Space–Time Scales of Shear in the North Pacific

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ABSTRACT

The spatial, temporal, and directional characteristics of shear are examined in the upper 1400 m of the North Pacific during late spring with an array of five profiling moorings deployed from 25° to 37°N (1330 km) and simultaneous shipboard transects past them. The array extended from a regime of moderate wind generation at the north to south of the critical latitude 28.8°N, where parametric subharmonic instability (PSI) can transfer energy from semidiurnal tides to near-inertial motions. Analyses are done in an isopycnal-following frame to minimize contamination by Doppler shifting. Approximately 60% of RMS shear at vertical scales >20 m (and 80% for vertical scales >80 m) is contained in near-inertial motions. An inertial back-rotation technique is used to index shipboard observations to a common time and to compute integral time scales of the shear layers. Persistence times are $O(7)$ days at most moorings but $O(25)$ days at the critical latitude. Simultaneous shipboard transects show that these shear layers can have lateral scales $\geq 100$ km. Layers tend to slope downward toward the equator north of the critical latitude and are more flat to its south. Phase between shear and strain is used to infer lateral propagation direction. Upgoing waves are everywhere laterally isotropic. Downgoing waves propagate predominantly equatorward north and south of the critical latitude but are isotropic near it. Broadly, results are consistent with wind generation north of the critical latitude and PSI near it—and suggest a more persistent and laterally coherent near-inertial wave field than previously thought.

1. Introduction

Shear instability is thought to drive most turbulence in the open-ocean thermocline, making characterization of the sources, scales, and variability of ocean shear an important goal of physical oceanography. Additionally, shear strongly modulates the propagation of internal gravity and Rossby waves through the Taylor–Goldstein family of equations. Away from major current systems and fronts, internal gravity waves give rise to most shear in the ocean, motivating a huge body of literature in the 1970s and 1980s to characterize the internal wave field with an empirical spectral description known as GM76 (Garrett and Munk 1972, 1975; Cairns and Williams 1976; Munk 1981) and to develop the theory (McComas and Müller 1981; Henyey et al. 1986) behind the means by which energy is transferred throughout the internal wave spectrum toward smaller shear-containing vertical scales. Taking a stochastic and isotropic view of the wave field, this work resulted in some key scalings that now give the basis for the so-called finescale parameterizations (Gregg 1989; Polzin et al. 1995; Gregg et al. 2003) by which turbulence can be estimated from measurements of the low-wavenumber aspects of the spectrum.

More recent work has focused on the peaks in the internal wave frequency spectrum, primarily the internal tides and near-inertial waves (NIW), which reside near the inertial frequency $f$. These motions are more deterministic and directional than the rest of the internal wave continuum and are thought to provide the energy for the continuum via nonlinear wave–wave interactions (Polzin and Lvov 2011). Though the GM76 spectrum contains an inertial peak, its variability and different wavenumber content than the rest of the spectrum are...
not captured (Garrett 2001). Because of their bluer wavenumber spectrum compared to internal tides (D’Asaro and Perkins 1984; Garrett 2001; Alford 2010), NIW contribute much more to shear and in fact dominate shear at most times and locations (though this has not been documented systematically). The question then becomes what leads to near-inertial shear in the ocean?

Here, we report rare simultaneous spatial transects and time series from an array of profiling moorings from the Internal Waves Across the Pacific Experiment (IWAP; Alford et al. 2007), in which we investigate the percentage of the total shear contained in NIW, the fraction going up versus downward, their lateral scales, persistence times, and propagation directions, with the goal of better understanding their behavior over a wide latitude range. The long-term goal is to better understand how and where they are generated, where they propagate, and the ultimate fate and effects of the energy they carry.

One of the most well-known generation mechanisms of NIW is by wind forcing at the ocean surface. Wind-forced NIW arise when traveling storms set up inertially rotating motions in the mixed layer (D’Asaro 1985). Spatial variability in the wind, mesoscale flows (Weller 1982), and the β effect (D’Asaro 1989) subsequently creates sufficiently large lateral gradients in mixed layer flows to pump slightly superinertial motions below the mixed layer. The modal distribution of these motions depends on the stratification during the storm, but much of the energy is generally in low modes (D’Asaro and Perkins 1984; Alford 2010), which can propagate large distances toward the equator (Nagasawa et al. 2000; Alford 2003a; Furuichi et al. 2005; Simmons and Alford 2012). [Motions can equivalently be described in terms of rays (Kroll 1975; Zervakis and Levine 1995; Garrett 2001)]. Because these low modes travel quickly, this energy is presumed to dissipate far from the storm that generated it, though the exact fate is unknown. The generation and evolution of these low modes was reasonably well characterized in the groundbreaking Ocean Storms Experiment (D’Asaro et al. 1995).

The slowly moving high modes (λc ≈ 50–300 m) that are left behind dominate the shear and are our focus here. Their evolution was poorly described by linear theory (D’Asaro 1995). They have since been observed many times propagating downward following winter storms (Hebert and Moun 1994; Alford and Gregg 2001; Alford et al. 2012). However, upward-propagating, near-inertial waves have been observed at a variety of depths (Alford 2010) and in general downgoing, near-inertial energy exceeds upgoing energy only little (Waterhouse et al. 2016), as found here. Since high-mode waves have a slow vertical group velocity and therefore would take many months to reach the deep sea and reflect back upward, these puzzling upgoing motions likely have other sources.

One such source of NIW is from parametric subharmonic instability (PSI) of the internal tide, which generates two waves near half the frequency (Müller et al. 1986). Importantly, one is traveling upward and the other downward, possibly providing an explanation for upgoing waves. For semidiurnal tides, this is possible at and equatorward of the critical latitude, λc = 28.8°, and expected to be more efficient at the critical latitude where the inertial frequency is exactly half the M2 tidal frequency (MacKinnon and Winters 2005; Simmons 2008; Carter and Gregg 2006; Alford et al. 2007). The observations reported on here span λc and were collected in part to observe this process. As documented in Alford et al. (2007) and MacKinnon et al. (2013a), shear increases sharply at and south of the critical latitude. At the same location, upgoing and downgoing shear become more equal, consistent with this hypothesis. MacKinnon et al. (2013b) demonstrated a statistically significant bispectrum between the internal tides and the near-inertial waves, proving the mechanism was active [however, see caveats by Chou et al. (2014)], though calculated transfer rates were not as great as predicted by model simulations by MacKinnon and Winters (2005), owing to multiple vertical modes, temporal and spatial incoherence, and other factors in the real ocean (Young et al. 2008; Hazewinkel and Winters 2011).

Several other mechanisms known to generate near-inertial waves are briefly mentioned here, which could complicate this simple two-mechanism view [see Alford et al. (2016) for a recent review]. Equatorward- and downward-propagating near-inertial waves have been observed in the region emanating from the Subtropical Front near 30°N (Alford et al. 2013), evidently generated through wave–mean current interactions at the front in like manner to that observed and modeled by Nagai et al. (2015). Geostrophic flow over bottom topography gives rise to lee waves, which lead to upward-propagating, near-inertial waves in 2D numerical simulations (Nikurashin and Ferrari 2010). Finally, wave–wave interactions among poleward-propagating waves can funnel energy toward the near-inertial peak (Fu 1981).

As a step toward the long-term aim of determining the degree and spatial distribution of mixing by NIW, we here report observations of NIW during the IWAP experiment (Alford et al. 2007). The experiment was designed to observe the propagation of internal tides northward from Hawaii but also provided an opportunity to study and document numerous NIW over a large
depth range (80–1400 m). Crucial aspects of the measurements are 1) simultaneous shipboard transects and moored profiler time series, allowing characterization of spatial structure and temporal evolution; 2) ability to interpolate shear onto isopycnals to minimize Doppler shifting by internal wave displacements; and 3) simultaneous shear and strain measurements allowing calculation of individual wave properties including lateral propagation directions following Winkel (1998).

In the following, we first outline the data and the experiment’s oceanographic context, including description of low-mode, near-inertial fluxes (section 2). Moored shear and strain are presented in section 3, and band-passing and Fourier filtering is used to determine the near-inertial portion of total shear, and the ratio of up-going to downgoing shear. Back rotation is used to compute the time scales of modulation of the near-inertial shear layers. Wavenumber frequency spectra of shear are presented next in isopycnal-following coordinates. In section 4, shipboard transects are presented showing the horizontal spatial structure of near-inertial waves. Their long lateral coherence scales are quantified with a feature-tracking algorithm. In section 5, we study the characteristics of upgoing and downgoing, high-mode NIW using a plane wave fitting method and estimate their lateral propagation directions from the phase between shear and strain (Winkel 1998). We end with a discussion and conclusions.

2. Data and oceanographic setting

a. Experiment

IWAP was designed to examine the northward progression of internal tide energy from its generation region near French Frigate Shoals, Hawaii, past the critical latitude and beyond. Besides the Hawaiian Ridge, the bathymetry in the area includes broad areas of rough bottom topography, notably the Musicians Seamounts (Fig. 1), but is relatively smooth north of 32°N. The experiment took place in two cruises spanning about 62 days from April to June 2006. The general approach was to obtain simultaneous spatial and temporal information by steaming along a main line with the Hydrographic Doppler Sonar System (HDSS; Pinkel 2012), a specialized, high-power Doppler sonar, while six moorings spaced along the line [McLane moored profiler (MP); MP1–MP6] profiled to ascertain temporal information. MP5 malfunctioned and is not included in this analysis. Analyses of internal tides during IWAP are reported in Alford et al. (2007), Zhao et al. (2010), and MacKinnon et al. (2013a,b); this paper focuses on the near-inertial signals.

b. Moored array

Six moorings of similar design were deployed spanning latitudes 25°–37°N, which are described in detail in Zhao et al. (2010). The primary measurement on each mooring was a McLane moored profiler, which climbed up and down through the water column along the mooring wire between 85 and 1400 m, completing one up or down profile each 1.5 h. This sampling pattern leads to a variable temporal resolution ranging from 3 h at 85- and 1400-m depth to 1.5 h at middepths. This is easily sufficient to resolve near-inertial motions, which have a period ranging from 20 to 28 h over the latitude range of the moorings, but still aliases the highest internal wave frequencies, as will be seen in the spectra. Each MP carried a Falmouth Scientific Instruments (FSI) CTD with vertical resolution of 2 m and an FSI acoustic current meter with vertical resolution of about 0.01 m s⁻¹ (Doherty et al. 1999; Silverthorne and Toole 2009; Alford 2010). Additional instrumentation on the moorings included a 300-KHz Teledyne RD Instruments (TRDI) ADCP at
50 m looking upward toward the surface and an Aanderaa RCM8 current meter and a Sea-Bird temperature logger at 3000 m. MP3 also had a TRDI 75-KHz ADCP looking downward from 50 m.

Data were corrected for mooring knockdown (<20 m at all sites, owing to generally weak currents and high wire tension) and interpolated onto a uniform 2-m, 1.5-h grid (Fig. 2). All records are at least 40 days long with the exception of MP4, which stopped profiling early; all told, the profilers collected 3184 profiles. Full-depth data are used to compute low-mode energy fluxes, as described in appendix A and presented briefly in section 2d; the remainder of the paper focuses on the upper 1400 m resolved by the ADCPs and MPs.

c. Hydrographic Doppler sonar system

Spatial transects of velocity and shear ($\partial u / \partial z$) in the upper 1000 m were measured along the mooring line (Fig. 1, dashed) with the Revelle’s HDSS (Pinkel 2012). HDSS consists of a 140- and a 50-KHz Doppler sonar reaching depths of approximately 250 and 1000 m, respectively. Velocity from the two sonars was merged and gridded in 8-m, 5-min bins (1.2-km horizontal resolution at a ship’s speed of 4 m/s$^{-1}$). Agreement between MP
and HDSS data is excellent, as demonstrated by a comparison during a close approach to MP3 (Fig. 3).

d. Oceanographic background

Mixed layer depth was shallow as typical of the region for this period, about 20–25 m toward the north and even shallower to the south (Klymak et al. 2015), with a seasonal thermocline below. Background currents (Fig. 2) were generally weak (<0.2 m s⁻¹), as expected given the region’s generally weak eddy kinetic energy (e.g., Rainville and Pinkel 2006b). Eastward flow normally seen around 30°N in association with the subtropical front is not clearly evident in the moored data, apparently because the front was closer to 32°N during the time of our measurements (Klymak et al. 2015).

Wind forcing during the experiment (not shown) was generally typical of late spring conditions, with generally light and variable winds with the exception of a storm on yearday 118. The work on inertial motions in the mixed layer, an estimate of the power available for wind-forced, near-inertial waves, is computed by forcing the Pollard–Millard slab model (Pollard and Millard 1970) with 6-h NCEP reanalysis winds (Kalnay et al. 1996) following Alford (2003b). Time-mean wind work during the experiment (Fig. 1) was weak compared with wintertime values and increased poleward approaching the storm track.

To contextualize the high-mode motions that are this paper’s focus, we compute low-mode, near-inertial energy flux at each mooring and present it here first while giving details of the calculation in appendix A. Time-mean profiles of near-inertial energy flux are shown below in Fig. A1, and depth-integrated values are plotted in Fig. 1. The low-mode (modes 1 and 2), near-inertial energy flux is northeast at MP1–MP3 and south at MP4 and MP6. Note that the diurnal and inertial frequencies are too close at these latitudes to separate with the bandpass filter. The northeast flux at MP1–MP3 is therefore likely dominated by the D1 internal tide propagating northward from Hawaii.

MP4 and MP6 are north of the D1 turning latitude, and the observed southward flux is likely near-inertial energy propagating equatorward. A rough comparison of the wind work to the observed low-mode flux at MP6 can be made with a simple box model following Alford (2003a). Taking a box from 36° to 40°N and assuming all wind work goes into equatorward-propagating waves results in an estimated depth-integrated flux of 270 W m⁻¹. The observed mean equatorward flux at MP6 was 130 W m⁻¹, 48% of the wind estimate. These percentages are approximately consistent with estimates from Alford (2003a) and from the Ocean Storms Experiment (D’Asaro et al. 1995). The wind-generated portion of the shear observed here can be thought of as what is left behind after these energetic low modes have left the region.

3. Temporal and vertical scales

a. Depth–time maps of shear and strain

Shear is computed by first differencing the velocity records at all moorings and smoothing over 10-, 20-, and 80-m vertical intervals. The 20 m interval is used for analyses unless otherwise noted. Total zonal shear (Fig. 4, left) is dominated at all moorings by inertially
oscillating rather than mean shear. Broadly, a tendency for stronger shear at (MP3) and south of the critical latitude (MP2) is seen, as stated in the introduction. Shear is visibly greatest up shallow, a consequence of WKB scaling (cf. Leaman and Sanford 1975). Superimposed on a continuum of motions, a variety of coherent upward- and downward-propagating features (down and upward phase propagation with time, respectively) is evident. Shear at MP3 is stronger and qualitatively different, with large regions of checkerboard shear that do not show clear upward or downward phase propagation. Alford et al. (2007) and MacKinnon et al. (2013b) interpreted these as evidence of PSI occurring near the critical latitude.

Though our focus is the shear field, the strain field will be required for our estimates below of lateral propagation direction and so is presented in Fig. 5 in parallel format with the raw strain on the left and the near-inertial filtered quantity at right. Strain is computed in like manner to shear as the vertical derivative of isopycnal displacement with subsequent smoothing with a 20-m vertical boxcar window. Strain magnitude does not decrease with depth as does shear, consistent with a $N^0$ scaling expected from WKB compared to $N^2$ or $N^3$ for
shear (e.g., Alford and Pinkel 2000). Near-inertial strain is a smaller percentage of the total than is shear, as expected from linear theory, which predicts a peak at \( f \) for shear and a null for strain, and because of its greater contribution from vortical motions (Pinkel 2014).

b. Persistence of near-inertial shear layers

A complex demodulation technique known as inertial back rotation (D’Asaro et al. 1995) is used next to remove the sense of inertial rotation in time. Backrotated shear has two advantages: first, it allows examination of the persistence or equivalently the time scales of modulation of the envelopes of near-inertial packets; the second is in interpretation of shipboard transects that take large fractions of an inertial period to conduct. Backrotated shear references measured shear to a common time, giving a closer approximation of a spatial snapshot.

If the measured shear field (written in complex notation \( U_z = u_z + iv_z \)) has a time and space structure given by

\[
U_z(x, y, z, t) = U_z(x, y, z)e^{-i\Omega t},
\]

then multiplication by \( e^{i\beta} \) gives the backrotated field \( U_z(x, y, z) \). If near-inertial waves dominate the shear

![Fig. 5. (left) Total and (right) near-inertial strain at IWAP moorings.](image-url)
field, then $\hat{U}_z(x, y, z)$ gives their slowly varying amplitude. Shear at other frequencies than the back-rotation frequency gives errors; for example, $\hat{U}_z(x, y, z)$ would beat with a time scale $\Delta \omega^{-1}$, where $\Delta \omega$ is the difference between the true and assumed frequencies. Hence, long persistence times in $\hat{U}_z(x, y, z)$ indicate both a dominance of near-inertial shear as well as a prolonged presence of a particular wave packet.

Applying the back-rotation operator at each mooring (Fig. 6, left) removes the near-inertial carrier wave, giving instead the envelope of the near-inertially rotating wave packets. The layers are advected vertically by the isopycnal displacements (black) and are much flatter when viewed in an isopycnal-following or semi-Lagrangian reference frame (right), which is accomplished by first computing a set of reference isopycnals with 1-m mean spacing and interpolating shear profiles onto them at each time (Pinkel 1985). Phase information is lost; for example, crests are no longer seen propagating upward or downward in time, allowing inference of vertical propagation direction, but wave duration is seen more clearly.

A persistence time of $O(5)$ days can be seen at most moorings. MP3 again stands out as qualitatively different than the other locations, with highly persistent features lasting $>20$ days. This tendency is quantified by calculating the autocovariance of isopycnal shear, for which examples at MP3 and MP6 are shown in Fig. 7. An integral time scale $\tau$ can then be computed as the first zero crossing, which gives $\tau = 25$ days for MP3 and 5 days for MP6, supporting visual conclusions from the maps. This calculation is repeated at each mooring and at all depths. Profiles of $\tau$ (Fig. 8) are $O(3–6)$ days for most moorings, increasing sharply to about 10–15 days at MP3.

These analyses suggest that NIW dominate shear in our data, since backrotated shear would not give long persistence times otherwise. To quantify, the near-inertial components of the shear field are isolated by means of bandpass filtering. Because vertical Doppler shifting of the shear layers by internal waves of other frequencies (primarily internal tides) transfers variance from the inertial peak to sum and difference frequencies (Pinkel et al. 1987; Alford 2001a; Pinkel 2008), it is important to do the filtering in the isopycnal-following frame. At each mooring, a fourth-order Butterworth filter with passband $(1 \pm 0.3)f$ is run over the time series for each isopycnal forward and backward, and the result is plotted to right in Fig. 4. The reduction of the frequency bandwidth of near-inertial shear due to the filter is apparent. However, also notable is a reduction in the vertical wavenumber bandwidth; that is, the temporal filtering has also removed the high-wavenumber signals, which would not be expected if the spectrum were separable in wavenumber and frequency as posited in the Garrett–Munk (GM) spectrum (Garrett and Munk 1979; Cairns and Williams 1976). The nonseparability of the observed internal wave spectrum has already been pointed out by Pinkel (1985) and will be examined in more detail below.

c. Fractions of up, down, and near-inertial shear

The up- and downgoing waves can be examined separately at each mooring by two means, which give nearly identical results in our dataset. Following Leaman and Sanford (1975), single profiles of complex shear $\hat{U}_c$ can be separated into components rotating counterclockwise (CCW) and clockwise (CW) with depth, which are up- and downgoing waves for a linear wave field. Alternately, the depth–time series of $\hat{U}_z$ can be 2D Fourier transformed and the four quadrants of the transform interpreted as up/down phase propagation and CW/CCW rotation in time. We employ the latter technique as we will be examining the 2D wavenumber/frequency spectra next. As an example, the zonal shear field at MP1 (Fig. 9, top) is decomposed into the constituents with upward (middle) and downward energy propagation (lower). While some regions are clearly a superposition of up- and downgoing waves, at other times and depths the separation filter successfully isolates packets of waves that appear to be propagating downward and upward. Their lateral propagation direction will be examined in section 5.

Conventional wisdom is that downgoing energy exceeds upgoing energy in the upper ocean as observed by Leaman and Sanford (1975) and Alford et al. (2012), consistent with the dominance of wind work at the surface as the forcing mechanism. This assumption is examined here by repeating the same separation as just done at MP1 on both the total and near-inertially filtered fields for all moorings and averaging in time (Fig. 10, left). Total shear (light gray) and the upward and downward components of near-inertial shear (darker grays) decrease with $N$ (black line) at all locations. The fraction of downgoing near-inertial shear is plotted in the middle columns of 10. A slight excess of downgoing shear is indeed seen at MP4 and MP6 as also found by Alford et al. (2007) and MacKinnon et al. (2013b), but the fractional excess is small compared with, for example, Ocean Station Papa (Alford et al. 2012), which is at much higher latitude and experienced much stronger wind forcing. There, down energy exceeded up energy by factors of 3–7 in the winter and 2–3 in the summer.
The total 20-m shear in near-inertial motions, given by the border between dark and light gray, accounts for approximately 60% of the total shear (Fig. 10, right). The fraction is greater, closer to 75%, at MP3, possibly suggestive of the generation of excess near-inertial shear by PSI. Because shear extends to smaller vertical scales than our measurements resolve, the ratio is somewhat sensitive to the amount of vertical smoothing. As will be clear in the spectra shown next, near-inertial motions do not contain very small vertical scales in our data, such that the ratio of near-inertial shear to the total decreases to about 50% when shear is differenced over a 10-m interval and increases to over 80% when an 80-m interval is used. Stated another way, larger scales are more dominated by near-inertial motions than the smallest scales.

d. Spectra

Extending the wavenumber spectra presented in MacKinnon et al. (2013a), we next document the vertical wavenumber and frequency characteristics of the shear measured at each mooring by means of wavenumber frequency spectra. Velocity was first WKB scaled (Leaman and Sanford 1975) as

![Fig. 6. (left) Depth-time maps of the real part of the backrotated shear \( u_{\phi} \) at all moorings. Isopycnals with a mean spacing of 30 m are superimposed in gray. (right) The \( u_{\phi} \) is plotted in an isopycnal-following frame.](image-url)
where \( N_o = 0.0042 \text{s}^{-1} \) is a reference buoyancy frequency and \( \overline{N}(z) \) is the time-mean stratification profile at each mooring (Fig. 9), and then placed into the isopycnal-following frame by interpolating onto isopycnals. Complex isopycnal-frame velocity \( \tilde{w} = u_{\text{wkb}} + i\tilde{v}_{\text{wkb}} \) is then demeaned and detrended in time and depth over the regions plotted in Fig. 4, and the 2D Fourier transform \( \tilde{w} \) is computed. The wavenumber frequency spectrum is then given as

\[
\Phi_v(\omega, m) = \left| \frac{\tilde{w} \tilde{w}^*}{\Delta \omega \Delta m} \right|, 
\]

where \( \Delta \omega = T^{-1} \) and \( \Delta m = Z^{-1} \) are the frequency and wavenumber bandwidths, respectively; \( T \) and \( Z \) are the length and vertical aperture of the records. The aperture \( Z = 1100 \text{ m} \) for all moorings; \( T = 44 \text{ days} \) for all moorings except MP4, which is 22 days because of its early failure.

Positive and negative frequencies then correspond to counterclockwise and clockwise rotation, respectively, while positive and negative vertical wavenumbers \( m \) correspond to upward and downward energy propagation, respectively. The shear spectrum \( \Phi_{\text{shear}}(\omega, m) \) can then be computed by multiplying at each frequency by \( (2\pi m)^2 \). Frequency spectra can then be computed by integrating over the wavenumber, and wavenumber spectra can be computed by integrating over frequency.

Shear wavenumber frequency spectra at all moorings (Fig. 11, left) show prominent peaks at \(-f\). The latitudinal shift of the \( f \) peak moving from MP6 to MP1 is clear, as seen before in work such as \( \text{van Haren (2005)} \). Tidal peaks are seen at \( M_2 \) and \( K_1 \) but are deemphasized by examining shear, owing to their red vertical wavenumber spectrum. There is no energy at zero wavenumber at any frequency, owing to detrending and the differentiation operator. Apparent in the Eulerian spectra (not shown) at all moorings to different degrees are prominent peaks at \( f \pm M_2 \), which are because of the vertical heaving of the near-inertial shear layers by tidal motions (Williams 1985; Anderson 1993; Alford 2001a). These are much less prominent when results are viewed here in an isopycnal-following frame.

The spectra indicate that the near-inertial motions have a dominant vertical scale, which is seen more clearly at the right where each spectrum is plotted versus normalized frequency \( \omega/f \) in the vicinity of the negative, near-inertial peak (black boxes at left in Fig. 11). Tidal peaks appear at different values of normalized frequency depending on the latitude; notably, the wavenumber-dependent stripe at MP3 is likely the diurnal tides, which cannot be separated from \( f \) there. All latitudes show a dominant peak near \( f \) at vertical scales around 100–200 m. Notable at all peaks is an absence of strongly dominant downward propagation that has been seen in summer by D’Asaro and Perkins (1984) and in winter at Ocean Station Papa (Alford et al. 2012). Pinkel (2008) argues that the hourglass shape of the spectra is owed to lateral Doppler shifting of the near-inertial layers by subinertial motions and waves of other frequencies. Even in a semi-Lagrangian frame, horizontal advection preferentially shifts the frequencies of the smaller-scale waves. Using a fixed frequency band to define the near inertial thus preferentially excludes the smaller-scale, near-inertial constituents. With two-dimensional data, hourglass-shaped wavenumber–frequency filters can be...
used to isolate the near-inertial band in a Doppler tolerant manner (Pinkel 2014).

Integrals over all frequencies and wavenumbers give the wavenumber and frequency spectra that are traditionally compared to the Garrett–Munk spectrum. The spectra are clearly nonseparable, violating one of the primary assumptions of the GM formulation, and it is interesting to see how the different bands conspire to give approximately the GM form, as originally pointed out by Pinkel (1984). Frequency spectra of velocity (Fig. 12, left) are plotted first as they are easily comparable with common single-depth records. The CW part of the spectrum quite closely follows GM76, with the inertial and tidal peaks superimposed. The bandwidth of the near-inertial peak is generally $\sim 0.2 f$, as has been observed in the past. Secondary peaks are seen slightly higher than this range at both MP1 and MP6, which could be interpreted as energy generated at higher latitudes and propagating to the mooring site.

Of particular interest are the wavenumber spectra of shear (Fig. 12, right), which are generally only slightly above GM76 (dashed), with a broad hump superimposed near 100–200-m scales, as has been observed in many other locations (e.g., Pinkel 1985; Alford and Gregg 2001; Polzin and Lvov 2011). Examining the wavenumber spectrum of only the near-inertial components (thick lines), it is clear that the hump is due to the near-inertial motions, which have a much narrower bandwidth than the rest of the internal wave continuum. Consequently, shear at 100–200-m scales is nearly all near inertial, while shear at 10–20 m scales is dominated by higher frequencies. When viewed in an Eulerian frame (not shown), the near-inertial peak drops off a factor of 2 sooner in wavenumber, owing to the increased spreading because of vertical Doppler shifting.

At MP3, and to a lesser extent MP2, narrower peaks are also seen at a lower wavenumber in addition to the 100–200-m hump. Of these, the downgoing peak is at a longer vertical scale. It is possible that these are the signatures of PSI. In this interpretation, consider an interacting wave triad $m_{IT} + m_{up} = m_{down}$, where $m_{IT}$ is the vertical wavenumber of a mode-1–3 internal tide, and $m_{up}$ and $m_{down}$ are an up- and downgoing, near-inertial wave. Then, taking for MP3 the observed $m_{up} = 3.5 \times 10^{-3}$ cpm and $m_{down} = 2.5 \times 10^{-3}$ cpm gives $m_{IT} = 1 \times 10^{-3}$ cpm or about 1000-m wavelength, a reasonable value.

### 4. Horizontal sections and lateral coherence scales

Having shown that shear in the region is dominantly near inertial, we now examine its horizontal structure by means of shipboard sections past the moorings. Profiles of velocity and shear from MP3 and the HDSS as the ship passed by the mooring agree well (Fig. 3), confirming that they are measuring the same features and allowing examination of their horizontal structure. The agreement is due in part to the dominance of near-inertial waves in this region and in part to their large horizontal scales.

Backrotated shear is computed from the shipboard data using the inertial frequency at MP3, $\omega = f(28.9^\circ)$. Since the operator is applied only over short intervals, the precise frequency used makes little difference.

We first present four transects conducted over a period of 8 days between 27.8° and 30.4°N, passing by mooring MP3. During this time period, velocity, shear, and strain at MP3 (Figs. 13a,b,d) showed several near-inertial waves including downgoing waves at 200 and 800 m and a vertically standing wave near 500 m, evident by its checkerboard pattern (Alford et al. 2007). The vertical heaving of the shear layers by the tidal displacements discussed previously is more evident now while zoomed in. Moored backrotated shear (Fig. 13c) removes the sense of phase propagation with depth but gives a sense of the time scale over which particular near-inertial shear features persist as shown above $[O(10)$ days at this location and time$]$ and also provides a key anchor station for comparison with the shipboard sections of backrotated shear (right).

Four sections were conducted past the mooring, at times indicated by the vertical dashed lines in the time series at left. The mooring location is shown at right with a vertical dashed line. In spite of the vertical advection of these features by internal tides and other motions, backrotated shear at a given location remains fairly constant for most features between the successive occupations, suggesting that they are the same near-inertial features sampled multiple times—a conclusion bolstered by the long coherence times shown in the moored backrotated shear in
Fig. 10. (left) Stacked profiles of time-mean RMS shear. Total shear is shown in light gray, with the downward and upward near-inertial portions plotted in black and dark gray, respectively. Buoyancy frequency $N$, scaled by a factor of 5 to fit on the same axes, is in black. (middle) Profiles of the time-mean ratio of downgoing near-inertial shear ($N_{I\text{down}}$) to total near-inertial shear ($N_{I\text{total}}$). (right) Profiles of the time-mean fraction of near-inertial shear ($N_{I}$) to the total.
Fig. 13c. Agreement between moored and shipboard backrotated shear at all flyby times is very good (as can be seen in Fig. 3 and by comparing times and locations of the dashed lines at left and right in Fig. 13), allowing interpretation of the sections as the spatial structure of the near-inertial motions seen at left.

Each of the layers of backrotated shear observed at the mooring (Fig. 13d) is seen sloping upward to the right in Figs. 13f–h, consistent with propagation downward and toward the equator. The slope of the features is somewhat less than that of a theoretical ray overplotted in Fig. 13h from Garrett (2001). (Note that rays project identically

Fig. 11. Wavenumber frequency spectra of WKB-scaled rotary shear, computed in an isopycnal-following frame from 120 to 1300 m. Right is the zoom in on the box at left, plotted vs normalized frequency $\omega/f$. Boxes at lower left in the panels at right show the spectral resolution in wavenumber and frequency.
onto latitude independent of their lateral propagation direction.) The downward sense of propagation is in agreement with the observed upward phase propagation at the mooring at 200 and 800 m. Together, the moored and shipboard sections at and north of the mooring are broadly consistent with generation of high-mode, near-inertial waves to the north, between 30° and 31°N and propagation downward and equatorward along wave...
characteristics to the mooring site. The phase between shear and strain (section 5 and appendix B) gives a propagation direction toward the southwest, consistent with these conclusions.

At and south of the critical latitude (dashed line), the features are not sloping, which could possibly be evidence that near-inertial energy at those locations is in part due to PSI, which occurs at the inertial frequency.
and therefore gives zero vertical group velocity. This tendency was also observed in our second set of shipboard transects, which were past MP2 and MP3 on yearday 140 (Fig. 14). Three of the transects spanned both moorings, while the fourth passed by MP3 only. The format of the figure is slightly different than previous since the transects passed both moorings, with the panels at the left representing shear and backrotated shear at the two moorings. MP2 contains more noninertial shear than MP3 (Fig. 9), and phase propagation at both moorings is complicated, lacking regular packets of clear upward or downward phase propagation. However, backrotated shear (Figs. 14b,d) shows similar long time scales, as before. Backrotated shear transects (right) again show features repeatedly observed in all transects and long lateral coherence scales, with some extending between moorings (e.g., 400–500 m). The features are everywhere fairly flat, consistent with the lacking sense of vertical propagation seen in the moored records.

Given the extremely short lateral coherence scales of $O(3)$ km noted between moored fixed-depth current meters by Garrett and Munk (1972), the long lateral coherence scales of the shear features is striking, with some features visibly extending at least 100 km. To quantify, we employed a feature-tracking algorithm developed for measuring the length of thermohaline intrusions (Shcherbina et al. 2009) to identify loci of laterally connected shear maxima and minima (gray lines in Figs. 13 and 14). The PDF of feature length (Fig. 15, thin black) peaks at horizontal scales of $\approx 10$–15 km, but its integral (thick gray) shows that about 20% of data are in features longer than 100 km.

We noted a visual tendency for layers in Figs. 13–14 to slope downward toward the equator north of $\lambda_c$ and to be more flat south of it. To quantify, the slope of each of the tracked layers is computed and its PDF plotted in Fig. 16 (gray). Shading represents one standard error in the probability distribution, estimated by randomly sampling many realizations of a Gaussian distribution with the same sample size. The maximum likelihood is at zero, indicating that most layers are flat. Near slopes of $\pm 0.002$, negative slopes are more common, suggesting a tendency for downward and equatorward propagation. Considering layers with mean location only north of (red) and south of $\lambda_c$ (blue), layers with slope magnitudes $0.001$–$0.004$ are significantly more likely to be sloping downward toward the equator north of the latitude than south of it; flat layers are significantly more likely south of $\lambda_c$ than north of it, confirming the visual conclusions drawn earlier.

5. Lateral propagation directions

We next employ a plane wave fitting method that allows us to study the characteristics of individual waves observed in the moored records. The simultaneous measurement of shear and strain then allows us to estimate lateral propagation directions for these waves based on their relative phase following Winkel (1998) and Alford and Gregg (2001) and described in appendix B.

Plane wave fits were performed on inertially filtered shear and strain in isopycnal coordinates using the MATLAB optimization toolbox, which minimizes the error between data and a plane wave model with variable parameters. The model used is a combination of two plane waves, one propagating upward and the other propagating downward:

$$X = a + A_u \cos(\omega_u t - m_u z + \phi_u) + A_d \cos(\omega_d t - m_d z + \phi_d),$$  \hspace{1cm} \hspace{1cm} (4)

where $a$ is a constant, $A_{u,d}$ are wave amplitudes, $\omega_{u,d}$ are the wave frequencies, $m_{u,d}$ are the vertical wavenumbers, and $\phi_{u,d}$ are the phases of downward and upward waves, respectively. The fits are applied in overlapping windows with vertical extent of 200 m and a temporal extent of three inertial periods. Fits done over a range of similar vertical and temporal extents indicate that the results are not very sensitive to these exact parameters. The phase difference between $\omega_u$ and $\omega_d$ is constrained to be $\pi/2$. The fit is applied to shear and strain together, such that frequency and wave-number are constrained to be the same for shear and strain.

Polar probability density functions are then computed for the upgoing and downgoing waves, which had a signal to noise ratio, defined as

$$\frac{\langle|X|^2\rangle}{\langle|u_z - X|^2\rangle},$$  \hspace{1cm} \hspace{1cm} (5)

greater than 1.6 for both shear and strain (20%–37% of the total waves detected; Fig. 17). Downgoing waves at moorings well away from the critical latitude (MP1, MP4, and MP6) show a clear tendency to propagate toward the equator, specifically, the southwest; consistent with wind generation at slightly higher latitude and subsequent propagation downward and equatorward past the mooring. Propagation at MP2 and MP3 is isotropic, possibly consistent with a lack of preferential lateral propagation direction for the PSI-generated waves. Propagation of the upgoing waves (blue) is isotropic at all moorings except MP3, which shows an equatorward tendency. A variety of reasons for isotropic propagation of the upgoing waves can be imagined (discussion).

6. Conclusions

We have presented simultaneous depth–time observations of shear and strain in the upper ocean from
profiling moorings and shear from shipboard ADCP transects. The data are analyzed with the goal of identifying the dominant source of shear in the central North Pacific and quantifying its spatial and temporal scales. The shear is found to be predominantly near inertial, and aspects of propagation and nonlinear interaction (PSI) are explored using a spatial array of five profiling moorings and simultaneous ship-base meridional transects. Our conclusions are as follows:

Fig. 14. (left) Moored data from MP2 and MP3 and (right) shipboard spatial transects past the moorings. The left panels show time series of raw and backrotated zonal shear at MP2 and MP3. Isopycnals are plotted at 60-m intervals. Vertical dashed lines indicate times that ship passed MP3 on each transect. The right panels show backrotated zonal shear measured on four consecutive shipboard transects, plotted vs latitude. Colored periods at the top of each ship transect correspond to time periods at the mooring panels.
1) 60% of RMS shear at scales $>20\text{ m}$ is near-inertial, increasing to about 70% at the critical latitude. For scales $>40\text{ m}$, the percentage increases to over 80% at all moorings. Shear is propagating upward and downward in approximately equal amounts, contrasting sharply with wintertime locations in the storm track that show a strong downward excess. Downward packets tend to be propagating equatorward except near the critical latitude where no preferential direction is observed. Propagation is isotropic at all latitudes for upgoing waves.

2) Shear measured at the critical latitude mooring MP3 stands out as greater, more dominated by near-inertial shear, and more temporally persistent than the other moorings, bolstering evidence presented in Alford et al. (2007) and MacKinnon et al. (2013b) of the importance of PSI at this latitude.

3) Shipboard transects of velocity and shear show the horizontal spatial structure of near-inertial waves. Simultaneous moored measurements confirm that the spatial features are near inertial. Much of the near-inertial shear is coherent over large horizontal distances, on the order of 80 km. A change in character is seen near the critical latitude for PSI of the M$_2$ internal tide, close to MP3. Between MP2 and MP3, features are more horizontal. Features observed north of MP3 show a tendency to slope more and are consistent with downward, equatorward-propagating, near-inertial waves.

7. Discussion

This paper has explored the upper-ocean shear field during weak, summertime wind forcing conditions, through the unusual lens of simultaneous transects and profiling moorings. The data show that even in summer, when near-inertial energy is at a minimum (Alford and Whitmont 2007), shear is dominated by near-inertial waves. This allows back rotation to be used to examine their lateral persistence, which is found to be tens of kilometers, in striking contrast to past moored estimates of very short horizontal coherence scales. It is clear from Figs. 13 and 14 that the slope of the layers and their heaving by displacements of other frequencies would give very short coherence scales estimated from laterally separated current meters at discrete depths (Garrett and Munk 1972).

The long lateral coherence and downward slope toward the equator of the shear layers are as expected for energy traveling downward and toward the equator along characteristics at the group velocity as depicted in Garrett (2001). The observed persistence times and lateral coherence scales are roughly consistent with this view. Taking a meridional group velocity of 0.05 m s$^{-1}$ for a wave of vertical wavelength 200 m at 1.03 f$_c$ at 31°N, a packet 50 km long would take 12 days to pass a fixed point on the ray, the same order as observed.

The longer persistence times at the critical latitude are suggestive of different physics, as concluded by MacKinnon et al. (2013b). In contrast to the features at higher latitude, which have been able to propagate meridionally because their frequency has exceeded f, PSI at $\lambda_c$ would be expected to generate motions at $f$, which would not propagate laterally or vertically. As with any motions at $f$ on a $\beta$ plane (D’Asaro 1989), the meridional variation of $f$ would allow meridional propagation as found by D’Asaro et al. (1995) for

![Figure 15](image1.png)

**Fig. 15.** PDF (thin black) and cumulative distribution function (CDF; thick gray dashed; axis at right) of lengths of shear features identified in Figs. 13 and 14 using a feature-tracking algorithm.

![Figure 16](image2.png)

**Fig. 16.** PDF of slopes of identified features. Thickness of lines is one standard deviation of the spread in Monte Carlo simulations of the PDF with the same sample size as the data. Blue and red lines are subsampled north and south of the critical latitude, respectively. The gray curve is all of the data.
FIG. 17. Polar probability density functions of lateral propagation direction, computed from the phase between shear and strain, for downgoing waves (red) and upgoing waves (blue) at all moorings. Azimuthal axis is degrees true, such that $180^\circ$ represents a wave traveling southward. One standard error in calculation of the PDF for this sample size as determined from Monte Carlo simulations is shown in the upper left panel in heavy black.
wind-generated motions, which might impose the observed tens-of-kilometers spatial scale on the PSI-generated motions. For example, motions 100 km north and south of $\lambda_c$ would have a frequency of $\sim (1 \pm 0.03)f$, sufficiently different from $f$ to allow propagation and decoherence. For comparison, Hazewinkel and Winters (2011) found lateral coherence scales of 30 km, consistent with our results.

The observation of dominant equatorward propagation for downgoing waves away from $\lambda_c$ but not near it supports the general picture drawn from the equatorward/downward slopes to the north and flatter to the south: midlatitude wind gives rise to a spectrum of near-inertial motions that are constrained to propagate down and toward the equator. At $\lambda_c$, PSI augments this energy source, providing upgoing and downgoing waves of approximately equal amounts that also propagate equatorward.

The rates of both of these processes (and therefore their relative importance) and their dependencies are not yet well constrained. Globally, the energy going into the mixed layer inertial motions has been estimated at 0.3–1.3 TW but still has a large uncertainty due to different wind sources (Alford 2001b, 2003b; Watanabe and Hibiya 2002; Jiang et al. 2005; Rimac et al. 2013) and uncertainty in the validity of the slab model (Plueddemann and Farrar 2006). Additionally, the fraction of this energy that propagates to depth is uncertain (Furuichi et al. 2008) but is known to depend sensitively on the mesoscale field (Weller 1982; Lee and Niiler 1998; Alford et al. 2012). Internal tide flux convergences near the critical latitude computed by Simmons (2008) are $5 \times 10^{-9}$ W kg$^{-1}$, giving a global conversion rate from internal tides to near-inertial waves of about 0.2 TW, the same order of magnitude as the wind work, suggesting that PSI is an important source of NIW (as well as a significant sink for the internal tide). The power input to the internal wave field (primarily NIW) from wave–mean flow interactions has been estimated at 0.36 TW by Nagai et al. (2015), also the same order. For progress, these processes, together with the nature of the upgoing near-inertial waves, need to be better understood.

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**APPENDIX A**

**Energy Flux**

Near-inertial energy flux is computed from the bandpass filtered velocity and displacement fields. The deep velocity and temperature measurements allow fitting of the data to the first two vertical modes for velocity and displacement. Mode shapes are determined from the full-depth World Ocean Database climatological stratification (Boyer et al. 2013), which compares well with measured stratification in the upper 100–1400 m. Energy flux is then computed using standard methods (Kunze et al. 2002). Energy flux is defined as the covariance of velocity and pressure perturbations:

$$ F = \langle u' p' \rangle . $$  \hspace{1cm} (A1)

Perturbation pressure $p'$ is computed as

$$ p' = p_{\text{surf}} + \int_0^z \rho' g \, dz , $$  \hspace{1cm} (A2)

where the $\rho'$ is the density perturbation.

The surface pressure $p_{\text{surf}}$ is not measured but can be inferred from the baroclinicity condition that the depth-averaged pressure perturbation must vanish:

$$ \frac{1}{H} \int_{-H}^0 p' (z,t) \, dz = 0 . $$  \hspace{1cm} (A3)

The mode fits to pressure are used to determine the correct offset to use since $p'$ cannot be integrated over the entire water column (Rainville and Pinkel 2006a). Total flux from (A1) (Figure A1, blue) then captures the higher-wavenumber structures missed by the modal fits (Figure A1, gray). Depth-integrated fluxes (Figure 1, arrows) are similar from the two methods.

**APPENDIX B**

**Determining the Propagation Direction of Near-Inertial Waves from the Phase between Shear and Strain**

Starting from the standard internal wave polarization relations for velocity and displacement, the polarization relations for shear and strain are
\[ u_z = \frac{(N^2 - \sigma^2)}{(\sigma^2 - f^2)} \left( -\frac{ik + \frac{lf}{\sigma}}{i} \right) w, \]  
\[ v_z = \frac{(N^2 - \sigma^2)}{(\sigma^2 - f^2)} \left( -\frac{kf}{\sigma} - il \right) w, \]  
\[ \gamma = \frac{\partial}{\partial z} \frac{\eta}{\sigma} = -\frac{m}{\sigma} w, \]

where the vertical velocity \( w = \exp[i(kx + fy + mz - \omega t)] \), \( \sigma \) is frequency, and \( k, l, \) and \( m \) are zonal, meridional, and vertical wavenumbers, respectively.

To match the sign convention used in the data, expressions for shear are multiplied by \(-1\). The polarization relations for shear and strain, matching the data conventions, are then

\[ u_z = \frac{N^2 - \sigma^2}{(\sigma^2 - f^2)} \left( \frac{ik}{\sigma} - \frac{lf}{i} \right) w, \]  
\[ v_z = \frac{N^2 - \sigma^2}{(\sigma^2 - f^2)} \left( \frac{kf}{\sigma} + il \right) w, \]  
\[ \gamma = \frac{\partial}{\partial z} \frac{\eta}{\sigma} = -\frac{m}{\sigma} w, \]
Computing the phase lag for all directions, we find expressions for the propagation direction $\theta$:

$$\theta_{\text{up}} = -\delta \phi_{\text{up}}, \quad \text{and} \quad (B7)$$

$$\theta_{\text{down}} = 180 - \delta \phi_{\text{down}} \quad (B8)$$

for upgoing and downgoing waves, respectively, where $\theta$ is propagation direction in degrees true (CW from north), and $\delta \phi$ is the phase lag between zonal shear $u_z$ and strain (positive if strain leads shear).

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