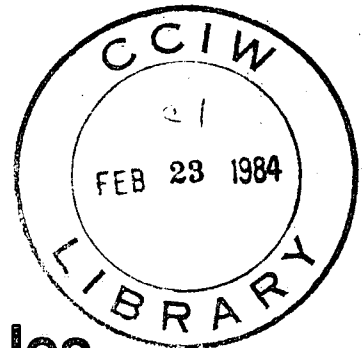**Environment Canada**

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DOES THE SCATTEROMETER SEE WIND SPEED OF FRICTION VELOCITY?

by

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LE DIFFUSIOMÈTRE VOIT-IL LA VITESSE DU VENT OU LA VITESSE DE FROTTEMENT?

par

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RÉSUMÉ

Les études de rétrodiffusion radar sur la surface de la mer renvoient
soit à la vitesse du vent, U, soit à la vitesse de frottement, u_*. D'après
la théorie de la diffusion de Bragg, les variations observées dans la
rétrodiffusion sont directement liées à la hauteur des ondes capillaires
et de gravité modulées par les plus grandes ondes en inclinaison et par
une déformation du champ d'ondes courtes. Une question se pose alors :
à quelle caractéristique du champ de vent le spectre des nombres d'onde
des ondes capillaires et de gravité est-il le plus probablement lié?
Dans le présent rapport, on examine pourquoi U est le paramètre météorologique
qui doit être associé à la rétrodiffusion de la bande L à la bande K_u.
En se basant sur la théorie et sur des résultats expérimentaux, on démontre
que le paramètre dominant est $U/C(\lambda)$ où U est la vitesse du vent à une
hauteur d'environ $\lambda/2$ pour des vagues dont la vitesse de phase est $C(\lambda)$.

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DOES THE SCATTEROMETER SEE WIND SPEED OR FRICTION VELOCITY?

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ABSTRACT

Studies of radar backscatter from the sea surface are referred either to the wind speed, U, or friction velocity, u_*. Bragg scattering theory suggests that these variations in backscatter are directly related to the height of the capillary-gravity waves modulated by the larger waves in tilt and by straining of the short wave field. The question then arises as to what characteristic of the wind field is most probably correlated with the wave number spectrum of the capillary-gravity waves. This study reviews the justification for selecting U as the appropriate meteorological parameter to be associated with backscatter from L-band to K_u-band. Both theoretical reasons and experimental evidence are used to demonstrate that the dominant parameter is $U/C(\lambda)$ where U is the wind speed at a height of about $\lambda/2$ for waves having a phase speed of $C(\lambda)$.

HISTORICAL REVIEW

The steps that lead to the development of the SEASAT-SASS go back to the early theories of radar backscatter from the sea surface and to measurements of the high frequency part of the spectrum of wind generated waves in wind wave flumes. Programs of the Advanced Applications Experiment Office of NASA were initiated in the middle 1960's to measure the normalized radar backscattering cross section from aircraft as reported by Moore and Bradley (1969). Measurements of the high frequency waves in wind-wave flumes as in Pierson and Stacy (1973) and Mitsuyasu and Honda (1974) were made in attempts to relate the water waves to the radar waves by means of various versions of Bragg scattering theory, as in Rice (1951), and others. There were other contending theories of backscatter that depended on the entire wave number spectrum as in Chia (1968). The full spectrum is still, in a sense, needed to correct Bragg theory and to account for low incidence angles.

To connect aircraft measurements at 500 m, or lower, to wind-wave flume measurements in flumes several meters high requires knowledge of the variation of the time averaged wind with height. Experiments in wind wave flumes usually provided the quantities, u_* and z_0, and aircraft measurements provided a mean wind at flight altitude or were made near some surface platform to obtain wind speed and direction, air temperature, sea temperature and, perhaps, relative humidity at some height above the sea surface.

The available theory for connecting wind wave flume measurements to free atmosphere measurements is that of Monin Obukhov (1954), which requires the empirical determination of the function, $\phi(z/L)$, where L is the stability length and of either a relationship between z_0 and u_* or the neutral drag coefficient, C_{DN}, defined by

$$C_{DN} = u_*^2/(\overline{UN}_{10})^2 \tag{1}$$

The generally accepted function for $\phi(z/L)$ is presently the Businger-Dyer function but there are no generally accepted functions for either $C_{DN}(\overline{UN}_{10})$ or $z_0 = z_0(u_*)$. It is, however, generally agreed that the neutral drag coefficient is not a constant.

The details on the development of the SEASAT-SASS can be found in the Skylab EREP Investigations Summary (NASA SP-399), the special issue in Science (Vol. 204, June 1979), the IEEE Journal of Oceanic Engineering (Vol. OE-5, April 1980) and two Journal of Geophysical Research special issues (Vol. 87 No. C5 April 1982 and Vol. 88, No. C3 Feb. 1983) and in the papers cited therein. It will be necessary to search rather diligently to find out which, of many, closure relations was used either to relate wind-wave flume data to the free atmosphere or to refer the wind to some constant height above the sea surface. It will also be difficult to find the rationale for the decision to relate backscatter to the mean wind speed and direction at a height of 19.5 m above the sea surface.

THE PRESENT SITUATION

The situation becomes even more confusing when one reads Jones, et al. (1977) Jones and Schroeder (1978), Ross and Jones (1978), Liu and Large (1981), O'Brien, et al. (1982), Brown (1983) and Pierson (1983). The 1977 reference finds a power law (to be defined later) relationship between σ^o(NRCS) and wind speed for upwind, downwind and crosswind and gives forms for σ^o versus χ for fixed $|\vec{V}|$ and θ. Jones and Schroeder (1978) use a z_o versus u_* relationship given by Cardone (1970) and find a power law relationship for u_*. Ross and Jones (1978) measured backscatter, both polarizations, $\overline{UN}_{10}$, u_* and z_o inferred from aircraft data at 150 m over a fetch from zero to 34 km and found essentially constant backscatter, wind speed, and u_* values for $\overline{UN}_{10}$ equal to 13.0 m/s and 9.0 m/s. The gravity waves increased in height over the fetch, but none of the other parameters changed by more than what would be expected as a result of mesoscale turbulence. Liu and Large (1981) tried to relate JASIN measurements of u_* to σ^o and found "no significant difference between the correlation of σ^o versus U_N and σ^o versus u_* "(see also Pierson (1983)). The S^3 group (O'Brien, et al. (1982)) believe that the vector wind stress, $\vec{\tau}$, is the primary quantity needed to study wind driven ocean currents but do not address either the relationship between the wind and the wind stress or the mechanism by which the wind actually generates ocean currents. Brown (1983) writes that "Since the establishment of good parameterizations between scatterometer signal and surface winds in regions of rapid change and strong air-sea interaction are hampered by lack of good wind data, the rationalization for wind rather than surface stress algorithms is not so strong. As an appropriate consort to the surface reflectivity, we might as well accept u_* rather than its ersatz companion, the wind. In regions where air-sea temperature differences are available, there is no difference, as [(1)] accurately relates U(z) to u_*. However, in dynamic regions, such as an intense cyclone with resultant large air-sea temperature difference, u_* can be expected to provide better correlations with the scatterometer". Pierson (1983) identifies the problem and argues that the wind is the more basic parameter and should be the quantity correlated with backscatter. An experiment to help clarify the problem is outlined for some future scatterometer such as NROSS.

STATEMENT OF PROBLEM

The various paper cited above document a variety of opinions on the relative importance of various parameters and measurements of the dependence of radar backscatter or some property of the atmosphere in motion, To do so, it is first necessary to state the problem according to our present understanding of it.

The primary quantity of meteorology is the air itself. It has mass, or density, and a wide variation in its ability to hold water vapor.

Air in motion is the wind, so wind is a secondary quantity. The speeds and directions of the winds are highly variable quantities near the Earth's surface, and even more so at the higher elevations of the atmosphere. For synoptic scale meteorological applications, the winds in a time averaged sense are not measured very accurately by presently available conventional and remotely sensed meteorological methods (with the possible exception of the SASS) as shown by Pierson (1983) and others.

Air in motion, the wind, is turbulent. For the study of turbulence, a tertiary quantity is the downward flux of turbulent momentum toward the sea surface in the planetary boundary layer as defined by (2).

$$u_*^2 = \tau/\rho = -<u' \, w'> \tag{2}$$

The quantity, u_*^2, is not routinely measured over the oceans. The quantity indicated by (2) is in essence the cospectrum of the fluctuations about some mean value of U as in $U = \bar{U} + u'$ averaged over an appropriate time interval at some reasonable height above the sea surface and the fluctuations in the vertical velocity, which are usually taken to have a zero mean. This cospectrum is shown by Large and Pond (1981) and high quality data on (2) can be found in Large's Thesis, Large and Pond (1981) and Smith (1980), among others. Near the sea surface for a height somewhere above the waves up to some higher elevation (2) is assumed to be a slowly varying quantity with time but more or less independent of height.

There are other ways that the wind affects the sea surface other than the downward flux of momentum implied by $<u' \, w'>$. One of the most important is the variable dynamic pressures applied to the moving water surface and the phase and magnitude of these varying pressures relative to the waves that are present. Fluxes of heat and water vapor are also involved as eddy quantities, each having problems associated with their bulk parameterization.

Somehow or other, air in turbulent motion over the ocean generates waves. These waves are more or less random as a time history at a point, short crested, and variable as to their properties as a function of the motion of the air over them, distance over which the air has moved and the time duration of the velocity of movement. These waves are described ultimately by non-linear equations, but present theories are at best third order approximations.

Within a linear theory, the spectrum of the waves generated by the moving air covers about three orders of magnitude in frequency and five, or so, in wave number. For waves shorter than about 5 cm, surface tension becomes important. Since the radar wavelength of the SASS on SEASAT is 2 cm, capillary waves near 2 cm in length are important in backscattering theories. Capillary waves have rounded crests and sharper troughs that point down into the water. According to Crapper (1957) the limiting form of pure periodic capillary waves produces waves with a very strange profile that can be as high as they are long.

The SEASAT-SASS estimated the power of the electromagnetic radiation backscattered from the sea surface. Actually the received power plus the noise in the system was estimated and then an estimate of the noise was subtracted. For low signal ($\sim 10^{-15}$ to 10^{-16} watts) to noise ratios (- 10 to -15 db), it was quite possible to obtain a negative estimate of the received power, which, unfortunately, was ignored in further data processing.

Pierson and Stacy (1973) showed that capillary waves are not generated in a

wind-wave flume for winds corresponding to friction velocities up to about 12 cm/sec, which yields a wind of about 4 m/s at 19.5 m. Liu and Lin (1982) have used laser methods to estimate the frequency spectrum of capillary waves. Their results differ from earlier studies, but still suggest that the spectrum increases more rapidly at higher frequencies than at lower frequencies in the capillary range. Since wavenumbers and not frequencies are the appropriate quantities in backscatter theories capillary waves advected forward and backward by the longer gravity waves may have their frequencies Doppler shifted so that the frequency spectrum cannot be simply transformed to the wavenumber spectrum.

The SEASAT-SASS (and by extention any other scatterometer) measures the roughness of the sea surface, including, perhaps, flying spray, and the sharp corners at the crests of gravity waves, and the Bragg scattering from a surface tilted by gravity waves and related to the capillary vector wave number spectrum. A scatterometer measures neither the wind nor u_*, or by extention, $\tau = \rho u_*^2$. The connection between the sea surface roughened by the moving air and the estimates of the normalized radar backscattering cross section measured by a scatterometer is thus built on a chain of theoretical and experimental results that depend on the true nature of the moving air in the planetary boundary layer and on how waves are generated. Neither the effective neutral wind at 19.5 m above the mean sea surface nor u_* may prove to be most directly related to backscatter measurements.

THE SASS-1 MODEL FUNCTION

The design of the experiments to determine the relationship between backscatter and moving air began several years before the launch of SEASAT. The worst case design was predicated on being able to determine a wind of 4 m/s within plus or minus 2 m/s based on the aircraft data of Jones, et al. (1977) and on Pierson and Stacy (1973). There were several dozens of published relationships for $C_{DN} = C_{DN}(U_{10})$, or variants thereof relating z_o and u_*. It was impossible to pick one and prohibitively expensive to measure u_* over a wide enough range of wind speeds during the anticipated lifetime of SEASAT.

The choice of the effective neutral wind at 19.5 m was based on two considerations. They were that anemometers on transient ships are at heights considerably in excess of 10 meters and that winds corrected to this height by means of Monin-Obukhov theory were relatively insensitive to which choice, of many, was made for equation (1) even after correction for stability effects.

Consider, the simplified situation for which the drag coefficient is constant and might equal 10^{-3}, or perhaps $2 \cdot 10^{-3}$ or perhaps $3 \cdot 10^{-3}$. The mean wind at 40 m can be referred to the wind at 19.5 m for neutral stability by

$$\bar{U}(19.5) = \frac{\bar{U}(40)(1 + 1.63\ C_{DN}^{\frac{1}{2}})}{(1 + 3.38\ C_{DN}^{\frac{1}{2}})} \tag{3}$$

and the wind at 5 m can be referred to 19.5 m by

$$\bar{U}(19.5) = \frac{\bar{U}(5)(1 + 1.63\ C_{DN}^{\frac{1}{2}})}{(1 - 1.69\ C_{DN}^{\frac{1}{2}})} \tag{4}$$

The wind stress computed for a mean wind at 40 m is given by

$$\tau = \rho\ C_{DN}\ \bar{U}_{40}^2\ (1 + 3.38\ C_{DN}^{\frac{1}{2}})^{-2} \tag{5}$$

and for 5 m by

$$\tau = \rho \, C_{DN} \, \bar{U}_5^2 \, (\, 1 - 1.69 \, C_{DN}^{\frac{1}{2}})^{-2} \tag{6}$$

For the same wind at 10 m, the stress varies by a factor of 3 and u_* varies by 73% under these assumptions. With these equations, and their obvious extentions, the wind measured at any height can be referred to 19.5 m. A wind for neutral stratification of 15 m/s at 40 m yields winds of 14.09, 13.77 and 13.53 m/s, for example, which for a factor of 3 in C_D produces a 56 cm/sec change in the wind at 19.5 m. Similarly a wind of 15 m/s at 5 m produces winds of 16.65, 17.4 and 18.00 m/s at 19.5 m for a 1.35 m/s change. Corrections for stability have roughly the same effect for a considerable variation in C_{DN}. The choices avoided a decision as to which of many drag coefficients should be used. However, it did not avoid the basic question of how the roughened sea surface is related to the turbulent moving air over the ocean.

When these decisions were made, the inherent difficulties of measuring the winds correctly by conventional meteorological means were not clearly in focus. The evolution of the understanding of the problem can be traced through the papers and special issues that have been cited above.

The SASS-1 model function is an attempt to relate the effective neutral wind speed at 19.5 m, which is (or was) the anemometer height on the British weather ships I and J, and wind direction, to backscatter, either vertically or horizontally polarized, by means of a relationship of the form

$$\sigma^o = \sigma^o \, (|\vec{V}|, \, \chi, \, \theta) \tag{7}$$

where $|\vec{V}|$ is the magnitude of the wind vector, χ, is the wind direction relative to the pointing direction of the radar beam and θ is the incidence angle. The model for the model function was simplified to the form

$$\sigma^o = K(\chi, \, \theta) \, V^{M(\chi, \, \theta)} \tag{8}$$

or to

$$\sigma^o_{db} = 10 \, \log_{10} \, \sigma^o = 10 \left[G(\chi, \theta) + H(\chi, \theta) \, \log_{10} V \right] \tag{9}$$

, hence the G-H tables of Schroeder, et al. (1983), and the concept, continued from earlier references, of a power law relation between wind and backscatter, which may or may not be correct. The objective is to find the vector wind, i.e. $\vec{V}$, given pairs of backscatter measurements 90° apart for two slightly different θ's, and one really needs to find a best fit to

$$|v| = V(\chi, \, \theta, \, \sigma^o) \tag{10}$$

if the data have random errors. The problem becomes trivial if (8) or (9) is used but non-trivial if a simple inverse of (7) to yield (10) is not available. The graph of σ^o in db versus $\log_{10} V$ is a straight line for any θ and χ.

If the SASS-1 model function is correct, if a correct way to recover the wind given the backscatter measurements is used and if the function giving C_{DN} as a function of $\bar{U}(10)$ is known, then τ can be found (plus ambiguities or aliases). If C_{DN} is a constant for all winds speeds, equation (9) becomes trivial since it becomes (11) from (1) with a minor correction for the difference between 19.5 and 10 m

$$\sigma^0 db = 10(G(\chi,\theta) + H(\chi,\theta) \log_{10}(u_*/C_{DN}^{\frac{1}{2}}))$$

$$= 10(G(\chi,\theta) - \frac{H(\chi,\theta)}{2} \log_{10}(C_{DN}) + H(\chi,\theta) \log_{10} u_*) \quad (11)$$

and a power law still results.

However, if for the higher wind speeds

$$C_{DN} = \alpha + \beta\bar{U}(10) \quad (12)$$

as in both Smith (1980) and Large and Pond (1981), then

$$u_*^2 = (\alpha + \beta\bar{U}(10))(\bar{U}(10))^2. \quad (13)$$

The value of u_* is then a rather complex function of $\bar{U}(19.5)$ and clearly a power law relation between u_* and σ^0 will not apply.

Conversely if the correct relation between backscatter and u_* was a power law as in

$$\sigma^0 = \alpha(\chi, \theta)(u_*^2)^{\beta(\chi, \theta)/2} \quad (14)$$

then

$$10 \log_{10} \sigma^0 = 10(\log_{10} \alpha(\chi,\theta) + \beta(\chi,\theta) \log_{10} u_* + \beta(\chi, \theta)\log_{10} \bar{U}(10))$$

$$= 10(\log_{10} \alpha(\chi,\theta) + \frac{\beta(\chi,\theta)}{2}(\log_{10}(\alpha + \beta\bar{U}(10))(\bar{U}(10))^2) \quad (15)$$

which is not a straight line on a $\sigma^0 db$ versus $\log \bar{U}(10)$ plot.

Inevitably, then, it is necessary to return to the fundamental questions of the way moving turbulent air roughens the sea surface. These are questions of how the wind varies with height very close to the sea surface, of the relation between $\bar{U}(z)$, u_*, and other parameters and of the basic physical effects involved. The problem is in one sense trivial, given an accepted equation for $C_{DN}(\bar{U}_{10})$ and that SASS-1, or something very similar, is correct. In another sense, it is extremely complex, given that some of the theories and assumptions that have been used may be incorrect and that the study of the generation of waves by the wind, especially capillary waves, is a difficult subject.

THE ROUGHNESS OF SHORT WAVES

Given that scatterometers respond to the roughness of waves with wavelengths in the neighborhood of the incident Bragg wavelengths, the fundamental question posed in the title to this paper separates neatly into two parts:

(a) the relationship between scattering cross-section and the vector wave number spectrum of the Bragg scatterers; and

(b) the parameterization of the spectrum of the Bragg scatterers in terms of characteristics of the wind and of the entire wave spectrum. The traditional approach has been to lump the two parts into a single "model function" relating the measured backscatter to either wind speed or friction velocity. There are so many sources of error in this approach and so few comprehensive sets of surface comparative data that no consensus has emerged regarding the source of backscatter variations. In this section we employ some field observations of frequency spectra of short waves to attempt to throw some light on the second of the points listed above.

A complete answer to part (b), above, requires detailed knowledge of the wave number spectrum of the Bragg scatterers and its dependence on external forcing, dissipation and modulation by longer wave components. This is a great deal more than the experimental data will support, and we must content ourselves here with a look at the integral of the spectrum or the variance for particular bands of components. Thus, we would like to explore the dependence of the spectral density of the gravity-capillary or, at least the very short gravity waves, on appropriate indicators of external forcing.

However, the Doppler shift of observed frequencies of short waves due to longer waves and currents limits the range of short waves for which the observed frequency spectrum yields any useful information about the desired wave number spectrum. For waves at the gravity-capillary range of direct interest, the orbital velocities of the long waves are generally sufficient to reverse their propagation direction, thereby making it difficult to deduce the wave-number spectrum from the observed frequency spectrum. On the other hand, longer waves near the peak of the spectrum cannot be expected to behave in an analogous way to the very short Bragg scatterers, their spectral density being affected by advection and by non-linear effects.

The spectral density of wave components on the rear face of the spectrum, sufficiently far from the peak, is a consequence of a balance between wind input and dissipation. The energy flux from the wind to the wave components is brought about by both normal (pressure) and tangential (shear) stresses. However, both calculations (Brooke Benjamin, (1959) and Miles, 1962)) and experiment (Kendall (1970)) indicate that normal stresses dominate the energy flux. In a recent numerical study, Al-Zanaidi and Hui (1984), using a two-equation closure model for the boundary layer turbulence, have shown that the energy input from the wind, $\partial E/\partial t$, is related to the wind speed and the wave phase speed, C by:

$$\frac{\partial E}{\partial t} = \omega E \frac{\rho_a}{\rho_w} \beta (\frac{U_\lambda}{C} - 1)^2 \qquad (16)$$

where β is a parameter, weakly dependent on the aerodynamic condition of the surface – smooth, transitional or rough, ω in the radian frequency of the wave component receiving wind input and, λ is its wavelength and U_λ is a reference height for the calculation.

The expression (16), supported by laboratory measurements in the wind-wave tank at the Canada Centre for Inland Waters, has the character of wind input induced by form drag. Inasmuch as the sea surface is aerodynamically smooth only at the lowest wind speeds, much of the momentum and energy flux from the wind must necessarily involve correlations between surface slope and pressure brought about by the flow near the surface in relation to the wavelengths of the surface roughness; i.e. the wave components inducing and absorbing the energy flux. Thus, the appropriate wind speed is the speed at a height corresponding to the wavelength (16) or a fraction thereof. Some further ideas of the nature of marine surface drag are given by Stewart (1974).

In an actively wind forced sea the gravity waves suffer dissipation largely through intermittent breaking. Breaking occurs when the local vertical acceleration at the crest exceeds g/2. Since the height of waves of a particular wavelength varies irratically in space and time, breaking is highly intermittent. However, the frequency of breaking events will clearly increase as the energy level in the spectrum increases. Thus the fractional rate of

dissipation per radian of waves at a particular frequency will depend on the spectral energy density $\phi(\omega)$ and the gravitational acceleration, g,

$$\frac{1}{\omega E} \frac{\partial E}{\partial t} = \delta\left(\frac{\omega^5 \phi(\omega)}{g^2}\right)^n \tag{17}$$

where δ in a dimensionless constant and n, an exponent to be determined by experiment.

Of course the choice of a power law for the functional form of (17) is purely a matter of convenience. By suitable choice of n, the dissipation rate can be made more or less sensitive to spectral density.

On the plausible assumption that the short gravity waves strike a balance between direct wind input and dissipation through breaking, (16) and (17) may be equated to yield:

$$\phi(\omega) = \gamma \, \omega^{-5} \, g^2 \left(\frac{\bar{U}\omega}{g} - 1\right)^{2/n} \tag{18}$$

where the dispersion relation for small amplitude gravity waves has been used to replace C.

For many years the rear face of the spectrum was thought to follow an ω^{-5} power law corresponding to the "hard saturation" of Phillips (1958) theory. In (18) this frequency dependence implies $n = \infty$. Now it appears (Forristall (1981), Kahma, (1981), Donelan, et al. (1983), Kitaigorodskii, (1983)) that ω^{-4} provides a better description of the energy containing (and not severely Doppler shifted) waves on the rear face of the spectrum. This corresponds to $n = 2$ or relatively "soft saturation", i.e. the rate of dissipation is not extremely sharply dependent on the spectral density.

After correcting for Doppler shift distortions to the observed spectrum, the normalized spectral density $\phi(\omega) \, \omega^5/g^2$ may be compared with the various candidates taken to be representative of wind forcing. These are: (a) the neutral equivalent wind speed at ship anemometer height $U_{19.5}$, generally favoured by meteorologists and (b) the friction velocity u_*; the hope of most oceanographers, and (c), $((U\omega/g) - 1)$ which is suggested by the rough analysis above.

DATA

The data set consits of 52 observations on Lake Ontario each 20 minutes long in which the waves were sensed by a capacitance wire and sampled at 20 Hz and the wind speed and turbulent components were sensed by a Gill anemometer bivane and sampled at 5 Hz. The average drag coefficient was estimated from the direct Reynolds flux using stresses averaged over 1 hour, or longer, if conditions were steady. The friction velocity was then computed from the 20 minute average wind speed using (1). The observed frequency of $\omega = 17.3$ sec^{-1} was selected because these waves were sufficiently far from the peak to allow the balance of wind input and dissipation and yet long enough ($\lambda = 20.7$ cm) so that the Doppler distortion was tolerable. Each spectral estimate had 1916 degrees of freedom and the frequency of 17.3 sec^{-1} was the energy weighted centroid of the band analysed, based on an ω^{-4} spectrum.

Each spectral estimate was corrected for the effect of the Doppler shift of the longer waves using an ω^{-4} spectral slope and orbital velocities with a Gaussian distribution of standard deviation $\sigma_v = \omega_p \sigma$; where ω_p is the peak frequency and σ^2 is the variance of the surface elevation. An ω^{-4} spectrum with sharp cut-off at ω_p yields $\sigma_v = \sqrt{3} \, \omega_p \sigma_\eta$. However, these generally short-

fetched spectra have most of the energy concentrated near the peak (Donelan, et al. (1983)); furthermore the attenuation of the velocities due to shorter waves and the increased spreading of wave energy of the shorter waves all tend to reduce the contribution of the waves above the peak to the root-mean-square Doppler shifting orbital velocities.

Observations of the bulk Richardson number were used to compute the Monin-Obukhov stability index (Donelan, et al. (1974)) and thus to compute the neutral equivalent wind speed at any height in the manner described by Large and Pond (1981).

RESULTS AND DISCUSSION

The Doppler shift corrected and normalized spectral estimates are compared with $\bar{U}(19.5)$, u_* and $((U(\lambda/2)\omega/g)-1)$ in Figures 1, 2 and 3 respectively. In each case there is some correlation of the spectral estimates with the wind forcing parameter.

The points of Figs. 1 and 2 are clearly stratified with respect to the parameter, U/C_p. Especially in Fig. 2, the spectral estimates corresponding to high U/C_p values are well below the other estimates. The explanation for this is probably that the friction velocity (or stress) receives contributions from the roughness due to the entire wave spectrum, while the spectral estimate reflects only the wind input to that part of the spectrum. For low values of U/C_p, most of the stress is carried by short waves (see for example, Kitaigorodskii and Volkov (1965) and Donelan (1982)) and therefore the spectral estimates on the rear face of the spectrum reflect the stress. On the other hand when U/C_p is large (small nondimensional fetch) a greater portion of the flux of momentum to the waves is absorbed directly by the waves near the peak of the spectrum. Under these conditions, the spectral levels on the rear face are poor indicators of the total stress. The equivalent neutral wind at an anemometer height, (say 19.5 m) is obtained from the usual logarithmic profile equation, $U_z = (u_* \ln(z/z_0))/\kappa$, where both u_* and z_0 are affected by the entire wave spectrum. The above comments regarding u_* consequently apply also to $\bar{U}(19.5)$.

When the spectral estimates are plotted in the context of (18) as in Fig. 3, the stratification U/C_p dissapears although some scatter remains. Much of the scatter is probably due to inexact correction for Doppler shift and other more complex aspects of the modulation of short waves by long waves (Irvine (1983)). Nontheless, the considerable improvement of the correlation in Fig. 3 over Figs. 1 and 2 suggests that, of the three choices explored, that given by (18) follows the data best. The least square regression line shown has been forced through the origin and has a slope of $4.66 \cdot 10^{-3}$, which is therefore the empirical value of γ in (18). A reasonable fit to the straight line in Fig. 3 implies $n = 2$ in (18). The choice of U at $\lambda/2$ is arbitrary and relatively insensitive to height within a wavelength, or so of the surface.

It might be argued that since the spectral estimates are related to u_*, except for high U/C_p, u_* may be inferred from backscatter at these wavelengths over most of the ocean much of the time where U/C_p is generally less than 2.0. Clearly this is unacceptable since during major storms the stress will be surely underestimated. Opposing swells complicate the situation even more.

The waves analysed here are an order of magnitude longer than those of the SASS. The response of 2 cm waves to some wind parameter is the question. Clearly these short waves are affected by other factors. However, the comments
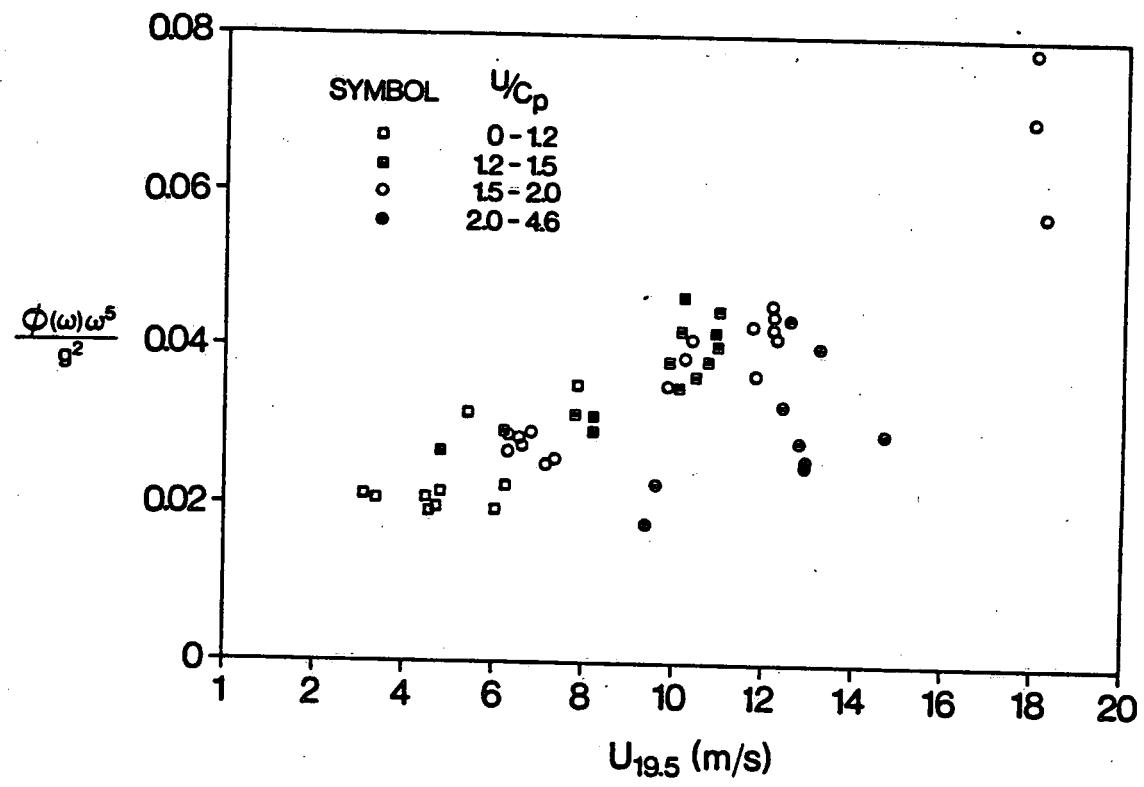
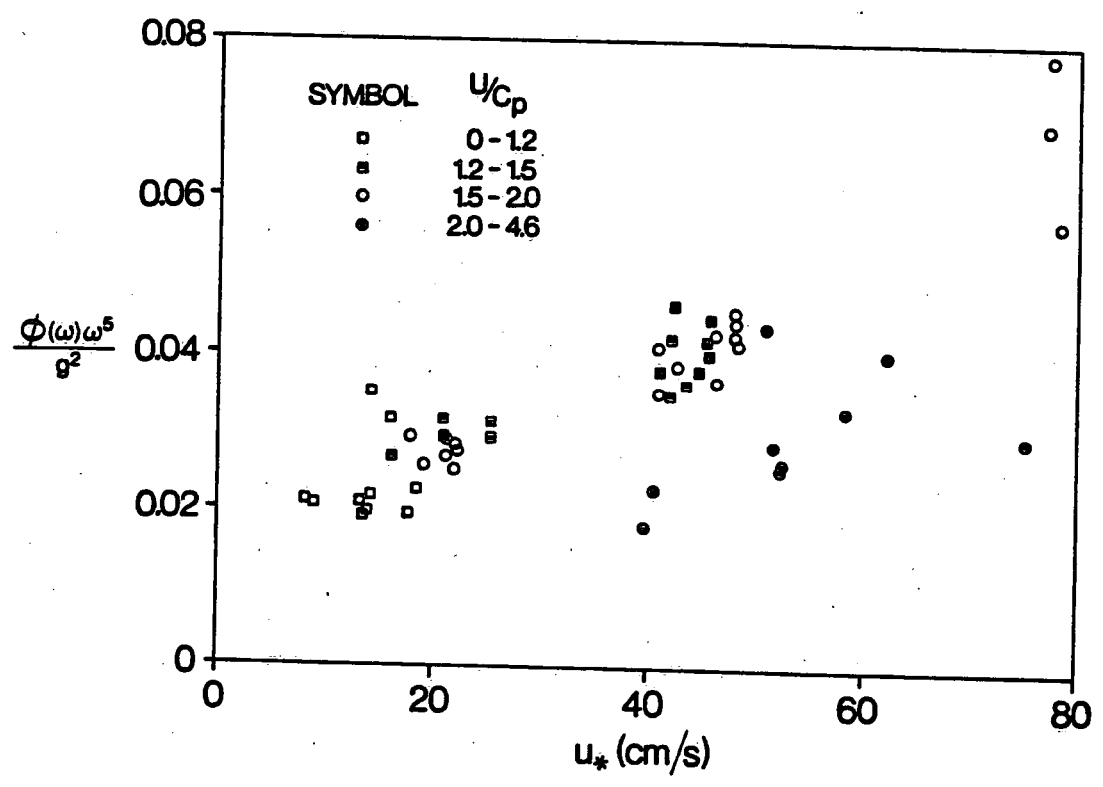FIG. 1 Normalized Spectral Density at $\omega = 17.3$ sec^{-1} Versus $\overline{U}(19.5)$.
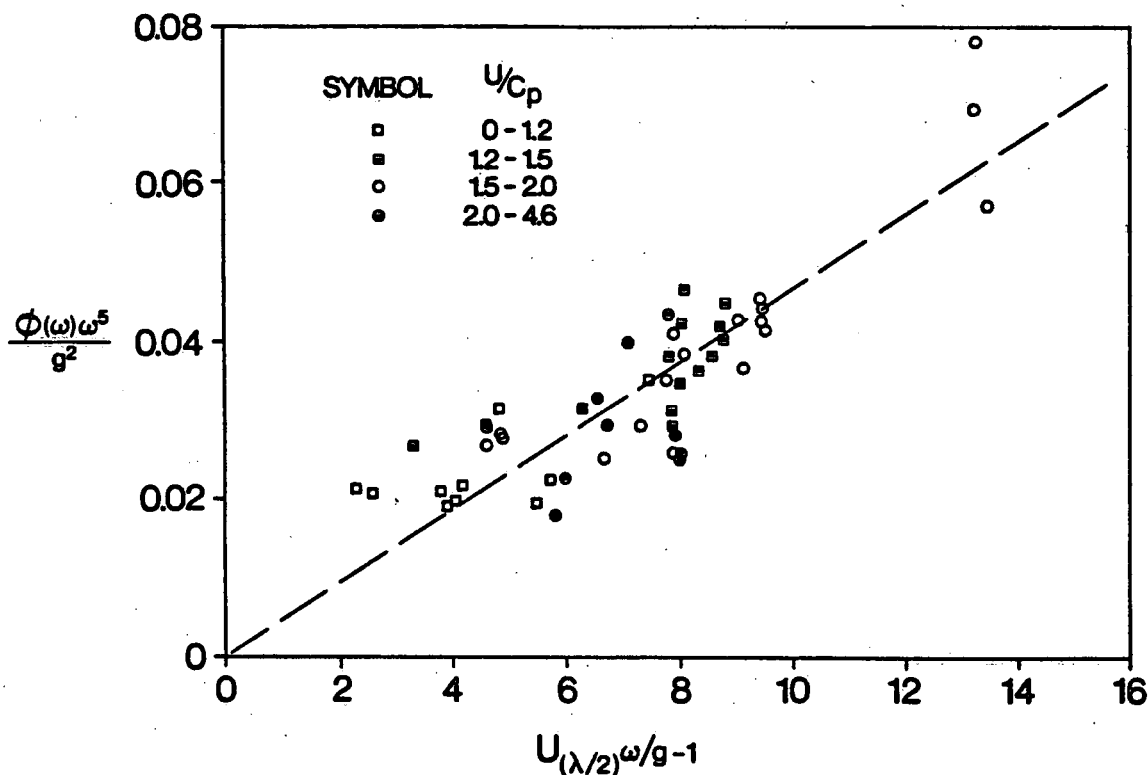

FIG. 2 Normalized Spectral Density Versus Friction Velocity.

FIG 3 Normalized Spectral Density Versus ((Uω/g)=1). (In these three figures, symbols correspond to various values of the ratio of the measured wind speed to the phase speed of the spectral peak.)

made above for longer waves might be expected to apply. Even though the spectral form may change, the relevant wind parameter will still be ((U/C(λ))-1) and not u_*.

CONCLUDING REMARKS

This complex problem has many facets, but we note that the wind profile near the surface is nearly neutral so that U at some height is not too illogical a starting point for improved models. Ultimately, though, it is to be expected that U(z), u_* and z_o will be related to the wave spectrum with perhaps a time lag and perhaps a dependence on high opposing swells and cross seas. Capillary waves for κ_u band will also be affected by surface tension and viscosity as well as possible breaking. That some sort of quasi-equilibrium exists over most of the ocean most of the time is demonstrated by the wind data from the SASS, but the reservations stated herein and the first sentence of the quotation from Brown (1983) must be kept in mind. There is, however, little hope that the tangential stress, u_*, will be more directly related to the waves than the normal stress (i.e. turbulent pressure variations). Direct measurements at sea of the vector wave number spectrum for waves from 20 to 1 cm in length will be needed to clarify these problems as well as more accurate data on the air sea temperature difference.

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