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Semidiurnal Baroclinic Tides on the Central Oregon Inner Shelf*

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ABSTRACT

Semidiurnal velocity and density oscillations are examined over the mid- and inner continental shelf near Heceta Bank on the Oregon coast. Measurements from two long-term observation networks with sites on and off the submarine bank reveal that both baroclinic velocities and displacements are dominated by the first mode, with larger velocities on the midshelf and northern parts of the bank. Midshelf sites have current ellipses that are near the theoretical value for single, progressive internal tidal waves compared to more linearly polarized currents over the inner shelf. Baroclinic current variability is not correlated to the spring-neap cycle and is uncorrelated between mooring locations. An idealized model of two internal waves propagating from different directions reproduces some of the observed variability in semidiurnal ellipse parameters. At times, the phasing between moorings along a cross-shelf transect are consistent with onshelf wave propagation, a characteristic also present in the output of a three-dimensional regional circulation model. Regional wind-driven upwelling/downwelling influences stratification at all shelf moorings. At locations north of the bank, stronger baroclinic velocities were found during periods of higher background stratification.

1. Introduction

Internal waves of tidal frequency (internal tides) are an important component of oceanic motions. Generated by the interaction of tides, stratification, and variable topography, internal waves can traverse ocean basins and affect processes far from their source regions (e.g., Alford 2003). Much of this energy is generated and dissipated on continental slopes, but it has been estimated that a global average of $\sim 20\%$ of this energy propagates onto continental shelves (Kelly et al. 2013). Near generation regions such as continental slopes, midocean ridges, and enclosed coastlines, there is evidence of coherence with barotropic tidal forcing (e.g., Rayson et al. 2012; Zhao et al. 2010; Scotti et al. 2008). Continental shelves that are open to ocean basins are

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susceptible to remotely generated as well as local wave sources, and their interaction can cause difficulty in long-term predictability of internal tidal strength (Nash et al. 2012a,b).

The inner shelf (the last few kilometers near the coastline in 10-50-m water depth) is far from deep-water internal tide sources, yet periods of internal wave activity remain important because they can coincide with the transport and mixing of heat, nutrients, and invertebrate larvae (Pineda 1991; Leichter et al. 1996; Lucas et al. 2011). Previous continental shelf observations of internal waves have concentrated on the midand outer shelf with some general findings: horizontal currents are clockwise rotating in the Northern Hemisphere (Rosenfeld 1990; Lerczak et al. 2003; Noble et al. 2009); energy density and flux is higher in deeper water and concentrated in low modes (e.g., Pringle 1999; Shearman and Lentz 2004); wave kinematics cannot be described as purely progressive because some portion of the energy is reflected from the coast (Rosenfeld 1990; Lerczak et al. 2003); in shallow water, wave propagation tends to be mostly oriented shore normal with a few exceptions (e.g., Cudaback and McPhee-Shaw 2009);

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FIG. 1. Example from summer 2011 of (a) temperature and (b) cross-shelf velocity from the inner shelf (15-m depth) at LB. Temperature contours in (a) are the 8.5° , 10.5° , and 12.5° isotherms.

and overall, there is little correlation between local sea level and baroclinic tides.

Over the continental shelf, varying bathymetry, timevarying stratification, and strong mesoscale currents can interact to create a complicated wave field. As internal wave speeds decrease over a shoaling shelf and become comparable to mesoscale flows, their interaction is expected to create large temporal variability and interference patterns of internal wave energy (Nash et al. 2012a). The relationship between internal tides and background shelf conditions such as coastal upwelling has been modeled in a twodimensional configuration (Kurapov et al. 2010) but has been difficult to observe because of the short duration of many time series. The classic response, observed over two-dimensional topography, was a decrease in stratification and semidiurnal energy in response to coastal upwelling and an increase in semidiurnal energy in the near surface (<40 m)following a wind relaxation (Hayes and Halpern 1976). A similar response was also observed on the northern California shelf (Rosenfeld 1990).

During summer, the Oregon shelf is characterized by predominantly equatorward winds that drive coastal upwelling (e.g., Huyer 1983). Shelf circulation generated by coastal upwelling interacts with a region of alongshelf varying bathymetry, Heceta–Stonewall Bank (e.g., Castelao and Barth 2005). The three-dimensionality of this large submarine bank influences the low-frequency wind- and pressure gradient–driven circulation over the inner shelf (Kirincich and Barth 2009). Initial observations of internal tide generation showed internal tides emanating as beams from the shelf break (e.g., Torgrimson and Hickey 1979). However, recently local generation of internal tides has been documented in deeper water over the Oregon continental slope (Kelly et al. 2012; Martini et al. 2011). Regions of the shelf with onshore internal tidal energy fluxes have been identified in numerical models (Kurapov et al. 2003; Osborne et al. 2011). Semidiurnal internal tides are weaker during the weakly stratified winter months (Erofeeva et al. 2003), and realistic simulations to isolate locations of barotropic-to-baroclinic semidiurnal energy conversion found that these locations showed little intraseasonal variation (Osborne et al. 2011).

In this work, moored velocity and density observations at mid- and inner-shelf locations are used to quantify internal tidal variability over Heceta-Stonewall Bank. Inner-shelf measurements of water column temperature and velocity frequently capture the large magnitude and temporal variability of the internal tide in this region (Fig. 1). The continental shelf over the bank is susceptible to multiple internal wave sources as well as spatiotemporal variability including background stratification variations due to wind-driven upwelling/downwelling. To characterize and quantify the relationship between internal tides and these background conditions, data from multiple mooring locations over six summers are analyzed. Although the focus here is the Oregon shelf, a similar degree of complexity is expected on most continental shelves on the open coast.

First, the different moored datasets are described and the background conditions typical of the Oregon summer are described (section 2). The primary results of the paper are then examined in three separate sections. Semidiurnal barotropic and baroclinic variability is quantified and differences between moorings identified (section 3). Next, the variability in observed wave parameters during a period of increased baroclinic velocity is described and compared to a simple model that consists of the linear superposition of multiple waves as well as a 1-km horizontal resolution Regional Ocean Modeling System (ROMS) simulation (section 4). Internal tidal events are identified from moored datasets that

each location. More than one value indicates multiple deployments. Depth of temperature sensors $T(z)$, and sampling period Δt .						
ADCP	Δt	Prin. Axis	T(z) (m)	Δt		
$300 \mathrm{kHz}_s$	4 min	$\sim 20^{\circ}$	2, 4, 6, 8, 10, 12, 15, 20, 25, 30, 40, 50, 60, 70	2 min		
$300 \mathrm{kHz}_b$	30 min	-5.7°	70	5 min		
$600 \mathrm{kHz}_b$	2 min	25.7°, 18.1°	1, 4, 9, 15	2 min		
$600 \mathrm{kHz}_b$	10 s	15°	1, 4, 7, 10, 13, 17, 20	5 s		
$600 \mathrm{kHz}_b$	2 min	5.7°, 4.5°, 10.3°	1, 4, 9, 15	2 min		
	$\begin{tabular}{ c c c c c c c c c c c c c c c c c c c$	ADCP Δt 300 kHz_s 4 min 300 kHz_b 30 min 600 kHz_b 2 min 600 kHz_b 10 s 600 kHz_b 2 min	ADCP Δt Prin. Axis 300 kHz _s 4 min ~20° 300 kHz _b 30 min -5.7° 600 kHz _b 2 min 25.7°, 18.1° 600 kHz _b 10 s 15° 600 kHz _b 2 min 5.7°, 4.5°, 10.3°	ADCP Δt Prin. Axis $T(z)$ (m) 300 kHz_s 4 min ~ 20° 2, 4, 6, 8, 10, 12, 15, 20, 25, 30, 40, 50, 60, 70 300 kHz_b 30 min -5.7° 70 600 kHz_b 2 min 25.7°, 18.1° 1, 4, 9, 15 600 kHz_b 10 s 15° 1, 4, 7, 10, 13, 17, 20 600 kHz_b 2 min 5.7°, 4.5°, 10.3° 1, 4, 9, 15		

TABLE 1. Instrumentation used in 2011. Station name with local water column depth H (m), ADCP frequency noting surface- and bottom-mounted configurations (subscripts *s* and *b*, respectively), sampling period Δt , and the principal axis of depth-averaged currents at each location. More than one value indicates multiple deployments. Depth of temperature sensors T(z), and sampling period Δt .

span multiple years, and their occurrences are compared to background oceanographic conditions (section 5). We follow these results with a discussion (section 6) and end with a summary and conclusions from this work (section 7).

2. Data and background conditions

Measurements consist of moored velocity from acoustic Doppler current profilers (ADCPs) and density from a variety of monitoring programs spanning the central Oregon shelf (Table 1). These include an offbank site to the north [Lincoln Beach (LB), 44.85°N], onbank sites off of Newport, Oregon [Newport hydrographic/Yaquina Head (NH/YH), 44.65°N], and a site at the southern end of Heceta Bank [Strawberry Hill (SH), 44.25°N] (see Fig. 2a). Long-term inner-shelf data (LB-15 and SH-15, 15-m water depth) are from two locations maintained by the Partnership for Interdisciplinary Studies of Coastal Oceans (PISCO), an interdisciplinary program to study the connections between inner-shelf oceanography and intertidal ecology with observations on the Oregon shelf since 1998 (e.g., Kirincich et al. 2005). Midshelf measurements are from a mooring on the Newport hydrographic line (NH-10, 83-m water depth), 10 nautical miles from shore (National Data Buoy Center Station



FIG. 2. (a) Central Oregon coast with mooring locations. Triangle is the location of wind measurements (NWP03 station). (b) Time line for instrument deployments that include both density and velocity measurements. Colors for the various moorings are the same as in (a). Vertical gray bars denote periods of spring tide. (c) Mooring configuration for standard PISCO inner-shelf monitoring site. See Table 1 for instrumentation details.

46094), with near-continuous sampling since 1999 (e.g., Huyer et al. 2007). Two other datasets are used to complement these observations from the 2011 upwelling season: a month-long deployment to detect highfrequency internal waves on the inner shelf near Newport (YH-20, 20-m water depth) and a multiyear study of the physical and biological process driving oxygen dynamics on the shelf using the Microbial Initiative in Low Oxygen Areas off Concepción and Oregon (MI_LOCO), which deployed a midshelf mooring at the southern location (SH-70, 70-m water depth) (Adams et al. 2013).

Wind measurements are from the Newport jetty, NOAA weather station NWP03, and sea level measurements are from NOAA tide gauge SBE03, located in Newport harbor. Wind measurements from this station have been shown to be representative of wind conditions at these mid- and inner-shelf mooring locations (Kirincich et al. 2005). The SBE03 hourly measurements of sea level are used for barotropic tidal reference at all locations as the difference in tidal timing is small (<4 min) between SH and LB.

All measurements of density and velocity are hourly averaged and low-pass filtered (Loess filter; Schlax and Chelton 1992) to retain periods > 40 h and define lowfrequency background stratification and velocity ($\rho_{\rm bg}$, $u_{\rm bg}$, $v_{\rm bg}$). Because of side-lobe reflection and signal blanking near the transducer head, moored ADCPs do not measure a full water column velocity profile. Velocities are extrapolated over the blanked regions to the surface and bottom and interpolated to an evenly spaced grid prior to filtering (Kirincich et al. 2005). Velocity measurements are then rotated into their principal axes defined by the low-frequency depth-averaged flow for each deployment (Table 1). These values all roughly correspond to local isobath orientations such that rotated velocity measurements (u, v) are assumed to be in the cross- and along-shelf directions (x, y). Most mooring locations have multiple deployments each year, but in general principal axes differ by $<5^{\circ}$ within each location.

To calculate isopycnal displacements ζ , temperature and salinity measurements are transformed to density and also interpolated to a uniformly spaced grid. NH-10 and YH-20 both have multiple measurements of temperature and salinity on the mooring line; at depths with only a temperature sensor, a linear temperature–salinity relationship is defined at each time step from the collocated temperature–salinity measurements. At mooring locations with only one salinity measurement, density is derived with constant salinity. Isopycnal displacements are calculated from the relation

$$\zeta(z,t) = \frac{\rho(z,t)_{\rm hp}}{\partial \rho(z)_{\rm bp}/\partial z},\tag{1}$$

where $\rho_{\rm hp}$ is the high-pass filtered density $\rho_{\rm hp} = \rho - \rho_{\rm bg}$.

The 2011 summer upwelling season

During the upwelling season on the Oregon shelf, lowfrequency wind relaxations and reversals directly drive shelf currents and alter shelf stratification (e.g., Barth et al. 2005). Similar to previous results, along-shelf currents were highly correlated at short lags with the alongshelf wind stress at all locations (Fig. 3) (Kirincich and Barth 2009). The strength of stratification at all mooring locations is also correlated but with a longer lag to the winds (14-30-h lag). From April to June, early in the upwelling season, there are significant differences between the mid- and inner-shelf stratification time series due to the influence of low-salinity Columbia River plume remnants that are advected past the midshelf site NH-10 (e.g., Huyer 1977). These effects are pronounced in the upper 10-m of the water column. The top-bottom density difference gives an estimate of a single value of stratification at the inner-shelf sites (Fig. 3b). At the midshelf site, density difference between 10 and 30 m is used to reduce the influence of near-surface freshwater events and focus on the variability of the seasonal pycnocline. The strength and vertical structure of shelf stratification is important for internal wave propagation as it affects both the phase speed as well as the vertical structure of the current and displacement modes (Fig. 4) (e.g., Zilberman et al. 2011).

3. Tidal band variability

The temporal variation of the dominant semidiurnal tidal signal M_2 is extracted from the high-pass filtered velocity and displacement data. Decomposition into vertical modes using background density profiles (Fig. 4) is combined with harmonic analysis in sliding 4-day windows to derive M_2 internal tide quantities (e.g., Martini et al. 2011; Rayson et al. 2012). Details of this modal harmonic decomposition are presented in the appendix.

a. Barotropic tides

The zero mode in modal harmonic velocity retrieves an estimate of the barotropic semidiurnal tide. A springneap cycle in the zeroth mode is observed at all moorings (Fig. 5a). The spatial variation in barotropic tidal amplitudes is similar to predicted by the regional TOPEX/Poseidon (TPXO) global inverse solution tidal model (Erofeeva et al. 2003). Only the along-shelf



FIG. 3. Background conditions from 2011. (a) North-south wind from NWP03 station. Double-sided arrow at top indicates the time period considered in text section 4. (b) Stratification (N) from NH-10 (black), LB-15 (blue), and SH-15 (red) moorings. (c) Depth-averaged along-shelf velocity with positive values toward the North. Colors are the same as (b). In all panels, vertical gray bars are periods of spring tide.

component of the barotropic flow is shown as crossshelf zero-mode flows are small. Though analysis with a 4-day window retrieves the combined effect of a semidiurnal band instead of individual tidal constituents, over longer time periods, the individual constituents also match TPXO predictions at these mooring locations (Adams et al. 2013). Mid- and inner-shelf onbank sites (SH-70 and SH-15) have higher barotropic semidiurnal velocity consistent with the intensification of the tides over the submarine bank (Erofeeva et al. 2003). North of the bank, at LB-15, the spring-neap cycle is not as pronounced, and barotropic semidiurnal velocity is small.

b. Baroclinic variability

While a spring-neap cycle is apparent in the barotropic currents, semidiurnal baroclinic variability is not related to the spring-neap cycle (Fig. 5b). Because the 2011 data cover a small number of complete springneap cycles (n = 11), the correlation between baroclinic currents and the spring-neap cycle are deferred to section 5b, where six summers of observations are combined. There is large temporal variability in baroclinic velocity with little overall correlation between locations (Fig. 6). In general, the first baroclinic mode is strongest, with larger total baroclinic velocity at midshelf (NH-10 and SH-70) compared to inner-shelf locations. The first mode is dominant at both mid- and inner-shelf locations on the Newport line (YH-20; NH-10), while locations to the north (LB-15) and the south (SH-15; SH-70) have more energy in higher vertical modes. Similar to findings from the New England shelf, these results also portray a temporally varying modal partition of baroclinic semidiurnal velocity (MacKinnon and Gregg 2003).

In addition to the magnitude of semidiurnal currents, internal wave kinematics can be diagnosed from the phase difference between vertical harmonic modes, such as bottom velocities and midwater column displacement (e.g., Lerczak et al. 2003), and from the parameters of harmonic ellipses (e.g., Levine and Richman 1989). Following the rotary current analysis of Foreman (1978), harmonic currents with both horizontal velocity components trace an elliptical hodograph that can be decomposed into two counterrotating circles of period $2\pi/\omega$, where ω is the cyclic frequency. Using the first-mode semidiurnal baroclinic velocities, ellipse properties of major u_{mj} and minor u_{mn} axis length, inverse ellipticity $u_{\rm mn}/u_{\rm mj}$, inclination β from the x axis, and phase ϕ_u are derived (Foreman 1978). The sense of rotation of the ellipse is determined by comparing the magnitudes of the counterrotating circles. If the clockwise (counterclockwise) circle is larger than the counterclockwise (clockwise) component, inverse ellipticity is negative (positive), and the ellipse rotates clockwise (counterclockwise) through



FIG. 4. Profile of density from 29 Jul 2011 at (a) LB-15 and (d) NH-10 used to solve for the (b),(e) first three vertical modes of displacement and (c),(f) horizontal velocity: mode 1 (blue), mode 2 (green), mode 3 (red). (top) For inner-shelf profile, only mode 1 is shown for clarity. (bottom) For midshelf profile, because of the lack of measurements below 70-m water depth, the density profile is assumed constant.

time. The horizontal velocity components of a single, freely propagating, linear internal wave in the Northern Hemisphere trace a clockwise-rotating ellipse with inverse ellipticity $u_{mn}/u_{mj} = f/\omega$ (0.73 for a semidiurnal internal wave at the latitude of Heceta Bank, where *f* is the Coriolis frequency) and a 180° phase difference between the horizontal velocity and vertical displacement modes (Gill 1982). The major ellipse axis u_{mj} will be interpreted as the current in the direction of wave propagation; μ_{mn} is the current perpendicular to wave propagation; β is the direction of propagation; and ϕ_u is the phase of the current in the direction of wave propagation relative to the onshore component of the current.

Histograms of ellipse properties over the 2011 season show their mean characteristics (Fig. 7). At all locations, the dominant sense of rotation of the horizontal currents is clockwise, consistent with Northern Hemisphere internal waves. About 20% of midshelf (NH-10 and SH-70) current observations have an inverse ellipticity close to this ratio. These distributions shift toward smaller values at the inner-shelf sites. Note that an inverse ellipticity of zero implies linearly polarized currents indicative of nonrotating waves. Histograms of the phase difference between first-mode horizontal currents and displacement indicate a preference for 180°, consistent with freely propagating waves, particularly at the NH-10 and LB-15 locations. Ellipse inclination β is interpreted as the propagation direction of the internal wave with eastward-oriented ellipses directed onshore. Because of the 180° ambiguity in ellipse orientation, rose histograms are repeated for the $\beta - 180^{\circ}$ directions. At the central and northern locations (NH-10, YH-20, and LB-15) the predominant direction of incident semidiurnal internal waves is from the north, while predominantly from the south at the southern locations (SH-70 and



FIG. 5. Lunar fortnightly M_f phase-averaged tidal velocities. (a) Zeroth mode (barotropic) semidiurnal along-shelf velocity from four mooring locations. Open symbols represent 2011 mean ±1 standard error bars. Spring (neap) tides are denoted by dark (light) shaded region. (b) First-mode baroclinic semidiurnal velocity. Symbols and shading are the same as in (a). (c) M_2 velocity magnitude predicted from TPXO. Velocity cophase lines in white; the 200-m isobath is marked in black.

SH-15). Although most ellipses align across bathymetry contours, there is also a fair amount of alongshoreoriented semidiurnal baroclinic variability.

4. Variability within a wave event

While the mean conditions described above (section 3b) show behavior consistent with theoretical single internal waves, there is temporal variability in semidiurnal baroclinic velocity, the goodness of the semidiurnal fit (regression skill; see the appendix), and ellipse properties (Fig. 8). During August 2011, both horizontal components of semidiurnal baroclinic velocity are apparent at most moorings (Figs. 8a-c), indicating wave propagation in both along- and cross-shelf directions. At each mooring, the variability of the first-mode displacement is correlated with the horizontal velocities, consistent with a progressive first-mode internal wave. While the average velocity decreases in shallower water, the average ratio of displacement to total water column depth increases dramatically between the mid- (\sim 5%) and inner-shelf sites (\sim 40%), indicating the increase in nonlinearity of the baroclinic tides as they transit the shelf.

The midshelf mooring, NH-10 shows consistently high semidiurnal fit skill (S_k) (Figs. 8d–f) in either velocity (red, blue) or displacement (black) with inverse ellipticity around the theoretical value (green) and a phase difference between first-mode displacement and velocity close to 180° (Figs. 8g-i). While wave parameters, particularly phase difference and inclination, stay fairly constant at LB-15 from 5 to 17 August, the other inner-shelf location shows more variability. At SH-15, there are shorter periods of high-skill events, extended deviations from the theoretical inverse ellipticity, and veering (temporal variation in ellipse inclination) seen in the records. For example, a 100° inclination variation is seen at SH-15 between 1 and 5 August (Fig. 81). Below, these behaviors are reproduced by windowed harmonic analysis of an idealized model consisting of the superposition of temporally varying, obliquely incident waves.

The temporal variation in phase difference between semidiurnal baroclinic currents and semidiurnal sea level $\phi_u - \phi_\eta$ indicates the degree of coherence between the baroclinic and barotropic tides (gray lines, Figs. 8g–i). For example, from 5 to 17 August at LB-15, the phase difference is fairly constant, indicating coherence



FIG. 6. Stacked bars of first three semidiurnal baroclinic velocity modes from all moorings as noted in panels. Gray vertical bars denote spring tides, double arrow in (a) is same as Fig. 3. Panels to the right are the time mean of each mode. Black lines denoting plus or minus one standard deviation from the mean.

between baroclinic and barotropic tides during this 12-day period.

a. An idealized model of a combination of two waves

While each vertical mode of a single, progressive, linear internal wave has a sense of rotation, ellipticity, and phasing between displacement and horizontal velocity that is constant in space, the combination of multiple waves creates interference patterns and spatial variability to these wave parameters (Rainville et al. 2010; Zhao et al. 2010) and can impact their extraction through windowed harmonic analysis (e.g., Martini et al. 2011). The observations above showed deviations from simple, single wave behavior such as time-varying wave ellipticities and veering of wave inclination. Here, we superpose a single wave and a second wave with lowfrequency temporal variability to demonstrate how ellipse parameters from windowed harmonic analysis are altered by this combined wave field.

A temporally varying first-mode internal wave can be written as an amplitude-modulated (AM) signal:

$$u_{\text{raw}}(z,t) = M(t)[\cos(k_z z)F(t)\mathbf{i} + (f/\omega)\cos(k_z z)F(t)\mathbf{j}],$$
(2)

where **i** and **j** are unit vectors in the along-beam and transverse-beam direction, respectively, and M(t) is the slowly varying modulating factor.

Two waves of the same carrier frequency ω are superposed propagating in different directions: a nonmodulated wave [M(t) = 1] propagating from the northwest at 135° from the positive x axis W1 and a temporally modulated wave W2 propagating from -135° to the x axis. For the second wave, the modulating factor is $M(t) = 1 - \sin(\omega_M t)$, with a modulating frequency $\omega_M = \omega/12.5$. The resultant wave field (Fig. 9a) oscillates between a pure standing wave in the along-shelf direction—a combination of two equal waves at T = 0, 6.25, 12.5, to being dominated by W1 ($T \approx 3, 15.5$), and finally to a partially standing wave when W2 > W1(T = 10). A comparison of elliptical parameters from windowed demodulation at two separate spatial locations (red and blue dots) illustrates a number of effects. First, near the quarter-wavelength location (red; neither a node nor an antinode; $\lambda_l \approx 0.25$), the skill of the fit (Fig. 9b), ellipse ratio (Fig. 9c), and change in dominant wave direction of propagation (Fig. 9d) are all reasonably reproduced by the windowed demodulation. However, near a velocity node (blue), the demodulated wave parameters oscillate in time. The skill of the fit decreases and there are significant departures from the theoretical inverse ellipticity, including periods where the wave ellipse rotates counterclockwise.

With a multidirectional, amplitude-modulated wave field, an interference pattern causes significant spatiotemporal dependence to the results of windowed



FIG. 7. Semidiurnal ellipse parameters for first baroclinic mode from all 2011 mooring deployments. (a–e) Histograms at each location are (left) inverse ellipticity (u_{mn}/u_{mj}) , values <1 indicate clockwise rotation; (center) phase difference in degrees, between velocity in the direction of wave propagation and displacement $(\phi_u - \phi_{\zeta})$; and (right) rose histogram of ellipse inclination to indicate propagation direction. Vertical lines in inverse ellipticity histograms are the theoretical ratio for a freely propagating, planar wave with (-0.72) and without (0) rotation. Vertical lines in phase difference histograms mark pure standing (90°) and freely propagating (180°) waves. In rose histograms, there is a 180° ambiguity in ellipse orientation, thus both inclinations within the ranges 0°-180° and 180°-360° are plotted.





FIG. 8. Time series of first-mode semidiurnal ellipse parameters for the period 20 Jul–30 Aug 2011 at the NH-10, LB-15, and SH-15 moorings. (a)–(c) First-mode baroclinic velocity oscillation in along-shelf (blue), and cross-shelf (red) directions. First-mode baroclinic displacement (black); see (c) for reverse axis. (d)–(f) Skill of semidiurnal harmonic fits to first-mode velocity structure in along- (blue) and cross-shelf (red) directions, and displacement (black). Dashed lines between 0.2 and 0.3 signify the minimum skill S_k defined in text [see (f) for skill axis]. Inverse ellipticity u_{nn}/u_{mj} (green), with theoretical free wave ratio plotted as dashed line [see (d) for axis]. (g)–(i) Phase difference between velocity in the direction of wave propagation and vertical isopycnal displacement (magenta). Phase difference between velocity in the direction of wave propagation and sea level displacement (gray). (j)–(l): Ellipse inclination, counterclockwise from the positive x axis. Dashed lines denote along-shelf (90°) and cross-shelf (0°, 180°) propagating waves.

demodulation. A number of behaviors are similar to observed real-world counterparts such as varying inverse ellipticity, decreases in the skill of harmonic fit, and smooth veering in directionality from one dominant wave to the other. For example, between 1 and 25 August, the NH-10 mooring has consistently high S_k with an approximately constant inverse ellipticity but slowly varying inclination (Fig. 8j). This is similar to the quarter-wavelength time series in Fig. 9 (red). A different example is from the SH-15 mooring. The period between 5 and 15 August sees skill fluctuations, changes in the inverse ellipticity, as well as veering similar to the time series from the idealized wave at the velocity node in Fig. 9 (blue).

While high-skill values on the Oregon shelf were ≈ 0.3 , a high-skill value for the idealized wave field was set at 0.75 to illustrate deviations from a high harmonic fit. In either case, slight decreases in harmonic skill do

not necessarily preclude large changes to the other wave properties, for example, the NH-10 period referred to above.

b. A comparison to ROMS

The geometry of Heceta Bank motivated the oblique internal wave inclination angles used in the idealized model above. In addition to observational evidence for a multidirectional wave field over Heceta Bank (section 3b), this variability is also present within solutions to a high-resolution numerical model. Internal wave parameters are compared to output from a ROMS hydrostatic model with 40 sigma levels, 1-km horizontal resolution, and M₂ tidal forcing. The model has been validated to data with harmonic analysis over long temporal windows (16 days) and reproduced internal tide dynamics at mid- and outer-shelf moorings (Osborne et al. 2011). In this work, we focus on a comparison of



FIG. 9. Idealized linear combination of multidirectional wave field and time-varying wave parameters from windowed demodulation. (a) Contours of onshore velocity field from the combined waves. Along-shelf coordinate in units of along-shelf wavelength λ_l vs time in units of carrier wave period. (b) Percent variance of onshore velocity explained by windowed harmonic regression (S_k). (c) Ratio u_{mn}/u_{mj} . Locations of time series in (b)–(c) are noted in red and blue circles at T = 0. (d) Inclination of wave ellipse.

model wave propagation through the phasing between mid- and inner-shelf moorings and along-shelf differences due to the three-dimensionality of Heceta Bank. The internal tide within this model is driven solely by local dynamics and will not reproduce any of the features of shoaling remotely generated internal tidal energy (e.g., Nash et al. 2012b).

For comparison purposes, a 4-day window of the yearlong model run is taken to be typical of internal tide propagation within the model. The modeled time period is representative of August conditions, with fully developed upwelling and a strong equatorward coastal jet to the north of Heceta Bank that has separated from the coast (e.g., Castelao and Barth 2005). Semidiurnal baroclinic ellipse parameters, energy, and energy flux are computed from the modeled fields to visualize locations of the offshore energy sources and propagation onto the bank and toward shore and facilitate comparison with the data. A rough estimate for the theoretical wavelength of a semidiurnal, mode-1 wave is about 5 km ($c_{\rm ph} = \lambda/T =$ NH/π , where λ is the wavelength, N is the Brunt–Väisälä frequency, and H is the water depth, using average values of $N = 0.003 \,\mathrm{s}^{-1}$ and $H = 100 \,\mathrm{m}$), while the wavelength calculated from modeled displacement phase difference computed at each grid point and referenced to the NH-10

location (Fig. 10) is about 10 km, similar to the cross-shelf distance between moorings and width of the shelf. A comparison of these modeled phase differences to data from the same locations for the August 2011 period show similar phase propagation between the NH-10 and YH-20 mooring (modeled phase difference, 228.5°; observation, 217.4°) but not between NH-10 and either the southern or northern inner-shelf sites. The discrepancy with the other locations is due to either the other mooring locations displaying weaker mode-1 energy at this time because of an oversimplified modeled wave field (e.g., M₂ forcing only) or remote internal tides in the observations.

Modeled energy flux vectors from water depths less than 200 m are all directed onshore and indicate two continental-slope sources of internal tide energy to the north and south of Heceta Bank that impinge on the shelf from these directions (Osborne et al. 2011). Moored baroclinic ellipse orientation at the northern sites (LB-15 and NH-10) suggested wave propagation from the northwest (Fig. 7), consistent with the modeled results shown here and with previous modeling studies in this location (Kurapov et al. 2003). The southern (SH) mooring ellipses suggested incoming wave propagation from the southwest, also comparable to the ROMS model output here (Osborne et al. 2011). In addition to



FIG. 10. Contours of internal tide phase difference in ROMS as determined from mode-1 baroclinic displacement referenced to the NH-10 location (black dot). Also plotted are modeled first-mode energy flux (yellow arrows) and unit vector energy flux estimates from data at the four mooring locations for the August 2011 period.

ellipse orientation, all modeled baroclinic velocities rotated clockwise with inverse ellipticity also decreasing toward shore at all mooring lines (not shown).

5. Multiyear comparison

Six summers of data from the PISCO inner-shelf and the NH-10 midshelf moorings are combined to summarize the strength of the internal wave field and identify patterns of co-occurrence between internal tides and background states including the spring-neap cycle, low-frequency stratification, and a seasonal progression. Velocity and density time series are gathered into internal tide events as described below.

a. Event definition

To identify strong internal tides, data from the three historical mooring locations are vertically decomposed and the 4-day harmonic analysis is conducted. These records are then subsampled to only include demodulated windows where the semidiurnal fit represents a statistically significant portion of the variance contained in the first vertical mode. Following Chelton [1983, his Eq. (18)], the $(1 - \alpha)$ critical value for the hindcast skill of a regression can be estimated by a null hypothesis test $(S_k = 0)$. For a large number of independent samples (N^*) , the test is

$$S_{kcrit}(\alpha) = A \frac{\chi_M(\alpha)}{M},$$
 (3)

where A is the average of many hindcasts conducted at long lags, with two degrees of freedom in the regression model (M = 2), and $\chi_M(\alpha)$ is the 100 α percentage point of the χ^2 distribution with M degrees of freedom. If a value of S_k exceeds the critical value given by Eq. (A7), the regression is statistically significant with $100(1 - \alpha)\%$ confidence. The effective number of independent observations $N^* \approx 20$ and critical regression skills for each demodulation window S_{kcrit} are given in Table A1. As seen here, the number of independent observations is much smaller than the total number of hourly measurements in each window because of autocorrelation within the time series. Time series from each location also have different statistics and different S_{kcrit} dependent on the autocorrelation at each location. With shorter window lengths, a higher amount of regression skill is required to significantly differentiate the regression from noise.

Consecutive significant windows, those which exceed the 4-day S_{kcrit} over a minimum of 4 days, are grouped into events. The distribution of events, coverage percentages, and background variable averages over six summers at the three locations are shown in Fig. 11. For the coverage percentage, a monthly percent of time occupied by events is calculated and then averaged (Fig. 11c). These are compared to monthly averages of background variables to reveal the seasonal progression. From April to July (yearday 100–200), as upwellingfavorable winds, incoming solar radiation, and background stratification increase, event percentages also increase. As the season progresses, the amount of upwelling-favorable winds and incoming solar radiation reduce. The accompanying reduction in background



FIG. 11. The 6-yr monthly mean values: (a) north–south (NS) wind from NWP03 and incoming solar radiation from Hatfield Marine Science Center; (b) surface stratification (0 to 15 m) from the three historical mooring records; and (c) fraction of time occupied by internal tide events. Monthly mean values are marked on the fifteenth of each month. (d) Internal tide events as defined by consecutive high-skill fits of first-mode semidiurnal baroclinic velocity and displacement for the long-term summertime mooring locations NH-10 (black), SH-15 (red), and LB-15 (blue). The 6 yr of data are shown (2006–11) spanning the upwelling season. Vertical gray bars denote periods of spring tide. Circle colors on left indicate year of occurrence for events in Fig. 13.

stratification and percent coverage becomes most apparent at the end of the season. In general, events can last up to 40 days on the midshelf and about half as long at inner-shelf locations; the total amount of time covered by these events is given in Table 2.

b. Events compared to background

To quantify the relationship between internal tides and the spring–neap cycle, we correlate individual time series of baroclinic velocity with the lunar fortnightly cycle. Each location and year have varying record lengths, and the number of degrees of freedom for this correlation is estimated as the total number of observed spring–neap cycles (about 50 cycles). Correlations between semidiurnal baroclinic velocity fluctuations and the spring–neap cycle were not significant at any of the mooring locations. The strength of coastal upwelling is often quantified by the relation to along-shelf wind stress. An alternate characterization is through an exponentially weighted accumulation of along-shelf wind stress, a measure that correlates with the position of the main upwelling front (Austin and Barth 2002)

TABLE 2. Fraction of total time that identified semidiurnal internal tide events occupied the summertime series (2006–11).

Year	NH-10	LB-15	SH-15
2006	0.59	0.34	0.54
2007	0.49	0.53	0.37
2008	0.23	0.11	0.34
2009	0.41	0.16	0.71
2010	0.84	0.59	0.42
2011	0.59	0.46	0.28
Avg	0.53 ± 0.20	0.37 ± 0.20	0.44 ± 0.16



FIG. 12. Comparison of daily averaged background stratification to weighted wind W_{5d} for six summers at (a) NH-10 and (b) LB-15. Colored contours indicate data density increasing from blue to red in 10% intervals from 10% to 90%.

$$W_{5d}(t) = \int_0^t \frac{\tau^s}{\rho_o \exp[(t' - t)/(5 \text{ days})]} dt', \qquad (4)$$

where wind stress is calculated following Large and Pond 1981 and 5 days is used for the weighting decay scale. The relationship between W_{5d} and shelf stratification shows that on average W_{5d} is negative, consistent with the predominant upwelling-favorable winds (Fig. 12). Midshelf stratification is higher than the inner shelf, but the inner shelf is never fully unstratified even during periods of extended upwelling. The strongest stratification values occur during wind relaxations ($W_{5d} \approx 0 \text{ m}^2 \text{ s}^{-1}$) and during weak upwelling or weak downwelling. The lowest inner-shelf stratification values are seen during strong wind reversals ($W_{5d} > 2 \text{ m}^2 \text{ s}^{-1}$), which are not observed at this midshelf mooring (Barth et al. 2005).

Internal tide events, with high baroclinic velocity, coincide with periods of stronger background stratification, consistent with a linear model (e.g., Gill 1982) (Figs. 13a,c,e). Here, all values of correlation skill R^2 are significant at the 95% confidence level for the number of individual events at each location. Instead of standard linear least squares regression, which minimizes error in the dependent variable y, a nonlinear regression is used following Fewings and Lentz (2010), which calculates a best-fit line using wave event standard deviations as a measure of uncertainty in both dimensions (x, y). During the Oregon summer, stronger stratification periods in general correlate with weak upwelling, relaxations in the wind, or short wind reversals ($W_{5d} < 1 \text{ m}^2 \text{ s}^{-1}$). Note that the relationship to baroclinic displacement is not shown because the displacement itself is calculated as a function of stratification. Significant relationships are found between the average first-mode baroclinic velocity and local background stratification at both the northern mid- and inner-shelf sites (Figs. 13a–d). At the southern site, SH-15, while correlations to wind and background stratification are significant, the magnitude of baroclinic currents during internal tide events is weak (Figs. 13e,f).

6. Discussion

Overall, temporal variability of inner-shelf baroclinic fields was challenging to explain. There was no consistent spring-neap cycle and only an intermittently consistent phase relation to semidiurnal sea level. Though semidiurnal baroclinic velocity fluctuations were higher during periods of high stratification, the overall correlation was low, suggesting a complicated relationship to upwelling/downwelling. In a region of relatively simple topography, Hayes and Halpern (1976) suggest that background stratification modulations during the transition from upwelling to downwelling shift the path of internal wave characteristics, thus altering the amount of wave reflection to the deep ocean. This regional control resembles similar mechanisms in other locations where a shoaled thermocline modifies wave propagation to the shelf (Petruncio et al. 1998; Noble et al. 2009). In southern California, internal tides were found to be coherent across the entire shelf only during periods of sustained equatorward flow, whose geostrophically balanced density field create a favorable waveguide (Noble et al. 2009).

We observe significant semidiurnal baroclinic ellipticity on the Oregon inner shelf. Possible explanations include the combination of multiple wave sources (explored in section 4a), Doppler-shifted upstream lowerfrequency variability (Pringle 1999), or the advection of a horizontally sheared background along-shelf flow by the shoreward-propagating internal tide. These lowfrequency currents can affect the propagation of internal



FIG. 13. (left) Occurrence of high-skill S_k first-mode semidiurnal baroclinic events vs background variables lowfrequency stratification (*N*, cycles per hour) and (right) weighted wind stress (W_{5d} , m²s⁻¹) at (a),(b) NH-10, (c),(d) LB-15, and (e),(f) SH-15 moorings. Event means are colored circles as in Fig. 11 (plus or minus one std dev). The number of events *n* are noted in each panel. Background shading indicates the data density of individual data windows used to create the mean. Thick dashed lines are the best fit nonlinear regression; thin lines are the 95% confidence levels for the regressions.

waves in a variety of ways, such as increasing the allowable frequencies of internal waves (Mooers 1975) or providing a source of energy to the semidiurnal band (Davis et al. 2008). In addition to changing stratification, regional-scale upwelling/downwelling on the Oregon shelf also creates strong along-shelf jets. To consider how the advection of background horizontal density and velocity gradients might affect our results, we consider the respective horizontal advection equations for wave-induced density and along-shelf velocity fluctuations:

$$\rho_t = -u\overline{\rho}_x, \quad v_t = -u\overline{v}_x. \tag{5}$$

Here, the magnitude of density ρ and along-shelf velocity v fluctuations are due to the advection of mean gradients

oriented in the cross-shelf direction by a cross-shelf semidiurnal baroclinic current u. Mean cross-shelf gradients are estimated by differencing the low-frequency density and depth-averaged along-shelf flow between the YH-20 and NH-10 moorings, and a typical semidiurnal baroclinic current of $0.1 \,\mathrm{m \, s^{-1}}$ is used. From these equations, approximate magnitudes for the rate of change of ρ and v due to this process can be estimated. Given the small magnitude of the observed background along-shelf velocity gradient (see Fig. 3c), this advection term is small compared to the observed semidiurnal values. Comparing this estimate to one derived from a well-resolved horizontal transect of the along-shelf flow (Castelao and Barth 2005) also yields small values of v. Inner-shelf moorings would have to be fortuitously located in regions of unresolved high horizontal shear in order to account for the observed baroclinic along-shelf velocity.

On the other hand, density fluctuations due to the horizontal density gradient advection are as strong as the observed density fluctuations on the inner shelf. This might explain the large displacement measurements and the need to reconsider the use of Eq. (1) to determine internal wave displacement over the inner shelf. These results are consistent with the background consisting of a geostrophically balanced coastal jet with no horizontal along-shelf velocity shear. The 83-m NH-10 mooring is located within the range of isobaths followed by this coastal jet as it deflects around the bank complex in the summer (Barth et al. 2005).

While internal waves are not supported over the continental shelf during the unstratified winter, the existence of even small amounts of stratification throughout the summer is apparently enough to support internal tide propagation on the Oregon shelf, though there is substantial year-to-year variability in the fraction of time occupied by wave events (Table 2). This potentially indicates year-to-year variability in the amount of incoming remotely generated semidiurnal internal wave energy (Nash et al. 2012a). The manifestation of remotely generated internal tides over the Oregon shelf will need future consideration. The Oregon ROMS model with local dynamics appears to simulate the upper slope wave sources and their shoreward propagation well, but not the extent of coherence between sites along the bank or the timing of low-frequency variability (Osborne et al. 2011). The addition of a remote wave field may improve our understanding of baroclinic tides on the Oregon shelf.

7. Conclusions

In this work we have combined observations from numerous moorings spanning the mid- and inner shelf over Heceta Bank to provide a picture of baroclinic variability. Barotropic currents on the Oregon inner shelf are consistent with the TPXO shallow-water tidal model in both magnitude and spatial variation (Erofeeva et al. 2003). Semidiurnal baroclinic current variability does not have a spring-neap relation nor is it overall correlated between the sites. At times, variability is well correlated along a single latitudinal line consistent with onshelf propagation, as seen in the relative phasing between observations as well as output from a 3D regional numerical model. Baroclinic velocities and displacement are both dominated by the first mode, with larger velocities at the midshelf and at northern sites. Midshelf sites also have ellipticities that are more consistent with single, progressive internal waves compared to the inner-shelf sites.

In contrast to unidirectional wave fields, continental shelf internal tides are highly variable, including time and space differences in ellipticity, inclination, and sense of rotation. Some of these features can be reproduced by an idealized linear model superposing two incoming amplitude-modulated waves. Over the Oregon slope Martini et al. (2011) also found that much of the variability was explained by a sum of multiple internal waves. Because of the smaller continental shelf spatial scales and the multiple wave sources over Heceta Bank, rapid evolution in spatial interference patterns could be expected (Nash et al. 2012a). The general behavior of signal extraction tools to amplitude modulation was explored in a highly idealized setting and reproduced some features of the observations. To make more accurate predictions of wave parameters over Heceta Bank, future steps will need to include the realistic effects of topography, background stratification, wave sources, and the appropriate temporal window that modulates the baroclinic tidal cycle. Another modeling approach would be to use a phase-modulated wave field (e.g., Colosi and Munk 2006). With the goal to resolve event-scale variability, amplitude modulation was chosen in favor of phase modulation because of the poor frequency resolution of short window harmonic analysis.

The spatial variability of velocity and density observations within a multidirectional wave field limit the use of density or velocity measurements alone. Significant baroclinic periods, termed events, can last days to weeks with consistent wave behavior such as elevated velocities and phase relations between internal and surface tides. To fully understand the propagation of internal tides to shallow coastal areas with smaller scales, 3D fields need to be better simulated. This is especially true as the coastline is approached and shallow-water depths limit the resolution of σ -level models. Fully nonhydrostatic dynamics will be needed to capture the observed higher-frequency waves with large vertical velocities at the front of the internal tides at these locations (e.g., Venayagamoorthy and Fringer 2007; S. Suanda and J. Barth 2015, unpublished manuscript).

At all moorings, variability in local stratification is related to wind-driven upwelling/downwelling, the dominant forcing mechanism at subtidal time scales. At locations to the north of Heceta Bank, stronger baroclinic velocities were significantly correlated to periods of high background stratification. Sustained observations over long periods were essential to this analysis. Finally, internal waves have important implications for environmental questions on the Oregon shelf with regards to intertidal invertebrate recruitment or coastal hypoxia. Both of these processes have been linked to low-frequency upwelling/downwelling as a physical driver. However, there are examples within these studies where the upwelling/downwelling hypothesis proved inconclusive (e.g., Shanks and Shearman 2009). Internal tides are an alternate physical driver of cross-shelf material transport, and their connection to upwelling/downwelling is indicative of the complexity of these processes and the work left to disentangle their effects.

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APPENDIX

Modal Harmonic Decomposition

Modal harmonic decomposition of the high-pass filtered velocity and displacement are used to isolate baroclinic semidiurnal variability. Data are projected onto dynamical vertical internal wave modes as defined by solution of the vertical structure function for linear flat bottom modes Ψ_n at each mooring location:

$$\frac{d^2\Psi_n(z)}{dz^2} + k_n^2 \frac{N^2(z) - \omega^2}{\omega^2 - f^2} \Psi_n(z) = 0.$$
 (A1)

The semidiurnal M₂ tidal frequency is ω , *f* is the Coriolis frequency and the eigenvalues k_n^2 are the squared wavenumber. Boundary conditions are $\Psi = 0$ at z = (0, -D), where *D* is the water depth. Background stratification N^2 is given by the vertical density gradient

$$N^{2}(z) = -\frac{g}{\rho_{o}} \frac{\partial \rho_{\rm bg}(z)}{\partial z}.$$
 (A2)

The vertical functions $\Psi_n(z)$ describe the structure of vertical velocity and displacement for each mode. The polarization relation for internal waves relates the displacement modal structures and the structure of horizontal velocity:

$$U_n(z) = \frac{1}{k_n} \frac{d\Psi_n(z)}{dz}.$$
 (A3)

At locations with many density measurements in the vertical, low-pass filtered density values $\rho_{bg}(z)$ at each mooring are used to determine N^2 and to solve for vertical modes at each time step (Fig. 4). Most of the 15-m PISCO sites have only four measurements in the vertical. At these locations, analytical functions are used for the vertical modes. A depth-uniform N^2 taken from the top-bottom density gradient yields symmetric structure functions [$\cos(nz/\pi)$]. As no vertical density gradients could be estimated at SH-70, analytical modes were used to define the horizontal velocity modes there as well.

The structure functions are used to project hourly, high-pass filtered velocity profiles $u_{hp}(z, t) = u(z, t) - u_{bg}(z, t)$ onto vertical modes:

$$u_{\rm hp}(z,t) = \sum_{n} \widehat{u_n}(t) U_n(z,t')$$
$$v_{\rm hp}(z,t) = \sum_{n} \widehat{v_n}(t) U_n(z,t'). \tag{A4}$$

Here, the vertical modes are slowly varying in time t', and the low-pass filtered density is used to solve Eq. (A1). Horizontal velocities are projected onto four



FIG. A1. Vertical mode fit from SH-70 cross-shelf velocity *u*. (a) Hourly, 40-h high-pass filtered data. (b) Fit using the barotropic and first three analytical velocity modes. (c) Absolute value of the difference between the hourly high-passed data and the modal fit. (d) Mean error at each depth during this period (envelope is plus or minus one std dev from the mean).

dynamical modes (n = 3), consisting of a barotropic zero mode ($U_0 = 1$) and the first three baroclinic modes.

The difference between the modally decomposed and original data is quantified by an RMS error that is small $(<0.02 \,\mathrm{m\,s^{-1}})$ throughout most of the water column (Fig. A1). As the PISCO inner-shelf sites have only four discrete measurements of density in 15-m of water, three modes are the most that can be resolved for displacement.

Temporal variability centered around the M₂ constituent is isolated for each of the above displacement $[\hat{\zeta}_n(t)]$ and velocity modes $[\hat{u}_n(t), \hat{v}_n(t)]$ by partially overlapped short window harmonic regression:

$$\begin{aligned} \widehat{\zeta_n}(t) &= A_{\zeta_n}(t_c) \sin\omega_{M_2} t + B_{\zeta_n}(t_c) \cos\omega_{M_2} t + \overline{\zeta_n} \\ \widehat{u_n}(t) &= A_{u_n}(t_c) \sin\omega_{M_2} t + B_{u_n}(t_c) \cos\omega_{M_2} t + \overline{u_n} \\ \widehat{v_n}(t) &= A_{v_n}(t_c) \sin\omega_{M_2} t + B_{v_n}(t_c) \cos\omega_{M_2} t + \overline{v_n}. \end{aligned}$$
(A5)

This yields an amplitude and phase for each modal harmonic constituent centered in time on the middle of the demodulation window t_c . For example, the amplitude and phase of the displacement modes are given by

$$\begin{aligned} \zeta_n(t_c) &= \left[A_{\zeta_n}(t_c) + B_{\zeta_n}(t_c)\right]^{0.5} \\ \phi_{\zeta_n}(t_c) &= \arctan\left[\frac{A_{\zeta_n}(t_c)}{B_{\zeta_n}(t_c)}\right], \end{aligned} \tag{A6}$$

with similar expressions for the horizontal velocity components. For each mode, a goodness of fit is defined by the fraction of variance within each window that is explained by the semidiurnal fit:

$$S_{k_n}(t_c) = \frac{\sigma_{n_{M_2}}^2(t_c)}{\sigma^2(t_c)},$$
 (A7)

where $\sigma^2(t_c)$ is the total variance within the demodulation window and $\sigma_{n_{M_2}}^2(t_c)$ is the variance from the semidiurnal harmonic. Depending on window duration, a number of

TABLE A1. The estimated degrees of freedom and minimum significance required for three different demodulation windows used on the first cross-shelf baroclinic velocity (subscript u) and displacement (subscript ζ) mode. All for S_{kcrit} at the 95% confidence level. Missing values for SH-70 $_{\zeta}$ are because no density data are available at this location.

	1 day	2 day	4 day
Location	$S_{kcrit} (N^*)$	$S_{kcrit}(N^*)$	$S_{kcrit}(N^*)$
NH-10 _u	0.54 (11)	0.33 (18)	0.22 (27)
SH-70 _u	0.50 (12)	0.32 (19)	0.22 (27)
$LB-15_u$	0.67 (9)	0.40 (15)	0.23 (26)
SH-15 _u	0.46 (13)	0.33 (18)	0.20 (30)
$YH-20_u$	0.46 (13)	0.35 (17)	0.26 (23)
NH-10 ζ	0.60 (10)	0.29 (21)	0.15 (39)
$SH-70\zeta$	_	_	_
LB-15 ζ	0.60 (10)	0.43 (14)	0.26 (23)
SH-15 ζ	0.60 (10)	0.54 (11)	0.32 (19)
$YH-20_{\zeta}$	0.99 (6)	0.67 (9)	0.46 (13)

semidiurnal frequencies besides M_2 are contained within the demodulated velocity and density fields. The 1-, 2-, and 4-day windows are all attempted, with slightly different characteristics (see Table A1).

REFERENCES

- Adams, K. A., J. A. Barth, and F. Chan, 2013: Temporal variability of near-bottom dissolved oxygen during upwelling off central Oregon. J. Geophys. Res. Oceans, 118, 4839–4854, doi:10.1002/ jgrc.20361.
- Alford, M. H., 2003: Redistribution of energy available for ocean mixing by long-range propagation of internal waves. *Nature*, 423, 159–162, doi:10.1038/nature01628.
- Austin, J. A., and J. A. Barth, 2002: Variation in the position of the upwelling front on the Oregon shelf. J. Geophys. Res., 107, 3180, doi:10.1029/2001JC000858.
- Barth, J. A., S. D. Pierce, and R. M. Castelao, 2005: Time-dependent, wind-driven flow over a shallow midshelf submarine bank. J. Geophys. Res., 110, C10S05, doi:10.1029/2004JC002761.
- Castelao, R. M., and J. A. Barth, 2005: Coastal ocean response to summer upwelling favorable winds in a region of alongshore bottom topography variations off Oregon. J. Geophys. Res., 110, C10S04, doi:10.1029/2004JC002409.
- Chelton, D. B., 1983: Effects of sampling errors in statistical estimation. *Deep-Sea Res.*, **30A**, 1083–1103, doi:10.1016/ 0198-0149(83)90062-6.
- Colosi, J. A., and W. Munk, 2006: Tales of the venerable Honolulu tide gauge. J. Phys. Oceanogr., 36, 967–996, doi:10.1175/JPO2876.1.
- Cudaback, C. N., and E. McPhee-Shaw, 2009: Diurnal-period internal waves near point conception, California. *Estuarine Coastal Shelf Sci.*, 83, 349–359, doi:10.1016/j.ecss.2008.12.018.
- Davis, K. A., J. J. Leichter, J. L. Hench, and S. G. Monismith, 2008: Effects of western boundary current dynamics on the internal wave field of the southeast Florida shelf. J. Geophys. Res., 113, C09010, doi:10.1029/2007JC004699.
- Erofeeva, S. Y., G. D. Egbert, and P. M. Kosro, 2003: Tidal currents on the central Oregon shelf: Models, data, and assimilation. *J. Geophys. Res.*, **108**, 3148, doi:10.1029/2002JC001615.
- Fewings, M. R., and S. J. Lentz, 2010: Momentum balances on the inner continental shelf at Martha's Vineyard Coastal Observatory. J. Geophys. Res., 115, C12023, doi:10.1029/2009JC005578.
- Foreman, M. G. G., 1978: Manual for tidal currents analysis and prediction. Institute of Ocean Sciences Pacific Marine Science Rep. 78-6, 57 pp.
- Gill, A. E., 1982: Atmosphere–Ocean Dynamics. Academic Press, 662 pp.
- Hayes, S. P., and D. Halpern, 1976: Observations of internal waves and coastal upwelling off the Oregon coast. J. Mar. Res., 34, 247–267.
- Huyer, A., 1977: Seasonal variation in temperature, salinity, and density over the continental shelf off Oregon. *Limnol. Oceanogr.*, 22, 442–453.
- —, 1983: Coastal upwelling in the California Current System. Prog. Oceanogr., 12, 259–284, doi:10.1016/0079-6611(83)90010-1.
- —, P. A. Wheeler, P. T. Strub, R. L. Smith, R. Letelier, and P. M. Kosro, 2007: The Newport line off Oregon—Studies in the North East Pacific. *Prog. Oceanogr.*, **75**, 126–160, doi:10.1016/ j.pocean.2007.08.003.
- Kelly, S. M., J. D. Nash, K. I. Martini, M. H. Alford, and E. Kunze, 2012: The cascade of tidal energy from low to high modes on a continental slope. *J. Phys. Oceanogr.*, 42, 1217–1232, doi:10.1175/ JPO-D-11-0231.1.

- —, N. L. Jones, J. D. Nash, and A. F. Waterhouse, 2013: The geography of semidiurnal mode-1 internal-tide energy loss. *Geophys. Res. Lett.*, **40**, 4689–4693, doi:10.1002/grl.50872.
- Kirincich, A. R., and J. A. Barth, 2009: Alongshelf variability of inner-shelf circulation along the central Oregon coast during summer. J. Phys. Oceanogr., 39, 1380–1398, doi:10.1175/ 2008JPO3760.1.
- —, —, B. A. Grantham, B. A. Menge, and J. Lubchenco, 2005: Wind-driven inner-shelf circulation off central Oregon during summer. J. Geophys. Res., 110, C10S03, doi:10.1029/ 2004JC002611.
- Kurapov, A. L., G. D. Egbert, J. S. Allen, R. N. Miller, S. Y. Erofeeva, and P. M. Kosro, 2003: The M₂ internal tide off Oregon: Inferences from data assimilation. *J. Phys. Oceanogr.*, 33, 1733–1757, doi:10.1175/2397.1.
- —, J. S. Allen, and G. D. Egbert, 2010: Combined effects of winddriven upwelling and internal tide on the continental shelf. J. Phys. Oceanogr., 40, 737–756, doi:10.1175/2009JPO4183.1.
- Large, W. G., and S. Pond, 1981: Open ocean momentum flux measurements in moderate to strong winds. J. Phys. Oceanogr., 11, 324–336, doi:10.1175/1520-0485(1981)011<0324: OOMFMI>2.0.CO;2.
- Leichter, J. J., S. R. Wing, S. L. Miller, and M. W. Denny, 1996: Pulsed delivery of subthermocline water to Conch Reef (Florida Keys) by internal tidal bores. *Limnol. Oceanogr.*, 41, 1490–1501.
- Lerczak, J. A., C. D. Winant, and M. C. Hendershott, 2003: Observations of the semidiurnal internal tide on the southern California slope and shelf. J. Geophys. Res., 108, 3068, doi:10.1029/2001JC001128.
- Levine, M. D., and J. G. Richman, 1989: Extracting the internal tide from data: Methods and observations from the Mixed Layer Dynamics Experiment. J. Geophys. Res., 94, 8125–8134, doi:10.1029/JC094iC06p08125.
- Lucas, A. J., P. J. Franks, and C. L. Dupont, 2011: Horizontal internal-tide fluxes support elevated phytoplankton productivity over the inner continental shelf. *Limnol. Oceanogr. Fluids Environ.*, 1, 56–74, doi:10.1215/21573698-1258185.
- MacKinnon, J. A., and M. C. Gregg, 2003: Shear and baroclinic energy flux on the summer New England shelf. J. Phys. Oceanogr., 33, 1462–1475, doi:10.1175/1520-0485(2003)033<1462: SABEFO>2.0.CO;2.
- Martini, K. I., M. H. Alford, E. Kunze, S. M. Kelly, and J. D. Nash, 2011: Observations of internal tides on the Oregon continental slope. *J. Phys. Oceanogr.*, 41, 1772–1794, doi:10.1175/2011JPO4581.1.
- Mooers, C. N. K., 1975: Several effects of a baroclinic current on the cross-stream propagation of inertial-internal waves. *Geophys. Fluid Dyn.*, 6, 245–275, doi:10.1080/03091927509365797.
- Nash, J. D., S. M. Kelly, E. L. Shroyer, J. N. Moum, and T. F. Duda, 2012a: The unpredictable nature of internal tides on continental shelves. *J. Phys. Oceanogr.*, **42**, 1981–2000, doi:10.1175/ JPO-D-12-028.1.
- —, E. L. Shroyer, S. M. Kelly, M. E. Inall, T. F. Duda, M. D. Levine, N. L. Jones, and R. C. Musgrave, 2012b: Are any coastal internal tides predictable? *Oceanography*, **25**, 80–95, doi:10.5670/oceanog.2012.44.
- Noble, M., B. Jones, P. Hamilton, J. Xu, G. Robertson, L. Rosenfeld, and J. Largier, 2009: Cross-shelf transport into nearshore waters due to shoaling internal tides in San Pedro Bay, CA. *Cont. Shelf Res.*, **29**, 1768–1785, doi:10.1016/ j.csr.2009.04.008.
- Osborne, J. J., A. L. Kurapov, G. D. Egbert, and P. M. Kosro, 2011: Spatial and temporal variability of the M₂ internal tide

generation and propagation on the Oregon shelf. J. Phys. Oceanogr., **41**, 2037–2062, doi:10.1175/JPO-D-11-02.1.

- Petruncio, E. T., L. K. Rosenfeld, and J. D. Paduan, 1998: Observations of the internal tide in Monterey Canyon. J. Phys. Oceanogr., 28, 1873–1903, doi:10.1175/1520-0485(1998)028<1873: OOTITI>2.0.CO;2.
- Pineda, J., 1991: Predictable upwelling and the shoreward transport of planktonic larvae by internal tidal bores. *Science*, 253, 548– 549, doi:10.1126/science.253.5019.548.
- Pringle, J. M., 1999: Observations of high-frequency internal waves in the coastal ocean dynamics region. J. Geophys. Res., 104, 5263–5281, doi:10.1029/1998JC900053.
- Rainville, L., T. M. S. Johnston, G. S. Carter, M. A. Merrifield, R. Pinkel, P. F. Worcester, and B. D. Dushaw, 2010: Interference pattern and propagation of the M₂ internal tide south of the Hawaiian Ridge. J. Phys. Oceanogr., 40, 311–325, doi:10.1175/ 2009JPO4256.1.
- Rayson, M. D., N. L. Jones, and G. N. Ivey, 2012: Temporal variability of the standing internal tide in the Browse basin, Western Australia. J. Geophys. Res., 117, C06013, doi:10.1029/2011JC007523.
- Rosenfeld, L. K., 1990: Baroclinic semidiurnal tidal currents over the continental shelf off northern California. J. Geophys. Res., 95, 22 153–22 172, doi:10.1029/JC095iC12p22153.
- Schlax, M. G., and D. B. Chelton, 1992: Frequency domain diagnostics for linear smoothers. J. Amer. Stat. Assoc., 87, 1070– 1081, doi:10.1080/01621459.1992.10476262.

- Scotti, A., R. C. Beardsley, B. Butman, and J. Pineda, 2008: Shoaling of nonlinear internal waves in Massachusetts Bay. J. Geophys. Res., 113, C08031, doi:10.1029/2008JC004726.
- Shanks, A. L., and R. K. Shearman, 2009: Paradigm lost? Crossshelf distributions of intertidal invertebrate larvae are unaffected by upwelling or downwelling. *Mar. Ecol. Prog. Ser.*, 385, 189–204.
- Shearman, R. K., and S. J. Lentz, 2004: Observations of tidal variability on the New England shelf. J. Geophys. Res., 109, C06010, doi:10.1029/2003JC001972.
- Torgrimson, G. M., and B. M. Hickey, 1979: Barotropic and baroclinic tides over the continental slope and shelf off Oregon. J. Phys. Oceanogr., 9, 945–961, doi:10.1175/1520-0485(1979)009<0945: BABTOT>2.0.CO;2.
- Venayagamoorthy, S. K., and O. B. Fringer, 2007: On the formation and propagation of nonlinear internal boluses across a shelf break. J. Fluid Mech., 577, 137–159, doi:10.1017/ S0022112007004624.
- Zhao, Z., M. H. Alford, J. A. MacKinnon, and R. Pinkel, 2010: Longrange propagation of the semidiurnal internal tide from the Hawaiian Ridge. J. Phys. Oceanogr., 40, 713–736, doi:10.1175/ 2009JPO4207.1.
- Zilberman, N. V., M. A. Merrifield, G. S. Carter, D. S. Luther, M. D. Levine, and T. J. Boyd, 2011: Incoherent nature of M₂ internal tides at the Hawaiian Ridge. J. Phys. Oceanogr., 41, 2021–2036, doi:10.1175/JPO-D-10-05009.1.