Global Observations of Open-Ocean Mode-1 M_2 Internal Tides

ZHONGXIANG ZHAO

Applied Physics Laboratory, University of Washington, Seattle, Washington

MATTHEW H. ALFORD

Scripps Institution of Oceanography, University of California San Diego, La Jolla, California

JAMES B. GIRTON AND LUC RAINVILLE

Applied Physics Laboratory, University of Washington, Seattle, Washington

HARPER L. SIMMONS

University of Alaska Fairbanks, Fairbanks, Alaska

(Manuscript received 6 June 2015, in final form 13 February 2016)

ABSTRACT

A global map of open-ocean mode-1 M_2 internal tides is constructed using sea surface height (SSH) measurements from multiple satellite altimeters during 1992–2012, representing a 20-yr coherent internal tide field. A two-dimensional plane wave fit method is employed to 1) suppress mesoscale contamination by extracting internal tides with both spatial and temporal coherence and 2) separately resolve multiple internal tidal waves. Global maps of amplitude, phase, energy, and flux of mode-1 M_2 internal tides are presented. The M_2 internal tides are mainly generated over topographic features, including continental slopes, midocean ridges, and seamounts. Internal tidal beams of 100–300 km width are observed to propagate hundreds to thousands of kilometers. Multiwave interference of some degree is widespread because of the M_2 internal tide's numerous generation sites and long-range propagation. The M_2 internal tide propagates across the critical latitudes for parametric subharmonic instability (28.8°S/N) with little energy loss, consistent with the 2006 Internal Waves across the Pacific (IWAP) field measurements. In the eastern Pacific Ocean, the M_2 internal tide loses significant energy in propagating across the equator; in contrast, little energy loss is observed in the equatorial zones of the Atlantic, Indian, and western Pacific Oceans. Global integration of the satellite observations yields a total energy of 36 PJ ($1 \text{ PJ} = 10^{15} \text{ J}$) for all the coherent mode-1 M_2 internal tides. Finally, satellite observed M_2 internal tides compare favorably with field mooring measurements and a global eddy-resolving numerical model.

1. Introduction

Internal tides are generated by barotropic tidal currents flowing over variable bottom topography in the stratified oceans (Wunsch 1975; Munk 1981; Baines 1982). Oscillating cross-isobath tidal currents disturb

isopycnal surfaces, and the disturbances radiate away as internal tides, that is, internal gravity waves at the tidal frequency. The global conversion rate from barotropic to baroclinic tidal energy is estimated to be 1 TW, mainly over continental slopes, midocean ridges, and seamounts (Baines 1982; Morozov 1995; Egbert and Ray 2000, 2001). In the past two decades, the revival of research interest in internal tides has been inspired mainly by new observations that 1) midocean ridges are powerful internal tide generators and 2) internal tides may transport the tidal energy over long distances (Dushaw et al. 1995; Ray and Mitchum 1996, 1997). The barotropic tidal energy scattered into internal tides is distributed into a set of freely propagating orthogonal baroclinic modes, dependent on a few dimensionless

Supplemental information related to this paper is available at the Journals Online website: <http://dx.doi.org/10.1175/JPO-D-15-0105.s1>.

Corresponding author address: Zhongxiang Zhao, Applied Physics Laboratory, University of Washington, 1013 NE 40th St., Seattle, WA 98105.
E-mail: zzhao@apl.washington.edu

parameters (Garrett and Kunze 2007, and references therein). High-mode waves are prone to dissipate in the vicinity of the conversion sites because of their low group velocity and high shear. Low-mode waves, on the other hand, may propagate hundreds to thousands of kilometers, carrying the majority of the baroclinic energy away from the conversion sites. The long-range propagation and evolution of internal tides has been the subject of a few recent field experiments (e.g., Alford et al. 2007; Nash et al. 2007; Mathur et al. 2014; Pinkel et al. 2015). Where and how they eventually dissipate remain open scientific questions in the oceanographic community (Rudnick et al. 2003; Alford et al. 2007; Waterhouse et al. 2014).

Internal tides have isopycnal displacements of $O(10)$ m in the ocean interior, with horizontal currents $O(1\text{--}10)\text{ cm s}^{-1}$, comparable to the barotropic tidal currents (Munk 1981). Internal tides affect a wide range of ocean processes of various spatiotemporal scales, such as vertical nutrient transport (e.g., Sharples et al. 2007), underwater sound transmission (e.g., Worcester et al. 2013), regional ecosystems (e.g., Jan and Chen 2009; Kurapov et al. 2010), and continental slope shaping (Cacchione et al. 2002). In particular, it is widely believed that internal tides provide significant mechanical energy for the abyssal ocean mixing that is the driving force of the global meridional overturning circulation (MOC) (Munk and Wunsch 1998; Webb and Sugimoto 2001; Wunsch and Ferrari 2004). The global MOC and climate are sensitive to the magnitude and geography of diapycnal mixing caused by internal wave breaking (Samelson 1998; Simmons et al. 2004; Jayne 2009; Melet et al. 2013). Therefore, it is important to better understand the generation, propagation, and dissipation of internal tides on the global scale.

Observing global internal tides is a challenging task for the following reasons. First, internal tides have much smaller horizontal scale than the barotropic tide. The first mode M_2 internal tide has a wavelength 100–200 km (appendix A), and higher modes have even shorter wavelengths. In addition, internal tides have rich vertical modal structures. Thus, both horizontally and vertically high resolution is required for quantifying internal tides. Second, there are usually multiple internal tidal waves at one location, and the complicated interference pattern makes it difficult to interpret single-station field measurements (Terker et al. 2014). Third, the internal tide field is temporally variable because the generation and propagation of internal tides are modulated by time-varying ocean environmental parameters such as ocean stratification, currents, and mesoscale eddies (Mitchum and Chiswell 2000; Zilberman et al. 2011; Nash et al. 2012; Zaron and Egbert 2014). It is thus difficult to

compare internal tide measurements from different observational periods.

Previous observations of internal tides have mainly been via field measurements of the internal tide-induced temperature, salinity, and velocity fluctuations (e.g., Wunsch 1975; Hendry 1977; Kunze et al. 2002). Acoustic tomography measures sound speed (thus travel time) fluctuations induced by internal tides (Dushaw et al. 2011). However, because of their high expense and logistical difficulty, the currently accumulated database of field measurements is insufficient for global internal tide mapping (Alford 2003). Fortunately, internal tides can be detected from their centimeter-scale sea surface height (SSH) fluctuations (Munk et al. 1965). Satellite altimetry thus provides a revolutionary technique for observing global internal tides from space (Ray and Mitchum 1996). Kantha and Tierney (1997, hereinafter KT97) estimated the global distribution of M_2 internal tidal energy and reported a global integration of 50 PJ ($1\text{ PJ} = 10^{15}\text{ J}$). They did not extract information on the internal tide's spatial propagation such as phase, horizontal propagation direction, and energy flux. These important quantities are now provided in our study.

Global mapping of internal tides from satellite altimetry has been hampered by the coarse sampling of altimeter satellites both in time and space and by the complex nature of the global internal tide field. To address these issues, a two-dimensional plane wave fit method has been developed (Ray and Cartwright 2001; Zhao and Alford 2009; Zhao et al. 2012). In this study, we apply this mapping technique to the SSH measurements from multiple altimeter satellites and construct a global map of M_2 internal tides. Like all other satellite altimetric internal tide products (KT97; Ray and Cartwright 2001; Dushaw et al. 2011), our results represent a 20-yr coherent field, neglecting the incoherent component resulting from temporal variability.

2. Data

In this study, we use the combined SSH measurements made by multiple altimeter satellites *European Remote Sensing Satellite 2* [ERS-2 (E2)], *Envisat* (EN), *TOPEX/Poseidon* (TP), *Jason-1* (J1), *Jason-2* (J2), and *Geosat Follow-On* (GFO). The SSH measurements are along four sets of satellite ground tracks, which are referred to as TPJ, TPT (TP tandem mission), ERS, and GFO, respectively (Fig. 1). Among them, the TPJ dataset is about 20 years long from 1992 to 2012, consisting of the SSH data from TP, J1, and J2 (Fig. 1, red). TPT consists of the SSH data along the interleaved tandem tracks of TP and J1, which are halfway between their original

FIG. 1. The spatial and temporal coverage of multisatellite altimeter data. (a) The spatial coverage. Shown are ground tracks of TPJ (red) and TPT (blue). Ground tracks of ERS (brown) and GFO (green) are not shown in (a), but are shown in two subregions at (c) high and (d) low latitudes. In (c) and (d), the black boxes indicate one wavelength of the local mode-1 M_2 internal tide, and the gray boxes the 160-km fitting window used in this study. Multiple satellites have denser ground tracks. (b) The temporal coverage. The observational periods along four sets of satellite ground tracks. The numbers of accumulated repeat cycles are given in parentheses.

tracks (blue). ERS includes 17 years of SSH data from E2 and EN (brown). GFO is about 7 years long from 2003 to 2009 (green). The accumulated observational time is about 50 satellite years. According to the Rayleigh criterion, all of these datasets are long enough to separate M_2 from other principal tidal constituents (Zhao et al. 2011; Ray and Zaron 2016). All SSH measurements have been processed by applying standard corrections for atmospheric effects, surface wave bias, and geophysical effects (AVISO 2012). The barotropic tide and loading tide are corrected using the global ocean tide model Global Ocean Tide 4.7 (GOT4.7; Ray 2013).

Globally there are 254, 254, 488, and 1002 ground tracks for TPJ, TPT, GFO, and ERS, respectively. Each dataset has sufficient along-track resolution (6–7 km) for resolving internal tides, but cross-track spacing varies among the different track patterns, with even the smallest spacing of 80 km for ERS (at the equator) being large relative to the internal tide wavelength. At each along-track point, tidal constants can be derived by point-wise harmonic analysis. Internal tides at off-track points must be inferred from neighboring on-track measurements (Figs. 1c,d). The large cross-track spacings rule out two-dimensional bilinear interpolation as a proper mapping method for any single satellite. Denser ground tracks of multiple satellites may just meet the

requirements for two-dimensional interpolation (Ray and Zaron 2016). However, the spatial scales of a multiwave interference pattern are smaller than those of one single wave (e.g., Rainville et al. 2010; Nash et al. 2012), and two-dimensional interpolation does not resolve the amplitude, phase, and direction of individual waves.

3. Methods

a. Two-dimensional plane wave fit

In the open ocean, a significant fraction of the internal tide is both temporally and spatially coherent (Ray and Mitchum 1996; Dushaw et al. 2011), as evidenced by the pronounced peaks in the frequency and wavenumber spectra of satellite SSH data (appendix C). We thus have evolved a two-dimensional plane wave fit method to extract internal tides (Ray and Cartwright 2001; Zhao et al. 2011). Plane wave fitting is a variant of traditional point-wise harmonic analysis. In this method, internal tides are extracted using all the available SSH measurements in a small window, instead of at one single site. Windows of 160 km $\times$ 160 km are employed in this study (Fig. 1, gray boxes). Internal tide parameters (amplitude A_m , phase ϕ_m , and direction θ_m) are to be determined by fitting plane waves to satellite SSH time series. There may be multiple internal tidal waves at any

FIG. 2. An example of the plane wave fit method. (a) A 160-km fitting window centered at 31°N, 196°E, showing all data locations in this region. (b) Amplitude (mm) vs direction obtained by the least squares fit. The first M_2 internal tidal wave is determined from the amplitude and direction of the biggest lobe (red arrow). (c) Residual variance (mm^2) vs direction in the least squares fit. The first M_2 internal tidal wave can be determined by residual minimum (red arrow). Amplitude maximum and residual minimum are identical in determining internal tidal waves. (d),(e) After removing the first internal tidal wave from the original SSH data, this procedure is repeated to determine the second M_2 internal tidal wave (blue arrow). (f),(g) As in (d) and (e), but for the third wave (green arrow). (h),(i) As in (d) and (e), but for the fourth wave (pink arrow). (j) Variances explained by these four M_2 internal tidal waves. (k) Amplitudes of these four M_2 internal tidal waves. The fourth wave is discarded because its amplitude is lower than the 95% CL. (m) Each internal tidal wave is refitted with the other waves temporarily removed, to reduce the cross-wave interference. Shown here are the finally determined three mode-1 M_2 internal tidal waves.

given site; therefore, harmonic constants of multiwave superposition follow

$$\text{SSH}(x, y, t) = \sum_{m=1}^3 A_m \cos(kx \cos \theta_m + ky \sin \theta_m - \omega t - \phi_m), \quad (1)$$

where x and y are the east and north in the Cartesian coordinates, and ω and k are the M_2 frequency and wavenumber, respectively. Wavenumber and phase speed can be theoretically calculated from climatological ocean stratification (appendix A). A previous study in the western Pacific Ocean reveals that the satellite-derived phase speeds agree with theoretical values (Zhao 2014).

Figure 2 illustrates the determination of the mode-1 M_2 internal tide at an off-track point (31°N, 196°E). The SSH measurements falling into a fitting window centered at 31°N, 196°E are used (Fig. 2a). The procedure is as follows. First, in each compass direction (angular increment is 1°), the amplitude and phase of one single plane wave are determined by the least squares fit. When the resultant amplitudes are plotted as a function of direction in polar coordinates, an internal tidal wave appears to be a lobe (Fig. 2b). The amplitude and direction of the first M_2 wave are thus determined from the biggest lobe (red arrow). Figure 2c shows the residual variance versus direction. Similarly,

the M_2 internal tide can be determined from the residual minimum. Amplitude maximum and residual minimum yield the same M_2 internal tide (Figs. 2b,c). Note that some lobes may be artifacts of the antenna rather than actual waves, because of the irregular distribution of multisatellite ground tracks. Side lobes are a ubiquitous feature in a wide range of research fields such as radar science and seismology (e.g., Rost and Thomas 2002; Chandran 2013). In this study, we predict the signal of the above determined wave and remove it from the original SSH data. This step removes the wave itself and suppresses its side lobes. This procedure can be repeated to extract an arbitrary number of waves. In this example, we repeat it three times to determine three more M_2 internal tidal waves. Finally, each wave is refitted with other waves temporarily removed, in order to reduce the cross-wave interference.

The 95% confidence level (CL) is calculated using the standard formula for linear regression of Gaussian-distributed data (via MATLAB's built-in function "regress") and shown in Fig. 2k. The 95% CL is only about 2 mm, because there are typically $O(10^4)$ independent SSH measurements in one fitting window. For comparison, point-wise harmonic analysis uses a time series of $O(10^2)$ SSH measurements at one single point. The fourth M_2 internal tide is discarded, because its amplitude is lower than the 95% CL (Fig. 2k). In the end, we obtain three mode-1 M_2 internal tidal waves, whose superposition gives the final internal tidal solution at 31°N, 196°E (Fig. 2m).

b. Parameters in the global mapping

Several parameters are subjectively determined in our global internal tide mapping. They may not be universally optimal choices because of the complexity of global internal tides. This study aims at a global view of open-ocean M_2 internal tides; therefore, we choose these parameters for robust mapping throughout the world oceans. Future improvements can be made by optimizing these parameters region by region.

1) HIGH-PASS FILTERING

A high-pass filter with cutoff wavelength 500 km is used to remove large-scale nontidal signals (appendix C). This step is necessary, because the orbit errors in altimeter missions cause large uncertainties in the fitted internal tides. In previous studies, high-pass filters with different cutoff wavelengths have been used (Dushaw 2002; Ray and Zaron 2011). Because the satellite tracks are generally in the south–north direction, a high-pass filter will suppress internal tides in the east–west direction. Zhao and D'Asaro (2011) showed that such westbound M_2 internal tides can be plane wave fitted

without prior high-pass filtering, but the noise level is higher.

2) POINT-WISE HARMONIC ANALYSIS

Prior point-wise harmonic analysis is not required by our plane wave analysis, because plane wave fitting extracts coherent waves both in time and space. We have tested mapping M_2 internal tides using two different datasets: the high-passed SSH data with/without prior point-wise harmonic analysis (appendix C). The results are almost the same. In this study, we use the previously harmonically fitted SSH data (appendix D) in order to better compare M_2 internal tides obtained by these two methods.

3) WINDOW SIZE

There is a trade-off in the size of the fitting window, over which we assume a uniform wave field and thus constant internal tide parameters. A bigger window has higher angular resolution. A smaller one has better spatial resolution (Zhao et al. 2011). We have tested different window sizes and chosen a 160-km fitting window. First, it is close to one wavelength of open-ocean mode-1 M_2 internal tides (Fig. 1, gray boxes). Second, it spans multiple ground tracks for robust plane wave fitting (Zhao et al. 2011).

All SSH data in water shallower than 500 m have been discarded. Data loss may be due to seasonal ice coverage at high latitudes. At the land–ocean boundary, we extract internal tides only if one fitting window has $\geq 50\%$ good data.

We should not conduct modal decomposition (as in appendix A) in a horizontally inhomogeneous ocean, which is mainly caused by bottom topography and boundary currents (Wunsch 1975; Kelly et al. 2012). In this study, we discard the internal tide solution at any grid point where the bottom slope is greater than 6/1000 (appendix B). Later, we will show that boundary currents induce energetic mesoscale eddies; therefore, internal tides in regions of boundary currents are also discarded using a noise level threshold (section 3c).

4) SPATIAL GRIDS

Our global mapping is conducted on a regular 0.2° latitude $\times$ 0.2° longitude grid. We have tested mapping on a $0.1^\circ \times 0.1^\circ$ grid in some regions (e.g., around the Luzon Strait), which led to a spatially high-resolution internal tide field (Zhao 2014). But it will take 4 times longer to run the global mapping.

5) THREE-WAVE FIT

We extract the three biggest internal tidal waves at each grid point. The nontidal noise in altimeter data

prevents us from extracting more waves. In Fig. 2, the fourth wave is discarded because its amplitude is lower than the 95% confidence level. In appendix E, we assess skill of the three-wave fit using the modeled internal tide data (without eddy noise) and find that the first three internal tidal waves account for about 95% of the total SSH variance.

6) LATITUDINAL RANGE

Our global mapping is within the latitudinal range 60°S–60°N. The ratio of the sea surface to interior amplitude of an M_2 internal tidal wave decreases poleward (appendix A). At high latitudes, an internal tide of the same interior amplitude has a smaller SSH amplitude and is more difficult to observe by satellite altimetry. According to theoretical and numerical models (Nycander 2005; Simmons 2008), there are no strong M_2 internal tides poleward of 60°S/N. Our latitudinal range is far away from the M_2 internal tide's turning latitudes 75°S/N. At the turning latitudes, the SSH fluctuation follows the Airy function (Dushaw 2006) and is also affected by the near-surface stratification and mixed layer (Wunsch 2013).

c. Variance analysis

In this section, we evaluate the performance of the plane wave fit method by comparing the original and residual variances. Figure 3a shows the variance of the harmonically fitted M_2 internal tides, extracted from all four datasets by harmonic analysis (appendix D) and spatially averaged in 160-km windows. Figure 3a shows that large variance appears mainly in two types of regions. The first is around major topographic features such as the Hawaiian Ridge and the French Polynesian Ridge. The second is in boundary currents such as the Gulf Stream, the Kuroshio Current, and the Antarctic Circumpolar Current (ACC), where the variances are mainly noise from the leakage of mesoscale eddies (appendix D).

In this study, the previously harmonically fitted M_2 internal tides are refined by the plane wave fit method. Figure 3b shows the total SSH variance of three M_2 internal tidal waves obtained by plane wave fitting. It reveals that the large variance appears mainly in the vicinity of topographic features, suggesting that they are real internal tide signals. Figure 3c shows the residual variance, that is, variance with temporal coherence but spatially inconsistent with a plane wave. The residual is mainly from leaked mesoscale eddies associated with boundary currents and high-mode M_2 internal tides near bottom topographic features.

Figures 3b and 3c have different spatial patterns, suggesting that mode-1 M_2 internal tides have been

successfully isolated from mesoscale eddies. The ratio of the plane wave fitted (Fig. 3b) to harmonically fitted (Fig. 3a) M_2 internal tides is shown in Fig. 3d, revealing that high ratios appear in regions of strong internal tides and weak mesoscale eddies.

In regions of extremely high mesoscale eddies, there is substantial noise from the broadband continuum spectrum. Our internal tide solutions in these regions are overwhelmed by mesoscale contamination. For the rest of the paper, regions with residual variance greater than 350 mm^2 are masked out (Fig. 3c, light blue; about 8% of the global ocean area). These regions include the Kuroshio, the Gulf Stream, the East Australian Current, the Agulhas Current, the Brazil Current, the Leeuwin Current, the loop current in the Gulf of Mexico, and the ACC. Xu and Fu (2012) label these high mesoscale regions as the type-1 or -2 regions, where the spectral slope is $-11/3$ or steeper (see their Fig. 4).

It is worth noting that the ERS and GFO data have been excluded in previous studies because of their high noise levels (e.g., Dushaw et al. 2011). Here their noise levels are reduced by the plane wave fit method. Higher spatial resolution is thus achieved by merging all four SSH datasets from multiple-mission satellite altimetry.

4. Global open-ocean mode-1 M_2 internal tide

a. Generation sources

Figure 4 presents the global maps of amplitude and phase of open-ocean mode-1 M_2 internal tides (i.e., three-wave superposition). We can better present the global internal tide field using the separately resolved internal tidal waves. In this study, we empirically divide the internal tide field into the northbound ($0^\circ < \theta < 180^\circ$) and southbound ($180^\circ < \theta < 360^\circ$) components, respectively. At one grid point, only the biggest wave is kept in each direction range. As shown in Fig. 5, the previously overlapped southbound and northbound internal tidal waves are now isolated. Thus, we can identify a number of internal tidal beams and determine their generation sites from their starting points. By an internal tidal beam, we mean the horizontal radiation of internal tides that have larger amplitude than background and linearly increasing phase with propagation.

Note that the southbound and northbound separation here is arbitrary—regional maps could be presented in any direction depending on the source's alignment and location. For example, in the western North Pacific Ocean, M_2 internal tides generally propagate in the east–west direction, so that the south/north separation is misleading (Fig. 5). In this region, we thus divide the field into the eastbound and westbound components (Fig. 6). We can

FIG. 3. Variance analysis of the plane wave fit method. (a) Variance of the harmonically fitted M_2 internal tide smoothed in 160-km windows. (b) Variance explained by the plane wave fitted M_2 internal tidal waves. (c) Residual variance. The light blue contours mark regions with high mesoscale eddies (residual variance $> 350 \text{ mm}^2$). (d) Ratio of the plane wave fitted (b) to harmonically fitted (a) M_2 internal tides. In all panels, the 3000-m isobath contours are in black.

FIG. 4. Global mode-1 M_2 internal tides from multisatellite altimetry. (a) SSH amplitude. (b) Phase. The 3000-m isobath contours are in black. The light blue masks the high mesoscale regions (see Fig. 3c). The spatial resolution is limited by the page size. A high-resolution figure is available as supplementary Fig. S1. The data file is available on request.

observe that internal tidal beams originate at topographic sites such as the Luzon Strait and the Izu-Bonin Ridge.

The satellite altimetric maps show that M_2 internal tides are ubiquitous in the world oceans and that they are geographically inhomogeneous (Figs. 4, 5). There are two relatively quiet regions: the eastern equatorial Pacific Ocean and the equatorial Indian Ocean. It is known that internal tides are mainly generated over strong topographic features. To demonstrate this point, the 3000-m isobath contours are superimposed on these maps (Figs. 4–6). These figures show that internal tidal beams mainly originate at tens of strong generation sites over topographic features. These generation sources have been previously identified in barotropic tide divergence (Egbert and Ray 2000, 2001), linear theory (Nycander 2005), and numerical models (Simmons et al. 2004; Arbic et al. 2010).

Baines (1982) estimated the semidiurnal internal tide generation along 230 continental sections worldwide using linear generation models and identified 12 major generation regions, nine of which can be confirmed in our satellite observations (different names may be used). They are the continental slope of western Europe,

the continental slope of East Asia (including the East China Sea), the continental slope of northeastern Brazil (around the Amazon River mouth), the continental slope of northwestern Australia (including the Timor Sea), the continental slope of southwestern India, the continental slope of Bangladesh and Burma (in the Bay of Bengal), the Gulf of Panama, the Gulf of Alaska, and the Mozambique Channel. In addition, the Mid-Argentine shelf and the Grand Banks of Newfoundland (including Cape Cod and Cape Sable) are selected out by Baines (1982); however, our internal tide solutions in these regions are fouled by mesoscale contamination. There is one exception: the North Bering Sea is estimated to be a strong generation site by Baines (1982), but is weak in our observations. Satellite altimetric results have thus largely confirmed the theoretical work of Baines (1982). Generation sites over continental slopes can also be observed in global numerical simulations (e.g., Arbic et al. 2012). However, the Baines (1982) model probably underestimated the global barotropic to baroclinic tidal conversion (14.5 GW), because it did not take into account along-slope barotropic tidal currents over the continental slope's three-dimensional features

FIG. 5. Global maps of the decomposed mode-1 M_2 internal tides. (a),(b) Amplitude and phase of the northbound component (propagation direction is within 0° – 180°). (c),(d) As in (a) and (b), but for the southbound component (propagation direction is within 180° – 360°). The 3000-m isobath contours are in black. The light blue masks the high mesoscale regions (see Fig. 3c).

FIG. 6. Snapshot SSH fields of mode-1 M_2 internal tides in the western North Pacific. (a) Westbound component (propagation direction is within 90° – 270°). (b) Eastbound component (propagation direction is within 270° – 360° or 0° – 90°). The 3000-m isobath contours are in black. The light blue masks the high mesoscale regions (see Fig. 3c). Remarkable generation sources include the Luzon Strait (LS), the Izu-Bonin Ridge (IB), the West Mariana Ridge (WM), the Mariana Arc (MA), and the Ryukyu Ridge.

such as canyons and escarpments (Cummins and Oey 1997; Lien and Gregg 2001; Althaus et al. 2003).

More recent studies revealed that midocean topographic features, such as submarine ridges, island chains and seamounts, are even stronger generation sources of internal tides (Morozov 1995; Ray and Mitchum 1996). Morozov (1995) listed about 40 ridges with strong M_2 internal tide generation using linear theoretical models and real bottom topography. Note that internal tide generation is sensitive to the spatial resolution of bottom topography (Kunze and Llewellyn Smith 2004; Carter et al. 2008; Niwa and Hibiya 2011). Morozov's generation map is limited by the low-resolution bottom topography used at that time. Nycander (2005) identified generation hotspots using 2-arc-min global topography (Smith and Sandwell 1997). It was later revealed that linear theory does not hold for steep submarine ridges (Pétrellis et al. 2006). Although internal tides are generated over most submarine topographic features, a number of strong generation sites (or sections) account for the bulk of global internal tide generation (Simmons et al. 2004; Niwa and Hibiya 2014).

In the Pacific Ocean, we can identify four strong generation regions: 1) the Hawaii region, including the Hawaiian Ridge, the Line Islands Ridge, and neighboring topographic features; 2) the French Polynesian Ridge in the South Pacific; 3) the western North Pacific, including a number of strong generation sites such as the Izu-Bonin Ridge, the Mariana Arc, and the Ryukyu Ridge (Fig. 6); and 4) the western South Pacific, covering submarine ridges and straits of Australia and New Zealand. Each of the four regions covers a large area, with many discrete generation hotspots. These regions also stand out clearly in global internal tide numerical model simulations (Shriver et al. 2012). There are several other energetic but isolated generation sites: the Mendocino Escarpment off the U.S. West Coast (e.g., Althaus et al. 2003), the east Pacific Rise, the Galapagos Oceanic Plateau in the east Pacific, the Easter Fracture zone along 25°S in the east Pacific Ocean, the Macquarie Ridge to the south of New Zealand, the Aleutian Ridge (also known as the Aleutian Islands chain), and the Kuril Strait connecting the Sea of Okhotsk and the

North Pacific Ocean. Note that the Emperor Seamounts Chain is generally in the north–south direction (along 170°E) and appears to be a weak generation site (Fig. 5). It could be underestimated, because our mapping tends to suppress the westward/eastward internal tides. In addition, there is strong M_2 internal tide generation in narrow straits in the Indonesian Archipelago.

In the Indian Ocean, most of the internal tide generation is in the Madagascar–Mascarene region. It consists of numerous generation sites, consistent with model simulations (Simmons 2008; Shriver et al. 2014). However, the nearby Carlsberg Ridge and the Mid-Indian Ridge generate weak internal tides. The Chagos-Laccadive Ridge is a noticeable generation source. The M_2 internal tides originate in several channels connecting the Bay of Bengal and the Andaman Sea. In the southern Indian Ocean, the Kerguelen Plateau (55°S, 71°E) is a remarkable generation site (e.g., Maraldi et al. 2011). However, the radiating internal tidal beams are quickly lost in the ACC. Another generation site is at the southern tip of the Southwest Indian Ridge (40°S, 45°E). However, the Ninety East Ridge does not induce much internal tide activity.

Internal tides in the Atlantic Ocean have long been studied (Hendry 1977; Pingree et al. 1986; Polzin et al. 1997; New and Da Silva 2002; van Haren 2004). Our satellite observations suggest that the Atlantic Ocean is full of M_2 internal tides. There exist numerous generation hotspots at the Mid-Atlantic Ridge (MAR), the Walvis Ridge in the southern Atlantic Ocean, and channels connecting the Caribbean Sea and the North Atlantic Ocean. Other remarkable generation sites include the Rio Grande Rise (31°S, 35°W), the Cape Rise Seamount (42°S, 15°E), the Azores Plateau in the North Atlantic Ocean, the Cape Verde Islands off West Africa, and the Sierra Leone Rise.

In summary, our satellite observations reveal that M_2 internal tides are ubiquitous in the world oceans. Strong generation regions (or sites) are observed and generally consistent with previous studies by barotropic tidal dissipation (Egbert and Ray 2000), theoretical (Baines 1982; Morozov 1995; Nycander 2005), and numerical (Simmons et al. 2004; Shriver et al. 2014) internal tide models.

b. Internal tidal beams

Discrete internal tidal beams can be clearly identified in the separately resolved internal tide maps (Fig. 5). Along each beam, phase increases linearly with propagation (Figs. 5b,d). In contrast, the internal tide's propagation direction cannot be well recognized in the three-wave superposed phase map (Fig. 4b). In the western North Pacific, westbound and eastbound beams are observed.

Most of the internal tidal beams do not spread much with propagation, and their widths are typically in the 100–300-km range. Some beams spread radially in propagation, such as the southbound beam from the Aleutians, whose width increases from about 300 to 2000 km over a propagation distance of 2000 km. Strong radial spreading is usually associated with point generation sources, such as the southbound internal tides from the Marquesas Islands (9°S, 139.5°W) in the South Pacific Ocean.

A number of internal tidal beams may travel across the ocean basins and reach the opposite continental slopes. This implies that a fraction of internal tidal energy eventually dissipates on the continental slopes, or loses coherence as they reflect (Alford and Zhao 2007a,b; Nash et al. 2012). For example, the northbound Hawaiian internal tides reach Alaska and the southbound Aleutian internal tides reach Hawaii (Zhao et al. 2012). In the Tasman Sea, an internal tidal beam from the Macquarie Ridge propagates over 1600 km and hits the eastern Tasmanian slope (Pinkel et al. 2015). In the Indian Ocean, the northwestward internal tides from the Mascarene Ridge may reach the Somali coast.

c. Internal tide energetics

The depth-integrated energy and flux of mode-1 M_2 internal tides can be computed from the SSH amplitude using the transfer functions derived from internal tide dynamics (appendix A). Both energy and flux are proportional to the SSH squared. Following these relations, energy and flux are computed and presented in Figs. 7a and 7b. They are the scalar (energy) and vector (flux) sum of the three separately resolved waves. Therefore, their spatial patterns are very similar, with some differences in regions of strong multiwave interference.

To investigate their latitudinal distributions, we calculate the zonally integrated energy and flux (Fig. 8). The earth's spherical shape has been taken into consideration in the zonal integration. The total energy is given in petajoules (10^{15} J) per degree (kinetic energy and potential energy; appendix A). The energy fluxes projected in the north and south directions are given separately (Figs. 8b,c). Note that the eastward and westward transports of the tidal energy are not shown here. The zonal integrations are made for the Atlantic, Indian, and Pacific oceans separately (Fig. 8, color lines). Figure 8 shows that energetic M_2 internal tides mainly occur at mid latitudes, consistent with their generation sites (see Fig. 4). Both energy and flux decrease poleward and equatorward from their generation sources. Figure 8 also shows that the equatorial zone (in particular the Pacific Ocean) is relatively quiet, mainly because of the lack of strong sources, but also partially because of the

FIG. 7. Global maps of depth-integrated (a) energy and (b) flux magnitude of mode-1 M_2 internal tides. The 3000-m isobath contours are in black. The light blue masks the high mesoscale regions (see Fig. 3c). A zoomed-in map in the green box is given in Fig. 9.

loss of coherence while internal tides propagate across the equator. In the equatorial Atlantic Ocean, there is strong M_2 internal tide generation along the northeast Brazilian continental slope (Fig. 4); therefore, outstanding peaks appear in energy and flux (Fig. 8).

Figure 8 shows that both the southbound and northbound internal tides sharply decrease when approaching the equator, so that not much internal tidal energy is transported from one hemisphere to the other. Figure 7 shows that the “equatorial barrier” mainly appears in the eastern Pacific Ocean. In the Atlantic, Indian, and western Pacific oceans, M_2 internal tides can travel across the equatorial zone with little energy loss. To better understand this process, we thus examine the cross-equator M_2 internal tides in the eastern Pacific Ocean (Fig. 9). The northeastward M_2 internal tides originate at the French Polynesian Ridge and travel across the equator. From 10°S to the equator, energy flux decays exponentially (Fig. 9d). At least a small fraction of energy flux (5% of the flux at 10°S) is transported across the equator. The significant energy loss may be due to 1) real dissipation or 2) undetectable incoherent internal tides. Specifically designed field experiments are needed to quantify the contributions from these mechanisms.

The M_2 internal tide may lose a significant fraction of energy at its critical latitudes (28.8°S/N ; MacKinnon and Winters 2005), whereby the M_2 tidal energy goes to shorter waves of the local Coriolis frequency by parametric subharmonic instability (PSI). Recent field observations reveal that PSI may significantly modulate the shear structure at the critical latitudes (Alford et al. 2007), but it does not deposit much of the tidal energy (MacKinnon et al. 2013). In our satellite results, no enhanced energy loss is observed at the critical latitudes (Figs. 8b,c), consistent with field measurements by MacKinnon et al. (2013).

We next compare our latitudinal distribution and global integration of M_2 energy with those of KT97 (their Fig. 6). Note that KT97 employed 4.5 years of TP data while we used 50 satellite years of multisatellite altimeter data. Our global integration is 36 PJ, compared to 50 PJ in KT97. Although rigorous error bars are difficult to calculate, we believe our estimate is more accurate, because of the bigger dataset (50 versus 4.5 satellite years). KT97 employed a globally uniform two-layer ocean model, thus lacking geographic variation in transfer function from SSH to energy (see appendix A). In addition, KT97 extracted M_2 internal tides using point-wise harmonic analysis, which may overestimate

FIG. 8. The latitudinal distribution of zonally integrated energy and flux. The integrations in the Indian, Pacific, and Atlantic Oceans are shown by color lines: (a) energy, (b) the north-bound energy flux, and (c) the southbound energy flux.

energy due to mesoscale contamination (appendix D). However, neither estimate contains the incoherent internal tide, which is hidden from satellite altimetry. The global internal tidal energy of all constituents has been estimated to be 100 PJ (see Fig. 5 in Wunsch and Ferrari 2004). For comparison, the global barotropic tidal energies are about 392 PJ for M_2 and 584 PJ for all tidal constituents (see Table 1 in Kantha 1998).

Residence time R of the mode-1 M_2 internal tide in the world oceans can be estimated from the globally integrated M_2 tidal energy E and the input rate from the barotropic tide W following $R = E/W$. Egbert and Ray (2001) estimated that the global total generation rate of the M_2 internal tide is 0.7 TW. The mode-1 M_2 input rate would be 0.42–0.56 TW, assuming mode 1 accounts for 60%–80% (Egbert and Ray 2003; Garrett and Kunze 2007; Zhao et al. 2010). We thus obtain a residence time of about 1–1.5 days, which suggests that, on global average, the M_2 internal tide dissipates within 400 km from the generation source (assuming a 160-km wavelength). However, some long-range internal tidal beams can transport the tidal energy over 3000 km (Zhao et al. 2010).

d. Interference pattern

The multiwave superposition forms a complex interference wave field (Nash et al. 2006; Martini et al.

2007), with a regular spatial pattern of nodes and antinodes. The energy and flux calculations from single-station measurements are misleading (Rainville et al. 2010; Zhao et al. 2010). The three-wave fit from satellite altimeter data enables us to evaluate the multiwave interference in the world oceans following Alford and Zhao (2007b). We quantify the degree of multiwave interference using the ratio $|\Sigma \mathbf{F}_n|/|\Sigma |\mathbf{F}_n||$, where $\mathbf{F}_n$ ($n = 1, 2, 3$) are the flux vectors of the separately resolved waves. The ratio ranges from 0 to 1, with 0 denoting a fully standing wave and 1 a pure progressive wave.

Figure 10 reveals the widespread multiwave interference of some degree in the world oceans. There usually exist multiple internal tidal waves at one location, likely because of the internal tide's numerous generation sources and long-range propagation. For example, in the central North Pacific, the southbound internal tides from the Aleutian Ridge and the northbound internal tides from the Hawaiian Ridge cause an interfering internal tide field (Fig. 10, blue box). A mooring deployed at 40°N, 198°E (Fig. 10, blue circle) observed an eastward flux (Alford and Zhao 2007a). In the case that one internal tide is dominant, the field is close to a pure progressive wave (Fig. 10). Thus, the single-station flux measurements are more informative in the case of one dominant wave (Kunze et al. 2002; Althaus et al. 2003; Lee et al. 2006).

FIG. 9. The cross-equator mode-1 M_2 internal tides in the eastern Pacific Ocean. See Fig. 7b for location. (a) SSH amplitude, (b) phase, (c) energy flux, and (d) zonally averaged flux due north. The northeastward internal tides from the French Polynesian Ridge decrease exponentially in the equatorial zone.

Alford and Zhao (2007b) diagnosed standing waves of the semidiurnal internal tide using mooring measurements. They examined the ratio of the moored flux (F) to energy (E) and compared it with the group velocity (C_g). Because of the influence of the transverse energy flux (see Fig. A1 in Alford and Zhao 2007b), the mooring method can only be applied in the deep ocean equatorward of 35°S/N . Thus, only 20–30 moorings are qualified for the diagnosis. Their results revealed both standing and progressive waves (see the top panel of Fig. 3 in Alford and Zhao 2007b). Thus, both satellite altimetry and field moorings suggest the widespread multiwave interference.

5. Comparison with moored observations

In this section, we assess the satellite altimetric internal tides using field mooring measurements. Alford and Zhao (2007a,b) examined a database of 2200 historical moorings and found that 80 moorings are qualified for calculating the internal tide energy and flux. Mooring instrumentation and deployment are listed in Table 1 of Alford and Zhao (2007a). Because most of the selected moorings have fewer than six instruments, uncertainties mainly result from vertical undersampling.

Moorings lying in the energetic mesoscale regions are not used in this study, because the satellite altimetric observations are fouled by mesoscale contamination (see section 3c).

In the 2006 Internal Waves across the Pacific (IWAP) experiment, six moorings were deployed in an internal tidal beam radiating from French Frigate Shoals, Hawaii (Alford et al. 2007). Each IWAP mooring was equipped with a McLane profiler, which crawled the cable from about 50 to 1450 m. These moorings have high vertical resolution in the upper 1450-m layer, equivalent to 62%–68% of the Wentzel–Kramers–Brillouin (WKB) stretched vertical coordinate, allowing robust and reliable resolution of the first five vertical modes (Zhao et al. 2010).

The SSH amplitude and phase are calculated from the mooring measured interior isopycnal displacements (appendix A; Zhao et al. 2010). Satellite altimetric amplitude and phase are determined using SSH data in a window centered at each mooring site. The comparison between the altimetric and moored M_2 internal tides is shown in Fig. 11.

Overall, the altimetric and moored results are in fair agreement. The amplitude and phase root-mean-square

FIG. 10. The degree of multiwave interference, described by $|\Sigma \mathbf{F}_n|/|\Sigma |\mathbf{F}_n||$, where $\mathbf{F}_n$ ($n = 1, 2, 3$) are energy fluxes of the plane wave fitted internal tidal waves. The blue box denotes a region of strong multiwave interference, as revealed by field measurements at one mooring site (blue circle). This map has been spatially smoothed by a $1^\circ \times 1^\circ$ sliding window. The 3000-m isobath contours are in black. The light blue masks the high mesoscale regions (see Fig. 3c).

(RMS) differences are 3 mm and 72° , respectively. Disagreement between the two products is understandable, because the altimetric results are 20-yr coherent internal tides in a spatial window, and the moored results are 0.5–2-yr-long coherent internal tides at single points.

Using moorings with SSH amplitude greater than 5 mm, the phase RMS difference is reduced to 40° . Furthermore, the IWAP moorings are along a strong internal tidal beam from Hawaii (Zhao et al. 2010), and their phase RMS difference is only 23° . Their amplitude RMS difference is 5 mm, compared to their average amplitude of 13 mm. The altimetric SSH amplitudes bias lower than the IWAP measurements (Fig. 11a), likely because of their different data lengths.

In summary, the satellite altimetric M_2 internal tides agree fairly well with the mooring measurements. We refrain from drawing a stronger conclusion, because of the small number of IWAP-type field moorings. Similarly, Chiswell (2006) compared the first mode M_2 internal tide from altimeter and current meter measurements and found good agreement in phase and a factor-of-2 difference in amplitude. In the future, it may be useful to compare the satellite results with the internal tide measurements from pressure inverted echo sounders (PIES) and long-term sea level time series from tide gauges (Mitchum and Chiswell 2000; Colosi and Munk 2006).

6. Comparison with numerical simulations

Internal tides have been simulated by global high-resolution, eddy-allowing numerical models (e.g., Simmons et al. 2004; Arbic et al. 2010; Müller et al. 2012;

Waterhouse et al. 2014). These models are forced by atmospheric fields and astronomical tidal potential, so that they can simulate the large-scale ocean circulation, mesoscale eddies, and internal tides simultaneously. These models can be sampled with high spatial and temporal resolution and resolve the incoherent internal tides missed by satellite altimetry.

In this study, we compare the satellite altimetric M_2 internal tides with the Generalized Ocean Layer Dynamics (GOLD) model simulation. The GOLD model is an isopycnal-coordinate ocean model developed at NOAA/GFDL and evolved from the Hallberg Isopycnal Coordinate Model (HIM; Hallberg 1997). The current version of the GOLD model has 50 layers in the vertical and a regular $1/8^\circ$ Mercator grid in the horizontal (Simmons and Alford 2012; Waterhouse et al. 2014). After the 5-yr spinup, the simulation is continued from 1 January 1995 and run for an entire year. The resultant M_2 internal tides are the year-long coherent components. The first two vertical modes are obtained by projecting the harmonically fitted density and velocity fluctuations onto the known vertical modal structures (see appendix A).

Figure 12 shows the satellite altimetric and GOLD-simulated mode-1 M_2 internal tides in the central North Pacific. Here the GOLD results are from three-wave superposition analyzed by the plane wave fit method (appendix E). For better comparison, the northbound (Figs. 12b,e) and southbound (Figs. 12c,f) components are shown. Both results reveal northward internal tides from the Hawaiian Ridge and southbound internal tides from the Aleutian Ridge (Zhao et al. 2010). The overall

FIG. 11. A comparison of mode-1 M_2 internal tides derived from satellite altimetry and field moorings. (a) SSH amplitude and (b) phase. Circles denote historical moorings. Filled circles denote six IWAP moorings. Moorings with SSH amplitudes greater than 5 mm are in red.

beam patterns are very similar for both components. However, there are a number of different aspects in the altimetric and modeled internal tide fields.

We further compare the amplitude and phase along two satellite tracks TPJ 234 and 249 (Fig. 13). The green lines denote the progressive components observed from the SSH measurements along one single track. They are extracted by an along-track progressive fit method, which is similar to the two-dimensional plane wave fit method (Zhao et al. 2011). Figure 13 shows that the plane wave fitted and along-track fitted internal tides are almost identical in phase.

Along the southbound Aleutian beam, the GOLD and satellite agree very well in amplitude, with an RMS difference of 0.6 mm (Fig. 13a). However, along the northbound beam, the GOLD amplitude is about 50% greater, yielding an energy flux that is twice as large as the satellite derived flux. Forward tide models often have overly large tides and require some sort of damping to match observations. The damping typically takes the form of a linear wave drag designed to parameterize unresolved barotropic to baroclinic conversion (e.g., Jayne and St. Laurent 2001; Arbic et al. 2004; Egbert et al. 2004). GOLD does not include such a wave drag configuration but is explicitly constrained by a global barotropic to baroclinic conversion rate of 1.5 TW.

The separately resolved northbound and southbound components enable us to compare phase as well (Figs. 13b,d). In general, the GOLD and satellite phases along both tracks agree well. Along TPJ 249, there is a systematic bias, which decreases northward from about 60° at 26°N and disappears at 36°N (Fig. 13d). We suggest that the bias results from the discrepancies in ocean stratification and thus phase speed.

The gray box in Fig. 13 marks a region of weak internal tides. Along this section, both the altimetric and modeled

amplitudes are very low, about 2–3 mm, close to the 95% confidence level (see Fig. 2k). Because these tides are weak, the associated phase has large uncertainties (Fig. 13b). This exercise suggests that the threshold level for observing and simulating internal tides is about 2–3 mm.

In summary, the satellite altimetric and model-simulated M_2 internal tides are in good agreement. The GOLD model is based on a geophysical fluid dynamics solution, without data assimilation. Thus, the favorable agreement with satellite observations is encouraging. Future improvements may be achieved by adjusting model parameters for better comparison with the satellite altimetric observations.

7. Summary

In the world oceans, low-mode M_2 internal tides are not only largely temporally coherent, but also spatially coherent. This feature warrants a two-dimensional plane wave fit method for mapping global internal tides using multisatellite SSH measurements. The plane wave fit method exploits M_2 internal tides with both spatial and temporal coherence, so that it greatly suppresses mesoscale contamination. By this technique, SSH measurements made by multiple altimeter satellites are merged to achieve higher spatial resolution in the internal tide mapping. The plane wave fit method separately resolves multiple internal tidal waves of different propagation directions. The separately resolved internal tidal beams enable us to better study their generation, propagation, and dissipation. Furthermore, using the climatological annual-mean stratification profiles in the *World Ocean Atlas 2013* (WOA2013; Locarnini et al. 2013; Zweng et al. 2013), we have computed the transfer

FIG. 12. Comparisons of mode-1 M_2 internal tides in the central North Pacific derived from (top) satellite altimetry and (bottom) GOLD simulation. (a),(d) Three-wave superposition. (b),(e) Northbound component. (c),(f) Southbound component. Comparisons along tracks TPJ 234 and 249 (black lines) are presented in Fig. 13.

functions from the SSH amplitude of an M_2 internal tide to its subsurface isopycnal displacement and depth-integrated energy and flux.

We have constructed a global map of open-ocean coherent mode-1 M_2 internal tides. Our results show that M_2 internal tides are generated over numerous topographic features, including continental slopes, narrow straits, midocean ridges, and seamounts. Discrete internal tidal beams originating at these generation hotspots propagate hundreds to thousands of kilometers. Their numerous generation sites and long-range propagation make multiwave interference a widespread feature. Thus, M_2 internal tides are ubiquitous in the world oceans, with remarkable geographic inhomogeneity. The M_2 internal tide loses little energy in propagating across its critical latitudes (28.8°S/N), consistent with field observations by MacKinnon et al. (2013). It propagates across the eastern equatorial Pacific with significant energy loss; however, in the Atlantic, Indian, and western Pacific oceans, it propagates across the equatorial zone without marked energy loss. The globally

integrated energy of the mode-1 M_2 internal tide gives a lower bound estimate of 36 PJ.

We have compared the M_2 internal tides derived from satellite altimetry with field mooring measurements. The former gives the 20-yr coherent internal tides, and the latter gives 0.5–2-yr-long coherent internal tides. Thus, the altimetric and moored results are in fairly good agreement. The M_2 internal tides derived from satellite altimetry and six IWAP moorings dedicated for measuring internal tides agree very well, with amplitude and phase RMS differences of 4.8 mm and 23° . Unfortunately, the limited number of such moorings prevents us from drawing strong conclusions.

We have compared mode-1 M_2 internal tides from satellite altimetry and the GOLD model simulation. In the central North Pacific, the altimetric and modeled internal tide fields have similar geographic patterns. Perhaps because GOLD does not include strong damping mechanisms, the GOLD simulated internal tides are stronger. Additionally, there are a number of inconsistent features such as a systematic phase bias.

FIG. 13. Along-track comparisons of satellite altimetric and GOLD simulated M_2 internal tides. (a),(b) The southbound M_2 internal tides along TPJ 234. (c),(d) The northbound M_2 internal tides along TPJ 249. See Fig. 12 for their geographic locations. The green lines denote the along-track progressive wave fitted results. The gray box marks a region of weak internal tides.

8. Perspectives

Satellite altimetry has its inherent shortcomings in mapping global internal tides. One major flaw is its long sampling intervals (or repeat cycles), so that a multi-year time series is required to extract internal tides. Furthermore, only the temporally coherent internal tide can be extracted. Thus, an important question is the degree of coherence of the global open-ocean internal tide. Recently, Shriver et al. (2014) addressed this question using a high-resolution global ocean circulation model (HYCOM; Arbic et al. 2010) and suggested that the internal tide is mostly coherent in regions with large amplitude. They also found that the internal tide becomes less coherent with propagation distance. There are a number of mechanisms responsible for the incoherent internal tide, including ocean stratification, background circulation, and mesoscale eddies (e.g., Rainville and Pinkel 2006; Zilberman et al. 2011; Zaron and Egbert 2014; Kelly et al. 2015). The incoherence caused by seasonal variation has been examined in a few recent studies (Ray and Zaron 2011; Müller et al. 2012; Shriver et al. 2014).

Satellite altimetry has low spatial resolution in that the SSH measurements are limited to hundreds of

predetermined ground tracks. In this study a higher spatial resolution is achieved by merging multisatellite altimeter data along four sets of ground tracks. In this regard, the next-generation satellite mission Surface Water Ocean Topography (SWOT) can make great improvements (Fu and Uebelmann 2014). SWOT will make SSH measurements in a swath of 120 km, instead of nadir looking.

Our global internal tide map may find immediate applications in a variety of oceanographic studies. It may provide guidance for seagoing oceanographers in designing experiment sites and interpreting field measurements in a basinwide geographic context. It may be used to examine the internal tide's along-beam decay and thus help quantify the relative importance of various dissipation mechanisms (Bühler and Holmes-Cerfon 2011; Mathur et al. 2014; Nash et al. 2012; Martini et al. 2013; Kelly et al. 2013b).

Internal tides have been simulated by global high-resolution, eddy-allowing numerical models including GOLD (Simmons et al. 2004; Waterhouse et al. 2014), HYCOM (Arbic et al. 2012; Shriver et al. 2014), STORMTIDE (Müller et al. 2012), and MITgcm (D. Menemenlis 2014, personal communication). To the best of our knowledge, all of these models can

successfully simulate the internal tide's generation hot-spots and long-range propagation. Further improvements may be realized by fine tuning parameters in these models using satellite altimetric observations. Currently, there are two other altimeter-based internal tide models, constructed by frequency–wavenumber analysis (Dushaw 2015) and harmonic analysis (Ray and Zaron 2016), respectively. Our understanding of the global internal tides will be improved with the concerted efforts of numerical modeling, field measurements, and satellite remote sensing.

Acknowledgments. This work was supported by NSF Grants OCE1129129 and OCE1130099 and NASA Grants NNX13AD90G and NNX13AG85G. The satellite altimeter products were produced by Ssalto/Duacs and distributed by AVISO (Archiving, Validation, and Interpretation of Satellite Oceanographic Data), with support from CNES (<http://www.aviso.altimetry.fr>). The satellite altimetric internal tide products in this study are available upon request. Discussions with Eric D'Asaro and Brian Dushaw are gratefully acknowledged. Detailed suggestions made by two anonymous reviewers greatly improved this manuscript.

APPENDIX A

Internal Tide Dynamics Parameters

Internal tides are long gravity waves, with their horizontal scale much greater than the ocean depth. A separation of variables technique is widely employed for simplification with the hydrostatic approximation (Wunsch 1975; Gill 1982). Thus, the solution can be expressed as a sum of orthogonal vertical modes, each of which has a fixed modal structure in the vertical and freely propagates in the horizontal like a wave in a homogeneous fluid. For example, its displacement $D(x, y, z, t)$ can be written as $D(x, y, z, t) = \sum A_n(x, y, t) \Phi_n(z)$, where the subscript n denotes the mode number. The amplitude $A_n(x, y, t)$ is determined using satellite altimeter data. In this study, only the first mode internal tide is considered ($n = 1$; omitted hereinafter).

Assuming that $\Phi(z)$ describes the vertical structures of displacement (η') and vertical velocity (w') and $\Pi(z)$ describes the vertical structures of baroclinic pressure (p') and horizontal velocity (u', v'), the orthogonal vertical modes $\Phi(z)$ can be obtained by the eigenvalue equation (Gill 1982; Pedlosky 2003)

$$\frac{d^2 \Phi(z)}{dz^2} + \frac{N^2(z)}{c^2} \Phi(z) = 0, \quad (\text{A1})$$

subject to the boundary conditions $\Phi(0) = \Phi(-H) = 0$, where H is the ocean depth, $N(z)$ is the buoyancy

frequency profile, and c is the eigenspeed. Additionally $\Phi(z)$ and $\Pi(z)$ are related via

$$\Pi(z) = \rho_0 c^2 \frac{d\Phi(z)}{dz} \quad (\text{A2})$$

and

$$\Phi(z) = -\frac{1}{N^2(z)\rho_0} \frac{d\Pi(z)}{dz}, \quad (\text{A3})$$

where ρ_0 is a reference water density.

The ocean stratification profiles are from the climatological annual-mean hydrographic data in the WOA2013 (Locarnini et al. 2013; Zweng et al. 2013). Figure A1a gives the buoyancy frequency profile at 25°N, 195°E (depth 4900 m). A set of baroclinic modes and their corresponding eigenvalues are obtained by solving Eq. (A1). The normalized $\Pi(z)$ and $\Phi(z)$ of the lowest three modes are shown in Figs. A1b and A1c, respectively.

The eigenvalue speed c is the phase speed in a non-rotating fluid. Under the influence of Earth's rotation (Ω), the dispersion relation is

$$\omega^2 = k^2 c^2 + f^2, \quad (\text{A4})$$

where $f [= 2\Omega \sin(\text{latitude})]$ is the inertial frequency and k is the wavenumber. The phase velocity c_p can be derived from c following Eq. (A4),

$$c_p = \frac{\omega}{(\omega^2 - f^2)^{1/2}} c, \quad (\text{A5})$$

where ω and f are the tidal and inertial frequencies, respectively. A global map of c_p of the mode-1 M_2 internal tide is given in Fig. A2a.

The baroclinic pressure fluctuation $p'(z)$ can be computed from its displacement fluctuation $\eta' [= \eta_0 \Phi(z)]$,

$$p'(z) = \rho_0 \eta_0 \int_{-z}^0 \Phi(\hat{z}) N^2(\hat{z}) d\hat{z} - p^{\text{surf}}. \quad (\text{A6})$$

Because $p'(z)$ must meet the orthogonal relation (see Fig. A1c), its depth integration should be zero (Kunze et al. 2002). Thus, p^{surf} is an integration constant, determined from the internal tide's interior amplitude η_0 ,

$$p^{\text{surf}} = \rho_0 \eta_0 \int_{-H}^0 N^2(z) \Phi(z) dz. \quad (\text{A7})$$

The sea surface displacement a is calculated from $p^{\text{surf}}/\rho_0 g$; therefore, the ratio of the sea surface to maximal interior displacements is

FIG. A1. An example of stratification and baroclinic modal structures. (a) Buoyancy frequency profile at 25°N, 195°E (depth 4900 m) from WOA2013. (b) Normalized baroclinic modes for displacement (η') and vertical velocity (w'). (c) Normalized baroclinic modes for pressure (p') and horizontal velocity (u', v').

$$\frac{a}{\eta_0} = \frac{1}{g} \int_{-H}^0 N^2(z) \Phi(z) dz. \quad (\text{A8})$$

Note that a and η_0 have a 180° difference in phase. As shown in Fig. A2b, the ratio (1/1000–1/200) does not violate the linear boundary conditions of Eq. (A1). The free surface dynamically acts as if it was rigid, and the eigensolutions are the same (Pedlosky 2003).

Depth-integrated available potential energy (PE) is computed from the interior amplitude (η_0):

$$\text{PE} = \frac{1}{2} \rho_0 \eta_0^2 \int_{-H}^0 N^2(z) \Phi^2(z) dz. \quad (\text{A9})$$

And depth-integrated kinetic energy (KE) is computed following

$$\text{KE} = \frac{1}{2} \rho_0 \mathbf{u}_0^2 \int_{-H}^0 \Pi^2(z) dz, \quad (\text{A10})$$

where $\mathbf{u}_0 = u_0 + iv_0$ is the horizontal velocity, which follows a polarization relation $u_0/\omega = v_0/f$.

Because of the influence of Earth's rotation, the energy partition between PE and KE follows

$$\frac{\text{KE}}{\text{PE}} = \frac{\omega^2 + f^2}{\omega^2 - f^2}. \quad (\text{A11})$$

The total energy E ($\equiv \text{PE} + \text{KE}$) can be calculated from Eqs. (A9)–(A11). Therefore, we build a transfer

function (E_n) from an internal tide's SSH amplitude a to its depth-integrated total energy (E),

$$E = \frac{1}{2} a^2 E_n(H, N, \omega, f), \quad (\text{A12})$$

where $E_n(H, N, \omega, f)$ is a function of water depth, stratification, tidal frequency, the local inertial frequency, and vertical mode number. Figure A2c gives the transfer function from SSH to energy for the mode-1 M_2 internal tide.

Integrating Eqs. (A9)–(A11) yields a relation between the magnitudes of horizontal velocity ($\mathbf{u}_0$) and vertical displacement η_0 . Typically, a mode-1 M_2 internal tide with 1-m interior displacement can cause horizontal velocity 0.3–1 cm s^{-1} at the sea surface. Thus, the depth-integrated energy flux ($\int_z p' u'$) is calculated from

$$F = \frac{1}{2} a^2 F_n(H, N, \omega, f). \quad (\text{A13})$$

Similarly, $F_n(H, N, \omega, f)$ is a function of water depth, stratification, tidal frequency, the local inertial frequency, and mode number. Figure A2d gives the transfer function from SSH to energy flux for the mode-1 M_2 internal tide.

We thus derived global transfer functions using the WOA2013 hydrographic profiles. Utilizing these parameters, we can calculate the internal tide's interior displacement (Fig. A1b) and velocity (Fig. A1c) and

FIG. A2. Parameters of the mode-1 M_2 internal tide from the climatological annual-mean hydrographic data in WOA2013. (a) Phase speed. (b) Ratio of the sea surface to maximal interior displacements. (c) Transfer function from SSH to depth-integrated energy. (d) Transfer function from SSH to depth-integrated energy flux. The 3000-m isobath contours are in black.

FIG. B1. (a) Bottom slope, calculated using bottom depths on a $0.2^\circ \times 0.2^\circ$ grid. (b) Horizontal gradient of phase speed of the M_2 internal tide (see Fig. A1a). Red lines mark regions of high mesoscale eddies associated with boundary currents (Fig. 3c). High gradients are mainly associated with rough topographic features and boundary currents.

depth-integrated energy density (Fig. A2c) and energy flux (Fig. A2d) from the satellite observed SSH parameters.

APPENDIX B

Horizontal Inhomogeneity

In a flat-bottom ocean with horizontally uniform stratification, an internal tidal wave can be decomposed into a series of orthogonal baroclinic modes (appendix A). More generally, modal decomposition can be applied to a gentle sloping bottom (Wunsch 1975; Kelly et al. 2012). However, in regions with sharp change of ocean bottom and/or stratification, the tidal equations cannot be decomposed into orthogonal modes (Wunsch 1975). And thus, modal coupling allows energy redistribution among baroclinic modes (Kelly et al. 2012, 2013a). In this section we examine where the ocean is inhomogeneous by studying the horizontal gradients of bottom and phase speed.

The bottom slope is calculated as follows. First, bottom depths on $0.2^\circ \times 0.2^\circ$ Mercator grids are extracted from the 2-arc-min topography database

(Smith and Sandwell 1997). Second, the zonal (s_x) and meridional (s_y) gradients are calculated and smoothed by three-point averaging. Third, the overall bottom slope (s) is calculated following $1/s^2 = 1/s_x^2 + 1/s_y^2$ (Fig. B1a). Following the same procedure, the horizontal gradient of phase speed of the M_2 internal tide is obtained (Fig. B1b). The phase speed is dependent on bottom depth and stratification; therefore, Figure B1b shows that large gradients of phase speed are associated with rough topographic features and boundary currents. It is assumed that boundary currents and their temporal variations cause sharp horizontal variation in stratification. Note that the effect of boundary currents is severely underestimated, because the phase speed is calculated using WOA2013 climatological hydrographic data.

In this study, we discard the internal tide solution where the bottom slope is steeper than an empirical threshold 6/1000. Based on this criterion, less than 2% of the global ocean area is masked out. Internal tides affected by boundary currents are not masked out using the criterion of horizontal gradient. In fact, the internal tide solution is contaminated by mesoscale eddies associated with boundary currents and is therefore discarded in examining the background noise level (Fig. 3c).

FIG. C1. The wavenumber and frequency spectra. (a)–(c) TPJ 223, (d)–(f) TPT 249, (g)–(i) ERS 157, and (j)–(m) GFO 230. The blue and red lines indicate spectra of the unfiltered (original) and along-track high-pass filtered SSH data, respectively. The green and brown lines indicate spectra of the harmonically fitted M_2 internal tides using the unfiltered and filtered SSH data, respectively. The top panels show the time-averaged wavenumber spectra. Spectral peaks corresponding to modes 1 and 2 are indicated by vertical lines (top panels). The middle panels show the space-averaged frequency spectra. Spectral peaks corresponding to the annual (vertical lines) and M_2 aliasing cycles are labeled. Gray boxes (whose widths are empirically selected) mark the aliasing frequencies of the M_2 internal tide, and zoomed-in views are shown in the bottom panels.

APPENDIX C

Wavenumber and Frequency Spectra

The wavenumber and frequency spectra can be calculated from the along-track SSH data. As examples, spectra from TPJ 223, TPT 249, ERS 157, and GFO 230 are displayed in Fig. C1. All these tracks cross the Hawaiian Ridge, and the SSH data between 0° and 40°N are used in the computation. For all cases, the wavenumber spectra are averaged in time (top panels), and the frequency spectra are averaged in space (middle panels). Spectra of the raw SSH data are shown by blue lines (top panels). All wavenumber spectra have an outstanding peak corresponding to the theoretical wavelength of the

mode-1 M_2 internal tide (appendix A) and a weak peak corresponding to mode 2. In the frequency spectra, all have outstanding peaks at annual and M_2 aliasing periods.

We employ a fourth-order Butterworth filter with a cutoff wavelength 500 km to remove barotropic tide residual and orbit error. Spectra of the high-pass-filtered SSH data are shown by red lines. Their wavenumber spectra show that the low-wavenumber component is removed (top panels). Their frequency spectra are lowered throughout the frequency range (middle panels). In particular, the spectral peaks at annual cycle are removed by along-track filtering.

From the raw and along-track filtered SSH data, the M_2 tidal constituent can be extracted by point-wise

FIG. D1. The harmonically fitted M_2 internal tides from satellite SSH measurements: (a) TPJ, (b) TPT, (c) GFO, and (d) ERS. The satellite SSH data are along-track, high-pass filtered using a fourth-order Butterworth filter. Then the M_2 internal tides are extracted by harmonic analysis. For clarity, the color-coded SSH amplitudes are shown for one out of every 10 points. Regions of energetic internal tides are evidently associated with topographic features like the Hawaiian Ridge and the French Polynesian Ridge. Mesoscale contamination is seen in regions such as the Kuroshio Current and the Gulf Stream.

FIG. E1. An evaluation of the plane wave fit method using the GOLD internal tide field. (a) Mode-1 M_2 internal tide from the GOLD simulation. (b) Three-wave superposition from plane wave fits. (c) Difference between (a) and (b).

harmonic analysis. The M_2 aliasing periods are about 62, 97, and 317 days for TPJ (TPT), ERS, and GFO, respectively. Spectra of the harmonically fitted M_2 internal tides are shown in brown and green, respectively (Fig. C1). They have outstanding spectral peaks corresponding to mode-1 and -2 internal tides.

The bottom panels give zoomed-in views of the frequency spectra around the M_2 aliasing periods. The window widths are empirically selected to cover the spectral peaks. Among these four datasets, TPJ is about 20 years long, so its frequency window is the narrowest (Fig. C1b). For comparison, the TPT frequency window is wider because of its shorter time series (Fig. C1e). GFO has the widest frequency window (Fig. C1k), because it has short time series (7 years) and a long M_2 aliasing period (317 days). The GFO results are affected by the nearby annual signals (Figs. C1k,m). This partly explains why the GFO-derived M_2 internal tides are noisier.

The wavenumber and frequency spectra have outstanding spectral peaks, suggesting that a significant fraction of the M_2 internal tide is temporally and spatially coherent, which warrants the application of the plane wave fit method.

APPENDIX D

Harmonically Fitted M_2 Internal Tides

The M_2 internal tides are extracted from the high-pass-filtered SSH data (appendix B) by point-wise harmonic analysis (Fig. D1). Each panel is for one SSH dataset. Note that the TPJ-derived M_2 internal tides (Fig. D1a) are the same as those in Müller et al. (2012) and Shriver et al. (2014), except that they used a different cutoff wavelength (400 km) in along-track

filtering. In all panels, energetic internal tides are observed over major topographic features such as the Hawaiian Ridge.

The harmonically fitted M_2 internal tides are apparently contaminated by leaked mesoscale signals, because the mesoscale motions have a broadband spectrum, overlapping M_2 internal tides both in the spatial and temporal domain. The degree of mesoscale contamination depends on the length of the SSH time series and the M_2 aliasing period. The longer aliasing period suffers from stronger mesoscale leakage, because the SSH frequency spectrum is red. TPJ is the least contaminated, as evidenced by the weak signals in energetic mesoscale regions such as the ACC and the Gulf Stream (Fig. D1a). The TPT results are also clean, except for high energetic mesoscale regions (Fig. D1b).

The GFO- and ERS-derived M_2 internal tides are noisy, because of their relatively shorter time series and longer aliasing periods. The noisy internal tides from ERS and GFO have been reported by Zaron and Egbert (2006). These datasets were previously excluded in internal tide mapping, because the addition of the GFO and ERS data may deteriorate the results (Dushaw et al. 2011). However, the advantage of ERS and GFO is evident: discrete internal tidal beams can be recognized, as a result of their denser ground tracks (Figs. D1c,d).

Multisatellite altimetry has admirable higher spatial resolution much needed in mapping global internal tides. Therefore it is a challenging and rewarding task to utilize all four datasets by suppressing mesoscale contamination. Plane wave fitting extracts internal tides with both spatial and temporal coherence. The strict requirement helps remove nontidal signals and thus yields a high-resolution global map through the merger of all four SSH datasets.

APPENDIX E

Three-Wave Fit of the GOLD Internal Tide Field

We here evaluate the performance of the plane wave fit method by applying it to the GOLD simulated internal tide field. Numerical model simulations like GOLD are sampled in high spatial ($1/8^\circ$) and temporal (2 h) resolution, so that the harmonically fitted M_2 internal tides are immune to mesoscale contamination. Figure E1a shows the mode-1 M_2 internal tide field in the central North Pacific, derived from the year-long harmonic component in the GOLD simulation. Shown in Fig. E1a is the SSH amplitude converted from the modeled sea surface pressure fluctuation.

Taking the GOLD simulated SSH data as satellite observations, we conduct three-wave decomposition by the plane wave fit method using the same parameters as described in section 3. The three-wave solution (Fig. E1b) accounts for about 95% of the GOLD internal tide. Large differences mainly occur in regions with rough topographic features, where the internal tide field is more complex (Fig. E1c). The difference may be further reduced by fitting more than three internal waves and using smaller fitting windows, but detailed examination is beyond the scope of this study. We thus conclude that the open-ocean internal tide field away from generation sites can be well described by the three-wave superposition.

REFERENCES

- 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.
- , and Z. Zhao, 2007a: Global patterns of low-mode internal-wave propagation. Part I: Energy and energy flux. *J. Phys. Oceanogr.*, **37**, 1829–1848, doi:10.1175/JPO3085.1.
- , and —, 2007b: Global patterns of low-mode internal-wave propagation. Part II: Group velocity. *J. Phys. Oceanogr.*, **37**, 1849–1858, doi:10.1175/JPO3086.1.
- , J. A. MacKinnon, Z. Zhao, R. Pinkel, J. Klymak, and T. Peacock, 2007: Internal waves across the Pacific. *Geophys. Res. Lett.*, **34**, L24601, doi:10.1029/2007GL031566.
- Althaus, A. M., E. Kunze, and T. B. Sanford, 2003: Internal tide radiation from Mendocino Escarpment. *J. Phys. Oceanogr.*, **33**, 1510–1527, doi:10.1175/1520-0485(2003)033<1510:ITRFME>2.0.CO;2.
- Arbic, B. K., S. T. Garner, R. W. Hallberg, and H. L. Simmons, 2004: The accuracy of surface elevations in forward global barotropic and baroclinic tide models. *Deep-Sea Res. II*, **51**, 3069–3101, doi:10.1016/j.dsr2.2004.09.014.
- , A. J. Wallcraft, and E. J. Metzger, 2010: Concurrent simulation of the eddying general circulation and tides in a global ocean model. *Ocean Modell.*, **32**, 175–187, doi:10.1016/j.ocemod.2010.01.007.
- , J. Richman, J. Shriver, P. Timko, E. Metzger, and A. Wallcraft, 2012: Global modeling of internal tides within an eddying ocean general circulation model. *Oceanography*, **25**, 20–29, doi:10.5670/oceanog.2012.38.
- AVISO, 2012: DT CorSSH and DT SLA Product Handbook. AVISO altimetry, 17 pp. [Available online at http://www.aviso.altimetry.fr/fileadmin/documents/data/tools/hdbk_dt_corssh_dt_sla.pdf.]
- Baines, P. G., 1982: On internal tide generation models. *Deep-Sea Res.*, **29A**, 307–338, doi:10.1016/0198-0149(82)90098-X.
- Bühler, O., and M. Holmes-Cerfon, 2011: Decay of an internal tide due to random topography in the ocean. *J. Fluid Mech.*, **678**, 271–293, doi:10.1017/jfm.2011.115.
- Cacchione, D. A., L. F. Praston, and A. S. Ogston, 2002: The shaping of continental slopes by internal tides. *Science*, **296**, 724–727, doi:10.1126/science.1069803.
- Carter, G. S., and Coauthors, 2008: Energetics of M_2 barotropic-to-baroclinic tidal conversion at the Hawaiian islands. *J. Phys. Oceanogr.*, **38**, 2205–2223, doi:10.1175/2008JPO3860.1.
- Chandran, S., Ed., 2013: *Adaptive Antenna Arrays: Trends and Applications*. Springer, 660 pp.
- Chiswell, S. M., 2006: Altimeter and current meter observations of internal tides: Do they agree? *J. Phys. Oceanogr.*, **36**, 1860–1872, doi:10.1175/JPO2944.1.
- 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.
- Cummins, P. F., and L.-Y. Oey, 1997: Simulation of barotropic and baroclinic tides off northern British Columbia. *J. Phys. Oceanogr.*, **27**, 762–781, doi:10.1175/1520-0485(1997)027<0762:SOBAPT>2.0.CO;2.
- Dushaw, B. D., 2002: Mapping low-mode internal tides near Hawaii using TOPEX/Poseidon altimeter data. *Geophys. Res. Lett.*, **29**, 1250, doi:10.1029/2001GL013944.
- , 2006: Mode-1 internal tides in the western North Atlantic Ocean. *Deep-Sea Res. I*, **53**, 449–473, doi:10.1016/j.dsr.2005.12.009.
- , 2015: An empirical model for mode-1 internal tides derived from satellite altimetry: Computing accurate tidal predictions at arbitrary points over the world oceans. Tech. Rep., Applied Physics Laboratory, University of Washington, 114 pp. [Available online at http://staff.washington.edu/dushaw/epubs/Dushaw_Global_IT_Predict_APL-TM1-15_v1.1.pdf.]
- , B. M. Howe, B. D. Cornuelle, P. F. Worcester, and D. S. Luther, 1995: Barotropic and baroclinic tides in the central North Pacific Ocean determined from long-range reciprocal acoustic transmissions. *J. Phys. Oceanogr.*, **25**, 631–647, doi:10.1175/1520-0485(1995)025<0631:BAITIT>2.0.CO;2.
- , P. F. Worcester, and M. A. Dzieciuch, 2011: On the predictability of mode-1 internal tides. *Deep-Sea Res. I*, **58**, 677–698, doi:10.1016/j.dsr.2011.04.002.
- Egbert, G. D., and R. D. Ray, 2000: Significant dissipation of tidal energy in the deep ocean inferred from satellite altimeter data. *Nature*, **405**, 775–778, doi:10.1038/35015531.
- , and —, 2001: Estimates of M_2 tidal energy dissipation from TOPEX/Poseidon altimeter data. *J. Geophys. Res.*, **106**, 22 475–22 502, doi:10.1029/2000JC000699.
- , and —, 2003: Semi-diurnal and diurnal tidal dissipation from TOPEX/Poseidon altimetry. *Geophys. Res. Lett.*, **30**, 1907, doi:10.1029/2003GL017676.
- , —, and B. G. Bills, 2004: Numerical modeling of the global semidiurnal tide in the present day and in the last glacial maximum. *J. Geophys. Res.*, **109**, C03003, doi:10.1029/2003JC001973.
- Fu, L.-L., and C. Uebmann, 2014: On the transition from profile altimeter to swath altimeter for observing global ocean surface topography. *J. Atmos. Oceanic Technol.*, **31**, 560–568, doi:10.1175/JTECH-D-13-00109.1.
- Garrett, C., and E. Kunze, 2007: Internal tide generation in the deep ocean. *Annu. Rev. Fluid Mech.*, **39**, 57–87, doi:10.1146/annurev.fluid.39.050905.110227.

- Gill, A. E., 1982: *Atmosphere-Ocean Dynamics*. Academic Press, 662 pp.
- Hallberg, R., 1997: Stable split time stepping schemes for large-scale ocean modeling. *J. Comput. Phys.*, **135**, 54–65, doi:10.1006/jcph.1997.5734.
- Hendry, R. M., 1977: Observations of the semidiurnal internal tide in the western North Atlantic Ocean. *Philos. Trans. Roy. Soc. London*, **286A**, 1–24, doi:10.1098/rsta.1977.0108.
- Jan, S., and C.-T. A. Chen, 2009: Potential biogeochemical effects from vigorous internal tides generated in Luzon Strait: A case study at the southernmost coast of Taiwan. *J. Geophys. Res.*, **114**, C04021, doi:10.1029/2008JC004887.
- Jayne, S. R., 2009: The impact of abyssal mixing parameterizations in an ocean general circulation model. *J. Phys. Oceanogr.*, **39**, 1756–1775, doi:10.1175/2009JPO4085.1.
- , and L. St. Laurent, 2001: Parameterizing tidal dissipation over rough topography. *Geophys. Res. Lett.*, **28**, 811–814, doi:10.1029/2000GL012044.
- Kantha, L. H., 1998: Tides—A modern perspective. *Mar. Geod.*, **21**, 275–297, doi:10.1080/01490419809388143.
- , and C. C. Tierney, 1997: Global baroclinic tides. *Prog. Oceanogr.*, **40**, 163–178, doi:10.1016/S0079-6611(97)00028-1.
- 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, and J. D. Nash, 2013a: A coupled model for Laplace's tidal equations in a fluid with one horizontal dimension and variable depth. *J. Phys. Oceanogr.*, **43**, 1780–1797, doi:10.1175/JPO-D-12-0147.1.
- , —, and A. F. Waterhouse, 2013b: The geography of semidiurnal mode-1 internal-tide energy loss. *Geophys. Res. Lett.*, **40**, 4689–4693, doi:10.1002/grl.50872.
- , —, G. N. Ivey, and R. J. Lowe, 2015: Internal-tide spectroscopy and prediction in the Timor Sea. *J. Phys. Oceanogr.*, **45**, 64–83, doi:10.1175/JPO-D-14-0007.1.
- Kunze, E., and S. G. Llewellyn Smith, 2004: The role of small-scale topography in turbulent mixing of the global ocean. *Oceanography*, **17**, 55–64, doi:10.5670/oceanog.2004.67.
- , L. K. Rosenfeld, G. S. Carter, and M. C. Gregg, 2002: Internal waves in Monterey submarine canyon. *J. Phys. Oceanogr.*, **32**, 1890–1913, doi:10.1175/1520-0485(2002)032<1890:IWIMSC>2.0.CO;2.
- Kurapov, A. L., J. S. Allen, and G. D. Egbert, 2010: Combined effects of wind-driven upwelling and internal tide on the continental shelf. *J. Phys. Oceanogr.*, **40**, 737–756, doi:10.1175/2009JPO4183.1.
- Lee, C. M., T. B. Sanford, E. Kunze, J. D. Nash, M. A. Merrifield, and P. E. Holloway, 2006: Internal tides and turbulence along the 3000-m isobath of the Hawaiian Ridge. *J. Phys. Oceanogr.*, **36**, 1165–1183, doi:10.1175/JPO2886.1.
- Lien, R.-C., and M. C. Gregg, 2001: Observations of turbulence in a tidal beam and across a coastal ridge. *J. Geophys. Res.*, **106**, 4575–4591, doi:10.1029/2000JC000351.
- Locarnini, R. A., and Coauthors, 2013: *Temperature*. Vol. 1, *World Ocean Atlas 2013*, NOAA Atlas NESDIS 73, 40 pp. [Available online at http://data.nodc.noaa.gov/woa/WOA13/DOC/woa13_vol1.pdf.]
- MacKinnon, J. A., and K. B. Winters, 2005: Subtropical catastrophe: Significant loss of low-mode tidal energy at 28.9°. *Geophys. Res. Lett.*, **32**, L15605, doi:10.1029/2005GL023376.
- , M. H. Alford, R. Pinkel, J. Klymak, and Z. Zhao, 2013: The latitudinal dependence of shear and mixing in the Pacific transiting the critical latitude for PSI. *J. Phys. Oceanogr.*, **43**, 3–16, doi:10.1175/JPO-D-11-0107.1.
- Maraldi, C., F. Lyard, L. Testut, and R. Coleman, 2011: Energetics of internal tides around the Kerguelen Plateau from modeling and altimetry. *J. Geophys. Res.*, **116**, C06004, doi:10.1029/2010JC006515.
- Martini, K. I., M. H. Alford, J. D. Nash, E. Kunze, and M. A. Merrifield, 2007: Diagnosing a partly-standing internal wave in Mamala Bay, Oahu. *Geophys. Res. Lett.*, **34**, L17604, doi:10.1029/2007GL029749.
- , —, E. Kunze, S. M. Kelly, and J. D. Nash, 2013: Internal bores and breaking internal tides on the Oregon continental slope. *J. Phys. Oceanogr.*, **43**, 120–139, doi:10.1175/JPO-D-12-030.1.
- Mathur, M., G. S. Carter, and T. Peacock, 2014: Topographic scattering of the low-mode internal tide in the deep ocean. *J. Geophys. Res. Oceans*, **119**, 2165–2182, doi:10.1002/2013JC009152.
- Melet, A., R. Hallberg, S. Legg, and K. Polzin, 2013: Sensitivity of the ocean state to the vertical distribution of internal-tide-driven mixing. *J. Phys. Oceanogr.*, **43**, 602–615, doi:10.1175/JPO-D-12-055.1.
- Mitchum, G. T., and S. M. Chiswell, 2000: Coherence of internal tide modulations along the Hawaiian Ridge. *J. Geophys. Res.*, **105**, 28 653–28 661, doi:10.1029/2000JC900140.
- Morozov, E. G., 1995: Semidiurnal internal wave global field. *Deep-Sea Res. I*, **42**, 135–148, doi:10.1016/0967-0637(95)92886-C.
- Müller, M., J. Y. Cherniawsky, M. G. G. Foreman, and J.-S. von Storch, 2012: Global M2 internal tide and its seasonal variability from high resolution ocean circulation and tide modeling. *Geophys. Res. Lett.*, **39**, L19607, doi:10.1029/2012GL053320.
- Munk, W. H., 1981: Internal waves and small-scale processes. *Evolution of Physical Oceanography*, B. A. Warren and C. Wunsch, Eds., MIT Press, 264–291.
- , and C. Wunsch, 1998: Abyssal recipes II: Energetics of tidal and wind mixing. *Deep-Sea Res. I*, **45**, 1977–2010, doi:10.1016/S0967-0637(98)00070-3.
- , B. Zetler, and G. W. Groves, 1965: Tidal cusps. *Geophys. J.*, **10**, 211–219, doi:10.1111/j.1365-246X.1965.tb03062.x.
- Nash, J. D., E. Kunze, C. M. Lee, and T. B. Sanford, 2006: Structure of the baroclinic tide generated at Kaena Ridge, Hawaii. *J. Phys. Oceanogr.*, **36**, 1123–1135, doi:10.1175/JPO2883.1.
- , M. H. Alford, E. Kunze, K. Martini, and S. Kelly, 2007: Hot-spots of deep ocean mixing on the Oregon continental shelf. *Geophys. Res. Lett.*, **34**, L01605, doi:10.1029/2006GL028170.
- , S. M. Kelly, E. L. Shroyer, J. N. Moum, and T. F. Duda, 2012: The unpredictable nature of internal tides on continental shelves. *J. Phys. Oceanogr.*, **42**, 1981–2000, doi:10.1175/JPO-D-12-028.1.
- New, A. L., and J. C. B. Da Silva, 2002: Remote sensing evidence for the local generation of internal soliton packets in the central Bay of Biscay. *Deep-Sea Res. I*, **49**, 915–934, doi:10.1016/S0967-0637(01)00082-6.
- Niwa, Y., and T. Hibiya, 2011: Estimation of baroclinic tide energy available for deep ocean mixing based on three-dimensional global numerical simulations. *J. Oceanogr.*, **67**, 493–502, doi:10.1007/s10872-011-0052-1.
- , and —, 2014: Generation of baroclinic tide energy in a global three-dimensional numerical model with different spatial grid resolutions. *Ocean Modell.*, **80**, 59–73, doi:10.1016/j.ocemod.2014.05.003.
- Nycander, J., 2005: Generation of internal waves in the deep ocean by tides. *J. Geophys. Res.*, **110**, C10028, doi:10.1029/2004JC002487.
- Pedlosky, J., 2003: *Waves in the Ocean and Atmosphere: Introduction to Wave Dynamics*. Springer, 264 pp.
- Pétrélis, F., S. Llewellyn Smith, and W. R. Young, 2006: Tidal conversion at a submarine ridge. *J. Phys. Oceanogr.*, **36**, 1053–1071, doi:10.1175/JPO2879.1.
- Pingree, R. D., G. T. Mardell, and A. L. New, 1986: Propagation of internal tides from the upper slopes of the Bay of Biscay. *Nature*, **321**, 154–158, doi:10.1038/321154a0.

- Pinkel, R., and Coauthors, 2015: Breaking internal tides keep the ocean in balance. *Eos, Trans. Amer. Geophys. Union*, **96**, doi:10.1029/2015EO039555.
- Polzin, K. L., J. M. Toole, J. R. Ledwell, and R. W. Schmitt, 1997: Spatial variability of turbulent mixing in the abyssal ocean. *Science*, **276**, 93–96, doi:10.1126/science.276.5309.93.
- Rainville, L., and R. Pinkel, 2006: Propagation of low-mode internal waves through the ocean. *J. Phys. Oceanogr.*, **36**, 1220–1236, doi:10.1175/JPO2889.1.
- , 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.
- Ray, R. D., 2013: Precise comparisons of bottom-pressure and altimetric ocean tides. *J. Geophys. Res. Oceans*, **118**, 4570–4584, doi:10.1002/jgrc.20336.
- , and G. T. Mitchum, 1996: Surface manifestation of internal tides generated near Hawaii. *Geophys. Res. Lett.*, **23**, 2101–2104, doi:10.1029/96GL02050.
- , and —, 1997: Surface manifestation of internal tides in the deep ocean: Observations from altimetry and island gauges. *Prog. Oceanogr.*, **40**, 135–162, doi:10.1016/S0079-6611(97)00025-6.
- , and D. E. Cartwright, 2001: Estimates of internal tide energy fluxes from TOPEX/Poseidon altimetry: Central North Pacific. *Geophys. Res. Lett.*, **28**, 1259–1262, doi:10.1029/2000GL012447.
- , and E. D. Zaron, 2011: Non-stationary internal tides observed with satellite altimetry. *Geophys. Res. Lett.*, **38**, L17609, doi:10.1029/2011GL048617.
- , and E. Zaron, 2016: M₂ internal tides and their observed wavenumber spectra from satellite altimetry. *J. Phys. Oceanogr.*, **46**, 3–22, doi:10.1175/JPO-D-15-0065.1.
- Rost, S., and C. Thomas, 2002: Array seismology: Methods and applications. *Rev. Geophys.*, **40**, 1008, doi:10.1029/2000RG000100.
- Rudnick, D. L., and Coauthors, 2003: From tides to mixing along the Hawaiian Ridge. *Science*, **301**, 355–357, doi:10.1126/science.1085837.
- Samelson, R. M., 1998: Large-scale circulation with locally enhanced vertical mixing. *J. Phys. Oceanogr.*, **28**, 712–726, doi:10.1175/1520-0485(1998)028<0712:LSCWLE>2.0.CO;2.
- Sharples, J., and Coauthors, 2007: Spring-neap modulation of internal tide mixing and vertical nitrate fluxes at a shelf edge in summer. *Limnol. Oceanogr.*, **52**, 1735–1747, doi:10.4319/lo.2007.52.5.1735.
- Shriver, J. F., B. K. Arbic, J. G. Richman, R. D. Ray, E. J. Metzger, A. J. Wallcraft, and P. G. Timko, 2012: An evaluation of the barotropic and internal tides in a high-resolution global ocean circulation model. *J. Geophys. Res.*, **117**, C10024, doi:10.1029/2012JC008170.
- , J. G. Richman, and B. K. Arbic, 2014: How stationary are the internal tides in a high-resolution global ocean circulation model? *J. Geophys. Res. Oceans*, **119**, 2769–2787, doi:10.1002/2013JC009423.
- Simmons, H. L., 2008: Spectral modification and geographic redistribution of the semi-diurnal internal tide. *Ocean Modell.*, **21**, 126–138, doi:10.1016/j.ocemod.2008.01.002.
- , and M. H. Alford, 2012: Simulating the long-range swell of internal waves generated by ocean storms. *Oceanography*, **25**, 30–41, doi:10.5670/oceanog.2012.39.
- , S. R. Jayne, L. C. St. Laurent, and A. J. Weaver, 2004: Tidally driven mixing in a numerical model of the ocean general circulation. *Ocean Modell.*, **6**, 245–263, doi:10.1016/S1463-5003(03)00011-8.
- Smith, W. H. F., and D. T. Sandwell, 1997: Global sea floor topography from satellite altimetry and ship depth soundings. *Science*, **277**, 1956–1962, doi:10.1126/science.277.5334.1956.
- Terker, S. R., J. B. Girton, E. Kunze, J. M. Klymak, and R. Pinkel, 2014: Observations of the internal tide on the California continental margin near Monterey Bay. *Cont. Shelf Res.*, **82**, 60–71, doi:10.1016/j.csr.2014.01.017.
- van Haren, H., 2004: Bandwidth similarity at inertial and tidal frequencies in kinetic energy spectra from the Bay of Biscay. *Deep-Sea Res. I*, **51**, 637–652, doi:10.1016/j.dsr.2004.01.006.
- Waterhouse, A. F., and Coauthors, 2014: Global patterns of diapycnal mixing from measurements of the turbulent dissipation rate. *J. Phys. Oceanogr.*, **44**, 1854–1872, doi:10.1175/JPO-D-13-0104.1.
- Webb, D. J., and N. Sugimotohara, 2001: Vertical mixing in the ocean. *Nature*, **409**, 37, doi:10.1038/35051171.
- Worcester, P. F., and Coauthors, 2013: The North Pacific Acoustic Laboratory deep-water acoustic propagation experiments in the Philippine Sea. *J. Acoust. Soc. Amer.*, **134**, 3359, doi:10.1121/1.4818887.
- Wunsch, C., 1975: Internal tides in the ocean. *Rev. Geophys. Space Phys.*, **13**, 167–182, doi:10.1029/RG013i001p00167.
- , 2013: Baroclinic motions and energetics as measured by altimeters. *J. Atmos. Oceanic Technol.*, **30**, 140–150, doi:10.1175/JTECH-D-12-00035.1.
- , and R. Ferrari, 2004: Vertical mixing, energy, and the general circulation of the oceans. *Annu. Rev. Fluid Mech.*, **36**, 281–314, doi:10.1146/annurev.fluid.36.050802.122121.
- Xu, Y., and L.-L. Fu, 2012: The effects of altimeter instrument noise on the estimation of the wavenumber spectrum of sea surface height. *J. Phys. Oceanogr.*, **42**, 2229–2233, doi:10.1175/JPO-D-12-0106.1.
- Zaron, E. D., and G. D. Egbert, 2006: Estimating open-ocean barotropic tidal dissipation: The Hawaiian Ridge. *J. Phys. Oceanogr.*, **36**, 1019–1035, doi:10.1175/JPO2878.1.
- , and —, 2014: Time-variable refraction of the internal tide at the Hawaiian Ridge. *J. Phys. Oceanogr.*, **44**, 538–557, doi:10.1175/JPO-D-12-0238.1.
- Zhao, Z., 2014: Internal tide radiation from the Luzon Strait. *J. Geophys. Res. Oceans*, **119**, 5434–5448, doi:10.1002/2014JC010014.
- , and M. H. Alford, 2009: New altimetric estimates of mode-1 M₂ internal tides in the central North Pacific Ocean. *J. Phys. Oceanogr.*, **39**, 1669–1684, doi:10.1175/2009JPO3922.1.
- , and E. A. D'Asaro, 2011: A perfect focus of the internal tide from the Mariana Arc. *Geophys. Res. Lett.*, **38**, L14609, doi:10.1029/2011GL047909.
- , M. H. Alford, J. A. MacKinnon, and R. Pinkel, 2010: Long-range propagation of the semidiurnal internal tides from the Hawaiian Ridge. *J. Phys. Oceanogr.*, **40**, 713–736, doi:10.1175/2009JPO4207.1.
- , —, J. Girton, T. M. S. Johnston, and G. Carter, 2011: Internal tides around the Hawaiian Ridge estimated from multisatellite altimetry. *J. Geophys. Res.*, **116**, C12039, doi:10.1029/2011JC007045.
- , —, and —, 2012: Mapping low-mode internal tides from multisatellite altimetry. *Oceanography*, **25**, 42–51, doi:10.5670/oceanog.2012.40.
- 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.
- Zweng, M., and Coauthors, 2013: *Salinity*. Vol. 2, World Ocean Atlas 2013, NOAA Atlas NESDIS 74, 39 pp. [Available online at http://data.nodc.noaa.gov/woa/WOA13/DOC/woa13_vol2.pdf.]