MIXING ALONG DEEP BOUNDARIES ON THE KAENA RIDGE

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ABSTRACT

Moorings deployed on the south (Aug-Nov 2002) and north (Nov 2002 - June 2003) flanks of the Kaena Ridge, Hawaii are used to examine current and temperature variability within 200 m of the steeply sloping bottom near the 2400 m isobath. On the south flank, horizontal currents and vertical displacements are dominated by the semidiurnal internal tide over the depths sampled. On the north flank, the semidiurnal tide is less energetic than on the south, with a different vertical structure as tidal amplitudes decrease toward the boundary. Near the boundary, near-inertial to diurnal oscillations associated with winter wind forcing north of the islands are present which were comparatively weak on the south flank. At both sites, strong temperature inversions are detected with vertical scales of ~ 100 m.

A Thorpe scale analysis of the overturns yields a time-averaged dissipation near the bottom of $1.2 \times 10^{-8} W/kg^{-1}$, 10 to 100 times higher than at similar depths in the ocean interior 50 km from the ridge at the south mooring. The dissipation at the north mooring is estimated to be an order of magnitude smaller at $1.8 \times 10^{-9} W/kg^{-1}$. At the south mooring, dissipation events much larger than the overall mean (up to $10^{-6} W/kg^{-1}$) occur predominantly during two phases of the semidiurnal tide: 1) during peak downslope flows when the tidally modulated stratification is minimum ($N = 5 \times 10^{-4} s^{-1}$), and 2) when the flow reverses from down to upslope flow as the tidally modulated stratification is ordinarily increasing ($N = 10^{-3} s^{-1}$). Shear instabilities, particularly due to tidal and near-inertial strain enhancements, appear to trigger downslope flow mixing at the south mooring. Convective instabilities are proposed as the cause for flow reversal mixing at the south mooring, owing to the oblique propagation of an internal tidal beam down the slope.

At the north mooring, dissipation occurs predominantly during periods of high strain at the maximum upward displacement of the semidiurnal tide. Overturns are particularly prevalent when the tidal strain is enhanced by near-inertial wave driven downwelling near the slope. The intermittency of near-inertial waves at the north mooring limits the predictability of mixing events.
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Chapter 1

Introduction

In this study, we investigate signatures of tide-induced mixing along the slopes of the Hawaiian Ridge. The observations represent some of the first long-term (3 and 8 months) measurements of current and temperature variability close to a steeply sloping (1/4 and 1/12) boundary that are suitable both for identifying specific dynamic processes (tides, inertial wave events) as well as signatures of mixing that are linked to these processes. In this section, we first review the overall context of "deep" mixing, that is mixing in the abyssal ocean below the main thermocline (1.1). The two primary energy sources for deep mixing, the tides (1.2) and wind-driven currents and waves (1.3) are briefly noted. Internal waves, both at tidal and near-inertial frequencies, are dominant in the observations and we note the importance of critical slopes in regard to internal wave reflections at boundaries (1.4). The way in which boundary mixing can influence the ocean interior is discussed, particularly the efficiency of sloping boundaries for exporting mixed fluid (1.5). This work was conducted as part of the Hawaii Ocean Mixing Experiment (HOME), and the relevant results of HOME colleagues are presented in 1.6, followed by an outline of the remainder of the dissertation.

1.1 Preamble : The Ocean Mixing Problem

The large-scale, meridional, overturning circulation in the ocean is driven by surface heat gain by the ocean in the tropics, and heat loss at high latitudes. Wind-driven currents support a net poleward advection of heat at the surface, while the formation of cold, dense waters at high latitudes feeds a thermohaline circulation that spreads throughout the abyssal ocean, eventually upwelling back to the surface, closing the circulation loop. To be in equilibrium with this circulation, the observed distribution of heat throughout the ocean requires mixing to occur well below the thermocline. Without abyssal mixing, the ocean basins would fill up with uniformly cold water underlying a thin layer of warm, sun-heated water.
Although there is great uncertainty in assessing the global heat budget, simple models can give intuitive insight into the complicated ocean mixing problem. In a landmark study, Walter Munk (Munk, 1966) proposed a steady one-dimensional balance between the upward advection and the downward diffusion of density via turbulent mixing. This simple model takes the form

$$ w \frac{\partial \rho}{\partial z} = \kappa \frac{\partial^2 \rho}{\partial z^2}, $$

where $w$ is the vertical velocity, $\rho$ is the density, and $\kappa$ is the vertical diffusivity. When this balance was applied to the observed stratification, and with suitable assumptions regarding $w$, a value of $\kappa = 10^{-4} m^2 s^{-1}$ was obtained. This corresponds to a supply of energy to the abyssal ocean at a rate of $2 \times 10^{12} W$, or $2 TW$, which is necessary to maintain the observed stratification.

Subsequent in situ measurements of the rate of mixing in the open ocean (Gregg, 1989; Ledwell et al., 1993) correspond to diffusivity values an order of magnitude weaker ($10^{-5} m^2 s^{-1}$) than predicted by Munk. Mixing rates are enhanced near rough topography such as the Mid-Atlantic Ridge (Polzin et al., 1997; Ledwell et al., 2000), seamounts (Lueck and Mudge, 1997), escarpments (Althaus et al., 2003), and other topographic features (Kunze and Sanford, 1996). The value of $\kappa$ was revisited more than 30 years later by Munk and Wunsch (1998), who incorporated in the new solution the large horizontal variations of mixing between the relatively quiet open ocean and the energetic dissipation occurring along deep, sloping, lateral oceanic boundaries as postulated by Armi (1978). Yet, the new solution still led to the same "average" value of $10^{-4} m^2 s^{-1}$. This canonical value has been so widely accepted and cited, that in October 2004 in Victoria, British Columbia, an assembly of Ocean Mixing specialists, the majority of them listed in the reference section of this dissertation, agreed to define the value of $1 \text{Munk}$ as equal to $10^{-4} m^2 s^{-1}$.

While there is still debate as to whether there is "missing mixing" in the ocean (Webb and Suginoohara, 2001), the primary energy source for deep ocean mixing is likely to be the tides and surface wind forcing (Munk and Wunsch, 1998; Wunsch and Ferrari, 2004). The overall tidal contribution has been accurately specified from astronomical measurements at $3.5 TW$ (see 1.2). The wind-driven contribution is still under investigation (see 1.3). Understanding and quantifying the pathways and processes that dictate the distribution of turbulent mixing in the ocean has kept oceanographers busy for the past 40 years, and it is a necessary step toward better predictions of the ocean circulation under climate change.

### 1.2 Tides

Ocean tides on Earth are due to the gravitational pull of the Moon, with a smaller contribution of the Sun. In 1695, Halley concluded that the Moon’s orbital velocity around the Earth was
increasing with time. To conserve the total angular momentum of the Earth-Moon system, Halley concluded that the distance between the Moon and the Earth was decreasing. The most precise measure of this distance comes from laser ranging observations made available after the installation of reflectors on the lunar surface by the Apollo mission in 1969 (Dickey et al., 1994). The Earth-Moon distance is in fact increasing, because the angular momentum of the system decreases due to tidal friction. Frictional dissipation occurs at a rate of $2.4 \, TW$ for the lunar semidiurnal constituent $M_2$ alone, and $3.7 \, TW$ for all constituents. Most of this energy loss (95%) is associated with ocean tides. It was first thought that nearly all the energy was lost by friction in the shallow seas where tidal currents and associated bottom drag are strong (Taylor, 1919). Using satellite altimetry and an inverse model of the tides, Egbert (1997) and later Egbert and Ray (2001) estimated that only $1.8 \, TW$ of $M_2$ energy is dissipated in shallow seas (out of $2.4 \, TW$), leaving $0.6 \, TW$ of energy to be dissipated in the deep ocean. Several numerical modeling experiments have since shown that most of the energy lost by the surface tide in the open ocean occurs in regions of rough topography such as the Hawaiian Ridge or the Mid-Atlantic Ridge (Simmons et al., 2004; Merrifield and Holloway, 2002; Niwa and Hibiya, 2001). At these sites, the surface or barotropic tide is converted into internal or baroclinic tides. The generated internal tide is either dissipated locally, or radiated away and subsequently dissipated. St Laurent and Garrett (2002) estimate that less than 30% of the internal tide energy is generated at small enough vertical scales suitable for local dissipation, while the rest is in the form of long-range propagating, low mode internal tides. At the Hawaiian Ridge in particular, Merrifield and Holloway (2002) found that $9.7 \times 10^9 \, W$ or $9.7 \, GW$ of tidal energy is radiated away from the ridge as internal tides while Niwa and Hibiya (2001) and Simmons et al. (2004) found an $M_2$ energy conversion from barotropic to baroclinic of $15 \, GW$ and $30 \, GW$ respectively. Egbert and Ray (2000) estimated from altimetry that $20 \, GW$ of barotropic energy is lost over the same area.

1.3 Wind

If almost $1 \, TW$ of power is contributed to mixing in the abyss by the tides (Egbert and Ray, 2001), the rest of the estimated $2 \, TW$ required to maintain the abyssal stratification may be provided indirectly by the wind. The wind can contribute to the abyssal mixing through the generation of large-scale and mesoscale currents or the generation of internal waves. The work done by the wind on the ocean circulation is estimated to be $1 \, TW$ (Wunsch, 1998). The ensuing large-scale currents create drag in areas of rough topography and in deep passages (Ferron et al., 1998) or over
sills (Lukas et al., 2001). This drag creates turbulent mixing and contributes an estimated 0.2 TW of power available for the deep ocean mixing (Wunsch and Ferrari, 2004). The wind also contributes to deep ocean mixing via the generation of internal waves, and their subsequent dissipation (Wunsch, 1998; Alford, 2001; Watanabe and Hibiya, 2002; Alford, 2003). The wind excites inertial motions in the surface mixed layer, and these motions pump internal waves in the stratified fluid below the mixed layer (D’Asaro, 1985, 1995; Nagasawa et al., 2000). These internal waves are generated at a frequency close to the inertial or Coriolis frequency and are referred to as near-inertial internal waves (NIWs). NIWs have been hypothesized (Garrett, 2001) and observed (Alford, 2003) to propagate predominantly equatorward away from their region of generation. Nonlinear interactions within the internal wave frequency band subsequently transfer energy to smaller and smaller scales, ultimately leading to dissipation as heat on molecular scales (Garrett and Munk, 1972). Alford (2003) calculated that 0.5 TW of power from the wind is injected in the internal wave field, which ultimately contributes to the deep ocean mixing budget, either in the open ocean or near boundaries.

### 1.4 Internal Wave Dynamics at Sloping Boundaries

The stably stratified interior of the ocean is never still due to the ubiquitous presence of internal inertia-gravity waves, referred to here simply as internal waves. These waves exist in a stratified fluid where motions are subject to both gravity and Coriolis forces. The dispersion relation for freely propagating internal waves is specified for frequencies, $\omega$, between the Coriolis frequency $f$ and the Brunt-Vaisala frequency $N$. The spectrum of these internal waves is similar throughout most of the ocean and is generally well described by the Garrett and Munk (1979) spectrum. Internal waves propagate at an angle, $\theta$, relative to the vertical, that depends on the wave frequency, the Brunt-Vaisala frequency and the Coriolis frequency such that $\omega^2 = N^2 \cos^2 \theta + f^2 \sin^2 \theta$. A peculiar aspect of internal wave dynamics arises upon reflection at a sloping boundary. Unlike light waves which undergo specular reflection at a smooth boundary, internal waves reflect so that the incident and reflected angle are maintained relative to the vertical, rather than to the normal angle of the boundary. In the extreme case where the topography and the wave propagation have the same slope, critical reflection occurs and linear theory predicts that the reflected wave has an infinite amplitude, an infinitesimal wavelength and a near-zero group velocity (Phillips, 1977).

In general, the reflected wave is steeper and produces greater shear than the incident wave, particularly around the critical frequency (Eriksen, 1982, 1985). Eriksen (1982) observes an enhancement in horizontal kinetic energy around the critical frequency within 100 m of the bottom,

\[ \text{...} \]
at a variety of island, seamount and continental slope sites, and which was absent 1000 m above the bottom. These enhanced levels of kinetic energy near the bottom can lead to enhanced turbulent dissipation and mixing.

The mixing associated with critical reflection on a non-planar slope can depend on the slope configuration above and below the critical region. Analytical solutions have shown that critical reflection produces more mixing in the case of a convex slope (Gilbert and Garrett, 1989; Müller and Liu, 2000) than in the case of a concave slope. In another study, Legg and Adcroft (2003) used a non-hydrostatic general circulation model to study the critical reflection of internal waves and concluded that the enhanced mixing associated with critical reflection did not depend on the curvature of the slope, but only on the critical character of the slope. To our knowledge, modeling studies have not been made of internal wave reflections on complex, finite amplitude, 3D bathymetry, such as at our study site.

Internal waves can also approach a sloping boundary obliquely, when the vertical plane containing the wave group velocity vector does not intersect the boundary along the line of greatest slope. Upon such a reflection, density fronts can form and propagate up the slope (Thorpe, 1999). These fronts are most likely to form when the incident wave is near critical and its obliqueness is limited to within 30° of the normal incidence.

Most of the studies on internal wave reflection consider internal waves coming from deep water and reflecting off a slope. Internal waves can also propagate down along a near-critical slope from their generation site, especially along continental shelves (Pingree and New, 1989). In the Bay of Biscay, Gemmrich and van Haren (2001) observed rapid temperature drops near the bottom occurring at semidiurnal periods. Gemmrich and van Haren (2001) attributed these temperature drops to fronts generated by the flow field of an obliquely downward propagating internal waves. We will consider this process further in section 6.1.2

1.5 Boundary Mixing

Munk (1966) noted that intense mixing near boundaries, along with an effective communication to the ocean interior, could be vital to the observed stratification. He envisaged internal wave breaking as an important source of mechanical energy for near-boundary mixing.

Armi (1978) observed 50 – 100 m thick well-mixed layers near sloping boundaries, attributed to turbulence in the bottom boundary layer. These layers can then intrude along constant density surfaces and spread into the interior ocean. The combined effect of these two processes
can have an equivalent effect to the vertical eddy diffusivity in the open ocean. As noted by Munk and Wunsch (1998), the "average" mixing in the ocean could be accounted for by energetic mixing at sloping boundaries, followed by spreading of the mixed fluid (but not the turbulence) into the quieter ocean interior.

Such intrusions were later observed as an Intermediate Nepheloid Layer (INL) (McPhee-Shaw et al., 2004) in which high concentrations of suspended sediments are advected laterally into deep water from nearby continental shelves and slopes. INLs have also been observed at semidiurnal periods in regions of near-critical slope (Moum et al., 2002), suggesting that the critical reflection or generation of internal \( M_2 \) tides provides mechanical energy for boundary mixing. In a laboratory experiment, McPhee and Kunze (2002) observe intrusions from a sloping bottom where internal wave reflection provides mechanical energy for the generation of turbulence. They relate the intrusion’s growth to the incidence angle of the internal wave, relative to the slope angle, and observe the strongest intrusions during critical reflection conditions.

To illustrate this process, we reproduced the laboratory experiment of McPhee and Kunze (2002) using a rectangular tank filled with salt-stratified water, a sloping boundary and an oscillating rough metal plate along this boundary to create turbulence (Figure 1.1). Fluorescine dye was slowly injected at the boundary to observe the intrusions into the tank interior. We first conducted the experiment with two layers of different density. Almost immediately after the turbulence started, an intrusion was formed that spread laterally at the interface between the two layers (Figure 1.2). We then extended the experiment to a more realistic ocean scenario, with a continuously stratified tank. After a few minutes, the same mechanism produced several evenly spaced intrusions spreading into the tank interior (Figure 1.1). The importance of mixing along sloping boundaries, as opposed to over a flat bottom, is that the mixed water can be exported into the ocean interior along constant density surfaces, thus enhancing mixing efficiency (Munk and Wunsch, 1998).

### 1.6 HOME Experiment

It is now well established from various in situ and remote observations that energetic semidiurnal internal tides originate from the Hawaiian Ridge (Ray and Mitchum, 1996; Dushaw et al., 1995). The Hawaii Ocean Mixing Experiment (HOME) was designed to investigate the details of the tidal mixing and dissipation along the Hawaiian chain (Rudnick et al., 2003). As part of the HOME modeling effort, the Princeton Ocean Model (POM) has been used to simulate barotropic to baroclinic tidal energy conversion along Kaena Ridge in the Kauai Channel (figure 1.3). The
simulations show internal tide energy originating on each side of the ridge near the nearly symmetric crest (Merrifield and Holloway, 2002). The energy propagates along internal wave characteristics; one beam propagates up and away from the ridge, a second propagates up and over the ridge, and a third propagates downward along the ridge flanks where there is either a near- or super-critical slope below the generation site (Figure 2.1). Observations from different instruments and platforms made during the HOME nearfield phase confirm this beam structure (Rainville and Pinkel, 2005; Martin et al., 2005; Nash et al., 2005).

Mechanical energy associated with this down-going beam as it propagates along the ridge flanks is available for turbulent mixing at the slope. These down-going beams along the north and south flanks of the ridge are of central importance to the present study. The numerical model (Merrifield and Holloway, 2002) indicates an $M_2$ baroclinic energy flux of $\sim 10\,GW$ radiating away from the entire Hawaiian Ridge. In their model, they use a bottom drag friction parametrization to account for the loss of energy of the combined barotropic and baroclinic tide at the bottom boundary. This parametrization accounts only for the turbulent dissipation created by shear stress at the boundary. They estimated that only 0.6 $GW$ of $M_2$ tidal energy is lost by frictional dissipation at the boundary, small compared to the $\sim 10\,GW$ of internal tidal energy radiated away.

Klymak et al. (2005), using direct dissipation measurements from 4 different instruments, estimate $3 \pm 1.5 \,GW$ of turbulent dissipation along the entire Hawaiian ridge. This estimate is based on rather sparse data collection around the Hawaiian Ridge system, and the authors acknowledge the fact that geographical variations of mixing intensity may bias their estimates one way or another.

To complement the sparse temporal resolution of HOME microstructure measurements near the bottom, deep moorings were deployed near the ridge top and over the flanks of Kaena Ridge, between the islands of Oahu and Kauai. These moorings were designed to sample temperature and current over the first hundreds of meters above the bottom. Indirect estimates of dissipation can be obtained through detection and analysis of turbulent overturns (Levine and Boyd, 2005).

Levine and Boyd (2005) deployed such a mooring near the ridge top ($1450\,m$) on the northern flank (Big Boy mooring, figure 1.3) and found significant overturns ($\sim 150\,m$) within $300\,m$ from the bottom that are linked to semidiurnal tidal phase and amplitude. Levine and Boyd (2005) describe a scenario where the internal tide first strains the mean density field, leading to regions of low $N$ that subsequently overturn. The phase of the tide when overturns occur varies with depth. They also found less frequent overturns occurring when the stratification is high, similar to the type of mixing events discussed in chapter 6 of this dissertation. Levine and Boyd (2005) found average dissipation levels of $\sim 10^{-8} \,Wkg^{-1}$ in the $150\,m$ thick layer for the entire experiment.
Rainville and Pinkel (2005) and Carter and Gregg (2005) observed high levels of near-diurnal internal wave energy at the nearby summit of Kaena Ridge. This diurnal energy flux was observed to vary more with the astronomical semidiurnal forcing than with the diurnal forcing. Parametric Subharmonic Instability (PSI) was proposed as a mechanism to transfer energy from the semidiurnal $M_2$ frequency to the near-diurnal $M_2/2$ frequency, but the exact mechanism describing this transfer still remains unclear (Rainville and Pinkel, 2005).

1.7 Thesis Outline

In this dissertation we examine temperature and current data from two deep ($2425m$) moorings on the southern (DS) and northern (DN) flanks of Kaena Ridge (Figure 1.3). These moorings are located at depths well below the main thermocline.

We use estimates of dissipation based on a Thorpe scale analysis to show that mixing associated with the semidiurnal internal tide is vigorous at the DS mooring (dissipation events $\geq 10^{-7} W kg^{-1}$). The time series provide an unique view of deep mixing driven by a tidal current. At the DN mooring, we observe a less energetic tidal environment than at DS, and enhanced variability in the inertial to diurnal band. The overall inferred mixing at the north mooring is one order of magnitude smaller than at the south mooring.

The dissertation is organized as follows. The mooring locations and instrumentation are first described in chapter 2. The observed flow characteristics and the tidal variability are discussed in section 3. The character and possible generation of near-inertial flow events are considered in chapter 4. The method for estimating turbulent overturns and the associated dissipation, along with an estimation of the uncertainties of the method are described in section 5. In section 6 we identify two types of mixing events at the south mooring, related to different phases of the semidiurnal tide, and investigate their relationship with observed shear and strain. We attribute the mixing events occurring at the north mooring to enhanced strain in the context of the co-occurrence of near-inertial waves near the bottom and the semidiurnal internal tidal beam higher in the water column away from the boundary. A discussion and summary follow in chapter 7.
Figure 1.1: Intrusion experimental setup, for a continuously stratified fluid, showing the oscillatory grid that generated turbulence at the sloping boundary. Several evenly spaced intrusions, marked by fluorescine dye, are visible on the right of the picture.

Figure 1.2: Snapshots of the intrusion experiment for a two-layer fluid, taken a few seconds after the onset of grid oscillation and generation of turbulence at the sloping boundary. The blue line scale is 10 cm in length.
Figure 1.3: Bathymetric map of the Kaena Ridge showing the locations of the DS and DN moorings (this study), the Big Boy mooring (Levine and Boyd, 2005), and FLIP (Rainville and Pinkel, 2005). The scale is in meters.
Chapter 2

Data and Experimental Setting

2.1 Moorings Description

Two moorings were deployed during the nearfield phase of the HOME experiment, at the same depth (∼ 2425 m), on the north and south sides of the Kaena Ridge, 40 km apart (Figure 1.3). Mooring Deep South (DS) was deployed on the south side of the ridge (158.65°W, 21.58°N), from August to November 2002. Mooring Deep North (DN) was deployed on the north side of the ridge (158.4°W, 21.87°N) between November 2002 and June 2003. The locations of the moorings were chosen in an attempt to sample downgoing semidiurnal tidal beams identified in the model runs of Merrifield and Holloway (2002) and Merrifield (2005) (Figure 2.1). It was hypothesized that the internal tide would be the primary energy source for near-boundary mixing at this depth.

The mooring designs were similar for DS and DN (Figure 2.2). Temperature was recorded every 3 and 5 minutes at mooring DS and DN, respectively, by Seabird Electronics temperature sensors (SBE 39). Sensor spacing varied between 16 and 24 m on mooring DS, while sensors were spaced every 8 m between 28 and 68 m above the bottom (mab) and every 24 to 32 m between 68 and 220 mab on mooring DN. According to the manufacturer, the SBE 39 sensors have a very small measurement error of 2 millidegrees, and a slow drift, typically less than 0.2 millidegree per month. One Seabird Electronics temperature and conductivity sensor (SBE 37) was located at 100 mab and 44 mab at mooring DS and DN, respectively.

Each mooring included an upward-looking RDI 300kHz ADCP, at 22 mab with a nominal range of approximately 100 m. To reduce the shock on the instruments during deployment, a nylon line was used between the anchor and the acoustic releases, and hence we were not able to sample closer to the bottom.
The ADCP data were 8 minute averages with a vertical resolution of 4 m. Ancillary data from the ADCP (i.e. amplitude and correlation of the acoustic return signal) indicate that the acoustic return from some depth bins was contaminated by the side lobe reflection off the mooring elements, which biases the current speed towards zero. These bins were not used in the analysis, although the data are included in some figures for reference (Figure 3.6 for example). Reliable current estimates (standard deviation of $\sim 0.02 \text{ ms}^{-2}$ for 40 pings averaged every 8 minutes) were obtained out to a range of only 40 mab to 60 mab, presumably due to a low concentration of acoustic scatterers at this depth.

One pressure sensor was included at the top of each mooring to estimate the effect of the mooring layover. We found that these lateral and vertical motions were negligible at DS, confirming the mooring line was near-vertical at all times. We found that the measured pressure does not differ from the predicted tidal pressure by more than 0.2 dbar $\sim 0.2$ m. This was consistent with the measurement of the ADCP tilt and roll, which never exceeded 5°. At DN, the pressure sensor failed, but considering the similarity of mooring designs and horizontal current speeds, we believe the mooring line at DN also remained near vertical at all times.

2.2 Topography and Stratification

The Kaena Ridge topography has been surveyed with multibeam sonars from different agencies and compiled into a bathymetric map with a horizontal resolution of 150 m (Figure 1.3). DS was located in a steep region on the south flank of the ridge, in a slight depression between two ridges running down the slope and separated by $\sim 10$ km (Figure 1.3). Averaged over a 1.5 km square, the topographic slope is 1/4.5 at DS, and oriented at a 35° angle relative to true north (NNE). Mooring DN was deployed on the north flank of the ridge, on a relatively flatter plateau, where the topographic slope is 1/11.6, and oriented 200° relative to true north (SSW).

At mooring DS, the overall vertical- and time-averaged buoyancy frequency is $\bar{N} = 1.02 \times 10^{-3}$ s$^{-1}$ or 14.5 cpd, and the Coriolis frequency is $f = 5.36 \times 10^{-5}$ s$^{-1}$ or 0.74 cpd. This equates to a critical frequency $\omega_c = \sqrt{f^2 \sin^2 \theta + \bar{N}^2 \cos^2 \theta}$ of 3.34 cpd averaged over a 1.5 km square around the mooring. $\omega_c$ varies depending on the area used to average the topographic slope (4.04 cpd for a 0.6 km square, 3.68 cpd for a 0.9 km square). In comparison, at mooring DN, $\bar{N} = 1.19 \times 10^{-3}$ s$^{-1}$ or 16.43 cpd, and the Coriolis frequency is $f = 5.43 \times 10^{-5}$ s$^{-1}$ or 0.75 cpd, leading to $\omega_c = 1.79$ cpd.
The difference in $N$ between the 2 moorings ($\sim 1.7 \times 10^{-4} \, s^{-1}$) is significantly larger than the 95% confidence interval of $1 \times 10^{-6}$ and $0.8 \times 10^{-6} \, s^{-1}$ for DS and DN, respectively. Changes in stratification related to mesoscale eddy activity along the Hawaiian Ridge likely would not affect the 3 and 8 month averages at each moorings.

Stratification averages were calculated for each month over the 16 years of CTD data at station ALOHA. No regular annual variations of stratification were observed at the depths of interest (2000 – 3000 m). We therefore suggest that the difference in $N$ between the 2 moorings is unlikely to be related to the different deployment times (August to November 2002 for DS and November 2002 to June 2003 for DN).

Time-average vertical temperature profiles along 160$^\circ$W, obtained as part of the World Ocean Circulation Experiment (WOCE), do not show a meridional change large enough on either side of the ridge, although the data are too sparse for a definitive comparison.

The implications of the different stratification at the two moorings will be discussed in section 7 in regards to turbulent dissipation and mixing. Also, at all depths, absolute temperatures at the moorings are within 0.03°C of the climatological absolute temperature values (16 years of repeat CTD profiles) at nearby station ALOHA (Figure 2.3). The Temperature-Salinity (TS) diagrams for the two moorings indicate the similarity of water mass properties on either side of the Ridge (Figure 2.4). The more elongated cloud of TS points at mooring DS compared to DN is related to the higher vertical excursions measured on the south flank. The narrow distribution of $T$ and $S$ allowed us to infer the salinity at depths where only temperature was measured. We used the measured temperature and the inferred salinity to calculate potential density. In chapter 6, we used these potential density profiles to detect statically unstable patches and to estimate turbulent dissipation.
Figure 2.1: Cross section of the ridge and $M_2$ modeled baroclinic energy fluxes from Merrifield (2005)

Figure 2.2: Position of the sensors on the DS (left) and DN (right) moorings. T indicates temperature measurement, C salinity measurement, and U current measurement.
Figure 2.3: Time-averaged vertical profiles of absolute temperature at DS (red) and DN (blue), corrected (dashed lines), and uncorrected (thick lines), (see chapter 6 for the details of the correction). Instrument locations are depicted by a ‘*’. The black line shows the climatological temperature profile at Station Aloha, located ~ 100 km to the NNE of mooring DN.
Figure 2.4: TS diagrams at mooring DS (left) and mooring DN (right). The data are organized in bins of 0.01°C and 0.001 PSU and the contours indicate the percentage of the data in each bins.
Chapter 3

Current and Temperature Observations

3.1 Description of Variability

Time series of current and temperature sampled near the bottom at DS are dominated by fluctuations at the semidiurnal frequency (Figure 3.1). Temperature changes $\geq 0.3^\circ C$ occur during spring tides, and $\leq 0.1^\circ C$ during neap tides. The horizontal tidal current is stronger in the across-slope direction than the alongslope, with peak speeds of $0.2 ms^{-1}$ during springs and $< 0.1 ms^{-1}$ during neaps. The alongslope flow is typically $< 0.1 ms^{-1}$, although the ratio of across-slope to alongslope tidal flow varies over the course of the experiment, due to changes in the baroclinic tidal component. The general variability at DN is similar to DS (Figure 3.2).

The standard deviation of the current velocity is similar at the two moorings (0.04 – 0.08 $ms^{-1}$), with an increase in amplitude away from the boundary (Figure 3.3). Over the entire experiment the overall ratio of across to along slope flow is 1.2 at DS (Figure 3.3). The standard deviations of the temperature oscillations and the corresponding vertical displacements are larger at DS than DN (120 $m$ compared to 60 $m$), consistent with the advection of the background stratification by similar currents up and down a steeper slope at DS than at DN. Besides the strength of the vertical displacements, the main difference between the two moorings is the partition of the variability between the tidal (detailed in the next section) and non-tidal contribution, namely the near-inertial to diurnal variability described in chapter 4. At DS, most of the variability can be explained by the tides, while a significant fraction of the variability at DN is related to near-inertial motions.

At both moorings, the long-term averaged current is weak, $< 0.01 ms^{-1}$ at all depths (Figure 3.3). At the lowest ADCP bin at DS, the long-term averaged current is anomalously large ($\sim 0.03 ms^{-1}$) and directed downslope. We checked the amplitude and correlation of the acoustic
return signal for this bin, and could not dismiss it based on sensor noise or mooring contamination. Nevertheless, with only one bin showing the anomalous mean current, we do not consider it further.

3.2 Tidal Variability

3.2.1 Mooring DS

Spectra of horizontal velocity, shear, and buoyancy frequency all show peak energy at the semidiurnal frequency band (Figure 3.4). Spectral peaks at the diurnal and inertial frequency bands are noticeably smaller than the semidiurnal peak. The semidiurnal band accounts for 60% of the total current variance at each depth, with the $M_2$ constituent alone accounting for 50% of the variance. Similar ratios are found for vertical shear and temperature. The temperature spectra are similar at all the measured depths (Figure 3.5), with slightly less energy in the semidiurnal bands as the distance from the bottom decreases (36% decrease between 220 $m_{ab}$ and 27 $m_{ab}$). Around the estimated critical frequency of $\sim 3.3$ cpd, the temperature oscillations are slightly more pronounced near the bottom, but not dramatically so as has been found in other observations of internal wave reflection (e.g. Eriksen (1982)).

A standard tidal analysis (Foreman, 1978) confirms that the semidiurnal constituents are approximately an order of magnitude larger than the diurnal (Table 3.1). In general, the $M_2$ and weaker $S_2$ constituents create a fortnightly spring-neap cycle with up to a factor of 4 difference in current amplitude. The current ellipses for the dominant $M_2$ tidal constituent are directed almost across the local isobaths (across-ridge). The angle between the topographic gradient and the semimajor axis of the $M_2$ current ellipses varies between $15^\circ$ and $30^\circ$ over the depth range examined (Figure 3.6). The semimajor axis of the $M_2$ current is $\sim 0.07 m s^{-1}$ with similar ellipse structure between 45 and 65 $m_{ab}$ (Figure 3.6). Although the current amplitudes of the other depth bins are suspect (see section 2.1), we include them in Figure 3.6 to show that the orientation and eccentricity of the current ellipses are similar over the depth range. The ellipses also tilt in the across-ridge/depth plane such that the across-ridge flow is parallel to the slope (not shown). Greenwich phases for the current ellipses range from $40^\circ$ to $50^\circ$.

The $M_2$ barotropic current, predicted either by the POM simulations (Merrifield and Holloway, 2002) or the TPXO model (Egbert, 1997), is not in phase with the measured semidiurnal current (Figure 3.6). When the semidiurnal surface elevation is high over the ridge, barotropic currents are directed southwestward (Merrifield and Holloway, 2002). At this phase of the tide, baroclinic currents along the south flank nearly oppose the barotropic current in model simulations (Merrifield
Table 3.1: The main diurnal and semidiurnal tidal constituents at mooring DS for horizontal velocities at 65 mab (top) and temperature at 60 mab (bottom)

<table>
<thead>
<tr>
<th>Tidal constituent</th>
<th>$O_1$</th>
<th>$K_1$</th>
<th>$N_2$</th>
<th>$M_2$</th>
<th>$L_2$</th>
<th>$S_2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>m/s</td>
<td>0.0055</td>
<td>0.003</td>
<td>0.025</td>
<td>0.074</td>
<td>0.01</td>
<td>0.029</td>
</tr>
<tr>
<td>Greenwich Phase</td>
<td>341</td>
<td>286</td>
<td>83</td>
<td>52</td>
<td>324</td>
<td>89</td>
</tr>
<tr>
<td>$\times 10^{-3}^\circ C$</td>
<td>7.1</td>
<td>2.2</td>
<td>27</td>
<td>63.5</td>
<td>4.6</td>
<td>14.2</td>
</tr>
<tr>
<td>Greenwich Phase</td>
<td>262</td>
<td>133</td>
<td>337</td>
<td>294</td>
<td>234</td>
<td>338</td>
</tr>
</tbody>
</table>

and Holloway, 2002) in agreement with the observed current (Figure 3.6). The phase comparison suggests that the baroclinic tidal current dominates over the barotropic tidal current at these depths, leading us to conclude that the dominant energy source for tidal motions at the mooring location is the baroclinic component of the tide.

The observed temperature variations (Figure 3.1) are due primarily to the vertical advection of the background stratification by the semidiurnal tide, with maximum temperatures associated with maximum downward displacements. Vertical displacements $\xi$ are computed as $\xi(t) = T(t)/\partial T/\partial z$ where $T(z)$ is the temperature variation and $\partial T/\partial z$ is the temperature gradient. To examine the sensitivity of estimating vertical displacements to the choice of $\partial T/\partial z$, the vertical displacement time series was first calculated for the sensor located 100 mab by using either the mean temperature gradient, or the temperature gradient low-pass filtered with a cut-off period of 48 hours. The standard deviation of vertical displacements is similar ($101 m$ and $98 m$) when using the mean and the low-passed $\partial T/\partial z$, respectively, and the RMS difference between the two estimates is $23 m$. For simplicity, we therefore use the overall mean $\partial T/\partial z$ to calculate vertical displacements from temperature at all the sensors.

The resulting displacement amplitudes exceed 100$m$ at all depths, corresponding to 200$m$ peak to peak changes in isotherm depths over a 12.42$h$ cycle. Because of the proximity of the bottom boundary, these displacements include a lateral component up and down the slope. Vertical displacement phases show a small increase with distance from the bottom, indicating that displacements near the bottom lead displacements near the top of the mooring by approximately 14$^\circ$ (30 minutes). Vertical displacements lag currents by 110$^\circ$, close to the 90$^\circ$ difference expected for a freely propagating internal tide, and also consistent with the advection of stratified water up and down the slope.
3.2.2 Mooring DN

Spectra of temperature at the north mooring (DN) are calculated for the 8 month long time series. The temperature spectrum at the highest sensor (220 mab) shows a sharp peak at the semidiurnal $M_2$ tidal frequency, as well as the $M_3$ and $M_4$ harmonics (Figure 3.7). In contrast to DS, at the sensor closest to the bottom (20 mab) the semidiurnal spectral peak is one order of magnitude smaller than at 220 mab, and the near-inertial/diurnal (0.7 – 1.1 cpd) band is much more energetic (Figure 3.7). There is also more energy in the high frequency band (3 to 16 cpd) as the distance from the bottom increases. The spectra at the other nine sensors show a smooth transition in energy with depth.

The kinetic energy spectrum shows the same features as the temperature spectra (Figure 3.7) at similar depths (Figure 3.8), i.e., a peak in semidiurnal energy and a wide peak in the near-inertial band. Due to the limited vertical coverage of current measurements, variations with depth of the energy in these bands were not observed. The kinetic energy at DN in the near inertial/diurnal band is one order of magnitude higher than at DS (Figure 3.9). The temperature variations in the near inertial/diurnal band are also more energetic at DN than at DS at all the depths sampled (Figure 3.9). We will examine this frequency band in more detail in chapter 4.

At DN, the critical frequency of $\sim$ 1.8 cpd is close to the semidiurnal frequency, yet the temperature spectra show a clear increase in energy with increasing distance from the bottom in this band. This is contrary to both the observations at DS (critical frequency of 3.34 cpd) where energy levels increase with decreasing distance from the bottom, and the linear theory of internal wave critical reflection (Eriksen, 1982), that predicts increased levels of energy around the critical frequency near the bottom. We conclude that the vertical structure of the semidiurnal band energy at mooring DN is inconsistent with wave reflection processes.

The DS and DN moorings were deployed in the path of expected tidal beams that originate near the edges of the ridge, and propagate down and away from the ridge flanks. In situ observations near the ridge edge confirm the presence of downward propagating beams (Martin et al., 2005; Nash et al., 2005; Rainville and Pinkel, 2005). We now consider whether the observed temporal and spatial variations of the semidiurnal band energy are consistent with these predictions. To isolate the astronomically coherent part of the variability, a harmonic tidal analysis (Foreman, 1978) is performed on the temperature and current speed records (Figure 3.10 and 3.11). All resolvable tidal constituents are used at each mooring (35 at DS, 59 at DN). On the north side of the ridge (DN), the percentage of total temperature variance explained by the harmonic tidal analysis increases from
20% near the bottom, to 50% at 200 mab. In contrast, 50 to 70% of the total variance is explained by the harmonic tidal analysis over the entire depth range at DS. We find similar differences between the moorings for the current data harmonic analysis (Figure 3.10). These differences are attributed to the weaker semidiurnal variability at DN compared to DS, especially near the bottom, and the higher near-inertial to diurnal variability at DN. At both moorings, a weakened acoustic return signal higher than 60 mab leads to an increasing number of missing data, which in turn artificially decreases the performance of the tidal analysis for the current data (Figure 3.10). The phase information from the harmonic analysis supports the notion of downward semidiurnal energy propagation (Figure 3.11). There is a steady increase in temperature phase with increasing depth at DN, and also at DS, indicating downward energy propagation. We assume that we are in the presence of internal tidal beams, and that we only sample the lower part of the beams. Given the short mooring lengths, we cannot infer the vertical wavelength of the internal tide from our measurements. The large confidence intervals for the phase near the bottom are a result of small temperature amplitudes at mooring DN. At all depths, the temperature amplitudes are a factor of 2-3 smaller at DN than at DS. The temperature amplitude increases tenfold between 30 and 220 mab at DN (from 3 to 30 millidegree), while the amplitude increase over the same depth range at mooring DS is only 30% (60 to 80 millidegree).

The measured $M_2$ current ellipses at DN are quite weak ($0.02 \text{ ms}^{-1}$) and are directed along the local topography (Figure 3.12), instead of across the topography as is found in other frequency bands (Figure 3.8). The observed ellipses are weaker in amplitude than either the barotropic ellipse predicted by the TPXO model, or the combined baroclinic and barotropic ellipse from the Merrifield and Holloway (2002) numerical model (Figure 3.12). In contrast, at DS Aucan et al. (2005) measured $M_2$ current ellipses directed across the slope that are much larger than the predicted TPXO barotropic ellipse, and equivalent in amplitude to Merrifield and Holloway (2002) numerical estimates. Nash et al. (2005) measured barotropic semidiurnal velocities of $0.04 \text{ ms}^{-1}$ near the 3000 m isobath on the south side of the ridge, with baroclinic velocities exceeding the barotropic velocity. We note that their measurement does not differentiate between $M_2$ and other semidiurnal constituents ($S_2$, $N_2$), that we are able to separate in our analysis.

Aucan et al. (2005) concluded that a semidiurnal beam propagates near the bottom at DS, with little variability in amplitude over the first 200 mab. From these considerations, we conclude that a semidiurnal beam propagates downward over DN, at a distance further from the bottom compared to DS. Assuming that the beams originate near the 1000 m isobath, we suggest that the beam
center is further from the bottom at DN than DS, and hence less tidal energy is present at the bottom at DS than DN.

We cannot resolve the barotropic tidal current directly from the data. Except near the boundary, the barotropic tidal velocities are constant with depth, and we do not believe that the weak tidal amplitudes at DN near the boundary are due to shears in a frictional bottom boundary layer since a similar effect was not observed at DS. In addition, the observed vertical displacement amplitude decreases near the bottom at DN (Figure 3.11), while the vertical displacement amplitudes of a barotropic flow up a slope increase near the boundary. We conclude that the barotropic tide alone cannot explain the observed vertical structure of the semidiurnal band variability at DN. Furthermore, the TPXO and POM models both predict similar barotropic amplitudes on either side of the ridge. This view is consistent with geometrical considerations of the local Kaena Ridge topography.

At DS, the local slope is super-critical with regard to $M_2$, allowing the propagation of a beam from an upper generation site, down and away from the ridge that grazes the mooring site (Figure 3.13). The mooring DN is located on a relatively flatter area on the ridge flank, and the slope is locally sub-critical for $M_2$. The geometry of the ridge may not allow for the propagation near the bottom of a downward propagating beam originating near the ridge top (Figure 3.14). Instead, the beam at mooring DN is likely to detach from the bottom, as suggested in Figure 7 in Nash et al. (2005). The displacement phase information 200 m above the bottom indicates that the $M_2$ beam at mooring DN is still propagating downward, however not necessarily right at the bottom as it does at DS.
Figure 3.1: Representative 14 day time series of temperature ($75mab$), and across-slope and along-slope currents ($65mab$) at mooring DS.
Figure 3.2: Representative 14 day time series of temperature ($52mab$), and across-slope and along-slope currents ($55mab$) at mooring DN.
Figure 3.3: Vertical profiles at DS (red) and DN (blue) of mean current speed (upper left), current speed standard deviation (upper right), across (thick) and along (thin) slope velocity standard deviation (middle left), ratio of along to across slope velocity standard deviation (middle right), temperature standard deviation (lower left), and corresponding vertical displacement standard deviation (lower right).
Figure 3.4: Power spectra at DS of a) horizontal current (65 mab), b) temperature (220 mab and 27 mab), c) vertical shear of the horizontal current (between 40 and 65 mab), d) depth-averaged buoyancy frequency (27 to 220 mab). Top abscissa is in cycles per day (cpd). The Coriolis, $M_2$ and critical frequencies are indicated.
Figure 3.5: Temperature spectra for all available sensors at DS. Corresponding sensor elevations in mab are indicated in the legend. Top abscissa is in cycles per day (cpd).
Figure 3.6: The $M_2$ (left) and $S_2$ (right) measured horizontal current ellipses at DS, with suspect depth bins in gray (i.e., side lobe contamination problems). Barotropic current ellipses predicted by the TPXO model (Egbert, 1997) are depicted at the top. The direction of the topographic gradient is indicated with the red arrow on the left.
Figure 3.7: Temperature spectra for all available sensors at DN. Corresponding sensor elevations in mab are indicated in the legend. Top abscissa is in cycles per day (cpd).
Figure 3.8: Rotary Velocity spectrum (top) and across and along slope current spectrum (bottom) at DN (48 mab).
Figure 3.9: Velocity spectra at DN (blue) and DS (red) moorings, 48 mab (top), temperature spectra at 220 mab (middle) and temperature spectra at 28 mab (bottom).
Figure 3.10: Percentage of variance explained by a harmonic tidal analysis for a) temperature, and b) current velocity, at mooring DS (red) and DN (blue)
Figure 3.11: $M_2$ phase (left) and amplitude (right) of the temperature (top) and the current velocity (bottom) at the DN mooring (blue), and at DS mooring (red). Results from the TPXO models are indicated as ‘x’
Figure 3.12: Bathymetry around mooring DN (top) and DS (bottom), with the measured (thin red) $M_2$ velocity ellipses, the predicted barotropic $M_2$ ellipses from TPXO (thick red), and the combined barotropic and baroclinic $M_2$ ellipses from POM (black, Merrifield and Holloway (2002)). Scale for bathymetry is in meters.
Figure 3.13: Ridge cross-section and theoretical $M_2$ ray paths at DS. Ray paths were constructed step by step by calculating the angle of propagation at each step by using the dispersion relation for a free $M_2$ frequency internal wave and the climatological stratification profiles from station ALOHA.
Figure 3.14: Same as figure 3.13 but for the north side of the ridge
Chapter 4

Near-Inertial and Diurnal Variability

4.1 Introduction

Striking differences between the variability of current and temperature on each side of the ridge are found in the inertial to diurnal band. There is one order of magnitude more energy in the $0.7 - 1.1$ cpd band at mooring DN compared to mooring DS. In this chapter, we consider whether incident NIWs, and/or the diurnal tide can account for the observations. As noted earlier, we discount the possibility that the reflection of internal waves at near-critical slopes alone can account for the increased Near-Inertial/Diurnal (NI/D) band at DN compared to DS. The critical frequencies are $1.79$ cpd and $3.34$ cpd at moorings DN and DS respectively. Velocity spectra at the 2 moorings (Figure 3.9) do not show any clear energy increase around these frequencies that is characteristic of critical reflections.

The intermittent nature of the near-inertial band energy leads us to believe that these are internal wave events forced by winter winds to the north of the ridge (described as the internal swell by Alford (2001)). This scenario is consistent with the higher energy observed on the north of the ridge compared to the south. These near-inertial waves propagate equatorward over the ridge and cannot reach the south mooring DS due to the steepness of the slope on the south side of the ridge, while there is no obstruction for these equator-ward propagating near-inertial waves to reach the north mooring DN. Chiswell (2003) hypothesized that such a shadowing effect occurs at his mooring site off the coast of New Zealand although data are not available to confirm this.

It is possible that the near-diurnal band energy is due to the diurnal tide. Diurnal internal tides can be generated at the ridge topography in the same manner as semidiurnal tides. However, we believe that the event like nature of the variability in this band is more consistent with the NIW band. Rainville and Pinkel (2005) and Carter and Gregg (2005) observed near-diurnal internal waves.
coupled with the semidiurnal tide at the nearby summit of Kaena Ridge. Here however, we note the absence of a peak at $M_2/2$ in our mooring data deep on the ridge north flank, suggesting that non-linear energy transfer from the dominant $M_2$ internal tide to $M_2/2$ internal waves is not the source of energy in the near-inertial to diurnal band (Figure 3.8).

4.2 Near-Inertial and Diurnal Band Temporal Modulation

To investigate the temporal variability in the different frequency bands, we arbitrarily define the near-inertial band from 0.7 to 0.9 $cpd$, the diurnal band from 0.9 to 1.1 $cpd$, and the semidiurnal band from 1.9 to 2.1 $cpd$. Time-depth plots of the variance in the three frequency bands are computed for temperature using 8 day subrecords. The variance time series show episodic events in the diurnal and the near-inertial frequency bands (e.g. days 340 or 375, figure 4.1c and d). The same events are also visible in the variance plots from the current record (Figure 4.2c and d). The semidiurnal band exhibits a fortnightly modulation, as expected for variability associated with the astronomical tide.

We use the results from the harmonic analysis and from the TPXO barotropic model of Egbert (1997) to compare the temporal evolution of the different frequency bands with the beating of tidal constituents (Figure 4.2). The TPXO model contains tidal frequencies $M_2$, $S_2$, $N_2$, $K_2$, $K_1$, $O_1$, $P_1$, $Q_1$, $MF$ and $MM$. We find that the measured semidiurnal band variability is well explained by the tidal constituents. The variability in the diurnal band, however, is not well explained by the beating of the diurnal tidal constituents alone (Figure 4.2c). We interpret the event-like character of the diurnal variability and the weak correspondence with "predicted" tidal behavior as an indication that the diurnal band defined here comprises high-frequency NIWs. Assuming the NIWs are generated at the local inertial frequency at their generation site, we expect the diurnal band waves to have originated north of the islands between $26.5^\circ N$ and $33.5^\circ N$ where $0.9 \leq f \leq 1.1$ $cpd$. This can explain the intermittent nature of the diurnal band, and the overall higher levels of diurnal variability during the winter months (Figure 4.2c).

As expected, the near-inertial band is completely unrelated to the tidal constituents (Figure 4.2d). Observed time-series of temperature (Figure 4.3 a and b) illustrates the depth-dependence of the near-inertial signal. At the top of the DN mooring, between 100 and 200 $mab$, (Figure 4.3a), the temperature oscillations are dominated by the semidiurnal tide, while near the bottom (Figure 4.3b), the temperature oscillates predominantly at periods between the diurnal and the inertial period of 32 hours. During this time period the vertical temperature gradient, calculated between pairs of sensors,
shows two minima separated by 26-28 hours (Figure 4.3c). During these stratification minima, large overturns are observed, leading to high estimated dissipation (see chapter 6 for the details of the overturns analysis). The velocity observed near the bottom, at the depths of the temperature sensors shown in Figure 4.3b, also undergo oscillations at diurnal to inertial periods. The direction of the velocity field and the temperature record are consistent in time with the advection of water along the slope.

4.3 Propagation of NIWs

We observe the largest near-inertial event in temperature oscillations between days 370 and 390 (January of 2003). We propose that these oscillations (as well as near inertial to diurnal band energy in general) are due to impinging internal waves that are forced by the surface winds to the north of the ridge, and propagate southward intersecting the Kaena Ridge. To support this hypothesis we first calculate the ray path of an idealized, equator-ward propagating, inertial wave following Garrett (2001) (his equation 16).

\[
\frac{dy}{dz} \simeq - \frac{N}{(2\omega \beta y)^{1/2}} \quad (4.3.1)
\]

where \(y\) is the meridional, equatorward distance of propagation of an inertial wave generated below the surface mixed layer at a latitude where its frequency is equal to the local Coriolis frequency \((\omega = f)\). \(z\) is the depth, \(\beta = \partial f / \partial y\) is the meridional rate of change of the Coriolis frequency \(f\), and the buoyancy frequency is approximated here by \(N = N_0 e^{-z/b}\). The values of \(N_0\) and \(b\) can be obtained from a best fit to the climatological vertical profile of \(N\) at station ALOHA between the surface and 3000 m depth. This leads to values of \(N_0 = 7.1 \times 10^{-3} \text{ s}^{-1}\) and \(b = 1.25 \text{ km}\), with an RMS difference between the approximate and measured \(N\) profiles of \(9 \times 10^{-4} \text{ s}^{-1}\).

Equation 4.3.1 can be integrated between the surface \(z = 0\) where \(y = 0\) and a depth \(Z\):

\[
Y = \frac{1}{2} \left( \frac{9N_0^2 \beta^2}{\omega \beta} \right)^{1/3} (1 - e^{-\frac{z}{b}})^{2/3} \quad (4.3.2)
\]

Setting \(Z = 2420 \text{ m}\) and the latitude to \(22^\circ N\), gives \(Y = 380 \text{ km}\). An inertial wave ray, if propagating equator-ward, would originate \(380 \text{ km}\) north of the mooring site, near \(25^\circ N\), if there were no previous bottom bounces. An inertial wave generated at this latitude has a period of \(28.2 \text{ hours}\) (or \(0.85 \text{ cpd}\)).
To estimate the propagation time, we use equation 25 from Garrett (2001):

\[
\frac{dy}{dt} = \left( \frac{N_0 b}{j \pi} \right)^2 \frac{\beta t}{\omega}
\]  

(4.3.3)

where \( j \) is the vertical mode number. Equation 4.3.3 can also be integrated between 0 and \( Y \) to give a travel time \( T \):

\[
T = \left( \frac{2 \omega Y}{\beta} \right)^{1/2} \frac{\pi j}{N_0 b} 
\]  

(4.3.4)

For our choice of parameters, this leads to a propagation time \( T \) of 5.8 days, where \( j \) is the wave vertical mode number.

We now try to relate the arrival of these waves deep along the ridge flank to the remote surface wind forcing. We consider the surface winds around the dates of interest in the expected area of generation, north of the mooring at 25° N (Figure 4.4). We examine the wind analysis product from the operational Global Data Assimilation Scheme (GDAS), with winds converted to 10 m height assuming neutral stability. The period covers the entire length of the mooring deployment, and the sampling is every 3 hours. We band-pass filter (periods between 24 and 32 hours) and separate the wind into clockwise and anticlockwise rotating components. The lagged correlation between the time series of the near-inertial clockwise component of the wind at 25°, 158° and the near-inertial kinetic energy remains statistically insignificant (0.38), but shows a small peak at a lag of 10.5 days. In comparison, the expected lag is \( \sim 6 \) days for a mode 1 wave, or \( \sim 12 \) days for a mode 2 wave (equation 4.3.4).

The relationship between wind stress to the north and near-inertial kinetic energy at Kaena Ridge is encouraging but inconclusive. A more elaborate slab layer model forced by the wind (D’Asaro, 1995; Alford, 2003), or a multilevel numerical model (Nagasawa et al., 2000) would be useful for investigating the timing between the generation of NIWs at the surface, and the observed near-inertial variability deep along the Kaena Ridge slope. Nonetheless, we believe that incident NIWs are still a plausible explanation for the intermittent energy in the near-inertial band, and consistent with the shadowing effect proposed for the southern ridge flank.

### 4.4 Interaction of NIWs with the Sloping Bottom

The near-inertial band vertical displacements at mooring DN have a rms value of 18 m. We seek to determine whether the observed vertical displacements are consistent with the horizontal motions of a freely propagating near-inertial internal wave, or if the vertical displacements are the result of the near-inertial horizontal motions directed across the slope, which must satisfy a no
normal flow condition at the boundary. The amplitude of the velocity field associated with a freely propagating internal wave can be expressed as

\[
    u = \pm \frac{A_0 \sqrt{m}}{k} \exp(i(kx + ly + \phi(z) - \omega t)) = \pm u_0 \exp(i(kx + ly + \phi(z) - \omega t)) \tag{4.4.1}
\]

\[
    w = \frac{A_0}{\sqrt{m}} \exp(i(kx + ly + \phi(z) - \omega t)) = u_0 \frac{k}{m} \exp(i(kx + ly + \phi(z) - \omega t)) \tag{4.4.2}
\]

The associated vertical displacement \( \zeta \) can be written as

\[
    \zeta = \frac{w}{i\omega} = u_0 \frac{k}{im\omega} \exp(i(kx + ly + \phi(z) - \omega t)) = u_0 \frac{1}{i\omega} \sqrt{\omega^2 - f^2} \frac{N^2 - \omega^2}{N^2 - \omega^2} \exp(i(kx + ly + \phi(z) - \omega t)) \tag{4.4.3}
\]

The amplitude of the vertical displacement for an idealized propagating near-inertial wave is therefore related to the horizontal velocity amplitude by

\[
    \zeta_0 = u_0 \frac{1}{i\omega} \sqrt{\omega^2 - f^2} \frac{N^2 - \omega^2}{N^2 - \omega^2} \tag{4.4.4}
\]

Alternatively, near the boundary, the flow has to satisfy a zero normal flow condition, so that \( u \) and \( w \) are related by

\[
    w = -u \frac{dH}{dx} \tag{4.4.5}
\]

Where \( z = -H \) is the ocean bottom, and the vertical displacement amplitude is then related to the horizontal velocity amplitude by

\[
    \zeta_0 = -u_0 \frac{1}{i\omega} \frac{dH}{dx} \tag{4.4.6}
\]

In our case, for a near-inertial wave of frequency 0.85 \( \text{cpd} \), a buoyancy frequency \( N = 16 \text{ cpd} \), a local Coriolis frequency \( f = 0.75 \text{ cpd} \), and a topographic slope of 1/10, the ratio of vertical displacement amplitude to horizontal velocity amplitude is \( \sim 450 \) for a free wave, and \( \sim 1600 \) for a wave encountering the slope. To compare this theoretical value of the ratio of vertical displacement to horizontal velocity to our measurements, we analyze in more detail the period of high near-inertial energy around day 375 (Figure 4.3 and 4.5). The current and temperature data (Figure 4.3b and d) are low-passed filtered between 0.7 and 0.9 \( \text{cpd} \) (Figure 4.5). We observe near-inertial across-slope velocities of 0.05 \( \text{ms}^{-1} \) amplitude, associated with a vertical displacement amplitude of \( \sim 60 \text{ m} \), leading to a ratio of displacement to horizontal velocity of \( \sim 1200 \), close to the theoretical ratio of \( \sim 1600 \) obtained above for a wave along the local topographical slope. The phase between temperature and velocity is such that downslope (upslope) flow is associated with increasing (decreasing) temperature. Unfortunately, no current data were available further than 70 \( \text{mab} \) to compare with the temperature data. Above 100 \( \text{mab} \), the low-pass filtered temperature shows an average upward phase (downward energy) propagation of 760 \( \text{mday}^{-1} \) (Figure 4.6b). The amplitude
of the temperature oscillations also decreases by a factor of 2 between 70 and 200 mab, consistent with the observations from Figure 4.1. We conclude that the incoming, downward propagating, near-inertial wave is constrained by the supercritical slope in the lower 100 m to produce vertical displacement larger than theoretically predicted.

We only have measurements for a very limited area of a complicated ridge slope system. Considering the locally supercritical slope, and the small horizontal scales (100’s of meters in the horizontal, 10’s of meters in the vertical) of the topographic features in the mooring area (Figure 3.12) compared to the horizontal scales of a typical near-inertial wave (10’s of km in the horizontal, 100’s of meters in the vertical), we did not consider here the theoretical treatment of the reflection of the NIW by a theoretical plane boundary as was done by Eriksen (1982) or Müller and Liu (2000).
Figure 4.1: Temperature variance, as a function of depth and time, calculated over 8 day intervals, for all frequencies (top), the semidiurnal (1.9-2.1 cpd, middle top), the diurnal (0.9-1.1 cpd, middle bottom), and the near-inertial (0.7-0.9 cpd, bottom) bands;
Figure 4.2: Velocity variance, calculated over 8 day intervals, for the measured current at 43 mab (blue), for the barotropic current from the TPXO model (green), and for the superposition of tidal constituents obtained from the measured current at 43 mab (red), for all frequencies (top), for the semi-diurnal (1.9-2.1 cpd, middle top), for the diurnal (0.9-1.1 cpd, middle bottom), and the near-inertial (0.7-0.9 cpd, bottom) bands.
Figure 4.3: 3 day time series of potential temperature between 68 and 220 mab (a), between 28 and 68 mab (colored lines in b), dissipation (black line in b), buoyancy frequency calculated between sensor pairs ($10^{-3}$ s$^{-1}$, c) overlaid with the top (crosses) and bottom (circles) of detected overturns (see chapter 6). Velocity, rotated so that the across-slope velocity is in the ordinate direction (d).
Figure 4.4: 10 m winds (vector direction and wind speed) obtained from the GDAS product. The circle has a radius of 320 km around the DN mooring location.
Figure 4.5: 5 day segments of temperature (a), vertical displacements (b) and across-slope current (c) at DN, band-pass filtered between 0.7 and 0.9 cpd. The temperatures (a) and displacements (b) are shown for the 6 sensors located between 28 and 68 mab, the current data (c) shown are for bins 43, 47 and 51 mab.
Figure 4.6: 5 day segments of vertical displacements (a), band-pass filtered between 0.7 and 0.9 cpd, and lines of constant phase (b). The displacements are shown for the 11 sensors located between 28 and 220 meter below sea level (mab). '.' and 'x' indicate local maxima and minima used to calculate the phase lines. Phase lines are line fits to the five top sensors (blue) and the bottom 6 sensors (black).
Chapter 5

Estimation of Dissipation and Mixing

5.1 Thorpe Analysis

The moorings provide a time series of coarsely resolved vertical temperature profiles. We use the collocated temperature and salinity measurement (Figure 2.4), to infer salinity at all the sensors depths and to calculate potential density profiles. A noticeable feature of these profiles is the frequent occurrence of statically unstable patches, or overturns. The detection and analysis of these overturns provides an estimate of mixing. Levine and Boyd (2005) used this method to estimate dissipation and mixing at a mooring near the Kaena Ridge crest, and Finnigan et al. (2002) performed this analysis on CTD data at two locations near the Hawaiian Ridge.

We detect an overturn using the following algorithm. First, inflection points in the density profile are detected where the density profile changes from stable to unstable (point A in Figure 5.1). The lower extent of the overturn (point B) is taken as the next point below point A where the density of A and B are equal. The point with the minimum density between point A and B is taken as point C, and the upper limit of the overturn (point D) is taken as the point above A that has the same density as C. Any small overturn contained within a bigger overturn is discarded, and overlapping overturns are combined into one. Similar results are obtained using potential temperature instead of potential density.

Within each detected overturn of vertical size $H$, the unstable profiles are reordered into stable ones. Each sample $\rho_n$ initially at a depth $z_n$ is assigned a new depth $z_m$ in the reordered profile. The difference $d'_n = z_m - z_n$ is called the Thorpe displacement (Thorpe, 1977), and the Thorpe scale is defined as the root mean square of this quantity for each re-ordered overturn:

$$L_T = \left(\overline{d'^2}\right)^{1/2}$$ (5.1.1)
The Thorpe scale can be related to another measure of turbulent dissipation, the Ozmidov length scale (Ozmidov 1965)

\[ L_0 = \frac{\epsilon^{1/2}}{N^{3/2}} \]  

(5.1.2)

Several studies have shown a linear relationship between \( L_T \) and \( L_0 \) (Dillon, 1982; Ferron et al., 1998). Dissipation can be estimated by

\[ \epsilon = a^2 N^3 L_T^2 \]  

(5.1.3)

where \( a = L_0 / L_T \). The value of \( a \) used in previous studies varies between 0.65 and 0.95 (see Finnigan et al. (2002) for a review). For the sake of consistency with the mooring study of Levine and Boyd (2005), we use \( a = 0.8 \) in this analysis. Levine and Boyd (2005) also discuss the validity of the Thorpe scale method for estimating dissipation.

For each detected overturn of size \( H_i \), we calculate the dissipation \( \epsilon_i \) using equation (5.1.3). For each profile, the average dissipation for the bottom layer is \( \bar{\epsilon} = \frac{1}{H} \sum \epsilon_i H_i \) where \( H = 200 m \).

In (5.1.2) and (5.1.3), \( N \) is obtained from the reordered profile, and therefore is always real. Following Osborn (1980), we also compute the vertical eddy diffusivity coefficient \( K_\rho \) by assuming that the turbulent kinetic energy balance is between shear production, buoyancy loss and turbulent dissipation,

\[ K_\rho = \Gamma \epsilon \bar{N}^{-2} \]  

(5.1.4)

The mixing efficiency \( \Gamma \) is taken equal to 0.2, and \( \bar{N} \) is now the background stratification.

## 5.2 Results

At mooring DS, 34\% of the profiles contain at least one overturn, significantly more than what we find at mooring DN (12\%), and what Levine and Boyd (2005) found near the ridge top at mooring Big Boy (\~10\%). The distribution of dissipation estimates follows an approximate lognormal distribution at both moorings, with more energetic overturns at DS than at DN (Figure 5.2). At DS, large overturns, corresponding to temperature differences of 30 millidegree or more, account for 15\% of the total dissipation, but are less frequent than the smaller overturns (Figure 5.3). These large overturns exceed 120 \( m \) in height (Figure 5.3b), and thus span several instruments. Inside the overturn, the vertical temperature profile displays a typical S shape (Figure 5.4). Including the detection of small overturns (\( \leq 4.10^{-3} \degree C \)) in our dissipation estimate cannot account for more than 10\% of the final estimate of dissipation (Figure 5.3c).
Mooring Profiles with overturns Average Dissipation Diffusivity
DS 34% $1.2 \times 10^{-8} W kg^{-1}$ $2 \times 10^{-4} m^2 s^{-1}$
DN 12% $1.8 \times 10^{-9} W kg^{-1}$ $2.5 \times 10^{-4} m^2 s^{-1}$

Table 5.1: Summary of the overturns analysis at both moorings

At mooring DN, fewer large overturns are detected, and the distribution of dissipation shows a higher percentage of dissipation contributed by smaller overturns compared to DS (Figure 5.5). Overturns at DN are in general less pronounced (Figure 5.6). The overall time-averaged dissipation at mooring DS is $1.2 \times 10^{-8} W kg^{-1}$ and $1.8 \times 10^{-9} W kg^{-1}$ at DN.

In sections 5.3-5.6, we discuss the sensitivity of our analysis and results to the different potential sources of errors. We will show that uncertainties in our analysis are greater at DN than DS, but still cannot account for the factor of 10 difference in dissipation between the two moorings (Table 5.1). A more detailed analysis of individual mixing events in the context of the observed flow will follow in chapter 6.

5.3 Temperature Bias Correction

For the detection of density inversions measured at independent instruments, it is important to account for any bias errors in the measurements. Averaged over the 3 months (8 months) of the experiment, temperatures show slight departures from a linear depth profile at mooring DS (DN) (Figure 2.3). Because the tidal vertical displacements are of the same magnitude as the height of both moorings, we assume that the sensors all sample water with similar properties, and so the departures from a smooth linear vertical profile are taken to be instrument bias errors. We therefore calculate corrections for each of the sensors so that the 3 months (8 months) averaged vertical profile is linear. The resulting linear profile (Figure 2.3) requires no more than a 5 millidegree shift in mean temperature for any given sensor. The manufacturer rating of the bias is 2 millidegrees.

For DS, we calculated the effects of the bias removal on the overturn detection and the estimated dissipation (Figure 5.3). After the bias removal, of the 41761 profiles analyzed, 14700 or 34 % contained at least one overturn $> 24 m$. The correction applied to the temperature profile minimizes the detection of spurious overturns arising from sensor errors. For comparison, without the temperature correction, 23000 profiles (or 54 % of the total) contain at least one overturn $> 24 m$, but most of these overturns are associated with small temperature differences approaching the accuracy of the sensors (Figure 5.3a). The temperature correction also increases the mean size...
of small overturns and decreases the size of large overturns (Figure 5.3b). The correction reduces the overall mean dissipation from $3.6 \times 10^{-8}$ to $1.2 \times 10^{-8} Wkg^{-1}$, while the large overturns are responsible for a relatively larger percentage of the dissipation (Figure 5.3c). At mooring DN, the effect of the temperature bias is essentially the same as at DS (Figure 5.5). We note however, that the number of large energetic overturns is significantly lower at mooring DN than at mooring DS, consistent with a lower overall average dissipation (Table 5.1).

5.4 Temperature Error Sensitivity

We test the sensitivity of the Thorpe scale analysis to random measurement errors. The Seabird SBE 39 thermistors have a nominal accuracy of 2 millidegrees. This error value is provided by the instrument manufacturer, and independent calibrations have shown this value to realistically describe this instrument performance (Lukas 2005 Pers. comm.). A significant fraction of the dissipation at both moorings is contributed by overturns spanning a temperature difference between instruments larger than the instrument error of 2 millidegree (Figures 5.3 and 5.5). For the largest overturns detected ($\geq 100 m$), at both moorings, this temperature error does not affect the detection of these overturns (Figures 5.4 and 5.6).

To quantify the sensitivity of our analysis to the instrument error, we perform a Monte-Carlo simulation on a series of profiles from mooring DN with a synthetic noise added. We use a sub-sample of the entire dataset containing 6000 vertical temperature profiles ($\sim 21 \text{ days}$). In this subsample, 12.5\% of the profiles contain at least one overturn. The average dissipation for the subsample is $1.91 \times 10^{-9} Wkg^{-1}$. We added synthetic instrument noise to these profiles by adding a normally distributed random number with a zero mean and a standard deviation equal to the instrument nominal accuracy of 2 millidegree. We infer salinity from the temperature after adding the noise. Averaged over 1000 such profile time-series, 21\% of the profiles contained one or more overturn (up from 12.5\%), and the average dissipation increased to $3.32 \times 10^{-9} Wkg^{-1}$ (up from $1.91 \times 10^{-9} Wkg^{-1}$).

For mooring DS, with a subsample of 10000 profiles, the dissipation increases from $8.05 \times 10^{-9}$ to $10.10 \times 10^{-9} Wkg^{-1}$, and the percentage of profiles showing overturns increases to 43\% from 31\%.

In conclusion, the synthetic instrument noise raised the mean dissipation by $1.4 \times 10^{-9} Wkg^{-1}$, a $\sim 50\%$ increase at DN and a 25\% increase at DS. The synthetic profiles contain high vertical wavenumber noise that is not a feature of the original profiles. We believe that the error estimate
is an upper bound on the effect of instrument noise on the Thorpe scale analysis. We conclude that random instrument noise alone cannot account for the strength of the inferred overturns. The uncertainty of the mixing estimates obtained with this analysis is much smaller than the factor 10 difference in the overall average of dissipation between the two moorings.

### 5.5 Salinity Compensation

We also investigate the sensitivity of our analysis to the presence of salinity compensated temperature inversions. We inferred salinity at the sensor depths from the approximately linear TS relationship measured at $mab$. If salinity compensated temperature inversions occur, the Thorpe analysis would predict spurious overturns. Overall, if salinity compensation occurred, it would lower the standard deviation of potential density. We calculate the time series of potential density at the collocated T/C sensor from the measured temperature and the measured salinity on one hand, and with the measured temperature and the inferred salinity on the other hand. We found that the standard deviation of the potential density is reduced by 2% when we used the inferred salinity for the density calculations. We conclude that we are not overestimating dissipation by detecting false density overturns created by a temperature measurement error or a salinity compensated temperature inversion.

### 5.6 Clock Drift and Sampling Time Difference

Temperature was recorded every 3 minutes at DS, and every 5 minutes at DN. We consider how the difference in sampling would affect the number of overturns detected, and the estimated dissipation. The temperature at DS is resampled every 5 minutes using linear interpolation, and subjected to the same overturn detection algorithm. We find that the resampling from 3 to 5 minutes slightly reduces the percentage of profiles containing at least one overturn (from 34 to 30%), and also reduces the estimated dissipation from $1.2 \times 10^{-8}W kg^{-1}$ to $1.02 \times 10^{-8}W kg^{-1}$.

According to the manufacturer, the instruments have a very slow temporal drift, typically less than 0.2 millidegree per month. We checked for differential sensor drift over time by repeating the bias corrections based on monthly averages. The results are similar to the 3 month (8 month) time averages, suggesting that differential drift errors are not an important factor. The sensor clock drifts were found to be less than $30 sec$ at any given sensor for DS, and less than $40 sec$ at DN, so we conclude that instrument clock drift has minimal effect on our 3 and 5 minutes averages. The
difference of sampling times between the moorings therefore cannot explain the large differences found in the estimated dissipation between the two moorings.

Figure 5.1: Schematics of the overturn detection algorithm, showing a vertical density profile (black line), see section 5.1 for the definition of the points.
Figure 5.2: Distribution of dissipation events at mooring DS (red) and DN (blue), as a percentage of the detected overturns (top) and as a percentage of the total number of profiles (bottom).
Figure 5.3: Mooring DS a) The distribution of the number of overturns, before and after the temperature bias correction, as a function of the observed temperature difference across the depths of the overturn. b) The distribution of overturn size as a function of the temperature difference across the overturn. c) The distribution of the percentage of the total dissipation contributed by the overturns, as a function of the temperature difference across the overturn.
Figure 5.4: a) Vertical profile at DS on day 249, hour 11.5 (Figure 6.4) of measured absolute temperature after removing the long term bias. *'s indicate the locations of the instruments. Thin lines represent the measurement plus or minus the nominal accuracy of the instruments. b) same as a) but with inferred potential density ($-1000\text{kgm}^{-3}$).
Figure 5.5: Same as figure 5.3, but for mooring DN.
Figure 5.6: same as 5.4 but at DN for day 375, hour 11.5 (Figure 4.3).
Chapter 6

Mixing Events and Mechanisms

6.1 Mixing Events at Mooring DS

6.1.1 The Two Types of Mixing Events

At mooring DS, the estimated Thorpe scales tend to be large/small during spring/neap tides (Figure 6.1a). Thorpe scales range from 20 to 100 m for overturns of 24 to 190 m. Dissipation values range from $10^{-7}$ to $10^{-6} W/kg^{-1}$ during spring tides, and from $10^{-8}$ to $10^{-7} W/kg^{-1}$ during neap tides (Figure 6.1b). The minimum attainable dissipation with this method is $1.2 \times 10^{-10} W/kg^{-1}$.

Because the observations are dominated by oscillations at the $M_2$ frequency, we divide the time series into 163 $M_2$ cycles to investigate the mixing events in relation to tidal phase. A complex demodulation of the horizontal velocity is used to assign each time step to an $M_2$ phase. With 8 minute sampling, 93 phases are specified in a 12.4 hour $M_2$ cycle. Temperature, current, stratification and dissipation occurring at the same phase are then averaged and displayed as a function of $M_2$ phase, which forms a composite tidal cycle (Figure 6.2). With this convention, upslope flow occurs for phases between $-90^\circ$ and $90^\circ$, downslope flow occurs during the rest of the cycle.

Enhanced dissipation occurs during two distinct phases of the tidal cycle (shaded areas on figure 6.2). The first is centered around $140^\circ$ and is associated with maximum downslope flows, increasing water temperatures, and low stratification. We refer to this mixing phase as "downslope flow mixing". Figure 6.3 is an example of a downslope flow mixing event. The current starts flowing downslope and the temperature increases near the bottom first, compared to at a distance from it, thus lowering the stratification. The temperature becomes nearly homogeneous over the 200m depth range during the event, followed by temperature inversions that affect the entire range.
The second dissipation peak, centered around $-90^\circ$, is associated with the flow reversal from downslope to upslope, maximum water temperatures, and increasing stratification with a pronounced local minimum. We refer to this phase as "flow reversal mixing". Figure 6.4 is an example of a flow reversal mixing event. At the end of the downslope flow, the temperature has reached a maximum, isotherms start to converge before overturning, and the event is followed by a sharp temperature drop and a subsequent restratification.

Although observed mixing events are linked to tidal phase, they do not occur as regularly as the tide itself. To illustrate the variability of tidal mixing, we visually inspect each $M_2$ cycle and assign the cycle to one of five categories: "downslope flow mixing only" when dissipation occurs near the maximum downslope current, "flow reversal only" when dissipation occurs at the reversal from down to upslope flow, "both downslope flow mixing and flow reversal" when both events occur during the same cycle, "no mixing" when dissipation failed to reach a threshold value of $0.5 \times 10^{-7} \text{W kg}^{-1}$, and "random mixing event" when dissipation events occur at other phases of the tide. We find that 63% of the tidal cycles contain flow reversal and/or downslope flow mixing, 21% have no significant mixing, and 16% show only odd mixing events. Tidal cycles containing flow reversal mixing events account for 40% of the total estimated dissipation, downslope flow events account for 20%, combined events 21%, and random events 16%. Although downslope flow mixing events appear throughout the experiment, flow reversal mixing events are nearly absent during the first spring-neap cycle, and their occurrence and amplitude appear to increase over time (Figure 6.5a). We attribute this increase in part to the increase in semidiurnal current amplitude (Figure 6.1). In addition, the first spring-neap cycle is characterized by more circularly polarized tidal currents than the other cycles (Figure 6.5b). We will return to this point when considering the cause of flow reversal mixing events in section 6.1.2.

We next consider whether tidal shear and strain act to trigger mixing events, in the manner of a shear instability. We calculate the shear over the depth range 41 to 65 m $\text{ab}$. Again, the ADCP only samples a small fraction of the 200 m length of the thermistor chain. $N$ is averaged over the same range as the ADCP, and also over the full range of the thermistors (27 to 220 m $\text{ab}$).

Representative downslope flow mixing events (Figure 6.6) occur during each semidiurnal cycle near the time of maximum downslope flow. Flow reversal mixing events are absent during the time period shown. Downslope flow mixing occurs when the stratification is at a minimum, or equivalently the strain is a maximum, over the tidal cycle. Tidal current shear tends to peak twice during the cycle during both up and downslope flow (Figure 6.6 b and c); however, the combination of high shear and strain during the downslope phase results in an inverse Richardson number highly
correlated with the dissipation events (Figure 6.6 b and d). The results are similar using different depth ranges for computing $N$ (i.e. 20 m, versus 200 m). Figure 6.6c shows the results using $N$ averaged over 200 m because it is less noisy than that based on a 20 m average. We conclude that downslope flow mixing events are the result of shear instability.

During other time periods, flow reversal mixing events are dominant and noticeable downslope flow events are absent (Figure 6.7). Strong flow reversal mixing events occur every other semidiurnal cycle. There are also other periods (not shown) when moderate flow reversal events occur at every semidiurnal cycle. In sharp contrast to downslope flow mixing, flow reversal mixing events do not coincide with elevated shear between 45 and 65 $mab$ or elevated inverse Richardson number (Figure 6.7 c and d). Similar results are obtained using $N$ averaged over the common 20 m depth range as shear, and also using shear and strain computed over 10 m spacings. Stratification minima occur during the events at a phase of the tidal cycle when the stratification would otherwise be increasing (Figure 6.7 b). The stratification begins to decrease approximately one hour before the main overturning event. This increased strain occurs because of a phase lag of the temperature signal with depth. Inspection of the temperature record for a typical event (Figure 6.7 a) shows that the isotherms begin to converge prior to mixing because temperatures decrease in the upper water column (150 to 220 $mab$) one hour before the lower water column (27 to 75 $mab$). In the lower water column, the measured currents and temperatures are consistent in that temperatures increase during downslope flows. In the upper water column, we do not have reliable current measurements to confirm that the early temperature decrease is due to a change to upslope currents.

The poor relationship between flow reversal mixing events and shear or Richardson number may be due to the lack of reliable current observations above 65 $mab$. For example, a shear instability, similar to the downslope flow mixing event and occurring above 65 $mab$, may create an overturn that is advected into the sample range. We believe this is unlikely, however, because the background currents below 65 $mab$ are near zero or directed upslope (i.e., at flow reversal or shortly thereafter), while the temperature record below 65 $mab$ shows a decrease much larger than the temperature changes observed during downslope flow mixing.

The character of the temperature signal strongly suggests that cold water downslope of the mooring has been lifted above the mooring, resulting in a convective instability. Levine and Boyd (2005) also observed mixing associated with the semidiurnal tide to occur at two distinct phases on the north side of Kaena Ridge at 1500 m depth (Big Boy mooring, Figure 1.3). In their case, overturns near 100 $mab$ occur 180° out of phase with overturns near 300 $mab$, which is attributed to a 180° phase shift in maximum tidal strain between these depth ranges.
For both mixing types, strain appears to be an important precursor to overturning. For downslope flow mixing, overturns occur during the maximum strain over the semidiurnal tidal cycle (Figure 6.2 and 6.6). For flow reversal mixing, overturns occur during a secondary strain maximum that is out of phase with the semidiurnal cycle. Over the entire dataset, the correlation between the strain and dissipation is insignificant (0.19); however, when calculated for a subrecord when downslope flow mixing is dominant (the 3 days shown on Figure 6.6), the correlation improves (0.45). For subrecords of similar length, at times when flow reversal mixing is dominant, the correlation between the strain and dissipation does not exceed 0.2. Tidal current shear, at least over the measured depth range, is not significantly correlated with dissipation, even over short periods when only one mixing type is observed (Figures 6.6 and 6.7). Combining the effects of shear and strain, in the form of an inverse Richardson number, does not improve the correlations obtained using strain alone, which again highlights the primary importance of strain for downslope flow events, and further discounts shear instability as a mechanism for flow reversal events.

6.1.2 Obliquely Propagating Internal Tides and Flow Reversal Mixing

Gemmrich and van Haren (2001) documented the occurrence of thermal fronts near the bottom boundary in ~ 850 m depth along the Bay of Biscay continental slope. They describe abrupt temperature changes, linked to the presence of an internal tide beam propagating downward at an oblique angle relative to the slope. Gemmrich and van Haren (2001) hypothesized that the obliquely propagating internal tide results in a variation of tidal phase along isobaths, particularly if the internal tide is in the form of a narrow beam. The variable phase of the tidal currents can advect cold water upslope above warm water once per tidal cycle, creating sharp thermal fronts and convective instability. The oblique propagation angle is crucial; an internal tide propagating directly downslope would not generate such unstable fronts. Gemmrich and van Haren (2001) observed sharp temperature drops associated with the collapse and/or passage of these fronts. In contrast to the studies of the reflection of internal tides normally incident on a sloping boundary (Nash et al., 2004; Legg and Adcroft, 2003), the mechanism identified by Gemmrich and van Haren (2001) relies on downgoing internal tides that propagate obliquely along the slope.

Similarities between the Bay of Biscay and the Kaena Ridge observations suggest that this mechanism may explain the mixing events documented here as flow reversal mixing. First, for both experiment sites, the observations were made over a supercritical slope for the semidiurnal tide, which allows a tidal beam generated at a ridge crest or shelf break to propagate downward without reflection. Such a beam would propagate farther from the slope with increasing depth;
however, Gemmrich and van Haren (2001) describe how obliquely propagating waves still can intersect the slope. This is apparently also the case at the Kaena Ridge. The observed tidal currents are oriented at an angle relative to the slope (Figure 6.2), consistent with a tidal beam with an oblique propagation azimuth. Following Gemmrich and van Haren (2001), this could lead to the generation of fronts with an angle $\varphi$ between 54° and 80° relative to the isobaths. Here we use equation 1) from Gemmrich and van Haren (2001) with $N^2 = 10^{-6} s^{-2}$, a topographic slope $\alpha = 18.6^\circ$, a $M_2$ vertical propagation angle of 7.5°, and an angle between the isobath and the bottom projection of the group velocity vector between 40° and 80°. During the first spring-neap cycle, when flow reversal mixing events are not detected, the current ellipse is more circularly polarized than during the subsequent cycles (Figure 6.5b). This suggests that a circular current ellipse, as opposed to a more unidirectional flow, is not conducive to this overturning mechanism.

Second, the temperature data from the Bay of Biscay and the Kaena Ridge bear a striking resemblance (Figures 6.4 and 6.8). The temperature drops sharply at the time of flow reversal from down to upslope, which Gemmrich and van Haren (2001) characterized as a passing thermal front. The similarities are particularly striking near the bottom (Gemmrich and van Haren (2001) observations were all below 50 mab). Higher in the water column (≥ 100 mab), above the elevations sampled in the Bay of Biscay, we see evidence for strong restratification following the front passage.

Third, tidally driven convection is an attractive mechanism for flow reversal events given that tidal current and current shears are weak during this phase of the tide. Enhanced strain, which is observed preceding mixing events (Figure 6.4), can also be a signature of a developing convective instability rather than a shear instability.

Lastly, the dependence of this mechanism on the location of the internal tide beam relative to the mooring (Figure 6.9) can also explain the intermittent nature of the observed mixing events. At our Kaena Ridge site, low dissipation is associated with low tidal amplitude, but high tidal amplitude is not always associated with strong dissipation (Figure 6.1). Our analysis has focused on the relationship of mixing events to the measured tidal currents. We emphasize that the measured tide is dominated by the baroclinic component at the mooring location (Figure 6.2 c). The observed increase in amplitude of the semidiurnal currents (Figure 6.1c), and the change in eccentricity over the duration of the experiment (Figure 6.5b) presumably are attributed to changes in the internal tide. Low frequency currents and changes in stratification higher in the water column, between the mooring location and the generation site near the ridge top, can cause temporal changes in
amplitude, phase, position or direction of the internal tide, which apparently influence the level of convective mixing observed at the fixed moored location (Figure 6.5a).

6.2 Mixing Events at Mooring DN

We found an overall time-averaged value of dissipation at mooring DN of $1.8 \times 10^{-9}$ $W kg^{-1}$, corresponding to an averaged eddy diffusivity of $2.5 \times 10^{-4}$ $ms^{-2}$. Due to the large spacing between sensors on the upper half of the mooring ($\Delta Z = 24 - 32 m$), these values are likely to be low estimates as overturns $\leq 32 m$ cannot be detected. In comparison, an averaged dissipation of $1.2 \times 10^{-8}$ $W kg^{-1}$ and an eddy diffusivity of $2 \times 10^{-3} ms^{-2}$ are estimated for mooring DS using the same method. Most of the dissipation events at the south mooring were linked to an energetic semidiurnal tidal beam propagating near the bottom compared to DS. At mooring DN, the semidiurnal tide is much weaker ($0.02$ for $M_2$ alone compared to $0.05 ms^{-1}$ at DS, Figure 3.11) and we postulated in section 3 that a beam is observed at mooring DN, but it is clearly detached from the bottom. Consistent with the notion that the tides are the dominant energy source for deep mixing at the Hawaiian Ridge, the average dissipation at the north mooring DN is one order of magnitude smaller than at the south mooring DS. At mooring DN, energetic dissipation events appear linked to near-inertial and diurnal events as well as the semidiurnal tide. We use the temporal variations of the semidiurnal and the combined near-inertial/diurnal variance (Figure 4.2) to examine the occurrence of dissipation (Figure 6.10). We find that dissipation is higher during periods when both semidiurnal and near-inertial/diurnal variances are high (Figure 6.10).

To examine the occurrence of mixing relative to the tidal phase, we compute a composite phase average, similar to the one obtained for DS, based on the semidiurnal oscillation of the upper most temperature sensor (Figure 6.11). Because the temperature oscillations near the bottom ($20 - 50 mab$) are weak, the average tidal strain between 20 and 220 mab is driven primarily by the temperature variations at 220 mab. Similar to DS, dissipation at DN tends to occur during a particular range of tidal phase. Unlike DS, only one dissipation “event” is identified. At DN, phase-averaged dissipation is maximum when the flow reverses from up to downslope, just the opposite of DS flow reversal mixing events when the flow switches from down to upslope. At this phase of the tide, however, tidal strain creates the weakest stratification (Figure 6.11) over the semidiurnal cycle. At DS, the timing of the weakest stratification over the cycle occurs near maximum downslope flow, hence the timing of downslope flow mixing. We conclude from these comparisons that tidal strain conditions the water column in a way that is favorable for mixing. Flow reversal mixing
at DS does not follow this pattern, and we attribute this to the importance of convective mixing for this particular type of mixing event.

Is the presence of near-inertial and diurnal energy important for mixing at DN compared to DS? In total, the effect cannot be large since mixing at DS is estimated to be an order of magnitude higher than at DN. Nevertheless, we find compelling evidence to suggest that near-inertial events do play a role in triggering mixing at DN. In chapter 3 we showed that the upper 100 m of the mooring (120 to 220 mab) is dominated by temperature variability in the semidiurnal band (Figure 4.1), attributed to a semidiurnal beam detached from the bottom. We also showed in chapter 4 that the lower part of the mooring (20 to 100 m above the bottom) is dominated by near-inertial and diurnal variability, in both temperature and velocity (Figures 4.1 and 4.2). These tendencies are also visible in details of the time-series (Figures 4.3 and 6.13). The upper temperature sensors show a general semidiurnal oscillation (Figures 4.3a and 6.13a), while the sensors close to the bottom oscillate at periods between the diurnal and inertial periods (Figures 4.3a and 6.13 b). At some phases, the semidiurnal oscillation at the top, and the near-inertial/diurnal oscillation near the bottom coincide to create times of high strain, i.e. the temperature (inertial to diurnal) in the lowest sensors is increasing, while the (semidiurnal) temperature at the higher sensors is decreasing. At these times, we observe overturns and associated mixing (Figure 4.3).

We now explore the occurrence of overturns and turbulent mixing in the context of combined near-inertial variability at the bottom and semidiurnal variability further up. The entire seven month temperature time series, measured at the upper most sensor (220 mab) is band-pass filtered around the semidiurnal frequency. The temperature record nearest to the bottom (27 mab) is band-pass filtered between the near-inertial and diurnal frequencies (0.7 – 1.1 cpd). All the recorded dissipation events are then plotted as a function of these two band-pass filtered temperatures (Figure 6.12). Similar to what we observe in Figure 4.3, the majority of dissipation occurs in the upper left quadrant, which corresponds to a low semidiurnal temperature near the top, and a high near-inertial temperature at the bottom. This is the configuration that leads to the weakest temperature gradient and the highest strain over the 200 m depth range. Thus we observe overturns and associated mixing during periods of high strain, which are dictated by the occurrence of semidiurnal upwelling (lowest tidal temperature) at some depth from the bottom (100 – 200 m), and near-inertial/diurnal downwelling (warmest temperatures) near the bottom. We conclude that the mixing at DN is strain related, and that this strain is a combination of both tidal and near-inertial internal wave fluctuations.
Figure 6.1: a) The estimated Thorpe scale ($L_T$), b) dissipation ($\epsilon$), and c) the measured semidiurnal tidal amplitude using horizontal currents at 65 mab at mooring DS.
Figure 6.2: A composite semidiurnal tidal cycle at DS, obtained using ensemble phase averages over the entire time series, of a) potential temperature recorded at 27, 43, 59, 75, 91, 107, 124, 147, 171, 195, and 220 mab, with blue lines corresponding to sensors closer to the bottom, depth averaged buoyancy frequency (27 to 220 mab, dashed black line), b) estimated dissipation ($\epsilon$, equation 5.1.3), and c) horizontal currents, rotated so that cross-slope velocity is vertical. Depth bins with suspect side lobe contamination are shown in gray. Estimates of the barotropic current (TPXO, (Egbert, 1997)) are included. Shaded areas indicate phases of intense mixing associated with downslope flow (dark), and flow reversal (light).
Figure 6.3: A characteristic overturn event depicted in 12 hour time series (day 233) of a) potential temperature recorded at 27, 43, 59, 75, 91, 107, 124, 147, 171, 195, and 220 mab, with blue lines corresponding to sensors closer to the bottom, b) estimated dissipation ($\epsilon$, equation 5.1.3, thin gray line) and depth-averaged buoyancy frequency (27 to 220 mab, thick black line), and c) horizontal currents, rotated so that cross-slope velocity is vertical. Depth bins with suspect side lobe contamination are shown in gray.
Figure 6.4: Same as figure 6.3 for day 249, showing an overturn during a different phase of the tide.
Figure 6.5: a) Dissipation ($\epsilon$) for the DS mooring, averaged over a semidiurnal cycle, is classified in terms of the dominant mixing types that occur during the cycle. b) The eccentricity of the semidiurnal current ellipse, quantified as the ratio of the minor over the major axis current amplitude. Ellipse amplitudes are obtained from a harmonic analysis of 7 day subrecords.
Figure 6.6: Downslope flow mixing events at mooring DS in 3 day time series of a) potential temperature, b) depth-averaged buoyancy frequency $N$ (27 to 220 mab, thick black line), and estimated dissipation (thin gray line), c) square of the horizontal current shear (from 41 to 65 mab), d) inverse Richardson number (from shear and $N$ above), and e) the horizontal velocity (at 65 mab and rotated so that across-slope velocity is vertical).
Figure 6.7: Same as figure 6.6 for a period of flow reversal mixing events.
Figure 6.8: Reproduced figure 4 from Gemmrich and van Haren (2001), Showing temperature (a) and velocity (b) measured 7.9 m above the bottom on the continental slope of the Bay of Biscay. The occurrence of thermal front passages are evident in a). The alongslope (thick line) and cross-slope (thin line) current component are shown.
Figure 6.9: Reproduced figure 6 from Gemmrich and van Haren (2001), showing a plan view of the idealized distortion of initially parallel isotherms due to cross-slope temperature gradients being advected by oblique internal waves. Note that for clarity the flow field (arrows) presents conditions at a quarter wave period prior to the instant of the depicted temperature field. Only a narrow beam, containing one wavelength of a periodic wave, is shown. Dashed lines mark locations where a thermal front is being generated. Here $T_0 < T_1$. The star represents the measurement site: (a) $t = t_0$, representing unstable stratification at measurement site, (b) $t = t_0 + \tau/2$, where $\tau$ is the wave period, representing stable stratification at measurement site.
Figure 6.10: 24 hours running average of dissipation (in color) as a function of the current variance in the inertial-diurnal frequency band (0.7 – 1.1 cpd, horizontal) and the semidiurnal frequency band (1.9 – 2.1 cpd, vertical). a) is the average dissipation per bin (top), b) is the total dissipation per bin (middle), and c) is the number of samples per bin (bottom). Color scale for the top 2 panels are log10 of the dissipation in $W/kg^{-1}$, log10 of samples in the bottom panel.
Figure 6.11: A composite semidiurnal tidal cycle at mooring DN, obtained using ensemble phase averages over the entire time series based on the temperature at 220 mab, of a) potential temperature recorded at 28, 36, 44, 52, 60, 68, 92, 125, 156, 188, and 220 mab, with blue lines corresponding to sensors closer to the bottom, b) estimated dissipation ($\epsilon$, equation 5.1.3), and depth averaged buoyancy frequency (27 to 220 mab, dashed black line), and c) horizontal currents, rotated so that cross-slope velocity is vertical.
Figure 6.12: Dissipation at mooring DN, as a function of the temperature at 27 mab band-passed filtered in the inertial-diurnal frequency band (0.7 − 1.1 cpd, ordinate) and at 220 mab band-passed in the semidiurnal band (1.9 − 2.1 cpd, abscissa). Pictured are the average dissipation per bin (top), the total dissipation per bin (middle), and the number of samples per bin (bottom). Color scale for the top 2 panels are log10 of the dissipation in W kg⁻¹, log10 of samples in the bottom panel.
Figure 6.13: 3 day time series at DN of potential temperature between 68 and 220 mab (a), of potential temperature between 28 and 68 mab (colored lines in b), and dissipation (black line on b), the buoyancy frequency calculated between sensor pairs ($10^{-3} \text{ s}^{-1}$), (c) overlaid with a measure of the top (crosses) and bottom (circles) of overturns. Velocity, rotated so that the across slope component is in the ordinate direction (d)
Chapter 7

Discussion

7.1 Mixing at Mooring DS

In the field experiment at mooring DS, we find that the semidiurnal tide dominates the current and temperature variability above the steep south flanks of the Kaena Ridge. In particular, near-inertial wave energy, diurnal tides, and subinertial currents, which might contribute to near boundary mixing due to bottom drag, are all much weaker than the semidiurnal tide. The predominance of the semidiurnal tide and lack of diurnal tide energy are consistent with their respective predicted ray paths from the model simulations of Merrifield and Holloway (2002). Barotropic to baroclinic semidiurnal tidal conversion near the top of Kaena Ridge results in downward propagating tidal beams that are the primary source of mechanical energy available for turbulent mixing deep along the ridge (Figure 3.13).

Overturns and the implied associated mixing and dissipation occur predominantly at two phases of the measured semidiurnal tidal cycle (Figure 6.2): when flows are near maximum in the down-slope direction, and at the flow reversal prior to upslope flow, at maximum downward isopycnal displacements. The mechanism causing the downslope flow mixing events appears to be a shear instability, triggered primarily by high strain, occurring when stratification is minimum over the tidal cycle. Levine and Boyd (2005) find a similar relationship between mixing events and strain at a shallower depth on the ridge. Flow reversal mixing occurs during high strain conditions that do not coincide with the tidal strain maximum. These events appear unrelated to shear, and the associated dissipation is poorly correlated with the inverse Richardson number, suggesting that shear instability is not the primary generation mechanism. Based on similarities with observations made by Gemmrich and van Haren (2001) in the Bay of Biscay, we suspect that the flow reversal mixing events are convectively driven. In this scenario, mixing develops because of the oblique
propagation angle of the downward propagating internal tide relative to the slope. This leads to
the advection of cold water above warm water along the slope, the generation of a sharp thermal
front, and eventually to statically unstable conditions and overturning. Evidence supporting this
hypothesis includes the observed orientation of the internal tide relative to the slope, the abrupt
temperature decrease associated with the overturn, and the poor correspondence with tidal shear.
The enhanced strain preceding these events may result from the development of a statically unstable
patch.

Ultimately, we seek a tidal mixing parametrization that can be incorporated into regional
numerical models, so that estimates of deep mixing can be extrapolated to the entire ridge system,
as well as to other locations. If tidal mixing were related to shear instabilities created by the tidal
flow at the boundary, we might expect that the mixing associated with the tide would have a high
predictability. The barotropic tide can be accurately predicted by a harmonic analysis. The phase of
the freely propagating internal tide, once generated, is decoupled from the astronomical forcing. At
this site, although the internal tide provides most of the mechanical energy, a few tidal constituents
explain a significant fraction of the observed variability, indicating a good coherence between the
phase of the internal tide and the astronomical forcing, presumably due to the proximity of the
mooring to the generation site(s).

Although both mixing events are linked to the tide, neither is as predictable as the tide
itself. Downslope flow mixing events exhibit a visual correspondence with inverse Richardson
number; however, at best the correlation is only 0.4. This may be due to limitations in our current
measurements, which only resolved shear over a small fraction of the overturning depth range.
Nonetheless, we believe that a traditional mixing parametrization based on tidal strain and shear
may be useful to explain downslope flow mixing.

If flow reversal mixing is caused by convective instabilities in the manner described by
Gemmrich and van Haren (2001), a mixing parametrization is more complicated. Predictability
would depend on the orientation of the internal tide relative to the topography, which in turn re-
quires detailed knowledge of how changes in the background stratification and currents affect the
generation and propagation of the internal tide. Small changes in the propagation azimuth of the
internal tide presumably would lead to significant changes in mixing.

In the context of the other HOME observations at Kaena Ridge, we find good agree-
ment between our inferred dissipation rates based on Thorpe scales and direct microstructure mea-
surements. Our estimated time-averaged dissipation is $1.2 \times 10^{-8} \text{W kg}^{-1}$, but can reach up to
$10^{-6} \text{W kg}^{-1}$. The corresponding time-averaged eddy diffusivity is $2 \times 10^{-3} \text{m}^2 \text{s}^{-1}$. For compari-
son, Levine and Boyd (2005) found comparable an average dissipation value of $2 \times 10^{-8} W kg^{-1}$ at the 1450 m isobath, based on a similar analysis and Klymak et al. (2005) found $4 \times 10^{-9} W kg^{-1}$ at 3000 m based on direct microstructure measurements. The microstructure measurements suggest that mixing rates are enhanced 100 to 200 m from the bottom (Klymak et al., 2005). Given the sporadic and event-like nature of the mixing, the value of continuous sampling over time for estimating mixing is highlighted by our observations. It is remarkable that the two methods of estimating mixing, microstructure profiles versus overturns from continuous temperature time series, give such consistent results.

7.2 Mixing at Mooring DN

We have argued that mixing along the flanks of the Kaena Ridge, at least at the mooring sites DS and DN, is dependent on the baroclinic tide as a source of mechanical energy. The internal tide takes the form of downward propagating beams at this depth in numerical model runs, and in observations close to the ridge crest. We infer that the orientation of the beam relative to the sloping bottom is a key consideration for mixing. In the case of DN, the physical separation of the beam from the slope appears to be an issue. The resulting $M_2$ tidal currents are 50% weaker at DN than at DS, where the beam appears to brush along the slope, and as a consequence mixing rates and dissipation are an order of magnitude smaller than at DS. We have identified both shear (downslope flow mixing) and convective (flow reversal mixing) instabilities as the triggers for mixing at DS. We do not find evidence for convectively driven mixing at DN, presumably due in part to the separation of the beam and slope. Tidal strain does appear to play a role in DN mixing events, consistent with a shear instability. We lack sufficient coverage of the current field with depth to make conclusive statements about the role of current shear. Whereas the maximum tidal strain occurred during the phase of strong downslope flows at DS, at DN the maximum strain occurs during maximum upward displacement of the isotherms when the flow reverses from up to downslope. We attribute the difference in phase of maximum strain between the two sites again to the details of the tidal beams. At DS, the strain phase is set by the kinematics of a downward propagating internal wave, which in turn will depend on the vertical wavenumber structure of the internal tide. At DN, we observe a decay of the tidal signal toward the boundary. This apparently causes the tidal strain to be set by the stronger tidal motions farther from the boundary; maximum strain occurs when the upper isotherms are displaced farthest from the boundary.
Returning to the issue of mixing parametrizations, the contrasts between DN and DS indicate the importance of specifying the altitude of the tidal beam above the topography. This in turn is likely to be sensitive to changes in stratification as the beam propagates downward from its generation region.

DN also differs from DS in the amount of near-inertial to diurnal band energy observed. The event-like behavior of this energy in the near-diurnal band suggests that this variability is not associated with locally generated diurnal internal tides. Moreover, diurnal internal waves generated near the ridge crest would have a shallow angle of downward propagation, making it unlikely that they would encounter the deep super-critical ridge flanks.

The smaller amount of near-inertial energy at DS is consistent with mooring DS being shadowed by the Kaena Ridge from the equatorward propagating internal near-inertial waves. These near-inertial waves appear to interact with the supercritical topography to produce vertical displacements near the bottom larger than for a freely propagating NIW. These vertical displacements near the bottom superpose with the displacements created by the semidiurnal beam further up to create periods of high strain, which presumably are prone to overturning and mixing under the action of a velocity shear. Here we observe energetic motions near the slope linked to incident, intermittent, near-inertial waves. The topographic slope around our observation site is supercritical for these waves. At other locations along the north side of the ridge, where the slope is critical for near-inertial to diurnal waves, mixing related to critical wave reflection is likely to play a role in the overall mixing budget of the ridge. On the south side of the ridge however, regardless of the slope, mixing linked to NIW is likely to be significantly lower due to sheltering effects.

7.3 Implications and Future Work

These results are not necessarily specific to Hawaii, or even ridge topographies. Internal tides are generated at many continental shelf breaks, and combined with supercritical or critical continental slopes, downgoing tidal beams are likely to interact with the topography. The relative importance of tidally-driven shear instability and convective mixing, as described by Gemmrich and van Haren (2001), remains to be determined. Given the large fraction of continental slope worldwide where downgoing tidal beams are potentially present, we speculate that these beams may account for a significant fraction of the mixing occurring at deep oceanic boundaries. This mechanism has received less attention than the critical reflections of incident internal tides. Furthermore, the exact position and orientation of the beam relative to the slope is essential to determine the associated
mixing, as was illustrated by the differences between the two moorings. How much of the beam is dissipated locally over the slope may be important for closing the overall baroclinic tidal energy budget at the ridge.

We note that our study site is below the main thermocline, within a depth range that has been a focus of abyssal mixing studies (Munk and Wunsch, 1998). The preliminary HOME results from Kaena Ridge encourage speculations that tidally-driven mixing mechanisms may differ above and below the main thermocline. The two moorings DS and DN were deployed at the same depths, on opposite sides of Kaena Ridge. The water mass at the two moorings is the same as shown by the TS diagrams and the time-averaged vertical temperature profiles. However, the stratification at the south mooring DS is weaker than at the north mooring DN (Figure 2.3 and section 2.2). Likewise, the estimated turbulent mixing at mooring DS is found to be one order of magnitude larger than at DN. This leads us to suggest that the more energetic mixing observed at mooring DS leads to a more well-mixed water column, and locally weaker stratification. Exchanges with the ocean interior may result in the export of this boundary mixed water along constant density surfaces (e.g. via stirring by mesoscale variability, Munk and Wunsch (1998)), and replenishment by water with the background ocean stratification. Clearly, the exchange rate with the open ocean is a crucial factor in determining how mixing modifies the properties of the fluid at the boundary. Our experimental setup was limited in space, and the lack of lateral coverage along a density surface did not allow us to investigate this rate of exchange.

The HOME experiment involved numerous platforms, instruments and investigators with the goal of understanding the overall tidal energy budget and the associated mixing around the Hawaiian Ridge, and at Kaena Ridge in particular. Preliminary results provide a detailed picture of the internal tide structure at and away from the ridge crest. However, this study identifies the need for additional work to gain a better understanding of the deep boundary mixing at the Kaena Ridge and elsewhere along the Hawaiian Ridge. The following tasks remain:

1. Observations over a greater depth range above the slope. Our current measurements did not extend far enough from the bottom to better assess the internal tide and NIW structure away from the bottom, in relation to our temperature observations. Current shears could not be reliably measured due to uncertainties regarding the effects of side lobe reflections. More reliable current and current shear information over a greater depth range will provide more details on the mechanisms that lead to energetic mixing.
2. Observations closer to the bottom. Due to mechanical constraints on the mooring, no instrument could be deployed below 22 mab. We therefore missed a significant fraction of the frictional bottom boundary layer.

3. Numerical models with improved resolution. The agreement between existing numerical models of the area and our observations is poor. This is likely due to the complex bathymetry that varies over scales much smaller than the model resolution. A better model resolution may help reconcile the model with the data, and allow the extrapolation of sparse temporal and spatial observations to the entire ridge.

4. Mixing associated with NIW. NIWs have been observed throughout the ocean. At our north flank site, we observe strong temporal variability of the NIWs, and mixing associated with the superposition of internal tides and NIWs. It is unclear how much mixing can be attributed to the NIWs alone. An experiment in a region subject to high NIW and limited internal tide energy (e.g., away from the shallow ridges identified as internal tide generation sites) may answer this question.

5. We could not positively confirm the link between surface wind generation to the north and NIWs events on the slope, nor could we calculate the modal content of these waves. Long term, full depth, monitoring of the current and density field at open ocean sites like station ALOHA could offer answers to these questions.


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