Observations of Small-Scale Processes Associated with the Internal Tide Encountering an Island

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ABSTRACT

Current-meter, temperature, and microstructure observations of the large-amplitude internal tide shoaling on the continental shelf of the east coast of northern New Zealand show the complexity of the internal kinematics and mixing. The propagation speed of the main internal wave was around 0.3 m s^{-1} , and nonstationary time series analysis was used to locate the trailing short-wavelength internal waves in frequency (periods of around 40 min) and tidal-phase space. The average energy dissipation rate ($5 \times 10^{-8} \text{ m}^2 \text{ s}^{-3}$) was an order of magnitude smaller than that observed on the open shelf in other studies, but peaks in dissipation rate were measured to be much greater. The vertical diffusivity of heat was around $10^{-4} \text{ m}^2 \text{ s}^{-1}$, comparable to, or greater than, other studies. Examples of the scale and sporadic nature of larger mixing events were observed. The behavior was complicated by the nearby steeply shoaling coast of the Poor Knight Islands. Consistent reflected wave energy was not apparent.

1. Introduction

The interaction of tides, stratification, and bathymetry at the shelf edge generate shoreward-propagating internal waves (Baines 1982). The waves have attracted a good deal of attention because, unlike many internal oceanographic processes, they are readily observed in satellite imagery (Liu et al. 1998). This attention is warranted as the internal waves are clearly common (e.g., Sherwin et al. 2002) and often result in some large isopycnal perturbations, sometimes reaching many tens of meters. Such large excursions force substantial velocities and mixing. Furthermore, during periods of favorable conditions for their creation, they are relatively frequent (i.e., 2 times per day rather than sporadic storm forcing).

The passage of these waves generates a rich variety of processes that have implications for many facets of shelf ecology. While containing far less energy than barotropic tides, the baroclinic tides generate sheardriven turbulence and entrainment that results in enhanced diapycnal transport within the interior of the water column (e.g., Sharples et al. 2001). This vertical mixing provides a significant source of nutrients to the photic zone. As well as affecting nutrient dynamics, the waves influence zooplankton, generating shoreward transport of larvae or providing mechanisms for con-

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centration of larvae (Pineda 1999; Lamb 1997). The passage of large internal waves can also drive resuspension of benthic sediments (Bogucki et al. 1997) and subsequent evolution of suspended sediment heterogeneity (Wang et al. 2001).

The internal waves usually form as waves of depression in regions where the seabed slope matches the internal characteristic slope. In waters where the pycnocline is around middepth, the leading edge of the shoreward-propagating wave flattens while the trailing edge steepens, eventually forming a sequence of waves (e.g., Orr and Mignerey 2003; Gerkema 1996). The trailing waves can generate significant internal shear and instability (Sandstrom and Oakey 1995). As the main wave (and accompanying higher-frequency waves) progresses into shallower water, the leading edge will flatten and the main wave will break (Helfrich 1992; Michallet and Ivey 1999), sometimes resulting in waves of elevation (Orr and Mignerey 2003).

Factors controlling wave evolution include nonlinear growth, bathymetry, rotation (Gerkema 1996), and stratification. Theory provides insight in slowly varying situations (Holloway et al. 1999), and a means of categorizing the waves. The linear small-amplitude mode-1 long-wave internal phase speed for a linearly density stratified water column is (e.g., Hutter 1984)

$$c_{pc} = \frac{NH}{\pi}, \qquad (1)$$

where *H* is the water depth and the buoyancy frequency $N \,(\text{rad s}^{-1})$ is given by $N = \sqrt{(g/\rho_0)(\partial \rho/\partial z)}$, where ρ_0 is a reference water density, *g* is gravitational acceleration, and $\partial \rho/\partial z$ is the vertical gradient of density. An equivalent rigid-lid, two-layer $(h_1, h_2$: upper- and lower-layer thicknesses) phase speed is

$$c_{p2} = \sqrt{g' \frac{h_1 h_2}{H}},\tag{2}$$

where $g' = (\Delta \rho / \rho_0)g$ is the modified gravity ($\Delta \rho$ is the density difference across the interface). Weakly nonlinear theoretical approaches (e.g., Lamb and Yan 1996; Holloway et al. 1999) allow the wave to evolve as it moves onshore along a two-dimensional, across-shore slice. Numerical wave-evolution approaches often follow this two-dimensional perspective.

Sherwin et al. (2002) showed that the two-dimensional "slice approach" potentially underestimates wave evolution. This highlights the need to examine wave propagation and dissipation in regions of alongshore bathymetric variation (e.g., Noble et al. 1988; Cresswell et al. 1996; Liu et al. 1998; Holloway and Merrifield 1999; Small 2001; Lynett and Liu 2002). These studies show how the internal wave sequence refracts as the phase speed changes with the progression of the waves across the region of bathymetric variation.

With increased amplitude comes the likelihood of velocity shear overcoming buoyancy-induced stability, whereby the local gradient Richardson number $Ri_g = N^2/U_z^2$ (U_z is the vertical gradient of horizontal velocity) decreases beneath a critical value Ri_{gc} ($R_{gc} = 0.25$) and the flow becomes unstable and stirring and mixing ensue. Observation suggest instability and turbulence at values of apparent $Ri_g > 0.25$ (Sandstrom et al. 1989) and that around 20% of energy in the internal tide-driven wave packet is lost to turbulent dissipation (Sandstrom and Oakey 1995).

This illustrates the importance of observational studies and the role of turbulent dissipation. Theory is useful as a starting point for wave categorization, and laboratory and numerical work illustrate the complex interior motion during wave evolution, especially from wave breaking (Taylor 1993; Michallet and Ivey 1999; Vlasenko and Hutter 2002). However, direct observation is required to guide understanding of how waves grow and dissipate at the field scale.

The present observations come from the Poor Knight Islands (Fig. 1), around 20 km off the east coast of Northland, New Zealand. The islands form an isolated outcrop in an otherwise gently sloping shelf area (Sharples et al. 2001) that sustains a significant diversity of marine life. The region is subject to substantial internal wave activity (Sharples et al. 2001). The objectives of the work were to (i) provide an understanding of the kinematics of the main internal wave and accompanying smaller-scale internal waves in the region, (ii) examine the effect of the island on the wave, (iii) quantify the related mixing in time and space, and (iv) identify linkage between the mixing and the internal wave processes.

2. Field experiment and techniques

a. Location

The Poor Knights Islands include two major islands, Tawhiti Rahi and Aorangi, separated by a shallow channel (Fig. 1). The volcanically formed islands rise to peaks over 200 m above sea level. The small-scale bathymetry shows the steep cliffs plunging more or less vertically to depths of around 80 m and the bed joins the general shelf slope quite rapidly, within a few kilometers. As shown in Fig. 1c, the 100-m contour follows the island quite well, while the 150-m contour shows the abrupt change in slope whereby the gradient is reduced to the south of the moorings. This is complicated by the



FIG. 1. Experiment location showing (a) the North Island of New Zealand, (b) northeast shelf coast with depth contours at 100-m intervals, and (c) The Poor Knights Islands in detail, showing mooring locations for this experiment as well as the 100, 125 (dashed), 150, and 160 m (dashed) contours. Small-scale bathymetric features A and B and the propagation arrow P are described in the text.

short ridge seen in the 125-m contour around halfway up the northernmost island, Tawhiti Rahi.

With regard to tide-driven internal wave formation, for typical stratification the shelf slope matches the critical slope for generation of internal tides at depths of around the 600-700 m (Sharples et al. 2001) and 1000-1500 m (Stewart, 2001). The islands are 30 km onshore from the shallowest of these regions and around 25 km from the coast. Satellite imagery clearly shows the internal wave refracting in this region (Fig. 2).

Maximum barotropic tidal range is around 2 m, decreasing to a neap tide of around 1 m. Our experiment, conducted in November-December 2000, captured a reasonable portion of the falling phase of this variation. The deep water around the island relative to barotropic tidal amplitudes implies that there will be only moderate barotropic tidal currents, with the prevailing shelf currents being of comparable significance. In the general vicinity of the islands mean flows of around 0.2 $m s^{-1}$ run toward the southeast (Sharples and Greig



FIG. 2. Internal wave fronts (WF) off the Poor Knight Islands from 17 Nov 1995, where the scale bar to the bottom left is 5 km. Detail of European Space Agency ERS-1 SAR image modified from Sharples et al. (2001).

1998). The M_2 tidal amplitude is in the range 0.05–0.1 m s⁻¹ aligned at around 320°, close to alongshelf, other than at times of strong baroclinicity (Sharples et al. 2001).

b. Moorings

Three moorings were placed in a triangle adjacent to the embayment between the two islands (Fig. 1). Two moorings (IW1 and IW3) were placed on the 110-m contour, 425 m apart, while another mooring (IW2) was placed inshore on the 70-m contour around 810 m from IW1 (Table 1). In addition to standard mooring equipment shown in Table 1, a NORTEK VECTOR acoustic velocimeter was placed on a gimbaled vane on IW1 at 30-m depth. On IW3 an Ocean Sensors autonomous profiling vehicle traversed between 11 and 100 m. This operated successfully early in the deployment but power failure resulted in little data being collected during the period of intense data collection. The ADP at IW1 suffered from loss of signal in the surface waters during daylight. It was determined that this was due to daytime migration of zooplankton scatterers to depth. The velocities were replaced with interpolated equiva-

Instrument	Depth (m)	Δt (s)	Notes
Mooring IW1: depth 110 m, lat	35 28.30°S, lon 174 44.98°E		
SONTEK ADP	104	60	5-m bins
Hugrun thermistors	10, 20, 30 80, 100	60	No instrument at 90 m
Aanderaa RCM	20, 104	120	
NORTEK VECTOR	30	0.06	3-min bursts
Mooring IW2: depth 70 m, lat 3	35 28.37°S, lon 174 44.57°E		
SONTEK ADP	64	60	5-m bins
Hugrun thermistors	35, 50	60	
Aanderaa RCM	20, 64	120	
Mooring IW3: depth 110 m, lat	35 28.07°S, lon 174 44.96°E		
Hugrun thermistor	11	60	
OS APV	11-100	—	Moored profiler

TABLE 1. Mooring details.

lents derived from the VECTOR (30 m) and the RCM current meter (20 m). The ADP at IW2 did not suffer from this, presumably because of the shorter acoustic pathway.

c. Profiling

The RV *Kaharoa* anchored near IW1 from the period from 26 November to 2 December 2000, inclusive. Most of 28 and 29 November were lost due to equipment problems. Repeated profiles with a SeaBird conductivity-temperature-depth (CTD) package provided data averaged over 2-m bins, from 5-m depth to typically within 5 m of the bottom. The CTD data were recorded on 26 and 30 November and 1 and 2 December 2000.

On 27 and 30 November and 1 and 2 December 2000, repeated profiles with a temperature gradient microstructure (TGM) profiler, Self-Contained Autonomous Microstructure Profiler (SCAMP; Precision Measurement Engineering, Encinitas, California), enabled independent estimates of energy dissipation rate (ε_T) and diffusivity of temperature (K_T) to be made (e.g., Ruddick et al. 2000). Because TGM requires slow (0.1 m s⁻¹) free-fall profiles, horizontal drift created linelength limitations so that it was generally only possible to profile to around 80 m. The TGM profiles were recorded every 35–45 min over 13-h periods.

d. Analysis

The VECTOR acoustic Doppler velocimeter instrument at 30-m depth on IW1 recorded three components of velocity at 16 Hz from a sample volume of around 1 cm³. The sampling regime recorded 1024 datapoint bursts every 3 min. Other parameters (e.g., pitch, roll, heading, pressure) were recorded every second. Spectra of vertical velocity (in the instrument reference frame) contained energy at wave frequencies due to the wave itself and, more likely, due to wave motion affecting the subsurface float (10 m). However, at times, a frequency^{-5/3} slope spectral region would be elevated above the noise floor. By isolating this portion of the spectrum it was possible to generate an inertial dissipation rate estimate (modified from Drennan et al. 1996) ε_{r} so that

$$\varepsilon_I = C \frac{[S(f)f^{5/3}]}{U^{2/3}},$$
(3)

where C = 2.7, U is the local advective velocity (Drennan et al. 1996), and $[S(f)f^{5/3}]$ is the average over the useable region of the frequency spectrum S(f). The spectra were derived from the 1024-point data bursts using a 128-point overlapped Hanning window. A threshold level was developed to remove data with an insufficient signal-to-noise ratio.

The frequency content of the internal motions recorded in the thermistor and ADP data was identified using nonstationary analysis. Multibank filter wavelet transform (WT) analysis can resolve wave events in thermistor time series (e.g., Stevens 1999). A wavelet amplitude parameter W is defined as

$$W(a,b) = a^{-1/2} \int_{-\infty}^{+\infty} \zeta(t) g\left(\frac{t-b}{a}\right) dt$$

where t is time, a and b are the temporal dilation and translation, respectively, and g is the wavelet basis function. The Morlet basis function is used as it has a simple waveform structure (Torrence and Compo 1998):

$$g(t) = e^{ikt}e^{-t^2/2},$$

where $i = \sqrt{-1}$. The technique is applied to time series of vertically averaged density rather than temperature from any single thermistor, as these would often provide no useable variation during the passage of the wave of depression. Results are presented as an image of W^2 , with dilations ranging in period from 20 to 166 min.

The Ri_g generated in this study used the ADP and thermistor results. Density was resolved from temperature by assuming a linear relationship between salinity and temperature determined from the CTD observations. This was quite linear throughout the period; however the slope of the relationship did change slightly. The average slope used for S/T is 0.085 and the standard deviation of S/T over all the profiles is 0.016 or less than 25%. The effect of this approximation is small relative to the effect of the coarse vertical resolution and the inherently noisy derivative. Vertical gradient quantities N^2 and U_z were calculated at every minute at the resolution of the observations then binned into 17-m bins.

A number of factors reduce the certainty of the critical value R_{gc} when applied to observations; Ri_g is calculated from the division of two derivatives (often very small gradient) making it susceptible to noise. It is resolved from discrete samples rather than quasicontinuous distributions. Even if it were calculated from high-resolution profiles, this does not simplify matters in that this type of profile contains instability (i.e., negative N^2), so options for sorting the profile lead to uncertainty. With the present measurements an O(1)critical value is appropriate (e.g., Peters 1999).

Direct instantaneous estimates of turbulent kinetic energy were resolved using the TGM-derived estimate ε_T of the energy dissipation rate ε and diffusivity of temperature K_T . Determination of ε provides an upper bound on the turbulent buoyancy flux (Osborn 1980). At the same time K_T indicates the rate at which a scalar (in this case temperature) is diffusing. The microstructure analysis divided profiles into 50% overlapped 1024-datapoint bins (approximately 1 m). Spectral fitting was applied to gradient spectra from two thermistors operating at different gains that had noise removed. The spectra for each were matched to both the Batchelor and Kraichnan universal spectral forms (Nash and Moum 2002). The overall best-fitting spectrum from both thermistors using both universal spectra was assumed "correct" unless it failed a global goodness-of-fit test.

The thermal diffusivity K_T was resolved from the dissipation rate of thermal variance χ_T (i.e., the area under the temperature gradient spectrum) so that $K_T = \chi_T / (2\langle \Theta_z \rangle^2)$, where the $\langle \Theta_z \rangle$ is the stochastic average of the background temperature gradient. There is a range of choices in specifying $\langle \Theta_z \rangle$ from field data (Sherman and



FIG. 3. Experiment average of the temperature profile with the ± 1 standard deviation shaded. Superposed upon this are the average potential density profile and an example temperature microstructure profile showing the extreme isotherm compression possible (1405 27 Nov 2000).

Davis 1995). Here we use the gradient of the local resorted temperature segment.

Temperature microstructure cannot resolve large energy dissipation rates nor can repeat profiles be executed rapidly (Gregg 1999). Consequently, comparison of ε_I and ε_T , especially with regard to quantifying intermittency, was not possible. They resolved different parts of the dissipation rate domain with ε_T being resolvable beneath values of 5×10^{-5} m² s⁻³, whereas ε_I , in this application, had a noise floor at around 1×10^{-6} m² s⁻³. The two complementary methods provide bounds on the energy dissipation rate.

3. Observations

a. Background conditions

The experiment-averaged stratification derived from the CTD was relatively linear (Fig. 3) resulting in a temporally and vertically averaged N = 0.005 rad s⁻¹. However, the high degree of continual and large amplitude variation in the stratification means that the usefulness of some static average is limited. The onestandard-deviation shaded region indicates the extent of the variability at each depth. There was a decrease in this variability shallower than 40 m. The standard deviation does not fall off near the bed, implying that there was significant along-slope variability in temperature.

The contribution to the density profile by salinity was small and counterstable. Thus, the averaged density profile (at atmospheric pressure) was also linear. The example microstructure profile in Fig. 3 illustrates the degree to which the mean profile was distorted during the passage of a wave, to the extent that the bulk of the top – bottom temperature difference was compressed into a 3-m gradient region. The microscale temperature example included in Fig. 3 captured a large old overturn above the gradient region (40–47 m) and a small region of newer small-scale overturns in the few meters directly below the interface. This microstructure profile was not typical; instead, it represents the extreme isotherm compression possible during the passage of the internal wave.

The internal fluctuations generated a continuous displacement spectrum (Fig. 4), which does have a weak



FIG. 4. Displacement spectrum based on a thermistor at middepth at IW1 compared with the Garrett–Munk spectrum and including limits due to rotation (f) and average buoyancy frequency (N). The 95% confidence interval is included (Emery and Thomson 1997).

roll-off near the average buoyancy frequency. Energy persists beyond this cutoff because locally N often is greater than the spatiotemporal average. While the Garrett–Munk (Munk 1981) empirical description of open-ocean internal wave energy is not applicable near coasts (Kantha and Clayson, 2000), it at least provides a guide for background levels of energy. Clearly the local energy levels exceed this background, except perhaps around 0.3 cph.

b. The main internal wave

The wave amplitudes were substantial; indeed, no single isotherm persisted between 10 and 100 m over the 11.5 days of observation (Fig. 5a). It is possible that the chosen isotherm in Fig. 5a (16.25° C) was at times located in the unmeasured regions above 10 m and below 100 m. However, at the very least this suggests that almost the entire water column was replaced at the observation location every half tidal cycle. The pressure recorded at IW1 (Fig. 5b) sustained some short-duration, high-frequency perturbations that corresponded to periods of rapid change in the temperature data. It could not be determined if the pressure perturbations were real or due to the mooring being affected by the flow. Either way, it is clear that there were some highly dynamical events controlled by the baroclinic structure.

The IW1 temperature and velocity data (Fig. 6) illustrate that the apparently well-behaved temperature signal concealed a complex velocity structure. The frontal structure typically arrived with a southwestward-moving surface flow leading a deeper strong flow by a period of around 2 h. The deeper flow was the dominant velocity feature and is seen in the U panel between 70 and 100 m at times 330.85, 331.4, 331.9, and 332.4. Velocities in this feature exceeded 0.3 m s⁻¹ flowing away from the island. The north–south velocity shear was not consistent; for example, the fronts at 330.85 and 331.9



FIG. 5. Time series recorded during late 2000 (day 330 is 26 Nov) at IW1 of (a) the measured 16.25° C isotherm position derived from the thermistors and (b) the pressure deviation (i.e., barotropic tide). The interface displacement (16.25° C) ranged over the full measurable depth (14–100 m). There was a moderate correspondence between internal and surface tidal amplitude in that the interface displacement decreased later in the time series.



FIG. 6. A 2-day section of time-depth contours from station IW1 (spanning noon 26 to noon 28 Nov). (a) Temperature contours at 0.5°C intervals (the deepest isotherm denotes 16°C). Horizontal velocities filtered with a time constant of 1 h showing (b) V (north-south) and (c) U (east-west) such that -U is a westward flow into the Poor Knights and +V is a northward flow. ADP signal dropout is augmented with current-meter (20 m) and vector (30 m) data. The zero velocity contour is white. (d) The vertical velocity (positive is upward) shows the regions of signal dropout left blank.

sustained significant shear in the north–south direction. This is contrary to the front at 331.4, which initially started out in the same way, but by the time the trough arrived at the sample location the flow had evolved a reasonably homogeneous northward component of around 0.07 m s⁻¹. The velocity signal in the deep strong flow preceded the isotherms, so it was cold water being squeezed beneath the warm core of water associated with the incoming wave of depression and forced offshore. In some instances the return flow at the base of water column after the warm core had passed by was nearly as strong (e.g., Fig. 6c, time 331.55).

The regions of ADP signal dropout, marked in Fig. 6d, were not simply a function of daylight. There was correlation with the arrival of the main internal wave, as there were significant periods during daytime when

surface water signal returned (e.g., time 332.4). Presumably mixing within the core of the internal wave was sufficient to counter light-avoidance migration by the zooplankton.

The weaker vertical velocity component due to the internal wave resulted in the W velocity in Fig. 6d. This shows the smaller scale features filtered from the U and V data and emphasizes the division between the main internal wave and the shorter waves. The former was the large-amplitude wave arriving roughly every 12 h, while the latter were the short waves of around 40 min period that evolved from the large-scale wave. The vertical velocity signal from Fig. 6 is presented in an expanded view in Fig. 7. Note that these were the largest events of this type observed during the sampling period, although several other packets came close to this in amplitude and clarity. The waves in this packet were around 40 min in period and, integrated, represent vertical excursions of around 50 m. There was a level of asymmetry as upward velocities were stronger than the downward velocities despite the strong offshore flow deep in the water column.

An additional perspective of the internal wave can be gained from looking at pairs of current meter vector time series (Fig. 8). The example 12-h segment illustrates the complexity of the response. There are no periods of strong purely northward current indicating the effect of the southward-flowing mean shelf current. The depth-averaged current was not removed because, on occasion, the internal wave response generates velocities extending over the full depth of the water column. Removing this depth-averaged current would modify the frequency response relating to baroclinic processes. In some instances the upper water column response is very weak. At other times the response is strongly and symmetrically sheared in an east–west sense (e.g., Fig. 8b). Figure 8b shows the arrival of the



FIG. 7. A detail of a section of the vertical velocity from Fig. 6d. The region of signal dropout is shown as cross-hatched. Also a line, 1 h in duration, is placed at the 5-m depth for scale. The zero velocity contour is white.



FIG. 8. A section of current-meter results from moorings IW1 and IW2 from 27–28 Nov. (top) Temperature time series from the current meters. The two uppermost lines that are intertwined are IW1–20 m (black) and IW2–20 m (gray). The intermediate line is IW2–64 m (gray) and the lowest line is IW1–104 m (black). A time segment is marked along the bottom of this panel. (b) and (c) Velocity traces at the outer IW1 and inner IW2 stations for the selected 12-h period. The time series have been low-pass filtered with a frequency cutoff of 2 h. The black and gray traces in (b) and (c) represent the upper and lower current meters, respectively. The squares denote the start of the time series.

main internal wave sending the surface waters shoreward and the deeper water offshore—this then reverses. The response maintains the flow rate while the direction turns. The inner station IW2 (Fig. 8c) has a smaller amplitude response and is only comparable in the broadest sense. A significant difference is that the depression often lasts longer at the inshore station (i.e., the IW2 deep thermistor stays warmer for longer). This is a consequence of the record coming from a shallower depth.

The proximity of the three moorings to one another enable lag analysis to identify the wave phase speed and direction (Fig. 9). Direct lagged correlation analysis indicates that there were 55- and 6-min lags for IW1–IW2 and IW1–IW3, respectively. The IW1–IW3 lag was of marginal quality because the available thermistor was so near the surface (Table 1). Direct correlation of the raw signals included the effect of any other waves that may have confounded the signal. A correlation between low-pass-filtered temperature data from the midpoint in the water column suggested a peak at a lag of 40 min.

Because the shape of the waves appeared to be evolving significantly between the two moorings (Fig. 8), then the most reliable signal to consider is the lag in the appearance of the front itself (Fig. 9). This was realized by generating time series containing the temporal derivative of temperature for all times when the



FIG. 9. The cross-correlation coefficient as a function of lag time for temperature time series at IW1 and IW2. Results are from the raw (solid), low-passed (dashed), and gradient-resolved time series (solid with symbols).

low-pass-filtered temperature was rising by more than a threshold amount $(+1.44^{\circ}C \text{ day}^{-1})$. This generated a spiky but explicable result (Fig. 9). Only about 12 events matched the criteria and these occurred at four different lag periods. The direction of propagation, estimated from the phase lags between IW1-IW2 and IW1-IW3, was at around 256° from true north at a speed of 0.29 m s⁻¹. This direction (arrow P in Fig. 1c) is normal to the offshore bathymetry like the 160-m contour but is around 15° off the local 100-m bathymetric contour. Potentially the waves travel in different directions and/or at different speeds at different times. Resolution at IW3 was insufficient to look at variation in propagation angle. Consequently, assuming the direction of propagation was constant then variations in lag were due to speed changes. The speed estimate above was based on the 40-min lag, which demonstrably varied by as much as 50% so this velocity could increase to 0.6 m s^{-1} . As these fronts are generated approximately every 12 h, the estimated mean wavelength for the main internal wave was 13 km. Sharples et al. (2001) identified a wavelength of 15 km for the open shelf. This wave speed lies between the slow speeds (0.1 m s^{-1}) found by Pineda (1999) and the faster waves (0.9 m s^{-1}) in the Shelfbreak Primer study (Colosi et al. 2001).

c. Smaller-scale waves

Figure 4 indicates that there is a spectrum of higherfrequency waves. The small-scale internal wave frequency is around one-half of that associated with the mean buoyancy frequency. The large waves shown in Figs. 6 and 7 arrived just after the face of a large internal wave. This matched the open-shelf paradigm. However, possibly because of their extreme amplitude, these waves were atypical of those seen throughout the experiment. In general in this dataset the small-scale internal wave packets appeared throughout the data record. Wavelet analysis was employed to quantify the tide–trailing wave phase relationship.

The wavelet analysis applied to the average density signal at each mooring showed that the small-scale internal waves were mainly in the <60 min energy band (Fig. 10) and persisted at the observation point only for a few periods (e.g., Fig. 7). The phase of the 12-h wave, shown in the middle panel of Fig. 10, is well correlated with the sawtooth average density signal at this time. Later in the time series, when the wave was less sawtoothlike, the maximum average density (i.e., the isotherms are maximally elevated) was located at a phase of $+90^{\circ}$. The event of Fig. 7 was clearly located in Fig. 10, around time 330.85.

A picture of the phase-energy relationship was con-



FIG. 10. A 3-day-long segment of wavelet analysis from station IW1 showing (a) isotherms, with the deepest isotherm being the 16.25° with 0.75° C spacing; (b) average potential density, and phase of the 12-h complex wavelet; and (c) wavelet amplitude for the higher-frequency components of the spectrum.

structed by distributing wavelet energy (energy summed over periods between 20 and 60 min) as a function of the semidiurnal phase. In addition, the wavelet calculation was applied to data from the nearby open shelf (recorded at a different time) as an illustration of the paradigm. These results were collated (Fig. 11) such that, if the main internal wave were a sawtooth in density, the step would appear at a phase of 180. The IW1 data peaked at a little under 180°, the "end" of the wave, just prior to the arrival of the next warm front (when the average density would drop again). This increases at IW2 with the bulk of the high-frequency energy appearing a little earlier in the phase but also at a much greater amplitude. The amplitude of the openshelf data is arbitrary because the stratification was different. However, it does show that the high frequencyphase relationship was around 90° out of phase with the waves seen on the shelf. On the open shelf, away from the Poor Knight Islands, the main internal wavetrailing waves phase relationship is as observed in many other open-shelf studies. In these situations solitons grow out of, and trail, the frontal structure. Consequently the phase relationship seen at the island is an effect of the island and not a peculiarity of the regional internal tides.

d. Mixing

While the wave processes were clearly energetic, it remains to be seen if they drove mixing directly. Figure



FIG. 11. Phase analysis showing the arbitrary amplitude of waves with frequencies between 20 and 133 min as a function of the 12-h phase. Results are shown for IW1 (solid), IW2 (dotted), and from a record taken from the open shelf during a different experimental program (dashed). The upper panel is a schematic of the phase position of a sawtooth wave.

12 shows the Ri_g distribution for day 335 (1 December 2000). The results were impressive in that much of the water column was identified as unstable in terms of Ri_g < Ri_{gc} . The deeper regions during the cold water inflow were unstable, as was the very trough of the thermocline depression. Other regions of instability were found sporadically in the warm core of the main internal wave. The main regions of stability were found around the thermocline when it approached the sur-

face. The large velocity signal seen at the base of the wave did not generate very low Ri_g because at any one time the region of vertical gradient was so small and the stratification was significant in this region also. Of course the Ri_g parameterization does not adequately describe bottom friction effects, which are likely to be important during the high flow rates during the passage of the maximum depression, as it does not include the direct influence of the boundary shear. Vlasenko and Hutter (2002) show the relatively ordered, but highly energetic, turnover processes possible in such a situation.

Considering the SCAMP data from the same day as the Ri_g of Fig. 12, Fig. 13 shows binned energy dissipation rate ε_T from 18 profiles. There were generally high levels of ε_T near the surface, especially early on in the sampling. What was clearly captured was the vertical band of high ε_T right as the warm core passed through the sample station (time 335.48). This amounts to two profiles being much more energetic than the surrounding sampling. There was also a region of elevated ε_T on the upstroke of the isotherm (times 335.48–335.6). The high dissipation rates rarely exceed 10^{-7} m² s⁻³. This differs somewhat from the laboratory experiments of Helfrich (1992) in that we saw significant amounts of turbulence in the upper water column.

The apparently well resolved picture of the smallscale energetics on day 335 (Fig. 13), whereby the turbulence is concentrated in the core of the main internal wave, was not consistently seen on other days. Vertically averaging the ε_T from 20 m to the deepest extent of each profile builds up a time series of average dissipation rate for the four sample days, as a function of time relative to the wave trough (Fig. 14). The sampling on 1 and 2 December provided the expected picture of the evolution of the energy dissipation rate. However, on 27 November (bottom trace) there was no clear





FIG. 12. The \log_{10} of the gradient Richardson number for a portion of day 335 (1 Dec 2000), the 16.5°C isotherm is marked. The white regions are where $Ri_g > 10$.

FIG. 13. Dissipation rate contours for day 335 (1 Dec 2000) estimated from 18 temperature microstructure profiles. The 17° C isotherm from the profile data is shown.





FIG. 14. Time sequences of vertically averaged ε_T for 4 days. Time is relative to the time of maximum isotherm depression (i.e., 0.5 in Fig. 13). The vertical averaging extends from a depth of 20 m to the full profile depth to avoid surface mixing effects. Each day is vertically offset by two decades so that the horizontal line shows the relative 10^{-8} m² s⁻³ level.

trend and on 30 November the signal actually opposed that seen in the December sampling.

An alternative view of trailing wave activity was provided by the spectra of turbulent fluctuations recorded by the moored VECTOR velocimeter (Fig. 15). The $f^{-5/3}$ slope inertial dissipation region was quite clear in the upper spectra allowing an estimate of ε_I . However, the region of turbulence was inseparable from the noise floor in the lower spectrum. Peaks in the resulting time series of reliable ε_I estimates (Fig. 16) are at least an order of magnitude greater than the ε_T measurement at 30 m and around two orders of magnitude greater than the vertical averages seen in Fig. 14. The initial high average level of ε_I was caused by wind-related mixing. This eased off by time 331.5. The resolved ε_I time series was very sparse around day 335. The larger ε_1 values appear to be correlated with the warm core of the main internal wave, except around times 332.6-333.0 and 333.8-334.0 when the VECTOR was not in the wellmixed core region but it was still recording high ε_I = $10^{-5} \text{ m}^2 \text{ s}^{-3}$.

4. Discussion

a. Large-scale kinematics

The main tide-driven internal wave propagates at 0.3 m s⁻¹. This implies a minimum travel time from the

FIG. 15. Spectra from the VECTOR velocimeter. The upper spectrum is from time 331.019 (27 Nov) and the lower spectrum (offset by two decades) is from time 332.192 (28 Nov). Also shown are the $f^{-5/3}$ slope reference lines. A representation of the \log_{10} – space 95% confidence limit derived from a chi-square distribution is shown by error bars (Emery and Thomson 1997).

closest generation region of around 30 h. While velocities within the core of the internal wave were on average less than the phase speed, they did sometimes exceed this speed (e.g., Fig. 8). The implied wavelength is around 13 km, which is comparable to Fig. 2. The linear phase speed estimate is (1) $c_{pc} \approx 0.18 \text{ m s}^{-1}$, which implies a wavelength of around 8 km. The rigidlid, two-layer phase speed (2) $c_{p2} = 0.29 \text{ m s}^{-1}$ is closer to that observed. This should not be regarded as strong evidence of the wave behaving purely in a two-layer fashion as there is substantial shoaling at this point.

Scaling from shallow-water theory implies that the horizontal displacement is $\approx aL/(2\pi h_1)$ (e.g., Young 1999) where a, L, and h_1 are the wave amplitude, wavelength, and layer depth, respectively. Taking a = 40 m, L = 13 km, and $h_1 = 50$ m then the horizontal displacement is around 1.7 km. This matches the estimate of the particle excursion scale derived using the observed average measured speed within the core of the main internal wave of 0.1 (m s⁻¹) sustained for one-half of the tidal period, implying a displacement of 2.1 km. These estimates indicate a horizontal displacement somewhere in the range of 1.7–2.1 km, a distance smaller that the scale of the island group but not significantly so. The comparable scales of the wavelength of the main internal wave and the island group suggest that the internal wave evolution is influenced by the island.

However, the wave is clearly not gradually varying, indicating that the horizontal scale of the frontal region



FIG. 16. Time series of results the from the VECTOR velocimeter (day 331 is 27 Nov). These include (a) the velocity magnitude U, (b) the temperature at 30 m and average potential density, and (c) estimated dissipation rate ε_{i} . Missing points in the dissipation rate record did not pass a noise threshold.

in the direction of wave propagation is also important. The frontal regions (i.e., when the isotherms drop very rapidly as the core of the main internal wave passes by) were observed to pass a single point over a period of around 150 min. This implies a frontal length scale of around 3 km, less than one-quarter of the wavelength. This frontal length scale is comparable to the island size and smaller than much of the bathymetric variability seen in Fig. 1c. It also suggests that the moorings were within the front at the time they encountered the island. This might increase the observed wave amplitude relative to that seen on the open shelf at a rate greater than that associated with the background change in slope.

As the trough of the main internal wave passed the observation point, the lower part of the water column is forced to flow underneath and offshore. As noted when describing the ADCP data, this generated some of the largest flows in the system, reaching 0.4 m s⁻¹ on occasion (e.g., Fig. 8). To sustain this flux, part of the supply of deep fluid must have come from around the island. The significant along-island velocities of Fig. 6 and Fig. 8 support this although there was presumably biased by the coastal shelf current U_c .

We could not quantitatively distinguish any partial reflection of the main internal wave. This is in agreement with Helfrich (1992). However there was evidence of some large-scale flow reversals (e.g., Fig. 8). This highlights the difficulty in looking for consistent correlative evidence for these shoaling processes when the response is being modulated by variations in the barotropic tide itself as well as the shelf circulation (Fig. 9), which has a comparable speed.

Because of the large amplitude of the main internal wave, the trailing end of the wave is most likely in the form of a broken internal bore (Vlasenko and Hutter 2002) accompanied by a group of solitons. Because of their short wavelength, the solitons generally sustain high shear and enhanced likelihood of generating mixing. We could not detect, in any quantitative correlative fashion, the propagation of solitons traveling either away or toward the island. This is not surprising as they are superposed upon the coastal shelf current U_c and the internal and external tidal currents, which would likely have confounded any correlation. However, the number of solitons observed at moorings IW1 and IW2 were similar.

b. Mixing and diapycnal transport

Laboratory experiments (Helfrich 1992) suggest that enhanced vertical mixing should occur everywhere inshore of the breaking point. The time-averaged K_T and ε_T (Fig. 17) characterize the effect of the wave processes upon the water column energy and transport. These average values of K_T and ε_T appear low given the scale of the internal dynamics. Thus the experimentaveraged rate of energy dissipation $\varepsilon_A = 5 \times 10^{-8}$ m² s⁻³ was less than observed on the open shelf (1 × 10^{-7} m² s⁻³), in both Sharples et al. (2001) and Sandstrom and Oakey (1995). The relatively calm weather experienced during the present experiment meant that



FIG. 17. Averaged TGM-estimated dissipation rate ε_i and diffusion of thermal variance K_T (dashed line) for the four microstructure days.

the near-surface levels of turbulent energy dissipation rate were not elevated by wind-forced turbulence. Also, it is possible that the observation location was inshore of a region of much higher dissipation rate at some break point farther offshore. If this were the case, one might have expected the isotherm displacements to have been reduced, which was not observed. It also suggests further study to look at variability of average dissipation rate with distance from the island.

An explanation for the apparent variation in vertically averaged ε_T (Fig. 14) may be contained in the following argument. The expected signal (i.e., larger dissipation rate around the time of the thermocline depression) assumes that the mixing was due to the main internal wave. However, if the small-scale internal waves generated a comparable amount of mixing, then this will have confused the average signal. The smallscale internal waves appeared at a wide range of times and were undersampled by the ε_T profiles. Possibly, as the thermocline displacement was smaller on 1 and 2 December (days 331 and 332; Fig. 5), there might have been less trailing wave instability so that the dissipation rate was more closely related to the main internal wave. Opposing this was the observation that average dissipation rates were no higher for the larger internal wave events. Furthermore, the present situation has longer period small-scale internal waves than say the solitons observed by Sandstrom and Oakey (1995), suggesting weaker variations in horizontal velocity.

The average energy dissipation rate in the present situation was less than observed in open-shelf experiments. However, the relationship between K_T and ε_T described in (1) is influenced by variation in N to the extent that in our present relatively weakly stratified observations the measured K_T was comparable to Sharples et al. (2001) and somewhat larger than Sandstrom and Oakey (1995). It was around two orders of magnitude smaller than the bottom-friction-driven island wake diffusivities observed by Wolanski et al. (1984) in 18 m of water moving at 0.6 m s⁻¹. This suggests that the present mixing was dominated by the baroclinic processes and not vortex structures shed from the island.

While the average dissipation rate and diffusivity are relatively well ordered (Fig. 17), Fig. 18 shows a snapshot of the complexity of the processes underlying this. This TGM profile shows the largest "overturn" event seen in around 100 profiles, with one or two other events coming close in size, but perhaps older in that the edges of the regions were more diffuse. A 3-m section of the profile has been lifted vertically by 6 m (to a new depth 51–54 m). This profile was taken in the rising thermocline section on 27 November 2000. The



FIG. 18. An example of large overturn event (day 331, 27 Nov 2000, 1331 h) and the accompanying finescale fluorometer record measured on the same package.

 ε_T estimates in the lifted section of data were around 7 $\times 10^{-8}$ m² s⁻³ whereas the dissipation rate in the main thermocline region was 3 $\times 10^{-7}$ m² s⁻³. The ε_T contours (Fig. 13) showed this portion to be one of the two main mixing regions. Hence, rather than the dominant mixing region occurring on the front face of the wave, potentially like a shoaling surface wave, any instabilities that start there are swept away to evolve elsewhere. Consequently most of the mixing occurs in the trailing wake.

Such snapshots of large-scale instability provided by the TGM profiles make interesting comparison with laboratory and numerical experiments (Helfrich 1992; Michallet and Ivey 1999; Vlasenko and Hutter 2002) where large and energetic perturbations clearly follow the trough. Although consideration of such results illustrate that it might be difficult to define the "front of the wave," the numerical results clearly illustrate the possibility of large-scale overturns as observed in the present field results.

Events like that seen in Fig. 18 may have a role in transport of biota and associated impact on populations. This provides perhaps a less subtle mechanism for the formation of thin layers than that identified by Wang et al. (2001) in that this region will collapse as it mixes to its new depth of neutral buoyancy. The fluorometer, which had a resolution of a few centimeters, clearly resolved the different properties inside and outside the patch. Presumably this structure will persist for some time. The subsequent profile, recorded 34 minutes later, showed no indication of the overturn region. However, this was possibly due to the horizontal advection processes rather than mixing of the overturn. A more direct observation of the influence of the mixing in the internal wave core was seen in the vertical ADP data (Fig. 6d, time = 331.4) when motion within the core was sufficient to return scatterers to the near surface. The VECTOR (Fig. 16a) average velocity (including the vertical component) shows a magnitude of around 0.15 m s⁻¹ at this time. This was apparently sufficient to overcome the swimming speed of the scattering zooplankton.

5. Conclusions

Observations of flow and mixing associated with a large-amplitude internal wave recorded relatively close to an island are described. (i) The propagation of the main internal wave was identified and found to be highly variable in nature, indicating a sensitivity not only to factors at the source of the internal tide but also to effects like the barotropic tide and the shelf circulation. It is clear that a better spatial appreciation of the evolution is required. (ii) The high-frequency internal wave activity was on average located toward the end of the main wave in phase space, somewhat different to open-shelf waves. There was a measurable increase in trailing wave activity closer to shore. The highfrequency internal wave propagation was affected by other advective processes. (iii) The measurements do not show any readily measurable reflected waves (either the main internal wave or smaller-scale processes). (iv) Peaks in the time series measurements of energy dissipation rate resolved using an acoustic velocimeter $(1 \times 10^{-5} \text{ m}^2 \text{ s}^{-3})$ were comparable with observations on the open shelf in other studies. Profiler results suggested lower levels of average energy dissipation rate, around 5 \times 10⁻⁸ m² s⁻³. Our data suggest that shear instability (Fig. 12) at this time of year will only be sporadic and not during the passage of the actual trough itself. However, as we see substantial mixing (Fig. 13) as the trough passes by, this suggests convective instability plays a significant role. (v) Temperature diffusivity was around $K_T = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, comparable to other studies. We were able to demonstrate the presence and scale of large discrete mixing events.

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