Energy Transport by Nonlinear Internal Waves

J. N. Moum1, J. M. Klymak2, J. D. Nash1, A. Perlin1, and W. D. Smyth1
1College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, Oregon
2University of Victoria, Victoria, British Columbia

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ABSTRACT

Wintertime stratification on Oregon’s continental shelf often produces a near-bottom layer of dense fluid that acts as an internal waveguide on which nonlinear internal waves propagate. Shipboard profiling and bottom lander observations capture disturbances that exhibit properties of internal solitary waves, bores and gravity currents. Wave-like pulses are highly turbulent (instantaneous bed stresses are 1 N m$^{-2}$), resuspending bottom sediments into the water column and raising them 30+ m above the seafloor.

The waves’ cross-shelf transport of fluid counters the time-averaged Ekman transport in the bottom boundary layer.

In the nonlinear internal waves we have observed, the kinetic energy is roughly equal to the available potential energy and is $O(0.1) \text{ MJ per m of coastline}$. The energy transported by these waves includes a nonlinear advection term $\langle uE \rangle$ that is negligible in linear internal waves. Unlike linear internal waves, the pressure-velocity energy flux $\langle up \rangle$ includes important contributions from nonhydrostatic effects and surface displacement. It is found that, statistically, $\langle uE \rangle \approx 2 \langle up \rangle$. Vertical profiles indicate that $up(z)$ is more important in transporting energy near the seafloor while $uE(z)$ dominates farther from the bottom. With the wave speed, $c$, estimated from weakly nonlinear wave theory it is verified experimentally that the total energy transported by the waves, $\langle up \rangle + \langle uE \rangle \approx c \langle E \rangle$.

The high but intermittent energy flux by the waves is, in an averaged sense, $O(100) \text{ W per m of coastline}$. This is similar to independent estimates of the shoreward energy flux in the semidiurnal internal tide at the shelfbreak.

1. Introduction

The circulation over Oregon’s continental shelf is principally wind-driven (Huyer et al. 1978). The large-scale cross-shelf circulation is thought to be determined by Ekman dynamics (Perlin et al. 2005a). Predominantly northerly winds in spring/summer drive near-surface fluid offshore, generating a cross-shelf density gradient that sustains a southward coastal jet, which in turn drives near-bottom fluid onshore. In fall/winter, winds are predominantly southerly, resulting in a northward coastal jet and effectively reversing the cross-shelf circulation pattern. As a consequence, there is a clear distinction in the cross-shelf structure of the stratification between upwelling and downwelling seasons (figure 1).

In spring and summer, stratification is concentrated near the surface. This provides a waveguide in which nonlinear internal waves of depression are commonly observed to propagate (Moum et al. 2003). In the left panel of figure 1 two examples of these are evident. At 20 km offshore is seen a large wavelength wave propagating onshore at about 0.3 m s$^{-1}$; the leading edge of this wave appears to be bore-like. A train of smaller scale waves inshore at 1.5 km is aliased in our profiling observations. In both cases, the influence of the waves is clear throughout the water column.

In winter, the water column is typically weakly-stratified, from the combination of wind-driven mixing, onshore transport of light near-surface fluid in the upper Ekman layer and offshore transport of dense near-bottom fluid in the bottom Ekman layer. In the right panel of figure 1, the lightest fluid is trapped inshore of 6 km. At times, this combination of motions leaves the middle portion of the shelf completely unstratified. However, at other times, a thin stratified layer exists near the seafloor (figure 1), upon which nonlinear waves of elevation propagate onshore (Klymak and Moum 2003). This particular example (right panel of figure 1) shows a train of waves propagating onshore at 12 km. Net fluid transport by these waves counters the offshore bottom Ekman transport.

When and where the internal tide steepens, it is frequently found that a significant portion of the tidal en-
ergy goes to generation of a nonlinear internal wave field (see Helfrich and Melville (2006) for a recent review). Since stratification is more typically strong near the surface and this is where most of our observations have taken place, the result we typically observe is a surface-trapped wave field (Sandstrom and Elliott 1984; Holloway 1987; MacKinnon and Gregg 2003). When stratification is concentrated near the bottom, tidal and mesoscale energy can also generate a bottom-trapped nonlinear internal wave field (Klymak and Moum 2003; Hosegood and van Haren 2003) that is similar to the form of nonlinear internal waves observed in the atmosphere (Smith et al. 1982).

Quantification of the energy transport in nonlinear internal waves has been limited by the resolution of the available observations (Pinkel 2000; Klymak and Moum 2003; Lien et al. 2005; Klymak et al. 2006), or has been based on a linear formulation of the energy flux (Chang et al. 2006). To define the energy transport mechanisms in the waves requires velocity measurements with such detail that they are difficult to obtain from a moving ship. Our recent observations from a bottom lander fixed to the seafloor provide sufficient detail to fully evaluate the wave energy transport terms.

The prevalence of bottom-trapped nonlinear internal waves over the Oregon shelf in winter downwelling season is evident from a set of observations made in January 2003. These include two sets of moored observations as well as shipboard profiling measurements both across the shelf and, for 36 h, at the location of the moorings (figure 2). These show both the cross-shore structure and temporal evolution of the waves.

Our objective here is twofold. Firstly, to provide a useful description of near-bottom flows over Oregon’s continental shelf from a series of complementary observations. Secondly, to quantify how the energy in nonlinear internal waves is partitioned and transported. This wave energy is compared to the energy available in the internal tide. We begin with the development of a theoretical framework to describe the energetics (§2). This is followed by a brief presentation of the measurements (§3) and a description of the relevant observations (§4). We then discuss how these observations are used to evaluate nonlinear internal wave energy and its transport (§5) and conclude with a discussion of some of the consequences of these results (§6) and a summary (§7). Details of our estimates of turbulence kinetic energy dissipation rate from bottom lander measurements is outlined in the Appendix.

2. Energetics: theoretical framework

In this section, we define the wave kinetic energy, available potential energy and the energy flux for a two-dimensional, nonhydrostatic, nonlinear internal wave. This results in an evolution equation for the wave energy that differs significantly from that which governs linear internal wave energetics (Kundu and Cohen 2004, for example). Following Venyagamoorthy and Fringer (2005), we form an energy equation (units [J m$^{-3}$]) from the sum of the dot product of the momentum equation with the velocity vector $u$, and $gz \times$ the mass conservation equation, where $g$ is the acceleration due to gravity and density is

$$
\rho(x, z, t) = \rho_\alpha + \rho_b(z) + \rho_w(x, z, t).
$$

(1)

The density is defined as the sum of a background density profile $\rho_\alpha + \rho_b(z)$, and a perturbation associated with the wave, $\rho_w(x, z, t)$.

The wave perturbation density, $\rho_w$, is defined by the requirement that the available potential energy of a wave,
and \( \rho \) means that \( \rho \) the difference between the observed density and undisturbed surface) reaches an asymptote. \( \rho = 0 \) averaged AP E This is a sufficiently large area to ensure that the spatially-
a distance of 50 wavelengths (\( L \)) to that rest state (resulting in horizontal isopycnals) over
terminated \( \rho \) et al. 1981; Hebert 1988; Winters et al. 1995). We de-
location of the shelfbreak mooring is 5 m greater.

Fig. 2. Location of the experiment off the Oregon coast in January 2003. Cross-shelf range of repeated transects is shown in yellow. Location of 36 h time series is shown as a white triangle. Depths are contoured at 50, 100 and 150 m. The inset shows relative locations of time series, bottom lander and shelfbreak mooring. The lander is 653 m due east of the mooring and shows relative locations of time series, bottom lander and shelfbreak mooring. The lander is 653 m due east of the mooring and 850 m due south of the profiling time series. The mean water depth at lander and profiling station is 124 m - the depth at the location of the shelfbreak mooring is 5 m greater.

\( \rho \) APE, be referenced to a thermodynamic rest state (Reid et al. 1981; Hebert 1988; Winters et al. 1995). We determined \( \rho_0 \) by redistributing the observed density field to that rest state (resulting in horizontal isopycnals) over a distance of 50 wavelengths (\( L \)) of the observed wave. This is a sufficiently large area to ensure that the spatially-averaged \( \rho \) (that is, \( \int_{-25L}^{25L} \int_0^H \rho \rho \) \( d\rho \) \( dz \) \( dx \), where \( z = 0 \) is the seafloor and \( z = H \) the height of the undisturbed surface) reaches an asymptote. \( \rho_w \) is then the difference between the observed density and \( \rho_0 \). This means that \( \rho_w > 0 \) in the vicinity of an elevation wave and \( \rho_w < 0 \), though small, elsewhere.\(^1\)

Neglecting mixing, mass conservation becomes

\[
\frac{\partial \rho_w}{\partial t} + u \frac{\partial \rho_w}{\partial x} + w \left( \frac{\partial \rho_b}{\partial z} + \frac{\partial \rho_w}{\partial z} \right) = 0, \tag{2}
\]

The pressure disturbance in a nonlinear internal wave is derived by Moum and Smyth (2006) as the sum of internal hydrostatic pressure (due to isopycnal displacement) plus external hydrostatic pressure (due to surface displacement) plus nonhydrostatic pressure (due to vertical fluid accelerations),

\[
p = g \int_z^H (\rho_o + \rho_b + \rho_w) d\zeta' + \rho_o g \eta_H + p_{nh}. \tag{3}
\]

Here, \( \eta_H(x, t) \) is the surface displacement, the nonhydrostatic pressure,

\[
p_{nh} = \rho_o \int_z^H \frac{\partial W_d}{\partial z} dz', \tag{4}
\]

and we refer to the internal hydrostatic pressures (the terms in the integrand of equation 3) separately as \( p_h^o \), \( p_h^w \) and \( p_h^\eta \).

We define

\[
E = KE + AP E, \tag{5}
\]

where

\[
KE = \frac{1}{2} \rho_o u_i u_i, \tag{6}
\]

and the available potential energy is

\[
AP E = \rho_w g z. \tag{7}
\]

The resultant energy equation is

\[
\frac{\partial E}{\partial t} + \frac{\partial}{\partial x} [u(KE + \rho_w g z + p_h^o + \rho_o g \eta_H + p_{nh})] + \frac{\partial}{\partial z} [w(KE + \rho_h^o + p_h^w + \rho_o g \eta_H + p_{nh})] + gz \frac{\partial}{\partial z} [w(\rho_b + \rho_w)] = -\rho_o \epsilon, \tag{8}
\]

or

\[
\frac{\partial E}{\partial t} + \frac{\partial f_E}{\partial x} + \text{vertical redistribution} = -\rho_o \epsilon, \tag{9}
\]

where the horizontal energy flux is

\[
f_E = u(KE + \rho_w g z + p_h^o + \rho_o g \eta_H + p_{nh}), \tag{10}
\]

and the RHS of (9) is the energy lost to turbulence, with \( \epsilon \) representing the viscous dissipation rate of the kinetic energy of the turbulence.\(^2\) The remaining terms represent energy transport in the vertical direction. While these are likely important for internal redistribution of energy, they

\(^1\)Note that \( \rho_w(z) \) differs from that defined by Moum and Smyth (2006) as it is referenced \( \rho_0 \) and not \( \rho_i \) (z). Moum and Smyth (2006) simply defined \( \rho_i \) (z) as the density profile ahead of the wave. This cannot be used as a reference to compute \( \rho \) APE.

\(^2\)Equation 8 ignores both the potential propagation loss due to local generation of small-scale but undetected internal waves from the disturbance (Crouch 1988) and the energy source associated with the density gradient across the disturbance, as suggested by Henyey and Hoering (1997). We can say nothing about the former and revisit the latter in discussion.
do not contribute quantitatively to the lateral transport, unless the waves are either rapidly decaying or rapidly growing. Our primary concern here is the lateral energy transport and we neglect the terms labeled “vertical redistribution” in (9).

The horizontal energy flux, \(f_E\), is the sum of nonlinear advection of wave energy (kinetic plus available potential energy) and a pressure-velocity work term, or

\[
f_E = uE + up. \tag{11}
\]

Our definition differs from that of Venyagamoorthy and Fringer (2005) in the absence of nonlinear advection and pressure-velocity work terms that involve the background density, \(\rho_0\), which is independent of \(x\) and hence does not contribute to lateral transport. (10) also differs from the rigid lid calculation of Venyagamoorthy and Fringer (2005) in the explicit inclusion of the external pressure caused by the disturbance of the free surface. Our observations indicate that \(\int_0^H udz \neq 0\), and the net result of the pressure-velocity work term involving \(\rho_0 g \eta_H\) is not zero.

The energy flux defined by (10) differs from that for a linear internal wave, for which \(f_E = u p_0^E\) (Nash et al. 2005), in two ways. The first is the presence of the nonlinear advection term, \(uE\), and the second is the nonhydrostatic part of the pressure-velocity work term. For a linear wave, it is simply shown (Kundu and Cohen 2004) that

\[
\langle up \rangle_{\text{linear}} = c_g \langle E \rangle, \tag{12}
\]

where \(\langle \cdot \rangle = \int_{-\infty}^{\infty} \int_0^H (\cdot) dz dt\), so that

\[
\langle f_E \rangle_{\text{linear}} = c_g \langle E \rangle, \tag{13}
\]

where \(c_g\) is the linear group wave velocity.

Given that a solitary nonlinear internal wave represents a quantity of energy \(\langle E \rangle\) confined to a compact region of space that moves with velocity \(c^3\), it seems appropriate that the resulting energy transport can be described by a flux equal to \(c \langle E \rangle\). That is,

\[
\langle f_E \rangle = c \langle E \rangle. \tag{14}
\]

Using data obtained from a relatively limited but well-defined and sufficiently well-observed set of individual nonlinear internal waves, we will demonstrate that (14) holds for these observations. We note that (14) is consistent with the following conditions;

- \(\rho_0 c\) can be neglected in (9),
- “vertical redistribution” does not contribute to lateral transport, and
- the wave is steady in a reference frame propagating horizontally at constant speed, \(c\), that is, \(\partial/\partial t = -c \partial/\partial x\).

3. Experimental Details

The observations discussed here were made from ship, from a rapidly-sampled bottom lander as well as a full water column mooring (locations shown in the inset of figure 2).

From shipboard profiling measurements using the freely-falling turbulence profiler, CHAMELEON, temperature, density, and \(\epsilon\) were computed (Moum et al. 1995). As well, an optical backscatter probe (880 nm) was deployed together with the microstructure sensors on the nose of CHAMELEON. An acoustic Doppler current profiler (ADCP; 150 kHz - sampled at 6 s and 4 m vertical bins) and high-frequency echosounder (120 kHz; ping- ing twice per s) were mounted in the ship’s hull. The echosounder provides a means of imaging the flow on fine vertical and temporal/horizontal scales. Even at a profiling rate of one CHAMELEON profile every 4 – 5 minutes, our perspective of the structure of the nonlinear internal waves is aliased. Detailed imaging of the waves was achieved using a high-frequency echosounder.

Over the period 20 January to 03 February, 12 cross-shelf profiling transects were executed with ship moving at 1-2 kts from offshore to inshore along the transect line shown on figure 2. As well, a profiling time series was conducted near the mooring locations (figure 2, inset) over a 36 h period from 30 January to 01 February, while R/V Revelle held position to within a few m using dynamic positioning.

A bottom lander was deployed for the extent of the shipboard observations. Instruments included an upward-looking ADCP (300 kHz - 5 s, 1 m ensemble averages recorded from 2 m above the sea bed), and two 5 MHz acoustic Doppler velocimeters (ADVs) sampled at 10 Hz, positioned to either side of the ADCP and 1 m above the bed. A SeaCat (pressure, temperature and conductivity) was located 1 m above the bed and sampled at 1 Hz. The ADCP and ADV were gimballed to ensure they remained horizontal. Unfortunately, a malfunction by a data storage device limited data from ADCP and ADVs to a short period from deployment on 20 January to 22 January. The SeaCat data extends from 20 January through 03 February.

The shelfbreak mooring included temperature measurements at 1 m above the bed, sampled at 2 minute intervals.

4. Observations

Our three different modes of observation (cross-shelf profiling, profiling time series and lander time series) offer unique perspectives of near-bottom flows. Our objective in this section is to provide a general description of the wave structure with the aid of all of these observa-
Typical density structure across the shelf during downwelling is represented by a series of three successive transects at 45°00′N, each separated by 24 h, following and after a period of sustained northward (downwelling-favorable) winds. The resultant cross-shore Ekman circulation includes a shoreward near-surface transport and offshore transport near the bottom. A consequence is a distinctive cross-shelf structure in density (represented here as $\sigma_\theta$; figure 3). Light fluid near the surface is concentrated at the inshore boundary, enhancing stratification there (this is seen most clearly in the right panel of figure 1). Denser fluid near the bottom is driven offshore. In between, at midshelf, the water column is weakly- or neutrally-stratified. This is a consequence of the combination of cross-shelf circulation and intense wintertime surface forcing (Perlin et al. 2006).

Offshore of the shelf break (nominally 200 m) the positions of isopycnals at depths greater than the mixed layer depth are controlled by mesoscale motions and/or the internal tide; this is independent of the downwelling circulation. In each example shown in figure 3, the isopycnals indicate the presence of an abrupt near-bottom front over the shelf (at 18 km in the first transect, 15 km in the second and 12 km in the third)\(^4\). Two common aspects of these fronts are high optical backscatter above the seafloor and high $\epsilon$ at the fronts’ leading edge. High optical backscatter at the seafloor indicates resuspension of sediments, likely caused by the elevated turbulence there.

The nature of these fronts differs between transects. The second is much sharper (and more turbulent) than the first; the third is clearly wavelike. At the head of the front in the third example appears to be a train of propagating internal solitary waves of elevation (Klymak and

\(^4\)Although we have no observations between these transects, there is no reason to think that this is a progression of the same feature. The small translation speeds inferred from figure 3 are $10 \times$ smaller than we would expect from accompanying analysis.
b. Profiling time series

In our 36 h profiling time series, a sequence of similar (though far from identical) near-bottom structures are revealed (denoted by the inverted triangles atop figure 4). These fronts are accompanied by vertical displacements of near-bottom isopycnals by 25 - 30 m over periods of about 5 minutes (corresponding to vertical velocities of $O(0.1 \text{ m s}^{-1})$). They are spaced in time by 9 h 15 min, 14 h 20 min and 8 h 10 min, successively, suggesting no simple correspondence with the dominant $M_2$ tide (the barotropic component of which dominates variations in sea surface height shown in the upper panel of figure 4). Like the fronts depicted in figure 3, each front in figure 4 is accompanied by high optical backscatter and high $\epsilon$.

These profiling data were obtained during a period of northward flow which is strongly modulated in amplitude (3rd panel; figure 4). The cross-shelf component of the flow is modulated in both amplitude and direction (2nd panel; figure 4), apparently a result of the internal tide. Unfortunately, current measurements from shipboard ADCP do not include the bottom 20 m and our bottom lander had expired by the time of this profiling time series. Given these shortcomings, together with the short record length, it is difficult to assess correspondences between the density fronts and currents.

Finer details of the structure at the front are provided by shipboard echosounder measurements aided by CHAMELEON profiles. Figure 5 shows the evolution of a scattering layer that defines the complex structure of the
Fig. 5. Expanded plot of the final density front shown in figure 4. Upper panel - acoustic image of internal bore with $\sigma_\theta$ contours at 0.1 kg m$^{-3}$ intervals. Lower panel - image plot of $\sigma_\theta$ with contours at 0.1 kg m$^{-3}$ intervals. To the right of the lower panel are vertical profiles of $\sigma_\theta$ and $\epsilon$ averaged (4-7 CHAMELEON profiles) at the times shown above the upper panel.

The final front observed in the 36 h series in figure 4. We presume that the scattering layer, which is initially located 1 – 5 m above the bottom, is a deep zooplankton layer. This layer is raised 30 m by the first undulation of the bore and an additional 10 m by the next undulation. Following this are a sequence of smaller undulations. The scattering layer remained 20 m above the bed more than 1 h following the passage of the initial front.

One aspect of the near-bottom scattering response shown in figure 5 seems to be consistent with most of the fronts we observed. Bright scatterers appear from the bottom following passage of the initial density front. This is possibly the emergence of bottom dwellers in response to the changing environmental conditions carried along with the fronts.

Contours of $\sigma_\theta$ plotted on the echosounder image confirm that the vertical displacement of the principal scattering layer reflects the vertical displacement of nearby isopycnals (figure 5). The density structure further reveals the continual near-bottom density increase behind the front (the near-bottom density increase of 0.5 kg m$^{-3}$ is seen in the difference between initial and final density profiles in figure 5). In detail, isopycnals indicate a bulbous structure at the leading edge of the densest fluid (note the black isopycnals just after 01:00). This is expected at the head of a gravity current flowing over the bottom as fluid flow above the bottom is reduced to zero at the sea floor (as depicted in the laboratory results of Britter and Simpson (1978); their figure 1a, for example).

The structure of the density fronts at 1 m height above the bottom over the 36 h period corresponding to our profiling time series from measurements at each of the three locations shown in figure 2 is seen in figure 6. Measurements of temperature and conductivity at the lander were restricted to a single measurement at 1 m height. As the nearest conductivity measurement from the bottom was at 10 m height at the shelfbreak mooring, our only complete comparison is potential temperature. The bores propagate principally onshore, so the 653 m east-west separation between lander and shelfbreak mooring provides a basis for estimating the speeds of the fronts (Moum and Smyth 2006). The signature of each bore is unique and evolves between the locations of lander and shelfbreak mooring. In fact, the only example that is unambiguously defined
at the both locations is the final bore. The inset in the upper panel of figure 6 shows the time of arrival of the front at these locations to differ by 1 h, giving an estimate of the frontal propagation speed as 0.18 m s\(^{-1}\).

c. Lander time series

Unfortunately, we have no velocity measurements from our lander that coincide with the 36 h profiling time series shown in figure 4. However, the lander was fully operational for the first 52 h of its deployment and the perspective offered is unique. The data are also sufficiently detailed to permit an assessment of the energy flux. During the 52 h deployment, six fronts passed the lander, traveling almost directly onshore (as determined from the direction of currents in the bore at 1 m height). Their spacing in time was 2 h 50 min, 8 h 15 min, 5 h 25 min, 8 h 45 min and 23 h 15 min.

A 22 h record from the bottom lander shows the first four of these fronts (figure 7). Both the rate of and the total density increase across each front differs considerably. At each front is an abrupt velocity pulse followed by a series of pulses. Each of these pulses is bottom-trapped and directed toward the shore, but the vertical displacements and separations between pulses varied considerably. The initial velocity pulse of fronts 1,3,4 coincides with the initial density increase. In front 2 the initial velocity pulse occurs before (ahead of) the large density increase; it does, however, coincide with a smaller, local density increase (figures 8, 9).

The density structure of the final front is not typical. In this final front the first velocity pulse coincided with a pulse of dense fluid at 1 m height. The initial velocity pulse was detected by the ADV at 1 m height, but the following pulses had velocity maxima more than 20 m above the bottom and no velocity signature at 1 m height. It is known that large amplitude mid-water column internal waves can be excited by an intrusive gravity current (Mehta et al. 2002). This may help to explain the lack of a bottom signature.

While no two fronts were the same, an example shows the salient features (figure 8). This front was accompanied by a density increase of 0.15 kg m\(^{-3}\) which occurred
Fig. 7. Time series of measurements from bottom lander. Upper panel - $\sigma_\theta$ at 1 m height above the bottom; Second panel - horizontal component of velocity (30 s ADCP averages) in the direction of wave propagation (due east); lower panel - current speed at 1 m height from ADV measurement. Vertical dashed lines indicate the arrival times of leading waves in each group as determined from the velocity records.

over a period of about 1.5 h. The flow at 1 m height was composed of an oscillatory component of $0.4 \text{ m} \text{ s}^{-1}$ peak-to-peak plus a mean component of $\simeq 0.15 \text{ m s}^{-1}$ during the passage of the density front. The direction of both the mean component of the flow and the pulses was due east. Eleven pulses are clearly visible in the record, each with the distinctive vertical velocity signature of an elevation wave (upward ahead and downward behind each wave). The first two pulses were separated by about 15 minutes and the rest appeared with a 10 minute separation. The onshore component of the flow extended almost 50 m above the bottom for the first three pulses.

The strong shear at the crest of the pulses is seen in the detail of the structure of the first pulse of figure 8 (expanded in figure 9). The shear is also intense at the bottom, where the fluid velocity $\rightarrow 0$. The reduction of velocity above the crest and at the bottom means there is a maximum in between, as seen in the vertical profile averaged at the crest in figure 9.

The internal structure of the wave pulse is revealed in some detail when the phase speed is subtracted from the local fluid velocity\(^5\). The phase speed estimate used here is the second order phase speed determined by solution of the Korteweg-de Vries equation (Gear and Grimshaw 1983); solutions for these particular examples are described by Moum and Smyth (2006). Removal of the phase speed provides a depiction of the fluid motion in the frame of reference of the wave (figure 10; here, $c=0.48 \text{ m s}^{-1}$). There is clear evidence of circulation in the $x-z$ plane. This must be part of a three-dimensional circulation within the wave, presumably an indication of instabilities within the wave core. In particular, there exist fluid parcels within the wave core that have velocities $> c$, an untenable scenario for a time period longer than that required for their passage through the wave.

d. Turbulence

The dissipation rate of turbulence kinetic energy was

\(^5\)We consider the pulse to be a wave to justify use of a phase speed.
computed from ADV measurements 1 m above the bottom (see Appendix). Intense turbulence was observed beneath and following the crest of the first wave (figure 9). The trailing edges of the first four waves in figure 8 were progressively more turbulent, after which the fluid passing the lander was continuously turbulent, and \( \epsilon \) was 100\( \times \) greater than before the wave passage.

The vertical structure of enhanced turbulence behind the front is indicated by the set of averaged \( \epsilon \) profiles shown in figure 5. At 10 m height, the increase in \( \epsilon \) was approximately a factor of 100, from \( 10^{-9} \) to \( 10^{-6} \) across the front. Behind the front \( \epsilon \) was relatively constant over the bottom 30 m.

The bottom stress, \( \tau_b \), was estimated using wall-layer scaling in the bottom boundary layer applied to the ADV measurement 1 m above the bottom; \( u_* = (\epsilon \kappa z)^{1/3} \), and \( \tau_b = \rho u^2 \) (Perlin et al. 2005b). Peak values of \( \tau_b \approx 1 \text{ N m}^{-2} \) coincide with peak speeds in the wave pulses. This demonstrates the intense and immediate interaction of the pulses with the bottom. Such high stresses at the periphery of the wave core represent a primary distinction between near-bottom waves of elevation and near-surface
Fig. 9. Time series of measurements from bottom lander translated to a spatial series via \( x = c_1 (t - t_0) \). Upper panel - horizontal component of velocity (5 s ADCP averages) in the direction of wave propagation; Second panel - \( \sigma_\theta \) at 1 m height above the bottom; Third panel - horizontal component of velocity (0.5 s averaged ADV) in the direction of wave propagation at 1 m height; Lower panel - \( \epsilon \) (1 minute averages) at 1 m height. To the right of the upper panel is a vertical profile of horizontal velocity averaged over the interval indicated by the horizontal black line at \( x = 0 \).

Fig. 10. Velocity vectors (\( x - z \) plane; 5 s averages) in a reference frame moving with the wave. The image plot represents the horizontal velocity component; warm colors indicate \( u > c \).
waves of depression which do not interact with a solid boundary.

e. Summary of observations

Cross-shelf transects clearly show high $\epsilon$ and high turbidity at sharp, near-bottom fronts where isopycnals approach the shelf (figure 3). As they approach the shore, these fronts have been observed to terminate in a series of solitary-like waves (Klymak and Moum 2003).

Our profiling time series (figure 4) offers a somewhat different perspective. In particular, the detailed imaging provided by high-frequency acoustics suggests the structure of the fronts is similar to what has been referred to as internal undular bores in atmospheric and laboratory flows. The structure outlined in figure 5 is similar to that observed by upward-looking echosounder images in northern Australia (figure 3 in Christie et al. (1981)), by downward-looking echosounder images showing depressions of scattering layers near the sea surface in Knight Inlet, British Columbia (by David Farmer and reproduced by Simpson (1997) as figure 7.16) and by dye in a laboratory experiment to replicate some of the properties of the geophysical examples (Simpson (1997) figure 13.5). In each of these cases, the signature is attributed to a combination of undular internal bore and internal solitary waves. Rottman and Simpson (1989) showed how these can be created in the laboratory at the leading edge of a gravity current and, indeed, density measurements shown in figure 5 demonstrate the signature of a gravity current behind the leading edge of the bore.

A third perspective is found in the detailed, near-bottom velocity profiling measurements from our lander time series. Here, the leading edges of the density fronts are observed to coincide with a velocity signature that would appear as undulations in either isopycnals or acoustic scattering layers. However, the velocity signature is clearly pulse-like (or solitary wave-like) in that individual velocity pulses (or wave cores) are distinct from each other and proceed with significantly greater speed than the remainder of the fluid behind the density fronts. Moum and Smyth (2006) have referred to these as trains of propagating waves. Whether the appearance of the waves ahead of the fronts is analogous to wave fission from a density current (Nash and Moum 2005) cannot be addressed from this set of measurements.

These observations are typical of those involving geophysical fluid flows in that they demonstrate a combination of traits that have each been clearly defined in the laboratory but are difficult to separate in geophysical observations. The features we observe have aspects that are both wave-like (i.e., symmetric perturbations about the disturbance), and bore-like (i.e., are associated with sloping isopycnals and horizontal pressure gradients). Henyey and Hoering (1997) coined the term solibore to represent the combination of features. For the remainder of this analysis, we consider the wave-like components of the flow in isolation from the background gradients. The objective is to define the range of magnitudes of kinetic and potential energy contained within the waves and to assess how the energy is transported by them. The bore-like aspects of the energetics are addressed by Shroyer et al. (2006). Ultimately, we must consider the interplay between the waves and the density fronts in which they are observed, as these fronts apply a pressure gradient and represent a potential source of energy (Henyey and Hoering 1997).

![Figure 11](image.png)

**Fig. 11.** Upper panel - $KE$ density through the wave shown in figure 9. Second panel - $APE$ density. Lower panel - vertically-integrated energies.

5. Energetics

The data obtained using the bottom lander provides a detailed measurement of the velocity field in the waves. In particular, the leading three waves of each of the four wave groups depicted in figure 8 have velocities that are large relative to both the background flows and also to any bore-like component of the flow. Moum and Smyth (2006) used the velocity fields as observed from the bottom lander to reconstruct the density and pressure field of each of these twelve waves. With the assumption that the waves are inviscid, two-dimensional and propagate without change in shape with known speed and direction, isopycnals parallel streamlines. Using a background density profile, isopycnals were mapped to streamlines computed from the velocity measurements as described by Moum and Smyth (2006).

Our purpose here is to quantify the total wave energy, the rate at which it is transported and to identify the role of individual terms in the energy transport as they appear in (10). Identical computations are applied to each of the twelve clearly-defined waves examined by Moum and Smyth (2006). From a time series, we cannot differentiate between the time dependence of $E$ and the spatial divergence of $f_E$, so we cannot make a direct assessment of (9). Instead, we assess $f_E$, $E$ and $\rho_0 \epsilon$ independently, find that $\rho_0 \epsilon$ is small over short propagation ranges and...
compare $f_E$ and $E$ via (14).

**Fig. 12.** $\langle KE \rangle_x$ vs $\langle APE \rangle_x$. A neutral regression fit (with 95\% confidence intervals), mean and median are indicated.

**a. KE, APE**

As an example calculation we show details of the first wave of the second group of waves depicted in figure 7\textsuperscript{6}. The energy density is shown in figure 11 in units of kJ m$^{-3}$. $KE$ is positive definite. $APE$ is positive in the vicinity of the wave (isopycnals are raised); $APE$ is slightly negative elsewhere, a consequence of the definition of $\rho_b$. This is reflected in the vertical integral of $APE$.

$\text{In the vicinity of the wave core, at least where } KE$

\[ \langle KE \rangle_x = 1.04 \text{ (0.89, 1.21)} \]

\[ \text{mean(KE) } / \text{ median(KE)} = 1.04 \text{ (0.89, 1.21)} \]

is non-zero, the vertical integral of $APE$ is about $4 \times$ that of $KE$. However, when integrated over the domain $(\langle \cdot \rangle_x = \int_{-\infty}^{\infty} \int_{0}^{H} (\cdot) dz dx = \int_{-\infty}^{\infty} \int_{0}^{H} (\cdot) cdz dt)$, estimates of central tendency indicate that $\langle KE \rangle_x \approx \langle APE \rangle_x$. The negative $APE$ remote from the wave core reduces the ratio of the integrated energy components by a factor of 3-4.

The ratio $\langle KE \rangle_x / \langle APE \rangle_x$ displays considerable scatter; this is particularly so among the low energy cases. The mean value is 1.48, with standard deviation 0.85 and standard error in the mean 0.25. The median value of $\langle KE \rangle_x / \langle APE \rangle_x$ is 1.20. Neutral regression indicates $\langle KE \rangle_x = 1.04(0.89, 1.21)\langle APE \rangle_x$, 95\% confidence limits in parentheses.

**b. Energy flux**

We first consider the three pressure-velocity flux (figure 13). As $\rho_{nh}$ is mostly negative (Moum and Smyth 2006, figure 9), so is the energy flux associated with it. The effect of the negative surface displacement above an elevation wave is similarly to cause a negative energy flux. The fact that the vertical integral is non-zero while $\rho_{0}g\eta_{H}$ is independent of $z$ is because $\int_{0}^{H} udz \neq 0$. This differs from the linear wave assumption (Kunze et al. 2002) and from that used for the rigid lid simulation of Venyagamoorthy and Fringer (2005). The hydrostatic pressure-velocity flux is $> 0$, and larger in magnitude than the sum of the other two terms. As a consequence, the net effect of pressure-work is a net positive (onshore, in the direction of wave propagation) energy transport (solid line in bottom panel of figure 14). The net pressure-work transport is of the same sign but smaller than the nonlinear advection term $uE$ in the example shown in figure 14. Statistically, over the twelve waves we have examined, $\langle uE \rangle \approx 2\langle up \rangle$. Again, this relation con-

\[ \langle uE \rangle = 1.04(0.89, 1.21)\langle APE \rangle_x, 95\% \text{ confidence limits in parentheses.} \]
tains considerable scatter, especially at low energies, the mean value of \( (uE)/(up) \) is 1.92 with standard deviation 1.84 and standard error of the mean 0.53. The median of the flux ratio is 1.54 and neutral regression yields \( \langle uE \rangle = 2.01(1.29, 3.28) \langle up \rangle \).

Depth-dependent time integrals through the waves provide a complementary view of the vertical structure of individual flux terms and of the net wave flux (figure 15). These flux profiles show consistent features common to all of the waves. Notably, the vertical distributions of advective and pressure-velocity terms differ significantly. The effect of the pressure-work transport term is to carry the energy close to the bottom. Dominated by the advection of \( A P E \), the nonlinear advection term acts farther away from the bottom.

c. Dissipative losses

Estimates of the energy dissipated via turbulence in the wave cores are derived from vertical profiles (figures 4,5) as well as from measurements at 1 m height above the seafloor from lander-based observations (figures 8, 9, A1). The profiles in figure 5 suggest that, in the wave cores, \( \epsilon \) is relatively uniform with height above the bottom and very close to its magnitude at 1 m height (figure A2). A scale estimate of \( 3 \times 10^{-6} \text{ m}^2\text{s}^{-3} \) seems like a reasonable upper bound to an average value in the wave cores. Integrated over the wave core, \( \langle \rho_0 \epsilon \rangle_x \) is \( \text{O}(1-10) \) \( \text{W m}^{-1} \). This is compared to the total wave energy, which is \( \text{O}(10^6) \text{ J m}^{-1} \) (figure 12). The ratio, \( \langle E \rangle_x / \langle \rho_0 \epsilon \rangle_x \) represents a dissipative decay time scale, which is \( \text{O}(10^{-8} - 10^0) \) s, or 1 - 10 d. If we take a typical wave period to be 500 s (figure 8, for example), this is equivalent to several hundred to several thousand wave periods. Or, at \( c = 0.5 \text{ m s}^{-1} \), a distance of 10s - 100s of km. This implies that, over short propagation ranges, we can neglect \( \rho_0 \epsilon \) in 9. However, it is quite likely that these wave propagate at least 100s of wavelengths, and the integral effect of dissipative losses must be considered over these longer propagation ranges. For our purpose, we consider (14) as a test of our estimates of \( f_E \) and \( E \).

We note here that we have not considered the energy required to propagate up the sloping shelf, which will appear as an energy sink in the long propagation problem. Evaluating this term is complicated by our incomplete knowledge of the mass transport by the waves. It is also complicated by the finite extent of the near-bottom waveguide. In fact, this waveguide may be partly due to the cumulative contributions of past wave trains that have transported dense fluid up the shelf.

d. Relating flux to energy

Direct comparison of \( \langle f_E \rangle \) to \( c \langle E \rangle \) reveals that they are nearly equal (figure 16)\(^7\). We have found it difficult to evaluate the uncertainty in these estimates, but it is doubtful that figure 16 suggests anything other than \( \langle f_E \rangle = c \langle E \rangle \). The mean value of \( \langle f_E \rangle / c \langle E \rangle \) is 0.88 with standard deviation 0.34 and standard error of the mean 0.10. The median value of this ratio is 1.04 and the neutral regression results in \( \langle f_E \rangle = 1.03(0.94, 1.14) c \langle E \rangle \).

\[ (f_E) = c(E) \]

\[ \langle f_E \rangle = 1.03(0.94, 1.14) \langle E \rangle \]

\[ \text{mean}(f_E/c(E)) = 0.88 \]

\[ \text{median}(f_E/c(E)) = 1.04 \]

FIG. 16. Comparison of \( \langle f_E \rangle \) vs \( c \langle E \rangle \). A neutral regression fit (with 95\% confidence intervals), mean and median are indicated.

6. Discussion

Analogies for the observed flows exist in atmospheric and laboratory flows. The “Morning Glory” in northern Australia has been considered a classic form of an internal undular bore in the atmosphere (Clarke et al. 1981). It has also been described in terms of solitary waves (Christie et al. 1981). The coincidence of undular bore and gravity current has been noted in the case of a thunderstorm outflow (Marks 1974) and depicted schematically by Rottman and Simpson (1989). The purpose of the experiments of Rottman and Simpson (1989) was to replicate the conditions for the generation of these atmospheric phenomena. Generation of intermittent turbulence resulting from gravity currents and solitary waves in the nocturnal atmospheric boundary layer has been observed by Sun et al. (2004).

The structure of near-bottom nonlinear internal waves must differ from that of waves propagating in a near-surface waveguide; the requirement that \( u=0 \) at the bottom means that there is an injection of vorticity into the wave core of one sign from below and of the other sign from above. A close look at figure 9 indicates an evolution of the near-bottom shear as the wave passed. A sharp reduction occurred at or following the increase in turbulence past the wave crest. This suggests that the strong interaction with the bottom may result in rapid diffusion of wave core momentum, introducing asymmetry across the wave.

In the case of surface-trapped waves, the wave-induced surface stress is minimal and the injection of vorticity from the surface must also be small. In surface-trapped
waves it may be that the primary source of turbulence is from shear instabilities at the bottom interface (Moum et al. 2003).

![Figure 15](image)

**Fig. 15.** Vertical profiles of fluxes integrated over the example wave shown in figure 9. The total transport is indicated as the vertical integral of the profiles.

The turbulence bursts in the waves are relatively short-lived (figure A1), at least in the leading waves. Turbulence appears at 1 m height within the wave cores and invariably disappears abruptly as the wave cores pass. The decay is much shorter than the local buoyancy time scale ((Smyth et al. 1997), indicating that it is not the stratification that attenuates the turbulence. This suggests that the turbulent fluid generated within the waves is trapped in and transported by the waves, not left behind to decay. It is consistent with the notion of "trapped cores" carried in the wave without contact with the ambient water (Lamb 2003) and evidence of acoustic scatterers trapped in wave cores (Klymak and Moum 2003).

There are several ways in which these near-bottom flows may impact shelf transport. High bed stresses act to resuspend seafloor particulate matter into the water column. These particles are raised high into the water column (many 10s of m) by vertical wave motions and thence irreversibly mixed by energetic turbulence. At the same time, we note that the direction of fluid transport by the waves must be in the direction of the energy transport (onshore). This is in the direction opposite to that of the Ekman transport in the bottom boundary layer beneath the northward flow that characterizes the shelf circulation during downwelling. Taken together, these motions may effect a significant vertical and cross-shelf transport which has not been previously been considered. So far as we know, none of these factors are incorporated into modern coastal circulation models.

Osborne and Burch (1980) assumed equipartition of kinetic and potential energies in their description of solitons in the Andaman Sea. They stated, but did not demonstrate, that equipartition holds "approximately" for KdV solitons in a 2-layer fluid. In fact, equipartition holds ex-
actly for KdV solitons independent of the upstream density and velocity profiles, provided that APE is computed with respect to the minimum potential energy in an infinite domain (Smyth and Klymak 2006). For geophysical solitary waves, however, the weakly nonlinear limit of KdV theory provides only an approximate description. Nonlinearity is generally not weak, and KdV theory can provide no guidance in this case. Here, we have suggested \( \langle K E \rangle \simeq \langle A P E \rangle \). For comparison, Klymak and Moum (2003) estimated \( \langle K E \rangle / \langle A P E \rangle \simeq 1.5 \) from a sample of three bottom-trapped waves with energy levels at the low end of those discussed here; this is a revision of the analysis presented in the original paper. Klymak et al. (2006) determined \( \langle K E \rangle / \langle A P E \rangle \simeq 1.4 \) for a single large amplitude, surface-trapped wave in the South China Sea with energy level four orders of magnitude greater than the waves analyzed here.

For internal waves that can be linearized, the energy flux is due solely to pressure-velocity work, the only pressure contribution being the internal hydrostatic component, \( p^h_\nu \) (Nash et al. 2005). It is straightforward to theoretically relate this to \( c_g(E) \) (Kundu and Cohen 2004), and this has been verified experimentally (Alford and Zhao 2006). The present observations indicate that an analogous result holds for finite-amplitude, nonhydrostatic waves with propagation speed \( c \): \( \langle f_k \rangle = c(E) \).

For nonlinear internal waves, the energy flux includes both a nonlinear advection term and a nonhydrostatic pressure-velocity work term that do not appear in the linear wave flux formulation. Yet the linear wave flux formulation has been applied to nonlinear waves (Chang et al. 2006). Since \( \langle uE \rangle \) is such a large component of \( \langle f_k \rangle \) (from figure 17, \( \langle uE \rangle \simeq \frac{3}{4} \langle f_k \rangle \)), this cannot be justified. It is significant that a larger fraction of the total energy flux is falsely represented by \( \langle uE \rangle \) (than \( \langle uP \rangle \)) because the sum of the nonhydrostatic plus surface displacement contributions to the pressure-velocity flux counters that due to \( p^h_\nu \) (as represented by the open circles and grey lines in figure 17). By neglecting the opposing contributions to the pressure-velocity flux and the nonlinear advection term, a closer approximation of the wave flux is achieved, albeit for the wrong reasons. Even then, this represents only 60% of the true value.

The cumulative energy transported by many wave trains is estimated by extrapolation from our small set of observations. We note that, while 4 wave trains were observed on 20 January (shown in figure 7), almost an entire day passed until the next wave train was observed. Our estimates of wave energy transport (neglecting the bore-like part) suggests \( \sum \langle f_k \rangle \simeq 20 \, \text{MJ m}^{-1} \), and that this occurred over a 48 h period. This in turn, implies an onshore energy transport by the waves alone of about 100 W m\(^{-1}\). Independent estimates of onshore energy transport indicate a similar energy flux by the \( M_2 \) internal tide (at 45°N, G. Avicola, pers. comm.).

Other analyses indicate that most of the energy at \( M_2 \) transported across the shelf in 120 m water depth during upwelling when stratification is significant across the shelf propagates shoreward through 80 m water depth to 50 m depth (120 m, 80 m and 50 m were the nominal depths of moorings at 45°N; G. Avicola, pers. comm.). On the other hand, during the downwelling experiment of January 2003, when a large portion of the shelf was unstratified, the proportion of \( M_2 \) energy reaching 50 m was reduced by a factor of 4. Is it possible that most of the energy in the \( M_2 \) frequency band at the shelfbreak went into nonlinear internal waves? Or that, at least the internal tide is an efficient generator of nonlinear internal waves at this time of year. Certainly the lack of stratification at midshelf (figure 1) indicates the problems inherent in internal wave propagation to the inner shelf.

The timing of the arrival of wave groups at the observing site is not \( M_2 \). However, it is possible that \( M_2 \) forcing is disguised by differential propagation speeds as stratification varies between shelf break and observational site (thus altering \( c \)) and that \( M_2 \) forcing is itself modulated by the variability of density interfaces at the shelf break caused by the mesoscale circulation.

In assessing the energetics of the waves we have reconstructed density from streamlines, which has produced nearly symmetric density structure across the waves. However, the waves (or wave groups) are clearly associated with a larger scale density front and the structure of density is not generally symmetric across the waves. The reasonable results we have obtained for the partitioning and transport of energy by the waves, however, suggests that the perspective we have taken is not greatly in error.

Our use of a symmetric density structure means that we have not attempted to address the more subtle problem posed by Henyey and Hoering (1997); that is, whether the potential energy difference across individual waves or wave groups is in balance with the energy dissipated by turbulence. It is clear though, that over short propagation ranges, dissipative losses are not important. So far as this particular geometry is concerned, most of the available wave energy will reach shallow water. Over long propagation ranges, say 100s of wavelengths, or 1000s of wavelengths (such as in the South China Sea; Ramp et al. (2004)), turbulence losses should be important, at least if our measurements of \( \% \) in these wave cores is representative.

7. Summary

Near-bottom flows over the Oregon shelf during winter downwelling exhibit characteristics of gravity current, internal bore and solitary wave. Groups of wavellike pulses with high-velocity cores \( (u \simeq 0.5 \, \text{m s}^{-1}) \) were observed at the leading edges of density fronts. These are associated with high bed stresses and internal turbulence that forms rapidly and disappears as the high-velocity wave cores pass.

Evaluation of nonlinear internal wave kinetic energy, available potential energy and energy flux indicates the following:
energy is roughly equipartitioned between kinetic and potential,  
\[ \langle KE \rangle \simeq \langle APE \rangle ; \]
- the energy flux is simply related to the total wave energy by the wave speed,  
\[ \langle fE \rangle \simeq c(E) ; \]
- the pressure-velocity energy flux includes significant contribution from both nonhydrostatic pressure and surface displacement, which, since they are of opposite sign to the hydrostatic pressure-velocity flux, reduces the total contribution of pressure-velocity work to the nonlinear wave energy flux; and
- nonlinear advection of energy is a significant component of the energy flux. Indeed it is a direct measure of the wave’s nonlinearity. For the waves we have observed,
\[ \langle uE \rangle \simeq 2\langle up \rangle . \]

The first two properties are consistent with linear waves. The latter two are unique to nonlinear internal waves.

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APPENDIX A Turbulence measurements

A spectrum of the cross-shore component of velocity recorded by an ADV at 1 m height with lowpassed current speed overplotted is shown in figure A1. Spectra were computed over 2048 points (204.8 s). The spectrum indicates one of the complications of quantifying turbulence from rapidly-sampled velocity records in shallow water. In 124 m water depth, the velocity signature of wind waves (2-5 s) at the bottom is negligible, but that of 12 s swell is not. The most energetic components of this signal are due to two swell components; 1 constant at about 12 s, the other varying in period from about 22 to 15 s over the record shown. The longer period swell component also increased in energy over the record. Turbulence in the spectrum record (figure A1) is identified by the broadband signal at frequencies greater than that of the swell. Turbulence associated with the wave pulses dominates the record. The rapid onset of turbulence as the velocity pulses pass indicates either rapid generation of turbulence within each pulse or advection of turbulence by the wave pulses. While this is important distinction to make, we do not have enough information from these data to tell the difference.

The problem of interpreting the high-frequency component of a spectrum when inertial subrange turbulence is advected by a random wave field past a point measurement of velocity was addressed by Lumley and Terray (1983) in order to interpret observations under much more extreme conditions in terms of the relative contribution of wave orbital velocities, very near the surface beneath wind waves in Lake Ontario. This was refined by Trowbridge and Elgar (2001) to apply to bottom ADV measurements in the surf zone. We use Trowbridge and Elgar (2001)’s method here to define the turbulence dissipation rate, \( \epsilon \), from our bottom ADV measurements. The full record of \( \epsilon \) computed over 204.8 s intervals is shown in figure A1.

A comparison of \( \epsilon \) determined from CHAMELEON measurements 1 m above the bottom and that determined from the ADV on the bottom lander is shown in histogram form (figure A2). These data represent measurements from the entire records shown in figures 4 and A1. Each data set includes four density fronts with sequences of wave pulses as well as bottom boundary layer turbulence of other origin. We note that the two distributions in figure A2 are similar. Mean and median values of the two distributions differ by about 50%, the lander values being larger (lander mean = \( 2.6 \times 10^{-6} \) m² s⁻³, lander median = \( 8.4 \times 10^{-7} \) m² s⁻³). Since the profile measurements under-resolve the time variability of the turbulence in the wave pulses, we might expect that the lander series samples more of the infrequent large values of \( \epsilon \), which could explain the difference in the distributions at high values of \( \epsilon \). Alternatively, this difference could be a real feature of the turbulence between these two periods. The differences at low values of \( \epsilon \) are explained by the differing noise levels of the two calculations (the lander ADV measurements demonstrated about 10× greater noise level).

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This is because the duration of the pulses is about 10 minutes while our profiling rate in this water depth was about 1 per 4 minutes.
FIG. A1. Spectrogram of cross-shore velocity measured by ADV on bottom lander. Individual spectra were computed over 2048 points of the 10 Hz-sampled signal (204.8 s). Ordinate is logarithm of frequency [Hz]. Abscissa is time. The third dimension is logarithm of spectral amplitude $[m^2 s^{-2}/Hz]$. The black line across the image plot is the current speed shown in figure 7; this is scaled to the ordinate axis in units of $[m s^{-1}]$. An example spectrum of vertical velocity is shown at the top of the figure with an inertial subrange scaled by measured dissipation and current speed ($V$); $a=12/55$, $\alpha = 1.5$. At the bottom is the time series of $\epsilon$ derived from the spectra.

FIG. A2. Comparison of $\epsilon$ estimates at 1 m height determined from the CHAMELEON profile measurements shown in figure 4 and from the bottom lander ADV measurements shown in figure A1. Median values of each distribution are indicated by the vertical lines at the top of the figure. The lander median is greater by about 50%.

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Written by J. M. Klymak
mailto:jklymak@ucsd.edu
http://opg1.ucsd.edu/jklymak/WorkTools.html