Seasonal and Spatial Variability of Near-Inertial Kinetic Energy from Historical Moored Velocity Records

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ABSTRACT

Temporal and spatial patterns of near-inertial kinetic energy (KE\textsubscript{moor}) are investigated in a database of 2480 globally distributed, moored current-meter records (deployed on 690 separate moorings) and compared with the distribution of wind-forced mixed-layer energy flux \( F_{\text{ML}} \). By computing KE\textsubscript{moor} using short (30 day) multitaper spectral windows, the seasonal cycle is resolved. Clear winter enhancement by a factor of 4–5 is seen in the Northern Hemisphere for latitudes 25°–45° at all depths \(<4500\) m, in close agreement with the magnitude, phase, and latitudinal dependence of the seasonal cycle of \( F_{\text{ML}} \). In the Southern Hemisphere, data coverage is poorer, but a weaker seasonal cycle (a factor of 2) in both KE\textsubscript{moor} and \( F_{\text{ML}} \) is still resolvable between 35° and 50°. When Wentzel–Kramers–Brillouin (WKB) scaled using climatological buoyancy-frequency profiles, summer KE\textsubscript{moor} is approximately constant in depth while showing a clear decrease by a factor of 4–5 from 500 to 3500 m in winter. Spatial coverage is sufficient in the Northern Hemisphere to resolve broad KE\textsubscript{moor} maxima in the western portion of each ocean basin in winter, generally collocated with \( F_{\text{ML}} \) maxima associated with storm forcing. The ratio of depth-integrated KE\textsubscript{moor} to \( F_{\text{ML}} \) gives a replenishment time scale, which is about 10 days in midlatitudes, consistent with 1) previous estimates of the dissipation time scale of the internal wave continuum and 2) the presence of a seasonal cycle. Its increase to \(~70–80\) days at lower latitudes is a possible signature of equatorward propagation of near-inertial waves. The seasonal modulation of the magnitude of KE\textsubscript{moor}, its similarity to that in \( F_{\text{ML}} \), and the depth decay and western intensification in winter but not in summer are consistent with a primarily wind-forced near-inertial field for latitudes poleward of \(~25°\).

1. Introduction

The intermittency of the near-inertial peak in spectra of horizontal current has long been noted. In contrast to the rest of the internal-wave spectrum, which displays a remarkable constancy in time and space (Garrett and Munk 1975), near-inertial energy rises and falls on short time scales and varies strongly with location. Since these motions make up about half of internal-wave energy, and since the work done on them by the wind (Alford 2001, 2003a; Watanabe and Hibiya 2002) is a substantial fraction of that argued to be needed to keep the deep oceans stratified (Munk and Wunsch 1998), a better understanding of their spatiotemporal distribution seems warranted.

We undertake this here using a database of 2480 historical moored current-meter records on 690 globally distributed moorings (Fig. 1), improving substantially on those used in Fu’s (1981) pioneering work. Specifically, we will examine the hypothesis that near-inertial motions in the deep ocean are primarily wind generated, by comparing the spatial and temporal patterns in deep inertial kinetic energy, KE\textsubscript{moor}, with those expected for a wind-driven field.

Strong circumstantial evidence supports this idea—namely, observations of (a) wind-driven mixed-layer motions (Pollard and Millard 1970; Weller 1982) and (b) downward-propagating near-inertial motions at greater depths (Leaman and Sanford 1976; D’Asaro and Perkins 1984; Hebert and Moum 1994; D’Asaro et al. 1995; Alford and Gregg 2001), which often have similar downward energy flux to that input by the wind.

The wind generates near-inertial motions by imparting momentum to the surface mixed layer, leading to an ageostrophic flow that subsequently adjusts and radiates waves (Rossby 1938). Observations show that the mixed-layer flow is often well modeled as a slab (Pollard and Millard 1970), in which case the flux from
The wind to mixed-layer inertial motions, $F_{\text{ML}}$, and their energy, $K_{E_{\text{ML}}}$, are straightforward to compute (D’Asaro 1985; Alford 2003b; Watanabe and Hibiya 2002).

The ocean’s adjustment can be treated efficiently by projecting the “step” velocity profile (slab flow in the mixed layer and zero below) onto normal modes (Gill 1984). The physics of this process are that lateral inhomogeneities in the mixed-layer motions lead to convergent flow at the mixed-layer base, with the resulting pressure gradients forcing waves that are then free to propagate into the thermocline. The factors affecting the speed of this process are not yet fully understood (D’Asaro et al. 1995), but include the $\beta$ effect (D’Asaro 1989) and interaction with the mesoscale flow (Young and Ben Jelloul 1997; Balmforth et al. 1998), both of which can greatly accelerate the shortening of scales needed for efficient energy transfer.

Subsequent propagation into the thermocline can be described in terms of normal modes (Gill 1984), or rays (Kroll 1975; Zervakis and Levine 1995; Garrett 2001). Single-depth measurements such as those presented here cannot distinguish vertical structure of individual waves. However, the ray approach enables prediction of expected amplitude versus depth, since the amplitude of a wave packet generated near the surface is expected to scale according to Wentzel–Kramers–Brillouin (WKB) theory. Thus, the WKB-scaled energy for a single event is, to a first approximation, constant [as observed by Fu (1981)]; dissipation and lateral dispersion would lead to a decay with depth. In addition, the geometry of rays can be examined for single storms (Garrett 2001). Importantly, since the waves are close to the inertial frequency, they are constrained to propagate equatorward toward lower inertial frequency. Thus, energy at a particular location can in principle be related, for a single event, to the mixed-layer energy at the generation latitude. The observed frequency or “blueshift” above $f$ would be expected to increase with increasing depth, up to the first bottom bounce (Garrett 2001). Important for this study, this bounce occurs within a few degrees, allowing us to compare $K_{E_{\text{moor}}}$ with $F_{\text{ML}}$ estimated at the mooring site rather than attempting to ray-trace backward to the generation site.

Several other mechanisms can also plausibly generate near-inertial motions. In theory, any ageostrophic
flow (of which the wind-forced mixed layer is one) tends to radiate inertial waves as it adjusts (Rossby 1938). Bottom boundary layer flows, for example, can radiate near-inertial energy in like manner to the surface mixed layer. In another mechanism, the original ageostrophic flow results from an instability or “loss of balance” (Molemaker et al. 2005) of the large-scale flow. The energy transfer in this mechanism has not been estimated, but could potentially drain significant energy from the general circulation (Wunsch and Ferrari 2004). This process is distinct from other wave-mean flow interactions such as trapping (Kunze 1985) (which merely focuses near-inertial energy rather than generates it) and critical layers (Kunze et al. 1995) (which can be a sink). Whether trapped or generated by the mean flow, near-inertial energy would be expected to covary with the large-scale shear, as has been observed by several investigators (Frankignoul and Joyce 1979; Brown and Owens 1981). For example, Kunze and Sanford (1984) observed heightened near-inertial waves near a front, but were unable to determine whether the energy had been trapped or generated by the front.

In an alternate view, Fu (1981) obtained good agreement with moored spectra using a model wherein the near-inertial peak results simply from higher-frequency motions propagating poleward on a β plane. The waves’ frequency remains constant, but f increases. When they approach the turning latitude where ω = f, the resultant amplification manifests itself as a near-inertial peak. However, observations (Alford 2003a) indicate a predominance of equatorward rather than poleward propagation (at least for the first two vertical modes), conflicting with this explanation.

In addition to radiation from bottom boundary layers as discussed above, various other interactions between deep flow and topography can likely lead to near-inertial waves. Observations of upward energy propagation above the Madeira Abyssal Plain (Saunders 1983), and above Caryn Seamount (Kunze and Sanford 1986), were both suggestive of bottom-generated waves, but the mechanisms and importance are not well understood.

Last, at latitudes ±28.8°, parametric subharmonic instability (PSI) can potentially transfer energy from the semidiurnal internal tide into inertial motions (MacKinnon and Winters 2005; H. Simmons 2005, personal communication). Van Haren (2005) observed a peak in near-inertial energy in a small set of moorings in the Atlantic, supporting this hypothesis.

To test the hypothesis that the wind is the primary driver of deep inertial motions, near-inertial kinetic energy, KE\text{moor}, is computed in 30-day blocks at all available moored records, and WKB scaled to enable evaluation of patterns in depth, space, and season. It is compared at each location with the mixed-layer energy flux and energy, F\text{ML} and KE\text{ML}, computed by forcing a slab mixed-layer model (Pollard and Millard 1970), tuned with observations, with National Centers for Environmental Prediction (NCEP) reanalysis winds (Kalnay et al. 1996). These three elements—namely, the large amount of data, the shorter block sizes (30 days instead of a year), and our comparison with mixed-layer flux and energy—allow testing of our hypothesis, and distinguish our study from previous work. We find 1) strong seasonal cycles in KE\text{moor} at most depths/latitudes, which are similar to those in F\text{ML} at the same locations, 2) spatial patterns in winter KE\text{moor} are as expected for wind forcing, and are markedly different from summer. Though the occurrence of other mechanisms is certainly not ruled out, these findings identify the wind as an important driver for deep near-inertial motions.

The remainder of the paper is organized as follows. Data and methods are described in section 2. Depth, time, and lateral plots of KE\text{moor}, F\text{ML}, and KE\text{ML} are presented in section 3. A summary and discussion follow. Errors are discussed in the appendixes.

2. Data and methods

a. Data

Velocity records from more than 2200 historical moorings (over 7000 instruments) were acquired from the Oregon State University Buoy Group, which maintains a database of 1064 deep-water moorings deployed from the 1970s to the 1990s, from the National Oceanographic Data Center (245 moorings), the World Ocean Circulation Experiment (446 moorings), from the Autonomous Temperature Line Acquisition System (ATLAS) database (416 moorings), and from the Woods Hole Oceanographic Institution Buoy Archive (445 moorings; kindly provided by C. Wunsch). These data were initially selected as per Alford (2003a, hereinafter A03) according to the following four criteria:

1) total water depth greater than 3200 m,
2) sampling interval less than 3 h,
3) record length at least six continuous months, and
4) latitude poleward of 2.5°.

A03 required full-depth coverage to resolve flux. Our present focus on energy, which does not, greatly increases the number of usable moorings (690 moorings as compared with A03’s 60).

Coverage (Fig. 1) is adequate in the North Atlantic, and poorer but acceptable in the North Pacific. Else-
where, substantial geographical voids are present, as to be expected, in the Southern Hemisphere and at low latitude. In addition, potential spatial biases result from the disproportionately large number of records in “interesting places,” for example, near strong currents or topography.

b. WKB scaling

Waves propagating vertically through the ocean’s depth-varying stratification undergo refraction, which affects their vertical wavenumber and amplitude. To account for this effect, velocity measurements are typically “WKB scaled” to enable comparison between different stratification environments (e.g., Leaman and Sanford 1976). Here, each time series is WKB scaled using annual-mean climatological buoyancy frequency, $N$, computed at each mooring location from the 2001 World Ocean Atlas (Conkright et al. 2002):

$$\tilde{u}_{WKB}(t, z_i) = \tilde{u}(t, z_i) \left[ \frac{N(z_i)}{\bar{N}} \right]^{-1/2}, \quad (1)$$

where $\tilde{u} = u + iv$ is the measured rotary velocity, $z_i$ is the measurement depth, and $\bar{N}$ is the depth-mean stratification at the site.

The effect of WKB scaling is shown for a typical North Atlantic site in Fig. 2. Climatological $N^2(z)$ (Fig. 2a) decreases greatly in depth, but velocity magnitude WKB scales as $N^{1/2}$, resulting in amplifications/reductions in $\tilde{u}_{WKB}$ by factors of 0.5–2 (Fig. 2b). Hence, unscaled energy (right, dashed), which scales as $N$, typically decays by a factor of $O(10)$, but is more nearly constant after scaling (solid). The depth averages of scaled and unscaled energy are equal in a perfectly sampled water column (finite sampling introduces only minor errors dependent on the geometry). Since a goal is the calculation of lateral distribution of depth-averaged energy, this is a desirable property. (An alternate approach, for example, would be to scale values at all sites by the same constant $N$. Since column-averaged $N$ varies by less than a factor of 2 over the range of sample locations, the choice has little effect on the results.)

c. Spectra

The time series of $\tilde{u}_{WKB}$ are next divided into half-overlapping 30-day blocks, and the rotary frequency spectrum computed for each. The choice of 30 days represents a balance between frequency and temporal...
resolution. To maximize spectral resolution and precision for our shortened block sizes, the sine multitaper method of Riedel and Sidorenko (1995) is employed. This estimator exhibits superior precision and low bias relative to Welch’s (1967) method for comparable block sizes, without appreciably affecting leakage.

For each block, the spectral estimate $\hat{S}(\omega)$ is given by the average over all $K$ (here $K = 3$ for this study) tapers:

$$\hat{S}(\omega) = \frac{1}{K} \sum_{k=0}^{K-1} \hat{S}_k(\omega), \quad (2)$$

where $\omega$ is cyclic frequency and the $K$th estimate,

$$\hat{S}_k(\omega) = \Delta t \left| \sum_{t=1}^{M} h_{t,k} \tilde{u}_{WKB}(t) e^{-i 2 \pi \omega \Delta t} \right|^2, \quad (3)$$

is computed by applying the time-domain taper

$$h_{t,k} = \left( \frac{2}{M+1} \right)^{1/2} \sin \left( \frac{(k+1) \pi t}{M+1} \right) \quad (4)$$

to the time series. Here $M$ and $\Delta t$ are the number of measurements and sampling interval. The resulting spectral resolution,

$$\Delta \omega = \frac{(K+1)}{T}, \quad (5)$$

where $T = M \Delta t = 30$ days is the record length, is 0.133 cpd.

Sample time series and spectra from a location at 42°N, 46°W are shown in Fig. 3. Time series of WKB-scaled zonal velocity are presented at left for the entire year (Fig. 3a), together with closeups (Figs. 3b–d) of the periods indicated with gray shading in Fig. 3a. Near-inertial motions are visible in Fig. 3a as “fuzz,” but are resolved as near-inertial motions in subsequent closeups, which depict periods in order of decreasing near-inertial energy.

The rotary spectrum computed with the entire-year record (Fig. 3e) shows the typical broad near-inertial peak (e.g., Fu 1981), together with a narrow peak at the $M_2$ frequency. Both peaks are well above the level of the internal wave continuum, which shows a slope

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![Image](image_url)
near −2, as also typical (Fu 1981). Clockwise energy (black) greatly exceeds counterclockwise (gray) at lower frequencies, as expected for linear internal waves and often observed. Typical records exhibit peak frequencies 1.02–1.2 times as great as the local inertial frequency (dashed line); the present record’s peak is near 1.02f.

The sine multitaper spectral estimate from each 30-day period in Figs. 3b–d is plotted in Figs. 3f–h. The spectral resolution $\Delta \omega \approx 1/30$ cpd, shown by the width of the gray box in Fig. 3f, is obviously coarser than the full-length record (shown by the box in Fig. 3e), as evident in the broadening of the $M_2$ “line.” However, resolution is still easily sufficient to resolve the near-inertial and tidal peaks (though, e.g., $M_2$ and $S_2$ are not separable). Precision (indicated by height of gray boxes) is also reduced somewhat relative to the full-length record, but is still sufficient to resolve all but the weakest peaks. That in Fig. 3h is barely significant.

d. Near-inertial kinetic energy

The kinetic energy of the near-inertial peak (J kg$^{-1}$) is calculated for each segment by integrating between the half-power points $\omega_{\pm 1/2}$ (Fig. 3, gray shading) on either side of the near-inertial peak,

$$ KE_{\text{moor}}^{\text{in}} = \frac{1}{2} \left[ u_{\text{WKB}}^{\text{in}} \right]^2 = \frac{1}{2} \int_{\omega_{-1/2}}^{\omega_{+1/2}} \tilde{u} \, d\omega. \quad (6) $$

Because our focus is exclusively on the inertial peak, we will henceforth drop the $\text{in}$ superscript and refer to the moored kinetic energy $KE_{\text{moor}}$. Motions with the correct polarization for each hemisphere are isolated by integrating only over negative/positive frequencies in the Northern/Southern Hemispheres, respectively (Figs. 3e–h, black). Owing to the typical steepness of the falloff on either side of $f$, the choice of the integration limits (e.g., half-power or quarter-power) does not appreciably affect the calculations.

Both the height and the width of the near-inertial peak vary appreciably (a confirmation of the intermittency of the near-inertial peak). The yearlong record is therefore highly statistically nonstationary in the near-inertial band. The spectra computed with shorter blocks suffer much less in this regard, at the cost of reduced resolution. Note that these observations indicate that both peak height and bandwidth must be taken into account in computing $KE_{\text{moor}}$.

Here, $KE_{\text{moor}}$ calculated from (6) is indicated at lower left in each panel. The 30-day periods shown (Figs. 3f–h) bracket the value for the yearlong record (Fig. 3e). The associated RMS velocity signals ($u_{\text{WKB}} = 3.3, 1.8, \text{and } 0.7 \text{ cm s}^{-1}$, respectively) are visually consistent with the panels at left (Figs. 3b–d).

As latitude increases, the near-inertial peak moves closer to the semidiurnal tidal peak. To avoid contamination by tidal motions, the integration is truncated at $M_2 - \Delta \omega = 1.8025$ cpd. At 50°, 55°, and 60°, this is 18%, 11%, and 6% above $f$, respectively; hence, some underestimation may occur if the peak is wider than these limits. Because the variance in most peaks is contained at frequencies <1.1f, contamination is minimal equatorward of 55°.

Latitudes 25.7°–34.6° [$\sin^{-1} (1/2 (K_1 \pm \Delta \omega))$] are, in like manner, potentially affected by diurnal tidal motions. However, visual inspection indicates that in the vast majority of cases the near-inertial peak in spectra from yearlong records dominates the $K_1$ peak. Hence, we have not attempted to truncate our integrations but acknowledge the possibility of slight overestimation in this latitude range.

As discussed in appendix A, mooring motion leads to errors in $KE_{\text{moor}}$ that are different for each mooring owing to a variety of designs and environment. The most serious potential systematic bias is a possible underestimation when instruments are dragged down by strong currents into depths where WKB refraction reduces observed $KE_{\text{moor}}$ by as much as 40%. In appendix B, confidence limits on $KE_{\text{moor}}$ are discussed. Since the 30-day spectral estimates have but few degrees of freedom, uncertainty is large for individual estimates. However, the large number of instruments in our study allows subsequent averaging and reduction of uncertainty in averages to about a factor of 2.

e. Mixed-layer flux and kinetic energy

We compare our estimates of $KE_{\text{moor}}$ with estimates of mixed-layer flux and energy computed by running the Pollard and Millard (1970) slab model forced with NCEP–National Center for Atmospheric Research (NCAR) reanalysis winds (Kalnay et al. 1996) at the location and period of each mooring. The model is run using a latitude-dependent damping, $r = 0.3f$, and climatological monthly-mean mixed-layer depth (Levitus and Boyer 1994). Flux, $F_{\text{ML}}$, is then computed as described in detail by Alford (2001, 2003b). Over these time scales, the integrated mixed-layer energy,

$$ KE_{\text{ML}} = \frac{1}{2} \int_0^H \tilde{u}_{\text{ML}}^2 \, dz, \quad (7) $$

is simply related to the flux via

$$ E_{\text{ML}} = r^{-1} F_{\text{ML}}. \quad (8) $$

Because $r$ varies by less than a factor of 2 over the range of latitudes sampled here, $KE_{\text{ML}}$ and $F_{\text{ML}}$ have identical statistics, seasonal cycle, and spatial dependence. In the following, calculations and plots are of $F_{\text{ML}}$ but we
include KE_{ML} axes using the value of $r$ at 40° for comparison.

The $F_{ML}$ depends only weakly on the choice of damping $r$; however, (8) implies that energy depends on the damping coefficient, $KE_{ML} \sim r^{-1}$. Both quantities are potentially underestimated by up to a factor of 2 owing to the finite temporal (Alford 2003b) and spatial resolution (Jiang et al. 2005) of the NCEP winds, or overestimated by approximately the same factor owing to inadequacies in the slab model (Plueddemann and Farrar 2006). To maximize quantitative certainty, the model was tuned by selecting $r = 0.3f$ [about 2 times that used by Alford (2001, 2003b)], maximizing agreement between slab-model and observed $KE_{ML}$ at the 14 moorings from the database with instruments shallow enough to be in the climatological mixed layer for part of the year. A sample calculation at 41°N (Fig. 4) indicates that agreement is good between model and observed kinetic energy, even though the instrument is within the climatological mixed layer for only about 100 days (Fig. 4b). As will be seen in section 3c, $KE_{ML}$ is also in good agreement with midlatitude values observed by Park et al. (2005), who estimate $KE_{ML}$ from many Argo floats (these drift while uploading data, during which time inertial loops can be fit to their trajectories). Hence, though a rigorous error bound is not possible, we feel $F_{ML}$ and $KE_{ML}$ are accurate to better than a factor of 2 (probably better in northern midlatitudes, which contain most of the tuning data). Importantly, the magnitude of $F_{ML}$ is only important for our estimates of the replenishment time scale and does not impact our directly observed KE_{moor} results.

Because waves propagate downward and equatorward from the mixed-layer forcing region, $KE_{ML}$ and $F_{ML}$ would be best estimated at the horizontal location determined by ray tracing from each observation depth back to the surface. However, for simplicity we compute the mixed-layer energy and flux at each mooring location. Because rays reach the bottom within 400 km of their source (Garrett 2001) except within 10° of the equator and poles (where we have no data), this is a justifiable simplification provided our focus is gross seasonal energy comparisons rather than event-by-event comparison (and far preferable to the uncertainties introduced by “reverse ray tracing” to determine the best forcing location for each mooring).

3. Results

a. Case studies

Sample 2-yr time series of near-inertial kinetic energy are shown in Fig. 5. Local winter, defined arbitrarily as November–February and May–August for the Northern/Southern Hemispheres, is shaded gray. In each panel, the WKB-scaled near-inertial energy $KE_{moor}$ (solid colors) and the energy flux from the wind to the mixed layer, $F_{ML}$ (dashed; note different units and scale factor), are plotted. To attempt to illustrate representative variability, examples are selected from the Northern (upper) and Southern (lower) Hemispheres, as well as from the Atlantic (left) and Pacific (right) Oceans. Ordinate limits are the same for all four panels.

Considerable variability in $KE_{moor}$ is seen both in individual records and as a function of location. In addition, the small number of degrees of freedom of these 30-day spectral estimates results in large formal error bars (upper left; see appendix B). However, the following qualitative observations may be made, which will be confirmed statistically (using all moorings) in the following sections:

- Winter enhancement of $KE_{moor}$ is evident at some depths at each location. It is often seen at great depth (most clearly in Fig. 5a, which shows a factor-of-10 seasonal modulation at all depths).
- The magnitude and phasing of seasonal modulation of mixed-layer energy (dashed) is very similar in each example to that observed in $KE_{moor}$. At one site (Fig. 5c), even shorter-period fluctuations in $KE_{moor}$ and $KE_{ML}$ appear correlated.
- In the Southern Hemisphere examples (bottom), seasonal modulation of both $KE_{moor}$ and $KE_{ML}$ appears weaker than in the Northern Hemisphere.
- When WKB-scaled, no obvious depth trend is discernible in $KE_{moor}$ in these examples (subsequent statistics will reveal a depth decay during winter).
The seasonal modulation of deep KE_moor at these sites, and its similarity in amplitude and phase to wind-forced mixed-layer energy, are suggestive that the inertial motions are associated with wind forcing. Because of the variability evident in these few examples, and large error bars in single spectral estimates, the remainder of the paper will focus on composites of all moorings. Because we wish to eliminate near-surface influences, we will henceforth consider only instruments deeper than 300 m.

b. Depth dependence

We first examine the depth dependence of WKB-scaled moored energy (Fig. 6). We first focus on the Northern Hemisphere, since coverage is so much better and to simplify the presentation. All values are plotted versus depth (Fig. 6a) and versus height above bottom (Fig. 6b) as dots, with darker grays indicating winter months. Oceanographers’ predilection to place instruments at standard depths is clear. The high degree of variability, which is greater at shallow depths, documents the well-known intermittency of near-inertial motions.

Data are next binned versus month and averaged in 500-m depth intervals (colored lines). The number of values in each month’s average, indicated in the legend (bottom left), varies somewhat from month to month since most records do not span exactly a year. The mean number in each depth bin is used to determine 95% confidence limits as described in appendix B (upper left, gray), which are less than a factor of 2 above 4000 m; below this, errors grow owing to the paucity of deep measurements.

The magnitude of bin-averaged KE_moor is \((1-10) \times 10^{-4} \text{ J kg}^{-1} (1.4-4.4 \text{ cm s}^{-1})\); individual values can be an order of magnitude higher or lower. In comparison with the unscaled quantity (Fig. 2), WKB-scaled KE_moor decreases much less in depth during all months. This lack of strong depth dependence, also observed by Fu (1981), greatly simplifies subsequent analysis, since energy at a given location and time can be efficiently characterized by the depth mean. However, in winter months (blue), a clear decay by a factor of \(4-5\) is significant at the 95% level (left, gray) from 500 to 3500 m, with more irregular and marginally significant structure below 4000 m. In summer, WKB-scaled energy is nearly constant, to within measurement error.

These conclusions are sharpened somewhat when plotted versus height above bottom instead of depth (Fig. 6b), perhaps owing to the more uniform sampling and resultant reduction of the “standard depth” biases.
in (Fig. 6a), in addition to hypsometry effects. Here, summer KE\textsubscript{moor} is constant or decreases by less than a factor of 2; a clear depth decay by about a factor of 4 from 5000 to 500 m above bottom is seen in winter. As a result of this decay, seasonal modulation is nearly undetectable within 500 m of the bottom. This reduction is a possible signature of topographic generation of inertial waves, which would not be expected to show a seasonal cycle. Above this, the observed winter decay with depth is consistent with an energy source at the surface (though meridional propagation of near-inertial waves complicates 1D interpretations). The observed seasonal cycle of not only the magnitude of KE\textsubscript{moor} but also its degree of depth decay are strongly suggestive that the surface energy source is the wind, whose forcing is strongest in the winter.

c. Seasonal cycle

To examine seasonal variability more closely, depth-mean KE\textsubscript{moor}, F\textsubscript{ML}, and their ratio (which yields a replenishment time scale) are averaged within 10° latitude bands and plotted by month to form a canonical seasonal cycle (Fig. 7). In contrast to the depth-month view just presented, which averaged all northern latitudes together, the seasonal cycle is here examined as a function of latitude. (We continue to focus for the moment on the Northern Hemisphere.) Each quantity is plotted as a thick line whose thickness indicates the 95% confidence limits.

Marked seasonal cycles well in excess of error bounds are seen in the latitude bands centered on 30°, 40°, and 50° (orange, yellow, green). The amplitude of
the former two is about 4–5, with the latter closer to 2–3. The highest latitude range (blue) shows a “two-
plateau” cycle, marginally resolved, where energy in
June–December exceeds February–April energy by a
factor of about 2. Energy in the 15°–25° band (which is
the most data poor, and strongly weighted toward lati-
tudes >20; see Fig. 1) is constant to within measure-
ment errors.

The same analysis is applied to $F_{\text{ML}}$ in Fig. 7b (note
wider ordinate scale). As indicated previously, monthly
mean $F_{\text{ML}}$ and KE$_{\text{ML}}$ differ only by a multiplicative
factor (8), allowing KE$_{\text{ML}}$ values to be indicated at right
using the value of $r$ at 40°. The conversion is correct at
40° (yellow), 10% too high at 50°, and a factor of 2 too
low at 20%, but these differences are nearly impercep-
tible in Fig. 7b. For comparison, KE$_{\text{ML}}$ measured by
fitting inertial loops to the drift tracks of Argo floats
and averaged from 30° to 60°N in the Atlantic (circles)
and Pacific (pluses) are overplotted. Since the floats
measure mixed-layer depth, they do not rely on a cli-
matology. Though the latitude range sampled by these
floats is not the same as here, their magnitude and sea-
sonal cycle agrees well with our 35°–45° curve (which
appears to contain most of their data). This agreement
gives confidence in the $F_{\text{ML}}$ and KE$_{\text{ML}}$ estimates.

As seen from the case studies, the magnitude of the
seasonal cycle of wind forcing in each band is generally
similar to that in KE$_{\text{moor}}$. The strongest modulation is
seen spanning 30° and 40°, as for KE$_{\text{moor}}$, with similar
magnitudes. At lower latitudes, $F_{\text{ML}}$ shows weak/ 
undetectable seasonal modulation, like KE$_{\text{moor}}$. An
exception is at the highest latitudes (55°–65°, blue), where
$F_{\text{ML}}$ shows a much stronger cycle than KE$_{\text{moor}}$ [How-
ever, its phasing (minimum in April) is the same as for
KE$_{\text{moor}}$.]

The ratio of depth-integrated KE$_{\text{moor}}$ to the energy
flux $F_{\text{ML}}$ gives a replenishment time scale,

$$\tau = \frac{\rho D(\text{KE}_{\text{moor}})}{F_{\text{ML}}}, \quad (9)$$

which can be interpreted as the time it would take to
substantially reduce KE$_{\text{moor}}$ (a factor of $e^{-1}$) if forcing
stopped. The mean depth of the ocean is $D$, taken as
4000 m. Here, we have taken advantage of the near
constancy with depth of KE$_{\text{moor}}$ (Fig. 6) to estimate its
depth integral in (9), which would not otherwise be
possible from these vertically sparse data.

The value of $\tau \approx 10$ days at latitudes 35°–45° (Fig. 7c,
yellow), increasing to 50–60 days in the Tropics (red).
Only at high latitude does $\tau$ exceed 100 days during a
portion of the year, associated with the $F_{\text{ML}}$ minimum
when climatological mixed layers in the North Atlantic
are at their deepest (and possibly poorest known owing
to the spatial patchiness of deep convection). At all
other latitudes, $\tau$ shows a weak or nonexistent seasonal
cycle, suggesting a temporally constant relationship be-
tween KE$_{\text{moor}}$ and $F_{\text{ML}}$.

For comparison, dissipation time scales computed in
like manner for the internal wave continuum bracket
these estimates (7 and 70 days for “ Abyssal Recipes” and
“pelagic” mixing rates of $K_p = 10^{-4}$ and $10^{-5}$ m$^2$
s$^{-1}$, respectively; Munk 1981). Though not definitive
evidence for wind forcing, the observed seasonal cycle
and calculated time scales are consistent with it; for
example, observation of a strong seasonal cycle and $\tau \approx 
1$ yr would be inconsistent, since energy reserves would
be too large to expect a seasonal cycle.

d. Probability density

We next present the statistics of KE$_{\text{moor}}$ and $F_{\text{ML}}$ and
at last consider the Southern Hemisphere. Probability
density functions (Fig. 8) indicate that both quantities
are qualitatively lognormal. The distributions of $F_{\text{ML}}$
and KE$_{\text{ML}}$ (right panels; axes on top and bottom, re-
spectively) are broader (more intermittent) than
KE$_{\text{moor}}$, as expected owing to their strong dependence
on wind speed (D’Asaro 1985; Alford 2001). Winter enhancement of KE$_{\text{moor}}$ is once again obvious in the
Northern Hemisphere (Fig. 8a). By contrast, austral
winter enhancement in the Southern Hemisphere (Fig. 8b)
can be barely discerned. The relative scarcity of Sou-
thern Hemisphere data is also apparent from the
insets, which indicate the number of observations in
each month.

Both the winter enhancement seen in the Northern
Hemisphere, and its relative lack in the Southern
Hemisphere, are replicated in the distributions of $F_{\text{ML}}$
and KE$_{\text{ML}}$. In the former case, $F_{\text{ML}}$ is clearly greater,
and more intermittent, during winter. In the Southern
Hemisphere, only a very weak seasonal cycle in KE$_{\text{moor}}$
and $F_{\text{ML}}$ is observable. (As seen next, this is partly due
to the strong weighting of the Southern Hemisphere
data toward tropical latitudes, which show weak sea-
sonal cycles in both hemispheres.) At the sampled lo-
cations, the seasonal cycle of midlatitude storms, which
force mixed-layer motions, is much weaker in the
Southern Hemisphere. These patterns are replicated in
the deep observed inertial kinetic energy.

e. Spatial structure

The spatial dependence of depth-mean KE$_{\text{moor}}$ is av-
ergaged over May–August (northern summer; Fig. 9a)
and November–February (northern winter; Fig. 9b).
The four-month averages, each of which is plotted as a
colored dot, show a great deal of variability. This again
reflects the intermittency; a single event occurring in
one record can strongly influence the mean. Nonetheless, some patterns are apparent. First, elevation of KE$_{\text{moor}}$ in each region during local winter is evident as seen in previous plots. Second, many of the most energetic northern winter values (Fig. 9b) occur in the western parts of ocean basins, where the strongest wind forcing is found (grayscale; Alford 2003b). This trend is marginally seen in the North Pacific, but is fairly well resolved in the North Atlantic where coverage is densest. Of course, mesoscale activity is strongest in the west as well, as would be expected for a generation by geostrophically adjusting mesoscale flows. However, the western intensification is absent during summer (Fig. 9a), again implicating the wind.

The latitude dependence of KE$_{\text{moor}}$ and its seasonal cycle is examined (Fig. 10a) by plotting energy versus latitude (dots; grayscale indicates month) and computing the zonal mean in 5° bins (colors). Supporting results in previous sections, a seasonal cycle is clearly evident for northern latitudes 25°–45°, with the greatest modulation (factor of 5) occurring near 35°–40°, consistent with Fig. 7a.

A seasonal cycle of a factor of 2–3 is barely resolvable at the 95% level in southern midlatitudes between 35° and 50°, but with smaller magnitude than in the Northern Hemisphere. As can be seen from Fig. 9 and the error bars in Fig. 10a, Southern Hemisphere data are weighted toward the Tropics. Hence, the last section’s probability density analysis (which did not separate versus latitude) showed a nearly undetectable seasonal modulation owing to the more numerous low-latitude records.

Broad winter maxima of KE$_{\text{moor}}$ ~ $5 \times 10^{-4}$ J kg$^{-1}$ are seen near 30°–40° in both hemispheres (coincident with the midlatitude storm track), decreasing by a factor of 3–4 toward higher latitudes. As discussed, some underestimation is possible for latitudes >55° in order to avoid the M2 peak, but the falloff occurs well before this. At lower latitudes, coverage is poor, but both hemispheres show a decrease near 20° relative to the midlatitude value, before rising again approaching the equator. In summer, the latitudinal distribution is much flatter.

High KE$_{\text{moor}}$ values are observed near 28°N, which elevate the mean in this latitude range by a factor of 2–3. As discussed, the possibility of contamination by diurnal tidal motions cannot be ruled out. It could also be the signature of PSI, as argued by van Haren (2005) from investigation of seven moorings along 20°W (those near 28°N are a subset of the moorings from this

![Fig. 8. Probability density of (a), (b) KE$_{\text{moor}}$ and (c), (d) F$_{\text{ML}}$ in the (a), (c) Northern and (b), (d) Southern Hemisphere. In (c), (d), the top axis corresponds to F$_{\text{ML}}$; the bottom axis corresponds to KE$_{\text{ML}}$ via (8) using r at 40° (see text). Colors represent the conditional population for each month, with the number of observations indicated in the legend.](image-url)
4. Summary

Temporal and spatial patterns of near-inertial kinetic energy $\text{KE}_{\text{moor}}$ were investigated in historical moored current meter records at 690 locations, with the aim of 1) documenting the spatiotemporal distribution of near-inertial energy and 2) determining whether extant data are consistent with a predominantly wind-forced near-inertial wavefield. Here $\text{KE}_{\text{moor}}$ is computed by integrating spectra computed over short (30 day) blocks, allowing the seasonal cycle to be resolved. It was compared with mixed-layer energy flux $F_{\text{ML}}$, computed from a data-tuned slab mixed-layer model forced with NCEP reanalysis winds at the time and location of each mooring. The data support the following conclusions:

- Northern Hemisphere $\text{KE}_{\text{moor}}$ exhibits a seasonal cycle at all depths <4500 m or >500 m above bottom (Fig. 6). Modulation is greatest (a factor of 4–5) near latitude 40°, but significant at the 95% level for latitudes 25°–45° (Figs. 7 and 10). A similar cycle, but substantially weaker, is barely resolved at the 95%
level in the Southern Hemisphere, despite poor data coverage (Figs. 7 and 8).

• Near-inertial signals are detectable in all seasons at all depths. WKB-scaled Northern Hemisphere KE_moor is approximately constant in summer, but in winter it decays in depth by a factor of 3–4 from 500 to 3500 m (Fig. 6), implying a surface, winter-enhanced energy source.

• The lateral distribution of Northern Hemisphere summer KE_moor is constant to within a factor of 2–3 in both longitude (Fig. 9), and latitude (Fig. 10). In winter, enhancement is seen preferentially from 30° to 50° in the western Atlantic and Pacific. In the Southern Hemisphere, the winter enhancement is in

the same latitude range, but data are not sufficient to resolve longitudinal structure.

• The magnitude and phase of the seasonal cycle, its western intensification, and its relative lack in the Southern Hemisphere are all mimicked in the distribution of wind forcing on the mixed layer $F_{ML}$.

• The ratio of KE_moor to $F_{ML}$, which can be interpreted as a replenishment time scale, is $\approx$ 10 days in midlatitudes (Figs. 7c and 10c), similar to time scales in the rest of the internal wave spectrum and consistent with the presence of a seasonal cycle. Its increase toward the Tropics possibly indicates that a substantial portion of inertial energy there has propagated from higher latitudes. Alternately, it may suggest the importance of PSI in generating near-inertial motions at and equatorward of latitude 29°.

5. Discussion

As discussed in the introduction, a variety of processes generate near-inertial waves, and we have ab initio admitted a certain bias toward the wind. The observed seasonal cycle, together with the wintertime presence of the expected patterns in depth (a decay, implying a surface source), longitude (western intensification), and latitude (peak at and equatorward of the storm track) strongly implicate the wind. However, the data do not rule out the importance of other mechanisms at specific times and places. For example, enhanced energy near 29°N [Fig. 10a; also seen by van Haren (2005)] and the increased time scale and reduced seasonal cycle to the south (Fig. 10c) may implicate the PSI mechanism at low latitudes. Likewise, the reduction of the seasonal modulation near the bottom may be evidence for topographic generation. In addition, the observed western intensification and slight decay with depth would be expected for generation by geostrophically adjusting mesoscale features (which are strongest near the surface and western boundary currents). However, the absence of these patterns during summer again implicates the wind.

The pathways of the energy from surface to the deep ocean are still far from clear. Energy is expected to propagate downward and equatorward from generation locations along characteristic rays (Garrett 2001). For general initial wavenumbers, the distance and travel time to the bottom are functions of the vertical mode number, $j$. In midlatitudes, these are 200–400 km and 8.3j days, respectively. Hence, near-bottom KE_moor of low-mode waves should be related, with short time lag, to surface forcing. Higher modes take proportionally longer (166 days for mode 20).

The existence of a seasonal cycle of deep KE_moor in
phase with that of $F_{ML}$ thus places constraints on the vertical wavenumber content of the global wavefield. Assuming that travel times a substantial fraction of a year would wash out the cycle or cause it to be out of phase with $F_{ML}$, modes less than about 20 must dominate. Otherwise, the deep seasonal cycle would be incoherent and/or out of phase with atmospheric forcing, contrary to observations. Estimates of the wavenumber spectrum of near-inertial waves are scarce, but D’Asaro and Perkins (1984) measured a red spectrum (spectral slope $\approx j^{-2}$) in the Sargasso Sea, in support of this hypothesis.

The wavenumber partitioning of storm-generated motions is therefore of key importance. Unfortunately, it is poorly known; linear theory described the evolution of low modes during the Ocean Storms Experiment (D’Asaro et al. 1995) fairly well but failed for the higher modes. A sizable portion ($\approx 35\%$) of energy input by storms escapes the region in the first few modes, which can be detected in full-depth historical moored records (Alford 2003a). Because these low modes potentially carry much of the energy input, and can propagate great distances, the ultimate fate of this energy (and resulting distribution of the dissipation) is important to the budget and geography of the deep ocean’s mixing.

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APPENDIX A

Mooring Motion

Mooring motion influences our measurements in three ways, which we discuss here in turn. Because mooring design and performance vary greatly over our database, we estimate the worst-case resulting errors rather than attempting to correct for the errors in each case.

First, a mean flow can cause a subsurface mooring to blow over, causing the current meters to measure flow at greater depth. Because velocity generally decays with depth, this can cause a low bias (Hogg 1986, 1991). Depressions of the top of the mooring by several hundred meters are observed in high-drag moorings in strong currents; this would push the meter to a region where $N$ was lower by up to 40%. Assuming WKB scaling, $KE_{moor}$ would be underestimated by the same factor during blowdown periods. Deeper meters are deflected proportionally less. Hence, a four-instrument mooring would be expected to underestimate the true depth-mean WKB-scaled kinetic energy by 10%–20%. This is a potential (slight) source of bias since near-inertial energy near strong flows (where generation by geostrophic adjustment would be most expected) is underestimated.

Second, the inertial motions themselves can accelerate the mooring. A steady inertial current would cause a mooring with infinitely fast response to describe a perfect circle, in phase with the currents and thus to underestimate the true inertial currents. Real moorings have a finite temporal response dependent on the depth-dependent inertial currents; likewise, real inertial currents are noncircular and nonsteady. Hence, the detailed modeling of the moorings’ inertial response would therefore be formidable even with detailed motion records. Instead, a rough upper bound is obtained by examining all available shallow pressure records from the database of moorings. In the inertial band, these should be associated with deflections of the mooring by inertial currents. The largest inertial pressure fluctuations we observed were $\pm 10$ dbar, corresponding to lateral deflections of 350 m for diurnal motions of a rigid 5000-m mooring, or $u_{err} = 2.5$ cm s$^{-1}$ for circular motions. Because these strong deflections occurred during strong inertial currents ($>10$ cm s$^{-1}$), and we expect $u_{err} \sim \bar{u}$, we estimate a worst-case associated error of $2.5/10 = 25\%$ in $\bar{u}$ (6% in $KE_{moor}$). Because the mooring response decreases with depth faster than the expected WKB decay, the fractional error decreases with depth.

Last, surface wave pumping can cause shallow instruments to overestimate the true current speed owing to a combination of several effects (Hamilton et al. 1997, and references therein). This class of errors can safely be ignored owing to our focus on $>300$ m.

APPENDIX B

Confidence Limits

Each moored $KE$ estimate is the mean of $N$ spectral estimates of complex velocity, where $v = \Delta\omega_{peak}/\Delta\omega_{spec}$. For our peak widths ($\approx 0.1$ cpd) and resolution ($\approx 0.03$ cpd), we expect $v \approx 3-4$. (Note we do not use $2\nu$ because we only sum over one sign of frequency.) If velocity is Gaussian, the distribution of scaled energy,

$$\overline{KE} = \frac{vKE}{E(KE)}, \quad (B1)$$
where $E(KE)$ is the expectation value, should be a $\chi^2$ variable with $n$ degrees of freedom. Subsequent averages of $n$ KE$_{moor}$ values are therefore $\chi^2$ variables with $nv$ degrees of freedom. The observed distribution is consistent with $v = 3$ (Fig. B1, black). Confidence limits are readily computed using standard methods (e.g., Bendat and Piersol 1986), as done for squared shear averages by Gregg et al. (1993).

Each mixed-layer $F_{ML}$ (or KE$_{ML}$) estimate is the average over 30 days of a number of individual values, $F_i$. Their distribution is not Gaussian (D’Asaro 1985; Alford 2001); hence, the distribution of the 30-day mean values more closely resembles a $\chi^2$ variable with one degree of freedom (Fig. B1, gray). This resemblance allows calculation of confidence limits in the same manner as for KE$_{moor}$, but using a $\chi^2$ distribution with $n$ rather than $3n$ degrees of freedom. Because 30-day mean $F_{ML}$ and KE$_{ML}$ are related by a constant factor $r$, their confidence limits are the same.

**REFERENCES**


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![Fig. B1. Probability density of scaled KE$_{moor}$ (black) and KE$_{ML}$ (gray), together with $\chi^2$ distributions for four and one degrees of freedom, respectively.](image-url)


