Nonlinear internal waves and plumes generated in response to sea loch outflow, AUV and time-lapse photography observations
Toberman, Matt; Inall, Mark; Dumont, Estelle; Griffiths, Colin

Published in:
Journal of Geophysical Research: Oceans
Publication date:
2017
Publisher rights:
©The Authors
The re-use license for this item is:
CC BY
The Document Version you have downloaded here is:
Publisher's PDF, also known as Version of record

The final published version is available direct from the publisher website at:
10.1002/2016JC012208

Link to author version on UHI Research Database

Citation for published version (APA):

General rights
Copyright and moral rights for the publications made accessible in the UHI Research Database are retained by the authors and/or other copyright owners and it is a condition of accessing publications that users recognise and abide by the legal requirements associated with these rights:

1) Users may download and print one copy of any publication from the UHI Research Database for the purpose of private study or research.
2) You may not further distribute the material or use it for any profit-making activity or commercial gain
3) You may freely distribute the URL identifying the publication in the UHI Research Database

Take down policy
If you believe that this document breaches copyright please contact us at RO@uhi.ac.uk providing details; we will remove access to the work immediately and investigate your claim.

Download date: 03. Nov. 2020
Nonlinear internal waves and plumes generated in response to sea-loch outflow, AUV, and time-lapse photography observations

Matthew Toberman1, Mark Inall1,2, Tim Boyd1, Estelle Dumount1, and Colin Griffiths1
1Scottish Association for Marine Science, Dunstaffnage Marine Laboratory, Oban, UK, 2Department of Geosciences, Grant Institute, University of Edinburgh, Edinburgh, UK

Abstract The tidally modulated outflow of brackish water from a sea loch forms a thin surface layer that propagates into the coastal ocean as a buoyant gravity current, transporting nutrients and sediments, as well as fresh water, heat and momentum. The fresh intrusion both propagates into and generates a strongly stratified environment which supports trains of nonlinear internal waves (NLIWs). NLIWs are shown to propagate ahead of this buoyancy input in response to propagation of the outflow water into the stratified environment generated by the previous release as well as in the opposing direction after the reflection from steep bathymetry. Oblique aerial photographs were taken and photogrammetric rectification led to the identification of the buoyant intrusion and the subsequent generation of NLIWs. An autonomous underwater vehicle (AUV) was deployed on repeated reciprocal transects in order to make simultaneous CTD, ADCP, and microstructure shear measurements of the evolution of these phenomena in conjunction with conventional mooring measurements. AUV-based temperature and salinity signals of NLIWs of depression were observed together with increased turbulent kinetic energy dissipation rates of over 2 orders of magnitude within and in the wake of the NLIWs. Repeated measurements allow a unique opportunity to investigate the horizontal structure of these phenomena. Simple metric scaling demonstrates that these processes are likely to be feature of many fjordic systems located on the west coast of Scotland but may also play a key role in the assimilation of the outflow from many tidally dominated fjordic systems throughout the world.

1. Introduction

The Earth’s coastal oceans form the frontline of terrestrial and oceanic interaction and are productive and dynamic environments. Coastal regions are typically fresher than the shelf seas or the abyssal ocean, and understanding how freshwater is assimilated into these areas is key to understanding them as a whole. Entrainment of fresh water is enhanced by turbulent processes, in particular velocity driven shear [Orton and Jay, 2005] and the presence of internal waves [Nash and Moum, 2005], generated in response to buoyant plumes. Geophysical scale buoyant plumes can be formed in a number of ways, perhaps most ubiquitously by fluvial estuarine systems [Nash and Moum, 2005], but they can also be generated by the tidally modulated outflow from sea lochs [Thorpe et al., 1983] or fjord sills [Farmer and Smith, 1980], via flow through constrained straits [Armi and Farmer, 1986] and even via anthropogenic buoyancy inputs such as the outflows from power stations [O’Callaghan et al., 2010; Chen and MacDonald, 2006]. Due to their very existence, buoyant plumes often propagate into stratified environments, allowing for two related phenomena: that of the layer that is intruding into, and the generation of internal bores which also require energy but propagate as a mass of water [Simpson, 1997]. Whether these waves radiate ahead of the gravity current as trains of nonlinear internal waves (NLIWs) depends on the ratio of the characteristic speed of the internal waves for the given stratification, to the propagation speed of the plume front. If the flow is supercritical, that is the plume front is faster than the waves then a train of internal waves may propagate...
behind the head of gravity current rendering it bore like. Whereas in the opposite subcritical regime, trains of NLIWs may be released and become freely propagating [Nash and Moum, 2005]. The tidally modulated outflow from Loch Etive, situated on the west coast of Scotland as shown in Figure 1, has been found to provide an ideal environment within which to observe both velocity shear and the genesis of internal waves [Thorpe et al., 1983; Boyd et al., 2010]. Loch Etive is surrounded by steep topography and is fed with fresh water from a substantial rainwater catchment of some 1400 km², the largest of all the Scottish sea lochs. This freshwater run off results in an annual fresh water input into the loch of some 3 × 10⁸ m³ [Edwards and Edelsten, 1977]. This low density fluid is restricted from entering Ardmucknish Bay by the disparity in tidal range between that of the bay and the loch interior, of some 2 m [Edwards and Edelsten, 1977], with the visually most impressive constriction being that known as the Falls of Lora, denoted sill 2 in Figure 1. As the tidal elevation in the bay drops below that of the loch interior, low density water is released and has been found to form buoyant gravity currents that result in the adjoining bay exhibiting intense near surface density stratification, which can support nonlinear internal waves (NLIWs) [Thorpe et al., 1983; Boyd et al., 2010].

During loch outflow, the plume is supplied with horizontal momentum associated both with the water density difference and that from a barotropic pressure gradient driven by the disparity in tidal elevation. The momentum supplied is great enough for the surface plume to encounter the headland at the north western extent of the bay. A portion of this momentum remains, which is then reflected back into this now highly stratified water column, generating an undular bore or a train of NLIWs that propagate away from the headland back toward the source of the buoyancy input as observed by Thorpe et al. [1983]. In the present study, evidence of the genesis and propagation of internal waves ahead of the plume, in a similar fashion to river plume systems [Nash and Moum, 2005], was also found. Examples of the surface signatures of both of these phenomena are shown in Figure 2. Ardmucknish Bay originally became an area of interest, with respect to internal wave activity, via the surface signatures that were first photographed by the seminal figure in

Figure 1. Study area location in a UK context, as demonstrated by red box in top right. The two outer most loch sills are bounded by red hatched areas and numbered. The locations of the moored CTDs and ADCP are demonstrated by the red circle and star, respectively. The orientation of a reference AUV track is demonstrated by the red line. The location of the time-lapse camera location is indicated by the white triangle.
gravity current research J. E. Simpson [Thorpe et al., 1983]. A portion of the present study sees a return to the use of standard optical photography, as well as the utilization of a velocity microstructure equipped autonomous underwater vehicle (AUV), to observe and track the dynamics of these features.

Provided internal waves are of sufficient amplitude, relative to their mean position below the sea surface, the velocity field generated extends up to the surface. This is manifested as stripes of alternating quiescent and rougher water forming at the convergent and divergent regions of the streamlines found beneath each peak and trough. During a calm sea state where capillary waves dominate, the convergent zones may be more quiescent that the surrounding water as at these points film accumulation is highest and a slick that damps capillary waves is formed [Thorpe et al., 1983]. In rougher sea states however, where wind driven gravity waves dominate over the surface tension, the convergence zones roughen the sea surface further by gathering together and thus intensifying surface waves, whereas the divergence has the opposite effect and leads to surface smoothing [Osborne and Burch, 1980]. The presence and dynamics of buoyant plumes can also be elucidated through this approach, due to the divergent zones at the frontal boundaries [Nash and Moum, 2005]. These patterns of surface roughness can be interpreted via the use of synthetic aperture radar (SAR) [Nash and Moum, 2005], and as in the present study, from oblique aerial optical photography [Thorpe et al., 1983; Wang and Pawlowicz, 2011; Richards and Bourgault, 2013].

The dynamic and evolutionary nature of buoyant gravity currents has provided a challenge to continuous direct observation, from their genesis mechanism through to the assimilation into the coastal ocean [Hetland, 2005]. Success has been achieved reproducing the key features of buoyant plumes in numerical studies, e.g., MacDonald et al. [2007], however, a high resolution fully nonlinear nonhydrostatic model is required in order to resolve the internal waves generated in these systems [Stashchuk and Vlasenko, 2009]. Many previous studies [Thorpe et al., 1983; Fong and Geyer, 2001; Orton and Jay, 2005; Nash and Moum, 2005; Pritchard and Huntley, 2006; Kilcher and Nash, 2010] demonstrate success in the observation of buoyant gravity currents either advecting past moored instrumentation, thus giving a measurement fixed in space, or making mobile observations with the use of a vessel. The first of these has shortcomings in its Eulerian nature, as each feature can only be observed once, although high vertical resolution can be achieved and measurements can be made very near the surface. The use of a vessel allows for repeated observations to be made on a time scale short enough to be quasi resampling the same features. However, ship time can prove costly, and more critically the draft of the ship may be deeper than that of the features themselves and thus the passage of the ship can lead to interference with the very processes which are hoped to be observed. An autonomous underwater vehicle (AUV) however, can offer the Lagrangian viewpoint of a moving vessel while being small and hydrodynamic enough not to interfere with the features through which it is moving. Gargett [1982] was perhaps the first to use an AUV to investigate gravity current related processes while investigating internal wave trains at the knight inlet, by...
“flying” the AUV through a train of waves. The author successfully elucidated the structure of the waves via the temperature and salinity signals, as well as enhanced turbulence in the troughs of the waves. Since this study, AUVs have been utilized in only a small number of investigations into buoyancy and internal waves phenomena. Levine and Lueck [1999] deployed an AUV in order to investigate an approximately 8 m thick thermal front, as well as to test the ability to accurately measure velocity microstructure from an AUV. MacDonald et al. [2007] used an AUV to investigate turbulent characteristics of the Merrimack River plume, Goodman and Wang [2009] observed the enhanced turbulence associated with trains of NLIWs, and Rogowski et al. [2014] deployed an AUV to investigate the buoyant discharge from the New River Inlet, North Carolina, however, the vehicle was not equipped with microstructure sensors.

To the best of the authors’ knowledge, this study is the first of its kind for which a velocity microstructure equipped AUV has been deployed simultaneously with that of the capture of aerial time-lapse images. This combination provides the potential to synchronously observe both the surface characteristics of the phenomena described as well the subsurface energetics and the resulting changes in water column structure and dynamics.

2. Methods

2.1. Observations

Observations were undertaken during two periods, from 8 to 10 February 2011, and on 6 December 2013. The first period consisted of the deployment of two stationary moorings, one equipped with a near bed and surface Seabird Microcat CTD, and the other a upward looking near bed, upward looking, 300 kHz RDI ADCP deployed for the duration of the study. Moored observations were augmented with daily AUV missions, sporadic vessel deployed CTD casts, and time-lapse photography. The locations of the two moorings, a reference AUV transect and the position of the time-lapse camera are shown in Figure 1. The timing of the three AUV missions undertaken on each of the 3 days, and the time-lapse photography are shown in Figure 3. During the first period, all instrumentation was deployed from the research vessel Calanus. The wind was light and predominantly from the south east throughout the observational campaign.

The second period of observations consisted of time-lapse photographs, from the same vantage point of the first period, and simultaneous vessel deployed CTD casts. On this occasion, live radio contact was possible between the vessel, and observers conducting the time-lapse photography, in order that direction may be given to the vessel to reposition with respect to surface features. The wind was moderate and from the

Figure 3. Time series of near surface and near bed temperature, salinity, density, and pressure from the stationary mooring, the location of which is shown in Figure 1. Grey boxes bound time between high and low water at the mooring location. Yellow box in temperature panel bounds the duration of the time-lapse photography sequence shown in Figure 6 and supporting information Movie S1. Yellow boxes in salinity panel bound the times of the three daily AUV missions. Yellow box in density panel bounds the section of data displayed in T-S space in Figure 16.
north west at during the morning but became light and turned toward the east during the duration of the
observations.

2.2. AUV Turbulence Data Processing
The instrumentation on board the SAMS REMUS AUV consists of a Neil Brown Ocean Sensors conductivity
and temperature probe as well as upward and downward facing 600 kHz RDI ADCPs. The AUV has been
adapted to allow for the inclusion of a Rockland Scientific International mASTP turbulence microstructure
package, equipped with two shear probes as well as pressure sensors and accelerometers. The computation
of turbulent kinetic energy dissipation rates, $\epsilon$ is undertaken via the integration of measured frequency
spectra following Boyd et al. [2010]. Corrections are made to account for contamination of the spectra from
mechanical noise arising from various sources, such as the vehicle propellers, via the cross correlation of the
frequency spectra with that from the on board three axis accelerometers [Goodman et al., 2006].

2.3. Time-Lapse Photogrammetry
Images were captured with an easily available, consumer quality digital camera, at intervals of 30 s. A
time-lapse composite of these standard images makes for a powerful tool to better understand the
dynamics of coastal systems, the real power of these photographs however lies in photogrammetric cor-
rection. This process allows for a quantification of the movement of surface features, and for complemen-
tary observations to be analyzed in a geographical and temporal context. The photogrammetric method
applied here is freely available as Matlab code courtesy of Daniel Bourgault and Rich Pawlowicz,
described in Bourgault [2008], and follows that of Pawlowicz [2003]. This is essentially a three stage pro-
cess involving the transformation from an image pixel coordinate system into that of a cartographic one.
This process requires knowledge of seven parameters, one that is intrinsic to the camera itself, the lens
focal length, and the remaining extrinsic parameters describing the position and focal direction of the
camera, specifically: latitude, longitude, elevation, angle from the horizontal, angle from the vertical, and
the direction relative to North. In this study, the focal length of the camera employed was known and the
camera position, longitude, latitude, and elevation were found via the use of a handheld GPS. The deter-
mination of the rotation angles is less straightforward as no in situ measurements are available. It is
though possible for these, or any of the other parameters to be estimated with the use of landmarks,
which appear in the image and of which the longitude and latitude are known. These could be islands or
headlands or even the known location of a ship or its wake. A nonlinear cost function can then be com-
puted in order to minimize the difference between the longitude and latitude of the known points and
that of the same points found via the photogrammetric transformation [Pawlowicz, 2003]. In order to
track surface features, Hovmöller diagrams are constructed by extracting a slice from each image within
the time-lapse sequence, which is chosen to be aligned as accurately as possible with the propagation
direction of the striped features described thus far. An average value of pixel intensity is calculated for
each of the five pixel sets perpendicular to its length and is stacked horizontally to form the Hovmöller
diagrams. These diagrams offer the ability, via the examination of characteristic gradients formed in the
composite diagram, to estimate propagation speeds of surface features.

3. Results
The reporting narrative of this work is led by the presence of the surface features visible in the photogra-
metrically rectified time-lapse photographs. The appearance and subsequent propagation of these fea-
tures is most powerful when viewed as a composite moving image, and the reader is thus directed
toward the three movies shown in the supporting information (S1–S3), in order to gain an overview of
the phenomena and to place them in geographical and tidally temporal context. There are however,
three feature types identifiable in the time-lapse photography, which are clearly demonstrated in Figure
2: (1) linear, plume features propagating away from the source of buoyancy, (2) NLIWs propagating ahead
of these features, and (3) NLIWs propagating away from the headland back toward the source of buoy-
ancy. This section begins by describing the multitidal cycle results obtained at the two mooring locations,
and a synthesis of the three AUV missions, in order to identify whether the conditions required for the
existence of these processes are satisfied. The remaining complementary observations will then be con-
sidered in order to investigate the characteristics of, and changes in water column structure in response
to, the phenomena listed above.
3.1. Near Surface Stratification

A prerequisite for the existence of any of the phenomena that form the focus of this study is near surface density stratification. The water column in Ardmucknish Bay was found to be persistently and stably stratified throughout the observation period. This is evident both at the mooring location, as demonstrated in Figure 3, as well throughout the area covered by the AUV transects, as demonstrated in Figure 4.

Throughout time series, shown in Figure 3, the water column is shown to be stratified with respect to both temperature and salinity. There is significantly larger variability at the surface with cooler 5.6–7°C, and fresher, S<20–29 near surface values compared to almost constant values of 7.6°C and 34.4 near the bed (variability <0.03°C, <0.08). The temperature stratification may instinctively appear to be unstable with cooler water at the surface, however, the density is almost entirely controlled by the salinity, with salinity changes describing 99% of the density variability. There is repeated tidal periodicity in both the near surface temperature and salinity, with rapid increases in both properties for approximately 2.5 h after each high water, followed by an almost equally rapid decrease to a similar values, followed by a period of constancy. This pattern is repeated in all but one of the five tidal cycles over which measurements were made (~day number 40.25), when in this case the increase in surface density begins at the time of high water.

The synthesis of density data obtained during all three AUV missions, shown in Figure 4, elucidates more of the vertical and horizontal structure of the density field. The near density stratification described thus far is clearly evident with the low density water being largely constrained to a pressure of less than 6 dbar with almost no density values of <1022 kg m⁻³ found further from the surface.

3.2. Tidal Velocity Structure

Reciprocating quasi barotropic currents, driven by the flood and ebb of the oceanic tide, can be seen in Figure 5 to extend up from the bed to a pressure of approximately 10 dbar. These currents exhibit values of less than 0.1 m s⁻¹ and are orientated along a North-East to South-West axis, aligned with the adjacent headland (see Figure 1). The strongest velocities, ≈0.3 m s⁻¹, are found nearer the surface and are thus thought to be driven by the buoyant plume dynamics. The direction of water flow near the surface is less coherent than at depth, although tidal periodicity is still present. The strongest currents recorded throughout the observational period are seen to flow toward the south west and are initiated at the end of the ebb phase. During each ebb phase however (shaded grey areas in Figure 5), there is complexity in the direction of the near surface flow, and there does not appear to be a pronounced and prolonged period of flow toward the north west, as might have been expected from the theoretical framework set out in section 1.
3.3. Plume and Wave Features Propagating Away From the Loch

The changes in near surface density in response to the arrival of the plume and wave-like features at the mooring location are demonstrated in Figure 6. This figure is composed of a selection of the photogrammetrically rectified images, synchronized with a subsection of the time series of near surface density observations, shown in Figure 3 (The full time-lapse sequence can be seen in supporting information Movie S1). The first image of the sequence, captured at 01:06 after local high water (AHW), shows the emergence of a semicircular stripe spanning the area between the small island to the west of the exit of the loch, and the spit to the north, which was observed to spread radially away from the source of buoyancy. This single stripe feature most likely represents a phenomenon that generates subsurface velocity convergence, such as the frontal zone at the head of a gravity current intrusion [Nash and Moum, 2005]. At 01:28 AHW, second part of Figure 6, the single stripe develops into a set of multiple stripes which are also seen to spread radially ahead of the gravity current. These multiple striped features are suggestive of a variety of phenomena, such as; the genesis of a train of nonlinear internal waves (NLIWs) supported by existing stratification ahead of the gravity current intrusion [Nash and Moum, 2005; Wang and Pawlowicz, 2011; Richards and Bourgault, 2013], the development of an undular internal bore, behind the head of a gravity current [Thorpe et al., 1983; Simpson, 1997] or the development of periodic multiple frontal structures [Garvine, 1984].
these phenomena relies on the existence of stratification ahead of the intrusion, the second can be manifested as a modification of the intrusion itself or of the existing stratification, whereas the third can exist as modification solely of the frontal intrusion itself. This emergence of multiple stripes ahead of the initial
single frontal stripe can also be seen clearly in the Hovmöller diagram constructed from the images captured during experiment one, shown in Figure 7. This diagram compares well with the conceptually equivalent plot shown in Nash and Moum [2005, Figure 4]. Nash and Moum [2005] describe, via Froude number criteria, the genesis and evolution of frontally trapped waves that are subsequently released when the plume becomes subcritical with respect to the first mode internal wave speed of the receiving waters. This Froude number is defined as $Fr = \frac{U}{c}$, where $U$ is the speed of the plume front and $c$ is the fastest internal wave speed possible in the fluid ahead of the front. Observations available for an analogous analysis to be applied to the results of experiment one, consist of the propagation speed of the plume and the NLIWs estimated from the Hovmöller diagram, Figure 7, and a CTD profile taken before the arrival of either features (shown in second part of Figure 6). The leading edge of the plume is visible in Figure 7 as a linear feature in time/distance space from the origin of the time axis, at 00:53 AHW, and at approximately 100 m from the south eastern end of the image transect. It is shown to propagate a distance of approximately 125 m in 11 min, resulting in a characteristic velocity of $\sim 0.2 \text{ m s}^{-1}$. Between 01:12 and 01:40 AHW, evidence of NLIWs emerge in succession, which are also shown to linear in time/distance space, and although it is not possible to trace every bright line from emergence to the end of the sequence, as a group they exhibit a range of velocities between 0.35 and 0.4 m s$^{-1}$, all of which are faster than that of the leading edge of the front. The vertical modal structure, and thus the possible phase speeds of internal waves can be calculated by solving the eigenvalue problem for the complex vertical velocity amplitude, formed as a function of the density structure. This is achieved by numerically solving the Taylor-Goldstien equation using an established code [Klink, 1999]. With respect to the density profile yielded from the CTD cast ahead of the waves, the primary mode and thus fastest possible wave speed was found to be $0.27 \text{ m s}^{-1}$. The comparison between the analytically derived internal wave speed and that estimated from the Hovmöller diagram therefore demonstrates that the plume is subcritical, that is, the propagation speed is less than that of the first mode internal wave speed, with respect to the ambient stratification. This density profile derived phase speed also compares reasonably well with that estimated from the Hovmöller diagram, although these observed speeds

![Figure 7](image-url)
are all greater than that of the first mode density profile derived values. This disparity may be explained by a potential nonlinearity of the wave forms. A common theory applied to describe oceanographic NLIWs is a modified form of the Korteweg-de Vries equation \[\text{Grimshaw, 1997}.\] Solutions to this equation, for the wave speed in an idealized two layer flow is given by \[c_{\text{NLIW}} = c_0 - \frac{2\pi}{3},\] where \(c_0\) is linear first mode internal wave speed, \(\eta\) is the wave amplitude and, \[x = \frac{3c_0}{2}(h_2 - h_1)/(h_1h_2),\] where \(h_1\) and \(h_2\) are the upper and lower layer thicknesses of the idealized two layer flow in question \[\text{Wang and Pawlowicz, 2011}\].

The observations available in the present study do not offer a trivial method to calculate either an exact upper layer thickness or the wave amplitude, thus we prescribe bounds that will result in the calculation of a range of \(c_{\text{NLIW}}\) values. The upper layer thickness limits based on the CTD profile shown in Figure 6, are chosen as \(3 < h_1 < 6\) m. The wave amplitude limits are based on a sensible lower limit of 0.5 m that will result in significant modification of the wave speed, while the upper limit is defined as the amplitude of waves observed by the AUV described in the subsequent section of 3 m. The resulting nonlinear waves speeds calculated using the linear wave speed above, are found to range to from 0.27 to 0.40 m s\(^{-1}\). This inclusion of the nonlinear term therefore allows the propagation speed estimated from the Hovmöller diagram to fall comfortably within the bounds of analytically derived values. Given the favorable visual comparison with previous studies, and the relative agreement with theoretically derived propagation speeds, we consider the stripes present in our time-lapse imagery to indeed be the surface manifestations of NLIWs, and will be referred to as such subsequently.

In terms of the near surface density, prior to the arrival of the waves the density at the easterly mooring location remained at an approximately constant value of 1018 kg m\(^{-3}\), as can be seen in Figure 6. As the waves propagated through the mooring location, see part 3 of Figure 6, no significant change in near surface density was observed, however immediately after the passage of these waves the arrival of the original plume front coincided with a brief and small, \(<1\) kg m\(^{-3}\), decrease in near surface density, see part 4 of Figure 6. Subsequently, a plume distinguished by linear flanks emerges, see parts 4 and 5 of Figure 6, which in the moving image sequence appears to be flowing strongly, parallel with these flanks, toward the north west (see supporting information Movie S1). The passage of this feature through the mooring location is accompanied by a period of approximately 1 h of a sustained, \(\approx 5\) kg m\(^{-3}\), increase in near surface density. This increase in density, coinciding with the propagation of surface features away from the loch, is counter intuitive with respect to canonical explanation given previously, of a low density buoyant plume emanating from the loch during ebb flow. However as can be seen in Figure 3 the near surface fluid remains less dense than that near the bed, thus these results do not necessarily preclude the existence of a near surface buoyant plume, but do point toward a more nuanced explanation, which is discussed further in section 4. The AUV transects undertaken simultaneously with the time-lapse photographic sequence (the path and timing of which are shown in supporting information Movie S1), demonstrate that the water within this plume is highly energetic, as can be seen in Figure 8. As the AUV passes into the plume, confirmed by a decrease in density of \(1\) kg m\(^{-3}\), a 3 orders of magnitude increase in \(\epsilon\) from \(\sim 10^{-8}\) W kg\(^{-1}\) to \(\sim 10^{-5}\) W kg\(^{-1}\) was observed, with respect to the ambient fluid outside of this feature.

Deeper water column modification in response to waves propagating away from the loch was investigated during the second observational period. Figure 9 displays water column density either side of a train of NLIWs identified in the time-lapse imagery. The first profile was undertaken at 02:28 AHW, loch-ward of a seaward propagating wave train. The second was conducted at 03:00 AHW, after vessel repositioning ahead of the waves. A clear difference is evident between the profiles. The pre wave arrival (second) profile demonstrates a thinner and more dense surface layer, whereas the profile conducted within the waves shows a thicker, less dense, mixed surface layer with a consistently less dense pycnocline below.

Further evidence of water column modification in response to internal wave phenomena is available in the form of surface stripes coincident with a quasi stationary group of CTD profiles. The reader is directed toward supporting information Movie S2 in order to place the CTD profiles in the temporal context of the time-lapse imagery, and Figure 10 for selected frames from this sequence. This group of 29 profiles was undertaken between approximately 02:50 and 03:40 AHW during the second observational period, in a quasi stationary fashion. The maximum separation is 120 m, however, the mean distance between consecutive profiles is 15 m. The mean location of selected profiles can be seen in Figure 10. To aid the interpretation of the multiple profiles, the values are interpolated onto a regular, 120 s by 0.5d b, time-pressure grid. This is shown alongside selected corresponding time-lapse images in Figure 10. The first of these examples
was captured before the arrival of the wave train that was seen to propagate though the location of the profiles. The water column is shown to exhibit near surface stratification, with water of density $<1021 \text{ kg m}^{-3}$ in the upper 1 dbar and thin pycnocline between 1 and 2 dbar, followed by a more diffuse density gradient, reaching a density of 1025.5 $\text{ kg m}^{-3}$ at the greatest pressure recorded. As the waves arrive at the location of the profiles, second part of Figure 10, there is a deepening of the surface stratified layer, and water of density $<1022 \text{ kg m}^{-3}$ is observed at a pressure of 4 dbar, however, the density measured nearest the surface remains unchanged. As the waves propagate through the profiling location, third part of Figure 10, there is a core of low density, $1020 \text{ kg m}^{-3}$, water present from the surface to a pressure of 4 dbar, below which the prewave arrival isopycnals are deflected downward.

Immediately behind the waves, with respect to their propagation direction, there is a region of indistinct water followed by a dark plume region (the edges of which are shaded red in Figure 10) similar to that described in the results of the time-lapse imagery during the first observational period. In between these two features, fourth part of Figure 10, there appears to be a return to the original structure with surface density increasing and isopycnals being deflected upward. As the dark linear patch moves through the profiling location, fifth part of Figure 10, the pycnocline becomes more diffuse and the near surface density does not
return to the low values observed at the start of the time series. As more of this patch passes the location of the profiles, last part of Figure 10, the isopycnals diverge further, leading to an increase in very near surface density, as observed previously, and a general deepening of the upper stratified layer, with density decreasing at a pressure of 7 dbar.

3.4. Waves Reflected Back Toward the Loch
Surface manifestations of NLIWs are evident in the time-lapse imagery, appearing as bright and dark bands that originate from the headland and propagate away toward the south east, as can be seen in Figure 11 (as well as supporting information Movies S1 and S3). The results of the concurrent CTD profiles, superimposed on these images in Figure 11, demonstrate vertical isopycnal displacements, in response to the bright bands in the imagery passing the sampling vessel, confirming that the surface signatures are indeed those of NLIWs. At 04:16 AHW, second part of Figure 11, as a bright band propagated through the profiling location there was a downward deflection of the isopycnals at a pressure greater than 2 dbar after which a deepening of the near surface low density layer was observed. As the bright patch behind the waves approached the profiling location, third part of Figure 11, there appeared to be a rethinning of the near surface layer coincident with the boundary of this patch passing through the profiling location. As the bright patch of water spread though the profiling location, sixth part of Figure 11, a deepening of this surface layer was then observed. The speed of these waves can be approximated from the Hovmöller diagram in Figure 12, in which they are clearly visible, and are shown to be propagating at $\sim 0.25 \text{ m s}^{-1}$. Using the same approach described previously to calculate the first mode linear internal wave speed via the density profile undertaken ahead of the waves (shown in Figure 11), yields a value of $0.24 \text{ m s}^{-1}$. In this case, the observed and analytically derived linear wave speeds are in good agreement. This result is encouraging in the support of the hypothesis that the surface stripes do indeed represent the surface signatures of internal waves. However, given the better agreement with the analytically derived nonlinear wave speed, found for the waves travelling in the opposing direction. It may be expected for the observed wave speed to also be in better agreement with a faster, nonlinear speed. One key feature of the waves propagating back toward the loch is that they are propagating in a velocity sheared environment. Although no direct velocity data are available for this period, the near surface flow is clearly evident in Figure 11, where the south easterly propagating waves can be seen to interact with the north westerly propagating plume. It may appear intuitive to propose that opposing surface flow would lead to a reduction in internal wave propagation speed, and thus explain the observed slow speed of the waves with respect to theory. However, an analytical study of an idealized two layer flow, with opposing velocity in the upper and lower layers, has shown that when the surface flow is opposing the internal wave propagation direction, an increase of propagation speed
should be observed [Choi, 2006]. Thus, in the case of the waves observed propagating back toward the loch, the opposing near surface velocity does not appear to explain the disparity between the observed wave speed and that derived from nonlinear theory.

Figure 10. Examples from photogrammetrically rectified digital photograph series (full movie shown in supporting information S2). Tidal phase relating to the image capture time is shown graphically, and in hours after high water in the top left corner. (bottom left) Displays a surface plot yielded from the interpolation onto a regular, 2 min by 0.5 dbar, grid of the quasi stationary profiles, the location of which are shown by white dots and timing by the solid black lines, with the image time demonstrated by the blue full depth line.
There is further direct observational evidence of NLIW phenomena, and the associated energetics, in the form of temperature, salinity and $\epsilon$ measurements made during 3 AUV transects undertaken on day number 39, shown in Figure 13. During these transects periodic depressions in density were observed. A decrease from the transect average of approximately 1026–1025.25 kg m$^{-3}$ was observed, accompanied by a greater than 2 orders of magnitude increase in $\epsilon$ from the ambient values of $\sim 10^{-9}$ to a maximum value of $4.9 \times 10^{-7}$ W kg$^{-1}$. The low density patches exhibit a separation of approximately 50 m, although there may be distortion in the apparent separation of these features due to their propagation (discussed below) relative to the moving AUV. Given the relative speed of the features, of $\sim 0.2$ m s$^{-1}$, to the AUV of 1.5 m s$^{-1}$, the distortion is estimated to be of the order of a 10%, either elongation or contraction, depending on whether the vehicle is traveling in the same direction, transect 8 or opposing direction, transects 7 and 9, to the features. The presence of structured patches of low density water, observed at a constant depth, coincident with enhanced levels of $\epsilon$ are strongly indicative of a train of nonlinear internal waves of depression, where less dense water is displaced from a region above. There is a small modification to the vehicles path as it travels though the wave troughs evident in the pressure record (not shown). This modification is manifested as a damping of the depth control loop, rather than a sharp decrease in depth that may have been expected if the low density observations were a result of the vehicle simply moving upward through the stratified environment into a region of less dense water. This modification to the vehicles path may in fact

Figure 11. Examples from photogrammetrically rectified digital photograph series (full movie shown in supporting information S3). Tidal phase relating to the image capture time is shown graphically, and in hours after high water in the top left corner. (bottom left) Displays a surface plot yielded from the interpolation onto a regular, 2 min by 0.5 dbar, grid of the quasi stationary profiles, the location of which are shown by white dots in the image panel. The timings of the profiles are shown by the black lines, with the image time demonstrated by the blue full depth line. (inset bottom left) Displays the vessel deployed CTD cast that best represents, in terms of time and space, the ambient density structure ahead of the wave features.
actually be a direct consequence of the vertical velocities associated with the internal wave themselves, as found by previous operators of an AUV within an internal wave train [Garrett, 1982]. The waves are preceded by a wider, but similar magnitude density depression that was accompanied by a more intensively region of turbulent kinetic energy dissipation, exhibiting values of $10^{-2}$ W kg$^{-1}$. The wider region of density change accompanied by higher values of $\epsilon$ may well be evidence of a turbulent region associated with the head of a gravity current.

A fundamental advantage of AUV-based measurements is the ability to pass through the same feature multiple times, allowing an insight into their temporal and spatial evolution. In the case of the observations shown in Figure 13, it can clearly be seen that the four density depressions move from left to right, corresponding to at least a proportion of the propagation direction having been toward the south east, as illustrated in geographical context in this same figure. The propagation speed of the waves can therefore be estimated via the distance travelled by the leading edge of the train of depressions in the time taken between observations, resulting in a value of $\sim 0.2$ m s$^{-1}$. In comparison, the first mode internal wave speed calculated from the CTD profile nearest in time and space to the observations (shown in Figure 13, and calculated following the method described previously) is found to be $0.26$ m s$^{-1}$. Although this speed is greater than that observed, the observed value is susceptible to the following possible sources of disparity. The AUV transects may not be perfectly parallel to the propagation velocity and thus only a component of the velocity is observed. Additionally there may be preexisting velocity structure within the water column, and the waves may also be somewhat aliased, resulting in the leading edge of the train not being observed during each AUV transect. The alignment angle of the AUV transect to the NLIW propagation direction can be investigated with reference to the wave fronts that are visible in Figure 11. Although the features do not represent the same set of waves as those observed by the AUV, their orientation does allow for an insight into the propagation direction of waves generated via interaction with the headland. In the last image of

Figure 12. Hovmöller Diagram generated from the image slices extracted from time-lapse sequence shown in Figure 11. Red dashed lines demonstrate gradients corresponding to a speed of $0.3$ m s$^{-1}$, toward the south east and north west, respectively. Striped surface features are shaded red in order to aid distinction. Inset shows a single example image, with the transect used to extract the image slices colored red.
the sequence, examples of which are shown in Figure 11, the wave fronts have propagated to as close to the position of those encountered by the AUV as observed. The waves are still well aligned with the headland from which they originated. The wave fronts are found to be orientated 8° from perpendicular to the AUV transect. Adjusting the estimated propagation speed of the AUV observed NLIWs for this misalignment would only therefore account for a less than 1% increase in speed. NLIWs will be simply advected by barotropic currents and it is therefore possible to eliminate this component of the movement if velocities are known. The barotropic velocities, shown in Figure 5, at the time of the NLIW passage are found to be 0.05 m s$^{-1}$ toward the south west. Therefore not only is the velocity small compared to that of the NLIW propagation speed, but the direction is also perpendicular to their propagation direction, and that of the AUV transect, and thus both the impact of the barotropic flow on the waves and our ability to measure it is deemed to be negligible in the context of the phase speed estimation method. As with the internal waves propagating toward the south east observed during the second, observational period, the near surface flow is in an opposing direction, toward the north west (see Figure 5). Therefore, this near surface flow does not unfortunately offer a solution to the observed wave speed being slower than those predicted by theory.

Figure 13. (top) Density and (middle) TKE, for three AUV transects undertaken on 8 August 2011. Transect number and average time after high water is shown in the top right and left corners, respectively. Grey lines mark the positions of the leading edge of the nonlinear waves, which are also displayed in geographical context in the bottom left, along with the corresponding AUV transect number. (bottom left) Displays the vessel deployed CTD cast that best represents, in terms of time (shown in profile panel) and space (shown by black marker in map panel), the ambient density structure.
Another fundamental property of these waves, the amplitude, can be estimated by considering isopycnal displacement. The density depressions of $\sim0.5\,\text{kg}\,\text{m}^{-3}$, observed at the constant pressure of 10 dbar, corresponds to water that has been displaced downward, with reference to the same CTD profile utilized to calculate the wave speed, by approximately 3 m.

### 3.5. Bulk Observations of Turbulence and Mixing

Thus far attention has been given to the existence of specific phenomena, but the recurrent nature of the AUV transects allow for an examination of bulk energetics and the associated water column modification. Figure 14 displays mean values of density and $\epsilon$ as a function of time, for each of the twelve transects undertaken at pressure 10 dbar on day number 39. As can be seen there is positive correlation with respect to time for the $\epsilon$ values and a negative correlation for density, suggesting that active mixing is resulting in the entrainment of the low density fluid from above the depth at which the AUV is measuring. It is possible that low density water is simply being advected into the control volume observed by the AUV. However, the existence of the linear mixing line relationship between the transect mean values of temperature and salinity, demonstrated in Figure 15 supports a hypothesis of conservative mixing.

Figure 14. Transect mean $\epsilon$ (left axis) and mean density (right axis) as a function of average time of AUV transect in AUV hours after high water, for each transect undertaken on 8 August 2011. Vertical error bars describe the 95% confidence intervals.

Figure 15. Transect mean temperature as a function of salinity for each transect undertaken on 8 August 2011. Marker size denotes time after high water.
4. Discussion

4.1. Increase in Near Surface Density During Ebb Flow

The increase in near surface density observed at the mooring location during every ebb of the first experimental period is counterintuitive. The expectation is for the onset of the ebb, and the subsequent outflow of low density water from the loch, to result in a decrease in near surface density. In order to investigate potential mechanisms responsible for this density change, a section of the temperature and salinity values observed at the near surface are shown in T-S space in Figure 16. During the periods of both increasing and decreasing density, the values can be seen to collapse well onto a linear mixing line, suggesting a conservative mixing process between two end members. This mixing line passes through a region of T-S space very close to that occupied by the deeper coastal water, and meets the zero salinity intercept at \( C \approx 3.5 \). Given that the minimum and maximum air temperatures for the month in question were 3 and 8°C, respectively, it is highly likely that the changes in water density are a result of the vertical mixing of riverine water and that of the coastal ocean. The sign of the near surface density change is therefore dictated by the competition between the entrainment of higher density water from below, and the introduction of new, in an advective sense, buoyancy supplied from the loch interior.

A canonical description of an ebb flow is thus as follow. During the early phase of the ebb, when the near surface density is seen to increase, the water encountered at the mooring location is arriving from seaward of the loch interior and is thus of similar density to that already present. As mixing occurs in response to the energetic plume, that can clearly be seen at this point in the ebb flow in Figure 6, denser water is entrained from below and an increase in near surface density is observed. As the ebb flow continues, water that has entered the loch proper, and therefore undergone vigorous vertical mixing with the riverine water, supplies enough buoyancy to overcome the entrainment of higher density water, and the expected decrease in near surface density is observed. The near surface density is therefore dictated by balance between the advection of a horizontal density gradient, \( \partial \rho / \partial x \) and the mixing across the vertical density gradient \( \partial \rho / \partial z \).

Figure 16. (top) Temperature as function of salinity, observed at the eastern mooring during a section of the ebb phase of the first observational period (the timing of which can be seen in Figure 3). Isopycnals are drawn with black lines. Grey line represents a linear regression demonstrating a proposed mixing line and grey square marker represents the average temperature and salinity measured near the bed for the same period. (bottom) Corresponding density time series.
4.2. Mechanisms Responsible for Enhanced $\varepsilon$

Given that the observations made by the AUV of a train of NLIWs, are synchronous with enhanced levels of $\varepsilon$, and that the water column density observations made during or after the passage of these and other trains of NLIWs, all demonstrated evidence of mixing. What are the possible mechanisms responsible for this supply of turbulent energy? There are two main candidates, one being that preexisting turbulent fluid is transported by the waves to the point of measurement, either horizontally in a trapped core, or vertically with the wave amplitude [Moum et al., 2007]. The other possibility is that the turbulence is generated locally through instabilities produced as a direct result of the velocity shear associated with the waves themselves [Moum et al., 2003]. It is perfectly possible for these two mechanisms to occur simultaneously and there may not always be a clear distinction between them, particularly if the NLIWs are thought to be undular through instabilities produced as a direct result of the velocity shear associated with the waves themselves [Boyd et al., 2010]. Prior to the AUV-based observations of the NLIW train, the moored velocity measurements demonstrate that a region exhibiting high levels of velocity shear (see Figure 5), and thus potentially turbulent fluid, is indeed available for transport. It is therefore possible that the downward displacement of fluid from this sheared region, at the interface of the near surface low density layer could be the source of the enhanced $\varepsilon$. The source of the turbulent fluid is therefore also that of the low density fluid and thus it may not be trivial to distinguish between the waves either transporting, or directly generating turbulence. Boyd et al. [2010] also reported enhanced levels of $\varepsilon$ associated with NLIWs, supported by a near surface layer interface in Ardmucknish Bay, and suggested that velocity shear at the base of this layer may be a source of this turbulence. Boyd et al. [2010] quantified this by calculating a bulk Richardson number, $Ri$ based on upper layer mean values of $N^2$ and $(\partial u/\partial z)^2$, of $Ri_b=1.2$. This value is above the 0.25 required to exceed the Miles Howard condition [Miles, 1961], however, the authors note that it is sufficiently close to this value, given the bulk nature of the Richardson number, to account for the observed mixing.

We can perform an equivalent analysis for conditions which best represent the bulk properties of the density stratification and velocity shear present at the time of the observations of enhanced $\varepsilon$ associated with NLIWs. The upper layer mean $N^2$ is estimated using the average profile that results from the synthesis of all available CTD profiles (not shown), undertaken on the day of the AUV derived NLIW observations, and is found to be 0.01 s$^{-2}$. An upper estimate for the velocity shear, $(\partial u/\partial z)^2$, in the near surface layer was estimated from the difference between the velocities found at a pressure of 3 dbar, shown in Figure 5 ADCP, and the bin 2 m below, and was found to be 0.03 s$^{-2}$, yielding a $Ri_b=0.3$. The values used may not be perfectly representative due the space/time disparity with respect to each other and the sparsity of both measurement and vertical resolution, but they do allow for an insightful scaling, and suggest that instabilities at the interface of the near surface low density layer and the more dense fluid below may well be a significant contributor to the observed elevated levels of $\varepsilon$. In support of this argument observations made when the AUV passed into regions of consistently lower density, than that of the surrounding water, also demonstrated enhanced levels of dissipation, as shown in Figure 8, confirming that this low density near surface region is indeed a turbulent one. In conclusion, whether the NLIWs are vertically transporting externally generated turbulence or generating turbulence in themselves has proven to be unclear. However, it has been shown that the presence of these waves results in enhanced levels of turbulence being found at greater depth than in their absence and that the deepening of the surface layer in response to the passage of NLIWs demonstrates that they are responsible for active mixing.

4.3. Scaling $\varepsilon$ Within Locally Observed Conditions

Using values that describe the local stratification and velocity shear, it is possible to parameterize $\varepsilon$, and thus with comparison, place the observations made within this study in the context of the conditions in which they were undertaken. Previous authors [Jurisa et al., 2016] have successfully applied the parameterization of Kunze [2014] to parameterize turbulent dissipation rates at the base of a river plume, given as

$$\varepsilon_{kunz} = \left(1 - Ri_b\right)N^2\frac{(Ri_b^{-1}-Ri_c^{-1})(Ri_b^{-1/2}-Ri_c^{-1/2})}{96},$$

where $Ri_b$ the flux Richardson number and $Ri_c$ the critical Richardson number are taken by Jurisa et al. [2016] to be constant values of 0.17 and 0.5, respectively. Jurisa et al. [2016] note that there is an inherent ambiguity in prescribing the length scale $L_s$, defined by Kunze [2014] as the thickness of the unstable shear layer. Jurisa et al. [2016] conclude that this length scale should primarily be a function of the maximum
overturning length scale, being that of the upper layer thickness, \( h \), thus \( L_s = h/\delta_h \) with \( \delta_h \) as a constant scaling factor. In the study of Jurisa et al. [2016], the upper layer thickness \( h \), was observed to be \( \sim 7-10 \text{ m} \) and the scaling factor \( \delta_H \) applied was 5. Here we apply a scaling factor to give an approximately equal ratio of \( h \) to \( \delta_h \). We prescribe the upper layer thickness bounds described previously, of \( 3 < h < 8 \text{ m} \). These bounds are then applied to equation (1) with a \( \delta_h \) of 3.5 and the \( N \) and \( R_h \) values calculated in the previous section. This results in an upper lower bound for the parameterized \( \epsilon_{kwb} \) of \( 3.5 \times 10^{-6} - 2.5 \times 10^{-5} \text{ W kg}^{-1} \). The values found within the troughs of the internal waves of \( 4.9 \times 10^{-7} \text{ W kg}^{-1} \) fall below this range, however, it should be noted that levels of turbulent dissipation can vary within the vertical structure of an internal wave form, and measurements for this study were only made at a single depth. The values observed within the plume itself, \( \sim 10^{-5} \text{ W kg}^{-1} \) (see Figure 8), which is an environment for which the parameterization is better suited, fall comfortably within this range. The values of \( \epsilon \) we found within the troughs of the NLIWs are also lower than those found during a previous study of Gargett et al. [1984], of between \( 4.7 \times 10^{-6} \) and \( 3.8 \times 10^{-5} \text{ W kg}^{-1} \), that calculated \( \epsilon \) from a heated platinum film probe mounted to an AUV that was flown through internal wave troughs. We do though appreciate that.

4.4. Sea-Loch Outflow in a Wider Geographical Context

The present and previous studies investigating sea-loch outflows [Thorpe et al., 1983; Boyd et al., 2010] have focused on Loch Etive. These and similar fjordic systems are however prevalent across the north west coastline of Europe and thus influence the fresh water input onto the entirety of the northwest European shelf. Given this importance, it is reasonable to ask: can the outflows be characterized in terms of simple and available metrics, that may allow for an insight into the likelihood of the presence of the energetic processes described thus far? In order to both characterize and nondimensionalize the outflow regimes, we follow Inall et al. [2015], after Farmer and Armi [1986] in defining a Densimetric Froude number, that compares the flow velocity to the speed of the fastest linear internal wave mode, of the form:

\[
Fr = \frac{U}{\sqrt{g'd}}
\]  

(2)

where \( U \) is a characteristic velocity of the outflow, \( d \) is the mean sill depth, and \( g' \) is the reduced gravity for two layer flow at the sill. This scaling offers an insight into whether the flow is undergoing internal

Figure 17. The components of Froude numbers composed of: average current speed at outermost sill, as a function of the first mode internal wave speed. For the outflow for 63 sea lochs obtained from Scottish Sea lochs: a catalogue [Edwards and Sharples, 1986].
hydraulically control as the Froude number approaches one Inall et al. [2015], and thus may exhibit the phenomena described thus far. The catalogue of sea lochs compiled by Edwards and Sharples [1986] offers various measured and estimated values describing more than 100 of these fjordic systems. For the characteristic velocity, \( U \), we choose the tidal mean current speed at the most seaward sill of each loch, which is calculated in the catalogue as a function of the tidal volume of the upstream area of the loch and the cross sectional area of the sill. The choice of buoyancy term, \( g' \), proves more challenging, as we desire a term that includes the density difference between the outflowing and receiving water, of which neither are directly available. However, the authors of the catalogue do estimate a salinity reduction of the water in the loch interior, relative to the incoming coastal water. This is calculated as a function of the ratio of the freshwater input into the loch to the tidal inflow. We apply this salinity reduction to a representative, oceanic salinity of 34 and utilize the resulting value with a temperature of 10°C in order to estimate the density of the outflowing water. This then allows for the calculation of values of \( g' \) for each of the 63 sea lochs for which all of the required values are available. The results of this process are shown in Figure 17 as a plot of \( U \), as a function of \( \sqrt{g'd} \). As can be seen the Froude number associated with Loch Etive is found to be greater than one, as is that of Loch Linnhe, which has been known to exhibit surface manifestations of internal processes (A. Dale, personal communication, 2014). We propose that a Froude number of greater than one suggests that the outflow from the loch in question is likely to be hydraulically controlled and thus exhibit the energetic phenomena described in this study thus enhance the assimilation of fresh water into the adjacent coastal ocean. More than half (33 of 63) of the sea lochs for which data were available exhibited Froude numbers of greater than one, suggesting that a significant proportion of fresh water assimilated into the coastal ocean over the north west European shelf occurs under the influence of these energetic phenomena.

5. Conclusions

It has been demonstrated that there is indeed a tidally pulsed release of brackish water from Loch Etive during each ebb phase. However, this is preceded by an increase in near surface density within Ardmucknish Bay. These density changes are dictated by the competition between the entrainment of deeper denser water and the stratifying effect of buoyancy supplied by water released from the loch interior.

Trains of NLIWs were observed to propagate away from the loch, generated by the interaction of the buoyant outflow with the existing remnant stratification. NLIWs were also seen to propagate in the opposite direction, generated by the reflection of this outflow into the strongly stratified near surface layer resulting from the outflow itself. In all of the three cases considered, perfect agreement has not been found in the comparison of observed wave speed with that of the analytically derived value. While the available supporting information observations do not offer a completely satisfactory explanation for this disparity, the agreement within a factor 1.5 of the observational value allows us to be convinced that internal wave processes have been observed.

All of the waves were demonstrated to have resulted in active mixing, and the reflected waves were observed to be associated with a 2 order of magnitude increase in the turbulent kinetic energy dissipation rate. The turbulent dissipation rates observed in the trough of the waves were found to be less than both those reported by previous authors, and a parameterization based on the local values of velocity shear and stratification, however, the values within the buoyant plume itself where found to be in agreement with parameterized values.

Simple scaling suggests that these processes are likely to be feature of many fjordic systems located on the west coast of Scotland and may also play a key role in the assimilation of the outflow from many tidally dominated fjordic systems throughout the world.

Both the use of an AUV to make turbulence measurements and oblique shore based photogrammetry are still relatively novel observational techniques, and the present work is thought to represent the only study for which these two techniques have been used concurrently.

The REMUS AUV has proven to be valuable tool for investigating near surface phenomena. It has been shown to be able to make reliable CTD measurements as well as velocity microstructure, and to be able to resample horizontally evolving flow features. The shore based oblique photogrammetry proved to be a
successful method for identifying flow features and the surface manifestations of NLIWs, as well as for the quantification of propagation characteristics.

Unfortunately, coincident observations of surface features captured in the time-lapse imagery and patterns of enhanced levels of $\epsilon$ observed by the AUV, associated with a train of internal waves, were not achieved. At the time the AUV passed under such surface features it was undertaking a transect at too great a depth to observe the turbulent signatures of the waves. This however does also strengthen the case for the use of an AUV, as conventional vertical profilers would stand little chance of making reliable measurements within the first few meters of the water column. Observations of the AUV passing into a buoyant plume identified in the time-lapse imagery were though, found to be coincident with enhanced levels of $\epsilon$.

In conclusion, these two observational methods have proved to be highly complementary. The interpretation of the cutting edge microstructure equipped autonomous underwater vehicle data was made far richer by the insight of the flow regimes gained through the relatively “low-tech” time-lapse photography that truly demonstrated the complexity of the hydrodynamics exhibited in the fjordic system investigated within this study.

Acknowledgment

This work is dedicated to the memory of Tim Boyd. This work was funded jointly by the National Environment Research Council and the Marine Alliance for Science and Technology for Scotland. Data presented in this manuscript are available from the British Oceanographic Data Centre (enquiries@bodc.ac.uk).

References

Nash, J. D., and J. N. Moum (2005), River plumes as a source of large-amplitude internal waves in the coastal ocean, Nature, 437(7057), 400–403.


