WAVE MODULATED TURBULENT FIELDS AT THE OCEAN SURFACE AND RELATED AIR-SEA FLUXES

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The upper most layers of the ocean, along with the lower atmospheric boundary layer, play a crucial role in the air-sea fluxes of momentum, heat, and mass, thereby providing important boundary conditions for both the atmosphere and the oceans that control the evolution of weather and climate. In particular, the fluxes of heat and gas rely on exchange processes through the molecular layers, which are usually located within the viscous layer, which is in turn modulated by the waves and the turbulence at the free surface. The understanding of the multiple interactions between, molecular layers, viscous layers, waves and turbulence is therefore paramount for an adequate parameterization of these fluxes.

We present evidence of a clear coupling between the surface waves, the surface temperature, and the surface turbulence. The modulation of the surface temperature by the waves lead to a measurable wave-coherent air-sea heat flux. When averaged over time scales longer than the wave period, the coupling between the surface temperature and turbulence leads to a spatial relationship between the temperature, divergence and vorticity fields that is consistent with spatial patterns of Langmuir turbulence. On time scales for which the surface wave field is resolved, we show that the surface turbulence is modulated by the waves in a manner qualitatively consistent with rapid distortion theory.

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1. Introduction

The upper most layers of the ocean, along with the lower atmospheric boundary layer, play a crucial role in the air-sea fluxes of momentum, heat, and mass, thereby providing important boundary conditions for both the atmosphere and the oceans that control the evolution of weather and climate.

The first models of boundary layers on both sides of the air-sea interface were developed from our understanding of the turbulent flow over rigid flat surfaces and extended to the field after the landmark Kansas experiment on the terrestrial boundary layer (Businger et al., 1971). Consequently, models of neutrally stratified flows are based on the well-known "law of the wall" which depends on the assumptions of a constant stress layer and horizontal homogeneity. When the flow is stratified, the "law of the wall" is further modified to include the so-called Monin-Obukhov similarity theory. The coupled air-sea boundary layers, however, are dynamically quite different from a solid flat surface. For one thing, the velocity field is not required to vanish at the interface. The ocean surface responds with drift currents, surface waves and turbulent eddies over a broad range of scales. Consequently, one would expect the dynamics of such an interfacial layer to be significantly different from that over a solid flat surface under similar forcing conditions. Indeed, while there is evidence that, in general, and on average, the Monin-Obukov similarity theory holds over the ocean (Edson and Fairall, 1998; Edson et al., 2004), there are also some notable differences. Over the last decade or so it has also become apparent that surface wave processes can play an important role in the kinematics and dynamics of the boundary layers (e.g. Janssen, 1989, 1999; Komen et al., 1994; Belcher and Hunt, 1998; Hristov et al., 1998; Edson and Fairall, 1998, Uz et al., 2002; Sullivan et al., 2004, 2007), and recent measurements and models of the drag of the sea surface on the atmosphere at moderate to high wind speeds in fact suggest that much of the momentum transfer at the surface is supported by the form drag of the waves. There is also evidence that the air-sea heat flux can be modulated by the wavy surface (Sullivan and McWilliams, 2002; Veron et al., 2008b). In addition, breaking waves generate turbulence (Rapp and Melville, 1990; Agrawal et al., 1992; Thorpe, 1993; Melville, 1994; Anis and Moum, 1995; Melville, 1996; Terray et al., 1996; Veron and Melville, 1999a), which along with small scale Langmuir circulations and coherent structures (Melville et al., 1998; Veron and Melville, 1999b; McWilliams et al., 1997; Veron and Melville, 2001; Sullivan et al., 2004, 2007) may lead to enhanced dissipation and mixing with significant departures from the "law

of the wall" and may also result in increased heat and gas transfer (Jähne et al., 1987; Hasse, 1990; Jähne and Haußecker, 1998; Zappa et al., 2001; Garbe et al., 2004; Schimpf et al., 2004; Turney et al., 2005).

With such richness in the phenomena and dynamics at the surface, one can also expect significant interactions between currents, surface waves, and turbulence. A well known example is Langmuir circulation that results from the interactions between vorticity and the Stokes drift generated by the surface waves (Leibovich, 1983; Thorpe 2004). On smaller time scales, there is also evidence that the turbulence can be significantly coupled with the surface waves (Lumley and Terray, 1983; Cheung and Street, 1988a, 1988b; Thais and Magnaudet, 1995; Teixeira and Belcher, 2002).

Besides, since the air-sea transfers of heat and gas rely in part on exchanges through the diffusive surface layers, and since these layers are typically smaller than the viscous sub-layer, our ability to adequately quantify these fluxes depends on our understanding of the small-scale turbulence that controls the dynamics of the molecular layers. In turn, our understanding of the small scale turbulence depends on our understanding of the multiple interactions between the turbulence, currents, and surface waves.

We present results from field experiments that show evidence of coupling between the surface temperature, the surface waves, the surface turbulent velocity fields, and the surface waves.

2. Experiments

The measurements described here were obtained from three different field experiments. The first was conducted from R/P FLIP, moored approximately 50 miles off the coast of California, San Diego $(32^{o} 38.43'N, 117^{o} 57.42'W, 302 \text{ m depth})$, during 21-29 July 2002. A second was from R/P FLIP, moored west of Tanner Bank $(32^{o} 40.20'N, 119^{o} 19.46'W, 312 \text{ m depth})$, during 20-26 August 2003. We also deployed the full system from Scripps pier in a reduced acquisition mode (two 20-minute records per day), for a period of approximately four months from 4 December 2003 to 6 April 6 2004.

The main instruments comprised an integrated active and passive infrared imaging and altimetry system (Veron et al., 2008a), and an eddy covariance atmospheric flux package. Both systems are described in more detail below.

2.1. Infrared imaging and altimetry system

The active and passive infrared imaging and altimetry system includes an infrared camera (Amber Galileo), a 60 W air-cooled CO_2 laser (Synrad Firestar T60) equipped with an industrial marking head (Synrad FH index) with two computer-controlled galvanometers, a laser altimeter (Riegl LD90-3100-EHS), a video camera (Pulnix TM-9701), a 6 degree-of-freedom motion package (Watson Gyro E604), and a single board computer (PC Pentium 4). All instruments were enclosed in a weatherproof, air-conditioned, aluminum housing. All instruments and computers were synchronized together to within 2 ms and also to GPS time. The infrared camera was set to record temperature images $(256 \times 256 \text{ pixels})$ at 60 Hz, with a 2 ms integration time, yielding better than 15 mK resolution. The footprint of the infrared camera was approximately 2 m by 2 m. The video camera (768 \times 484 pixels) was synchronized to the infrared camera and acquired full frames at 30 Hz. The foot print of the infrared camera was contained within the foot print of the video camera. The infrared CO_2 laser and accompanying marking head were used to actively lay down patterns of thermal markers on the ocean surface in order to study the rate of decay of an imposed surface temperature perturbation while tracking the Lagrangian velocity, shear, and vorticity at the surface. Finally, the laser altimeter measured the distance to the water surface at 12 kHz (averaged down to 50 Hz) with a foot print of 5 cm diameter, contained within both the infrared and video images. The system, among other things, yields the velocity at the surface by tracking active thermal markers laid down with the CO_2 laser (Thermal Marking Velocimetry, TMV) or by performing cross-correlation analysis on the passive surface temperature fields (Particle Image Velocimetry, PIV). The detailed performance of the passive and active IR measurement system for ocean-surface kinematics is described in Veron et al. (2008a).

2.2. Eddy covariance system

In addition to the optical infrared system, we used an eddy covariance system to acquire supporting meteorological and atmospheric boundary layer flux data. The eddy covariance system included a three-axis anemometer/thermometer (Campbell CSAT 3), an open path infrared hygrometer/ CO_2 sensor (Licor 7500), a relative humidity/temperature sensor (Vaisala HMP45), and a net radiometer (CNR1). Turbulent fluxes of momentum, heat and moisture were calculated using the covariance method over 30-min averages. The sonic temperature was corrected for humidity and pressure, Rotation angles for correcting the orientation of the anemometer were obtained from the 30-min averages of the velocity components, and the latent heat flux was corrected for density variations (Webb et al., 1980). For the purposes of this paper, the good agreement between the flux covariance data and estimates using the Toga Coare 3.0 algorithm (Fairall et al., 1996; Fairall et al., 2003) supports the use of the Monin-Obukhov similarity theory for these open ocean conditions (Edson et al., 2004) and gave us confidence that our covariance measurements were adequate measurements of the total flux above the diffusive and wave boundary layers (see Veron et al., 2008b).

The instruments were deployed at the end of the port boom of R/P *FLIP* approximately 18 *m* from the hull at an elevation of 13 *m* above mean sea level (MSL).

The infrared optical system was set-up with the view ports clear of the end of the boom. The meteorological package was placed 15.5 m from the hull on the port boom with all instruments facing upwind with the exception of the net radiometer which was deployed with its axis downwind.

We also deployed 2 fast response, high resolution, subsurface thermistors (RBR ltd -1040 95 ms - 1 Hz sampling rate) placed at 1.2 m and 2 m from the mean sea level and fixed to the hull of R/P FLIP. An upward looking waves-enabled ADCP (RDI Workhorse 600kHz), was also rigidly mounted to FLIP at 15m depth and yielded directional wave spectra and significant wave height for the duration of the experiment. Finally, GPS position and R/P FLIP heading were sampled at 50 Hz and used to correct the ADCP data for FLIP motion and to align all other directions to true north.

3. Results

3.1. Surface waves - Surface temperature interactions

As described above, the footprint of the laser altimeter was located within the infrared image. This also allowed us to examine the modulation of the surface temperature by the waves. From the infrared images, we have generated time series of the temperature averaged over the footprint of the altimeter. To avoid sky reflectance and other effects, only night-time temperature time series were used. Longwave donwelling and upwelling measurements also indicate that contamination by "warm" clouds is negligible. We have made sure that no contamination by the active IR spot was present.

Figure 1a shows frequency spectrograms of the surface displacement for





Fig. 1. Spectrograms of (a) the surface displacement and (b) the surface temperature, at night for the 2003 experiment. (c) and (d) show the squared coherence and phase between the surface displacement and temperature, respectively. (e) shows wind speed and direction for the duration of the experiment.

the duration of the R/P FLIP experiment in August 2003. The swell peak is clear, and also a wind-wave peak appears with the wind event (figure 1e), showing the classical downshift in wind-wave peak frequency with time. The quality of the near-IR laser wave gauge data is dependent on the roughness of the sea surface with drop outs increasing for lower wind speeds. The data shown here were good for frequencies up to 1 Hz and low-pass filtered at that frequency. The surface displacement spectrum is unremarkable, showing both wind-wave and swell peaks at 0.16 and 0.08 Hz, respectively, and a ω^{-4} slope above the wind-wave peak. The temperature data on figure 1b are considered accurate for the full range of frequencies shown: up to 10 Hz. The temperature spectrum displays a peak at the wind-wave peak frequency and an ω^{-1} slope above the wind-wave peak, up to 2-3 Hz, and then $\omega^{-9/2}$ up to 10 Hz. Less apparent, but present nonetheless, are maxima in the temperature spectra at the swell peak frequencies. Figures 1c and 1d show that the low-frequency long-waves (swell and wind-waves) correlate well with the temperature fluctuations, while the coherence rapidly goes to zero at frequencies above approximately 1 Hz. The phase between the temperature and surface displacement near the peak of the coherence is negative, indicating that the maximum in temperature lags the maximum surface displacement. These results are qualitatively consistent with the results of Simpson and Paulson (1980) and Miller and Street (1978), but depart from those of Jessup and Hesany (1996), at least at the higher wind speeds. The reason for this discrepancy is still unclear. (However, more recent measurements by Jessup (personal communication) are consistent with our results and others cited above.) While we find that the phase between surface displacement and temperature waves at the surface is negative, it shows a trend with the wind speed.

In the absence of direct measurements of the thermal boundary layers in the immediate neighborhood of the surface, we cannot use the surface temperature and surface displacement to directly measure the wave-modulated heat flux; however, we could estimate the wave-modulated heat flux in the atmospheric boundary layer at elevations of O(10) m based on linear coherence techniques. We found that the fraction of the heat flux coherent with the surface waves scales with variance of the wave slope (Veron et al. 2008b). However, there is a lot of scatter in the data, implying other dependencies not considered here. Some may be due to the z-dependence of the wave induced flux, which may be expected to have both oscillatory and monotonically-decaying components having a surface wavenumber dependence (c.f. Sullivan & McWilliams, 2002), whereas these measurements were made at a fixed height above MSL.

3.2. Surface turbulence - Surface temperature interactions

As stated in section 2, the infrared images can be used to calculate the two-dimensional surface velocity \mathbf{u} , and other kinematic variables from infrared surface temperature images (Veron et al., 2008a). The velocity at the surface can further be projected onto the directions parallel and perpendicular to the wind or wave directions. Here, we employ an algorithm typically employed for Particle Image Velocimetry (PIV) and calculate the

surface velocity by performing a running normalized cross-correlation on sub-windows of surface temperature images separated by 33 ms. Each cross correlation yields a local average displacement (over the sub-window) which is then used in conjunction with the instantaneous image resolution and time interval between the successive images to estimate an average surface horizontal velocity. In most cases, we used 8x8 pixel sub-images with 50%linear overlap and image pairs separated by 166 ms yielding 6 velocity maps per second, each having a spatial resolution of approximately 3 cm. Details on the performance of this technique can be found in Veron et al. (2008a). The surface velocity field \mathbf{u} obtained is the sum of velocities associated with the mean currents $\mathbf{u}_{\mathbf{c}}$, the orbital velocity of the surface waves $\mathbf{u}_{\mathbf{w}}$, and the turbulence \mathbf{u}' . The benefit of this PIV technique is the ability to simultaneously study processes involved in the evolution of the surface temperature and surface velocity fields including mean currents, wave orbital motions, and turbulence. In particular, direct surface velocity measurements, along with single point altimetry, allow us to measure surface wave directional spectra, and the spectra of the surface kinematics, as illustrated below.

Here, we examine the spatial distribution of the kinematic fields along with that of the surface temperature. Figure 2 shows the two-dimensional wavenumber spectra for the divergence and vorticity which were obtained from the PIV processing of the temperature images. Figure 2 also shows the two-dimensional wavenumber spectrum for the surface temperature. The spectra were obtained by averaging 120 individual spectra of a single velocity or temperature image, taken randomly during one minute of data. The data shown here were collected under wind speeds of 8 $m s^{-1}$ at the beginning of the wind event at 05:03 UTC, on 23 Aug 2003, during the development of the wind sea. First, we note that the two dimensional spectra for the surface divergence, vorticity, and temperature, exhibit some anisotropy indicating that there is some degree of directionality in the spatial structures of these fields. Secondly, we note that the spectrum of the vorticity (2b) is approximately aligned with that of the temperature (2c). This is a consequence of the coupling between surface temperature and velocity. We know from previous work (Veron and Melville, 2001) that small scale Langmuir circulations and coherent turbulence (streaks) are present at these scales over this range of wind speeds. Even if instantaneous realizations of the surface temperature and velocity do not clearly exhibit such structures, time averages can reveal their presence. Here, the twodimensional spectra are averaged over one minute of data. Colder streaks in the temperature images correspond to water parcels that have been or-



Fig. 2. Two-dimensional wavenumber spectra of the surface divergence (a), vorticity (b) and surface temperature (c) for a wind speed of 8 $m s^{-1}$. The spectra were calculated from an average of 120 individual spectra taken over one minute of data starting at 05:03 UTC on 23 Aug 2003. (d) Shows the directions for the wind, from 305°, the wind waves, to 117°, the swell, to 97°, the vorticity spectral peak, 52°, the divergence spectral peak, 122°, and the temperature spectral peak, 43°.

ganized by the underlying Langmuir cells and have collected in windrows at the surface. These water parcels have been in contact with the air the longest and therefore have had time to exchange significant heat, leading to colder surface temperatures. In addition, these parcels of water also have been exposed to the wind stress the longest and therefore, the "cold" surface temperature streaks also correspond to surface "jets" in which the surface velocity is larger than that of the surrounding fluid. A schematic of these surface jets can be found in Leibovich (1983). As a consequence, there are two horizontal shear layers on each side of the temperature streaks where vertical vorticity is enhanced. Thus, when Langmuir circulations are present, the surface exhibits streaks of enhanced velocity, or surface jets,

9

along with streaks of colder temperature and streaks of vorticity, all in the same direction and aligned with the wind. This is precisely what is shown in figure 2. The reader is reminded that the two-dimensional spectrum will show the largest signal in the direction of the largest variance, i.e. the direction perpendicular to the streaks. For example, in the data shown in figure 2, the vorticity (temperature) spectrum peaks at 52^{o} (43^{o}) indicating that the streaks in the vorticity fields are oriented along the direction $322^{o} - 142^{o}$ ($313^{o} - 133^{o}$), roughly aligned with the wind.

Finally, we note that the spectrum of the divergence is perpendicular to that of the vorticity. This is also consistent with the presence of Langmuir circulations. Indeed, Langmuir circulations are generated from the positive feedback between the surface jets, vertical vorticity, and the shearing and stretching of the vertical vortex lines by the Stokes drift (Craik 1977, Leibovich 1977). This means that the surface waves are propagating in the downwind direction, with their wavenumber vector approximately aligned in the direction of the surface streaks. Accordingly, regions of divergence and convergence associated with the non-linear surface wave field and even perhaps breaking waves, will be aligned parallel to the wave fronts, perpendicular to the jets of colder surface temperature and enhanced vertical vorticity. Indeed, for the data shown in figure 2, the spectral peak for the divergence is approximately in the direction of the wind and wind waves indicating that the structure in the surface divergence field is perpendicular to the predominant wind and wind-wave directions, and also perpendicular to the streaks in the surface temperature and surface vorticity. Hence, when Langmuir circulations (or Langmuir turbulence) are present, the twodimensional wavenumber spectra for the vorticity and divergence are perpendicular to one another with the spectrum for the vorticity parallel to that of the surface temperature. The presence of developing Langmuir circulations during the early stage of this wind event was also confirmed by visual inspection of the temperature images and visual observation of the surface during the experiment.

3.3. Surface turbulence - Surface waves interactions

We have seen in the preceding section that the surface kinematics is coupled with the developing wind seas, on time scales longer than approximately a minute. This raises the question of the coupling between the surface turbulence, waves, and currents on shorter time scales, and in particular the coupling between turbulence and surface waves.

As described above, the footprint of the laser altimeter was located

within the infrared image and therefore within the surface velocity and surface kinematic fields. This allows us to examine the modulation of the surface velocity and kinematic fields by the waves. From the infrared images and the resulting velocity images, along with the wave directional spectra obtained from the processing described in section 3c, we have decomposed the surface velocity field into two components u_1 and u_2 , which, respectively, are aligned with, and orthogonal to, the direction of propagation of the wind-waves. To avoid sky reflectance and other effects in the infrared images, only nighttime data were used. Also, we have subtracted imagemean velocity from every velocity vector of that particular PIV estimate. This operation subtracts both currents $\mathbf{u_c}$ and wave orbital velocity $\mathbf{u_w}$, and yields the surface turbulent velocity field, $\mathbf{u'}$. From these we can generate time series of the statistics of the turbulent surface kinematics variables averaged over the footprint of the infrared imager.

Figure 3a shows frequency spectrogram of the surface displacement η for the duration of the *FLIP* experiment in August 2003. Figure 3b shows that the frequency spectra for the turbulent velocity fluctuations in the direction of the wave propagation, u'_1 , exhibit peaks at the local maxima of the wind-wave spectra. Figures 3c and d show the squared coherence $C_{u'_{1}\eta}$ and phase $\Phi_{u'_1\eta}$, between u'_1 and η , respectively. It shows a peak in the squared coherence of approximately 0.2 - 0.3, rapidly going to zero at frequencies above approximately 1 Hz. The phase between the turbulence and surface displacement near the peak of the coherence is slightly negative, indicating that the maximum in turbulence intensity lags the maximum surface displacement. This is presumably a consequence of the enhanced wind-stress and turbulence on the windward side of the wave. This is corroborated by identical results for the other kinematic variables (not shown). Incidentally, this is consistent with observations of warmer skin temperature also on the windward side of the waves that results from enhanced wind stress and the resulting turbulence which leads to the disruption of the aqueous viscous layer and the thermal molecular layer, i.e the cool skin (Miller and Street, 1978; Simpson and Paulson, 1980; Veron et al., 2008b).

Increased turbulence and vorticity near the wave peak could also be a result of the shearing effects on the turbulence and the local stretching of vortex lines by the passing surface waves (Teixeira and Belcher, 2002). We examine next the effect of the waves on the turbulence over short, wave resolved, time scales.

We wish to extract the wave-coherent turbulent quantities. Following the standard Reynolds decomposition $\mathbf{u} = \overline{\mathbf{u}} + \mathbf{u}'$ where the overbar repre-





Fig. 3. Spectrograms of (a) the surface displacement η and (b) the along wave turbulent surface velocity u'_1 , at night. (c) and (d) show the squared coherence and phase between the surface displacement and the turbulent surface velocity, respectively.

sents spatial averages over the PIV estimates (which are distributed over the footprint of the infrared imager). The primes indicate fluctuating quantities taken as deviations from the mean which can be further decomposed into wave coherent, $\tilde{\mathbf{u}}$, and turbulent components, \mathbf{u}'' . Here we use a technique similar to that presented in Veron et al., (2008b) to extract the averaged wave coherent quantities. We use the coherence and phase of each variable with the surface displacement, η , due to the waves. Denoting the spectrum of u'_1 by $S_{u'_1u'_1}$, we find that the spectrum of u'_1 coherent with the waves

(i.e. the spectrum of \tilde{u}_1) is given by $S_{\tilde{u}_1\tilde{u}_1} = S_{u_1'u_1'} \times C_{u_1'\eta}$.

It follows that the variance of the phase-coherent velocity in the direction of the wave propagation, (\tilde{u}_1) , is given by:

$$\overline{\tilde{u}_1^2} = \int S_{\tilde{u}_1 \tilde{u}_1} d\omega, \tag{1}$$



Fig. 4. Variance of the wave-coherent, along wave, surface velocity turbulence with the wind speed (a) and the rms surface wave slope (b). The large symbols show the bin-averaged data and the solid lines are the fits to the bin averaged data.

Figure 4 shows the variance of the wave-coherent, along wave, surface velocity turbulence as a function of the wind speed (a) and the surface wave slope^a (b). While noisy, the data show that there is some fraction of the

^aSince the surface slope, (η_x, η_y) , was not directly measured in the experiments, the root mean square wave slope is instead approximated from the wave elevation frequency spectrum by $S = \left(\int_{0.5f_p}^{1.5f_p} S_{\eta\eta}k^2 df\right)^{\frac{1}{2}}$, where f_p is the peak frequency and k the wavenumber given by the dispersion relationship. Finally, we estimate $ak = \sqrt{2}S$. This is valid for the gravity wave range only.

surface turbulence that is coherent with the surface waves. The data show that the variance of the wave coherent surface turbulence increases with wind speed and wave slope, with perhaps a tighter fit with the wave slope. Figure 4 also shows data taken in shallow water and at lower wind speeds from Scripps Pier during the winter of 2003 (see Veron et al., 2008a for details). The two data sets merge continuously and the large symbols are the bin averaged data. The solid lines show the linear fit through the data (forcing a zero intercept for figure 4b).

Of further interest is the total fraction of this wave coherent turbulence to the total turbulent intensities. Following the work of Towsend (1976, pp71-77) and Texiera and Belcher (2002), we use the frame work of rapid distortion estimate this ratio. Assuming initially isotropic turbulence and plane straining by a unidirectional, monochromatic linear wave field of amplitude a and wavenumber k, then

$$\frac{\tilde{u}_1^2}{u_1'^2} = -\frac{4}{5} \frac{\left(\beta - \beta^{-1}\right)}{\left(\beta + \beta^{-1}\right)} + \frac{3}{35} \left(\beta - \beta^{-1}\right)^2 \tag{2}$$

where, β denotes the strain ratio, i.e. the ratio of the velocity gradient tensor (or vorticity components) with and without distortion. In this case, β varies as:

$$\beta \approx 1 - ak. \tag{3}$$

Neglecting terms higher than quadratic in the wave slope ak, we find that:

$$\frac{\tilde{u}_1^2}{u_1'^2} \sim \frac{4}{5}ak + \frac{26}{35}\left(ak\right)^2 \tag{4}$$

which indicates that the normalized wave-modulated turbulence in the direction of the wave propagation is of order O(ak). It should be noted that this theoretical estimate relies on assumptions of isotropy and homogeneity in the turbulence. These can be violated near the surface (Belcher et al., 1994) and in the case where the turbulence is injected by localized breaking events; although, as noted by Texiera and Belcher (2002), even in the latter case, the integral length scale of the turbulence is small compared to the wavelength of the surface breaking wave, hence satisfying the scale separation necessary for rapid distortion theory to be applicable. On that note, we do not expect the swell to significantly affect the surface turbulence, at least in a rapid distortion sense, since the strain rate from the orbital motion of the swell is generally too weak compared to that of the turbulence on itself. In fact, figure 3c shows no coherence between the waves and the surface turbulence at the swell frequency. Finally, the estimate above is developed for monochromatic linear waves and not for a full wave spectrum, thus we expect differences between this idealized theoretical prediction and the observed field data. Nevertheless, we feel that it is valuable and interesting to compare the data with the available theoretical estimates. Figure 5 shows the ratio of the variance of the wave coherent turbulence (\tilde{u}_i^2) to the total variance in the turbulence $(\overline{u_i'^2})$ as a function of the rms wave slope. The variance ratio in the direction of the waves increases with the wave slope. We should note here that we expect this ratio to be relatively noisy, especially at low wind speeds and wave slopes as $\overline{u_i^{\prime 2}}$ becomes smaller. In figure 5, we also show the theoretical estimate from equation 4. While there is a large scatter in the data, and an expected flattening due to noisy data for low surface wave slopes, the theoretical estimate agrees reasonably well with the data at the larger slopes. These data appear to be the first field observations qualitatively supporting the use of rapid distortion theory to predict the modulation of surface turbulence by surface waves.



Fig. 5. Fraction of variance of the wave coherent turbulence $(\overline{\tilde{u}_1^2})$ to the total variance in the turbulence $(\overline{u'_1}^2)$ as a function of the wave slope. The large symbols show the bin-averaged data and the solid gray line shows the rapid distortion theory prediction of equation 18.

4. Conclusions

We have shown that the surface wave modulate the surface temperature and that the phase relationship between elevation and temperature waves depends on the wind speed. The peak in temperature is located on the rear face (upwind) of the surface waves (phase between -180° and 0°) for wind speeds larger than 1 to 2 m s⁻¹. This appears to agree with the suggestions

15

of Simpson and Paulson (1980) that the increase in surface temperature was a result of the thinning of the thermal boundary layer by the wind stress on the rear face of waves. At very low wind speeds, the temperature peak is located on the front face (downwind) of the surface waves. This is consistent with the theoretical results of Witting (1972) and the laboratory data of Miller and Street (1978). The wind speed dependence of the phase is also in agreement with the laboratory data of Miller and Street (1978). At very low wind speeds, the modulation of the surface temperature is likely caused by the mechanical straining of the surface, inducing a peak temperature on the front face of the wave, while at higher wind speeds, the wind stress contributes to the thinning of the thermal boundary layer. At higher wind speeds yet, micro-breakers, parasitic capillary waves and eventually larger breaking waves locally destroy the surface thermal diffusive layer creating an enhanced surface temperature on the front face of the waves. This could in fact explain part of the data of Jessup and Hesany (1996), who found a positive phase between the surface waves and temperature fluctuations when the wind and waves were co-propagating. Indeed, their data were collected at relatively low wind speeds $(U_{10} < 6 \ m \ s^{-1})$ in the presence of large swell. It is then possible that the temperature modulations that they observed were the result of mechanical straining. However, this does not explain their observations of the phase dependence on the relative direction of the wind and swell.

Also, we have shown that there are periods of strong coupling between the surface wave field and wind forcing, and both the surface kinematic and temperature fields. During the early phase of the wind event described here, the two-dimensional wavenumber spectra of the surface vorticity and surface temperature, when averaged over time scales of one minute and longer show that these mean fields are organized in longitudinal structures aligned with the wind direction. In addition, we also show that the vorticity and divergence structures at the surface are orthogonal to each other. This coupling, which is revealed when the surface fields are averaged over many wave periods as is done with the CLII approach, is consistent with the presence of Langmuir circulations (or Langmuir turbulence). As the wind stabilizes, additional order in the surface kinematic fields is found and the structures in the normal deformation and shear are aligned with that in the fields of surface divergence and vorticity, respectively. This orthogonatily can be explained by a surface velocity field of hyperbolic form. In turn, hyperbolicity leads to significant stretching and convergence in the mean surface velocity field, once again consistent with the effects of Langmuir

circulations.

Over short time scales where the waves are resolved, we observe the modulation of the surface turbulence by the surface waves. We show that there is a maximum wave-coherent turbulent intensity near the crests of the surface waves. These results are consistent with the modulation of the skin temperature by the surface waves where warmer temperatures, resulting presumably from the destruction of the molecular thermal layer by the turbulence, occur near the crest of the waves. Our data shows that the relative wave-coherent turbulent intensities vary with both wind speed and surface wave slope. The data are qualitatively consistent with rapid distortion theory but more work is needed for more complete quantitative comparisons.

acknowledgments

We thank Peter Matusov for his help in the design, construction, testing and deployment of the original instrumentation. We thank Bill Gaines, and the captain and crew of the R/P FLIP for all their help in the field experiments. We thank John Hildebrand and Gerald D'Spain for sharing space and facilities on R/P FLIP. This research was supported by an NSF grant (OCE 01-18449) to WKM & FV and an ONR (Physical Oceanography) grant to WKM. Finally, we wish to thank Mahmood Mohammad for organizing the Regional NSF-CBMS Conference on Water Waves.

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