Temperature and temperature-gradient records made using thermistors and differential thermopiles on a freely-falling instrument package in 100 meters of water on the Oregon shelf are used to describe the finestructure and microstructure associated with the following hydrographic features: a well-stirred surface layer with an equilibrium thickness of about 8 m; a seasonal thermocline with transition regions above and below; a cool ribbon of water below the thermocline at about 20 m depth; a salinity-compensated warm-water anomaly in the range 25-35 m; and a stable temperature inversion at about 60 m. The temperature gradient spectra, when fit to the universal spectra, yield estimates of thermal and mechanical dissipation rates as high as $10^{-5} \, \degree C^2 \, s^{-1}$ and $10^{-2} \, cm^2 \, s^{-3}$, respectively in patches of turbulence in the surface layer and thermocline; and lower than $10^{-7} \, \degree C^2 \, s^{-1}$ and $10^{-5} \, cm^2 \, s^{-3}$ deeper in the water column. By assuming a local balance between production of thermal fluctuations and their dissipation, the vertical eddy diffusivity for heat is estimated to vary from $50 \, cm^2 \, s^{-1}$ in the surface mixed-layer to about $1 \, cm^2 \, s^{-1}$ or less in the seasonal thermocline, to less than $0.1 \, cm^2 \, s^{-1}$ in some of the temperature inversions. Systems
of layers and interfaces produced by the mechanism of diffusive double-diffusive convection were not found on the inversions. Some non-persistent layers and interfaces were observed throughout the water column and it is conjectured that they were produced by an interaction of the shear with the ambient stratification. The interfaces seem to degenerate into billow turbulence initiated by Kelvin-Helmholtz instabilities.
Temperature Finestructure and Microstructure Observations in the Coastal Upwelling Region off Oregon during the Summer of 1974

by

George Otto Marmorino

A THESIS
submitted to
Oregon State University

in partial fulfillment of the requirements for the degree of

Doctor of Philosophy

June 1977
APPROVED:

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Thesis presented on November 12, 1976

Typed by Rebecca Rakish for George Marmorino
This work is dedicated to my mother and father.
ACKNOWLEDGEMENTS

Dr. Douglas Caldwell has been a source of both encouragement and inspiration throughout this work, from the initial stages of development work in Green Peter Lake, to the final stages of figure drawing and manuscript editing; and in addition, allowing me to work in his office, provided me with ready access to the computer terminals and drafting room.

Professor Robert L. Smith kindly made available the current meter records and made possible the collection and processing of the indispensable CTD data. Help at sea was provided by John Brubaker, Dennis Barstow, Tom Yao, Steve Wilcox, Stuart Eide, Priscilla Newberger, and Mark Matsler. The CTD work was supervised by Des Barton. I wish to thank my committee members -- Drs. Clayton Paulson, Bob Smith, and Henry Crew -- for their comments on the material.

The research was sponsored by the Oceanography Section, National Science Foundation under Grant GA-23336 and DES74-17968.
# TABLE OF CONTENTS

I. INTRODUCTION

II. FINESTRUCTURE AND MICROSTRUCTURE OBSERVATIONS DURING A PERIOD OF VARIABLE WINDS
   Observations
   - Surface mixed-layer
     - temperature microstructure
     - mixed-layer deepening
   - Cool ribbon
     - heat budget
     - temperature inversion
   - Warm anomaly
     - fineststructure
     - microstructure
   - Deep inversion
   Conclusions

III. DISSIPATION OF TEMPERATURE FLUCTUATIONS IN THE SEASONAL THERMOCLINE
   Theory
   - Temperature gradient spectrum
   - Estimating $\varepsilon$, $\varepsilon_\theta$, and $K_H$
   Observations
   Results
   - Spectra
     - Turbulent patches
     - Lengthscales and magnitudes of the stable gradients
   Discussion
   - A mixing model
   - Heat budget
   - Comparison with other measurements
   Conclusions
IV. LAYERS AND INTERFACES: SMALL-SCALE IMPRINTS ON THE TEMPERATURE STRUCTURE

Observations

Double-diffusive layering 102

Non-double-diffusive layering 108

Kelvin-Helmholtz instabilities? 111

Conclusions 118

V. SMALL-SCALE TEMPERATURE STRUCTURE AS MEASURED BY THERMOCOUPLES

The Thermocouple Transducers

Construction details 121

Sensitivity and spatial resolution 124

Difficulties encountered 126

Data Analysis

Theory 128

Results

In situ sensitivity 130

Use of a thermopile in conjunction with a thermistor 131

Use of two thermopiles simultaneously 144

A horizontal thermopile 144

Conclusions 149

VI. CONCLUSIONS 151

VII. REFERENCES 153
LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>Parameters for the resolved surface mixed-layers</td>
<td>19</td>
</tr>
<tr>
<td>II</td>
<td>Parameters in the warm anomaly</td>
<td>51</td>
</tr>
<tr>
<td>III</td>
<td>Parameters in several regimes</td>
<td>63</td>
</tr>
<tr>
<td>IV</td>
<td>Microstructure drops and surface conditions</td>
<td>72</td>
</tr>
<tr>
<td>V</td>
<td>Large-scale observations just above, within, and just below the thermocline</td>
<td>76</td>
</tr>
<tr>
<td>VI</td>
<td>Calculations for observations just above, within, and just below the thermocline</td>
<td>85</td>
</tr>
<tr>
<td>VII</td>
<td>Calculations for turbulent patches</td>
<td>89</td>
</tr>
<tr>
<td>VIII</td>
<td>Interface, layer, and stability characteristics</td>
<td>109</td>
</tr>
<tr>
<td>IX</td>
<td>Thermopile characteristics</td>
<td>123</td>
</tr>
</tbody>
</table>
## LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Location of 100 m station</td>
<td>5</td>
</tr>
<tr>
<td>2</td>
<td>Summer 1974 north-south winds as recorded on the jetty at Newport, Oregon (44°38'N, 124°04'W).</td>
<td>9</td>
</tr>
<tr>
<td>3</td>
<td>Temperature, plotted against depth and time, from CTD casts made within 0.5 km of the 100 m station (45°00.0'N and 124°10.0'W).</td>
<td>11</td>
</tr>
<tr>
<td>4</td>
<td>Current stick-diagram for the current meters at 25, 50, 75, and 90 m. The Newport-jetty wind is shown at the top.</td>
<td>12</td>
</tr>
<tr>
<td>5</td>
<td>CTD and MSP measurements of properties of the surface layer.</td>
<td>14</td>
</tr>
<tr>
<td>6</td>
<td>Net heat flux into (positive) the ocean surface</td>
<td>16</td>
</tr>
<tr>
<td>7</td>
<td>Gross MSP temperature profiles, 6-min. apart, through the well-mixed surface layer and thermocline (1700 on Sept. 8, Y3 Drop 7).</td>
<td>17</td>
</tr>
<tr>
<td>8</td>
<td>Surface-layer microstructure (Y3, Drop 34A).</td>
<td>22</td>
</tr>
<tr>
<td>9</td>
<td>(a) Temperature-gradient spectra of the mixed-layer record shown in Figure 8. (b) The gradient spectra corrected in various ways.</td>
<td>23</td>
</tr>
<tr>
<td>10</td>
<td>Log-log plot of surface mixed-layer thickness (h) against time during an upwelling-favorable wind event.</td>
<td>26</td>
</tr>
<tr>
<td>11</td>
<td>Salinity profiles (light lines) before the deepening, 0-10 hr. on Figure 10, and (heavy lines) after the deepening, 35-37 hr.</td>
<td>28</td>
</tr>
<tr>
<td>12</td>
<td>(a) Two gross temperature profiles during relatively calm conditions (Y3, Drops 23A and 26A). (b) Sequence of gross temperature profiles through the surface mixed-layer and seasonal thermocline (Y3, Drops 35A-L).</td>
<td>29-30</td>
</tr>
<tr>
<td>13</td>
<td>T-S curves during (CTD 7-15), and after (CTD-18-25) the presence of the cool ribbon.</td>
<td>34</td>
</tr>
<tr>
<td>14</td>
<td>(Top) Cool ribbon temperature contours based on MSP profiles. (Bottom) Temperature gradient profiles.</td>
<td>35</td>
</tr>
<tr>
<td>Figure</td>
<td>Page</td>
<td></td>
</tr>
<tr>
<td>--------</td>
<td>------</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>Changes in the cool-ribbon core during the observation period.</td>
<td>38</td>
</tr>
<tr>
<td>16</td>
<td>Temperature profiles through the cool ribbon and the underlying warmer water.</td>
<td>40</td>
</tr>
<tr>
<td>17</td>
<td>Plot of ΔT vs. ΔS across the inversion at 28 m during Y2 and at 25 m (below the cool ribbon) during Y3.</td>
<td>43</td>
</tr>
<tr>
<td>18</td>
<td>T-S diagram for CTD casts 30 (1237 PDT, 15 Sept. 1974) and 31 (1412 PDT).</td>
<td>44</td>
</tr>
<tr>
<td>19</td>
<td>(Top) Temperature profiles through the warm-water anomaly (Y2, Drops 17A-L).</td>
<td>45</td>
</tr>
<tr>
<td>20</td>
<td>Plot of temperature, salinity, and depth ranges of the anomaly vs. time.</td>
<td>48</td>
</tr>
<tr>
<td>21</td>
<td>Ensemble-averaged temperature gradient spectra from the warm-water anomaly.</td>
<td>49</td>
</tr>
<tr>
<td>22</td>
<td>Temperature and salinity profiles from CTD 42 (PDT 1047, 10 Sept.)</td>
<td>53</td>
</tr>
<tr>
<td>23</td>
<td>(a) T-S diagram for CTD 42 (PDT 1047, 10 Sept.) and CTD 43 (PDT 1437). (b) T-S diagram showing the warm state (CTD 33-40) and the cooler state (CTD 44-51).</td>
<td>54</td>
</tr>
<tr>
<td>24</td>
<td>(Top) Temperature and (Bottom) temperature gradient profiles through a deep inversion with several isotherms indicated (1117-1218 PDT, 10 Sept.).</td>
<td>55</td>
</tr>
<tr>
<td>25</td>
<td>Histograms of 1 m averaged gradients from gross temperature profiles through the relatively isothermal core.</td>
<td>56</td>
</tr>
<tr>
<td>26</td>
<td>Ensemble-averaged temperature gradient spectra above, within, and below the core.</td>
<td>58</td>
</tr>
<tr>
<td>27</td>
<td>Plot of kurtosis and skewness (both are zero for a normal distribution) for gradient records above, within and below the core.</td>
<td>59</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
<td></td>
</tr>
<tr>
<td>--------</td>
<td>-------------</td>
<td></td>
</tr>
<tr>
<td>28</td>
<td>T-S diagram for (top) the August cruise and (bottom) the September cruise.</td>
<td></td>
</tr>
<tr>
<td>29</td>
<td>Temperature profiles through the thermocline. MSP 21 A-E; Ri ~ 300.</td>
<td></td>
</tr>
<tr>
<td>30</td>
<td>Temperature profiles through the thermocline. MSP 17 A-L; Ri ~ 50.</td>
<td></td>
</tr>
<tr>
<td>31</td>
<td>Temperature profile and (uncorrected) gradient profile MSP 32; Ri ~ 35.</td>
<td></td>
</tr>
<tr>
<td>32</td>
<td>Turbulent patch within the thermocline (MSP 31C)</td>
<td></td>
</tr>
<tr>
<td>33</td>
<td>Normalized, ensemble-averaged spectra (left) above (middle), and (right) below the thermocline (MSP 19, 21).</td>
<td></td>
</tr>
<tr>
<td>34</td>
<td>Normalized thermocline spectra (MSP 31, 32)</td>
<td></td>
</tr>
<tr>
<td>35</td>
<td>Gradient spectrum of turbulent patch in the thermocline.</td>
<td></td>
</tr>
<tr>
<td>36</td>
<td>Gradient spectrum from the MSP 17D thermocline and the normalized, ensemble-averaged spectrum for turbulent patches in 17 D-L.</td>
<td></td>
</tr>
<tr>
<td>37</td>
<td>Plot of the average of the maximum (stable) vertical temperature gradients observed in &quot;interfaces&quot; of the given thickness.</td>
<td></td>
</tr>
<tr>
<td>38</td>
<td>Histograms of observed interface thickness from above, within, and below the thermocline.</td>
<td></td>
</tr>
<tr>
<td>39</td>
<td>Plot of vertical eddy diffusivity $K_H$ against the square of the buoyancy frequency.</td>
<td></td>
</tr>
<tr>
<td>40</td>
<td>Temperature, plotted against depth and time, from CTD casts made within 0.5 km of a 100 m station (45°00.0'N and 124°10.0'W) off the Oregon coast in September 1974.</td>
<td></td>
</tr>
<tr>
<td>41</td>
<td>Temperature profiles through the temperature inversion below the cool ribbon.</td>
<td></td>
</tr>
<tr>
<td>42</td>
<td>Profile (Drop 26A) through a deep-water temperature inversion. The inset shows data from a nearby CTD.</td>
<td></td>
</tr>
<tr>
<td>Figure</td>
<td>Page</td>
<td></td>
</tr>
<tr>
<td>--------</td>
<td>------</td>
<td></td>
</tr>
<tr>
<td>43</td>
<td>110</td>
<td></td>
</tr>
<tr>
<td>44</td>
<td>112</td>
<td></td>
</tr>
<tr>
<td>45</td>
<td>114</td>
<td></td>
</tr>
<tr>
<td>46</td>
<td>115</td>
<td></td>
</tr>
<tr>
<td>47</td>
<td>122</td>
<td></td>
</tr>
<tr>
<td>48</td>
<td>133</td>
<td></td>
</tr>
<tr>
<td>49</td>
<td>134</td>
<td></td>
</tr>
<tr>
<td>50</td>
<td>135</td>
<td></td>
</tr>
<tr>
<td>51</td>
<td>137</td>
<td></td>
</tr>
<tr>
<td>52</td>
<td>139</td>
<td></td>
</tr>
<tr>
<td>53</td>
<td>140</td>
<td></td>
</tr>
<tr>
<td>54</td>
<td>141</td>
<td></td>
</tr>
<tr>
<td>55</td>
<td>142</td>
<td></td>
</tr>
<tr>
<td>56</td>
<td>143</td>
<td></td>
</tr>
</tbody>
</table>

43 Shear-induced layering in the seasonal thermocline (Drop 8A).

44 Temperature profiles through an interface at about 73 m (Drop 6A, B, C, D).

45 Profiles through a 10 m thick temperature inversion (Drop 25 A-G, 1117-1218 PDT, 10 September).

46 An example of the early "roll-up" stage of Kelvin-Helmholtz initiated billow turbulence.

47 Thermopiles, of 3 and 9 cm diameters, mounted concentrically at the lower end of free-fall instrumentation.

48 An example of in situ comparisons between the largest diameter thermopile (9 cm) and a differentiated thermistor.

49 Simultaneous records of the differentiated thermistor and thermopile (3 cm) obtained from the ocean (Y3-7A, 76-81 m) and from a fresh-water lake (GP7-5, 50-75 m).

50 Profiles obtained using two thermopiles simultaneously (N = 0.003 hz and $\frac{\partial T}{\partial z} = 0.02$ °C/m): (a) 1.5 cm diameter $T_p$ vs. 5 cm; (b) 1.5 cm vs. 9 cm.

51 Coherence between thermopile (3 cm) and differentiated-thermistor records from the ocean, and (O) a fresh water lake.

52 Ensemble-averaged (heavy curve) and individual coherence and phase plots for seven thermopile (9 cm)/thermistor yo-yos through the thermocline shown in Figure 48.

53 Thermopile and differentiated-thermistor normalized, ensemble-averaged, autospectra for the thermocline profiles shown in Figure 48.

54 Coherence and phase plots for seven thermopile/thermistor yo-yos through a stable temperature inversion.

55 Squared norm of the frequency response function.

56 Coherence between thermistor and thermopile of different diameters.
Dependence of coherence between a thermopile and thermistor upon water-column stability.

Coherence between two thermopiles with $N = 0.003$ Hz but with $D$ variable.

Differentiated thermistor vs a horizontal thermopile.
TEMPERATURE FINESTRUCTURE AND MICROSTRUCTURE OBSERVATIONS IN THE COASTAL UPWELLING REGION OFF OREGON DURING THE SUMMER OF 1974

I. INTRODUCTION

Vertical profiles of temperature made with high-resolution instrumentation have shown that the stratification in geophysical fluids is not smooth. In the ocean, various processes -- interleaving of locally created water masses of slightly different densities, thermohaline double-diffusion, or weak shearing, for example -- can create "ragged" profiles with typical vertical scales of variation of one to ten meters; this is finestructure. Strong shears, produced for example by intrusions, boundary-processes or interfacial waves, stir the finestructure producing finer-scale variations on the order of 10 cm or less; this has come to be called microstructure. (When a weak stirring acts in a region that is well-mixed in temperature already there may not be any temperature microstructure created.) If the stirring action is prolonged, a turbulent condition may develop and the temperature fluctuations may be used as an indicator of the intensity of the turbulence. In all these cases, high-resolution temperature profiles alone can be used to identify the type of process at work and to estimate rates of vertical transport. This dissertation reports the first such measurements in the upwelling region off Oregon.

The data were obtained on three short cruises in the summer of 1974, but only two cruises, one in August and the other in September, are discussed here. In addition to the high-resolution
temperature profiles, conductivity-temperature-depth (CTD) casts provided information on the density distribution, and the vertical shear of the horizontal flow was calculated using time series from a nearby current meter mooring. The methods used involved some drawbacks -- working from a drifting ship, sometimes in fairly rough conditions; entanglement or breakage of flimsy conductor links; and recording data on chart paper which later had to be digitized by hand -- which somewhat limited the amount of data that could be collected, but still a large amount of good representative data were obtained.

Some mention should be made here with regard to the estimation of the dissipation rates of thermal and mechanical energy via comparisons between the "observed" temperature-gradient spectra and the "universal" spectra. As will be explained in some detail later, there are fairly large uncertainties in both the frequency response function for the transducer used and in the values of the "universal constants". Thus, the estimates derived here may be meaningful only in a relative sense, and must of course be used with this in mind. It is interesting perhaps to point out that many of the shortcomings encountered in the present investigation have since been overcome -- for example, a magnetic tape data-acquisition system is now in use; a smaller transducer with better-known response is available; and the new kind of conductor link and instrument-package
design allows repeated profiles to be made much more easily -- and future work, which hopefully will be encouraged, should certainly help settle some of the uncertainty in the absolute value of the estimates.

The organization after this introductory chapter is as follows: Chapter II defines the hydrographical setting and presents an overview of the finest structure and microstructure; Chapter III discusses the summer thermocline and the transition regions just above and below it; Chapter IV documents a search for double-diffusive convective layers and Kelvin-Helmholtz instabilities; Chapter V reports the use of a new transducer for microstructure work -- an array of thermocouples; and Chapter VI contains some concluding remarks.
II. FINESTRUCTURE AND MICROSTRUCTURE OBSERVATIONS DURING A PERIOD OF VARIABLE WINDS

Observations of temperature finestructure and microstructure have been made in the ocean off the Oregon coast (Figure 1) in a regime strongly influenced by seasonal coastal upwelling and possessing a variety of hydrographic phenomena. The main hydrographic feature is the gradual upward slope of the isopycnals shoreward. Transient phenomena due to changes in the wind from upwelling-favorable to non-favorable include nearshore, surface, and bottom responses. In this paper we describe the temperature finestructure (1-m scales) and microstructure (1-cm scales) during such a transient period, in the following regimes: the surface mixed-layer and the transition to the seasonal thermocline, the core of relatively cool fresh water found within or just below the thermocline, the warm anomaly, and deep temperature inversions. The seasonal thermocline is examined in the next Chapter; the bottom boundary layer has been recently discussed by Caldwell (1976).

While these are the first published temperature microstructure measurements made in these regimes, much previous work has been done. For example, Halpern (1976) recently discussed the structure and stability of the density field during a coastal upwelling event as measured in water 100 m deep, including estimates of the vertical mixing in the wind-drift layer; Huyer, Smith and Pillsbury (1974) discussed the upwelling region during a period of variable winds; Huyer and Smith (1974) discovered the cool ribbon, and they, Wang (1976) and Wright (1976) discussed the ribbon and warm anomaly below it. The warm anomaly has been found to be coincident with a zone of low dynamic stability (Mooers, Collins, and
Figure 1. Location of 100-m station. The circle has a 1-km diameter.

The descriptive questions to answer are:

(1) If there is any temperature microstructure present is it persistently associated with particular larger-scale (1-10 m) structure?

(2) Can episodes of activity be correlated with current shear or intrusive events?

(3) Is the small-scale structure of the layer-interface type? Are double-diffusive processes active where they could occur? Is there evidence of billow turbulence or buoyant convection?

The classical oceanographic problem of obtaining estimates of vertical eddy diffusivities is of great concern here because of the very large ratio of horizontal ($L_H$) to vertical ($L_V$) dimensions of the features of interest. For example, the warm anomaly has a ratio of $L_H/L_V = 10^3$. The total vertical transport from such a feature then can be quite large even if the fluxes are small. Diffusivities can be obtained by estimating rates of mechanical energy dissipation $\varepsilon$ and the dissipation rate of thermal fluctuations $\varepsilon_\theta$, estimates which rest upon a knowledge of the thermistor response function at high wavenumbers (0.5 to 5 cm/cm) and the assumption that the appropriate turbulence theory—essentially the universal shape of the temperature or temperature
gradient spectrum--can be used.¹

Three "microstructure cruises" were conducted (Figure 2), but only results from the second (Y2) and third (Y3) will be discussed here. Cruise Y2 followed an intense period of upwelling but before measurements could begin a strong thermocline had already developed and the warm water anomaly had weakened considerably. The data from the surface layer, cool ribbon, and deep inversion come from Y3. The fact that we were not out for the entirety of an active upwelling event should not be taken to imply a lack of sea-worthiness of the authors but was due, of course, to ill-timed ship scheduling.

The microstructure profiler (MSP), discussed by Caldwell, Wilcox, and Matsler (1975) and Caldwell (1976), uses a 0.057-cm bead thermistor and occasionally a horizontal array of thermocouples (see Chapter V). Vertical profiles (drops) were made at fall speeds of 10-50 cm/sec. Chart records of temperature and temperature derivative could later be digitized. The convention used in the figures is that "z" is positive downwards and a negative temperature gradient is stabilizing. Repeated profiles ("yo-yos") are designated by a letter following the

¹How to correct the thermistor record for encapsulation and water boundary-layer signal attenuation is still not satisfactorily known. We use the two-pole power correction advocated by Gregg (1976b)--this may be an over-correction (Gregg, 1976a)--taking care to observe the change in time constant with fall speed. Just as important are the uncertainties in the shape of the universal, one-dimensional, isotropic-turbulence, temperature-gradient spectrum. It is crucial, for instance, to know where the inertial subrange intersects the viscous-convective subrange in order to estimate $\varepsilon$ and $\phi$ but the intersection wavenumber depends upon "universal constants" which current workers believe are not constant but depend upon the Reynolds number (Clay, 1973; Williams and Paulson, 1976).
Figure 2. Summer 1974 north-south winds as recorded on the jetty at Newport, Oregon (44°38'N, 124°04'W). The three cruises made on the R/V Yaquina are indicated. Negative wind-speed-squared indicates an upwelling-favorable north wind.
Large-scale temperature and salinity data were obtained with a CTD. Data from a current-meter mooring near the 100-m station was provided by Professor R.L. Smith.

OBSERVATIONS

The winds, measured from the south jetty at Newport, Oregon, were predominantly southward through the summer but during times the fine-structure measurements were made significant departures from the mean occurred (Figure 2).

A summary of the temperature structure from CTD casts (Figure 3) shows a surface layer of variable thickness, a thermocline centered at about 20 m, the cool ribbon bisecting the thermocline on the 7th and 8th, some temperature inversions at about 30 m, the bulk of the water column at 8-9 °C, and some pockets of relatively warmer water at 70 m on the 10th and 11th. The mean period of the oscillations in the 8°C isotherm through the 9th is 12±2 h, and could represent the semi-diurnal tide. The oscillations in the water column between the surface and 40 m are of shorter period—about 7.5 h; their 10-15 m amplitude is similar to that seen in a 1973 100-m anchor station (Holbrook and Halpern, 1974). Spectral analysis of 54-day temperature records from the 100-m current-meter mooring shows significant energy at the semi-diurnal period at 90-m, and at shorter periods (5-10 h) at 25 m.

A weak upwelling event preceded our arrival on station (Figures 2 and 4). During the period of weak winds and the onset of northward winds
Figure 3. Temperature, plotted against depth and time, from CTD casts made within 0.5 km of the 100-m station (45°00.0'N and 124°10.0'W). Times of CTD casts used and MSP profiles are shown at the bottom.
Figure 4. Current stick-diagram for the current meters at 25, 50, 75, and 90 m. The Newport-jetty wind is shown at the top. An upwards-pointing stick indicates a northward flow. The isotherms are reproduced from Figure 3. Periods without sticks at 25 m are due to faulty direction data. Sea/land breeze effects evident in the wind-sticks were not seen on station.
which followed (September 7 and early on the 8th), the cool ribbon and
temperature inversions were present, the currents were southward except
near the surface, and the coldest and saltiest bottom water ($T \approx 7.65^\circ C,
S \approx 33.82 \, ^\circ /\circ_0$) was observed. The northward winds continued for two
days, driving the dense bottom water and ribbon further out on the shelf.
The temperature inversions weakened and eventually disappeared. The cur-
rents changed to northward first at 50, 75, and 90-m meters—the poleward
undercurrent—and later at 25 m. Toward the end of the cruise southward
winds reversed the current to a southward flow again, the near surface
flow—the alongshore jet—being stronger than the deeper current. Again
the bottom water becomes cold and salty. This sequence of events is
quite similar to observations by Huyer, Smith, and Pillsbury (1974)
during a period of variable winds in July 1972.

Surface mixed-layer

The thickness and mean temperature of the surface mixed-layer (sm1),
though quite variable over the cruise, seem to have been consistently
measured by both the MSP and CTD (Figure 5). The CTD cast was made amid-
ships while station position was maintained with bow-thrusters (the ship's
draft was almost 3 m)$^2$; and the MSP profiles were always made away from
the stern while the ship drifted broadside to the wind. It is believed

$^2$Measurements of surface layer properties from large ships are apparently
valid, at least for low wind speeds. STD casts from a much larger ship,
NOAA's Oceanographer, anchored in 103 m of water off the Oregon coast,
were used to study the changes in heat content of an 8-m deep surface
mixed-layer (Reed and Halpern, 1975).
Figure 5. CTD and MSP measurements of properties of the surface layer.

The winds shown on the wind stick-diagram were observed on the R/V Yaquina while on station (45°N, 124°10'W).
that there was no ship-induced mixing or contamination sampled by the MSP.

During the period of northward winds there occurred first a rapid increase in sml thickness (to over 10 m) late on the 8th, and then on the 9th an increase in temperature to 15°C. Although the deepening could have been due to local wind-mixing, advective effects are more probable. For example, the temperature increase on the 9th is accompanied by a 0.9‰ salinity decrease, to 31.3‰. It is possible that a river plume, perhaps from the Columbia, was advected past the station. The gradual warming trend over the 9th and 10th can be explained by a net heat gain due to insolation (Figure 6) of about 300 cal cm\(^{-2}\) which, when distributed over the 3-m sml, yields about a 1°C rise.

During the period of strong south winds, the sml was relatively well mixed, often with a sharp interface, with a fairly complex microstructure and overall vertical gradient at its base very much larger than in the thermocline below (Figure 7). There is some evidence for entrainment of water from below and occasionally the sharp interface breaks down, presumably due to a mixing event as apparently happened in 11E.

Toward the end of the cruise there occurred an sml cooling and deepening associated with an impulsive north wind. This deepening event will be discussed in detail later.

Temperature microstructure

Because the sml was so thin--the 2-m long MSP often being of greater extent--and because gains were kept low so the thermocline signals would
Figure 6. Net heat flux into (positive) the ocean surface.
The temperature scale applies only to profile A, the other profiles are offset.
remain on scale, few MSP profiles yielded records suitable for spectral analysis. A total of eight drops, however, did resolve much of the sml temperature microstructure and these are summarized in Table 1. Unfortunately, these cannot be thought of as an ensemble because they sample the surface under non-steady conditions; also, the length of record analyzed was necessarily short (less than 50% of the sml thickness). Nevertheless, a study of this small sample provides some interesting results.

In Table 1, the Monin Obukhov length $L_b$ is reckoned negative when the surface bouyancy flux leads to unstable conditions (Turner, 1973). Thus, $L_b$ is typically $-15$ m at night, implying that buoyant convection could mix downwards to 15 m if the sml were that thick; and during the day, $L_b \approx +7$ m due to the stabilizing effect of insolation. Hourly ship observations and the methods in James (1966) were used for the necessary calculations.

The temperature gradient records from the sml show relatively strong thermal activity in 50-100% of the layer. The smallest fluctuating gradients resolved were less than $10^{-3}$ °C/cm (c.f. the adiabatic gradient, $10^{-6}$ °C/cm), and the largest were greater than 0.02 °C/cm. A particularly active section is shown in Figure 8 along with the (integrated) temperature profile.

The flatness factors (kurtosis) for the resolved sml (Table 1) are large and positive ($K = 0$ for a normal distribution) as expected for a turbulent signal. The kurtosis, the variance $\sigma^2$, and skewness $S$ were computed from the gradient records uncorrected for thermistor response; if corrections had been applied all three statistics would increase.
Table 1. Summary of parameters for the resolved surface mixed-layers.

<table>
<thead>
<tr>
<th>Drop</th>
<th>Fall-speed (cm/s)</th>
<th>Length of section analyzed</th>
<th>Mixed-layer thickness</th>
<th>Monin-Obukhov length</th>
<th>Friction velocity $u^*$ (mm s$^{-1}$)</th>
<th>Variance $\sigma^2$ $[10^{-5}(^\circ C),^2]$</th>
<th>Skewness $S$</th>
<th>Kurtosis $K$</th>
<th>Transition wavenumber $k_*$ (cy/cm)</th>
<th>$\epsilon$ (cm$^2$ s$^{-3}$)</th>
<th>$\epsilon_0$ ($^\circ C$)$^2$ s$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>4C</td>
<td>34</td>
<td>3.72</td>
<td>8.5</td>
<td>-82</td>
<td>10</td>
<td>3.4</td>
<td>+.12</td>
<td>3.7</td>
<td>0.02</td>
<td>&gt;10$^{-5}$</td>
<td>3x10$^{-7}$ - 4x10$^{-7}$</td>
</tr>
<tr>
<td>5B</td>
<td>43</td>
<td>2.34</td>
<td>8.5</td>
<td>-82</td>
<td>10</td>
<td>0.1</td>
<td>-.48</td>
<td>6.4</td>
<td>0.02</td>
<td>&gt;10$^{-5}$</td>
<td>9x10$^{-9}$ - 2x10$^{-8}$</td>
</tr>
<tr>
<td>5C</td>
<td>40</td>
<td>3.72</td>
<td>8.5</td>
<td>-82</td>
<td>10</td>
<td>0.01</td>
<td>-.65</td>
<td>4.8</td>
<td>0.02</td>
<td>&gt;10$^{-5}$</td>
<td>7x10$^{-9}$ - 1x10$^{-8}$</td>
</tr>
<tr>
<td>11D</td>
<td>35</td>
<td>3.86</td>
<td>8.0</td>
<td>+37</td>
<td>10</td>
<td>0.6</td>
<td>+.10</td>
<td>8.7</td>
<td>0.02</td>
<td>&gt;10$^{-5}$</td>
<td>5x10$^{-9}$ - 7x10$^{-8}$</td>
</tr>
<tr>
<td>14</td>
<td>11</td>
<td>2.38</td>
<td>12.0</td>
<td>-162</td>
<td>11</td>
<td>0.04</td>
<td>+.18</td>
<td>2.7</td>
<td>0.04</td>
<td>&gt;10$^{-4}$</td>
<td>4x10$^{-9}$ - 1x10$^{-8}$</td>
</tr>
<tr>
<td>23A</td>
<td>15</td>
<td>1.6</td>
<td>4.0</td>
<td>-4</td>
<td>3</td>
<td>4.4</td>
<td>-.45</td>
<td>4.4</td>
<td>0.01</td>
<td>10$^{-2}$ - 10$^{-3}$</td>
<td>4x10$^{-7}$ - 3x10$^{-6}$</td>
</tr>
<tr>
<td>31A</td>
<td>14</td>
<td>1.5</td>
<td>8.0</td>
<td>-17</td>
<td>7</td>
<td>0.4</td>
<td>+.66</td>
<td>3.1</td>
<td>not resolved</td>
<td>&gt;3x10$^{-8}$</td>
<td></td>
</tr>
<tr>
<td>34A</td>
<td>15</td>
<td>1.5</td>
<td>8.0</td>
<td>+4</td>
<td>7</td>
<td>5</td>
<td>-.30</td>
<td>5.2</td>
<td>0.12</td>
<td>10$^{-2}$ - 10$^{-3}$</td>
<td>4x10$^{-7}$ - 6x10$^{-6}$</td>
</tr>
<tr>
<td>15</td>
<td>1.5</td>
<td>8.0</td>
<td>+4</td>
<td>7</td>
<td>1.2</td>
<td>-.05</td>
<td>2.7</td>
<td>0.07</td>
<td>10$^{-3}$</td>
<td>9x10$^{-8}$ - 6x10$^{-7}$</td>
<td></td>
</tr>
</tbody>
</table>
The skewness values are not large and fluctuate in sign indicating that temperature fluctuations of either sign are equally probable. The mixed layers were generally not extremely well-mixed but appear to have been in a state of active stirring, without a mean gradient.

If the layers were turbulent, the spectra of temperature gradient $G_\theta(k)$ should exhibit the universal, one-dimensional form:

$$G_\theta(k) = \beta_1 \epsilon_\theta k^{-1/3} e^{-1/3} k + 1/3, \quad L_\sigma^{-1} < k < k_* \quad (1)$$

$$G_\theta(k) = q_\theta (\nu/\epsilon)^{1/2} k^{1/3}, \quad k_* < k < k_K < k_B \quad (2)$$

where $k$ is radian wavenumber, $\int_0^\infty G_\theta(k) \, dk = \frac{2}{3} \frac{\beta_1}{\beta_2} = \frac{2}{3} (\nu/\epsilon)^{1/2} k_\sigma$ is the rate at which the temperature fluctuations are smoothed by diffusion, $\kappa$ is the heat diffusivity, $\epsilon$ is the rate at which kinetic energy is dissipated by viscosity, $\nu$ is the kinematic viscosity, $k_B = (\epsilon/\nu^2)^{1/4}$ is the Batchelor wavenumber, $k_K = (\epsilon/\nu^3)^{1/4}$ is the Kolmogorov wavenumber, and $k_* = (\beta_1/q)^{3/2} k_K$ is the transition wavenumber between the inertial subrange [Eq. (1)] and the viscous-convective subrange [Eq. (2)]. For the inertial subrange to exist the Reynolds number must be large enough to allow a separation between the scale $L_\sigma$ of the energy-containing eddies and the smaller isotropic scales. In the viscous-convective subrange the temperature fluctuations are strained by a velocity field which itself is decaying due to viscosity. The local maximum of the gradient spectrum occurs at $k = k_B$; for $k > k_B$ the spectrum falls off exponentially as (thermal) diffusion smooths the smallest temperature fluctuations. The spectrum for $k > k_*$ is called the Batchelor spectrum (see Gibson and Schwarz, 1963; or
Grant, Hughes, Vogel, and Moilliet, 1968, for details). The values we have chosen for the "universal constants" are: $\beta_1 = 0.41$ and $q = 2$; thus $k_* = 0.1 k_K$.

The spectra of the gradient profile shown in Figure 8, observed after the wind had deepened the sml to about 8 m, show (Figure 9a) that the variance decreases slightly with depth in this particular example, and although the confidence intervals are quite large (a result of the small sample used) a +1 slope (the viscous-convective subrange slope) seems to describe about one decade in frequency.

The spectra are calculated using an FFT and a box-car data-window. This causes a sinc$^2$-type spectral window to be (automatically) applied to the spectral estimates and because this window has substantial side-lobes, leakage can occur if the true spectrum is red. Ordinarily no correction is made for this effect. The effects of the various forms of thermistor corrections and the extent of the leakage can be seen in Figure 9b. Here the spectra (of only the higher-variance section of Figure 8) uncorrected for thermistor or spectral window effects, is nearly flat for two decades and then begins to roll off at a wavelength of about 2 cm. Applying no extraneous spectral window and either the one-pole thermistor-power correction of Aagaard (1969) or the two-pole corrections used by Gregg, Cox, and Hacker (1973) causes the spectrum to rise at about a +1 slope. When a Hann spectral window—a filter with very small side lobes—is used in conjunction with the two-pole power response once again the +1 slope appears.

Is the +1 slope a viscous-convective subrange and is the intersection of the +1 slope with the flatter region at low wavenumbers the $k_*$ tran-...
Figure 8. Surface-layer microstructure (Y3, Drop 34A). Winds were from the north at 6 m/sec; there was a 7-sec, 5-foot swell. The fall speed of the instrument was 14.9 cm s\(^{-1}\).
Figure 9. (a) Temperature-gradient spectra of (●) the upper 1.5 m and (○) the lower 1.5 m of the mixed-layer record shown in Figure 8. The dashed lines indicate thermistor corrections applied to the least count noise. The confidence interval is 95%. (b) The gradient spectra of the upper 1.5 m with (----) no thermistor-attenuation corrections applied; (-----) corrected a la Gregg, Cox, and Hacker (1973); (○) corrected as before and filtered with a Hann spectral window; and (●) corrected a la Aagaard (1969). The heavy lines are universal spectral curves (see text).
sition? Shown in Figure 9b are two universal curves showing the inertial subrange and the Batchelor spectrum for $\varepsilon = 10^{-3}$ and for $\varepsilon = 10^{-2}$ calculated using the appropriate Prandtl number ($\text{Pr} \equiv \nu/\kappa$ of 9.3). The universal spectra were moved vertically so as to envelope the corrected data points assuming the $+1$ region is a viscous-convective subrange and at least part of the lower-wavenumber region is an inertial subrange.

In the positions shown the curves correspond to the following estimates of the dissipation: $10^{-3} < \varepsilon < 10^{-2}$ cm$^2$/sec$^3$ and from Eq. (1) or Eq. (2), $5.5 \times 10^{-6} < \varepsilon_0 < 5.9 \times 10^{-6}$ (°C)$^2$/sec. The estimate of $\varepsilon_0$ can be compared with a lower bound estimate obtained directly from the variance of the signal as recorded and digitized: $\varepsilon_0$ (lower bound) = (2) (3) ($\kappa$) ($\sigma^2$) = $4.2 \times 10^{-7}$ (°C)$^2$/sec, where the factor of 3 arises because it is assumed $(\nabla \tau)^2 = 3(\partial T/\partial z)^2$, a reasonable assumption in a patch of isotropic turbulent fluid. This lower bound is shown in Table 1 along with the estimates obtained by fitting the universal curves to the observed spectra.

Just how representative are the estimates of $\varepsilon$ and $\varepsilon_0$ observed in the sme? Our observations (Table 1) show $10^{-5} \leq \varepsilon \leq 10^{-2}$ and $5 \times 10^{-9}$ < $\varepsilon_0$ < $5 \times 10^{-6}$ with little correlation with $u_*$ or $L_b$. (The lack of correlation with larger-scale conditions is probably not real, but due to the infrequency and relatively poor resolution of the sampling.) Recent work in a fully-turbulent 70-m thick mixed-layer (Gregg, 1976b) shows similar values of $\varepsilon$ but very much lower values of $\varepsilon_0$: $10^{-4} \leq \varepsilon \leq 3.3 \times 10^{-3}$ cm$^2$ s$^{-3}$, and $1.5 \times 10^{-10} \leq \varepsilon_0 \leq 1.9 \times 10^{-8}$ (°C)$^2$ s$^{-1}$. The reason our $\varepsilon_0$ are larger than Gregg’s may be that the source of temperature fluctuations (either the surface or the sme transition) was much closer--and
much closer—and in the case of the transition, much stronger—in our work. In the case of $\varepsilon$, both Gregg's observations and ours occurred under similar wind-stress conditions ($u_* = 1 \text{ cm s}^{-1}$) but the energy available for mixing had to be distributed over a factor-of-ten thicker sml, thus one would expect our $\varepsilon$ to be an order-of-magnitude larger. For MSP 34A, one of the better resolved profiles, $\varepsilon = 10^{-2}$ and this is indeed the case. Also, in our work the base of the mixed layer was much closer to the surface where the local sea could directly stir and mix the layer.

Mixed-layer deepening

A short period of calm conditions on the 10th was followed by an impulsive upwelling-favorable wind event extending through the 11th. Measurements made during this time up until we left station to return home show that the surface layer quadrupled its thickness in relatively short order (Figure 10). The theory of Pollard, Rhines, and Thompson (1973) predicts a deepening within one inertial period to a depth $h_{\text{max}} = \frac{2u_*}{\sqrt{Nf}}$ where $N$ is the buoyancy (Brunt-Väisälä) frequency of the seasonal thermocline into which the sml advances and from which it entrains saltier colder fluid (in this case). Although they did not have such shallow mixed layers in mind—their theory entirely neglects surface wave influences—it is interesting that $h_{\text{max}}$ is a very good estimate of the observed mixed layer depth. Before the deepening event $L_b = 0 \text{ m}$ and neither wind shear nor buoyant convection processes were active. Soon after the onset of the winds ($t = 13 \text{ hr}$ on Figure 10) $L_b$ decreased to $-15 \text{ m}$ so convection may have played a role in the deepening process.
Figure 10. Log-log plot of surface mixed-layer thickness (h) against time during an upwelling-favorable wind event. Hour 1 corresponds to 0100 PDT Sept. 19, 1974. The north winds commence at 1300 and reach 9 m sec\(^{-1}\) by 2300. Depths shown were measured by (o) the CTD and (●) MSP. The prediction of \(h_{\text{max}}\) by Pollard, Rhines, and Thompson (1973) is shown. The bar extends one inertial period.
At \( t \approx 30 \text{ hr}, L_b + 4 \text{ m} \) and the buoyant forces apparently opposed any further deepening. At the end of our second cruise a similar deepening to again 8-9 m was seen. Under similar wind conditions and in the same water depth, Reed and Halpern (1975) observed the bottom of the mixed layer to be just below 8 m.

Is the deepening due to a local mixing or is advection important also? A local deepening would conserve the integrated (between 0 and 15 m say) salt content before and after the mixing. This is partially true in this example (Figure 11), but once the well-mixed surface layer reaches an apparent equilibrium depth its salinity continues to increase at a rate two orders-of-magnitude faster than evaporative losses would explain. The explanation is the replacement of surface water by an offshore transport of newly upwelled saltier water as predicted by the Ekman-Thorade model of coastal upwelling as discussed by Halpern (1974), a simple horizontal balance giving a rate of increase of the correct order: 0.05 °/oo h⁻¹.

The change in the temperature finestructure caused by the deepening event can be seen in Figure 12. During relatively calm conditions temperature profiles show a very thin surface layer (Figure 12a). The interface (strong gradient) in 23A at about 12 m is not the base of the (2-m thick) surface layer but is a remnant of previous conditions. It persists in profiles 23 B, C, and D (not shown) but is being eroded by a mixing localized to within a few meters above or below. Fifteen hours later 26A shows a thinner and warmer surface layer, and the interface at 12 m is weakened. Both surface layers appear to have at their bases weak interfaces with \( \Delta T \approx 0.1 \text{ C} \) or less. This is to be compared with \( \Delta T \geq 1 \text{ C} \) for 11 A-F (Figure 7) and 35 A-L (Figure 12b) during an order-
Figure 11. Salinity profiles (light lines) before the deepening, 0–10 hr on Figure 10, and (heavy lines) after the deepening, 35–37 hr.
Figure 12. (a) Two gross temperature profiles during relatively calm conditions (Y3, Drops 23A and 26A). The absolute temperature scale applies to 23A; 26A is offset by 2°C.
(b) Sequence of gross temperature profiles through the surface mixed-layer and seasonal thermocline (Y3, Drops 35A-L). The absolute temperature scale applies only to Drop A. The other profiles are offset by 1°C. The time between drops is 3-4 minutes. Ship drift over the sequence was 1.5 km due south. The profiles have been aligned by fixing the 10.3°C isotherm at 20m. The dashed lines are interpolations, made necessary by chart recorder offset adjustments.
of-magnitude higher wind stress. In Figure 12b twelve "yo-yos", just a few minutes apart in time but about 100 m in space, show an actively mixed surface layer 5 to 10 m thick with strong evidence of entrainment. It should be noted that the first six profiles (A, B, ...F) show sml temperature fluctuations of about the same size or smaller than those in 11 A-F. The large entrainment events are not common. Pollard (1973) encountered rapid variability, similar to that observed here, in a 5-m zone above the base of a 40-m deep layer during 10-15 m s⁻¹ winds. He attributed the fluctuations to turbulent eddies or breaking internal (interfacial) waves.

Probably the easiest way to estimate energy dissipation in the mixed layer is to use the inviscid estimate \( \varepsilon \sim u^3/\ell \) where \( u = u_\ast \) is a characteristic turbulent velocity and \( \ell = D \) is the size of the largest eddies. Thus, \( \varepsilon \sim (1 \text{ cm s}^{-1})^3/10^3 \text{ cm} = 10^{-3} \text{ cm}^2/\text{sec}^3 \), a perfectly reasonable value. The entrainment events can provide an estimate of \( \varepsilon \) if it is assumed that local kinetic energy was used to do work and lift colder, saltier blobs from the base of the sml into the layer. The potential energy increase is approximately \( (\Delta \rho)gh \sim (1.8 \times 10^{-4} \text{ gm/cm}^3)(g)(2 \text{ m}) = 35 \text{ ergs/cm}^3 \) of water swept up. If large entraining events occur every 20 min (Figure 12b), \( \varepsilon \sim 3 \times 10^{-2} \text{ cm}^2/\text{sec}^3 \), in rough agreement with the spectral estimate for Y3-34A.

A striking similarity exists between the event in 35H (Figure 12b) and some of the results of a study by Thorpe (1973) of Kelvin-Helmholtz instabilities. Based on a formula he developed by Thorpe, the mean \( \varepsilon \) during the transition of the event from its initial to final flow (estimated to take about 30 min) is about 0.015 cm² s⁻³, in agreement with the other estimates.
An examination of a possible balance for thermal and kinetic energy can be made with the help of the flux-gradient relationships used by atmospheric boundary-layer dynamicists (see Dyer, 1974). For Y3-34A:

$L_B = +4.5 \text{ m} ; Z - D = 8 \text{ m} ; u - w - u_ = 0.7 \text{ cm/sec} ; \text{ and } \theta - \theta_0 = (\text{surface heat flux})/\rho C_p u_0 = 0.04^\circ \text{ C}, \text{ where } k = 0.41. \text{ A constant-flux, turbulent shear-flow under steady, horizontally homogeneous conditions is assumed. Then the functions } \phi_H \text{ and } \phi_M = 1 + 5 \left( \frac{Z}{L_B} \right) = 10; \frac{36}{3Z} = \theta_0 \phi_H / Z = 4 \times 10^{-4} ^\circ \text{C/cm}; \text{ and } \frac{3U}{3Z} = u_0 \phi_M / k_0 Z = 0.02 \text{ sec}^{-1}. \text{ Balancing production against dissipation in the mixed layer we have (Z positive upwards):}

\[
-w\theta \frac{\partial \theta}{\partial Z} = \frac{\varepsilon + a \theta}{2}, \quad (3)
\]

\[
[-(-1 \times 10^{-5}) \neq (3 \times 10^{-6})], \text{ and}
\]

\[
-w^u \frac{\partial U}{\partial Z} + \alpha g w\theta = \varepsilon \quad (4)
\]

\[
[-(-10^{-2}) + (-5 \times 10^{-3}) \neq (6 \times 10^{-3})]
\]

Though possibly just coincidental, there is an approximate balance in (4); but it seems we have over-simplified the temperature equation. Eddy diffusivities of heat and momentum can also be calculated:

$K_H = K_M = k_0 u_* Z/\phi_H = 23 \text{ cm}^2 \text{ sec}^{-1}; \text{ and the gradient Richardson number}$

$Ri = \left( \frac{Z}{L_B} \right) \phi_H / \phi_M^2 = 0.2. \text{ It is interesting to compare these numbers with Halpern's (1976) observation of a 15-m thick wind-drift layer in the summer of 1973 at the same location as the present work (but under higher wind-stress) for which he obtained } K_H = 55 \text{ cm}^2 \text{ sec}^{-1} \text{ and } Ri = 0.2 \text{ to } 0.4. \text{ An additional estimate of } K_H \text{ follows from applying (eddy) diffusion
theory to the "smoothing" of the base of the sml in Y3-11E (Figure 7). A value \( K_h = 0(50 \text{ cm}^2/\text{sec}) \) is obtained.

Cool ribbon

Huyer and Smith (1974) have seen a sub-surface ribbon of relatively cool water, probably the result of southward advection of subarctic water by the alongshore coastal jet associated with upwelling. They give the following characteristics of the cool ribbon: (1) it is several tenths of a degree cooler than the water below it; (2) salinities are in the range 32.5 to 33.0 °/oo, mostly between 32.6 and 32.8°/oo, and can be higher (near 33.6 °/oo along the shoreward edge); (3) the ribbon can be shallower than 20 m; and (4) it is associated with southerly surface flow, its location often being coincidental with the axis of the coastal jet. The ribbon can be present during active upwelling or during the relaxation phase after a period of north winds as we have here (Wang, 1976).

We observed the cool ribbon (Figure 3) at a depth of about 20 m on September 7 and 8, at which time the flow may have had a southward component (Figure 4). The T-S properties of the ribbon (Figure 13) agree with the criteria of Huyer and Smith. The T-S characteristics of the water above and below the ribbon are fairly constant whether the ribbon is present (CTD 7-15) or absent (CTD 18-25).

Temperature contours and gradient profiles from the MSP (Figure 14) indicate that the ribbon has a "core", bounded by the 7.9 C isotherm, in which there is little temperature microstructure relative to the core boundaries. (Note that for purposes of display the ribbon's "shape" in
Figure 13. T-S curves during (CTD 7-15), and after (CTD 18-25) the presence of the cool ribbon. Isograms of sigma-t are also shown.
Figure 14. (Top) Cool ribbon temperature contours based on MSP profiles. (Bottom) Temperature gradient profiles. Times are approximate.
Figure 14 is different from Figure 3; also there is a 0.1°C consistent offset between CTD and MSP absolute temperature.) Only one CTD profile showed the core to be well mixed in both T and S: \(\overline{dT/dZ} \leq 2 \times 10^{-5}\) °C/cm and \(\overline{dS/dZ} \leq 2 \times 10^{-5}\) °C/cm over a 4-m thick core. More often the mean gradients were an order-of-magnitude greater. There was no evidence of entrainment of warmer water into the core on the gross T profiles although there was some microstructure and evidence of weak convection in the core (Figures 7, 14, and 16).

Using a value for the mean shear of \(7 \times 10^{-3}\) s\(^{-1}\) (based on a surface wind-drift current and the 25-m current) the following macroscopic Richardson numbers were calculated: above the core, \(7 < Ri < 30\); within, \(Ri > 0.4\); and below, \(8 < Ri < 20\).

Spectra were not computed for data from within the core because of the extremely low signal levels. In the thermocline, the following maximum values were computed: \(\varepsilon_{\text{s}} \approx 5 \times 10^{-3}\) cm\(^2\) s\(^{-3}\), \(\varepsilon_{\text{c}} \approx 2.8 \times 10^{-5}\) (°C)\(^2\) s\(^{-1}\), and \(K_H \approx 0.5\) cm\(^2\) s\(^{-1}\); but immediately above the core the microstructure levels are much weaker. Below the core, in the inversion: \(\varepsilon_{\text{s}} \approx 10^{-5}\) cm\(^2\) s\(^{-3}\), \(\varepsilon_{\text{c}} \approx 3 \times 10^{-7}\) (°C)\(^2\) s\(^{-1}\), and \(K_H \approx 0.05\) cm\(^2\) s\(^{-1}\). Large fluctuations in the microstructure levels do occur (Fig. 14) but an average eddy diffusivity for the boundary of the ribbon may be about 0.1 cm\(^2\) s\(^{-1}\).

Heat budget

A simple vertical model of the heat and salt transports into the ribbon is proposed in which the ribbon is heated both from above and
below while salt is added from below and lost out the top. Thus, the salt flux tends to stratify an initially well-mixed core, tending to make the salt profile linear while the heat flux tends to warm the core symmetrically from above and below, weakening the intensity of the core while only slowly changing its distinctive signature on the temperature profile. For the September 1974 ribbon the salt gradient was about 50% larger below the core than above; the temperature gradients were more nearly the same. The salt gradient within the core was never larger than that above or below and usually was a factor-of-two smaller; once, near the beginning of the 24-hour observation of the ribbon, the core was well-mixed in salinity. If the core was restratified from an initially well-mixed condition to the observed mean salinity gradient over this period of time by small-scale mixing processes which can be parameterized by an eddy diffusivity $K_H$, $K_H$ must have been about 0.1 cm$^2$/sec. Using this value the net heat flux into the core (of 3-m average thickness) was $6 \times 10^{-5}$ cal/cm$^2$ sec (from above; this yields a temperature increase of about $0.02\pm 0.1^\circ$ C in one day. Similarly, a net salt flux into the core (from below) of about $2 \times 10^{-8}$ gm/cm$^2$ sec. gives a salinity increase of about $0.005\pm 0.005$ o/oo. These increases are roughly those observed (Figure 15).

Historical data (Reed and Halpern, 1973; Holbrook, 1975) show the cool ribbon to be of large horizontal extent. Observed off the Washington coast in September and October of 1972 and 1973 at a mean depth of 40-100 m, it had a thickness of 10-15 m over the shelf increasing seaward, and $S \sim 32.9$ (about the same as observed here) and $T \sim 7.2$ C. Assuming our model to apply continuously as the ribbon moves southward from above 48°N to 45°N (a range of 355 km) at 10 cm/sec (a reasonable value for September,
Figure 15. Changes in the (o) CTD temperature and (●) salinity of the cool-ribbon core during the observation period. The correlation coefficient $r$ describes the goodness of linear regression.
see Huyer, 1976) with a mean thickness ~5 m and a net inwards heat flux of the observed value, a value $K_H \sim 0.06 \text{ cm}^2/\text{sec}$ is required to account for the temperature increase of 0.8 C, very close to the previous estimate of 0.1 cm$^2$/sec. (It is quite possible that on average there is no net salt flux into the core, explaining why the salinity remains about the same.) The value 0.1 cm$^2$/sec, two orders-of-magnitude larger than the molecular diffusivity, is evidence of a milder mixing than that occurring in the surface layer where $K_H$ was estimated to be two orders-of-magnitude greater.

Temperature inversion

Below the cool ribbon is a large but stable temperature inversion. The inversion can be caused by the ribbon alone or by a simultaneous intrusion of anomalously warm water (the so-called warm temperature anomaly) below the ribbon. The observed profiles (an example is shown in Figure 16) are very similar in appearance to those observed by Huyer and Smith at a time when both features, the ribbon and the anomaly, were present (cast 44 m in Anonymous, 1972). However, the signature of the anomaly on the T-S diagram that they observe (Figure 9 in Huyer and Smith, 1974) does indeed appear anomalous, while on our T-S diagram (Figure 13) what looks like anomalously warm water when the cool ribbon is present appears to be only part of the normal stratification when the ribbon is absent. [The peculiar double inversion (Figure 16) and a more probable example of the warm anomaly will be discussed in the next section.]
Figure 16. Temperature profiles through the cool ribbon and the underlying warmer water. CTD 3 (unprocessed for calibration corrections and pressure effects) and MSP 4A were made on station in 100 m of water; the ship drifted 300 m to the NE when MSP 4B was made, and another 300 m for MSP 4C. Notice the horizontal preservation of the small spot of cold water in the core and some of the step-like features in the water below.
Can the temperature inversion region, where warm salty water underlies cooler, fresher water, support (diffusive) double-diffusive convection? If $\Delta T$ and $\Delta S$ are the changes across the inversion then the stability number $R_p = \beta \Delta S / \alpha \Delta T$, where $\beta = \frac{1}{\rho} \frac{\partial \Delta \rho}{\partial \Delta S}$ and $\alpha = \frac{1}{\rho} \frac{\partial \Delta \rho}{\partial \Delta T}$, is seen to be in the correct range (Figure 17) for diffusive-type layered convection. On the temperature profile (Figure 16) there appear to be a few layers on the inversion. Closer examination, (see Chapter IV) however, shows that these are not consistent from profile to profile, and that probably shear-induced mixing processes dominate over the double-diffusive mechanism.

Warm anomaly

Huyer and Smith (1974) describe the formation of the warm temperature anomaly this way: "cold saline water is brought to the surface by intense upwelling, is then subject to solar heating, and is carried offshore by the Ekman transport. When it approaches lighter surface water (usually lighter because of lower salinity) it tends to sink and to be mixed with the surrounding water. The warm temperature anomaly results from the onshore-offshore component of the flow. The resulting warm tongue does not appear to extend seaward of the axis of the cool ribbon." The anomalously warm water generally has salinities between 33.0 and 33.5 $^\circ/_{oo}$, the anomaly can exceed 1 C, and occurs above the 26.0 $\sigma_T$ surface, and is often associated with high turbidity. The anomaly is a recurring summer feature and sometimes occurs in winter (Huyer, 1976). However, the temperature inversion associated with the anomaly need not
be present during active upwelling nor is it a summertime climatological-mean condition (Halpern, 1976).

Such an anomalous mass of water, was found on the second cruise soon after the end of a strong upwelling event (see Figure 2) and observed for over a day. A current meter at 25 m showed the current to be offshore and to the south as expected. The T-S curves (Figure 18) from CTD casts 1.5-hr apart show the anomalously warm water but at a lower salinity and sigma-t than expected.

Finestructure

Temperature profiles through the anomaly (Figure 19) often show a peculiar double inversion (second "mode") structure (see also Figure 16). The second mode appears in most of the MSP profiles but only in a few of the CTD casts, possibly because of the lower vertical resolution of the latter. When present in the CTD temperature profile there is a compensating second mode in the salinity profile. [Turbidity profiles have shown a midwater turbid layer with a second mode appearance (see Figure 3 in Hartlett and Kulm, 1973).] Higher modes are also seen occasionally. Cross-shelf hydrographic sections in the summer, 1972, showed the warm anomaly to have a complex temperature structure (Huyer, 1973, p. 35 and 57).

It is interesting to speculate about the origin of the modal structure. Experiments with sinking "jets" discharged horizontally into a cross-flowing stream (Ayoub, 1973) have shown the jet to form twin vortices which give a second mode structure when sampled vertically, with
Figure 17. Plot of $\Delta T$ vs. $\Delta S$ across the inversion at 28 m during Y2 and at 25 m (below the cool ribbon) during Y3. Using the values $\alpha = 0.00015 \, (^{\circ}C)^{-1}$ and $\beta = 0.00076 \, (^{\circ}/o)^{-1}$ the lines $R_\rho = 2, 4, \text{ and } 15$ have been drawn. The small numbers are CTD cast numbers. There does appear to be some trend with time for the Y3 observations.
Figure 18. T-S diagram for CTD casts 30 (1237 PDT, 15 Sept. 1974) and 31 (1412 PDT). Two sigma-t lines are indicated.
Figure 19. (Top) Temperature profiles through the warm-water anomaly (Y2, Drops 17A-L). The repetition time was 6 minutes and the ship drifted 0.7 km to the WSW while the data were being taken: thus, the drift between samples was about 50 m. (Bottom) Gradient records and the 8.5°C contour.
the scalar concentration about 50% higher in the vortices than in the core axis (c.f. Figure 19, A-F). These observations may be relevant to the case of sinking, anomalously warm water, moving westward in to a faster southward current if the two flows are dynamically similar. The important non-dimensional groups are a velocity ratio $K = \text{jet velocity}/\text{ambient cross-stream velocity}$, and a densimetric Froude number $Fr$. Unfortunately, estimates of $K$ and $Fr$ for the oceanic case are outside the range experimentally studied and this mechanism remains only a possible explanation.

Again, the temperature inversion (only the first, more persistent inversion, at about 28 m, was examined) has stability numbers in the correct range for double-diffusive convection (Figure 17) but no convincing example of layered structure caused or maintained by that mechanism has been found. A detailed examination of the gradients in the inversion (Caldwell, et al., 1975, Figures 9-11) has shown some to persist horizontally while others do not, implying a weak stirring to be present.

The shear based on measurements at 25 and 75 m (the 50-m current meter malfunctioned) was 0.008 s$^{-1}$. The shear through the anomaly (27-37 m) could very well have been larger than this. Estimates of the overall Richardson number vary between one and ten. This anomaly does not seem to be as subject to dynamic instability as the anomaly seen by Johnson, Van Leer, and Mooers (1976) which occurred in a zone for which the instantaneous Richardson number was less than one for 80% of the measurements. Their measurements were made during an upwelling event when strong near-inertial motions were mainly responsible for the
shear; such was not the case for our measurements.

As previously, a vertical heat budget can be performed on the warm water mass. Assuming the observed temperature decrease of about 0.5 °C in one day (Figure 20) to be caused by transport vertically out of the anomaly via "eddy diffusion" along the observed mean gradients, \( K_v \) turns out to be 0.5 \( \text{cm}^2 \text{s}^{-1} \). Using this value, a maximum change of only 0.03 °/°o would be expected in salinity.

Microstructure

In order to analyze the MSP records the anomaly was subdivided into an "above" region (the first inversion at about 28 m), a "pre-mixed" region (the water cooler than 8.5 °C at about 30 m on A-F on Figure 19), changing to a "mixed" region (G-L at about 30 m), and a "below" region. The gradient records from these regions were digitally low-pass filtered to suppress leakage effects, then used to compute spectra (which were adjusted for the filtering). Despite this precaution the spectra (Fig. 21) show a predominance of finestructure (wavelengths larger than 10 cm). The low wavenumber temperature signal especially evident on A-F with a wavelength of about three meters \( (k_o=0.0033 \text{cy cm}^{-1}) \) leaves its imprint on the gradient records producing a high estimate of gradient power in a

\( ^3 \text{From the measured ship drift (16 cm s}^{-1} \text{SWS) and 25-m current meter (25 cm s}^{-1} \text{SW) it would appear the profiles sample the anomaly closer and closer to its presumed source region nearer shore. On the other hand, the profiles seem to indicate a downstream development with the anomaly more uniform in the later profiles. Accepting for the time being the latter evidence, the interior region, at first labeled "pre-mixed", changes to a "mixed" region.} \)
Figure 20. Plot of temperature, salinity, and depth ranges of the anomaly vs. time.
Figure 21. Ensemble-averaged temperature gradient spectra from the warm-water anomaly. The dashed line shows the thermistor attenuation correction, and the bars show ± one standard deviation.
broad band of wavenumbers $k < k_0$, effectively masking any inertial subrange. In the "above" and "premixed" spectra a small viscous-convective subrange (+1 slope) may be seen in the range $0.2 < k < 0.4$ and $0.1 < k < 0.25$ cm$^{-1}$ respectively. All the spectra show a strong roll-off of gradient signal prior to an upward slope due to digitizer noise. Fitting the universal curve to the local maximum and roll-off, estimates of $\varepsilon$ can be made; these are shown in Table 2 along with the statistics and a production-dissipation balance estimate of $K_H$. Using the calculated $K_H$ and observed temperature gradient the maximum vertical heat flux from the anomaly is $\leq 4.4 \times 10^{-4}$ cal cm$^{-2}$ s$^{-1}$ which would, over the 1-hour period of the observations, lower the mean anomaly temperature 0.003°C, an amount not resolved by the measurements.

Why is there a factor-of-ten difference in $K_H$ between the 24-hour CTD measurements and the 1-hour MSP measurements? Most probably we may be hoping for too much agreement between rather limited observations from a very energetic environment in which the terms in vertical balance equations may be overwhelmed by advective effects. For example, the time rate-of-change of the perturbation temperature variance, $\bar{\varepsilon}(\theta^2)/\bar{\varepsilon}t$, computed from integrating the temperature spectrum for wavelengths less than 10 cm is $0[-10^{-8} (^\circ C)^2 s^{-1}]$ through the presumed transition from the pre-mixed to mixed interior. How then could mixing—which increases $\theta^2$—have occurred? The change in the interior shape of the anomaly must be due to our having sampled different water. The conclusion from the brief MSP sampling is that the anomaly interior is very weakly, intermittently mixed.
Table 2. Summary of parameters in the warm anomaly.

<table>
<thead>
<tr>
<th>Region</th>
<th>$\epsilon$ (cm$^2$s$^{-3}$)</th>
<th>$\epsilon_0$ ($^\circ$C$^2$s$^{-1}$)</th>
<th>$K_H$ (cm$^2$s$^{-1}$)</th>
<th>$\sigma^2$ ($^\circ$C/cm)$^2$</th>
<th>Skewness</th>
<th>Kurtosis</th>
</tr>
</thead>
<tbody>
<tr>
<td>&quot;above&quot;</td>
<td>$0(10^{-6})$</td>
<td>$6\times10^{-7}$ to $2\times10^{-6}$</td>
<td>$.02$ to $.05$</td>
<td>$7.3\times10^{-5}$</td>
<td>$2.4+1$</td>
<td>$10+9$</td>
</tr>
<tr>
<td>&quot;pre-mixed&quot;</td>
<td>$0(10^{-6})$</td>
<td>$7\times10^{-8}$ to $2\times10^{-7}$</td>
<td>$.003$ to $.02$</td>
<td>$1.4\times10^{-5}$</td>
<td>$1.6+1$</td>
<td>$8+5$</td>
</tr>
<tr>
<td>&quot;mixed&quot;</td>
<td>$&lt;10^{-6}$</td>
<td>$3\times10^{-10}$ to $1\times10^{-9}$</td>
<td>$.004$ to $.01$</td>
<td>$1.1\times10^{-7}$</td>
<td>$0.2+2$</td>
<td>$7+8$</td>
</tr>
<tr>
<td>&quot;below&quot;</td>
<td>$0(10^{-6})$</td>
<td>$5\times10^{-8}$ to $2\times10^{-7}$</td>
<td>$.0003$ to $.02$</td>
<td>$1.0\times10^{-5}$</td>
<td>$-1.4+5$</td>
<td>$4+4$</td>
</tr>
</tbody>
</table>
Deep Inversion

Some relatively thick but weak temperature inversions appeared in the water column below 40 m on the 10th and 11th of September (see Figure 2). Temperature and salinity profiles from CTD 42 (Figure 22) show a 10-m thick temperature inversion with an overall stability number of 2.5. The water below the inversion is isothermal but shows a small stable salinity gradient. The appearance of the temperature profile could lead one to speculate that some abnormally cool water is sliding over some unusually warm water. [The current meters at 50, 75, and 90 m show a broad northward flow, decreasing with depth (see Figure 4), with a mean shear of \((3-7) \times 10^{-3} \text{ sec}^{-1}\).

The T-S diagram for CTD 42 (Figure 23a) shows the inversion to be weakly stratified with a hint of some instability (caused by salinity inversions of the order of the CTD least count noise). Also, the T-S curve is in an "transitional state" between two more "stable states." A relatively warm state (the upper group of T-S curves in Figure 23b) exists prior to CTD 42, while a cooler state exists afterwards (the lower group in Figure 23b). In addition, there is a mass of anomalously warm water present during CTD 42 (and CTD 43) at \(S -33.66^\circ/\). This is the warm isothermal mass below the inversion in Figure 22.

Immediately after CTD cast 42 was made a sequence of MSP yo-yos (MSP 25 A-G), an average of 10-min apart, was made (Figure 24). The temperature inversion zone is quite active. The region below, an isothermal "core" bounded by the 8.10 C isotherm, shows little activity by comparison. As in the case of the cool ribbon, the core is bounded
Figure 22. Temperature and salinity profiles from CTD 42 (PDT 1047, 10 Sept.). The stability number $R_p = 2.5$ over the 10-m depth range indicated.
Figure 23. (a) T-S diagram for CTD 42 (PDT 1047, 10 Sept.) and CTD 43 (PDT 1437).
(b) T-S diagram showing the warm state (CTD 33-40) and the cooler state (CTD 44-51).
Figure 24. (Top) Temperature and (Bottom) temperature gradient profiles through a deep inversion with several isotherms indicated (1117-1218 PDT, 10 Sept.). The mean current at 50 m and the ship drift speed were 27 cm s⁻¹, both the the NE.
Figure 25. Histograms of 1-m averaged gradients from gross temperature profiles through the relatively isothermal core. The expected adiabatic gradient and the least-count gradient due to digitizer noise are indicated.
above and below by a relatively intense thermal activity. While it itself shows no entrainment of water, the boundary zones are rich in entrainment events. A rather crude overall Richardson number (Ri) calculation shows that above the core $0.4 < Ri < 4.0$, within $Ri = 1 + 0.7$, and below $Ri < 0.2$ during the time of MSP 25 and CTD 42.

If the core were very well-mixed it would have an adiabatic vertical temperature gradient (about $10^{-4} \, ^\circ C/m$). Although some parts of the core's temperature profile had about the right sign and magnitude to be adiabatic, the most common magnitude was about four times too large and of either sign (Figure 25).

The power in the gradient varies by more than three decades above, below, and in the core (Figure 26). For the region above the core, a good fit to the universal curve is obtained. With $k_x = 0.022 \, cy \, cm^{-1}$, $\varepsilon = 10^{-5} \, cm^2 \, sec^{-3}$ and $\varepsilon_\theta = 1.6 \times 10^{-7} \, (^\circ C)^2 \, sec^{-1}$. Applying the production-dissipation balance, $0.7 < K_v < 2.2 \, cm^2 \, sec^{-1}$. A convincing fit to the spectra from in and below the core was not obtained but from the statistical variance $\varepsilon_\theta > 2.7 \times 10^{-10} \, (^\circ C)^2 \, sec^{-1}$ within and $\varepsilon_\theta > 4 \times 10^{-9} \, (^\circ C)^2 \, sec^{-1}$ below the core. The large kurtosis values (Figure 27) indicate that the regions above and below the core could be turbulent; the skewness reflects the small but significant mean gradients. It is possible that some of the temperature fluctuations below represent fossil turbulence; this may explain why a viscous-convective subrange is not seen.

The sharp persistent interface bounding the core, the lack of entrained cold water within the core, and the statistics imply that the core is being eroded away at its boundaries by a relatively intense
Figure 26. Ensemble-averaged temperature gradient spectra above, within, and below the core computed from differentiated temperature records. The dashed portion shows the thermistor attenuation correction, and the bars give ± σ. The noise level is shown as "X's."
Figure 27. Plot of kurtosis and skewness (both are zero for a normal distribution) for gradient records above, within, and below the core.
stirring while the core itself is not active. As before we can look at an entraining interface—one of several in this case since each profile has at least one—from above the core to estimate $\varepsilon$: $\varepsilon = \Delta PE/\Delta t = (\beta \Delta S - \alpha \Delta T) \rho gh/\Delta t \approx 5 \times 10^{-4} \text{ cm}^2 \text{ sec}^{-1}$. Since this estimate is localized near the interface, about 10% of the total volume of the inversion zone, this is consistent with the spectral estimate of $10^{-5} \text{ cm}^2 \text{ sec}^{-1}$. (In Chapter IV it is shown that the interfaces probably become unstable via Kelvin-Helmholtz instabilities.)

CONCLUSIONS

Observations of the small-scale temperature structure in a 100-m deep water column off the Oregon coast, made amidst a background of CTD profiles and near a moored current-meter string, show that, in general:

(1) There is a characteristic finestructure and microstructure associated with the large-scale features which were advected past the observation site.

(2) In general, temperature "interfaces" pervade the water column. These have vertical gradients an order-of-magnitude larger than the mean gradient and a vertical extent of many centimeters.

(3) There can be a conspicuous absence of microstructure over several meters extent. Examples occur within the core of the cool ribbon, within the warm anomaly, and deeper in the water column.

(4) Temperature overturns on the order of 0.5 m vertical extent are common in regions of relatively strong and large-scale vertical gradient.
Fully turbulent patches with no mean gradient are not an uncommon occurrence in temperature profiles. More specifically:

(6) The surface mixed layer can deepen to about 8 m in an inertial period under a stress of about 0.5 dynes cm\(^{-2}\). When well stirred, the layer is bounded below by a strong gradient across which large entrainment events can occur. The average temperature-gradient fluctuations are about 0.003 °C cm\(^{-1}\) (r.m.s), and well-developed turbulence with inertial and viscous-convective subranges exists.

(7) The cool ribbon of water at about 20-m depth has a 3-m thick core which is relatively devoid of temperature microstructure. It is hypothesized that mixing within the core is infrequent and so it would be devoid of velocity microstructure (turbulence) as well.

(8) The temperature inversions found below the cool ribbon or as part of the deeper "warm water anomaly" have stability numbers, \(\beta \Delta S/\alpha \Delta T\), between 2 and 15, in the correct range for stable systems of double-diffusive convective layers to exist, but none are found.

(9) The warm anomaly tends to have a peculiar double-inversion structure. Temperature gradient spectral analysis yields surprisingly low estimates of mechanical energy dissipation.

(10) A deep inversion at about 60 m caused by relative motion of water masses at depth shows evidence of shear instabilities and isotropic turbulence. The scarcity of microstructure below the inversion is probably representative of the deep water in the complete absence of intrusions.
A summary of the rates of dissipation of energy and temperature variance and the vertical eddy diffusivity for heat (Table 3) shows high rates of energy dissipation and large eddy transport in the surface and bottom boundary layers.

Further microstructure work with a full complement of sensors—velocity, salinity, temperature, optical density, and of course pressure—should be undertaken with an aim to study the time-history of significant features, such as the ribbon and the warm anomaly. A need for microstructure measurements of the stability of the offshore flow has recently been expressed (Halpern, 1976). A small freely drifting submersible would be ideal for these studies. Vertical profiling while following a transponder embedded in the flow of interest would be another approach (see for example Voorhis, Webb, and Millard, 1976). As long as both small-scale structure and the larger-scale structure (including local circulation and meteorological forcing) are simultaneously obtained, the results of microstructure programs should be applicable to other areas where dynamically similar large-scale features are observed.
Table 3. Summary of the rate of dissipation of mechanical energy $\varepsilon$ and of temperature fluctuations $\varepsilon_\theta$, and the vertical eddy diffusivity $K_H$ in several regimes in the 100-m water column.

<table>
<thead>
<tr>
<th>Region</th>
<th>$\varepsilon$</th>
<th>$\varepsilon_\theta$</th>
<th>$K_H$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(cm$^2$ s$^{-3}$)</td>
<td>((°C)$^2$ s$^{-1}$)</td>
<td>cm$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>Well-stirred surface layer</td>
<td>$6 \times 10^{-3}$</td>
<td>$5 \times 10^{-6}$</td>
<td>50</td>
</tr>
<tr>
<td>Cool ribbon inversion</td>
<td>$10^{-5}$</td>
<td>$3 \times 10^{-7}$</td>
<td>0.1</td>
</tr>
<tr>
<td>Warm anomaly inversion</td>
<td>$10^{-6}$</td>
<td>$1 \times 10^{-6}$</td>
<td>0.05</td>
</tr>
<tr>
<td>Deep inversion</td>
<td>$10^{-5}$</td>
<td>$2 \times 10^{-7}$</td>
<td>1</td>
</tr>
<tr>
<td>Bottom layer</td>
<td>$10^{-4}$</td>
<td>$10^{-10}$</td>
<td>20</td>
</tr>
</tbody>
</table>

(from Caldwell, 1976)
III. DISSIPATION OF TEMPERATURE FLUCTUATIONS IN THE SEASONAL THERMOCLINE

The summer thermocline is a zone of very strong vertical density gradient in the ocean, in lakes, and in reservoirs and represents a barrier to the vertical transport of heat, salt, momentum and nutrients. At least part of the thermocline lies within the photic zone, and so the structure of the thermocline, often observed to be a conglomerate of "layers" of relatively large horizontal extent, may very well influence the spatial extent of plankton patches. Another reason for interest in this region is that if we can establish some facts about the small-scale structures, structures which have been shown to be relevant in the vertical mixing process (e.g., Woods and Wiley, 1972), and rates of dissipation of mechanical and thermal energy in the seasonal thermocline they may be applicable to other stratified parts of the water column.

In this paper we use observations of temperature finestructure and microstructure from the summer thermocline off the Oregon coast to estimate (by the computation of temperature gradient spectra) the dissipation rates and (by a production-dissipation balance) the vertical eddy diffusivity $K_H$. Gregg, Cox, and Hacker (1973), using vertical profiles of the temperature structure in the main thermocline in the North Pacific found $K_H$ to be very much smaller than the widely quoted value of 1 cm$^2$ s$^{-1}$. Dillon, Powell, and Myrup (1975), using a similar microstructure approach in the seasonal thermocline in Lake Tahoe calculated values of $K_H$ which were consistent with long-term estimates derived from heat budget calculations. The reason for (apparent) disagreement in the ocean may be one of data-sample representativeness. As Woods (1973) notes: "Every study
(of the seasonal thermocline) is limited to relatively narrow ranges of frequency and wavelength. These ranges constitute a spectral window through which the investigator views the thermocline. He is ignorant (though not necessarily unaware) of the motions whose spectral positions lie outside his window." Aware of this problem, we examine many profiles through the thermocline which are closely spaced in time and space, thereby obtaining statistical degrees of freedom; and use data one month apart to achieve at least some degree of representativeness.

Because of the strong density gradient, the seasonal thermocline is usually observed to have a large gradient Richardson number and hence, is dynamically stable. Thus, at any moment most of the thermocline is probably non-turbulent. It is only in small patches within the thermocline that (possibly isotropic) turbulence is found. Applying turbulence theory to temperature gradient records from these patches the dissipation rates will be found. By measuring the distribution of the patches in space and time, estimates of an effective time-averaged $K_H$ for the entire thermocline can be made. In addition, some comparisons with empirical laws relating $K_H$ to the buoyancy (Brunt-Väisälä) frequency are made.
Temperature gradient spectrum

In order to use the observed temperature microstructure (lengthscales smaller than about 10 cm) as an indicator of the velocity microstructure (turbulence) the following assumptions are made. Temperature is a passive property: the fluctuations are small enough so that the associated buoyancy is dynamically unimportant. The Reynolds number \((Re \equiv u \ell / \nu)\) and the Péclet number \((Pe \equiv u \ell / \kappa)\) are large enough for an equilibrium range to exist in the kinetic energy spectrum. Because the temperature is passive, there is an equilibrium range in its spectrum (in which temperature variance is transferred from low to higher wavenumbers with negligible molecular dissipation) which, if the turbulence is isotropic, consists of the convective and viscous-convective subranges. Most of the dissipation of the temperature variance occurs in the high wavenumber (lengthscales about 1 cm and smaller) diffusive subrange. Under these conditions, the one-dimensional temperature gradient spectrum should exhibit the universal form:

\[
\begin{align*}
G_\theta(k) &= \beta_1 \epsilon_\theta \epsilon^{-1/3} k^{1/3}, \quad L_0^{-1} < k < k_* \\
G_\theta(k) &= q \epsilon_\theta (\epsilon) \epsilon^{-1/2} k^{1/2}, \quad k_* < k < k_K < k_B
\end{align*}
\]

where \(k\) is radian wavenumber, \(\int_0^\infty G_\theta(k) \, dk = \left(\frac{3T}{2z}\right)^2\), \(\epsilon_\theta = 2k(\nabla T)^2\), is the rate at which the temperature fluctuations are smoothed by diffusion, \(\kappa\) is the heat diffusivity, \(\epsilon\) is the rate at which kinetic energy is dis-
sipated by viscosity, $\nu$ is the kinematic viscosity, $k_B = (\varepsilon/\nu^2)^{1/4}$ is the Batchelor wavenumber, $k_K = (\varepsilon/\nu^3)^{1/4}$ is the Kolmogoroff wavenumber, and $k* = (\beta_1/q)^{3/2}k_K$ is the transition wavenumber between the inertial subrange [Eq. (1)] and the viscous-convective subrange [Eq. (2)]. For the inertial subrange to exist, the Reynolds number must be large enough to allow a separation between the scale $L_o$ of the energy-containing eddies and the smaller isotropic scales. In the viscous-convective subrange the temperature fluctuations are strained by a velocity field which itself is decaying due to viscosity. The local maximum of the gradient spectrum occurs at $k = k_B$; for $k > k_B$ the spectrum falls off exponentially as (thermal) diffusion smooths the smallest temperature fluctuations. The spectrum for $k > k*$ is called the Batchelor spectrum (see Gibson and Schwarz, 1963; or Grant, Hughes, Vogel, and Moilliet, 1968, for details). The values we have chosen for the "universal constants" are: $\beta_1 = 0.41$ and $q = 2$; thus $k* = 0.1 k_K$.

For the smaller wavenumbers, $k \leq L_o^{-1}$, temperature (either the fluctuating or the mean part) can no longer be considered passive; and the turbulence, modified now by buoyant forces, becomes anisotropic, being

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4 The local maximum occurs in general at (a diffusive lengthscale)$^{-1}$ given by $k_d = (\gamma/\kappa)^{1/2}$ where $\gamma$ is an inverse time scale (for eddy overturning). Under turbulent conditions, $\gamma = (\varepsilon/\nu)^{1/2}$ is the effective rate of strain in which case $k_d = k_B$, the Batchelor wavenumber. If the flow is non-turbulent, $\gamma$ can be taken as about equal to $N$, the buoyancy frequency. In this case $k_d$ varies from 0.6 to 1.2 $\text{cm}^{-1}$ in the present work. Since no turbulence measurements were made we cannot tell if viscosity may have stopped the velocity fluctuations before diffusion destroyed the temperature fluctuations (the problem of the so-called fossil turbulence). Thus, there is no way to tell for certain if $k_d = k_B$ or $k_d = (N/\kappa)^{1/2}$.

5 There is evidence that the "constants" depend upon Reynolds number (see, for example, Williams and Paulson, 1976).
isotropic at most in the horizontal (really the plane perpendicular to the mean vertical density gradient). There then may exist a buoyancy range in the energy spectrum, in which case the temperature gradient spectrum decays as $k^{-1}$ (see Moseley and Del Balzo, 1976). The intersection of the buoyancy range ($k^{-1}$) with the convective range ($k^{1/3}$) occurs at about

$$k_0 = L_0^{-1} = (N^3/\varepsilon)^{1/2}$$

(3)

Estimating $\varepsilon$, $\varepsilon_\theta$, and $K_H$

Fitting the universal curve (Eqs. 1 and 2) to the observed spectra provides estimates of $\varepsilon$ and $\varepsilon_\theta$. If a $-1$ range is observed at low wave-numbers $\varepsilon$ can also be estimated by using Eq. (3). In a turbulent patch with a characteristic turbulent velocity and lengthscale $u$ and $l$ (the patch size), the inviscid estimate

$$\varepsilon \sim u^3/l$$

(4)

can be used.

To estimate $K_H$, the following assumptions are made to simplify the equation for the time rate-of-change of $\bar{\theta}^2$: (1) the mean vertical velocity $\bar{w}$ is zero; (2) conditions are horizontally homogeneous; (3) conditions are in a steady state; (4) $-\bar{w}\bar{\theta} = K_H \frac{\partial \bar{\theta}}{\partial z}$ (eddy diffusivity definition). Performing a vertical integration (to eliminate the vertical divergence term) there remains

$$K_H \frac{\partial \bar{\theta}}{\partial z} = \frac{\varepsilon_\theta}{2}$$

(5)

from which $K_H$ can be obtained. Equation (5) represents a balance between production from the stirring of the vertical gradient and dissipation of
temperature fluctuations, and has been used by previous workers (Gregg, Cox, and Hacker, 1973, for example).

The data to be discussed were obtained over a time interval during which the overall conditions $\frac{\partial \theta}{\partial z}$, $(\frac{\partial \theta}{\partial z})^2$, surface mixed layer depth, wind speed, etc. were not too variable, in accordance with assumption (3). Also, upwelling was not active, so $\bar{w}$ was presumably very small (assumption 1). The vertical integration is represented by considering (5) as a volume average, where the particular volumes used will be discussed later. Ensemble averaging is accomplished by averaging over many sequential profiles obtained in a short time interval.

To demonstrate the validity of assumption (2), one must show that the production of temperature variance by horizontal stirring is much smaller than that due to vertical processes, that is, show that

$$r \equiv \frac{K_M \left( \frac{\partial T}{\partial z} \right)^2}{K_R \left( \frac{\partial T}{\partial R} \right)^2}$$

is greater than unity. The distance $R$ is measured along a sigma-t (constant density) surface and $K_R$ is the eddy diffusivity along that surface. From the observed T-S properties and using an R-directed speed of 20 cm s$^{-1}$, $\frac{\partial T}{\partial R}$ is likely to be smaller than about $7 \times 10^{-6}$ °C cm$^{-1}$ while $\frac{\partial T}{\partial z}$ is about $10^{-2}$ °C cm$^{-1}$. Therefore for vertical production to dominate, $K_R < (2 \times 10^6) K_M$. Estimates of horizontal eddy coefficients off Oregon have been made in the surface layer (Reed and Halpern, 1975) for which $K_R \sim 3 \times 10^6$ cm$^2$ s$^{-1}$ and at 60 m (Mooers, Collins, and Smith,
1976) for which $10^5 < K_R^{(6)} < 10^7$ cm$^2$ s$^{-1}$. If one assumes that $K_R$ for the thermocline is much smaller than these values then assumption (2) is valid; if, on the other hand, $K_R > 2 \times 10^6$ then the calculated $K_H$ are in error, being too large by an unknown amount.

---

3Actually the horizontal eddy viscosity was measured.
OBSERVATIONS

The microstructure profiler (MSP), discussed in detail by Caldwell, Wilcox, and Matsler (1975), was used to sample the seasonal thermocline in a total water depth of 100 m on the continental shelf off the Oregon coast during three cruises in the summer of 1974. The data used in this paper are from the August-September cruises. The profiles (drops) are denoted by numbers; a sequence of rapidly repeated profiles ("yo-yos") are denoted by the drop number and a letter (see Table 4). Selected chart recordings of (gross) temperature and (high-gain) temperature derivative were digitized for analysis. Further discussion of cruise results and instrumentation can be found in Chapters II, IV, and V.

The conditions at the surface were not too different between the two cruises (Table 4). In each cruise the winds began to blow from the north during the measurements, but the data were obtained before the seasonal thermocline could be significantly altered by the horizontal and vertical advection associated with the ensuing coastal upwelling. The T-S diagrams (Figure 28) show the thermocline to be near maximum stability, and departures from the mean are fairly small.

A sequence of profiles (Figure 29) shows the general character of the seasonal thermocline. In this paper, the thermocline is defined as the zone of maximum temperature gradient (on average about 1°C m⁻¹). Bordering the thermocline is an "above" region extending into the surface mixed layer and a "below" region both with relatively smaller gradients. The stability of each of these regions is shown in Table 5. The eddy diffusivity will be evaluated in these regions. Beyond the change in hydro-
Table 4. Summary of microstructure drops and surface conditions.

<table>
<thead>
<tr>
<th>Date and drop</th>
<th>Time (PDT)</th>
<th>Wind speed (m s⁻¹)</th>
<th>Surface mixed-layer depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aug. 15, 1974</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>17 A-L</td>
<td>1257-1351</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>19 A-G</td>
<td>1825-1843</td>
<td>6</td>
<td>4</td>
</tr>
<tr>
<td>21 A-E</td>
<td>2122-2138</td>
<td>6</td>
<td>4</td>
</tr>
<tr>
<td>Sept. 11, 1974</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>31 A-C</td>
<td>0905-0926</td>
<td>4</td>
<td>7</td>
</tr>
<tr>
<td>32 A,B</td>
<td>1018-1032</td>
<td>4</td>
<td>7</td>
</tr>
</tbody>
</table>
Figure 28. T-S diagram for (top) the August cruise and (bottom) the September cruise. Isolines of sigma-t are shown.
Figure 29. Temperature profiles through the thermocline. MSP 21 A-E; $\overline{\text{Ri}} \sim 300$. The scale applies only to 21A, the other profiles are offset by 1.5 C. The time between profiles was 4 minutes. The ship drifted 1.4 km to the SSW while the data were being taken; thus, the distance between profiles could have been as much as 200 m. The dashed lines are interpolations, made necessary by chart recorder offset adjustments.
static stability it is physically reasonable to separate the above region as it is a buffer zone between the wind-stirred surface layer and the thermocline. During the day under calm conditions this region may vanish completely. Further justification for analyzing the below region separately comes from a belief by limnologists that it is more closely connected with the deeper water than with the physical processes occurring in the thermocline (Hutchinson, 1975, p. 428).

The ratio of the stabilizing effect of the salt gradient to that of the temperature gradient $G_o \equiv (\rho \frac{dS}{dz})/(\alpha \frac{dT}{dz})$ is greater than one in all three regions. Thus, both salt and heat contribute about equally to the density gradient. For the case $G_o \gg 1$ heat would play no dynamic role at all, but for the present work (see Table 5) it will at times be a poor assumption to say that temperature fluctuations are passive contaminants of the flow.

An important measurement is that of the gradient Richardson number $R_i$. (For $R_i$ greater than a critical value of about 1 turbulence is suppressed.) Unfortunately the best we can do is compute a macroscopic Richardson number $R_i \equiv N^2/S^2$ where $N$ is the buoyancy frequency and $S = \Delta U/\Delta Z$ is the mean shear. A nearby current-meter mooring provided the velocity at 25 m and a near-surface flow was computed from the surface winds. The mean shear is assumed to be constant with depth, but the variation in $N$ is taken into account. While not very much confidence can be attached to absolute values the relative values of $R_i$ may be meaningful. The calculations show (Table 5) that on the average the thermocline is very stable in a macroscopic sense whereas the transition regions above and below are less stable. A large value of $R_i$ does not
Table 5. Summary of large-scale observations just above, within, and just below the thermocline.

<table>
<thead>
<tr>
<th></th>
<th>( \frac{dT}{dz} )</th>
<th>( \frac{dS}{dz} )</th>
<th>Buoyancy Frequency N</th>
<th>Richardson Number Ri</th>
<th>Gp</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABOVE</td>
<td>( 2.4 \times 10^{-5} )</td>
<td>( 3.2 \times 10^{-5} )</td>
<td>0.0036</td>
<td>20+14</td>
<td>1.36</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(assumed)</td>
<td></td>
</tr>
<tr>
<td>WITHIN</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aug.</td>
<td>( 2.8 \times 10^{-4} )</td>
<td>( 3.8 \times 10^{-4} )</td>
<td>0.013</td>
<td>250+180</td>
<td>1.36</td>
</tr>
<tr>
<td>Sept.</td>
<td>( 8.6 \times 10^{-5} )</td>
<td>( 8.6 \times 10^{-5} )</td>
<td>0.007</td>
<td>35</td>
<td>1.01</td>
</tr>
<tr>
<td>BELOW</td>
<td>( 2.3 \times 10^{-5} )</td>
<td>( 2.4 \times 10^{-5} )</td>
<td>0.0033</td>
<td>17+12</td>
<td>1.04</td>
</tr>
</tbody>
</table>

\( \alpha = \frac{1}{\rho} \frac{\partial \rho}{\partial T} \), \( \beta = \frac{1}{\rho} \frac{\partial \rho}{\partial S} \), \( N^2 = g(\beta \frac{dS}{dz} - \alpha \frac{dT}{dz}) \), \( z \) is positive downwards
necessarily mean that all turbulence is suppressed since by having the gradients very non-uniform any (local) value of $\text{Ri}_g$ is compatible with any (overall) value of $\text{Ri}$ over the interval $\Delta Z$ (Stewart, 1969). However, the probability of patchy turbulence probably increases with decreasing (but still relatively large) values of $\text{Ri}$, and profiles with relatively low $\text{Ri}$ do seem more "turbulent". For example, at the time of MSP 17, $\text{Ri}$ (-50) was a factor of 10 lower than during MSP 19 or 21, and the MSP 17 profiles (Figure 30) do show much more finestructure and microstructure. During MSP 32B, $\text{Ri}$ was again fairly small and again the temperature profile (Figure 31) shows a "steppy" appearance. MSP 32B shows a layer (a region of relatively small vertical gradient) of about 3-m thickness bounded above and below by very sharp, strong, stabilizing gradients. An examination of the temperature gradients in layers such as this indicates the presence of active turbulence (Figure 32). Later, spectral analysis will confirm this.
Figure 30. Temperature profiles through the thermocline. MSP 17 A-L; $R_i \sim 50$. The scale applies only to 17A, the others are offset by 1.5°C. The time between profiles was 6 minutes. The ship drifted 0.7 km to the WSW while the data were being taken; thus, the distance between profiles could have been as much as 50 m. The dashed lines are due to offset adjustments.
Figure 31. Temperature profile and (uncorrected) gradient profile MSP 32; $Ri \approx 35$. Note the relatively isothermal patch between 13 and 15.5 m.
Figure 32. Turbulent patch within the thermocline (MSP 31C). Total patch thickness is about 1 m; the section shown was used to compute the spectrum shown in Figure 35. The data are uncorrected for thermistor response.
RESULTS

Spectra

One-dimensional spectra of temperature gradient were computed using $2^n$ data points with a Fast Fourier Transform algorithm. A Hanning filter was applied to suppress leakage. The spectral estimates were band averaged yielding estimates at almost equally spaced intervals along the wavenumber axis. The spectra are shown both corrected and uncorrected for signal attenuation by the transducer (see Chapter I for details). Where possible, the spectra are first normalized by dividing each estimate by the (uncorrected) record variance and then ensemble-averaged. Spectra of salinity fluctuations could not be calculated because of the poor vertical resolution of the CTD.

Spectra from above, within, and below the thermocline are compared with the universal spectra in Figures 6 and 7. There may be a viscous-convective subrange present in all the spectra, but it is difficult to tell because there is so much power in the low wavenumber bands. An electronic filter cut the signal off sharply at 30 Hz ($0.7 < k < 2.0 \text{ cy/cm depending on fallspeed}$). The spectral estimates after this point represent the blow-up (by digital-domain corrections for the thermistor attenuation) of digitizer least-count noise. If the spectra are corrected for this they all roll-off at 0.7 to 2 cy/cm and this has been taken into account when fitting the universal curves (by eye). Occasionally the roll-off occurs at much lower wavenumber, implying a local spectral maximum at the diffusive lengthscale. The vertical bars
Figure 33. Normalized, ensemble-averaged spectra (left) above (middle), and (right) below the thermocline (MSP 19, 21). The dashed line shows estimates corrected for thermistor response. Suggested fits to the universal curves are shown for various ε. The vertical lines show plus and minus one standard deviation.
Figure 34. Normalized thermocline spectra (MSP 31, 32). The increased slope after about 1 cy/cm is caused by blow-up of the least-count digitizer noise and should be ignored when comparing the observed spectra to the universal curves, shown for $\epsilon = 10^{-5}$ and $10^{-4}$. 
showing ± (one standard deviation) indicate that there can be considerable variability between samples and that curve-fitting, with ε as parameter, is not unambiguous.

Results of various calculations are shown in Table 6. The skewness values are all negative reflecting the fact that the temperature distribution is contributing to the stable stratification. The kurtosis can be extremely large in the thermocline due to the intermittency of the small-scale fluctuations. The variance σ², uncorrected for thermistor attenuation (as are the skewness and kurtosis), provides a lower bound estimate of ε₀ (not shown in Table 6): ε₀ = (2 + 1) 2ρσ², where the value of the factor (2 + 1) depends upon isotropy assumptions (for isotropic turbulence ε₀ = 6ρσ²). The upper bound for ε₀ comes from (the largest ε) curve-fit. K_H is calculated from Eq. (5) using only the tabulated range in ε₀.

The spectra from MSP 31-32 and those from MSP 21 (below) seem to show a buoyancy range. Using Eq. (3), ε ~ 4 x 10⁻⁴ and 6 x 10⁻⁵ cm² s⁻³ respectively, which are slightly higher than the previous estimates.

Turbulent patches

A spectrum (Figure 8) for the gradients in a single patch seems to have a decade-long viscous-convective subrange as well as a smaller inertial subrange. An ensemble spectra (Figure 9, for patches occurring in closely spaced profiles shows a similar shape. The rates of energy dissipation are very large. Calculations (using \( \frac{\partial^2 T}{\partial z^2} \) for the thermocline)
Table 6. Summary of calculations for above, within, and below the thermocline. (The standard deviation is shown in parentheses.)

<table>
<thead>
<tr>
<th>Data</th>
<th>Length of record analyzed</th>
<th>Mean Gradients</th>
<th>Variance</th>
<th>Skewness</th>
<th>Kurtosis</th>
<th>( \frac{\partial^2 T}{\partial z^2} )</th>
<th>( \frac{\partial^3 \theta}{\partial z^3} )</th>
<th>( N^2 )</th>
<th>( \sigma^2 )</th>
<th>( \varepsilon^2 )</th>
<th>( \varepsilon )</th>
<th>( K_H )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Above (19,21)</td>
<td>2.5</td>
<td>0.20</td>
<td>0.04</td>
<td>5.1x10^-4</td>
<td>-1.7</td>
<td>11.5</td>
<td>3.0x10^-4</td>
<td>(1-4)x10^-5</td>
<td>10^-4-10^-3</td>
<td>0.3-5.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(19,21)</td>
<td>0.6</td>
<td>(0.10)</td>
<td>(1.5)</td>
<td>(13.5)</td>
<td>(5.8x10^-4)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Within (19,21)</td>
<td>3.0</td>
<td>1.5</td>
<td>0.5</td>
<td>6.4x10^-3</td>
<td>-2.0</td>
<td>9.7</td>
<td>4.4x10^-4</td>
<td>(0.7-2)x10^-5</td>
<td>10^-5-10^-4</td>
<td>0.008-0.05</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(19,21)</td>
<td>0.6</td>
<td>(0.2)</td>
<td>(1.0)</td>
<td>(8.2)</td>
<td>(3.0x10^-4)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(31,32)</td>
<td>10.4</td>
<td>0.5</td>
<td>0.12</td>
<td>1.7x10^-3</td>
<td>-3.5</td>
<td>64.3</td>
<td>3.1x10^-4</td>
<td>(2-5)x10^-6</td>
<td>10^-5-10^-4</td>
<td>0.03-0.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(3.0)</td>
<td>(3.2)</td>
<td>(77)</td>
<td>(2.3x10^-4)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Below (19)</td>
<td>2.7</td>
<td>0.35</td>
<td>0.032</td>
<td>7.3x10^-4</td>
<td>-1.6</td>
<td>10.1</td>
<td>3.6x10^-5</td>
<td>(2-5)x10^-6</td>
<td>5x10^-5</td>
<td>0.06-0.18</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(0.4)</td>
<td>(0.12)</td>
<td>(1.6)</td>
<td>(6.8)</td>
<td>(2.5x10^-5)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(21)</td>
<td>6.1</td>
<td>0.15</td>
<td>0.032</td>
<td>4.2x10^-4</td>
<td>-1.1</td>
<td>1.8</td>
<td>1.3x10^-6</td>
<td>&lt;1x10^-8</td>
<td>10^-7-10^-5</td>
<td>&lt;0.002</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1.2)</td>
<td>(0.01)</td>
<td>(0.7)</td>
<td>(2.8)</td>
<td>(7x10^-7)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 35. Gradient spectrum of turbulent patch in the thermocline (see Figure 32. The roll off at $k = 2$ cm$^{-1}$ is caused by an electronic filter which low-passes at 30 Hz. The blow-up of least-count noise can be seen (after 3 cm$^{-1}$). The vertical lines are the 95% confidence limits.
Figure 36. The upper spectrum is the corrected version from the MSP 17D thermocline. The lower spectra are the (dashed) corrected and (solid) uncorrected normalized, ensemble-averaged spectra for turbulent patches in 17 D-L. The vertical lines show plus and minus one standard deviation. The heavy curves are universal curves for $\varepsilon = 10^{-3}$ and $10^{-2}$. 
show (Table 7) $K_H$ to be an order-of-magnitude larger than in the whole of the thermocline.

Using the observed patch size, $l \sim 70$ cm, and $u \sim U/30 \sim 1$ cm s$^{-1}$, the patch Reynolds number is $u_l/v \sim 5000$, just large enough for an inertial subrange to exist (Tennekes and Lumley, 1972, p. 266). By Eq. (4), $\varepsilon \sim 0.014$ cm$^2$ s$^{-3}$, in rough agreement with the other estimates.

Figure 36 also shows the spectrum of a 4-m record from the 17D thermocline. Using the observed mean gradient for this profile (0.37 °C m$^{-1}$) and the calculated $\varepsilon_\theta$ ($1.4 \times 10^{-3} < \varepsilon_\theta < 4.4 \times 10^{-3}$) one obtains $32 < K_H < 100$ from Eq. 5.

Lengthscales and magnitudes of the stable gradients

It is interesting to compare the regions studied with regard to the vertical thickness and magnitude (uncorrected for attenuation effects) of the observed stabilizing, relatively long-lived temperature gradients. Let the vertical distance $l$ be the length over which the stable temperature gradient was significantly larger (greater than the least count gradient) than the mean. Each fluctuation deviating from the mean in this way can be thought of as an interface characterized by $l$ and, for convenience, the maximum gradient over the length $l$, $\left(\frac{\partial T}{\partial z}\right)_{\text{max}}$. A plot of $\left(\frac{\partial T}{\partial z}\right)_{\text{max}}$ against $l$ (Figure 37) shows that within and below the thermocline $\left(\frac{\partial T}{\partial z}\right)_{\text{max}} \sim 5 \left(\frac{\partial T}{\partial z}\right)$ and above $\left(\frac{\partial T}{\partial z}\right)_{\text{max}} \sim 25 \left(\frac{\partial T}{\partial z}\right)$, in each case there being little dependence on $l$. [The ratio $\left(\frac{\partial T}{\partial z}\right)_{\text{max}}/\partial z$ is in some way similar to $K_H = (\nabla \theta)^2/\nabla^2 \theta$ (Eq. 5).] Histograms of the number of observations
Table 7. Summary of calculations for turbulent patches

<table>
<thead>
<tr>
<th>Data</th>
<th>$\Delta z$ (cm)</th>
<th>Skewness</th>
<th>Kurtosis</th>
<th>$\sigma^2$ $(^{\circ}\text{C cm}^{-1})^2$</th>
<th>$\epsilon$ $(\text{cm}^2 \text{s}^{-3})$</th>
<th>$\epsilon_0$ $[^{\circ}\text{C}^2 \text{s}^{-1}]$</th>
<th>$K_h$ $(\text{cm}^2 \text{s}^{-1})$</th>
</tr>
</thead>
<tbody>
<tr>
<td>31C</td>
<td>68</td>
<td>0.01</td>
<td>0.12</td>
<td>$5.8 \times 10^{-4}$</td>
<td>$10^{-2}-10^{-1}$</td>
<td>$(0.9-3) \times 10^{-4}$</td>
<td>2-6</td>
</tr>
<tr>
<td>17 D-L</td>
<td>67</td>
<td>0.16</td>
<td>1.6</td>
<td>$1.6 \times 10^{-3}$</td>
<td>$10^{-3}-10^{-2}$</td>
<td>$(1.1-3.6) \times 10^{-4}$</td>
<td>0.2-0.8</td>
</tr>
<tr>
<td></td>
<td>(23)</td>
<td>(0.40)</td>
<td>(1.0)</td>
<td>$(1.5 \times 10^{-3})$</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 37. Plot of the average of the maximum (stable) vertical temperature gradients observed in "interfaces" of the given thickness. The average maximum plus one standard deviation is also shown. The heavy line represents the mean large-scale gradient observed in each regime. (Data: MSP 19 A-G, 21 A-E.)
of an interface with thickness \( \ell \) (Figure 38) show a strong dependence on \( \ell \), a more active (higher \( \varepsilon \) and \( \varepsilon_0 \)) region having a higher percentage of very small lengthscale fluctuations. Interfaces, defined as above, with \( \ell > 4 \) cm, are relatively infrequent in each region.
Figure 38. Histograms of observed interface thickness from above, within, and below the thermocline. Lengths of records were about the same, with the exception of 21 Below, so no normalization was done.
DISCUSSION

A mixing model

The results of the calculations of $K_H$ raise the question of what the downward heat transport $F_H = \rho c_p K_H \frac{\partial T}{\partial z}$ really is. One of the most active thermoclines, 17D, yields $F_H$ as large as $0.47 \text{ cal cm}^{-2} \text{ s}^{-1}$; but analysis of other thermoclines gives $F_H$ as small as $1.2 \times 10^{-4} \text{ cal cm}^{-2} \text{ s}^{-1}$. Are these variations real? What is the average flux?

Simple models of vertical mixing propose that heat progresses downwards spasmodically by passing through relatively thin, short-lived layers which form via a shear instability, possibly of the Kelvin-Helmholtz type (Stommel and Federov, 1967; Woods, 1968). If we apply the Stommel-Federov model to 17D, which possibly shows about five or six layers, and require that $F_H = 0.47 \text{ cal cm}^{-2} \text{ s}^{-1}$, then a shear instability must occur on each interface, causing a new layer to form, within a period of a few minutes. This is roughly the length of the sampling period between the MSP 17 profiles and may explain the large changes seen from one profile to the next (although ship drift provides an alternate explanation of why successive profiles are dissimilar).

All of the MSP 17 profiles show a very "active" temperature structure. MSP 13, 14, and 15 (MSP 16 did not use a thermistor) made several hours before MSP 17, show "quieter" temperature records in the thermocline, similar to MSP 19 and 21. (These earlier profiles were not analyzed in detail because "yo-yo"sequences were not obtained.) It is conjectured that the bulk of the vertical heat transport occurs through such "active"
thermoclines. As 35 profiles were made in a 12.5 hour period and 12 of them (17 A-L) were active, large vertical heat transport occurs at most 12/35 = 33% of the time. Using $K_H = 32 \text{ cm}^2 \text{s}^{-1}$ (the lower bound in 17D) as typical of an active thermocline, the effective, mean eddy diffusivity for the thermocline ($\overline{K}_H$) must be less than $(0.33) (32) = 10 \text{ cm}^2 \text{s}^{-1}$.

An analysis of those parts of MSP 17 profiles which seem to be patches of turbulence has shown that $0.5 < K_H < 5 \text{ cm}^2 \text{s}^{-1}$ within the patches. Even in the quieter profiles patches seem to exist although their vertical extent is much smaller. A very rough estimate of the value of the fraction ($f$) of the typical thermocline which contains patches is 0.1, and this can be used to estimate an effective overall diffusivity:

$$\overline{K}_H = f \cdot K_H \approx (0.1) (0.5 \text{ to } 5) = (0.05 \text{ to } 0.5) \text{ cm}^2 \text{s}^{-1}.$$  

This is smaller than $10 \text{ cm}^2 \text{s}^{-1}$, as expected, but within the range derived using the universal curve fits to the thermocline spectra (Table 6). It seems likely then that $\overline{K}_H < 1 \text{ cm}^2 \text{s}^{-1}$ for the summer coastal thermocline under meteorological conditions similar to those in this study.

Heat budget

Since the currents were mostly alongshore, it may be reasonable to neglect horizontal advection and to attempt a heat budget for the slab-like volume of water occupied by the thermocline. Denoting variables above (below) by superscript "a" ("b"), the predicted rise in the thermocline is

$$\frac{\delta T}{\delta t} \approx [K_a \left( \frac{\delta T}{\delta Z} \right)^a - K_b \left( \frac{\delta T}{\delta Z} \right)^b] D^{-1},$$

where $D$ is the thermocline thickness. Using $\left( \frac{\delta T}{\delta Z} \right)^a = \left( \frac{\delta T}{\delta Z} \right)^b = 0.002 \degree \text{C} \text{ cm}^{-1}$, and $\Delta K = K_a - K_b \approx 1 \text{ cm}^2 \text{s}^{-1}$ and $D = 600 \text{ cm}$, $\delta T = 0.1 \degree \text{C}$ in 520 minutes, an increase which may or may
not have been observed. Therefore, neither the assumption of a vertical balance nor the value of $\Delta K$ is unreasonable.

**Comparison with other measurements**

How do the estimates of $\varepsilon$, $\varepsilon_\theta$, and $K_H$ compare with results from other studies in similar regions? Belyaev (1975) obtains $0.06 < \varepsilon < 0.15$ cm$^2$ s$^{-3}$ at the base of a surface mixed layer; Gibson (1976) finds at 23 m off San Diego that the $\varepsilon$ in a turbulent patch can be as high as $0.1$ cm$^2$ s$^{-3}$, but that average values are more like $\varepsilon \sim 10^{-3}$ cm$^2$ s$^{-3}$ and $\varepsilon_\theta \sim (5 - 34) \times 10^{-8}$ (°C)$^2$ s$^{-1}$; Woods (1968) finds $K_H = 0.3$ cm$^2$ s$^{-1}$ in the summer thermocline in calm weather but $K_H = 5$ cm$^2$ s$^{-1}$ during a heating, restratification period (Woods and Wiley, 1972); and Gregg (1976) finds $7 \times 10^{-10} < \varepsilon_\theta < 2 \times 10^{-7}$ (°C)$^2$ s$^{-1}$ in the transition region at the base of a 70-m surface mixed layer, and $2 \times 10^{-9} < \varepsilon_\theta < 4 \times 10^{-8}$ (°C)$^2$ s$^{-1}$ in the upper thermocline. Generally then, our estimates of $\varepsilon_\theta$ are very much larger than those measured by previous workers. Our choice of $K_H < 1$ cm$^2$ s$^{-1}$ is not inconsistent with the work of Woods.

Various authors (see Powell, 1975) have proposed the following expression for the vertical eddy diffusivity:

$$K_H = (\text{constant}) \ (N^2)^{-m},$$

where $m$ varies from 0.25 to 2. Ozmidov (1965), through dimensional arguments, finds that $K_H$ can attain a maximum value for isotropic turbulence of

$$K_H = 0.1 \ \varepsilon \ N^{-2}$$

Sarmiento, Feely, Moore, and Broecker (1975) propose an $N^{-2}$-type expression
for the vertical eddy viscosity $K_v$ based on many measurements in the deep-ocean benthic mixed-layer.

Our measurements of $K_H$, along with estimates of $\epsilon$, are compared with the above work in Figure 39. Unfortunately, estimates of $\epsilon$ are unavailable for the other investigations. Assuming that for lakes $\epsilon$ would be about the same or smaller than the oceanic $\epsilon$ and assuming $\epsilon \approx 3 \times 10^{-5}$ cm$^2$ s$^{-3}$ in an extrapolation of the benthic mixed layer study, Figure 39 shows that the idea of $K_H$ being proportional to $(N^2)^{-m}$ with $\epsilon$ as a parameter is not totally unreasonable. For $\epsilon$ about $10^{-4}$ cm$^2$ s$^{-3}$, our data show $m$ to be about $2/3$. 
Figure 39. Plot of vertical eddy diffusivity $K_v$ against the square of the buoyancy frequency. Our measurements are shown as heavy bars along with estimates of $\varepsilon$. The work of Hutchinson in Lake Mendota and in Linsley Pond, and that by Powell and Jassby in Castle Lake, all as reported in Powell (1975), is indicated by the dotted regions. Measurements by Dillon, Powell, and Myrup (1975) in the September thermocline in Lake Tahoe are also indicated. Sarmiento, Feely, Moore, and Broecker (1975) calculated $K_v$ in the benthic mixed layer using measurements made in the GESECS program. These are shown in the upper left. The expression they propose is coincident with Eq. (7) if $\varepsilon = 3 \times 10^{-5}$ and this is shown as a solid line. Parallel lines, for other values of $\varepsilon$, could be drawn through our measurements.
1. The summer thermocline in 100-m of water off the Oregon coast has a very non-uniform temperature profile.

2. Small-scale stable temperature interfaces typically of less than four centimeters vertical extent have temperature gradients 5-25 times higher than the local mean vertical gradient.

3. A viscous-convective subrange is seen in some of the temperature gradient spectra.

4. In the transition region between the surface mixed layer and the thermocline the rate of dissipation of thermal fluctuations, $\varepsilon_\theta$, is about $(1 - 4) \times 10^{-5} \, (\circ C)^2 \, s^{-1}$; in the thermocline $\varepsilon_\theta \sim (0.2 - 2) \times 10^{-5} \, (\circ C)^2 \, s^{-1}$; and below the thermocline, still in a region a relatively rapid temperature decrease but in a transition to the deeper shelf water, $\varepsilon_\theta < 5 \times 10^{-6} \, (\circ C)^2 \, s^{-1}$.

5. A convective subrange (equivalent to the -5/3 inertial subrange in the velocity spectrum) is seen only in turbulent layers (patches) of about one meter or less thickness within the thermocline. Here the rate of dissipation of mechanical energy $\varepsilon$ is estimated from the shape of the temperature gradient spectra to be about $10^{-2} \, cm^2 \, s^{-3}$; and the Reynolds number (Re) in the patch is calculated to be greater than $5 \times 10^3$. Thus, turbulence in the seasonal thermocline is not totally inhibited by small Re or by large (20-400) macroscopic Richardson number.

6. It is not really clear what the effective, time-averaged vertical eddy diffusivity $K_H$ for the thermocline is. Calculations for one profile show $K_H$ as large as 100 cm$^2$ s$^{-1}$ while for another, several hours later,
$K_H$ can be four orders-of-magnitude smaller. It is estimated that on the average $K_H \leq 1 \text{ cm}^2 \text{ s}^{-1}$.

7. The measurements of $K_H$ are not inconsistent with the proportionality $K_H \propto (N^2)^{-m}$, where $m$ is about 2/3. It is suggested that $K_H$ also depends on $\varepsilon$. 
IV. LAYERS AND INTERFACES: SMALL-SCALE IMPRINTS ON THE TEMPERATURE STRUCTURE.

Mixing in stratified natural fluid systems with large macroscopic Richardson numbers tends to occur intermittently, often in thin patches and occasionally in a more organized vertical pattern. Fine-resolution vertical profiles of temperature or salinity record the imprint of the mixing, frequently showing layers of relatively homogeneous water separated by thinner interfaces across which density and velocity change abruptly. The abrupt changes lower the local Richardson number to turbulence-sustaining levels. In this paper, some observations of layers and interfaces made with a vertically profiling temperature microstructure instrument are discussed relative to some mechanisms for layer and interface formation which have been studied in the laboratory.

The mechanisms are of two types. The first relies on the difference between molecular diffusivities of heat and salt: double-diffusive convection (Turner, 1973). No ambient shear is needed for layer formation by this process, potential energy in the unstably stratified component (heat in this case) driving the convection. A second group of mechanisms depends upon a mean vertical shear for a local source of kinetic energy and includes: (1) billow turbulence (Woods and Wiley, 1972); (2) a mechanism recently discovered by Long (1973) which can produce more than one interface from a stratified shear flow when the overall Richardson number is somewhere between 1 and 2.5; (3) a momentum-density
diffusive instability in a geostrophic shear flow (Calman, 1976).

The shear mechanisms require the ambient shear to be large enough to lower the local Richardson number below a critical value (0.25 for many flows) for a time long enough for shear instabilities to grow and produce unstable density gradients (local overturning) and a well-mixed turbulent (if the Reynolds number is high enough) layer. For the simple case of an interface between two homogeneous layers, the initial instability is a Kelvin-Helmholtz (K-H) instability. The K-H instability grows large and becomes a billow, and the ensuing mixing has come to be called billow turbulence. Billow turbulence has been proposed as the major interior mixing mechanism in all statically stable fluids, including the atmosphere (Woods and Wiley, 1972). The effect of the turbulence is to rapidly increase the thickness of the interface, causing a rapid vertical diffusion, raising the center of gravity of the mixed fluid, and creating two interfaces where before there was but one. When the turbulence subsides, the layer can collapse, generating internal waves which can trigger instabilities in other parts of the water column. We should expect, then, to see ample evidence of these processes in the temperature structure. As will be seen, some of the temperature features are very striking. Although only a few examples are shown here, others can be found in Chapters II, III, and V which further describe the structure in the region.
OBSERVATIONS

All the data reported here were taken during a four-day cruise off the Oregon coast near a current-meter mooring in 100 m of water. One hundred and eight temperature microstructure profiles [denoted here by a drop number with a letter following to indicate a sequence of repeated (yo-yo) profiles] were made amidst 56 conductivity-temperature-depth (CTD) casts. Many of the small-scale (<1 m) layers and interfaces were embedded in larger-scale (>1 m) gradients that could be identified in both the microstructure profiles and the CTD casts. The instrumentation used is described by Caldwell, Wilcox and Matsler (1975).

Double-diffusive layering

A system of double-diffusive layers might be expected on the relatively persistent inversion found in the Oregon coastal waters (Mooers, Collins, and Smith, 1976). This inversion was present on the 7th and 8th of September 1974 at about 25 m (Fig. 40). It was caused by the presence of the cool ribbon, a core of relatively cool, fresh water at about 20 m. It may have been intensified by a simultaneous intrusion of relatively warm, salty water (the so-called warm-water anomaly) at about 30 m (Chapter II, hereafter II). Whatever the cause, the situation, relatively warm, salty water below colder, fresher water, was favorable for
Figure 40. Temperature, plotted against depth and time, from CTD casts made with 0.5 km of a 100 m station (45°00.0' N and 124° 10.0' W) off the Oregon coast in September 1974. The shaded areas are regions of temperature inversions (temperature increasing with depth). The times of the microstructure profiles are indicated at the bottom.
the occurrence of a system of layers caused and/or maintained by diffusive double-diffusive convection. Furthermore, the layering should be expected to persist in time and space on the basis of the stability analysis of Huppert (1971).

However, profiles through the inversion (Figure 41) do not show the uniformly stepped temperature structure which is so characteristic of the more well-established observations of diffusive convection in the ocean (e.g., Neshyba, Neal, and Denner, 1971). The few layers and interfaces which were present did not persist from one profile to the next. Since the ship drift between profiles was about 300 m (the instrument drift was probably much less than this) the horizontal extent of any layer must be less than 300 m.

\[ \frac{R_p}{G_p} \equiv \frac{\beta}{\alpha} \frac{\partial S}{\partial Z} / \frac{\partial T}{\partial Z} \]

where \( Z \) is positive downwards, \( \alpha = -\frac{1}{\rho} \frac{\partial \rho}{\partial T} \) and \( \beta = \frac{1}{\rho} \frac{\partial \rho}{\partial S} \). For the inversion region \( 2 < G_p < 4 \) (II), if it is assumed that the diffusive layering produces interfaces with interfacial stability numbers \( R_p \) about equal to \( G_p \) then \( 2 < R_p < 4 \) which puts the system in Huppert's stable regime. (The definition of \( R_p \) is \( R_p \equiv \beta \Delta S / \alpha \Delta T \) where \( \Delta \) denotes a change across the interface.)
Figure 4.1 Temperature profiles through the temperature inversion below the cool ribbon. The curves are integrations of temperature gradient records. The temperature scale applies only to 4A; the others are offset by 0.8 °C. Drop 4A was made at the 100 m station; the ship had drifted 300 m to the NE when 4B was made, and another 300 m for 4 C.
A search of the many other temperature inversions in the water column (Fig. 40) resulted in only one example of a quasi-layered structure: a series of three layers on a two-meter thick inversion about 15 m above the bottom (Fig. 42). Over the whole inversion $G_{po} = 1.8$, putting possible diffusive interfaces in Huppert's unstable regime. Immediately after the profile in Figure 42, three more profiles about 12 minutes and less than 150 m apart were made and showed no layering on the inversion.

If the diffusive layer-forming mechanism is assumed to be at work in the two inversions mentioned above, estimates of the layer size and the time needed for the layers to form could be obtained by using results (Eqs. 6 and 20 in Turner, 1968) from laboratory experiments in which an initial, stabilizing salinity gradient \( \frac{\partial S}{\partial z} \) is heated from below at a constant rate $H$. For \( \frac{\partial S}{\partial z} \_o \) we use the observed mean density gradient. For the externally imposed heat flux we use $H = 3 \times 10^{-4}$ cal cm\(^{-2}\) s\(^{-1}\) for the inversion at 25-30 m and $H = 2 \times 10^{-5}$ cal cm\(^{-2}\) s\(^{-1}\) for the deep inversion. The first value is derived in II on the basis of a heat budget of the cool ribbon and also from an estimate of the eddy diffusivity in the inversion using turbulence theory; the second value is an estimate of the heat flux out of the bottom boundary layer (Caldwell, 1976).

The predicted layer depths, 0.2 to 0.4 m and 1.7 m for the

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\(^3\)This is not necessarily the case elsewhere. Gregg and Cox (1972) found that every prominent, stable temperature inversion had a "family" of sharp interfaces where they concluded double diffusive convection was active.
Figure 42. Profile (Drop 26A) through a deep-water temperature inversion. The inset shows data from a nearby CTD.
shallow and deep inversions respectively, agree roughly with the observations. However, the formation times, 260 min and 80 min respectively, are very much larger than an overturning time-scale based on the mean vertical shear (Table 8). It seems likely that the diffusive mechanism, if at work at all, is only secondary to layer formation by some type of shear-induced mixing process.\(^9\)

**Non-double-diffusive layering**

An example of layering in the seasonal thermocline where the temperature and salinity distributions are both stabilizing—and double-diffusive mechanisms can thus be ruled out—is shown in Figure 43. Ten or so layers can be seen. There are large fluctuations within the layers and the bounding interfaces are about 2 cm thick (Table 8). The ratio of the mean interface thickness to the mean layer depth is about 0.05. Long (1973) obtained 0.16 for this ratio from his laboratory experiments. Woods (1968) in the seasonal thermocline around Malta usually observed 10 cm interfaces between 1-2 m layers, giving a ratio of 0.05 to 0.10.

It is interesting to compute for the seasonal thermocline the Ozmidov length \(L_O = (\varepsilon/N^3)^{1/2}\), a lengthscale representative of the

\(^9\)Neshyba, Neal and Denner (1971) found that a shear of 0.01 to 0.02 s\(^{-1}\), slightly larger than that estimated here, was enough to dominate the mechanism (presumably double diffusive convection) responsible for the layered system.
Table 8. Summary of interface, layer, and stability characteristics. \( H \) (h) is the observed layer (interface) thickness; \( L_o = (\epsilon/N^3)^{1/2} \) is the Ozmidov length; \( \Delta T \) is the temperature difference across the interface; \( G_p \equiv \beta_3 S_N^O/\alpha_3 \Delta T \) is the large-scale "stability" number; \( \Delta S \) and \( \Delta \rho \) are estimates of the salinity and density differences across the interface, where \( \Delta S \equiv G_p \Delta T \alpha/\beta; \ N^{-1} \) is the buoyancy period; \( S^{-1} \) is the period of overturning due to the mean shear; and \( \overline{RI} \equiv N^2/S^2 \) is the macroscopic Richardson number.

<table>
<thead>
<tr>
<th>Regime</th>
<th>( H ) (m)</th>
<th>( L_o ) (m)</th>
<th>h (cm)</th>
<th>( \Delta T ) (°C)</th>
<th>( G_p )</th>
<th>( \Delta S ) (gm cm(^{-3}))</th>
<th>( \Delta \rho ) (gm cm(^{-3}))</th>
<th>( N^{-1} ) (min cy(^{-1}))</th>
<th>( S^{-1} ) (min cy(^{-1}))</th>
<th>( \overline{RI} )</th>
</tr>
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<tbody>
<tr>
<td>Double Diffusive</td>
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<tr>
<td>4A, B, C</td>
<td>0.2-0.5</td>
<td>0.02-0.2</td>
<td>4-20</td>
<td>0.1-0.3</td>
<td>2-4</td>
<td>0.04-0.2</td>
<td>(0.2-1.4)\times10^{-4}</td>
<td>2-4</td>
<td>15</td>
<td>8-20</td>
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<tr>
<td>26A</td>
<td>0.5-1</td>
<td>1.0</td>
<td>4-9</td>
<td>0.008</td>
<td>1.8</td>
<td>0.003</td>
<td>1\times10^{-6}</td>
<td>32</td>
<td>21</td>
<td>0.2-1.2</td>
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<tr>
<td>8A</td>
<td>0.4</td>
<td>0.2-0.5</td>
<td>2</td>
<td>0.14</td>
<td>-0.4</td>
<td>0.011</td>
<td>3\times10^{-5}</td>
<td>4</td>
<td>15</td>
<td>13</td>
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<tr>
<td>Single interfaces</td>
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<tr>
<td>6A-D</td>
<td>--</td>
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<td>10-120</td>
<td>0.031</td>
<td>-2.1</td>
<td>0.012</td>
<td>1.4\times10^{-5}</td>
<td>21</td>
<td>38</td>
<td>3.2</td>
</tr>
<tr>
<td>25D</td>
<td>--</td>
<td>--</td>
<td>&lt;20</td>
<td>0.07</td>
<td>+1.5-2.5</td>
<td>0.02-0.03</td>
<td>(0.5-1.5)\times10^{-5}</td>
<td>20-40</td>
<td>39</td>
<td>0.9-4.0</td>
</tr>
</tbody>
</table>
Figure 43. Shear-induced layering in the seasonal thermocline (Drop 3A). Roughly ten layers can be seen. (The isothermal region at 13-14 m persists in neighboring profiles; its origin is unknown.)
largest isotropic eddies in a stratified shear flow, where \( \epsilon \) is the rate of dissipation of mechanical energy and \( N \) is the buoyancy (Brunt-Väisälä) frequency. Estimates of \( \epsilon \) obtained from the convective-to-viscous-convective transition in the temperature gradient spectra in the seasonal thermocline vary between \( 10^{-5} \) to \( 10^{-4} \) cm\(^2\) s\(^{-3}\) during periods without intrusions below (see Chapter III); an analysis of the thermocline at the time of the inversion showed \( \epsilon \) to be as large as \( 10^{-2} \) cm\(^2\) s\(^{-3}\). Using \( 10^{-4} < \epsilon < 10^{-3} \) and \( N = 0.025 \) s\(^{-1}\), we calculate \( 16 < L_0 < 50 \) cm, very close to the observed layer thicknesses. A somewhat independent estimate of \( L_0 \) is given by the wavelength at which the transition from the buoyancy subrange to the convective (inertial) subrange in the temperature gradient spectrum occurs (e.g., Moseley and Del Balzo, 1976). At the time of the inversion the transition was at about 50 cm, again remarkably close to the layer sizes. Is this merely coincidental?

**Kelvin-Helmholtz instabilities?**

Almost every temperature profile showed at least one interface which was not obviously associated with other interfaces. Figure 44 shows four profiles made over a 30-min interval through a 0.03 C interface which is somewhat typical of those found below about 40 m at this time. The ship drifted about 0.7 km during the measurements. The depth of the interface increased about 2 m towards shore. The nearest neighboring interfaces were 8 m above and 13 m below. Each
Figure 44. Temperature profiles through an interface at about 73 m (Drop 6A, B, C, D). The ship drifted 0.7 km during the drops: thus, each was at most 240 m from the last.
had a different $\Delta T$. There is little change in the mean vertical gradients above or below the interface; but there is a large variation in its thickness, probably due to a localized intermittent mixing. Both shear and buoyancy effects are important (Table 8), and evidence will be given that the vertical transport takes place via turbulence initiated by K-H instabilities.

A series of seven profiles, about 9 min apart and perhaps very close together in space, through a 10-m thick inversion shows many interfacial mixing events (Fig. 45). The similarity of some of these interfacial instabilities to observations of K-H instabilities made both in the laboratory (Thorpe, 1973) and in the field (Woods, 1968; Thorpe, 1974) is striking. The instability in the fourth profile, Drop 25D, can be explained by assuming that the temperature probe cut through a K-H billow in its roll-up stage (Fig. 46). Various stages of billow turbulence appear elsewhere in the inversion; but in other profiles, where both temperature and salinity fields were stabilizing, additional examples of the easily recognized roll-up stage have been found in the temperature traces (see II for an example at the base of the surface wind-mixed layer).

Following Thorpe (1973), the sequence of events on an interface like that in 25D is probably similar to the following: (1) At the onset of the K-H instability the interface has a thickness $2d = 4$ cm. The velocity difference across it is estimated as follows: at 50 m, $(u, v) = (2.8, 26.3 \text{ cm s}^{-1})$; at 75 m, $(u, v) = (-0.6, 20.5 \text{ cm s}^{-1})$. Assuming the largest possible velocity difference exists there is a
Figure 45. Profiles through a 10 m thick temperature inversion (Drop 25 A-G, 1117-1218 PDT, 10 September). The mean current at 50 m and the ship drift speed were 27 cm s$^{-1}$, both to the NE. The dashed lines are interpolations made necessary by offset adjustments in the chart recordings.
Figure 46. An example of the early "roll-up" stage of Kelvin-Helmholtz initiated billow turbulence. On the left is a sketch showing the vertical path of the thermistor through the billow; on the right is the actual temperature profile, Drop 25D (also shown in Fig. 45).
relative flow $U_o = +3$ cm s$^{-1}$ above (say) the interface and an
opposing flow $U_o = -3$ cm s$^{-1}$ below the interface; (2) The instability
grows, becoming a billow with an amplitude of about 20 cm (Fig. 46.
Laboratory observations show that the maximum vertical velocities
are about $2/3 U_o = 2$ cm s$^{-1}$ after the billows overturn. Gradients
between the billows can be expected to be quite large, about (20) x
$(0.07^\circ C/4 \text{ cm}) = 0.35^\circ C \text{ cm}^{-1}$; (3) The flow within the billows becomes
turbulent when the billow height is about one-third its wavelength $\lambda$
(the distance between billows): thus, $\lambda \geq (3) x (40 \text{ cm}) = 1.2 \text{ m}$. The
billows amalgamate to form a turbulent layer, the boundaries of which
spread at a rate of about $(0.1) U_o = 0.3$ cm s$^{-1}$; (4) The spreading
stops when the nondimensional time $\tau \equiv g\Delta\rho t/2U_o = 3$, where $t$ is time
since the flow first became turbulent. For the present example this
occurs after about 30 min. The final layer thickness $(2D)$ should thus
be expected to be about 10 m$^{11}$. The layer Reynolds number $Re = U_o D/v$
$-10^5$; the layer Richardson number is expected to be 0.4 on the basis
of Thorpe's laboratory work; (5) Most of the turbulence has subsided
when $\tau = 15$. Dimensionally, the collapse time is about 1 hour

$^{10}$ Thorpe (1974) observed billows with $\lambda = 3.4$ m in Loch Ness. Woods
(1968) gives $\lambda = 0.8$ m in the seasonal thermocline from visual
observations by divers.

$^{11}$ Woods and Wiley (1972) suggest that the final thickness is only
four or five times the initial thickness. In this case, we should
expect $2D \sim 20$ cm.
for this case; (6) The approximate mean rate of energy dissipation in the layer, $\bar{\varepsilon}$, is [from Thorpe's Eq. (15)]: $1.2 \times 10^{-4} < \bar{\varepsilon} < 3.5 \times 10^{-4} \text{ cm}^2 \text{ s}^{-3}$. This compares with an independently estimated value of $\varepsilon$ in the whole inversion of about $10^{-5} \text{ cm}^2 \text{ s}^{-3}$ (II).
1. Interfaces separating layers of homogeneous water (or zones of different temperature gradient) as thin as a few centimeters were found in various parts of the water column during the observation period. Some, fairly deep, with changes across them of about 0.03°C and 0.001°/oo, seemed to persist over large nearly horizontal distances.

2. Where the temperature field was destabilizing and double-diffusive layering was possible, it was not found. The coastal upwelling region may be too mechanically energetic for persistent systems of layers to form.

3. Layering occurred when temperature, in addition to salinity, was stabilizing. In the seasonal thermocline for example, where the large-scale Richardson number was about ten, the layer size was about 50 cm.

4. When layers were observed, their thicknesses were within a factor of two of the Ozmidov lengthscale $L_o \equiv (\varepsilon / N^2)^{1/2}$.

5. Vertical transport across many interfaces seemed to occur via the mechanism of Kelvin-Helmholtz initiated billow turbulence.
V. SMALL-SCALE TEMPERATURE STRUCTURE AS MEASURED BY THERMOCOUPLES.

Thermistors have been the usual transducer for small-scale measurements of the vertical component of the temperature gradient in lakes (e.g., Dillon, Powell, and Myrup, 1975) and in the ocean (e.g., Gregg, 1976). In addition, thermistors have been used to measure temperature differences between points 10-50 cm apart in the vertical (Woods, 1968) and to obtain vertical profiles of quasi-horizontal temperature differences between points several meters apart (e.g., Gregg, Cox, and Hacker, 1973). To achieve better dynamic range in the presence of a varying mean signal, it is common to use the high-passed (differentiated) part of the temperature signal; and so it seems natural to use from the start a differential temperature transducer—a thermocouple. Thermocouple materials are readily available at reasonable cost, thermocouples of the same type are completely interchangeable, and for the temperature ranges of interest, have a well-defined sensitivity. The thermocouple is not subject to self-heating problems as are thermistors. Furthermore, if noise can be reduced to an acceptable level, a single junction thermocouple can be made to have a faster response time than even the smallest flake or bead thermistor, thus providing useful information at high wavenumber. 12

12 Thermocouples with junction diameters smaller than 0.002 in. (.005 cm) are commonly used in psychrometers (Lopushinsky, 1972).
Thermocouples with fixed reference-junction temperature have been used in the ocean by Urich and Searsfoss (1948) and Liebermann (1951) and in fresh water by Speranskaya (1960). Shirtcliffe (1969) for example, has used single thermocouples in "floating" or differential operation to study the onset of small-scale convection in the laboratory.

In this chapter we report on the use of thermocouples with floating reference-junction temperature (differential thermopiles) in both a freshwater lake and in the ocean. Construction details, including a description of the various geometrical configurations used, signal interpretation, and detailed analyses of selected records are described. The associated free-fall instrument package and its use in the field are described in Caldwell, Wilcox, and Matsler (1975).

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13 It is perhaps interesting to note that these were the first records of small-scale temperature fluctuations made in the sea (Benilov, 1969).
THE THERMOCOUPLE TRANSDUCERS

The electronics package permitted the use of either (a) two thermopiles or (b) one thermopile and one thermistor on any given profile (drop). All transducers were mounted at the bottom of the free-fall package. The thermopiles, built on thin cylindrical tubes of nylon, were mounted concentrically, with a thermistor at the center (Figure 47.)

Construction details

An individual thermocouple pair is made from 0.005-in. (0.13-cm) diameter (unsheathed) chromel and constantan wires. A junction is formed by twisting the wires together, applying soldering-flux, applying a small amount of solder, and then cutting the junction down leaving enough twists to provide good mechanical contact between the wires. (A better method would be butt welding which produces a small bead-like junction.) The diameter of the junction is twice the diameter of the individual wires or about 0.010 in. (0.26 cm). We found it unnecessary, as did Liebermann (1951), to apply electrical insulating material to the junctions.

The standard thermopile (see Table 9) consisted of forty thermocouples (two junctions per pair) connected in series and constructed so that alternating junctions are uniformly radially-spaced on one of two

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14 Omega Engineering, Inc., Box 4047, Springdale Station, Stamford, Connecticut 06907.
Figure 47. Thermopiles, of 3 and 9-cm diameters, mounted concentrically at the lower end of free-fall instrumentation. A 0.020-inch (.05 cm) glass-encased thermistor is mounted at the center. Also shown is the protective "cage".
Table 9. Thermopile characteristics.

<table>
<thead>
<tr>
<th>Effective tube diameter, D (cm)</th>
<th>Row spacing, d (cm)</th>
<th>Number of pairs of thermocouples</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>0.8</td>
<td>20</td>
</tr>
<tr>
<td>3</td>
<td>0.8</td>
<td>40</td>
</tr>
<tr>
<td>3</td>
<td>none</td>
<td>4</td>
</tr>
<tr>
<td>5</td>
<td>0.8</td>
<td>40</td>
</tr>
<tr>
<td>9</td>
<td>0.8</td>
<td>40</td>
</tr>
</tbody>
</table>
parallel planes (Figure 47). If viewed from the side, one sees two rows of junctions separated by a distance $d$, and if the fluid has a constant, horizontally-homogenous (vertical) temperature gradient the output from the electronics would be a constant voltage, the sum of the outputs of the individual thermocouples. Flow studies \(^{15}\) indicated that "blocking" of the flow due to the tube does not occur for typical drop speeds, but that the lower row of junctions may perturb the flow enough to warrant the following design: the junctions of the lower row are bent inwards, around the edge of the support tube, while those of the upper row simply extend radially outwards (Figure 47). The wires between junctions went through small holes drilled in the tube and a thin layer of epoxy cemented them flat along the inside tube surface. Two insulated copper leads, connected to the first and last junctions in the series, communicated the thermopile voltage to the overboard instrument package. Total construction time for one thermopile was about a day.

**Sensitivity and spatial resolution**

The chromel-constantan thermocouple (ANSI Type E) has the highest emf output of any standard metallic thermocouple, $60 \times 10^{-6} \text{ V} (\circ\text{C})^{-1}$ at relevant temperatures. The gain used in the overboard electronics is

\(^{15}\) The flow was visualized using slow-motion photography and strongly illuminated aluminum flakes suspended in the flow. The flakes were perturbed by the bottom row of thermocouples but since the flakes were comparable in linear size to the wire diameter it is quite possible that the fluid flow was not itself affected. In any case, the thermocouples of the lower row should in future work be offset with respect to those of the upper row.
1000 so that for a 40-pair chromel-constantan thermopile with a total sensitivity of 2.4 mV °C⁻¹, the output is 2.4 V °C⁻¹. The gain is increased from this level, when needed, by onboard electronics. The noise level is set by the amplifiers used. Over a 30-Hz bandwidth, the amplifier noise represents about 4 x 10⁻⁵ °C.

The thermopile sensitivity was checked in the laboratory by calibration in a two-fluid "bath". Each row was in fluid of uniform but different temperature and the two fluids were separated by a thin layer of immiscible fluid. The measured sensitivity was about 15% lower than expected, an effect caused most probably by the rather crude set-up. The effects of axial heat conduction between junctions is calculated to be negligible on the basis of available formulae (Benedict, 1969; Blackwell and Moffat, 1975).

The spatial resolution of the thermopile depends upon the temperature structure of the fluid being measured, upon the response characteristics of each thermocouple junction and of course on the row-spacing and tube diameter. The former effect results from the fact that the temperature difference at the output is the sum of the temperature differences for each pair. Thus, small-scale variations in the fluid may cancel between pairs and not be seen at the output.

The nonideal frequency response of each junction is the result of time lags introduced by the thermal inertia of the junction, spatial averaging over the length of the junction, and signal attenuation in the boundary layer of water around each junction, the thickness of which depends upon the fall-rate of the probe. For 0.020 in. thermistors, time constants including the effect of attenuation due to the boundary
layer and electrical encapsulation are measured to be in the range 0.060 sec for a head-on speed of 8 cm/sec (Gregg, Cox, and Hacker, 1973) to 0.095 sec for 10 cm/sec (Aagaard, 1969). These figures should serve as upper bounds for the time constant of a 0.010-inch diameter junction. Balko and Berger (1968) give a formula for the time constant of very small junctions moving from air into water. Extrapolating their results to a 0.010-in. junction, a time constant of about 0.055 sec is found, which of course does not include the effect of the fluid boundary layer.

Difficulties encountered

Three difficulties arose in the field. The first was breakage of the 0.005-in. wires due to handling and "corrosion". Minor repairs can be made successfully by resoldering wires and corrosion can be minimized by washing the thermopiles in distilled water and drying them before storage. A 40-pair thermopile of 0.002-in. wire was judged marginal in terms of handling. A standard 40-pair probe of 0.005-in. wire is estimated to have a **minimum** life expectancy of thirty hours of field use.

The second problem, fouling by collection of gelatinous macroplankton, was encountered in the coastal waters off Oregon in the summer. It is unknown whether fouling occurred during the drop or while the instrument was being pulled back to the ship during recovery. The material was hosed off with a strong jet of water without doing damage to the junctions.

The cause of the final problem is not known for certain. The symptom is a slowly time-varying "offset voltage" appearing at the output
introducing a more-or-less linear trend into the chart recordings. Though an intermittent effect in any one drop, most drops do not have the problem, and in a series of repeated profiles the offsets can occur on one drop but not at all on the next few. It is possible that fouling could be involved.
In the following $T(z)$ will be the temperature, assumed to be dependent only on the vertical coordinate $z$ (positive downwards). Let $\Delta T(z)$ be the temperature-difference signal from the thermopile. Ideally,

$$\Delta T(z) = T(z+d) - T(z),$$

(1)

where $d$ is the (vertical) row-spacing. The sign convention used is that a stable temperature (temperature decreasing with depth) will yield $\frac{dT(z)}{dz} > 0$. The thermopile is a differentiator only for wavelengths $\lambda$ much larger than the row spacing, i.e., $\Delta T/d = \frac{dT}{dz}$ when $d << \lambda$. The temperature profile $T(z)$ can be reconstructed from either the differentiated thermistor signal $\frac{dT}{dz}$ or the thermopile signal $\Delta T(z)$; but it is of more interest to establish the relation between the respective (power) spectra.

The Fourier transform of (1) by the "shift theorem" is

$$\hat{T}(k) = T(k) (e^{i2\pi kd} - 1)$$

(2)

where the $\hat{}$ denotes the Fourier transform and $k$ is the vertical wave-number (cycles/cm). The transform of the temperature gradient is

$$\hat{\frac{dT}{dz}} = i2\pi k \hat{T}(k).$$

(3)

Comparing the absolute values of (2) and (3) it is found that

$$|\hat{\Delta T}(k)| = 2 \sin (\pi kd) |\hat{T}(k)|$$

(4)

The conversion between an estimate of the power spectra of $\Delta Y$, denoted $D_\theta(k)$, and the power spectra of $\frac{dT}{dz}$, $G_\theta(k)$, is:
Theoretically, the thermopiles give no output for wavelengths \( \lambda_n \) such that \( n\lambda_n = d, \, n=1,\, 2,\ldots \) [The \( \lambda_n = \frac{1}{k_n} \) produce the zeroes in denominator of (5).] For a typical row-spacing of 0.8 cm, \( \lambda_1 = 0.80 \text{ cm}, \, \lambda_2 = 0.40 \text{ cm}, \ldots \) The factor \( \sin^2(\pi kd)/(\pi k)^2 \) is the squared norm of the frequency response function (or the squared gain) of the thermopile output relative to the derivative. For \( d = 0.8 \text{ cm} \) the thermopile signal theoretically rolls off beginning at about 0.15 cy/cm (see Figure 55).
RESULTS

Thermopile transducers have been used successfully both in fresh water, Green Peter Lake, Oregon, U.S.A., and in the ocean off the Oregon coast. The water depth was about 100 m in each case and both had well-developed summer thermoclines. The lake had a typical lower-stability hypolimnion, while the ocean showed a rapidly changing, complex structure below the thermocline. Signals were recorded on either a two or four channel high-speed ($2.5 \text{ cm s}^{-1}$) chart recorder, in the latter case on two different sensitivities per signal which, in addition to on-board gain changes, helped make up for the lack of dynamic range. Select records were later digitized at the Oregon State University Computer Center.

(A summary of the transducers used is given in Table 9).

In situ sensitivity

Thermopile (Tp) sensitivity was determined in situ by comparison with the response of the differentiated thermistor (Th) to a thermal feature in the water column. The particular features chosen were of relatively large vertical extent (very much larger than the row spacing so the thermopile would act like a differentiator; this also assures that high-frequency signal attenuation not be a consideration for either transducer) and in the most stably stratified parts of the water column (to obtain horizontal homogeneity). These comparisons confirm
the expected sensitivity for the thermopile (examples are shown in Figure 48 and 49).

Records made using two concentric thermopiles also show coherent signals, independent of the tube diameter, the medium, i.e., fresh or salt water, and the drop speed. Figure 50 is an example from the ocean.

**Use of a thermopile in conjunction with a thermistor**

Discrepancies between simultaneous signals from a differentiated thermistor and a thermopile can be caused by (a) biological fouling--a problem recognizable at least after a drop; (b) poor water flow--eliminated; (c) vertical lengthscales comparable to the row-spacing--obvious from analysis; and (d) horizontal non-uniformity. The latter can result from relatively large-scale isotherm slopes; horizontal isotherms but a tilted instrument package, caused by recovery-line drag for instance; or horizontal thermal variability on a scale smaller than D, the tube diameter. If a Th record which shows turbulent small-scale fluctuations is filtered so that the smallest vertical wavelength is much larger than D, the correlation between fluctuations on a simultaneous Tp record remains poor, leading to the conclusion, which will be further supported by subsequent data, that lack of horizontal coherence is caused by small-scale thermal variability.

To see exactly to how small a wavelength the two signals are coherent, plots of coherence and phase against wavenumber can be examined. The coherence squared, $\gamma(k)^2$, can be interpreted as the fractional portion of the mean square value or variance contributed by the Th (input)
Figure 48. An example of *in situ* comparisons between the largest diameter thermopile (9 cm) and a differentiated thermistor. The data, cruise Y3 and drop 35 (Y3-35), is from the seasonal thermocline off the Oregon coast. The thermocline was sampled every few minutes, twelve times in all, (only seven shown) while the ship drifted a total of 1.5 km due south. For each "yo-yo" (B, C, E, F, G, I, J), the left trace is the Th and the right is the Tp. The scales shown apply to all the Tp traces but only approximately to all the Th traces. The depth variations of features from drop to drop are real.
Figure 49. Simultaneous records of the differentiated thermistor and thermopile (3 cm) obtained from the ocean (Y3-7A, 76-81 m) and from a fresh-water lake (GP7-5, 50-75 m). The mean gradient and buoyancy (Brunt-Väisälä) frequency for the ocean data are $5.2 \times 10^{-5}$ °C/cm and $7.6 \times 10^{-4}$ Hz (22 minute period) respectively, $3.8 \times 10^{-4}$ °C/cm and $6.7 \times 10^{-4}$ Hz (25 minute period) for the lake.
Figure 50. Profiles obtained using two thermopiles simultaneously ($N = 0.003$ Hz and $\frac{dT}{dz} \approx 0.02 ^\circ C/cm$):

(a) 1.5-cm diameter $T_p$ vs. 5-cm; (b) 1.5-cm vs. 9-cm. The temperature difference scale applies to all four traces and can be converted into an approximate vertical derivative scale of $\pm 0.02 ^\circ C/cm$ by dividing by the thermopile row-spacing (0.8 cm).
which appears in the Tp (output) at wavenumber k. If the records are completely unrelated, \( \gamma = 0 \); if \( \gamma = 1 \), they are completely coherent; and if \( 0 < \gamma < 1 \), there is some "noise" present in the system. Noise results from the digitization; some may be electronic if the gains were very high; and some is caused by the physical processes of mixing and stirring.

In Figure 51, for example, the Th-Tp coherence is generally high until a wavelength of about 10 cm is reached, at which point the coherence falls. Notice that the coherence is low when the wavenumber = \( 1/(\text{the thermopile averaging length, } D) = 1/3 \text{ cm} = 0.33 \text{ cy/cm} \). The ocean data has a slightly higher coherence at small vertical wavelengths, possibly because of the ocean's slightly higher stability \( (N = 6.7 \times 10^{-4} \text{ Hz}) \), but more likely because there is little power at small wavelengths at depth in the lake record.

The large confidence intervals at low wavenumbers for estimates of coherence and phase, shown only in Fig. 51, result from the limited depth range (about 10 m) of a profile through a specific regime (surface layer, thermocline, etc.). The 95% bias or significance levels (not shown in the figures) are typically 0.2 at the higher wavenumbers where a large number of degrees of freedom are available.

A series of "yo-yos" with the largest diameter thermopile through the highly stratified, seasonal, coastal thermocline (shown in Figure 48) is examined in Figures 52, 53 and 55; a series through a less well stratified, stable temperature inversion, possibly associated with the salt-stabilized warm anomaly which can exist below the thermocline (see
Figure 51. Coherence between thermopile (3 cm) and differentiated-thermistor records from the (•) ocean, and (○) a fresh-water lake (see Figure 49). Length of records used: 5.0 m and 23.1 m, respectively. The 95% confidence limits are given for the fresh-water data.
Huyer, Smith, and Pillsbury, 1974 for example) is examined in Figures 54 and 55.

For the therinocline data, the records are very coherent (Figure 52 until about 0.10 cy/cm at which point the variance levels in Th and especially Tp spectra (Figure 53 fall rapidly. Again the coherence is low when k = 1/D = 1/(9cm) = 0.11 cy/cm, supposedly due to mixing on length-scales smaller than D. The Tp and Th are in phase as expected (Figure 52), until k - 0.2 cy/cm at which point the records contain little meaningful signal. The gradient spectra (Figure 53 are typical of data from the thermocline in showing a nearly -1 slope at low k. Similar remarks apply to data from the temperature inversion (Figure 54). A comparison with the theoretical transfer function (Figure 55) shows for both data sets a premature roll-off caused by the large averaging length (D = 9 cm) of the thermopile.

As the averaging length is decreased, the coherence between Tp and Th should remain high until larger wavenumbers are reached, other things being equal. This is indeed the case in both the thermocline and the temperature inversion (Figure 56). On the other hand, it should not be expected that the roll-off wavenumber will decrease indefinitely as D is increased, since thermal features which have horizontal extent of 5 cm may very well extend 10, 20 . . . , cm. Thus, there appears to be little significant difference in the coherence for D = 5 cm vs D = 9 cm (Figure 56).

The horizontal coherence should be expected to be low where the probability of occurrence of mixing events is high (in the surface wind-stirred layer for example), and where large rates of energy dissipation
Figure 52. Ensemble-averaged (heavy curve) and individual coherence and phase plots for seven thermopile (9 cm)/thermistor yo-yos through the thermocline shown in Figure 48. \( N = 4.5 \times 10^{-3} \text{ Hz and } \frac{\partial \theta}{\partial z} = 0.4 ^\circ \text{C/m}. \)
Figure 53. Thermopile and differentiated-thermistor normalized, ensemble-averaged, autospectra for the thermocline profiles shown in Figure 48. Neither spectrum is corrected for attenuation effects. The thermopile spectra, $G_\theta(k)$, is offset one decade lower. The error bars are ± (one standard deviation).
Figure 54. Ensemble-averaged (heavy curve) and individual coherence and phase plots for seven thermopile (9 cm)/thermistor yo-yos (Y2-7D, E, F, G, H, I, J) through a stable temperature inversion ($N \geq 1 \times 10^{-3}$ Hz and $\frac{dT}{dz} \geq 0.05$ °C/m).
Figure 55. Squared norm of the frequency response function for (o) the thermocline profiles (Figures 48, 52 and 53) and (●) the inversion profiles (Figure 54). The heavy curve is the theoretical prediction for \( d = 0.8 \text{ cm} \).
Figure 56. Coherence between thermistor and thermopile of different diameters in: (top) the thermocline, (——) 3 cm (Y3-7A); (——) 5 cm (Y3-23A, B); (——) 9 cm (Y3-35B, C, E, F, G, I, J); (N=0.005 Hz and $\frac{3\pi}{3z} = 0.5 \, ^\circ C/m$), and (bottom) the temperature inversion, (——) 3 cm (Y3-7A, B); (——) 9 cm (Y2-7D, E, F, G, H, I, J); (N=0.002 Hz and $\frac{3\pi}{3z} = 0.1 \, ^\circ C/m$).
(large mixing intensity) create small turbulent length scales. Using the 3-cm Tp in the surface well-mixed layer downwards, first through a transitional zone, and then the thermocline, where infrequent mixing is expected due to high stability, and then in the deeper inversion, shows the expected behavior (Figure 57).

**Use of two thermopiles simultaneously**

The use of two thermopiles eliminates the effect of a (possibly) different time constant for the thermocouple junctions than for the thermistor bead. Plots show that as D is decreased the signal activity increases (Figure 50). Again, high coherence is expected only for those (vertical) wavelengths that are comparable to or larger than the (larger) thermopile diameter (see Figure 58), assuming a structure more-or-less isotropic on this scale.

**A horizontal thermopile**

A special-purpose thermopile was constructed with a group of four junctions diametrically opposed to another group of four so that a temperature difference is measured in a plane perpendicular to the thermistor. The temperature difference divided by the separation distance gives an estimate of the local horizontal gradient, assuming the instrument is vertically aligned. An examination of a typical section of data (Figure 59) shows that the plane of the thermocouples is inclined at an
Figure 57. Dependence of coherence between a thermopile (3 cm) and thermistor upon water-column stability (Y3-7A): (........) surface mixed layer, $N = 0$ Hz; (-----) transition to the thermocline, $N = 0.003$ Hz; (-----) thermocline, $N = 0.005$ Hz; (-----) warm-anomaly temperature inversion, $N = 0.002$ Hz.
Figure 58. Coherence between two thermopiles with $N = 0.003$ Hz (see Figure 50) but with D variable:

(----) 1.5 cm vs. 9 cm (Y3-22), and (——) 1.5 cm vs. 5 cm (Y3-21C).
Figure 59. Differentiated thermistor vs a horizontal thermopile (3 cm).

The noise level is evident on the Tp. (Data: YJ-19A, depth -13 m.) Five seconds is the rotation period of the instrument package.
angle in the range 4 to 12° to the isotherms (as measured by the ther-
mistor).

If the axis of the vehicle were tilted off vertical at a constant angle, the Tp output would be sinusoidal in a constant gradient region due to the (5-sec period) vehicle rotation. (For a variable gradient, the effect of rotation would still be evident in the records.) In those cases for which line drag was of negligible concern, no clear example has been found in which the Tp signal could be explained solely by considering the Tp to have sliced through a horizontal field of isotherms at some constant angle. The conclusion is that quite often, especially in recently mixed patches of water with vertical extent of order one meter, the vertical gradient is not horizontally homogeneous but instead the isotherms are tilted over rather short length scales and the tilt is typically 10 ±5°.
CONCLUSIONS

Unencapsulated thermocouple junctions can be used in serial differential fashion to give a measurement of the horizontally-averaged vertical temperature derivative in either freshwater or the ocean. The time response of the 0.026-cm junctions compare favorably with that of a small glass-encased bead thermistor. Comparisons between simultaneous thermopile and differentiated-thermistor records indicate good coherence for wavelengths typical of thermal finestructure, 10 cm to 10 m, but that coherence on the smaller microscales depends systematically upon the physical regime: the stability of the water column, and the probability of occurrence of small-scale mixing and stirring.

For relatively high values of the stratification parameter, the Brunt-Väisälä frequency, $N$, of the order $10^{-3} \text{ Hz}$ or larger, thermal features with vertical wavelengths $\lambda_V \sim 5 \text{ cm}$ or higher have a horizontal extent $\lambda_H \geq 9 \text{ cm}$, while for $\lambda_V \sim 2 \text{ cm}$, $\lambda_H \geq 3 \text{ cm}$. This is consistent with the work of Elliott and Oakey (1975) who find $\lambda_H \sim 50 \text{ cm}$ for $\lambda_V \sim 5 \text{ cm}$ in the less stratified Denmark Strait.

Loss of horizontal coherence is caused by thermal variability on scales smaller than the horizontal averaging length of the transducer, caused by active or recently active (fossil) mixing, or perhaps by a larger scale tilting of the isotherms. The tilt is measured to be on the order of $10^\pm5^\circ$ over horizontal scales of several centimeters. Elliott and Oakey (1975) measure an rms tilt of about $25^\circ$ over about half a meter, while Hacker (1973) observes an rms slope of $3^\circ$ over a horizontal scale of almost two meters.
Somewhat related measurements using two thermistors arranged parallel to one another but separated horizontally have been obtained by Gregg (1976, his Figure 18) and by us. We used as small a separation as possible (a few mm) but, not surprisingly, there were some active regions in the water column in which the correlation between signals was poor. Generally however, as was found in this paper, his coherence improves as \( N \) increases (and microscale mixing events are suppressed).

An interesting application of the horizontally-averaging thermopile might be their use in determining the horizontal covariability between temperature and for example chlorophyll concentration. Observations on large scales (greater than 40 m) indicate phytoplankton behave as a passive contaminant of the fluid motion (e.g., Fasham and Pugh, 1976). Our work has shown that strong vertical temperature (and presumably salinity) gradients with small vertical extent continue in the "horizontal" over the thermopile diameter (up to 9 cm) and often from profile to profile (up to 50 or 100 m) to form rather long-lived interfaces. In the epipelagic zone plankton may react to variability of this order and group themselves in accordance with these interfaces to form micro-distributions (Cassie, 1959) and a thermopile as described herein may be useful in a study of such phenomena.
VI. CONCLUSIONS

This study has been a first attempt at a description of some of the smaller-scale mixing processes active in the upwelling region off Oregon as "imprinted" in the temperature field. As expected, a single temperature profile is not smooth. The presence of thin, statically stable interfaces with temperature gradients very much larger than the local vertically averaged gradient, and rapid time-space changes in the thermal structure are evidence of stirring processes. On the other hand, there can be a conspicuous absence of thermal "activity" in regions extending many meters in the vertical. The explanation for this is the suppression of turbulence by very strong stability (in parts of the seasonal thermocline for example), or, in homogeneous regions away from boundaries, the presence of a stirring so weak that entrainment of water of a different temperature into the region is not possible (the cool ribbon "core" is apparently an example of this).

Shear-induced stirring seems to dominate throughout the water column -- in the surface layer to the bottom boundary (turbulent Ekman?) layer. (Double-diffusive layering has been found to not be a significant process in this area.) Moderate-Reynolds-number turbulence occurs in patches of relatively small vertical extent and produces a narrow inertial subrange and decade long viscous-convective subrange in the temperature gradient spectra. In many cases, the billow turbulence may be initiated via growing Kelvin-Helmholtz instabilities.
Relatively homogeneous layers are observed to have a vertical scale about equal to the Ozmidov length.

Unencapsulated thermocouple junctions used in differential fashion are a usable tool in this kind of work. When arrange in a horizontal array they show that the horizontal coherence of thermal features with a vertical scale of less than about ten centimeters is limited by small-scale mixing processes, the occurrence of which depends on the water column stability and larger-scale stirring processes.

Further microstructure work with a full complement of sensors -- velocity, salinity, temperature, optical density, and of course pressure -- should be undertaken with an aim to study the time-history of significant features, such as the cool ribbon and the bottom boundary-layer. A need for microstructure measurements of the stability of the offshore flow (e.g., the warm anomaly) has recently been expressed. A small freely drifting submersible would be ideal for these studies. Vertical profiling while following a transponder embedded in the flow of interest would be another approach. As long as both small-scale structure and the larger-scale structure (including local circulation and meteorological forcing) are simultaneously obtained, the results of microstructure programs should be applicable to other areas where dynamically similar large-scale features are observed.
VII. REFERENCES


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