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Turbulent mixing induced by nonlinear internal waves in Mono Lake, California

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Abstract

Describing the spatial and temporal occurrence of turbulent mixing and the mechanisms of turbulence production are fundamental problems in physical limnology and oceanography. Here we present results from Mono Lake, California, that illustrate pronounced steepening of the pycnocline upon cessation of strong wind forcing and the enhanced turbulence resulting from this nonlinearity. Three events occurred with wind speed of >10 m s⁻¹. Upwind, the upper pycnocline upwelled, but on relaxation of the wind, it downwelled 6–10 m and compressed. The similarity of wave amplitude during compression to mixed layer depth and Lake numbers <4 were indicative of significant nonlinearity. Energy in the internal wave field increased by two orders of magnitude, but ~50% of the energy was lost within 24 h. The largest energy losses occurred at low frequencies rather than via an energy flux from low to moderate and high frequencies. Eddy diffusivities, computed over 6-d intervals using the heat budget method, averaged ~10⁻⁶ m² s⁻¹ in the energized pycnocline but were five to 10 times higher during the initial upwelling and downwelling periods. Microstructure casts, taken 1–2 d after the first event, revealed that the percentage of the pycnocline that was turbulent varied with bottom slope, being 80% turbulent where bottom slopes were highest. The threshold for turbulent transport was exceeded only 10% of the time and only where bottom slopes were highest.

Knowing when and where turbulent mixing occurs in the water column is essential for a wide diversity of limnological and oceanographic problems. For instance, nutrient fluxes through the pycnocline support new production in the upper water column (Lewis et al. 1986), and the meridional overturning circulation in the ocean depends on mixing occurring in the abyssal ocean (Munk 1966; Munk and Wunsch 1998). Linking enhanced turbulence to the internal wave field has been a major advance (Saggio and Imberger 1998; MacIntyre et al. 1999; Rudnick et al. 2003). It is now well established that considerable turbulence occurs near lateral boundaries. For instance, recent oceanographic and limnological studies have demonstrated enhanced turbulence near sloping boundaries and rough topography near seamounts and ocean ridges (Toole et al. 1994; Lueck and Mudge 1997; Polzin et al. 1997) and near the benthic boundary (Wuest et al. 1996; Goudsmit et al. 1997) and rough topography in lakes (MacIntyre et al. 1999).

Considerable effort is now being expended to link turbulence production with different types of nonlinear waves. Steepened internal waves with amplitude similar to the thermocline depth have long been noted as a striking feature of long, narrow lakes (Thorpe et al. 1972; Hunkins and Fliegel 1973; Farmer 1978). Thorpe et al. (1972) and Mortimer and Horn (1982) identified these waves as surges; they form after metalimnion compression and propagate as progressive waves across a lake. Their role in turbulence generation has been unclear. Hunkins and Fliegel (1973) noted isotherm deepening in conjunction with surges and attributed the deepening to intense mixing. Surges also occur in small lakes (Lemmin 1978; Weigand and Chamberlain 1987), but the Richardson numbers calculated from estimated current speeds have been above the threshold for inducing turbulence. Horn et al. (1986) found that the enhanced current speeds associated with surges lowered Richardson numbers to values close to critical. In response to strong wind forcing in a much larger lake, Lake Biwa, Saggio and Imberger (1998) and Boegman et al. (2003) noted the steepening of the basin-scale Kelvin wave, an order-of-magnitude increase in energy in the basin-scale waves, but a two–order-of-magnitude increase in energy at higher frequencies. The energy in the internal wave field rapidly dissipated, and they hypothesized that the higher frequency waves shoaled nearshore and contributed to enhanced mixing near the boundary. Lorke (2007) has demonstrated enhanced turbulence due to the breaking of high-frequency waves at the boundary. At present, however, field studies have not quantified the turbulence associated with large nonlinear internal waves.

Laboratory and modeling studies have demonstrated several types of internal waves that form after wind forcing and illustrate the energy transfers when waves become nonlinear (Michallet and Ivey 1999; Boegman et al. 2005a). After an impulsive wind, several different mechanisms may distort the internal wave field. The standing wave may reflect from the boundary with no change in shape or the thermocline may steepen and become a surge, after the wind ceases or after one or more reflections of the standing wave, with a train of solitons forming on its rear face (Horn et al. 2001; Boegman et al. 2005a). For strong tilting, supercritical waves known as internal bores or hydraulic jumps may form. The degree of thermocline tilting as a function of stratification, wind forcing, and basin geometry can be predicted from the Wedderburn number (W) (Monismith 1986; Imberger and Patterson 1990). For large W (W > 10), the thermocline does not tilt. For 1 < W < 10, partial upwellings occur; for W = 1, the pycnocline

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upwells, and for \( W \ll 1 \), the lake can fully mix. Horn et al. (2001) used theory and experiments to develop a phase space based on the inverse \( W \) and the ratio of mixed layer depth to basin depth to predict the type of internal waves and their degree of nonlinearity. Nonlinear steepening of the thermocline and formation of surges and solitons is expected when \( 1 < W < -3 \) and when the ratio of mixed layer depth to basin depth is \( < 0.5 \); internal bores or hydraulic jumps are likely as \( W \) drops below 1. When graphed in the phase space of Horn et al. (2001), the nonlinear waves described in the earlier field studies were surges. The surges may or may not be dissipative, but whether the solitons that form on their rear will shoal nearshore depends on bottom slope and wave type (Vlasenko and Hutter 2002; Boegman et al. 2005a). Based on the field studies showing elevated energy in the internal wave field at frequencies higher than that of basin-scale waves and based on laboratory studies showing the energy losses associated with solitary waves, a considerable fraction of the mixing nearshore is attributed to shoaling of waves with frequencies higher than basin-scale waves (Boegman et al. 2003). Other losses are assumed to come from shear as internal wave-induced currents interact with the bottom boundary (Lorke et al. 2002; Lemckert et al. 2004; Marti and Imberger 2006).

Despite these laboratory and field studies, our understanding of nonlinear waves and resulting mixing is incomplete. In particular, we do not know how the more complex stratification typically found in lakes affects the type of nonlinear waves and mixing dynamics. The laboratory experiments discussed above, with the exception of some in Monismith (1986), were done in two-layer flows. Similarly, surges have been noted in lakes primarily in fall (Thorpe 1974), when the metalimnion has thinned as a result of autumn cooling and when the thermal structure is primarily two layered. In lakes, the metalimnion can be extensive or multilayered. With respect to linear waves, not only can the first vertical wave mode be supported, as in two-layer flows, but higher vertical internal wave modes may be present (LaZerte 1980; Weigand and Chamberlain 1987) and their prevalence may change with changing stratification over a season (Vidal et al. 2007). As the metalimnion thickens, the transition from linear vertical mode 1 waves to higher vertical mode waves leads to the question of whether the modes of nonlinear waves would similarly shift from vertical mode 1 to a higher order.

Only a few field studies have addressed the changes in the internal wave field and turbulence in lakes with decreases in the \( W \) or related Lake number. The Lake number \( (L_N) \) is an integral form of \( W \) that takes into account basin morphometry (Imberger and Patterson 1990). Whether the two indices provide different information about the degree of tilting of the upper and lower metalimnion has not yet been settled (Stevens and Imberger 1996; MacIntyre et al. 1999). MacIntyre et al. (1999) found increased internal wave activity and turbulence that were several orders of magnitude higher inshore than offshore when \( L_N < 2 \). Internal waves increased in amplitude and were followed by a train of moderately high-frequency waves, similar to those observed by Lemmin (1978). In studies in a 1.5-km² kettle lake, MacIntyre et al. (2006; in press) showed that low-\( L_N \) events accompany the passage of fronts. The coefficient of eddy diffusivity \( (K_u) \) increases to order \( 10^{-6} \) when \( L_N < 3 \) and to \( 10^{-5} \text{m}^2 \text{s}^{-1} \) when \( L_N < 1 \). The metalimnion abruptly descends when \( L_N < 1 \), a behavior expected for hydraulic jumps. Despite the relation of \( K_u \) to \( L_N \) in those studies and others (Romero et al. 1998; MacIntyre and Romero 2000; Yeates and Imberger 2004), only a few field studies have illustrated the pathways of energy flux from the internal wave field to turbulence in lakes (Antenucci and Imberger 2001; Gomez-Giraldo et al. 2006; Lorke 2007).

Developing predictive models for the onset of turbulence and its intensity are major goals in physical limnology and oceanography. Success requires understanding the role of internal wave dynamics in turbulence production. The next step is to quantify the relation. When nonlinear waves are present, isotherms undergo abrupt changes in elevation, with some isotherms ascending or descending more rapidly than others, and, at times, the metalimnion can split (Boegman et al. 2003). The internal wave field is being strained. Changes in the effective rate of strain, \( dW/dz \), where \( w \) is vertical velocity computed from the rate of upward or downward movement of isotherms, have been linked to overturning in the internal wave field (Alford and Pinkel 2000). Further, they found reasonable agreement between an estimate of \( \varepsilon \) based on rate of strain squared, \( \varepsilon_{str} = 500 \langle (dW/dz)^2 \rangle \) (where braces indicate averaging) to estimates of \( \varepsilon \) based on shear and overturning. Thus, it is valuable to determine whether rate of strain is a useful metric for turbulence production in lakes.

The goal of this article is to describe the response of the internal wave field to three periods with strong wind forcing in Mono Lake, California, and to evaluate likely mechanisms for turbulence production. Our measurements were made in early spring, and the stratification in the upper water column transitioned from two-layered to nearly linear stratification, thus allowing us to see the shifts in the dominant vertical mode internal waves. Measurements included surface meteorology; time-series temperature measurements at a station where the pycnocline intercepted the boundary; frequent conductivity, temperature, depth (CTD) profiles; and, following the first wind event, temperature-gradient microstructure profiles. We calculate the vertical modes that can be supported by the density stratification and verify the ones present using spectral analysis, and we describe the nonlinearity of the wave field by the \( W \) and Lake numbers. Further, we calculate the potential energy in the internal wave field, rate of strain, and \( K_u \) using the heat budget method based on frequent CTD casts at multiple stations and from the spread of a conservative tracer, sulfur hexafluoride \((SF_6)\). Microstructure profiles illustrate the spatial distribution of turbulence during a period of moderate to low internal wave activity following the first wind event. These data allow us to illustrate and quantify internal wave dynamics leading to enhanced turbulence in lakes.

**Study site**

Mono Lake (38°N, 119°W; surface area, 150 km²; maximum depth, 45 m) is a terminal, saline lake located
in the Great Basin east of the Sierra Nevada. Descriptions of the lake’s mixing dynamics are provided in Romero and Melack (1996), MacIntyre et al. (1999), and MacIntyre and Jellison (2001). Descriptions of its biology and geochemistry are provided in Jellison and Melack (1993a,b) and Jellison et al. (1993). Large freshwater inputs due to above-average snowpack in the Sierra Nevada can lead to chemical and thermal stratification that persists throughout the year and even for several years with concomitant reduction in nutrient fluxes and primary production. In such years, mixing near sloping boundaries or topographic features may be a critical mechanism for nutrient supply to the euphotic zone and maintenance of the lake’s productivity. Strong winds occur in association with the passage of fronts and as a result of diurnal differences in the heating of local land masses. Long-term wind records indicate that the winter period with frontal systems with high winds persists into April and May. Winds typically exceed 10 m s⁻¹ ten to twelve times per month in April and May. Our study was conducted in April 1998 to capture the internal wave dynamics and turbulence associated with the strong wind forcing.

**Methods**

Meteorological and radiation measurements are described in MacIntyre et al. (1999). These include wind speed and direction, relative humidity and air temperature, rainfall, and photosynthetically available radiation measured at the southern shore of Paoha Island (Fig. 1). Attenuation of photosynthetically available radiation within the water column was measured with a submersible LiCor sensor (model LI-192S sensor, model LI-185 meter) at 0.5-m intervals.

Time series of temperatures between 5 m and ∼20 m were obtained with self-contained temperature loggers (WaDaRs, TSKA) deployed on a subsurface mooring just north of Sta. 5. Each WaDaR has an accuracy of 0.01°C, a resolution of 0.0014°C, and a time constant of 1–3 min. Spectral roll-off for the former would begin at 720 cycles per day (cpd) and for the latter at 240 cpd. We sampled at 1.2-min intervals. No spectral roll-off was evident at the high-frequency limit of our power spectra, 600 cpd. The units were calibrated before and after deployment against a platinum resistance thermometer with an accuracy of 0.006°C. The uppermost logger was located at 5.2 m, the second at 8.2 m, and subsequent loggers were 2 m apart to 20.2 m.

Microstructure profiling is described in MacIntyre et al. (1999). Microstructure sampling was designed to capture the effects of variations in bottom slope. Mono Lake’s bathymetry slopes steeply along the southwest shore and along the southern and eastern sides of Paoha Island (http://pubs.usgs.gov/mf/2002/2393/), with slopes ranging from 10% to less than 1%. We had two sampling locations near the island with bottom slopes of 4% and 5%. Our inshore site, Sta. 5, was located where bottom slopes were 3%. This site is in the lee of a promontory for cyclonically traveling disturbances, where bottom slopes were 10%. Our offshore site, Sta. 6, was located at 38 m in depth in the deepest basin of the lake, where the bottom was essentially flat (slope ∼ 0.1%). We also profiled opportunistically at Sta. 4 and Sta. 7 as well as at several unmarked stations. Locations for all microstructure casts as well as the profile dates are indicated in Fig. 1.

A CTD (Sea-Bird Electronics model Seacat SBE 19) was used to profile temperature and conductivity at several sampling stations around the basin (Fig. 1). Additional details on data processing are provided in MacIntyre et al. (1999).

**Tracer experiment**—Following a modified procedure developed by Maiss et al. (1994) and Wuest et al. (1996), SF₆ was injected into the seasonal thermocline at 9.5 m. Because we were working in the seasonal thermocline, gas exchange with the atmosphere was not important and SF₆ behaved conservatively. We suspended an enclosed gas equilibrator, which consisted of a 1-m–long Plexiglas cylinder (radius = 10 cm), at the injection depth from a buoy for a period of 5 h on day 106. During the injection, lake water from the mid-injector depth was pumped at a rate of about 7 L min⁻¹ through the injector. SF₆ (99.8% pure) was bubbled through a diffusion stone suspended within the cylinder and vented via a hose directly to the atmosphere. Laboratory tests using a 0.005-Pa partial pressure standard (50 ppbv) demonstrated that water leaving the equilibrator had SF₆ concentrations of >50% of the equilibrium value, indicating that approximately 20 g of tracer was injected. Approximately every 30 min during the injection CTD casts were made so that the temperature and specific conductivity could be evaluated at the injection depth.

Following the injection, water samples were collected in 60-mL biological oxygen demand (BOD) bottles using a series of eight or nine discrete BOD samplers that were attached to a single line at known intervals between 5 and 15 m. After collection, the BOD bottles were stored cold in the field to prevent degassing. Within a few hours of collection, a known volume of water (~30 mL) was extracted from the BOD bottles into 50-mL glass syringes. After adding a known volume of ultra–high-purity nitrogen gas (~20 mL), the syringes were shaken for 3 min to transfer the SF₆ from the water to the headspace. This gas was either injected into a 20-mL Vacutainer™ for storage and later analysis using the method described by Clark et al. (2004) or it was analyzed using the headspace method of Wanninkhof et al. (1987) at a temporary laboratory setup near the lake. The detection limit using these methods is ~0.05 pmol L⁻¹, and the precision, determined earlier through duplicate measurements, is ±3%.

The tracer was injected at 9.5-m depth at Sta. 4 on day 106 from 09:00 h until 14:00 h; temperature at depth varied from 6.25°C to 5.25°C. Lake surveys of the vertical distribution of tracer were completed 1, 2, 4, 6, 8, 12, and 17 d after injection at five to 11 stations. CTD profiles were obtained on the day of injection and on each sampling day, among others, at the same stations (Fig. 2).

The distribution of tracer in the vertical can be described by the following dispersion equation: \( \partial C / \partial t = K_z \partial^2 C / \partial z^2 \), where \( K_z \) is the eddy diffusivity in the vertical direction, \( z \).
and C is the SF$_6$ concentration. An estimate of K$_z$ can be obtained from the change in the second moment of the distribution with time, $2K_z = \epsilon \sigma^2 t$, where $\sigma^2$ is the variance of concentration at a particular time (Fisher et al. 1979). We assumed spreading was up or down from the depth of maximum concentration, calculated variance for each profile in which concentration exceeded the limits of detection, summed those variances, and then calculated $K_z$. Internal wave motions contributed to the variance in the distribution, particularly early in the study when mixing was low; thus, we only calculated $K_z$ from day 112 to day 118 (days 6–12 of the tracer experiment), when mixing had increased. Unfortunately, on day 118 concentrations of SF$_6$ at Sta. 5, our boundary station, were only just above the limits of detection and were nearly uniformly distributed in the vertical. We did not include the data from Sta. 5 in our analysis.

**Calculations**—Rates of dissipation of turbulent kinetic energy ($\epsilon$) were computed by a least-squares fit of the power spectral densities of the temperature-gradient signal to the Batchelor spectrum (Dillon and Caldwell 1980; Imberger and Ivey 1991). Segmenting was performed as in MacIntyre et al. (1999). Values of $K_z$ were calculated using the heat-flux method (Jassby and Powell 1975) from CTD profiles taken from six to 11 stations in the southern portion of the basin on average every 2–6 d during the experiment and from lakewide profiles obtained at the beginning and end of the experiment. Following Jellison and Melack (1993b), we used Sweers averaging to correct...
for differences in thermocline depth due to internal wave activity. Hypsographic data derived from detailed bathymetry were used to compute heat content and fluxes within different volumes of the lake. Ninety-five percent confidence limits were computed using a bootstrap approach (Dixon 1993). For depths below 6 m, at most 15% of the heating was due to direct solar radiation; the rest was due to turbulent mixing.

The temperature and conductivity data from the CTD casts were used to compute density, with an equation of state developed for Mono Lake water (Jellison et al. 1999), and to calculate buoyancy frequency, \( N = \left( -g/\rho \right) \delta \left( p/c \rho \delta z \right)^{1/2} \), where \( \rho \) is density, \( g \) is gravity, and \( z \) is depth. Isotherm depths were determined using linear interpolation between the depths of the WaDaRs. As a result of the presence of a temperature minimum at 15 m, we computed isotherm displacements above that depth. Power spectra of isotherms were computed using discrete fast Fourier transforms (fft.m in Matlab) following Bendat and Piersol (1986), with averaging in the frequency domain (cosine taper window) to preferentially filter higher frequencies and to improve statistical confidence. Power spectral densities were multiplied by the mean buoyancy frequency at depth so the results would be proportional to energy. The 5% and 95% confidence intervals are computed following Bendat and Piersol (1986, p. 524) and use a Chi distribution. As a result of the large depth fluctuations of isotherms and relatively infrequent CTD profiles, it was difficult to reliably describe a time series of \( N \) for the full water column. Hence, we limited ourselves to analysis of the 3–6°C isotherms. Flux of wind energy was calculated as \( P = \rho_a C_d U^3 \), where \( U \) is wind speed corrected for 10-m height, \( \rho_a \) is air density, and the drag coefficient, \( C_d \), equals 0.0013. Potential energy (PE) was calculated following Moum et al. (1992); \( PE = N^2 \left( \langle d^2 \rangle \right) \), where \( d \) is isotherm displacement and braces indicate averaging. The \( W \) and Lake numbers (Imberger and Patterson 1990) were computed as in MacIntyre et al. (1999). Wind speeds were filtered over 6-h periods, an interval which corresponds to one quarter of the period of the second vertical mode first horizontal (V2H1) internal wave. We calculated \( W \) for the upper and lower pycnoclines and \( L_N \) using Sweers averaged density profiles from days 106, 112, 114, 118, and 124 and the time series of wind. The length of the lake in the direction of prevailing winds was 13.5 km. \( L_N \) for the time series based on sequential density profiles only varied by 10%, so results are only presented based on the density structure on day 112; the time series of \( W \) for the lower pycnocline was the same for each profile. Effective rate of strain was calculated following Alford and Pinkel (2000); theoretical vertical modes were calculated following Gill (1982).

**Results**

Temperature, density structure, and meteorological conditions—Because Mono Lake was chemically stratified, winter mixing did not lead to holomixis (Fig. 3). Instead, temperatures in the upper 15 m in winter dropped to 2.5°C, with warmer water below. By the time of our study, heating had begun in the upper water column such that there was a stratified upper layer to ca. 10 m, which continued to warm during our study (Figs. 3, 4), a temperature minimum at 15 m, and bottom temperatures of 5°C. \( \sigma_t \) increased from 66 kg m\(^{-3} \) in the upper water column to 79 kg m\(^{-3} \) at depth. Initially pycnoclines occurred between ~10 and ~14 m as a result of the initial thermocline, and a third resulted from the chemocline between 18 and 22 m.

![Fig. 3. Profiles of temperature, \( \sigma_t \), and buoyancy frequency, \( N \), at Sta. 6 on day 112 (22 April 1998).](image)

As a result of the passage of fronts, three events occurred in which 10-min average winds persisted above 8 m s\(^{-1} \) for 1.5–3 d (Fig. 5A). Wind speeds and direction were not steady during these events. The strongest wind event occurred on day 113, when 10-min average wind speeds...
exceeded 14 m s$^{-1}$ and gusts exceeded 20 m s$^{-1}$. The upper pycnocline upwelled on initiation of each wind event (Fig. 5B). Cessation was followed by downwelling, with compression of the pycnocline. Compression was particularly noticeable for the second event. The 2-m descent of the 3$^\circ$C and 5$^\circ$C isotherms after the second wind event is indicative of strong mixing and is suggestive of conditions when a hydraulic jump is present. That is, the heat that had been introduced into the upper water column is rapidly mixed downwards. Each wind event was followed by several days with elevated internal wave activity. In the following section, we first illustrate the predicted vertical modes and use spectral analysis to assess whether they are present. Then we discuss subsections of the internal wave record in the context of wind forcing and internal wave dynamics.

**Theoretical modes**—Vertical and horizontal velocity structure and phase speeds are presented in Fig. 6 and Table 1. The predicted modal structure applies to standing waves (seiches). The first vertical mode will tilt along the chemocline, with horizontal velocities in opposite directions above and below. The higher vertical modes all have maxima in horizontal velocity in the upper water column. Shear associated with the second vertical mode wave will be highest across the thermocline and chemocline. The period of first mode and higher order vertical standing internal waves is given by $2L n^{-1} c_i^{-1}$, where $n$ is the number of horizontal nodes, $c_i$ is the phase speed of each mode $i$, and $L$ is the length of the lake at the depth where the amplitude of each mode is maximal. The predicted frequency of the waves depends on the axes on which they oscillate. As a result of shifts in wind direction, we compute frequencies along the longer northeast–southwest axis of the lake and along the shorter axis between Sta. 5 and the island and assume these provide upper and lower limits. Along the long and short axes, respectively, predicted frequencies of the first vertical first horizontal (V1H1) internal wave are 3–9 cpd, and frequencies of the V2H1 internal wave are 1–3 cpd (Table 1). Modes 3–6 ranged from 0.6 to 0.3 cpd along the long axis but from 2 to 1 cpd along the shorter northern axis. The internal Rossby radius, $R = c/f$, where $f$ is inertial frequency, is $\sim 8$ km for the first vertical mode wave and $\sim 3$ km for the second vertical mode wave. The Burger number, $S = R/R_L$, where $R_L$ is radius of the lake, is thus 1.3 and 0.5 and indicates that the lake will be weakly influenced by rotation (Antenucci and Imberger 2001). Using the geometries in MacIntyre et al. (1999), the frequency of a vertical mode 1 Kelvin wave would be 1.3–1.9 cpd and that of a vertical mode 2 Kelvin wave would be 0.46–0.65 cpd.

**Power spectral analysis**—Wind forcing occurred with peaks at frequencies centered on 1 cpd and 2.2 cpd, with
slight enhancement at frequencies from 4 to 8 cpd (Fig. 7A). Considerably more energy is concentrated at diurnal frequencies. Power spectral density of the 5°C isotherm shows two regions with enhancement, the first ranged from 1 to ~2 cpd and the second from 2.8 to 6 cpd (Fig. 7B). Enhancement occurred at similar frequencies for the 3.5°C isotherm (not shown). The first frequency range corresponds to that calculated above for the V2H1 internal wave along the long axis of the lake, a vertical mode 1 Kelvin wave, and to the first six vertical modes along the shorter northern axis. The second range corresponds with the V1H1 internal standing wave along the long axis of the lake. The correspondence between the periodicity of wind forcing and the V2H1 mode and the vertical mode 1 Kelvin waves indicates these waves would in part be wind forced. The overlap of enhancement for the wind and isotherm spectra between ~3 and 6 cpd indicates that wave motions with these frequencies would also be wind forced. The spectra fall off with a slope of ~2, similar to the Garrett and Munk spectrum and to the results of Saggio and Imberger (1998).

Internal wave dynamics—The internal waves were highly responsive to the magnitude of wind forcing and to shifts in direction (Figs. 8–10). High winds, predominantly from the west or southwest, led to upwelling (Fig. 8, day 101, 102; Fig. 9, day 113; Fig. 10, all days). Relaxation of the wind, or a shift to a more northerly direction, led to pronounced
pycnocline compression. Because the wind was gusty, each event had multiple low values of Lake and W numbers. The low Lake numbers were 3, 2.3, and 4 for the three wind events, respectively (Fig. 11). The lows for the W numbers were 0.4, 0.4, and 1 for the upper pycnocline and ~10, 6, and 10 for the lower pycnocline for the three wind events.

Prior to day 115, when the upper water column was two layered—that is, when the upper mixed layer was weakly stratified—the internal waves that followed thermocline compression were first vertical mode with movements of the upper and lower boundaries of the pycnocline in phase (Fig. 8; Fig. 9, day 114). The waves were often steep-sided, with higher frequency waves on the crests and troughs. Some of these changes were in response to shifts in wind magnitude or direction. After day 115, when the upper water column had become linearly stratified as a result of the combination of heating and wind mixing, second and higher vertical mode responses occurred throughout the pycnocline (Figs. 9, 10). For example, relaxation of the winds at day 115 was accompanied by isotherms above 10-m depth upwelling and those below 10-m downwelling. Several cycles of expansion and contraction occurred over the next 2 d. A third vertical mode was evident on day 115.8, with isotherms upwelling above 8 m, downwelling between 8 and 10 m, and upwelling between 12 and 14 m. A fourth vertical mode wave was evident on day 123.45, with isotherms upwelling above 7 m, downwelling below 9 m, upwelling above 12 m, and downwelling below 12 m. Second vertical mode responses also occurred with a vertical scale of 2–4 m. These localized expansions accompanied upwelling on days 113, 121, and 123 and following compression on day 114.

Both W and L indicate the waves produced by the high winds were nonlinear, although interpretation of these indices in Mono Lake, with its three-layered stratification, is unclear. Values of W for the upper water column (\(W_U\)) for the first two wind events are in the range predicted by Horn et al. (2001) for bore formation, whereas the third was on the boundary between that for bore formation and surge and soliton formation. In contrast, values of L were in the range indicative of surge and soliton formation. Given the uncertainty in characterizing wave types based on W and L, we compare the amplitude \(\zeta\) of the disturbance relative to the mixed layer depth, \(h_l\), as \(\zeta_{\text{max}}/h_l = 1/W\) (Horn et al. 2001). The thermistor chain was sufficiently upwind that the amplitude of the internal waves is close to maximal. Assuming the mixed layer depth was 10 m prior to day 118 and 6 m afterwards (Fig. 5), the ratio was on the order of 1 for all three events. Thus, the downwelling we observed could either have been the

Table 1. Phase speed, length of the lake along southwest axis (L) at depth of maximum amplitude, frequency, and wave period for the first six vertical mode first horizontal standing internal waves. Frequency and wave period are computed for southwest (SW) and northerly (N) axes. Northerly axis (L = 4 km) is between Sta. 5 and the island.*

<table>
<thead>
<tr>
<th>Mode</th>
<th>Phase speed (m s(^{-1}))</th>
<th>L and depth (km, m)</th>
<th>Frequency (SW) (cpd)</th>
<th>Wave period (SW) (d)</th>
<th>Frequency (N) (cpd)</th>
<th>Wave period (N) (d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.81</td>
<td>11.8, 17.5</td>
<td>3.0</td>
<td>0.34</td>
<td>8.7</td>
<td>0.11</td>
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<td>1.01</td>
<td>3.07</td>
<td>0.33</td>
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<td>1.76</td>
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<td>1.57</td>
<td>0.64</td>
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<tr>
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<td>0.11</td>
<td>12.6, 11</td>
<td>0.38</td>
<td>2.65</td>
<td>1.18</td>
<td>0.84</td>
</tr>
<tr>
<td>6</td>
<td>0.09</td>
<td>12.7, 8</td>
<td>0.30</td>
<td>3.33</td>
<td>0.95</td>
<td>1.05</td>
</tr>
</tbody>
</table>

* cpd, cycles per day.

Fig. 7. Spectral density and 95% confidence intervals (dotted lines) for days 102–124 of (A) wind speed showing diurnal and semidiurnal periodicity and (B) °C isotherm displacement. Vertical lines identify peaks in wind field.
Fig. 8. (A) North–south (thin line, N positive) and east–west (thick line, E positive) hourly averaged wind vectors and (B) 0.5°C isotherms for days 101–112. Upwelling is indicated with a U and downwelling with a D, with the duration of these events indicated by the length of the solid lines. The 4.5°C isotherm is in bold.

Fig. 9. Time series as in Fig. 8 but for days 112–117. The 4.0°C isotherm is in bold. C and E in the upper part of (B) indicate contractions and expansions, respectively, from second vertical mode motions extending throughout the upper water column. Embedded Es mark localized second vertical mode expansions. First arrow on x axis indicates nonlinear second vertical mode wave, which looks like a shock wave. Second arrow indicates third vertical mode internal wave.
signature of an internal bore or the nonlinear steepening
that leads to surge formation. Our observations are similar
to those at the upwind site in Mortimer and Horn (1982)
and Horn et al. (1986). They similarly noted thermocline
compression after cessation of the wind. Their cross-basin
thermistor arrays indicated that a surge subsequently
propagated across the lake. However, because pycnocline
compression in our study was greater for the events when
\( W_U \) was less than 1, and because pycnocline depression, as
occurred with the second event, is a sign of the intense
mixing associated with a hydraulic jump (Horn et al. 2001)
(as were the 2-m scale temperature inversions that occurred

Fig. 10. Time series as in Fig. 8 but for days 120–125. The 3.5°C and 8.5°C isotherms are in bold. Times when third and fourth
vertical mode internal waves occurred are indicated on the x axis.

![Graph](image1)

Fig. 11. Time series of Lake number, \( L_N \); Wedderburn number, \( W_U \), for upper pycnocline;
and Wedderburn number, \( W_L \), for lower pycnocline for days 100–125.
within it), it is likely that the first two events were internal bores. Regardless of which type of nonlinear wave was induced, they were likely to have propagated across the lake.

The internal waves that occurred within 4 d of the first low-L\textsubscript{N} event were first vertical mode and often steep-sided (Fig. 8), as were those that initially followed the second low-L\textsubscript{N} event (Fig. 9, day 114). Holloway (1987) noted similar features on the continental shelf and attributed them to hydraulic jumps. Thorpe (1974) noted similar features in Loch Ness and attributed them to the passage of a reflected surge. Appt et al. (2004) similarly found steep-sided waves due to the reflection of the surge in Lake Constance. Regardless of their exact nature, their form indicates they are nonlinear and a likely source of turbulence.

Second vertical mode internal waves formed after the second and third low-L\textsubscript{N} events, when wind forcing had decreased (Figs. 9, 10). These were also often steep-sided. The expansion on day 115.2 was so abrupt that it appears to have been a shock wave. Three cycles of expansion and contraction followed the second low-L\textsubscript{N} event. Their average frequency was 4.25 cpd, higher than the 3 cpd predicted for the second vertical mode response (Table 1). Thus, these mode 2 waves were likely progressive waves. The second mode compressions on days 121.2 and 123.4 occurred within 3 h of pycnocline downwelling. The periodicity of the second vertical mode wave on day 121.2 was \( \sim 8 \) h and may have been indicative of a V2H1 seiche on the lake’s northern axis. The cycles of expansion and contraction on day 123 were not sustained because of increased winds from the southwest. Overall, the abrupt formation and the steep-sided upwellings and downwellings that characterize these second vertical mode waves are indicative of a nonlinear response to the wind and, again, are a likely source of turbulence.

Changes in energy in the internal wave field in response to wind—The daily averaged wind energy flux was 0.4, 1, and 0.9 W m\(^{-2}\) for the three wind events (Fig. 12A). The 5\(^\circ\)C isotherm was band-passed filtered to illustrate the contribution of different frequencies (Fig. 12B). Amplitudes of waves in each frequency band increased with wind forcing. The deflections of the isotherm as a whole were largest, 10 m, for the wind event with L\textsubscript{N} = 2.3. Displacements for waves with frequencies between 0.72 and 2.4 cpd (corresponding to the first peak in wind forcing and of the V2H1 internal wave) were up to 8 m in amplitude and were larger than those with frequencies corresponding to the higher frequency wind events and wave motions (2.4–8 cpd, not shown).

The potential energy in the wave field increased with wind forcing and decreased over time (Fig. 12C). For the first two events, the potential energy at its maximum was 0.05–0.08 m\(^2\) s\(^{-2}\). The \( \sim 2 \) order-of-magnitude increase in potential energy in the wave field in the second event corresponded to a similar increase in energy flux from the wind. Potential energy decreased an order of magnitude within 1 d and another order of magnitude over the next 5 d for the first event. Potential energy decreased by a factor of 8 after 1 d and by a factor of 30 over 2 d for the second event. Potential energy increases and decreases in other frequency bands were proportional. Potential energy of waves at higher frequencies than illustrated here were several orders of magnitude lower. Results were similar for the 4.5\(^\circ\)C and 5.5\(^\circ\)C isotherm displacements (not shown).

Power spectra of isotherm displacements show the 100-fold increase in energy in the internal wave field induced by the high winds in the second low-L\textsubscript{N} event (Fig. 13). In contrast to results in Saggio and Imberger (1998) and in a second analysis by Boegman et al. (2003), the energy increased uniformly at all frequencies. The subsequent
Based on the heat changes from days 107 to 124 computed from the lakewide CTD casts (Fig. 1), the mean \( K_z \) was \( 7 \times 10^{-6} \text{ m}^2 \text{s}^{-1} \) in the upper pycnocline between 9 and 12 m. Logarithmically averaged \( K_z \) values at those depths were a factor of two lower than the average in the period of lower internal wave energy between days 106 and 112 but only slightly less than the average between days 112 and 118 and between days 118 and 124 (Fig. 4A,C,D). In contrast, from days 112 to 114, when the largest internal wave deflections occurred, logarithmically averaged \( K_z \) from 9 to 12 m was 4.7 times the average and was seven times higher than during the period with low internal wave activity. Where the water column was most stratified, 11–12 m, values of \( K_z \) increased five to 10-fold (Fig. 4A,B).

**Horizontal and vertical movement of SF\(_6\)**—For the first few days after SF\(_6\) injection on day 106, the tracer patch was found at only a few stations. On day 110, tracer was found at five of the 10 stations sampled (Fig. 2). The patch had moved southeast a maximum distance of 6 km. Horizontal advection was too slow to evenly mix the tracer throughout the basin at this time. Through day 112, the tracer was concentrated in layers that were 2–3 m thick and detectable between 7 and 13 m. The tracer patch was more extensive following the wind event. Tracer was found at every station sampled on days 114, 118, and 124 (not shown). After the storm, the maximum peak concentrations had decreased by a factor of 20. The greatest dispersion occurred near Sta. 5, where the tracer was nearly uniform in the vertical on day 114 and was further diluted by day 118 and found to 15 m in depth. At a near-boundary station on the west shore, the tracer was in a 5-m–thick layer on day 118. In contrast, at offshore stations sampled on day 118, most of the tracer was concentrated in 3–4 m–thick layers. On day 124, SF\(_6\) concentrations at two of the eight stations sampled were still layered. Based on the variance in distribution, without using the nearshore data from Sta. 5, where concentrations were near the limits of detection, average \( K_z \) in the pycnocline for days 112–118 was \( 2 \times 10^{-6} \text{ m}^2 \text{s}^{-1} \). This value, four times less than that computed for all sampling sites at depths from 7 to 13 m using the heat budget method (Fig. 4C), primarily describes mixing at offshore sites or at sites with smaller bottom slopes than Sta. 5.

**Temperature-gradient microstructure**—The frequency of turbulent events within the pycnocline varied with time after wind forcing and by location (Fig. 14). Approximately 80% of the pycnocline above the temperature inversion was turbulent at the island station, 60% at the inshore station, and less than 10% offshore within 2 d of the end of the first low-L\(_N\) event (days 104 and 105). In contrast, on days 110 and 111, about 30% of the water column was turbulent near the island, 15% inshore, and 5% offshore. The percentage of the water column that was turbulent at a given location was greater on days with higher potential energy in the internal wave field (Fig. 12C). Sites with greater bottom slopes or greater proximity to topographic features were more turbulent than those with smaller bottom slopes. The threshold for diapycnal transport decay does as well (not shown, but see Fig. 12C) and implies that turbulence is being induced by attributes of internal waves on the basin scale as well as on smaller scales. Because the energy in the larger scale waves was much larger than that of the moderate and higher frequency waves, the contribution of the low-frequency waves to turbulence production is greater. These results are quantitatively different from those observed in Lake Biwa (Saggio and Imberger 1998; Boegman et al. 2003), despite the similar amplitude of the disturbance relative to mixed layer depth (Boegman et al. 2005a). They observed a 1 order-of-magnitude increase in energy content at frequencies near 1 cpd and nearly a 2 order-of-magnitude increase at moderate to high frequencies in response to the high winds induced by a typhoon on Lake Biwa. Hence, while some of the turbulence production in Lake Biwa did occur at low frequencies, they infer a considerable transfer of energy from low- to moderate- and high-frequency waves and the shoaling of those waves nearshore. In our case, most of the energy dissipation occurred within the low-frequency, large-amplitude waves.

\( K_z \) _from spatially distributed CTD casts_—Eddy diffusivities in the upper water column at depths above 10 m and in the upper pycnocline below were high throughout the study period (Fig. 4). \( K_z \) exceeded \( 10^{-5} \text{ m}^2 \text{s}^{-1} \) at all times in the upper water column. Values were highest, \( 10^{-4} \text{ m}^2 \text{s}^{-1} \), from day 112 to day 114 when wind forcing was highest. As the stratification shifted at depths above 15 m from two layered to linear as a result of heating, \( K_z \) values showed a greater depth dependence. \( K_z \) values in the upper pycnocline were lower than in the upper mixed layer but always at least an order of magnitude higher than molecular rates.

**Fig. 13.** Spectral density \( \times N \) for 5°C isotherm displacements for days 110–112 and 113–115.
Turbulence and rate of strain—Depth-averaged rate of strain in this study was highest during the first two wind events, with highest values during upwelling and downwelling on day 103 (Fig. 16A). The decay in rate of strain mirrored the decline in wind energy flux and in potential energy of the internal wave field (Fig. 12A, C). In the manner of Alford and Pinkel (2000), we computed vertically integrated rate of dissipation of turbulent kinetic energy from rate of strain, \( \epsilon_{str} = \epsilon \times \exp(\text{mean}(\ln(\text{d}W/\text{d}z)^2)) \), where \( \epsilon \), to be dimensionally consistent, has units \( \text{m}^2 \text{s}^{-3} \). We averaged over 4-h periods. As our calculations are designed to indicate whether rate of strain is predictive of turbulence, we multiplied by a factor of 10 and we did not square a second time because, as will be seen below, doing so provides more reasonable estimates of \( K_z \) than does the algorithm used by Alford and Pinkel (2000). Using this ad hoc procedure, \( \epsilon_{str} \) had order \( 10^{-6} \text{ m}^2 \text{s}^{-3} \) during the high wind events and decayed to order \( 10^{-8} \text{ m}^2 \text{s}^{-3} \) as the wind tapered (Fig. 16B). In comparison to measured values during the period of moderate winds on days 105–108 and of low winds on days 108–112 (Figs. 14, 15), these values are somewhat high. However, when considering the likely log-normal distribution of \( \epsilon \) (Baker and Gibson 1987) and fivefold to 20-fold increases in \( \epsilon \) when corrected for log-normality (MacIntyre et al. 1999), these values are not unreasonable. We further followed Alford and Pinkel (2000) and computed the 4-h averaged coefficient of eddy diffusivity as \( K_{z, str} = \kappa_{\text{mix}} \epsilon_{str} \text{ N}^{-2} \) (Fig. 16C), where \( \kappa_{\text{mix}} \), the mixing efficiency, is assumed to be 0.2. When computed over the upper pycnocline (10 and 12 m), agreement is within a factor of 2 with our measurements of \( K_z \) from the heat budget method for those depths.

Discussion

This study demonstrates the critical importance of the initial period with high winds in terms of inducing nonlinearity in the internal wave field and turbulence in lakes. If wind forcing is strong relative to stratification and persistent, pronounced thermocline tilting occurs. On
Fig. 15. Profiles of temperature and $\varepsilon$ (gray histograms) at Sta. 6 and near the island on day 105 and at Sta. 5 on day 104.

Fig. 16. Time series of (A) absolute value of rate of strain, (B) rate of dissipation of turbulent kinetic energy estimated from rate of strain, $\varepsilon_{\text{str}}$, and (C) coefficient of eddy diffusivity estimated from rate of strain, $K_{z,\text{str}}$. Values are depth-averaged from 7.5 to 12 m over 4 h.
relaxation of the wind, the pycnocline at the upwind end of the lake rapidly downwells and compresses. Eddy diffusivities averaged over 6-d periods, including large-scale pycnocline deflections and subsequent enhanced internal wave activity, were $>10^{-5}$ m$^2$s$^{-1}$ in the upper water column above 10 m and $10^{-6}$ m$^2$s$^{-1}$ in the upper pycnocline ($10 < N < 40$ cph) below 10 m. Eddy diffusivities, averaged over a 2-d period when $W = 0.4$ and $L_N = 2.3$, were $10^{-4}$ m$^2$s$^{-1}$ above 10 m and $10^{-5}$ m$^2$s$^{-1}$ in the upper pycnocline. The fivefold to 10-fold increase in $K_z$ in the initial phase of upwelling and downwelling indicates that the initial period was the most important for generation of turbulence. Based on the magnitude of the pycnocline displacement relative to the mixed layer depth, values of the $W$ and Lake numbers, and comparisons with other studies, the pronounced pycnocline compressions are either the signatures of internal bores or the first stage of formation of a surge that will subsequently propagate across the lake. The enhanced rate of strain of the first two events and the fivefold to 10-fold increases in $K_z$ for the second event indicate that the first two events were more likely internal bores.

Several mechanisms have been proposed for turbulence production by wave breaking at boundaries with the current paradigm indicating an energy transfer from steepened basin-scale waves to higher frequency waves that break nearshore (Saggio and Imberger 1998; Horn et al. 2001). Much of the laboratory work supporting this paradigm has been done using $W$ numbers that are in the range in which surges and solitons are predicted, but oceanic studies of internal bores show solitons developing as a response to the nonlinearity and an appreciable energy loss associated with them (Holloway 1987). We did not observe enhancement of energy for any of the high-wind events in the frequency band at which solitary waves are expected (Boegman et al. 2003, 2005b). Note that the time response of our temperature loggers allows us to capture some of this frequency range but precludes observations to the buoyancy frequency (0.001 cpd). However, we found a similar lack of enhancement when we deployed loggers with a time constant of $<5$ s in later studies in Mono Lake in which the $W$ number also decreased to order 1 (E. McPhee-Shaw, W. J. Shaw, and S. MacIntyre unpubl.). Although wind forcing caused increases in energy in the internal wave field at all frequencies, the greatest decay of energy was observed at lower frequencies. Thus, the observations in this study indicate that turbulence production in the nearshore can be dominated by energy losses associated with the steepening of the large-amplitude waves generated by the wind, as opposed to breaking of moderate- to high-frequency waves. Further, the internal waves that degenerated into turbulence were forced waves, not standing waves.

The timescale for mixing across the pycnocline can be approximated as $\tau_{mix} = l/K_z$, where $l$ represents the thickness. The compression of the thermocline, which occurred during the low-$L_N$ events, reduced $l$ from ca. 10 m to 2 m or less. For the increased values of $K_z$, that were observed during high winds, $\tau_{mix}$ decreased from 100 d to 4 d or less. Our estimates of $K_z$ during the windy period include data from nearshore and offshore sites; assuming that $K_z$ in the nearshore is an order of magnitude higher (MacIntyre et al. 1999), the timescale for vertical flux in the nearshore is on the order of half a day. Thus, considerable vertical flux is possible in nearshore waters as a result of the initial compression of the pycnocline. Whether such vertical fluxes are possible offshore depends upon the amplitude and nonlinearity of the disturbance. Appt et al. (2004) have found smaller amplitude disturbances offshore. Our finding of layers of elevated concentration of introduced tracer offshore for up to 10 d after strong wind forcing attests to reduced vertical fluxes away from the boundaries.

Our calculations link rate of strain in the pycnocline with measured values of dissipation and eddy diffusivity. Rate of strain was most enhanced when strong wind forcing steepened the pycnocline and remained elevated when steep-sided vertical internal wave modes were present. The discrepancy between the coefficient used to compute $e$ in this study and that in Alford and Pinkel (2000) indicates that more effort is required to understand internal wave dynamics as they relate to rate of strain and dissipation. To first order, 2 order-of-magnitude increases in flux of wind energy, potential energy in the internal wave field, and in rate of strain correspond to order-of-magnitude increases in the coefficient of eddy diffusivity. Microstructure data taken on days 104–105, when rate of strain was enhanced in the pycnocline (Figs. 8, 16), show that (1) the thermal structure was steppy, particularly inshore (Fig. 15), (2) the thickness of steps and temperature gradients between steps changed between sequential microstructure casts (data not shown), and (3) that turbulence was enhanced both on the steps and between steps (Fig. 15). We infer that the turbulence was induced by the straining of the pycnocline by the steep-sided waves as they and the wind-induced pycnocline deflections differentially compressed and expanded the water column at different depths. Few of our measured values of $e$ exceeded the threshold for vertical transport within a few days of the first event with low $L_N$. Thus, internal wave straining may need to exceed a certain magnitude in order to induce vertical fluxes.

As heat was mixed downward in spring and as the upper pycnocline thickened, the internal waves supported transitioned from vertical mode 1 to vertical mode 2 and higher. This transition is not unexpected, since a thickened pycnocline supports higher internal wave modes (Gill 1982), and similar observations have been made in the coastal ocean in spring and attributed to an increase in wind energy supporting higher modes as stratification increased, and to a transfer of energy from mode 1 to mode 2 waves as a result of bottom drag (MacKinnon and Gregg 2005). However, our results indicate that the mode of nonlinear waves that formed after the initial disturbance changed as the pycnocline thickened. That is, we observed steep-sided vertical mode 1 waves after the low-$L_N$ events early in the study. Abrupt, steepened vertical mode 2 waves occurred after low-$L_N$ events when the pycnocline had thickened. These waves may be progressive waves resulting from the disturbance (Maxworthy et al. 1998). We can infer, based on our rate of strain calculations, that wherever the steep-sided waves occur, rates of dissipation of turbulent kinetic energy will be elevated. Thus, the
temporal and spatial distribution of turbulence within the pycnocline may change with seasonal thinning and thickening of the pycnocline, as it affects the modes of nonlinear internal waves in lakes.

Because the waves that formed immediately after the strong wind events were nonlinear, as were the waves that followed, transport of fluid from nearshore to offshore waters likely occurred (Maderich et al. 2001; Boegman et al. 2003; Appt et al. 2004). The localized 2–4-m expansions of the pycnocline may also contribute to transport of water from inshore to offshore waters (Lawson 2007). Whether the mixed fluid from the boundaries mixes with the ambient fluid offshore will depend on the turbulence induced as the nonlinear waves move offshore. Since the extent of mixing is larger inshore than off, these offshore transports and associated mixing may be a major supply route of nutrients for the biota offshore.

While our study emphasized mixing within the upper pycnocline, we also observed $\sigma_z$ to decrease from 79.5 kg m$^{-2}$ to 79.2 kg m$^{-2}$ below 22 m when $L_N$ decreased to 2.3 and $W$ for the chemocline dropped to 6. We infer that when wind forcing was sufficient to reduce $L_N$ to $\sim 2$ for the lake as a whole, nonlinear steepening of the deeper pycnocline occurred and induced significant vertical fluxes.

Spatial and temporal variability in mixing—Determining when turbulent events induce diapycnal transport is a critical problem in physical limnology and oceanography. Results from this study indicate that eddy diffusivities in the nearshore increase by nearly an order of magnitude when the pycnocline at the upwind end of the lake rapidly downwells on the cessation of strong wind forcing. As a result of the nonlinearity of the basin-scale wave, potential energy in the internal wave field decreases rapidly and rate of strain reaches maximal values. Thus, vertical fluxes of solutes and particulates occur rapidly over short periods of time. Comparisons of dissipation rates with the threshold for diapycnal transport showed that within a few days of high winds, few turbulent events were sufficiently energetic to induce vertical flux. Persistence of layers of SF$_X$ offshore over a $\sim$10-d period that included considerable wind forcing provides evidence for spatial variations in diapycnal mixing and indicates that thin layers of elevated biomass of bacteria, phytoplankton, and zooplankton, as observed in tidal fjords (McManus et al. 2003), have the potential to be persistent features in environments with well-developed internal wave fields. Slow horizontal dispersion further allows time for development of different biological communities.

Further insight into turbulence production will be obtained by a comparative approach, be it by modeling (Vlasenko et al. 2005), laboratory studies (Horn et al. 2001; Boegman et al. 2005a), or field studies (Gomez-Giraldo et al. 2006) that address the role of changing water depth and density stratification on modal structure and rates of internal wave steepening. Further work is warranted to evaluate the utility in using rate of strain to compute $\varepsilon$ and $K_Z$. Of critical importance for fluxes is the fact that unless upwelling reaches the surface, the timescale for vertical transport across the metalimnion is $\tau_{mix} = \varepsilon/K_Z$. Pronounced pycnocline compression, as observed here, enables rapid fluxes of heat, solutes, and particles across the metalimnion in nearshore waters. Timescales of mixing offshore will depend on the attributes of the nonlinear waves as they propagate offshore.

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