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THE MECHANICAL COUPLING BETWEEN AIR AND SEA --- AN EVOLUTION OF IDEAS AND OBSERVATIONS

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MANAGEMENT PERSPECTIVE

Advances in numerical weather modelling and improved prediction of global changes are critically dependent on the mechanical coupling between the air and the sea. This coupling has traditionally been described by the roughness of the surface. This review paper traces our evolution of ideas on how best to understand the roughness and how best to describe it in terms of wind speed and sea state.

Major advances in weather modelling require more data from remote areas of the ocean surface, which can be done best via remote sensing. This technique requires an empirical description of wind speed as a function of the very short (centimetric) surface waves. It is pointed out that the empirical relations derived from data in moderate winds will likely fail in high winds because the short waves are limited in steepness by wave breaking.

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SOMMAIRE À L'INTENTION DE LA DIRECTION

Les progrès de la modélisation des conditions météorologiques et de meilleures prévisions des changements à l'échelle du globe sont essentiellement fonction du couplage mécanique entre l'air et l'océan. Depuis toujours, ce couplage a été décrit par la rugosité de la surface. Le présent document d'étude décrit l'évolution des idées sur la meilleure façon de comprendre la rugosité et de la décrire en fonction de la vitesse du vent et de l'état de la mer.

Les progrès importants de la modélisation des conditions météorologiques exigent plus de données provenant des zones éloignées de la surface des océans, lesquelles peuvent être obtenues par télédétection. Cette technique exige une description empirique de la vitesse du vent comme fonction des vagues de surface très courtes (centimétriques). On a signalé que les relations empiriques développées à partir de données s'appliquant à des cas de vents modérés ne couvrivont probablement pas les cas de vents forts parce que le déferlement limite la cambrure des vagues courtes.

ABSTRACT

The roughness of a water surface is parametrized in terms of the wind speed and wave age. Although the variation of equivalent drag coefficient over typical wave age ranges is not large, the distribution of momentum input to the spectrum moves towards the peak in young seas. The implications of this and the inevitable high wind speed saturation of short waves are discussed in the context of scatterometry.

RÉSUMÉ

La rugosité à la surface d'une étendue d'eau est paramétrée en fonction de la vitesse du vent et de l'âge des vagues. Même si la variation du coefficient de résistance équivalente par rapport aux plages caractéristiques de l'âge des vagues est faible, la répartition de l'apport dynamique au spectre se déplace vers le point maximal dans les océans jeunes. Les effets de cette caractéristique et la saturation inévitable des vagues courtes par la vitesse élevée du vent sont traités dans le contexte de la diffusiométrie.

1. INTRODUCTION

In assessing the progress in a subject of such long-standing interest as air-sea interaction, one is faced with the choice of where to begin. To begin at the beginning would entail a trek through early paths, perhaps already well known and certainly well documented elsewhere. The space available for more recent advances would necessarily be reduced. On the other hand, one would prefer to begin at a beginning --- at the point of infusion of new ideas or revealing observations --- from which stems a new view of the subject. Fortunately, such a beginning occurred in the mid-to-late fifties when the significance of waves in the mechanical coupling of air and sea began to be realized. At the same time, stimulated by Ursell's (1956) critical review, new ideas of wave drag began to emerge. Soon after, the theory of weak non-linear transfers among surface waves was developed --- it was to have far-reaching effects on the understanding of the evolution of wind-driven seas and ultimately on the methodology of numerical wave prediction. This is an appropriate beginning too, because it marks the awakening of Klaus Hasselmann's continuing interest in waves and air-sea interaction.

2. AN HISTORICAL GLIMPSE

Most general circulation modellers would like to use the simplest air-sea transfer coefficients that reflect all the necessary physics and not an iota more. The conventional wisdom in the fifties was that the drag (wind-stress) coefficient C_D was a constant, or perhaps, weakly dependent on the surface wind. So, with apologies to the dimensionally pure, it was generally represented by:

$$C_D = \frac{\tau}{\rho U_{10}^2} = A + B\,U_{10} \qquad [1]$$

where τ ($=\rho u_*^2$) is the surface stress, ρ the air density, u_* the friction velocity and U_{10} the wind speed at 10 m height (hereafter indicated by U). A and B are constants.

Since 1960 there have been more than thirty observational studies of the surface drag coefficient leading to a recipe of the type [1]. Most of these have been summarized by Geernaert (1990) in a recent review. Figure 1 has been constructed from the data given in his Table 1. Each point represents a published investigation of C_D in the form of [1] and is plotted at the publication date. Figure 1a shows the estimated drag coefficient at a surface (10 m) wind speed of 7.5 m/s --- about the global marine average. Evidently, in the view of air-sea interactionists, the magnitude of the drag coefficient in moderate winds has remained quite stable at about 1.3×10^{-3}. This is not so for high winds when a clear increase with time is apparent (Figure 1b). The reason for this appears to be due largely to increased efforts to acquire data in high winds. Before 1971, no studies reported wind speeds in excess of 13 m/s and few detected any trend in C_D with U amidst the usual considerable observational scatter.

The consensus now appears to be that the drag coefficient increases with wind speed, at least for speeds in excess of about 10 m/s. Although there is little reliable field data on winds less than 4 m/s, few will argue that at very low wind speeds, when the surface is hardly ruffled by the wind, the drag coefficient will correspond to that in smooth flow; i.e., decreasing with increasing wind speed. The observed increase with wind speed was anticipated by Charnock (1955), who argued, on dimensional grounds, that for rough flow the roughness length, z_0, should be proportional to u_*^2/g. These two asymptotic limits for smooth and rough flow describe the general trend of open ocean drag coefficient estimates rather well (Donelan, 1990):

$$\text{Smooth:} \quad z_0 = \frac{0.11\nu}{u_*}; \quad u_* < 2(\nu g)^{1/3} \qquad [2a]$$

$$\text{Rough:} \quad z_0 = \frac{0.014u_*^2}{g}; \quad u_* \geq 2(\nu g)^{1/3} \qquad [2b]$$

The roughness length and drag coefficient are connected via:

$$C_D\,(z) = \kappa^2\left(ln\frac{z}{z_0}\right)^{-2} \qquad [3]$$

where ν is the kinematic viscosity of air, κ is von Karman's constant and z is the height of measurement of τ in [1]. The drag coefficients corresponding to [2] are graphed in Figure 2.

It is widely recognized that the spectrum of roughness elements is distributed preferentially over short gravity waves by reason of their steepness and reduced celerity in the wind direction (Munk, 1955). [Phillips (1977) has provided a physical basis for

Charnock's formula [2b] on the assumption that only waves having phase speeds, $c < 5u_*$ contribute to the surface roughness].

Most of the drag coefficient estimates in the sixties and seventies were made in steady winds and open ocean conditions. The results, usually expressed in the form of [1], gave little reason to seek a formulation different from [2]. However, at short fetch and in laboratory tanks higher roughness lengths were observed (Kunishi, 1963; Hidy and Plate, 1966), and it became apparent that the roughness length parametrization should take explicit account of the "mobility" of the roughness elements. Kitaigorodskii and Volkov (1965) approached the problem in a reference frame moving at the wave phase speed, c(k), so that contributions to the roughness length are weighted by the ratio of u_* to c(k). For a continuous spectrum (Kitaigorodskii, 1968):

$$z_0 = \alpha\left[\int_0^\infty S(k)\, exp\{-2\kappa c(k)/u_*\}\, dk\right]^{1/2} \qquad [4]$$

where S(k) is the wavenumber spectrum of surface elevation, k is the wavenumber and α an empirical constant.

This approach requires knowledge of the wavenumber spectrum in some detail and a simpler contracted version based on the phase speed at the peak of the spectrum, c_p, is a more practical tool (Kitaigorodskii, 1970):

$$z_0 = 0.3\sigma\, exp(-\kappa c_p/u_*) \qquad [5]$$

where σ is the rms deviation of the surface. This works well at short fetch but greatly underestimates the stress near full development (Donelan, 1990).

Various attempts to include wave properties in the specification of surface roughness include those of Hsu (1974), Byrne (1982) and Donelan (1990). At short fetch, when the dominant waves are relatively steep and actively breaking, Donelan found values of the drag coefficient to be about twice as large as those given by [2]. By analogy with flow over a rough wall, the roughness length to mean wave height ($\sqrt{(2\pi)}\sigma$) ratio was investigated. For waves near full development, this ratio was less than one hundredth of that due to typical sand grain roughness; whereas at very short fetch the ratio approached that observed due to sand grains (Nikuradse, 1932), suggesting very strongly separated flow around the dominant waves. Figure 3, taken from Donelan (1990), shows the strong dependence of the relative roughness (z_0/σ) on the inverse wave age U/c_p. The field and laboratory data show the same trend but the points are not contiguous. The regression lines to the two data sets are given as:

$$\text{Field:} \quad \frac{z_0}{\sigma} = 5.5\times10^{-4}\left(\frac{U}{c_p}\right)^{2.7} \qquad [6a]$$

$$\text{Laboratory:} \quad \frac{z_0}{\sigma} = 9.8\times10^{-6}\left(\frac{U}{c_p}\right)^{3.5} \qquad [6b]$$

(Throughout this paper empirical coefficients have been rounded to 2 significant figures where no greater precision is warranted.)

These relationships indicate that the fraction of the spectrum contributing to the roughness is a strong function of wave development. In order to provide a complete description of the roughness of a pure wind sea in various states of wave development, we include the empirical relationship between σ and U/c_p from Donelan et al., (1985):

$$\sigma = 5.5\times10^{-2}\,\frac{U^2}{g}\left(\frac{U}{c_p}\right)^{-1.7} \qquad [7]$$

so that for aerodynamically rough field conditions,

$$z_0 = 3\times10^{-5}\,\frac{U^2}{g}\,\frac{U}{c_p} \qquad [8]$$

In Figure 4 the roughness length for rough flow has been graphed versus U^2/g for various values of the inverse wave age parameter U/c_p. The intersection with the smooth flow curve [2a] determines the minimum roughness length. Full development (Pierson and Moskowitz, 1964) corresponds to $U/c_p = 0.83$. In the open ocean U/c_p is generally in the vicinity of unity; i.e. the wind sea is usually not far from equilibrium. However, near coastlines (fetch-limited waves) and in localized storms (duration limited waves) the waves may be quite under-developed and U/c_p may be 3 or 4. Over-developed waves ($U/c_p < 0.83$) may exhibit considerably smaller roughness lengths and are not well described by [8].

Most open ocean experiments on the aerodynamic roughness of the sea conclude with a relationship of the form of [1]; i.e. no discernible effect of wave development on the drag coefficient. The reason for this is an unfavourable balance of range of C_D and its measurement error. Typically C_D is estimated with an r.m.s. error of about 20% (Donelan, 1990). In the usual range of C_D, this corresponds to a factor of 2 to 3 in z_0. At any given wind speed this sampling error would generally exceed the natural variability of z_0 caused by the changes in wave age. It is thus no accident that the wave age dependence described by [8] was determined from very fetch-limited observations which covered a wide range of wave development.

3. ROUGHNESS SCALES

If all that one needed to know about air-sea interaction were the numerical values of the transfer coefficients for momentum, heat and mass, then a description such as [2] or [8] for the appropriate roughness length (momentum, z_0; heat z_t; mass z_m) would suffice and it would matter little how the transfers were effected and which scales were important in determining the coupling. However, there are many practical reasons for striving for a more detailed understanding; among these are: (1) understanding the energy, momentum or action balance for surface waves and predicting their evolution on the ocean; (2) calculating the energy supply to the upper ocean, which in fully rough flow is delivered from the wind via waves of all scales; (3) developing a fuller understanding of electromagnetic and acoustic reflectivity and emissivity of the surface --- a crucial step in improving oceanic remote sensing.

The evolution of ideas of the energy balance in the wind sea spectrum was given substantial thrust by the development of the theory of weak non-linear interactions. Phillips (1960) obtained the conditions for resonance among quartets of gravity waves and Hasselmann (1962, 1963) developed a general perturbation theory for the non-linear resonant interaction of free waves in a homogeneous random sea. The energy balance in deep water is determined by the wind input, S_{in}, dissipative processes including viscosity and wave breaking, S_{ds}, and by non-linear transfers among different wavenumbers, S_{nl}:

$$\frac{\partial F}{\partial t} + v_i \frac{\partial F}{\partial x_i} = S_{in} + S_{nl} + S_{ds} \qquad i = 1,2 \qquad [9]$$

where F is the directional spectrum and $v_i = \left(\frac{g}{k}\right)^{1/2} \frac{k_i}{2k}$ is the (linear theory) group velocity.

The development of the wind sea spectrum depends on the balance of the "source functions" on the right hand side of [9]. They are broadly distributed across the short wave part of the spectrum, but S_{nl} is believed to be the dominant term near the peak. Figure 5, taken from Hasselmann et al.,(1973), gives a general picture of their distribution.

The spectral balance [9] was first explored in a comprehensive fetch-limited experiment (Hasselmann et al., 1973). A sequence of spectra at various fetches (Figure 6, from this experiment) reveal the evolution of the peak to lower frequencies with fetch, the establishment of a "quasi-saturated" high frequency (equilibrium) range, and the enhanced peak. The energy at a particular frequency rises to a maximum as fetch increases and falls again to an equilibrium level further downfetch. This "overshoot" phenomenon, described before by Barnett and Sutherland (1968), was therefore clearly identified with an "enhanced" spectral peak. Donelan et al., (1985) later showed a clear connection between degree of enhancement and wave age (Figure 7).

The two principal effects of the non-linear transfers near the peak are (Hasselmann et al., 1985: (1) energy is moved from higher frequencies to frequencies below the peak, thereby producing the evolution of the peak to lower frequencies with time or fetch; (2) energy is removed from or added to the peak region depending on the sharpness of the peak, thereby acting to stabilize the shape of the spectrum.

The greatly enhanced peak at short fetch, and the pronounced overshoot observed, indicate that the average steepness of waves at the peak may be greater than that in the quasi-saturated region ($1.5\omega_p$ to $3\omega_p$). Such steep, large waves may contribute significantly to the surface roughness. Indeed, wind tunnel measurements (Donelan, 1990), of direct wind input through differential pressure on upwind and downwind slopes, show that about 50% of the total stress was supported by these short-fetch waves. In this case $z_0/\sigma \sim 1/30$ or 100 times larger than would be observed near full development.

The fact that waves near the spectral peak can contribute to the aerodynamic roughness of the sea surface and do directly absorb momentum and energy from the wind, has important consequences in all three aspects of air-sea interaction listed at the start of this section. The wind input source function can be quite different from the usually assumed form, which is linear in both the spectrum and the inverse wave age. The effect of breaking on producing separated air flow and greatly enhanced drag (Banner & Melville, 1976) may make an important contribution to the total drag. The energy input is heavily weighted towards the long, faster travelling waves and greatly increased kinetic energy input to the surface may occur in storm conditions or at short fetch. This is supported by recent observations of greatly enhanced kinetic energy dissipation rates near the surface (Drennan et al., 1991). Perhaps most significant, from the point of view of weather or climate prediction, are the implications for remote sensing --- particularly scatterometry --- that the stress is not confined to the very short waves. Interpretation of microwave backscatter from centimetric waves in terms of τ or u_* is therefore on far shakier ground than usually assumed. Instead of a localized response (in wavenumber space) of the short microwave scatterers to the wind stress directly, one must consider the energy balance of the entire wave spectrum. Some of the energy delivered to the vicinity of the spectral peak will be lost directly by large wave breaking, some will be transferred to shorter waves where the dissipation is most rapid, and some will be added to the forward face of the spectrum, producing net growth. The estimation of surface stress from microwave backscatter therefore requires a complete picture of the energy balance in the wave spectrum as well as the usual phase effects of hydrodynamic modulation and purely geometric effects associated with tilt of the surface by the long waves. All of the source and modulation functions are sensitive to the details of the wavenumber spectrum, about which we lack much hard observational information.

4. REFLECTIONS ON SPECTRAL SHAPE AND SCATTEROMETRY

The needs of remote sensing have placed particular emphasis on obtaining a detailed description of the wavenumber spectrum. Observations of wave spectra from point gauges provide reliable information on the spectrum near the peak, but the conversion of observed frequency spectra to wavenumber spectra is confounded by Doppler shifting of shorter components by long wave orbital velocities and currents. Various attempts to deduce the wavenumber spectra by invoking appropriate balances of the source functions have been put forward by, Kitaigorodskii (1983), Phillips (1985), Plant (1986) and Donelan and Pierson (1987). Banner (1990) has demonstrated the effect of Doppler shifting on frequency spectra and made some interesting inferences of the underlying wavenumber spectra for equilibrium ranges in the gravity wave spectrum. However, the range of wavelengths where both gravity and surface tension are important --- the capillary-gravity waves --- is central to microwave scatterometry. Various microwave backscattering experiments seem to show that the spectral levels of these waves are sensitive to wind. Wu (1990) has summarized the wind speed dependence of backscattered power, σ_0, at a particular Bragg wavelength $\sigma_0 (k) \sim U^n$. He finds that the summary of all available data in the Bragg wavelength range of 0.87 cm to 70 cm suggest a variation of wind sensitivity given by:

$$n = 0.23\, k^{1/3}, \quad k \text{ in m}^{-1}. \qquad [10]$$

The wind sensitivity varies from roughly U^5 to $U^{2.5}$ over the wavelength range covered. Taken at face value, this implies a continuous wind dependence over a wide range of wind speeds. In fact the data were gathered in the usual rather restricted range of field data --- about 5 m/s to 20 m/s. It seems unlikely that the spectral levels will increase indefinitely with wind speed. Rather, at sufficiently high winds the high wavenumber spectra will become fully saturated and show no further appreciable wind dependence. The large tank wavenumber spectra of Jähne and Riemer (1990) are very revealing in this regard. Their downwind wavenumber spectra for one decade on either side of the capillary-gravity transition are reproduced in Figure 8. The variation in wind sensitivity with wavenumber is apparent. While the short gravity waves on the left of the diagram show only weak sensivitity, the capillary-gravity region unfolds under the action of the wind providing, in this tiny corner of the wave spectrum, the essential key to scatterometry. The purely capillary waves to the right of the diagram are quite wind-sensitive but are also strongly attenuated by viscosity.

The ordinate of Figure 8 is the spectrum times k^4 --- Phillips' (1985) degree of saturation. A vertical slice through Figure 8 at any wavenumber reveals the tendency to saturation at the highest wind speeds. The saturation level increases from gravity to capillary gravity waves in a manner that suggests that the limiting process is wave breaking, as originally suggested by Phillips (1958). If this were so the asymptotic spectral dependence in both short gravity and capillary ranges would be as γk^{-4} but the degree of saturation γ would be somewhat greater for capillary waves --- roughly proportional to the square of the ratio of their limiting steepness, i.e., ~ 3. Of course, the purely capillary waves are limited by viscosity and the wave-breaking asymptote is never reached on the right side of Figure 8, but the tendency is apparent.

5. CONCLUDING REMARKS

I have not attempted to cover all aspects of the development of air-sea interaction and its present status, but rather, I have chosen to focus on two aspects of air-sea interaction that are of particular interest to general circulation and wave modellers and which bear directly on weather and climate modelling.

The selection of a drag coefficient based on wind speed alone is fundamentally wrong and should be replaced by [8], which relates the roughness length to both wind and wave parameters in a dimensionally consistent way. When the mean square slope is dominated by the locally generated wind sea the surface stress will be accurately estimated in this manner. In special circumstances, such as during intense storms, the surface wind may be counter to recent swell and far greater roughness may be expected.

The use of scatterometry to estimate global marine winds holds great promise for better general circulation and wave modelling. However, in high winds the approach to saturation of the normally wind-sensitive capillary-gravity region of the spectrum may complicate the estimation of storm winds. It is well-known that other scattering mechanisms --- besides Bragg scattering --- become important in high winds (see for example, Banner and Fooks, 1985 and Donelan and Pierson, 1987) and extend the wind sensitivity of microwave scatterometers to higher wind speeds. The azimuthal dependence of other scattering mechanisms, such as breakers, may be quite different from that of the

Bragg waves and may therefore add to the difficulty of recovering the surface wind vector.

Clearly, we have come a long way in understanding, parametrizing and monitoring the mechanical coupling between air and sea. Even more clearly we have a long way to go. We should not, however, be discouraged. That nature has opened a window in the wave spectrum through which we may view the surface winds from 1000 km up is altogether marvellous. That the view from the window is incomplete and many things yet remain hidden is an incentive to explore further.

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FIGURE CAPTIONS

1. An historical record of published marine drag coefficients at 7.5 m/s (a) and 20 m/s (b). The drag coefficients are referred to an anemometer height of 10 m. The plotted points are taken from a summary of drag coefficient formulae given by Geernaert (1990).

2. Drag coefficients plotted versus wind speed but derived from dimensionally consistent roughness length formulae for smooth flow [2a] (---) rough flow [2b] (.....).

3. The ratio of measured roughness length z_0 to root-mean-square wave height σ versus inverse wave age U_{10}/c_p. The straight lines are regression lines to the laboratory (O) and field (Δ, □, ▲) data separately . Error bars are two standard deviations. The solid bars are the estimated sampling errors. The broken bars are the deviation of z_0/σ about the regression line. The striped bar on the ordinate represents the Charnock relation [2b] for the wind speed range of 7.5 to 20 m/s. (From Donelan, 1990.)

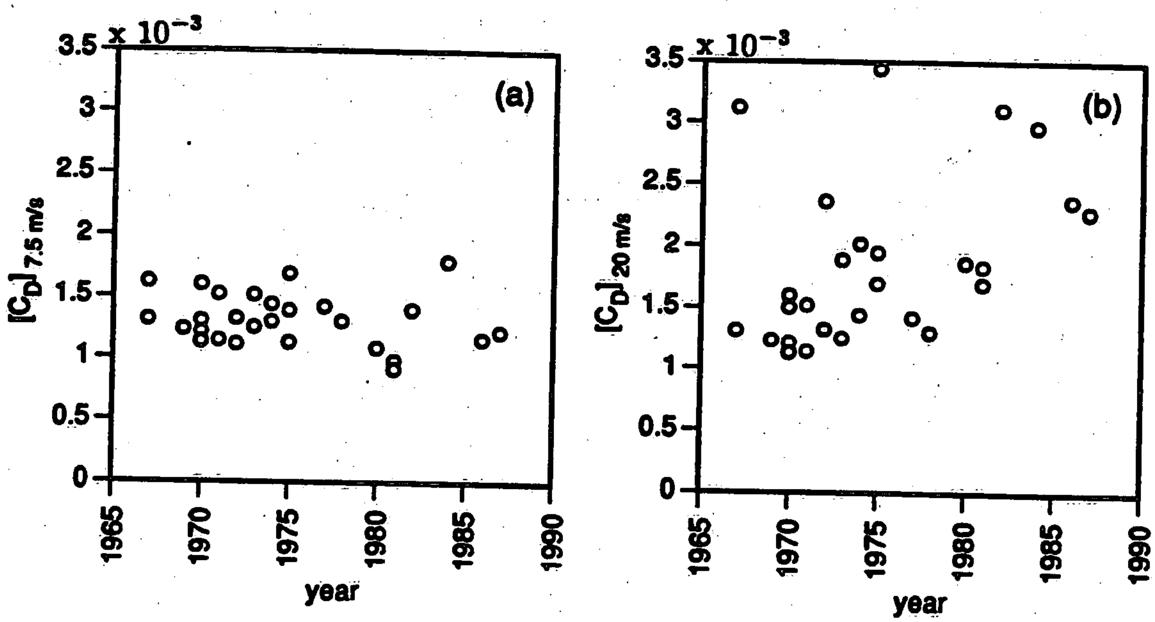
4. Roughness length versus wind speed squared for various values of U/c_p. The dashed line is the smooth flow condition [2a]. The family of solid lines represents rough flow [8] for various values of U/c_p: 0.83, 1, 1.5, 2, 2.5 in order of increasing z_0.

5. Schematic energy balance for the case of negligible dissipation in the main part of the spectrum. (From Hasselmann et al., 1973.)

6. Evolution of wave spectra with fetch for offshore winds (11^h-12^h, Sept. 15, 1968) The spectra are labelled with the fetch in kilometres. (From Hasselmann et al., 1973.)

7. Frequency spectra times ω^4 normalized by the rear face $[\omega^4\Phi(\omega)]_{r.f.}$ which is the average of $\omega^4\Phi$ (ω) in the region of 1.5 $\omega_p < \omega < 3\omega_p$. The lines corresponding to ω^{-5} and ω^{-3} are also shown (--- - ---). The effect of a 10 cm s^{-1} ambient current with or against the waves is also shown (-- -- --) as is the effect of wind drift in a 10 m s^{-1} wind (--- ---). The spectra are grouped in classes of U/c_p. (From Donelan et al., 1985.)

8. A slice through the wavenumber spectrum in the downwind direction at the wind speeds shown (in m/s). The spectra have been multiplied by k^4 to show the degree of saturation B(k). (From Jähne and Riemer, 1990.)



x 10⁻³
(a)
(b)
[C_D]_7.5 m/s
[C_D]_20 m/s
3.5
3
2.5
2
1.5
1
0.5
0
1965
1970
1975
1980
1985
1990
year

Fig. 1

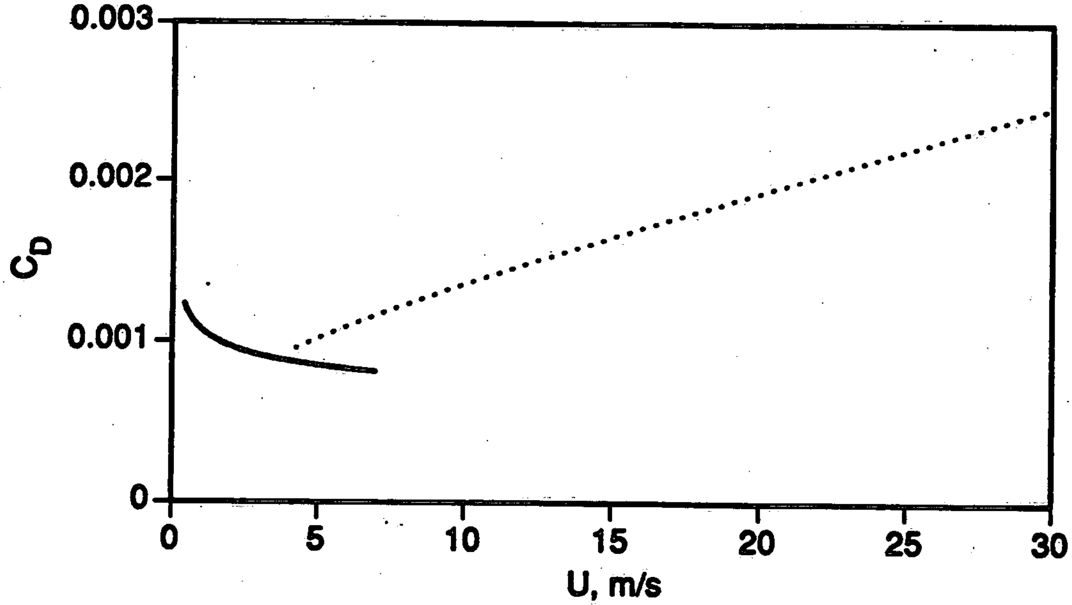


Fig. 2

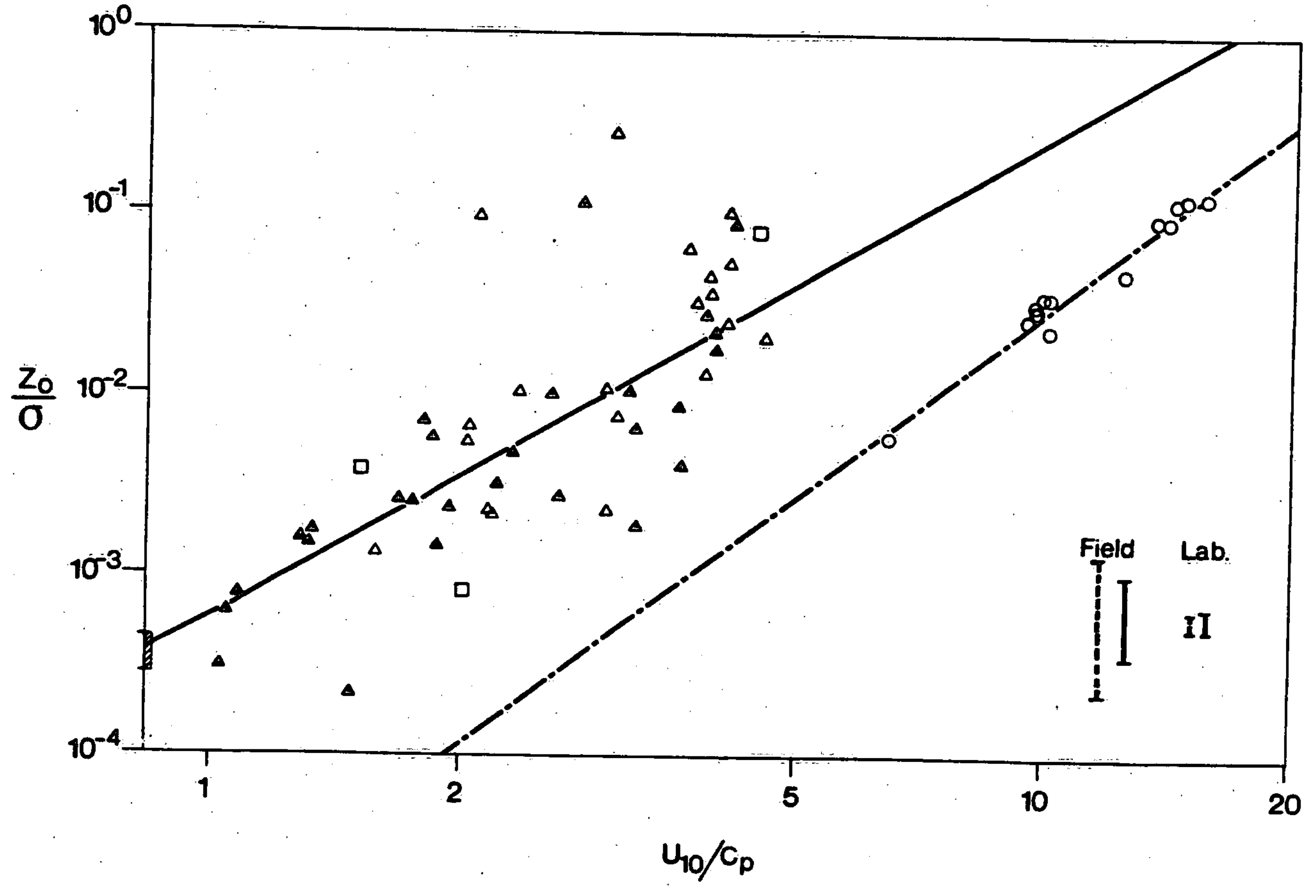


Fig. 3

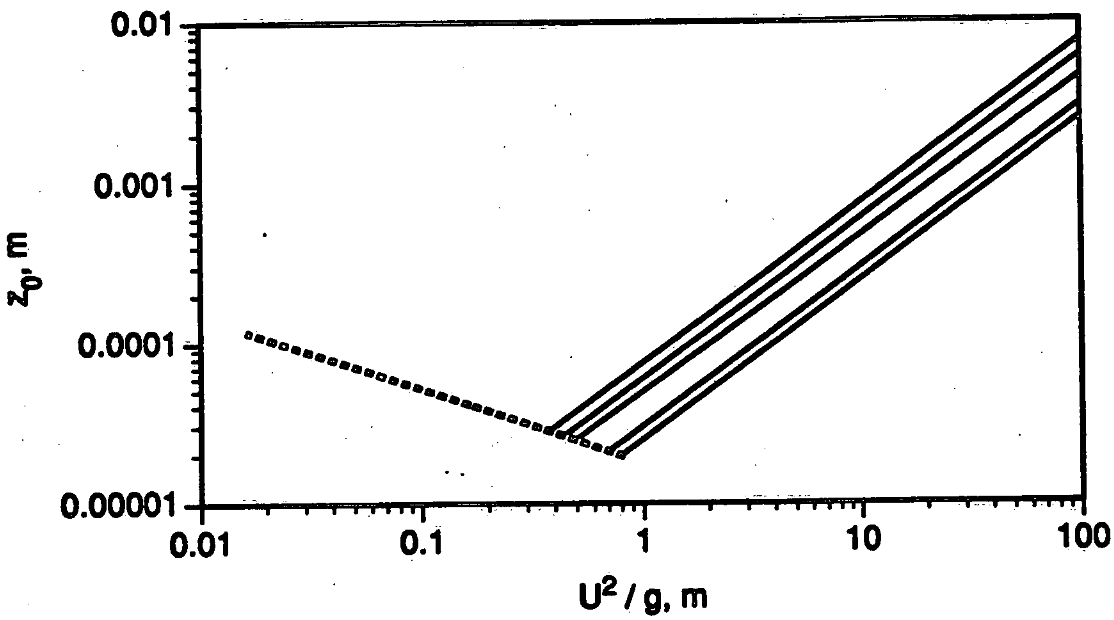


Fig. 4

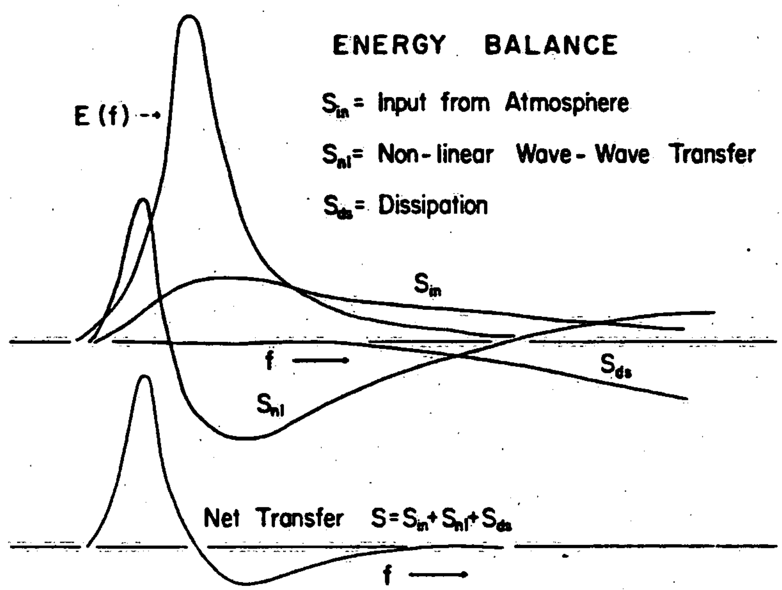

ENERGY BALANCE
S_{in} = Input from Atmosphere
S_{nl} = Non-linear Wave-Wave Transfer
S_{ds} = Dissipation
E (f)
S_{in}
f
S_{ds}
S_{nl}
Net Transfer $S = S_{in} + S_{nl} + S_{ds}$
f

Fig. 5

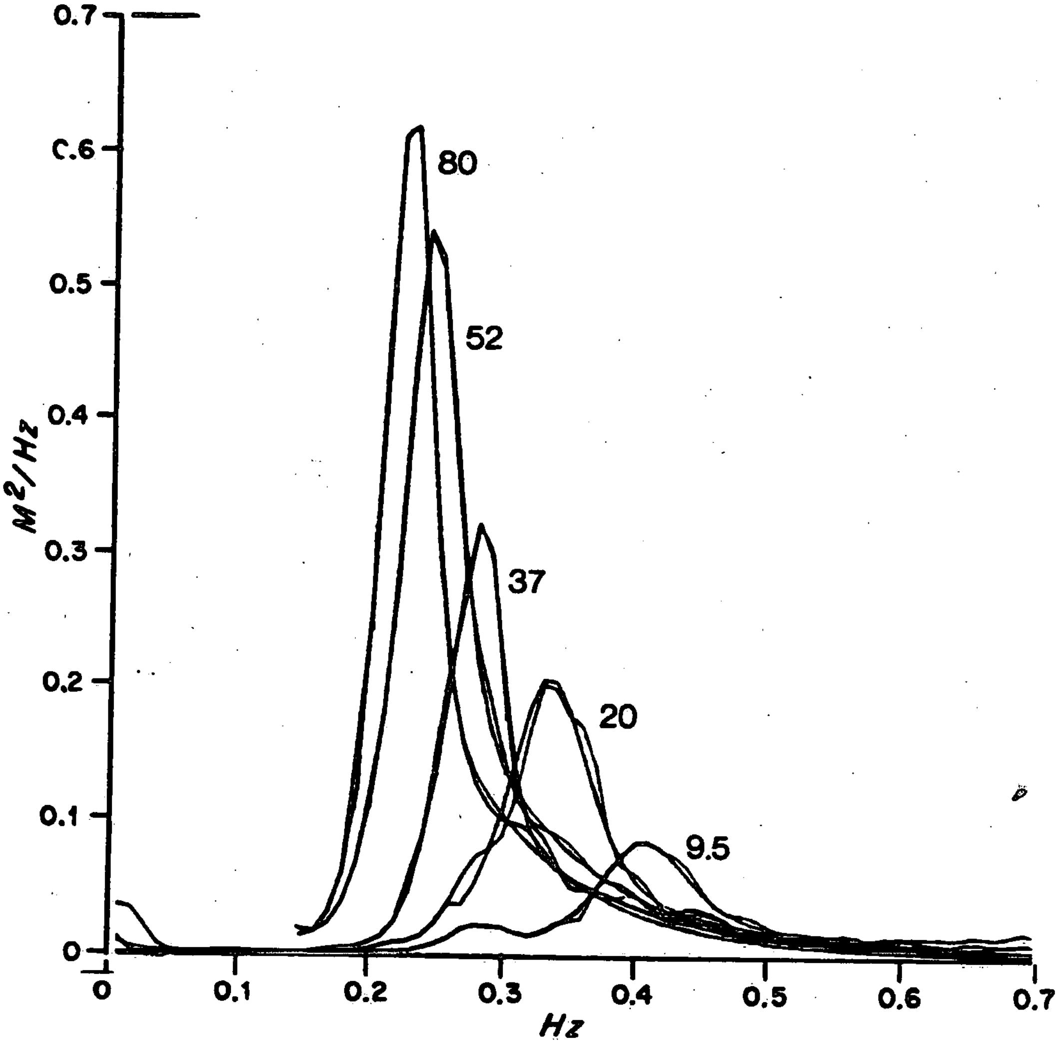
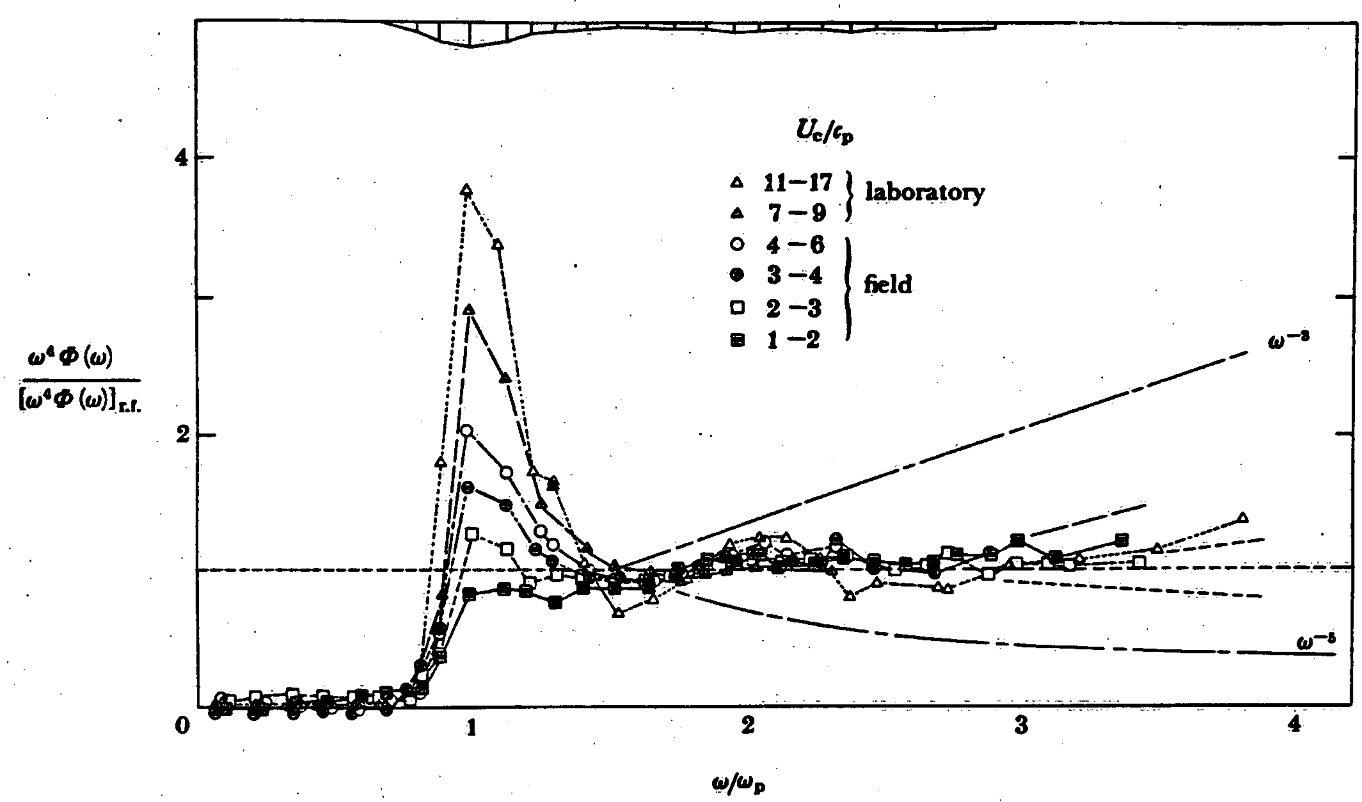
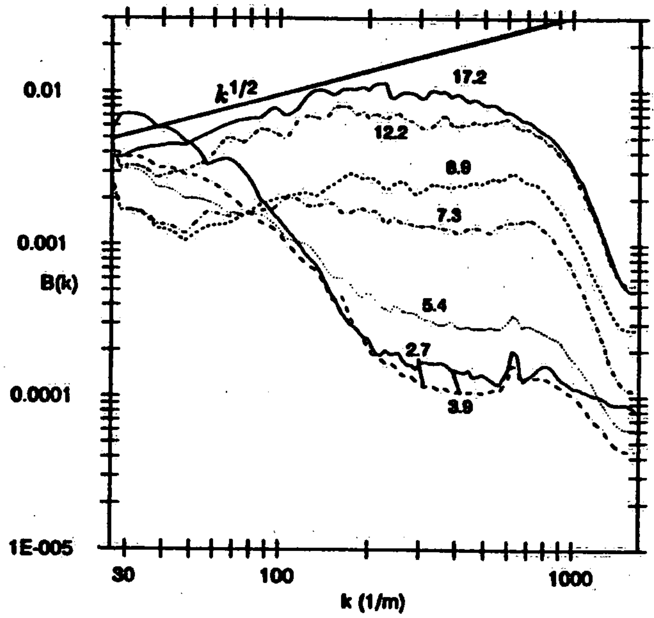


Fig. 6



U_c/c_p
11–17
7 – 9
laboratory
4 – 6
3 – 4
2 – 3
1 – 2
field
ω^{-3}
ω^{-5}
$\frac{\omega^4 \Phi(\omega)}{[\omega^4 \Phi(\omega)]_{r.f.}}$
0
2
4
1
3
ω/ω_p

Fig. 7



0.01
0.001
B(k)
0.0001
1E-005
30
100
1000
k (1/m)
$k^{1/2}$
17.2
12.2
8.9
7.3
5.4
2.7
3.9

Fig. 8

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Environment Canada Environnement Canada

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