Dissipation parameterizations in spectral wave models and general suggestions for improving on today's wave models

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ABSTRACT

Since the WAM-Cycle 3 'consensus', a wide variety of parameterizations have been proposed for spectral wave models, in particular for the dissipation source functions. With some of these parameterizations and using the same wind fields, at the global scale, random errors have been reduced by at least 30% for the significant wave height, which is the most accurately measured wave parameter. Some of this improvement is related to a better understanding of wave evolution, that can be quantitative in the case of swell dissipation or semi-quantitative, in the case of threshold effects on breaking-induced dissipation for which the dissipation rates are still poorly known. A wide range of observations provide further semi-quantitative or qualitative constraints on the magnitude and variability of wave growth and dissipation rates. These include spectral tail levels measured by buoys, estimates of mean square slopes from active or passive microwave instruments, breaking statistics from video, and acoustic noise variability. All these measurements – and others – have something to tell us about the realism of our parameterizations, but they all require, to a varying degree, a better understanding of what is measured. With these caveats, it is shown that the energy level at frequencies between 0.2 and 1 Hz can be better represented by a saturation-based dissipation, than by the Komen-type parameterization based on a spectrally-integrated steepness.

1 Introduction

My intention in participating in this workshop is to find out how we can make numerical wave models more accurate and useful for as wide a variety of uses as possible. Some of these future developments require some understanding of where we are today, where and why different models disagree and what data or analysis of existing data is needed to decide on specific issues what the the best parameterization option. This present document thus contains explicit recommendation to operational forecasting centers, be it ECMWF, NCEP or Meteo-France, and may contain some interesting topics for academic researchers that may want to apply their science to the noble goal of improving numerical wave forecasts. In doing so, I will try to bring as much evidence as necessary, and recognize when the evidence is thin or missing.

A reasonably rich literature has shown that the accuracy of numerical wave models, be it for marine safety or geophysical research purposes, is controlled by

- the accuracy of the forcing fields, essentially the wind speed and direction (Cavaleri and Bertotti, 2006; Janssen, 2008), but also the sea ice (Tolman, 2003) and icebergs (Ardhuin et al., 2011), and water levels and currents (Ris et al., 1999; Ardhuin et al., 2012b).
- the quality of the parameterizations of generation, non-linear evolution and dissipative processes
- the accuracy of the numerical schemes used to integrate the different pieces of the wave energy balance, be it the source term integration (Hargreaves and Annan, 2000; Tolman, 2002b) or the

spatial propagation (Tolman, 2002a; Roland, 2008; Ardhuin and Roland, 2012), or the interaction of these different components.

These three categories were listed here in order of decreasing importance, an importance that is very subjective and strongly varies with the scale considered and the application context, and the actual model chosen. To give a few counter-examples to the hierarchy given above, both numerical propagation issues and dissipation are very important for global swell fields(Wingeart, 2001; Tolman, 2002a; Ardhuin et al., 2009). Coastal reflection is generally negligible for accurate estimates of wave heights, but it is very important for reproducing seismic noise records (Ardhuin et al., 2012a; Ardhuin and Roland, 2012). Some wave models, due to parameterizations or numerical choices are more prone to errors in some conditions. For example SWAN artificially broadens spectra in complex coastal areas due to 'renormalization' (a redistribution of negative action values in directional space), which may be a problem for some applications. In the same context, an accurate WAVEWATCH III application (i.e. one in which refraction limiters do not get activated and splitting errors is small) may require a time step too small to be computationally feasible.

Here we shall focus on dissipative processes in deep water. This is thus only a piece of this more general puzzle. The present paper summarizes findings already published or under review. More details can be found in Ardhuin et al. (2010b), Filipot and Ardhuin (2012).

It is interesting to note that all dissipation parameterizations discussed here include a quasi-linear term with a coefficient that multiplies the frequency-directional power spectrum of the surface elevation $F(f,\theta)$. This coefficient is proportional to a wave steepness ε to the fourth power or a higher power in the case of Alves and Banner (2003). However, this steepness is parameterized very differently.

In Komen et al. (1984), it is defined from the full wave spectrum

$$\varepsilon^{\mathrm{KHH}} = k_r H_s, \tag{1}$$

giving a dissipation source term

$$S_{\mathrm{oc}}^{\mathrm{KHH}}(f,\theta) = C_{\mathrm{ds}}\sqrt{g k_r}\,(k_r H_s)^4 \left[(1-a)\frac{k}{k_r} + a\frac{k^2}{k_r^2}\right] F(f,\theta), \tag{2}$$

where H_s is the significant wave height, and k_r is a representative mean wavenumber defined by

$$k_r = \left[\frac{16}{H_s^2}\int_0^{f_{\max}}\int_0^{2\pi} k^r E(f,\theta)\,\mathrm{d}f\,\mathrm{d}\theta\right]^{1/r}, \tag{3}$$

with $r = -0.5$ and $a = 0$ used by the WAMDI Group (1988), while Bidlot et al. (2005) used $r = 0.5$ and and $a = 0.6$.

Phillips (1984) introduced a steepness that is local in frequency. This local steepness $\varepsilon^P(f)$ is proportional to $\sqrt{B(f)}$, where the non-dimensional energy level $B(f)$ at that frequency (also called saturation) is defined by

$$B(f) = \int_0^{2\pi} k^3 F(f,\theta') C_g/(2\pi)\,\mathrm{d}\theta'. \tag{4}$$

Such a local steepness only makes sense for a smoothly varying spectrum (Phillips, 1984, page 1428, column 2). Indeed for monochromatic waves of very small amplitudes $B(f)$ can be very large but is not associated to steep waves.

Several parameterization inspired by Phillips (1984) have been proposed, and they mostly differ in the choice of the threshold B_r. In Alves and Banner (2003) S_{oc} is proportional to $(B/B_r)^4$, so that it increases steeply as B becomes larger than the threshold B_r, but it starts dissipating for $B < B_r$.

In the dissipation source functions of Ardhuin et al. (2010b) and Babanin et al. (2010), B_r acts more like a switch and $S_{\rm oc}(f,\theta)$ is not such a high power of B,

$$S_{\rm oc}(f,\theta) \quad = \sigma \frac{C_{\rm ds}^{\rm sat}}{B_r^2} \left[\max \left\{ B(f) - B_r \right\}^2 \right] F(f,\theta) \tag{5}$$

where $C_{\rm ds}$ is a non-dimensional constant, B_r is a threshold for the saturation and $F(f,\theta)$ is the spectral density of wave energy. The minor differences between Babanin et al. (2010) and Ardhuin et al. (2010b) include a different effect of wave directional distribution in the exact definition of B, and a different formulation of the cumulative effect. In Babanin et al. (2010) this cumulative effect may dominate at lower frequencies than it does in Ardhuin et al. (2010b). We also note that Ardhuin et al. (2010b) is mostly derived from Banner and Morison (2006, 2010), except for the smoothing of B over frequencies. Finally, in Ardhuin et al. (2010b) B is also a function of the wave direction, leading to a maximum dissipation in the mean wave direction, whereas Babanin et al. (2010) used a prescribed directional distribution of the dissipation which has a local minimum in the mean wave direction.

In a recent study of current effects, Ardhuin et al. (2012b) have shown that the two types of parameterizations, following Komen et al. (1984) or Phillips (1984), can be adjusted to give very similar behaviour, except in in strong opposite currents, where the change in steepness is much less severe when it is defined from the full spectrum, which gives much less dissipation compared to parameterizations with a locally defined steepness.

Eventually, the most compelling evidence a rather local dissipation is the evolution of swells over large distances and wind seas in the presence of swell. Recent measurements by Hwang (2008) and García-Nava et al. (2012) have shown a reduction of wind stress and wind sea energy in the presence of swell that is not yet accounted for in saturation-based parameterization, and that is contrary to the results of Komen-type dissipation terms, that typically give an enhanced wind sea growth in the presence of swell (Ardhuin et al., 2007, 2010b). This erroneous behaviour of the parameterization is probably the source of larger errors for mean periods $T_{m0,-2}$ when using Komen-type parameterizations.

This and other patterns of the wave model results will now be illustrated based on global wave hindcasts.

2 Global hindcasts

The results presented here were obtained with the WAVEWATCH III$^{(R)}$ modelling framework, implemented on a global grid with a 0.5 degree resolution and no data assimilation. The model is forced by 6-hourly ECMWF operational analyses. The model uses 32 frequencies from 0.037 to 0.72 Hz, and 24 directions. The non-linear interactions are parameterized using the Discrete Interaction Approximation (DIA) by Hasselmann et al. (1985). We have tested four different wind input and dissipation combinations, given by (Tolman and Chalikov, 1996) - TC, Janssen et al. (1994) - WAM4, Bidlot et al. (2005) - BJA, Ardhuin et al. (2010b)-TEST441, and a recent revision of that parameterization with a smoothing of the swell dissipation threshold - TEST451. The global distributions of errors are illustrated in figure 1 for the year 2007.

2.1 Wave heights

We first looked at the model results in terms of significant wave heights H_s. When averaged over the globe the normalized r.m.s. error is reduced from 15% with TC, to 14.2% with WAM4, 13.4% with BJA, 11.3% with TEST441 and 10.9% with TEST451. In these latter parameterizations, the single most sensitive parameter in the model is the swell dissipation term. Given the very poor knowledge of swell dissipation processes, possibly due to air-sea friction (Ardhuin et al., 2009) while it has been suggested

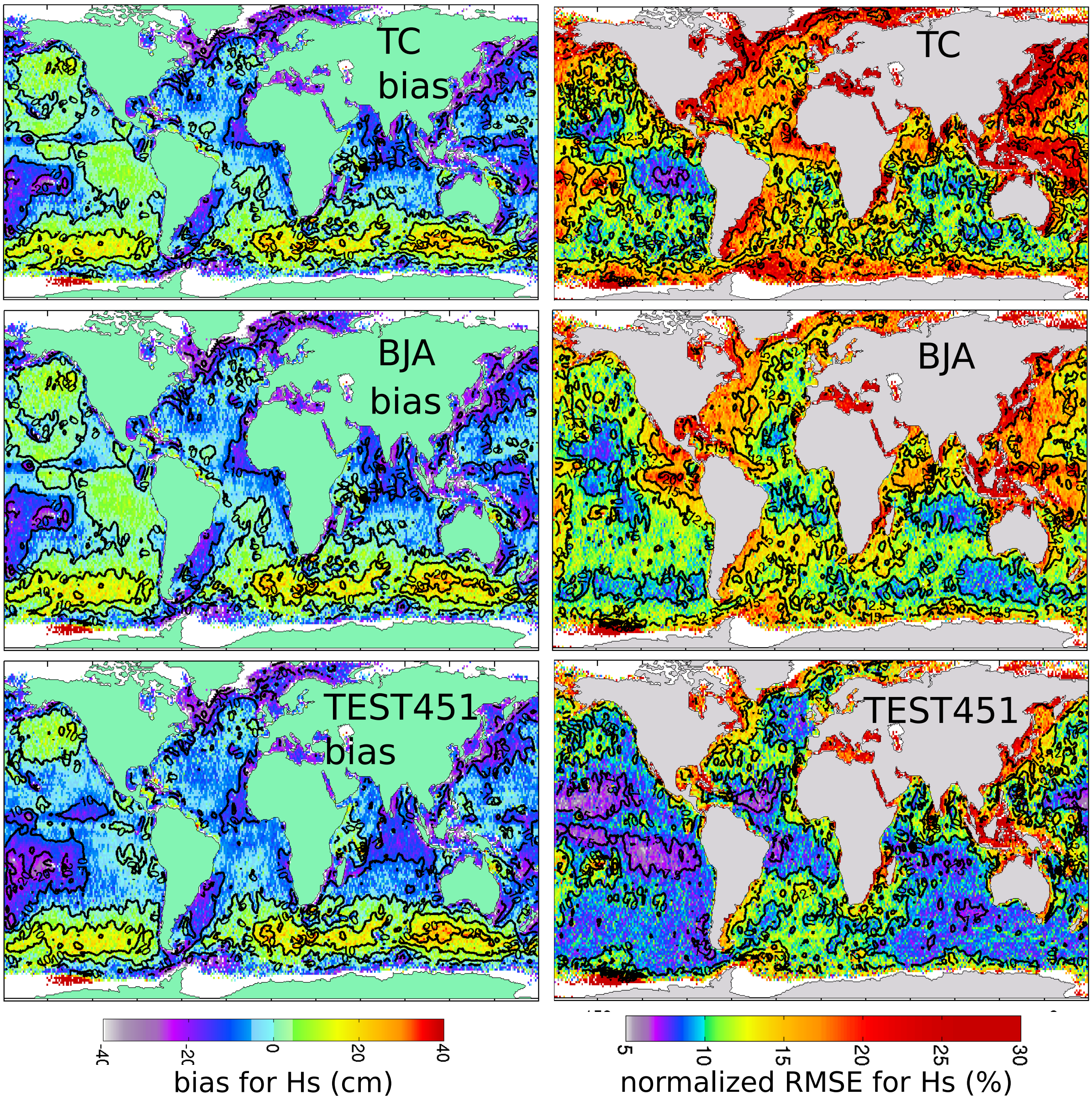

Figure 1: Bias and normalized RMSE error for wave heights for the year 2007.

that it could be associated to wave-turbulence interactions in the water (Ardhuin and Jenkins, 2006; Kantha et al., 2009), there is certainly more room for progress on significant wave heights in the future.

2.2 High frequency tail and mean square slopes

The lowest order statistics of ocean surface slopes are defined by the mean square slopes, or mss in short, and are important for a wide range of applications. First of all, radar backscatter from the sea surface at near vertical angles (nadir) is nearly inversely proportional to the mss, which offers a global dataset for validation using satellite altimeter data.

Microwave radiometric data that measure the emissivity of the sea surface are also strongly related to the mss, at least for low winds when foam is not present on the sea surface (e.g. Martin, 2004). At a fixed wind speed, for both X-band data and L-band data, from AMSR-E and SMOS, there is a very clear dependence of the emissivity with wave height, that is very similar to what is inferred from altimeter cross sections. Here we show data from a wave buoy, for which the spectrum was integrated only up to 0.4 Hz, showing again the same dependence.

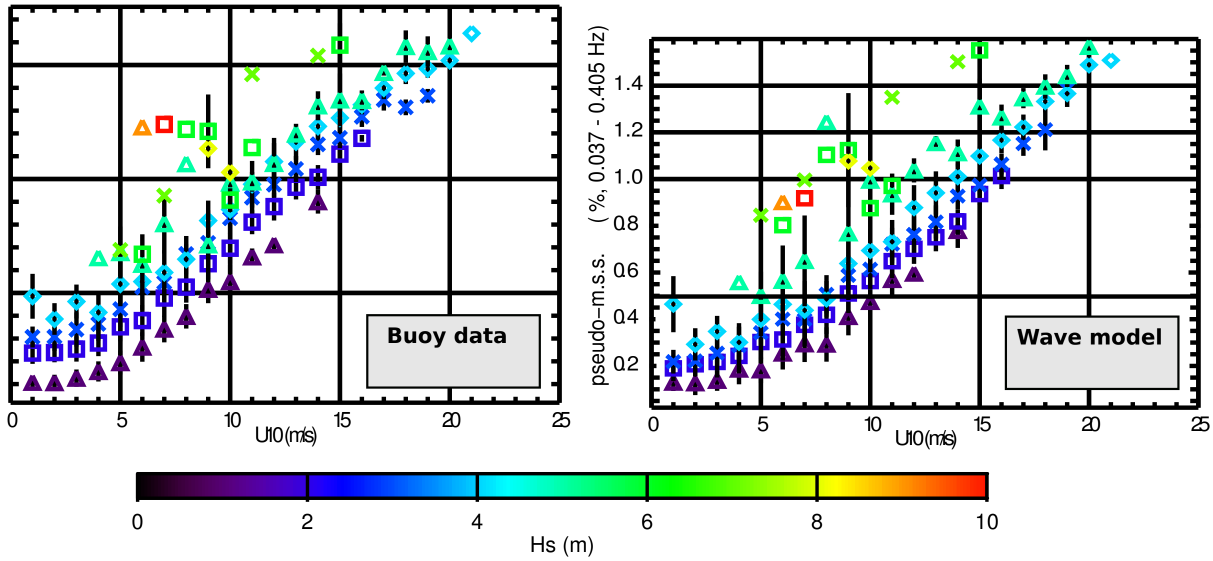

Figure 2: Example of mean square slopes estimates from buoy data and model output, here using TEST441, and its variability with wind speed and H_s.

Obviously, not all wave buoys do not measure slopes. Here the slope is estimated from the heave spectrum, and obtained by converting the measured frequency spectra into wavenumber spectra, assuming a linear dispersion relationship. That linear transformation breaks down for relatively high frequencies (e.g. Janssen, 2009) and also the buoy response is not linear. Still the resulting distribution of this pseudo-slope with wind speed and wave height is consistent with all the remote sensing data and supports a general increase of the mss with the wave development.

Because this integrated parameters sums the contributions from a wide range of scales, it is not simple to infer whether the increase in mss is due to the broadening of the spectrum towards lower frequencies or also possibly an increase in the spectral tail level. The spectra produced by the TEST441 and TEST451 parameterization have a tendency to fall off like $f^{-4.5}$ and, up to 0.4 Hz, are generally consistent with the spectral levels reported by Forristall (1981). These spectra are, however, probably too energetic at frequencies at frequencies above 0.5 Hz.

2.3 High frequency tail and directional parameters

Ardhuin and Roland (2012) have shown that the directional spreading of dominant waves is generally well predicted with the TEST441 parameterization with the necessity to include coastal reflections for buoys close to continents. A similar validation for shorter waves is made difficult by the poorer quality of buoy data in that range, but the data suggests that the directional spectra at 0.4 to 0.5 Hz are systematically too narrow. This means that the input and dissipation parameterizations must have an important spurious narrowing effect, because it is well know that the DIA tends already to make a spectrum too broad in directions.

Another observation of interest is the acoustic noise associated to second order wave-wave interactions that produces either bound acoustic-gravity modes, or that resonantly forces seismic Rayleigh waves. With the wave directional distribution M such that the frequency-directional ocean wave spectrum $E(f,\theta)$ is expressed as $E(f,\theta) = E(f)M(f,\theta)$, we define the directional integral $I(f)$ that represents the net effect of all waves traveling in opposite directions,

$$I(f) = \int_0^\pi M(f,\theta)M(f,\theta+\pi)\mathrm{d}\theta,. \tag{6}$$

In the case of measurements near the surface, the measured noise level is dominated by acoustic-gravity modes for which the theory is well established, at least up to frequencies of 0.2 Hz with a well known relation between the noise level and $I(f)$ Herbers and Guza (1994); Ardhuin and Herbers (2012). In the case of deeper water measurements, they are generally dominated by seismic Rayleigh waves, and there is a poorly known linear propagation effect that makes the absolute level of the spectra difficult to interpret. However, the noise spectrum at a frequency $2f$, should be proportional to the directional integral $I(f)$. It is thus possible to use the variability of the recorded noise level at frequencies of the order of 1 Hz, to diagnose possible problems in the wave model at frequencies around 0.5 Hz. Here we use data from the ALOHA cabled observatory, that was analyzed in detail by Duennebier et al. (2012). These acoustic measurements, from 4700 m depth, show a remarkable variability with the wind speed, that are probably attributable, at least up to an acoustic frequency $f_s = 1$ Hz, to the wave-wave interaction effect described by Hasselmann (1963) and thus proportional to the directional integral $I(f_s/2)$. We have applied the same numerical modelling technique previously used for seismic noise (Ardhuin et al., 2011) to reproduce these observations. Namely, the theoretical seismic source is integrated over the ocean to produce the noise at one location. However, for the frequencies considered here, this spatial integration gives the same result as the use of the local value of $I(f_s/2)E^2(f_s/2)$, and multiplication by a constant. The results in figure 3 show a very good reproduction of the variability of the noise level at $f_s = 0.3$ Hz, but a very poor result for $f_s = 1$ Hz.

Given the detailed comparison of the wave model against buoy data, with a typical 20% r.m.s. error and correlation $r = 0.88$ on $E(f)$ in the open ocean for $f = 0.4$ Hz (this value is for the buoy 51001, 180 nm to the north-west of Oahu) and 30% error in coastal waters for $f = 0.5$ Hz (this is for the Waimea buoy, off the north shore of Oahu) we expect that the relative changes in $I(f)$ can be obtained from the measured noise level and our modeled $E(f)$. This may not be true with any model parameterization. Using the BJA parameterization from the ECMWF wave model(Bidlot et al., 2005), the correlation between modeled and measured wave spectral level $E(f = 0.4$ Hz$)$ at the buoy 51001 is reduced to $r = 0.78$ with an r.m.s. error of 18%. Also our parameterization yields a 7% positive bias on this energy level, whereas the BJA parameterization has a 5% negative bias. Another parameterization, by Tolman and Chalikov(Tolman and Chalikov, 1996) (hereinafter TC) which was used in another seismic noise study(Kedar et al., 2008) gives a 8% positive bias but a r.m.s. error of only 16%.

The correlation between the locally modeled $E^2(f)I(f)$ and our noise model increases from $r = 0.54$ at $f_s = 0.13$ Hz, to $r = 0.86$ at $f_s = 0.15$ Hz, and reaches $r = 0.96$ for $f_s = 0.8$ Hz, which gives an idea of diminishing importance of the noise sources coming from regions beyond the spatial correlation distance of the modeled wave field. For f_s above 0.8 Hz, the measured noise level correlates better with the wind speed, as found in previous studies(Duennebier et al., 2012; Zhang et al., 2009), or even better with the wave spectral density $E(f)$.

This correlation is illustrated by figure 3. The degradation of the model results for $f_s > 0.6$ Hz can be interpreted as a poorly modeled directional spectrum, giving errors on the variability of $I(f)$. Although multiplication of $E^2(f)$ by $I(f)$ clearly improves the correlation for $f_s < 0.6$ Hz, this is not the case for the higher frequencies. In particular, most of the recorded data at 1 Hz ranges from -10 to 0 dB, while the model gives a much narrower range from 2 to 4 dB. If the theory is correct, this means a very strong variability of the product $I(f_s/2)E^2(f_s/2)$, that the model does not reproduce.

We have also tried the BJA and TC parameterizations(Bidlot et al., 2005; Tolman and Chalikov, 1996) which give larger errors for other wave parameters(Ardhuin et al., 2010b). For these, the correlation of $I(f)E^2(f)$ with the measured noise level becomes insignificant.

From the good correlation of the noise level with $E^{2.3}(f)$ for $f = 0.5$ Hz (i.e. $f_s = 1$ Hz) we would

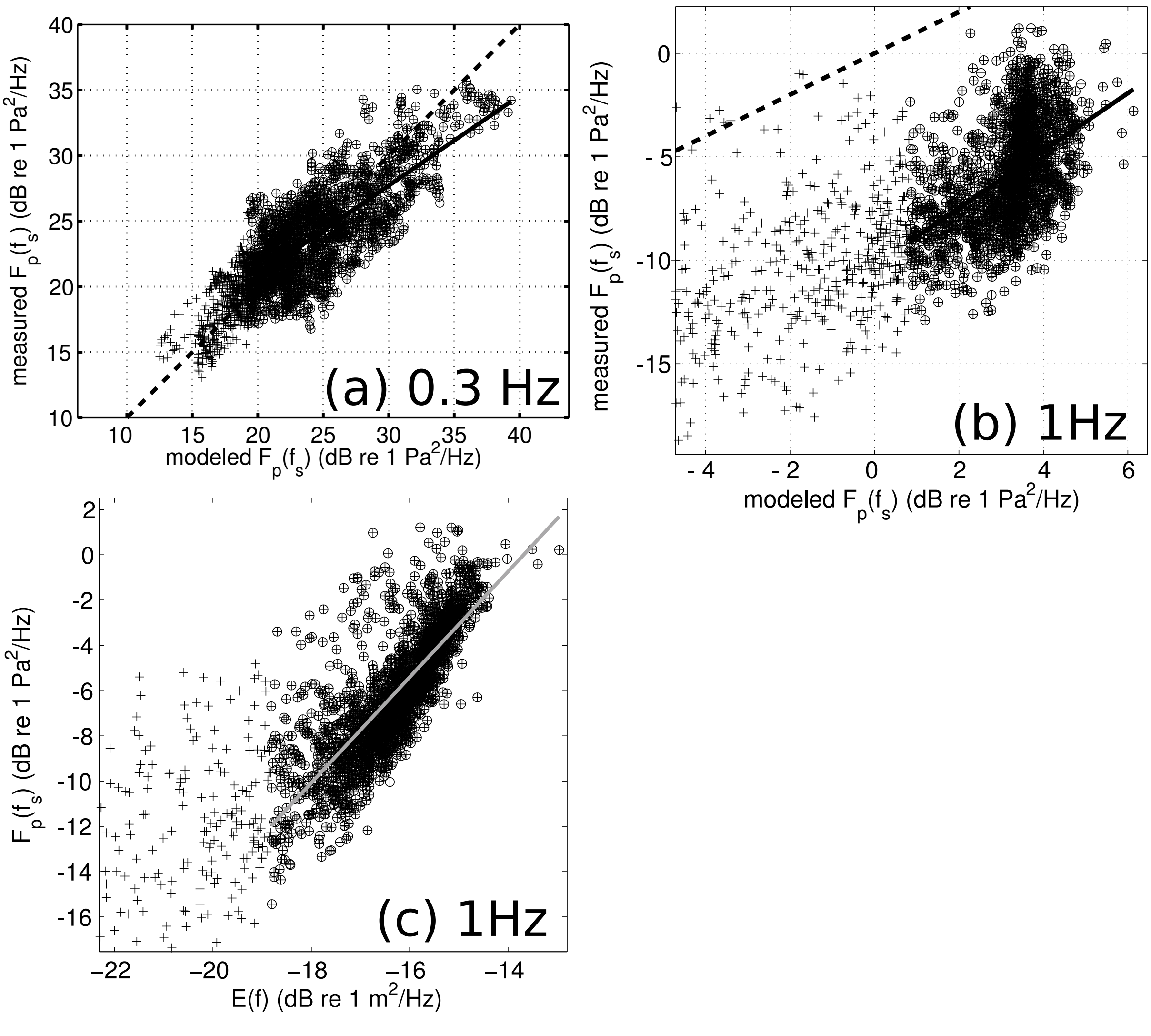

Figure 3: Modeled versus observed spectral densities at the ALOHA cabled observatory, located on the sea floor, 100 km north of Oahu. .(a) $f_s = 0.3$ Hz and (b) $f_s = 1$ Hz. At this higher frequency, the observed noise level correlates better with the wind speed or (c) with the modeled wave spectral density $E(f)$. Data points without circles correspond to the lowest 10% in the modeled values, which were excluded when computing the best fit exponents p between the noise level and the wave spectral density $E(f)$ which are $p = 0.38$ and $p = 2.3$ at 0.3 and 1 Hz, respectively.

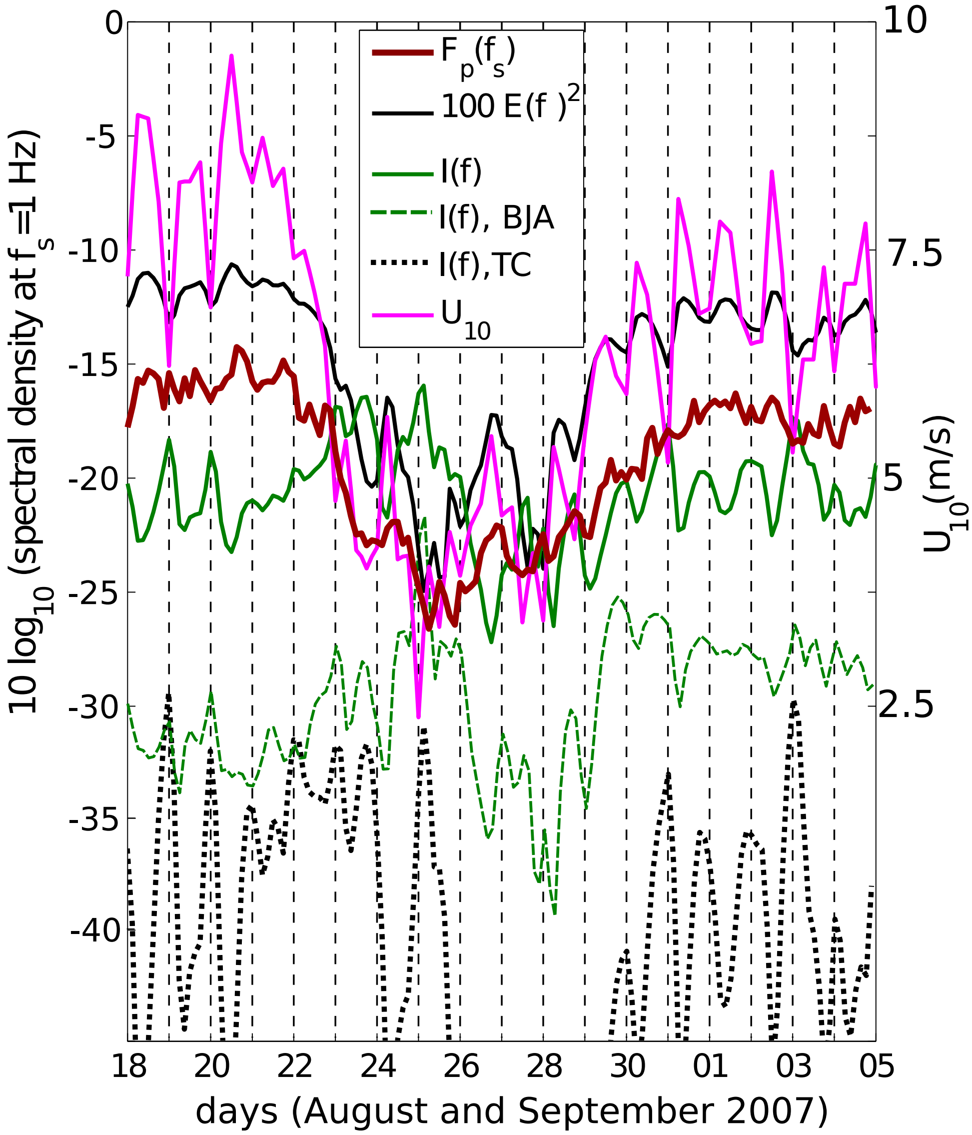

Figure 4: Example of noise level and wave parameters evolution in August 2007. $F_p(f_s = 1Hz)$ is the noise level measured at ACO, on the sea floor. The other lines are modelled wind or wave parameters for $f = 0.5$ Hz, expected to correspond to 1 Hz noise.

expect that $I(f)$ should vary like $E^{0.3}(f)$. On the contrary, all the parameterizations tested here give a decreasing $I(f)$ when $E(f)$ increases, as shown in figure 4. This inverse variation of $I(f)$ and $E(f)$ is correct around the peak of the wind sea (see also Ardhuin et al., 2012a).

The wave model parameterization are thus inappropriate to describe the variability of $I(f)$ for these high frequencies. Figure 4 also shows the 20 dB difference in the mean value of $I(f)$ for these frequencies between our parameterization and TC, which was already discussed by Ardhuin et al.(Ardhuin et al., 2011).

Although our parameterization was an attempt to fit the observed broad directional spectra (e.g. figure 2 Ardhuin et al., 2010b) it still underestimates the spectral width at wave frequencies f above 0.4 Hz. It is thus likely that some processes in the wind-wave spectral evolution are poorly represented or missing. In particular, wave breaking has been found to generate short gravity waves, an effect that is not taken into account in the wave models tested here, but which was found necessary to reproduce the azimuthal variability of radar observations(Kudryavtsev et al., 2005). This indirect effect of breaking may be an important process for the noise levels at acoustic frequencies above 0.6 Hz.

3 Conclusions

Recent tests of parameterizations in numerical wave models have shown that the errors at the scale of the global ocean are most sensitive to the parameterization of swell dissipation. That aspect can be improved by trial and errors, but it certainly requires more observations and theoretical work to guide further improvements. My experience is that nonlinear swell dissipation term generally perform better than linear ones, especially when a threshold is included to make almost zero the dissipation rate of low steepness swells.

Another aspect in which improvement is badly needed is the spectral shape of the inertial and high frequency ranges, in particular the directional distribution. Spectral moments such as the mean square slope clearly increase with wave development, but this still needs to be reconciled with observations of swell effects on the energy level at high frequency. Also, modelled wave spectra at frequencies higher than 0.4 Hz tend to be too narrow in their directional distribution. This is particularly true when the wind speed increases, as shown by underwater acoustic data. That variability of the wave spectrum will have to be further confirmed with more direct measurements, for example using stereo video imagery, which can now be used to map the evolution of meter-scales waves (e.g. Gallego et al., 2008; Ardhuin et al., 2010a). One should note that, in both modelled and real waves, possible modification in the tail level will also feedback on the wind-wave growth and wind stress. Testing these new parameterizations in a coupled wind-wave model will become more and more necessary to avoid errors and unwanted effects.

All these developments will require a better understanding of measurements, as well as new theoretical work and parameterization efforts. An important area that can lead to progress is the analysis and modelling of wave breaking statistics, assuming that, similarly to freak waves, they can be associated to the shape of the wave spectrum. Efforts in this direction by Banner and Morison (2010) and Filipot et al. (2010) show some clear potential.

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