Tidal and near-inertial peak variations around the diurnal critical latitude

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[1] Spectra from historic long-term open-ocean moored current meter data between latitudes $0° < |\varphi| < 45°$ reveal a significant drop in semidiurnal tidal band (D_2) energy by $\sim$50% at $|\varphi| \approx 25–27°$, whilst the peak near the local inertial frequency f is increased by a factor of $\sim$10 up to the level of D_2-energy at $|\varphi| \approx 28–30°$, where f coincides with diurnal frequencies. The increase in f-energy is accompanied by a red-shift of the peak frequency to 0.97 ± 0.01f, or a poleward spreading of enhanced energy. This contrasts with more common blue-shift. The enhancement may be the result of sub-harmonic instability, as supported by sparse significant bicoherence at half-D_2, although i) systematic enhancement of diurnal tidal frequencies, notably M_1, was not observed, ii) the latitudes of low D_2-energy and high f-energy do not coincide. This may be due to a mix of coupled and independent waves, whilst the poleward trapping of sub-f energy suggests non-traditional effects. **Citation:** van Haren, H. (2005), Tidal and near-inertial peak variations around the diurnal critical latitude, *Geophys. Res. Lett.*, *32*, L23611, doi:10.1029/2005GL024160.

1. Introduction

[2] The contribution of breaking internal waves to deep-ocean mixing is largely dependent on the existence of short vertical length scales (high vertical modes) of the motions that induce shear instabilities. Whilst kinetic energy is dominated at semidiurnal ('D_2') tidal frequencies (most at harmonic frequencies M_2, S_2, N_2), debate is ongoing how motions at these frequencies transfer their energy to small scales such that large vertical current shear is established that induces the breaking of internal waves near the buoyancy frequency N.

[3] Observations [*Leaman and Sanford*, 1975] show that vertical scales at the inertial frequency $f = 2\Omega\sin\varphi$, the vertical component of $\mathbf{\Omega}$, the Earth rotational vector, at latitude φ, are quite short O(100 m) in the open ocean. A resonant wave-wave interaction model has been proposed [*McComas and Bretherton*, 1977] for the generation of small vertical scales following 'parametric sub-harmonic instability' (PSI), a term used for a process in which energy is transferred from energetic large-scale waves to small-scale waves at half the original frequency, until the energy level of the latter is of the same order as that of the former. Although historic estimates of the transfer rate yielded O(100 days) for weak interactions, renewed interest for the importance of strong non-linear transfer was recently aroused following numerical estimates O(10 days) for the transfer rate from a random internal wave field [*Hibiya et al.*, 2002] and $\sim$5 days for a mode-1 internal tidal wave [*MacKinnon and Winters*, 2005]. The recent models focus on transfer of energetic D_2-motions to smaller f-scales, so that e.g. at $|\varphi| \approx 29°$ $2f \approx 1.4\cdot10^{-4}\mathrm{s}^{-1} \approx M_2$.

[4] As traditionally free internal gravity waves exist at frequencies (σ) in a band between $f < \sigma < N$, dominant D_2-waves are thought to transfer energy via PSI to free propagating waves only when their diurnal half-frequencies D_1 fall within the internal wave band; that is at $|\varphi| < 29.91°$ for $S_2 \rightarrow S_1$ generation, $|\varphi| < 28.80°$ for $M_2 \rightarrow M_1$ and $|\varphi| < 28.21°$ for $N_2 \rightarrow N_1$. As a result, one expects to observe due to PSI across the above 'critical' latitudes φ_c i) a transition in tidal energy with attenuation of D_2-energy on the equator-side of φ_c, ii) a transition in f/D_1-shear and associated mixing, with enhanced mixing on the equator-side of φ_c, iii) largest f-energy and shear around φ_c.

[5] So far, most of the evidence of PSI comes from numerical modeling, which shows local enhancement of near-inertial energy around $|\varphi| \approx 29°$ [*Hibiya et al.*, 2002] and a dramatic decrease in amplitude but only between $27.5° < |\varphi| < 29.5°$ [*MacKinnon and Winters*, 2005]. Multi-mode (beam) D_2-waves seem to transfer to subharmonic waves as well as to higher harmonics [*Lamb*, 2004].

[6] Open ocean observations of PSI are mainly limited to tethered falling XCP and XBT observations from 0$\sim$1500 m above topographic ridges, notably in the Pacific, from which enhanced mixing by a factor of 2–3 is estimated between $22° < |\varphi| < 32°$ compared to poleward observations [*Hibiya and Nagasawa*, 2004]. Short (4.5 days) shipborne ADCP data from 0–800 m above Hawaiian Ridge suggest a very local non-linear coupling between D_2 and D_1 at $\sim$520–580 m [*Carter and Gregg*, 2005], but no conclusions could be drawn on specific individual frequencies within D_1, due to the shortness of record.

[7] In this paper, more than half-year long deep-ocean moored current meter records are used to further investigate possible importance of PSI for open-ocean internal wave induced mixing. The focus is on the latitudinal variation of: i) f-energy; ii) D_2-energy, and iii) the energy at non-tidal diurnal harmonic frequencies like M_1 and N_1 (S_1 was not distinguishable from astronomically forced K_1). Moorings above large topography are avoided when possible because of difficulty of tracking with a few current meters tidal beams that may vary in space and time [*van Haren*, 2004a].

[8] This implies certain assumptions about generation and propagation of near-inertial internal waves in the open-ocean. 'Global' models of f-enhancement all account for the beta effect. Ambiguous is the direction of meridional propagation, as some models describe a broadband generation of internal waves that focus their energy at their

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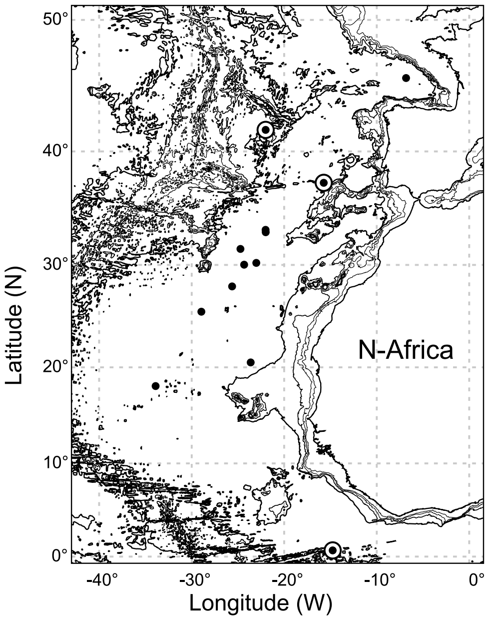

Figure 1. North-Atlantic Ocean with mooring positions (•), sometimes encircled for clarity.

critical latitude following poleward propagation [*Munk*, 1980; *Fu*, 1981]. [*Garrett*, 2001] discusses near-surface generation and subsequent equatorward propagation, yielding a blue-shift of the spectral f-peak. Local modifications in peak frequency and peak height may occur, with trapping of f-energy in regions of sub-f negative vorticity [*Mooers*, 1975; *Xing and Davies*, 2002].

[9] The critical latitude may be just poleward of the latitude of f when inertio-gravity waves on a spherical shell are considered so that a red-shift of the f-peak can be observed [*Maas*, 2001]. *Gerkema and Shrira* [2005] discuss trapping of poleward propagating sub-f energy beyond the f-latitude in layers of small N (>~f). This poleward trapping is predicted including the horizontal component of the Earth's rotational vector $2\Omega\cos\varphi$ that yields inertio-gravity wave band limits $\sigma_{min} < \sigma < \sigma_{max}$, with $\sigma_{min} <$ f and N < σ_{max}. These limits substantially extend the traditional internal gravity wave limits when N is small: e.g. around $|\varphi| = 30°$, N = 2.5f yields 20% smaller, larger limits than f, N, respectively.

2. Data

[10] The OSU deep-water archive is the main supplier of current meter data investigated here, supplemented with data from the Canary Basin [*Siedler and Paul*, 1991; *van Haren*, 2004b] and the Bay of Biscay. Focus is on data roughly along 20°W between $0° < \varphi < 45°$N (Figure 1 and Table 1). The data were searched following the criteria: sampling rate ≥once/2 hours, water depth ≥4000 m, away from large topography, dense coverage between $20° < |\varphi| < 35°$, more or less along a meridian, data from mid-depth. The latter criterion was difficult to hold, as too few moorings resolved the water column well. Also, the initial criterion of ≥1 year of data could not be maintained, and had to be changed to ≥7 months.

[11] The data are sub-sampled to once/2 hours and arbitrary sub-sets are taken equal to the length of the shortest record. The data are not sub-divided for year and season of observations. Also, data are not selected for current meter type, most of them were Aanderaa RCM5, 7 or 8. Except for amplitude correction of a single data record (Table 1), no corrections have been applied.

[12] Instead of scaling internal wave kinetic energy by N, which may be applicable for tidal energy but less clearly for f-energy [*van Haren et al.*, 2002], data are grouped in three depth ranges across which N varied by less than a factor of two (Table 1). Within each group the high-frequency non-tidal harmonic internal wave band energies were expected to fall within the range of 95% statistical significance, which implied a relative "error" of ~30%.

[13] Energy densities are determined for two frequency bands by summing the spectral contents of $0.97f < \sigma < 1.06f$ and $0.955M_2 < \sigma < 1.055M_2$, using the mean relative bandwidth $\Delta\sigma/\sigma = 0.09$ that captured most tidal and inertial motions for Bay of Biscay and Canary Basin data [*van Haren*, 2004a]. This summing of spectral contents yielded energy estimates to within a relative error of ~12%. It is

Table 1. Moored Current Meter Data Along ~20°W (North-Atlantic Ocean)[a]

Position N	Position W	Length,[b] yr, days, yr	H, m	'surface' z, m, $N/f_D = 49 \pm 6$	'1000–2000 m' z, m, $N/f_D = 26 \pm 10$	'≥3000 m' z, m, $N/f_D = 7 \pm 4$
45°48.0′	06°50′	330,1995/6	4800	–	–	3800
41°44.9′	21°57′	370,1980/1	3840	600	1500	3000
37°21.6′	15°43.3′	330,1982/3	4950	–	–	4300
33°13.0′	22°00.3′	225,1980/1	5300	–	–	4600
33°00.1′	21°59.8′	240,1992/3	5274	200	–	–
31°28.8′	24°43.8′	216,1985/6	5444	146[c]	1016	4368
30°10.9′	23°01′	540,2003/4	5158	–	1525	3070
30°01.4′	24°20′	240,1987/8	5300	141	1000	5175
27°58.4′	25°38′	220,1987/8	5000	261	1130	5190
25°31.9′	28°57.2′	240,1992/3	5700	110	1500	3500
20°29.6′	23°37′	370,1986/7	4540	405	1255	4505
18°05.6′	33°54.0′	220,1992/3	5300	200	–	–
00°39.8′	14°45.8′	710,1992/4	4325	–	2000	3200

[a]H denotes water depth, z instrument depths that are grouped in three depth categories. In the deepest category data are used from the uppermost current meter. Stratification is normalized by $f_D = f(30°)$.
[b]Of each mooring the shortest length is indicated.
[c]'Anomalous' record: amplitude was smaller by a factor of 2.5 compared to records at nearby latitude, depth.

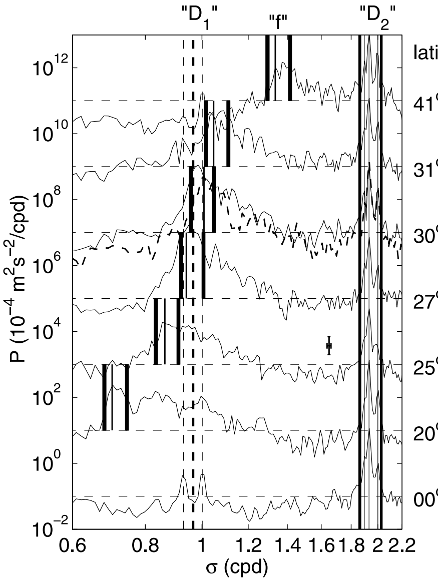

Figure 2. Kinetic energy spectra from 8 current meter records between 1000–2000 m (details in Table 1). Spectra are offset by two decades, except those from the two moorings near 30°N (30°11′N is dashed). The vertical dashed lines indicate diurnal (D_1) frequencies from left-right: O_1, M_1 (heavy dashed) and K_1. The heavy solid borders marking D_2 are at $0.965M_2$ and $1.055M_2$. The height of the bars indicating f-bands is always 2 decades for easy referencing peak heights, whilst the bandwidth is always 0.09f. The positioning in the frequency domain depends on the peak frequency. One fundamental bandwidth is indicated with the 95% statistical significance level.

repeated twice for the tidal band, with and without the spectral peaks at M_2, S_2 and N_2 removed. Without these peaks, the D_2-band is flat and comprises only 'incoherent' internal tidal motions, whilst time series lack a spring-neap cycle.

3. Observations

[14] The spectral detail of the near-inertial/tidal band (Figure 2) shows a large change in near-inertial peak for latitudes $28° < \varphi < 30°$N: its height was about 10 times larger than at low- and mid-latitudes, whilst nearly equal or slightly larger than the M_2-peak, and its central frequency was significantly *red*-shifted with respect to local f. This was in contrast with more common blue-shift, which was also observed here at other latitudes. Unsmoothed spectra did not reveal whether the local enhancement of f-energy was due to enhancement at M_1 and N_1, at which the energy was neither peaking nor throughing with respect to that at surrounding frequencies (not shown). Likewise, D_1-constituents O_1, K_1 could also not unequivocally be distinguished at these 'critical' latitudes. All seemed part of a flat f-band, which is suggested to have a finite width due to interactions with low-frequency variations in the environment [*van Haren*, 2004a].

[15] D_1 basically shows the same variation in energy levels as the f-band. D_1 shows not very energetic peaks (<0.1 times the energy at f(30°)) at both low ($\varphi \sim 0°$: O_1, K_1) and mid-latitudes ($\varphi \sim 42°$: K_1 mainly). At latitudes where D_1 can be separated from the f-band, at $\varphi \sim 20°$, weakly at 31.5°, pronounced near 25°, a moderate enhancement of kinetic energy beyond these tidal harmonic peak levels is found in a broad band $0.99O_1 < \sigma < 1.01\ K_1$, with $\Delta\sigma = 0.09\ K_1$. This enhancement by a factor of ~2 in D_1 partially coincides with a latitudinal change in D_2-energy.

[16] D_2-energy is observed nearly halved at latitudes just south of the D_1-critical latitudes (Figure 3). Focusing on intermediate depths (1000–2000 m) where near-surface and near-bottom effects are negligible, total D_2-energy shows a significant depression at φ = 25°, 28°N, whilst being remarkably uniform otherwise. This depression is not exactly of the same relative amount for total and incoherent energies, but the ratios of incoherent/total D_2-energy are not significantly different between the latitudes (not shown). At and just poleward of 28°N one observes the large increase in f-energy, which drops to 'background' levels north of 31.5°N.

[17] The above observations are also found at near-surface (<1000 m) and deeper (>3000 m) levels, although somewhat less pronounced than at intermediate depths. Despite the slightly larger 'noisy' variations, possibly as a result from unknown near-bottom and near-surface processes, the D_2-depression and the f/D_1-peak are visible in all plots of Figure 3. Similar analyses were also performed for Pacific Ocean data, but the set was thus limited equatorward of $|\varphi| = 25°$ that only a suggestion for a f-peak and for D_2-energy loss were observed, with no evidence for latitudinal dependence of peaks.

4. Discussion

[18] The observation of a D_2-depression just south of the D_1-critical latitudes alone cannot be unequivocally attributed to PSI, because total and incoherent D_2-energy dropped at nearly the same rate. Full proof would be the observation of unchanged barotropic tidal energy with

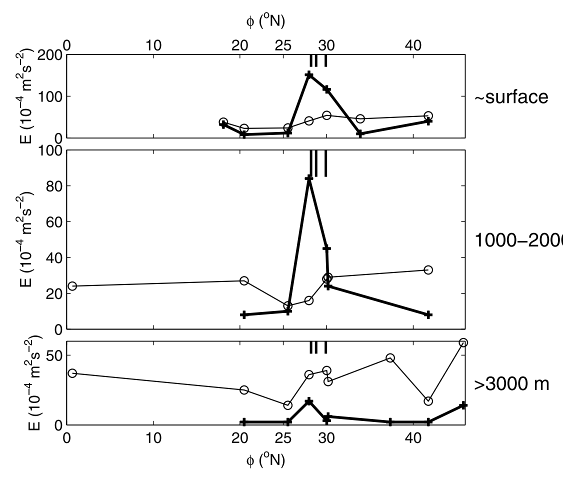

Figure 3. Summed energies densities $E = \Sigma P$/band observed at three different depth levels for the inertial band (+ heavy solid line) and total D_2 (o). The relative error is always 12% (see text). The short solid lines indicate critical latitudes for (left-right) N_1, M_1 and S_1.

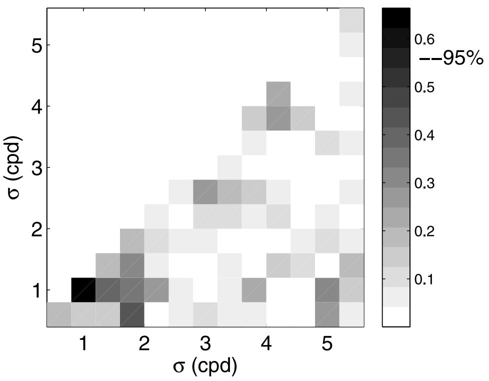

Figure 4. Bicoherence for first 15 days of total N-S current observed at 27°58.4′N (1130 m). Using 2.5 days half-overlapping and Hanning-tapered sections, the single statistically significant point was found at [1, 1] ± 0.16 cpd, indicating a coupled wave triad of D_1, D_1 with D_2.

latitudinal variations only in baroclinic energy. However, such proof can only be obtained after comparison of data from the internal tide source with ocean interior data. Alternatively, proof of PSI may come from independent estimates of mixing and from bicoherence: a statistical analysis to distinguish the extent of quadratic phase coupling in a (deterministic) signal from independently generated waves [e.g., *Kim and Powers*, 1979].

[19] As for mixing, a very limited transect, along 28°W between $28° < \varphi < 31°$N, of upper ocean (<1500 m) XBT and XCP yielded no significant latitudinal variations in eddy diffusivity K (Nagasawa, in the work by *van Haren* [2004b]). However, K estimated from lowered ADCP/CTD data, along 23°W between $26° < \varphi < 31°$N showed significant variations in the upper (<1000 m) and, especially, the deep (<4000 m) layers, with largest values at $28° < \varphi < 29°$N (up to $\log K = -3.5 \pm 0.5$) and $29° < \varphi < 31°$N (up to $\log K = -1.5 \pm 0.5$), respectively (C. Veth, personal communication, 2005). No significant variations were observed between 1500–3000 m. It is noted that these profiles were just single snap-shots and more profiling is requested.

[20] Bicoherence analysis results mid-depth are ambiguous: although significant non-linear coupling of D_1 and D_2-waves was indeed only found at $28° < \varphi < 30°$N (Figure 4), evidence of tidal PSI, it was only found for five 2-weeks periods, of which four for meridian currents, and never for entire records. These limited significant results resemble those of [*Carter and Gregg*, 2005]. As bicoherence cannot distinguish waves that propagate separately from their coupling area this requires further investigation.

[21] Perhaps the most clear evidence of possible PSI influence in the present observations are the f-(D_1-) energy variations with latitude, including the rare red-shift of peak frequency. However, one could question why PSI does not stop poleward of 29.91°, as we did observe some enhancement at 31.5°. In traditional theories, a reason for spread of f-energy across the critical latitude is trapping by negative sub-f vorticity. However, the possible influence of negative sub-f vorticity by meddies, which are found in the Canary Basin, extends only between ~700–1400 m, not in deep layers. Likewise, eddies shed by islands [*Sangrà et al.*, 2005] are mainly surface layer (<1000 m) phenomena, whilst their main shedding area is to the south, so that eddies shed from the Canaries do not reach the center of the Canary Basin poleward of 26°.

[22] In contrast, considering non-traditional theories for which $\sigma_{min} < f$, free propagating f-inertio-gravity waves generated at $|\varphi| = 20°(26°)$ will hit their $|\varphi_c| = 28.8°$, when the background stratification amounts $N \approx 2f(4f)$. Likewise, trapping of D_1-(PSI-induced)-f energy may be found poleward of the D_1/f-latitude, e.g. generation at 28.8° becomes trapped at 31.5°(30°) for $N \approx 4f(6f)$. Such low values are observed in smoothed ($\Delta z > 100$ m) N in the Canary Basin, but only at great depths (>4800 m). However, throughout the water column many small-scale layers (typically 10–25 m vertical extent) show such low N-values and these layers may significantly alter the fate of f-motions, as was suggested from observations in the Mediterranean Sea [*van Haren and Millot*, 2004]. Hypothesizing, non-traditional theory and thin homogeneous layers aid in separating PSI-induced enhanced f/D_1-energy from latitudes of reduced D_2-energy.

[23] In the present study D_2-energy at 25–28° is observed lower by about 50% compared to its environment, whilst f/D_1-energy at 28–30° is more than doubled, at all depth levels. This suggests that in addition to PSI other mechanisms may be at work, for example trapping of D_1-energy coming from a larger area than the area of observed loss of D_2-energy, and a limited loss of D_1-energy due to mixing. This requires further study and more data.

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References

Carter, G. S., and M. C. Gregg (2005), Persistent near-diurnal internal waves observed above a site of M_2 barotropic-to-baroclinic conversion, *J. Phys. Oceanogr.*, in press.

Fu, L.-L. (1981), Observations and models of inertial waves in the deep ocean, *Rev. Geophys. Space Phys.*, *19*, 141–170.

Garrett, C. (2001), What is the "Near-Inertial" band and why is it different from the rest of the internal wave spectrum?, *J. Phys. Oceanogr.*, *31*, 962–971.

Gerkema, T., and V. I. Shrira (2005), Near-inertial waves on the "nontraditional" β plane, *J. Geophys. Res.*, *110*, C01003, doi:10.1029/2004JC002519.

Hibiya, T., and M. Nagasawa (2004), Latitudinal dependence of diapycnal diffusivity in the thermocline estimated using a finescale parametrization, *Geophys. Res. Lett.*, *31*, L01301, doi:10.1029/2003GL017998.

Hibiya, T., M. Nagasawa, and Y. Niwa (2002), Nonlinear energy transfer within the oceanic internal wave spectrum at mid and high latitudes, *J. Geophys. Res.*, *107*(C11), 3207, doi:10.1029/2001JC001210.

Kim, Y. C., and E. J. Powers (1979), Digital bispectral analysis and its applications to nonlinear wave interactions, *IEEE Trans. Plasma Sci.*, *PS-7*, 120–131.

Lamb, K. (2004), Nonlinear interaction among internal wave beams generated by tidal flow over supercritical topography, *Geophys. Res. Lett.*, *31*, L09313, doi:10.1029/2003GL019393.

Leaman, W. H., and T. B. Sanford (1975), Vertical propagation of inertial waves: A vector spectral analysis of velocity profiles, *J. Geophys. Res.*, *80*, 1975–1978.

Maas, L. R. M. (2001), Wave focusing and ensuing mean flow due to symmetry breaking in rotating fluids, *J. Fluid Mech.*, *437*, 13–28.

MacKinnon, J. A., and K. B. Winters (2005), Subtropical catastrophe: Significant loss of low-mode tidal energy at 28.9°, *Geophys. Res. Lett.*, *32*, L15605, doi:10.1029/2005GL023376.

McComas, C. H., and F. P. Bretherton (1977), Resonant interaction of oceanic internal waves, *J. Geophys. Res.*, *82*, 1397–1412.

Mooers, C. N. K. (1975), Several effects of a baroclinic current on the cross-stream propagation of inertial-internal waves, *Geophys. Fluid Dyn.*, *6*, 245–275.

Munk, W. H. (1980), Internal wave spectra at the buoyant and inertial frequencies, *J. Phys. Oceanogr.*, *10*, 1718–1728.

Sangrà, P., et al. (2005), Life history of an anticyclonic eddy, *J. Geophys. Res.*, *110*, C03021, doi:10.1029/2004JC002526.

Siedler, G., and U. Paul (1991), Barotropic and baroclinic tidal currents in the eastern basins of the North Atlantic, *J. Geophys. Res.*, *96*, 22,259–22,271.

van Haren, H. (2004a), Band-width similarity at inertial and tidal frequencies in kinetic energy spectra from the Bay of Biscay, *Deep Sea Res., Part I*, *51*, 637–652.

van Haren, H. (2004b), LOCO-IW04-Canary Basin: R.V. Pelagia cruise 64PE231, 14 October–08 November 2004, cruise report, 39 pp., R. Neth. Inst. for Sea Res., Den Burg, The Netherlands.

van Haren, H., and C. Millot (2004), Rectilinear and circular inertial motions in the western Mediterranean Sea, *Deep Sea Res., Part I*, *51*, 1441–1455.

van Haren, H., L. Maas, and H. van Aken (2002), On the nature of internal wave spectra near a continental slope, *Geophys. Res. Lett.*, *29*(12), 1615, doi:10.1029/2001GL014341.

Xing, J., and A. M. Davies (2002), Processes influencing the nonlinear interaction between inertial oscillations, near inertial internal waves and internal tides, *Geophys. Res. Lett.*, *29*(5), 1067, doi:10.1029/2001GL014199.

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