Spectral Wave–Turbulence Decomposition

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ABSTRACT

A new method of wave–turbulence decomposition is introduced, for which the only instrument required is one high-frequency pointwise velocity sensor. This is a spectral method that assumes equilibrium turbulence and no wave–turbulence interaction. Nonetheless, laboratory and field experiments show that the new method produces results in good agreement with the results of established wave–turbulence decomposition methods. Therefore, this spectral method proves useful when neither a synchronized wave gauge, nor a second velocimeter, is available. Furthermore, this study indicates that uncertainty in velocimeter probe orientation is responsible for most of the wave bias occurring in turbulent velocity data, so that an accurate measurement of this orientation makes wave–turbulence decomposition unnecessary.

1. Introduction

Pointwise velocimeters allow us to determine benthic and water column shear stresses via the direct calculation of Reynolds stresses from fluctuating velocities. However, because variance associated with waves is often much larger than that associated with turbulence, some form of wave–turbulence decomposition must be used (Jiang and Street 1991; Thais and Magnaudet 1995; Trowbridge 1998).

In a flow with both waves and currents, the instantaneous horizontal velocity \( u \) can be written as

\[
    u = \bar{u} + \bar{u} + u',
\]

where \( \bar{u} \) is the horizontal component of mean velocity, \( \bar{u} \) is the wave-induced orbital velocity, and \( u' \) is the turbulent velocity. After Reynolds averaging of the mean momentum equation using (1), the Reynolds stress \( \tau \) becomes (see, e.g., Jiang and Street 1991)

\[
    -\frac{\tau}{\rho} = \bar{u}\bar{w} + \bar{u}'\bar{w}' + u'\bar{w}',
\]

where \( \rho \) is the density of water and \( w \) is vertical velocity. For irrotational, progressive waves (Dean and Dalrymple 1991), the first term on the rhs of (2) is zero. Furthermore, when waves and turbulence coexist, the latter is defined as motions that do not correlate with waves (Jiang and Street 1991; Thais and Magnaudet 1995), so the second and third terms on the rhs of (2) are also zero. Thus, under these conditions, the Reynolds stress is the same as that which is found for steady flows:

\[
    -\frac{\tau}{\rho} = u'\bar{w}'.
\]

However, as shown by Trowbridge (1998), small uncertainties in instrument orientation or a gently sloping bed can bias velocity measurements such that in practice \( \bar{u}\bar{w} \) may not be exactly zero.

Various methods of wave–turbulence decomposition can be used to remove this wave contamination from a turbulence dataset. One is that of Benilov and Filyushkin (1970), in which motions that correlate with displacement of the free surface are considered to be due to waves, and those that do not correlate are turbu-
lence. Apart from the high-frequency measurement of 
three components of velocity, this method also requires 
the simultaneous measurement of instantaneous free-
surface position, via a pressure- or capacitance-type 
wave gauge.

Another method of decomposition is that of Trow-
bridge (1998) and Shaw and Trowbridge (2001), which 
uses two velocity measurements spaced farther apart 
than the largest turbulence scale [approximately one-
quarter the water depth, according to Shaw and Trow-
bridge (2001)] but are well within one wavelength of 
each other. Motions that correlate between the sensors 
are waves, while motions that do not correlate are tur-
bulence. This method requires the use of two high-
frequency velocimeters, synchronized with each other.

We now introduce a new method, which we call the 
"Phase" method. Assuming equilibrium turbulence and 
no wave–turbulence interaction, in this method the 
phase lag between the \( u \) and \( w \) components of surface 
waves is used to interpolate the magnitude of turbu-
ulence under the wave peak within the inertial subrange 
of the spectral domain. This method is useful when 
neither a synchronized wave gauge nor a second syn-
chronized velocimeter exists.

2. Method

The method of wave–turbulence decomposition we 
introduce here is useful when wave and turbulence time 
scales overlap with each other in the spectral domain, 
but it assumes that waves and turbulence do not inter-
act (i.e., no stretching of turbulence at the wave time 
scales due to the strain field the waves create). As is the 
case with the Benilov and Filyushkin (1970) method, 
the goal of this method is to calculate wave stresses 
through the spectral sum

\[
\overline{\bar{u}\bar{w}} = \int_{-\omega_{\text{Nyquist}}}^{\omega_{\text{Nyquist}}} S_{\bar{u}\bar{w}}(\omega) \, d\omega,
\]

where \( S_{\bar{u}\bar{w}} \) here is the two-sided cross-spectral density 
(CSD) of wave-induced orbital velocities (Bendat and 
Piersol 2000) and \( \omega \) is frequency. Again, as is the case 
with Benilov and Filyushkin’s method, the turbulence 
spectrum can be expressed as a difference between the 
spectrum of raw velocities and that of wave-induced 
orbital velocities:

\[
S_{\bar{u}\bar{w}}(\omega) = S_{\bar{u}\bar{w}}(\omega) - S_{\bar{u}\bar{w}}(\omega).
\]

The same holds true for the integrals of these spectra:

\[
\overline{\bar{u}\bar{w}'} = \overline{\bar{u}\bar{w}} - \overline{\bar{u}\bar{w}}.
\]

For finite data series, the wave stress becomes

\[
\overline{\bar{u}\bar{w}} = \sum_{j=1}^{N/2} \bar{U}_j \bar{W}_j,
\]

where \( \bar{U}_j = U(\omega_j) \) is the Fourier transform of \( u(t) \) at 
frequency \( \omega_j \), \( N \) is the number of data points used in 
the Fourier transform, and \( t \) is time.

The question that remains is how to separate wave-
induced velocities from turbulence. Separation in the 
spectral domain is difficult, because the wave peak is 
often found at wavenumbers at which turbulence is also 
ergetic. Therefore, it is not valid to simply bandpass 
out the wave peak from the velocity series. Further-
more, interpolating the turbulence CSD under the 
wave peak is not practical, because the real part of the 
CSD has both positive and negative values. The au-
tospectral density (PSD) of \( u \) or \( w \) takes only real, posi-
tive values, so interpolating the turbulence spectrum 
below the wave peak of a PSD is an accomplishable 
task (see Fig. 1).

The remaining task is thus to express the wave 
stresses or CSD of (7) in terms of the PSDs \( S_{uu} \) and \( S_{ww} \). 
Since

\[
S_{uu} = \frac{1}{d\omega} |U_j|^2,
\]

we choose to find the CSD in terms of the magnitudes 
of the Fourier coefficients \( U_j \) and \( W_j \).

The first step is to find the phase of each Fourier

FIG. 1. PSD of horizontal velocity \( S_{uu} \) for a representative data 
series in the field. Pluses (+) represent data points within the 
wave peak, while dots (●) are data points outside the wave peak. 
The solid line is a least squares fit to the data outside the wave 
peak (as discussed in section 3c) and is used for estimating the 
magnitude of the wave-related Fourier coefficients \( (\bar{U}_j) \) under 
the wave peak, for use in (14). Data in this figure are from the ADV 
at 20 cmab during the June–July 2002 experiment.
coefficient. Since the Fourier coefficients can be written in phasor notation (Hecht 1987) as

$$U_j = |U_j|e^{i\angle U_j}$$ and $$W_j = |W_j|e^{i\angle W_j},$$

the phase of each takes the form

$$\angle U_j = \arctan \left[ \frac{\text{Im}(U_j)}{\text{Re}(U_j)} \right].$$

Via (9) and (10), each spectral component of the two-sided CSD takes the form

$$U_j^* W_j = |U_j||W_j|[\cos(\angle W_j - \angle U_j) + i \sin(\angle W_j - \angle U_j)].$$

Recalling the Euler relation leads to

$$U_j^* W_j = |U_j||W_j|\cos(\angle W_j - \angle U_j).$$

Summing over the two-sided spectral domain, the odd (imaginary) component gives no contribution, so

$$\overline{uw} = \sum_{j=\text{wave peak}} U_j^* W_j = \sum_{j=\text{wave peak}} |U_j||W_j|\cos(\angle W_j - \angle U_j).$$

In lieu of wave–turbulence interaction, this gives a reasonable estimate of the wave stress, where the magnitude of $\bar{U}$ is the difference between the raw $U$ and the turbulence $\bar{U}$ interpolated below the wave peak (via the straight line in Fig. 1). Since waves dominate the signal under the wave peak, this method assumes that the phase of (11) is dominated by waves as well, and thus (14) is a valid approximation of the wave-induced stresses. After finding the wave stress by summing (14) over the width of the wave peak, turbulent stresses (throughout the full spectral domain) are found via (6).

3. Discussion

a. Laboratory flume experiment

To compare the Phase method to Benilov and Filyushkin’s method, we ran experiments in the Stanford University Environmental Fluid Mechanics Laboratory’s wave-current flume [see Pidgeon (1999) for a description of this flume]. Partially progressive mechanical waves of 1–2-s period were generated on a mean current of 5–10 cm s$^{-1}$. A Richard Brancker Research WG-50 capacitance-type wave height gauge and NorTek Vector Velocimeter recorded data simultaneously, at 25 Hz, under the control of LabView. Using both wave–turbulence decomposition methods, we calculated Reynolds stresses for ensembles of $2^{13}$ data points (about 5½ minutes of data).

Figures 2 and 3 show the Reynolds stresses observed during many runs under differing current and wave conditions. Under these narrowbanded waves, both the phase method and Benilov and Filyushkin’s method resulted in nearly the same turbulent stresses. Most of the
time, the decomposed stresses were very small, with squared shear velocity \( u^2 \) on the order of \( 10^{-5} \) m\(^2\) s\(^{-2}\). For a flow speed at 1 m above the bed (mab) of 10 cm s\(^{-1}\), and assuming a canonical drag coefficient \( C_D \) of 0.0025 (Dronkers 1964), the quadratic drag relation \( u^2 = C_D U^2 \) predicts a squared shear velocity of \( 2.5 \times 10^{-5} \) m\(^2\) s\(^{-2}\), which is akin to the observed results. Sporadically, however, the decomposed \( u^2 \) was much greater than \( 10^{-5} \) m\(^2\) s\(^{-2}\), but this was an effect of reflection of waves from the downstream end of the tank. Standing waves were thus generated, causing the first term on the rhs of (2) to become nonzero and resulting in real wave stresses (not just a wave bias). Nonetheless, both wave–turbulence decomposition schemes behaved similarly.

b. Field survey

In the field, however, the wave peak is necessarily broader than in the laboratory, and wave–turbulence interaction resulting from the overlap of wave and turbulence time scales can cause the Phase method to inaccurately estimate the Reynolds stress. In the field experiments described in Bricker et al. (2005), we applied the Phase method, alongside the methods of Benilov and Filyushkin (1970) and Shaw and Trowbridge (2001), to data acquired by SonTek Field Acoustic Doppler Velocimeters (ADVs) and a NorTek Vector Velocimeter under broadbanded wind waves at Coyote Point, located on the shoals of southern San Francisco Bay. Experiments were carried out for two weeks during June 2000 and June–July 2002, in a location where a diurnal sea breeze generated wind waves of approximately 50-cm height and 2-s period. During spring tide, water column depth varied with the tide from about 1 to 4 m, and tidal currents had an amplitude of about 30 cm s\(^{-1}\). As in the laboratory experiment described above, we acquired data at 25 Hz, using LabView to synchronize velocity data with WG-50 surface height data. Again, Reynolds stresses were calculated for 5½-min ensembles of velocity data, as this is much longer than the time scale of the individual waves, yet shorter than the time scale over which the sea state and tidal regime varied. Due to the large amount of data acquired, we then averaged these ensembles into sets of 10 (each encompassing about 1 h of data) for presentation in Figs. 4–8, in order to make these figures easier to read.

1) COMPARISON OF DECOMPOSITION METHODS

To quantify the degree of agreement between different methods of wave–turbulence decomposition, we define the normalized root-mean-square (RMS) deviation \( \sigma \) as

\[
\sigma = \sqrt{\frac{\sum (\tau_2 - \tau_1)^2}{N}}.
\]

where \( \tau_1 \) is the stress as determined via one given wave–turbulence decomposition method, \( \tau_2 \) is the stress determined by another method, \( \bar{\tau}_1 \) is the mean of the

Fig. 4. Reynolds stresses resultant from various methods vs Reynolds stresses resultant from the Shaw and Trowbridge method. Squares (□) are raw stresses (not decomposed). Crosses (×) are stresses decomposed via the Phase method. The solid line has a slope of 1. Data are from a SonTek ADV at 20 cmab during the June 2000 experiment.

Fig. 5. Reynolds stresses resultant from various methods vs Reynolds stresses resultant from the Shaw and Trowbridge method. Squares (□) are raw stresses (not decomposed). Crosses (×) are stresses decomposed via the Phase method. The solid line has a slope of 1. Data are from the SonTek ADV at 62 cmab during the June 2000 experiment.
stress $\tau_1$ over the full length of the dataset, and $N$ is the number of shear stress data points (each of which is an average over ten 5-min ensembles of velocity data, as discussed above). Table 1 presents the normalized RMS error between the method of Shaw and Trowbridge and the Phase method for the June 2000 experiment. This table also compares the results of the Shaw and Trowbridge method with simple removal (bandpassing) of the wave peak entirely and with raw (not decomposed) stresses. At both locations, stresses resulting from the Phase method (as opposed to bandpassed or raw data) show the best agreement with those from Shaw and Trowbridge's method (see also Figs. 4 and 5). However, in all cases, the error is smaller near the seabed (20 cmab) than at 62 cmab.

Table 2 (and Figs. 6 and 7) shows similar results but for the June–July 2002 experiment. In addition to comparing the Phase method to the method of Shaw and Trowbridge, the use of a wave gauge allows comparison to Benilov and Filyushkin's method as well. The two ADVs (at 20 and 153 cmab) show that the Phase method (as opposed to bandpassed or raw data) produces the best comparison to both the Shaw and Trowbridge method and to Benilov and Filyushkin's method. Also akin to results from the June 2000 experiment, the error is lowest near the bed and increases at locations higher in the water column.

Results from the NorTek Vector (Table 3 and Fig. 8) are different, however, because this instrument was fitted with an accurate internal compass and tilt meter ($\pm 0.5^\circ$ for direction and tilt), while SonTek ADV orientation was measured manually (with a compass and level). The vector therefore did not experience a large tilt error, causing the wave bias in the raw data to be very low. This resulted in a good agreement between Benilov and Filyushkin's method and raw data for the
vector and also indicates that the majority of the wave bias experienced by the SonTek ADVs was due to inaccurate measurement of these instruments’ orientations.

To prove this, we artificially tilted the data from the NorTek Vector during postprocessing and show the results of this tilting in Table 4. With no tilt, the raw data agree well with data decomposed by Benilov and Filyushkin’s method, indicating that the wave bias was very small. As tilt increases, however, the wave bias in the raw data grows. At a tilt of 3° or greater, Phase-decomposed data agree better with Benilov and Filyushkin–decomposed data than raw data does. Figure 9 shows how bad the agreement is between raw data and Benilov and Filyushkin–decomposed data at a tilt of 10°.

Because of the accuracy of the raw data in the case of the nontilted NorTek Vector, the raw power spectrum truly represented the effects of turbulence without a wave bias, and the Phase method, which interpolates through a predetermined section of this already-accurate power spectrum, gave worse results than no filtering at all (raw data) did. Basically, the Phase method smoothed the section of the power spectrum through which it interpolated and therefore removed accuracy from the result.

Only in the case of an instrument with precisely known orientation, such as the NorTek Vector, might raw data be used with confidence. In all other cases, the Phase method produces a smaller error than simple bandpassing or use of raw data does. This means that, even for relatively short-period wind waves such as the ones we studied, the amount of turbulent stress overlapping with the wave peak is not negligible. Under longer-period waves, such as ocean swell, the wave peak overlaps with yet more energetic scales of turbulence. In the analysis of such swell, we thus expect that the improvement of the Phase method over bandpassing would be even greater than it is under wind waves.

2) WAVE–TURBULENCE INTERACTION

Since the Phase method agrees well with Benilov and Filyushkin’s method and with Shaw and Trowbridge’s method near the seabed, the assumption made in the Phase method, that turbulence under the wave peak in the spectral domain is still equilibrium turbulence that falls on the straight line of Fig. 1, appears valid. Further up in the water column, however, waves stretch turbulence that falls within the wave frequency band, making the straight-line assumption of the Phase method invalid. The Phase method thus attributes such stretched turbulence, which falls within the spectral domain of the wave peak, to the wave bias and not to turbulence, thereby incorrectly estimating the turbulent stress.

As is discussed in Bricker et al. (2005), the extent to which waves stretch turbulence, and thus the extent to which turbulent stresses deduced from the Phase method diverge from stresses resultant from either Shaw and Trowbridge’s or Benilov and Filyushkin’s methods, can be expressed by the rapidity parameter \( R \) (Monismith and Magnaudet 1998), which expresses the ratio of wave-induced strain to turbulence-induced strain:

\[
R = \left( \frac{\partial u}{\partial z} \right)_\text{wave} / \left( \frac{\partial u}{\partial z} \right)_\text{turbulence},
\]

where \( z \) is the vertical coordinate. The Phase method is valid at locations where \( R \ll 1 \), as wave-induced strain is weak here, and turbulence intensities within the spectral region occupied by the wave peak are not amplified above their equilibrium level. In locations where \( R \) grows larger than 1, however, the Phase method cannot distinguish between turbulence stretched by the waves and the waves themselves.

<table>
<thead>
<tr>
<th>Decomposition method 1</th>
<th>Phase</th>
<th>Bandpass</th>
<th>None (raw data)</th>
<th>Decomposition method 2</th>
<th>Phase</th>
<th>Bandpass</th>
<th>None (raw data)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shaw</td>
<td>0.71</td>
<td>0.76</td>
<td>0.56</td>
<td>Shaw</td>
<td>0.59</td>
<td>0.67</td>
<td>0.92</td>
</tr>
<tr>
<td>20 cmab</td>
<td>0.74</td>
<td>n/a</td>
<td>n/a</td>
<td>20 cmab</td>
<td>0.67</td>
<td>n/a</td>
<td>n/a</td>
</tr>
<tr>
<td>153 cmab</td>
<td>0.75</td>
<td>0.84</td>
<td>0.81</td>
<td>153 cmab</td>
<td>0.84</td>
<td>0.95</td>
<td>n/a</td>
</tr>
</tbody>
</table>
Bricker et al. (2005) assumes that wave-induced strain can be found from linear wave theory and that the turbulent strain field derives from the law of the wall. For the conditions at Coyote Point, the result is that at $z/H = 20$ cm (near the bed), $R = 0.3$, which indicates that the wave-induced strain field is weak here, and the Phase method thus works well. At $z/H = 62$ cm, however, $R = 3.0$, indicating that the wave-induced strain field is strong, stretching the turbulence that overlaps with the wave peak, and causing turbulent stresses decomposed by the Phase method to stray away from the results of the other methods. At $z/H = 95$ cm, $R = 8$, and at $z/H = 153$ cm, $R = 21$, causing the Phase method to stray even farther as instrument height increased.

c. Possible sources of error

Our dataset was too large to delineate the wave peak by eye for each spectrum, so the bounds of the wave peak were defined empirically. For each ensemble $S_{uu}$ or $S_{ww}$ of ADV data from the June 2000 experiment, we assumed the wave peak in Fourier space spanned the region between 0.2 Hz on the left and, on the right, 1.0 Hz higher than the highest-energy frequency of the wave peak. For the June–July 2002 data, the wave peak generally did not extend as high in frequency as it did in the June 2002 dataset, so we set the right-side limit to 0.5 Hz above the highest-energy frequency of the wave peak instead. We then used these frequencies as the bounds for the wave peak in (14). Visual inspection of many spectra similar to that in Fig. 1 showed this method to reasonably delineate the bounds of the wave peak. However, the estimate of the bounds of the wave peak can affect the results of the Phase method (discussed below).

Another possible source of error comes from the line we fit (via linear least squares regression) to the turbulence spectrum, used in determining the Fourier coefficients for wave stress $\tilde{U}_j$ (Fig. 1). In fitting this line to the spectrum, we used mostly the part of the PSD at frequencies lower than those at which the wave peak was present. Specifically, we used all the data points at frequencies less than that of the left-hand limit of the wave peak and only half that number to the immediate right of the wave peak. We chose to do this because Lumley and Terray (1983) and Gross et al. (1994) found that the PSD at frequencies higher than the wave frequency can contain more energy than the turbulence itself contains, because waves advect eddies at these frequencies past the sensor, effectively aliasing wave energy into these higher frequencies. Eddies at lower frequencies, however, are not similarly affected. Nonetheless, during our analysis, the PSD at frequencies to the right of the wave peak (though still at lower frequencies) was used when fitting the line to the turbulence spectrum.

| Table 3. Normalized RMS deviation between series of shear stresses as determined via decomposition method 1 compared with those determined via decomposition method 2 for the NorTek Vector Velocimeter during the June–July 2002 experiment. |
|---|---|---|---|
| Decomposition method 1 | Phase | Bandpass | None (raw data) |
| Decomposition method 2 | Shaw | None (raw data) | Benilov |
| 95 cmab | n/a | n/a | 0.60 |
| Table 4. Normalized RMS deviation between series of shear stresses as determined via decomposition method 1 compared with those determined via decomposition method 2 for the NorTek Vector Velocimeter, with the instrument orientation in the data artificially tilted to various angles during postprocessing. |
| Decomposition method 1 | Phase | None (raw data) |
| Decomposition method 2 | Benilov | Benilov |
| Tilted $0^\circ$ | 0.60 | 0.29 |
| Tilted $1^\circ$ | 0.59 | 0.32 |
| Tilted $3^\circ$ | 0.58 | 0.59 |
| Tilted $5^\circ$ | 0.57 | 0.85 |
| Tilted $10^\circ$ | 0.54 | 1.15 |

Fig. 9. Reynolds stresses resultant from various methods vs Reynolds stresses resultant from Benilov and Filyushkin’s method. Squares (□) are raw stresses (not decomposed). Crosses (×) are stresses decomposed via the Phase method. The solid line has a slope of 1. Data are from the NorTek Vector Velocimeter at 95 cmab during the June–July 2002 experiment, with an artificial tilt of $10^\circ$ applied to instrument orientation in data during postprocessing.
Table 5. Normalized RMS deviation between series of shear stresses as determined via decomposition method 1 compared with those determined via decomposition method 2 for the SonTek ADV at 20 cmab during the June–July 2002 experiment. L indicates the low-frequency bound of the wave peak, and R indicates the high-frequency bound.

<table>
<thead>
<tr>
<th>Decomposition method 1</th>
<th>Phase</th>
<th>Bandpass</th>
<th>Phase</th>
<th>Bandpass</th>
</tr>
</thead>
<tbody>
<tr>
<td>Decomposition method 2</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>L: 0.1 Hz</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>R: peak + 0.25 Hz</td>
<td>0.19</td>
<td>0.27</td>
<td>0.56</td>
<td>0.75</td>
</tr>
<tr>
<td>R: peak + 0.5 Hz</td>
<td>0.19</td>
<td>0.32</td>
<td>0.63</td>
<td>0.84</td>
</tr>
<tr>
<td>L: 0.1 Hz</td>
<td>0.20</td>
<td>0.38</td>
<td>0.67</td>
<td>0.93</td>
</tr>
<tr>
<td>R: peak + 1.0 Hz</td>
<td>0.19</td>
<td>0.19</td>
<td>0.56</td>
<td>0.59</td>
</tr>
<tr>
<td>L: 0.2 Hz</td>
<td>0.19</td>
<td>0.25</td>
<td>0.60</td>
<td>0.76</td>
</tr>
<tr>
<td>R: peak + 0.25 Hz</td>
<td>0.18</td>
<td>0.19</td>
<td>0.59</td>
<td>0.67</td>
</tr>
<tr>
<td>R: peak + 0.5 Hz</td>
<td>0.18</td>
<td>0.25</td>
<td>0.60</td>
<td>0.76</td>
</tr>
<tr>
<td>L: 0.2 Hz</td>
<td>0.20</td>
<td>0.19</td>
<td>0.55</td>
<td>0.55</td>
</tr>
<tr>
<td>R: peak + 1.0 Hz</td>
<td>0.20</td>
<td>0.19</td>
<td>0.55</td>
<td>0.55</td>
</tr>
<tr>
<td>L: 0.3 Hz</td>
<td>0.51</td>
<td>0.51</td>
<td>0.72</td>
<td>0.72</td>
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<tr>
<td>R: peak + 0.25 Hz</td>
<td>0.51</td>
<td>0.51</td>
<td>0.72</td>
<td>0.72</td>
</tr>
<tr>
<td>R: peak + 0.5 Hz</td>
<td>0.51</td>
<td>0.51</td>
<td>0.72</td>
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</tr>
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</table>

4. Conclusions

We formulated a new, spectral method of wave–turbulence decomposition (which we call the Phase method) based on the assumption that waves and turbulence do not interact. Via laboratory and field experiments, comparisons with other wave–turbulence decomposition methods show that the Phase method produces good results where the wave-induced strain field is weaker than the turbulence-induced strain field. Where the wave-induced strain field dominates, however, turbulence that overlaps with the wave peak in the frequency domain is stretched, and the Phase method incorrectly attributes such turbulence to the wave bias instead. Even in this case, though, the Phase method produces better results than those produced by simple removal of the wave peak via bandpassing. Therefore, the Phase method proves useful when neither a synchronized wave gauge, nor a second velocimeter, is available. The benefit of the Phase method over bandpassing should be even more pronounced under ocean swell than under the wind waves measured in our study, because ocean swells are longer and thus overlap with more energetic scales of turbulence (which are responsible for most of the Reynolds stress in the flow) than wind waves do.

Our results furthermore show that most of the wave bias existing in turbulent velocity measurements under wind waves over a flat bed may be a result of inaccurate measurement of sensor probe orientation, as suggested by Trowbridge (1998). Using a velocimeter equipped with a precise compass and tilt/roll sensor, we found the wave bias in recorded data to be negligible, in which case wave–turbulence decomposition during postprocessing of data was not necessary.

Acknowledgments. The authors are thankful to Gene Terry for his suggestion that direct examination of spectra might provide a means of separating waves and turbulence. Mark Stacey, Lucinda Shih, and Satoshi Inagaki also provided valuable input. Funding for this work was provided by the UPS foundation and by the Office of Naval Research (ONR Grant N00014-99-1-0292-P00002 monitored by Dr. Louis Goodman).

REFERENCES


