Turbulence Observations in the Northern Bight of Monterey Bay from a Small AUV

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1. Introduction

Modern, in situ, measurements of turbulence in a marine environment are generally attributed to Grant and colleagues (Grant et al., 1959) who used hot-film anemometry to estimate the turbulent dissipation rate from a velocity spectrum. These measurements were performed either in a bottom mounted or in a towed configuration. However, because of various deployment based contamination effects, this approach was quickly abandoned in favor of using vertical free-fall profilers. These profilers were equipped with the Siddon (1965) thrust probe to estimate directly the turbulent dissipation rate (Osborn, 1974). This latter approach to turbulence measurements provided very high vertical resolution (scales of order meters and less) and was often made concomitantly and on the same platform with sensors capable of obtaining finescale density and velocity.

The free fall profiling instruments were considerably simpler than the previously used towed bodies and were...
completely decoupled from ship motion. This resulted in a much higher quality of turbulence data. However, they were very limited in the type of data which they could obtain and provided essentially a one dimensional, one time vertical characterization of the turbulent field. Increasingly, observations began to suggest that ocean turbulence was intermittent in time and space and, in regimes of strong stratification, tended to have a noncontiguous, “pancake-like” structure (Woods, 1968; Gregg, 1987) with the conclusion that the vertical profiler based measurements would not always be able to resolve such structures (Yamazaki et al., 1990).

Efforts were made in the 1990s as a part of the Coastal Mixing and Optics experiment to use loosely tethered rapidly cycling profilers (Oakley and Greenan, 2004) to obtain the horizontal spatial structure of the turbulent field. In the past, horizontal sampling, using towed bodies and submarines, had provided unique views of the phenomena of internal waves (Gargett, 1982), salt fingers (Fleury and Lueck, 1992), and, in some limited cases, turbulence (Osborn and Lueck, 1985).

Recently, a number of alternative techniques and platforms have been developed and used to measure ocean turbulence. For a review of such techniques see Lueck et al. (2002). One platform which is particularly useful for inferring the turbulent dissipation rate, $\varepsilon$, to a very high resolution, both horizontally as well as vertically, is an AUV (Dhanak and Holappa, 1999; Levine and Lueck, 1999).

A number of such mobile platforms have now been developed. These include: (1) the Large Diameter Unmanned Underwater Vehicle (LDUUV) (Levine and Lueck, 1999); (2) the Ocean Explorer (Dhanak and Holappa, 1999); (3) T-REMUS AUV Gen I (Levine et al., 2001; Goodman et al., 2006); (4) AutoSub (Thorpe et al., 2002); (5) the Bluefin Robotics Odyssey III AUV (Wijesekera et al., 2003); and (6) T-REMUS AUV Gen II (MacDonald et al., 2007). With the exception of the LDUUV and the AutoSub, these vehicles are small, of order 2 to 4 m in length.

The first effort using an AUV for turbulence measurements involved the Large Diameter Unmanned Underwater Vehicle (LDUUV) (Levine et al., 1996; Levine and Lueck, 1999). This vehicle is 7.6 m long with a maximum diameter of 0.66 m and mass of 2000 kg. The noise floor of dissipation rate obtained with this vehicle was $10^{-9}$ W/kg (Lueck et al., 2002). Following this work, three smaller AUVs were developed for measuring turbulence. These were: the Ocean Explorer, REMUS AUV Gen I, and the Bluefin Robotics Odyssey III AUV. Observational results showed that these smaller vehicles could also be used to make high quality turbulence measurements with dissipation rate noise floor as low as $10^{-9}$ W/kg (Goodman et al., 2006). The small AUV has the additional advantage over larger ones in being more maneuverable and being able to be deployed in both a constant depth as well as a yoyo mode.

The AUV used in this manuscript is the second generation T-REMUS. Like T-REMUS Gen I, this vehicle is outfitted with turbulent shear probes, FP07 thermistors, and a variety of “hotel” sensors. By employing a procedure which corrects sensor vibration and vehicle motion effects arising from all directions, Goodman et al. (2006) have shown that with this configuration a dissipation rate noise floor of $\varepsilon \approx 10^{-9}$ W/kg can be achieved.

In this manuscript we demonstrate the usage of a small AUV to obtain the spatial structure of turbulence, both in the horizontal as well as vertical, in a shallow coastal environment. Direct estimates not only of the turbulent dissipation rate but also of the buoyancy Reynolds number are made. The latter quantity is an indicator of the spatial dynamic range of the turbulent field. These turbulent quantities are obtained simultaneously with high resolution measurements of temperature, salinity, and current. This allows a detailed description of the turbulent field and its relationship to the surrounding larger scale environment.

We will examine a specific set of AUV based measurements obtained as a part of the 2006 Layer Organization in the Coastal Ocean (LOCO) experiment. These were made on 19:00 PDT 17 July 2006 to 03:00 PDT 18 July 2006 in a shallow region of Monterey Bay. Three large scale features are identified which contributed to the development of the turbulent field. These are: (1) a warm near surface intrusion; (2) the bottom mixed layer; (3) a large amplitude, nonlinear internal wave train. We explore the spatial structure of the turbulence in these three regimes. Particular emphasis will be given to the nature of the turbulence induced by the internal wave train. To that end we will develop a simple theory, following the approach of Moum et al. (2003), which examines the role of the internal wave induced vertical strain gradient on producing the turbulent field.

In Section 2 we describe the T-REMUS vehicle and its sensor systems. Section 3 gives an overview of the deployment strategy. In Section 4 we discuss the local LOCO environment with particular emphasis on the internal wave field. Section 5 contains the description of the T-REMUS data obtained. In Section 6 we discuss and interpret the turbulence data in terms of surrounding larger scale environment. We examine in some detail our observation that intense turbulence in the thermocline was associated with the trailing edge of an internal wave train. We present an explanation of why this occurred. Section 7 contains the summary and conclusions.

2. T-REMUS and its sensor suite

The observational approach is to use the Autonomous Underwater Vehicle, T-REMUS, shown in Fig. 1. T-REMUS is a custom designed REMUS 100 vehicle manufactured by Hydroid Inc., containing the Rockland Microstructure Measurement System (RMMS), an upward and downward looking 1.2 MHz ADCP, a FASTCAT Seabird CTD, and a WET Labs BB2F Combination Spectral Backscattering Meter/Chlorophyll Fluorometer. In

![Fig. 1. The SMAT T-REMUS Autonomous Underwater Vehicle. It is 2 m long with a mass of 63 kg. Vehicle based sensors are indicated in the figure.](image-url)
addition, the vehicle contains a variety of “hotel” sensors which measure pitch, roll, yaw, and other internal dynamical parameters. Table 1 shows the sensor location and sampling rate of the major sensors of the T-REMUS vehicle.

This suite of sensors on T-REMUS allows quantification of the key dynamical and kinematical turbulent and finescale physical processes. The turbulence measurements are made concomitantly with very high spatial resolution measurements of temperature, salinity, and depth. The BB2F sensor system measures chlorophyll fluorescence and optical backscattering at 470 nm and 700 nm wavelength. The turbulent and finescale parameters which can be estimated from the data collected by the T-REMUS include: the turbulent dissipation rate, the buoyancy Reynolds number, finescale velocity shear, and finescale stratification. We have also shown (Goodman et al., 2006) that heat flux can be calculated directly from the RMMS derived vertical velocity and temperature data. Below we provide more details on the T-REMUS measuring system suite.

2.1. Rockland Microstructure Measurements System (RMMS)

The RMMS package designed by RGL Consulting Ltd. was mounted to the left side of the nose of the vehicle. It is housed in a pressure case of 6 cm diameter and is approximately 0.5 m long. It consists of two fast response thermistors (Thermometers FP07) to measure temperature and its derivative and two shear probes, mounted 26.94 mm apart from each other on the front end of the pressure case. The shear probes provide an output proportional to the cross-stream velocity gradient $\partial v/\partial s$ and vertical velocity gradient $\partial w/\partial s$, where $s$ is the along track distance. The electronics board of the RMMS supports a set of tri-axial accelerometers oriented along the three principal axes of the pressure case, and a pressure transducer that provides both pressure and its derivative. All the microstructure (turbulence) data are collected at a sampling rate of 500 Hz. Goodman et al. (2006) describe a technique of coherently subtracting all three components of accelerometer data from the shear data to obtain an optimal estimate of the dissipation rate. Using this technique the underway effective noise floor for the dissipation rate $\epsilon$ is found to be less than $10^{-9}$ W/kg.

2.2. Conductivity, Temperature, and Depth (CTD)

The SBE 49 CTD by Sea-Bird Electronics, Inc was mounted on the right front of T-REMUS. It is an integrated CTD sensor system specifically designed for usage by either a towed body or an AUV. The SBE-49 CTD has a pump-controlled, TC-ducted flow which minimizes salinity spiking. The 16 Hz sampling rate results in very high spatial resolution, typically of order centimeters. The in situ CTD collected temperature and depth data are used to calibrate temperature and depth data obtained from the RMMS probes.

2.3. Acoustic Doppler Current Profilers (ADCP)

The REMUS 100 AUV comes standard with a 1.2 MHz 20° beam angle RDI Acoustic Doppler Current Profiler/Doppler Velocity Log (ADCP/DVL) with up and down looking transducers. This system provides both the vehicle speed and direction over the bottom for the dead reckoning navigation system as well as the water column velocity. Both of the ADCP systems have a blanking distance of approximately 1.1 m. With the vehicle diameter of approximately 0.2 m there is a 2.4 m region surrounding the vehicle where velocities cannot be measured. The vertical bin size used in the experiment was 0.5 m. Velocity measurements near the bottom and water surface are not attainable due to reflections that contaminate the water ping. Fong and Jones (2006) evaluated AUV based ADCP measurements by comparing vehicle based data with that from a nearby bottom mounted ADCP. They found good overall agreement, although they did observe a significant bias error in the direction of vehicle motion, of order 10 cm/s. This bias error was also found not to be depth independent and thus is very significant for in situ velocity shear estimates. Fong and Jones (2006) were not able to identify the source of these errors. However, they recommended that vehicle transects be located perpendicular to the dominant flow direction to minimize this bias. For weak current measurements of order 10 cm/s and less they recommended taking repeated transects in opposing directions to compute an averaged unbiased velocity. We employ this method to minimize this bias effect.

2.4. Combination backscattering meter and fluorometer (BB2F)

This is a combination sensor, manufactured by WET Labs, for concurrent determination of optical backscattering at 470 nm and 700 nm and chlorophyll fluorescence. These are all performed within the same scattering volume. This sensor system played a critical role in the AUV based LOCO field experiments in identifying regions of thin plankton layers and

<table>
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<td>RMMS</td>
<td>Thrust probe</td>
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<td></td>
<td>FP07 thermistor</td>
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<td></td>
<td>Pressure transducer</td>
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<td>Accelerometer</td>
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<td>Seabird CTD</td>
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<td>BB2F</td>
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T-REMUS sensors. The nomenclature used are: s, along track distance; u, along track velocity; v, athwartship velocity; w, vertical velocity; T, temperature measured; P, pressure measured; ax, ay, az, acceleration in the along track, athwart track, and vertical direction; S, salinity; D, depth; Beta470 and Beta700, optical backscattering at 470 and 700 nm in units of B/m/sterv; Chl $a$, chlorophyll a in units of $\mu g/l$. |
relating these to the turbulent and fine scale. We will be reporting on these measurements in a subsequent manuscript. The BB2F system is also very useful as a surrogate tracer to help identify small scale intrusions.

3. Experimental site and deployment strategy

The location of the experiment within Monterey Bay is shown in Fig. 2a with the detailed tracks of the T-REMUS AUV shown in Fig. 2b. The experiment was centered at 36.93N, 121.92W, in the northern bight of Monterey Bay. Average depth of the water in the area was 19 m. There were 12 continuous AUV tracks which ran parallel to the fixed LOCO observatory stations K0 to K4. These tracks were approximately perpendicular to isobath contours. Each track, 2.5 km in length, took 40 min and alternated between an outbound one and an inbound one. The AUV operated in a 5 degree yoyo mode with a ground speed of 1.2 m/s. AUV depth ranged from 1.0 m from the surface to 4.0 m above the bottom. Each track consisted of approximately 16 profiles with the horizontal distance between profiles being on average 150 m apart. Observations began 19:00 PDT 17 July 2006 and ended at 03:00 PDT 18 July 2006. The bathymetry of the experimental site, Fig. 2b, shows water depth ranging from 16 m to 23 m, uniformly sloping across the shelf.

4. The local environment

Monterey Bay is one of the strongest upwelling regions along the west coast of Northern America. The LOCO 2006 T-REMUS measurements occurred in July at the end of the upwelling season, which occurs from late February until late July (Graham, 1993).

In Fig. 3a, b, and c we show for the experimental period, respectively, air temperature and sea surface temperature, wind speed, and direction. These data were obtained from the nearby (within 18 km) R/V Thompson, which also participated in the LOCO experiment. The meteorological data in Fig. 3a, b, c, are courtesy of T. Cowles. Fig. 3d is the calculated net heat flux through the air–sea interface using the data from Fig. 3a–c and assuming an average incident short wave radiation for that location and for the time of the day of the experiment. Positive values indicate gaining heat from the air and negative values indicate losing heat. The heat flux quickly drops to negative values after sunset at about 20:00 PDT. Fig. 3e is the tidal height obtained from the tidal station “Santa Cruz”, located at 36.96N, 122.02W. Fig. 3a shows that the difference between air and sea surface temperature is very small, less than 1.5 °C, over the 8-hour experimental period. Fig. 3b and c show very low wind speed and a near steady direction from the southwest. This results, along with the small air water temperature difference, in the net air/sea heat flux after sunset being very low, although slightly negative.

The tides in Monterey Bay are mixed. The K1, O1, M2 and S2 constituents contribute approximately 80% of the tidal variation (Breaker and Broenkow, 1994). The tidal height in the LOCO site region on July 17th–18th 2006 is about 1.5 m changing from high tides to low tides (Fig. 3e) during the experimental period. Tidal currents at the LOCO site cannot be determined from our ADCP measurements but are expected to be very small (<5 cm/s).

From ADCP measurements, McManus et al. (2005) found at the LOCO site that the total non de-tided currents to be relatively weak, with maximum values less than 12 cm/s. Fig. 4 shows that the magnitude of the current obtained from the T-REMUS upward and downward ADCP to be of order of magnitude of this previous result. In Fig. 4a we present the mean current profile, obtained by averaging over the entire data of 12 T-REMUS track legs. The left hand plot in Fig. 4a is

![Fig. 2.](image-url)
the across slope current, the right hand plot the along slope current. The error bars shown in Fig. 4a represent 1 standard deviation. Note that error bar magnitude decreases with increasing depth with a minimum occurring near mid depth and showing a maximum near the bottom. This is a result of our yoyo sampling strategy with more data points near mid depth. To obtain a time series of current magnitude and direction, we average over sequential outbound and inbound tracks to minimize along track bias (Fong and Jones, 2006).

Fig. 4b and c show, as a function of time, vertically averaged current magnitude and direction (relative to north). Because of the inbound and outbound averaging, each point in Fig. 4b and c are 1 h 20 min apart. The vertically averaged current magnitude is noted to increase slightly, from 0.08 m/s to 0.10 m/s during the eight hour time period of the experiment. With the very low wind speed and air-sea interaction, these results suggest that advective effects are a result of the larger scale circulation and not locally wind driven.

Fig. 5 shows the temperature field of the experimental site, obtained from the T-REMUS CTD measurements averaged over the 12 T-REMUS tracks shown in Fig. 2b. The black vertical line is the across isobath location of a thermistor chain located at site K1 of Fig. 2b, approximately 500 m perpendicular to the AUV tracks to the northwest. Note the downward sloping of the isotherms and the corresponding warming of the water shoreward. Maximum salinity variation over the LOCO site is of order 0.06 psu and thus density variation is principally determined by temperature variation. Fig. 6a and b shows the average density profile and buoyancy frequency profile over the site. The black dashed line is the vertical structure of the lowest internal wave mode calculated from linear theory using the average buoyancy profile shown in Fig. 6b.

4.1. Internal waves

The LOCO site is northeast of the Monterey Canyon, a source of large amplitude internal solitary like waves, (Lien and Gregg, 2001; Carter et al., 2005; McManus et al., 2005) and thus is expected to be a region of intense internal wave activity. In Fig. 7 we show a time series of isotherms spanning temperatures of $T = 11.2 \, ^\circ C$ to $T = 14 \, ^\circ C$, spaced every 0.5 $^\circ C$. This data comes from the thermistor chain at station K1. The depth of each thermistor is indicated by the green dots. Data courtesy of D.V. Holliday. The sampling rate is 1/30 cycle/s and covers the exact same time period as that of the T-REMUS observations.

Note the intrusion of warm water beginning at approximately $t = 20:45$ PDT, clearly seen in the downwelling of isotherms $T = 12 \, ^\circ C$ to $T = 14 \, ^\circ C$ of Fig. 7. The appearance of a large amplitude, isolated, internal solitary wave, which we label A, first occurs at $t = 21:37$ PDT and is of order 8 min in duration with a maximum downward displacement of 3.2 m. (The usage of the term “large” for amplitude is somewhat arbitrary but is
based on the magnitude of the amplitude relative to the water depth.) A second large, isolated, internal solitary wave, labeled, B, of order 7 min in duration and 4 m maximum downward displacement, occurs at $t = 22:25$ PDT. Following event B, beginning at $t = 23:05$ PDT, there is a continuous train of large amplitude, downwardly displaced, internal solitary waves, on average 25 min apart. In Fig. 7 we label 9 of these wave features, which have maximum displacements of order 1.5 m. They first occur at $t = 23:05$ PDT and last until almost the end of the experiment at $t = 02:34$ PDT. These wave train events vary in their individual duration from 7 min to under 5 min, slightly decreasing with time. In Fig. 8 we show the vertical profile of the downward displacement of event A, blue circles, event B, green circles, and the mean downward displacement of the 9 members of the identified wave train, black circles. Event B and the wave train show a near monotonic decrease of amplitude with depth. Event A shows a maximum at $z = 6$ m close to the

**Fig. 4.** Averaged current and direction obtained from the upward and downward ADCPs mounted on the T-REMUS. a) Average along isobath current (right hand profile) and across isobath current (left hand profile) shown with the one standard deviation (67%) error bars. Averaging is over the duration of the experiment, 8 h; b) vertically averaged current speed, and c) current direction, both as a function of time. Data in b) and c) are averaged over sequential outbound and inbound legs to minimize the along track bias (Fong and Jones, 2006).

**Fig. 5.** Map of water temperature averaged over the 12 AUV tracks. The abscissa is the horizontal distance, $s$, offshore relative to the initial reference point at (36.94N, 121.91W), shown as the filled black circle in Fig. 2b. The ordinate is water depth in meters. The scale to the right represents temperature (°C). The black lines are isothermals spaced at an interval of 0.2 °C. The numbers at the top of the contour map indicate profile number, located at their average location. The vertical thick black line shows the location of fixed K1 station. Bottom is indicated by the thick black line.

**Fig. 6.** Profile of sigmat, $\sigma_t$ (left panel), and buoyancy frequency (right panel) obtained from the data in Fig. 5. The light dotted line is the lowest internal wave mode (eigenfunction) obtained using linear internal wave theory with the buoyancy profile in (b).
location of the maximum in the lowest mode wave function shown in Fig. 6b.

5. T-REMUS hydrography and turbulence observations

We show in Fig. 9 temperature contour plots obtained from the T-REMUS CTD, with each of the 12 panels corresponding to the respective 12 track legs of Fig. 2b. The vehicle underwent a 5-degree yoyo, resulting in the profiles being on average 150 m apart horizontally. The initial time of each track leg is shown at the bottom of the figure in each of the 12 panels. The profile number of each T-REMUS yoyo is shown at the top of each figure, referenced to the first profile taken along the track. Thus, the first profile will alternate from being the most shoreward to the most seaward. The temperature contour map shown in Fig. 5 was obtained by averaging the 12 contour maps of Fig. 9.

A number of features are apparent in these panels of contour maps. First, note the warm near surface $T = 14$ °C layer which intrudes seaward with increasing panel (leg) number. With increasing panel number, the observations are occurring at a later sampling time. The use of Taylor's hypothesis and the observed along isobath current structure shown in Fig. 4 can be used (within the caveats of its assumption of a “frozen”
Fig. 9. Twelve contour plots of temperature (°C) as a function of across isobath range, $s$, and depth, $z$. Each contour was obtained on the 12 legs shown in Fig. 2b. Black contour lines show isotherms with 0.2 °C spacing. Profile number notation is the same as that in Fig. 5. The leg number and beginning time of each leg are showed at bottom of each figure. Isotherms of $T = 11, 11.2, 11.4$ and 14 °C are emboldened in white. The across isobath location of the K1 thermistor chain is shown by the thick black line.
Fig. 10. Contour plots of log$_{10}$(\(\varepsilon\)), where \(\varepsilon\) (W/kg) is the turbulent kinetic energy dissipation rate. Black contour lines show the isotherms with 0.2 °C spacing. The same isotherms are shown as in Fig. 9, with the same notation.
field) to associate time with an upstream distance. For an average current speed of 9 cm/s, panel #12 would then represent observations made 2.5 km upstream to that of panel #1. Note the intrusion of warm surface water at the location of K1 at approximately the same time, \( t = 20:45 \) PDT, when it was observed in the isotherm time series shown in Fig. 7. Panel #4 of Fig. 9 shows that at approximately 21:30 PDT a very strong sharpening of the warm water intrusion occurred. This resulted in a very sharp, small scale frontal feature characterized by a temperature gradient of 2 °C/300 m.

Within the thermocline, panels 5 and 6 of Fig. 9 show large amplitude isolated isotherm displacements near the location of K1, at a depth of \( z \approx 8 \) m. The subsequent panels show increasingly large amplitude variability at the base of the thermocline.

Finally, we note the relatively thick bottom mixed layers present in all of the panels. In particular, note the progression upslope, with increasing panel number, of the deep water isotherms 11.2 °C and 11.4 °C, outlined in white and located in the lower left hand corner of the panels. These isotherms eventually move of order 500 m upslope. This feature is suggestive of enhanced boundary layer mixing and an upslope buoyancy flux (e.g. Trowbridge and Lentz, 1998).

In Fig. 10, we show dissipation rate, \( \varepsilon \), measurements obtained from the on board RMMS system. These are obtained by using all three components of accelerometers to minimize vehicle motion and vibrations in estimating the turbulent shear spectra and then fitting the corrected spectra with a universal Naysmith spectrum (Goodman et al., 2006). With this technique we achieve a noise floor less than \( \varepsilon \approx 10^{-9} \) W/kg. Estimated \( \varepsilon \) are obtained by averaging over 20 samples, each of horizontal length 0.25 m. This corresponds to a 5 m horizontal and a 0.5 m vertical averaging distance. For a discussion of this averaging technique and its associated statistics including effective number of degrees of freedom see MacDonald et al. (2007). The panels in Fig. 10 show 3 regimes of high dissipation rate with \( \varepsilon > 10^{-7} \) W/kg. These are: (1) a near surface regime associated with the warm water intrusion; (2) a bottom mixed layer regime; (3) a mid depth regime of strong vertical density (temperature) gradient. Note the very strong isolated patch of turbulence in what looks to be in the lee of an internal solitary wavelike feature. This location of a turbulence patch relative to the sharpening of the near surface warm water intrusion.

In Fig. 11 we show the corresponding panels of the buoyancy Reynolds number defined by

\[
Re_b = \frac{\varepsilon}{\nu N^2} = \left( \frac{\Gamma_b}{\Gamma} \right)^{\frac{4}{3}}
\]

where \( \nu \) is kinematic viscosity,

\[
\Gamma_b = \left( \frac{\varepsilon}{\nu} \right)^{\frac{1}{3}}
\]

is the buoyancy or Ozmidov length scale and

\[
\Gamma = \left( \frac{\varepsilon}{\nu} \right)^{\frac{1}{3}}
\]

the Kolmogorov length scale. \( Re_b \) is a measure of the turbulent spatial dynamic range and is a better indicator of the strength of the turbulent field in a stratified fluid than \( \varepsilon \). For example, from Fig. 6b, at the depth of maximum of the buoyancy frequency at the top of the thermocline we have \( N_{max} = 0.04 \) s\(^{-1}\). A dissipation rate of \( \varepsilon = 3 \times 10^{-8} \) W/kg would correspond to \( Re_b = 20 \), a value sufficiently low where turbulence is thought not to exist because of the strong damping affect of buoyancy (Itweire et al., 1986). We will use the term "strong" turbulence to refer to turbulence characterized by \( Re_b > 200 \) and \( \varepsilon > 10^{-8} \) W/kg. The criterion of \( Re_b > 200 \) is used since under this condition isotropy of the turbulent field is thought to occur (Yamazaki and Osborn, 1990). The additional usage of the dissipation rate criterion is somewhat arbitrary and used to rule out large values of \( Re_b \) due to near isothermal conditions. Using the definition of a mixing efficiency

\[
\Gamma = \frac{B}{\varepsilon}
\]

where \( B \) is the turbulent buoyancy flux we can express the vertical diffusivity \( \kappa_p \) as

\[
\kappa_p = \frac{B}{N^2} = \nu \Gamma Re_b
\]

Thus the panels in Fig. 11 can be also interpreted as that of vertical diffusivity. If we use the standard value of \( \Gamma = 0.2 \) (Gregg, 1987; Osborn, 1980), we see that \( \nu \Gamma = 2 \times 10^{-7} m^2/s \) and the scale for \( \kappa_p \) of the data in Fig. 11 would span four orders of magnitude from \( \kappa_p = 2 \times 10^{3} m^2/s \) to \( \kappa_p = 2 \times 10^{7} m^2/s \), the latter near the molecular floor value of thermal diffusivity.

Figs. 10 and 11 show that the strongest values of turbulence occur near the bottom and appear throughout all of the panels. There is also very strong turbulence near the sharp surface frontal feature of panel #4, which resulted from the sharpening of the near surface warm water intrusion. Strong turbulence in the thermocline mainly occurs in panels 7 to 12.

6. Discussion

The finescale AUV measurements, collocated and made simultaneously with the turbulence measurements, allow examination of the local processes which produce and maintain the turbulent field. The experiment took place in very light winds with virtually no air–sea interaction. As indicated by the ADCP measurements in Fig. 4 local barotropic tidal variation is also expected to be small. Figs. 9–11, along with the thermistor chain derived data shown in Figs. 7 and 8, indicate that there are three features in which the strongest turbulence occurred. These are: (1) the near surface warm water intrusion; (2) the bottom boundary layer; (3) the large amplitude internal wave train. We will examine the spatial structure of each of these regimes and identify their most probable mechanism of generation and maintenance. Particular attention will be given to the turbulent field arising from the internal wave train.

6.1. Near surface warm water intrusion

One of the most striking aspects of this data is the intrusion of near surface warm water. This intrusion is very shallow and, at times, did not appear in the T-REMUS
Fig. 11. Contour plots of buoyancy Reynolds number \( \log_{10}(Re_b) \). These plots also can be considered to be that of \( \log_{10}(\kappa) \) where \( \kappa \) is the eddy diffusivity, related to \( Re_b \) by \( \kappa = \Gamma v Re_b = 2 \times 10^{-7} Re_b \) (m²/s), where \( \Gamma \) is the mixing efficiency and \( v \) is the molecular viscosity. The same isotherms are shown as in Fig. 9, with the same notation.
hydrography data shown in Fig. 9, since the shallowest depth reached by the vehicle was 1 m below the surface. Nevertheless, it is clear that water of temperature $T = 1 \, \degree C$ intruded offshore with time, which by Taylor’s hypothesis can be identified with increasing upstream distance. The northwest mean surface flow direction (Fig. 4) had a significant cross isobath component in the same direction as that of this intrusion. Downwelling of the $T = 12 \, \degree C$ to $T = 14 \, \degree C$ isotherms in Fig. 6 show that the intrusion begins to occur at site K1 at approximately 20:30 PDT, around the same time that it was observed in the T-REMUS hydrography data in panel #3 of Fig. 9. The near surface warm water intrusion extends entirely across panel #12 by 02:10 PDT having moved 1.7 km during the time of the experiment. From this information we can estimate an advection velocity $v_1$ of

$$v_1 \approx \frac{1.7 \, \text{km}}{5.7 \, \text{h}} = 0.3 \, \text{m/s},$$

a value somewhat larger but close to the estimated ADCP across isobath near surface velocity of $v = 5 \, \text{cm/s}$, shown in Fig. 4. This relative agreement of estimates seems to support the idea of a strong across isobath surface flow driving this intrusion. However, we cannot rule out the possibility that this occurrence of warm water appearing increasingly offshore with time was the result of an along isobath downstream directed advection of warm water which happened to vary in the cross shore direction. We do not have data on upstream temperature sections to distinguish these two possibilities for the source of the warm water intrusion.

Careful inspection of Figs. 10 and 11 shows that strong turbulence occurs at the base of this intrusion. To illustrate this, we show in Fig. 12 density and turbulent shear profiles used to obtain the contour maps of panel #8 of Figs. 10 and 11. Note the large values of turbulent shear at the base of the warm layer indicated in profiles 6 through 17, of Fig. 12, supporting the idea of a shear driven interfacial instability associated with the warm water intrusion. From Fig. 4 we observe that the mean across isobath surface to be

$$\frac{\text{d}u}{\text{d}z} = 4 \times 10^{-2} \, \text{m/s}^{-1} \quad \text{at} \quad 1 \, \text{m},$$

which along with the corresponding mean value of $N$ immediately below the surface layer, given in Fig. 6, results in

$$R_i = \frac{N^2}{(\text{d}u/\text{d}z)^2} = \left(\frac{0.04}{0.04}\right)^2 = 1,$$

which, within the uncertainty of this calculation and taking into account that these values are averaged over the entire data set, is not far from the theoretical value of shear induced instability at $R_i = 0.25$.

6.2. Bottom mixed layer

From Fig. 11, using as a criterion the buoyancy Reynolds number, the strongest turbulence observed throughout this experiment was in the offshore location of the bottom mixed layer. Panels 7–12 of Figs. 10 and 11 show that toward the end of the experiment bottom mixed layers heights tended to increase. During the middle of the experiment from $t = 21:00$...
PTD to \( t = 23:00 \) PTD, as shown panels 4–6 of Figs. 10 and 11, the weakest bottom mixed layers occurred.

Note in Fig. 9 the apparent movement upslope with increasing panel number of the deeper bottom mixed layers. This is readily seen by following a near bottom isotherm, for example, the \( T = 11.2 \) °C isotherm, initially located in panel #1 at the very lower left hand side of the panel. Consider the along slope location in which this isotherm intersects the bottom. Let \( \bar{d}r \) be the across isobath distance which the isotherm bottom intersection point travels over time period \( \bar{d}t \). Using the data in Fig. 9 we calculate the speed, \( V_i \), in which the \( T = 11.2 \) °C isotherm moves along the bottom and find that

\[
V_i = \frac{\bar{d}r}{\bar{d}t} = \frac{900 \text{m}}{7.75 \text{h} \cdot 3600 \text{s/h}} = 3 \text{ cm/s}
\]

This upsource movement of the isotherm is a result of a change in temperature \( \bar{d}T \) over some volume \( V \) due to a flux of heat downslope or, equivalently, a flux of buoyancy upslope. Let the volume of such a temperature anomaly \( \bar{d}T \) be given by

\[
\bar{d}V = l_p l_d \bar{d}r
\]

where \( l_p \) is the along isobath length of the temperature anomaly. The rate of heat transferred, \( \bar{d}q/\bar{d}t \), upslope is then given by

\[
\frac{\bar{d}q}{\bar{d}t} = \rho_0 c_p \bar{d}V \frac{\bar{d}T}{\bar{d}t}
\]

where \( c_p \) is the specific heat. If a near steady state condition exists then the heat fluxed into the interior should be of order \( \bar{d}q/\bar{d}t \) whence it follows that the turbulent vertical diffusivity would be given by

\[
\kappa_T = \frac{\bar{d}q}{\rho_0 c_p \bar{d}V \left( \frac{\bar{d}r}{\bar{d}t} \right)_{\text{bottom}}} = \frac{l_p}{\left( \frac{\bar{d}r}{\bar{d}t} \right)_{\text{bottom}}} \frac{\bar{d}T}{\bar{d}t} \]

Using observed values of \( \bar{d}T = -0.2 \) °C, \( \bar{d}T = 7.75 \), \( l_p = 7 \), \( \bar{d}T / \bar{d}z = -0.02 \) °C/m, yields \( \kappa_T = 2 \cdot 10^{-3} \text{ m}^2/\text{s} \), a value which is of order that shown in panel 7–12 of Fig. 11, where the upslope movement of the bottom mixed layer is clearly observed. Recall that we use the conversion \( \kappa_T = \nu V \eta_0 = 2 \cdot 10^{-7} \eta_0 \text{ m}^2 \text{s}^{-1} \) in Fig. 11. Several simplifying assumptions have been made in obtaining Eq. (5) but the approximate agreement of Eq. (5) with observations does support the idea of an upslope transfer of buoyancy (negative flux of heat) being of order and near balancing the cross isopycnal turbulent buoyancy flux. Presumably this has resulted from shear at the top of the bottom mixed layer. Note that there is some indication of an enhanced shear near the bottom in the mean across isobath velocity profile shown in Fig. 4a.

6.3. Large amplitude internal wave region

The thermistor chain data of Fig. 7 show large amplitude isotherm displacements beginning with the occurrence of the feature which we have termed event A, at \( t = 21:37 \) PTD, and followed closely by event B, at 22:25 PTD. These two features have the characteristics of large amplitude, nonlinear, internal solitary waves of depression. They have maximum downward displacements, \( \eta \), of order 3.2 and 4 m, respectively, and decrease with depth, as seen in Fig. 8. A wave train of downwardly displaced isotherms follows event B. In Fig. 7 we have numbered 9 such individual features which have amplitudes greater than 1 m. These begin to occur at \( t = 23:05 \) PTD, persisting until nearly the end of experiment, with the last one occurring at \( t = 02:34 \) PTD. The mean displacement of these 9 features as a function of depth is shown in Fig. 8 as the black circles.

Isolated, large amplitude depressions in the T-REMUS derived isotherms can also be seen in panels 5 and 6, of Fig. 9, near the location of K1 (black vertical line) and occurring, respectively, around the same time as events A and B of Fig. 7. The isotherms in Fig. 9 are estimated from the T-REMUS yoyo profiles, which are on average 150 m apart. Spatial aliasing is expected for estimating variability whose horizontal length scale \( l \) is such that \( l < 300 \) m. The large depression features shown in panels 5 and 6 of Fig. 9 are the result of interpolating adjacent pairs of isotherms and thus are most likely aliased.

We can estimate the propagation speed of internal solitary waves using the approach of Stanton and Ostrovsky (1998) who used the two layer large amplitude internal wave model CombKdV (Lee and Beardsley, 1974; Ostrovsky and Stepanyants, 1989) to obtain the nonlinear propagation speed \( \nu \) from the expression

\[
\nu = c - \frac{\alpha_1^2 \nu^2}{6 \alpha_0}
\]

where

\[
c = \sqrt{g h_1 h_2 / \nu}
\]

is the linear phase speed, \( \nu \) a nonlinear adjustable parameter expected to be between \( 0 < \nu < 1 \), \( h_1, h_2 \) the depth of the upper and lower layers, respectively, with \( d = h_2 + h_1 \) the water depth and \( \alpha, \alpha_0 \) given by

\[
\alpha = \frac{3c(h_1 - h_2)}{2 h_1 h_2}
\]

\[
\alpha_1 = \frac{3c}{h_1^2 h_2} \left( \frac{7}{8} (h_2 - h_1)^2 - \frac{h_2^2 + h_1^2}{d} \right)
\]

From the averaged density profile shown in Fig. 6a using \( g' = 0.05 \text{ m}^2/\text{s} \), \( h_1 = 5 \), \( h_2 = 14 \), results in \( c = 0.14 \text{ m/s} \) and, with \( \nu = 1 \), an upper bound propagation speed \( \nu = 0.17 \text{ m/s} \). If we were to employ the eKdV modal approach of Grimshaw et al. (1997, 2001) using an observed wave amplitude of 3.2 and the buoyancy profile in Fig. 6b, we find 20% lower values, with the lowest mode linear propagation speed \( c = 0.11 \text{ m/s} \) and nonlinear propagation speed \( \nu = 0.14 \text{ m/s} \).

In Figs. 13 and 14 we show in the upper panels the temperature profiles which were used to make the contour plots of panel #5 and panel #6 of Fig. 9. The lower panel of these figures show the corresponding profiles of turbulent shear used to make the contour maps of Fig. 10. (Fig. 11 used both of these types of data to make the contour maps of \( \eta_0 \) ) Upper panel
profile #7 of Fig. 13 clearly shows a strong displacement relative to its two adjacent profiles. If we associate this displacement with event A and assume that the internal wave propagates nearly parallel to the T-REMUS tracks then from the observation time by the T-REMUS of profile #7 of Fig. 13 at $t = 22:05$ PDT, it would have taken the wave 28 min to propagate 300 m to the observed position on the AUV track, resulting in a wave speed of $v = 0.18$ m/s, close to the estimated theoretical value given by Eq. (6). The profiles in the upper panel of Fig. 14 show several displacement anomalies, not allowing us to identify which one might be event B.

It is interesting to note that in panel #5 of Figs. 10 and 11 we see a patch of turbulence in what appears to be the trailing edge of the internal solitary wave event A. This is also seen in

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**Fig. 13.** Sequence of profiles through Leg # 5. (a) Temperature (°C), (b) Turbulent velocity shear $dw/ds$ (l/s). Same notation as in Fig. 12.

**Fig. 14.** Sequence of profiles through Leg # 6. (a) Temperature (°C), (b) Turbulent velocity shear $dw/ds$ (l/s). Same notation as in Fig. 12.
the lower profile #8 of Fig. 13 indicated by the large turbulent shear variance there. The upper profile #8 of Fig. 13 also shows strong temperature finestructure between \( z = 6 \) m and \( z = 10 \) m at the depth range of maximum displacement of event A, shown in Fig. 8. Note that profile #8 is offshore of the profile showing the maximum displacement, profile #7, which supports the conclusion that the turbulence is located at the trailing edge of an internal wave propagating upslope. This is a similar observation made by Moum et al. (2003) off the coast of Oregon, who found significant turbulence in the trailing edge of large amplitude internal solitary waves. The dissipation rate of this patch is on the high side of observations with \( \varepsilon \approx 10^{-7} \) W/kg but its buoyancy Reynolds number is \( Re_b \approx 10^2 \) indicating only moderately intense turbulence. This arises because of the very high value of stratification there.

Except for a few isolated patches of turbulence, strong spatially continuous turbulence of \( Re_b \geq 200 \) does not arise in the thermocline until panel #7 of Fig. 11 at approximately 23:00 PDT, where it is observed to be located at \( z = 5 \) m depth in the most offshore part of the experimental region. With time we see that this intense turbulence eventually extends throughout the observational area, trending to occur between the 11.8 °C and 12 °C isotherms on a downwelling trend, going shoreward from \( z = 5 \) m to \( z = 10 \) m. Toward the end of the experiment in panel 12 of Fig. 11 we also see strong turbulence formed between the 12.8 °C and 13 °C isotherms located at \( z = 5 \) m. Using Figs. 4 and 6, we calculate a mean Richardson number and find that at \( z = 5 \) m, \( \text{Ri}_z = 4 \), increasing to \( \text{Ri}_z = 8 \) at \( z = 10 \) m. Thus, this enhanced thermocline turbulence occurs in a region indicating stability with respect to the mean flow.

As we have noted from the thermistor chain data, a train of internal solitary like waves is seen at K1 at 23:00 PDT and continues until near the end of the experiment. Some evidence of their effect can be seen in panels 7–12 of Fig. 9, where in the very shallowest regions of the thermocline there is increased isotherm variability. The question arises whether the arrival of this wave train is related to the occurrence of the thermocline turbulence discussed above. The internal waves could lower the value of the local Richardson number (and thus increase the likelihood of flow instability) in two ways: (1) either by producing enhanced shear, or (2) by vertically stretching isopycnals and lowering the value of \( N \). Following the modal approach of Grimshaw (2001) for internal solitary waves of second order (whose maximum amplitude \( A \) satisfies the eKdV equation) we can write a kinematic relationship for the internal wave induced shear, \( U_z \), and strain, \( \eta_z = \partial \eta / \partial z \), in terms of the maximum internal wave displacement \( A \) as

\[
U_z = \nu A \frac{\partial^2 \phi}{\partial z^2} = - \nu N^2 \frac{\partial}{\partial z} A \phi
\]

\[
\eta_z = A \phi_z
\]

where \( \nu \) is the nonlinear wave speed, \( c \) the linear wave speed, \( \eta \) the internal wave displacement, \( N \) the buoyancy frequency, and \( \phi \) the wave (eigen) function which we take as the lowest mode and which is normalized such that its maximum value is set equal to 1. In Fig. 15 we show profiles of \( U_z \) and \( \eta_z \) for the mean maximum amplitude of the wave train of \( A = 1.5 \) m and using the previously calculated lowest mode values of \( c = 0.11 \) m/s, \( \nu = 0.14 \) m/s with \( \phi \) and \( N \) given in Fig. 6b. Although the maximum value of shear shown in Fig. 15 is of order that of the mean shear given by the current profile in Fig. 4a, the buoyancy frequency, \( N \), at the depth where \( U_z \) is a maximum, i.e. \( z = 4 \) m, is also maximum there. This would result in a maximum value of \( \text{Ri}_z = 2 \) at \( z = 4 \) m with \( \text{Ri}_z \) sharply increasing with depth to the mean buoyancy value. From Fig. 15, note the strain rate near the surface could result in as much as a 50% decrease in the local Richardson number, but this again is too small a change in the Richardson number to affect stability. Thus we conclude that the direct effects of internal wave shear and straining will not produce the instability necessary to generate the observed turbulence and will have minimal effect over the depth range where the turbulence has been observed, \( z = 5 \) to \( z = 10 \) m.

A new mechanism for the generation of turbulence by large amplitude internal solitary waves has been recently proposed by Moum et al. (2003). They have suggested, and have shown supporting observational evidence, that internal solitary wave straining can significantly compress isopycnals to the point where the resulting strongly enhanced shear can produce instability and turbulence. See Fig. 22 of Moum et al. (2003) for an illustration of the process. In their model they assume a two dimensional flow field with a background flow having some component in the direction of internal wave propagation and streamlines aligned with isopycnals. Let \( z \) be the equilibrium depth of an isopycnal (isotherm in our case) and let \( \eta \) be the displacement of an internal wave from \( z \). If \( u_0 \) is the background laminar flow in the direction of propagation of the waves then using conservation of mass flow, a change in the spacing between isopycnals will change the velocity \( u_0 \) by

\[
u' = \frac{u_0}{1 + \eta^2}
\]

resulting in an internal wave induced shear of

\[
U_z = \frac{\partial u'}{\partial z} = \frac{u_0^2 \eta^2}{(1 + \eta^2)^2}
\]
In Fig. 16 we plot the vertical profile of the mean internal wave train displacement, open black circles, originally shown Fig. 8, along with the internal wave strain induced shear enhancement, which we define from Eq. (11) as \( E_N = U_z / u_0 \). Note that \( E_N \) has units of m \(^{-1}\). We see the largest values of \( E_N \approx 0.5 \) m \(^{-1}\) at \( z = 4, 8, 9, 11 \) m. If from Fig. 4a we use the along track component of current of \( u_0 \approx 0.2 \) m/s at \( z = 8, 9 \), this would result in an enhanced along track shear of 0.01 s \(^{-1}\) which would result in reducing the local Richardson number to of order 0.5. These calculations are based on the first and second derivatives of the average displacement of the wave train and thus instantaneous estimates of \( E_N \) are expected to have considerable variability, at times reducing the local Richardson number much more. It should be noted that the maxima in \( [E_N] \) at 8 and 9, are located at the approximate depths where we see the occurrence of strong turbulence in panels 10–12 of Fig. 11. The occurrence of the strong turbulence somewhat shallower in panel 12 of Fig. 11 occurs near the depth of the shallow maximum in \( E_N \) at \( z = 4 \) m.

Finally, let us assess the potential importance of the turbulence to the evolution of the internal wave train field. If we assume kinetic and potential energy are approximately equal (which was observed in a strong internal wave field by Moum et al., 2003) we can estimate the total internal energy by:

\[
E_{IW} = \int_{\text{depth}} dz \left( N^2 \eta^2 \right)
\]

where the integral is taken over the water depth. Using data from Figs. 6 and 8, we find \( E_{IW} \approx 0.01 \) J/m\(^2\). At the very end of the experiment in the most offshore location shown in panel 12 of Fig. 10, turbulence of order \( \varepsilon = 10^{-7} \) W/kg extended from the surface to a depth of \( z = 10 \) m. For this situation the characteristic turbulent decay time would be given by

\[
\tau = \frac{E_{IW}}{\varepsilon} = \frac{0.01 \text{m}^3 \text{s}^{-2}}{10^{-7} \text{m}^2 \text{s}^{-3} \cdot 10 \text{m}} = 10^4 \text{s}
\]

where \( h \) is the vertical extent of the turbulence, taken as 10 m. If we use for the wave train members the previously estimated nonlinear wave propagation speed of \( v = 0.14 \) m/s, this results in a decay distance of 1.4 km with a significant decay of the wave train expected over the extent of observations. From Fig. 7 decrease in amplitude of the wave train is noted at the very end of the experiment.

7. Summary and conclusions

We have used an AUV equipped with microstructure and finescale structures to examine the vertical and horizontal spatial structure of a turbulent field in a very shallow water environment. The vehicle performed a 5 degree yo-yo pattern which resulted in sampling most of the water column over a 2.5 km across isobath swath. Data were collected for 8 h, repeating the same across isobath track by reversing direction upon the completion of each track leg. This resulted in 12 continuous parallel across isobath AUV tracks of data. Estimates of turbulent dissipation rate and finescale buoyancy were obtained on a vertical scale of 0.5 m and an average horizontal scale of 150 m. The experiment took place under very light wind and weak air–sea interaction conditions and thus the environment was near ideal to isolate the mechanisms of generation and maintenance of the turbulence occurring within the water volume.

The region sampled, the bight in northeastern Monterey Bay, is known to be characterized by strong internal wave activity much of which is thought to be associated with tidal generation from the edges of Monterey Canyon and then propagating into shallow water. Salinity variation in the area is very weak with stratification principally determined by temperature. The AUV data was supplemented by data from a nearby thermistor chain, which sampled temperature every 30 s and throughout most of the water column.

With this data set we were able to determine the principal factors affecting the turbulent field, the most likely mechanisms of its generation, and the character of its spatial variation both horizontally as well as vertically. As an indicator of the strength of the turbulent field we use both the dissipation rate, \( \varepsilon \), and the buoyancy Reynolds number \( R_{b} \). The latter spanned values from 10 to 10,000. Values of order \( R_{b} \approx 20 \) are less thought to be associated with fluid velocity variability too weak to overcome buoyancy effects (Itsweire et al., 1986). For \( R_{b} > 200 \), the turbulence can be considered strong enough to be fully isotropic (Yamazaki and Osborn, 1990). From this criterion, because of the occurrence, at times, of extremely large values of stratification, i.e. \( N > 3 \times 10^{-2} \) s\(^{-1}\), values of dissipation rate of order \( \varepsilon = 10^{-8} \) W/kg were sometimes observed to be associated with variability too weak to be considered turbulent. Conversely within the bottom mixed layer, which was characterized by low stratification, there were dissipation rate values of the same magnitude, \( \varepsilon = 10^{-8} \) W/kg, but with \( R_{b} \) three orders of magnitude larger. Values as high as \( R_{b} = 10^{4} \) were observed in the experiment in the bottom mixed layers, near the surface immediately below a warm water intrusion, and, at times, in the thermocline.

The overall flow field of the experimental site is to the northwest, which according to classical Ekman dynamics, because the coastline is to the right of this flow, would result in a vertical circulation characterized by downwelling. However, there was a very strong across isobath seaward surface flow with a near bottom return flow. The experiment took place during ebb tide and it is possible that this flow structure is
tidally coupled. We have insufficient supporting current data to examine this hypothesis further.

This flow exhibited a very strong near surface across isobath shear resulting in intense turbulence at the depth of the interface, clearly seen in the lower panel of Fig. 12 in profiles 6–17. The mean flow field also showed an along isobath shear which extended into the thermocline to a depth of 12 m. However the simultaneous occurrence of strong stratification (Fig. 7) resulted in a mean background Richardson number of \( R_i \approx 6 \) in the thermocline, indicating flow stability there.

Using the buoyancy Reynolds number of \( Re_b > 200 \) and \( \varepsilon > 10^{-8} \) W/kg as a criterion for strong turbulence, three regimes of strong turbulence are identified. These are: (1) a near surface warm water intrusion; (2) the bottom mixed layer; (3) a nonlinear internal wave train which followed two isolated internal solitary wave events. During the middle of the experiment at \( t = 21:30 \) PDT the warm water surface intrusion was observed to evolve into a very strong frontal feature in which very strong turbulence occurred nearby. Both the warm water surface intrusion and upslope movement of bottom mixed layer resulted in strong shear driven turbulence. The upslope transport of buoyancy in the most offshore location of the bottom mixed layer was approximately balanced by cross isopycnal turbulent buoyancy flux.

Using data from a nearby thermistor chain, large amplitude downwardly displaced isothermal variability suggested the occurrence of nonlinear solitary-like waves. These waves first appeared as two isolated events, labeled A and B, which were then followed by a near continuous train of similar but somewhat smaller amplitude waves. The AUV based hydrography showed a large downwardly displaced isolated feature which appeared to be associated with event A. A propagation speed of \( v = 0.18 \) m/s was estimated for event A, based on the times of observation by the thermistor chain and the T-REMUS and the distance it would have traveled. This was in close agreement with an estimate of the propagation speed of a nonlinear internal solitary wave for this environment using higher order KdV theory. Event A showed moderately strong turbulence of \( Re_b \approx 200, \varepsilon \approx 10^{-7} \) W/kg in what looked to be in the lee of its propagation direction.

Strong turbulence in the thermocline was noted towards the end of the experiment. Using the modal eKdV model of Grimshaw (2001), along with the observed values of internal wave displacement and local buoyancy frequency, indicated that internal wave induced shear were not sufficiently strong to affect the mean local Richardson number. Moreover the largest internal wave induced effects in lowering the local Richardson number were predicted to occur at depths much shallower than that where the turbulence was observed. However, a theory was developed based on the approach of Moum et al. (2003) to examine the role of internal wave induced vertical strain gradient on inducing shear. We derived a quantity termed the enhancement factor, \( En \), which could be used to quantify the amount of internal wave induced shear induced from the mean flow. It was shown that this effect of internal wave induced vertical strain was sufficiently strong to lower the local Richardson number below 1. Maxima of \( En \) occurred at depths where the largest values of thermocline turbulence were observed.

At the end of the experiment strong turbulence was noted in the upper 10 m of the water column. Calculation of the effect of this turbulence on the internal wave train suggested a turbulent decay distance of 1.4 km.

The AUV measurements have been unique in providing a detailed description of both the turbulent as well as larger scale flow field. These measurements, along with the supplementary thermistor chain data, allow interpretation of the turbulent and finescale data in a larger scale context. A number of interesting questions have arisen from this analysis, in particular with regard to the nature of the larger scale flow field and its relationship to the structure and intensity of the turbulence. These include the following:

1. What is the nature of the vertical circulation in this region? Is it tidally coupled? What role does the near surface across isobath warm water intrusion play in the overall apparently downwelling state of the flow field? What role would this play in enhancing near bottom shear and thus in enhancing the turbulence within the bottom mixed layer?

2. What is the physics of the intense frontal structure produced by the sharpening effect of warm water intrusion? What are its space and time scales? What role does turbulence play in its evolution?

3. What is the role of the near surface offshore across isobath advection play in internal wave strain induced shear enhancement?

We hope to begin to address these questions by the use of modern high resolution numerical modeling as well as supplementary data obtained along the LOCO fixed stations.

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