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ABSTRACT

A laboratory experiment is conducted to determine the direct attenuation of waves in an opposing wind. Upwind travelling waves are generated by a submerged wave-maker and a wave-follower borne pressure probe is used to obtain the amplitudes and relative phase of surface pressure and surface elevation. It is shown that the resulting attenuation rates are substantial even when the slope is only 2.5%. Independent verification of these measurements is provided by the spatial changes in amplitude of the waves as they propagate against the wind.

The attenuation rates are proportional to $(U/c-1) \cdot |U/c-1|$ as would be expected from Jeffreys' (1924) theory, and the sheltering coefficient is about 0.11.

RÉSUMÉ

Une expérience en laboratoire sert à déterminer l'atténuation directe des vagues par un vent contraire. Les vagues qui se propagent contre le vent sont produites par un générateur de vagues immergé; une sonde de pression portée par la vague permet de mesurer les amplitudes et les phases relatives de la pression superficielles et du niveau. On a montré que les taux d'atténuation qui en résultent sont importants même quand la pente n'est que de 2.5%. Les variations spatiales de l'amplitude des vagues qui se propagent contre le vent permettent de vérifier ces mesures de façon indépendante.

Les taux d'atténuation sont proportionnels à $(U/c-1)\cdot|U/c-1|$, conformément à la théorie de Jeffrey (1924); le coefficient d'abri est d'environ 0.11.

MANAGEMENT PERSPECTIVE

The art and practice of wave forecasting/hindcasting depends in large measure on the sources and sinks of wave energy, i.e., the input from the wind and wave dissipation. In most models operating today, the wind input is based on a highly idealized mathematical theory which predicts that no direct attenuation occurs when the wind opposes the wave direction. This paper demonstrates that wave attenuation by an opposing wind is as efficacious as wave amplification by a following wind and therefore calls into question the conventional wisdom. The consequences of this have implications in practical wave prediction and in an understanding of the physics of wave generation by wind.

Publication of this information may lead to a recommendation of wave prediction techniques significant for control processes and energy development offshore and in the north.

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PERSPECTIVE DE GESTION

La théorie et la pratique de la prévision et de la prévision à posteriori des vagues dépendent dans une grande mesure des sources et des puits d'énergie de vagues, c'est-à-dire l'apport du vent et la dissipation des vagues. Dans la plupart des modèles d'aujourd'hui, l'apport du vent est calculé à partir d'une théorie mathématique hautement idéalisée qui postule qu'aucune atténuation directe ne se produit lorsque le vent va en sens contraire des vagues. Le présent article remet cette opinion répandue en question en démontrant que l'atténuation des vagues par un vent contraire est aussi efficace que l'amplification des vagues par un vent d'arrière et a donc des implications sur la prévision des vagues et sur la compréhension de la physique de la génération des vagues par le vent.

La publication de cette information pourrait conduire à des recommandations concernant les techniques de prévision des vagues, importantes pour les processus de contrôle et pour l'exploitation du potentiel énergétique au large et dans l'Artique.

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1.0 INTRODUCTION

Following the publication of Ursell's (1956) critical review of our understanding of the physics of wave generation by wind a great deal of experimental and theoretical effort has been devoted to the problem. Three aspects of the process of wave generation by wind may be identified and each of these is believed to be dominant at some stage in the excitation or growth of water waves. Two of these have to do with the initiation of waves where none exist and the third addresses the question of the amplification of existing waves. Phillips (1957) demonstrated that a flat water surface may be excited by "intrinsic" turbulent pressure fluctuations in the wind. The excitation is preferred for wavenumbers that obey a resonance condition and the resulting growth rate is linear in time.

Valenzuela (1976) and Kawai (1979) have described the initiation of capillary-gravity wavelets due to instabilities of the coupled laminar shear flow on both sides of the air-water interface. The growth of small disturbances is exponential and the theory appears to be in reasonable accord with laboratory experiments on the growth of wavelets immediately following the turning on of the wind.

These mechanisms are interesting of themselves and of practical importance because they describe the genesis of the process of wave generation by wind. However, most of the transfer of energy and momentum to waves from the wind occurs in the amplification of existing waves by a following wind. Hence from a practical point of view it is the most significant aspect of wave generation by wind. We are concerned here with the reverse of this process - the attenuation of existing waves by an opposing wind. Apart from its own usefulness in modelling waves for predictive purposes, the question of how (and

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indeed, if) waves are attenuated directly by wind is likely to throw some light on the, certainly more significant, problem of wave amplification by wind.

2.0 SOME THEORETICAL NOTES

Inasmuch as surface water waves are nearly irrotational, the energy flux from wind to waves is brought about mainly by surface pressure fluctuations in phase with the wave slope. Longuet-Higgins (1969) has shown that the existence of a thin boundary layer beneath the water surface allows the wave induced tangential stresses to play a part in the wave generation process. Essentially the action of a tangential stress causes differential thickening of the boundary layer thereby introducing an additional pressure distribution which, if the induced stress is in phase with the surface elevation, acts to amplify existing waves in a following wind and to attenuate them in an opposing wind.

Rough calculations of the energy flux due to tangential stress variations indicate that they are one-to-two orders of magnitude smaller than the observed flux due to normal stresses and, indeed, laboratory measurements by Kendall (1970) and Hsu et al. (1981) are in accord with this estimate.

Evidently most of the energy flux between wind and waves is brought about by the component of surface pressure in phase with the wave slope, which for a component of frequency, ω is:

$$\frac{\partial E(\omega)}{\partial t} = \overline{p_o \frac{\partial \eta}{\partial x}} \cdot c(\omega) \qquad (1)$$

$$E = \rho_w g \overline{\eta^2}$$

$$\eta_{(x,t)} = a \cos (kx - \omega t)$$

where E is the wave energy density, $\eta(x,t)$ the surface elevation, p_o the pressure exerted by the air on the surface, c the phase speed, k the wavenumber and ρ_w the water density. The overbar denotes an average over one wavelength.

In the definition sketch of Figure 1 the wave component $\eta(x,t)$ progresses to the right with favourable wind in the top part of the figure and with adverse wind in the bottom.

The simplest description of flow near a water surface driven by an air stream is one in which the flow is irrotational above and below the interface and the air velocity, averaged over one wavelength, is constant with height. In such a "potential flow" (Lamb, 1932) the pressure follows from Bernoulli's equation and is in exact anti-phase with the surface elevation so that (1) is identically zero. In order to do work on the wave the pressure must be shifted relative to the potential solution. The harmonic of the wave-induced surface pressure of wavenumber k is shown in Figure 1. At the top of the figure the pressure is shown shifted by ϕ_g so that there is some component in phase with the wave slope and consequently wave growth. Whereas at the bottom of the figure the pressure is shifted the other way (ϕ_a) corresponding to wave attenuation.

The central question, which has occupied a great many researchers over the past sixty years, is by what mechanism can such pressure shifts occur? Phillips (1977) has summarized the major theoretical and experimental attempts to answer the question. It is sufficient to point out here that there are two theories which are still in contention and whose relative merits may be tested experimentally by measuring the righthand side of (1) for waves progressing against the wind.

The first theory was advanced by Jeffreys (1924, 1925). The idea is that the downwind slope of the wave is to some degree sheltered from the wind so that the pressure is reduced there and increased on the exposed upwind face. The resulting pressure difference causes a form drag, and the corresponding growth rate of the waves is given by:

$$\frac{\partial E}{\partial t} = \rho_a s \, (U - c)^2 \, \overline{\left(\frac{\partial \eta}{\partial x}\right)^2} \cdot c \qquad (2)$$

where ρ_a is the air density and s the form drag or "sheltering" coefficient. (Jeffreys originally called it the "exposure" coefficient). Written to reflect the change in sign of the energy flux as the direction of U is reversed relative to c (Figure 1) (2) becomes:

$$\frac{\partial E}{\partial t} = \rho_a s \, \overline{\eta^2} \, k^2 \, c^3 \left|\frac{U}{c} - 1\right| \cdot \left(\frac{U}{c} - 1\right) \qquad (3)$$

It is customary to discuss the amplification of waves in the non-dimensional form of the fractional energy increase per radian, ζ

$$\zeta = \frac{1}{\omega E} \frac{\partial E}{\partial t} = s \frac{\rho_a}{\rho_w} \left| \frac{U}{c} - 1 \right| \cdot \left(\frac{U}{c} - 1\right) \qquad (4)$$

The second theory, due to Miles (1957, 1959a, 1959b, 1962), requires no hypothesis regarding the sheltering effect of steep waves and is a correctly argued mathematical theory. However, it is rather idealized being both linear, in the sense that the feedback to any wave component is independent of all others, and quasi-laminar, in that the effects of atmospheric turbulence are considered only in the establishment of an appropriate boundary layer mean velocity profile. It may be that Miles' theory is relevant to wave amplification only under quite restrictive conditons. In this theory the existence of a pressure shift required for wave amplification hinges on the behaviour of the flow in the vicinity of the "matched" layer where the wind speed and wave phase speed are equal. In Figure 2 the atmospheric boundary layer profiles are shown in a reference frame at rest with respect to the wave form. It is clear that a matched layer can be said to exist in a favourable wind but not in an adverse wind.

The streamline pattern around the matched layer is closed and leads to the formation of a vortex force (Lighthill, 1962) which, acting on the mean velocity field, induces energy flux to the wave induced motion provided that the ratio of curvature to gradient of the velocity profile is negative at the matched layer. A logarithmic velocity profile satisfies this condition and (Miles, 1959) the resulting fractional energy increase per radian is given by:

$$\zeta = \frac{\rho_a}{\rho_w} \left(\frac{u_*}{\kappa c}\right)^2 \cdot \beta \qquad (5)$$

where κ is the von Kármán constant, u_* the friction velocity ($\rho_a u_*^2$ is the tangential wall stress) and β is related to the profile curvature and is about 3 for strongly forced waves and drops to zero as the wave speed approaches the wind speed. The reduction in β as the waves develop comes about because the matched layer rises and, correspondingly, the curvature-to-gradient ratio and the wave-induced fluctuations diminish rapidly. On the other hand, if the waves are too young and the boundary layer not fully rough the matched layer could descend to the viscous sub-layer where the profile curvature vanishes and with it β. This mechanism is therefore likely to be effective only in the range $1.5 < U/c < 4$. Indeed, the well-known Bight of Abaco experiments (Snyder et al. 1981) show good agreement with Miles' calculations for two of his sample profiles and values of U/c in this range. However, when Riley et al. (1982) repeated the calculations for profiles appropriate to the conditions of the Bight of Abaco experiment the theory was able to account for only 40% of the observed growth rates.

In an ingenious experiment Kendall (1970) explored the flow over and the pressure on a rubber wall on which waves could be made to progress upwind or downwind. Although he was able to verify some of the consequences of Miles' theory - in particular the amplitude of the wave-induced wall pressure - the phase shifts and consequent growth rates were more than twice as large as those predicted by Miles' theory. More significantly, Kendall found that the phase shift was non-zero for stationary waves and waves moving upwind for which there is no matched layer and hence no momentum flux associated with Miles' theory. It would appear that Miles' theory is unable to account fully for the observed growth rates in a favourable wind and, of course, does not address the problem of wave attenuation in an adverse wind.

Kendall's experiments, although very revealing, do not provide reliable attenuation rates for water waves in an adverse wind. Among the considerable differences, carefully listed by Kendall (1970), between his simulated 'waves' and oceanic swell, the large steepness ($ak \approx 0.2$) used may be the weakest link in applying his results to the swell attenuation problem.

Our purpose here is to examine the attenuation of water waves in an adverse wind and to explore the effect of steepness on the normalized growth rates.

3.0 THE EXPERIMENT

The wind-wave flume and relevant instruments are illustrated schematically in Figure 3. In order to generate waves progressing against the wind, a wave-maker was installed at the downwind end of the tank. The wave-maker was submerged, standing 1.04 m high in 1.20 m of water, so that wind-generated waves would pass over it and lose their energy on the beach behind.

The paddle-generated waves, of course, propagated in both directions and were eventually absorbed by one or other beach. The wind-tunnel was closed and the air flow could be driven at speeds up to 16 m/s. Capacitance wave staffs at eight stations were used to estimate the attenuation of upwind propagating waves for comparison with the surface pressure measurements obtained from an electro-hydraulic wave-follower at the location shown.

The pressure probe was of the 'disk' type designed by Elliott (1972) and differs from his design "F" only in that it incorporates an additional port on the perimeter of the probe pointing upwind. This yields a measure of the dynamic pressure and enables a correction to be made for the small sensitivity of the actual pressure measurement to ambient flow speed. Pressure fluctuations were converted to electrical signals by an MKS Baratron (Model 223AH) differential pressure transducer. The probe was connected directly to the positive port of the transducer and through a pneumatic low pass filter to the reference port and the complete system mounted on the wave-follower with the axis of symmetry of the disk horizontal and transverse to the flow. A pair of capacitance wave staffs 10 cm on either side of the disk provided concurrent surface elevation and also yielded some directional information and estimates of the long-crestedness of the waves under the disk.

Estimates of the righthand side of (1) are particularly sensitive to the phase between waves and surface pressure. Therefore the phase distortion introduced by the probe was measured in the manner described by Snyder et al. (1974) and is illustrated in Figure 4. We are concerned here primarily with waves of frequency 0.527 Hz, for which a 1° phase correction is necessary, and peripherally with waves of 1.054 Hz, for which a larger (7°) phase correction will be made.

4.0 RESULTS AND DISCUSSION

Pressure measurements were made under a wide variety of conditions using monochromatic or random paddle excitation at various frequencies, amplitudes, wind speeds and heights of the pressure probe above the surface. A small subset of these has been analyzed to-date and forms the basis for this paper. Several additional aspects of the problem will be addressed following the analysis of many more runs, but the four runs discussed herein appear to be sufficient to demonstrate that moderate laboratory "swell" is strongly attenuated by an adverse wind.

The overall conditions of the runs are summarized in Table 1. For all these runs the paddle was excited with a sinusoid of frequency

0.527 Hz. The observed spectrum of surface elevation, $S_{\eta\eta}$ at the wave-follower station for run 82 is shown in Figure 5. The two narrow lower frequency peaks are paddle-generated and progress against the wind, while the broader higher frequency peak is wind generated. Phase differences between a pair of wave staffs separated by 10 cm in the x direction demonstrate that most of the energy near 1.0 Hz is free and evidently a harmonic of the excitation frequency. Henceforth the paddle-generated waves are termed "swell" for brevity. The highest wind speeds (runs 80 and 83) cause the wind sea spectrum to merge with the higher frequency component of the swell and so only the fundamental (0.527 Hz) swell is considered in these two runs.

The corresponding pressure spectrum, S_{pp} measured 5 cm above the moving surface is shown in Figure 6. Pressure fluctuations are related to the wave slope and to the square of the difference between wind and phase velocities. The 1.05 Hz swell is steeper than the fundamental, so that the difference in spectral levels is largely due to the larger velocity difference of the wind with the 0.527 Hz swell compared to that with the 1.05 Hz swell. The pressure spectrum corresponding to the wind sea is relatively low, partly because the observing height is a significant fraction of these wavelengths but largely because, wave and wind being in the same direction, the velocity difference is much lower.

Coherence and phase for run 82 are illustrated in Figures 7 and 8. The phase corrections (Figure 4) have not been included. The arrows indicate the swell components and the forward face of the spectrum of the wind "sea". Positive phase angles correspond to pressure leading surface elevation by less than 180° and both swell components show differences of about 20° from 180°. (The 7° phase correction to the 1.05 Hz component reduces its phase difference to about 25°.) These phase shifts correspond to direct attenuation of the swell components (Figure 1), whereas the opposite shift of the pressure over the wind sea corresponds to wave amplification. It is worth noting that the phase corrections (Figure 4) would considerably increase the phase shifts and corresponding growth rates for the wind sea, but since we are concerned here with swell attenuation no attempt has been made to apply these corrections.

Since the pressure fluctuations decay exponentially with height without change of phase (Snyder et al. 1981) the fractional energy exchange per radian is given by:

$$\zeta(\omega) = \frac{[Qu(\omega)]_{p\eta}\, e^{kz}}{\rho_w\, g\, S_{\eta\eta}(\omega)} \tag{6}$$

where $[Qu(\omega)]_{p\eta}$ is the quadrature spectrum between pressure, p and surface elevation, η measured at height z. The exponential correction e^{kz} amounts to 10% for the low frequency swell and 25% for the shorter wavelength component. The values of ζ are listed in Table 1 and all are negative corresponding to attenuation of the swell through the direct action of pressure on the surface.

An independent check of these measurements may be made using the changes in wave energy with distance from the board. An example is shown in Figure 9 using the 0.527 Hz swell in a strong adverse wind (run 83). As described in Mitsuyasu and Honda (1982), the slope of the regression line, α yields the fractional energy decrease per radian:

$$\zeta = \alpha\, C_g/\omega \tag{7}$$

where the group velocity, C_g from linear theory is 1.74 m/s. Since run 83 reflects the decrease due to other causes in addition to the wind, we compare the rate of decrease for this run with another (run 107) in which there was no wind and the paddle was driven with the same amplitude and frequency as in run 83. The difference in slope $1.04 \times 10^{-3}\ m^{-1}$ yields a fractional energy decrease per radian of

5.45×10^{-3} which is in reasonable agreement with the direct (pressure-slope) estimate of 6.45×10^{-3}.

Finally in Figure 10 the values of ζ for the principal swell component are graphed against $(U_{10}/c -1)^2$ as suggested by (4). The amplification rates in a favourable wind recently obtained by Hsiao and Shemdin (1983) in a field experiment are indicated by their fitted line (dashed). As pointed out by Hsiao and Shemdin, the growth rates obtained by Snyder et al. (1981), in a much narrower range of U/c, are consistent with their results. It is indeed striking that, when plotted in the form suggested by Jeffreys' theory, amplification and attenuation rates are quite comparable. The sheltering coefficients are listed in Table 1 as are the slopes of the swell components. From this limited subset of the data it is difficult to draw firm conclusions about the effect of slope on the sheltering coefficient. Suffice it to say that for quite gentle waves ($ak \approx 0.025$ corresponding, for example, to 10 s, 156 m swell of 62 cm amplitude) the attenuation rate is substantial and the sheltering coefficient is nearly constant, thereby providing substantial support to Jeffreys' theory.

The attenuation rates for the higher frequency swell are listed in Table 1 but are not included in Figure 10. If, as suggested by Al-Zanaidi and Hui (1984), the appropriate wind speed for comparison of different wavelengths is the wind speed at a height related to the wavelength, then the sheltering coefficients for the 1.05 Hz swell approach more closely those of the fundamental component. However, the attenuation of the small secondary swell may be affected by the disturbance caused by the much larger principal component and it is expected that further analysis will be revealing in this regard.

Using a well-tested two-equation (closure) model for the boundary layer turbulence, Al-Zanaidi and Hui (1984) have found that their numerical model for flow over small amplitude water waves yields comparable growth and attenuation rates but the equivalent sheltering coefficients for attenuation, although also nearly constant, are a factor of three smaller than those obtained here.

In general, boundary layer models are unable to deal with separation and it is tempting to suggest that the large observed energy exchange rates between wind and waves owe their origins to some form of flow separation. Measurements of mean surface stress (e.g., Kendall, 1970) indicate that separation does not occur although, as Kendall pointed out, the flow is more likely to separate when moving against the waves than with them. In fact the likelihood of separation increases with the ratio of pressure perturbation to average tangential stress or roughly as $ak(1 - c/U)^2$ over a single sinusoid. It is entirely possible that separation occurs intermittently leaving the average tangential stress everywhere positive but greatly enhancing the momentum and energy fluxes. Banner and Melville (1976) have demonstrated that separation greatly increases (almost fifty times) the momentum transfer and argue that air flow separation and wave breaking are intimately connected.

Two other experiments designed to measure the normal pressure over waves in an adverse wind are known to us. Mizuno (1976) reported wave attenuation directly through pressure-slope correlations and, although his results are rather scattered, the attenuation rates are comparable to those reported here. Young and Sobey (1984), on the other hand, find that their pressure measurements "closely follow the predictions of potential flow theory, with the pressure in anti-phase with the water surface".

The debate continues but it is becoming increasingly clear that Miles' theory is insufficient to account for the observed growth or attenuation rates, and some other mechanism must also come into play. Perhaps Jeffreys was, after all, very nearly correct.

REFERENCES

Al-Zanaidi, M. and W.H. Hui. 1984. Turbulent Flow Over Water Waves - a Numerical Study. J. Fluid Mech. 148, pp. 225-246.

Banner, M.L. and W.K. Melville. 1976. On the Separation of Air Flow Over Water Waves. J. Fluid Mech. 77, pp. 825-842.

Elliott, J.A. 1972. Instrumentation for Measuring Static Pressure Fluctuations within the Atmospheric Boundary Layer. Boundary-Layer Meteorol. 2, pp. 476-495.

Hsiao, S.V. and O.H. Shemdin. 1983. Measurements of Wind Velocity and Pressure with a Wave Follower During MARSEN. J. Geophys. Res., 88, No. C14, pp. 9841-9849.

Hsu, C-T., E.Y. Hsu and R.L. Street. 1981. On the Structure of Turbulent Flow over a Progressive Water Wave: Theory and Experiment in the Transformed, Wave-Following Co-ordinate System. J. Fluid Mech. 105, pp. 87-117.

Jeffreys, H. 1924. On the Formation of Waves by Wind. Proc. Roy. Soc. A, 107, pp. 189-206.

Jeffreys, H. 1925. On the Formation of Waves by Wind. II. Proc. Roy. Soc. A, 110, pp. 341-347.

Kawai, S. 1979. Generation of Initial Wavelets by Instability of a Coupled Shear Flow and their Evolution to Wind Waves. J. Fluid Mech. 93, pp. 661-703.

Kendall, J.M. 1970. The Turbulent Boundary Layer over a Wall with Progressive Surface Waves. J. Fluid Mech. 41, pp. 259-281.

Lamb, H. 1932. Hydrodynamics, 6th ed. Cambridge University Press.

Lighthill, M.J. 1962. Physical Interpretation of the Mathematical Theory of Wave Generation by Wind. J. Fluid Mech. 14, pp. 385-398.

Longuet-Higgins, M.S. 1969. Action of a Variable Stress at the Surface of Water Waves. Phys. Fluids, 12, pp. 737-740.

Miles, J.W. 1957. On the Generation of Surface Waves by Shear Flows. J. Fluid Mech. 3, pp. 185-204.

Miles, J.W. 1959a. On the Generation of Surface Waves by Shear Flows. Part 2. J. Fluid Mech. 6, pp. 568-582.

Miles, J.W. 1959b. On the Generation of Surface Waves by Shear Flows. Part 3. J. Fluid Mech. 6, pp. 583-598.

Miles, J.W. 1962. On the Generation of Surface Waves by Shear Flows, Part 4. J. Fluid Mech. 13, pp. 433-448.

Mitsuyasu, H. and T. Honda. 1982. Wind-Induced Growth of Water Waves. J. Fluid Mech. 123, pp. 425-442.

Mizuno, S. 1976. Pressure Measurements Above Mechanically Generated Water Waves (I). Reports Res. Inst. Appl. Mechanics, Kyushu University, 23, No. 75, pp. 113-129.

Phillips, O.M. 1957. On the Generation of Waves by Turbulent Wind. J. Fluid Mech. 2, pp. 417-445.

Phillips, O.M. 1977. The Dynamics of the Upper Ocean, 2nd ed. Cambridge University Press.

Riley, D.S., M.A. Donelan and W.H. Hui. 1982. An Extended Miles' Theory for Wave Generation by Wind. Boundary-Layer Meteorol. 22, pp. 209-225.

Snyder, R.L., F.W. Dobson, J.A. Elliott and R.B. Long. 1981. Array Measurements of Atmospheric Pressure Fluctuations Above Surface Gravity Waves. J. Fluid Mech. 102, pp. 1-59.

Snyder, R.L., R.B. Long, J. Irish, D.G. Hunley and N.C. Pflaum. 1974. An Instrument to Measure Atmospheric Pressure Fluctuations Above Surace Gravity Waves. J. Mar. Res. 32, pp. 485-496.

Ursell, F. 1956. Wave Generation by Wind. Surveys in Mechanics (ed. G.K. Batchelor), pp. 216-249. Cambridge University Press.

Valenzuela, G.R. 1976. The Growh of Gravity-Capillary Waves in a Coupled Shear Flow. J. Fluid Mech. 76, pp. 229-250.

Young, I.R. and R.J. Sobey. 1984. Measurements of the Wind-Wave Energy Flux in an Opposing Wind. J. Fluid Mech. (in press).

TABLE 1. **Summary of Runs**

Run No.	Wind Speed at z = 15 cm (m/s)	Paddle Freq. (Hz)	Swell Freq. (Hz)	Height of Pressure Probe (cm)	Swell Slope ak	ζ x10^3	S
80	9.5	0.527	0.527	8	0.025	-6.48	0.113
81	3.3	0.527	0.527	4	0.051	-1.30	0.110
82	6.6	0.527	0.527	5	0.050	-3.38	0.102
83	9.3	0.527	0.527	9	0.044	-6.45	0.117
81	3.3	0.527	1.054	4	0.058	-2.51	0.089
82	6.6	0.527	1.054	5	0.039	-5.64	0.063

FIGURE CAPTIONS

1. Illustrative sketch showing the relationship between pressure and surface elevation components for wave amplification (top) or attenuation (bottom).

2. Illustrative sketch showing wind profiles over waves in a frame of reference fixed with respect to the wave form. At the top the waves are in a favourable wind; at the bottom in an adverse wind. The region of sign change in the velocity is termed the "matched" layer.

3. Sketch of the wind-wave flume showing the locations of the wave-maker and wave staffs. The asterisk indicates the wave-follower borne pressure probe. Vertical exaggeration is x5.

4. Phase lag introduced by the pressure probe-transducer system. The arrows show the "swell" frequencies analyzed in this paper.

5. The spectrum of surface elevation η for run 82.

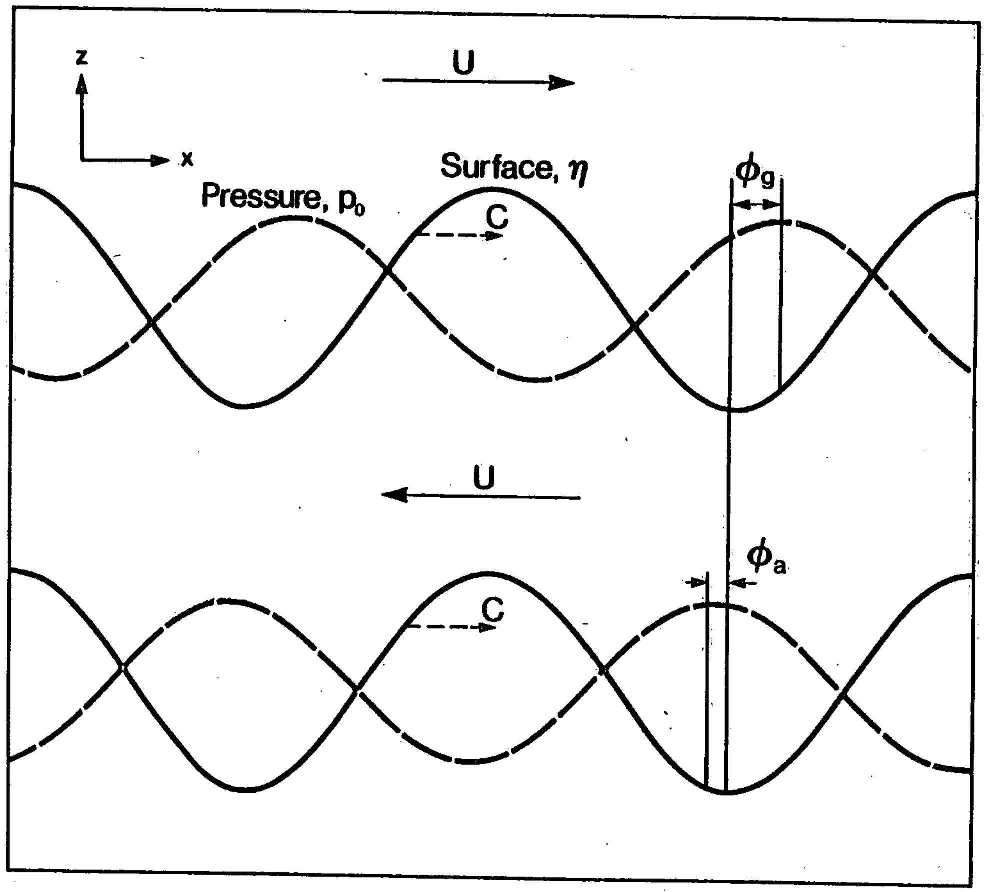
6. The spectrum of pressure p for run 82 measured from the wave-follower 5 cm above the surface.

7. Coherence between pressure and surface elevation for run 82.

8. Measured phase between pressure and surface elevation for run 82. Positive values means η lags p. At this stage the phases have not been corrected for the instrument lag shown in Figure 4.

9. Changes in the spectral density of the 0.527 Hz "swell" as it progresses. (-O- --) run 107 with no wind; (——●——) run 83 with strong adverse wind.

10. The magnitude of the fractional energy change per radian. Circles, with fitted solid line, are attenuation data from Table 1, i.e., ζ is negative. ($\bullet$) ak ≈ 5%; (o) ak ≈ 2.5%. The dashed line is the result of Hsiao and Shemdin (1983) for wave growth, i.e., ζ is positive.



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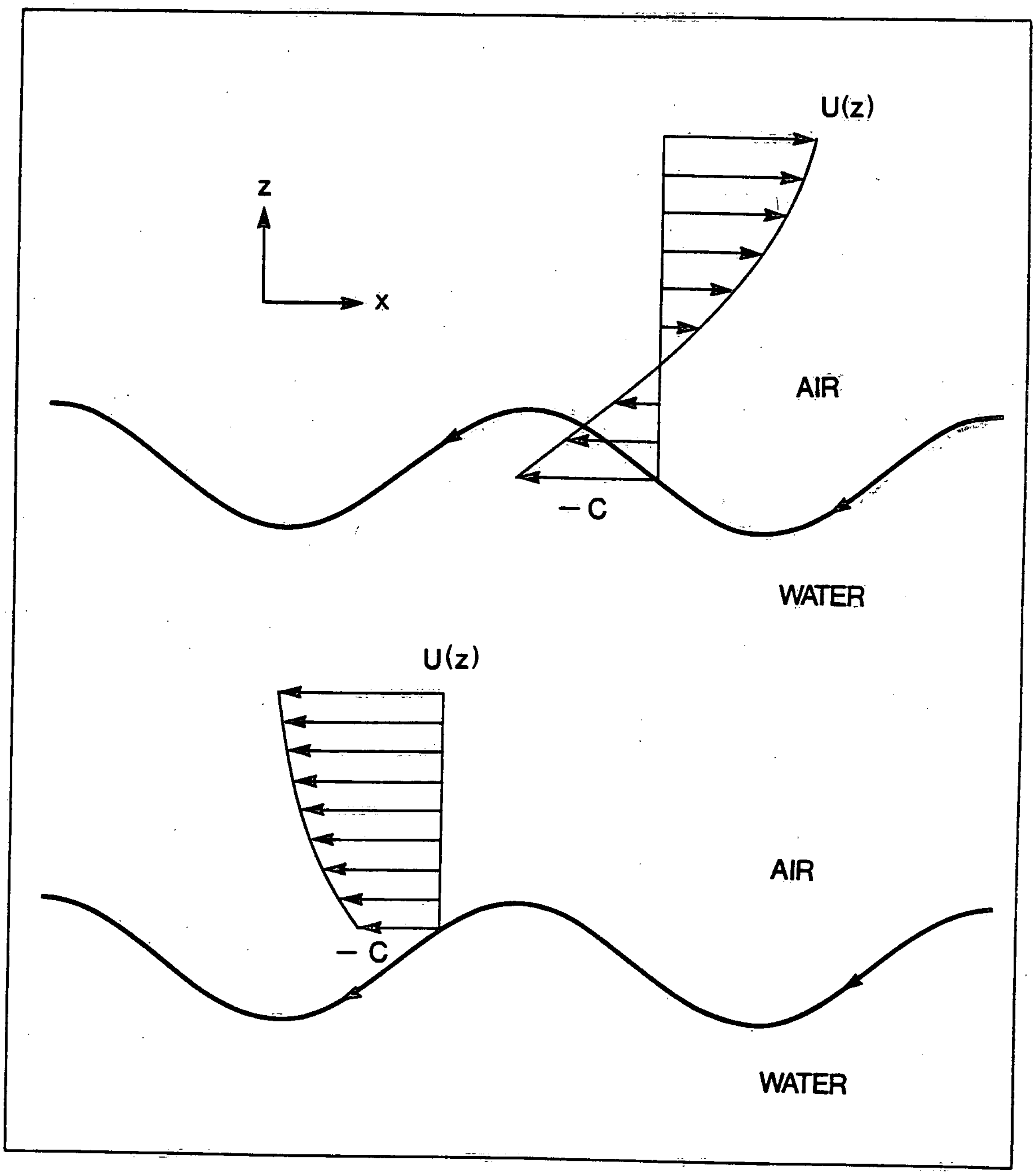
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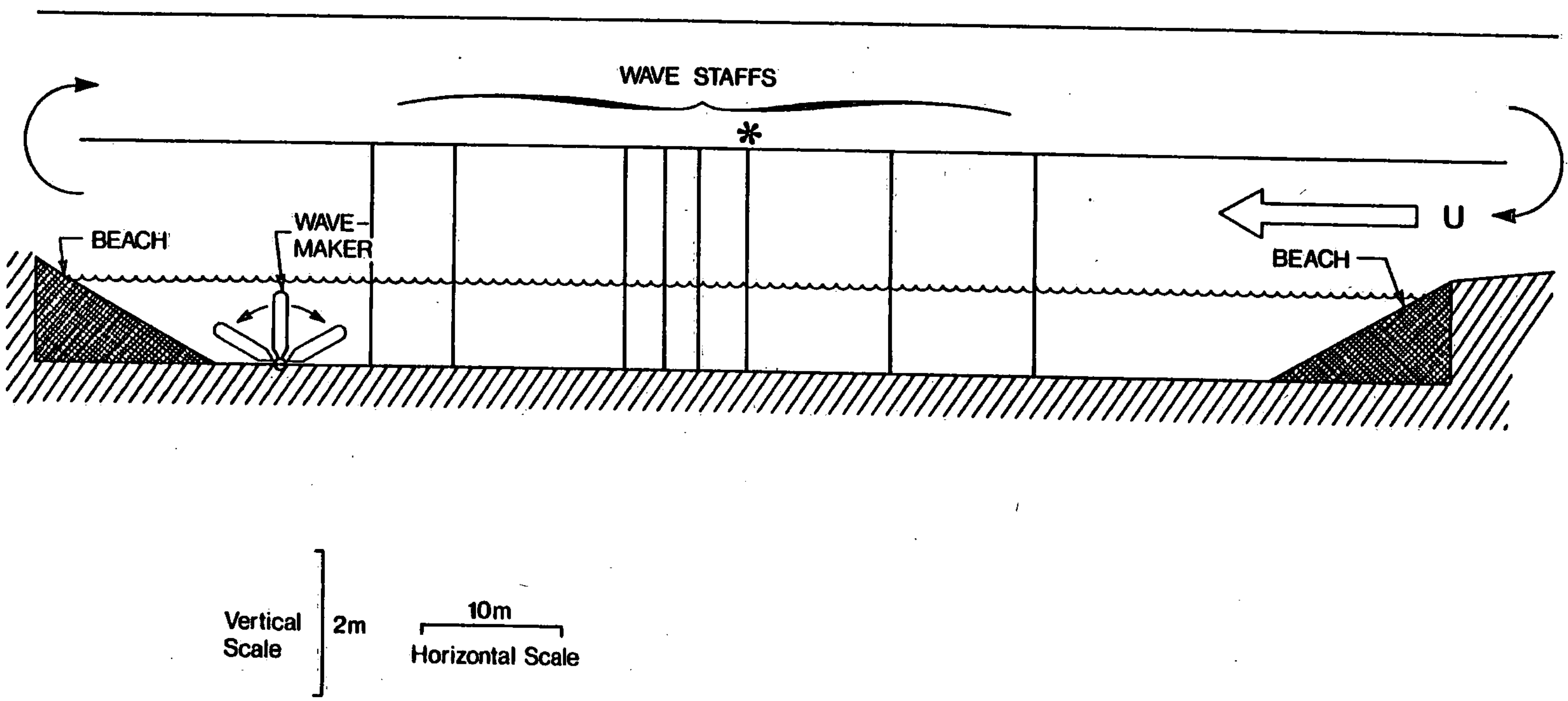

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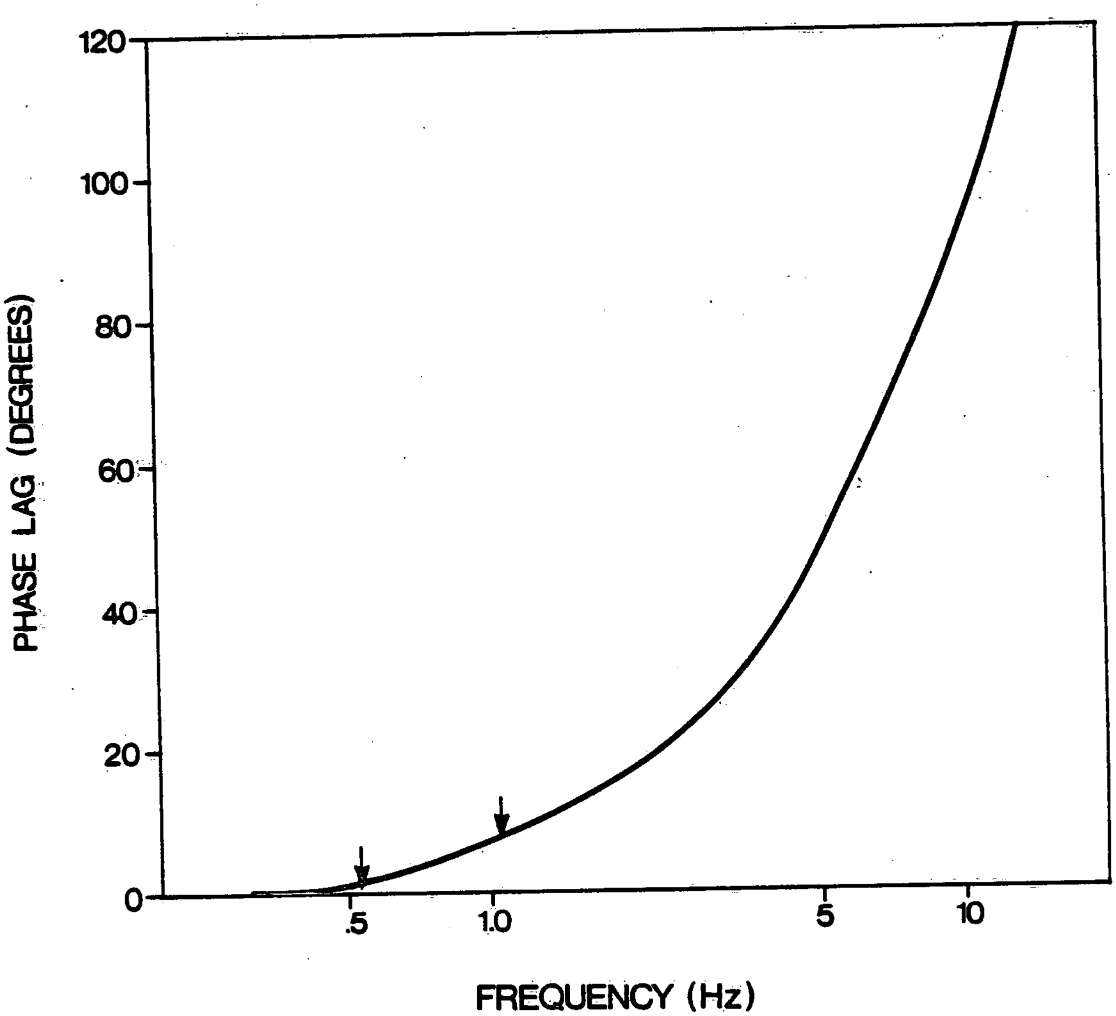

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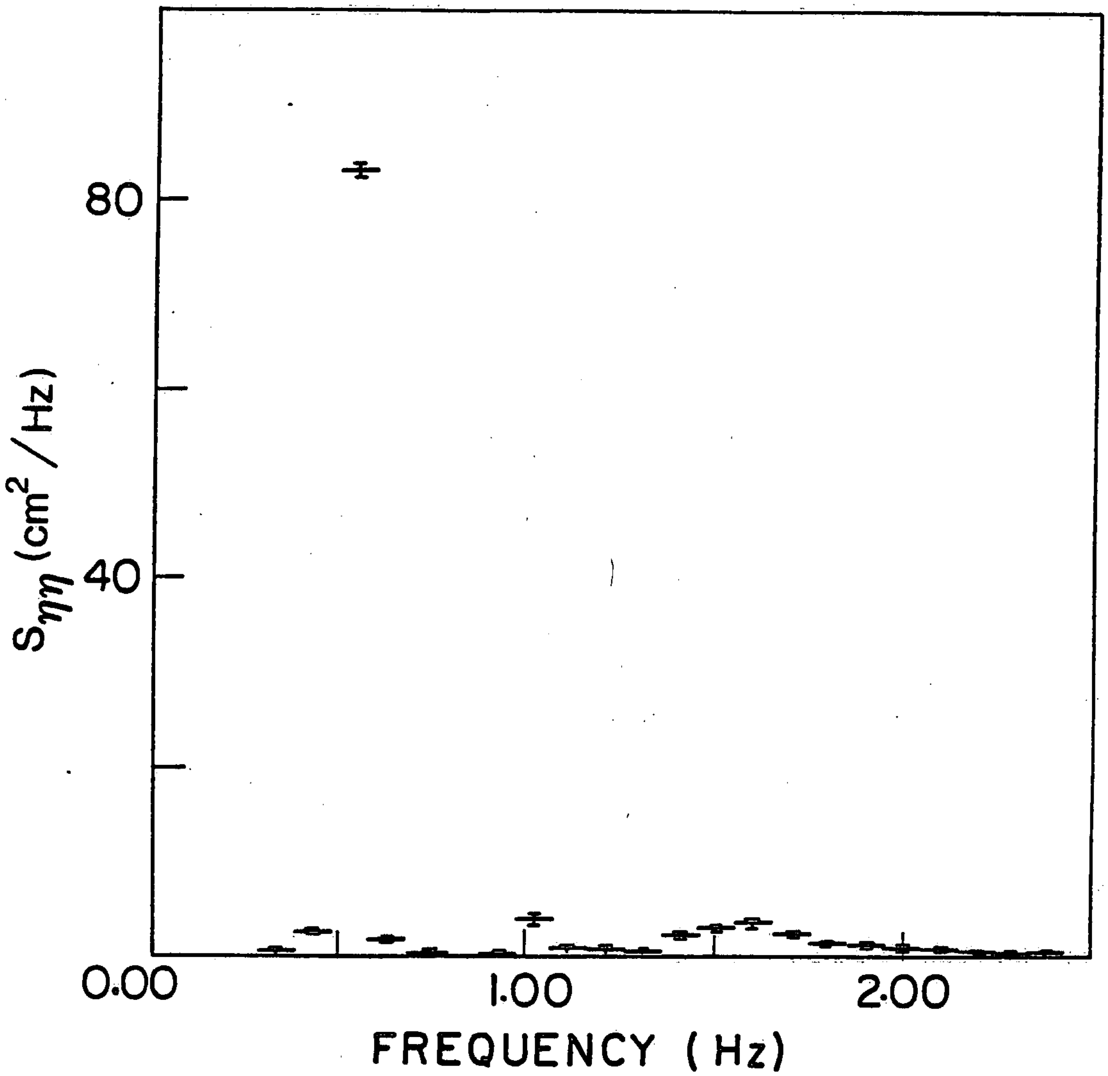

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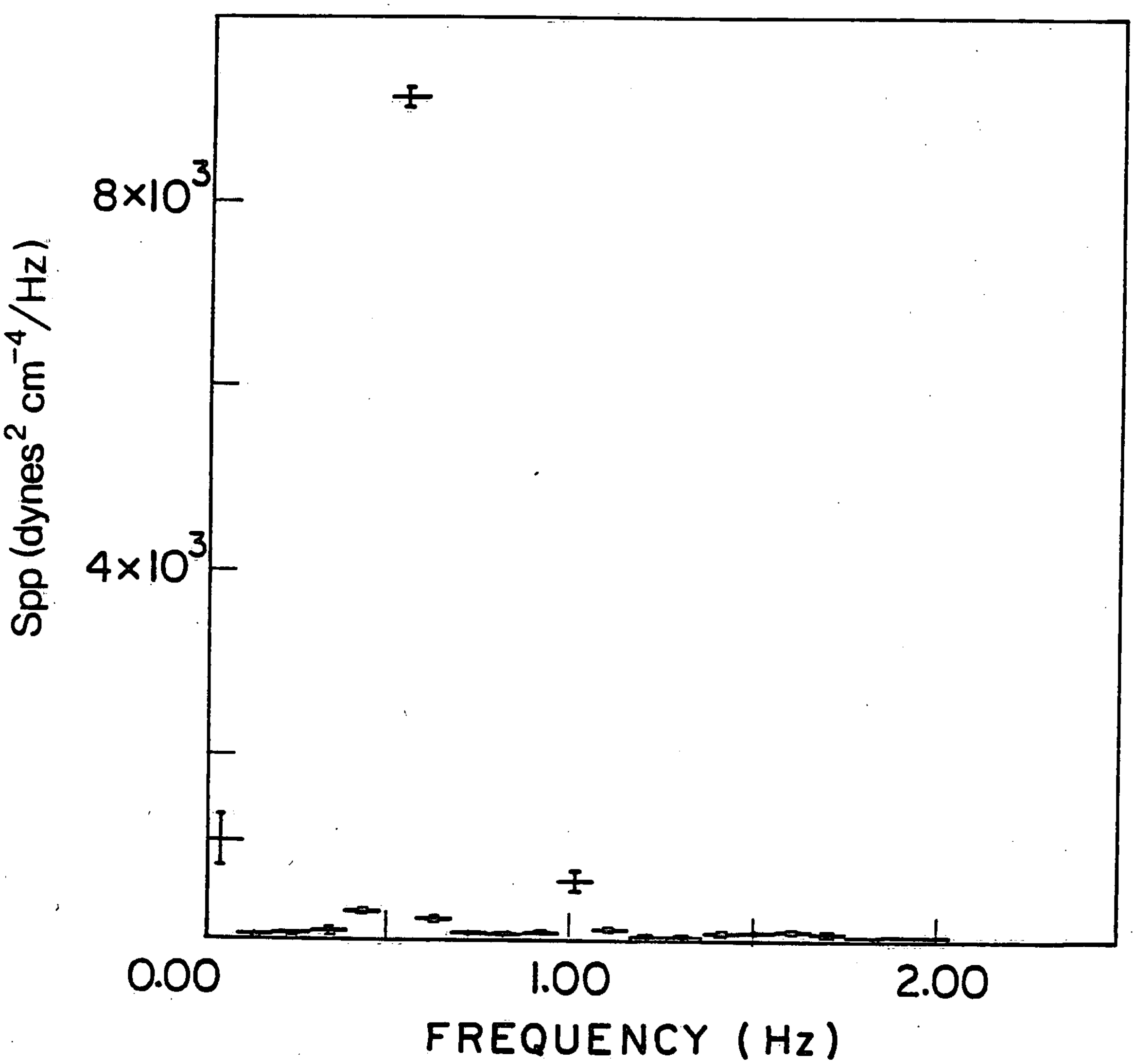
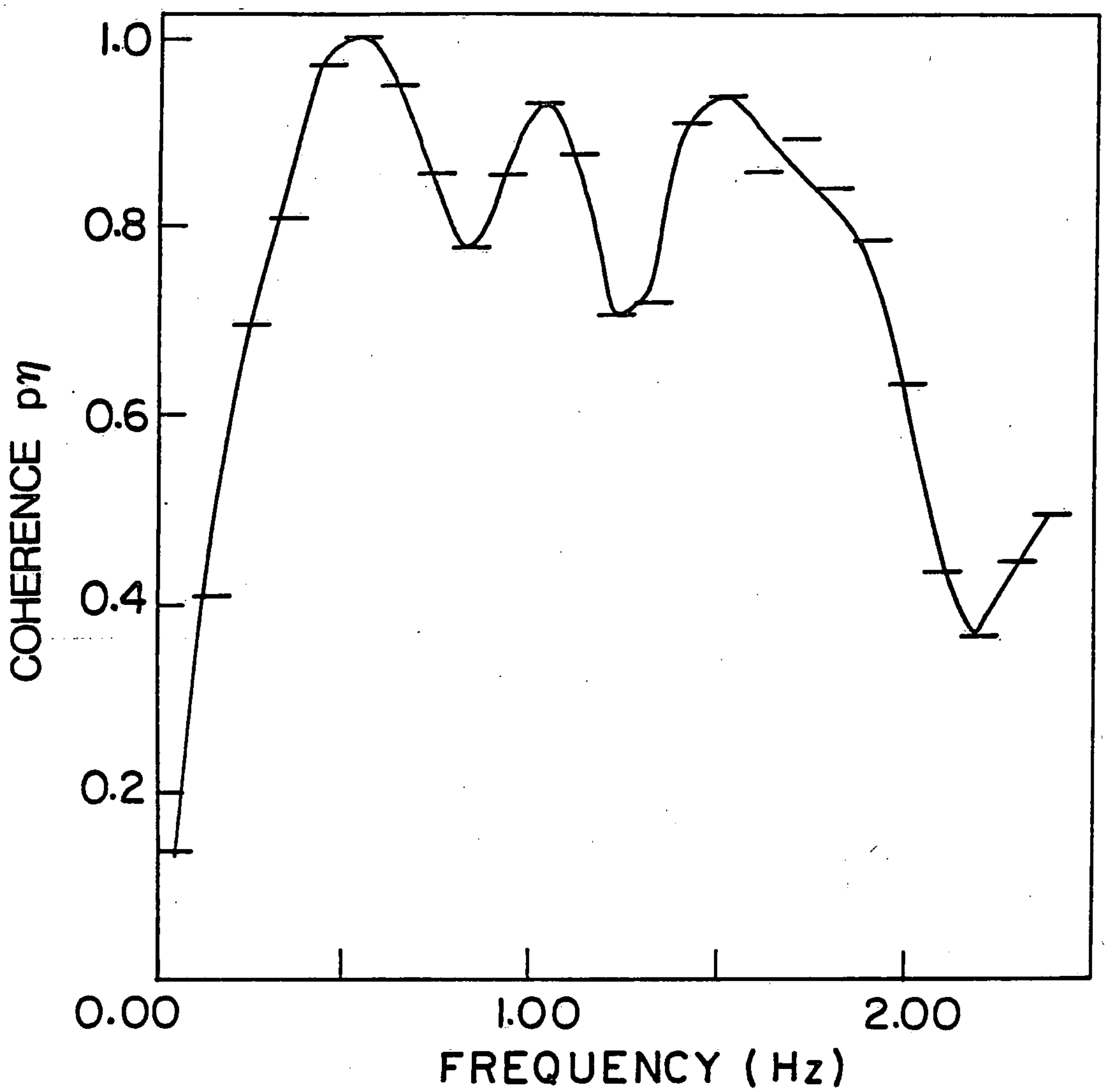
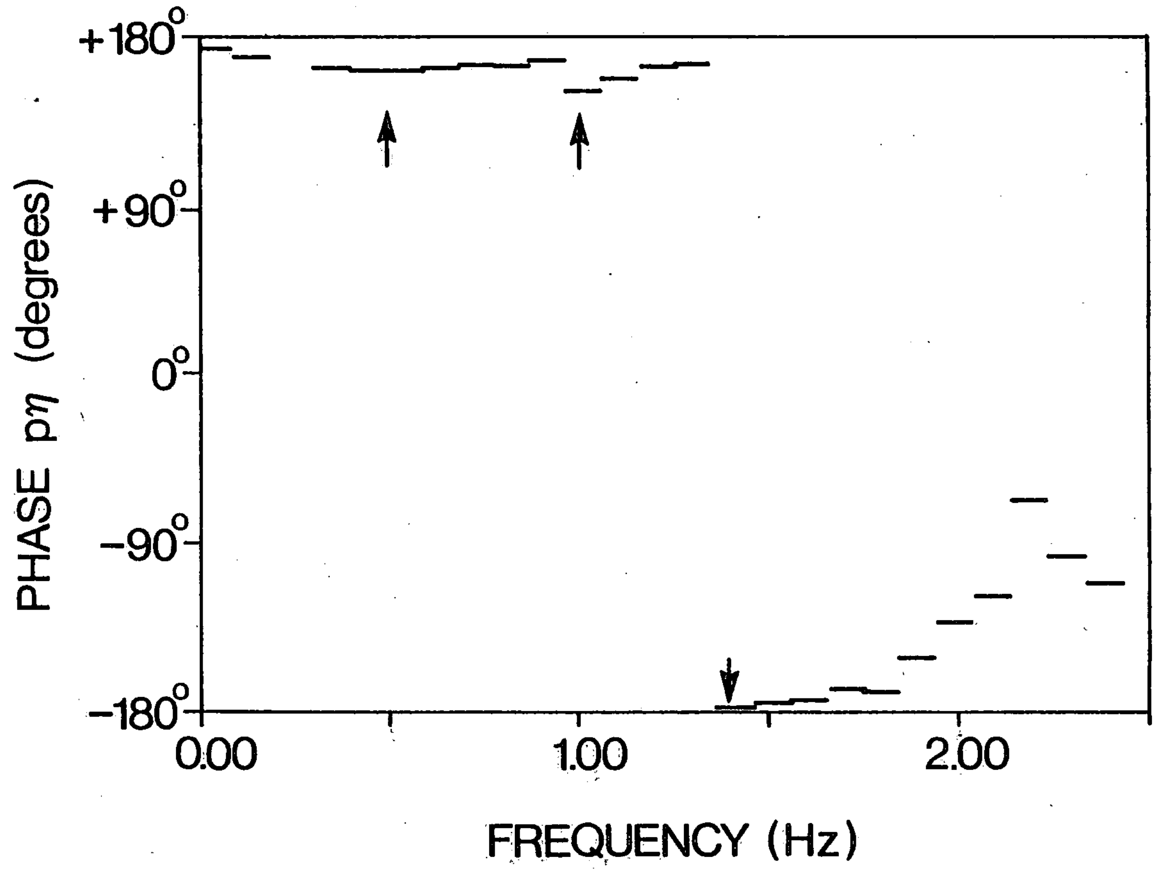


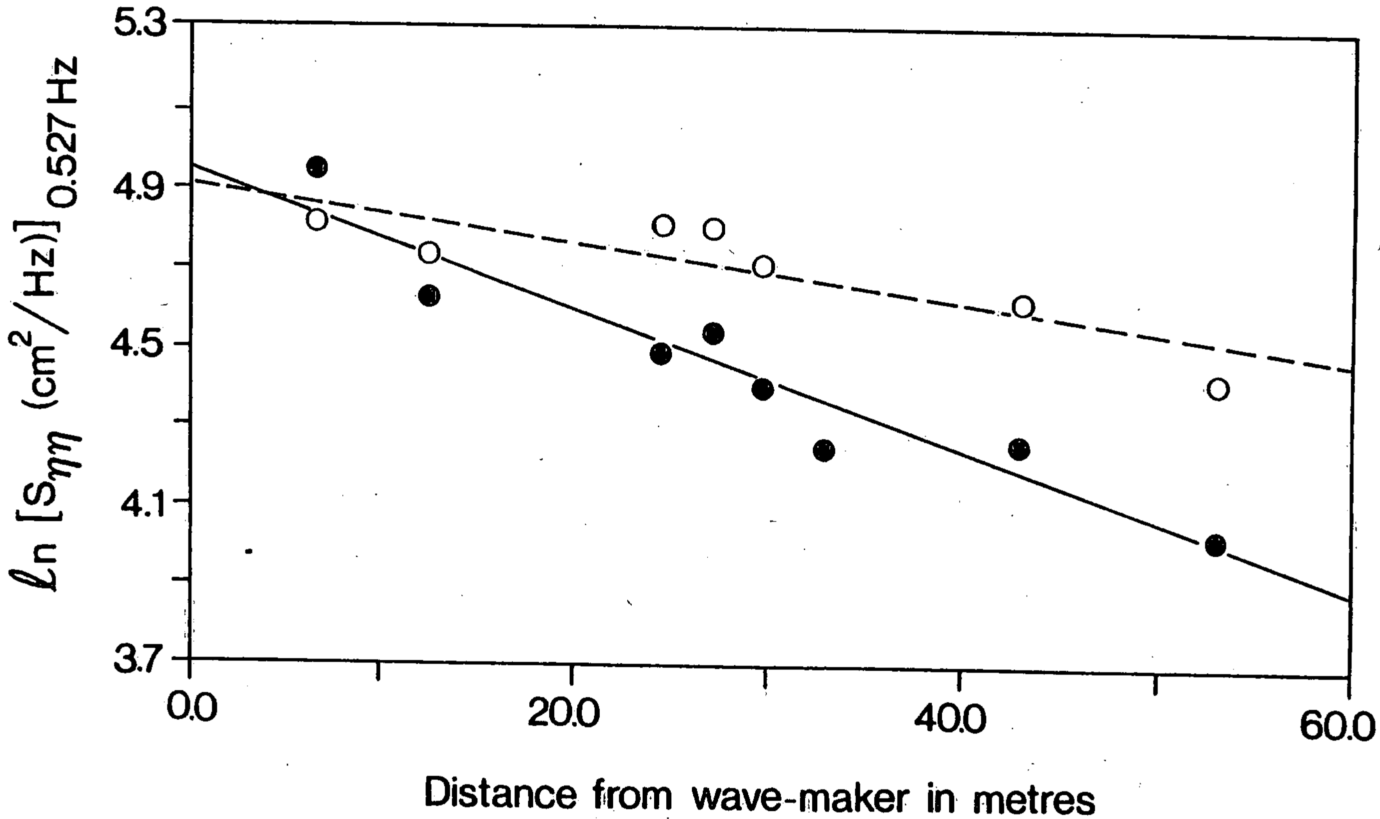
6

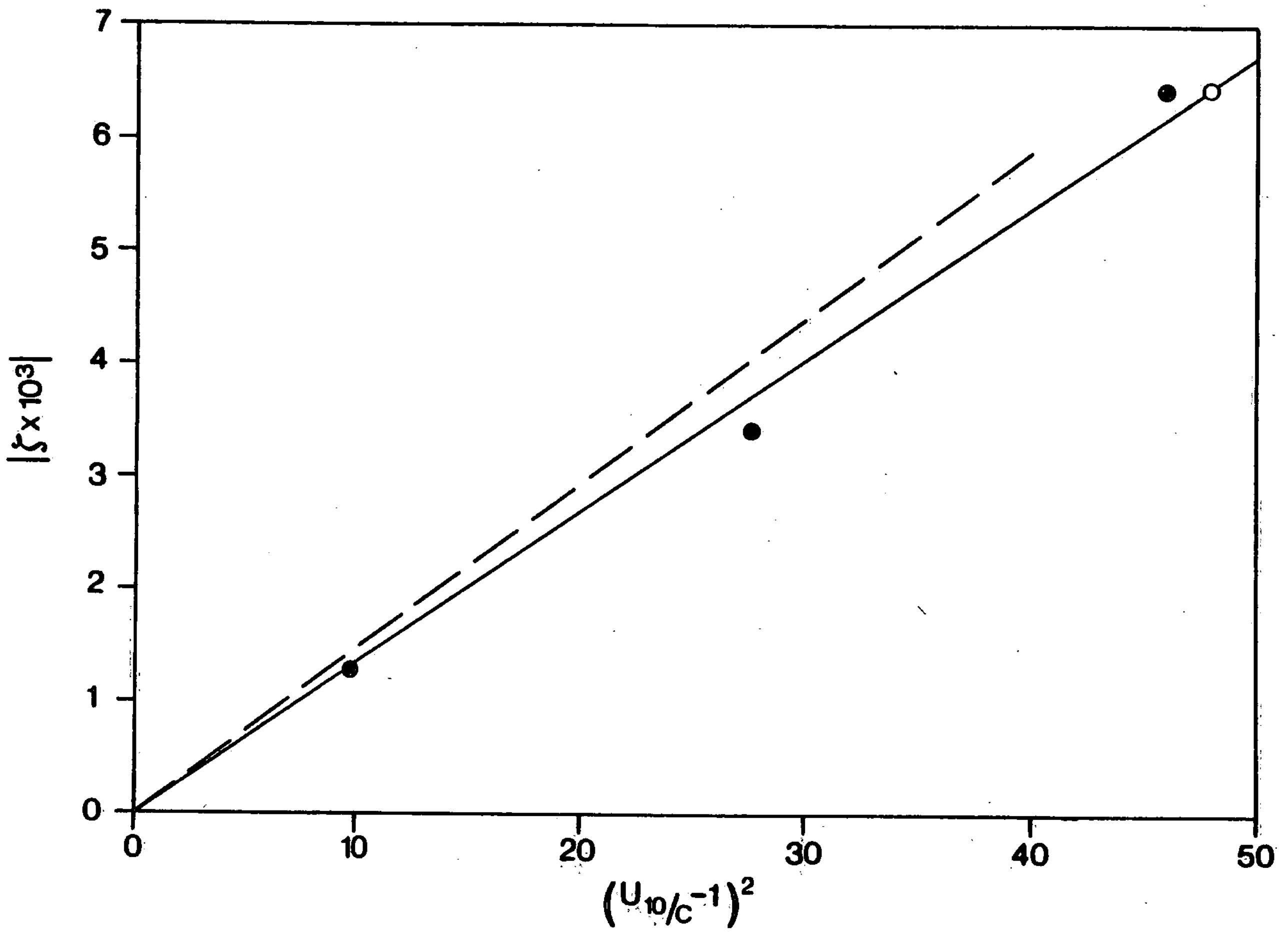


7



8


5.3
4.9
4.5
4.1
3.7
$\ell n\ [S_{\eta\eta}\ (cm^2/Hz)]_{0.527\,Hz}$
0.0
20.0
40.0
60.0
Distance from wave-maker in metres



0
1
2
3
4
5
6
7
|ζ x 10³|
0
10
20
30
40
50
(U10/c -1)²