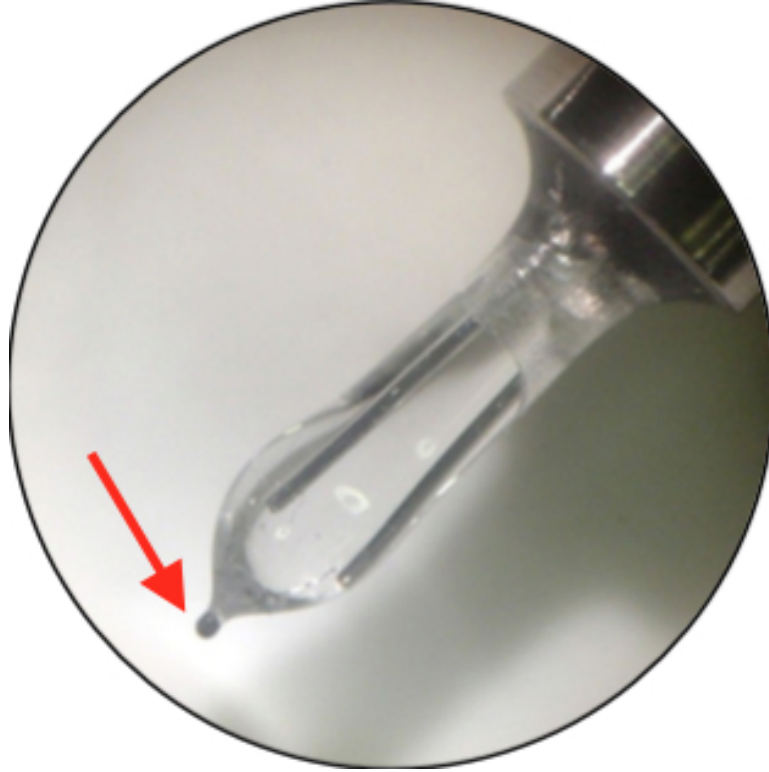


Figure 16: A microscope image of the tip of the FP07 thermistor probe made by Rockland Scientific. The small sensing tip is indicated by the red arrow. The water proofing material at the stainless-steel shaft has been removed.

The FP07 is the smallest and, consequently, the fastest thermistor-type temperature sensor that is commercially available. The FP07 is also popular because thermistors have a high temperature coefficient of resistance,

$$\alpha = \frac{1}{R_T} \frac{dR_T}{dT} \approx -0.04^\circ\text{C}^{-1}, \quad (98)$$

where R_T is the resistance of the thermistor. They are also available in a wide range of nominal resistance.

Frequency and spatial response of the thermistors

The linear dimensions of the thermistor are small enough for microstructure measurements – i.e. it has adequate spatial resolution to resolve typical Batchelor length scales of 0.1 mm to 1 mm (see Table 1). However, the frequency response of the FP07 is poorly known and reports

⁵Made by Amphenol-Thermometrics, USA.

have given ambiguous results with respect to both the time-constant of the sensor response and whether the response is single- or double-poled (Goto et al., 2016; Sommer et al., 2013b; Nash et al., 1999; Hill, 1987; Vachon and Lueck, 1984). These ambiguities result partly because these sensors are “hand made” and hence differ in their physical dimensions, but primarily because calibrating their frequency response is difficult and no commercial facility exists for this service.

By profiling slowly, e.g. $0.05 - 0.1 \text{ ms}^{-1}$, it is possible to resolve the higher wavenumbers (Sherman and Davis, 1995) because, by Taylor’s Hypothesis (Section 2.1), $k = Uf$, where U is the profiling speed. This is shown graphically in Figure 17, where the temperature gradient spectra (originally shown in Figure 9) are plotted as a function of frequency at two different profiling speeds. While a reduction in the profiling speed improves the ability to resolve the higher wavenumbers, it also restricts the number of profiles and thus the amount of sampling achievable. The lack of sensor resolution is also especially difficult for high-energy turbulent events.

Two possible corrections for the thermistors are illustrated in Figure 17b. The first technique proposed by Vachon and Lueck (1984) is less aggressive and speed-dependent. It takes the form of a double-pole transfer function, i.e. :

$$H_{VL}(f) = \frac{1}{[1 + (2\pi\tau f)^2]^2} \quad (99)$$

with a time constant $\tau = \tau_o (U/U_o)^{-1/2}$ where $\tau_o = 4.1 \times 10^{-3} \text{ s}$, $U_o = 1.0 \text{ ms}^{-1}$ and U is the speed of the flow past the sensor (i.e. profiling speed). The second technique proposed by Sommer et al. (2013b), denoted as H_S , is more aggressive and independent of U i.e., $\tau = 1 \times 10^{-2} \text{ s}$. At typical profiling speeds of approximately 0.75 ms^{-1} ($\bar{u}_3$ in Figure 17b), the thermistor’s response correction with either technique is more than a factor of 3 at the peak of the viscous-diffusive subrange for $\epsilon \approx 10^{-7} \text{ W kg}^{-1}$ (Figure 17b), thus the FP07 cannot resolve all existing scales in energetic environments.

3.3 Micro-conductivity sensors

The electrical conductivity of seawater depends on temperature, salinity, and pressure. It increases with respect to all three of its dependent parameters, but the pressure dependence is small. The conductivity-coefficients for seawater,

$$\frac{\partial C}{\partial S} \approx 0.1 \text{ S/m/PSU}, \quad \frac{\partial C}{\partial T} \approx 0.1 \text{ S m}^{-1} \text{ }^\circ\text{C}^{-1}, \quad (100)$$

are numerically comparable for salinity and temperature. The most commonly used sensor for the measurement of conductivity fluctuations in the microstructure range is the two-electrode SBE7 sensor⁶ (Figures 14 and 18).

The wavenumber response of the SBE7 is only partially documented (Hill and Woods, 1988) and somewhat controversial (Sommer et al., 2013a), but it appears that it is better than 100 cpm. The conductivity sensor is usually considered a purely spatially averaging sensor, with properly designed electronics, but Sommer et al. (2013a) found evidence for a minor time-dependent response.

In the open ocean, conductivity fluctuations result almost entirely from temperature fluctuations, because the background gradient of salinity is small compared to the large-scale gradient

⁶Manufactured by Sea-Bird Electronics, USA.

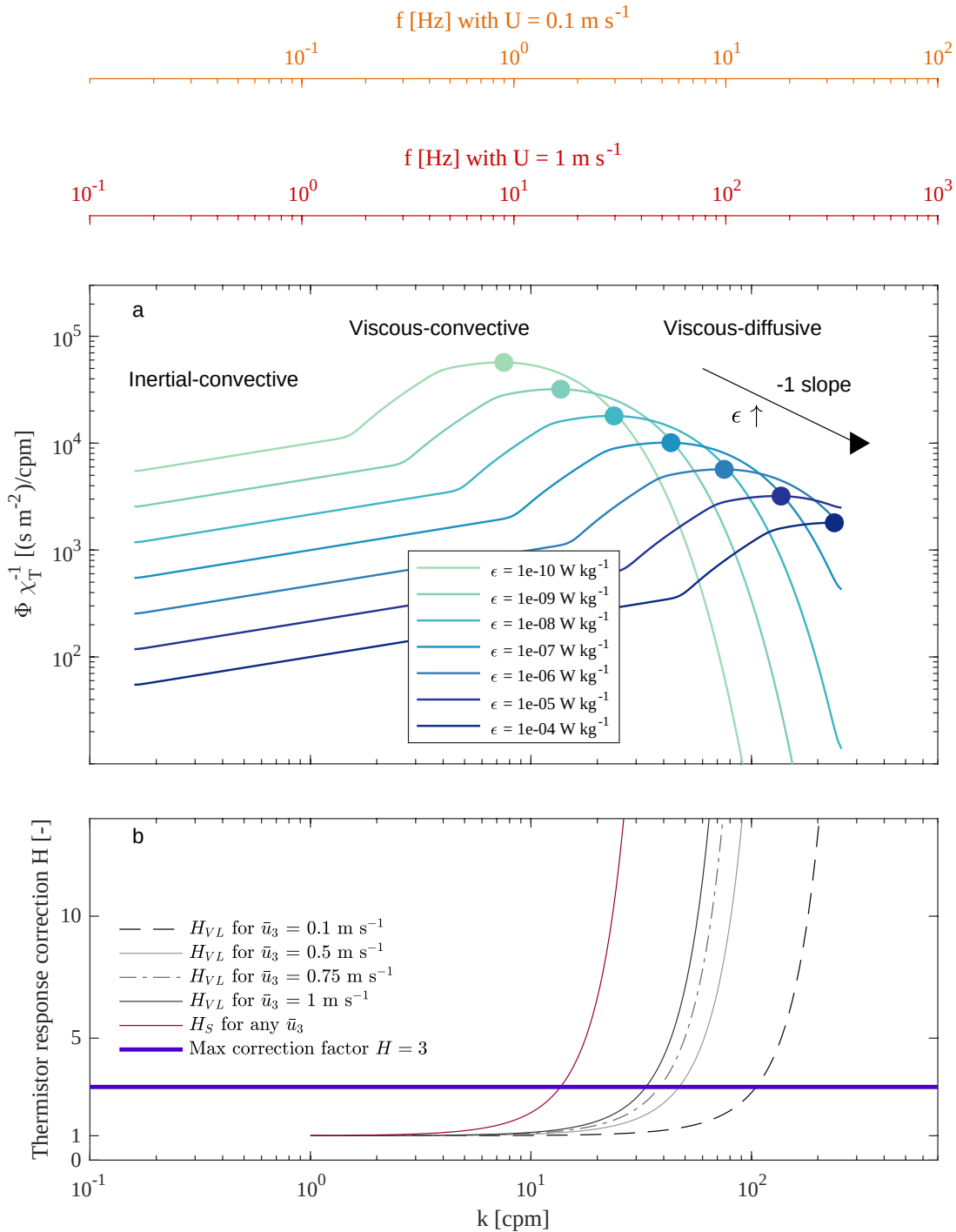


Figure 17: (a) Kraichnan's model spectra for the temperature gradient $\partial T/\partial x_3$ spectra Φ for varying ϵ with the universal constant set to $q = 5.25$ (Bogucki et al., 2012). The circles represent the wavenumber where 25% of the variance (i.e., χ) is resolved. The frequencies corresponding to the wavenumbers are illustrated above for slow and fast profiling speeds, U . (b) The thermistor's response correction H_{VL} uses the definition of Vachon and Lueck (1984), which is dependent on the mean flow speed past the thermistor $\bar{u}_3$, while H_S is the aggressive frequency correction of Sommer et al. (2013b) that is independent of $\bar{u}_3$ (vertical profiling velocity). Note: Figure adapted from Bluteau et al. (2017). Both U and $\bar{u}_3$ are used to represent the speed of the flow past the sensor.

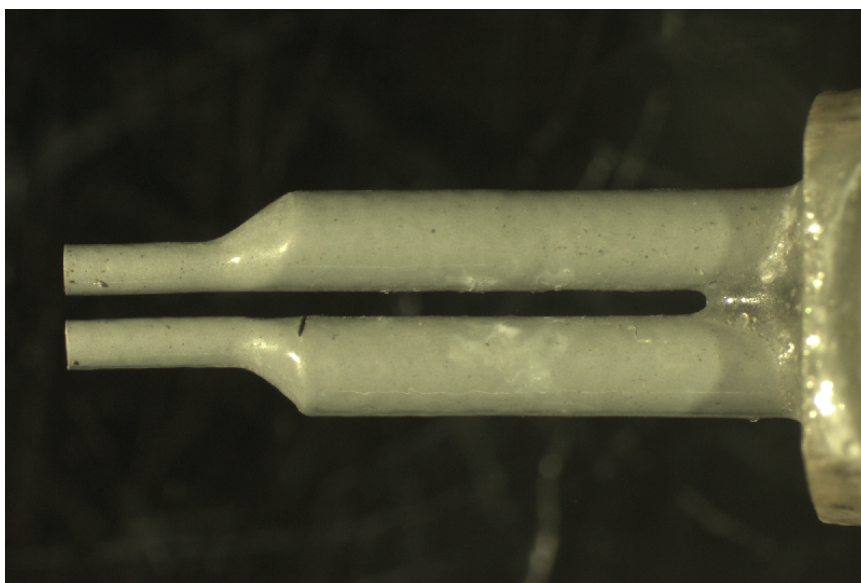


Figure 18: A microscope image of the tip of the SBE7 micro-conductivity sensor. The two needles are separated by 1 mm centre-to-centre. The electrodes are on the front face and are not visible in this image.

of temperature. Thus, the micro-conductivity sensor is mostly a substitute for a fast thermometer. In coastal regions that may be substantially influenced by fresh water from rivers, the conductivity fluctuations may be mostly due to salinity fluctuations. In tropical lakes, where the salinity is small ($S < 6$ PSU), conductivity fluctuations are almost entirely due to salinity and, consequently density.

4 Introduction to Vertical Turbulence Profilers

4.1 Basic Mechanics of a Vertical Profiler

A profiler falls because it has a *net* weight in water. Its weight is

$$mg = \rho_I g V , \quad (101)$$

where m is its mass, ρ_I is its density, V is its non-flooding volume, and g is the gravitational acceleration. The buoyancy of the instrument is

$$B = \rho g V , \quad (102)$$

where ρ is the density of the water. The net weight of the profiler is $mg - B$. The instrument falls if $\rho_I > \rho$ and rises if the reverse is true. In both cases the centre of buoyancy must be above the centre of mass for the instrument to be statically stable. The centre of buoyancy is at the centre of volume if the fluid density is uniform over the length of the profiler.

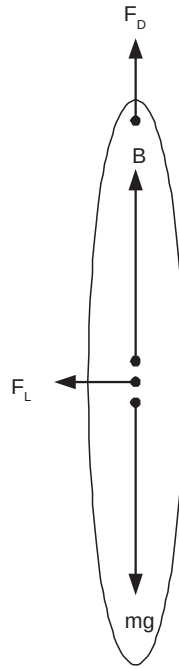


Figure 19: The basic forces on a vertical profiler – weight mg , buoyancy, B , drag, F_D and lift, F_L .

Once the instrument starts to move, there are the additional forces of drag, F_D and lift, F_L . A well designed profiler will have no lift when its longitudinal axis is vertical. The drag on the instrument is

$$F_D = \frac{1}{2} \rho C_D A W^2 = C W^2 , \quad (103)$$

where C_D is the drag coefficient ($\sim 0.1 - 1$), A is the cross-sectional area, W is the velocity through the water (and essential the fall rate), and all coefficients can be lumped in to a single dimensional coefficient of drag, C , if they are constant. Even though every conceivable shape of instrument has some drag, it is good practice to add a significant amount of additional drag so that the fall rate is independent of water density. This is usually done with brushes so that there are no large coherent eddies on the lee side of the instrument that can perturb its motion.

Ignoring lift, the steady force balance is

$$-\rho_I gV + \rho gV + CW^2 = gV(\rho - \rho_I) + CW^2 = 0 . \quad (104)$$

The additional drag due to an increased C then permits the density of the instrument to be significantly larger than the density of the water, so that the density difference, $\rho - \rho_I$, is less dependent of the natural increase of ρ with depth. The result is a fall-rate, W , that is fairly independent of depth. A typical profiler density is $\rho_I \approx 1150 \text{ kg m}^{-3}$. Using two to three drag brushes gives a fall rate of 0.5 m s^{-1} . A VMP-200, that was specifically modified for use in tidal channels, by an additional stainless-steel shell of 3 kg attached near the bottom, falls at 1.4 m s^{-1} with a single brush (Figure 20). Faster profiling is an advantage in a highly turbulent environment. Some poorly designed instruments have stalled their descent (hung up) due to the increasing density of seawater with depth, $\rho = \rho(z)$. That is, they stop a depth where $\rho - \rho_I = 0$. If it is tetherless and has no time-out on its ballast-release mechanism, the profiler may stay at that depth forever.



Figure 20: A VMP-250 modified for use in swift tidal channels by the addition of a weight collar.

The complications

For shallow work, it is highly desirable to have a tether attached to the vertical profiler, so that it can be retrieved quickly for subsequent profiling. The tether will add drag that increases in proportion to the length of tether deployed. The tether can also add weight or buoyancy to the instrument depending on the density of the tether relative to that of the water. Most synthetic lines have a density close to that of water and, at considerable cost, they can be modified to match a specified density.

The amount of line that must be deployed does not depend solely on the depth of profiling. The tether must be kept slack so that the ship's motions do not transmit to the profiler (Figure 21). The vertical shear of horizontal current will displace the ship from the profiler. So will windage. The result is that the length of deployed tether is always greater than the depth

of the profiler, and the length of tether required to reach a target depth is very site specific. The deepest tethered profiling has been to 1200 m. For deeper work it is common to use an untethered profiler - an instrument that drops ballast weights so that it rises back to the surface using its own buoyancy. Of course, this slows down the profiling process because time has to be spent to find, retrieve, and re-arm the profiler for its next deployment.

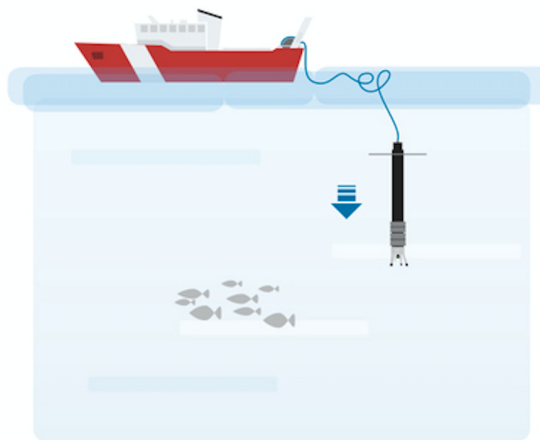


Figure 21: A cartoon of a vertical profiler deployed from a ship. The tether is kept intentionally slack so that the motion of the ship is decoupled from the motion of the profiler. Two or three turns of loose tether in the water near the surface is sufficient for decoupling.

4.2 History

The very first ocean turbulence measurements were taken with a horizontal profiler – a minesweeping paravane converted to a towed vehicle – in a tidal channel by Grant et al. (1962) in the late 1950s (Figure 22). However, using a towed vehicle in the open ocean is difficult and the heated anemometers used by Grant et al are easily contaminated by temperature fluctuations. Additional problems, in those days, were data recording and data communication.

The first vertical profiles of ocean turbulence were conducted by the research group of Charles Cox at the Scripps Institute of Oceanography in the mid 1960. These profilers carried only temperature sensors (usually thermistors) because a suitable velocity sensor (such as the shear probe) had not yet been developed. Data were recorded on audio tape recorders by converting each signal into a sinusoidal voltage with a frequency in the audio band, a technique known as frequency modulation. After the profiler is recovered the audio tape is played back and the audio signal is passed through a ‘frequency discriminator’ (FM-demodulator) to recover the very low-frequency (including DC) signal so that it could be sampled by an analog-to-digital converter that was typically the size of a small house. Some of these profilers descended to full ocean depth (Gregg, 1977).

The development of the airfoil probe by Siddon and Ribner (Siddon, 1971; Siddon and Ribner, 1965), and its subsequent adaption for ocean use by Osborn (Osborn, 1974) in 1972 opened the flood-gates to velocity turbulence measurements in the ocean. Within a decade, a large number of different vertical profilers were developed by research groups in Canada (Figure 23) and the USA (Figure 24). This was followed by profilers developed in Europe (Figure 25) and Japan (Figure 26). The first commercially produced profiler was the Fly (Figure 23, bottom right), but it did not see extensive production. The second commercially produced profiler

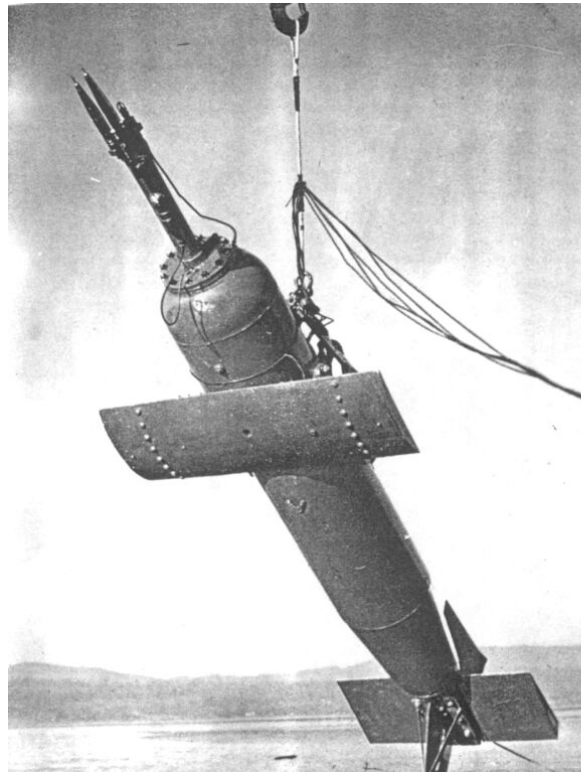


Figure 22: The mine-sweeping paravane used by Grant et al. (1962). Effectively, the first ocean turbulence profiler.

was the TurboMAP (Figure 26, right) and many units have been produced, but it is now out of production. The Micro-scale profiler (Figure 26, left), never functioned properly. The descendants of the MSS series of profilers (Figure 25, second from right) are still commercially available. These, and other profilers, are discussed in Lueck et al. (2002).

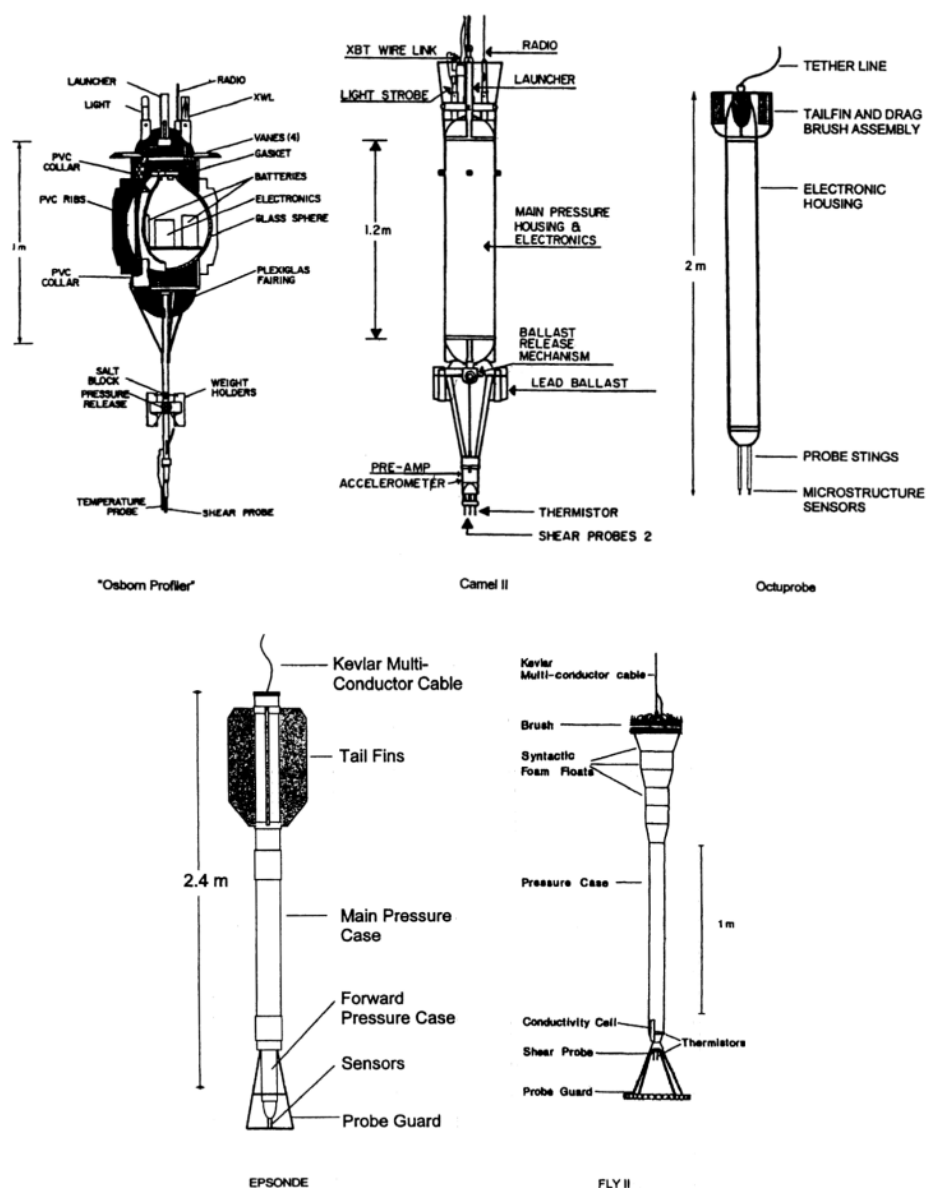


Figure 23: Sketches of the vertical microstructure profilers developed in Canada between 1972 and 1985. From left to right – the first profiler that carried the shear probe (Osborn, 1974) – the subsequent series of “Camel” profilers by Osborn, University of British Columbia (Crawford and Osborn, 1980; Crawford, 1980; Lueck and Osborn, 1982; Lueck et al., 1983; Moum and Osborn, 1986) – the Oakey profilers Octuprobe and Epsonde, Bedford Institute of Oceanography, Dartmouth, (Oakey, 1977, 1988) – the FLY profiler developed at the Institute of Oceanography in British Columbia, Sidney (Dewey et al., 1987).

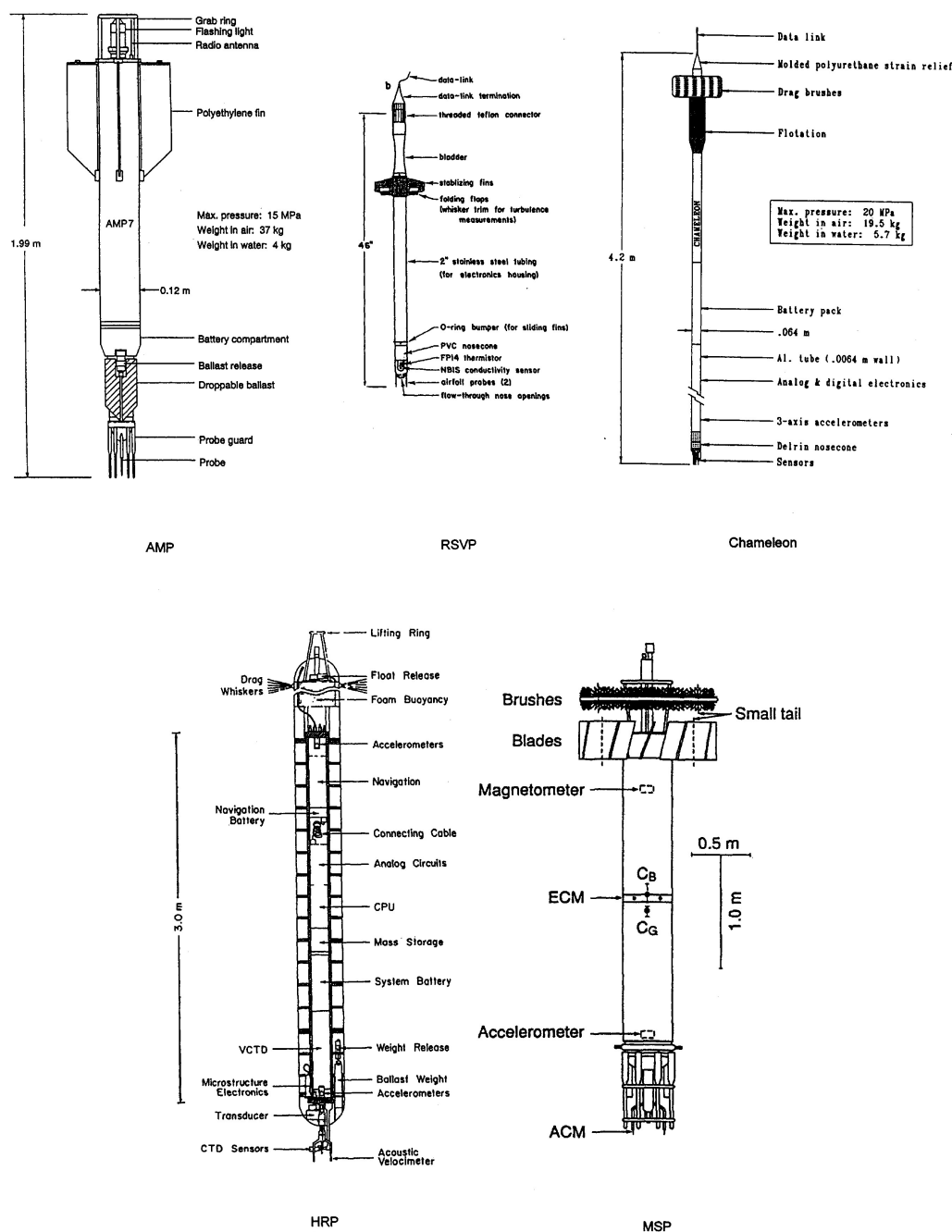


Figure 24: Sketches of vertical microstructure profilers developed in the USA in the 1980s. From left to right – AMP, by Gregg, Applied Physics Laboratory, University of Washington, Seattle, WA (Gregg et al., 1982) – the Rapid-Sampling Vertical Profiler, by Caldwell, Oregon State University (Caldwell et al., 1985) – Chameleon by Caldwell and Moun, OSU (Moun et al., 1995) – the XDP (not shown) (Lueck and Osborn, 1985; Johnson et al., 1994) – High Resolution Profiler, by Schmitt and Toole, Woods Hole Oceanographic Institute, (Schmitt et al., 1988; Hayes et al., 1984) – MSP, by Gregg and Sanford, APL, UW (Winkel et al., 1996).

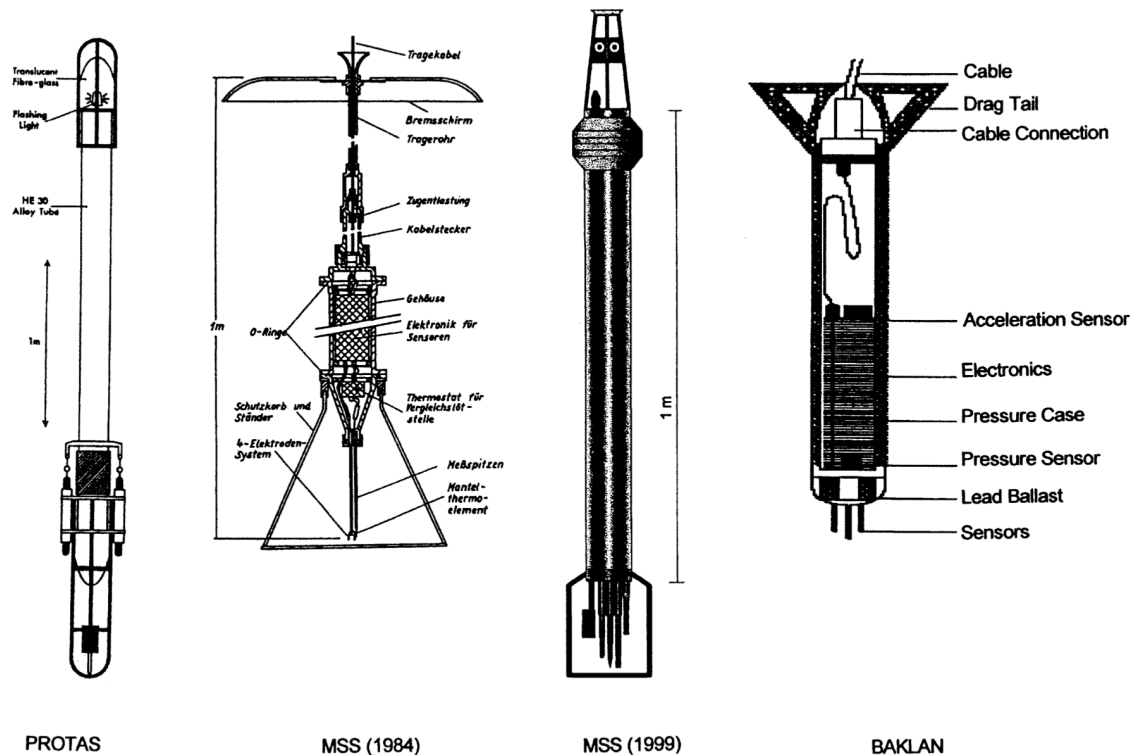


Figure 25: Sketches of vertical microstructure profilers developed in Europe between 1972 and 1999. From left to right – PROTAS by Simpson, University of Wales, UK, (Simpson, 1972) did not carry shear probes but was the first profiler with a long and slender shape – the MikroStrukturSonde by H. Prandke, Institute for Marine Research, Warnemuende, Germany (Prandke et al., 1985; Stips, 1992; Prandke et al., 2000) – the BAKLAN developed by Paka, Shirshov Institute of Oceanology, Kaliningrad, Russia (Arvan et al., 1985; Paka et al., 1999).

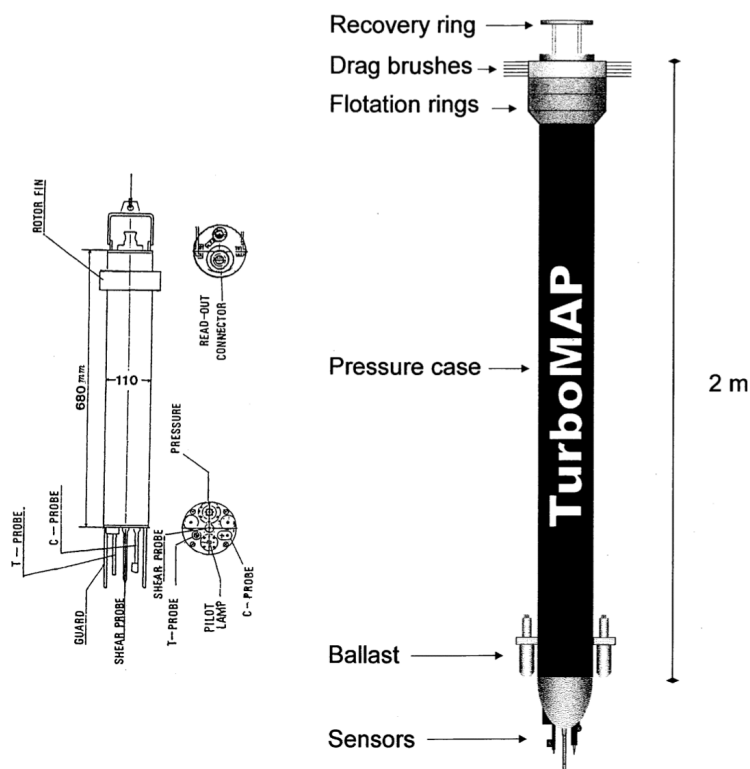


Figure 26: Sketches of two vertical microstructure profilers that were commercially developed in Japan in the 1990s by Alec Electronics in Kobe. From left to right – the Micro-Scale Profiler for Kanari at the University of Hokkaido (Kanari, 1991) – the TurboMAP produced for H. Yamazaki at the Tokyo University of Marine Sciences (Wolk et al., 2002).

5 The VMP-250

5.1 Overview

The VMP-250 is a coastal-zone, vertical profiler for the measurement of turbulence microstructure. It is designed for operation from small vessels with limited deck space (e.g. Zodiacs), or where electrical power facilities are limited or missing (e.g. ice camps). It is light in weight (~ 110 N), simple to deploy (Figure 28) and rated to a safe depth of 500 m, with optional upgrade to 1000 m.



Figure 27: The VMP-250.



Figure 28: An internally recording VMP-250 about to be deployed from a small vessel, by tossing it overboard.

The VMP-250 supports up to two shear probes, two FP07 thermistors, one micro-conductivity sensor, and a compact, high-precision conductivity-temperature (CT) sensor (Figure 29). Optional external sensors include a chlorophyll-turbidity (CLTU) sensor and a dissolved oxygen sensor. Internally, the instrument has a precision tilt sensor, two 1-axis vibration sensors, and a pressure transducer. See the instrument manual for the detailed specifications of these sensors.

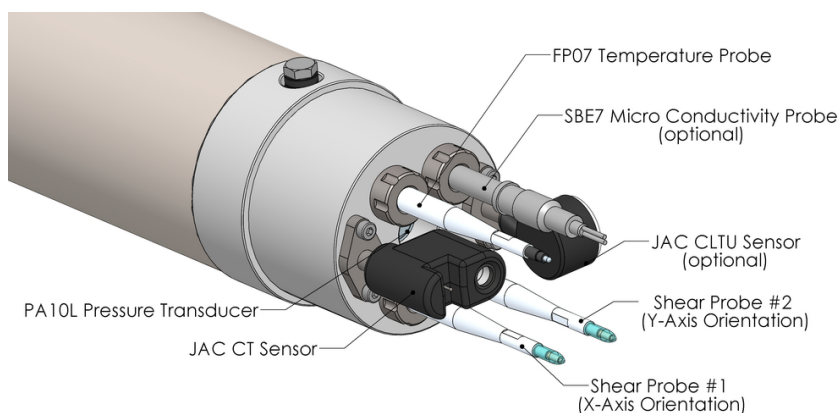


Figure 29: The sensors that are mounted to the front bulkhead of the VMP-250.

The x -direction of the VMP-250 is directed through the ON/OFF hole on the VMP-250, and through the pressure port on the VMP-6000. The y -direction is to “port” and the z -direction is vertical, positive pointing up.

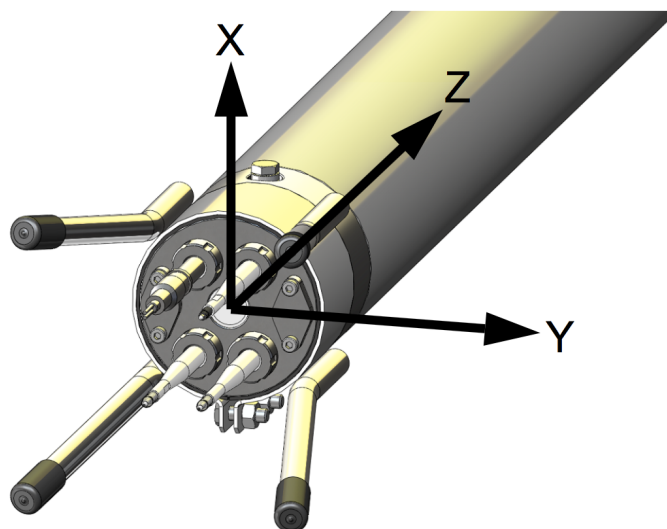


Figure 30: The coordinate system for the VMP-250.

The VMP-250 is available with an internal data logger and, in this configuration, it requires only a line for recovery. It is also available with real-time telemetry, which requires an electro-mechanical cable and a winch. All Rockland instruments create identical data files and the Rockland-supplied data processing software can be used with any instrument.

5.2 Installation and Removal of Probes

You should always install test sensors when storing or shipping your instrument, and only install real probes when you are about to use the instrument. The test sensors have internal resistors that bring the sensor channel readings to approximately their mid scale. For this reason, the test probes are stamped with their sensor type: T1, T2, S1, S2, and C1. The name on the test probe should match the name on the SMC cable.



Refer to the instrument User Manual for detailed steps for probe installation.

Important things to keep in mind when installing and/or removing probes are:

- When you withdraw a test probe from its holder, it will be attached to a cable that you should pull out only far enough to access the connector at the end of the test probe.
- When you disconnect a probe from the cable, be careful that the cable does not accidentally slip inside the probe holder.
- The mating connector on the cable identifies the input connector on the internal electronics (for example T1).
- When tightening the cable connectors to the probes, *hand-tightness* is sufficient.
- When you push the probe into its holder, you will feel (and hear) it being stopped by metal-to-metal contact.
- The probe holder uses an O-ring and a ferrule to make a seal that is good to extremely high pressures ([Figure 31](#)). If you reverse the position of the O-ring and the ferrule or over tighten the probe holder, the seal will be useless at any pressure.
- Orient the shear probes to the x - and y -axis of the instrument ([Figure 30](#)), to match the orientation of the accelerometers. Note: For the shear probe, the direction of sensitivity is perpendicular to the flat cut-out surface, which identifies the probe's serial number ([Figure 32](#)). Correctly mounted, the shear probes will be orthogonal to each other ([Figure 33](#)).
- Thermistors and micro-conductivity probes do not need alignment.

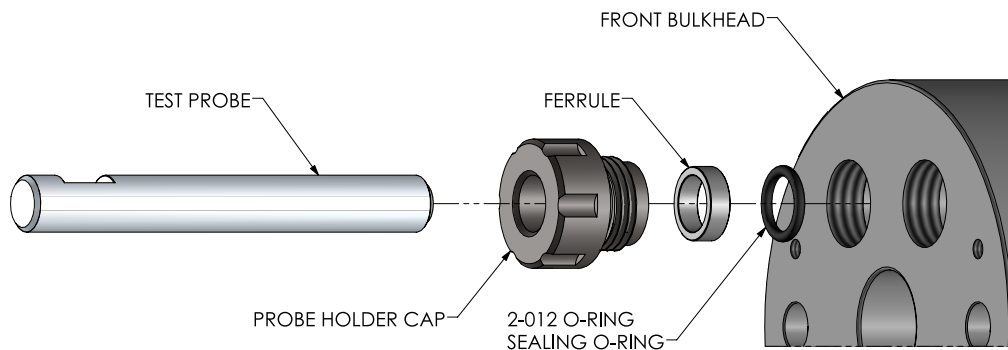


Figure 31: VMP-250 holder for shear probes, FP07 thermistors, and micro-conductivity sensors.

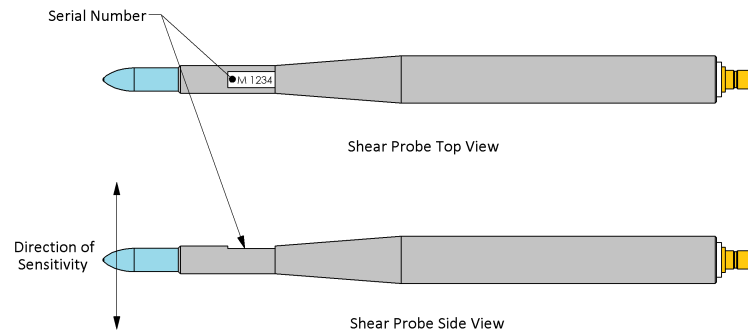


Figure 32: Shear probe direction of sensitivity.

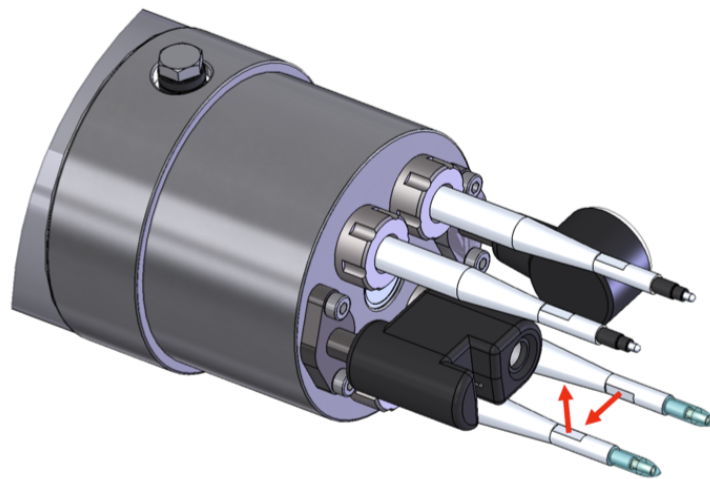


Figure 33: Shear probe alignment.

5.3 Power Consumption

The VMP-250 is powered internally by a 1.8 Ah, Polymer Lithium-ion rechargeable battery with a nominal voltage of 14.8 V. A fully charged Li-ion battery has a voltage of 16.8 V and the battery itself will turn off if its voltage falls below 12.6 V, but this should never happen during data acquisition. Instead, when the voltage drops to about 13 V (i.e. remaining charge is less than 5%), the instrument will shut down the data acquisition process, close all data files, and shutdown the power supply board.

At room temperature, the battery has a capacity of 26.6 Wh (i.e. $14.8 \text{ V} \times 1.8 \text{ Ah}$). A VMP-250 has a base power consumption of 1.7 W providing approximate 12 hrs of operation time. The power consumption increases with additional sensors and the operation time is reduced. [Table 4](#) provides details of power consumption with each additional sensor.

Table 4: Estimated power draw and expected battery life for various configurations. Note: “Standard Sensors” include two shear probes, one FP07 thermistor and the CT. The optional chlorophyll-turbidity and dissolved oxygen sensors are labelled as CLTU and DO, respectively.

Configuration	Power Draw [W]	Battery Life [hours]
Standard Sensors (i.e. Shear, FP07, CT)	1.7	12
Standard Sensors + SBE7	2.3	9
Standard Sensors + CLTU	2.3	9
Standard Sensors + SBE7 + CLTU	2.9	7
Standard Sensors + SBE7 + CLTU + DO	3.3	6

5.4 Data Storage

Data acquired by the instrument are recorded internally on a CompactFlash card. The rate of data recorded is approximately:

$$\text{Data Rate [bytes/s]} = \text{columns in address matrix} \times 2 \text{ [bytes/column]} \times \text{Sampling Rate [Hz]}$$

For the VMP-250-IR, with a typical address matrix of 8 columns and a sampling rate of 512 Hz, this works out to just greater than 8 kB/s. In this configuration, a 4 GB CompactFlash card would allow for approximately 100 hours of recorded data.

5.5 Internal Electronics

The internal electronics are contained within the pressure case of the VMP-250. The electrical system is designed to take signals measured by the sensor and transmit them to the Persistor computer where the data acquisition is controlled. A schematic of the electronics are shown in Figure 34.

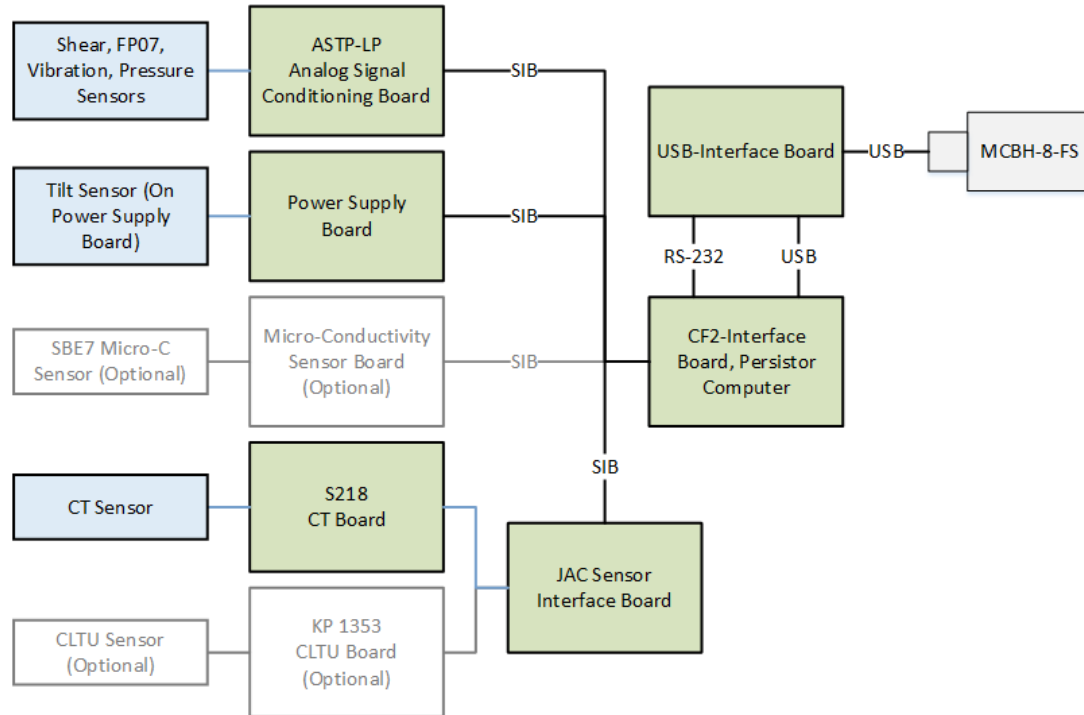


Figure 34: System block diagram for a VMP-250

The basic functionality of the electronic boards are as follows:

ASTP Board: The ASTP (i.e. accelerometer, shear, temperature, pressure) board receives analog signals from the piezo-accelerometers, shear probes, temperature probes, and the pressure sensor. These analog signals are conditioned and then sampled. The digitized data are sent via the serial instrument bus (SIB) to the CF2 interface board.

Micro Conductivity Board: The micro conductivity board receives raw power from the power supply board and generates an isolated +5 V supply using an onboard converter. Analog signals from the SBE7 micro conductivity probe are conditioned and then sampled. The data samples are transferred via the serial instrument bus to the CF2 interface board.

Power Supply Board: The power supply board receives power from an internal battery and provides power rails to all analog and digital components. The board also monitors the ON/OFF switch, and carries the 2-axis inclinometer and 3 V CR123 Lithium battery. The CR123 battery powers the monitoring circuit that is used to turn the instrument on and off.

JAC Sensor Interface Board: The JAC Sensor Interface board serves two purposes. First, it converts the digital outputs from the JFE Advantech Co. electronics boards (i.e. S218 for the CT sensor and KP1353 for the CLTU sensor) so that they are compatible with the Serial Instrument Bus (SIB). Secondly, the JAC Sensor Interface Board modifies the

voltage supplied by the power supply board so that it is the correct voltage required by each JAC sensor. Note: For each JFE Advantech sensor installed, there will be a small daughter (P088A) between the corresponding JFE Advantech board (i.e. S218 and/or KP1353) and the interface board (P058B).

CF2 Interface Board: The CF2 interface board handles all communication of data samples from the other boards and provides connection to the RS-232 and USB lines. It contains the Persister CF2 computer. Data is logged onto a Compact Flash card.

USB Interface Board: The USB Interface board acts as both a USB hub and serial-to-USB converter. This board takes the USB and RS-232 outputs from the CF2-Interface board and allows both to be transmitted along one USB cable.



Details of all the electronics boards are given in the VMP-250 Instrument Manual.

6 Data Acquisition Software (ODAS5-IR)

Data acquisition is conducted by the Persistor data logger using the program `odas5ir`. The data are recorded on a CF card. Acquisition is controlled by the contents of a configuration file named `setup.cfg`. Using the deck cable, the user can interact with the data logger via a RS-232 serial link operated in console mode. This mode is also called a “terminal” connection. The logger uses the Pico-DOS operating system, a subset of the DOS operating system. The bi-directional transfer of files (up- and down-loading) is achieved via a separate USB link. Thoroughly review the ODAS5-IR 4.0.0 User Guide for full details. Note: If your instrument has a USB-Interface Board, the RS-232 connection is established through the USB connector on the deck cable.

6.1 Console Control

To establish a serial console connection you must have the deck cable kit and the VMP itself. This deck cable kit includes:

- 1 deck cable that connects to the rear bulkhead of the VMP
- an AC-to-DC adapter that takes Mains power (from a wall outlet) and charges the battery in the VMP
- 1 USB port to connect on computer to connect deck cable.

Connect the power adapter of the deck cable to a Mains power outlet. It accepts 50/60 Hz and voltages from ~ 90 to ~ 240 VAC. To establish a connection to the VMP connect the deck cable to the rear bulkhead. Install the necessary drivers on your computer. Start `motocross` (supplied by Rockland⁷) or any other suitable serial terminal software⁸. The instrument will turn on when deck cable is inserted into the USB port on your computer. Within a few seconds you should see a response in the console. If not, consult the ODAS5-IR Manual. The most frequent problem is an incorrect COM-port number. Stop data acquisition, or the countdown to the start of data acquisition, by pressing the Q key. You are now in control of the data logger and should see the prompt `C:\>`. Thoroughly review the ODAS5-IR 4.0.0 User Guide for full details.

By typing `dir` you will see a listing of the files in your current directory which should be the root directory (designated by `C:\>`). You should see the following list of files.

- `\data` – the directory containing the data files,
- `autoexec.bat` – a script that is run upon power-up of the data logger,
- `usb1.run` the program for connecting via USB,
- `setup.cfg` – the user-configurable ASCII file containing the information that controls data acquisition and subsequent data processing.

Type

```
>> odas5ir -help
```

and see if you get a help listing. If so, the program is embedded in the flash memory of the data logger, which makes it invisible to the `dir` command, but it is still available. If not, you are missing the data acquisition program. See the ODAS5-IR User Guide for instructions on transferring it to your data logger. If you do not see the file `autoexec.bat` then your logger

⁷This Windows-only program is produced by Motorola and is automatically licensed to any user of the Persistor data logger.

⁸Mac users can use `CoolTerm`

will not start automatically on power-up. If the file `autoexec.bat` is missing but needed, see the ODAS5-IR User Guide for the appropriate command string that must be in this file and transfer it to the data logger. The free space on your disk is displayed by the `dir` command and can be used to estimate the size of the disk.

Now check that the data logger can collect data. Issue the command

```
>> odas5ir -f setup.cfg -c all
```

The data logger will sequentially collect data from every channel listed in your configuration file and provide the statistics of the minimum, maximum, mean and standard deviation in counts of the *raw* binary data (they are not in physical units). When test sensors are installed in the profiler, the mean and the standard deviation of most channels are predictable. Once you are familiar with what to expect for all channels, you can quickly determine if your data acquisition is working properly.

The next test is to start data acquisition manually to check for error messages. Issue the command

```
>> odas5ir -f setup.cfg -l 3000 -n
```

Data acquisition should start within a couple of seconds, and a line of information will appear for every data record, approximately every second ([Listing 1](#)).

Listing 1: Example Console Output

```
starting DAQ
free sectors: 2045280
created input buffer of size: 10240 bytes
L  0 N  1 p -1.05
L  1 N  2 p -1.05
L  2 N  3 p -1.05
L  3 N  4 p -1.05
L  4 N  5 p -1.05
L  5 N  6 p -1.05
L  6 N  7 p -1.05
L  7 N  8 p -1.05
L  8 N  9 p -1.05
L  9 N 10 p -1.05
L 10 N 11 p -1.05
```

Each line is formatted using,

```
L  <LAST_RECORD_COUNT> N  <NEXT_RECORD_COUNT> p <PRESSURE>
```

where `LAST_RECORD_COUNT` and `NEXT_RECORD_COUNT` are record numbers for the last and next records. The current pressure is displayed following the `p` specifier in dbar.

Stop data acquisition by pressing the `Q` key.

Next, test automatic data acquisition by toggling the VMP-250-IR ON/OFF using the switch on the deck cable, or the magnet and not issuing any commands to the instrument. When the VMP turns on, you should see a startup script on the terminal window.

6.2 RSILink 4.0.0 and File Transfers

Thoroughly review the RSILink 4.0.0 User Guide for full details.

To down- and up-load files you must connect the USB cable to your notebook computer. In MotoCross, issue the command

```
>>usb1
```

and then start the application **RSILink** on your notebook computer. Details are in the ODAS5-IR User Guide and the RSILink User Guide. Hit the **connect** button to establish the bi-directional USB connection. You can transfer any type of file in either direction. ‘Downloading’ moves files from the data logger to your notebook computer; this feature is primarily used for moving data files to your notebook computer. ‘Uploading’ moves files from your notebook computer to your data logger; this feature is used to move the configuration file, **setup.cfg**, to the data logger after editing it on your notebook computer (see [Section 6.3](#)). You can also use **RSILink** to move executable files to your data logger – if, and only if, the **usb1.pxe** program is on your data logger. Otherwise, you have to use **motocross** to move executable files. For example, after re-formatting the disk drive (which may be necessary if the CF card becomes corrupted) there will be no files on the disk.

6.3 The Configuration File

The configuration file, **setup.cfg**, controls the data acquisition and is also used in post-cruise data processing to convert your data into physical units⁹. The user must make sure that this file is correct, or as close as possible to correct, before collecting data.



The configuration file must be in pure ASCII. If using a non-standard English text editor, invisible characters can become embedded in your ASCII file and will corrupt data acquisition.

Information is grouped into sections that are identified by a **[label]**, except for the start section **[root]** which exists by default at the beginning of the file and ends at the start of the next section. The root section (Listing 2) contains parameters used to configure the VMP and software, including sampling rate, record size, and data directory. The **rate** is the rate at which lines are sampled, **prefix** is the string to which the file number is appended to create the file name. **disk** indicates the directory where the data is stored, and **recsize** is the size of a record, in seconds. **profile** indicates the direction of profiling, **no-fast** defines the number of ‘fast’ columns in the address matrix, and **no-slow** is the number of slow columns in the address matrix.

The **[matrix]** section follows the root section ([Listing 2](#)). The matrix section identifies the channels that will be sampled and the order of sampling. It consists of one or more rows, where each row has an identical number of columns. First, the number of rows in the address matrix is indicated. The rest of the lines are the rows of the address matrix and start with a row-identifier. The order of data sampling is from left to right, starting with the first row, then the second row, etc., until you get to the end of the last row. When you get the end of the last row, you return to the beginning of the first row and continue sampling.

The channels are split into “fast” and “slow” channels to save disk space. The slow channels are low-bandwidth signals, such as pressure, and do not need to be sampled as quickly as the turbulence channels. The slow channels are only sampled once every iteration through the entire

⁹The configuration file is crucial for using your VMP-250-IR. Details are described in the ODAS5-IR User Manual. The user must become completely familiar with this manual and be confident in their configuration file.

address matrix. Fast channels are turbulence signals, such as shear. Fast channels are sampled on every row of the address matrix. The channels are identified by a number in the range of 0 to 255. These channels must exist in an instrument, or else data acquisition will fail. Rockland will identify the channels that are in your instrument by supplying an initial configuration file.

Listing 2: Root section and matrix section of the configuration file.

```
rate=512
prefix=dat_
disk=c:/data
recsize=1
no-fast=9
no-slow=2

[matrix]
num_rows=8
row01= 255  0  1  2  5  8  9  52  53  64  65
row02=  32 40  1  2  5  8  9  52  53  64  65
row03=  41 42  1  2  5  8  9  52  53  64  65
row04=   4  0  1  2  5  8  9  52  53  64  65
row05=  10 11  1  2  5  8  9  52  53  64  65
row06=  12 48  1  2  5  8  9  52  53  64  65
row07=  49 50  1  2  5  8  9  52  53  64  65
row08=   0  0  1  2  5  8  9  52  53  64  65
```

The [instrument_info] section has one required parameter, **vehicle**, but can be used to provide additional information such as instrument model, serial number, etc (Listing 3). The conversion of data into physical units, the method of estimating the speed of profiling, and data visualization are vehicle-dependent (see Technical Note 039).

Listing 3: Instrument info section of the configuration file.

```
[instrument_info]
vehicle=VMP
model=VMP-250
sn=231
```

The remaining sections of the configuration file are the [channel] sections (Listing 4). These sections are not used for data acquisition but are required for converting your data file into a Matlab mat-file, for converting your data into physical units, and for processing your data. A copy of the configuration file is embedded near the front of your data file so that it can be used for data processing with the Rockland ODAS Matlab Library of functions. These functions make assumptions about the names of the channels, so care must be taken in editing this portion of the configuration file. The first part of each channel section is specific to your instrument and need not be changed. The last part of a channel section must be changed whenever you change a calibrated sensor, such as the shear probe or thermistor, because this part is specific to your sensor. Note, thermistors and micro-conductivity probes are not always calibrated, in which case use nominal values.

Listing 4: A channel with board-specific and probe-specific calibration values

```
[channel]
id=4
type=therm
name=T1
; board-specific calibration information
adc_fs=4.096
adc_bits=16
a=-12
b=0.99855
G=6
E_B=0.68222
; probe-specific calibration information
SN=T944
beta_1=2976.47
beta_2=256353.27
T_0=280.265
units=[degC]
```

6.4 Data File Formats

Data files are structured as a series of records. The first record is a configuration record constructed by prepending a header to the configuration string. Subsequent records are data records constructed by prepending a header to a data block (Figure 35).

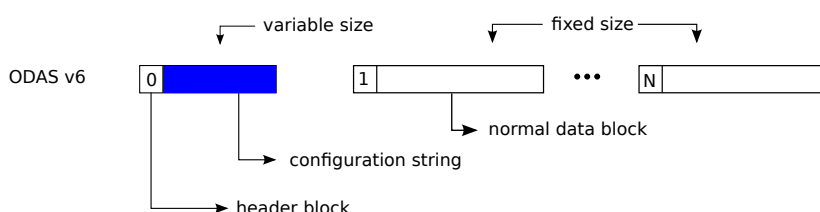


Figure 35: The format of a data file. The very first record (labelled 0 and blue) holds the configuration information. All subsequent records hold data and are of identical size.

Both configuration and data records start with a 128-byte header comprised of 64, unsigned 16-bit words. This header contains the information needed to correctly read the data file. Details about the contents of the header are found in ODAS5-IR User Guide.

The first record within a data file is a copy of configuration file. The record consists of a header followed by an ASCII string that is the content of the configuration file. The record number is always equal to 0 because it is the first record of the file. The size of the first record depends on the size of the configuration file.

Data records consist of a header (briefly discussed above) and a data block. The data block consists of unprocessed data. Data words are stored in the order provided by the address matrix. The number of words within a data block is a multiple of the size of the address matrix. The total size of a data record (header + data block) is stored in the header. All data records are the same size.

6.5 Log File

The log file, `logfile.txt` is found in the same directory as the data files. It contains a record of events that occur when the data acquisition software is running. Each event entry includes the time, event type, record number where the event occurred, and data file name. The log file provides the history of data acquisition and is a tool for troubleshooting problems.

Further details about the data acquisition software are found in the ODAS5-IR User Guide.

7 Pre-deployment Checks

7.1 Electronics Check: The Bench Test

The purpose of the bench test is to confirm that your VMP is recording data and functioning properly. To conduct a bench test:

- With the test probes installed, connect the VMP to your computer using the deck cable.
- Lay the VMP horizontally on a foam cushion (to minimize vibrations), or stand it up vertically.
- Turn the instrument “ON” and record about 60s of data. If error messages appear in the console, consult the ODAS5-IR User Manual.
- Download the data file.
- Use either Zissou Essentials of the Matlab function `quick_bench.m` to display the data. The Matlab syntax is as follows:

```
>> quick_bench('DAT_001', 'SN')
```

where 'DAT_001.p' is the name of the bench test data file, and 'SN', which is an optional input, is a string of the serial number of your VMP. The software will generate at least two figures: (1) a time series of the raw data collected by the ASTP and inclinometer channels and, (2) the spectra of the ASTP channels. Additional figures will be generated if other sensors are present on your instrument.

- Use the generated figures to complete the ‘Bench Test Checklist’ included with your instrument manual or available on Rockland Scientific’s website.



The bench test spectra represent the electronic noise of your instrument and – for most channels – should be similar to those in the ASTP calibration certificate shipped with the VMP. Notably, the spectra of the vibration sensors (A_x and A_y) that you obtain will differ from the spectra in your calibration report because the accelerometers are disconnected when we perform the factory bench test.

7.2 Mechanical Check

Before deploying your VMP, check all external fasteners and ensure they are secured firmly. Check that all probes are properly seated and ensure all probe holders are adequately tightened. Ensure all O-rings and sealing surfaces (including probe bodies) are clean and free of scratches and cuts/nicks. Note: all o-rings should be replaced annually.

7.3 Pre-deployment Checklist

The following pre-deployment checklist should be completed before every deployment:

- ☐ Read and understand the instrument manual
- ☐ Inspect all sealing surfaces on instrument and confirm their integrity
- ☐ Complete bench test checklist with test probes and confirm data acquisition
- ☐ Ensure all probe ports have a greased O-ring and white ferrule
- ☐ Ensure O-ring and ferrule are in correct order (consult manual for diagram)
- ☐ Ensure that all bolts and connectors are secure (consult manual for safe torque rating)
- ☐ Securely attach the drag brush, tail assembly and buoyancy collar (if applicable)
- ☐ Tie tether securely to eyebolt
- ☐ Ensure all probes are fully inserted
- ☐ Confirm shear probes are oriented correctly
- ☐ Ensure probe holders are tightened correctly (consult manual)
- ☐ Turn the instrument on and monitor LED for confirmation



A copy of this checklist is included in your instrument manual, or available on Rockland Scientific's website.

8 VMP Deployment

8.1 Cruise Outline

A typical cruise outline is as follows:

- Load ship
- Identify work areas - VMP rest area and laboratory space for looking at data.
- Place and secure winch or line-handling system.
- With test probes installed, connect deck cable to VMP and confirm its proper operation.
- Install the aft section of the VMP and attach the tether.
- Install shear probes and thermistors.
- Discuss station plan with scientific staff and crew. Make the officers familiar with the VMP and its mode of deployment.
- Decide whether deployment will be over the side or off the stern.
- Discuss with officers the issue of the propeller. It must be truly static when the instrument is deployed because it can cut the cable instantly.
- Go to first station and make several deployments that end well above the bottom to build experience and confidence.
- Retrieve VMP to deck and download the data files.
- Run Zissou Essentials or `quick_look.m` to assess the instrument and the deployment method (i.e. is the tether slack during free-fall?)
- Explain the many figures produced by `quick_look.m`.
- Run the station plan.
- Download the data files.
- Rinse the VMP with fresh water.
- Remove thermistors and shear probes.
- Disassemble the tail section.
- Dry the instrument and place it into its shipping container.
- Return to port and off load the ship.

8.2 Tethered Vertical Profilers

When deploying your instrument with a tether, the goal is to produce a smooth fall rate and minimal tilting that will result in high quality data¹⁰. To achieve this, the instrument must be decoupled from the motion of the deployment platform by using a slack tether.

The progression of a **downward profile** is shown in [Figure 36](#). The specific steps are as follows:

1. Deploy the profiler and hold it below the surface for 30 s to 90 s, releasing any air caught in the tail assembly.
2. To profile, let the instrument fall through the water column, keeping 1 to 2 coils of tether on the surface of the water ([Figure 36a](#)).
3. When the profile is complete¹¹, pull the instrument up to the surface ([Figure 36b](#)).

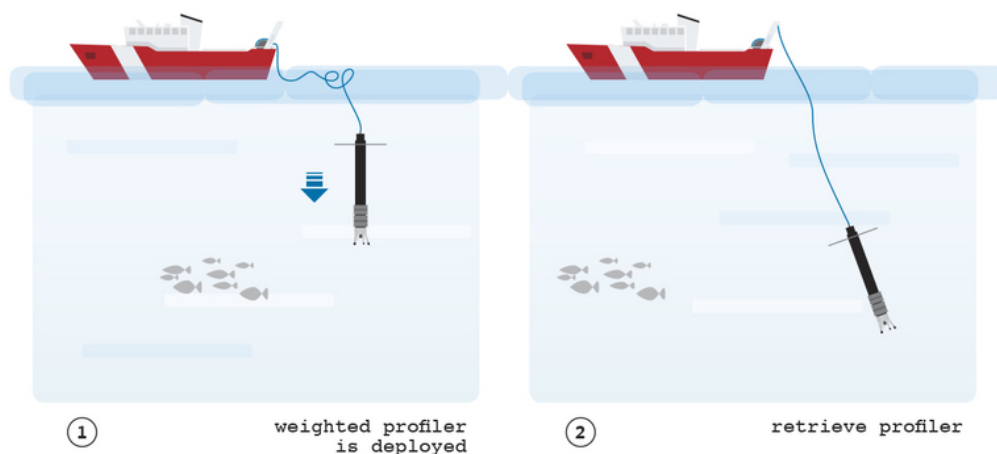


Figure 36: Downward Profile

¹⁰High quality data is achieved when the fall rate is constant (approximately 0.75 ms^{-1}) and pitch is small (5°). Details are found in Rockland Technical Note 039 – A Quick Guide to `quick_look.m`.

¹¹There are several ways to estimate the depth of your profile, including marks on the cable and time of flight, which is described in the Instrument Manual.

The progression of an **uprising profile** is shown in Figure 37. The deployment technique is similar to that for a downward profile, however, two people are needed – one for the instrument tether and one for the release tether. **The two people should position themselves as far apart as the vessel will allow.**

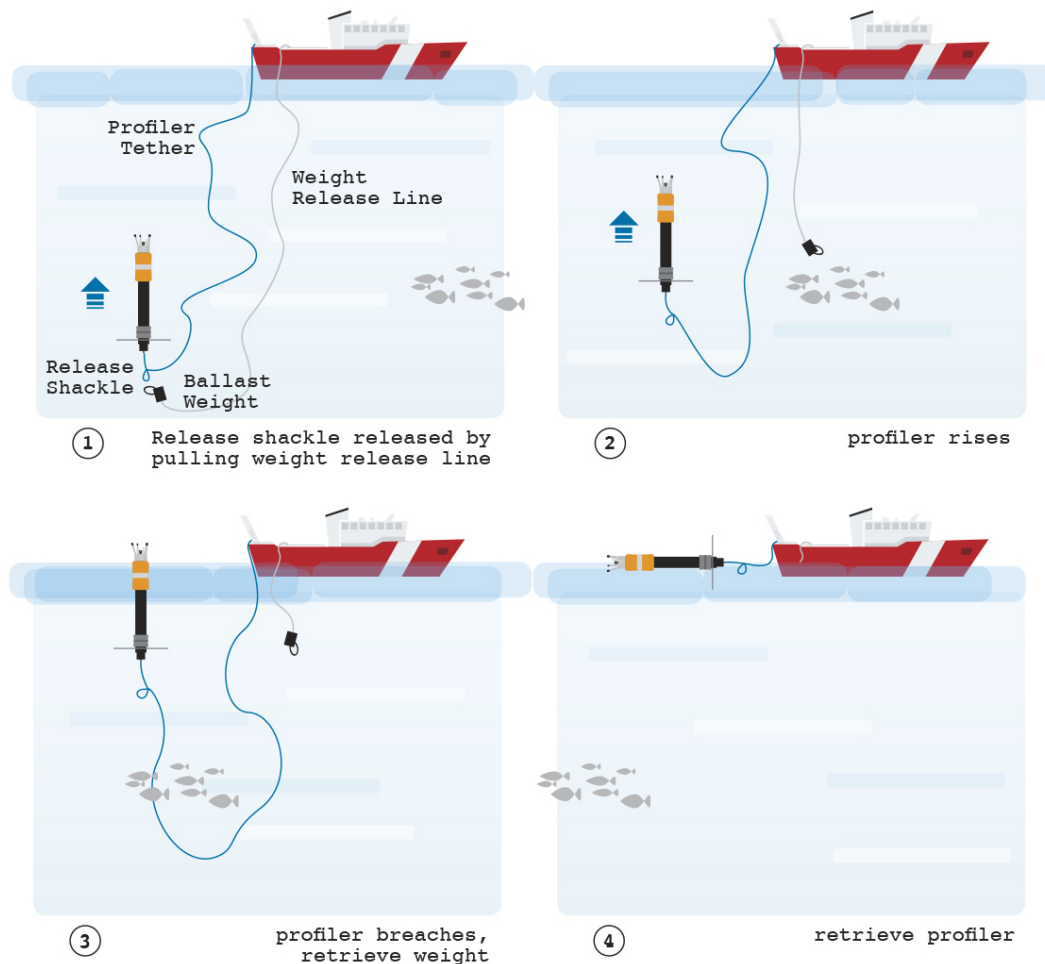


Figure 37: Schematic of an uprising profile. Note: In step one, the release is actually triggered by pressing the button on the release controller (i.e. not by pulling on the release line as indicated in the figure).

8.3 Winch Operations

Winches can be equipped with a rope tether for Internally Recording (IR) instruments or an electro-mechanical tether and a slip ring for Real Time (RT) systems. Winches such as the CSW-7 and PID-02 (Figure 38) have an IP55 rating. They can withstand gentle spray of water from a hose, but cannot withstand immersion or impact from waves. It is therefore recommended that you store your winch inside, or cover it, whenever possible. This will increase the lifespan of the winch.



Please reference the Winch Manual for complete information about the operation and maintenance of the winch. Only general guidelines are provided here.

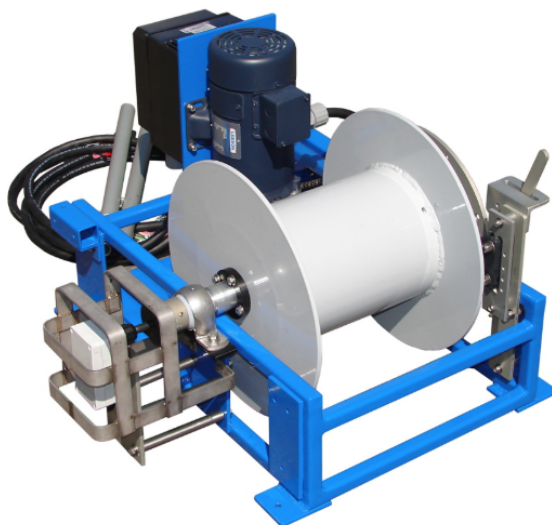


Figure 38: PID-02 electric winch recommended for VMP-250-IR deployment.

Installation

To install the winch on a vessel please consider the following:

- Can the vessel provide appropriate power to the winch? Check the winch specifications for more information about power requirements.
- Can the winch be mounted on the deck in an appropriate location?
 - When possible, winches should be bolted to the deck.
 - For smaller winches, it is appropriate to strap the winch to the deck.
 - Position the winch near an A-frame, crane or davit equipped with a sheave or block.
 - If using a Real Time system, and if possible, position the winch close enough to the lab to connect the winch to a computer with the deck cable
- Can the winch be covered with a tarp, stored in a crate on deck or moved inside during storms? The winch is not rated to withstand waves or immersion in water.

Operation

The following are basic recommendations for deploying and recovering a vertical profiler using a winch. These instructions do not include considerations of ship positioning, weather conditions and vessel type. Always ensure you have a plan and good communication with the vessel's crew before any attempt to deploy the instrument.

- Make sure the rope is securely tied to the winch when installing new rope.
- Mark the tether 5 or 10 m from the VMP so that you know when it is nearing the surface.
- Mark the tether at intervals of your choice so you know how much tether has been deployed.
- When deploying the instrument, remove the tether from the sheave or block to aid in throwing the tether into the water. Disengage the clutch and pull the tether off the drum by hand, ensuring there is one coil of tether on the surface at all times. It is recommended that the deployment is completed by two people: one who removes the tether from the drum, and the second who throws the tether into the water.
- Always leave a minimum of 10 wraps of the tether on the drum. The friction from the tether wrapped on the drum is stronger than any knot.
- When recovering the instrument, place the tether through the sheave or block. If using a Real Time system, ensure the sheave or block has an appropriate radius based on the minimum bend radius of the electro-mechanical cable.
- It is recommended to recover your VMP at a **max winch speed of 60%**. Increased speed will increase load on the tether and increase risk of the tether breaking.
- At the end of each deployment session, gently rinse the rope with fresh water, especially if operating in a dirty or saltwater environment. If salt water dries on the rope the salt crystals may cut the fibers and decrease the strength of the rope. Rinsing will also help remove unwanted odours.
- Dry the motor, brake and gear box with a towel after rinsing.
- Cover winch or bring inside ship if heavy seas are expected.

Care and Maintenance

The following are instructions for the care and maintenance of your winch:

Pre-deployment (at least one month ahead of deployment):

- Confirm that the winch is functioning correctly
 - Provide power to the winch
 - Rotate drum forward and backwards
 - Test emergency stop
 - Listen for unexpected noises
- Apply LPS 3 Rust Inhibitor to the steel components especially the sprockets.
- Re-apply grease to the chain if needed. Corrosion Block Grease is recommended.

- If using a Real Time instrument, test communication between VMP and computer via winch cable.

Post Cruise:

- Give the winch a final gentle rinse with fresh water. Allow the winch to dry.
- Inspect for rust or damage.
- Re-apply rust inhibitor spray.
- Apply grease to chain as needed.
- Store in a dry location indoors.

9 Instrument Maintenance

It is absolutely imperative that you follow a rigorous maintenance schedule to avoid damage and ensure the longevity of your VMP. The following procedures are recommended:

- After every use, rinse with fresh water.
- After every cruise, inspect and clean all sealing surfaces and O-rings.
- Once a year completely disassemble the instrument, clean every sealing surface, and replace all O-rings.

At all times, check for any signs of corrosion or moisture damage.



Thoroughly review and reference the Maintenance section of the Instrument Manual for complete maintenance details. The content provided here is simply a summary of the information.

9.1 Cleaning the VMP

While still fully assembled, rinse the VMP thoroughly with clean fresh water. Dry off the instrument and disassemble. Wipe every sealing surface clean with isopropyl alcohol and inspect for debris and damage. Carefully check o-rings for hair, lint, or nicks and scratches. The smallest disturbance to the sealing surface can allow water to leak into the pressure tube. If the o-ring is damaged, replace with a new o-ring. In the absence of any damage, replace o-rings annually. Lightly grease the o-rings before replacing.

Remove connectors on the rear bulkhead, clean the surfaces and inspect the o-rings. Disassemble the probe holders, and pay special attention to the base of the probe holders for any signs of corrosion or scratches. Parts without any electronic components (nuts, probe holders, etc.) should be rinsed in fresh water and completely dried. Inspect the entire instrument for signs of wear or corrosion ([Figure 39](#)). Reassemble the instrument. Always insert test probes when storing the instrument.

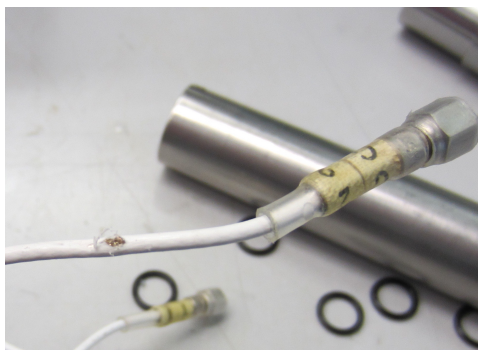


Figure 39: Inspect your SMC cable for any nicks or damage. A compromised SMC cable will produce noisy data. Check the SMC connectors for signs of water damage or tarnishing.

Further details and instructions on cleaning the VMP are outlined in detail in your Instrument Manual.

9.2 Corrosion and Moisture Damage

Check the VMP frequently for corrosion, in the form of pitting (Figure 40). Although pitting on the side of the nose cone (Figure 41) is not a show-stopper, any pitting on sealing surfaces (Figure 42) can be detrimental to your deployment. Inspect your instrument carefully and contact us if you have any questions or concerns.



Figure 40: Severe pitting on the nose cone.



Figure 41: Pitting on the surface on the nose cone is less concerning than pitting on a sealing surface.



Figure 42: Pitting on the nose cone impinging on the face-seal of the probe holders. This instrument should not be deployed.

Inspect the instrument for moisture damage. When opening the VMP, watch for any signs of moisture or condensation inside the pressure tube. Green or blue corrosion residue indicate that moisture has been present (Figure 43 and Figure 44). Any yellowing on the clear plastic that encases the SMC connectors indicates that moisture has wicked its way behind the probes. Please consult us before deploying an instrument that has any corrosion.

Remove the desiccant bag. If the moisture indicator has turned pink, the desiccant bag must be re-activated¹². To reactivate, place the desiccant bag in a 275 °F oven for 16 hours. Replace the desiccant bag in the VMP, tightly securing it to the rail.

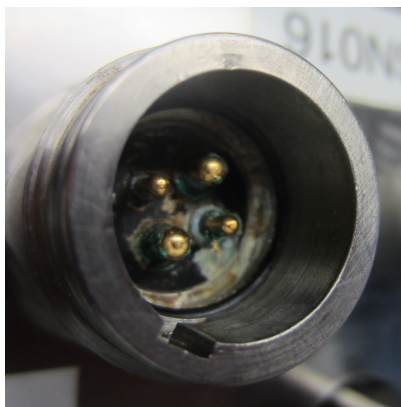


Figure 43: Green corrosion on the Impulse connector pins



Figure 44: Green corrosion on the pin of the SMC cable

¹²With time, the moisture indicator will activate due to moisture in the air. This does not mean that the instrument has been flooded.

10 Data Conversion and Processing

10.1 Software Tools

Rockland Scientific offers two software tools to process your data:

- *Zissou Essentials*: A standalone software program designed to be a user friendly tool to inspect data acquired with your instrument, primarily for quality control purpose. The software is used for conversion of data to physical units, data inspection and basic data visualization. It is intended to quickly visualize the data to ensure that an instrument is performing as expected. This software is designed to open and process any Rockland binary data file (p-file). Details of the software can be found in the Zissou Data Processing Manual.
- *ODAS Matlab Library*: A library consisting of a set of Matlab functions that provide the tools to process data collected with all instruments made by Rockland. A valid license for Matlab is required to use the functions. The library is more complete than Zissou Essentials in that it is intended for both data visualization and data analysis. The Matlab Library allows for more automated processing and enhanced control of the processing parameters. Details of the software can be found in the ODAS Matlab Library Manual.

10.2 Technical Note 039 – “A Guide to Data Processing”

A comprehensive overview of the routines and parameters that are used to analyze Rockland data files is provided in Technical Note 039. The document is primarily focused on the ODAS Matlab Library and covers the following essential topics:

- Configuration files
- Conversion of the data to physical units
- Data visualization and profile selection
- Calculation of the rate of dissipation of turbulent kinetic energy¹³

The following advanced data processing techniques are also outlined in the technical note:

- Hotel-files
- *It situ* Thermistor Calibration
- Despiking shear data
- Inspecting spectra

A full review of this technical note is highly recommended.

¹³For details of the calculation algorithm, refer to Technical Note 028 (“Calculating the Rate of Dissipation of Turbulent Kinetic Energy”)

10.3 Computing the Dissipation Rate of TKE

A schematic of the method used to estimate the dissipation rate of TKE, ϵ , from shear data is provided in Figure 45. Details of the method are described in Technical Note 028, available on Rockland's website¹⁴.

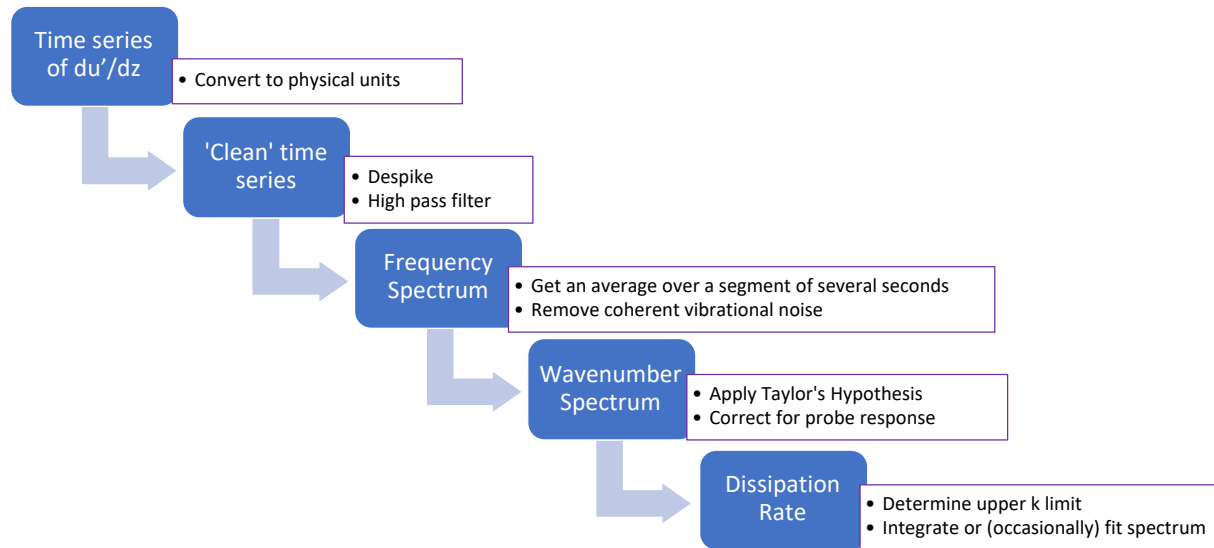


Figure 45: Overview of the method to compute the dissipation rate of TKE. ϵ .

10.4 Choosing Dissipation Rate Processing Parameters

The functions within Zissou Essentials and the ODAS Matlab library use default parameters that are best suited for downward vertical profiles, and it is likely that you will want to change these parameters when processing the data for your specific instrument, and your scientific objectives. In particular, you may want to modify the spatial resolution of the dissipation rate. Higher resolution would enable the detection of smaller scale variability in ϵ , whereas lower resolution would increase the statistical reliability of the estimate and the effectiveness of the noise removal algorithms. Therefore, there is always a compromise to consider when choosing your processing parameters. Ultimately, the choice is dependent on your scientific objectives, but a good understanding of the parameters can help guide the decision.

Within the software packages, the resolution of the dissipation rate is controlled by the following parameters:

- **diss_length:** The time span [s] over which each dissipation estimate is made. Default = 8;
- **overlap:** The overlap [s] of the dissipation estimates. Default = 4.
- **fft_length:** The length [s] of the fft-segments that will be ensemble-averaged into a spectrum. Default = 2.

The influence of the **diss_length** and **overlap** is depicted graphically in Figure 46 for a 24 s segment of data. Using the default parameters, there is an estimate of ϵ every 4 s (i.e. **diss_length-overlap**), but the estimate is computed from an 8 s (i.e. **diss_length**) segment of data. For a given profiling speed, W , this will result in an estimate of ϵ corresponding to a space interval

¹⁴www.rocklandscientific.com

of $W \times (\text{diss_length} - \text{overlap})$, which equates to 3 m for a typical vertical profiler travelling at 0.75 m s^{-1} . For a glider with an along-path profiling speed of 0.3 m s^{-1} , the default parameters would yield an along-path resolution of 1.2 m, which would correspond to a vertical resolution of 0.4 m if $dP/dt = 0.1 \text{ m s}^{-1}$.

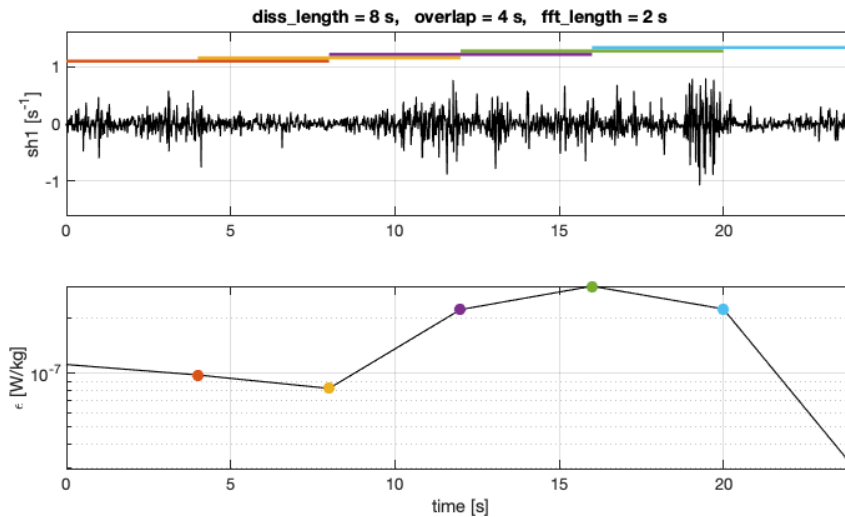


Figure 46: Estimating dissipation rates using the default parameters. The top panel shows the $sh1$ data as a function of time and the lower panel shows the corresponding dissipation rates. The coloured bars in the top panel show the segment of data that was used to generate the corresponding point in the lower panel.

The `fft_length` parameter does not affect the resolution of ϵ . Instead it controls how many spectra are ensemble averaged into the final spectrum that is used to make the estimate of ϵ ¹⁵. The more spectra that are averaged, i.e. the higher $\text{diss_length}/\text{fft_length}$, the more degrees of freedom you have and the better your statistical reliability. When using Rockland’s software, the number of spectra that are averaged is given by $2 \times \text{diss_length}/\text{fft_length} - 1$ because the `fft`-segments are always overlapped by 50%. This implies that when using the default parameters, seven spectra are averaged for each estimate of ϵ . The processing code requires that a minimum of three `fft`-segments are used, i.e. $\text{diss_length}/\text{fft_length} \geq 2$, and an error will be issued if the chosen parameters do not satisfy this criterion.

With these key ideas in mind, the simplest way to increase the spatial resolution in ϵ is to decrease the `diss_length`. Correspondingly, you should reduce the `overlap`, so that 50% overlap is maintained¹⁶, i.e. $\text{diss_length}/\text{overlap} = 2$. The consequence of doing this is reduced statistical reliability because there will be fewer `fft`-segments used for each ϵ estimate (i.e. fewer degrees of freedom) – but that might be a price you are willing to pay.

By now, you might be thinking, if I reduce the `fft_length` in addition to the `diss_length` and the `overlap`, I can get a higher resolution in ϵ and maintain the number of degrees of freedom. For example, if you set the `diss_length`=4, `overlap`=2, and `fft_length`=1, the ratios of the default parameters are maintained. This is indeed true, but comes at a consequence of the frequency resolution – and hence, the lowest frequency value – of the spectrum since $\Delta f = 1/\text{fft_length}$ (Figure 47). And to make matters more complicated, the necessary frequency

¹⁵See Technical Note 028 for details on the estimation of ϵ

¹⁶Note: Another way to increase the vertical resolution – assuming a fixed `fft_length` and `diss_length` – is to increase the `overlap`. An overlap of greater than 50% can be used, however, the assumptions of statistical independence start to be violated and having an overlap greater than 50% is not recommended.

resolution is dependent on the turbulence level, and hence on the dissipation rate (See item 2 below).

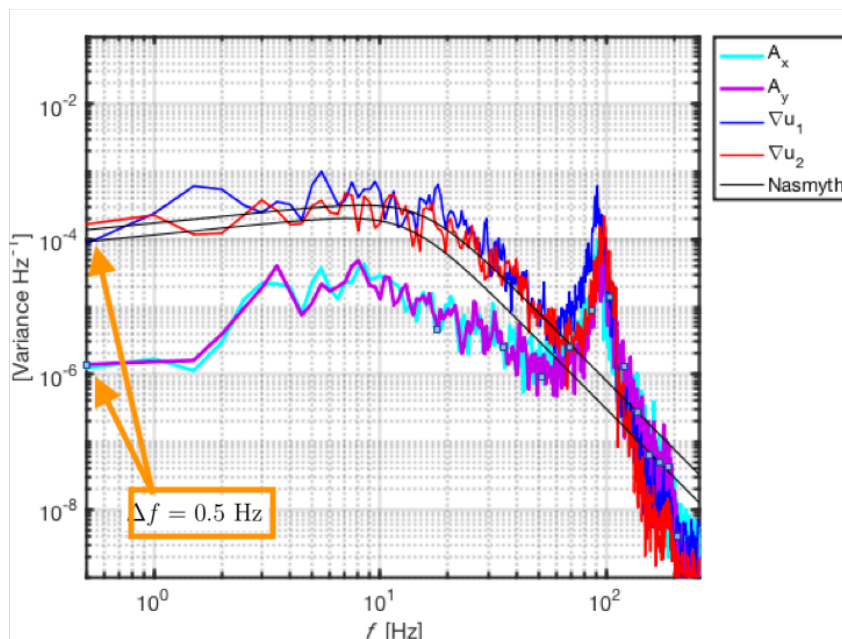


Figure 47: Example of a frequency spectrum computed with the default `fft_length= 2s`, giving a frequency resolution of 0.5 Hz.

So, how does this minimum dosage of time series analysis translate to analyzing your microstructure data? The recommended approach is as follows:

1. Use the default parameters – or some knowledge of your study site – to get an estimate of the order of magnitude of the dissipation rate.
2. Using the theoretical Nasmyth spectra (Figure 7), compute your desired wave number resolution. To obtain a meaningful estimate of the integral of the spectrum you want to resolve at least 90% of the theoretical variance (Figure 5, grey region) A good rule of thumb is to try to have the first wavenumber bin about half way to the peak (or smaller). For example, for a dissipation rate of $10^{-7} \text{ W kg}^{-1}$, a value of $\Delta k = 2 \text{ cpm}$ would be a good choice. Hence, lower dissipation rates will require smaller Δk values than higher dissipation rates.
3. Use your vehicle speed, W (i.e. dP/dt for a vertical profiler or along path speed for a glider), to estimate Δf , i.e.

$$\Delta f = W \Delta k \quad (105)$$

and hence `fft_length= 1/Δf`. Hence, lower dissipation rates will require longer `fft`-segments than shorter higher dissipation rates.

4. Set the `diss_length` to be a multiple of the `fft_length` and at least $2 \times \text{fft_length}$. The `overlap` should be `diss_length/2`. Shorter `diss_length` segments will yield greater spatial resolution in ϵ , but lower statistical reliability. This compromise is something for you to decide upon based on your scientific objectives.

10.5 Analysing Turbulence Scalar Spectra

The one-dimensional scalar (temperature) gradient spectrum can be directly integrated to obtain χ (Oakey, 1982a):

$$\chi = 6\kappa \overline{\left(\frac{\partial T'}{\partial x_3}\right)^2} = 6\kappa \int_0^\infty \Phi_{\partial T/\partial x_3}(k) dk \quad (106)$$

provided both the viscous-convective and viscous-diffusive subranges are resolved. Turbulence is likely more isotropic at these high wavenumbers but resolving the entire viscous-diffusive subrange is difficult. When ϵ is unknown, the wavenumbers associated with these subranges must be identified by the shape of the measured temperature gradient spectra. The high wavenumbers, and the spectral roll-off in the observed spectra, may be “drowned” by instrument measurement noise. These measured spectra must also be substantially corrected (sometimes up to a factor of $\sim 10^3$) due to the poor response of temperature sensors at high frequencies. The poor thermistor response will also affect the salinity gradient spectrum since it depends on the conductivity and temperature respective spectra and their cospectrum as shown in eq. 10.5.

Methods for obtaining χ

With those technological and theoretical considerations in mind, Bluteau et al. (2017) detailed how to estimate χ from temperature gradient spectra once ϵ is determined from the shear spectrum. By obtaining ϵ first, the theoretical wavenumbers associated the viscous-convective and viscous-diffusive subranges can be determined using the Kolmogorov wavenumber $k_\eta = 1/L_K$ and the Batchelor wavenumber k_B [rad m⁻¹] instead of relying on identifying a spectral roll-off in the temperature gradient spectra. Identifying the viscous subrange from the spectra itself is challenging given the ambiguities with the thermal frequency response and instrument noise as illustrated in Figure 48 for relatively low ϵ (see also Section 4 of Bluteau et al., 2017).

The techniques detailed by Bluteau et al. (2017) involve integrating the resolved portion of the viscous subrange when a spectral roll-off is observed (e.g., low ϵ and noise levels). In more energetic environments, the smallest and largest wavenumbers become increasingly separated, and the inertial-convective subrange can be fitted to obtain χ .

This strategy has many advantages over integrating the viscous subranges. The inertial-convective subrange is located at lower wavenumbers than the viscous-diffusive subrange, and thus more readily resolved with the FP07 thermistor. Even at the fast profiling speeds, there is no correction for the thermistor’s response required within the inertial-convective subrange except perhaps for the largest wavenumbers when $\epsilon \gtrsim 10^{-5}$ W kg⁻¹ (Figure 17). The universality of the theoretical form in this subrange is also less debated than that in the viscous subranges. The technique’s main caveat is that with decreasing L/L_K the lowest wavenumbers are increasingly impacted by the mean flow. These wavenumbers must be avoided, and in very weak turbulent environments there is insufficient bandwidth within the inertial-convective for spectral fitting. For stratified-shear flows, the buoyancy Reynolds number:

$$Re_b = \frac{\epsilon}{\nu N^2} \quad (107)$$

can be used to evaluate this bandwidth (see Gargett et al., 1984; Bluteau et al., 2011). This parameter measures the ratio between the large overturns in a stratified flow i.e., Ozmidov length scale L_O and the smallest overturns η such that $Re_b = (L_O/L_K)^{4/3}$.

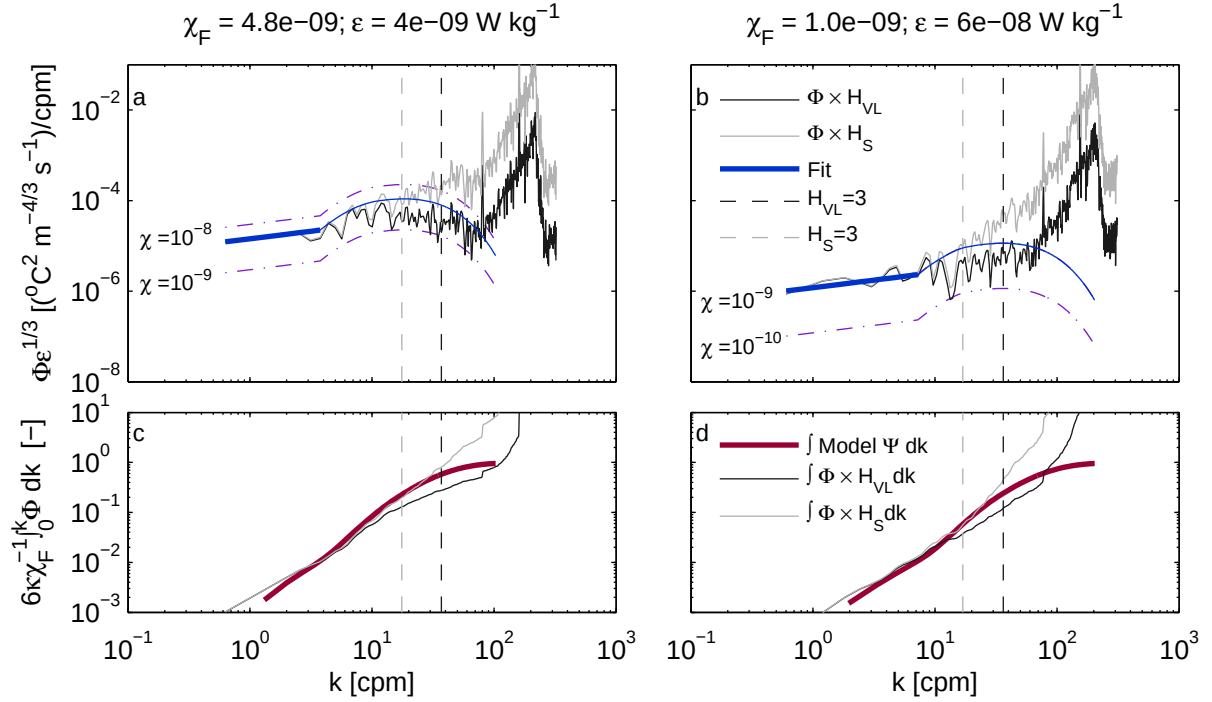


Figure 48: (a and b) Examples of the vertical temperature gradient $\partial T/\partial x_3$ spectra for two low ϵ . The observed spectra Φ have been corrected for the thermistor frequency response using the definition of Vachon and Lueck (1984), H_{VL} , and the more aggressive definition of Sommer et al. (2013b) H_S . The panels directly beneath (c and d) show the cumulative integrated spectra for the model and the observed spectra, both are non-dimensionalized by χ_F — the fitted estimate from the inertial-convective subrange. The effects of applying different thermistor correction factors are also illustrated. The violet dotted-dashed lines are the Kraichnan spectrum for χ rounded to the nearest order of magnitude above and below χ_F . Reproduced from Bluteau et al. (2017).

Analysis and algorithm techniques

We summarize the algorithms and thresholds for estimating χ from the scalar gradient spectra, which are detailed by (Bluteau et al., 2017). As a rule of thumb, the integration method is usable when the scalar gradient spectra roll-off is resolved. For typical VMP profiling speeds of $\sim 0.6 - 0.8 \text{ m s}^{-1}$, the integration method is restricted to $\epsilon \lesssim 5 \times 10^{-8} \text{ W kg}^{-1}$ given the thermistor's response time (Figure 17). The fitting method is applicable when $Re_b \gtrsim 450$ since anisotropy restricts the spectral bandwidth of the inertial subrange (of velocity at least). This threshold can be relaxed to $Re_b \lesssim 180$ for moored measurements since the longitudinal turbulence characteristics are less impacted by stratification (Gargett et al., 1984). Both the integration and fitting methods were usable over the range of conditions when $Re_b \gtrsim 450$ and $\epsilon \lesssim 5 \times 10^{-8} \text{ W kg}^{-1}$ (Bluteau et al., 2017), given the FP07's response time and the large χ observed i.e., noise floor did not adversely impact their estimates.

Instead of prescribing which technique to apply χ , Bluteau et al. (2017) attempt both techniques on each scalar gradient spectra estimated over the same segments as the shear spectra used for obtaining ϵ . To attempt spectral fitting of the inertial-convective subrange, the spectra must resolve lower wavenumbers than necessary for estimating ϵ from the shear probes. For these low ϵ , however, the integrated χ_I is typically more reliable than χ_F because the temperature gradient spectral roll-off is normally resolvable (Figure 17a). If working with good quality tem-

perature (FP07) measurements, similar segment and spectral averaging strategies can be used for integrating the scalar spectrum as integrating the shear spectrum to estimate ϵ . Integration of the scalar spectrum is however limited to weaker ϵ since the viscous subranges of the scalar spectrum are located at higher wavenumbers than the shear spectrum. For salinity, an integrated estimate χ_I will theoretically rarely be possible so the chosen segment length and spectral averaging strategy must ensure lower k are resolved to obtain χ_F by fitting the inertial-convective subrange. Irrespective of the Re_b and the scalar (salt vs temperature), $kL_K \approx 0.01$ must be resolved to ensure enough spectral bandwidth for fitting. For low $\epsilon < 10^{-9} \text{ W kg}^{-1}$, this criteria translates to $k \sim 2 \text{ rad m}^{-1}$ (0.3 cpm), while for larger $\epsilon \sim 10^{-7} \text{ W kg}^{-1}$ the spectra must resolve $k \sim 6 \text{ rad m}^{-1}$ ($\hat{k} \sim 1 \text{ cpm}$, Figure 17).

Spectral fitting algorithm for χ The theoretical temperature (or salinity) gradient model spectrum in the inertial-convective subrange is given by (Tennekes and Lumley, 1972):

$$\Psi_{\partial T/\partial x_i}(k) = C_T \chi \epsilon^{-1/3} k^{1/3} \quad (108)$$

where k is the wavenumber expressed in rad m^{-1} and C_T is the Obukhov-Corrsin universal constant C_T . Sreenivasan (1996) compiled observations from diverse published sources, which yielded C_T between 0.3 and 0.5. From this analysis, he recommended $C_T = 0.4$ and so like others (e.g., Geyer et al., 2008; Holleman et al., 2016), we set $C_T = 0.4$ and note that the possible range for C_T between 0.3 and 0.5 may bias χ by 25%.

We use the maximum likelihood estimator (MLE) (e.g., equation 3 of Bluteau et al., 2016) to fit the inertial-convective subrange model (eq. (108)) to the observed wavenumber spectral observations $\Phi_{\partial T/\partial x_i}(k)$. Given the larger number of spectral observations with increasing k , spectral fitting is biased towards the higher and hence more isotropic wavenumbers k . The lower k , potentially more adversely impacted by the mean flow, are excluded by applying the mean absolute deviation (MAD) misfit criteria proposed by Ruddick et al. (2000):

$$\text{MAD} = \left| \frac{1}{n} \sum_{i=1}^n \left(\frac{\Phi(k_i)}{\Psi(k_i)} - \left\langle \frac{\Phi(k_i)}{\Psi(k_i)} \right\rangle \right) \right| \quad (109)$$

to short subsets of the spectra approximately 0.5 to 0.7 of a decade long. Here, n refers to the number of individual i spectral observations $\Phi(k)$ used to fit the theoretical spectra $\Psi(k)$ in eq. (108). If all subsets of the spectra yield a $\text{MAD} > 2(2/d)^{1/2}$ (Ruddick et al., 2000), the spectra is completely discarded. Otherwise, the final wavenumber range used to obtain χ_F starts at the lowest wavenumber at which $\text{MAD} < 2(2/d)^{1/2}$ and ends at $k \lesssim 0.1\eta^{-1}$ or the wavenumber at which the frequency response correction exceeds 3, whichever is smallest.

Integration algorithm for χ To obtain χ_I , eq. (106) is integrated to a maximum wavenumber k_{mI} in the temperature gradient spectral observations. For typical VMP profiling speeds, this maximum is largely controlled by the thermal frequency response. This technological limit is set to the wavenumber where the frequency response correction reaches 3, corresponding to $\hat{k} \sim 30$ -50 cpm for the correction of Vachon and Lueck (1984), and $\hat{k} \sim 10$ cpm for the more aggressive correction of Sommer et al. (2013b) — both illustrated in Figure 17b. In some instances, such as at low turbulence levels, these wavenumbers may be dominated by measurement noise. To avoid setting k_{mI} within the noise-dominated wavenumbers, we identify the local minima of the smoothed (band averaged) spectra between $0.2 k_B$ and k_B . If none is found, the local minima is set to k_B . The final k_{mI} is the smallest of this local minima and the maximum usable k according to the correction for the thermistor's response.

The integrated χ_I estimate is deemed acceptable if $k_{mI} > 0.3k_b$ i.e., if the spectral peak (roll-off) is resolved. The variance is then adjusted to account for the unresolved variance according to the proportion of variance that should be resolved as a function of k/k_B using Kraichnan's model spectrum with the same universal constant $q = 5.25$ recommended by Bogucki et al. (2012).

Estimating the salinity gradient spectrum

Estimating the dissipation of salt variance from the salt gradient spectrum is somewhat more complicated than working with temperature. The conductivity sensor can resolve more accurately higher wavenumbers than the FP07 but the salinity signal depends on both the measured conductivity and temperature. The thermal response thus hinders resolving the viscous-diffusive subrange of the salinity gradient spectra, which are theoretically located at even higher wavenumbers than temperature. Not all is lost since the inertial-convective subrange of the salinity gradient spectrum exists over similar wavenumbers than that of temperature gradients. The inertial-convective subrange for salinity can therefore be used to estimate χ_S provided the salinity gradient spectra can be computed.

To compute the salinity gradient spectrum, we linearise the salinity relationship:

$$S(t) = aC(t) + bT(t) + \lambda \quad (110)$$

where λ is considered constant given the small pressure variations over the short segments of data used for computing turbulence quantities. The salinity gradient spectrum thus depends on the conductivity gradient and temperature gradient spectrum, and their cospectrum C_{CT} :

$$\Phi_S = a^2\Phi_C + b^2\Phi_T + 2abC_{CT}. \quad (111)$$

The parameters a and b vary weakly with temperature and salinity, such that they are assumed constant over the short segment lengths used to compute the scalar gradient spectra. Therefore, the number of degrees of freedom obtained from the spectral averaging must be increased given the use of a cospectra in the estimation of $\Phi_{\partial S/\partial x_i}$.

11 Essential Troubleshooting

Rockland is committed to helping you troubleshoot any problem with your instrument or probes. Do not hesitate to contact our support team with questions or to request a review of a bench test or data file.

Poor quality data collected with a Rockland instrument could be caused by several reasons:

1. The probe could be broken. The microstructure probes ([Figure 14](#)) used on Rockland instruments are very fragile and are designed to only make contact with fluids. The tips of the probes should NOT be touched or wiped with a cloth, and care should be taken when installing and removing probes. Each probe undergoes thorough pressure and functionality testing before being approved for use. However, despite best practices, it is still possible that the probes can become damaged or broken during deployment and recovery operations, or even from contact with debris or plankton.
2. The `setup.cfg` file could have an error. This could cause the data to be improperly converted from raw counts into physical units. An error in the address matrix could result in 'bad buffers', check the `logfile.txt` in the data directory on the instrument for evidence of bad buffers.
3. There could be a problem with the instrument hardware. Check the `logfile.txt` for any records of Bad Buffers. If you have a internally recording instrument, bad buffers are either a mistake in the `setup.cfg` file, or a problem with the instrument electronics. Contact Rockland for assistance troubleshooting bad buffers.
4. The data could be contaminated by external electronic or mechanical sources.

The following sections outline several steps to try to identify the cause of the poor data quality for each of the sensor types.



See the VMP-250 User Manual for more detailed troubleshooting information

11.1 Shear Probes

Identifying a Broken or Damaged Probe

The shear probe is composed of a silicone tip surrounding a Teflon mantle that houses a piezo-ceramic beam ([Figure 13](#)). The beam is as fragile as a potato-chip. If you suspect damage to your shear probe, begin troubleshooting by visually inspecting the probe ([Figure 49](#)). In rare cases, seawater can leak past the epoxy filling and create a short in the probe.



Figure 49: Visibly Damaged Shear Probe

The health of shear probes can also be assessed with the use of a Giga-Ohm meter and high quality Digital Multi-Meter. NOTE: Be very careful **not to damage the pin in the SMC cable when testing probes** – it is easy to damage a probe while testing if you use improper equipment. Rockland Scientific recommends using an SMC adaptor cable between the probe and the test equipment.

Shear probes should have:

- A resistance above 50 G Ω . A high resistance indicates that the biomorphs are well insulated and hence will yield high signal to noise ratios. Values below 50 G Ω could indicate water ingress and a short in the probe caused by seawater, which is a problem that cannot be fixed.
- A capacitance above 0.7 nF. A lower capacitance value could indicate that the probe was cracked and therefore would no longer yield accurate signals in response to the turbulent fluctuations.



To check the resistance, use a Giga-ohm meter with an input voltage of no more than 50 V. Otherwise you will damage the probe.



To avoid electric shock do not hold the shear probe when testing the resistance with a Giga-ohm meter.

Reviewing data

If data from two (or more) shear probes deployed on an instrument at the same time consistently do not agree across one or more profiles there is cause to investigate. Begin by confirming that the correct sensitivity value (**sens**) for each probe has been entered in the **setup.cfg** file. If the shear probe is more than a year old and has been well used, the probe may need to be re-calibrated. Rockland Scientific recommends that shear probes be calibrated once a year. If the correct **sens** value has been entered in the **setup.cfg** file and a difference of greater than 20% is still observed between probes then the probe may be damaged.

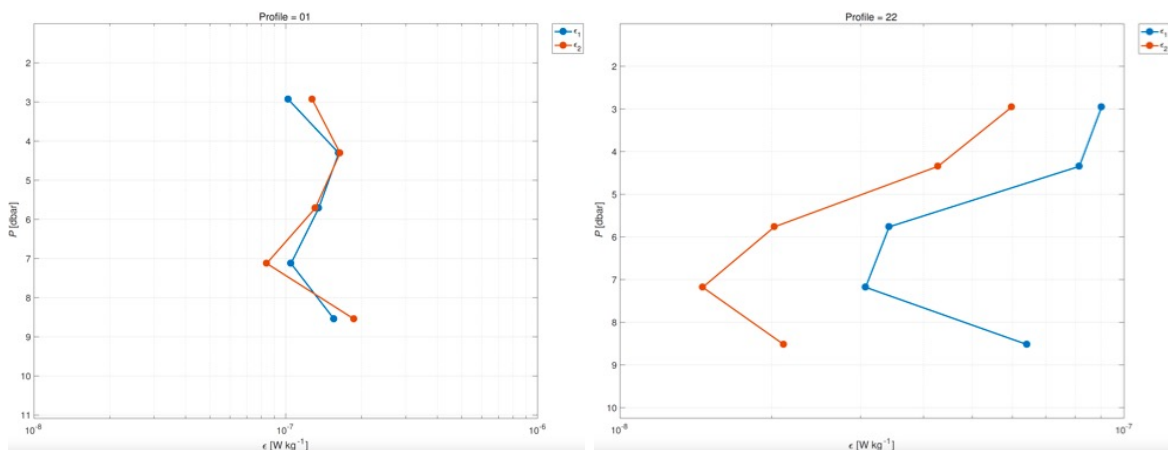


Figure 50: Comparison of dissipation rate estimates for two profile. In profile 01 (left), the estimates from the two probes agree well, whereas in profile 22, the estimates consistently disagree.

It is difficult to determine if a shear probe is broken or damaged by reviewing a bench test file with the probe installed. Evidence of damage can be better observed by reviewing data files taken during deployment. The time series data (in raw counts) can reveal several issues. Possible signs of concern and their associated explanations are as follows:

- *Repeating and regular spikes may be caused by electronic or mechanical noise.* If the same signal appears in the Ax and Ay channels there is a good chance that the noise is from a mechanical source such as loose components on the VMP or platform (i.e glider). Seaweed attached to the shear probe is an other possible source. If the spike is very consistent in frequency and amplitude it is more likely to be noise from an electronic source.
- *Increased signal in one shear probe compared to the other shear probe may indicate a damaged probe or SMC cable.* If the piezo-ceramic beam is badly damaged then it may cause increased noise in the signal. If the beam has a small crack, the increased noise may be intermittent. Increased signal in one shear channel may also be caused by corrosion on the SMC cable. Run an electronics bench test with test probes to confirm if the shear channel is behaving normally. Remember that the test probes can become corroded too. If you suspect the test probe has corrosion causing noise in the channel, try swapping the S1 and S2 test probes to see if the noisy signal changes channels as well.
- *Signals that are consistently one-sided may indicate that there is seawater leaking inside the sting and creating a short.* Shear probes are an AC signal and you should always expect to see a plus and minus side to any signal.
- *A reduced or minimal signal may be due to a poor connection between the probe and the SMC cable.* Check to see that the SMC cable is properly connected to the probe. If the problem persists check to see if the SMC cable is properly connected to the ASTP board.

11.2 FP07 Thermistors

Identifying a Broken or Damaged Probe

The FP07 temperature probe is extremely fragile. The sensing tip is coated in a thin layer of glass that can easily break if touched or if accidentally brushed with loose clothing. [Figure 16](#) shows an FP07 under a microscope. If the glass is cracked, the probe will experience water ingress at the probe tip and cause an intermittent short resulting in spikes the data.

The FP07 is a negative coefficient thermistor meaning low resistance corresponds to a high temperature, whereas high resistance corresponds to a low temperature. That being said, our circuitry is designed so that increasing raw counts is an increase in temperature and a decrease in raw counts is a decrease in temperature. The relationship between raw counts and temperature (°C) is roughly linear. The two extremes are:

1. The least possible resistance is caused by a short via seawater. This will show an increase in raw counts and an increase in the temperature value.
2. The highest possible resistance is caused by an open loop, which occurs when the small black sensing tip is broken off ([Figure 51](#)). This will show as a decrease in raw counts and a decrease in the temperature value (typically -17°C).

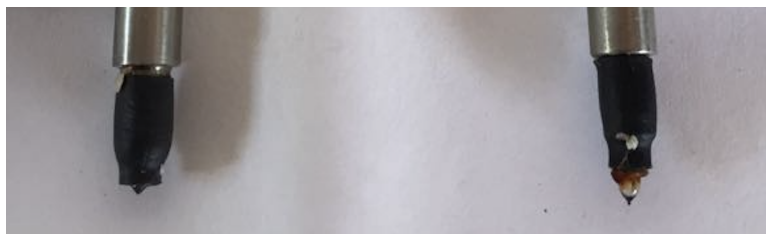


Figure 51: FP07 probe missing sensing tip (left) and FP07 in good condition (right).

The health of temperature probes can be assessed with a high quality digital multimeter (Fluke brand). **Note:** It is easy to damage a probe while testing if you use improper equipment. **Be very careful not to damage the pin in the SMC cable when testing probes.** Rockland Scientific recommends using an SMC adaptor cable between the probe and the test equipment.

Temperature probes should have a resistance between 1.5 k Ω and 2.5 k Ω . If a temperature probe has a resistance less than 1.5 k Ω there is most likely a short in the probe caused by seawater. Temperature test probes, on the other hand, contain a 3 k Ω resistor.

In rare cases, stray voltage from a broken probe tip has led to a small amount of calcification forming on the tip of the FP07 probe. If you observe a small white bead on the probe after a longer deployment (such as a glider deployment) contact Rockland customer support with a photo.

Reviewing Data

Reviewing data files from the previous deployment is recommended to confirm the FP07 temperature probe is behaving correctly. If a crack in the glass tip exists, the temperature probe may behave normally in air when dry; however once a probe with a crack in the glass tip is deployed, seawater will eventually leak inside the probe and create intermittent spikes. Comparing the data to a reference signal, such as that from a JFE Avantech temperature sensor (T_{JAC}), is the easiest way to detect problems. Three figures are included to highlight three scenarios:

- [Figure 52](#) shows a healthy FP07 probe. The offset between $T1_{fast}$ and T_{JAC} is simply because the thermistor is not yet calibrated. Note that the water temperature at the deployment site is high and near the upper range of the FP07 probe.
- [Figure 53](#) shows a FP07 probe with the black sensing tip completely removed. A complete short is made providing no resistance and resulting in the maximum possible negative number.
- [Figure 54](#) shows a FP07 probe with a suspected crack in the glass tip. The intermittent spiking to lower temperatures indicate shorting caused by seawater slowly leaking into the glass tip of the probe. Data appears to be valuable between 4 and 10 dbar.

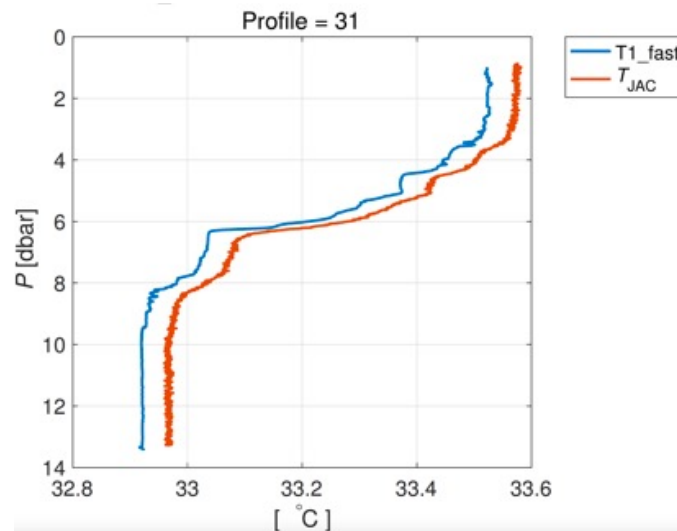


Figure 52: Temperature profiles from a healthy FP07 probe (blue) and reference JAC-T (red).

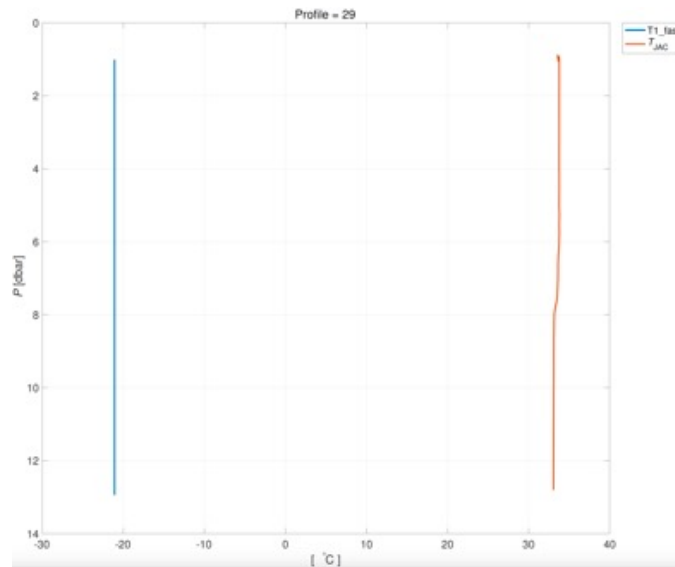


Figure 53: Temperature profiles from a broken FP07 probe in T1 (blue) and reference JAC-T (red).

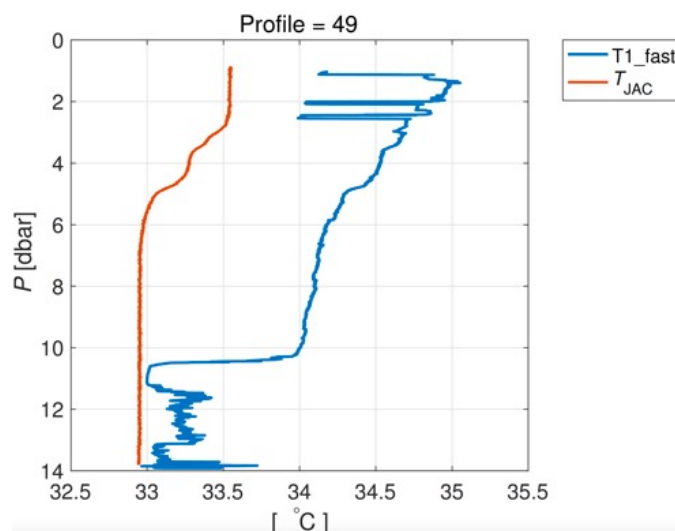


Figure 54: Temperature profiles from an FP07 probe with a suspected crack in the glass tip (blue) and reference JAC-T (red).

11.3 Micro-Conductivity Probes

Identifying a Broken or Dirty Probe

The SBE7 Microstructure Conductivity Probe¹⁷ is fragile and should not be touched or wiped with a cloth. Damage to the SBE7 Microstructure Conductivity Probe can typically be observed visually. Bent or scrapped sensing tips indicate damage. The sensing tips are coated in platinum black which appears jet-black in colour. If the tips appear discoloured, soak the probe tip with 1% Triton X-100 solution for ten minutes. For a detailed description of cleaning, refer to Rockland Scientific's Technical Note 033 "Probe Preparation and Treatment of the SBE7 Micro-Conductivity Sensor".

¹⁷Manufactured by Sea-Bird Electronics, USA.

12 Additional Materials

12.1 Rockland Web Resources

Please visit www.rocklandscientific.com to learn more about available web resources.

12.2 Manuals

The Rockland technical manuals pertinent to your instrument are

- VMP-250-IR Instrument Manual
- ODAS5-IR User Guide
- ODAS Matlab Library User Guide
- Zissou Essentials User Guide (also embedded in software)

12.3 Technical Notes

The Rockland Scientific website provides “Technical Notes” on a range of relevant topics. Technical Notes of particular interest to this application include:

- *TN-002: Digital Signal Processing to Enhance Oceanographic Observations*
A discussion of the theory behind pre-emphasis.
- *TN-005: Converting signals into physical units*
An explanation of how to use the shear probe calibration, electronics calibration, and other information to produce signals in physics units.
- *TN-011: Serial Instrument Bus Protocol*
A description of the signalling portal used with the 16-bit bi-directional Serial Instrument Bus of the ODAS system.
- *TN-015: Modeling the Spatial Response of the Airfoil Shear Probe*
A discussion of the wavenumber response of the shear probe.
- *TN-028: Calculating the Rate of Dissipation of Turbulent Kinetic Energy*
A method for calculating the rate of dissipation of turbulent kinetic energy from the shear measured with an air-foil shear probe.
- *TN-030: On the Forms of the Velocity, Shear, and Rate-of-Strain Spectra*
- *TN-039: A Guide to Data Processing*
A guide to using the ODAS Matlab library to process Rockland Scientific data files, including how to perform an *in situ* calibration of the FP07 thermistor using data from a Sea-Bird CT or JFE Advantech CT sensor.
- *TN-040: Noise in Temperature Gradient Measurements*
Provides a method for estimating the noise in measurements of the gradient of temperature fluctuations obtained with a Rockland instrument.
- *TN-042: Noise in Shear Probe Measurements*
Provides a method for estimating the noise in measurements of the velocity shear obtained with a Rockland instrument.

- *TN-046: Rockland Inclonometers*
Describes the functionality of the inclinometer in Rockland instruments.
- *TN-047: Why Calibrate the FP07 Thermistor?*
Describes the reasons for and against calibrating the FP07 thermistor.
- *TN-048: Interpreting the Results of Calibrate All*
Provides a guide to interpreting the results of the `cal -all` function for common instrument channels found on standard Rockland Instruments.
- *TN-049: Tow-Yoing with the VMP-250-IR*
Describes deployment technique of tow-yoing that is possible with the VMP-250-IR

12.4 Turbulence Literature

Additional reading material of high-quality include:

- “Marine Turbulence” by Baumert, Simpson and Sündermann (Baumert et al., 2005) .
This is a very comprehensive text on ocean turbulence, covering theory, techniques, observations and models. It is the result of the European CARTUM project, and has chapters authored by 53 active scientist from Europe, North America, Asia and Australia. Expensive but worth it.
- “The Turbulent Ocean” by Thorpe (Thorpe, 2005).
A comparatively short treatise on ocean mixing including the role of internal waves in driving ocean mixing.
- “Turbulence” by Hinze (Hinze, 1959).
A nice book on the fundamental equations of turbulence. Highly focused on laboratory measurements.
- “Turbulent Flows” by Pope (Pope, 2009).
A comprehensive treatment of turbulence theory and measurements in the laboratory, atmosphere, and ocean.

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