Observations of Internal Gravity Waves by Argo Floats

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ABSTRACT

This study examines the global variability of the internal wave field near a depth of 1000 m using data from a set of 194 Argo floats equipped with Iridium communications, capable of measuring hourly temperature and pressure during the park phase of their 10-day cycles. These data have been used to estimate vertical isotherm displacements at hourly intervals, yielding a global measure of the heaving due to internal gravity waves. The displacement results have been employed to examine the global variability of these waves and how the displacement power spectrum compares to the canonical Garrett–Munk spectrum. Using the data, the authors find correlations between internal wave intensity and seafloor roughness, proximity to the seafloor, and the magnitude of the local barotropic velocity. The measurements also show large seamount-trapped waves at high latitudes and coastally trapped subinertial waves. These observations provide a rough global census of the nature of these waves that can ultimately be used in studies of ocean mixing.

1. Introduction

It has been suggested for some time that ocean stratification and meridional overturning are related to the strength of abyssal mixing (Munk 1966). More recently acknowledged has been the role of the spatial variability of mixing on these processes (Samelson 1998; Simmons et al. 2004). Despite this importance, global patterns of mixing are still relatively poorly understood. Internal waves are thought to be a key contributor to abyssal mixing (Munk and Wunsch 1998). They overturn and break, much like surface waves. Via turbulence, this dissipates some of their energy, which leads to diapycnal mixing. Hence, understanding the variability of internal waves can yield clues to the distribution of dissipation and mixing throughout the ocean (Gregg 1989).

Recent studies (Kunze et al. 2006; Whalen 2012) have attempted to assess the global distribution of diapycnal mixing. In this work we present broad observations that complement these works by examining the global behavior of internal waves using data collected from a suite of profiling floats deployed as part of the global Argo array. The general behavior of these floats has been described in a number of places (Roemmich et al. 2004). The floats used in Argo typically cycle vertically between 2000 m and the sea surface on approximately 10-day cycles, collecting profiles of temperature and salinity as a function of pressure during their ascent. In many cases the floats also collect periodic observations of temperature during their park phase at 1000 m. In this work we examine the park-phase temperature variability from a suite of 194 floats deployed over the globe by the University of Washington (UW) during the period 2004–13. These floats were all equipped with Iridium communications and thus were capable of storing and transmitting considerably more data than most other active Argo floats during this time period. In the suite of floats used in this work, temperature and pressure observations were collected at 1-h intervals during the park phase, allowing variability in the internal gravity wave frequency band along the float paths to be examined.

As will be shown in section 2, the park-phase temperature data from the floats can be used to examine the
vertical displacement spectrum of internal gravity waves at the park depth. Because of the quasi-Lagrangian nature of the float motion during their park phase, the effect of advection on Argo floats can be assumed to be minimal. Thus, even in regions with fast currents [e.g., the Antarctic Circumpolar Current (ACC)], the data can be used to make credible estimates of the displacement spectrum uncontaminated by Doppler shifts in frequency.

In a prelude to our broad analysis, we focus on two case studies of specific floats (section 3). Following the two case studies, we will turn our attention to observations made by our dataset (section 4) and comparisons to the empirically derived Garrett–Munk (GM) spectrum (Garrett and Munk 1975). While the energy spectra of internal waves in much of the world’s oceans have properties similar to the GM spectrum, regions whose spectra depart markedly from this model are clearly of interest, perhaps as sources or sinks of internal wave energy (Wunsch 1975). Deviations could also point to areas where the GM description is inadequate (Levine 2002).

Also in section 4, we examine the effects of seafloor roughness, height above the seafloor, and barotropic velocity magnitude on internal wave intensity. Internal waves are one of the primary mechanisms for mixing the deep ocean (Munk and Wunsch 1998), and because diapycnal diffusivity is known to be elevated above rough bathymetry (Polzin et al. 1997; Ledwell et al. 2000), we examine the possibility internal waves might be excited in these areas as well. We also examine the properties of internal gravity waves at high latitudes.

2. Methods

a. Displacement estimates

The data presented in this work come exclusively from a set of 194 University of Washington Argo floats using Iridium communications, which have a broad spatial coverage (shown in Fig. 1). The primary mission of these floats is to collect vertical profiles of temperature and salinity in the upper 2000 m of the ocean, nominally at 10-day intervals. The floats drift freely during a quasi-isobaric park phase at 1000 dbar between profiles. These floats collect hourly measurements of temperature and pressure during the park phases. Thus, from these hourly measurements we can observe the temperature variability representing isotherms heaving vertically past the quasi-isobaric float. As an example, Fig. 2 shows hourly temperature measurements during both a single park phase for float 5135 and over the entire float lifetime.

With the knowledge of the local temperature gradient obtained from the vertical profiles, we can translate the temperature fluctuations into isotherm displacement. It is assumed that isotherm displacements are primarily due to internal wave motion, at least for periods less than a few days.

For individual park phases, we estimate the hourly vertical isotherm displacement observed by the float \( \eta_T \) as

\[
\eta_T(t) = \frac{T_p(t) - \bar{T}_p}{(dT/dP)_{1000}}.
\]

Here \( \bar{T}_p \) is the time average of the hourly park-phase temperature measurements, \( T_p(t) \) over one park phase,
and \((dT/dP)_{1000}\) is the temperature gradient at 1000 dbar, estimated using the profile measurements within 100 dbar of the parking pressure. The temperature gradient is calculated from the average of the vertical profiles immediately before and after the particular park phase.

If the floats were perfectly isobaric, \(\eta_T\) would be the only estimate needed to determine vertical isotherm displacement. However, the floats are imperfect and do typically experience small vertical motions while in the park phase, so the float’s displacement \(\eta_P\) needs to be taken into account and removed from the total isotherm displacement. We define the float displacement as

\[
\eta_P(t) = P_p - P_{p'}(t).
\]

Here \(P_p\) is the time average of the hourly park-phase pressure measurements \(P_{p'}(t)\), taken over one park-phase cycle. The float motion \(\eta_P\) is typically less than 20% of \(\eta_T\). By taking the motion of the float into account, we arrive at a corrected estimate for isotherm displacement and hence vertical internal wave displacement \(\eta(t)\), given by

\[
\eta(t) = \eta_T(t) + \eta_P(t).
\]

An example of the various components of \(\eta(t)\) is shown in Fig. 3. In our analyses here, we have chosen to high-pass \(\eta(t)\) using a fourth-order Butterworth filter in order to remove all variability below 0.3 cycles per day (cpd). This filtering removes error from slow temperature trends during the park phase that is likely due to low-frequency spatial variability and water-mass changes. Poleward of about 10° of latitude the inertial frequency exceeds 0.3 cpd, so except near the equator the inertial peak is not removed by this filter.

**b. Combining park phases for longer records**

In analyzing the internal wave displacement, it can sometimes be advantageous to examine each set of park-phase data as a discrete time series. This gives good temporal resolution (~10 days) of the variability of the spectrum, but it comes at the expense of frequency resolution in spectral analysis. In other cases it is acceptable to increase the length of the park-phase record in order to have better frequency resolution, but this requires time series that are longer than the 10-day park phases.

To examine longer records, we have stitched together all the individual park-phase records for \(\eta(t)\) to create a single multiyear time series for each float. We linearly interpolate \(\eta(t)\) of all park phases in order to create a set of hourly samples that is the length of the operative mission of the float, typically several years (an example is given in Fig. 4). After interpolation, hourly bins that correspond to times when the float is in its profiling phase (about 6 h every 10 days) are set to zero (the park-phase displacement average, adding no variance). Additionally, all \(\eta(t)\) for park phases that do not meet quality controls are set to zero (see section 2c for details on quality control). The advantage of having one long time series for each float is that it can be broken down into uniform segments that are longer than 10 days.
this work we used nonoverlapping 30-day segments, as this considerably increases the frequency resolution without sacrificing too much temporal and spatial resolution. Constructing 30-day segments of 10-day park phases by filling the time (typically 6 h) between park phases with zeroes does not in practice distort the shape of the power spectrum to a degree that affects our analysis.

c. Quality control

We have implemented several quality controls in order to avoid unreasonable estimates for internal wave displacement. For each park phase we have required the following:

- The temperature profile in the vicinity of 1000 dbar is nearly linear. We find the least squares best-fit line to

![Fig. 3](image)

**Fig. 3.** (a) Raw park-phase temperature during cycle 22 for float 5135. (b) Raw park-phase pressure during cycle 22. (c) Processed isotherm displacement as seen by the float (gray), the float displacement (dashed), and the actual isotherm displacement (thick black) during the park phase of cycle 22.

![Fig. 4](image)

**Fig. 4.** The synthesized time series of vertical internal wave displacement for float 5135, located in the Indian Ocean. The small gaps of data (zeroes) correspond to park phases where the vertical temperature profile was not linear enough to meet quality control (see section 2c).
the temperature profile within 100 dbar of the parking depth. This best-fit line must account for at least 90% of the variance of the local temperature profile for the displacement data to be included in our analysis. This is the most common quality criterion that is violated (in just over 10% of park phases), occurring most often at high latitudes where density is increasingly dependent on salinity instead of temperature.

- The parking pressure is within 50 of 1000 dbar. Occasionally a float will park at a pressure significantly different than 1000 dbar. For consistency, data from these park phases are not included in the analysis.
- The vertical float movement is minimal. There can be no more than one pressure measurement in each park phase that deviates more than 50 dbar from the average parking depth. Float motion is accounted for in Eqs. (2) and (3), but park phases with large deviations are eliminated to avoid excessive error.
- Park phases have at least 24-hourly measurements. The number of samples in most park phases greatly exceeds this number.

If any of these conditions are violated for a particular park phase, the corresponding hourly bins for the entire park phase are set to zero.

d. Power spectra

Whether we use the park-phase segments or the 30-day segments (described in section 2b), the power spectrum for segment $j$, $S_j(\omega)$, is defined for a discrete time series as

$$S_j(\omega) = \frac{2}{M_j \Delta t} \left[ \sum_{m=0}^{M_j-1} \eta_{j,m} e^{-i2\pi \omega \Delta t} \right]^2.$$  \hspace{1cm} (4)

Here $M_j$ is the number of hourly measurements in the $j$th park phase or 30-day segment, $\Delta t$ is the time between measurements (1 h), and $\omega$ is the frequency of oscillation. The factor of 2 comes from confining frequencies to positive values only, considering the symmetry of the displacement spectrum in positive and negative frequencies. The total variance in $S_j(\omega)$ is exactly equal to the variance in the $j$th temporal segment, in accordance with Parseval's theorem.

The spectrum from each of the 30-day segments for a single float can be formed into a mean power spectrum, which is useful for accurately quantifying typical internal wave behavior for a float that did not drift over vast distances. We define the average power spectrum $\tilde{S}(\omega)$ as the mean of all $S_j(\omega)$ for the 30-day segments. Choosing 30-day segments gives a frequency resolution of 0.03 cpd for the power spectrum. The hourly measurements during each segment result in a Nyquist frequency of 12 cycles per day, and thus the major tidal and inertial frequencies are generally well resolved with this method. Segments that have zeroes (corresponding to times of vertical profiling or data that did not meet quality control criteria) in more than 50% of their hourly bins are not included in the estimate for $\tilde{S}(\omega)$. An example of $\tilde{S}(\omega)$ is shown in Fig. 5 and will be discussed in section 3a.

In this study we calculate spectra that are obtained from individual park phases as well as the longer 30-day segments. We use park phases in sections 4a and 4c–f where the precise location is more important than accurate frequency resolution. Conversely, in section 4b, the frequency resolution is of greater importance to us than the floats’ location, so the 30-day segments are used for analysis. Section 3b uses the 30-day segments to describe general trends (Figs. 6, 7) and then uses park phases to explore the data with greater temporal/spatial resolution.

Generally, we used the 30-day segments to calculate $\tilde{S}(\omega)$ in favor of the park phases because the individual
park phases can vary in temporal duration, thus causing the frequency grids for the spectral estimates to be non-uniform. However, as a consistency check we compared the power spectrum for each float using the park phases with the 30-day segments (we truncated some park phases to yield a uniform number of hourly measurements to ensure a consistent frequency grid). There was little quantitative difference between the average power spectrum from the park-phase segments and the 30-day segments, which suggests that the spectrum of the 30-day segments is not severely distorted. Examples of this can be seen in Figs. 5 and 6.

![Graph](image)

**FIG. 6.** As in Fig. 5, but for float 6097 (WMO 5902118), located in the Southern Ocean near the Amundsen Ridge. The quantity $\mathcal{S}(\omega)_{park}$ is formulated from 5-day park phases for this float. The lower-right plot shows the surface locations of the float (the white star is the first profile and the black star is the final profile).

![Graph](image)

**FIG. 7.** Temporal evolution of the displacement power spectrum over the lifetime of float 6097. Each vertical section represents one power spectrum calculated from a 30-day segment. There are two gaps that did not have sufficient data that met quality control, and hence no power spectrum is calculated. The horizontal lines show frequencies of interest ($M_2$, $K_1$, and average inertial).
section 2c), so the use of a constant gradient should not introduce a large error. Therefore, the majority of the variability seen in temperature and pressure is real, rather than measurement error.

Equation (2) assumes a constant vertical temperature gradient, which introduces additional error. However, we only analyzed park phases where the temperature profile was nearly linear in the vicinity of the parking depth (see section 2c), so the use of a constant gradient should not introduce a large error.

We assume a chi-square distribution on our spectral estimates $S_\omega$. The integrated internal wave variance during park phases is a prevalent quantity in our paper (see section 4). During a park phase that lasts 4–9 days, there are approximately 100–200 degrees of freedom that make up each integrated internal wave variance estimate. This leads to an error of 20%–30%, which is likely the largest source of error on our estimates of internal wave variability.

Given the variability shown in subsequent sections is much larger than these combined sources of error, we are confident that the trends presented by this paper are an accurate representation of real-world phenomena.

3. Single float case studies
a. Common features

Most floats used in this paper operated for at least 3 yr and completed at least 100 cycles. Float 5135 [World Meteorological Organization (WMO) 5901312] had a lifetime of nearly 5.5 yr and collected profile and park-phase data from over 200 cycles; the averaged power spectrum $S_\omega$ from this float is shown in Fig. 5.

Its location for the entire deployment was over a relatively small region just west of Australia. Given the length of deployment and limited spatial range, we would expect $S_\omega$ to be generally representative of the internal wave behavior in the area.

Some of the most obvious features in Fig. 5 include the inertial peak and the diurnal and semidiurnal tidal peaks. The semidiurnal peak is ubiquitous for all floats in the dataset, although this peak sometimes is blended with the near-inertial peak at higher latitudes. The diurnal peak is a common feature equatorward of 30° Poleward of 30°, the diurnal peak usually vanishes, though important exceptions exist that will be discussed later in the paper. The near-inertial peak is also a nearly ubiquitous feature in the dataset. For simple linear internal gravity waves, vertical motion (and thus temperature variability) vanishes at the inertial frequency; thus, the observed peak is likely due to contributions from near-inertial motion, when the waves oscillate with some vertical component. The dispersion relation for linear, freely propagating internal gravity waves confines the waves to exist in a band of frequencies bounded below by the inertial frequency and above by the buoyancy frequency. Here we define the buoyancy frequency $N$ as

$$N^2 = \frac{g}{\rho} \frac{\partial \rho}{\partial z},$$

where $g$ is the force of gravity, and $\rho$ is potential density. In Eq. (5), $N$ has units of radians per unit time, but in this study we convert to cpd by introducing a factor of $(2\pi)^{-1}$. A cutoff in the spectrum at frequencies below the inertial frequency can be seen in Fig. 5. The buoyancy frequency for float 5135 is about 30 cpd, well above the Nyquist frequency for our data (12 cpd), so the theoretical high-frequency cutoff cannot be seen in this case. The buoyancy frequency generally exceeds 12 cpd for floats at low- to midlatitudes. Only at high latitudes do some floats experience buoyancy frequencies that are lower than 12 cpd. In these conditions, we generally observe a steep drop-off in spectral density above the buoyancy frequency as predicted by simple linear theory. Less obvious harmonic signals are also noticeable in Fig. 5. There is a peak near the $2M_2$ and $M_2 + K_1$ frequencies ($M_2$ and $K_1$ are prominent semidiurnal and diurnal tidal frequencies, respectively). Although not present in Fig. 5, significant peaks at $M_2 + f$ are discernible for a number of floats (where $f$ is the local inertial frequency). For float 5135, the spectral density at frequencies above the semidiurnal peak is quite consistent with the GM spectrum in both magnitude and slope, as can be seen in Fig. 5 (spectra from other floats are also generally consistent with the GM spectrum and are examined in further detail in section 4b).
b. Observations of seamount-trapped waves

A second case study we present is notable for its uniqueness in the dataset. Float 6097 (WMO 5902118) was subjected to very large diurnal motion while in the vicinity of two seamounts around the Amundsen Ridge (see Fig. 6 for the trajectory). For freely propagating internal gravity waves, diurnal motion is generally not permitted poleward of $30^\circ$, as beyond these latitudes the diurnal frequencies are below the inertial frequency. Yet the averaged power spectrum from this float shows a strong peak at the diurnal frequency (see Fig. 6). As can be seen in Fig. 7 (the temporal evolution of the spectrum), however, this peak is not steady in time. The elevated diurnal activity only appeared when the float was in the immediate vicinity of the two seamounts. To examine this further, we quantify our observations by defining diurnal variance or semidiurnal variance $\sigma^2$ during $j$th park phase as

$$\sigma^2_j = \int_{\omega_j - \Delta\omega}^{\omega_j + \Delta\omega} S_j(\omega) \, d\omega,$$

where $\omega_j$ is the $K_1$ or $M_2$ frequency used for calculating the diurnal or semidiurnal variance, respectively. We have chosen $\Delta\omega$ to be 0.3 cpd in order to capture as much as possible of the variance in and around the semidiurnal and diurnal peaks. The semidiurnal variance is calculated to provide a standard with which we can compare changes in diurnal variance. As can be seen in Fig. 8, it would seem that there is a connection between proximity to the seamounts and the strength of the diurnal signal.

The variance in the semidiurnal and diurnal bands can be examined as a function of seafloor depth, which in this case can be thought of as a proxy for proximity to the seamounts. As can be seen in Fig. 9, the semidiurnal peak increases by about an order of magnitude over the seamounts compared to the open ocean. While a significant increase, this is greatly exceeded by the change in the diurnal signal, which increases by three to four orders of magnitude near the two seamount apexes. The vertical amplitude of the seamount-trapped waves measured by the floats is often around 100 m, and during early 2011, amplitudes in excess of 200 m were observed, as shown in Fig. 10.

In a modeling study, Brink (1990) found similar results, with subinertial variance near a seamount increasing several orders of magnitude above the open-ocean levels. Around the same time, moored observations showed that the diurnal tidal amplitude increased significantly near Fieberling Guyot in the eastern North Pacific (Eriksen 1991). Eriksen also found that there was a spring–neap cycle to the diurnal amplitude, a pattern we also observe for a month-long period as shown in Fig. 10. The data suggest that these two seamounts in the Amundsen Ridge could be sites of elevated mixing due to the large internal waves that are consistently present there; unlike superinertial internal gravity waves, these diurnal waves are trapped and not free to propagate away from the seamounts, implying that any energy initially input into baroclinic diurnal motion will ultimately be dissipated locally.

4. Multifloat observations

a. Global observations

While the float ensemble used here does not have complete spatial coverage over the world’s oceans, there are large regions of the ocean that are well sampled. These regions are diverse enough that we can begin to quantify how internal waves behave under a variety of conditions. Figure 11 shows a global plot of internal wave displacement variance derived from each park phase of every instrument included in this ensemble. The internal wave displacement variance during park phase $j$ is

$$\sigma^2_j = \int_{f}^{N} S_j(\omega) \, d\omega,$$

where $N$ and $f$ are the local buoyancy and inertial frequencies. These limits are chosen based on the properties of the dispersion relationship for linear, freely propagating
internal waves. In practice, it is not possible to integrate to the buoyancy frequency when it is above the Nyquist frequency. In these cases, we integrate only to the Nyquist frequency.

b. Comparisons to the Garrett–Munk spectrum

We examine the relationship between the observed internal wave spectrum and the expected Garrett–Munk displacement spectrum, originally formulated by Garrett and Munk (1975) and modified by Cairns and Williams (1976). To quantify the comparisons, we estimate the variance of the Garrett–Munk spectrum (hereinafter GM76, after the Cairns and Williams modifications) integrated between 3 and 6 cpd, and the variance of a fit of the float-derived spectrum integrated between 3 and 6 cpd. We choose to integrate over this particular frequency range because 3 cpd is well above the main tidal/inertial frequencies, while 6 cpd is almost universally lower than the buoyancy frequency observed by all the floats in the ensemble. Thus, this frequency domain should extend over a fairly consistent region of the observed spectrum.

The expected variance EV is established by integrating the local GM spectrum $GM_i(\omega)$ of each 30-day segment of each float from 3 to 6 cpd:

$$ EV_i = \int_{3 \text{ cpd}}^{6 \text{ cpd}} GM_i(\omega, f, N) \, d\omega. \quad (8) $$

Here $GM_i(\omega, f, N)$ depends on the local inertial and buoyancy frequencies and has been evaluated via MATLAB using the GM76 Toolbox (Klymak 2013, unpublished data; available online at http://hornby.seos.uvic.ca/~jklymak/GarrettMunkMatlab/). The subscript $i$ denotes the $i$th 30-day segment. To characterize the observed spectrum, we obtain a least squares fit of $S_i(\omega)$ between 3 cpd and the Nyquist or buoyancy frequency (whichever is lower). The fit is referred to as $F_i(\omega)$, and takes the form

$$ F_i(\omega) = C_1 \omega^{C_2}, \quad (9) $$

where $C_1$ and $C_2$ are constants defined by the best fit ($C_2$ is typically around $-2$, consistent with the GM expectations). Not all frequency bins of $S_i(\omega)$ above 3 cpd and below the buoyancy/Nyquist frequency are included in the fitting process, however. The frequency bins within 5% of $M_2 + K_1$, $M_2 + f$, and integer multiples of $M_2$ are not included in fitting $F_i$ so that it only characterizes the background slope of the spectrum, rather than include the effects of these peaks. Once $F_i(\omega)$ is obtained from $S_i(\omega)$, it is integrated from 3 to 6 cpd to calculate the observed variance $OV_i$:

$$ OV_i = \int_{3 \text{ cpd}}^{6 \text{ cpd}} F_i(\omega) \, d\omega. \quad (10) $$

The ratio of $OV_i/EV_i$ provides an estimate of the similarity between the observed internal wave spectrum and the GM spectrum for each 30-day segment; a global map of this ratio is shown in Fig. 12. The ratio exceeds one in typically energetic regions of the ocean such as the ACC,
the Hawaiian Ridge, and portions of the Gulf Stream. It is not surprising that the observed ratio is approximately one near the northwest Atlantic continental slope (the Woods Hole Oceanographic Institution site D, located at roughly 39°N, 70°W), the location of numerous long-term site moorings in the 1970s that had a large influence on the development of the GM models (Garrett and Munk 1975). There are also large regions where the internal wave displacement spectrum falls far below the expected GM spectrum. These regions often appear to

FIG. 10. (top) The spring–neap envelope of the barotropic zonal velocity (gray shading) and the internal wave displacement (black line) observed by float 6097 from November 2010 to March 2011. The dashed line plots seafloor depth, but only after mid-December, as the float was ice covered prior to this (preventing a satellite fix on location). Barotropic zonal velocity was obtained from the TPXO7.2 tidal model. Several large gaps are due to quality control criteria being violated. Note the two different scales for the (left) barotropic tide and the (right) internal wave displacement. (bottom) As in the above plot, but zoomed in on the month of January, where the diurnal motion is exceptionally large and dominates the displacement signal.

FIG. 11. Estimates of internal gravity wave vertical displacement variance from the 194 profiling floats used in this study. Values are averaged into 2° by 2° bins.
correspond to areas with smooth bathymetry or regions at high latitudes. Data within 10° of the equator were excluded as the GM spectrum from 3 to 6 cpd approaches zero at the equator (forcing $O_{V}/E_{V}$ to infinity).

c. Effects of local bathymetry and the barotropic tide

The float dataset used here is unique in its ability to observe the broad relationship between internal waves and bathymetric variability over large spatial scales. Although the float ensemble does not provide truly global coverage, the floats used have been deployed over diverse bathymetric conditions and thus provide a wealth of knowledge of internal wave intensity at a depth near 1000 m over both rough and smooth conditions. The distance from 1000 m to the seafloor is quite variable, so we can observe the effects of proximity to the seafloor on internal wave displacement. With the use of the tidal model, Ocean Topography Experiment (TOPEX)/Poseidon Global Inverse Solution 7.2 (TPXO7.2; Egbert et al. 1994), examination of the effects of the barotropic velocity on internal wave intensity can also be examined.

d. Internal waves versus roughness and proximity to bottom

We characterize the local bathymetric conditions (roughness and float height above bottom) using the 1-min gridded elevations/bathymetry for the world (ETOPO1; Amante and Eakins 2009) data within 100 km of the surface location. ETOPO1 has 1 arc-minute resolution of Earth’s land and seafloor topography and is constructed from a number of datasets that are primarily based on bathymetric soundings and satellite gravimetry. The 100-km radius is chosen as it is on the order of the horizontal wavelength of the first baroclinic mode internal waves of the semidiurnal frequency; our results are not particularly sensitive to this choice. The ETOPO1 data within 100 km of a float position are weighted in a Gaussian manner so that the bathymetry directly under the float is most heavily weighted, yet the influence of bathymetry at the fringe of the 100-km radius is still included (again, the results are not particularly sensitive to the weighting scheme). The weighting factor is

$$w_i = e^{-(L_i/100 \text{ km})^2},$$

where $L_i$ is the horizontal distance of the $i$th ETOPO1 data point from the location of the float park phase. We calculate a weighted average of depth $\bar{D}$ by weighting all individual ETOPO1 depths $D_i$ by $w_i$. Because floats park at approximately 1000-m depth, the height in meters above the bottom $H$ is then

$$H = \bar{D} - 1000.$$ 

The bottom roughness $R$ is the weighted standard deviation of depth around $D$:

$$R^2 = \frac{1}{J} \sum_{i=1}^{J} {\left[ \bar{D} - D_i \right]}^2 w_i,$$

where $J$ is the total number of depth measurements within the 100-km radius of the float. Although not shown in Eq. (13), we have removed the best-fit plane (by least squares) from the data so that smooth, sloping bathymetry (e.g., some continental slopes) is not parameterized as rough. From each park phase, we obtain three important estimates for comparing internal waves
to local bathymetry: 1) the internal wave variance $\sigma^2$ [Eq. (7)]; 2) the weighted bottom roughness $R$; and 3) the weighted height above the bottom $H$.

As an example, we show in Fig. 13 a time series of all three quantities for float 5061 (WMO 1900413). Float 5061 was deployed in the middle of the Indian Ocean, and over the course of its multiyear lifespan traveled westward across the Southwest Indian Ridge. During the period that the float crossed the ridge, $H$ decreased and $R$ increased, indicating shallower, rougher local conditions. Internal wave variance correspondingly increased significantly, which can be seen in Fig. 13. Similar correspondences can be seen throughout the float ensemble, as is shown in Fig. 14, showing internal wave variance as a function of height above the seafloor and bottom roughness for each park phase of every float.

e. Internal waves versus roughness and barotropic tides

As the barotropic tide generates fluid oscillations over bathymetric irregularities, internal waves are created. Thus, it is reasonable to hypothesize that the intensity of the internal waves generated will not only depend on the roughness, but also the strength of the barotropic tide. Here we examine the relationship between internal wave variance, barotropic velocity strength $U$ (as calculated by TPXO 7.2), and the bottom roughness (ETOPO1). Floats in regions of fast lateral advection have been removed from this analysis, because, as will be discussed (section 5a), internal lee waves can be a significant source of internal waves. This source of internal waves obscures the barotropic tidal signal we wish to examine; hence, these floats are not included. We remove floats that had a mean horizontal velocity (calculated from distance and time between profiles) greater than 0.1 m s$^{-1}$. These floats came primarily from the ACC, the Gulf Stream, the Kuroshio, and the equator and accounted for 18% of the total number of floats.

The definitions for internal wave variance $\sigma^2$ and bottom roughness $R$ are again defined by Eqs. (7) and (13), respectively. Both are calculated once for each park phase. The strength of the barotropic tidal velocity $U$ is also calculated once per park phase and is defined as

$$U^2 = \frac{1}{K-1} \sum_{k=1}^{K} [u_k - \bar{u}]^2 + [v_k - \bar{v}]^2. \quad (14)$$

Here $U^2$ is the combined variance of the east–west and north–south barotropic velocities $u$ and $v$, respectively. The velocities are computed onto an hourly grid during each park phase by the TPXO model 7.2 (Egbert et al. 1994), and $K$ is the total number of hourly velocities during the park phase. Figure 15 shows internal wave displacement variance plotted against $R$ and $U$. Internal wave variance appears to have noticeable correlations with both bottom roughness and the barotropic velocity. This suggests that strong tides oscillating over coarse bathymetry present favorable conditions to internal wave generation. Combined, Figs. 14 and 15 illustrate that, in a broad sense, internal wave intensity is strongly dependent on seafloor roughness, proximity to the seafloor, and the strength of the local barotropic tides. While these findings are not a revelation, it is still informative to observe general trends of internal waves.
over the wide range of conditions the float database encompasses.

f. Subinertial diurnal waves

The dataset provides broad float coverage poleward of 30°, especially in the Southern Hemisphere, and the measurements indicate a number of cases of notable diurnal intensification above the turning latitude (~30°), as shown in Fig. 16. Each point in Fig. 16 represents the difference of the diurnal peak from the background of the spectrum, rather than the total diurnal variance. To calculate the difference, first we integrate over three regions of each park-phase spectrum, including (i) the three frequency bins closest to one cycle per day; (ii) the three
frequency bins immediately above those in (i); and (iii) the three frequency bins immediately below those in (i).

The integrated variance arising from (ii) and (iii) are then averaged to estimate the background variance of the spectrum just outside the diurnal range. This average is then subtracted from (i) to estimate the degree to which diurnal variance exceeds the background variance of the spectrum (if at all), which we refer to as the diurnal prominence ($P$, in Figs. 16, 17). The park phases do not all have uniform duration. While most are close to 9 or 10 days, some are closer to 5 days, which causes the frequency resolution to change depending on the length of the park phase. This means that the three frequency regions [(i)–(iii)] do not occupy a completely uniform range for all floats (the frequency resolution is usually between 0.1 and 0.2 cpd). Because we are examining qualitative trends only, the moderate differences in frequency resolution do not significantly affect our conclusions.

We can also estimate the difference of the diurnal variance from the spectral background in an effort to negate any spectral broadening that may occur in areas of high overall variance (the ACC for example). In some cases there exists a relatively high diurnal variance even when no clear diurnal peak is discernible. While prominent diurnal peaks are not uncommon equatorward of 30°, poleward of this latitude discernible diurnal peaks become much less frequent and more localized. This topic is examined in detail in the following section.

5. Discussion of multifloat observations

a. Global observations

Perhaps the most striking feature in Fig. 11 is the intensification of internal wave variance in the path of the ACC east of Drake Passage. The pattern of intensification

![Figure 16](image1.png)

**Fig. 16.** A spatial plot of the diurnal prominence $P$ (see section 4f for definition) of each park phase for each float in the ensemble used here. Variance has been classified by different variance bins, indicated by the different sized/colored dots. The turning latitudes for diurnal waves are indicated by the black dotted lines. Park phases occurring when a float is under ice are not included, as locations are not accurately known.

![Figure 17](image2.png)

**Fig. 17.** As in Fig. 16, but focused on a region of the Southern Ocean (the southern Indian Ocean). The shaded contours indicate depth from ETOPO1. The brown shaded region is land.
is qualitatively similar to recent estimates of energy flux from geostrophic flows into internal lee waves (Nikurashin and Ferrari 2011). This suggests that lee waves could be a significant contributor to internal waves in the ACC east of Drake Passage. Both Whalen et al. (2012) and Wu et al. (2011) present detailed observations of diapycnal diffusivity in the Southern Ocean. Each study shows that, qualitatively, the spatial pattern of diffusivity is comparable to the pattern of internal wave variance in Fig. 11. Coupled with these recent studies, our observations help corroborate the notion that breaking internal waves play a key role in mixing the interior of the Southern Ocean.

Other regions with relatively large internal waves include the Hawaiian Ridge and the Gulf Stream. A number of works have shown Hawaii to be a strong source of internal tides (e.g., Alford and Zhao 2007). We also observe internal wave intensification in the vicinity of the Gulf Stream. Rainville and Pinkel (2004) found that internal wave intensity in the North Pacific can be significantly enhanced on the coastal side of the Kuroshio. The Kuroshio has strong horizontal velocity gradients, and the corresponding increase in vertical vorticity creates a barrier that low-frequency internal waves cannot cross (Kunze 1985). We hypothesize the same process may be trapping internal waves on the coastal side of the Gulf Stream, resulting in the elevated internal wave variance.

Curiously, internal waves appear to be relatively weak over much of the subtropical Indian Ocean. The Southwest Indian Ridge is a prominent bathymetric feature in this region, and because of its roughness, it is surprising that there is little or no enhanced internal wave energy.

The intensification of internal waves near the equator is intriguing. Whalen et al. (2012) estimates higher than average dissipation at the equator but acknowledges that this intensification could be a sampling artifact of features such as equatorial jets. We observe that internal waves are indeed more intense at the equator, lending credence to the notion of elevated dissipation observed by Whalen et al. (2012). There are also many other similarities between the global results presented by Whalen et al. (2012) and the internal wave intensity in Fig. 11.

There is no obvious seasonal cycle in the internal waves observed. This suggests that the internal waves arise from a fairly consistent forcing (at least on the time scale of months). Alford and Whitmont (2007) used horizontal velocity observations from moorings to show a factor of 2 to 3 increase in wintertime near-inertial energy compared to summertime estimates. The observed wintertime intensification often resulted from intermittent events, rather than being consistently more energetic than the summertime. Our float ensemble should be capable of observing such intermittency due to the hourly temporal resolution and high percentage of time spent in the park phase, yet generally we do not observe a seasonal dependence of the inertial peak. This is possibly because the float temperature technique used here is only capable of inferring vertical motions, thus skewing our results toward the steadier tidal signals rather than the predominantly horizontal motion of near-inertial waves.

b. Garrett–Munk comparisons

Caution is required in making generalized conclusions about how the GM spectrum compares to the observed spectrum from this dataset. Floats are often deployed at scientifically interesting locations, thus potentially biasing the results and conclusions. The GM ideas were formulated using primarily midlatitude data, while most of the floats in our ensemble are not at midlatitude. Furthermore, a large portion of our midlatitude observations were collected around Hawaii and the Gulf Stream, sites that are not representative of most of the ocean.

With these caveats noted, the mean of all OV/EV, shown in Fig. 12 is slightly less than one (0.7), which is fairly close to what would be expected. This low value could at least partially be explained by the single-depth measurements. The float data used here are all collected in the vicinity of a depth of 1000 m, thus making the modal structure of the internal wave field impossible to determine. If 1000 m lies at, or near, one of the vertical modal nodes (primarily the low modes), then the displacement inferred by the floats could be a significant underestimate of the maximum displacements elsewhere in the water column.

Another source of error could be the latitudinal dependence of the GM spectral amplitude. There is no such dependency observed in the real-world spectra of internal waves (Levine 2002). This could at least partially explain why the ratio of the observed spectrum to the GM spectrum appears to be so low (often an order of magnitude less than GM) at many high-latitude locations, thus contributing to lowering the global average.

c. Effects of local bathymetry and barotropic tide

Results shown in Fig. 14 suggest that when floats are over smooth bathymetry and far above the seafloor (the upper-left portion of the figure), they tend to observe relatively low-amplitude internal wave fields. When the floats are over rougher bathymetry, and closer to the seafloor (lower right), internal wave variance increases dramatically. This result is analogous to Polzin’s 1997 observations of diffusivity in the Brazil basin (Polzin et al. 1997), where diffusivity was dependent on proximity to the seafloor and seafloor roughness.

An analogous examination of the dependence of internal wave amplitude on both seafloor roughness and
the strength of the barotropic tide (Fig. 15) also indicates a relationship between both bottom roughness and the barotropic tide. This dual dependency is clearest for the roughness bins less than 400 m, and the barotropic velocity bins less than 2 cm s\(^{-1}\) (where most bins have well over 100 individual park-phase estimates). Figures 14 and 15 show that internal wave displacement is dependent on seafloor roughness, proximity to the seafloor, and forcing from the barotropic tide. It is striking and informative to observe the expected trends throughout this large dataset that covers such a diverse range of conditions.

d. Coastally trapped waves

In addition to seamount-trapped waves at the Amundsen Ridge (section 3b), we observe significant diurnal displacement at many locations poleward of 30°. Most of the discernible diurnal peaks poleward of 30° (see Fig. 16) occur near continental rises or other sloping bathymetry, suggesting that these diurnal waves are coastally trapped. We find little evidence of diurnal excitation in the pelagic ocean. While the seamount-trapped waves discussed in section 3b were fairly ubiquitous, the observations of coastally trapped waves in general are quite patchy. Figure 17 highlights a region in the Southern Ocean, on the shelf between 75° and 104°E, that showed significant diurnal activity, but not much else in the surrounding area. While there is considerable uncertainty in the cause of such shelf intensification, we posit that the patchiness of the diurnal signal near coastal rises might be partially explained by sporadic wind events. Storms could provide the necessary energy to excite coastally trapped waves. Intermittent wind intensification could also explain why we observe significant diurnal energy at one location and no such intensification at another nearby sites.

6. Conclusions

The combination of quasi-Lagrangian drift trajectories, relatively high-frequency sampling, and multiyear operating lifetimes make the iridium Argo floats a unique platform to observe internal waves. The float data are most instructive on global-scale internal wave behavior as demonstrated in section 4. There is significant similarity between the direct observations of internal waves presented here and studies that employ mixing parameterizations (section 5a). The wide range of geographic and oceanic conditions encompassed by our work helps to further validate such parameterization methods.

The nature of the Argo program is global in its scope; however, section 3b demonstrates that internal waves on a finer scale can be investigated from data collected from an individual float. There are almost undoubtedly individual iridium floats that could contribute significant information to smaller regional studies of internal waves.

The gaps in spatial coverage will slowly be filled, as there is a steady supply of new iridium floats. As the coverage improves, so too will our ability to address the questions examined in this work. A complete global coverage would also open up many more lines of inquiry.

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