TURBULENT ENERGY DISSIPATION
OVER THE CONTINENTAL SHELF

by

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Abstract

A free-falling instrument was used in coastal waters to measure turbulent velocity fluctuations, temperature, conductivity and pressure from the near surface to 15 cm above the bottom. A probe guard system has been developed that protects the delicate temperature and shear sensors from bottom sediments and minimizes instrument vibrations that would otherwise contaminate the shear signal. From the shear signal the viscous dissipation rate of turbulent kinetic energy is calculated.

A new technique is presented for the analysis of shear spectra for dissipation rate calculations. The identification and elimination of noise, both at low and high frequencies, is accomplished by a positive feedback loop during analysis and insures more accurate estimates of the microstructure shear variance, \( \left( \frac{\partial u}{\partial z} \right)^2 \). The technique improves the confidence in the dissipation rate estimates and results in a noise level of \( 3.0 \times 10^{-7} \text{ W m}^{-3} \). This noise level is low, considering the structural modifications made to the profiler for near-bottom sampling.

The microstructure instrument was used for 670 profiles over the continental shelf west of Vancouver Island in June, 1985. From the dissipation rates near the bottom, within the constant stress layer, values of the bottom stress are calculated. Variations in the bottom stress and the height of the turbulent bottom boundary layer correlate with the diurnal tidal currents that dominate the flow near the bottom. The height of the bottom well mixed layer was found to be nearly independent of the height of the turbulent bottom boundary layer. Over most of the shelf, vertical density variations are attributable to advection rather than local mixing. Near shore, in depths less than 100 m, the tidally driven turbulent bottom boundary layer extends throughout most of the
water column during periods of maximum tidal current.

Seaward of the 100 m depth contour the current and density measurements above the bottom boundary layer, 40 to 50 m above the bottom, reveal the mean structure of the Tully eddy. Contours of constant density show that the structure is an upwelling centre confined to a region over part of the Juan de Fuca Canyon system. Turbulent mixing within the core of the eddy was found to be weak.

Oxygen samples indicate that the wind-induced upwelling brings slope water up the canyons to the shallow (<100 m) banks near shore. Nutrients in the slope water are mixed vertically by the tidally driven bottom boundary layer over these banks. Flux rates for $NO_3$ of 387 mmole s$^{-1}$ per metre of coastline are estimated during the strong upwelling conditions in June, 1985.

From the turbulent dissipation rate measurements within the bottom boundary layer an estimated lower limit to the decay scale for the $K_1$ period shelf wave is roughly put at 1100 km. This is in good agreement with the model of Brink (1982a). From the dissipation rate measurements above the bottom boundary layer, a friction decay time scale for the Tully eddy is estimated to be 231 hours. This is supported by the observations of Freeland and McIntosh (1987, personal communication) that show large, frequent fluctuations in the circulation at periods of ~ 20 to 330 hours. A global dissipation of $4.8 \times 10^{10}$ W is estimated for the tides over all continental shelf regions, only 2.5% of the total tidal kinetic energy dissipated by friction in all the oceans and seas.
# Table of Contents

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abstract</td>
<td>ii</td>
</tr>
<tr>
<td>List of Tables</td>
<td>vii</td>
</tr>
<tr>
<td>List of Figures</td>
<td>viii</td>
</tr>
<tr>
<td>Acknowledgements</td>
<td>xii</td>
</tr>
<tr>
<td>1. Introduction</td>
<td>1</td>
</tr>
<tr>
<td>2. Background</td>
<td>4</td>
</tr>
<tr>
<td>2.1 Oceanic Turbulence</td>
<td>4</td>
</tr>
<tr>
<td>2.2 The Continental Shelf Bottom Boundary Layer</td>
<td>4</td>
</tr>
<tr>
<td>2.3 Regional Oceanography of the</td>
<td>6</td>
</tr>
<tr>
<td>Continental Shelf West of Vancouver Island</td>
<td></td>
</tr>
<tr>
<td>3. Microstructure Measurements</td>
<td>12</td>
</tr>
<tr>
<td>3.1 The FLY II Profiler</td>
<td>12</td>
</tr>
<tr>
<td>3.2 Processing the FLY II Signals</td>
<td>24</td>
</tr>
<tr>
<td>3.3 Spectral Analysis and Energy Dissipation Rates</td>
<td>30</td>
</tr>
<tr>
<td>3.4 Noise Reduction in FLY II</td>
<td>42</td>
</tr>
<tr>
<td>3.5 Dissipation Rate Profiles</td>
<td>49</td>
</tr>
<tr>
<td>3.6 Additional Measurements</td>
<td>51</td>
</tr>
</tbody>
</table>
4. The Bottom Boundary Layer: Theory .................................................. 55

4.1 Introduction ................................................................. 55

4.2 Structure and Dynamics ..................................................... 56

4.3 Eddy Coefficients ............................................................. 66

4.3a Eddy Coefficient for Momentum, $K_m$ ................................ 66

4.3b Eddy Coefficient for Density, $K_{\rho}$ ............................... 69

5. Observations of the Bottom Boundary Layer .............................. 72

5.1 Introduction ................................................................. 72

5.2 Bottom Stress and Drag Coefficient Estimates ....................... 72

5.3 The Time Dependent Boundary Layer at M2 .......................... 93

5.4 The Constant Dissipation Rate Layer ................................... 108

5.5 Conclusions ................................................................. 112

6. Distribution of Microstructure Over the Continental Shelf .......... 115

6.1 Introduction ................................................................. 115

6.2 The Distribution of Kinetic Energy Dissipation ..................... 115

6.3 The Distribution of Mixing .............................................. 125

7. Tidal Dissipation ............................................................. 146

7.1 Introduction ................................................................. 146

7.2 Shelf Wave Decay ......................................................... 146

7.3 Secular Deceleration Due to Tidal Friction ......................... 150
8. Upwelling and The Tully Eddy ............................................ 153
   8.1 Introduction ......................................................... 153
   8.2 Nutrient Flux Rate ................................................. 153
   8.3 Eddy Decay .......................................................... 156

9. Conclusions ............................................................... 19
   9.1 The Bottom Boundary Layer ...................................... 159
   9.2 The Continental Shelf ............................................. 161

10. References ............................................................... 163

Appendix A: Error Analysis for Shear Probe $\epsilon$ and $u_*$ Calculations .............. 174

Appendix B: Calculations for Temperature, Conductivity, Salinity,
Depth, Density and $\sigma_t$ ............................................ 184

Appendix C: Tidal Constituents form Current Survey ........................................ 188
List of Tables

1. Instrumentation on the **FLY II** profiler. ........................................ 14

2. Composition of the bottom sediments. .............................................. 53

3. Individual $\epsilon$ values from profile 506 at M2, and the calculated $u_*$ and CSL height values. ....................................................... 83

4. Comparison between CSL (5.2.7) and CDRL (4.3.6) analysis at station M2. . 112

5. Summary of the average dissipation rates and eddy coefficients $\overline{K}_p$ over the shelf west of Vancouver Island. .......................... 122
List of Figures

1. The continental shelf region west of Vancouver Island. .......................... 7
2. Colour enhanced infrared AVHRR image taken at 1258 PST June 16, 1985. ... 9
3. Colour enhanced infrared AVHRR image taken at 0257 PST June 17, 1985. .... 10
4. The FLY II microstructure profiler. ...................................................... 13
5. a) The bottom portion of the FLY II profiler
    b) The flow \( \mathbf{V} \) seen by the shear probe is composed of the fall speed \( \mathbf{W} \) and the horizontal turbulent velocity \( \mathbf{u} \). ................................. 16
6. A short section of the calibrated conductivity and slow response temperature signals from FLY II ................................................................. 21
7. Examples of temperature, salinity and \( \sigma_t \) profiles calculated from data obtained with the FLY II profiler. ..................................................... 23
8. Nasmyth's (1986, personal communication) spectral data fitted with a cubic spline and the cumulative area under the spline. ......................... 34
9. An example spectrum (authors' conception) showing the selection of the best integration band \( k_i^1 \) to \( k_i^n \). ....................................................... 35
10. Band-averaged spectral values from an average of 21 individual 1024 point spectra from the range of dissipation rates between \( 10^{-5} \) to \( 10^{-4} \) W m\(^{-3}\). .................. 39
11. Band-averaged spectral values from an average of 36 individual 1024 point spectra from the range of dissipation rates between \( 10^{-6} \) to \( 10^{-5} \) W m\(^{-3}\). .................. 40
12. Band-averaged spectral values from an average of 41 individual 1024 point spectra from the range of dissipation rates below \( 10^{-6} \) W m\(^{-3}\). .................. 41
13. A spectrum calculated from the shear probe signal when the profiler was fitted with the original probe guard. .......................................... 43
14. a) The first probe guard design, which proved to be very noisy.
    b) The second and smaller probe guard which was then modified and used for all the bottom boundary layer studies. .......................... 44
15. The lower portion of the smaller probe guard (Figure 14b) showing the fairing added to the back of the cylinder and the double helical strakes. ......... 47
16. a) Log(ε) values calculated from the shear time series. 
b) The same data in a), with a log ordinate scale. 50

17. The 42 stations of CTD and Oxygen data, 14-15 June 1985. 52

18. The Hecate Strait region east of the Queen Charlotte Islands. 55

19. a) A dissipation rate profile of average dissipation rate values from three consecutive profiles through the bottom boundary layer. 
b) The same data in a), with a log ordinate scale. 86

20. Two dissipation rate profiles calculated from two shear probes mounted side-by-side during a single profiler descent. 87

21. An example of an averaged dissipation rate profile. 88

22. Time series of the average friction velocity $\overline{u}_r$ and the average current magnitude $|\overline{U}|$ at 3 m height from stations M2 and C2 respectively. 89

23. A $\sigma_t$ profile showing the layer of weakly stratified water immediately above the bottom. 94

24. a) A 36 hour portion of the current records from mooring C2. 96

b) A 36 hour portion of the predicted tidal currents from the constituents extracted using harmonic analysis. 98

c) The residual currents after subtracting the predicted tidal component (Figure 24a) from the raw current data (Figure 24b). 99

25. Time series of the well mixed layer (wml) height and the turbulent bottom boundary layer height. 100

26. a) Time series of average dissipation rate profiles, and 
b) individual $\sigma_t$ profiles at station M2. 103

27. a) An example of an average dissipation rate profile with (4.3.6) fitted to the lower portion. 110

b) Another example profile, similar to that shown in a). 111

28. Contour plot of the log of dissipation rates from the time series of profiles made at M1. 116
29. Contour plot of the log of dissipation rates from the time series of profiles made at M2. .................................................. 117
30. Contour plot of the log of dissipation rates from the time series of profiles made at M3. .................................................. 118
31. Contour plot of the log of dissipation rates from the time series of profiles made at M5. .................................................. 119
32. Contour plot of the log of dissipation rates from the time series of profiles made at M6. .................................................. 120
33. Contour plot of the log of dissipation rates from the time series of profiles made at B2. .................................................. 121
34. The dynamic heights calculated between 40 and 100 m depth from the CTD data collected during the CTD and Oxygen survey. .................................................. 123
35. The $\sigma_t$ contours at a depth of 5 m. .................................................. 127
36. The $\sigma_t$ contours at a depth of 25 m. .................................................. 128
37. The $\sigma_t$ contours at a depth of 50 m. .................................................. 129
38. The $\sigma_t$ contours at a depth of 75 m. .................................................. 130
39. The $\sigma_t$ contours at a depth of 100 m. .................................................. 131
40. The $\sigma_t$ contours at a depth of 125 m. .................................................. 132
41. The $\sigma_t$ contours at a depth of 149 m. .................................................. 133
42. The $\sigma_t$ contours at a depth of 170 m. .................................................. 134
43. The $\sigma_t$ contours at a depth of 175 m. .................................................. 135
44. The $\sigma_t$ contours at a depth of 200 m. .................................................. 136
45. The dissolved oxygen contours at a depth of 50 m. .................................................. 137
46. The dissolved oxygen contours at the bottom. .................................................. 138
47. Contours of the log of the vertical eddy coefficient for density $\bar{K}_\rho$ at M3. .................................................. 139
48. Contours of the log of the vertical eddy coefficient for density $\bar{K}_\rho$ at M5. .................................................. 140
49. Contours of the log of the vertical eddy coefficient for density $\bar{K}_\rho$ at M1. .................................................. 141
50. Contours of the log of the vertical eddy coefficient for density $\overline{K}_\rho$ at M2. . . 142

51. Contours of the log of the vertical eddy coefficient for density $\overline{K}_\rho$ at M6. . . 143

52. Contours of the log of the vertical eddy coefficient for density $\overline{K}_\rho$ at B2. . . 144
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1. Introduction

The continental shelf is that part of the ocean that extends seaward from the shore for approximately 50 km, and has a gradual bottom slope from 0 to 200 m depth. The edge of the shelf is called the shelf break and is well approximated by the 200 m contour. Beyond the shelf break the bathymetry drops away rapidly, a region called the continental slope. The continental shelf west of Vancouver Island supports a variety of active oceanic flows. The currents are influenced by the earth’s rotation, the density stratification, the atmospheric wind, the bottom topography and the tidal forces.

The physical oceanography on the shelf influences several important oceanic processes. Of key importance to biological processes is the fact that much of the primary productivity in the oceans occurs in continental shelf waters. This is a result of the mixing and advection of nutrients onto the shelf. The shallow depth of the shelf region means that friction and boundary effects have an increased influence on the flow. Similarly, the tidal currents are faster over the shelf than in the deep ocean, and the entire water column can be dominated by the effects of friction. The mixing of quantities, and the importance of friction on the flow are related to the intensity of the turbulence.

The study of fluid flow over a solid surface, or near the interface of two fluids is the study of boundary layer dynamics. Geophysical boundary layers include the flow of the atmosphere over the land and sea, the dynamics of the upper ocean and the oceanic bottom boundary layer. Much research has been done on each of these boundary layers, both theoretically and experimentally. The bottom boundary layer over the continental shelf is perhaps one of the most complicated and diverse boundary layers and is certainly the most dynamic region of the benthic environment.
This work is an attempt to study the turbulent structure over the continental shelf west of Vancouver Island, with particular attention to the structure and role of the turbulent bottom boundary layer. The approach is essentially observational, with theoretical considerations added to entrench the measurements into the framework of previous studies and to relate the observed structures to basic physical principles. The adaptation of a free-falling profiler for near-bottom studies of oceanic turbulence has provided a unique set of measurements of the tidally driven bottom boundary layer. The measurements of turbulent fluctuations and water properties provided by the profiler reveal the full vertical structure of the entire water column, and of the boundary layer down to a height of 15 cm above the bottom. Repeated profiles in time provide information on the evolution of the time-dependent boundary layer and assist in understanding the physical importance of the turbulent boundary layer to the general oceanography of the shelf region.

The mechanics of turbulence is not completely understood and I shall draw on both early theories and recent technologies in an attempt to measure and interpret the microstructure. Of primary importance is the quantitative evaluation of the dissipation rate of turbulent kinetic energy, the distribution of turbulent stresses and the turbulent diffusion of ocean scalars. The distribution of turbulent dissipation over the shelf is unknown. It is suspected that turbulence within the bottom boundary layer is one of the controlling mechanisms of the large scale coastal circulation. The turbulent Reynolds stress \(-\rho \overline{w'w'}\) is the quantity responsible for extracting momentum from the mean flow, but it is difficult to measure. The dissipation rate of turbulence \(\varepsilon\) is the easiest property of turbulence to measure, and can be used to estimate mixing rates and energy levels of turbulence.
The circulation on the continental shelf southwest of Vancouver Island is composed of two currents: a quasi-geostrophic current established by wind and estuarine forcing, and a tidal current. The quasi-geostrophic current is dominant in the surface layer (< 75 m). Quasi-geostrophic currents are slowly varying, frictional geostrophic currents, and the non-tidal circulation will be called geostrophic for short throughout the remainder of this thesis. The tidal currents are predominantly diurnal (Crawford and Thomson, 1984) and dominate the motion at depth. Of primary concern is the inter-dependency of the turbulence and the circulation, and the distribution of turbulent mixing at all depths over a shelf during active upwelling. These and a detailed observation of a time-dependent boundary layer and the microstructure profiler FLY II will be presented.

Of key importance to the final results are; the design and evaluation of several probe guard systems, the completion of two pilot cruises to the shelf in 1984 to determine efficient data sampling procedures and to find appropriate locations for microstructure stations, the design and implementation of a multi-phased oceanographic survey in 1985, the development of detailed spectral analysis techniques for analyzing turbulence data, and the interpretation and presentation of unique microstructure data time series. The microstructure survey in 1985 consisted of; 4 current moorings with a total of 24 current meters, a CTD survey from 42 stations, an oxygen survey with 256 samples, and 676 microstructure profiles from 7 locations. In addition to processing the 676 microstructure profiles collected in 1985 from the continental shelf west of Vancouver Island, 265 microstructure profiles were processed from the pilot cruises in 1984, and 185 profiles from a 1983 cruise to Hecate Strait. The complete data set is very extensive, and analysis of all these data has not been attempted in this dissertation.
2. Background

2.1 Oceanic Turbulence

The role of turbulence in the ocean is to simultaneously act as a moderator of the large scale dynamics and as an agitator of the small scale structure. Turbulence is the friction that works against ocean currents, gyres and eddies. It extracts and redistributes momentum from these large structures, cascading the energy to smaller scales. The chaotic motions of turbulence mix fluids and dissipate energy, bringing a homogeneity to the ocean that would otherwise be lacking.

It is only recently that techniques have been developed for extensive measurements of oceanic turbulence. Improvements in instrumentation, data recording and spectral analysis have permitted oceanographers to observe the distribution of turbulence and the temporal variations of the structures and dynamics governing that distribution. Free falling profilers are perhaps one of the most successful developments. Modifications to the FLY profiler at the Institute of Ocean Sciences are discussed. These modifications have permitted some of the first measurements of the highly variable turbulent bottom boundary layer over the continental shelf west of Vancouver Island.

2.2 The Continental Shelf Bottom Boundary Layer

The structure of the bottom boundary layer is strongly dependent on the dynamics and the distribution of water masses, and inversely, the dynamics and structure of the shelf waters are determined by the mixing and dissipation that take place within the bottom boundary layer. These two main properties of turbulence, its ability to mix and redistribute water properties, and its ability to dissipate kinetic energy from the large scale flow, will be discussed with respect to the oceanography over the continental shelf.
In a detailed review of the continental shelf bottom boundary layer, Grant and Madsen (1986) express repeated need for research such as that conducted as part of this thesis. They note, "No complete data set on stably stratified continental shelf bottom boundary layers has been published to date." Measurements described here, made with a free-falling turbulence profiler and current meter moorings, reveal a highly complex vertical boundary layer structure. The observations provide a unique look at the distribution of turbulence from the near surface (depth ~ 10 m) to within 15 cm of the bottom at depths ranging from 60 to 200 m. These measurements are some of the first to reveal the vertical structure of the turbulent bottom boundary layer, and show the influence of advection of horizontal density variations on the bottom well mixed layer.

The measurements and analysis presented are intended to provide a badly needed investigation into the full structure of the turbulent boundary layer. Previous work has been limited to near-bottom measurements of velocities within the inertial subrange (Grant et al. 1984; Cacchione and Drake, 1982; Grant and Madsen, 1986) or to outer-layer measurements without direct stress or near-bottom measurements of the turbulent velocities (Weatherly and Van Leer, 1977; Dickey and Van Leer, 1984). Many models of the stably stratified oceanic boundary layer exist (Businger and Arya, 1974; Weatherly and Martin, 1978; Weatherly, Blumsack and Bird, 1980; Mellor and Yamada, 1974; Long, 1981; Richards, 1982; Soulsby, 1983) but have few measurements to support and justify the closure techniques and vertical scaling used. Data presented here shed light on what was previously either assumed, or ignored.
2.3 Regional Oceanography of the Continental Shelf

West of Vancouver Island

The region of study is the southern portion of the continental shelf west of Vancouver Island (Figure 1). No microstructure measurements were made with the present system on the continental slope, west of the shelf break (200 m), but Lueck, Crawford and Osborn (1983) report on 13 profiles made over the slope in 1980 using Camel II. The microstructure measurements made with FLY II, and the current observations provided by the 4 moorings C1-C4, were collected in June, 1985. The conditions over the shelf may be substantially different at other periods of the year, as the circulation is known to vary considerably between different seasons (Freeland, Crawford and Thomson, 1984). The role of the turbulent bottom boundary layer and the distribution of mixing will therefore be different in the fall and winter, and the measurements presented here may not represent the conditions during these periods.

The circulation over the shelf region is dominated by two types of currents: 1) a slowly varying, quasi-geostrophic current (called geostrophic for short), and 2) the tidal currents. The geostrophic current is established by wind forcing and estuarine fluxes along the shore. The wind driven currents and the freshwater outflow have seasonal variations (Freeland et al. 1984). The mean (monthly) near-shore component of flow (e.g. near B2) is always to the northwest. The mean currents at the shelf break are towards the northwest during the winter months (October - March), and to the southeast in the summer (May-August). The months of April and September are transition periods between these two conditions. The tidal currents are dominated by diurnal fluctuations (Crawford and Thomson, 1984) and are typically responsible for 50% of the observed
Figure 1 The continental shelf region west of Vancouver Island. The microstructure stations used in June 1985 are identified by (○) and the current moorings by (▲).
current. This proportion increases nearer shore, and decreases seaward of the shelf break.

The influence of freshwater runoff on the circulation and its variability on the shelf is substantial (Freeland et al. 1984, Denman, Freeland and Mackas, 1982; Freeland and Denman, 1982; Emery, Thomas, Collins, Crawford and Mackas, 1986). The density structure is most variable during the summer, when the outflow of brackish water from the Strait of Juan de Fuca is a maximum and the wind-driven currents provide favourable conditions for coastal upwelling. The reader is referred to the colour-enhanced infrared AVHRR images on pages 13,085-13,086 of Emery et al. (1986). Similar images are reproduced here (Figures 2 and 3). These images were taken on June 16 and 17, 1985, during the period when microstructure data were collected at stations M6 and M2 respectively (Figure 1).

Freshwater runoff from the Fraser River reaches its maximum during June. The Fraser runoff, via the Strait of Juan de Fuca, accounts for more than 80% of the estuarine flow onto the shelf west of Vancouver Island in June (Freeland et al. 1984). The remaining freshwater sources are scattered along the west coast of Vancouver Island and in Barkley Sound and have peak outflow in November. Establishing the distribution of turbulence in the vicinity where this water meets the saline ocean is important in determining the controlling dynamics and the manner in which the fresh water is mixed and redistributed on the shelf.

One of the most important features of the circulation on the Vancouver Island shelf is the seasonal upwelling that brings nutrients onto the shelf during the summer months. The wind-driven southerly flow along the shelf break in the summer induces favourable upwelling conditions. During the week of June 16, 1985, the upwelling index (Bakun,
Figure 2 Reproduction of colour-enhanced infrared AVHRR image taken at 1258 PST June 16, 1985. The shades represent surface temperature variations where lighter is colder. Of interest is the cold water along the shore, the cold estuarine water within the Strait of Juan de Fuca, and the cold water identifying the Tully eddy just west of the enterance to the Strait of Juan de Fuca. The image was processed at the UBC Department of Oceanography Satellite Laboratory. The surface temperature scale is indicated in Fig. 3.

1973) reached a maximum for 1985, with a one-week average of approximately 100 m³/s per 100 m of coastline (Emery et al. 1986). The surface waters, bounded by the Vancouver Island shoreline and extending seaward for about 10 km, persistently show the presence of cold water, either upwelled by advection or by vertical mixing (Figures 2 and 3). Identifying the distribution of vertical mixing in the region and the circulation of the upwelled water is one of the goals of this thesis.
Figure 3 Reproduction of colour-enhanced infrared AVHRR image taken at 0257 PST June 17, 1985. The image was processed at the UBC Department of Oceanography Satellite Laboratory. Cold water near Vancouver Island is approximately 9.5 °C. A general shift to warmer temperatures from Fig 2 is a result of the combination of very calm seas and surface heating by solar radiation.

An interesting feature that occurs every summer on the shelf west of Vancouver Island is the Tully eddy. It is so named after the study of Tully (1942), who first reported its physical structure. Since then it has received much attention (Denman, Freeland and Mackas, 1982; Freeland and Denman, 1982 and 1983; Weaver and Hsieh, 1987). The actual driving mechanism for the eddy is not well understood, as it is most likely a combination of many mechanisms. Features of the Tully eddy are: 1) that it is seasonally driven and its appearance coincides with both the onset of strong upwelling conditions
and the maximum estuarine outflow, 2) it is topographically restricted to the shelf region at the entrance to the Strait of Juan de Fuca and south of La Pérouse Bank, and 3) it appears to be a centre for upwelling, distinct from the near-shore region mentioned above. The driving mechanisms associated with the Tully eddy will not be addressed here. Mixing within the Tully eddy and the role it plays in the upwelling of slope water will be discussed in §§6 and 8.

The distributions of density, oxygen and turbulent kinetic energy dissipation will be presented for the period of June 10 - 20, 1985. The microstructure instrument FLY II was modified for the work described here; in particular, a probe guard was added for near-bottom sampling. These modifications are outlined, as well as details of the data processing and calculations of the turbulent kinetic energy dissipation rate, in §3. A review of boundary layer dynamics is presented in §4. The observed density, current and turbulent distributions within the bottom boundary layer are presented in §5. A summary of the distributions of turbulence and mixing over the shelf region is presented in §6. Estimates of the kinetic energy dissipated from the tidal flow are made in §7 and from the Tully eddy in §8. Also presented in §8 are estimates of the rates of upward flux of nutrient due to turbulent mixing for the active upwelling period of June 10-20, 1985. Conclusions are given in §9.
3. Microstructure Measurements

3.1 The FLY II Profiler

Most measurements of oceanic fine structure (ie. temperature, salinity and velocity fluctuations with scales between 1 m and 100 m) and microstructure (scales smaller than 1 m) obtained using tethered profilers are limited to depths above the bottom boundary layer. A new profiling instrument has been developed which simultaneously measures temperature, salinity, turbulent shear \( \frac{\partial u}{\partial z} \) and depth from near the ocean surface to within 15 cm of the bottom. The notation \( u \) and \( u' \) will be used interchangeably to indicate turbulent, fluctuating horizontal velocity, upper case letters will denote mean, large scale velocities or the flow relative to the falling probe. Also, the notation \( u, v \) and \( w \) will indicate turbulent velocities in the \( x, y \) and \( z \) directions respectively, as will \( u_1, u_2 \) and \( u_3 \) if indices are used.

A probe guard permits this free-falling instrument to "land" on the ocean floor while protecting the sensitive probes from bottom sediments (Figure 4).

The probe guard design suppresses the shedding of large eddies which cause mechanical vibrations and are interpreted as noise in the turbulent shear signal. Following a profile, the profiler is retrieved using a neutrally buoyant Kevlar\textsuperscript{TM} line. The water column may be sampled quickly and without risk to probes, permitting repeated profiles in space and time. Only two previous studies have sampled turbulent features near the bottom using profilers. Lueck and Osborn (1985) used expendable dissipation profilers to obtain single turbulent profiles through the bottom boundary layer in a submarine canyon. Caldwell, Wilcox and Matsler (1975) used a tethered profiler to measure temperature microstructure, but not velocity microstructure, on the continental shelf.
Figure 4 The FLY II microstructure profiler. The probe guard fitted to the lower end protects the delicate temperature and shear sensors from bottom sediments.

The microstructure profiler FLY II (Fast Light Yo-yo) is illustrated in Figure 4 and discussed by Dewey, Crawford, Gargett and Oakey, (1987). FLY I was designed and built by N.S. Oakey and A.E. Gargett. Sensors on FLY II include airfoil shear probes (Osborn and Crawford, 1980), slow and fast response thermistors, a conductivity cell, two tilt gauges and a pressure gauge. Added for this study were the slow response thermistor,
the conductivity cell and the probe guard. Specifications of the probes are outlined in Table 1. The pressure case contains the signal amplifiers, analog-to-digital electronics and the battery-driven power supply. A 12-bit digitizer samples each of two "fast" channels and one of eight "slow" channels 274 times a second, for a data transmission rate to the surface of 822 data points per second. The net sampling rate for each of the two fast channels is 274 times a second and for each of the eight slow channels is 34 times a second. The order in which the slow channels are sampled is selected before profiling.

Digital signals are transmitted to the ship where they are recorded on cassette or 9-track tape and converted to analog signals for display on a chart recorder. Either two shear signals or one shear and one fast temperature signal are transmitted on the fast channels; combinations of fast and slow temperature, conductivity, two components of tilt, pressure and the internal battery voltage are transmitted on the slow channels. If two shear probes are used on the two fast channels, (as was the case in 1983 in Hecate Strait) then the fast temperature signal is sent on more than one of the slow channels to increase the sampling rate.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Manufacturer</th>
<th>Range</th>
<th>Precision</th>
<th>Response</th>
</tr>
</thead>
<tbody>
<tr>
<td>Conductivity</td>
<td>Sea-Bird Electronics, Inc.</td>
<td>26 to 60 mmho cm(^{-1})</td>
<td>±0.05 mmho cm(^{-1})</td>
<td>0.34 s</td>
</tr>
<tr>
<td>Thermistors</td>
<td>Thermometrics Inc.</td>
<td>1.5° to 13°C</td>
<td>±0.003°C</td>
<td>0.3 s</td>
</tr>
<tr>
<td>slow: P60</td>
<td></td>
<td>1.8° to 17°C</td>
<td>±0.004°C</td>
<td>0.018 s</td>
</tr>
<tr>
<td>fast: P20</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pressure gauge</td>
<td>Humphrey Inc.</td>
<td>0 to 250 m</td>
<td>±0.5 m</td>
<td>1.0 m</td>
</tr>
<tr>
<td>Tilt (pendulum)</td>
<td>Humphrey Inc.</td>
<td>0 to 45°</td>
<td>±0.5°</td>
<td>0.2 s</td>
</tr>
<tr>
<td>Shear probe</td>
<td>Under Sea Technology</td>
<td>0 to 4 s(^{-1})</td>
<td>±5%</td>
<td>1-2 cm*</td>
</tr>
</tbody>
</table>

* (Ninnis, 1984)

The probe guard is attached at the lower end of the profiler as shown in Figure 4,
and may be removed for studies where near-bottom sampling is not required. The shear probe(s) and fast response thermistor are located at the lower end of the pressure case (Figure 5a). Further along the profiler are the conductivity cell and the slow response thermistor. Two tilt gauges are mounted inside the case and a pressure sensor is mounted at the top of the profiler. External floats of syntactic foam are used to reduce the fall speed and to maintain a vertical orientation. A circular brush at the top is used to control further the fall speed and orientation of the profiler and to break up the larger eddies in the wake which would otherwise generate low frequency, large scale vibrations and contaminate the shear probe signal(s).

The **FLY II** profiler described here is a tethered instrument and to date has completed over 1500 surface-to-bottom profiles. No shear probes and only one of the fast response thermistors have been damaged during these boundary layer profiles. Fall speed, determined by the amount of floatation and drag, is typically set between 0.5 and 1.0 m s\(^{-1}\) and is calculated from the pressure record. The shock caused by the impact of the probe guard with the bottom appears as a spike in the shear signal and is used (together with the pressure signal) to identify the bottom during analysis. Once on the bottom, **FLY II** is winched to the surface, or to any intermediate depth, where it is ready for further profiling. Valid microstructure signals are measured only during the descent.

The shear probes resolve horizontal velocity fluctuations whose vertical wavelengths are approximately between \(\sim 0.01\) and 1.0 m. At a fall speed of 0.6 m s\(^{-1}\), these scales correspond to fluctuations at frequencies between 0.6 and 60 Hz. Although the response of the shear probe drops off at the higher frequencies (smaller wavelengths) due to spatial averaging over the probe, the true response to \(\sim 0.6\) cm wavelength is recovered by using
Figure 5  a) The bottom portion of the FLY II profiler. The conductivity cell and slow response thermistor were added for this study. The shear probe (left) and fast response thermistor (right) are located at the leading edge of the pressure case. b) The flow $\mathbf{V}$ seen by the shear probe is composed of the fall speed $\mathbf{W}$ and the horizontal turbulent velocity $\mathbf{u}$. 
a transfer function derived by Ninnis (1984). The resolution at low frequencies (longer wavelengths) is determined by the configuration of the pre-amplifier, the pitch and roll characteristics of the profiler system and a very slight temperature dependence in the shear probe calibration constant (Osborn and Crawford, 1980).

The shear signal is differentiated before digitization and represents microstructure shear fluctuations. From the differentiated shear signal and the fall speed as determined by the rate of change of pressure, the dissipation rate of turbulent kinetic energy can be calculated. The first step is to estimate a single turbulent shear component from the shear probe signal. Equation (3.1.1) shows the dependence of the microstructure shear estimate $\frac{\partial u}{\partial z}$ on the fall speed $W$, the differentiator gain $G$, the calibrated shear probe sensitivity $S$ and the output voltage $V_0$ (Crawford, 1976).

$$\frac{\partial u}{\partial z} = \frac{1}{W} \frac{\partial u}{\partial t} = \frac{V_0}{G2\sqrt{2SW^2}}. \tag{3.1.1}$$

The total viscous dissipation rate is a sum of all the microstructure shear components, given by

$$\epsilon = \frac{1}{2\mu} \overline{\left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right)^2}, \tag{3.1.2}$$

where repeated indices indicates summation and the overbar represents the mean value with respect to time. Averaging in time is appropriate if the process is ergodic, where it is assumed that fluctuations at a fixed point are stationary random functions of time and a time average is identical to an ensemble average.

For isotropic turbulent regimes the total viscous dissipation rate per unit volume $\epsilon$ can be estimated by assuming that for isotropic turbulence (Hinze, 1975)

$$\left( \frac{\partial u}{\partial y} \right)^2 = \left( \frac{\partial u}{\partial z} \right)^2 = \left( \frac{\partial v}{\partial x} \right)^2 = \ldots$$
Thus a measurement of the single component \( \frac{\partial u}{\partial z} \) provides a direct estimate of a known fraction of \( \epsilon \), and by assuming isotropy, an estimate of the remaining contributions can be made. The result is the total dissipation rate given by,

\[
\epsilon = 7.5 \mu \left( \frac{\partial u}{\partial z} \right)^2,
\]

(3.1.3)

where \( \mu \) is the dynamic viscosity of sea water.

This is the rate of kinetic energy loss from the turbulent motions due to viscous forces. I have chosen to use the dynamic viscosity \( \mu \) and represent \( \epsilon \) as the dissipation rate per unit volume \( (W \ m^{-3}) \). This quantity is appropriate in a stratified environment where the density varies both in the horizontal and vertical directions. It also allows direct comparison with energy fluxes at the surface, or from known energy sources that are in units of Watts. Other researchers may choose to use the dissipation rate per unit mass (by dividing my \( \epsilon \) by the local density \( \rho \)), with the result in dimensions of \( L^2/T^3 \), (Watts/unit mass).

When using (3.1.1) and (3.1.3) to calculate the dissipation rate, the fall speed \( W \) must be known. To estimate \( W \) the pressure signal from a profile is fit to a least squares cubic curve, which is differentiated and the resulting quadratic gives the fall speed at every depth to within \( \sim 5\% \). The calibration constant \( S \) of the shear probe is sensitive to the angle of attack \( \alpha \) (i.e. the response is not a true cosine). Flow past the shear probe \( V \) is composed of the parallel component along the axis of the shear probe \( W \) (fall speed) and the fluctuating turbulent velocity component \( u \) perpendicular to \( W \). The angle of
attack \( \alpha \) is then the angle between \( W \) and \( V \) (Figure 5b). Deviation in the constant \( S \) is a few percent for \( \alpha = 5^\circ \) and \( \pm 5% \) for \( \alpha = 10^\circ \). The two tilt gauges located within the case monitor the amount of absolute tilt, which varies slowly and has never exceeded \( 5^\circ \). Strong turbulent velocities (\( u \sim 2.0 \text{ cm s}^{-1} \)) cause a value \( \alpha \sim 2^\circ \).

The variance \( \left( \frac{\partial u}{\partial z} \right)^2 \) required in (3.1.3) is normally calculated from the power spectrum of the microstructure shear time series given by (3.1.1). By calculating the power spectrum rather than simply summing the variance of the shear signal, the non-turbulent component of the shear signal (noise) can be identified (§3.3 and Appendix A) and losses due to the roll-off of the shear probe response at high frequencies can be corrected (Ninnis, 1984). Instrument vibrations, detected by the shear probe, that are within the turbulence bandwidth will be interpreted as small horizontal velocity components. These effects are considered noise and have been described in detail by Moum and Lueck (1985). Several modifications to FLY II were investigated to limit these vibrations and are discussed in §3.3.

There are two thermistors mounted on FLY II: a fast response thermistor (Thermometrics P20) near the shear probe(s) and a slower response thermistor (Thermometrics P60) near the conductivity cell. The faster response thermistor has a manufacturer's specified cold-plunge response time of 0.018 s, and may be used to measure temperature microstructure and turbulent overturns. For stability the fast thermistor signal was low-pass filtered at 10 Hz and the temperature is then resolved to approximately 6 cm in the vertical. The circuit has a precision of 4 millidegrees. The second thermistor is mounted near the conductivity cell (Figure 5a) and is used with the conductivity signal to determine salinity. The specified response time for this thermistor is 0.3 s, and the
circuit has a precision of 3 millidegrees. The position and characteristics of the slower thermistor were selected to provide accuracy, stability and a response similar to that of the conductivity cell. The slower thermistor is also more robust than the delicate fast thermistor and less susceptible to damage.

The conductivity cell is a Sea-Bird Electronics 4-4/A model. The accompanying electronics (Sea-Bird Eletronics 4-03) are mounted inside the FLY II pressure case. The manufacturer's quoted response time and precision of the conductivity signal are 0.34 s and ±0.05 mmho cm⁻¹ respectively for a flow rate of 1.0 m s⁻¹. Figure 6 shows a section of the conductivity and slow thermistor signals for a fall speed of 0.66 m s⁻¹.

The vertical shift in the conductivity signal must be taken into account when calculating the salinity from the conductivity and temperature signals. A vertical shift of ~ 10 cm is required to match the conductivity signal to the slow temperature signal. This shift in the conductivity signal corresponds to a relative delay of ~ 0.15 s in response. The absolute response time of the conductivity cell is then ~ 0.5 s at a fall speed ~ 65 cm s⁻¹.

For a free-falling profiler with a constant fall speed, the vertical displacement is nearly constant, permitting an accurate compensation when processing the signals. In a conventional CTD unit, winched from a rolling ship, the shift is not constant; therefore, free-falling profilers have the potential to produce more accurate salinity and density measurements with better vertical resolution. Temperature, salinity and σₜ (Figure 7) are calculated as outlined by Millero, Chen, Bradshaw and Schleicher (1980) and in Appendix B. The quantity σₜ indicates density (ρ) anomalies and is defined by σₜ = ρ - 1000 (in kg/m³). Unfortunately the gain in the conductivity circuit was set too coarse in 1985,
Figure 6 A short section of the calibrated conductivity and slow response temperature signals. The conductivity signal has been shifted 10 cm upward to compensate for the slower response of the conductivity cell.
and the precision in $\sigma_t$ is only $\pm \sim 0.02$. 
Figure 7 Examples of temperature, salinity and $\sigma_t$ profiles obtained with the FLY II profiler. The data is from station B2 on the west of Vancouver Island (Figure 1).
3.2 Processing the FLY II Signals

The stages involved in processing the FLY II signals are outlined here for completeness, and to provide a manual for future work. The first stage in the processing of the microstructure data collected with FLY II, after it has been transferred from cassette or Hewlett-Packard 9-track data tapes to Univac 1600 bpi tapes (a Sperry 1100/60 has been used for processing), is the trimming of excess data from the top and bottom of each profile and the reversing of the profiles. The profiles are trimmed because the data acquisition system is often turned on before the profiler is actually released from the surface and is turned off after the profiler has landed on the bottom. The profiles are reversed such that the data collected at the bottom occur first in the data file. Each scan of 24 channels must be reversed separately so that the order of the data remains sequential.

The top of the profile is identified in the pressure signal. After the pressure signal has increased by 1.5 decibars, the start of the profile is identified. All preceding data are truncated. The bottom of the profile is identified by two processes. First the pressure signal is monitored and the bottom is identified by the absence of further increase in the pressure. Secondly, the shear signal is analyzed for the spike which occurred upon impact of FLY II with the bottom. Since the shear channel is sampled at a rate of 274 readings per second and the pressure at only 34 readings per second, the "bottom spike" in the shear signal is a more accurate identifier than the pressure record. In addition, the pressure gauge only changes with every decibar so that the level at which pressure becomes a constant only identifies the bottom to within 1 metre. A large spike in the shear signal followed by a constant pressure identifies the bottom. Data beyond the last zero crossing preceding this spike are truncated.
The reversing of a profile and the sequencing of data is then only a book-keeping problem. The original channel sequence is now reshuffled to keep the sequential order. Each scan of 24 readings of the 8 channels then has a new order associated with it. This order change is summarized below.

New: 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24
Old: 22 23 24 19 20 21 16 17 18 13 14 15 10 11 12 7 8 9 4 5 6 1 2 3

The original data (old channel sequence) was stored in 1985 in the following channel order for each scan of 24 readings;

Channel

<table>
<thead>
<tr>
<th>Channel</th>
<th>Old</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shear 1:</td>
<td>readings 1, 4, 7, 10, 13, 16, 19, 22</td>
</tr>
<tr>
<td>Fast Temp:</td>
<td>readings 2, 5, 8, 11, 14, 17, 20, 23</td>
</tr>
<tr>
<td>Conductivity:</td>
<td>readings 3, 9</td>
</tr>
<tr>
<td>Slow Temp:</td>
<td>readings 15, 21</td>
</tr>
<tr>
<td>Pressure:</td>
<td>reading 6</td>
</tr>
<tr>
<td>Tilt 1 and 2:</td>
<td>reading 12 and 18</td>
</tr>
<tr>
<td>Reference voltage</td>
<td>reading 24</td>
</tr>
</tbody>
</table>

Following the trimming and reversing, the new data sequence per scan is;

<table>
<thead>
<tr>
<th>Channel</th>
<th>New</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shear 1:</td>
<td>reading 1, 4, 7, 10, 13, 16, 19, 22</td>
</tr>
<tr>
<td>Fast Temp:</td>
<td>reading 2, 5, 8, 11, 14, 17, 20, 23</td>
</tr>
<tr>
<td>Conductivity:</td>
<td>readings 18, 24</td>
</tr>
<tr>
<td>Slow Temp:</td>
<td>readings 6, 12</td>
</tr>
<tr>
<td>Pressure:</td>
<td>reading 21</td>
</tr>
<tr>
<td>Tilt 1 and 2:</td>
<td>readings 15 and 9</td>
</tr>
<tr>
<td>Reference voltage</td>
<td>reading 3</td>
</tr>
</tbody>
</table>

This newly ordered data is then stored on magnetic tape.

Spikes are removed from the shear time series after the digital signals have been separated, trimmed and reversed. The spikes are believed to be caused by internal and external events during the profiler's descent. External sources are most likely particles that strike the shear probes directly or any other part of the profiler including the probe...
guard. Internal sources are also suspected and may arise from power supply fluctuations, analog-to-digital errors, crosstalk between the various sensors and electronic components.

Spikes are removed for more than cosmetic reasons. As part of the spectral analysis of the shear time series, Fourier transforms are calculated from sections of the time series. A spike, similar to a delta function, contains components of all frequencies and therefore contributes to every Fourier component. The variance of the time series is calculated from the integral of the squared Fourier transform, or power spectrum. By contributing to all components a spike may contribute significantly to the calculation of variance.

The identification of a spike and its removal is a tricky process, and the thresholds quoted here may require adjustment if the equipment is different, or if the FLY II circuitry is modified. Turbulent shear data are inherently intermittent, and contain bursts of intense signals which contain numerous frequency components. Strong, high frequency components due to turbulence have almost all the spectral characteristics of a series of spikes. A fortunate consequence of the structure of turbulence is the nature of microstructure velocity spectra. They are band limited and have a well known shape that drops off at high frequencies. These two properties allow the detection of spikes.

The first process (suggested by R. Lueck, 1983, personal communication) is to pass the time series through both high and low pass filters and compare the ratio of the two resultant time series with a threshold ratio. This threshold must be determined for each set of measurements made with particular pre-amplifier and filter characteristics, since these settings will determine the ratio that identifies spikes. The variance of the high-passed time series is also checked against a variance threshold. This process identifies small spike-like signals in weak turbulent regions that would otherwise trigger the high-
low pass ratio, but in fact are signals at the noise level. A final check is to determine the portion of variance contributed by the region immediately around the spike. In areas of very strong turbulent signals, there may be a large high-low pass ratio and high mean variance. A spike is then identified as contributing over 10% of the variance from only 1% of the time series.

These checks are summarized below. The choice of the thresholds is dependent on the offsets and gains of the amplifiers and must be reselected when new probes and circuits are used. The thresholds given here were selected after extensive testing to ensure that spikes were removed for a range of signal strengths. A 545 point time series (2 seconds) section is passed through the high and low frequency filters, each with a cutoff frequency of 20 Hz, and will be denoted by \( HP_i \) and \( LP_i \) respectively. The first check is to compare the ratio,

\[
Ratio = \left| \frac{HP_i}{LP_i} \right|
\]  

against the ratio threshold of 4.5. If \( Ratio \) is larger than 4.5, then the variance defined by,

\[
Variance = \overline{HP_i^2} = \frac{\sum_{i=1}^{545}(HP_i)^2}{545},
\]  

is checked against a variance threshold of 3.5. This variance threshold corresponds to a dissipation rate of \( \epsilon \sim 10^{-6} \) W m\(^{-3}\), near the resolution of the profiler. Spikes are therefore not removed if the signal strength is below this level. If the variance is above this noise threshold then the final check is to determine how much of the variance is accounted for by the spike at point \( i \) identified by (3.2.1). This check is denoted by,

\[
Variance - \frac{\sum_{j=i-2}^{i+2}(HP_j)^2}{545} > 0.9 \times Variance.
\]
If (3.2.3) is satisfied, after passing the previous checks, then a spike has been found. The ten points centered about the spike are then replaced with the running mean calculated over the previous 137 points or approximately $\frac{1}{2}$ second of data.

The fast response temperature data collected by FLY II in 1985 have been averaged in groups of 4 readings to provide a single value for each data scan. The resistor bridge portion of the FLY II circuitries can be modified so that the range and resolution of the readings will suit the local environment. The temperature range selected in 1985 for the slow thermistor was 1.5-13.0 °C with a digital resolution of approximately 0.003 °C. The temperature range for the fast thermistor was set to 1.8 - 17.0 °C with a resolution of 0.004°C. Details of the temperature, conductivity, salinity, density and $\sigma_t$ calculations are given in Appendix B. The formulae are those of Millero et al. (1980). The probes were calibrated on October 17, 1985 in the Ocean Physics calibration tank at the Institute of Ocean Sciences. The local dynamic viscosity $\mu$, required for each $\epsilon$ calculation, is obtained using the local temperature and an average salinity of 32.5 °/oo in the formula provided by Miyake and Koizumi (1948), shown in Appendix B.

Once the local density, viscosity and the dissipation rate $\epsilon$ (§3.3) have been calculated, the isotropic-intensity parameter $I$ (Gargett, Osborn and Nasmyth, 1984) can be estimated. Values of $I > 50$ suggest conditions of oceanic turbulent isotropy at viscous scales (Gargett et al. 1984). The parameter $I$ is defined by,

$$I = \frac{k_s}{k_b},$$

(3.2.4)

where

$$k_s = \left(\frac{\epsilon}{\rho \nu^3}\right)^\frac{1}{2},$$

28
is the Kolmogoroff wavnumber and

\[ k_b = \left( \frac{\rho N^3}{\epsilon} \right)^\frac{1}{2}, \]

is a buoyancy wavenumber related to the stratification through the buoyancy frequency defined as \( N \equiv \left( -\frac{g}{\rho} \frac{\partial \sigma}{\partial z} \right)^{\frac{1}{2}} \). The parameter \( I \) is informative in that it compares the separation of scales between the largest vertical scale which might be considered to be isotropic in a stratified environment and the viscous dissipation scales. However, it is not an absolute indicator of turbulence isotropy. The \( \Delta z \) chosen for the \( N \) calculation is \( \sim 10 \text{ cm} \), and thus \( N \) represents the local density gradient against which the viscous length scales are straining. This point becomes more important when considering local eddy coefficients for density (§4.2). The parameter \( I \) is used only to identify regions where local isotropy is not strictly valid due to increased stratification. Only in a few rare instances was \( I < 50 \), in which case the estimate of \( \epsilon \) using (3.1.3) is slightly suspect since the turbulence is anisotropic at dissipation scales. These occurrences are associated with weak turbulence (\( \epsilon \sim 10^{-6} \text{ W m}^{-3} \)) in the region of a pycnocline. For further details on the interpretation of \( I \) see Appendix A.
3.3 Spectral Analysis and Dissipation Rates

Once the shear data have been despiked, the dissipation rate per unit volume can be calculated using spectral techniques. The dissipation rate $\epsilon$ is calculated from the variance of the time series $\frac{\partial u}{\partial z}$ using (3.1.3). This method was first introduced for shear probe measurements by Osborn (1974). Spectral techniques are used to calculate the variance $\left(\frac{\partial u}{\partial z}\right)^2$ so that noise in the signal can be isolated, and the spatial response of the shear probe and attenuation in the differentiator gain $G$, can be corrected for at high frequencies.

Once the shear $\left(\frac{\partial u}{\partial z}\right)$ time series is obtained, it is subdivided into overlapping sections from which power spectra are calculated. To improve the spatial resolution in $\epsilon$ and to approximate stationarity better, the sections are shorter near the bottom where theory predicts (cf §4.3) that the dissipation rate $\epsilon$ has the vertical structure approximated by

$$\epsilon = \rho \frac{u_*^3}{\kappa z},$$

(3.3.1)

where $u_*$ is the friction velocity, $\kappa$ is von Kármán’s constant ($\kappa \approx 0.41$), $z$ is the distance from the bottom and $\rho$ is the local density. The sections are overlapped by 50% as outlined below,

![](image)

128 points
The minimum number of points per section is \( J = 128 \) and the maximum is \( J = 1024 \). The overlapping smooths out the \( \epsilon(z) \) profile, as redundant portions of time series are analyzed. This is desirable, since the beginning and end 5% portions of each section are cosine tapered before passing the \( J \) points to the fast Fourier transform routine.

Before the \( J \) shear data points are tapered and passed to the Fourier transform, the linear trend in the data is removed. The resultant series has both the mean and the trend removed. Removing the trend reduces the low frequency energy that can not be resolved by the Fourier transform analysis.

A cosine taper is applied and the time series is passed to a fast Fourier transform routine as the real part, along with a corresponding zero imaginary part. The Fourier routine returns the real \( (X_i) \) and imaginary \( (Y_i) \) parts to the Fourier transform (Stearns, 1978). The power spectrum \( \Psi_i \) is then given by,

\[
\Psi_i = 1.1025 \left( X_i^2 + Y_i^2 \right) \frac{dt}{J},
\]

where \( 1.1025 = (1.05)^2 \) is the correction applied to compensate for the cosine tapering and \( dt \) is the sampling rate (~ \( \frac{1}{274} \) s). The corresponding frequency associated with the \( i^{th} \) power spectrum component \( \Psi_i \) is,

\[
f_i = (i - 1)df,
\]

where \( df = \frac{1}{Jdt} \).

The power spectrum \( \Psi_i \) is then corrected for the roll-off in the shear probe response at small scales, or high frequencies. The probe response can only be corrected to wavelengths \( \lambda > 0.606 \) cm, so the spectrum above the frequency associated with this wavelength is truncated \( (f_i > \frac{W}{\lambda}) \). The quartic transfer function to correct for the spatial response is
given by (Ninnis, 1984),

\[ T_i = 1.003 - 0.164 \left( \frac{k_i}{k_0} \right) - 4.537 \left( \frac{k_i}{k_0} \right)^2 + 5.503 \left( \frac{k_i}{k_0} \right)^3 - 1.804 \left( \frac{k_i}{k_0} \right)^4, \]

(3.3.4)

where \( k_i \) is the cyclical wavenumber associated with a given frequency \( (k_i = \hat{k}_i) \) and \( k_0 \) is the zero response cyclical wavenumber of the particular probe used. For the probes used in the present study \( k_0 = 1.7 \text{ cm}^{-1} \). The spatially corrected power spectrum \( \Psi_i^s \) is then

\[ \Psi_i^s = \frac{\Psi_i}{T_i}, \]

(3.3.5)

The power spectrum is also corrected for the roll-off in the differentiator gain at frequencies above 60 Hz. This correction takes the form

\[ \Psi_i^{sg} = \Psi_i^{s} 10.0(\beta \exp(\alpha f_i)), \]

(3.3.6)

where \( \beta = 0.00326, \alpha = 0.0524 \) and \( f_i \) is the frequency \textit{above} 60 Hz only.

The next task is to calculate the dissipation rate \( \epsilon \) from the corrected power spectrum \( \Psi_i^{sg} \). The variance \( \left( \frac{\partial u}{\partial z} \right)^2 \) is calculated using (Jenkins and Watts, 1968),

\[ \left( \frac{\partial u}{\partial z} \right)^2 = \int_{k_l}^{k_u} k_3^2 \phi_{11}(k_3)dk_3 = \sum_{i=1}^{i_u} \Psi_i^{sg} df, \]

(3.3.7)

where \( k_l \) and \( k_u \) are the lower and upper limits of integration in wavenumber space, \( k_3 \) is the vertical wavenumber and \( \phi_{11}(k_3) \) is the velocity spectrum of the \( u_1 \) component.

The integration limits are selected to minimize the contributions to the integral from regions of the spectrum, both at low \( (k < k_l) \) and high \( (k > k_u) \) wavenumbers, that are due to noise. The following technique for selecting \( k_l \) and \( k_u \) is believed to be unique and was developed and implemented by the author. If the indices associated with
wavenumbers $k_l$ and $k_u$ are $i_l$ and $i_u$ respectively, then the recoverable dissipation rate is given by,

$$\epsilon_{lu} = 7.5\mu \sum_{i=i_l}^{i_u} \Psi_i^{ag} df. \quad (3.3.8)$$

The contributions lost by selecting the wavenumber band $k_l$ and $k_u$ are then added to the integrated dissipation rate $\epsilon_{lu}$,

$$\epsilon = \epsilon_{lu} \left(1 + \frac{P_l}{100} + \frac{P_u}{100}\right), \quad (3.3.9)$$

where $P_l$ is the estimated percentage lost below $k_l$ and $P_u$ is the estimated percentage lost above $k_u$. The values of $P_l$ and $P_u$ have been estimated from a spline fit to Nasmyth's spectral data (Nasmyth, 1986, personal communications). The cumulative area under the Nasmyth universal spectrum $G_2(k_3/k_s)$ is calculated from a spline fit and is shown in Figure 8.

The selection of an individual spectrum's integration band $k_l$ and $k_u$, is a positive feedback process where repeated estimates of $\epsilon$ are made until a convergence is found. The process is outlined here for completeness, as I feel it is a crucial stage of estimating accurate dissipation rate values from shear probe measurements. If, for example, the normalized spectrum

$$G_2(k_3/k_s) = \left(\frac{\epsilon\mu^5}{\rho^6}\right)^{-1/4} \frac{k_3^2}{k_s^2} \Phi_{11}(k_3)$$

$$= \left(\frac{\epsilon^3}{\rho^2\mu}\right)^{-1/4} \left(\frac{W}{2\pi}\right) \Psi^{ag}, \quad (3.3.10)$$

is the solid line in Figure 9, then the selection of the final integration band $k_l$, $k_u$ would proceed as follows.

The universal dissipation spectrum calculated from Nasmyth's (1986, personal communication) $F(k_3/k_s)$ data using,
Figure 8 Nasmyth's (1986, personal communication) spectral data fitted with a cubic spine and the cumulative area under the spline. The wavenumbers $k_5$ and $k_{90}$ represent the wavenumbers where 5% of the area lies below $k_5$ and 10% above $k_{90}$ respectively. These values (e.g. 5% and 90%) have been chosen only as examples. The real integration limits for a measured spectrum are selected to minimize the contribution to the variance from noise.
Figure 9 An example spectrum (authors' conception) shown as the solid line, and Nasmyth's data (c) (1986, personal communication) showing the selection of the best integration band $k_l^i$ to $k_u^i$. By selecting a band too wide (e.g. $k_l^i$ to $k_u^i$) noise is included in the integration (3.3.8) and the dissipation rate recovered by (3.3.9) is overestimated. When normalized by this overestimate ($\epsilon^1$), the spectrum (dashed line) lies well below the universal shape (Nasmyth's data). Integration bands narrower than $k_l^i$ to $k_u^i$ result in dissipation rates similar to $\epsilon^i$, the indication that an appropriate band ($k_l^i$ to $k_u^i$) has been found.
\[
G_2(k_3/k_s) = \frac{k_s^2}{k_3^2} \frac{1}{2} \left[ F(k_3/k_s) + k_3 \frac{\partial F(k_3/k_s)}{\partial k_3} \right],
\]

(3.3.11)
is identified by the (o) in Figure 9. The first choice of \( k_l \) and \( k_u \) for the measured spectrum, identified as \( k_l^1 \) and \( k_u^1 \), are crudely approximated by \( k_l^1 = 0.14 \text{ cm}^{-1} \), and \( k_u^1 = k_n \), the upper wavenumber where the spectrum begins to actually increase due to noise. By selecting \( k_l^1 \) and \( k_u^1 \) for this spectrum, a significant portion of noise has been included in the summation (3.3.8) and the resulting dissipation rate estimate \( \epsilon_1 \) from (3.3.9) would be substantially overestimated. If the spectrum were normalized using \( \epsilon_1 \) in (3.3.10), the resultant \( G_2(k_3/k_s) \) spectrum would lie far below the universal shape, as indicated by the dashed line in Figure 9. The universal shape is used only as a guide and was most helpful in the development of this technique.

A second integration band is then selected with the new limits, \( k_l^2 \) and \( k_u^2 \), independently estimated such that the integration band is approximately 10% narrower than the previous band. If the new estimate of \( \epsilon_2 \) (3.3.9) is less than \( \epsilon_1 \), then noise has been eliminated from the summation part of the calculation (3.3.8), and \( \epsilon_2 \) is more accurate than \( \epsilon_1 \). This process is repeated until \( \epsilon^{i+1} \approx \epsilon^i \) and the narrower band, \( k_l^{i+1} \) and \( k_u^{i+1} \) has not excluded further noise from the band \( k_l^i \) and \( k_u^i \). Because the integration has been corrected to compensate for power outside the integration limits using (3.3.9), choosing narrower limits should not alter the dissipation rate. This process breaks down when the integration band is so narrow that only a few spectral values contribute to the initial estimate (3.3.8). The spectrum is then normalized using \( \epsilon^i \) in (3.3.10) (solid line in Figure 9) and can be compared to the universal shape.

The results indicate that this process is capable of estimating dissipation rate values
down to $\epsilon_0 \simeq 10^{-6} \text{ W m}^{-3}$ with an approximate uncertainty of 50% at this level (Appendix A). At $\epsilon_0$, the bandwidth becomes very small and only a few spectral estimates contribute to the variance, and $\epsilon_{lu}$ (3.3.8). Below $\epsilon_0$, noise masks the shear spectrum, and the $k_3^2\phi_{11}(k_3)$ peak cannot be identified.

The dissipation rate values calculated using (3.3.9) are the final estimates, with the accuracy of individual estimates being dependent on the magnitude of $\epsilon$ in comparison to the background vibrational noise ($\S$3.4), and the uncertainty of the parameters in (3.1.1). These last uncertainties are discussed in Appendix A. The cumulative uncertainty in the $\epsilon$ values range between 20% for high $\epsilon$ values ($\epsilon \sim 10^{-3} \text{ W m}^{-3}$) to 50% for very low $\epsilon$ values ($\epsilon \sim 10^{-6} \text{ W m}^{-3}$). These error estimates have been obtained by adding the square root of the individual error contributions in (3.1.1) which are believed to be random. Higher uncertainties (30%—70%) are estimated by assuming that all errors add linearly (Appendix A).

For each profile, the 1024 point power spectra are averaged into four groups, normalized and plotted. If the dissipation rate values have been accurately estimated using (3.3.8) and (3.3.9), then the normalized spectra (3.3.10) should match the universal spectrum in the vicinity of $\frac{k}{k_0} \simeq 0.13$. This wavenumber represents the peak in the $G_2$ universal dissipation spectrum, where the rate of kinetic energy loss due to viscosity is a maximum. This region of the spectrum must be recovered for an accurate estimate of the viscous dissipation rate. The spectra are sorted and averaged into four groups, associated with energetic spectra for which $\epsilon > 10^{-4} \text{ Wm}^{-3}$, an intermediate range $10^{-5} < \epsilon < 10^{-4} \text{ Wm}^{-3}$, a weak range $10^{-6} < \epsilon < 10^{-5} \text{ Wm}^{-3}$ and the spectra of very weak turbulence near the noise level of the instrument $\epsilon < 10^{-6} \text{ Wm}^{-3}$. This separation
ensures that noise of weakly dissipative spectra does not contaminate the higher dissipation rate spectra, since the noise is normalized by the individual dissipation rates for each spectrum. Examples of the normalized universal spectra are shown in Figures 10, 11 and 12. The universal shear, or dissipation spectra $G_2 (k_3/k_4)$ are calculated by (3.3.10), and the corresponding universal velocity spectra can be calculated using (3.3.11).
Figure 10  Band-averaged spectral values from an average of 21 individual 1024 point spectra from the range of dissipation rates between $10^{-5}$ to $10^{-4}$ W m$^{-3}$. All data shown in Figures 10, 11 and 12 are from a single profile (number 535) from station M2. This range represents weak to moderate turbulence. Both the universal velocity $F_2 (k_3/k_s)$ [a] and shear $G_2 (k_3/k_s)$ [b] spectral representations are shown ($\Delta$). Nasmyth’s data (o) are shown only as a reference. The confidence intervals are 96% (~ two standard deviations). The average dissipation rate and the associated Kolmogoroff wavenumber are also shown.
Figure 11 Band-averaged spectral values from an average of 36 individual 1024 point spectra from the range of dissipation rates between $10^{-6}$ to $10^{-5}$ W m$^{-3}$. This range represents dissipation rates between the lower limit of resolution and weak turbulence.
Figure 12  Band-averaged spectral values from an average of 41 individual 1024 point spectra from the range of dissipation rates below $10^{-6}$ W m$^{-3}$. This range represents very weak turbulence near the noise level of the present profiler and all spectra at the noise level.
3.4 Noise Reduction in FLY II

Mourn and Lueck (1985) estimate the lowest noise level for shear probes mounted on the free-falling profilers Camels II and III, in terms of a dissipation rate per unit volume, to be $\epsilon_n = 10^{-7}$ W m$^{-3}$. (This value of $\epsilon_n$ is chosen as a reference dissipation rate in the following discussion.) The estimate $\epsilon_n$ was obtained from an untethered, streamlined profiler without floats, brushes, or a probe guard. Noise due to horizontal accelerations, tilting and vibrations was considered. Broadband noise within the shear signal $V_0$ is most likely due to mechanical vibrations and instrument tilting (Mourn and Lueck, 1985) and will be addressed here. A higher noise level is expected for FLY II, due to the probe guard installed for near-bottom sampling and the tethering cable.

The initial design of FLY II consisted of untapered cylindrical floats and probe guards. Noise levels in terms of the dissipation rate ranged from $5.0 \times 10^{-5}$ to $8.0 \times 10^{-4}$ W m$^{-3}$, or 27 dB to 39 dB above $\epsilon_n$. These represented unacceptably high noise levels and would have limited accurate measurements to only the most energetic turbulent regimes. Spectral analysis of the shear probe signal indicated two major sources of noise. Electrical noise at 73 Hz was of some concern, but noise at such high frequencies does not contribute to the dissipation rate in regions of weak turbulence. Mechanical vibrations caused significant noise within the turbulence bandwidth at frequencies between 4 and 20 Hz (Figure 13). Two sources were suspected: 1) the structural resonance of the probe guard-profiler system, or 2) the vibrations caused by the shedding of eddies.

Mechanical vibrations are most likely generated by the shedding of eddies at blunt surfaces, from the probe guard and at the upper end of the profiler. In an attempt to reduce the eddy shedding, the syntactic foam floats were tapered and a circular brush
Figure 13 A spectrum calculated from the shear probe signal when the profiler was fitted with the original probe guard. There is substantial vibrational noise within the turbulence band (4–30 Hz) and electrical noise at higher frequencies (∼70 Hz).

added at the trailing end. The brush ensures that the profiler sheds small eddies which are less likely to vibrate the profiler in the turbulence bandwidth. These modifications lowered the noise to 23 dB above $\epsilon_n$. Initially two probe guard designs were used. The two-stage design shown in Figure 14a produced noise at 4 and 18 Hz, and the resulting dissipation rate associated with this noise was 39 dB above $\epsilon_n$. The simpler probe guard shown in Figure 14b generated less vibrational noise, only 27 dB above $\epsilon_n$. Tests with the smaller probe guard at a fall speed of 80 cm s$^{-1}$ revealed that most of the noise was near 18 Hz. Lowering the fall speed to 60 cm s$^{-1}$ revealed a peak due to noise at 13 Hz. Coherent eddy shedding from the lower ring of the probe guard was suspected as a major source of this noise.

A cylinder falling through a fluid sheds eddies to form a von Kármán vortex street
whose shedding frequency is determined by the Strouhal number (Every, King and Weaver, 1982). This frequency is given by,

\[ f_s = \frac{0.212W}{d}, \]

where the Strouhal number is 0.212 for \( Re > 10^3 \), \( W \) is the velocity of the flow past the cylinder and \( d \) is the diameter of the cylinder. As a vortex is shed from the cylinder the local pressure disturbance is altered and the structure experiences a time-varying lateral force at the shedding frequency. If the structure has a resonant frequency near the shedding frequency then the system experiences sustained vibrations. However, even
in the absence of resonance, vibrations are present. A 0.95 cm diameter cylinder falling at 80 cm s\(^{-1}\) has a Reynolds number of \(6.0 \times 10^3\) and oscillates at 17.8 Hz. At a fall speed of 60 cm s\(^{-1}\) the frequency is 13 Hz; therefore the noise peaks, noted above, at 13 and 18 Hz are attributed to these eddies. The structural resonance of the probe guard was estimated to be at frequencies higher than 100 Hz by accoustical tests in the large water tank located at the Institute of Ocean Sciences.

To eliminate the von Kármán vortex street and the vibrations related to the eddy shedding, several modifications to the probe guard were investigated. Every \textit{et al.} (1982) suggest two methods of reducing the vibrations of a system: 1) to alter the transfer function or structure of the system and 2) to modify the magnitude and frequency of the forcing function. Since the turbulent component of the shear probe signal is band-limited, attempts were made to modify the magnitude and frequency of the eddies shed by the probe guard rather than redesign another guard system.

A teardrop fairing along the upper rim of a cylinder is similar in effect to a splitter plate as discussed by Apelt, West and Szewczyte (1973). By adding such a feature several flow parameters are modified. Apelt \textit{et al.} (1973) observed that the Strouhal number decreases as the ratio of splitter plate length \(L\) to cylinder diameter \(d\) is increased from \(\frac{L}{d} = 0\) to \(\frac{L}{d} = 1.0\). The shedding frequency on the other hand was observed to increase by up to 10\% for \(\frac{L}{d} = \frac{1}{4}\) but to decrease by 10\% for \(\frac{L}{d} = 1.0\). In addition to changing the frequency of the eddy shedding, the net form drag on the cylinder is reduced and the diameter of the wake downstream of the cylinder decreases. A narrower wake may imply less energy in the eddies and less horizontal forcing in the vibrations. The width of the probe guard wake is an important consideration since the shear probes are located
centrally, downstream of the probe guard ring and must not be permitted to enter the wake. Calculation after Schlichting (1960) indicates that the maximum wake diameter 15 cm downstream of the probe guard ring is only \( \sim 3.0 \) cm. Confirmation that the shear probes are not in the probe guard wake is also provided by estimating the dissipation rate within the wake, nearly 36 W m\(^{-3}\), many orders of magnitude larger than what is actually measured.

A teardrop fairing was added to the lower ring of the smaller probe guard (Figure 14b) with \( \frac{L}{d} = 1.44 \) (Figure 15). A significant reduction in the mechanical noise was not evident despite the fact that Every et al. (1982) indicate vortex-excited oscillations should be eliminated. Roshko (1954) on the other hand found that a splitter plate with \( \frac{L}{d} = 1.14 \) was insufficient at \( Re = 10^5 \) but with \( \frac{L}{d} = 5 \), the vortex motion was suppressed. A variety of fall speeds (and therefore \( Re \)) were investigated by adding or subtracting the tapered floats. It was observed that the noise increased with fewer floats and higher fall speeds. With six floats the fall speed was 60 cm s\(^{-1}\) and the noise (13 Hz) translated to a dissipation rate 20 dB above \( \epsilon_n \).

To further modify the magnitude and frequency of the vortex shedding, the flow around the probe guard must be completely disrupted. The most successful device for disrupting the flow around cylinders is the helical strake (Every et al. 1982). Every and King (1979) found that a 50% reduction in the vibration amplitude could be obtained with round strakes of diameter 0.05\( d \). In addition to modifying the amplitude and frequency of the vortex street, the helical strake increases the drag of the system. The wake however contains relatively smaller, more isotropic eddies and does not increase significantly in total size.
Figure 15 The lower portion of the smaller probe guard (Figure 14b) showing the fairing added to the back of the cylinder and the double helical strakes.

With the fairing on the ring portion of the probe guard, single helical strakes of $\frac{1}{3}d$ nylon line were added to the four support rods (Figure 5a). A single helical strake of $\frac{2}{3}d$ polypropylene line was also wrapped around the faired ring. Test profiles with this arrangement indicated a much reduced noise level. Unfortunately the single strake on the faired ring caused the profiler to rotate during descent. The tilt signals showed a distinct oscillatory motion during ascent and the tethering cable became twisted. A second helical strake was added, overlapping the original strake on the faired ring (Figure 15). The two strakes were wound in opposite directions and formed a wider more complex pattern, eliminating the tendency for the profiler to rotate.

In many coastal locations the ocean bottom is composed of soft mud and it was feared that under such conditions the probe guard might sink into the sediments, possibly damaging the sensitive probes. Adding the alternating helical strakes increases the net surface area of the probe guard by 55% and reduces the possibility of the probe guard from sinking into the mud. The increased drag can be compensated by adjusting the
amount of floatation. The eddies shed by this complex faired-straked arrangement are highly uncorrelated in space and are of sufficiently small scale and of such a broadband nature as to cause minimal contamination of the shear signals.

The net result is a noise level only 5 dB above $\epsilon_n$ for fall speeds near 60 cm s$^{-1}$. This corresponds to a "noise" dissipation rate of $3.0 \times 10^{-7}$ Wm$^{-3}$. Computed dissipation rates at this low level do not represent turbulent fluctuations, and little confidence can be put in the actual turbulent dissipation rate values below $10^{-6}$ W m$^{-3}$. The coherent vibrations at the Strouhal frequency are eliminated and vibrational contributions to the spectrum $\Psi^{sg}$ are small and broadband (1-60 Hz). The noise level in $\epsilon$ is also dependent on the selection of the integration limits $k_l$ and $k_u$. Figure 11 shows the average of 41 spectra near this limit. Noise is present at low and high frequencies, but $k_l$ and $k_u$ have been selected to crop out as much noise as possible for the integration (3.3.8). Dissipation rate estimates near the noise level ($\epsilon < 10^{-6}$ W m$^{-3}$) are obtained from only a few ($\sim 10$) spectral values within a narrow band (Figure 12). An accuracy of 50% is estimated for dissipation rates near $10^{-6}$ W m$^{-3}$ (Appendix A). This accuracy is confirmed by the fact that the measured spectra, when normalized by the estimated dissipation rate $\epsilon$, match the universal spectra near the dissipation peak (Figure 11b). This accuracy would not be obtainable if spectral shapes (i.e. $-\frac{5}{3}$ slope) were fitted in the inertial subrange ($\frac{k}{k_z} < 0.1$) which is contaminated by noise. For a more detailed description of the estimation of the integral (3.3.8) and the selection of the integration limits (a crucial procedure for accurate $\epsilon$ estimates) see §3.3 and Appendix A.
3.5 Dissipation Rate Profiles

Once the shear time series is subdivided and the dissipation rate values ($\epsilon$) computed from the power spectra, a vertical profile of the dissipation rate values may be plotted. Figure 16a is an example of such a profile. Due to the large range of dissipation rates measured in the ocean (typically 4 orders of magnitude), the logarithm values are plotted. The ordinate in Figure 16a is the linear distance to the bottom ($z$). In Figure 16b the logarithm of both $\epsilon$ and $z$ is plotted, and the structure in the bottom portion ($z < 10$ m) of the profile can be seen more easily. Shown in Figure 16b is the fit of (3.3.1) to the lower $\epsilon$ values in an attempt to identify the constant stress layer and the bottom stress $\tau_0(= \rho u_*^2)$. This is explained in detail in §5.2.

The profiles shown in this thesis, either from individual dissipation rate values (Figure 16) or from averages of several profiles, are plotted as solid lines. This is in contrast to the stick plots more common in the past (Osborn, 1978; Crawford and Osborn, 1979a; Lueck et al. 1984). Although the individual dissipation rate values are separated in the vertical by as much as 1.2 m, they are calculated from nearly continuous ($\sim 457$ pts per m) shear data and are overlapped by 50%. Thus a solid line is more representative of the measurements and is also more pleasing to the eye.

Due to noise in the shear signal, the dissipation rates never drop to zero, or even significantly below the estimated noise level of $\epsilon_n = 3.0 \times 10^{-7}$ W m$^{-3}$. Very quiet (non-turbulent) regions can be expected some of the time, in which case the reported dissipation rate is entirely a result of vibrational noise in the profiler system. The estimate of $\epsilon_n$ has been chosen by the inspection of many dissipation rate profiles and the calculated spectra from the quietest regions.
Figure 16  a) Log dissipation rate values calculated from the shear time series. The $\epsilon$ values within 5 m of the bottom are calculated from time series sections of less than 1024 points. b) The same data in a), with a log ordinate. The dashed line is the distribution found by fitting $\epsilon = \rho u_*^3/\kappa z$ within the constant stress layer, and the average value of $u_*$ shown is estimated from the dissipation rates in the constant stress layer only. The data was collected at M2, 17 June 1985. Total depth 135 m.
3.6 Additional Measurements

Data collected for this dissertation were part of the "Bottom Boundary Layer Experiment" and include; a microstructure survey (temperature, conductivity and velocity shear) from seven stations M1, M2, M3, M5, M6, M7 and B2 (Figure 1), a CTD (Conductivity, Temperature and Depth; Guildline model) and dissolved oxygen survey from 42 stations (Figure 17), and a current survey with 24 current meters on 4 moorings, C1-C4 (Figure 1). The current moorings were deployed on June 10, 1985. The microstructure survey commenced on June 11, 1985 at station M1. Data were collected at each microstructure station for approximately 22 hours, nearly one diurnal period. Following M1, data were collected at B2 on June 11 and 12, at M3 on June 12 and 13, at M6 on June 16 and 17, at M1 on June 17 and 18, at M5 on June 18 and 19, and at station M7 on June 19 and 20. The CTD and oxygen survey were conducted on June 14 and 15, starting at station 42 and continuing in decending order to station 1.

The current moorings were designed to resolve the current shear near the bottom with the meters at heights of $z = 3, 5, 10$, and 30 m above the bottom. The meters at 5 m above the bottom were vector averaging AML–Geodyne type (Savonius rotor and direction vane modified to record on 9-track tape), set at a sampling rate of 1 s. The remaining meters near the bottom were Anderaa–RCM4 type, recording at intervals of 5 minutes. Each of the three moorings near the shelf break, C1–C3, also had an InterOcean–S4 vector averaging current meter located at a depth of 40 m below the surface. The sampling rate for these meters was set at 1.5 minutes. The AML–Geodyne meters have a digital resolution which translates to 2.3 cm/s. This turned out to be too coarse to resolve the small variations near the bottom. These data will not be presented. The tidal
Figure 17 The 42 stations of CTD and Oxygen data, 14-15 June 1985. The survey actually started in the southern region at station 42, and proceeded northward in descending order.

constituents were extracted from the remaining data by filtering to one hour samples and applying the harmonic analysis techniques of Foreman (1979). The constituents are tabled in Appendix C.

The Oxygen data were processed by B.G.Minkley of the Institute of Ocean Sciences. The oxygen content was estimated by titration of water samples brought up in Niskin bottles. Samples were collected near the bottom at each station, and at six depths at a selection of 23 stations.
The CTD data were contoured by H.J. Freeland using the objective analysis technique of Denman and Freeland (1985), and will be discussed in §§5 and 6. The temperature and conductivity signals recorded by the FLY II profiler were processed using the formulae provided in Appendix B.

Also collected were 5 bottom grabs near stations M1, M2, M5 M6 and B2. The grabs were analyzed by R. Kool of the B.C. Provincial Museum, the results of which are listed in Table 2. Clear weather during the cruise allowed the acquisition of satellite imagery by the satellite receiving facility at the Department of Oceanography, University of British Columbia. The June 16 and 17 AVHRR images have been processed to reveal the surface temperature variations and are shown in Figures 2 and 3 respectively. Andy Thomas of the Department of Oceanography completed the image processing. Colour reproductions can be found in Emery et al. (1986)

<table>
<thead>
<tr>
<th>Station</th>
<th>Depth</th>
<th>Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>154 m</td>
<td>sand and coarse gravel</td>
</tr>
<tr>
<td>M2</td>
<td>131 m</td>
<td>sticky mud with some sand in mud</td>
</tr>
<tr>
<td>M5</td>
<td>168 m</td>
<td>ooze</td>
</tr>
<tr>
<td>M6</td>
<td>142 m</td>
<td>sand and very fine gravel</td>
</tr>
<tr>
<td>B2</td>
<td>62 m</td>
<td>coarse and large gravel</td>
</tr>
</tbody>
</table>

Also processed and analyzed are some microstructure profiles from station W05 in Hecate Strait (Figure 18). These data were collected in July, 1983, before the FLY II profiler was modified to include salinity measurements. The probe guard of Figure 14b
was used without the strakes, therefore noise levels are high. The total water depth at W05 is 35 m, so repeated profiles every 2 minutes allowed 20 profiles to be made in only 40 minutes. The profiler was also fitted with 2 shear probes, so that 2 dissipation rate profiles are obtained. These data are discussed in §5.2 with respect to bottom stress estimates.
Figure 18 The Hecate Strait region east of the Queen Charlotte Islands. Microstructure profiles were collected at station W05 in July, 1983. The FLY II profiler did not carry a conductivity cell nor a slow response thermistor at this time. Also, the smaller probe guard had not yet had the modifications to reduce the vibrational noise.

4. The Bottom Boundary Layer: Theory

4.1 Introduction

A review of bottom boundary layer theory will be presented so that the dissipation rate measurements may be interpreted in terms of the structures found in models and
laboratory studies. Much work has been done on boundary layer flow and it is important to know what information can be gained from the profiles of turbulent dissipation rate through the boundary layer over the continental shelf. It is also useful to understand the dynamics governing simple boundary layer structures so that the observations of a complicated oceanic boundary layer may be interpreted correctly.

4.2 Structure and Dynamics

The bottom boundary layer plays an important role in the dynamics of flow over the continental shelf. The vertical structure and temporal evolution of the flow are determined by local sources of mixing due to shear-induced turbulence near the bottom, and to the structure and history of local currents and water masses. A recent review of the dynamics of continental shelf bottom boundary layers has been presented by Grant and Madsen (1986). Near-bottom current, density and turbulence measurements have been made by Weatherly and Martin (1978) and Soulsby (1983); and near-bottom current-wave interactions have been studied by Grant and Madsen (1979). Models of the time-dependent, three-dimensional current structure are discussed by Soulsby (1983) and Richards (1982). The models depend, for the interpretation of turbulent mixing, on representations of the turbulence in terms of second order closure theories. Soulsby (1983) models the diffusive terms with a linear eddy viscosity and Richards (1982) considers two eddy viscosities similar to that of Gibson and Launder (1978) and Mellor and Yamada (1974). When stratification is included in these models, a layer of well-mixed, neutrally stratified water develops at the bottom, and the turbulent velocity fluctuations and Ekman veering are constrained to this layer.

Before we proceed it may be useful to establish what is meant by the terms homo-
geneity and stationarity. Homogeneity implies uniform conditions in space, unchanged by a translation. Stationarity refers to changes in time. Under certain conditions a flow can be assumed to be both homogeneous and stationary when small space and time scales are considered. Conversely, real flows are always inhomogeneous and non-stationary when large distances and times are considered. At certain stages of the analysis either, or both homogeneity and stationarity will be assumed, and attempts to justify these assumptions will be made. The terminology is further complicated by the fact that Taylor’s frozen turbulence hypothesis is invoked to get vertical shear ($\frac{\partial u}{\partial z}$) from the time-differentiated probe voltage (3.1.1). This operation transforms time variations into space variations, and considerations of stationarity into considerations of homogeneity. In other words, the problems of non-stationarity in time series analysis become problems of inhomogeneity in space series analysis. The vertical microstructure “space series” will however, still be referred to as “time series”.

In the turbulent flow of an unstratified, non-rotating and laterally homogeneous fluid over a flat bottom, we can identify three distinct regions. Immediately next to the boundary is a layer dominated by viscous stresses. For a smooth bottom, this layer forms a continuous viscous sub-layer; on a rough bottom, the roughness elements protrude out of the viscous sub-layer into the turbulent region of the flow. Above the viscous region, shear stress is sustained by turbulent fluctuations and a nearly logarithmic velocity profile is found. Finally, above the logarithmic layer, the current shear decreases to zero as the velocity approaches its asymptotic value.

The current structure of the second region can be modeled by assuming that the turbulent interactions can be represented by a linear eddy viscosity $K_m = \kappa z u_*$. The
result is the law of the wall, or logarithmic velocity profile,

$$\overline{U} = \frac{u_*}{\kappa} \ln \frac{z}{z_0},$$

(4.2.1)

where $u_*$ is the friction velocity defined by $u_* = \sqrt{\tau_0/\rho}$, (where $\tau_0$ is the bottom stress,) $\overline{U}$ the mean current, and $z_0$ is an integration constant.

The turbulent kinetic energy equation for an unstratified flow is (Hinze, 1975),

$$\frac{\rho}{2} \frac{\partial q^2}{\partial t} + \rho U_i \frac{\partial q^2}{\partial x_i} + \rho \frac{\partial}{\partial x_i} \left( \frac{\partial u_i}{\partial x_i} \tau_{ij} \right) + \rho \overline{u_i} \overline{u_j} \frac{\partial \overline{U_j}}{\partial x_i} =$$

$$- \frac{\mu}{\partial x_i} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \frac{\tau_{ij}}{\partial x_i} \partial x_j - \frac{\mu}{\partial x_i} \left( \frac{\partial u_i}{\partial x_i} + \frac{\partial u_j}{\partial x_i} \right) \frac{\partial u_j}{\partial x_j} + \frac{\mu}{\partial x_i} \left( \frac{\partial u_i}{\partial x_i} + \frac{\partial u_j}{\partial x_i} \right) \frac{\partial u_j}{\partial x_j},$$

(4.2.2)

where the terms represent, in order: the mean rate of change of total turbulent kinetic energy ($q^2 = u_i u_i$), the advection of turbulent kinetic energy by the mean flow, the mean turbulent convection by turbulent pressure and velocity fluctuations, the mean rate of kinetic energy production by the work of the Reynolds stress on the mean shear; and on the righthand side, the work expended by the viscous shear stresses and the viscous dissipation rate of turbulent kinetic energy ($= \epsilon$ (3.1.2)). Repeated indices represent the accepted notation for summation.

In most turbulent flows, (4.2.2) can be greatly simplified (Townsend, 1976). To a good approximation, the rate of change of turbulent kinetic energy due to advection by the mean flow is negligible (Townsend, 1976). Similarly, it can be shown that the mean turbulent convection terms and the viscous shear stresses can often be omitted (Osborn, 1980). The resulting turbulent kinetic energy equation for homogeneous turbulence may be written,

$$\frac{\rho}{2} \frac{\partial q^2}{\partial t} + \rho \overline{u_i} \overline{u_j} \frac{\partial \overline{U_j}}{\partial x_j} = \epsilon.$$

(4.2.3)

This can be simplified further when the flow is reasonably steady, or stationary, and the rate of change of turbulent kinetic energy is small in comparison to the rate of production.
and dissipation. The validity of this assumption will be estimated for the constant stress layer in §4.3. The turbulent kinetic energy balance can now be written for unstratified flows as,

$$-\rho u'w' \frac{\partial U}{\partial z} = \epsilon,$$

(4.2.4)

where the component of the Reynolds stress that works on the mean shear $\frac{\partial U}{\partial z}$ is $(-u'w')$ and the overbars indicate an appropriate average such that the quantities represent the steady conditions. For measurements at a fixed height in the bottom boundary layer the appropriate time interval over which the mean quantities are to be averaged is estimated by Grant et al. (1984) to be larger than 10 minutes (to minimize the influence of surface wave motion in shallow regions) and approximately less than 40 minutes (to minimize the influence of internal wave motion and tides).

Equation (4.2.4) implies that a vertical profile of $\epsilon$ provides information on the distribution of source as well as sink terms in the turbulent kinetic energy equation in the turbulent bottom boundary layer (tbbl). The height of the tbbl is the distance from the bottom where the dissipation rate (and therefore the shear production rate) drops to some interior, background level.

To this reasonably simple structure (equations (4.2.1) and (4.2.4)) we add the effects of time-dependence, mainly in the form of tidal currents. In a time-dependent or accelerating turbulent flow, (4.2.4) may no longer strictly hold. The production and dissipation of turbulent kinetic energy may no longer be locally balanced, although they might balance on average over a tidal cycle. An increasing shear may thus generate turbulence faster than it is locally dissipated, and advect the energy downstream, in which case the mean advection term in (4.2.2) must be re-instated. It should nevertheless be noted that
the time-dependent tidal flow varies and accelerates slowly in comparison to the turbulent fluctuations, which react quickly to changes in the mean flow (Soulsby, 1983), so that (4.2.4) is expected to hold in all but extreme tidal accelerations. The tidally driven tidal boundary layer grows and decays both in height and intensity with the time-dependent current shear but in general with a phase lag. Richards (1982) predicts a lag of up to 11° for the tidal boundary layer flow. Also, the phase of the oscillatory current in deep water (>50m) is depth-dependent because inertial forces are comparable with the frictional forces away from the boundary and the currents there lag those within the tidal boundary layer, where frictional forces dominate (Soulsby, 1983).

To this flow we then add the effects of rotation, and present the geostrophic equations with the addition of a friction term. The vertical structure of oscillatory currents in a rotating reference frame, namely tidal flow, is modeled by Prandle (1980) with the shallow-water equations. He considers an eddy viscosity of the form

\[ K_m = \alpha + \beta z, \]  

(4.2.5)

where \( \alpha \) and \( \beta \) are constants to be determined. (A detailed discussion of eddy viscosities is provided in the following section, §4.3.) The vertical structure of the currents is found by solving,

\[ \frac{\partial \bar{U}}{\partial t} - f \bar{V} = -g \frac{\partial \zeta}{\partial x} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \rho K_m \frac{\partial \bar{U}}{\partial z} \right), \]  

(4.2.6a)

\[ \frac{\partial \bar{V}}{\partial t} + f \bar{U} = -g \frac{\partial \zeta}{\partial y} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \rho K_m \frac{\partial \bar{V}}{\partial z} \right), \]  

(4.2.6b)

where \( \bar{U} \) and \( \bar{V} \) are the horizontal, non-turbulent current velocities, \( f \) the Coriolis parameter and \( \zeta \) the surface elevation. The solution to (4.2.6) is critically dependent on the vertical distribution of \( K_m \). Prandle (1980) chooses the linear form (4.2.5) so that
an analytical solution can be found. To this end the currents are rewritten in the form of counter-rotating vectors, as are the surface elevation gradients. The solution is then a combination of Kelvin functions (Abramowitz and Stegun, 1972, p379), and can be written (Prandle, 1980),

\[
R_j = \frac{G_j}{i\beta p_j^2} + L_j \left( \text{ber} \left( 2p_j \left( z + \frac{\alpha}{\beta} \right)^{1/2} \right) \pm i \text{bei} \left( 2p_j \left( z + \frac{\alpha}{\beta} \right)^{1/2} \right) \right) + B_j \left( \text{ker} \left( 2p_j \left( z + \frac{\alpha}{\beta} \right)^{1/2} \right) \pm i \text{kei} \left( 2p_j \left( z + \frac{\alpha}{\beta} \right)^{1/2} \right) \right),
\]

(4.2.7)

where \( R_j \) are the current magnitudes, \( G_j \) the surface elevation gradients, \( p_1 = [(f + \omega)/\beta]^2 \), \( p_2 = [(f - \omega)/\beta]^2 \), \( L_j \) and \( B_j \) are constants, ber, bei, ker, kei are the Kelvin functions, the (-) is taken when \( f < \omega \), \( i = \sqrt{-1} \), and \( j = 1, 2 \) represent anti-clockwise and clockwise rotating components respectively (Prandle, 1980). This solution is only presented to give an indication of the complexity of solutions to the mean motion when a simple, homogeneous flow is considered. No provision has been provided in this model for the influence of an inhomogeneous density distribution, which will be shown (§5.3) to be substantial for continental shelf bottom boundary layers. The distribution of turbulence (friction) in this model depends on the constants \( \alpha \) and \( \beta \), but is not limited in height due to the uniform density of the fluid. As the flows becomes even more complex (below) a simple solution to the mean flow (as in (4.2.7)) is not possible.

With rotation now affecting the mean current structure the turbulent energy equation is now approximated by the two-dimensional extension of (4.2.2)

\[
-\rho \left( \frac{u'w'}{\partial z} + \frac{v'w'}{\partial z} \right) = \bar{\epsilon},
\]

(4.2.8)

where I have returned to the Reynolds stress notation \((u'w')\) instead of the eddy viscosity model \((K_m)\) which will be dealt with in §4.3. Equation (4.2.8) assumes that the turbu-
lence reacts quickly enough to variations in the mean flow $\bar{U}, \bar{V}$ to use a local energy production–dissipation balance (Soulsby, 1983). A profile of $\epsilon$ still provides information on both the steady state rate of turbulent kinetic energy production and dissipation.

To this turbulent, time-dependent boundary layer, we add density stratification. At first, we allow the density to vary only in the vertical. Models such as (4.2.7) are still applicable, but the turbulent fluctuations and current veering are usually confined to the well mixed, weakly stratified layer that forms immediately above the bottom. An appropriate eddy viscosity in the stratified water above the well mixed layer is not obvious. In a similar model by Weatherly and Martin (1978), the current (Ekman) veering is found to be concentrated near the top of the well mixed layer.

The turbulent kinetic energy equation must now include the effects of the stratification on the velocity fluctuations. For steady state conditions the terms in the turbulent kinetic energy equation must be measured over periods that are long compared to the local decay time for turbulence (in a stratified flow outside the $bbl \sim 0.1$ to $0.2 \times \frac{2\pi}{N}$, Crawford, 1986) and short with respect to variations in the mean flow. The rate of production of turbulent kinetic energy will now be balanced at all depths by viscous dissipation and work done against gravity,

$$-\rho \left( u'w' \frac{\partial \bar{U}}{\partial z} + v'w' \frac{\partial \bar{V}}{\partial z} \right) = \epsilon + g \rho \bar{w}' . \tag{4.2.9}$$

The stratification suppresses the vertical turbulent velocities ($w'$) which must now do work against gravity as well as viscous stresses. The result is a reduction in the Reynolds stress and therefore the amount of turbulent kinetic energy produced. The viscous dissipation rate is therefore reduced because the production is down and there is an additional sink for the turbulent energy not included in (4.2.2). The reduction in Reynolds stresses
reduces the height of the tbbl from the value it would have in the absence of stratification.

In spite of the work that must be done against gravity in a stratified flow, the shear stress ($\tau_0$) near the bottom is usually strong enough to generate a limited region of isotropic turbulence. This turbulent mixing gradually forms a layer of almost neutrally stratified water immediately above the bottom. Richards' (1982) model predicts that the height of this well-mixed layer ($\text{wml}$) in an oscillatory, initially stratified boundary layer grows indefinitely, with the growth rate diminishing, but never becoming zero. A buoyancy flux downwards towards the boundary is required to limit the height of the wml.

There are now two distinct regions of the turbulent bottom boundary layer distinguished by the density distribution; the wml and the stratified water above it. These regions are governed by different dynamics, the difference being the influence stratification imposes on the turbulent fluctuations.

Internal waves can play a significant role in the energy balance above the wml and can radiate energy away from the boundary layer (Townsend, 1965). The waves can be generated by turbulent fluctuations at the top of the wml, or by shear instabilities at the interface between the wml and the stratified water. The important property of internal wave generation is that the wave energy is extracted from the mean flow where the waves are produced, and dissipated by diffusion and breaking elsewhere. This radiative energy loss by internal waves from the boundary layer can be significant and large enough to affect the dynamics of the layer itself (Townsend, 1968). In this case, the turbulent kinetic energy equation (4.2.8) becomes insufficient, and a term representing the radiative energy loss, $W_R$, must be included in (4.2.8). The form of $W_R$ depends on
the exact means of wave production (Turner, 1973). The generation of internal waves may be significant enough to limit the growth of the turbulence and dominate the turbulent kinetic energy equation. Internal waves generated near the boundary are also important to mixing at mid-depths where they dissipate.

Scorer (1985) presents a colourful discussion on turbulence in a stratified flow, and argues that the most important alteration of the fluid dynamics when stratification is included is the ability of the medium to radiate energy in the form of internal waves. The suppression of vertical velocities has a net result on the turbulence of mainly reducing the vertical diffusion. The intrinsic property of turbulence to have a tendency towards equipartition causes the turbulence to be nearly unchanged at the dissipation scales.

In a three-dimensional, stratified flow the turbulent energy balance (4.2.9) becomes,

$$\overline{\rho} \frac{d \overline{U}}{dz} + \frac{d \overline{V}}{dz} = \epsilon + g \rho \overline{w^2} + W_R,$$

(4.2.10)

and represents turbulence production balanced by viscous dissipation, buoyancy flux and internal wave radiation. The actual mechanism for the generation of internal waves by turbulence is not clear, but it is believed to rely on resonant displacement of isopycnals (Townsend, 1965). Kelvin-Helmholtz (K-H) instabilities are often a more effective generation source of internal waves (Thorpe, 1975) and conditions could arise (i.e. very stable stratification and strong shear) whereby energy might be more easily transferred to internal waves than to turbulence. Under these conditions the amount of turbulent kinetic energy can be reduced, since the instability mechanism of generating waves (K-H) does not enter the energy balance (4.2.10), but can reduce the energy available for turbulent shear production. Thorpe (1975) notes that of the kinetic energy extracted from the mean flow, the part due to Kelvin-Helmholtz instabilities can be distributed such that
approximately 10% goes into turbulent mixing and as much as 16% can be radiated away by internal waves.

Finally, we reach a level of complexity commensurate with that of the real ocean by considering the influence of local topography and horizontal density variations on the structure of the bottom boundary layer. Sloping bottoms, ridges and canyons are common-place on continental shelves and can steer the flow, modify the vorticity and allow focussing of internal waves. A variety of different water masses also meet on continental shelves. Coastal upwelling brings cold, nutrient rich water up from the continental slope, and river runoff can introduce freshwater near shore. This three-dimensional density structure is continuously transformed and advected by the time-dependent currents. The horizontal and vertical shears shuffle the various water masses, forming fronts and plumes. Vertical profiles of temperature, salinity and density become time-dependent on time scales similar to those of the fluctuating currents. Under these conditions, advection can dominate the density structure and local mixing is only important if it is sustained for periods longer than associated with advection of different properties.

The typical continental shelf bottom boundary layer will evolve over a tidal cycle through varying levels of turbulence in a changing density structure arising from contributions of all above effects. Direct measurements of turbulent energy dissipation, obtained over a tidal cycle, with density and current measurements have allowed us to recognize and resolve some of the physics of the bottom boundary layer at a level not hitherto possible. The measurements can also be compared to the various structures predicted by the models. Boundary layer measurements made on the continental shelf off Vancouver Island are described in §§ 5 and 6.
4.3 Eddy Coefficients: Theory

Most models of the turbulent structure in the oceanic bottom boundary layer adopt "second order" closure theories to describe the interaction of the turbulence and the mean flow (Prandle, 1980; Weatherly and Martin, 1978; Richards, 1982; Soulsby, 1983). These models assume forms of eddy coefficients that are tractable and provide mathematical solutions to the dynamical equations (i.e. 4.2.6). For stratified flows, the two vertical eddy coefficients of primary importance are momentum ($K_m$) and density ($K_\rho$). They are defined by,

$$K_m = \frac{-\overline{u'w'}}{\overline{\partial u/\partial z}},$$  \hspace{1cm} (4.3.1)

and

$$K_\rho = \frac{g\overline{p'w'}}{\overline{\rho N^2}},$$  \hspace{1cm} (4.3.2)

where $g$ is the acceleration due to gravity and $N$ is the local Brunt-Väisälä (buoyancy) frequency. The eddy coefficient for momentum as defined by (4.3.1) is also referred to as the turbulent, vertical eddy viscosity. The physical interpretation of $K_m$ stems from the analogy made by Boussinesq (1877) that turbulent stresses act like the viscous stress, in that they are directly proportional to the local velocity gradient. The eddy coefficient for density ($K_\rho$) is a measure of the vertical diffusivity in a stratified turbulent flow.

4.3a Eddy Coefficient for Momentum, $K_m$

In the following discussion, the mechanisms responsible for the loss of energy, or the transfer of momentum due to the radiation of internal waves will not be considered (or is at least beyond the scope of this work and unobtainable with the present measurements) and (4.2.8) is assumed to describe the turbulent kinetic energy balance. Within the wml, buoyancy effects and current veering are minimal, and the turbulent kinetic energy
equation (4.2.8) is further simplified to (4.2.4). In this layer, it is possible to model the eddy viscosity by,

\[ K_m = l u_*, \quad (4.3.3) \]

where \( l \) is a mixing length and \( u_* \) the friction velocity. The depth dependence of \( l \) determines the vertical distribution of \( K_m \) and I shall investigate two forms; a purely linear mixing length \((l = \kappa z)\) and a modified linear mixing length given by (Blackadar, 1962),

\[ l = \frac{\kappa z}{1 + \frac{\kappa z}{l_0}}, \quad (4.3.4) \]

where \( l_0 \) is an asymptotic mixing length to be determined. Intuitively, one might expect \( K_m \) (and \( l \)) to decrease in the outer region of the turbulent boundary layer, and both the linear and constant mixing length theories must become inappropriate at some distance from the bottom. Also, at large \( z \) (above the wml), \( w' \) is likely to be affected by buoyancy and (4.2.4) no longer approximates (4.2.8). Both mixing length theories are therefore restricted to the wml.

The substitution of (4.3.1) with \( -\overline{u'w'} = u_*^2 \) and (4.3.3) into (4.2.4) gives,

\[ \epsilon = \rho \frac{u_*^3}{l}, \quad (4.3.5) \]

representing a turbulent kinetic energy balance in the turbulent wml. Substitution of (4.3.4) into (4.3.5) gives,

\[ \epsilon = \rho \frac{u_*^3}{\kappa z} + \rho \frac{u_*^3}{l_0}. \quad (4.3.6) \]

This model predicts a production and dissipation rate distribution with a term inversely proportional to \( z \) and a constant, asymptotic value. The linear approximation \( l = \kappa z \)
results in the distribution of $\epsilon$ with only the first term on the lefthand side of (4.3.6). Mellor and Yamada (1974) suggest a value for $l_0$ given by,

$$l_0 = \gamma \frac{\int_0^\infty z q \, dz}{\int_0^\infty q \, dz},$$

where $\gamma$ is an empirical constant (in the literature $\gamma$ ranges from 0.05 to 0.3) and $q^2 = (u^2 + v^2 + w^2)$. This form has also received much attention by modelers (Weatherly and Martin, 1978; Mofjeld and Lavelle, 1984; Richards, 1982) but hinges on the choice of $\gamma$. In (4.3.7), $l_0$ represents a scaled "centre of mass" length scale for the distribution of turbulent kinetic energy, and determines the asymptotic dissipation rate. No measurements of $l_0$, and therefore of $(\rho u_*^3/l_0)$ have been reported in the literature for continental shelf bottom boundary layers.

The form of (4.3.6) suggests two regions that can be identified in a profile of $\epsilon$ through the turbulent $\text{wml}$. Very near the bottom ($\kappa z < l_0$) the first term dominates, and the dissipation rate is inversely proportional to $z$. This layer is called the constant stress layer and will receive special attention in §5.2. Further from the bottom ($\kappa z > l_0$) the second term dominates and the dissipation rate approaches a constant value $(\rho u_*^3/l_0)$. These two regions will be referred to as the constant stress layer (CSL) and the constant dissipation rate layer (CDRL), respectively. Dissipation rate estimates in the CSL are used to estimate the friction velocity $u_*$ and the bottom stress $\tau_0$. For some of the dissipation rate profiles presented here, the height and magnitude of the constant dissipation layer, and $l_0$ can, for the first time, be estimated directly from measurements.
4.3b Eddy Coefficient for density, $K_\rho$

The vertical eddy coefficient for density $K_\rho$ is a measure of the vertical diffusivity in a stratified turbulent flow. The stipulation that the flow be both turbulent and stratified for $K_\rho$ to be meaningful follows from the definition (4.3.2). If the fluid is neutrally stratified, then the fluctuating density anomaly $\rho'$ and the buoyancy frequency $N$ are both zero. The eddy coefficient for density is therefore undefined.

The well mixed layers observed over the continental shelf west of Vancouver Island where always weakly stratified with the temperature slowly decreasing to the bottom. Mixing in the wml may be important to quantities other than density ($\rho$), such as oxygen, in which case the eddy coefficient $K_\rho$ is an indicator of the local vertical diffusivity of any scalar (Turner, 1973).

Mixing within and above the bbl is important to the redistribution of quantities over the continental shelf. The following discussion then applies to upper and mid-depth regions as well as to the stratified part of the bbl.

If, in contrast to the neutrally stratified case, the fluid is so strongly stratified that turbulence cannot be sustained, then $K_\rho$ approaches zero due to the reduction of the correlation $\overline{\rho'w'}$. An internal wave field may have $\rho' \neq 0$ and $w' \neq 0$, but the product $\overline{\rho'w'} \approx 0$, since the fluctuations are out of phase. This seems reasonable, since internal wave motion is distinctly different from turbulence and does not induce vertical mixing, unless of course, the waves break.

To estimate $K_\rho$ using (4.3.2) directly, one requires accurate measurements of $\overline{\rho'w'}$, a quantity not obtainable from FLY II measurements. To this end we introduce the flux Richardson number, $R_f$, defined to be the ratio of the rate of turbulent energy loss
through buoyancy effects to that of turbulent energy production. Given the turbulent kinetic energy equation ((4.2.8) without Ekman veering),

$$-\bar{\rho}u'w' \frac{\partial \bar{U}}{\partial z} = \epsilon + g\rho'w', \quad (4.3.8)$$

representing the rate of production, viscous dissipation and the loss to buoyancy, the flux Richardson number is,

$$R_f = \frac{g\rho'w'}{-\bar{\rho}u'w' \frac{\partial \bar{U}}{\partial z}}. \quad (4.3.9)$$

For the existence of turbulence, and therefore a non-zero viscous dissipation rate, $R_f$ must be less than 1.0. It can be argued that for steady state turbulence, $R_f$ must be significantly (~ an order of magnitude) less than 1.0 (Stewart, 1959). The reasoning is that the turbulence generated by a shear $\frac{\partial \bar{U}}{\partial z}$ is initially in the $u'^2$ component, from which the $v'^2$ and $w'^2$ components are redistributed by turbulent pressure-velocity correlations. The dissipation by viscosity ($\epsilon$) occurs through all three components, while the loss to buoyancy is through $w'$ only. As $w'$ is suppressed by the buoyancy forces, all three terms in (4.3.8) are reduced, and turbulence becomes a less efficient sink for momentum from the mean flow. At a certain $R_f$ there is sufficient partition of energy between the two sinks and the source for the maintainance of steady state turbulence. This condition occurs at the critical flux Richardson number, $R_{fcrit}$.

Values for $R_{fcrit}$ in the literature include Ellison’s (1957) theoretical prediction of $\sim 0.15$, Britter’s (1974) measurements where $R_{fcrit} \simeq 0.18 - 0.2$, and Oakey’s (1985) analysis of oceanic data, (where to a good approximation $\bar{R}_f \simeq R_{fcrit}$ (Crawford, 1986)) which gives $R_{fcrit} \simeq 0.21(\pm 0.16)$.

The flux Richardson number and the eddy coefficient for density are related upon
substitution of (4.3.2) and (4.3.9) into (4.3.8):

\[
K_p = \frac{R_f \epsilon}{\bar{\rho}(1 - R_f)N^2}.
\]  

(4.3.10)

For steady state turbulent flows we may assume that, if \(R_{fcrit}\) represents the flux Richardson number, then an average \(\overline{K_p}\) can be estimated by substituting \(R_{fcrit}\) into (4.3.10) (Osborn, 1980). With \(R_{fcrit} = 0.21\) we have, for steady state conditions,

\[
\overline{K_p} \approx \frac{0.265\epsilon}{\bar{\rho}N^2}.
\]  

(4.3.11)

It should be noted that (4.3.11) is valid for mean, steady state conditions only. Bursts of turbulence can occur in strongly stratified flows, resulting in locally high, short-lived \(R_f\) and \(K_p\) values. However, there is some evidence for geophysical flows that, even in these bursts, a production - dissipation balance may be appropriate, since the decay time of turbulence is much less than the time scale of the forces generating the bursts. If so, the flux Richardson number may be near the critical value much of the time (Crawford, 1986).

The distribution of \(K_p\) over the shelf and the temporal variations at fixed locations will be discussed in §6.2. From the time series of \(\epsilon\) and \(N^2\), the time series contour plots of \(K_p\) at the microstructure stations can be estimated (4.3.11), and these distributions will be discussed with respect to the general circulation over the shelf.
5. Observations of the Bottom Boundary Layer

Introduction

From the theory discussed in the last section it is clear that there are some fundamental features of the bottom boundary layer that are revealed by measurements of the vertical distribution of density and of turbulent dissipation rate. These features are: the distribution of sink terms in the turbulent kinetic energy equation (from which we may infer source terms), the distribution of mixing throughout the turbulent boundary layer, the distinction between the turbulent bottom boundary layer (\( \text{tbbl} \)) and the well mixed layer (\( \text{wml} \)), the magnitude of the bottom stress, the height of the constant stress layer (CSL), and the magnitude and height of the constant dissipation rate layer (CDRL). In §5.2 the dissipation rate values in the constant stress layer are used to evaluate the bottom stress and a drag coefficient. In §5.3 the evolution of a tidally driven \( \text{tbbl} \) is presented, and in §5.4, estimates of the constant dissipation rate layer are presented.

5.2 Bottom Stress and Drag Coefficient Estimates

There are three physical processes from which the bottom stress may be derived: 1) the mean velocity profile near the bottom, 2) direct measurements of the Reynolds stress \( \rho \overline{u'w'} \), and 3) the turbulent dissipation rate measured within the constant stress layer. All three techniques have physical limitations. Reported here are bottom stress values obtained from near-bottom dissipation rate values calculated from turbulent fluctuations at viscous scales. The main differences between the three techniques are outlined below, with a discussion of their assumptions and restrictions.
a) Velocity Profile Technique

The structure of a neutrally stratified bottom boundary layer is determined by the magnitude of the flow far from the boundary and by the roughness of the bottom. These two parameters determine the form of the bottom shear stress. In the following discussion the shear stress at the bottom will be denoted by $\tau_0$ and the roughness of the bottom surface will be characterized by $z_0$. The importance and influence of wave-induced stresses are acknowledged, but in the following discussion it will be assumed that current shear alone dominates. This is valid in deeper regions where surface wave motion does not penetrate to the bottom, except perhaps during storms. For more details on the influence of waves and swell in the bottom boundary layer the reader is referred to Grant and Madsen (1979). The mean velocity distribution in the boundary layer is then a function of the shear stress $\tau_0$, the roughness parameter $z_0$, the distance to the bottom $z$ and the fluid properties $\rho$ and $\mu$, where $\rho$ is the seawater density and $\mu$ is the dynamic viscosity. The characteristic velocity scale in the bottom boundary layer is called the friction velocity $u_*$,

$$u_* = \sqrt{\frac{\tau_0}{\rho}}. \quad (5.2.1)$$

It has been observed that the Reynolds stress near the bottom is constant over a layer (the constant stress layer) and is equal to the bottom stress $\tau_0$. It can be shown (Hinze, 1975) that with a linear mixing length $l = \kappa z$ the mean shear in the constant stress layer is

$$\frac{\partial \bar{U}}{\partial z} = \frac{u_*}{\kappa z}, \quad (5.2.2)$$

(the law of the wall) and upon integration we obtain the logarithmic velocity profile,

$$\bar{U} = \frac{u_*}{\kappa} \ln \frac{z}{z_0}, \quad (5.2.3)$$
where $z_0$ is the constant of integration and represents a length scale proportional to the roughness elements at the bottom. A similar profile is obtained for a turbulent flow over a hydrodynamically smooth bottom. Under such conditions $z_0$ in (5.2.3) is replaced by a viscous length scale proportional to the height, $\delta$, of the viscous sublayer and the constant of integration is chosen to match the velocity at the top of the viscous sublayer ($\delta u^*_2/\nu$),

$$\bar{U} = \frac{u_*}{\kappa} \ln \frac{z}{\delta} + \frac{\delta u^*_2}{\nu}.$$  

(5.2.4)

Tennekes and Lumley (1972) employ a velocity deficit law to obtain a similar velocity distribution. They consider the reduction in the mean geostrophic velocity ($U_g$) due to shearing stresses introduced into the flow as friction becomes important near the bottom. This logarithmic velocity profile is given by,

$$\frac{U - U_g}{u_*} = \frac{1}{\kappa} \ln \frac{z}{\delta_p} + B,$$  

(5.2.5)

where $U_g$ is the far field velocity (assumed to be geostrophic), $\delta_p$ is the height of the turbulent planetary boundary layer (defined as the region where friction is important) and $B$ is the constant of integration. The constant stress layer is then only the lower portion of the velocity distribution (5.2.5). The choice for $\delta_p$ and the constant of integration $B$ in the velocity deficit case is not obvious (Yaglom, 1979).

By measuring the mean velocity at a number of heights above the bottom one can do a linear least squares fit between $\bar{U}$ and $\ln z$, from which $u_*$ and $z_0$ in (5.2.3) can be calculated from the slope and intercept respectively. The roughness parameter $z_0$ can also be estimated from the bottom microtopography, but it is not clear how this calculation can be successfully repeated from one environment to another (Grant and Madsen, 1986).

Grant et al. (1984) measure the velocity at heights of 28, 53, 103 and 203 cm above the bottom and find that in order to maintain uncertainties in $u_*$ of less than $\pm 25\%$, 

74
they require linear regression coefficients from their least squares fit to (5.2.3) of better than $R^2 = 0.993$. Therefore, despite the fact that a logarithmic profile is predicted by theoretical assumptions, extremely accurate current measurements are required to obtain only adequate estimates in $u_\ast$. This is partly due to the fact that the logarithmic approximation is valid to a much greater height in the boundary layer than the approximation of a constant stress layer (Soulsby, 1983). Another reason is that the logarithmic velocity profile predicts small shears at the operating heights of current meters, while simultaneously predicting large shears as $z$ approaches $z_0$, where current meters work less well (Gust, 1985; Griffin, 1987). It is the measurements in this lower region (the constant stress layer) that provide the most accurate values of $\tau_0$. Accurate measurements of the mean current ($\overline{U}$) close to the bottom ($< 1$ m) are difficult due to the large ratio of fluctuating turbulent velocities to the mean flow, the influence of form drag on the current (Chriss and Caldwell, 1982) and the low overall speeds which are often near the stall speed and resolution of the current meter (Gust, 1985). It follows that a friction velocity ($\text{and } \tau_0$) may be calculated from current measurements made outside the constant stress layer and fitted to a logarithmic function, but that the estimate of $u_\ast$ obtained in this fashion may not be representative of the actual bottom stress.

b) Reynolds Stress Estimates

The most direct method of estimating the bottom stress $\tau_0$ is to assume a layer of constant Reynolds stress near the bottom,

$$\tau_0 = -\rho u'w',$$  \hspace{1cm} (5.2.6)

and measure $u'$ and $w'$ in this layer. Field measurements of $u'w'$ have been attempted by Soulsby (1983) and Gross and Nowell (1985). Soulsby measures $u'$ and $w'$ at fixed
heights above the bottom using electromagnetic current meters. The product $u'w'$ is calculated over 8 to 12 minute time series. These measurements are restricted to wavelengths in the inertial subrange and are very sensitive to the orientation and tilting of the instrumentation ($\tau_0$ varies $\pm 10\%$ per degree). Also, in order to recover the full covariance, theoretical spectral shapes are used to extrapolate the measured spectra to higher wavenumbers, above the inertial subrange. The behaviour of the product $u'w'$ is inherently intermittent and Soulsby (1983) found that 90% of the calculated variance in $u'w'$ occurs during only 26% of the time. Direct Reynolds stress measurements are therefore very sensitive to the period over which the average is calculated and to the tilt of the instrument.

c) Dissipation Rate Technique

Within many stationary turbulent flows, and certainly within the well-mixed portion of a non-accelerating bottom boundary layer ($< 10 \text{ m}$) there is a local balance between the rate of production of turbulent kinetic energy through the Reynolds stress working on the mean shear and the rate of viscous dissipation of that energy (4.2.4). Within the constant stress layer the Reynolds stress is equal to the bottom stress and is given by (5.2.6). If there is a balance between production and dissipation rate of turbulent kinetic energy (4.2.4), then in the constant stress layer,

$$\epsilon = \frac{u_3}{\kappa z}. \quad (5.2.7)$$

If $\epsilon$ can be measured within this layer, at a distance $z$ from the bottom, then $u_*$ can be calculated. The cubic dependence implies that errors in $\epsilon$ translate to smaller errors in $u_*$ and intermittency in $\epsilon$ results in only slight fluctuations in $u_*$ and $\tau_0$. 

76
c.1) Inertial Dissipation Rate Technique

Previous estimates of the dissipation rate $\epsilon$, for bottom stress analysis (e.g. Grant et al. 1984; Gross and Nowell, 1985; and Huntley and Hazen, 1987) have been obtained from measurements of turbulent velocities at inertial scales, and will be denoted as the inertial dissipation technique. This calculation follows from the theories of Kolmogoroff (1941). Briefly, Kolmogoroff proposed that if the Reynolds number is sufficiently high then the fluctuating velocities at scales much smaller than the energy-containing eddies are locally isotropic. Kolmogoroff’s first hypothesis then asserts that the kinetic energy spectrum in this isotropic range will depend only on $k$ (the three-dimensional wavenumber associated with the fluctuations), $\nu$ (the kinematic molecular viscosity) and the rate at which energy is passing through the spectrum from larger to smaller scales. This rate of energy transfer must equal the dissipation rate $\epsilon$ of turbulent kinetic energy at the smallest (viscous) scales.

From dimensional analysis (Grant et al. 1982) the three-dimensional, isotropic, kinetic energy spectrum takes the form,

$$E(k) = \left( \frac{\epsilon}{\rho} \right)^{\frac{2}{3}} k^{-\frac{5}{3}} F \left( \frac{k}{k_s} \right),$$

(5.2.8)

where

$$k_s = \left( \frac{\epsilon}{\rho \nu^3} \right)^{\frac{1}{4}},$$

(5.2.9)

is the Kolmogoroff wavenumber and $F \left( \frac{k}{k_s} \right)$ is a universal, non-dimensional function. Kolmogoroff’s second hypothesis further simplifies (5.2.8) by proposing that at even higher Reynolds numbers there exists a range of wavenumbers between the energy containing eddies and the viscous scales in which viscous forces are unimportant. This range of
wavenumbers is called the inertial subrange and within it, (5.2.8) takes the form,

\[ E(k) = A \left( \frac{\epsilon}{\rho} \right)^{\frac{3}{5}} k^{-\frac{5}{3}}, \quad (5.2.10) \]

where \( A \) is the three dimensional Kolmogoroff constant \( (A \approx 1.589, \text{ Nasmyth, 1986, personal communication}) \). If the Reynolds number is sufficiently large and an inertial subrange exists then the friction velocity can be found from (5.2.7) and (5.2.10),

\[ u_* = \left\{ \frac{E(k)k^{\frac{5}{3}}}{A} \right\}^{\frac{1}{2}} (\kappa z)^{\frac{1}{3}}, \quad (5.2.11) \]

where \( k \) is taken within the inertial subrange.

Although this theory is rather straightforward, the existence of an inertial subrange in the ocean can not always be assumed. In stratified flows, the assumptions in Kolmogoroff’s theories can break down, as energy is extracted from the flow due to buoyancy forces. The presence of stable stratification will suppress the larger eddies and reduce the Reynolds number, limiting the inertial subrange. An inertial subrange only exists if the Reynolds number is very large and turbulent kinetic energy is neither added nor extracted from the flow at inertial scales. If turbulent kinetic energy is extracted at inertial scales by buoyancy effects, then the dissipation rate calculated from inertial velocity spectra (5.2.11) will tend to overestimate the true dissipation rate of turbulent energy at viscous scales. The spectral slope may also be affected by energy loss in the inertial subrange, as the -5/3 slope only represents a transfer of energy between scales without loss of energy (Turner, 1973).

Breakdowns in the assumptions for an inertial subrange have been observed in the ocean. Gargett et al. (1984) show significant departures from a \( k^{-\frac{5}{3}} \) dependence for moderate turbulence \( (\epsilon \sim 4.0 \times 10^{-3} \text{ to } 3.0 \times 10^{-4}W \text{ m}^{-3}) \) in stratified flow \( (N \sim \)
4.2 \times 10^{-2} \text{ to } 1.1 \times 10^{-2} \text{ s}^{-1}). \text{ Armi and Flament (1985) further point out the extreme caution required when extrapolating spectral slopes with turbulent velocity spectra and fitting theoretical spectral slopes to real flows, as in (5.2.10). Huntley (1987) finds that, in most constant stress layers, the assumptions in Kolmogoroff’s theory are not satisfied. He concludes that } u_* \text{ estimates calculated from velocity measurements within the inertial subrange are only valid for energetic flows where } u_* > 0.8 \pm 0.2 \text{ cm s}^{-1} \text{ or when corrections are applied that extrapolate the inertial subrange theories below depths where there is no separation of the production and dissipation scales. Gross and Nowell (1985) calculate the friction velocity from velocity measurements at inertial scales and find that they required } u_* > 2.0 \text{ cm s}^{-1} \text{ in order to resolve the full dissipation rate within 1 metre of the bottom. Most of their estimates are therefore calculated from dissipation values obtained above } z = 1 \text{ m where an inertial subrange is more likely to exist, but where a constant stress layer may not.}

Measurements of inertial scale velocity fluctuations within the constant stress layer are further restricted by the inhomogeneity of the flow at small distances from the boundary. Inhomogeneous turbulent flows will have non-zero divergence and as such are not represented by the production – dissipation balance (5.2.7). The assumptions of homogeneity and isotropy are valid for different scales of the velocity fluctuations at different heights } z \text{ from the boundary. To a first order approximation, turbulence is isotropic at wavenumbers } k > \frac{2\pi}{z} \text{ (Pond, 1965). The constant stress layer, as suggested by current measurements (Hinze, 1975), could be restricted to heights } z < \frac{\nu^{2000}}{u_*} \text{ for hydrodynamically smooth flow (see page 89), so, as the Reynolds number and } u_* \text{ increase and the inertial subrange extends to larger scales, the height of the constant stress layer simul-}
taneously decreases; in which case an isotropic inertial subrange may never exist within the constant stress layer. Typical oceanic values of $\nu$ and $u_*$ ($\nu \approx 1.5 \times 10^{-6} \text{ m}^2/\text{s}$ and $u_* \approx 4.0 \times 10^{-3} \text{ m/s}$) predict a constant stress layer less than 0.75 m in height (Hinze, 1975). Inertial subrange velocity fluctuations at these distances from the boundary will be highly anisotropic, let alone homogeneous. Spectral shapes are therefore unlikely to exhibit universal forms in the inertial subrange and dissipation rates estimated from such measurements will not represent the true viscous dissipation rate.

c.2) Viscous Dissipation Rate Technique

The dissipation technique is not limited by many of the factors noted above if the dissipation rate $\epsilon$ is calculated from velocity measurements at higher wavenumbers, within the dissipation, or viscous subrange. These fluctuations are more likely to be isotropic, theoretical spectral shapes and slopes are not required and contamination by wave-induced velocities and form drag will be weak. Successful measurements of the dissipation rate in oceanic turbulence have been obtained using shear probes as described by Osborn (1974), Osborn and Crawford (1980) and Oakey and Elliott (1982). Shear probes are used to measure velocity fluctuations at scales between $\sim 0.01 \text{ m}$ and $\sim 1.0 \text{ m}$, a band which usually spans the viscous subrange. The viscous subrange is the band of scales (wavelengths) where the kinetic energy is being dissipated directly by viscosity. Therefore, velocity measurements at scales within the viscous subrange (dissipation scales) may be used to estimate contributions to the total dissipation rate $\epsilon$, without the need to fit the data to particular spectral shapes. The component of turbulent shear measured by a shear probe $\left( \frac{\partial u}{\partial z} \right)$ is fully recovered, and the assumption of isotropy is only used to approximate the remaining contributions to $\epsilon$ (4.2.2) (Monin and Yaglom, 1975). This $\epsilon$
is then used in (5.2.7) to estimate the friction velocity and therefore the bottom stress. This is the dissipation technique, distinct from the inertial dissipation technique of Grant et al. (1984) and Gross and Nowell (1985).

Shear probes are usually mounted on vertical free-falling microstructure vehicles (Elliott and Oakey, 1979; Osborn and Crawford, 1980; Lueck et al. 1983; Caldwell et al. 1985; Dewey et al. 1987), or on horizontally traversing platforms (Gargett et al. 1984; Lueck, 1987, personal cummunication). Using Taylor's frozen turbulence hypothesis (3.1.1), a time series of turbulent microstructure shear is converted to a space series of the signal \( \frac{\partial u}{\partial z} \) sampled at discrete depths. The power spectrum of this signal is the one dimensional spectral function \( k_3^2 \phi_{11}(k_3) \), where \( k_3 \) represents the vertical wavenumber and \( \phi_{11}(k_3) \) is the spectrum of the horizontal velocity \( u_1 \). The dissipation rate per unit volume \( \epsilon \) for isotropic turbulence can be expressed as integrals over the three and one dimensional spectral functions (Monin and Yaglom, 1975),

\[
\epsilon = 2\mu \int_0^\infty k^2 E(k) dk = \frac{15}{2} \mu \int_0^\infty k_3^2 \phi_{11}(k_3) dk_3. \tag{5.2.12}
\]

Although the integrals in (5.2.12) are theoretically taken from 0 to \( \infty \), Figure 8 shows that integration up to \( (k_3/k_s) \approx 0.56 \) will recover more than 90% of the area under the normalized, universal form of the \( k_3^2 \phi_{11}(k_3) \) spectrum given by (3.3.10).

The actual integration limits chosen for each spectrum are determined so that a minimal amount of low and high frequency noise (Dewey et al. 1987) is included in determining the variance (5.2.12). The small percentage of the variance assumed lost is then added back as explained earlier (§3.3).

Velocity measurements at dissipation scales \( (k > 0.03 k_s) \) are also less susceptible to contamination from oscillating currents due to surface waves, which contribute to
measurements of velocities within the inertial subrange \((k < 0.03 k_s)\). The \(k_2^2 \phi_{11}(k_3)\) spectra presented here have been calculated from sections of the microstructure shear profile < 3.8 s long (< 2.5 m in depth) and velocity contributions due to wave-motion are therefore not resolved. If the wave stresses actually generate turbulence, then this contribution to the dissipation rate is measured. In conditions where there is significant wave-motion near the bottom (ie. during storms), free-falling shear probe apparatus would detect the wave-induced shear at a vertical scale approximately > 2 m and such components are eliminated when the local trend is removed from the signal (§3.3).

Homogeneity at the energy containing eddy scales is not a requirement for spectral analysis of viscous scale shear measurements since the variance \(\overline{\frac{\partial u}{\partial z}}^2\) may be estimated directly from the shear time series. Spectral techniques are employed to increase the signal to noise ratio, and to allow various wavenumber-domain corrections to the signal to compensate for the spatial limitations of the probe (§3.3).

The dissipation rate values obtained within the constant stress layer using (5.2.12) are then used to estimate the friction velocity and the bottom stress predicted by (5.2.7).

A slight error (over estimate) is introduced in \(u_*\) by assuming that the time series in the constant stress layer is homogeneous (stationary) and that the recovered variance \(\overline{\frac{\partial u}{\partial z}}^2\) represents the value at the center of the time series. If (5.2.7) represents the true constant stress layer (CSL) distribution of \(\epsilon\), then the variance calculated by (3.3.7) represents the dissipation rate at a slightly lower height than the center of the time series, \(z\). Individual microstructure shear time series are intermittent and non-stationary to a greater degree than the \(z^{-1/2}\) dependence predicted by (5.2.7) and therefore no depth corrections within the constant stress layer were applied to compensate for this non-
stationarity. The maximum error in $z$ introduced by this assumption is 9.1% for the first 128 point time series, which represents a 3.0% error in $u_*$. This error is reduced by $2/5$ for each subsequent estimate.

Figure 16 shows a dissipation rate profile ($\#506$) from station M2. The friction velocity has been found using (5.2.7) within the constant stress layer. The following discussion outlines the selection of the CSL height and the calculation of the friction velocity in that layer.

Determining the height of the constant stress layer (CSL) is rather subjective and the criterion used is the height above which the standard deviation in the running average $\bar{u}_i|_{-3}$, calculated over the preceding 4 $u_*$ values, is a minimum. This technique identifies the height at which the $u_*$ values calculated using (5.2.7) start to deviate from the nearly constant values immediately next to the bottom. Table 3 shows the $u_*$ estimates from the near-bottom $\epsilon$ values and the identified CSL height for the $\epsilon$ profile shown in Figure 16.

Table 3: Table of individual $\epsilon$ and $u_*$ values and estimated CSL height for profile 506 at M2.

| $z_i$ (m) | $\epsilon_i$ (W m$^{-3} \times 10^{-4}$) | $u_{*i}$ (cm s$^{-1}$) | $\bar{u}_i|_{-3}$ (cm s$^{-1}$) | Standard Deviation in $\bar{u}_i|_{-3}$ | $\bar{u}_i|_3$ (cm s$^{-1}$) |
|-----------|-----------------------------------|-----------------------|---------------------------------|---------------------------------|-------------------|
| 0.32      | 7.765                             | 0.463                 |                                 |                                 | 0.378             |
| 0.49      | 2.631                             | 0.372                 |                                 |                                 | 0.378             |
| 0.65      | 1.056                             | 0.301                 |                                 |                                 | 0.378             |
| 0.82      | 1.638                             | 0.377                 | 0.378                           | 17.5%                           | 0.378             |
| 1.15      | 1.181                             | 0.379                 | 0.357                           | 10.4%                           | 0.378             |
| 1.49      | 0.703                             | 0.347                 | 0.351                           | 10.3%                           | 0.373             |
| 1.82      | 1.15                              | 0.437                 | 0.385                           | 9.7% $\Rightarrow$             | 0.382             |
| 2.16      | 0.543                             | 0.361                 | 0.381                           | 10.4%                           | 0.380             |
| 2.49      | 0.246                             | 0.291                 | 0.359                           | 16.8%                           | 0.370             |
| 2.82      | 0.354                             | 0.342                 | 0.357                           | 16.9%                           | 0.367             |

$\Rightarrow u_* \simeq 0.382$ cm s$^{-1} \pm 14.1\%$

CSL height $\simeq 1.8$ m $\pm 0.16$ m
The CSL heights estimated using this technique are greater than those predicted by the velocity measurements of Hinze (1975) where \( z_{CSL} \approx \frac{\nu^2}{\overline{u}_*} \). These could then be overestimates or Hinze's values may be low. The gradual departure of the measured \( \epsilon \) profile from the CSL distribution (5.2.7) implies that there is no abrupt transition in the stress at the top of the CSL and the Reynolds stress \( \tau \) deviates slowly from the constant stress value \( \tau_0 \). Figure 16b shows the same profile but with a logarithmic ordinate. The dashed line represents the theoretical dissipation rate profile as calculated using \( \epsilon = \rho \overline{u}_*^3 / \kappa z \) with the friction velocity obtained from the measured dissipation rate values. The cubic relation between \( \epsilon \) and \( \overline{u}_* \) fixes the slope of the theoretical \( \epsilon \) profile (dashed line in Figure 16b). The estimated \( \overline{u}_* \) is the average obtained from the \( \epsilon \) values within the CSL. The height of the CSL is theoretically arbitrary and depends on the choice of a threshold, say \( \gamma \), at which point \( |\tau - \tau_0| \approx \gamma \). The CSL height estimates will vary significantly from profile to profile due to the intermittency of turbulence. Confidence in the CSL height can be improved by averaging several dissipation rate profiles.

Once the height of the constant stress layer has been estimated the dissipation rate values in that layer, say “\( n \)” individual \( \epsilon_i \) values, are used to calculate “\( n \)” \( \overline{u}_* \) estimates. The average of these “\( n \)” values is then the \( \overline{u}_* \) estimate for that profile, and the standard deviation of the “\( n \)” estimates can be calculated. For a single profile the standard deviation is typically of order 20%. Confidence in a calculated value of \( \overline{u}_* \) is increased by averaging several consecutive \( \epsilon \) profiles. (Anywhere from 3 to 20 \( \epsilon \) profiles where averaged depending on the profiling rate, the depth and the stationarity of the mean flow.) The standard deviation in \( \overline{u}_* \) calculated from the average \( \overline{\epsilon} \) values is typically 10%. Figure 19 shows the average dissipation rate values from three quasi-synoptic profiles (obtained in
less than 10 minutes) through the bottom boundary layer with improved confidence in
the $\overline{u}_*^+$ and CSL height estimates.

The CSL is resolved much better when many profiles can be averaged. Turbulent
dissipation rate profiles from station W05 in Hecate Strait (Figure 18) were collected
in July 1983 in water 35 m deep. Series of 20 profiles where obtained over periods of
approximately 40 minutes. For these measurements the profiler carried two orthogonally
aligned shear probes so that two dissipation rate profiles could be calculated for each
profiler descent. Therefore, 40 individual dissipation rate profiles were collected in \( \sim 40 \)
minutes. The orthogonal arrangement of the shear probes confirms local isotropy in the
horizontal plane between the \( \frac{\partial u}{\partial z} \) and \( \frac{\partial v}{\partial z} \) turbulent shear components. Although this is
informative and does add confidence to measurements made with only one microstructure
shear component \( \frac{\partial u}{\partial z} \), it is a weak test for true, three-dimensional isotropy. Figure 20
shows two individual dissipation rate profiles from the two shear probes measured during
a single profiler descent.

From measurements made during non-accelerating phases of the current, it is possi­
ble to average consecutive profiles over periods as long as 20 minutes without violating
stationarity (Grant et al. 1982). Figure 21 shows one such average profile of 20 individual
profiles sampled in \( \sim 18 \) minutes. The averaging smooths the intermittency and reveals
the mean turbulent structure. These average profiles also show the increased accuracy
in resolving the constant stress layer. Standard deviations in the friction velocities cal­
culated from the average dissipation rates are consistently of order 10%, and have been
as low as 3%.

The dissipation rate measurements made in 1985 on the continental shelf west of
Figure 19  a) A dissipation rate profile of average dissipation rate values from three consecutive profiles through the bottom boundary layer.  b) The same data shown in a) but now with a log ordinate axis. The data is from station B2 (Figure 1) in water 68 m deep.
Figure 20 Two dissipation rate profiles calculated from two shear probes mounted side-by-side during a single profiler descent. The data were collected in Hecate Strait at 0557 PST July 8, 1983 at station W05 (Figure 18). The water depth was 30 m.

Vancouver Island at M2 (Figure 1) were collected over a complete diurnal tidal cycle. A time series of the average friction velocity and the average current magnitude $U(3)$ at 3 metres above the bottom is shown in Figure 22. The current mooring was approximately 1 km north of the microstructure observations, therefore local fluctuations in the current data over short periods (10 minutes) may not represent flow fluctuations at
Figure 21 An example of an averaged dissipation rate profile. The profile is the average of 20 individual dissipation rate profiles calculated from 10 profiler descents. The averaging smooths out any intermittency and shows the logarithmic layer immediately above the bottom. Above the constant stress layer is a layer of constant dissipation. The data is from Hecate Strait, station W05, collected in July 1983. The water depth at W05 is 30 m.
the microstructure site, and visa versa. The values $u_*$ and $\bar{U}(3)$ in Figure 22 are one hour averages to smooth out any rapid fluctuations. The correlation coefficient between the friction velocity and current magnitude at 3 m is $R^2 = 0.94$ at zero phase lag. A similar comparison was not possible for the Hecate Strait data due to a malfunction in the current meter placed near the bottom.

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**Figure 22** Time series of the average friction velocity $u_*$ and the average current magnitude $|\bar{U}|$ at 3 m height from stations M2 and C2 respectively. The correlation coefficient between these series is $R^2 \approx 0.94$. The observations were made between 0900 PST June 17 and 0600 PST June 18, 1985. The water depth was approximately 138 m.

An estimate may now be made of the accuracy in assuming that the acceleration term $\frac{\rho \partial q^2}{2 \partial t}$ in the turbulent kinetic energy equation (4.2.3) may be neglected in (4.2.4) and (5.2.7). Soulsby (1983) indicates that within the CSL the turbulent kinetic energy per unit mass can be approximated by $\frac{1}{2} q^2 \approx 5.0 u_*^2$. In which case

$$\frac{\rho \partial q^2}{2 \partial t} \approx \rho 5.0 \frac{\Delta u_*^2}{\Delta t},$$
where $\Delta u_2^2$ and $\Delta t$ can be calculated from Figure 22. During maximum acceleration (between hours 0900 and 1100) the approximate rate of change of turbulent kinetic energy per unit volume within the CSL is $\sim 6.0 \times 10^{-6}$ W m$^{-3}$. This is a factor of ten less than the mean dissipation rate within the CSL over the same period. The assumption of omitting the acceleration term in (5.2.7) appears to be valid.

The drag coefficient $C_D$ is defined here as

$$C_D = \frac{u_*^2}{\overline{U(1)}}. \quad (5.2.14)$$

The current at one metre above the bottom is extrapolated from the data shown in Figure 22, where a least squares fit of (5.2.4) to the measured values of $\overline{u_*}$ and $\overline{U(3)}$ predicts the $\overline{U(1)}$ values. The average drag coefficient at M2 is found to be

$$C_D = 0.82 \times 10^{-3} \ (\pm 14\%),$$

where the uncertainty is the standard deviation from 11 individual $C_D$ estimates. This $C_D$ value implies that the flow is hydrodynamically smooth (Sternberg, 1968). From (5.2.4) the height of the viscous sublayer, $\delta$, is calculated; it varies between 0.7 cm and 1.2 cm over a diurnal cycle, in good agreement with the measurements of Caldwell and Chriss (1979). The height of the viscous sublayer is found to depend on the viscous length scale ($\frac{\nu}{u_*}$) as

$$\delta \simeq (24.2 \pm 3.4) \frac{\nu}{u_*}.$$ 

The constant of proportionality (24.2) is twice that found from laboratory experiments (Monin and Yaglom, 1975, p277), but close to the values measured on the Oregon shelf by Chriss and Caldwell (1984). The latter believe that the constant is not universal. A bottom grab at M2 revealed that the sediments where made up of a sticky mud, a
surface that is very likely to be "smooth". No information about bed forms and local topographic irregularities was available and an estimate of the actual bottom "roughness" is not possible.

Chriss and Caldwell (1982) indicate that under hydrodynamically smooth conditions the actual bottom stress can be as much as four times lower than the stress calculated from current measurements in the logarithmic portion \((z > 15 \text{ cm})\) of the boundary layer. When the stress is calculated from the slope of (5.2.4) fitted to the current measurements made at 3, 5 and 10 metres above the bottom, the values are consistently 3.8 times larger than the estimates found from the dissipation rate profiles. In predicting a higher bottom stress \((\text{higher } u_*)\), the method of fitting a logarithmic function to current measurements in this region also predicts a rougher bottom \((\text{smaller } \delta)\) and a higher drag coefficient.

The discrepancy between the bottom stress estimates calculated from dissipation rate profiles and those from the current velocity data has two significant consequences. First, the current data is outside the constant stress layer and although the current profile may be approximated by a logarithmic function \((\text{average correlation coefficient of } R^2 \sim 0.91)\), the extrapolation that gives the bottom stress from that function is inaccurate. The second is the influence of form drag on the current meter measurements (Chriss and Caldwell, 1982), which is important when large irregularities in the boundary influence the total stress farther from the boundary, outside the constant stress layer. Skin friction dominates the Reynolds stress within the constant stress layer and determines the sediment transport conditions. Therefore, bottom stress estimates made from inertial velocity or current measurements outside the constant stress layer could be erroneous due to the influence of form drag. Dissipation rate profiles through the CSL calculated from
viscous scale measurements provide bottom stress estimates that are directly related to the Reynolds stress within the constant stress layer, and thus provide a measure of the skin friction. Since sediment transport is determined by skin friction and the resulting Reynolds stress, dissipation rate estimates from viscous scale measurements allow a more accurate calculation of the bottom stress than near-bottom current measurements (Chriss and Caldwell, 1982).
5.3 The Time-Dependent Boundary Layer at M2

The microstructure measurements from station M2 and the current measurements from C2 will be discussed together to show the evolution of the time-dependent bottom boundary layer. The microstructure measurements were collected over a 22 hour period, nearly one diurnal tidal cycle. Three properties of the bottom boundary layer structure will be analyzed; the distribution of the rate of turbulent kinetic energy dissipation, the density structure and the current structure.

Bottom stress ($\tau_0$) values have been calculated from the dissipation rate values within the constant stress layer, immediately above the bottom (§4.3). Variations in the bottom stress are shown to correlate well with the diurnal current near the bottom (Figure 22). The constant stress layer at M2 is less than 3m in height throughout the tidal cycle. The results indicate that the flow at M2 is hydrodynamically smooth, with a drag coefficient of approximately $C_D \approx 0.82 \times 10^{-3}$.

Density profiles, calculated from the temperature and conductivity signals recorded by the microstructure profiler are used to identify the bottom well mixed layer (wml), which is that portion of the density profile immediately above the bottom where the water is weakly stratified (Figure 23). From the FLY II data a typical buoyancy frequency in the wml is found to be $N \approx 8.0 \times 10^{-4}$ s$^{-1}$. A wml was observed on almost every profile made on the Vancouver Island shelf (over 1200 profiles between 1983-85).

The current meter records can be used to identify a number of features. The instantaneous current magnitudes and directions at the fixed heights identify the current shear and therefore regions of turbulence production. The current measurements can also be used to estimate where the water is coming from, relative to the topography, providing a
Figure 23 A $\sigma_t$ profile showing the layer of weakly stratified water immediately above the bottom, approximately 36m deep. This layer will be referred to as the well mixed layer (wml). The data is from station M2 and was collected on 17 June 1985. The water depth is approximately 138 m. An over-shoot in the conductivity signal results in what appears to be small, negative density gradients within the wml. These are believed to be erroneous.

rough description of the three-dimensional water movement. The current meter records also contain temperature measurements and these can identify the advection of different water bodies past the mooring.
Figure 24a shows a 36 hour portion of the current records from the meters at 3, 10, 30 and 88 m above the bottom. The current meters at 3, 10 and 30 m from the bottom were Aanderaa RCM4s sampling at 5 minute intervals, and the meter at 40 m under the surface (88 m from the bottom at C2) was an InterOcean S4 sampling at 1.5 minute intervals. Vectors are averaged and shown at 10 minute intervals. The periods over which simultaneous microstructure measurements were obtained are indicated (a,b,c,...). The current magnitude at $z = 3$ m is well correlated with the bottom stress (Figure 22a). The distribution of shear as revealed by the current records will assist in the analysis and interpretation of the turbulence and density data. The current record at 5 m above the bottom is not shown because the currents below $z \leq 10$ m were always well correlated and the shear in this region is well described by the two records at $z = 3$ m and $z = 10$ m.

Significant current veering was observed only outside the wml. It is likely that most of the boundary layer current veering takes place within the pycnocline at the top of the wml, as predicted by Weatherly and Martin (1978). This is difficult to confirm in the records presented, since there is only one meter in the vicinity of the pycnocline, at $z = 30$ m. The record does show that when the wml is deeper than 30 m, the current at $z = 30$ m is in the same direction as the bottom currents. When the wml is shallower than 30 m, the current at $z = 30$ m veers from those within the wml.

The current data were also filtered (to one sample per hour) and passed through the Foreman (1979) tidal current harmonic analysis programs. Twenty two tidal constituents were either extracted directly, or inferred from each current record (Appendix C). These constituents were then passed back through the Foreman (1979) programs to generate
Figure 24  a) A 36 hour portion of the current records from mooring C2. The meter heights are 3, 10, 30 and 88 m. The total depth at C2 is approximately 135 m. North is directed upward and the time interval between vectors is 10 minutes.
synthetic tidal currents. Figure 24b shows these predicted currents for the same 36 hour period as the currents in Figure 24a. The differences between Figures 24a and 24b are then residual fluctuations at non-tidal frequencies and are shown in Figure 24c. The tidal component of the current at $z = 88$ m accounts for approximately 40% of the observed current, while nearer the bottom, as much as 80% of the current is accounted for by the tides. The turbulent structure of the bottom boundary layer reacts quickly to current changes and in the following analysis, the full unfiltered currents (Figure 24a) will be considered.

Figure 16 shows a profile of dissipation rates calculated from a microstructure shear profile. The turbulent bottom boundary layer ($t_{bbl}$) is identified by the layer of strong turbulent dissipation rates of approximately $\epsilon > 10^{-6}$ W m$^{-3}$, which in Figure 16 is below $z \approx 26$ m. The height of the $t_{bbl}$ varies over the diurnal tidal cycle, but the structure is dependent on variations in the stratification as well as on the distribution of shear. The stratification and the height of the $w_{ml}$ are revealed by $\sigma_t$ profiles (Figure 23). A simple relation between the $t_{bbl}$ and the $w_{ml}$ was not observed. They are often modeled to be the same height, a situation rarely observed in almost 1200 profiles on the continental shelf. The reasons for the observed differences will be analyzed in the remainder of this section.

Figure 25 shows time series of the height of the $w_{ml}$ and the $t_{bbl}$ over the 22 hour microstructure observation period at M2. There are two periods when the height of the $w_{ml}$ is significantly different from the height of the $t_{bbl}$. In most turbulent boundary layer models (Soulsby, 1983; Richards, 1982), this condition cannot occur. The $w_{ml}$ and $t_{bbl}$ are typically modeled to be dependent on one another in such a way that the height
Figure 24 b) A 36 hour portion of the tidal predictions calculated from the tidal constituents extracted by harmonic analysis of the filtered (1 hour interval) data shown in Figure 24a.
Figure 24  c) The residual currents after the tidal component plotted in Figure 24b has been subtracted from the raw data shown in Figure 24a. The non-tidal currents are dominant in the upper layer and weak near the bottom.
of the **tbbl** is the height of the **wml**. The boundary layer observed at M2 clearly shows that this is not the case (Figure 25).

---

**Figure 25** Time series of the well mixed layer (**wml**) height (●) and the turbulent bottom boundary layer (**tbbl**) height (▲) at M2. The times are PST for June 17 to June 18, 1985. The height of the **wml** is the height of the weakly stratified water as revealed by σ₁ profiles (e.g. Figure 23). The height of the **tbbl** is the height where the turbulent dissipation drops below $10^{-6}$ W m$^{-3}$. These heights are shown on the individual profiles in Figure 26.

---

**Weatherly et al.** (1980) model a bottom boundary layer driven by mean and diurnal currents. They attempt to show that if the diurnal tidal currents are of similar magnitude to the mean currents, the bottom boundary layer thickness will be at least twice that when only the mean current is considered. Their **tbbl** is never deeper than the **wml** plus the thickness of the pycnocline at the top of the **wml**. Pynoclines at the top of the **wml** were rarely deeper than 2 to 3 metres. Figure 25 indicates that the **tbbl** can be
substantially deeper than the \textit{wml}.

There are four mechanisms which could result in simultaneously different \textit{wml} and \textit{tbbl} heights.

1) During the decelerating phase of an oscillating current, the \textit{tbbl} diminishes, leaving a \textit{wml} that represents an integrated history of previous mixing. The \textit{tbbl} will shrink faster than the \textit{wml} provided the rate of viscous dissipation of turbulence is more rapid than the rate of mixing which would erode the \textit{wml} from above.

2) If a strong density front (i.e. dense water displacing a less dense boundary layer) is advected past the observation site, then the observed density structure represents events (mixing) that occurred upstream of, and prior to the observations. In this case the \textit{wml} again represents an integrated history of previous events while the active turbulent velocities measured by a sensor represent the instantaneous mixing process. Density structures can be advected without significant change, while turbulence decays rapidly and can be advected only short distances (Crawford, 1986).

3) Ekman pumping could induce a convergence at the bottom, thus causing the denser water in the \textit{wml} to accumulate and form a thick \textit{wml} without the presence of turbulence. This mechanism is a localized phenomena and can only occur near the centre of a cyclonic circulation.

4) The final mechanism for producing different \textit{wml} and \textit{tbbl} heights occurs when current shear associated with the bottom boundary layer extends above the \textit{wml} and turbulence is generated and dissipated in the stratified water found there. This type of current shear could be associated with Ekman veering or a wall jet (Townsend, 1976).

To determine which, if any, of these processes is responsible for the observed differ-
ences in the \textit{wml} and \textit{tbbl} heights (Figure 25), a careful analysis of the simultaneous
current, density and turbulence time series is required. Figure 26 shows time series of
dissipation rate and $\sigma_t$ profiles from M2. The dissipation rate profiles in Figure 26 are
the average of 4 deep (surface - to - bottom) profiles from each set of 10 profiles collected
within the 50 minute sampling period. The sampling procedure consisted of alternating
one deep profile through the entire water column with two short profiles through the
bottom boundary layer only. The first profile of series 1 at M2 was obtained at 0900
hours on June 17, 1985, with following series starting at 2 hour intervals, on the odd
hour. The \textit{wml} and \textit{tbbl} heights are indicated on both the $\epsilon$ and $\sigma_t$ profiles.

The $\sigma_t$ profiles in Figure 26 are individual profiles calculated from the slow tem­
perature and conductivity signals (Appendix A) and each profile is representative of the
mean density structure throughout a 50 minute sampling period. Due to a gain problem
in the conductivity circuit, the negative gradients frequently observed near the top of
the \textit{wml} are believed to be erroneous. Such structure could be caused by internal wave
billowing, but this cannot be determined by these measurements.

Determining the height of the \textit{tbbl} often relies on rather subjective criteria. Here I
have chosen the \textit{tbbl} height to be the height at which the $\epsilon$ values drop to $10^{-6}\,\text{W m}^{-3}$.
This represents a reasonably low level of turbulence, but is above the noise level of the
instrumentation. For some dissipation rate profiles the \textit{tbbl} is easily identified as the
only turbulent layer immediately adjacent to the bottom (e.g. profile number 2a). For
dissipation rate profiles with a more complicated structure (e.g. profile number 8a), it
is necessary to study the current records for a better understanding of the structure of
the mean flow and the distribution of shear (and therefore turbulence production). The
Figure 26  a) Time series of average dissipation rate profiles, and b) individual $\sigma_t$ profiles at M2. The average dissipation rate profiles are from 4 profiles through the entire water column. The (●) indicate the height of the wml as revealed by the $\sigma_t$ shown in b). Similarly, the (▲) indicate the approximate height of the tbbl ($\epsilon \approx 10^{-6}$ W m$^{-3}$) as revealed by the $\epsilon$ profiles in a). Data were collected on 17 and 18 of June 1985. The total water depth at M2 is 138 m.
confidence intervals about the tbbl heights in Figure 25 reflect the difficulty encountered during these more complicated periods. The tbbl can be composed of several turbulent regions (e.g. profile 8a, Figure 26). An alternative to identifying one turbulent bottom boundary layer would be to define multiple turbulent boundary layer heights, each associated with some part of the boundary layer mean flow. This approach seems equally as subjective and would rarely simplify the analysis.

The first occurrence of different wml and tbbl heights (Figure 25) is during the start of the ebb portion of the diurnal bottom current (hours 0900-1100 Figure 24a). The tbbl is a single layer of turbulence increasing in e towards the bottom (profile 2a, Figure 26). The wml is more than twice the height of the tbbl.

During the next 6 hours, an anomaly in the boundary layer density structure is advected past M2. Between hours 1200 and 1600 (profiles 2 and 4) the wml height grows from approximately 26 m to 40 m. During this period, the tbbl grows from 14 m to 29 m. The wml grows without simultaneous mixing associated with the tbbl! There is no change in the density of the wml during this event. After 1600 PST the tbbl height exceeds the wml height until they are the same at 0530 PST on 18 June.

The conditions outlined above indicate the advection past M2 of a deep wml. The observed changes in the density structure are associated with advection, rather than local mixing (see also §5.2). To illustrate this we shall consider the equation for the conservation of mass.

The conservation of mass for a volume of fluid at a fixed depth (or height) is

\[
\frac{\partial \rho}{\partial t} + U_i \frac{\partial \rho}{\partial x_i} = K_{\rho,i} \nabla^2 \rho, \quad (5.3.1)
\]

where \(K_{\rho,i}\) is the eddy coefficient for density in the \(i^{th}\) direction and \(\nabla^2 = \frac{\partial^2}{\partial x_i \partial x_i}\). The
terms in (5.3.1) represent, respectively, the net rate of change of density, the rate of change due to mean advection and the rate of change due to mixing. The magnitude of the two last terms can be roughly estimated and will indicate the relative importance of advection to mixing.

If we consider a unit volume of fluid at the top of the wml at 1330 PST \((z = 35\) m, profile 3b, Figure 26), estimates of the advection and mixing terms in (5.3.1) can be approximated.

The magnitude of the advection terms can be approximated from the CTD data (Figure 41) where the horizontal density gradient near the bottom is approximately \(\frac{\Delta \rho}{\Delta z} \approx 2.3 \times 10^{-5} \text{ kg m}^{-4}\). The component of the current along this gradient is estimated from the current data at C2 to be \(\overline{U} \approx 8.0 \text{ cm s}^{-1}\). The rate of change due to advection is approximately

\[
\overline{U} \frac{\partial \rho}{\partial x} \approx 1.84 \times 10^{-6} \text{ kg(m}^3\text{s})^{-1}
\]

Of the mixing terms in (5.3.1), the contributions from vertical mixing dominate over the horizontal terms even though horizontal eddy coefficients are typically \(10^6\) times larger than the vertical eddy coefficient (Pond and Pickard, 1981). This is due to the very weak horizontal density gradients. The vertical eddy coefficient for density is calculated using (4.3.11) and has a maximum in the region of the pycnocline of \(\overline{K_{\rho,3}} \approx 10^{-5} \text{ m}^2\text{s}^{-1}\) (see §6.3 and Figure 50). The mixing contribution to the conservation of mass is approximately

\[
\overline{K_{\rho,3}} \frac{\partial^2 \rho}{\partial z^2} \approx 1.5 \times 10^{-7} \text{ kg(m}^3\text{s})^{-1},
\]

over a magnitude less than the contribution from advection. This result indicates that the time-dependence of the tbbl and wml is more complicated than the models discussed
if §4 would predict.

Weatherly and Martin (1978) observe a wml that grows when the current direction within the bbl is down slope and advection brings warmer water into the observation area. The wml observed at M2 suggests a similar structure, but is due to the advection of denser (colder) water at mid-depths (profile 4b, Figure 26), rather than the advection of a less dense (warmer) wml as observed by Weatherly and Martin (1978). The total mass in a vertical water column at M2 has increased due to the increased density of the water at mid-depths. The increased density at depths below \( z = 80 \) m represents an increase, averaged throughout the water column, of \( 0.127 \text{ kg/m}^3 \). The density of the wml is virtually unchanged during this process although its increased thickness represents an additional increase in the total mass of the region.

The strong pycnocline at the top of the wml after 2130 hours (profile 9b, Figure 26) separates the bottom boundary layer into three distinct regions; the turbulent wml, the turbulent pycnocline at the top of the wml, and the turbulent stratified outer boundary layer between the top of the bbl and the top of the wml. In stratified turbulent flows, turbulent stresses are inefficient in transferring momentum across isopycnals (and pycnoclines). Entrainment of turbulent kinetic energy from the boundary can not extend beyond the wml due to the suppressed turbulent stresses at the pycnocline. The turbulence in this outer layer is therefore locally generated, and dissipated. Diffusive eddy viscosity models do not predict this outer turbulent layer.

Turbulence within the wml maintains a nearly neutral density stratification. The turbulence in the stratified water above the wml, although restrained by the work done against gravity, gradually mixes and erodes the stratification. This mixing produces
a density structure, similar in appearance to that which results from double diffusion (Phillips, 1977), with step gradients above the wml (profile 8b, Figure 26). It is important to mention that double diffusion is not expected to be significant within the continental shelf bottom boundary layer and is not responsible for the observed density structure. The formation of steps implies horizontal layers of turbulent mixing. This layering could also be a result of turbulence generated by Kelvin-Helmholtz billowing (Turner, 1973), but this process is unsubstantiated by the present data. Turbulence above the wml persists for 12 hours, from 1700 to 0500 hours (Figure 25), during which time the wml height does not increase.

During the period when the tbbl is deeper than the wml there is a distinct reduction in the viscous dissipation rate at the height of the pycnocline (profile 9a, Figure 26). The amount of energy lost to buoyancy may be significant and the flux Richardson number (§4.2) must be very near the critical value $R_f = R_f \text{crit} \simeq \frac{1}{5}$. If this is the case, then $\frac{4}{5}$ of the turbulent kinetic energy is being dissipated by viscosity and $\frac{1}{5}$ by the work done against gravity. The observed $\epsilon$ values decrease by more than $\frac{1}{5}$ (in fact by an order of magnitude) and imply a layer near the pycnocline of either reduced turbulent kinetic energy production due to the suppression of the vertical component of turbulent velocity by the stratification, or an additional energy sink, or both.

Thorpe (1975) notes that at pycnoclines, Kelvin-Helmholtz instabilities transfer approximately 10% of the energy removed from the mean flow into mixing (not included in (4.1.10)) and as much as 16% might be radiated away via internal waves. Townsend (1968) on the other hand finds that the amount of energy radiated away by internal waves in the outer regions of a boundary layer may be more than 50% of the energy lost from
the mean flow in that region. Carruthers and Hunt (1985) model a boundary layer with a turbulent layer underlying a stratified layer, and show that when the dissipation rate is independent of height, the vertical flux of internal wave energy is approximately 10% of the total energy dissipated by viscous stresses within the turbulent bottom layer. The evidence thus suggests that internal wave generation is important under certain conditions. An estimate of the rate of energy radiation can be made following the method of Carruthers and Hunt (1985) for boundary layers with constant dissipation rate layers.

5.4 The Constant Dissipation Rate Layer (CDRL)

The mixing length theory (4.3.4) predicts two regions that can be identified by a profile of $\epsilon$ through the $wml$. In §5.2 the bottom stress ($\tau_0$) and friction velocity ($u_*$) were estimated from the dissipation rate values within the constant stress layer ($z < 3.0 m$). Above the constant stress layer, (4.3.6) predicts a constant dissipation rate layer (CDRL), where the mixing length is $l_0$ and the dissipation rate is $\epsilon = \rho u_*^3 / l_0$. At large distances from the boundary a constant mixing length $l_0$ is more realistic than a linear model $l = \kappa z$, for which $l$ increases unrealistically for large $z$. The height of the CDRL and the region where (4.3.6) is valid are restricted to within the $wml$, since $l_0$ is an inappropriate mixing length when stratification restricts vertical excursions. Since (4.2.2) represents the balance of turbulent energy production and dissipation in the $wml$, a CDRL implies a constant production layer. Oceanic models implementing the mixing length (4.3.4) have been successful in predicting a turbulent $wml$ (Weatherly and Martin, 1978; Mofjeld and Lavelle, 1984; Richards, 1982) but are unsupported by measurements of real dissipation rate distributions.

Observations in the atmospheric boundary layer by Caughey and Palmer (1979) and
Kaimal et al. (1976) show that under certain conditions the dissipation rate is approximately constant with height. These observations were made in convective boundary layers. Such conditions are rarely found near the ocean bottom (Weatherly and Martin, 1978). Nonetheless, I believe some of the dissipation rate profiles presented here are the first oceanic bottom boundary layer measurements to show a dissipation rate that is nearly constant with height above the CSL.

In §5.2 the bottom stress is determined from dissipation rate values within the CSL. If the mixing length is modeled by (4.3.4), the turbulent kinetic energy equation is approximated by (4.3.6), which is valid in and above the CSL but not outside the wml. At certain periods of the diurnal \textbf{tbbi} the vertical dissipation rate profiles do exhibit CSL and CDRL distributions (profile 9a, Figure 26, and Figure 21). Figure 27 shows two average dissipation rate profiles with (4.3.6) fitted to the lower portion. The value of $u_*$ estimated from (4.3.6) is within 10% of the value estimated by (5.2.7) fitted only within the CSL.

The CDRL predicted by (4.3.6) is not always present. The time series of average microstructure profiles at M2 (Figure 26) suggests that the modified mixing length theory (4.3.4) is valid after the \textbf{tbbi} is fully developed, up to 7 hours after the start of the flood tide, and as much as 4 hours after the near-bottom currents reach a maximum. This observation is a result of the turbulence farther from the boundary lagging the bottom stress by 4 to 6 hours (30° to 50°) (Figures 22 and 26). Once the \textbf{tbbi} is fully developed, the dissipation rate distributions persistently show the two-layer structure predicted by (4.3.6). For these periods, the values of $u_*$, $l_0$ and the estimated height of the CDRL are shown in Table 4. Also shown are the $u_*$ estimates found using (5.2.7) and the
Figure 27 a) An example of an average dissipation rate profile with (4.3.6) fitted to the lower portion. Equation (4.3.6) models the \( \epsilon \) distribution with a constant stress layer near the bottom and a constant dissipation rate layer further from the bottom, above the constant stress layer. Also shown are the \( u_\ast \) and \( l_0 \) values calculated using (4.3.6). This is an average of all 10 profiles centered at 2330 PST, June 17, 1985.

approximate CSL heights.

The model of Carruthers and Hunt (1985) predicts that when the dissipation rate is independent of height, and the turbulent boundary layer is capped by a layer of strong stratification, internal waves can be generated in the stratified layer. They estimate
that the total amount of turbulent kinetic energy dissipated by viscosity throughout the turbulent layer is one order of magnitude greater than the energy radiated away by the locally generated internal waves. Integrating the dissipation rate throughout the CDRL of profile 8a (Figure 26 and Figure 27a) gives an approximate maximum radiation rate of

\[ W_R \approx 4.0 \times 10^{-5}, \]
Table 4: Comparison between CSL (5.2.7) and CDRL (4.3.6) analysis at M2

<table>
<thead>
<tr>
<th>Time (PST)</th>
<th>Δ T</th>
<th>Analysis using (5.2.7)</th>
<th>Analysis using (4.3.6)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Time After Minimum Current</td>
<td>u_0 (cm s^-1)</td>
</tr>
<tr>
<td>23:30 June 17, 1985</td>
<td>14 hrs</td>
<td>0.380 (±0.023)</td>
<td>0.96 (±0.13)</td>
</tr>
<tr>
<td>01:30 June 18, 1985</td>
<td>16 hrs</td>
<td>0.340 (±0.077)</td>
<td>1.0 (±0.12)</td>
</tr>
<tr>
<td>03:30</td>
<td>18 hrs</td>
<td>0.265 (±0.026)</td>
<td>1.1 (±0.11)</td>
</tr>
<tr>
<td>05:30</td>
<td>20 hrs</td>
<td>0.133 (±0.031)</td>
<td>1.2 (±0.12)</td>
</tr>
</tbody>
</table>

in W m\(^{-3}\), where it is assumed that 1/10 of the integrated dissipation rate is the vertical flux of internal wave energy (Carruthers and Hunt, 1985).

5.5 Conclusions

In conclusion, there are some very interesting features of the continental shelf bottom boundary layer that are revealed by time series of vertical dissipation rate profiles. Of primary importance is the distinction between the turbulent bottom boundary layer (\textit{tbbl}) and the well mixed layer (\textit{wml}). The two are not the same, and conditions of identical \textit{tbbl} and \textit{wml} heights could be the exception rather than the rule. This observation would indicate that stratification is not a dominant factor in determining the height of the \textit{tbbl} and therefore the frictional or Ekman layer. The reason for this is that the \textit{wml} represents a history of mixing and the observed variations at a fixed site will be dominated by the mean advection. The active \textit{tbbl} on the other hand reacts to the local bottom current magnitude and shears between the \textit{wml} and interior flows. The advective terms in a conservation of mass estimate are found to be more than 10 times larger than the contribution from local mixing.
From analysis of the limited current data, there appears to be a relation between the height of the wml and the distribution of current veering. Within the wml the current magnitude increases with height \( z \), but the current direction is depth-independent. When current veering was observed, it was concentrated within the pycnocline at the top of the well mixed layer. This agrees with the observations of Weatherly and Martin (1978). Conditions were also observed when the current magnitude decreased with increasing height \( z \), \( \left( \frac{\partial U}{\partial z} < 0 \right) \). This condition was usually accompanied by turbulence dissipation in the stratified water above the wml.

The choice of an appropriate turbulent kinetic energy equation for the tbbl depends on the density structure of the bbl and the region of the bbl under investigation. Within the wml it is appropriate to choose (4.3.2). Within a few metres of the bottom (approximately \( z < 3 \) m) the Reynolds stress is very nearly equal to the bottom stress, and (4.3.4) can be approximated by (5.2.7). No significant Ekman veering was detected within the wml (at C2) and (4.2.8), which includes two components of shear \( \frac{\partial U}{\partial z} \) and \( \frac{\partial V}{\partial z} \), would appear to be inappropriate. When the tbbl height is equal to, or greater than the wml height, the buoyancy terms must be included in the overlap region. In this case (4.3.10) is appropriate, with the internal wave radiation loss \( W_R \) suspected to be significant in regions of sustained turbulence and strong stratification.

The difference between turbulence in a neutrally, or weakly stratified fluid and in a stably stratified fluid is discussed in §4.2. Suffice to say that there are three sinks for the turbulent kinetic energy (viscous dissipation, loss to buoyancy, and the generation of internal waves) in the stably stratified case, and only one (viscous dissipation) for neutral stratification. The work done against buoyancy forces and the generation of internal waves
in fact involve similar terms. The distinction being that the energy lost through single
turbulent displacements in the vertical \((g\rho'w')\) does not radiate energy away, as is the
case with energy loss due to the generation of internal waves. An estimate of the rate of
energy flux away from the CDRL due to internal waves is put at \(W_R \simeq 4.0 \times 10^{-5} \text{W m}^{-3}\). 
6. Distribution of $\epsilon$ and Mixing Over the Continental Shelf

6.1 Introduction

The distribution of kinetic energy dissipation and the distribution of mixing over the shelf west of Vancouver Island can be identified from the six microstructure stations M1, M2, M3, M5, M6 and B2 (Figure 1). The distribution of $\epsilon$ indicates areas where kinetic energy is being lost from the mean flow to turbulent kinetic energy and dissipated by viscosity. Similarly, the distribution of $\overline{K_p}$ identifies regions of active mixing. These distributions will be presented, along with interpretations of how they are related to the local dynamics.

6.3 The Distribution of Kinetic Energy Dissipation

The microstructure stations (M1-M6 (less M4) and B2, Figure 1) were selected to provide an indication of where active mixing is occurring on the shelf. Current moorings (C1-C4) were deployed next to stations M1, M2, M3 and M7 respectively. As of yet no microstructure data have been processed from M7, because the FLY II pressure channel went offscale at 215 m, well above the bottom at 270 m.

From the time series of profiles at each station a contour plot of the dissipation rate can be made. The 4 profiles in each one hour series through the entire water column have been averaged (as in Figure 26). These average profiles have then been contoured using NCAR routines on a VAX-1185. Contours are shown in Figures 28 to 33 for stations M1, M2, M3, M5, M6 and B2, respectively. The ordinate is height above the bottom, the abscissa is time. For all the stations where sufficient temporal data are available, there is a diurnal signal in the near-bottom dissipation rates due to the diurnal tidal currents.

The analysis presented in §5.3 of the dissipation rate profiles at station M2 indicates
that the turbulent bottom boundary layer (tbbl) is almost entirely driven by the diurnal tides. This variation is now contoured in Figure 29. The tbbl can now be approximated by the log(ε) = −5.0 contour. This choice is different than that used in §5.3 due to the smoothing and extrapolation that takes place when the data are contoured.
Figure 29 Contour plot of the log of dissipation rates from the time series of profiles made at M2. The contour intervals are one factor of ten from $10^{-3}$ to $10^{-6}$ W m$^{-3}$. Data were collected between 0900 June 17 and 0600 June 18, 1985. Total water depth is $\sim$ 140 m.

The series of contour plots from stations M1, M2, M3 and B2 (Figures 28, 29, 30 and 33 respectively) indicate an average increase in the rate of turbulent dissipation, from the shelf break towards the shore. This increase is perhaps most prevalent in the tbb1, likely due to an increase in tidal current magnitude in the shallower regions.
The contour plot of $\epsilon$ at M6 (Figure 32) shows the arrival of a "turbulent front" between 1500 and 1600 PST. The $\epsilon = 10^{-5}$ contour is seen to rise sharply, indicating either the advection of a deep layer of active turbulence, or a drastic increase in the local current shear at mid-depths. In §5.3, the observation of a density front (deep wml) was discussed. Figure 32 indicates that "turbulent fronts" also exist. With the increased $\epsilon$ at
mid-depths is a corresponding decrease in the density. Transects of $\epsilon$ profiles across the shelf might reveal the structure and importance of such features.

Table 5 summarizes the dissipation rate measurements by dividing the water column into two regions; those dissipation rates within the bottom boundary layer ($z < 30-40 \text{ m}$)
Figure 32 Contour plot of the log of dissipation rates from the time series of profiles made at M6. The contour intervals are one factor of ten from $10^{-3}$ to $10^{-6}$ W m$^{-3}$. Data were collected between 0745 June 16 and 0645 June 17, 1985. Total water depth is $\sim 130$ m.

The dissipation rates have been averaged over the observation periods (diurnal) at the six stations. Also shown are the average eddy coefficients for density, as calculated using (4.3.11). These averages make it easier to see the variation of the turbulence over
Figure 33 Contour plot of the log of dissipation rates from the time series of profiles made at B2. The contour intervals are one factor of ten from $10^{-3}$ to $10^{-6}$ W m$^{-3}$. Data were collected between 2230 June 11 and 1900 June 12, 1985. Total water depth is $\sim$ 70 m.

Two important observations can be made from this summarized version of the data. First, the average dissipation rates are always larger in the bottom layer than at interior depths. Second, the average dissipation rates increase from the shelf break towards the shore in both regions (within the bbl, and above the bbl). Due to the large range of dissipation rate values typically encountered over a diurnal period, it is likely that short
Table 5: Average dissipation rates and $K_p$ within 2 layers: in the maximum bbl, and above.

<table>
<thead>
<tr>
<th>Station</th>
<th>Dissipation Rate</th>
<th>$K_p$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>BBL (Thickness)</td>
<td>Above BBL (Thickness)</td>
</tr>
<tr>
<td>M1</td>
<td>$6.34 \times 10^{-6} \pm 47%$ (40 m)</td>
<td>$2.01 \times 10^{-6} \pm 6.3%$ (90 m)</td>
</tr>
<tr>
<td>M2</td>
<td>$3.07 \times 10^{-5} \pm 30%$ (40 m)</td>
<td>$4.39 \times 10^{-6} \pm 9%$ (80 m)</td>
</tr>
<tr>
<td>M3</td>
<td>$4.00 \times 10^{-5} \pm 91%$ (30 m)</td>
<td>$6.02 \times 10^{-6} \pm 5%$ (70 m)</td>
</tr>
<tr>
<td>M5</td>
<td>$1.01 \times 10^{-5} \pm 42%$ (40 m)</td>
<td>$2.71 \times 10^{-6} \pm 12%$ (120 m)</td>
</tr>
<tr>
<td>M6</td>
<td>$9.90 \times 10^{-5} \pm 114%$ (40 m)</td>
<td>$1.12 \times 10^{-5} \pm 29%$ (70 m)</td>
</tr>
<tr>
<td>B2</td>
<td>$7.73 \times 10^{-5} \pm 13%$ (30 m)</td>
<td>$6.66 \times 10^{-5} \pm 26%$ (20 m)</td>
</tr>
</tbody>
</table>

periods of high dissipation rates dominate the average values. This would explain the high standard deviations shown in Table 5.

Two possible mechanisms could account for the increased turbulent activity at mid-depths. One is an increase in the baroclinicity of the mean flow, resulting in increased current shear (and therefore turbulence production). The other could be an increase in the internal wave activity and the resultant increased energy dissipation due to viscous dissipation of the waves. If the turbulent bottom boundary layer is a major source of internal waves, then the increase in near bottom turbulence towards the shore may be accompanied by an increase in internal wave activity at depths above the bottom boundary layer.

The presence of internal wave and internal shear-generated turbulence cannot be substantiated with the present data. Internal wave dissipation is extremely difficult to identify (Thorpe, 1975) and although the current records (e.g. Figure 24a) often show
periodic fluctuations that are likely due to internal waves, estimates of the distribution of internal wave kinetic energy is not possible. The current meter data (Figures 24a) and the dynamic heights calculated from the CTD data (between 40 and 100 m) (Figure 34) both suggest mid-depth shear.

**Figure 34** The dynamic heights calculated between 40 and 100 m depth from the CTD data collected during the CTD and Oxygen survey, and from CTD casts made at the start of each microstructure series. The units are in dynamic metres as described by Pond and Pickard (1981). The calculations and contouring was done by H.Freeland using the objective analysis technique of Denman and Freeland (1985). The spacing of contours is related to the geostrophic current magnitude by the given scale, assuming that the geostrophic current is zero at 100 m depth. Flow is along contours of constant dynamic height.
For a current shear to generate turbulence, the Richardson number

\[ R_i = \frac{N^2}{\left(\frac{\partial U}{\partial z}\right)^2} \]

should be less than the critical value 0.25 (Turner, 1973). To see if this criterion is met, the data from M2 (microstructure) and C2 (current) are considered. If the shear observed between \( z = 30 \) m and \( z = 88 \) m at C2 (2300 PST Figure 24a) is uniform over the 58 m interval, then \( R_i \approx 19.4 \). Even if the shear is concentrated between \( z \approx 70 \) m and \( z \approx 90 \), near the high dissipation zone revealed by profile 8a in Figure 26, the Richardson number is \( R_i \approx 2.5 \), still above the critical value required to initiate turbulence. It is likely that the vertical scale of the shear is less than the spacing between current meters; therefore the actual Richardson number is smaller.

To interpret the dissipation rate measurements in terms of the rate of kinetic energy loss from the mean flow, it is desirable to know what portion of the dissipation rate is due to the tidal current and what portion is due to the non-tidal current. Unfortunately, the relation between the current magnitude and the dissipation rate is non-linear (\( \epsilon \propto |U|^3 \)). Although the various contributions to the current magnitude (e.g. tidal and non-tidal) can be added linearly, the contributions to the kinetic energy dissipation rate can not. The problem of partitioning the turbulent dissipation rate among the mean current fields has no accurate solution. An approximate solution is possible due to the result that the tidal currents over the shelf region are nearly barotropic (Figure 24) and (Crawford and Thomson, 1984) and that the geostrophic current (and Tully eddy) is baroclinic and confined to the upper 100 m (Freeland and Denman, 1982).

The harmonic analysis of the current records (Appendix C) indicates that the tidal constituents vary only slightly with depth and that most tidal components are barotropic.
The dynamic heights calculated from the CTD data (Figure 34) represent the geostrophic shear between 40 and 100 m depth (assuming no geostrophic current at 100 m depth).

If the tidal current is barotropic, then the region of greatest kinetic energy loss from the tides is near the bottom where the current shear of a barotropic current is concentrated. Similarly, if the non-tidal current is baroclinic and restricted to depths above the bottom boundary layer, then the shear associated with the non-tidal current is above the bottom boundary layer. The distribution of kinetic energy dissipation is often similar to the distribution of current shear, and under the above conditions, the average dissipation rate in the bottom boundary layer may be associated with the tides and the average dissipation rate above the bottom boundary layer may be associated with the non-tidal current. This partition is based on a rather bold simplification but allows us to proceed with an analysis which would otherwise be impossible. To this end, §7 is devoted to the analysis of tidal energy dissipation and §8 to the energy dissipation from the mean or non-tidal flow.

6.3 The Distribution of Upwelling and Mixing

One of the objectives of this research was to determine regions of active mixing within the upwelling centre known as the Tully eddy. The term “eddy” refers to the circular structure and the “closed” density contours that are measured, and does not imply exact dynamical similarity to structures such as warm or cold core eddies found near the Gulf Stream. In fact, one may define an eddy as a structure that “traps” water and maintains its shape due to a rotational motion. In that case, the Tully eddy found in June 1985 may not be a true eddy, but merely a cyclonic circulation or meander that is topographically restricted and fed by processes particular to the local region. Nonetheless, I will continue
to call it the Tully eddy, for lack of a more precise characterization.

The Tully eddy is known to be seasonal, appearing some time around June. The driving mechanisms are not completely understood, but its coincidence with the southerly wind-driven shelf break current that develops in June and the maximum runoff from the Fraser River through Juan de Fuca Strait, suggest that these are two dynamically important considerations. From the measurements presented here an indication of the distribution of turbulence is possible, with the result that we can identify regions of high kinetic energy loss, and also regions of active mixing.

The Tully eddy is a highly variable structure (McIntosh, 1987, personal communications) and was present during our cruise between June 10-20, 1985. The CTD survey conducted between June 14-15 revealed a signature similar to those found by Freeland and Denman (1982) and Tully (1942). The $\sigma_t$ contours are shown in Figures 35 to 45 and were calculated by Freeland (1984, and personal communications) from the CTD data collected on June 14-15 using the objective analysis techniques of Denman and Freeland (1985). Figure 45 shows the dissolved oxygen contours (ml/l) at 50 m depth and Figure 46 shows contours of dissolved oxygen content at the bottom (this time drawn by hand).

The satellite images (Figures 2 and 3) and the $\sigma_t$ contours at all depths between 25 and 100 m (Figures 36 – 39) show that the isopycnals are displaced upward in the eddy centre, and that there is advective upwelling. The presence of advective upwelling as opposed to local vertical mixing is indicated by the distribution of mid-depth turbulence. The time series contour plots of $\overline{K}_p$ at stations M3 and M5 (Figures 47 and 48 respectively) show depths where there is very little vertical mixing. In order to mix the cold upwelled water from the bottom at 170 m to the surface, there must be active mixing.
Figure 35  The $\sigma_t$ contours at a depth of 5 m. The contours were calculated from the CTD data (42 stations, Figure 17) using the objective analysis technique of Denman and Freeland (1985). Data were collected between June 14 and 15 1985. Units are kg m$^{-3}$. Throughout the water column. Figure 48 shows that near surface (depths less than 70 m, or heights greater that $z \sim 105$ m) mixing near the eddy core is considerable.

The bathymetry of the area shows that the core of the upwelling is located over a bottom feature unofficially named the Spur Canyon (Figure 1). Freeland and Denman (1982) believe the Spur canyon to be an influential structure in the eddy formation. The canyons play an additional role to the current and vorticity steering proposed by Freeland and Denman (1982) by providing a channel for the upwelled slope water.
The upwelling conditions produced by the mean wind-driven along-shore current bring slope water up onto the shelf. The actual path taken by the slope water is along the Spur canyon floor, which joins the slope region to the middle of the shelf. This route is suggested by the oxygen values measured at the bottom during the June 1985 cruise (Figure 46). Six of the lowest values recorded follow the Spur canyon floor, from the start of the Juan de Fuca canyon to the south, up to the head of the Spur canyon to the north. This route is also suggested by the microstructure data collected at M5.
Figure 37 The $\sigma_t$ contours at a depth of 50 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are kg m$^{-3}$.

The density profiles here showed a layer immediately adjacent to the bottom where the density increases significantly as the bottom is approached. This indicates the presence of dense slope water flowing along the bottom.

Figures 49 to 52 are the contour plots of $\overline{K}_\rho$ from stations M1, M2, M6 and B2 respectively. Table 5 also summarizes the $\overline{K}_\rho$ distributions. As was the case with the dissipation rate distributions, mixing is also seen to increase shoreward of the shelf break. The mixing efficiency, or vertical eddy coefficient for density, reaches a maximum above
Figure 38 The $\sigma_t$ contours at a depth of 75 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are kg m$^{-3}$.

the bottom boundary layer at B2, over Swiftsure Bank. The average $\bar{K}_p$ there is extremely large (1.63 m$^2$s$^{-1}$), and is 4 orders of magnitude ($10^4$) larger than the average values reported by Osborn (1980) for an upper shear zone. This $\bar{K}_p$ distribution indicates that surface-to-bottom mixing is very active over the shallow banks near shore.

Mixing within the Tully eddy is less efficient than the mixing over the shallow banks, and does not extend throughout the entire water column. If the eddy plays an active role in the upwelling of nutrient-rich slope water, it is most likely in the form of advection,
Figure 39 The $\sigma_t$ contours at a depth of 100 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are kg m$^{-3}$. 
Figure 40 The $\sigma_t$ contours at a depth of 125 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are kg m$^{-3}$. 
Figure 41  The $\sigma_t$ contours at a depth of 149 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). The odd depth was chosen to include an additional data value that would not have been included had 150 m had been used. Units are kg m$^{-3}$. 
Figure 42 The $\sigma_t$ contours at a depth of 170 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are kg m$^{-3}$. 
Figure 43 The $\sigma_t$ contours at a depth of 175 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are kg m$^{-3}$. 
Figure 44 The $\sigma_t$ contours at a depth of 200 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are kg m$^{-3}$. 

\[ \Delta = 0.02 \]
Figure 45 The dissolved oxygen content contours at a depth of 50 m. The contours were calculated using the objective analysis technique of Denman and Freeland (1985). Units are ml l⁻¹.
Figure 46 The dissolved oxygen content contours at the bottom. The contours were fit by hand to the oxygen values measured at the bottom. Units are ml l$^{-1}$.
Figure 47 Contours of the logarithm of the vertical eddy coefficient for density $K_p$ (m²s⁻¹) at M3 calculated using (4.3.11). The data has been calculated from the dissipation rate and the density profiles obtained with FLY II. Contour intervals are two orders of magnitude, or two factors of ten, ranging from $K_p$ values of less than $10^{-5}$ to greater than $10^1$. Mixing throughout the water column was not observed and is therefore not responsible for the high $\sigma_t$ values near the centre of the Tully eddy (Figure 34). The dates of these observations are identical to the $\epsilon$ contours. The total water depth at M3 is ~120 m.
Figure 48 Contours of the logarithm of the vertical eddy coefficient for density $\overline{K_\rho}$ (m$^2$s$^{-1}$) at M5 calculated using (4.3.11). Contour intervals are two orders of magnitude, or two factors of ten, ranging from $\overline{K_\rho}$ values of less than $10^{-5}$ to greater than $10^1$. The total water depth at M5 is $\sim$ 180 m.
Figure 49  Contours of the logarithm of the vertical eddy coefficient for density $K_\rho$ ($m^2s^{-1}$) at M1 calculated using (4.3.11). Contour intervals are two orders of magnitude, or two factors of ten, ranging from $K_\rho$ values of $10^{-5}$ to $10^1$. The total water depth at M1 is $\sim 150$ m.
Figure 50  Contours of the logarithm of the vertical eddy coefficient for density $\overline{K}_\rho$ ($m^2s^{-1}$) at M2 calculated using (4.3.11). Contour intervals are two orders of magnitude, or two factors of ten, ranging from $\overline{K}_\rho$ values of $10^{-5}$ to $10^1$. The total water depth at M2 is $\sim 140$ m.
Figure 51 Contours of the logarithm of the vertical eddy coefficient for density $K_\rho$ (m$^2$s$^{-1}$) at M6 calculated using (4.3.11). Contour intervals are two orders of magnitude, or two factors of ten, ranging from $K_\rho$ values of $10^{-5}$ to $10^1$. The total water depth at M6 is $\sim 130$ m.
Figure 52 Contours of the logarithm of the vertical eddy coefficient for density $\overline{K}_\rho$ (m$^2$s$^{-1}$) at B2 calculated using (4.3.11). Contour intervals are two orders of magnitude, or two factors of ten, ranging from $\overline{K}_\rho$ values of $10^{-5}$ to $10^1$. The total water depth at B2 is $\sim$ 70 m.
as the mixing near the eddy centre is limited to near surface depths (depth < 70 m) and does not extend into the upwelled slope water at the bottom. Vertical flux rates for $NO_3$ at M5 are discussed in §8.2.
7. Tidal Dissipation

7.1 Introduction

There are two applications of an estimate of the tidal kinetic energy dissipation rate from tidal flows over the continental shelf. First is a determination of the decay rate for the diurnal shelf waves that dominate the tidal signal over most of the shelf west of Vancouver Island (Crawford and Thomson, 1984). Second is an extrapolation to the global rate of kinetic energy loss from the Moon - Ocean - Earth system due to the tidally generated turbulence over continental shelves (Lambeck, 1977).

7.2 Shelf Wave Decay

The generation of shelf waves and their propagation along continental shelves is well documented (Cutchin and Smith, 1973; Gill and Schumann, 1974; Mysak, 1980; Brink, 1982b; Hsieh, 1982). The continental shelf west of Vancouver Island is no exception, with the $K_1$ period shelf wave often dominating the diurnal currents. Crawford and Thomson (1984) compare the observed shelf waves along Vancouver Island to the theoretical model of Brink (1982a) and find that the diurnal-period currents are almost entirely associated with first-mode shelf wave oscillations. The analysis of Crawford (1984) shows that these waves are generated near the mouth of Juan de Fuca Strait and that they propagate along the shelf to the northwest.

A detailed analysis of shelf wave dynamics will not be presented, and I feel it will be sufficient to note that the observed horizontal current fluctuations are found to be well explained by wave solutions to the unforced baroclinic wave equations (Leblond and Mysak, 1978). These current fluctuations are at known frequencies, (the largest are $K_1(1.0027 \text{ cyc/day})$ and $O_1(0.9295 \text{ cyc/day})$) and can be identified in current records.
using harmonic analysis (Foreman, 1979). The current variations associated with each constituent are usually represented in the current ellipse notation. The four parameters required to trace out the ellipse of a tidal current are the major and minor axes (Ma, Mi), the inclination (θ) of the northern major axes from due east and the phase lag (g) of the maximum current behind the maximum tidal potential (Foreman, 1979). Results of the harmonic analysis of the current data collected as part of this study are summarized in Appendix C.

The current meter moorings C1-C3 and the microstructure stations M1, M2, M3, M5, M6 and B2 are located near the Carmanah line of current observations analyzed by Crawford (1984). His analysis shows that the average energy flux of the $K_1$ shelf wave between the shore and the 200 m depth contour is approximately 68 MegaWatts ($10^6$ Watts). He also finds that the waves decay by about 90% in 300 km, ten times less than the frictional decay scale modeled by Brink (1982a). From the measurements presented here it is possible to estimate the decay length scale for the $K_1$ shelf wave.

As indicated in §6, the tidal currents are responsible for approximately 40% of the observed speeds above the bottom boundary layer ($z > 40$ m) and approximately 80% within the bottom boundary layer. Since the tidal constituents are uniform with depth (Crawford and Thomson, 1984) one may expect the tidal current shear $\frac{\partial U_T}{\partial z}$ to be concentrated near the bottom. The mean (geostrophic) currents are baroclinic with most of the mean shear $\frac{\partial U_g}{\partial z}$ occurring above the bbl between depths of 40 and 100 m (Figure 34).

From Table 5 an average dissipation rate within the bottom boundary layer is found to be $\bar{\epsilon}_{bbl} \simeq \bar{\epsilon}_T \simeq 4.2 \times 10^{-5}$ W m$^{-3}$, with an average bottom boundary height of 37 m. Assuming that all the dissipation is from the $K_1$ shelf wave will give a lower limit to the
decay length scale.

The cross-sectional area of the bottom boundary layer is approximately $37 \times 40,000$ m² and the shelf wave energy loss per metre of coastline in then $\epsilon_{SW} \simeq 62$ W. The length scale associated with frictional decay is then

$$L_F \simeq \frac{6.8 \times 10^7}{62} \simeq 1100 \text{ km}$$

(7.2.1)

This estimate is a lower limit, as I have assumed that all the dissipation within the bbl is energy loss from the single $K_1$ shelf wave component. The observed decay length scale is smaller ($\sim 300$ km) which is likely a result of dispersion (Crawford, 1984).

Brink (1982a) derives an expression for the decay length scale due to bottom friction (turbulence) using a bulk bottom stress estimate ($u_* \simeq 0.03 U_{rms}$), a drag coefficient of $C_D = 1.5 \times 10^{-3}$, and assumes a boundary layer thickness given by Weatherly and Martin (1978) of $h_{MB} = 1.3 u_* f^{-1} \left[1 + \left(N_0^2 / f^2 \right) \right]^{-1/4}$. For $U_{rms} = 10$ cm s⁻¹ and $f = 3.39 \times 10^{-5}$ s⁻¹ Brink (1982a) finds a frictional layer 10m deep and an associated decay scale for the shelf waves of $L_F \simeq 5300$ km. The difference between (7.2.1) and this value could be due to the smaller boundary layer height and larger drag coefficient used by Brink (1982a), or more likely an over estimate of the shelf wave dissipation rate per metre $\epsilon_F \simeq 62$ W. Again, a more accurate estimate of $\epsilon_F$ is not possible due to the non-linear dependence of the dissipation rate on the absolute current magnitude.

Webster (1985) considers a linear bottom-friction term for the shallow water wave equations of the form $\frac{rU}{h}$, where $r$ is related to the drag coefficient, $r = C_D U_{rms}$. In his model he uses a value $r = 0.1$ cm s⁻¹ and finds a decay length scale of approximately 1000 km for the first and second mode shelf waves. Mitchum and Sturges (1982) find
values of $r$ ranging between 0.01 and 0.02 from current measurements on the shelf west of Florida. A mid-shelf value of $r \approx 0.012$ is obtained from the estimate of $C_D$ made in §5.2 and the value $U_{rms} \approx 14.4 \text{ cm s}^{-1}$ from the current data at C2. This is in good agreement with Mitchum and Sturges (1982), but much lower than the value used by Webster (1985) who obtains similar decay length scales.

These results indicate that the drag coefficients typically used in models may be overestimates but the depth of the energy dissipation layer is underestimated. Another possibility is that a large portion of the wave energy is extracted in the shallower, near shore regions of the shelf. Unfortunately this region was under-sampled by this survey.
7.3 Secular Deceleration due to Tidal Friction

From the dissipation rate measurements on the continental shelf, an estimate of the local contribution to the secular acceleration of the Moon - Ocean - Earth (MOE) system can be made. Such an estimate is historically important as it fills a void of direct measurements in the analysis of the planetary tidal system. This result may therefore be interesting to oceanographers, astronomers and geophysicists. A very brief introduction of the subject is necessary, but for a more complete derivation and history see Lambeck (1977), Defant (1961), Jeffreys (1976) and Schwiderski (1985).

The frictional effects of the ocean tides play a number of important roles in the celestial MOE system. Of key importance is that the form of the frictional terms in the dynamical equations is unknown, as is the distribution of the friction and mechanical energy loss. Friction, both in the form of form drag and skin friction causes the gravitational bulge in the oceans to lag behind the meridian of the forces. In other words, there is a phase lag between the time of high water and the passing of the Moon overhead. The net result is that the gravitational attraction between the ocean and the Moon produces a torque. This torque, which in the case of the Moon is nearly 5 times larger than that due to the Sun, simultaneously acts to decelerate the revolution of the Moon and decelerate the rotation of the Earth. This simplified argument does not take into account the complicated shapes of the oceans, but does indicate the nature of the secular acceleration. Broshe and Sundermann (1971) show that not all the torques are in the same direction, and that some tidal currents will actually accelerate the Earth's rotation and accelerate the Moon's revolution.

It has recently been found (Schwiderski, 1985) that the form drag (pressure) dissipation...
pations approximately \( \sim 47\% \) of the work done on the ocean and bottom friction (turbulence) dissipates the remaining \( \sim 53\% \). Turbulence within the interior of the ocean is believed to be negligible. The form drag is further found to account for almost all of the secular decelerations, while the bottom friction transforms the kinetic energy entirely into heat (Schwiderski, 1985). Estimates of the rate of energy dissipation due to bottom friction come from astronomical calculations, satellite observations and tidal models (Schwiderski, 1985), and are consistently of the order

\[
-\frac{dE}{dt} \simeq 1.9 \times 10^{12} \text{W}.
\]

The distribution of this dissipation is still a mystery.

From the measurements on the continental shelf west of Vancouver Island, a very rough estimate for the tidal dissipation for similar shelf regions can be made. The average dissipation within the bottom 37 m attributed to the tides is \( \bar{\epsilon}_{BL} \simeq 4.3 \times 10^{-5} \) W m\(^{-3}\), and per unit area of shelf is \( \bar{\epsilon}_T \simeq 1.59 \times 10^{-3} \) Wm\(^{-2}\). These measurements were made in depths ranging between 60 and 200 m. Neglecting regions shallower than 60 m likely biases this estimate. It is therefore most likely an underestimate. The total area of all continental shelf regions is approximated by Lambeck (1977) at \( 3.0 \times 10^{13} \) m\(^2\), which results in a total dissipation of

\[
-\frac{dE}{dt} \simeq 4.8 \times 10^{10} \text{W},
\]

(7.3.1)

and implies that continental shelves (if similar to the shelf west of Vancouver Island) contributes insignificantly to the total dissipation. This dissipation is only 2.5% of that required.

Previous estimates for the tidal dissipation in shelf regions by Taylor (1919), Jeffreys (1976) and Lambeck (1977) have been made using a bulk relation for determining energy
dissipation rates per unit area

$$\bar{\epsilon}_T \simeq \alpha \rho \overline{|U|^3},$$  \hspace{1cm} (7.3.2)

where $\alpha$ is a drag coefficient (the value 0.002 is most often used), $\rho$ is the density of the fluid and $\overline{U}$ is the mean tidal velocity. From the current meter moorings C1-C3 an rms tidal current amplitude of $\overline{U} = 14.4$ cm s$^{-1}$ was found in the bottom 30 m. If we use the measured $\bar{\epsilon}_T$, $\bar{\rho} \simeq 1026.0$ kg m$^{-3}$, and $\overline{U}$ in (7.3.2) a value for $\alpha$ is found to be 0.0005, four times less than the value used to estimate the tidal dissipation by Taylor (1919), Jeffreys (1976) and Lambeck (1977). This follows directly from the estimated drag coefficient reported in §5.2, which is lower than the values previously adopted.

Equation (7.3.2) has also been used (Lambeck, 1977) to estimate the dissipation from the interior regions of the ocean. From the Laplace tidal equations an average tidal current of 1 cm s$^{-1}$ is found and with $\alpha = 0.002$, (7.3.2) predicts a mid-ocean contribution of $-\frac{dE}{dt} \simeq 10^8$ W. An extremely small rate of energy loss. This has led to the assumption (Lambeck, 1977) that boundary effects and the shallow shelf regions must account for a majority of the dissipation. The result (7.3.1) implies that shelf regions similar to that west of Vancouver Island may only contribute a small fraction (2.5%) of the total.

Two likely sources exist for the unaccountable energy dissipation. As was done in the past, we may draw on the non-linear nature of the turbulent energy sink and push the tidal dissipation into regions even shallower than the continental shelf. The data presented shows that the amount of dissipation increases as the water depth decreases. Extrapolating the dissipation rate data into shallower regions (total depths < 30 m) accounts for only another $\sim 4.1\%$. The sparsity of data in such depths does not make this extrapolation conclusive.
8. Upwelling and the Tully Eddy

8.1 Introduction

From the measurements of the turbulent dissipation rate it is possible to estimate the rate of kinetic energy loss from the Tully eddy and the flux rate of nutrients over the shallow, near-shore banks. The distributions of $\epsilon$ and $\overline{K_\rho}$ outlined in §6 and in Table 5 indicate that the dissipation rates observed in the bottom boundary layer are predominantly due to tidal forcing and those above the bottom boundary layer may be attributed to the mean flow. This, once again, is not entirely true, but appears to be a satisfactory partition considering the non-linear relation between the superimposed current fields and the turbulence. Mixing and energy dissipation associated with the eddy will then be restricted to depths above the bottom boundary layer. Justification for this is suggested by the $\sigma_z$ contours (Figures 35 to 41) that show the eddy structure is confined to the upper 100 m of the water column.

8.2 Nutrient Flux Rate

Although no direct measurements of the nutrient concentrations were made during the cruise in June, 1985, estimates of the nutrient flux rates can be made from the vertical eddy coefficient estimates and the measured oxygen concentration distribution. This follows from the tight relation between the dissolved oxygen content and the concentration of nutrients for water below the euphotic zone, approximately 50 m deep (Denman, 1987, personal communications). Above the depth of 50 m, oxygen and nutrient concentrations are no longer related. In the following discussion, nitrate ($NO_3^-$) will be used as a nutrient indicator, although phosphate and silicate are also important and will have similar flux distributions.
Previous studies of the distributions of nutrients, temperature, salinity and dissolved oxygen in the area west of Vancouver Island include that of Mackas, Louttit and Austin (1980); Denman, Mackas, Freeland, Austin and Hill (1981); Denman, Forbes, Mackas, Hill and Sefton (1985) and Mackas, Denman and Bennett (1987). Well documented in these studies is the observation of high nutrient concentrations in the near shore region, over Swiftsure Bank.

Mackas et al. (1980) estimate a seasonal upward flux of $NO_3$ over the Swiftsure Bank region of $\sim 4.4 \times 10^{-4}$ mmol m$^{-2}$ s$^{-1}$ (or 7 mmol s$^{-1}$ per metre of coastline). This estimate is obtained from an average upwelling velocity of $10^{-5}$ m s$^{-1}$ and a near-bottom $NO_3$ concentration of 35 mmol m$^{-3}$. They note that this estimate may be 1 to 2 orders of magnitude less than the flux rate during peak upwelling events. Similarly, Denman et al. (1981) estimate an average upward flux of 9.8 mmol s$^{-1}$ per meter of coastline for the period between 20 May and 5 July, 1979, a period when the upwelling is substantial.

An estimate of the upward flux rate of nutrients due to turbulent mixing can be made from the vertical eddy coefficient ($\overline{K}_p$) estimates and dissolved oxygen ($O_2$) measurements made at B2. The close relation between dissolved oxygen and $NO_3$ for depths greater than 50 m permits a rough calculation of the vertical $NO_3$ gradient near the bottom. From the $O_2$ survey conducted in June 1985 the dissolved oxygen gradient in the bottom 30 m is approximately $\frac{\Delta O_2}{\Delta z} \simeq -4.4 \times 10^{-2}$ ml l$^{-1}$ per metre. This is then converted into a vertical gradient for $NO_3$ from the tabulated data of Denman et al. (1985), where the linear fit between $O_2$ and $NO_3$ gives, $\frac{\Delta NO_3}{\Delta O_2} \simeq -5.0$. The vertical gradient in $NO_3$ is then found using

$$\frac{\Delta NO_3}{\Delta z} \simeq \frac{\Delta O_2}{\Delta z} \times \frac{\Delta NO_3}{\Delta O_2},$$

(8.2.1)
assumes linear relations between the $O_2$ and $NO_3$ concentrations, an assumption that can be valid only beneath the euphotic zone. A linear fit of the $O_2$ and $NO_3$ concentrations provided by Denman et al. (1985), gives a correlation coefficient of $r^2 \simeq 0.959$ is found, suggesting that the relation is linear. The vertical gradient of $NO_3$ is found using (8.2.1) to be $\frac{\Delta NO_3}{\Delta z} \simeq 0.22$ mmol m$^{-4}$ (±28%). The uncertainty represents an 11% uncertainty in the estimate of $\frac{\Delta NO_3}{\Delta z}$ and a 17% in $\frac{\Delta O_2}{\Delta z}$ (1 standard deviation).

From Table 5, the vertical eddy coefficient in the bbl at B2 is found to be $K_p \simeq 1.1 \times 10^{-1}$ m$^2$ s$^{-1}$ (±43%) and the $NO_3$ flux rate is then

$$K_p \frac{\Delta NO_3}{\Delta z} \simeq 2.42 \times 10^{-2} \text{mmol m}^{-2} \text{s}^{-1},$$

(8.2.2)

with an uncertainty of ±71%. This is almost 2 orders of magnitude larger than the seasonal average estimated by Mackas et al. (1980) and the 7 week average estimated by Denman et al. (1981). Even if a lower limit is chosen, the flux is still 11 times that found previously. Using the shelf width adopted by Denman et al. (1981) of 65 km, a net vertical flux of $NO_3$ of $2.52 \times 10^7$ mmol s$^{-1}$ is obtained.

A similar calculation at M5, near the centre of the Tully eddy, indicates that the vertical flux of nutrients near the bottom is of the order $\sim 1.6 \times 10^{-3}$ mmol m$^{-2}$ s$^{-1}$. At M6 the flux rate is $\sim 1.1 \times 10^{-3}$ mmol m$^{-2}$ s$^{-1}$. These are much less than the estimate for B2, but 4 times the seasonal average over the shallow banks. During strong upwelling conditions the nutrient flux rates over most of the shelf region are higher than the near-shore seasonal average. The Tully eddy, a dominant feature of the circulation, does not seem to be a significant region of vertical mixing.

The estimated $NO_3$ flux rate over the Swiftsure Bank, although much higher than the previous estimates, may be of the correct magnitude for such an active upwelling period
(Denman, 1987, personal communication). If the seasonal flux is indeed $4.4 \times 10^{-4}$ mmol m$^{-2}$ s$^{-1}$, the variance in the signal must be large, with extensive periods of virtually no flux. The importance of this observation is that the replenishing of nutrients may occur during short periods of high flux, and that the stability of the dependent ecosystems could hinge critically on these periods.

8.3 Eddy Decay

From the microstructure observations in the vicinity of the Tully eddy (M2, M3, M5 and M6), an estimate of the frictional decay time scale can be made from the average dissipation rates above the bottom boundary layer. An estimate of the kinetic energy in the eddy is obtained from the dynamic height contours in Figure 34. The spacing of the contours represents the mean current speed at 40 m relative to 100 m. This is calculated to be $|\bar{U}| \sim 12 (\pm 0.9)$ cm s$^{-1}$, in a region bounded by a 15 km radius from the centre of the eddy. From the current data of Freeland and McIntosh (1987, personal communications) and the drifter data of Crawford (1986, personal communications) it appears that the mean currents at a depth of 40 m are similar to those near the surface. Integrating over the upper 100 m, an average mean velocity is estimated to be 8.4 cm s$^{-1}$. The average kinetic energy per m$^{-3}$ is roughly $\sim 4.8$ J m$^{-3}$, and the total kinetic energy of the eddy is approximately $\sim 3.4 \times 10^{11}$ J.

The average rate of dissipation of turbulent kinetic energy above the bbl in the region of the eddy (from Table 5) is $\bar{\epsilon}_E \simeq 5.76 \times 10^{-6}$ W m$^{-3}(\pm 41\%)$ where the uncertainty is the standard deviation in all the $\epsilon$ values. The estimate of $\bar{\epsilon}_E$ is obtained from the average of all $\epsilon$ values above the bottom boundary layer at stations M2, M3, M5 and M6.
The frictional decay time scale for the eddy is then

\[ T_F \simeq \frac{4.8}{5.76 \times 10^{-6}} \text{ s} \]

\[ \simeq 231 \text{ hours}, \]  

or approximately $9\frac{1}{2}$ days ($\pm 41\%$). The frictional decay time scale represents an estimate of the time required for internal friction due to turbulent stresses to dissipate the kinetic energy of the mean flow. This estimate is a lower limit, as I have assumed that all the dissipation above the bbl is dissipation from the geostrophic flow and that there is no potential energy in the eddy (this alone could double the decay time).

From the current data of Freeland and McIntosh (1987, personal communication) an estimate of the average observed decay rate can be made. McIntosh (1987, personal communications) has calculated the circulation \( C = \oint \vec{U} \cdot d\vec{l} \) (Gill, 1982) at 12 hour intervals over a 105 day period (January–June, 1985) (with the tides removed). From the time series of the circulation \( C \), an estimate of the “e” folding decay rate can be made. The average “e” folding decay rate is then found using

\[ \overline{\Delta t_e} \simeq \frac{C(1/e - 1)}{\partial C/\partial t}, \]  

(8.3.2)

where \( \partial C/\partial t \) is calculated from the first difference, “e” = 2.7183 and only contributions when \( \partial C/\partial t < 0 \) (decay) are considered. The average is found to be \( \overline{\Delta t_e} \simeq 95 \pm 119 \text{ hours} \) (± standard deviation). The large standard deviation is due to a few extremely long decay times when the slope is near zero. Eliminating data values outside 2 standard deviations about the mean \( (\Delta t_e > 334 \text{ hours}) \) gives a new mean of

\[ \overline{\Delta t_e} \simeq 68 \text{ hours}. \]

(8.3.3)

A total of 106 12 hour observations contributed to this mean, representing exactly half the data.
These different decay scales indicate some interesting features. The fact that frictional effects predict a decay time over 3 times longer than the observed “e” folding decay time, suggests that friction, in the form of turbulence or an eddy viscosity, is not an important factor in the eddy dynamics. The observed decay rate from the current data further implies that eddy decay is dominated by external forcing, through torques opposing the cyclonic circulation. These torques are most likely form drag (pressure) and surface stresses (wind). Also, the eddy is in a state of decay exactly 50% of the time.
9. Conclusions

9.1 The Bottom Boundary Layer

From the diurnal time series of turbulent microstructure profiles on the continental shelf west of Vancouver Island some important features of the turbulent bottom boundary layer and mid-depth mixing can be identified. The free-falling profiler FLY II was fitted for near-bottom sampling of turbulent shear \( \frac{\partial u}{\partial z} \), temperature and conductivity. The measurements presented here indicate that the instrument can measure accurate low-level oceanic dissipation rates in spite of the increased vibrations caused by the probe guard. Part of this is due to the attention paid to the spectral analysis of the shear probe time series and the minimization of noise within the turbulence bandwidth.

Possibly the most interesting feature revealed by the microstructure profiles is the distinction between the time-dependent turbulent bottom boundary layer (tbbl) and the bottom well mixed layer (wml). The two are rarely the same height and occurrences of either a deeper turbulent layer than well mixed layer, or a deeper well mixed layer than turbulent layer were observed. The reasons proposed for these different conditions stem from the dependency of the particular structures on the local dynamics. The intensity of the near-bottom turbulence reacts quickly to the magnitude of the near bottom current. The turbulence within the constant stress layer and the estimated friction velocities are shown to correlate with the magnitude of the bottom current. The height and interior structure of the tbbl develop more slowly with the extent of this development depending most strongly on the duration of the diurnal current. Turbulence above the well-mixed layer is modified by the effects of stratification, but is not eliminated when current shear exists above the wml and turbulence is locally generated in this region.
A well-mixed layer was observed on almost every microstructure profile made on the shelf in June 1985. The height of the layer was observed to vary considerably (10-40 m). This variability rarely correlated with the local mixing rates, and is believed to be due almost entirely to advection. The thickness of the well-mixed layer decreased slowly when its height exceeded the local turbulent boundary layer height. Conversely, the well mixed layer height did not always grow when the turbulent boundary layer height exceeded that of the well mixed layer. In the later case, the erosion of the statification above the well mixed layer produced narrow mixed layers.

In addition to the turbulent and density structures observed with the microstructure profiles, the near-bottom currents were measured with fixed moorings. Current veering was concentrated within the pycnocline at the top of the well-mixed layer. Little or no current veering was observed within the well mixed layer, even when the well-mixed layer height exceeded 30 m. These observations agree with the model of Weatherly and Martin (1978), but no provisions are made in their model for turbulence production above the pycnocline at the top of the well-mixed layer.

In agreement with the Weatherly and Martin (1978) model are the observations that advection within the bottom boundary layer dominates the density and well-mixed layer structures. The advection of “turbulent fronts” was also observed. Turbulence levels during such events can account for a significant portion of the total dissipation over a tidal cycle.

The turbulent kinetic energy equation has two forms applicable in two regions of the turbulent bottom boundary layer. Within the well-mixed layer the production of turbulence through turbulent stresses working on the mean shear is balanced by viscous
dissipation (4.2.4). This can be simplified to (4.3.7) in the constant stress layer by assuming a linear eddy viscosity. Outside the well-mixed layer stratification becomes important in two ways. First, the vertical turbulent velocity \( w' \) is suppressed by working against gravity, and second, internal waves can propagate, allowing a radiation sink for kinetic energy. Current veering is concentrated within the pycnocline at the top of the well-mixed layer. The resultant turbulent kinetic energy equation for the stratified bottom boundary layer is (4.2.10).

A shelf wave decay scale due to turbulent dissipation is estimated to be 3200 km. This is in reasonable agreement with the model of Brink (1982a). The dissipation of tidal kinetic energy over all shelf regions is estimated to be \( 4.8 \times 10^{10} \) W, only 2.5% of the dissipation required to account for a balanced energy budget. The turbulent dissipation in very shallow regions may account for the missing energy loss, but more measurements are required to determine the exact distribution.

9.2 The Continental Shelf

The dynamics of the shelf regions are subject to three main driving mechanisms; the wind, tides, and estuarine outflow. The current structure that develops is a combination of quasi-geostrophic and tidal velocities. Near the surface (< 50m) the non-tidal component is dominant (60%), and nearer the bottom the tides dominate (80%). The contribution of internal waves was not recoverable from this data. The onset of upwelling conditions by the wind-driven flow in June brings slope water onto the shelf. This cold water has low dissolved oxygen and high nutrient content. The mixing and redistribution of the slope water and the fresh water outflow are governed by both the geostrophic and tidal motion.
The distribution of turbulent mixing has been revealed by microstructure time series at seven locations over the shelf. Three distinct features can be identified by these time series. First, the turbulent bottom boundary layer is tidally driven at the diurnal frequency and varies between 5 and 50 m in depth. In the shallower locations (depths < 60 m) the turbulent boundary layer can occupy the entire water column. Second, mid-depth mixing increases towards the shore, eastward of the shelf break and is believed to be generated by regions of current shear and by the dissipation of locally generated internal waves. Finally, the region of most importance to the upwelling of nutrient-rich slope water is identified as the shallower banks near shore. Vertical mixing over the banks occurs at all depths during peak tidal currents and enables the nutrients to penetrate into the euphotic upper layers.

An upward nutrient flux rate due to turbulent mixing in the bbl of $2.42 \times 10^{-2}$ mmol m$^{-2}$ s$^{-1}$ is estimated for the highly turbulent water near shore, two orders of magnitude larger than the seasonal average. A decay length scale of the $K_1$ period shelf wave is estimated at 1100 km. Finally, an estimated lower limit of the friction decay time scale for the Tully eddy is found to be $9\frac{1}{2}$ days. The observed circulation fluctuates much more rapidly, indicating that friction (turbulent dissipation) is not important in the eddy dynamics.

In summary, it appears that the role of turbulent dissipation is not directly important to the dynamics of local large scale circulation. The ability of turbulence to mix quantities is important, but relies on the large scale circulation to bring the nutrients into the shallow banks where the turbulent bottom boundary layer bridges between the benthic and euphotic regions.
10. References


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Appendix A: Error Analysis for Shear Probe $\epsilon$ and $u_*$ Calculations

The accuracy in a calculation of $\epsilon$ within the bottom boundary layer depends on six factors. Five of these (1-5) are discussed by Soulsby (1980) for $u'w'$ measurements and the sixth addresses the assumption of isotropy. These factors are:

1) The loss of low-frequency contributions due to the length of time series used to calculate the shear spectrum: If the section is too short then low-frequency information will be lost. Near the ocean bottom this loss becomes a trade-off with depth resolution and stationarity.

2) Loss of high-frequency contributions due to the digitization rate: If the digitization rate is too slow, high-frequency contributions will not be resolved. The digitization rate is usually limited by the data acquisition system, the sensor and the amount of computer power available during analysis.

3) Stationarity and homogeneity: The spectral methods used to calculate the power spectrum $k_3^2 \phi_{11}(k_3)$ assume that the section of time series is stationary, in that the variance does not change from the start of the record to the end (both in time and space). Sufficiently short records may be assumed to be stationary, and in regions where the turbulent kinetic energy does not vary significantly with depth, longer records may be used. Turbulent flows are locally homogeneous when the divergence terms in the turbulent kinetic energy equation are negligible compared to the source (Reynolds stress-shear production) and sink (viscous dissipation) terms (Hinze 1975). These conditions will be satisfied when the mean turbulent kinetic energy quantities are considered at fixed heights, over periods that are long compared to the turbulence decay time scale ($0.1$ to $0.2 \times 2\pi/N$ Crawford, 1986) and short compared to variations in the mean flow ($\sim 40$
minutes, Grant et al. 1984) due to internal waves.

4) Sampling variability: The random error associated with making single estimates in a stochastic field. This may be reduced by increasing the number of estimates in an ensemble average, but with too many samples the series tends to be too long in time to satisfy stationarity.

5) Sensor response: The sensor used to measure the turbulent fluctuations will likely have spatial and temporal response limitations. The small-scale spatial response is usually determined by the size of the sensor and the range of fluctuations that it can physically resolve. The high-frequency temporal response is a measure of how quickly the instrument can respond to a change in the fluctuating field. The response transfer function which corrects for some of these shortcomings is available for some sensors (Soulsby 1980; Ninnis 1984).

6) The isotropy of the flow: It is very often assumed that the turbulence being measured is isotropic. Theories leading to a spectral shape, as in (4.3.10) and (4.3.12), and theories which relate different spectra, as in (4.3.14), all assume isotropy. The validity of this assumption will depend on the intensity of the turbulence, the strength of stratification and the scales at which the turbulence is measured. In the bottom boundary layer there is also a height restriction which, to first order (Pond 1965), states that fluctuations are isotropic for wavelengths smaller than the height above the bottom at which the measurements are made (i.e.: up to wavenumbers $k \simeq \frac{2\pi}{z}$).

I will now address these factors with respect to the analysis of shear probe measurements. First, I shall review the methods used to calculated the dissipation rate per unit volume $\varepsilon$ from shear probe measurements. The differentiated voltage signal $V_0$, 175
proportional to $\frac{\partial u}{\partial z}$, is related to the shear $\frac{\partial u}{\partial z}$ by Taylor's hypothesis,

$$\frac{\partial u}{\partial z} = \frac{1}{W} \frac{\partial u}{\partial t} = \frac{V_0}{G^2 \sqrt{2SW^2}},$$

where $W$ is the fall speed of the profiler, $G$ is the differentiator gain, $S$ is the shear probe calibration (Osborn and Crawford 1980), $2\sqrt{2}$ a calibration constant and $V_0$ the differentiated output voltage. The dissipation rate $\epsilon$ is then calculated from the variance of $\frac{\partial u}{\partial z}$, which under the assumption of isotropy is given by (Monin and Yaglom 1975, p353),

$$\epsilon = 7.5\mu \left( \frac{\partial u}{\partial z} \right)^2 = 7.5\mu \int_{k_1}^{k_u} k_3^2 \phi_{11}(k_3) dk_3,$$

where $\mu$ is the dynamic viscosity and the variance $\left( \frac{\partial u}{\partial z} \right)^2$ is the area under the power spectrum of the $\frac{\partial u}{\partial z}$ time series integrated from the lower limit $k_1$ to some upper limit $k_u$.

In theory, $k_1 = 0$ and $k_u = \infty$.

We shall now consider errors associated with calculating the shear $\frac{\partial u}{\partial z}$ using (A.1). The fall speed $W$ changes very slowly with depth and is calculated at each depth from the differentiated pressure record. An estimated error of 5% in $W$ results in a 10% error in $\frac{\partial u}{\partial z}$ and a 20% error in $\epsilon$. The differentiator gain $G$ is also known to 2%. The shear probe calibration $S$ is known to approximately 7%, which reflects the standard deviation in the constants obtained from multiple calibrations over time. The cumulative error estimated in the shear $\frac{\partial u}{\partial z}$ is then 19%, which translates to an approximate uncertainty in the variance $(\frac{\partial u}{\partial z})^2$ of 38%. The dynamic viscosity $\mu$ is predominantly temperature dependent and is calculated for a mean salinity of 32.5°/o0 from the formula of Miyake and Koizumi (1948). An estimated uncertainty of 5% in $\mu$ adds to give a total uncertainty in $\epsilon$ of 43%. This is a conservative estimate for the uncertainty, in that all the errors have been added linearly. Most of these contributions are random, and a more realistic estimate
of the uncertainty is obtained by adding the square root of the individual uncertainties. This approach leads to an uncertainty in $\epsilon$ of approximately 20%, but further assumes that the full variance is recovered by the integration of $k_3^2\phi_{11}(k_3)$ from $k_l$ to $k_u$ (A.2).

In order to estimate the uncertainty in $\epsilon$ calculated from (A.2) we must consider each of the six factors mentioned above.

1) Length of Record, and 2) Digitization rate

The first consideration is to prove that the records selected are long enough to resolve the low-frequency portion of the spectrum. Figure 8 shows the cumulative area under the universal shear spectrum $G_2\left(\frac{k_3}{k_s}\right)$. The wavenumber indicated by $k_l = k_5$ is the wavenumber such that only 5% of the area under $G_2\left(\frac{k_3}{k_s}\right)$ lies below this limit. This occurs at approximately

$$k_5 \approx 0.04k_s,$$  \hspace*{1cm} (A.3)

where $k_s$ is the Kolmogoroff wavenumber. The number of points in the time series section required to recover the spectrum down to $k_5$ is then given by

$$J_5 = \frac{2\pi}{W k_5 \Delta t} = \frac{2\pi}{W(0.04)k_s \Delta t},$$  \hspace*{1cm} (A.4)

where $\Delta t$ is the digitization rate. If $J$, the number of points in the power spectrum is larger than this lower limit $J_5$, then less than 5% of the spectrum will be lost. If, for example we choose $\epsilon = 5.0 \times 10^{-4}$ W m$^{-3}$ and $\mu = 1.547 \times 10^{-6}$ kg (m s)$^{-1}$, then $k_s = 615$ m$^{-1}$ and if $\Delta t = 3.646 \times 10^{-3}$ s and $W \approx 0.625$ m s$^{-1}$ as is the case with the present profiler, then $J_5 = 111$ points. As we will see later, this is smaller than the minimum number ($J = 128$) of points used in this paper. For the values of $\epsilon$ presented here, the actual percentage lost by selecting $k_l > 0.0$ as predicted by Figure 8, has been added to the original estimate (3.3.9).
The cumulative area curve in Figure 8 has also been adapted to estimate the variance lost by integrating \((A.2)\) to the upper bound \(k_u\) to eliminate high frequency noise. For example, the 90% recovery wavenumber of \(k_{90} \approx 0.56 k_s\) is also shown in Figure 8. Using the same values outlined above, this translates to an upper critical frequency of \(f_{90} \approx 34.1\) Hz, where \(f = \frac{W_k}{2\pi}\). For each spectrum, \(k_u\) is selected as the wavenumber above which noise begins to dominate the variance. This upper bound was below the 90% wavenumber \(k_{90}\) for less than 10% of the nearly 30 thousand spectra calculated from over 300 microstructure profiles. The actual percentage lost by choosing \(k_u < \infty\) has been calculated for each \(\epsilon\) value as predicted by Figure 8, and added to the original estimate (3.3.9).

3) Stationarity and Homogeneity

The next factors which must be satisfied are the assumptions of homogeneous turbulence and the assumption of stationarity. In an inhomogeneous turbulence the mean divergence and acceleration terms not included in (4.1.4) can be non-zero (Hinze, 1975). If, on average, these terms are significant at the scales of interest, then (4.3.7) is inapplicable. The homogeneity of turbulence is dependent on the length scale of the observations. The largest scales are always inhomogeneous, while the smallest (viscous) scales are almost always locally homogeneous (Batchelor 1957). Measurements of turbulence at viscous scales are more likely to satisfy homogeneity than measurements within the inertial subrange. The production term in the turbulent kinetic energy equation (4.1.4) works at larger than viscous scales and homogeneity can only be assumed for mean, steady state conditions. If, in the neutrally stratified constant stress layer the turbulence is observed to be steady over periods longer than the local decay period (Crawford, 1986), then there
must be an eventual balance of turbulence production and dissipation.

The dissipation rate values presented here have been calculated from vertical profiles of microstructure shear $\frac{\partial u}{\partial z}$. If a vertical dissipation rate profile has a constant stress layer distribution given by (4.3.7), then the variance $(\frac{\partial u}{\partial z})^2$ will have a similar distribution and

$$
\left( \frac{\partial u}{\partial z} \right)^2 \propto z^{-1}.
$$

To minimize the errors in spectral shape introduced by this non-stationarity we have chosen short sections ($J = 128$) near the bottom, with increasing length with $z$ ($J = 256, 512$ and $1024$). This technique also provides greater depth resolution in $\epsilon$ and $u^*$ near the bottom at the expense of reduced confidence in individual values of $\epsilon$.

For time series analysis, stationarity refers to variations in time. Homogeneity refers to variations in space. By employing Taylor’s frozen turbulence hypothesis (A.1), a “time series” becomes a “spatial series”, and considerations of stationarity become considerations of homogeneity. Stationarity can still refer to variations in a quantity in time at a fixed height above the bottom.

Soulsby (1980) considers non-stationarity due to the changes in the mean flow. The concern is that the physical processes driving the turbulence within the boundary layer must be constant over the observation period. Non-stationary turbulent flows may be expected during rapid acceleration or deceleration phases of a tide. Internal waves may also cause the turbulence to be non-stationary at time scales of 30 to 60 minutes (Grant et al. 1984). Individual microstructure shear profiles are not subject to this non-stationarity. Dissipation rate values at fixed heights $z$ above the bottom from consecutive profiles have been averaged under the assumption of stationarity. Consecutive profiles obtained over periods longer than 10 minutes have only been averaged when the supporting current
data confirm that the flow is reasonably stationary. If the flow is accelerating or decelerating, time-dependent terms omitted in (4.1.4) may be important in the balance between turbulent production and dissipation.

4) Sampling Variability

Sample variability arises when single estimates of an intermittent or stochastic process are obtained. Soulsby (1983) found that from $u'w'$ measurements, 90% of the stress was accounted for during only 26% of the time. This was due to the bursting nature of the boundary layer turbulence, and the inherent intermittence of the $u'w'$ product. To resolve any bursts of turbulent kinetic energy in my data, the vector current meter at 5 metres above the bottom on the mooring at C2 (Figure 1) was set to sample at one second intervals. Strong bursting as observed by Soulsby (1983) was not detectable in our records. This could be due to the relatively flat topography in the vicinity of station C2, or it could be that kinetic energy of the large scale flow may not burst in the same way as the Reynolds stress. Vertical profiles from regions where there is active bursting, accelerating flows or uneven turbulence production will require an increased sampling rate in order to resolve the intermittency. Consecutive profiles show some intermittency, but estimates of $u_*$ from consecutive profiles vary by less than 10% (standard deviation). Averaging consecutive profiles reduces the effects of intermittancy and the standard deviation in the turbulent dissipation measurements is $\sim 15\%$.

5) Sensor Response

The small-scale spatial response of shear probes has been investigated by Nininis (1984). The probes are initially calibrated by rotating them in a water jet at various angles, or by comparing the probe response against a probe of known calibration in a tur-
bulent water tunnel. Ninnis (1984) then cross-calibrates shear probes with a laser-Doppler velocimeter to determine the response at high wavenumbers. The resulting spatial transfer function extends spatial response up to wavenumbers \( k \approx 10.36 \text{ cm}^{-1} \) (wavelengths down to \( \lambda \approx 0.606 \text{ cm} \)), much higher than what is typically required to recover \( \approx 98\% \) of the area under the shear spectrum (Figure 8) and well into the background spectral noise caused by instrument vibrations (Dewey et al. 1987).

6) Isotropy

Finally we address the question of isotropy. Gargett et al. (1984) investigate the extent of isotropy for turbulence in a stratified fluid. They derive a parameter which can be used to estimate the degree of isotropy or anisotropy for oceanic turbulence. Their parameter \( I \) is defined by,

\[
I = \frac{k_s}{k_b},
\]

where

\[
k_b = \left( \frac{N^2 \rho}{\epsilon} \right)^{\frac{1}{2}},
\]

is the buoyancy wavenumber, \( k_s \) is the Kolmogoroff wavenumber and \( N \) is the buoyancy frequency defined as \( N \equiv \left( -\frac{g}{\rho_0} \frac{\partial \sigma_z}{\partial z} \right)^{\frac{1}{2}} \). To a first approximation they conclude that in cases where \( I > 500 \) the turbulence has an inertial subrange at wavenumbers below the viscous subrange. For cases where \( 50 < I < 500 \) an inertial subrange does not exist but the turbulence is isotropic at dissipation scales within the viscous subrange. In all the microstructure profiles presented in \( \xi 5 \) there is a well mixed, weakly stratified portion of water occupying at least the bottom 10 metres. The isotropy parameter \( I \) in this bottom layer never dropped below 50, and would do so only for very weak turbulence (\( \epsilon < 10^{-6} \text{ W m}^{-3} \)) near the noise limit of the present profiler (Dewey et al. 1987). The dissipation
rate values were always well above the noise level within the neutrally stratified bottom few metres, with the result that $I$ was always $> 100$ within this layer.

At interior depths, above the wml, a rough estimate of the local value for the turbulent dissipation rate can be made for conditions when stratification becomes important, and isotropy may not exist at the dissipation scales. If we use $I > 50$ as a guide, (although Gargett (1987, personal communications) indicates the value $I > 50$ is indeed a very rough test for isotropy) then A.5 yields,

$$\epsilon > 184\mu N^2,$$

for true isotropy at dissipation scales. For typical values of $\mu$ and $N^2$ at mid-depths ($\mu \approx 1.4 \times 10^{-3}$ kg/ms, and $N^2 \approx 10^{-4}$ s$^{-2}$) we find that for $\epsilon < 2.5 \times 10^{-5}$ W m$^{-3}$ the turbulence may not be isotropic at the dissipation scales. In these conditions (A.2) may overestimate the true dissipation rate by assuming isotropy. This result adds further uncertainty to the dissipation rate values in stratified regions where the reported $\epsilon$ values are below $\sim 10^{-5}$ W m$^{-3}$. Once again, this is only a guide and indicates that caution is required when discussing low dissipation rate values from stratified regions. The turbulence could still be isotropic at the dissipation scales, but stratification has become a significant restraint on the vertical fluctuations.

The other factor which limits isotropy is the distance from the boundary with respect to the scales of the fluctuations that are measured. The present system stops (lands) with the shear probes 0.15 m above the bottom. The first $\epsilon$ estimate is centered at $z \sim 0.3$ m, and unless the turbulence is very weak ($\epsilon < 10^{-5}$ W m$^{-3}$), the viscous subrange is recovered and is isotropic at all distances $z > 15$ cm. This follows by assuming that wavelengths smaller than $(2\pi/z)$ are isotropic (Pond, 1965), and that the peak in the
dissipation spectrum is at \( k_p \simeq 0.13k_z \). After substitution (\( \nu \simeq 1.36 \times 10^{-6} \text{ m}^2/\text{s}, \rho \simeq 1026.0 \text{ kg/m}^3 \)) we find the depth restriction (and this time the restriction is quite implicit) for recovering the peak in the dissipation spectrum is

\[
\epsilon > \frac{1.42 \times 10^{-8}}{z^4}, \tag{A.8}
\]

where \( \epsilon \) is in units of \( \text{W m}^{-3} \). The righthand side diminishes quickly with increasing height \( z \). At \( z \sim 0.15 \text{ m} \) the limit for isotropy occurs at a dissipation rate of approximately \( \epsilon \simeq 2.8 \times 10^{-5} \text{ W m}^{-3} \). A restriction of this magnitude on the dissipation scales negates any attempt to measure an inertial subrange at similar distances from a boundary.

The turbulent, well mixed bottom boundary layer measured by the FLY II shear probes is therefore isotropic at dissipation scales. The above factors are satisfied by shear probe turbulence measurements within most of the bottom boundary layer and reasonable confidence can be placed in the calculation of \( \epsilon \) using \( k_3^2 \phi_{11}(k_3) \) spectra and (3.3.9). The estimated uncertainty in individual dissipation rate values depends on the intensity of the turbulence and the selection of the integration limits \( k_l \) and \( k_u \), both of which are discussed in §3.3 and §3.4 with respect to actual measured spectra.
Appendix B: Calculation of Temperature, Conductivity, Salinity, Depth, Density and $\sigma_t$

The calibrations of the slower response thermistor and the conductivity cell were conducted in the Ocean Physics calibration tank at the Institute of Ocean Sciences in October, 1985. The procedure consisted of slowly increasing the temperature of the tank water, of a known salinity, and recording both the calibrated water temperature, and the raw voltage outputs from the thermistor and conductivity circuits. The lower half of the profiler FLY II was completely immersed in the tank so that any temperature effects on the circuitry itself were accounted for. The temperature was varied from 1.8 °C to 16.0 °C. The salinity of the water only varied between 33.096 °/oo to 33.187 °/oo. The actual conductivity, as recorded by a calibrated meter, varied between 29.18 to 41.86 mmho/cm. A fast response thermistor was also calibrated, with the result only 1.2% different from the 1983 calibration conducted by Sy-Tech Research.

The calibration data was plotted and all probes were found to have a very linear response. The data were then fit by a least squares technique to quadratic functions. The pressure calibrations from 1983 were also analyzed to provide the depth as a quadratic function of voltage as well. The digital data recorded range between -2048 to 2048 which correspond to voltages ranging between -5 and +5 volts. The conversion to volts is accomplished by multiplying the recorded digital value by $B = 10.0/4096.0$. The ranges shown are the result of pre-selected offsets in the circuitries and are not the ranges of the probes. The depth, slow response temperature and conductivity are then given by;
Depth

\[ Z_i = A_z V_z^2 + B_z V_z + C_z \quad [\text{m}] \quad (B.1) \]

Range: 1.0 - 214.0 m

\[ A_z = 2.69177 \]
\[ B_z = 36.832 \]
\[ C_z = -6.541 \]

Slow Temperature

\[ T_i = A_t V_t^2 + B_t V_t + C_t \quad [\text{°C}] \quad (B.2) \]

Range: 1.5 - 13.0 °C

\[ A_t = 0.002115 \]
\[ B_t = -1.13495 \]
\[ C_t = 7.061956 \]

Conductivity

\[ C_{STP} = A_c V_c^2 + B_c V_c + C_c \quad [\text{mmho/cm}] \quad (B.3) \]

Range: 26.1 - 65.7 mmho/cm

\[ A_c = 0.104545 \]
\[ B_c = 3.959 \]
\[ C_c = 43.2973 \]

To calculate the salinity, the formulae given by Millero et al. (1980) are used. First the conductivity ratio, \( R_T \) is found using,

\[ R_T = \frac{C_{STP}}{C_{35,15,0} \ t_T \ R_p}, \quad (B.4) \]
where
\[ C_{35,15,0} = 42.92, \]
\[ r_T = E_0 + E_1 T + E_2 T^2 + E_3 T^3 + E_4 T^4, \]  
(B.5)

and
\[ E_0 = 6.766097 \times 10^{-1} \]
\[ E_1 = 2.00564 \times 10^{-2} \]
\[ E_2 = 1.104259 \times 10^{-4} \]
\[ E_3 = -6.9698 \times 10^{-7} \]
\[ E_4 = 1.00131 \times 10^{-9} \]

and the pressure correction term \( R_p \) is equal to 1.0 ± 0.2%, for depths less than 400m.

The salinity is then given by,
\[ S = \sum_{j=0}^{5} a_j R_T^{j/2} + \frac{(T - 15.0)}{1.0 + k_1(T - 15.0)} \sum_{j=0}^{5} b_j R_T^{j/2}, \quad [^{o}/^{oo}], \]
(B.6)

where
\[ k_1 = 0.0162 \]
\[ a_0 = 0.008, \quad a_1 = -0.1692, \quad a_2 = 25.3851 \]
\[ a_3 = 14.0941, \quad a_4 = -7.0261, \quad a_5 = 2.7081 \]
\[ b_0 = 0.0005, \quad b_1 = -0.0056, \quad b_2 = -0.0066 \]
\[ b_3 = -0.0375, \quad b_4 = 0.0636, \quad b_5 = -0.0144 \]

The density (in gm/cm\(^3\)) is calculated by,
\[ \rho = \sum_{j=0}^{5} d_j T^j + S \sum_{j=0}^{4} e_j T^j + S^{3/2} \sum_{j=0}^{3} f_j T^j + S^2 \sum_{j=0}^{3} g_j T^j, \quad [\text{gm/cm}^3] \]  
(B.7)

where
\[ d_0 = 0.9998416, \quad d_1 = 6.793952 \times 10^{-5}, \quad d_2 = -9.09529 \times 10^{-6} \]
\[ d_3 = 1.001685 \times 10^{-7}, \quad d_4 = -1.120083 \times 10^{-9}, \quad d_5 = 6.536332 \times 10^{-12} \]
\[ e_0 = 8.25917 \times 10^{-4}, \quad e_1 = -4.449 \times 10^{-6}, \quad e_2 = 1.0485 \times 10^{-7} \]
\[ \epsilon_3 = -1.258 \times 10^{-9}, \quad \epsilon_4 = 3.315 \times 10^{-12} \]

\[ f_0 = -6.33761 \times 10^{-6}, \quad f_1 = 2.8441 \times 10^{-7} \]

\[ f_2 = -1.6871 \times 10^{-8}, \quad f_3 = 2.83258 \times 10^{-11} \]

\[ g_0 = 5.4705 \times 10^{-7}, \quad g_1 = -1.97975 \times 10^{-8} \]

\[ g_2 = 1.6641 \times 10^{-9}, \quad g_3 = -3.1203 \times 10^{-11} \]

The quantity \( \sigma_t \) is calculated by,

\[ \sigma_t = (\rho - 1.0) \times 1000.0 \]

\( (B.8) \)

Although \( \sigma_t \) is not the actual density, differences in \( \sigma_t \) denote actual density differences in units of \( \text{kg/m}^3 \). Therefore the buoyancy frequency \( N \) is correctly calculated using

\[ N^2 = -\frac{g}{\rho} \frac{\partial \sigma_t}{\partial z}. \]

\( (B.9) \)

The local dynamic viscosity \( \mu \) in \( \text{kg (ms)}^{-1} \) is calculated from the local average temperature \( \overline{T} \) over the depth range of the dissipation rate estimate using (Miyake and Koizumi, 1948),

\[ \mu = (1.87 \times 10^{-3}) \left( 1.0 + 0.0323T + 0.00023T^2 \right)^{-1}. \]

\( (B.10) \)

where 0.0323 and 0.00023 are appropriate constants for salinities between 30.5 and 34.5 \( \% \).
Appendix C: Tidal Constituents from Current Survey

Results of the harmonic analysis using the programs outlined by Foreman (1979) are presented as they were output from the I.O.S. Tides and Currents Section's HP-1000 mini-computer. The programs were run by K. Lee and A. Douglas. The raw current data was filtered and sub-sampled to a data rate of one sample per hour. This is a one-hour filter, passed over the data 3 times to eliminate high frequency noise. The Foreman (1979) program compensates for the effect of this filter on the amplitudes of the tidal constituents. The mooring identifications are noted differently than the C1–C4 notation indicated on Figure 1. The notation is:

\[
\begin{align*}
SB1 &= C1 \\
SB2 &= C2 \\
SCI &= C3 \\
JFR &= C4
\end{align*}
\]

Some of the constituents (Q1, P1 and N2) have been inferred from recoverable constituents (O1, K1 and M2 respectively). The number of recovered constituents is dependent on the length of the record and the choice of the Rayleigh coefficient. Selecting a Rayleigh coefficient less than 1 recovers more constituents, but the accuracy is reduced (Foreman, 1979).
**Stn: SB1  Depth: 0041m  NITINAT CANYON**  
**Lat 48 15.9W  Long 125 40.6W**

**Start** = 18.00 Hrs 10/6/85  **Mid** = 1.00 Hrs 17/6/85  **End** = 8.00 Hrs 23/6/85  PST  
**No. Pts** = 303  **No. Valid X Pts** = 303  **No. Valid Y Pts** = 303

**Ellipse Amplitudes in METRES/SEC**  
**Phase [g]** is GREENWICH PHASE LAG adjusted to local time

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**X Average** = 0.134  **X Standard Deviation** = 0.093  **RMS(Residual Error)** = 0.063  **X Matrix Condition** = 0.7160

**Y Average** = 0.187  **Y Standard Deviation** = 0.092  **RMS(Residual Error)** = 0.060  **Y Matrix Condition** = 0.7160

**RAYLEIGH Coefficient** = 0.83

File: /DATA/KEITH/*SB1_004185161A5.TR60

Constituents saved in file /DATA/CONST/*SB1_004185161A.CURR:::3:1

Tidal constituents for mooring C1, total depth 151 m. Meter 1 (depth 41 m)
Tidal constituents for mooring C1, total depth 151 m. Meter 2 (depth 121 m)
Stn: SB1 Depth: 0141m NITINAT CANYON

Lat 48 15.9N Long 125 40.6W

Start = 18.00 Hrs 10/6/85 Mid = 3.00 Hrs 19/6/85 End = 12.00 Hrs 27/6/85 PST
No. Pts = 403 No. Valid X Pts = 403 No. Valid Y Pts = 403

Ellipse Amplitudes in METRES/SEC
Phase[g] is GREENWICH PHASE LAG adjusted to local time

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Y Average = -.022 Y Standard Deviation = .100 RMS(Residual Error) = .050 Y Matrix Condition = .8464
RAYLEIGH Coefficient = 1.00

Tidal constituents for mooring C1, total depth 151 m. Meter 3 (depth 141 m)
Stn: SB1  Depth: 0148m  NITINAT CANYON
Lat 48° 15.9’N  Long 125° 40.6’W

Start = 18.00 Hrs 10/6/85  Mid = 3.00 Hrs 19/6/85  End = 12.00 Hrs 27/6/85 PST

No. Pts = 403  No. Valid X Pts = 403  No. Valid Y Pts = 403

Ellipse Amplitudes in METRES/SEC  Phase(g) is GREENWICH PHASE LAG adjusted to local time

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X Average = 0.021  X Standard Deviation = 0.099  RMS(Residual Error) = 0.039  X Matrix Condition = 0.8464
Y Average = -0.020  Y Standard Deviation = 0.079  RMS(Residual Error) = 0.058  Y Matrix Condition = 0.8464
RAYLEIGH Coefficient = 1.00

Tidal constituents for mooring C1, total depth 151 m. Meter 4 (depth 148 m)
**Stn:** SB2  Depth: 0040m  **Nitinat Shelf**  
**Lat:** 48.186N  **Long:** 125.306W  
**Start =** 16.00 Hrs 10/ 6/85  **Mid =** 10.00 Hrs 15/ 6/85  **End =** 4.00 Hrs 20/ 6/85  **PST**  
**No. Pts =** 229  **No. Valid X Pts =** 228  **No. Valid Y Pts =** 228

**Ellipses Amplitudes in METRES/SEC**  
**Phase[°]** is GREENWICH PHASE LAG

**adjusted to local time**

| Name | Cyc/Hr | Major | Minor | Inc | g | g+ | g- | CX | ERCK | CY | ERCY | SX | ERSX | SY | ERSY |
|------|--------|-------|-------|-----|---|----|----|----|------|----|------|----|------|    |      |
| 1 Z0 | 0.0000000000 | .143 | 0.000 | 127.5 | 180.0 | 52.5 | 307.5 | .087 | .007 | -.114 | .005 | -.000 | .005 | .000 | .003 |
| 2 MSF | .0028219325 | .046 | -.022 | 165.9 | 354.1 | 188.2 | 159.9 | -.027 | .011 | -.012 | .007 | .035 | .006 | .020 | .004 |
| 3 Q1 | .0372185040 | .006 | .001 | 151.3 | 299.3 | 148.0 | 90.6 | -.006 | -.015 | -.015 | .034 | .018 |
| 4 O1 | .0387306510 | .034 | -.014 | 164.9 | 262.6 | 97.7 | 67.5 | .033 | .007 | .018 | .007 | .000 | .005 | .018 | .005 |
| 5 P1 | .0415525880 | .021 | -.100 | 152.7 | 271.8 | 119.1 | 64.5 | .017 | .050 | .077 | .020 |
| 6 K1 | .0417807470 | .079 | -.038 | 158.7 | 261.8 | 103.2 | 60.5 | .033 | .007 | .075 | .007 | .049 | .005 | .005 | .005 |
| 7 N2 | .0789992510 | .023 | -.009 | 32.9 | 29.3 | 356.4 | 62.2 | .029 | -.063 | -.070 | .001 |
| 8 N2 | .0805114060 | .086 | .054 | 34.7 | 64.5 | 29.7 | 99.2 | .008 | .008 | .075 | .008 | .061 | .005 | .020 | .005 |
| 9 S2 | .0833333280 | .028 | -.024 | 96.0 | 315.2 | 224.6 | 45.8 | .007 | .008 | .023 | .008 | .027 | .005 | .007 | .005 |
| 10 K2 | .0835616950 | .020 | -.010 | 99.5 | 272.6 | 173.2 | 12.1 | .003 | -.027 | -.024 | .003 |
| 11 M3 | .1207651000 | .009 | -.007 | 169.4 | 124.1 | 314.7 | 293.5 | .003 | .007 | -.006 | .005 | -.008 | .007 | -.002 | .005 |
| 12 SK3 | .1251140800 | .008 | -.003 | 11.4 | 271.8 | 260.4 | 283.3 | .008 | .007 | -.003 | .005 | -.003 | .003 | -.003 | .005 |
| 13 M4 | .1610228100 | .011 | -.007 | 89.7 | 176.6 | 86.9 | 266.3 | .001 | .009 | -.011 | .006 | -.007 | .009 | .002 | .006 |
| 14 MS4 | .1638443700 | .014 | -.011 | .3 | 52.3 | 52.1 | 52.6 | .004 | .009 | -.010 | .007 | .013 | .010 | .003 | .007 |
| 15 S4 | .1666666600 | .012 | -.003 | 147.4 | 352.2 | 204.8 | 139.6 | .002 | .009 | -.005 | .006 | -.010 | .009 | .005 | .006 |
| 16 2MK5 | .2028035500 | .009 | -.006 | 9.2 | 140.7 | 131.5 | 149.8 | -.010 | .007 | .001 | .005 | -.000 | .007 | .000 | .007 |
| 17 2SK5 | .2084474300 | .005 | -.002 | 123.0 | 233.4 | 110.4 | 356.5 | -.003 | .007 | .003 | .005 | -.000 | .007 | .004 | .005 |
| 18 M6 | .2415342000 | .008 | -.003 | 66.5 | 126.1 | 59.6 | 192.6 | -.004 | .009 | -.002 | .006 | -.002 | .009 | .007 | .006 |
| 19 2MS6 | .2445561300 | .010 | -.003 | 174.9 | 285.9 | 111.0 | 100.8 | -.009 | .010 | -.000 | .007 | -.003 | .010 | .003 | .007 |
| 20 2SM6 | .2471780800 | .006 | -.004 | 127.9 | 353.3 | 225.4 | 121.2 | -.002 | .009 | -.004 | .006 | -.004 | .009 | .003 | .006 |
| 21 3MK7 | .2833149000 | .003 | .001 | 8.9 | 112.7 | 103.8 | 121.6 | -.003 | .007 | .000 | .005 | .002 | .007 | .001 | .005 |
| 22 MB | .3220456200 | .002 | .001 | 157.0 | 321.2 | 164.2 | 118.2 | .001 | .007 | .000 | .005 | .001 | .007 | .001 | .005 |

**X Average=** .078  **X Standard Deviation=** .115  **RMS(Residual Error)=** .072  **X Matrix Condition=** .1048  
**Y Average=** -.121  **Y Standard Deviation=** .078  **RMS(Residual Error)=** .050  **Y Matrix Condition=** .1048  
**RAYLEIGH Coefficient=** .63

File: *SB2_004085161A.TR60*

Constituents saved in file /DATA/CONST/*SB2_004085161A.CURR:3:1

**Tidal constituents for mooring C2, total depth 129 m. Meter 1 (depth 40 m)**
Stn:SB2 Depth:0041m NITINAT SHELF       Lat 48 18.3N Long 125 30.6W

Start = 17.00 Hrs 10/ 6/85  Mid = 1.00 Hrs 17/ 6/85  End = 9.00 Hrs 23/ 6/85 PST
No. Pts=  305  No. Valid X Pts=  304  No. Valid Y Pts  304

Ellipse Amplitudes in METRES/SEC      Phase[g] is GREENWICH PHASE LAG adjusted to local time

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Y Average = .121  Y Standard Deviation = .104  RMS(Residual Error) = .064  Y Matrix Condition = .7254
RAYLEIGH Coefficient = .03

File : /DATA/KEITH/*SB2_004185161A.TR60
Constituents saved in file /DATA/CONST/*SB2_004185161A.CURR:::3:1

Tidal constituents for mooring C2, total depth 129 m. Meter 2 (depth 41 m)
Stn:S82 Depth:0099m  NITINAT SHELF  Lat 48 18.3N  Long 125 30.6W

Start = 16.00 Hrs 10/ 6/85  Mid = 1.00 Hrs 19/ 6/85  End = 10.00 Hrs 27/ 6/85  PST
No. Pts=  403  No. Valid X Pts=  403  No. Valid Y Pts= 403

Ellipse Amplitudes in METRES/SEC  Phase[g] is GREENWICH PHASE LAG adjusted to local time

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Y Average=-.016 Y Standard Deviation=.083 RMS(Residual Error)=.051 Y Matrix Condition=.8464
RAYLEIGH Coefficient=1.00

Tidal constituents for mooring C2, total depth 129 m. Meter 3 (depth 99 m)
Stn: SB2  Depth: 0119m  NITINAT SHELF  Lat 48 18.3N  Long 125 30.6W

Start = 16.00 Hrs 10/6/85  Mid = 1.00 Hrs 19/6/85  End = 10.00 Hrs 27/6/85 PST
No. Pts = 403  No. Valid X Pts = 403  No. Valid Y Pts = 403

Ellipse Amplitudes in METRES/SEC  Phase [g] is GREENWICH PHASE LAG adjusted to local time

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X Average = .124  X Standard Deviation = .091  RMS(Residual Error) = .053  X Matrix Condition = .8464
Y Average = -.015  Y Standard Deviation = .060  RMS(Residual Error) = .036  Y Matrix Condition = .8464

Rayleigh Coefficient = 1.00

Tidal constituents for mooring C2, total depth 129 m. Meter 4 (depth 119 m)
Tidal constituents for mooring C2, total depth 129 m. Meter 5 (depth 126 m)
**Stn:SC1 Depth:0043m NITINAT SHELF**

**Start = 16.00 Hrs 10/6/85  Mid = 1.00 Hrs 17/6/85  End = 10.00 Hrs 23/6/85 PST**

**No. Pts= 307  No. Valid X Pts= 299  No. Valid Y Pts 299.**

**Ellipse Amplitudes in METRES/SEC Phase[1] is GREENWICH PHASE LAG adjusted to local time**

**Tidal constituents for mooring C3, total depth 116 m. Meter 1 (depth 43 m)**
Ellipse Amplitudes in METRES/SEC  Phase[g] is GREENWICH PHASE LAG adjusted to local time

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Y Average= .008 Y Standard Deviation= .105 RMS(Residual Error)= .022 Y Matrix Condition= .9898
RAYLEIGH Coefficient= 1.00

Tidal constituents for mooring C3, total depth 116 m. Meter 2 (depth 86 m)
Stn:SC1  Depth:0106m  NITINAT SHELF

Lat 48 22.0N  Long 125 20.1W

Start = 14.00 Hrs 10/6/85  Mid = 0.00 Hrs 19/6/85  End = 10.00 Hrs 27/6/85 PST

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Ellipse  Amplitudes in METRES/SEC  Phase[g] is GREENWICH PHASE LAG adjusted to local time

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RAYLEIGH Coefficient= 1.00

Tidal constituents for mooring C3, total depth 116 m. Meter 3 (depth 106 m)
### Tidal constituents for mooring C3, total depth 116 m. Meter 4 (depth 113 m)

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RAYLEIGH Coefficient= .90

File : /DATA/KEITH/*JFR_024585161A5.TR60
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Tidal constituents for mooring C4, total depth 255 m. Meter 2 (depth 245 m)
Stn: JFR Depth: 0250m JUAN DE FUCA ENTRANCE

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Y Average = -.002 Y Standard Deviation = .104 RMS(Residual Error) = .044 Y Matrix Conditions = .9071
RAYLEIGH Coefficient = .90

File: /DATA/KEITH/*JFR_025085161A5.TR60

Constituents saved in file /DATA/CONST/"JFR_025085161A.CURR::3:1

Tidal constituents for mooring C4, total depth 255 m. Meter 3 (depth 250 m)
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X Average= 0.092  X Standard Deviation= 0.235 RMS(Residual Error)= 0.043  X Matrix Condition= 0.9071
Y Average= -0.000  Y Standard Deviation= 0.093 RMS(Residual Error)= 0.039  Y Matrix Condition= 0.9071
RAYLEIGH Coefficient= 0.90

File: /DATA/KEITH/*JFR_025285161A5.TR60

Constituents saved in file /DATA/CONST/*JFR_025285161A.CURR:::3:1

Tidal constituents for mooring C5, total depth 255 m. Meter 4 (depth 252 m)