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Edited by Bernard LeMaute and Daniel Hanes

AIR-SEA INTERACTION

by

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MANAGEMENT PERSPECTIVE

This work reviews and summarizes the principal advances in understanding air-sea interactions during the last two decades. It emphasizes the new understanding of the effect of waves on the interfacial transfers of momentum, heat and mass. These are important aspects of the coupling between atmosphere and oceans or lakes, which is a critical aspect in understanding atmospheric and oceanic circulation, the hydrological cycle and climatic change.

Dr. J. Lawrence
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PERSPECTIVE-GESTION

Ce document passe en revue et résume les principaux progrès réalisés dans la connaissance des interactions air-mer au cours des deux dernières décennies. Il souligne la nouvelle compréhension de l'effet des vagues sur les transferts interfaciaux de quantité de mouvement, de chaleur et de masse. Ce sont des aspects importants de l'interaction entre l'atmosphère et les océans ou les lacs, qui est elle-même un facteur critique de notre compréhension des circulations océanique et atmosphérique, du cycle hydrologique et de changements climatiques.

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ABSTRACT

This paper explores advances in air-sea interaction in the last two decades, especially aspects related to the exchange of momentum, energy and mass. The modern view of the mechanical coupling between air and sea stems from the pioneering work of Kitaigorodskii, who advanced the idea that the roughness of the sea surface should be related not only to the wind but also to the state of wave development. The failure of various field observations to clarify the matter is ascribed to measurement and sampling errors and the tendency for individual experiments to be confined to a rather narrow range of wave development. A carefully chosen fetch-limited data set is used to revisit the problem and it is shown that the aerodynamic roughness of a wind excited water surface depends on the state of wave development.

Once the roughness of the water surface is known, the question of heat and mass transfer amount to understanding the dependence on the roughness of the thin diffusive boundary layers near the interface. Various models are discussed and compared with field observations. The general tendency appears to be that the Prandtl number dependent heat and mass transfer coefficients (for gas-phase limited substances) are very insensitive to surface roughness. On the other hand, there is considerable evidence from laboratory measurements that water-phase limited substances are strongly dependent on the degree of small scale wave breaking and the consequent mixing of the sub-aqueous diffusive boundary layer.

The effect of a density gradient on the character of the flow and its exchange properties is also discussed with reference to the air boundary layer.

RÉSUMÉ

Cette présentation étudie les progrès réalisés au cours des deux dernières décennies dans le domaine de l'interaction air-mer, en particulier des aspects liés aux échanges de quantité de mouvement, d'énergie et de masse. Cette vision moderne de l'interaction mécanique entre l'air et la mer est née des travaux avant-gardistes de Kitaigorodskii, qui a émis l'idée que la rugosité de la surface de la mer devrait être corrélée non seulement au vent mais aussi à l'état de développement de la vague. On pense que, si les diverses observations sur le champ n'ont pas permis d'élucider la question, c'est en raison des erreurs de mesure et d'échantillonnage, ainsi qu'à la tendance des expériences prises individuellement à être limitées à une gamme plutôt étroite de développement de la vague. On utilise un ensemble soigneusement choisi de données à fetch limité pour revoir le problème, et on montre que la rugosité aérodynamique d'une surface de l'eau agitée par le vent dépend de l'état de développement de la vague.

Une fois la rugosité de la surface de l'eau connue, la question du transfert de chaleur et de masse revient à comprendre comment il dépend de la rugosité des fines couches limites diffusives près de l'interface. La tendance générale semble être que les coefficients de transfert de chaleur et de masse dépendants du nombre de Prandtl (pour les substances limitées en phase gazeuse) ne sont pas affectés par la rugosité de la surface. Par contre, les mesures en laboratoire étayent considérablement l'hypothèse que les substances limitées en phase aqueuse sont très dépendantes du degré de déferlement de la vague à petite échelle et du mélange subséquent de la couche limite diffusive sub-aqueuse.

L'effet d'un gradient de densité sur le caractère du flux et ses propriétés d'échange sera également abordé en relation avec la couche limite dans l'air.

I N D E X

AIR-SEA INTERACTION

Air-Sea Interaction

1. Introduction

In the context of ocean engineering science the principal processes that occur at the air-sea interface are those involving exchanges of momentum, energy and mass. On average the ocean absorbs more than 2-1/2 times the incoming solar energy than does the atmosphere. The energy from the warmed ocean surface is transferred to the atmosphere by infra-red radiation and by sensible and latent heat transfer, so that the atmosphere is to a large extent heated from below. Zonal differences in heating and the earth's rotation drive the large scale atmospheric circulation from which the oceans acquire most of their energy and momentum. Thus the sun's energy in one form or another, crosses the air-sea interface three times before becoming the kinetic energy of oceanic waves and currents. (Benton et al, 1962).

On a global scale the two surficial geophysical fluids act as a coupled thermodynamical system, in which the response of one fluid to the forcing imposed by the other leads to a change in the interfacial conditions and therefore to the exchange rates (degree of forcing). We are not concerned here with the general coupled ocean atmosphere interaction problem, (see for example, Gill, 1982) but rather with the specific air-sea interaction problem of defining the boundary conditions that are appropriately applied at the air-water interface given the mean

conditions in the thin [O(10 m)] boundary layers on either side of the interface. In the main these are boundary constraints on the turbulent fluxes that arise in considering the mean motion or properties of a turbulent fluid. In the mean horizontal momentum equation it is the vertical flux of horizontal momentum or surface stress. In the equation for the mean concentration of a property or contaminant it is the vertical flux of that property at the interface.

In this chapter we explore recent advances in understanding the physics of the air-water interface insofar as it pertains to the question of the specification of boundary conditions on the turbulent fluxes of momentum, heat and mass. For a thorough discussion of the radiative balance at the interface, the reader is referred to Katsaros (1989).

2. The Roughness of the Sea Surface

The degree of mechanical coupling between atmosphere and ocean is conveniently described in terms of the characteristic roughness of the denser fluid. Such an approach derives its antecedents from the extensive body of knowledge of fluid flow over solid walls (see, for example, Monin and Yaglom, 1971). The sea surface, though roughly eight hundred times denser than the overlying air flow, is in motion and its shape evolves on the characteristic time scale of the surface waves. Insofar as the time scale of evolution of the turbulent shear flow near the surface $\left[O\left(\frac{dU}{dz}\right)^{-1}\right]$ is much smaller than the changes in the roughness pattern (a few wave periods of the roughness elements) [1] an analogy with a solid wall may be drawn for momentum transfer (Riley et al, 1982). In other circumstances, for example, the enhancement of moisture flux due to the production of droplets and certain aspects of gas transfer, a porous solid wall is an inadequate model. (See section 5.8)

The surface motion also imposes severe practical constraints on detailed measurements at the surface, so that the surface fluxes must be inferred from measurements [2] made in the fully

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[1] Those waves contributing to the roughness of the sea surface.

[2] Various wave following devices have been constructed to make observations close to the surface (Dobson and Elliott, 1978; Baldy et al, 1978; Hsaio and Shemdin, 1983) but following errors of less than 1% of maximum stroke (travel) are beyond the reach of such devices so that the thin viscous layer [O (1mm)] near the surface is inaccessible via such techniques. Laser-Doppler systems using small moving optical parts to cause the illuminated volume to track the optically sensed surface have

turbulent outer boundary layer. However, the characteristics of the flow are to a large extent determined in the very thin layer [O (1mm)] adjacent to the surface.

The surface boundary layer [O(10m)] is traditionally treated as a layer of constant stress for stationary and homogenous mean[3] conditions (Monin and Obukhov, 1954). In this layer the vertical transport of horizontal momentum (or stress) is effected by molecular viscosity and by turbulent mixing. The former is negligible except very near the surface where the turbulence is suppressed. In the constant stress layer we have:

$$\tau_s/\rho = u_*^2 = \nu \frac{dU}{dz} - \overline{uw} = \text{constant} \qquad [1]$$

where τ_s is the surface stress, ρ the air density, u_* the friction velocity (defined by eq.1), ν the kinematic viscosity of air, u and w are the downwind and vertical velocity components with upper and lower case letters denoting means and

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been devised but the size of the illuminated volume is typically larger than the viscous sub-layer and in any case the method is essentially a laboratory method.

Since the stress is continuous across the air-water interface, the surface stress may be inferred from measurements in either fluid. With few exceptions however, such measurements are made in the air, because in the water, both the mean flow and turbulence are relatively weak compared to the orbital wave velocities and in addition some of the momentum transfer is absorbed, retained and exported by the growing wave field. This last requires knowledge of the spatial gradients in the wave field so that an estimate of the total surface stress from single point measurements in the water is not possible in principle. In view of this, we confine our remarks to the air boundary layer unless otherwise stated.

[3]These terms are qualified in a later section when we come to grips with geophysical data.

deviations from the mean respectively. The overbar indicates a suitable averaging process.

When the flow is aerodynamically smooth a very thin viscous sub-layer exists adjacent to the surface wherein the turbulence is suppressed so that the term $\nu \frac{dU}{dz}$ accounts for all the stress and the velocity profile is linear. Far from the surface turbulence dominates the mixing process and the velocity profile is logarithmic [4]. In wall coordinates $(z u_* / \nu)$ the asymptotes of the velocity profile may be described thus:

$$\frac{U_z}{u_*} = \frac{z u_*}{\nu} \; ; \qquad \frac{z u_*}{\nu} < 5$$

$$\frac{U_z}{u_*} = A_* \ln \frac{z u_*}{\nu} + C_* \; ; \qquad \frac{z u_*}{\nu} > 30 \qquad [2]$$

This is illustrated schematically in Figure 1. For smooth flow the constants A_*[5] and C_* have been determined experimentally by several researchers (c.f. Monin and Yaglom, 1971) and, within quite close tolerances, they are 2.5 and 5.5. The viscous sub-layer thickness δ_ν is taken to be the height at which the linear and logarithmic asymptotes intersect.

$$\delta_\nu = \frac{11.6\,\nu}{u_*} \qquad [3]$$

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[4]Provided the density stratification is negligible.

[5] $1/A_*$ is referred to as von Karman's constant, κ .

For a global average marine u_* value of about 30 cm/s, the viscous sub-layer is less than 1mm thick. Consequently estimates of τ_s are based on measurements in the logarithmic layer, from which one may obtain the friction velocity, u_* and virtual origin of the velocity profile or the roughness length, z_0 .

$$\frac{U_z}{u_*} = \frac{1}{\kappa} \ln \frac{z}{z_0} \qquad [4]$$

For smooth flow it therefore follows that $\frac{1}{\kappa} \ln \frac{z_0 u_*}{\nu} = -5.5$ or $z_0 u_* / \nu = 0.11$.

The drag coefficient $\left[C_z = \tau_s / \rho U_z^2\right]$ provides a convenient parameterization of the surface stress in terms of the mean boundary layer wind at typical anemometer heights. It is an equivalent measure of the roughness of the surface

$$\left[C_z = \kappa^2 \left(\ln z/z_0\right)^{-2}\right]$$

except that it has the added disadvantage of being dependent on the height of measurement. For smooth flow then:

$$C_z = \kappa^2 \left(\ln \frac{z u_*}{0.11 \nu}\right)^{-2} \qquad [5]$$

Oceanic estimates of C_{10} (the drag coefficient measured at $z = 10$ m) range from values less than given by eq.[5] to much larger values showing a general increase with the mean wind speed, U_{10} . The wind excited air-sea interface may be anything from ultra-smooth to fully rough having drag coefficient that may vary by an order of magnitude (3×10^{-4} to 3×10^{-3}) and corresponding roughness

lengths from 10^{-6} to 10mm. Thus an important and persistent task of air-sea interaction has been to find a consistent functional description of z_0 or C_z on more readily observable parameters of the air-sea interface. Have we succeeded in this?

The last article on air-sea interaction to appear in this compendium some twenty-seven years ago (Deacon and Webb, 1962) contained a summary of drag coefficient estimates, in which the distribution of points suggest approximate aerodynamic smoothness below 2.5 m/s and general agreement with Charnock's conjecture [6] above 6 m/s. Conventional practice today follows the results of carefully executed open ocean experiments such as those due to Smith (1980) and Large and Pond (1981) and differs in no essential way from that offered by Deacon and Webb (1962). This suggests that the roughness of the sea surface may be parameterized solely on [7] u_*, ν and g.

Smooth $$z_0 = \frac{0.11\,\nu}{u_*} \quad ; \quad u_* < 2\,(\nu g)^{1/3} \qquad [6a]$$

Rough $$z_0 = \frac{0.014\,u_*^2}{g} \quad ; \quad u_* \geq 2\,(\nu g)^{1/3} \qquad [6b]$$

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[6] On dimensional grounds, Charnock (1955) has suggested that $z_0 = m_0 u_*^2/g$, where g is the gravitational acceleration and the constant of proportionality m_0 must be established empirically. In a comprehensive review of drag coefficients Garratt (1977) selected m=0.0144.

[7] This amounts to parameterization on u_* only since g is a virtual constant and ν varies by at most 30% over typical extremes of marine atmospheric temperatures.

Anticipating a later result, Charnock's formula [6b] aappears to represent the surface roughness near full development when most of the stress is supported by short gravity waves. Yet for these very short waves, other physical parameters besides u_* and g are certainly important.

Among these are the surface tension γ, and water viscosity ν_w that affect the characteristics of the short waves principally but might also affect the breaking characteristics of longer waves.[8] The addition of artificial slicks to wind-excited water bodies, thereby attenuating short waves preferentially, (Lamb, 1932, Phillips, 1977, Ermakov et al, 1986), has demonstrated their importance in establishing both the mean square slope of the surface (Cox and Munk, 1954) and the aerodynamic roughness (Van Dorn, 1951). Naturally occurring surface active materials undoubtedly alter the short wave characteristics and with them the roughness, but the degree to which this occurs is unknown[9] and no systematic exploration has yet been attempted.[10] In

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[8]It has been demonstrated both by numerical modelling (Gent and Taylor, 1976) and by experiment (Banner and Melville, 1976) that strong air-flow separation from a breaking crest greatly enhances the surface stress.

[9]The global sea surface temperature range is about -2°C to 30°C or about a factor of two in ν_w. No evidence for this effect on z_o has yet been accumulated although Kahma and Donelan (1988) have demonstrated its effect on the initiation of waves in a laboratory tank.

[10]In a recent paper Geernaert et al (1988) have attempted to relate C_{10} to γ. It is doubtful that the observed change in γ (corresponding to less than 4% change in the minimum phase speed) is itself the cause of the observed trend. Rather the reduction of γ below the clean water value signals the presence of surface active contaminants whose horizontal concentration

strong winds the surface active materials will be mixed into the body of the fluid and the clean water roughness will be recovered. Scott's (1972) experiments indicate that even a moderate wind of 5-6 m/s was adequate to clean the surface. Thus it might be expected that surface contamination will contribute little to the variability of the z_o for fully rough flow (U_{10}>7.5 m/s).

How can the dependence of z_o on u_* described by equation [6] be reconciled with the well-known roughness characteristics of solid walls, in which three distinct regimes are distinguished depending on the thickness of the viscous sub-layer vis-a-vis the height h_o of the surface roughness elements [11] ? In smooth flow the roughness elements are buried within the viscous sub-layer and the outer flow remains unperturbed by them so that the roughness depends only on the imposed stress and the fluid viscosity $[z_o = fcn(u_*, \nu)]$ as described above. With increasing u_* the viscous sub-layer thins until the roughness elements begin to interact directly with the turbulent outer flow causing some additional form drag.

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variations, produced by the passage of a wave, cause gradients of surface tension. The interface tries to recover a uniform surface tension and acts as a visco-elastic membrane. The relevant parameter in the associated wave damping is the surface dilational modulus $\epsilon_\gamma = d\gamma / d(\ln A)$, where A is the surface area per molecule of the contaminants (Hogan, 1986; Scott, 1986).

[11]Roughness elements are surface features with sufficient steepness to cause flow separation on their lee faces and hence form drag. The flow may remain attached to quite large features having gentle slopes, but separate from small abrupt roughnesses.

In this transitional regime the roughness length first decreases more slowly with increasing u_* then begins to increase until form drag on the roughness elements accounts for virtually all the stress. The flow is now fully aerodynamically rough and the observed roughness length bears a constant relation to the height of the roughnesses $z_0 \sim h_0$ [12] (c.f. Monin and Yaglom, 1971).

Constraints on stationarity and homogeneity required for the existence of a constant stress layer pre-dispose open ocean (fetch essentially unlimited) stress measurements to conditions approaching full development of the wave field. According to Pierson and Moskowitz (1964) at full development wave height [13] is quadratically related to wind speed [14]:

$$\sigma = 0.0608\, U_{10}^2 / g \qquad [7]$$

so that Charnock's relation may be written:

$$z_0 / \sigma = 0.23\, C_{10} \qquad [8]$$

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[12] Measurements over rough solid surfaces show that the constant of proportionality depends on steepness and spacing of roughness elements. For example Nikuradse's (1932) pioneering experiments yielded $z_0 = h_0/30$ for sand grain roughness elements, while measurements over 18 cm high wheat stubble (Businger et al., 1971) yielded $z_0 = h_0/7.5$. See also Lettau (1969).

[13] We use throughout the root-mean-square surface deviation σ as a measure of wave height. For narrow spectra the significant height $H_{1/3} \doteqdot 4\sigma$ (Longuet-Higgins, 1952; Goda, 1970).

[14] Pierson and Moskowitz (1964) used the wind speed at the height of the ship's anemometers used (19.5m). We use the more conventional wind speed at 10m height (U_{10}) following Bretschneider (1973).

Since C_{10} changes by only a factor of two for open ocean fully rough flow $(U_{10} > 7.5\ m/s)$ in the reliably observed wind speed range (up to 25 m/s), the ratio of roughness length to wave height is for practical purposes constant [15] for rough flow over fully developed waves. Thus the tendency is in agreement with rough flow over a wall, but the ratio (z_0/σ) is 100 times smaller than typical solid wall roughnesses. The obvious inference is that the large waves do not contribute to the surface roughness both because they are not steep and they travel at speeds c approaching or exceeding the wind speed [16]. Munk (1955) argued along these lines on the basis of Jeffreys' (1924, 1925) sheltering ideas. More recently Phillips (1977), in the light of Miles' (1957) shear flow instability mechanism for momentum transfer between wind and waves, limits the roughness elements to those waves for which $c/u_* < 5$. Assuming that the spectral density of these short waves is "saturated" (wind speed independent) he shows that the root-mean-square height of these short waves is proportional to u_*^2/g . This reconciles [17] Charnock's

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[15] A factor of two change in z_0 is reflected in a 15% change in C_{10} in the observed z_0 range for these conditions $(0.1 < z_0 < 1.7\ mm)$

[16] Full development corresponds to wave age of 1.2 (Pierson and Moskowitz, 1964 via Bretschneider, 1973). Wave age = c_p/U_{10} . c_p is the phase speed of the waves at the peak of the spectrum.

[17] The idea of a saturated wind sea spectrum above the peak frequency has been dealt a severe blow by recent observations (Toba, 1973; Forristal, 1981; Kahma, 1981; Donelan et al, 1985), which support a wind dependent region of the spectrum at frequencies just higher than the peak frequency ω_p, $(1.5\omega_p < \omega < 3\omega_p)$.

formula eq. [6b] with rough flow over a solid wall provided that the energy containing waves $(0.6\,\omega_p < \omega < 3\,\omega_p)$ support a negligible part of the momentum transfer to the surface. This is very probably the case near full development but the strongest forcing of the oceans occurs in storms wherein the intense winds are generally localized and vary appreciably in direction over distances of a few hundred kilometers, so that the waves are likely to be relatively undeveloped (or "young" i.e. wave age $c_p/U_{10} << 1.2$). In these situations the wind forcing is stronger at the peak of the spectrum and this coupled with energy transfer between wave components leads to an "enhanced" peak (Hasselmann et al, 1973). A clear progression of degree of peak enhancement with inverse wave age (or degree of wind forcing U/c_p) is seen in Figure 2 from Donelan et al (1985), who combined observations of fetch-limited field waves with laboratory waves. The waves at the spectral peak in the most fetch-limited field cases $(U/c_p \sim 4 \text{ to } 6)$ fall within Phillips' bounds $(c/u_* < 5)$ and are in fact about as steep as the waves at $3\,\omega_p$, which may be saturated or nearly so.

. Beyond $3\,\omega_p$ such "frequency-of-encounter" spectra are somewhat whitened by Doppler shifting (Ataktürk and Katsaros, 1987) and the underlying wave number spectrum may be saturated or nearly so. Indeed, Kitaigorodskii (1983) has advanced a theoretical argument for transition to saturation in the short gravity wave region and Jackson et al (1989) point out that indefinite extension of the observed wind dependent region to higher frequencies would lead to mich higher mean square slopes than observed. Recently Banner (1989) has reconciled much of the observational evidence with a proposed spectral form showing saturation of the short gravity waves.

Direct measurements of the momentum transfer (via pressure-slope correlations, $p\frac{\partial\eta}{\partial x}$) to very fetch-limited laboratory waves (Figure 3) demonstrate that the transfer at the spectral peak can, in these circumstances, be a significant fraction of the total stress. Here $p\frac{\partial\eta}{\partial x}$ accounts for about 50% of the total stress measured independently with an X-film hot wire anemometer. In this case $z_0/\sigma \sim \frac{1}{30}$ or one hundred times larger than would be observed over a fully developed sea. Thus the idea that the energy containing waves near the peak of the spectrum do not contribute to the stress (i.e. act as roughness elements) clearly loses validity when the waves are strongly forced. The roughness length cannot then be determined solely by ν, u_* and g but other physical parameters or characteristics of the wave field, such as peak wave length λ_p and phase speed c_p, must enter the problem.

Various schemes to account for the long wave contribution to the roughness have been suggested. The first and most elegant of these was proposed by Kitaigorodskii and Volkov (1965). They argued that if we view the waves in a frame of reference moving with the phase speed c, the appropriate logarithmic law is:

$$\frac{U_z - c}{u_*} = \frac{1}{\kappa} \ln \frac{z}{a} \qquad [9]$$

or

$$\frac{U_z}{u_*} = \frac{1}{\kappa} \ln \frac{z}{a \exp(-\kappa c/u_*)} \qquad [10]$$

thus $$z_0 \propto a \exp(-\kappa c/u_*) \qquad [11]$$

where a is identified with the amplitude of the wave. In a continuous wave spectrum this is readily extended to (Kitaigorodskii, 1968):

$$z_0^2 = \alpha^2 \int_0^\infty S(k) \exp[-2\kappa c(k)/u_*] \, dk \qquad [12]$$

where k is the wavenumber and α must be determined from data.

This approach treats the "mobility" of the roughnesses but does not account for effects of varying steepness across the spectrum. It requires knowledge of the wave number spectrum to quite high wavenumbers and a simpler contracted version based on the peak of the spectrum has been offered (Kitaigorodskii, 1970):

$$z_0 = 0.3 \, \sigma \exp(-\kappa c_p/u_*) \qquad [13]$$

No additional effect of wave steepness on z_0 was determined by Kitaigorodskii (1970) perhaps because steepness and wave age (c_p/U_*) are already quite well correlated (Huang, 1981).

In a recent paper Geernaert et al (1986) compare the performance of several models for C_{10} incorporating wave effects and find that Hsu's (1974) formula (extending Charnock's to allow m_0 to be a function of the wave steepness) performs best[18] but

[18]This refers to the comparison made with coefficients determined from previous data. Geernaert et al (1986) also adjust the coefficients to find best agreement with their data. It is not clear that this is a useful procedure since if adjustments are required to force agreement to a particular data

only slightly better than Charnock's (with m_o =0.0185 as deduced by Wu, 1980). In fact Charnock's formula seems to fit particular data sets well, but the value of the proportionality factor m_o varies from case to case (Kitaigorodskii and Volkov, 1965; m_o =0.035; Garratt, 1977, m_o =0.0144; Wu, 1980, m_o = 0.0185; Geernaert, 1986, m_o =0.0192). Kitaigorodskii and Volkov (1965), Donelan (1982) and Merzi and Graf (1985), show that data collected over a wide range of wave ages do not show good correlation between u_* and z_o . Evidently z_o and u_* are not uniquely related in general, although they are well correlated in those data sets in which the wave age is not very variable. Many studies (Smith and Banke, 1975; Garratt, 1977; Smith, 1980; Large and Pond, 1981) conclude that the neutral drag coefficient is best parameterized on the wind speed[19] alone. Others (Kitaigorodskii and Volkov, 1965; Hsu, 1974; Donelan, 1982; Merzi and Graf, 1985), find that parameters of the wave field are required in addition. The formulae of Kitaigorodskii (1970) and Hsu (1974) are discussed further in Section 4. It may be argued that the first group of researchers have drawn their experimental samples from a population in which the appropriate wave parameters cover too small a range to make their presence felt amidst the noise of the stress estimates. The second group,

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set, then the universality of the method cannot be defended.

[19]This distasteful dimensional inconsistency is avoided by using Charnock's approach for rough flow patched to the smooth flow drag coefficient (eq.[6]). The result (Figure 4) may not be distinguished from, say, Large and Pond (1981).

on the other hand, see the wave related signal in the measured stress but not sufficiently clearly to agree on its form, much less its size. A consistent parameterization of the roughness of the sea surface in terms of the wave field will only be possible when we are able to strengthen the wave related signal (by widening the range of wave development in our observations) and to suppress the noise in our measurements. Given the requirements for stationarity and horizontal homogeneity in the wind, fetch-limited studies are probably the only way to achieve the former goal. In Figure 4, some estimates of C_{10} (from Donelan, 1982) for two distinct age groups of waves ("very young" and "adolescent" are suitably descriptive terms) are compared against the "mature" wave results of Large and Pond (1981) and Garratt's (1977) version of Charnock's formula. The signature of wave development in C_{10} cannot be denied, but accurate parameterization of it requires minimization of the measurement errors. The next section is devoted to identifying the sources of error in estimates of the surface fluxes.

3. The estimation of surface fluxes.

There are many indirect methods of estimating the wind stress at the air-sea interface. Most have been critically reviewed by Deacon and Webb (1962) and Kraus (1968). In particular Kraus finds the wind profile quite unsatisfactory in principle and a recent careful error analysis by Blanc (1983) suggests that further understanding of surface fluxes will not likely emerge from profile data alone. The only other indirect method in common use is the so-called "inertial dissipation" method. This method is derived from the turbulent energy balance equation (see, for example, Wyngaard and Cote, 1971; Kraus, 1972; Fairall and Larsen, 1986) and depends on the dominance of terms involving the production and dissipation of turbulent energy. The method has the merit of being considerably easier to execute on a sea-borne platform and appears to agree rather well on average with direct measurements of $\overline{uw}$ for steady state, open ocean conditions in moderate and strong winds (e.g. Large and Pond, 1981). In light winds Geernaert et al (1988) find much poorer agreement. Essential assumptions about the residual (ignored) terms in the energy balance are based on data over solid surfaces and may be quite wrong for flow over actively developing waves.

It would seem that our best hope of understanding and parameterizing the turbulent exchanges at the air-sea interface lies in focussing on the one direct measure of the fluxes --- the so-called eddy correlation method. Even this is fraught with

errors of various kinds. These are essentially of three types: (a) measurement errors, (b) inaccurate assumptions, (c) sampling variability.

3.1 Measurement errors.

A full discussion of such errors may be found in Kaimal and Haugen (1969) and Dobson et al (1980). One of the most difficult measurement errors to avoid is that due to imperfect levelling of the coordinate system of the measurements (see, for example, Deacon, 1968; Rayment and Readings, 1971; and Wieringa, 1972). In some mechanical anemometers mounted on a rigid support after the fact levelling may be achieved via the assumption that the mean vertical wind component must vanish over averaging times corresponding to wind runs[20] two or more orders of magnitude larger than the distance from the boundary. Instruments in which the component wind speed is recovered only after significant electronic processing may not be stable enough to allow one to estimate the mean vertical velocity with sufficient accuracy to level adequately. In such cases every effort is made to level the instrument on installation and to keep track of subsequent tilts. However, such geometric assiduity may be to little avail in such instruments as an acoustic anemometer (described by Kaimal, 1980), wherein the uncertainty in the true acoustic path about the centreline of the geometric axis can be of the order of the ratio of transducer diameter to path length --- typically a

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[20]The length of the total passage of air running through the sensor, in practice the mean wind times the elapsed time.

few degrees. Initially, the acoustic path is much more closely aligned with the centreline of the transducers, but this may change with aging and working of the piezo-electric transducer material. Kaimal and Haugen (1969) have shown that in unstable air over land a 1° tilt error can account for a 25% error in the stress.

3.2 Inaccurate assumptions

The second type of error arises from the assumption of a constant stress layer and the consequent practice of inferring the surface stress from measurements at heights of 10 metres and more (e.g. 12.5 and 22 m, Large and Pond, 1981; 33 m, Geernaert et al, 1987). The idea of a constant stress layer is based on fairly confining assumptions (Tennekes, 1973a). To see where these might be likely to fail we examine the mean horizontal momentum equation

$$\frac{\partial \underset{\sim}{V}_h}{\partial t} + (\underset{\sim}{U} \cdot \nabla_h)\underset{\sim}{V}_h = f_c \underset{\sim}{V}_h \times \hat{k} - \frac{1}{\rho}\nabla_h P + \frac{1}{\rho}\frac{\partial \underset{\sim}{\tau}_h}{\partial z} \qquad [14]$$

where $\underset{\sim}{U}$ is the mean velocity vector, $\underset{\sim}{V}_h$ is its horizontal component, f_c is the Coriolis parameter (f_C=1.454x10^{-4} sin (Latitude) sec^{-1}), $\hat{k}$ is the vertical unit vector, ∇_h the horizontal gradient operator, $\underset{\sim}{\tau}_h$ is the horizontal stress vector and P the pressure. A truly constant stress layer implies that all terms but the last in eq.[14] sum to zero. In general this does not occur, but under steady-state and horizontally homogenous conditions the left hand side vanishes leaving a

balance between the Coriolis term, the horizontal pressure gradient and the vertical stress gradient. Near the surface the stress decays with height and finally vanishes at the top of the planetary boundary layer ($\sim 0.25\, u_* / f_c$ Wyngaard, 1973). Near the surface the stress gradient is balanced by the horizontal pressure gradient.

$$\frac{\partial \overline{\rho u w}}{\partial z} = \frac{\partial P}{\partial x} \quad ; \quad (z << u_* / f_c) \qquad [15]$$

where U is now aligned with the surface wind so that the Coriolis term is negligible near the surface for this component.

At the top of the planetary boundary layer we have geostrophic balance:

$$f_c V_{gs} = \frac{1}{\rho} \frac{\partial P}{\partial x} \quad ; \quad (z > 0.25\, u_* / f_c) \qquad [16]$$

where V_{gs} is the component of the geostrophic wind normal to the surface wind. Ignoring the small density changes near the surface:

$$u_*^2 = f_c V_{gs} z + u_{*s}^2 \qquad [17]$$

or

$$u_*^2 = u_{*s}^2 \left(1 - \frac{\alpha_0 f_c z}{u_{*s}}\right) \qquad [18]$$

where use has been made of the result from planetary boundary layer theory and observation (Clarke, 1970; Panofsky, 1973) that $|V_{gs} / u_{*s}| = \alpha_0$ is a constant which for neutral

conditions is about 12. Thus the assumption that u_*^2 (measured) is the surface stress (u_{*s}^2) always introduces a systematic (and wind speed dependent) underestimate of the surface stress, which in some cases is in excess of 30%. Of course, eq.[18] may be used to correct this bias, although to my knowledge this is seldom, if ever, done in surface layer stress studies.

The constraints on time and space derivatives of the horizontal wind (left hand side of eq.[14] are also seldom checked. In fact the notion that a disequilibrium wave field could modify the surface roughness has led several researchers to examine the measured stress (at substantial heights) in conditions that are either unsteady and inhomogenous, or both--- across intense fronts for example. With the approximation eq.[18] of a quasi-constant stress layer, the constraint on time and space variability must be at most $f_c V_{gs}$,which at mid-latitudes is $\sim 5 \times 10^{-5}$ U or about 20% per hour or 5 m/s per 100 km. These are rather stringent constraints and if properly applied, would considerably reduce the number of acceptable estimates of the sea surface roughness --- and with it, no doubt, a good deal of the scatter! Of course, measurements made very close to the surface (for example from small buoys rather than ships) are less subject to errors of this sort that stem from the generally unjustified assumption of a constant stress layer.

3.3 Sampling Variability

The averaging process that leads to a mean equation of the type [14] is, strictly speaking, an ensemble average over a large number of realizations of a particular flow. In geophysics this is clearly impossible and we instead average over space or time and assume ergodicity, i.e. that for sufficiently wide averaging windows the time or space average approaches the ensemble average with any desired accuracy. As shown by Lumley and Panofsky (1964) the ratio of the variance of the estimates of the mean to the mean squared of some flow property y depends on the averaging time Ω (or distance), the integral time scale of the property $\mathcal{T}$ and the ensemble variance of y about its ensemble mean, $\overline{y'^2}$:

$$\epsilon^2 = \frac{Var}{\bar{y}^2} = \frac{2\,\overline{y'^2}\,\mathcal{T}}{\bar{y}^2\,\Omega} \qquad [19]$$

In practice the averaging time Ω cannot be arbitrarily large because the required conditions of stationarity and homogeneity will be violated. However, the process over which we are averaging must be contained within the time or space scales of the average if we are to have any hope of obtaining a meaningful average. This amounts to a requirement for a separation of scales of the boundary processes we wish to study and the large scale flows that drive them. Figure 5 (from Pierson, 1983) shows the spectra of downwind velocity fluctuations over water on time scales of tenths of a second to days --- a similar composite

spectrum was compiled by Van der Hoven (1957) over land. In both cases the microscale turbulence, produced by interaction of the flow with the boundary, is separated from the large scale flow variations by a distinctly weaker band of fluctuations. Taking the air-water interface results from Pierson (1983), obtained at 11.5 metres height, the best averaging time for air-sea turbulence studies appears to be about 20 minutes. The spectra of horizontal and vertical velocity fluctuations over water (adapted from Miyake et al., 1970), are shown in Figure 6. The spectra scale with distance from the boundary so that when plotted in "universal" coordinates, they collapse on a common curve. Since the scales of the boundary- generated turbulence increase with height, while the larger scales do not, the spectral gap tends to fill in as the measurement height increases, thereby exacerbating the difficulty of obtaining a convergent average. The large scale fluctuations of vertical velocity are suppressed
(Figure 6), but significant downstream velocity fluctuations remain at scales 100 times the anemometer height. Note that the appropriate velocity scale in the boundary layer similarity theory (Monin and Obukhov, 1954) is the friction velocity u_*, yet Miyake et al., (1970) scale their spectral ordinates by $\overline{w'^2}$, reflecting their feeling of the statistical uncertainty in the u_* estimates.

These "universal" spectra (see also Figures 7 and 8) compliment

those of Panofsky and Mares (1968) over land. The cospectra of momentum and sensible heat flux are shown in Figures 7 and 8. For a more complete discussion of universal spectral shapes over land and water, see Busch(1973)..

Equation [19] may be used to estimate the uncertainty in the estimates of various means and variances. Sreenivasan et al (1978) have collected appropriate data from an offshore tower and assessed the ratio $\overline{y'^2}/\bar{y}^2$ and the integral scales $\mathcal{T}$ for various flow variables and their moments. Their results may be written in the form $\epsilon(y) = \alpha_y\left(\frac{z}{U\Omega}\right)^{1/2}$. Their measured values of α_y for fluxes and variances are given in Table 1. Sample errors for typical values of height (10m) and wind speed (10 m/s) are also shown.

Table 1 Expected errors %

parameter, y	α_y	Expected % error, $100\epsilon(y)$ U=10m/s, Z=10m	
		10 min average	20 min average
$\overline{u^2}$	3.5	14	10
$\overline{w^2}$	1.7	7	5
$\overline{\theta^2}$	4.2	17	12
$\overline{q^2}$	3.3	13	9
$\overline{uw}$	5.5	22	16
$\overline{w\theta}$	8.0	33	23
$\overline{wq}$	6.6	27	19

In the following, we test the error in estimates of u_*^2 using a sequence of twenty minute averages drawn from seven hours of data in very steady conditions. Various meteorological and wave records are graphed in Figure 9. Each point represents a 20 minute average derived from samples obtained 5 times a second at a height of 10.8m. The average wind speed is 10.6 m/s and the extremes are within 11% of this. All other conditions show minor and smooth variations. In particular the water temperature changes by less than 0.5°C and the boundary layer stability (indicated by the bulk Richardson number, Rb---see below) remains essentially neutral throughout. By contrast, the measured drag coefficient shows pronounced fluctuation ($\pm$31% about an average of 0.00172). We wish to examine the variability of the $-\overline{uw}$ estimates. Since the mean wind does vary a little during the seven hour period we make use of the essentially quadratic behaviour of $-\overline{uw}$ on U (in this small range of U) to adjust the estimates of the former to that appropriate to the average wind speed of 10.6 m/s. Division of these by the average wind speed squared yields the C_{10} curve shown. The ratio of standard deviation to mean is 0.19 compared to 0.16 from Table 1.

It is apparent that single twenty minute averages yield rather inaccurate stress estimates. The accuracy is particularly poor in light winds and at substantial heights. This illustrates why measured u_* values are seldom used as a scaling variable; instead one seeks a parametric dependence of u_* on more easily and accurately measured variables such as the wind speed, wave

age et cetera. Once such a relationship has been found u_* may be calculated from the measured mean variables.

These error estimates may be tested against a large sample of stress estimates such as that collected by Large and Pond (1981) using the inertial-dissipation method (Fairall and Larsen, 1986). Large and Pond summarized 1591 hourly averaged stress estimates in 2 meter/second wind speed bins (their Figure 6). Two thirds of their data was obtained from a stable tower at 12.5 m and the rest from a ship at 22 m. The error estimates from eq.[19] for these two heights straddle their plotted standard deviations. Their data were obtained in "open ocean" conditions so that fetch (wave development) effects are not important. Stability corrections have been made so that the remaining scatter is largely due to sampling variability and is in close agreement with the estimates based on the work of Sreenivasen et al (1978). As we have pointed out, the most appropriate sampling interval for marine boundary layer fluxes is 20 minutes, but these are so inaccurate that it is wise to pool as many consecutive 20 minute samples as the constraint of stationarity will allow (Large and Pond pooled three 20 minute samples for their hourly averages). The height dependence of ϵ is the principal reason that laboratory measurements are subject to less sampling variability. For example, laboratory measurements (taken from Donelan, 1979) of $\overline{uw}$ averaged over 4-1/2 minutes at a height of 26.2 cm in an average wind of 6.54 m/s showed a standard error of 5.2% of the mean, whereas eq[19] yields 6.7%.

Businger et al (1971) noted the very much lower variability of surface drag plate stress estimates compared to those from acoustic anemometers at substantial heights over Kansas wheat stubble.

One of the reasons for the large sampling variability of estimates may be the formation of organized structures in the boundary layer such as roll vortices aligned with the wind (Tennekes, 1973b). These may be seen in Figure 10 in which very cold dry air flowing over water causes the evaporating water vapour to condense in the air. The downwind organization is a possible cause of the large sampling variability since it tends to increase the downwind coherency and hence the integral time scale $\mathcal{T}$ of the sampling process rendered by point observations in an advected mean flow. As remarked by Kraus (1968), one may obtain more rapidly convergent statistics from a crosswind travelling platform (ship or aircraft), but this does not seem to have been exploited. Fast travelling platforms have a potential advantage over fixed platforms in terms of the sampling error since U in eq.[20] is the fluid speed relative to the measuring apparatus which, for aircraft, may be an order of magnitude faster than the fluid speed.

4. Parameterizing the surface roughness

An important goal of air-sea interaction research---that of parameterizing the surface roughness---remains unrealized after more than three decades of quite intense field work. The notion that all the stress is supported by the high wavenumber part of

the spectrum, leading to a strictly wind speed (or friction velocity) dependent roughness length, cannot be correct in general although it does appear to be adequate for many open ocean cases. However, given the sources of error and sampling inaccuracies that plague even the most careful investigator, it is not altogether surprising that no consensus has been reached.

It has often been remarked that the widespread habit of relating the drag coefficient (a dimensionless number) to the wind speed is inconsistent and physically meaningless. Unfortunately, several attempts to rectify this have led to dimensionally correct relations which are flawed in another more subtle way. The problem arises when one attempts to deduce a relationship between two non-dimensional ratios both of which contain a measured variable. The severity of the problem depends on the level of measurement error of the common variable and large errors can lead to very impressive but essentially spurious correlations between the non-dimensional variables (Hicks, 1978, 1981; Kenney, 1982). The traditional velocity scale is the friction velocity u_* and, therefore, it frequently appears in non-dimensional combinations. As we have shown, the sampling error is particularly large for $\overline{uw} = -u_*^2$ and the literature abounds with correlations flawed on its account; for example: the drag coefficient, $(u_*/U_z)^2$ versus c_p/u_*. Many attempts to find the dependence of the sea surface roughness on wave development have been distorted in this way. The problem is exacerbated in several studies by failure to measure certain

aspects of the wave field and instead, to deduce them from the wind speed and fetch.

In the following, we attempt to avoid all these pitfalls with a selected data set in which the range of wave development is unusually wide. We will demonstrate that with sufficient care, the effect of the long waves on the roughness length may be determined and that the effect diminishes with increasing wave age, until at full development, Charnock's (1955) formula is recovered. The concept of an asymptotic surface roughness condition at full development, to which the dimensional arguments of Charnock apply, was advanced by Kitaigorodskii (1968). The data set is the neutral subset of that described in Donelan (1982), in which the stress was estimated by the eddy correlation method, using a Gill anemometer-bivane mounted at about 11 meters on a fixed platform in Lake Ontario. Here, neutral is taken to include cases in which the bulk Richardson number R_b is less than 0.01 in magnitude. R_b is defined by:

$$R_b = \frac{g z (\theta_z - \theta_s)}{\theta_z U_z^2} \qquad [20]$$

where θ is the potential virtual temperature, ^{o}K. The subscripts s and z refer to the surface and the measurement height respectively.

The bulk Richardson number may be expressed in terms of the Monin-Obukhov (1954) stability index, $\zeta = z/L$ via the bulk aerodynamic coefficients for the fluxes of momentum, heat and

water vapour: drag coefficient, C_D; Stanton number, C_H; and Dalton number, C_E.

$$\gamma = \kappa C_H C_D^{-3/2} R_6 \qquad [21]$$

in which C_E is assumed to be equal to C_H. The loss of accuracy in making this assumption is generally not significant. (See section 5.4)

By restricting the data to essentially neutral cases we avoid substantial corrections to the measured variables and are, therefore, less dependent on the accuracy of the stability functions (see section 5.7). A further requirement was that the 20 minute averaged wind speed exceed 5 m/s. This excludes the smooth cases (eq.[6]) and avoids the difficulties associated with propellor-vane systems in light winds (Busch et al, 1980). Non-overlapping groups of three consecutive twenty-minute averages were pooled to yield 52 independent sixty-minute averages. The measurements were made at heights of 10.8 to 11.6m and covered the (60 minute average) wind speed and inverse wave age range of 5.2 to 17.3 m/s and 0.8 to 4.6. So that the expected sampling variability of $\overline{uw}$ estimates varied between 9% and 16% (ratio of standard deviation to mean), while that of U was between 1.6% and 2.9%. Thus about 1/4 of the error in C_{10} estimates arises from inaccurate measurements of the mean wind speed. A further correction, to allow for the change of stress

with height (eq.[18]), amounted to an increase of 2% to 9% in the estimated surface value of $\overline{uw}$.

In general this correction is larger at low wind speeds. This field data set was supplemented with 14 $\overline{uw}$ measurements in a large laboratory tank (Donelan, 1979). The laboratory measurements were made a height of 26.2 cm and were corrected to the surface assuming a constant vertical stress gradient, i.e. a uniform increase of 40% was applied to all the laboratory $\overline{uw}$ estimates. The laboratory data covers a range of wind speed (extrapolated to 10m height) and inverse wave age of 5.5 to 21 m/s and 6.5 to 15.4.

The laboratory data and two groups of the field data, drawn from the tails of the wave age distribution, are compared in Figure 4 with Charnock's deduction (using m=0.014 i.e. Garratt's, 1977 result - rounded). For the same wind speed the young field waves are aerodynamically much rougher than the considerably larger mature (closed triangles) and fully developed (Large and Pond, 1981) waves. Yet the youngest of the lot, the laboratory waves, though generally rougher than the Charnock formula would indicate, are considerably smoother than the young field waves. The range of significant heights covered by these data varies from 3 cm to 4 metres. Figure 4 illustrates that Charnock's approach, while apparently a good model for open ocean data, does not fare at all well when a wide range of wave parameters is considered.

In order to illustrate the analogy to rough wall flow, Kitaigorodskii (1968) compared the results of two laboratory experiments (Kunishi, 1963; Hidy and Plate, 1966) and showed that for fully rough flow, the roughness length is proportional to the root-mean-square wave height, both non-dimensionalized via u_* and ν. As discussed above, this choice of non-dimensional variables will introduce some artificial dependence of one on the other, but the range of the plotted points (reproduced in Figure 11) is well beyond that which would be produced by sampling errors in u_* only. Our laboratory data (open circles) have been added and are in keeping with the trend of the original data. The field data (open and filled triangles and dashed line --- Large and Pond, 1981) follow the same trend ($z_0 \sim \sigma$) but the groups of points are displaced to the right, more and more with increasing wave age. The wind speed parameterization of Large and Pond (1981) has been transposed to Figure 11 via Bretschneider's (1973) full development relation ($\sigma = 0.0608\, U_{10}^2/g$). Evidently for aerodynamically rough flow, the roughness length and wave height are proportional and the constant of proportionality varies over several orders of magnitude depending on wave age. This matter has also been explored by Kondo et al., (1973) and Kuznetsov (1978). The data described above, covering a wide range of wave age, are graphed in Figure 12 in a manner designed to reveal the dependence of roughness length on both wave height and wave age. The three anomalously high points (more than two standard deviations above the regression line) are treated as

outliers and are not included in determining the regression lines of Figures 12, 13 and 14. The correlation coefficients are high but the two sets of data (laboratory and field) are not to be reconciled in this manner.

A possible cause of the difference between field and laboratory roughnesses (Figure 12) is the use of the wind speed referred to 10m, which certainly has no relevance for the laboratory work. A more consistent choice is the wind speed at a height commensurate with the wavelength of the waves at the spectral peak (Al-Zanaidi and Hui, 1984; Donelan and Pierson, 1987). In Figure 13 we have replaced U_{10} with $U_{\lambda/2}$ or the wind speed at half a wavelength above the mean surface. At this height, the pressure disturbance due to that wavelength (observed to be exponential in z/λ by Snyder et al, 1981) has nearly vanished, so that $U_{\lambda/2}$ is an appropriate U_∞ or reference velocity for both field and laboratory studies. This brings the laboratory and field data closer together but they remain distinct. In fact, no systematic choice of the height of U_z reconciles the laboratory and field data.

An indication of the scatter is provided by the vertical bars, which are two standard deviations in extent. The solid bars are the expected sampling variability, while the dashed bars are the vertical deviation of the points about the regression lines shown. In Figure 12, the Charnock relation (m_o = 0.014) is shown by the striped bar on the ordinate, which is at U_{10}/c_p =0.83 i.e. full development. The vertical extent of Charnock's z_o/σ

arises because of the observed variation in u_* with U_{10}; the extent of the bar covers the range of wind speeds for which $\overline{uw}$ has been measured for fully rough conditions (7.5 to 20 m/s). In Figures 13 and 14, Charnock's relation is indicated by a short dashed line, being the locus of points for which $U_{10}/C_p=0.83$ and U_{10} varies between 7.5 and 20 m/s. The very strongly forced waves in both field and laboratory yield roughness lengths of about the same fraction of mean roughness height ($=\sqrt{2\pi}\,\sigma$) as sand grains (Nikuradse, 1932) --- suggesting very strongly separated flow around the dominant waves in the system.

The same data sets have been replotted in Figure 14 in which the mean wind at some height has been replaced by the friction velocity. Correlations are substantially better, but a good deal of the improvement is a result of the common variable u_*, ($z_o = fcn[u_*/U_z]$). This is indicated by the tilt of the error bars derived from the sampling variability of u_* in both u_*/c_p and z_o.

Thus one may parameterize the sea surface roughness, for fully rough flow, in a manner consistent with the body of open ocean measurements, with the following simple regression formula:

$$\frac{z_o}{\sigma} = A_o \left(\frac{V}{c_p}\right)^{B_o} \qquad [22]$$

The values of A_o, B_o and the correlation coefficients for $V = U_{10}, U_{z/2}$ or u_* are listed in Table 2.

Table 2 Regression coefficients for surface roughness

	V	A_o	B_o	Corr.coeff.
Field	U_{10}	5.53×10^{-4}	2.66	0.83
Field	$U_{\lambda/2}$	3.70×10^{-4}	3.38	0.83
Field	u_*	1.84	2.53	0.92
Laboratory	U_{10}	9.76×10^{-6}	3.48	0.98
Laboratory	$U_{\lambda/2}$	1.31×10^{-5}	4.01	0.98
Laboratory	u_*	2.05×10^{-1}	2.18	0.99
Hsu (Field and laboratory)	u_*	0.637	2.00	

We have followed the approach of Kitaigorodskii and Volkov (1965) in parameterizing the roughness length in terms of "mobility" of the surface roughness or wave age. Hsu (1974) argued that the overall steepness σ/λ_p is the appropriate parameter. Both Hsu's work and this are based on deepwater waves primarily in which c_p and λ_p are deduced from the peak frequency and the choice of non-dimensional parameters from the set ($z_o, \sigma, \omega_p, U_z$) is a matter of physical reasoning. Hsu (1974) constructed a modification to Charnock's formula based on overall steepness, whereas we have used the state of wave

development to modify a solid wall model. In the end, Hsu obtains an expression of the form of [22], and his fit to a collection of various field and laboratory data is shown in Figure 14 and included in Table 2. Hsu's results appear to be in general agreement with the data of Figure 14, but the slope is somewhat less than suggested by these data and the roughness at full development is somewhat larger than that given by [6b]. Possibly the inclusion of laboratory data with field data reduces the slope of Hsu's regression line.

Kitaigorodskii's simplified formula [13] is also graphed on Figure 14. Most of the field data are quite well represented by [13], but near full development [13] underestimates z_0 considerably. This comes about through the exponential dependence on wind forcing at the peak in the simplified formula [13]. The spectral calculation [12] more correctly represents the distributed contribution to roughness, but few experiments yield sufficient data to compute the integral across the spectrum.

The use of laboratory experiments in air-sea interaction studies is very valuable in extending the range of governing variables and in reducing the sampling variability problem to any level desired since stationarity may be prescribed. However, one must interpret the results cautiously since there are many fundamental differences in the flow, the most obvious of which are the existence of side walls and a return flow in the water (Wang and Wu, 1987). Various attempts to include laboratory wave

data with field data have demonstrated substantial differences in the scaling properties of the two sets (Phillips, 1977; Donelan et al, 1985). Such differences may provide clues to the understanding of the underlying physics. It is instructive to compare the steepness of the large waves in both field and laboratory (Figure 15). The vertical coordinate of these time series has been normalized by the wavenumber of the peak deduced from linear theory ($k_p = \omega_p^2/g$), while the time has been normalized by ω_p . (The actual k_p may be slightly smaller --- up to 10%). Since the wind-sea spectrum is quite narrow, only the waves near the peak are obvious and the laboratory waves frequently approach the limiting steepness of Stokes' waves (Longuet-Higgins and Cokelet, 1976). With increasing U/c_p from field to laboratory, the steepness and number density of steep waves increases uniformly. The close packing of these steep, large waves reduces their effective height as roughness elements. It may well be that the groupiness of the waves contributes to their performance in establishing the aerodynamic roughness. The importance of three dimensional effects in the momentum input to waves has been explored by Stewart (1974) and by Csanady (1985); such effects are clearly inhibited in wind- wave tunnels.

In spite of considerable progress in understanding the mechanics of wave generation by wind (Phillips, 1957; Miles, 1957, 1959 a,b, 1962; Banner and Melville, 1976;Valenzuela, 1976; Plant, 1982; Landahl, 1985) we are yet unable to deduce the

surface roughness from the momentum transfer between wind and waves. The principal problem is that the momentum is distributed over the entire spectrum and our knowledge of the wavenumber spectrum of the short waves is rather rudimentary, although recent work (e.g. Banner et al, 1989; Shemdin and Hwang, 1988, and Shemdin et al, 1988) promises to correct this.

Lacking an adequate theoretical approach our estimates of the roughness of the sea surface must rest on experimental evidence guided by dimensional arguments and previous experience with boundary layers over solid surfaces.

The roughness of the sea surface for aerodynamically smooth and fully rough conditions may be described by eq.[6a] and eq.[22] respectively. The transitional regime between smooth and rough must, for the moment, be described by matching smooth and rough, so that roughness length is the larger of that given eq.[6a] and eq.[22].

This approach ignores the often observed ultra-smoothness of the sea-surface. The condition appears to occur in light winds and may, in some instances, be due to sampling variability which, as we have pointed out, is particularly large in light winds. However, in the vicinity of 5 m/s (see Phillips, 1977, Fig.4.27) the surface does appear to be smoothest and, indeed, appreciably smoother than a featureless plate. At least two explanations for this have been offered. Csanady (1974) has argued that the ultra smoothness of the surface in light winds arises because of "thickening of the laminar sub-layer due to energy transfer

associated with surface tension variations". While this is a possible explanation for ultra smoothness, it does not help in understanding the very occasional observations of upward momentum transfer (e.g. Davidson and Frank, 1973). These appear to be associated with loss of momentum from over-developed waves (i.e. waves travelling well beyond the speed of the wind in the boundary layer) leading to a "wave-driven wind" (Harris, 1966; Holland, 1981). An unusually clear observation (Holland, 1981) of this phenomenon is reproduced in Figure 16.

Although much remains to be explored both theoretically and experimentally, a consistent picture is emerging of the roughness of the wind-driven sea surface. In very light winds, the surface is aerodynamically smooth but the capillary-gravity ripples generated by shear flow instabilities (Valenzuela, 1976); Kawai, 1979; van Gastel et al, 1985) soon become large enough to disturb the viscous boundary layer (Kahma and Donelan, 1988). The aerodynamic characteristics are now transitional between smooth and rough but the partitioning of the momentum flux between viscous stress and form drag is not understood.

Through nonlinear wave-wave interactions (Phillips, 1960, 1961; Hasselmann, 1962,1963a, 1963b), the spectrum broadens and, being limited by dissipation at the short wavenumber end, the peak shifts to successively longer and faster wave components. The faster they travel, the weaker the direct interaction with the wind and so the burden of supporting the stress is borne by the relatively short waves. On the assumption that the limiting

ratio of roughness length to mean height of roughness elements for slow moving wave components is about 1/30 (Figure 13) and the high wavenumber spectrum is quasi-saturated (Banner, 1989), one finds that the bulk of the stress is supported by waves travelling at phase speeds $c < 5u_*$, as suggested by Phillips (1977). For full development c_p is about $30u_*$, so that the stress supporting waves are indeed very short compared to the energy containing waves.

Since the spectral density decreases continuously toward high wavenumbers and the phase speed increases beyond $k_{min}=(g/\gamma)^{1/2}$, it is probable that the capillary waves $k >> k_{min}$ contribute relatively little to the roughness and k_{min} can be regarded as an approximate upper limit to the roughness related wavenumbers. The lower limit is wind speed dependent ($k_o \sim g/25u_*^2$) so that changing the wind speed alters the width of the band of quasi-saturated waves contributing to z_o and yields Charnock's formula as shown by Phillips (1977), provided the waves are mature. Stronger forcing ($U/c_p >> 1$) has two effects: (a) the wavenumbers contributing to the roughness approach the peak wavenumber; (b) the steepness of the energy containing waves increases steadily (Figure 2, Figure 15 and Huang et al, 1981) so that even for moderate forcing ($U/c_p \sim 2$) there may already be a contribution to the roughness from the waves near the spectral peak as well as a contribution at higher wavenumbers ($k > 10\ k_p$). For sufficiently strong forcing

($U/c_p > 5$) these contributions overlap and the stress is probably supported largely by the energy containing waves (see Figure 3). Laboratory wind generated waves are very young ($U/c_p > 5$) and the principal cause of changes in z_0/σ probably arises from changes in the steepness of the energy containing waves as U/c_p is altered. At the other end of the development scale, the variations in z_0/σ probably arise mainly from the changes in width of the part of the short wave spectrum through which most of the stress is transferred. Young field waves include both these effects.

These comments are indirect inferences from observed aerodynamic properties of the air sea interface. We know that in fully rough flow, all of the momentum flux is transmitted to waves of various lengths (Stewart, 1961), but the actual distribution across the spectrum is unknown --- field experiments (Snyder et al, 1981; Hsiao and Shemdin, 1983; Hasselmann et al, 1986), have so far managed to explore only the region near the peak. Yet the matter is of the utmost importance in understanding and predicting the evolution of wind waves and the rate of kinetic energy input to the oceans (Mitsuyasu, 1985; Komen, 1987). Momentum transfer to the faster waves near the spectral peak implies considerably larger kinetic energy fluxes than if the very short waves are the principal stress receptors.

5. Heat and Mass Transfer

5.1 Introduction

Once the roughness characteristics of the air-sea interface are known, what can be said of its resistance to the transfer of heat and mass (including water vapour and various gases)? Air-sea interactionists are charged with the task of determining the interfacial fluxes in terms of mean variables i.e. finding appropriate parametric descriptions of the transfer coefficients in the so-called bulk aerodynamic formulae:

$$\overline{w\theta} = \frac{H}{\rho c_1} = -C_H(\Theta_z - \Theta_s)(U_z - U_s) \qquad [23a]$$

$$\overline{wq} = \frac{E}{\rho} = -C_E(Q_z - Q_s)(U_z - U_s) \qquad [23b]$$

$$\overline{wm} = \frac{F}{\rho} = -C_F(M_z - M_s)(U_z - U_s) \qquad [23c]$$

where H is the sensible heat flux, c_1 the specific heat at constant pressure, E the evaporation, F the gas flux; the subscripts z and s denote the measurement height and the surface respectively; θ is the potential temperature ($= t + \gamma_1 z$ where γ_1 is the adiabatic lapse rate 9.8×10^{-3} °c/m), q the specific humidity (Q_s its saturated value at the surface at temperature Θ_s), m the gas concentration. The bulk transfer coefficients are C_H (Stanton number) for heat, C_E (Dalton number)

for water vapour and C_F for gas flux. The fluxes are positive upwards.

The surface drift velocity U_S is usually ignored, thereby reducing the transfer coefficients by 2-3%. Of course, significant ambient currents other than wind drift should be included in U_S (Geernaert et al, 1986).

As before we assume that conditions are steady and homogeneous so that the surface boundary layer is a constant flux layer. The assumption breaks down for the heat flux in light winds, low temperature gradient and high humidity when the radiative heat flux is significant. Further, in strong winds the evaporation of spray in the boundary layer will affect both heat and mass fluxes (Bortkovskii, 1987).

Various attempts to describe the transfer coefficients, C_i have been based on hypotheses of mixing length, or surface renewal or other simple models of the flow adjacent to a solid wall or fluid of far greater density. Such heat and mass transfer "laws" seek a functional dependence of C_i on Z/Z_i; R_{e*}, Pr and γ; (where Z_i is the roughness length for heat or mass, $Re_* = Z_o\ u_*/\nu$ is the roughness Reynolds number, $Pr = \nu/\kappa_i$ is the molecular Prandtl or Schmidt number or the ratio of diffusivities of momentum and heat or mass. Generally these transfer laws have been based on careful laboratory experiments (e.g. Owen and Thompson, 1963; Kader and Yaglom, 1972; Yaglom and Kader, 1974; Brutsaert, 1975; Deacon, 1977; Liu et al, 1979; Kitaigorodskii and Donelan, 1984). In general they predict strong Prandtl

number dependence of the heat and mass coefficients and weaker dependence on roughness Reynolds number than shown by the drag coefficient C_D.

On the other hand, observations of heat and mass transfer from natural water surfaces provide a much less clear picture. By and large, transfer coefficients derived from such observations are widely scattered for many of the reasons given above for Reynolds stress. In addition, instrumental difficulties abound (Schmitt et al, 1978; Large and Pond, 1982) and the evaporation of spray in strong winds may drastically alter both sensible and latent heat fluxes. In a recent review of marine water vapour fluxes, Smith (1989) reported that the Dalton number dependence on roughness Reynolds number (graphed as wind speed) from several experiments may variously decrease, not change or increase. The observations fall in the general range of 0.001 to 0.002 in the wind speed range of 5 to 14 m/s. Relying strictly on observational evidence, Smith advocates a neutral Dalton number referred to 10m height in this wind speed range of 0.0012 ± 0.0001. Preliminary results from HEX05 (Katsaros et al, 1987), a recent experiment devoted to a better parameterization of C_E, show no change in C_E up to wind speeds of 18 m/s.

Field estimates of the Stanton number appear to be similarly scattered with no clear indication of a roughness Reynolds number dependence. Heat flux measurements are complicated by the difficulty of measuring the surface temperature, which may be

appreciably different from the bulk water temperature (Hasse, 1971), radiative effects and the problem in marine atmospheric boundary layers of faulty temperature readings due to accumulation of (hygroscopic) salt residues on the temperature sensor (Schmitt et al, 1978).

The practice of using field data to find relationships between the bulk coefficients and the wind speed, although dimensionally inconsistent, finds some justification in the case of the momentum transfer coefficient via Charnock's similarity arguments relating Z_o to u_*, but only for fully developed waves. No such direct correspondence between u_* and Z_i is possible because the mechanism for the transfer of heat or contaminants is quite different from that of momentum. Ultimately heat or mass transfer must occur by molecular contact between the fluid and the surface. Whereas, in transitional or rough flow, momentum is also transferred directly to the roughness elements via pressure differences between their windward and leeward faces and there is a concomitant increase in turbulence near the surface. The eddy diffusivity of momentum and all contaminants is thereby increased at least near the tops of the roughnesses, while the spaces between roughnesses are sheltered to some degree, and the diffusive boundary layers could be even thicker there than they would be on a smooth surface with the same friction velocity. Thus, depending on the relative importance of these two competing effects, resistance to heat and mass transfer may increase or decrease with increasing roughness Reynolds number, R_{e*}.

The difficulties and hazards of gathering field data are such that a complete data set, covering a wide variation of all parameters of interest to heat and mass transfer (Re_*, Pr, τ, γ, ν_w breaking waves of various scales, bubble and spray production etc.), may be long indeed in coming. On the other hand, laboratory experiments are incapable of simulating all aspects of air-sea interaction at once, but, properly designed, may yield valuable clues to certain aspects of the behaviour of natural air-water interfaces. For example, the importance of anthropogenic contaminants in affecting environmental quality demands a fuller understanding of the rates of transfer of such substances across natural air-water interfaces. Unlike temperature and water vapour, many of these substances have Prandtl numbers quite different from unity in the phase (air or water) that limits their transfer. Clearly, studying the Prandtl number, dependence of the bulk transfer coefficients is much more easily approached in the laboratory. Similar comments can be made about the effect of surface tension, wave breaking, bubbles, etc. One can envisage a process of laboratory experiments in which various aspects of the problem are isolated and examined and finally the results synthesized into a form that can be subjected to appropriate statistical tests using field data. Given the difficulty of covering a wide range of conditions in the field and the various sources of error and sampling variability therein, the alternative goal of using field

data directly to determine the heat and mass transfer characteristics of natural air-water interfaces appears to be beyond our present and (foreseeable) future wit.

5.2 Sub-layers

The heat and mass transfer problem is essentially one of determining the resistance to transfer of various contaminants imposed by the boundary layers on either side of the interface. Within each boundary layer there are three sub-layers in which different mechanisms operate. Adjacent to the interface molecular processes dominate in the viscous (momentum) or diffusive (heat or contaminant) sub-layer. The thickness of the diffusive and viscous sub-layers will be different when κ_i differs from ν ; i.e. Prandtl number differs from unity. For further information see Schlichting (1968). In general, this first sub-layer is very thin (of the order of a few millimetres or less) and not accessible to measurements in the field. At larger distances from the boundary turbulent mixing, produced by mechanical shear, dominates the transport processes. In this intermediate sub-layer (the "dynamic sub-layer") the profiles of momentum and all passive contaminants are logarithmic with distance from the interface. In the absence of density stratification, the dynamic sub-layer extends to measurement height and throughout the constant flux layer. Departures from the neutral logarithmic profile occur when the density stratification is sufficiently strong. Since the production of mechanical turbulence ($\overline{uw}\,\frac{\partial U}{\partial z}$) diminishes with distance

from the boundary (as the shear weakens), whereas that due to buoyancy ($\overline{w\theta_v}g/\Theta_v$) does not, the dynamic sub-layer gives place to a "diabatic" layer in which buoyancy forces act to increase (lapse or unstable stratification) or decrease (inversion or stable stratification) the shear induced turbulence. A commonly used measure of the outer limit of the dynamic layer (the Monin-Obukhov length, L) is obtained via a balance of the production/suppression of turbulence via buoyancy versus the production due to mechanical shear: $\zeta = \frac{z}{L} = -z\kappa g\,\overline{w\theta_v}/\Theta_v u_*^3$. Businger (1973), using normalized velocity shear data from Businger et al (1971), showed that buoyant and shear productions are equal (in the unstable case) when z = 0.57 L. We may take the outer limit of the dynamic sub-layer to be one fifth of this i.e. within the dynamic sub-layer buoyancy generated turbulence is at most 20% of that generated by shear. By virtue of [21] and [22] the dynamic sub-layer lies within δ_ν and δ_D .

$$\delta_D = \frac{0.29\ \Theta_z\ U_z^2}{C_H\ C_D^{-3/2}\ g\ (\Theta_z - \Theta_s)} \qquad [24]$$

For typical marine boundary layer wind speeds and air-water temperature differences this is of the order of metres. Within a distance δ_D of the boundary buoyancy effects may be ignored as is done in the following discussion of resistances to heat and mass transfer. We later return to a fuller discussion of the diabatic layer.

5.3 Resistance to heat and mass transfer

The molecular conductivity of heat or diffusivity of mass χ_i may be very different from the molecular viscosity ν, so that even in smooth flow, when the surface transfers of momentum, heat and mass are all effected entirely by molecular processes, the resistance to transfer in the molecular sub-layers may vary greatly from contaminant to contaminant. On the other hand, in the dynamic sub-layer, where turbulent mixing dominates the diffusivity of various passive contaminants, resistance is independent of Prandtl number. In order to separate the molecular processes from those due to turbulence, it is convenient to consider the resistance to transfer across the neutral air-water interface as the sum of four series resistances: two due to the diffusive sub-layers in the air and water respectively, and two due to the corresponding dynamic sub-layers. Using the subscripts s, δ and z to indicate the surface, the height of the junction of diffusive and dynamic sub-layers, and the measurement height respectively, the generalized resistances (difference/flux) in the diffusive (primed) and dynamic (double primed) layers above or below the interface may be written:

$$r_i = r_i' + r_i'' = \frac{\rho u_* (M_s - M_h)}{F} + \frac{\rho u_* (M_h - M_z)}{F} \qquad [25]$$

$$= \frac{M_h - M_s}{m_*} + \frac{M_z - M_h}{m_*}$$

where a scaling contaminant concentration m_* (or velocity, temperature or humidity) has been introduced [$\theta_* = -H/\rho c_p u_*$; $q_* = -E/\rho u_*$; $m_* = -F/\rho u_*$]. For clarity we restrict our attention to the boundary layers above the interface.

Field measurements in the diffusive sub-layers are generally unattainable so that attention is focused on the dynamic sub-layer where the profiles are logarithmic:

$$r_i = \frac{M_z - M_s}{m_*} = \frac{Pr_t}{k} \ln \frac{z}{z_i} \qquad [26]$$

The virtual origin of these profiles, Z_i is termed the roughness length for the property M and with Z_o (for momentum) is a sufficient description of the heat or mass transfer of passive contaminants in a neutrally stratified flow:

$$C_z = r_o^{-2} = k^2 / (\ln (z/z_o))^2 \qquad [27a]$$

$$C_i(z) = (r_o r_i)^{-1} = k^2 / \left(\ln\left(\frac{z}{z_o}\right) \cdot Pr_t \cdot \ln(z/z_i) \right) \qquad [27b]$$

$$i = \theta, q, m$$

Pr_t is the turbulent Prandtl number or turbulent Schmidt number, being the ratio of eddy diffusivities for momentum and heat or mass. The value of Pr_t has been explored in a wide range of experiments and appears to be in the range of 0.7 to 0.9 in the dynamic sub-layer. Kader and Yaglom (1972) find that the

consensus from the most reliable data yields $Pr_t = 0.85$. In the same spirit there appears to be little justification for allowing κ to stray much from the traditional value of 0.4.

In a neutral boundary layer at large distances ($Z \gg z_i$) from the surface, all passive (i.e. having no effect on the dynamics) admixtures are transferred by the same process of turbulent mixing and therefore must have the same eddy diffusivity, $K_i = K_p$ (where the subscript denotes the eddy diffusivity of a passive admixture). Furthermore, in the part of the boundary layer in which $U(z)$, $\Theta(z)$, $Q(z)$ and $M(z)$ are logarithmic Pr_t cannot be a function of Z. Momentum is transferred both by turbulent mixing and by pressure gradients and, therefore, the turbulent Prandtl number (K/K_i) need not be unity. Further, in a diabatic boundary layer, the buoyancy forces produced by temperature and humidity differences may make K_θ and K_q different from other K_i.

5.4 <u>Some models of heat and mass transfer.</u>

While it is simpler to make observations in the dynamic sub-layer, the resistance to transfer in the diffusive sub-layer r_i' is Prandtl number dependent and generally much larger than r_i''. Thus several laboratory results are couched in terms of sub-layer charactistics; e.g. the sub-layer transfer coefficient (Dipprey and Sabersky, 1963; Owen and Thompson, 1963).

$$C_i' = \frac{F}{\rho u_*(M_s - M_\delta)} = (r_i')^{-1} \qquad [28]$$

By virtue of [25], [26] and [27], the relationship between the bulk and sub-layer transfer coefficients C_i and C_i' is:

$$C_i = \left[r_D \left\{ (C_i')^{-1} + r_D P_{r_t} - \frac{U_\delta}{u_*} P_{r_t} \right\} \right]^{-1} \qquad [29]$$

$$= C_D \left[\sqrt{C_D} / C_i' + P_{r_t} \left(1 - U_\delta / U_z \right) \right]^{-1}$$

The last term depends on the lower limit of the dynamic sub-layer and is subject to some choice. The sub-layer resistance r_i' depends on Prandtl and roughness Reynolds numbers and is usually expressed in the form: $r_i' = \beta R_{e_*}^m P_r^n$; where β may depend on the type and spacing of the roughness elements. Various attempts to establish β, m and n for smooth or fully rough flow have yielded similar Prandtl and roughness Reynolds number tendencies but with differences in detail. Using the concept of random renewal of surface material by intermittent ejection and inrush of fluid from and to the surface (as observed by Grass, 1971)[21], Brutsaert (1975), argued that r_i' should be proportional to $Pr^{2/3}$ for smooth flow and $Pr^{1/2}$ for rough. He was able to reconcile the laboratory data from several sources by judicious choice of U_δ, β, m and n. His results for the sub-layer resistance are:

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[21] Papadimitrakis et al (1987) have observed similar 'bursting' phenomena in the boundary layer over progressive waves in a tank.

$$r_i' = 13.6\ Pr^{2/3} \quad ; \quad Re_* = 0.135$$

[30]

$$r_i' = 7.3\ Re_*^{1/4}\ Pr^{1/2} \quad ; \quad Re_* > 2$$

and corresponding bulk transfer coefficient:

$$C_i(z) = \frac{C_D}{13.6\ C_D^{1/2}\ Pr^{2/3} + Pr_t\left(1 - 13.5\ C_D^{1/2}\right)} \quad ; \quad Re_* = 0.135$$

$$C_i(z) = \frac{C_D}{7.3\ C_D^{1/2}\ Re_*^{1/4}\ Pr^{1/2} + Pr_t\left(1 - 5\ C_D^{1/2}\right)} \quad ; \quad Re_* > 2$$

[31]

Liu et al (1979) using a similar approach find that $r_i' = 16\ Pr^{1/2}$ for smooth flow and $r_i' = \beta\ Re_*^{1/4}\ Pr^{1/2}$ for rough flow. β is acknowledged to be dependent on roughness characteristics and a value of 9.3 agrees well with the laboratory heat and water vapour transfer measurements of Mangarella et al (1973) over paddle generated waves. The models of Brutsaert and Liu et al are in close agreement for rough flow and, although the Prandtl number dependencies for smooth flow are different, the differences are not significant for heat and water vapour transfer - Prandtl numbers of 0.7 and 0.6 respectively.

Kader and Yaglom (1972) and Yaglom and Kader (1974), for smooth and rough flow respectively, use similarity arguments to postulate essentially $Pr^{2/3}$ dependence of the sub-layer resistance for large Prandtl numbers. For rough flow they find

that $r_i' \propto Re_*^{1/2}$. Dipprey and Sabersky (1963) and Owen and Thompson (1963) both explored the sub-layer resistance in rough flow and found that $r_i' \propto Re_*^{0.2} Pr^{0.44}$ and $r_i' \propto Re_*^{0.45} Pr^{0.8}$ respectively.

There seems to be general agreement that the sub-layer resistance to transfer increases with Prandtl number (exponents of 0.44 to 0.8) and roughness Reynolds number (exponents of 1/5 to 1/2). Garrett and Hicks (1973), however, showed that a weak negative dependence of r_i' on Re_* is observed over natural vegetation, where the mechanics of the flow may be modified by the orientation and flutter of the individual elements (leaves, twigs, etc). Undoubtedly, the aerodynamic characteristics of such surfaces are quite different from those of a wavy water surface, which may find a better analog in the work over bluff bodies reported above. Nonetheless, a great deal of boundary layer research over natural vegetation is transported to the marine boundary layer, where, evidently, it should be applied with some caution.

The principal shortcoming of these semi-empirical approaches to determining the resistance to heat and mass transfer is that they have been devised for smooth or fully rough flow, whereas the ocean surface roughness is transitional a good deal of the time (2.8 m/s < U_{10} < 7.5 m/s for fully developed and neutrally stratified conditions). To fill this void Kitaigorodskii and Donelan (1984) proposed a mixing length model for heat and mass

transfer for those gases in which the principal resistance is in the air phase. Their model uses the formulation of Riley et al (1982) for momentum and extends it to allow the surface mixing length for contaminants l_{ms} to depend on Prandtl number. In addition to covering all dynamic roughness ranges continuously, this model, based on Van Driest's (1956) profile, treats the various sub-layers as a continuum, thereby eliminating the rather arbitrary choice of U_δ in the sub-layer resistance models. In their model, the contaminant mixing length, l_m is given by:

$$l_m = l_{ms} + Pr_t^{-1/2} \kappa z \left[1 - \exp(-z u_* / 13\nu)\right]^2 \qquad [32]$$

where the mixing length l_m is defined in the usual way:

$$\rho u_* m_* = -F = (K_i + K_i) \frac{dM}{dz}$$
$$= \left(K_i + l_m^2 \frac{dU}{dz}\right) \frac{dM}{dz} \qquad [33]$$

Kitaigorodskii and Donelan found that

$$l_{ms} = 0.54 \, Pr^{-0.39} \, l_{us} \qquad [34]$$

from the data of Moller and Schumann (1970) obtained in a wind-water tank over a wide range of Prandtl number (0.6 to 8500) and only one roughness Reynolds number ($Re_* = 5.3$). Kitaigorodskii and Donelan (1984) did not explore the possibility that roughness Reynolds number might influence the ratio of surface mixing

length for a contaminant, l_{ms} to that for momentum, l_{us}. Such a dependence is certainly supported by all the heat and mass transfer work discussed above and may be readily incorporated in the surface mixing length:

$$l_{ms} = \beta \, R_{e_*}^{m} \, Pr^{-0.39} \, l_{us} \qquad [35]$$

with the constraint that $\beta \, (5.3)^m = 0.54$. The choice of $m = -0.5$ yields good agreement with the field estimates of Dalton numbers.

The Dalton numbers from several of these semi-empirical models of heat and mass transfer are compared with the results of various field experiments summarized by Smith (1989). The roughness of the sea surface for fully developed conditions is as given by a smooth curve joining the smooth and rough asymptotes of equation [6]. The observations tend to support the view that increasing surface roughness, while increasing the turbulence levels near the surface, also increases the fractional area protected in the lee of the roughnesses. Thus, heat and mass transfer are accelerated near the crests and windward sides of the waves and retarded in the troughs and on the leeward sides. The net effect appears to be fairly constant heat and mass

transfer coefficients at least at moderate (4 m/s to 18 m/s) wind speeds. There is a clear need for further laboratory testing of heat and mass transfer at the air-water interface over a wide range of Prandtl and roughness Reynolds numbers.

5.5 The effect of spray on heat and mass transfer.

It has often been argued that the evaporation of water droplets in the near surface layer will lead to significant enhancement of the Dalton number (Wu, 1974); Ling and Kao, 1976; Ling et al, 1978; Resch and Selva, 1978; Bortkovskii, 1987) but there are dissenting opinions (e.g., Street et al, 1978). During moderate winds, the principal source of liquid water in the atmospheric boundary layer appears to be associated with the bursting of air bubbles formed during wave breaking both in the laboratory (Toba, 1962) and in the field (Monahan, 1968). In strong winds (25 m/s and greater) the water at the crests of steep waves is sometimes detached and blown into spume.

The presence of water droplets in the air stream certainly exposes more water surface to direct contact with the air, but the overall evaporation rate may not increase correspondingly unless there is an adequate supply of heat to replace the latent heat of vapourization of the droplets. If the droplets are large, they may provide the necessary heat of vapourization to the mass lost during the relatively short time they are suspended in the air stream. They return to the surface somewhat cooler so that the latent heat transfer is principally from the surface to

the vapour released during the flight of the large droplets. At the other end of the scale, the very small droplets evaporate completely during flight and thereby cool and moisten the air. This tends to increase the upward sensible heat flux in unstable conditions (or reduce the downward flux in stable conditions) and to reduce the water vapour flux directly from the surface. In this context, the small droplets are those with settling speeds w_f smaller than the root-mean-square vertical turbulence velocity. Byutner (1978) finds that particles remain suspended in a turbulent flow if $u_* > 2.5\, w_f$. The relative importance of small diffusing particles and larger ones that return to the surface with mass deficit and momentum excess may be deduced from the observed size distribution of spray droplets. According to Bortkovskii (1987) the distribution follows a Nukiyama-Tanasawa relation (Wallis, 1969) of the form (Borisenkov et al, 1974):

$$F(R) = 4 \frac{R^2}{R_m^3} \exp\left(-\frac{2R}{R_m}\right) \qquad [36]$$

where R is the drop radius and R_m its value at the mode of the distribution.

Bortkovskii (1987) has carried out a rather complete analysis of the dynamics and thermodynamics of spray formation and evaporation, and concludes that spray evaporation is insignificant below 9 m/s, doubles the Stanton and Dalton numbers at about 18 m/s and overwhelms the surface transfers in gale force winds. The field measurements of evaporation in open sea

conditions show essentially no wind speed dependence of the Dalton number (Figure 17). While most measurements have been at relatively low wind speeds, those of Large and Pond (1982) cover the range of 4 to 14 m/s and the recent concerted efforts of many air-sea interaction researchers (Smith and Anderson, 1988) have extended the results to 18 m/s.

Bortkovskii (1987) treats the problem of spray evaporation as a statistical one in which the volumetric concentration of droplets is sufficiently low that they may be treated separately with regard to their trajectories and heat exchanges. He further finds that the electric forces acting on charged droplets are insignificant. The radii of droplets fall in the range of 10 μm to $10^3 \mu m$ (Wang and Street, 1978). Bortkovskii suggests that 30 $\mu m < R_m < 50 \mu m$ and that the principal evaporation enhancement comes from droplets that do not evaporate completely but return to the surface after a short time of order 0.3 seconds. All but the largest droplets equilibrate more rapidly than this so that the exchange of heat between the droplet and surroundings is principally by forced convection with radiation being relatively unimportant.

The action of ejecting droplets of water into the air can be viewed as a mechanism for bypassing the relatively high resistance of the diffusive sub-layer by placing evaporating surfaces directly in the dynamic sub-layer. For example, a saturated layer of depth 5 cm could cause an increase in the Dalton number (equation [27b]) of more than a factor of 2. This

does not appear to happen, perhaps because bulk transfer coefficients are computed using mean values in the outer flow on each side of the boundary and the evaporating spray reduces the air temperature near the surface and, therefore, the saturation humidity. Eventually the continuing heat drain from air to airborne spray must be drawn from the water surface[22]. The surface undergoes further cooling and consequently reduced upward flux of both sensible and latent heat. The system is to some degree self-limiting and a full understanding of the thermodynamics of the wind blown ocean surface remains a challenge deserving of vigorous pursuit.

5.6 The cool skin of the ocean

The calculation of air-sea exchanges via bulk coefficients (equation [23]) has been discussed in terms of differences of means between some measurement height and the surface. Surface measurements are difficult to obtain and one is generally forced to work with "bucket" values; that is, samples obtained from some shallow depth traditionally by means of a bucket. These values are, of course, not the surface values and a suitable correction must be made to allow appropriate use of the various heat and mass transfer models discussed. Of course, one could model the coupled boundary layers (see, for example, Street et al, 1978) so that the surface values would not be required as input to the

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[22]The heat capacity of a 10m column of air is about the same as that of 3 mm of water.

calculation, but instead, the bulk values in both fluids would suffice. This approach will eventually be commonplace, but, at present, understanding of the heat and mass transfer properties of the aqueous boundary layer lag far behind that of the air boundary layer.

Thus in the calculation of heat and vapour fluxes across the air-sea interface using [23] one should adjust for the difference of the sea surface temperature from the bulk water temperature. The surface may be warmer or cooler than the bulk depending on the direction of flow of the near-surface heat flux (incoming solar, sensible, latent and net long wave radiative). In clean water this is generally upward (yielding a cooler surface) because the divergence of incoming solar radiation is small. This "cool skin" (Woodcock, 1941; Montgomery, 1947; Hasse, 1963; Katsaros, 1977) may be sufficiently different (several tenths of a degree Celcius) from the bulk to bias the heat flux appreciably.

Various attempts to explore the magnitude and causes of the surface-to-bulk temperature difference include those by Saunders (1967), Hasse (1971), Wu (1971), Paulson and Parker (1972) and Katsaros (1977). The general goal is to produce a prescription for correcting the observed bulk water temperature to the surface. Hasse (1971) has succeeded in providing the simplest relation that appears to be in general accord with field observations. Hasse's formula is given as the sum of two terms with coefficients that are dependent on the depth of the bulk

measurement. It may be written in dimensionally consistent form using Saunders' (1967) dimensional analysis:

$$(T_s - T_z)_w = -\left(\frac{\nu}{\kappa_t u_* \rho c_1}\right)_w \left[\alpha_1 T_s + \alpha_2 Q_I\right] \qquad [37]$$

where z is the depth of the bulk measurement, κ_t is the thermal conductivity, T_s the sum of sensible, latent heat and net long wave radiation from the surface and Q_I is the short-wave (sun plus sky) direct radiation; T_s and Q_I are positive upwards; the subscript w denotes water. The coefficients α_1 and α_2 are determined empirically and are dependent on bulk measurement depth: α_1 because the temperature gradient, although concentrated near the surface, is not confined to it; α_2 because of the divergence of short-wave flux as the incoming radiation is absorbed. The numerical values in Table 3 are adapted[23] from Hasse (1971) and are for clear water.

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23 α_1 and α_2 were computed from Hasse's coefficients using $C_{10}=1.21\times10^{-3}$ and values of κ_t and ν corresponding to water temperature of 10°C. Note that ν and κ_t change in opposite directions with temperature such that the Prandtl number $(\nu/\kappa_t)_w$ changes by more than a factor of 2 over the full range of ocean water temperatures, principally through the variation of viscosity ν.

Table 3

Variation of the coefficients α_1 and α_2 with depth z of bulk water temperature measurement.

Depth z (m)	-0.25	-0.5	-1.0	-2.5	-5.0	-10.0
α_1	6.9	7.0	7.2	7.5	7.7	7.8
α_2	1.2	1.3	1.4	1.6	1.7	1.8

The model is based on forced convection in a turbulent wall layer, and therefore it will not be accurate for very low wind speeds (<2 m/s) or for high wind speeds, where wave breaking disturbs the surface layer (>8 m/s).

It is of interest to examine the error in using [23] for bulk heat flux calculations without compensating for the skin temperature [37]. For illustration, we ignore (both short and long) radiative effects, so that $T_s = H(1+B^{-1})$ where B is the Bowen ratio (sensible heat flux/latent heat flux). The error is $Pr_w\ C_H\ C_D^{-1/2}\ c_i (c_i)_w^{-1}\ (\rho/\rho_w)^{1/2}\ \alpha_1\ (1+B^{-1})$ and at moderate wind speeds, where C_{10} and C_H are roughly 0.0012, this may be as small as 8% at high latitudes where the Bowen ratio is about 0.45 (Sverdrup, 1951) and as large as 13% in the tropics where a Bowen

ratio of 0.1 is more typical[24]. The error in the latent heat flux calculation is smaller and usually negligible.

5.7 The diabatic profile

So far we have treated the neutrally stratified boundary, i.e. one in which the effects of buoyancy are negligible. When there is a mean density difference across the constant flux layer, the relative importance of buoyancy increases with height as the mechanical shear weakens. The balance between shear generated and buoyancy generated or suppressed turbulence introduces a new length scale, the Monin-Obukhov (1954) length. Including both temperature and humidity related buoyancy effects, this may be written (Zilitinkevich, 1966)

$$L = \frac{-u_*^3}{\kappa g \left[\overline{w\theta}/\Theta + 0.61\,\overline{wq}\right]} \qquad [38]$$

At heights greated than about 0. 1 L the dynamic sub-layer evolves into the "diabatic sublayer". Here, according to the

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24 α, for bulk water temperature at 1.0m depth was used. The variation in Prandtl number with temperature reduces the range of error. In addition, the wind stress is assumed to be communicated to the surface currents. A fraction (up to 20%, Donelan, 1979) may remain in the wave field leading to smaller $(u_*)_w$ and somewhat larger errors in bulk heat flux calculations.

similarity theory of Monin and Obukhov, the mean gradients[25] are universal functions of $\zeta = \frac{z}{L}$:

$$\frac{\partial U}{\partial z} = \frac{u_*}{\kappa z} \Phi_u(\zeta) \qquad [39]$$

$$\frac{\partial \Theta}{\partial z} = \frac{\theta_*}{\kappa z} \Phi_\theta(\zeta) \qquad [40]$$

$$\frac{\partial Q}{\partial z} = \frac{q_*}{\kappa z} \Phi_q(\zeta) \qquad [41]$$

$$\frac{\partial M}{\partial z} = \frac{m_*}{\kappa z} \Phi_m(\zeta) \qquad [42]$$

The non-dimensional gradients $\Phi(\zeta)$ must be determined empirically from the flux-profile relations [39] to [42]. They are related to the eddy diffusivities K_i by:

$$K_i = u_* \kappa z / \Phi_i(\zeta) \qquad [43]$$

so that the turbulent Prandtl number is:

$$Pr_t(\zeta) = \Phi_i(\zeta) / \Phi_u(\zeta) \qquad [44]$$

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[25]All single-point turbulence parameters are universal functions of ζ . We restrict our attention here to the mean values, since they are essential to the determination of boundary fluxes. For a further discussion of other turbulence parameters see, for example, Busch (1973), Businger (1973), Panofsky (1973) and Wyngaard (1973). These are part of a set of excellent papers on micrometeorology edited by Haugen (1973).

by virtue of [4] and [26] the neutral values of the non-dimensional gradients are $\Phi_u(0)=1$; $\Phi_\theta(0) = \Phi_q(0) = \Phi_m(0) = 0.85$.

The differences may be obtained by integration between the roughness height z_i and the measurement level z:

$$U_z - U_s = \frac{u_*}{\kappa}\left[\ln(z/z_0) - \Psi_u(\zeta)\right] \qquad [45]$$

$$M_z - M_s = \Phi_m(0)\frac{m_*}{\kappa}\left[\ln\left(\frac{z}{z_i}\right) - \Psi_m(\zeta)\right] \qquad [46]$$

where

$$\Psi_u(\zeta) = \int_{\zeta_0}^{\zeta} \frac{[1 - \Phi_u(\zeta)]}{\zeta}\, d\zeta \qquad [47]$$

$$\Psi_m(\zeta) = \int_{\zeta_i}^{\zeta} \frac{[1 - \Phi_m(\zeta)/\Phi_m(0)]}{\zeta}\, d\zeta \qquad [48]$$

and $\zeta_0 = z_0/L$; $\zeta_i = z_i/L$

Many forms of the non-dimensional gradients have been proposed; most based on measurements over land. Although the accuracy of most measurements is inadequate to permit the selection of one form over another, common usage seems to favour the approach of Dyer and Hicks (1970)[26]. The most complete measurements, including both stable and unstable conditions were made by Businger et al (1971). However, their conclusion that

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[26]Other flux-profile relationships are reviewed by Dyer (1974), Yamamoto (1975) and Yaglom (1977).

the von Karman constant κ and neutral turbulent Prandtl number P_{rt} were 0.35 and 0.74 has not been accepted, and some doubt regarding effects of flow distortion on the flux measurements remain. Forced agreement with the generally accepted values of 0.4 and 0.85 yields the following non-dimensional gradients (Kitaigorodskii and Donelan, 1984):

$$\left.\begin{aligned} \Phi_u(\zeta) &= (1-17\zeta)^{-1/4} \\ \Phi_m(\zeta) &= 0.85\,(1-10\zeta)^{-1/2} \end{aligned}\right\} \quad \zeta < 0 \qquad [49]$$

$$\left.\begin{aligned} \Phi_u(\zeta) &= 1 + 5.4\,\zeta \\ \Phi_m(\zeta) &= 0.85\,(1 + 7.3\,\zeta) \end{aligned}\right\} \quad \zeta > 0 \qquad [50]$$

These expressions for Φ_u and Φ_m fall within the range of commonly used expressions summarized by Yaglom (1977).

The profile parameters Ψ are therefore (Paulson, 1970):

$$\left.\begin{aligned} \Psi_u(\zeta) &\doteqdot 2\ln\left(\frac{1+X}{2}\right) + \ln\left(\frac{1+X^2}{2}\right) - 2\tan^{-1}X + \pi/2 \\ \Psi_m(\zeta) &\doteqdot 2\ln\left(\frac{1+Y}{2}\right) \end{aligned}\right\} \quad \zeta < 0 \qquad [51]$$

where $X = (1-17\zeta)^{1/4}$

and $Y = (1-10\zeta)^{1/2}$

$$\left.\begin{aligned}\Psi_u(\zeta) &\doteq -5.4\,\zeta \\ \Psi_m(\zeta) &\doteq -7.3\,\zeta\end{aligned}\right\}\ \zeta > 0 \qquad [52]$$

The approximations in [51] and [52] are nearly exact since z_{δ} and z_i are very small.

These relations [39] to [52] provide the necessary information to deduce the boundary stress and fluxes from observed differences provided the neutral bulk transfer coefficients are known. A typical approach is given by Large and Pond (1982).

In this discussion of the diabatic layer all contaminants have been assumed to behave like passive admixtures and hence to have identical properties above their diffusive sub-layers. This is very definitely not so in the trade wind measurements of Phelps and Pond (1971) where the differences are attributed to radiative effects in a moist boundary layer. Another possible explanation is that the large scale convective motions impose downdrafts of warm dry air on a surface boundary layer in which the local gradients (negative for both temperature and humidity) create warm moist correlations of the smaller scale motion (Donelan and Miyake, 1973). Consequently, the spectra and fluxes of temperature and humidity are very dissimilar and the simplified diabatic profiles discussed here are not applicable. The density gradient in the surface layer is determined largely by humidity and the spectrum of humidity and of the moisture flux are of the "universal" form (see Figure 8), while the temperature spectrum and flux are very different. Furthermore, local

gradients of temperature yield poor estimates of the heat flux (Paulson et al, 1972). Thus one must be careful about the application of this simplified similarity theory to real marine atmospheric boundary layers, where processes aloft can affect the gradients associated with the larger scale motions and thus decouple the fluxes from the local surface boundary gradients. A further source of concern is the evaporation of spray droplets in the surface boundary layer. This may make the heat and moisture fluxes quite height-dependent and invalidate the assumptions that lead to the diabatic profile [45], [46]. Careful measurement of the flux-profile relations above the air-water interface remains a priority of air-sea interaction research.

The diabatic drag coefficient is from [45]:

$$C_D(\zeta) = \kappa^2 \left[\ln z/z_0 - \Psi_u(\zeta) \right]^{-2} \qquad [53]$$

and the bulk coefficient for passive admixtures based on the sub-layer models [29] is:

$$C_i(\zeta) = \kappa^2 \left[\left\{ \ln z/z_0 - \Psi_u(\zeta) \right\} \cdot \left\{ \tau_i' + Pr_t(0) \left\{ \ln \frac{z}{z_0} - \frac{U_\delta \kappa}{u_*} - \Psi_m(\zeta) \right\} \right\} \right]^{-1} \qquad [54]$$

In the mixing length model the diabatic forms are given by virtue of [43]:

$$l_u = l_{us} + \Phi_u^{-1}(\zeta)\, \kappa z \left[1 - \exp(-z u_*/13\nu) \right]^2 \qquad [55]$$

$$l_m = l_{ms} + \left(\Phi_u(\gamma)\cdot\Phi_m(\gamma)\right)^{-1/2} \kappa z \left[1-\exp\left(-\frac{z u_*}{13\nu}\right)\right]^2 \quad [56]$$

5.8 The aqueous boundary layer

While considerable success has been achieved in treating the air-sea interface viewed from above by analogy with a rigid porous wall, such an approach is not fruitful in modeling the charactistics of the other side of the interface - the aqueous boundary layer. Even in relatively calm conditions, it is not clear that a classical wall layer will exist beneath the interface for, while the large air-water density contrast supresses vertical motion adjacent to the interface, the horizontal water motion is essentially unconstrained. Thus, instead of simple wall layer scaling, based on a single velocity (u_*) and length scale (Z), another velocity scale, associated with the horizontal unconstrained motions, is required.

In all but the lightest winds, the wave induced velocities are substantially larger than the turbulence or even (except at very short fetches) the mean flow. The interaction between the shear-induced turbulence and the largely irrotational wave motion, though weak, introduces additional wave-related length and velocity scales. Kitaigorodskii and Lumley (1983) and Kitaigorodskii et al (1983) have explored the effects of wave turbulence interactions both theoretically and experimentally.

Finally, the onset of wave breaking in moderate winds brings about a dramatic change in the characteristics of the aqueous boundary layer. The breaking process injects a sudden burst of kinetic energy and momentum beneath the surface (Donelan, 1978; Melville and Rapp, 1985; Longuet-Higgins, 1988) and produces a "cloud" of bubbles, which are believed to be the principal source of marine salt aerosols (Blanchard, 1983; Resch, 1986; Monahan et al, 1986) and to play an important role in the transfer of many gases across the air-water interface (Thorpe, 1982; Memery and Merlivat, 1986).

The air-water transfer of many important gases in the biogeochemical cycle is controlled by resistance in the water phase. The diffusion of gases in air is typically 10^4 times that in water so that unless the gas is highly soluble or reactive in water, it is water phase controlled. Laboratory wind-water tunnel experiments have demonstrated that, while gas phase controlled fluxes are approximately proportional to the wind speed (i.e. constant bulk transfer coefficient), water phase controlled fluxes are more strongly dependent on wind speed. The diffusive sub-layer in the water is so thin that the breaking of even the small gravity-capillary waves that form at quite low wind speeds (Kahma and Donelan, 1988) is sufficient to weaken its resistance. Kerman (1984) estimates that these ubiquitous small breakers cover much more surface area than the visible whitecaps.

Exploring the aqueous boundary layer and its resistance to mass transfer are currently very active research areas (see for

example: Broecker and Hasse, 1980; Liss and Slinn, 1983; Brutsaert and Jirka, 1988; and Brumley and Jirka, 1988). Much of the work to date is not directly applicable to the effects of wave breaking, but the pioneering work of Kitaigorodskii (1984) is a substantial step in the right direction.

In recent years there has been a considerable increase in the thrust to understand the structure of turbulence in the near surface waters, both in the laboratory and the field. A good sample of this work is reported in Toba and Mitsuyasu (1985). Some investigators find that the turbulent characteristics are in keeping with wall layer scaling both in the field (Jones, 1985) and the laboratory (Mitsuyasu and Kusaba, 1985) while others find significant differences (Cheung and Street, 1988 - laboratory; Terray et al, 1989 - field).

The study of wave breaking and near surface turbulence is probably the most exciting and rapidly expanding aspect of air-sea interaction. It derives its new-found impetus from its application to many issues of ocean science or engineering interest: gas transfer and aerosol production; many aspects of active and passive remote sensing (see Toba and Mitsuyasu, 1985; Phillips and Hasselmann, 1986; Geernaert and Plant, 1989); oceanic acoustics (see Kerman, 1988); wave prediction (Komen et al, 1984); upper ocean dynamics (Huang, 1986). A parallel interest in the effects of surface contaminants is also expanding (Scott, 1986; Hogan, 1986; Bortkovskii, 1987; Alpers and Huhnerfuss, 1989). Surface contaminants preferentially

attenuate shorter waves and thus alter the characteristics of breaking and influence the many issues listed above.

The last twenty years have seen substantial progress in understanding and parameterizing the roughness of the sea surface and the characteristics of the air boundary layer. While there is yet much to be learned here, the widest gaps in our knowledge of air-sea interaction are in the water boundary layer and its intimate relationship with surface waves. Fueled by the requirements of remote sensing, ocean acoustics, gas transfer and wave prediction, and stimulated by the inherent fascination of the truly complex and intricate interplay of many physical processes, the exploration of wave breaking and its effects on the aqueous boundary layer is growing rapidly. It is here that one should expect to see the major advances in air-sea interaction in the next ten or twenty years.

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SYMBOLS

Subscripts

z	at height Z
s	at surface
w	water
p	peak of spectrum
h	horizontal component
$\sim$	vector
u	momentum
θ	temperature
q	specific humidity
m	gas concentration
D	momentum flux, drag
H	heat flux
E	evaporation
i	generic substance or property

Physical properties

g	graviational constant
γ	surface tension
ν	molecular viscosity of air
ν_w	" " of water
K_i	Diffusivity of substance i in air
K_{iw}	" " " i in water

ρ Density of air

ρ_w " " water

$Pr = \nu/\kappa_i$ Prandtl or Schmidt number of substance i in air

$Pr_w = (\nu/\kappa_i)_w$ " " " " " " i in water

c_1 Specific heat at constant pressure

f_c Coriolis parameter

Flow variables

(Upper case denotes mean and lower case, turbulence)

U, u downwind velocity

V_h horizontal velocity

V_g geostrophic wind velocity

w vertical velocity

w_f particle settling velocity

T, t temperature

Θ, θ potential temperature

γ_1 adiabatic lapse rate

Θ_v, θ_v potential virtual temperature

Q, q specific humidity

M, m gas concentration

P, p pressure

τ_h horizontal stress

T_s net long wave and turbulent heat flux from surface

$-Q_I$ incoming short-wave radiation

R radius of droplets

R_m mode radius of droplets

$F(R)$ droplet distribution function

$\hat{k}$ vertical unit vector

Turbulence properties

B Bowen ratio = sensible heat flux/latent heat flux

C bulk transfer coefficient

E evaporation rate

F gas transfer rate

H sensible heat flux

$\mathcal{T}$ integral time scale

K turbulent viscosity

K_i turbulent diffusivity

$Pr_t = K/K_i$ turbulent Prandtl number

R_b bulk Richardson number

$Re_* = u_* z_0/\nu$ roughness Reynolds number

L Monin-Obukhov length

z_0 roughness length for momentum

z_i roughness length for passive scalers

Ω averaging time

ℓ mixing length

$m_* = -F/\rho u_*$ a concentration scale

$\theta_* = -H/\rho c_p u_*$ a temperature scale

$q_* = -E/\rho u_*$ a specific humidity scale

$u_* = (\tau/\rho)^{1/2}$ a velocity scale - friction velocity

r non-dimensional resistance

r' resistance of diffusive sub-layer

r''	resistance of dynamic sub-layer
δ_ν	height of the viscous sub-layer
δ_D	height of dynamic sub-layer
ϵ	sampling error
$\zeta = z/L$	Monin-Obukhov stability index
κ	von Karmans constant
Φ	non-dimensional gradient
Ψ	integrated profile parameter
τ	momentum flux or stress

Surface and wave characteristics

A	surface area per molecule of contaminant
a	proportional to wave amplitude
c	phase speed of waves
h_o	height of surface roughness
k	radian wave number
m_o	Charnock's parameter
$S(k)$	wavenumber spectrum
$S_{\eta\eta}(f)$, $\phi(\omega)$	frequency spectrum
α	Kitaigorodskii's parameter
ϵ_γ	surface dilational modulus
$\partial\eta/\partial x$	surface slope
λ	wave length
σ	standard deviation of surface elevation
ω	radian wave frequency

KEY WORDS

Air-sea interface
Air-water interface

Boundary layers
Bubbles
Bulk Richardson number

Capillary waves
Constant stress layer
Cool skin

Dalton number
Diabatic profiles
Diffusive sub-layer
Drag coefficient
Droplet distribution
Dynamic sub-layer

Eddy correlation

Fetch-limited waves
Friction velocity

Gravity waves

Heat flux
Heat transfer coefficient

Integral time scale
Logarithmic profile

Mass transfer
Mass transfer coefficient
Measurement errors
Mixing length
Momentum transfer
Monin-Obukhov stability index

Non-dimensional gradients

Prandtl number
Profiles

Rough flow
Roughness height
Roughness length
Roughness Reynolds number

Sampling variability
Spectra
Sub-layer resistance
Spectral peak enhancement
Spray
Stanton number
Surface tension
Smooth flow

Turbulence
Turbulent fluxes
Turbulent Prandtl number

Viscosity

Wall coordinates
Waves
Wave age
Wave breaking

Figure Captions

1. Schematic diagram showing the balance of viscous and turbulent stresses and the velocity profile in the surface boundary layer. The height scale on the right is for u_* =30 cm/s or U_{10} of about 7.5 m/s, which is the global marine average wind speed.

2. Frequency spectra times ω^4 normalized by the rear face $[\omega^4 \phi(\omega)]_{r.f.}$ which is the average of $\omega^4 \phi(\omega)$ in the region $1.5\omega_p < \omega < 3\omega_p$. The lines corresponding to ω^{-5} and ω^{-3} are 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/c_p. U_c is the component of the 10m wind in the direction of the waves at the spectral peak. (from Donelan et al. 1985).

3. A laboratory wave spectrum $S_{\eta\eta}(f)$ (broken line) and associated pressure-elevation quadrature spectrum (solid line). The quadrature spectrum has been adjusted by the exponential decay to yield the momentum transfer to the waves. Further normalization by ρU^2 yields the spectrum of contributions to the drag coefficient.

4. The drag coefficient versus wind speed. Smooth flow (- - - -), Charnock's formula with m = 0.014 (——), Large and Pond (—— — ——). Symbols correspond to data gathered at various stages of wave development. Very young laboratory waves ($6.5 \leqslant U_{10}/C_p \leqslant 15.4$ — open circles; young field waves ($3.5 \leqslant U_{10}/C_p \leqslant 4.6$ — open triangles; mature field waves ($0.8 \leqslant U_{10}/C_p \leqslant 2.0$ — closed triangles). The vertical bars are a measure of the sampling error, being $\pm$ 1 standard deviation about the mean for wind speeds of 10.6 and 6.5 m/s (field and laboratory respectively). The sampling error is inversely proportional to the square root of the wind speed.

5. Horizontal downwind spectra over water from an anemometer at 11.5 metres, showing the separation of meso-and micro scales. (Redrawn from Pierson, 1983).

Top: Spectrum of the wind for an average wind of 14 m/s for a 17-hour, 4-min sample of 1-min averages. Composited with an 18,000-point 5-Hz sample.

Middle: Spectrum of the wind for an average wind of 11.5 m/s for a 1-day, 1-hour, 36-min sample of 1-minute averages combined with an estimate of the high-frequency part based on measured values of u at 5 Hz.

Bottom: Spectrum of the wind for an average wind of 6.6 m/s for a 5-day, 16-hour, 32-min sample of 1-min averages combined with an estimate of the high-frequency part based on measured values of u at 5 Hz.

6. Spectra of down-stream (u) and vertical (w) velocity components on logarithmic axes. The spectra are normalized by the variance of the vertical velocity component. The measured frequency and mean wind are f and U. Eight runs are superimposed to illustrate the similarity of the spectra in these "natural" coordinates. Agreement in slope to the -2/3 line shown suggests the existence of an inertial sub-range. (Redrawn from Miyake et al., 1970).

7. Cospectra of momentum transfer i.e. between u and w normalized by $-\overline{uw} = u_*^2$. (Redrawn from Miyake et al., 1970).

8. Cospectra of sensible heat flux i.e. between temperature fluctuations t and w normalized by $\overline{wt}$ (Redrawn from Miyake et al, 1970).

9. Seven hours of very stationary mean conditions used to estimate the sampling variability of $\overline{uw}$ over water. The measurements were made from a research tower in Lake Ontario (see Donelan et al., 1985). σ and T_p are the standard deviation of surface elevation and the peak period of the waves respectively. Each point represents a 20-minute average centered on the point. The points are connected by straight lines for reasons of clarity.

10. Photograph taken looking west from the seventh floor of the Canda Centre for Inland Waters during a sudden cold spell. The steam fog on the water reveals the pattern of longitudinal organization.

11. Roughness Reynolds number versus normalized root-mean-square surface elevation from various laboratory experiments (open squares — Kunishi [1963]; solid squares — Hidy and Plate [1966]; other symbols as in Figure 4). The broken line corresponds to the wind speed dependent drag coefficient from Large and Pond [1981] in which σ is computed from the Bretschneider [1973] full development relation at each wind speed.

12. The ratio of measured roughness length, z_o to root-mean-square wave height σ versus inverse wave age U_{10}/c_p. The straight lines are regression lines to the laboratory and field data separately (o and □ Bulk Richardson number R_b less than 0.002 in magnitude; ▲ $0.002 < R_b < 0.011$; Δ $-0.011 < R_b < -0.002$). Error bars are two standard deviations. The solid bars are the estimated sampling errors. The broken bars are the deviation of z_o/σ about the regression line. The striped bar on the ordinate represents the Charnock relation (with m=0.014) for the wind speed range of 7.5 to 20 m/s.

13. As in Figure 12 except that the wind speed at 10m has been replaced by the calculated wind speed at one half wavelength of the peak waves. Here the Charnock relation is the heavy dashed line in the neighbourhood of $U_{\lambda/2}/c_p$ of unity. The roughness of sand grains is shown also.

14. As for Figure 12 except that U_{10} has been replaced by the friction velocity u_*. The Charnock relation (m=0.014) is the heavy dashed line at the lower left. The error bars are now tilted because the sampling variability of u_* affects both abscissa and ordinate. The curved line (- - - -) is Kitaigorodskii's (1970) simplified formula [13]; the straight double dotted (—— - - ——) line is Hsu's formula [22].

15. Time series of surface elevation normalized by the theoretical wave number of the peak $k_p = \omega_p^2/g$. The abscissa is in radians of the peak wave frequency. The lower three traces are from field data, the upper from laboratory data.

16. Wind and temperature profiles measured over Lake Ontario showing formation and decay of a "wave-driven" wind. The profiles are running averages over 30-min plotted at 10-min intervals. The humidity difference is expressed in buoyancy equivalent degrees Celsius. The time series of wind and air-water differences are obtained from measurements at the top level and the surface water temperature. (From Holland, 1981). Note that in the profiles the dots representing the measurements at the centre level (5.3m) are equally spaced and correspond in time to the time series of wind and temperature shown above.

17. Dalton number versus wind speed. The various lines are derived from the semi-empirical theories of Brutsaert (—— - - ——); Owen and Thompson (- - - -); Kitaigorodskii and Donelan (— — —); Kitaigorodskii and Donelan modified (___ ___ ___) using the drag coefficient curve shown. The modified version of Kitaigorodskii and Donelan incorporates an effect of roughness Reynolds number on the surface mixing length - in equation [35] β = 1.24, m = -0.5.
The shaded area encompasses most of the open ocean field observations reported by Smith (1987).

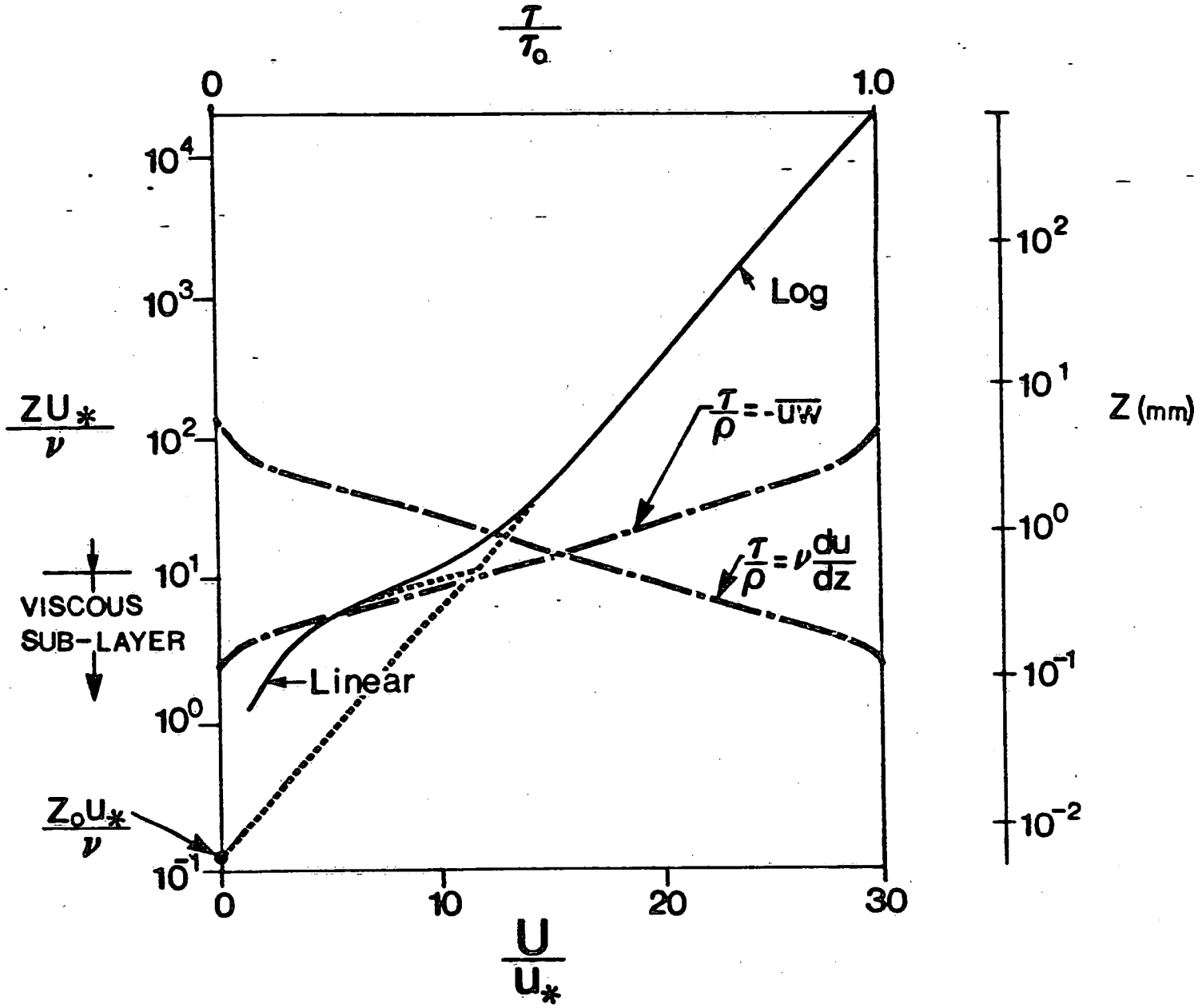

$\frac{\tau}{\tau_0}$
0
1.0
10^4
10^3
10^2
10^1
10^0
10^{-1}
$\frac{ZU_*}{\nu}$
VISCOUS
SUB-LAYER
$\frac{Z_0U_*}{\nu}$
Log
$\frac{\tau}{\rho} = -\overline{uw}$
$\frac{\tau}{\rho} = \nu\frac{du}{dz}$
Linear
0
10
20
30
$\frac{U}{u_*}$
10^2
10^1
10^0
10^{-1}
10^{-2}
Z (mm)

FIG. 1.

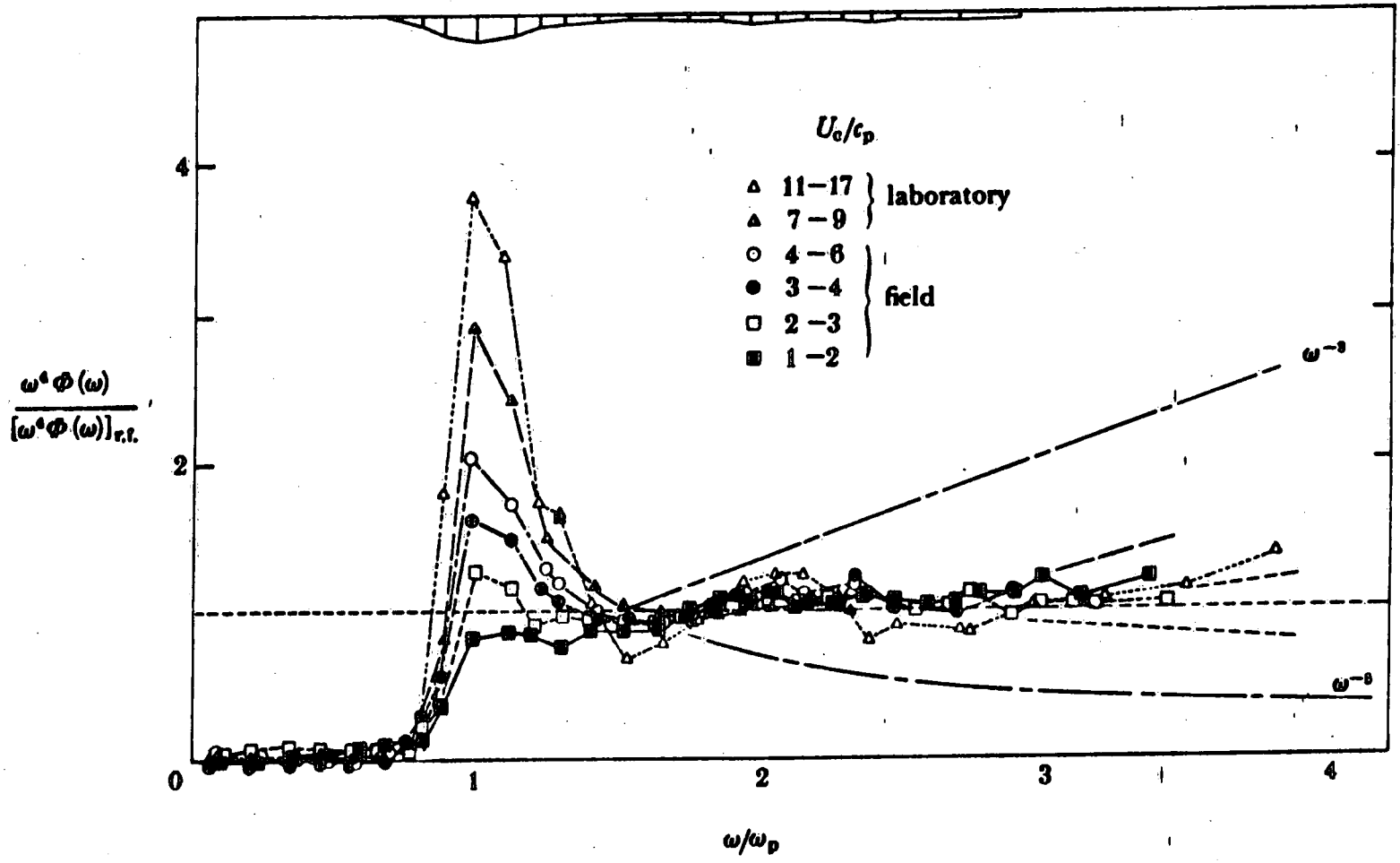

U_c/c_p
11−17
7−9
laboratory
4−6
3−4
2−3
1−2
field
ω^{-3}
ω^{-3}
$\frac{\omega^4 \Phi(\omega)}{[\omega^4 \Phi(\omega)]_{r.f.}}$
ω/ω_p
0
1
2
3
4
2
4

FIG. 2

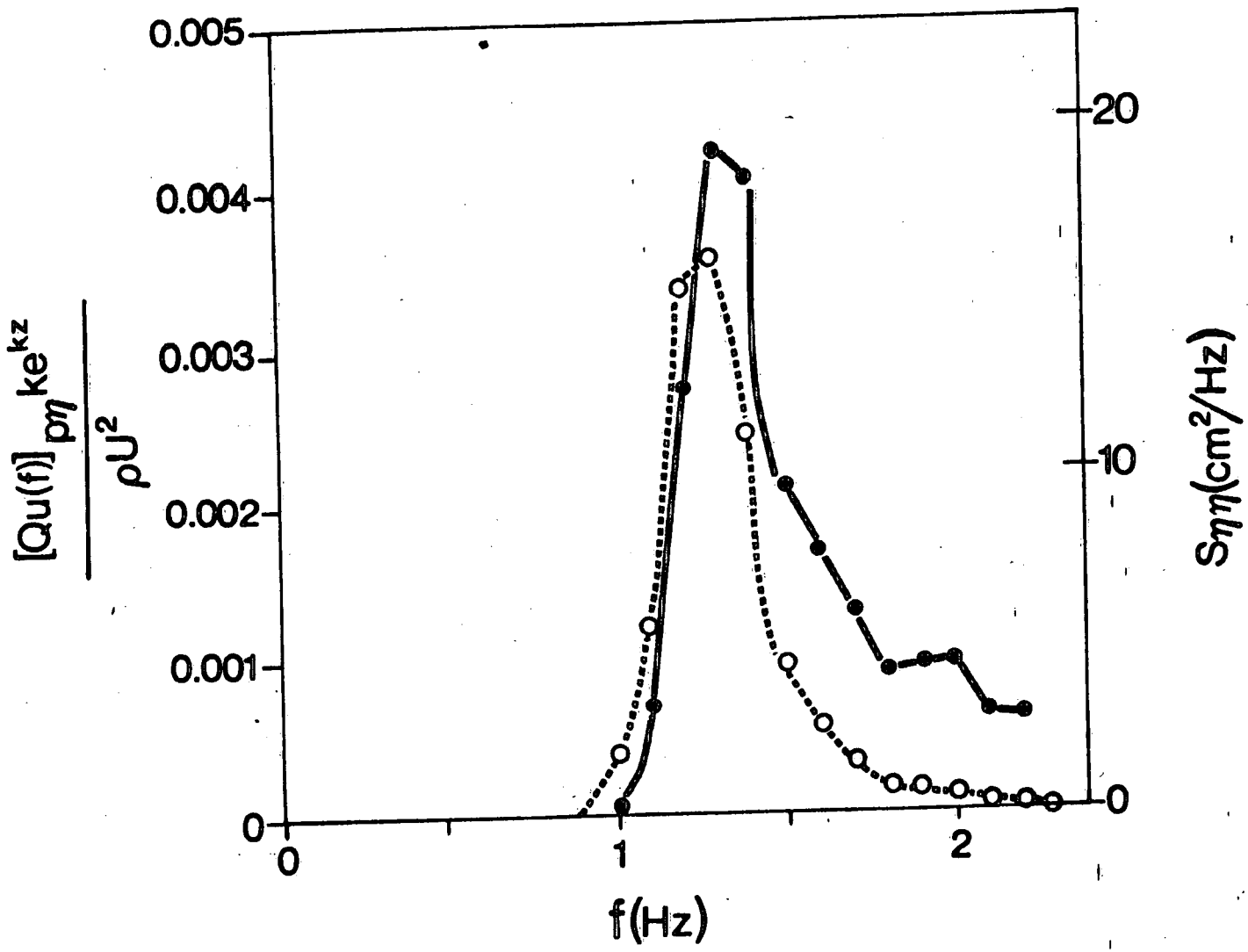

0.005
0.004
0.003
0.002
0.001
0
$\frac{[Qu(f)]_{p\eta} ke^{kz}}{\rho U^2}$
20
10
0
$S_{\eta\eta}(cm^2/Hz)$
0
1
2
f(Hz)

FIG. 3

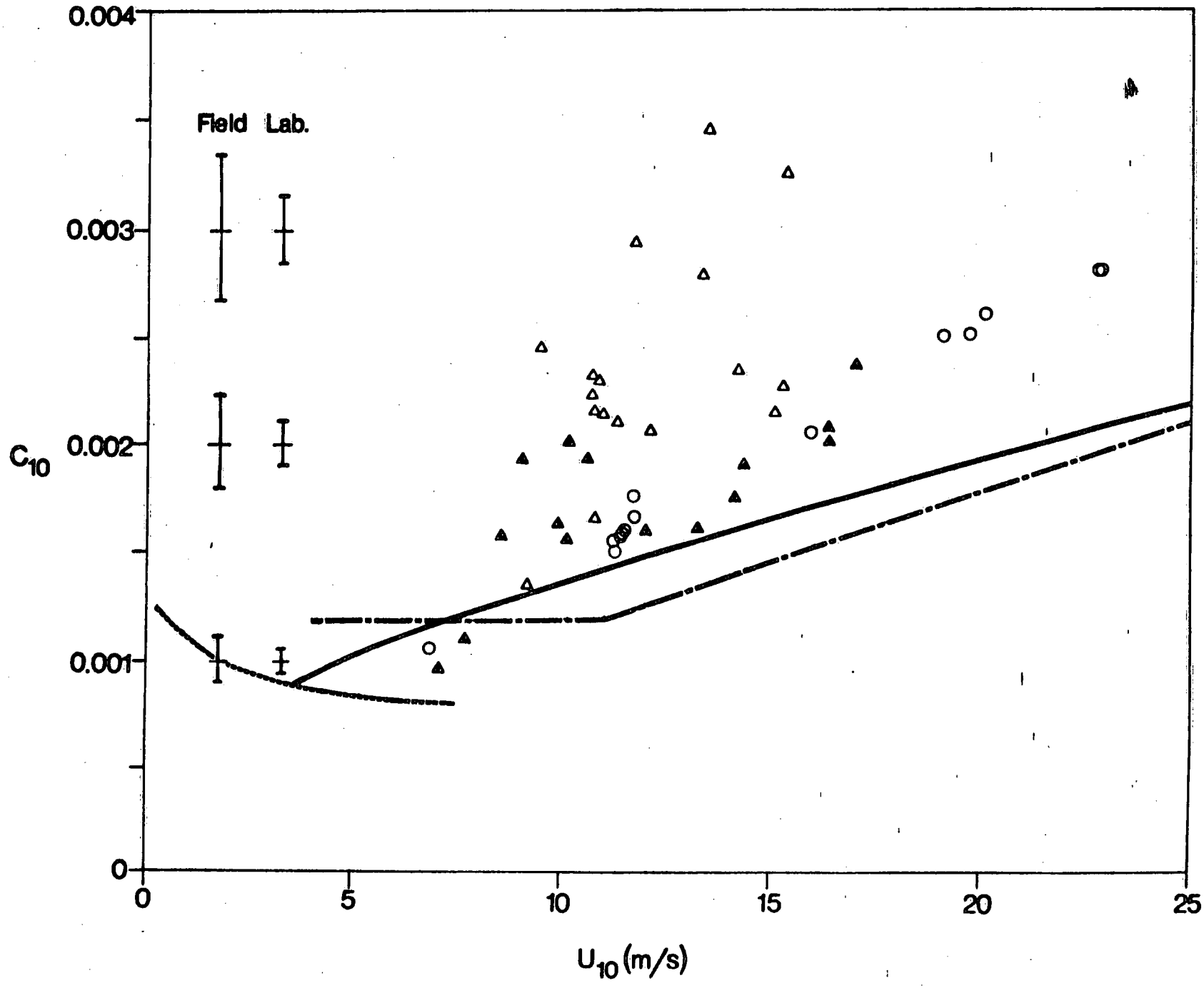


FIG. 4.

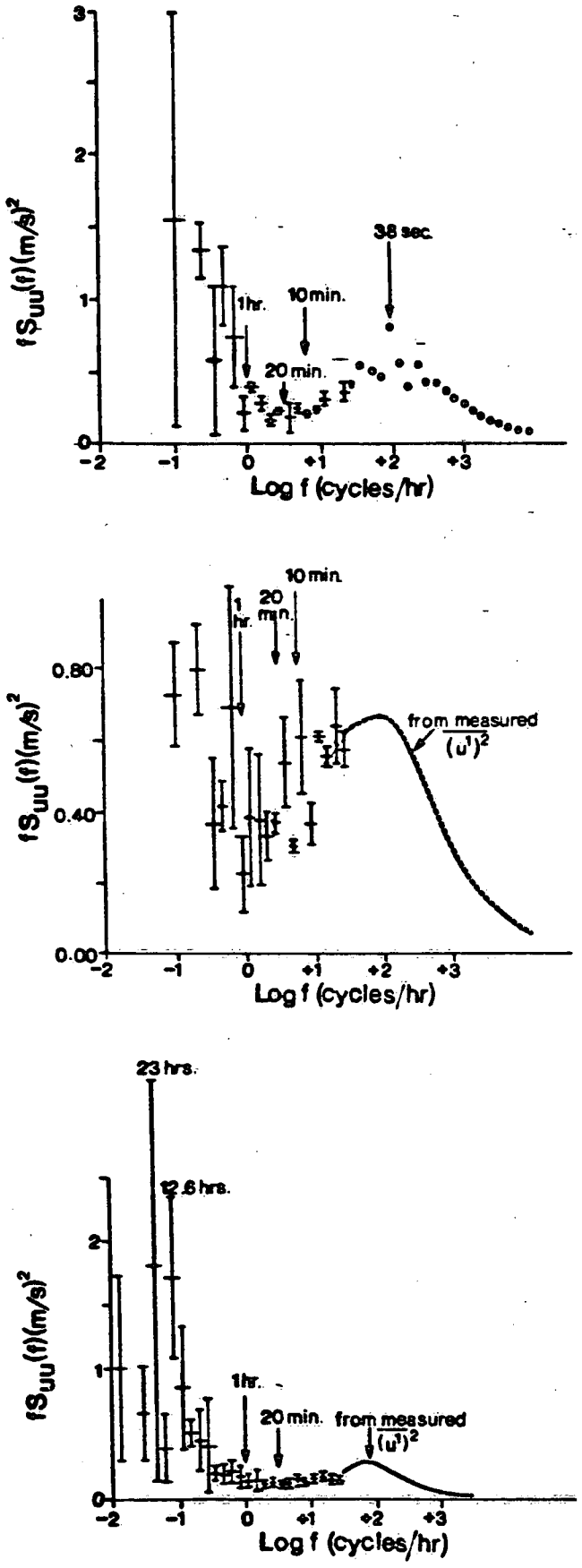

$fS_{UU}(f)(m/s)^2$
38 sec.
10 min.
1 hr.
20 min.
Log f (cycles/hr)
10 min.
1 hr.
20 min.
from measured $\overline{(u')^2}$
23 hrs.
12.6 hrs.
1 hr.
20 min.
from measured $\overline{(u')^2}$

FIG. 5

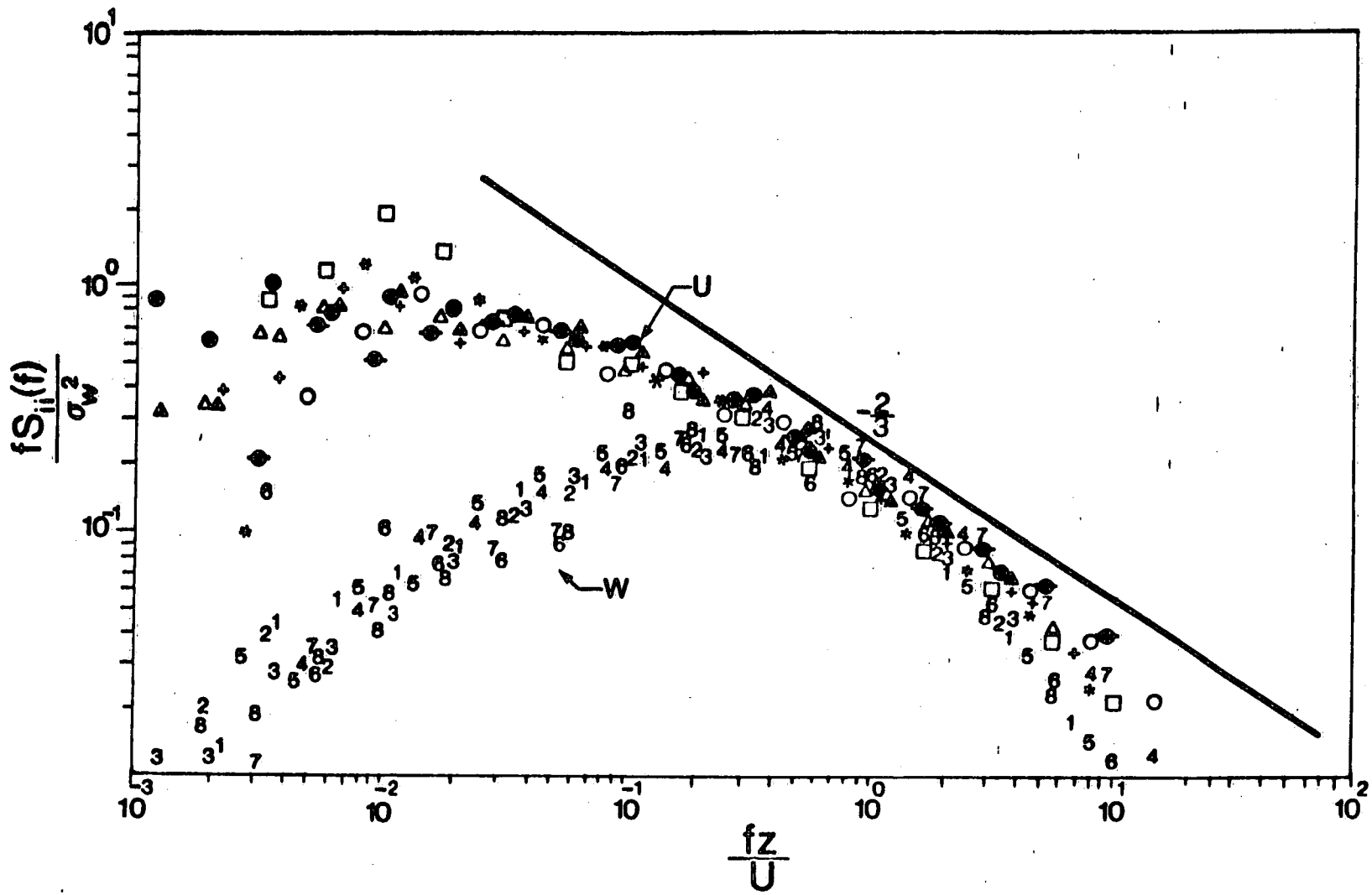

$\frac{fS_{ii}(f)}{\sigma_w^2}$
$\frac{fz}{U}$
U
W
$-\frac{2}{3}$
10^{1}
10^{0}
10^{-1}
10^{-3}
10^{-2}
10^{-1}
10^{0}
10^{1}
10^{2}

Fig. 6

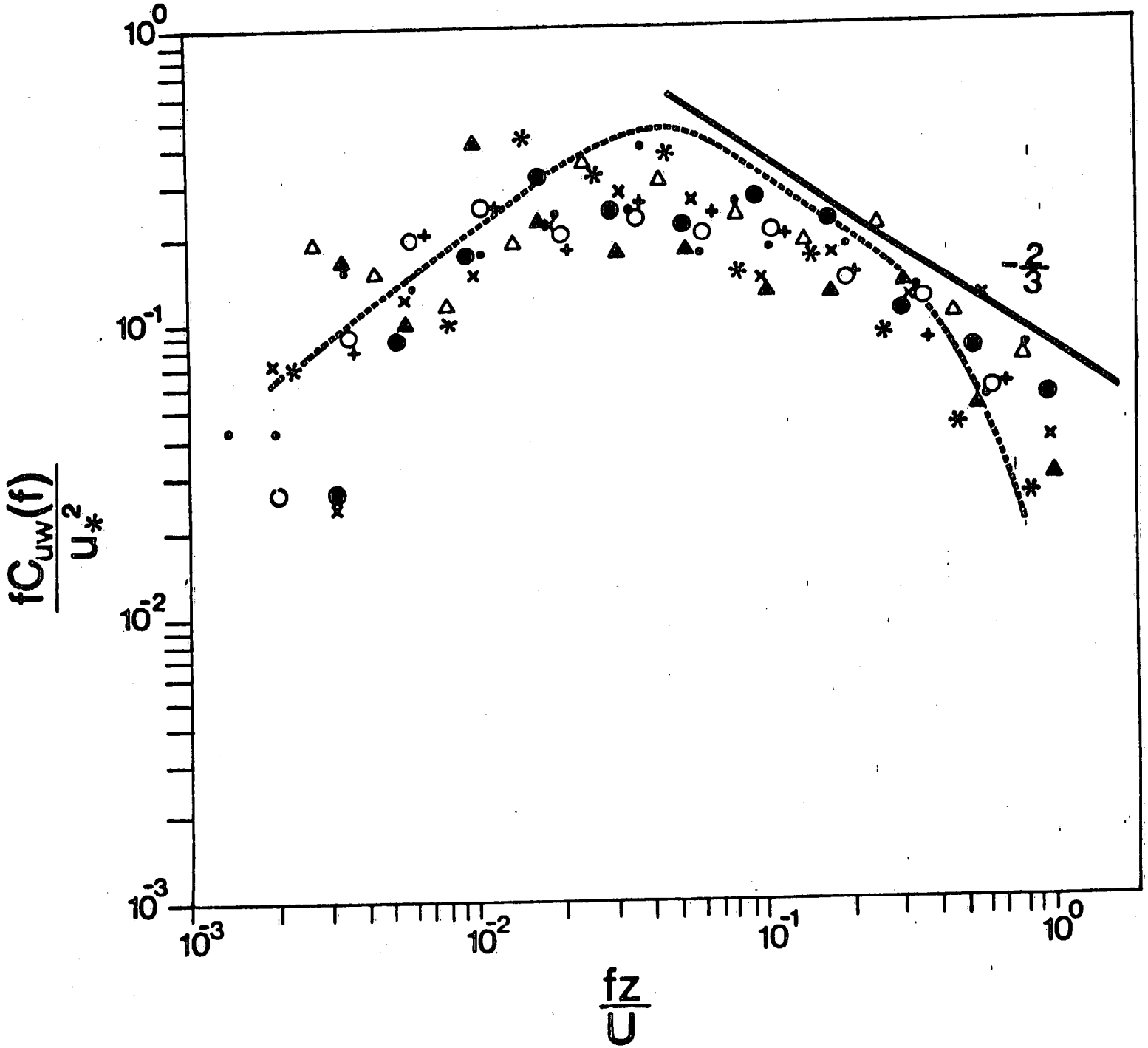


FIG. 7.

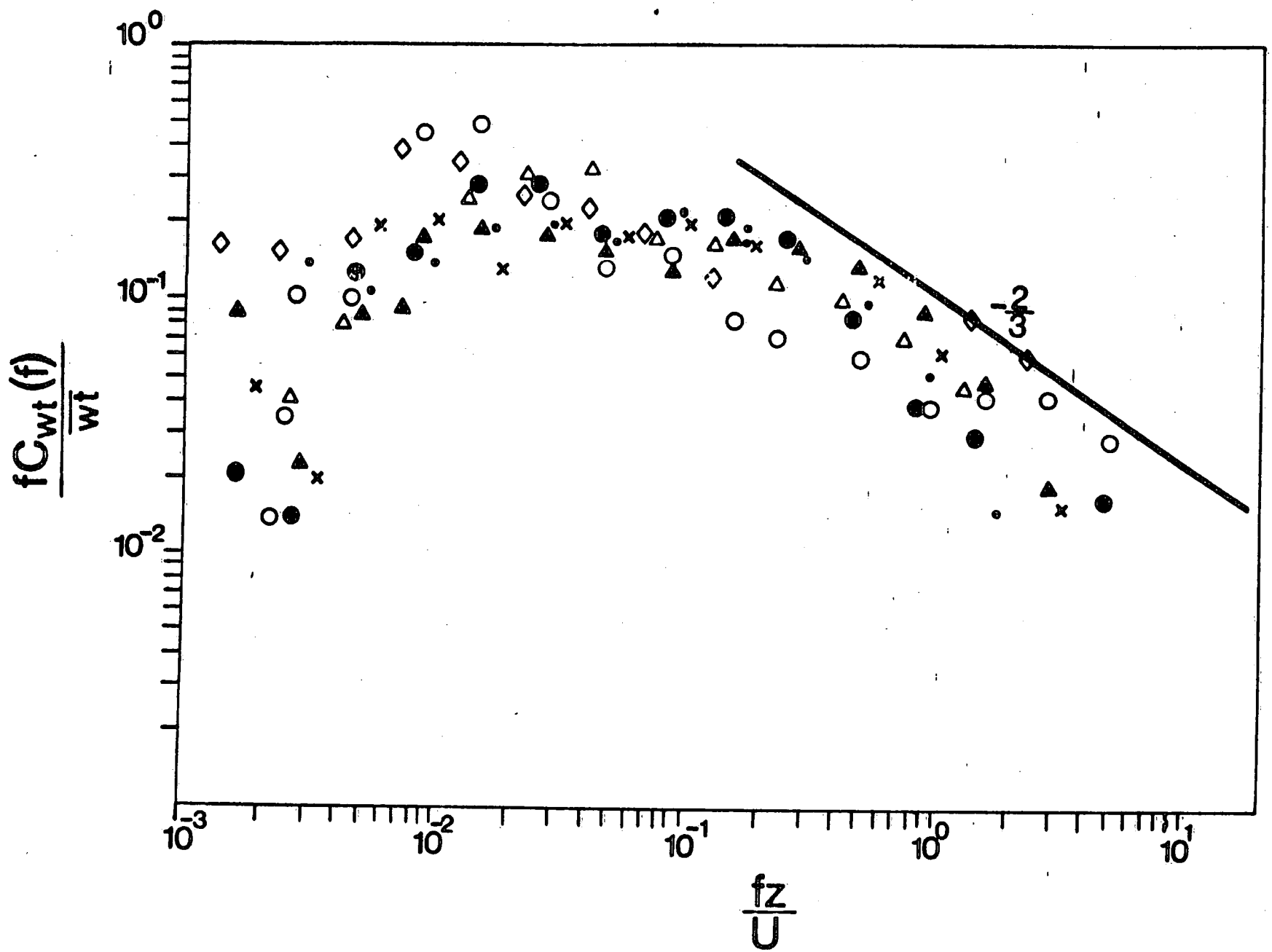

10^0
10^{-1}
10^{-2}
$\frac{fC_{wt}(f)}{\overline{wt}}$
$-\frac{2}{3}$
10^{-3}
10^{-2}
10^{-1}
10^0
10^1
$\frac{fz}{U}$

FIG. 8.

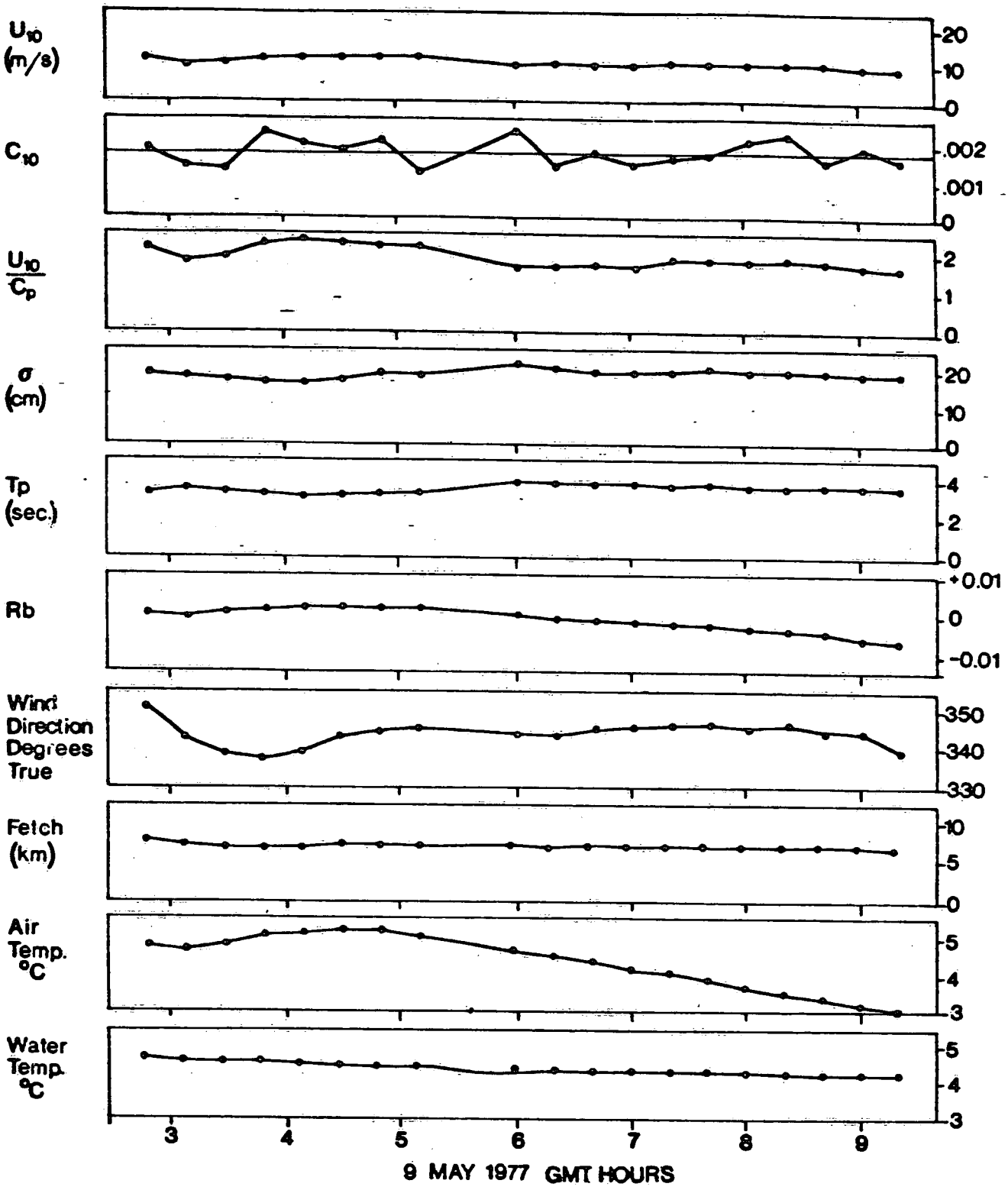
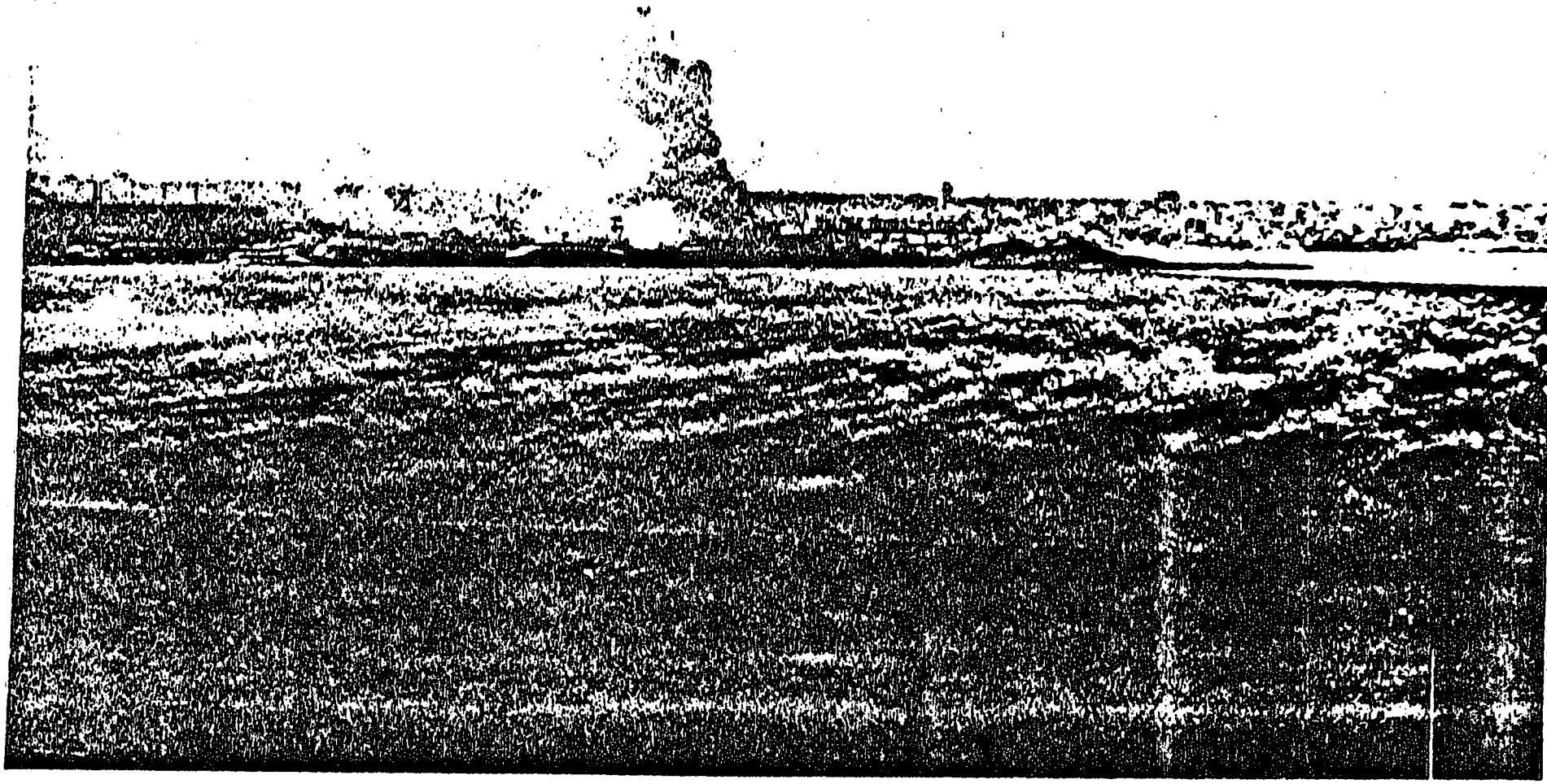

U_{10} (m/s)
20
10
0
C_{10}
.002
.001
0
$\frac{U_{10}}{C_p}$
2
1
0
σ (cm)
20
10
0
Tp (sec.)
4
2
0
Rb
+0.01
0
-0.01
Wind Direction Degrees True
350
340
330
Fetch (km)
10
5
0
Air Temp. °C
5
4
3
Water Temp. °C
5
4
3
3
4
5
6
7
8
9
9 MAY 1977 GMT HOURS

FIG. 9



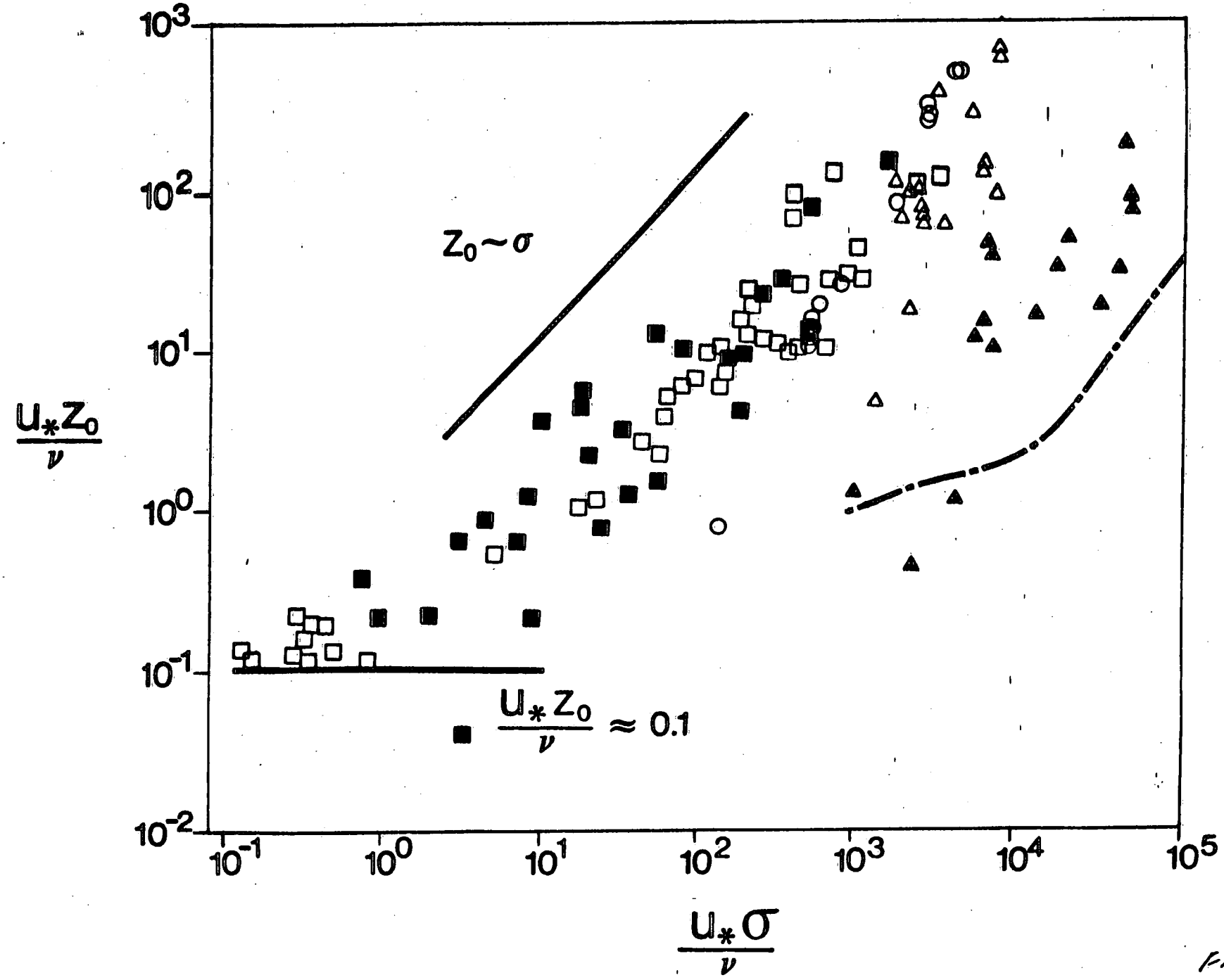

10^3
10^2
10^1
10^0
10^{-1}
10^{-2}
$\frac{u_* z_0}{\nu}$
$z_0 \sim \sigma$
$\frac{u_* z_0}{\nu} \approx 0.1$
10^{-1}
10^0
10^1
10^2
10^3
10^4
10^5
$\frac{u_* \sigma}{\nu}$

FIG. 11

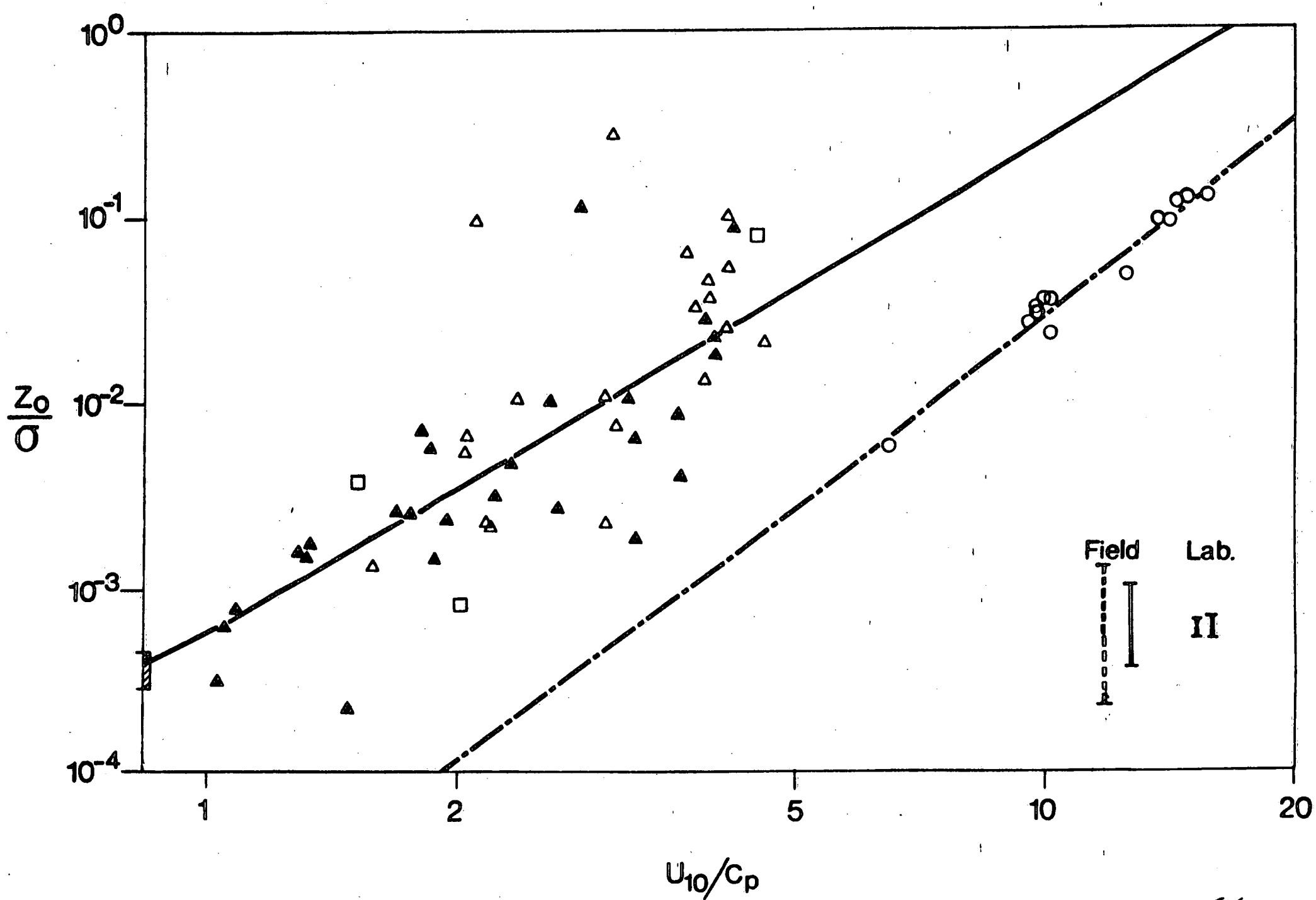


FIG. 12.

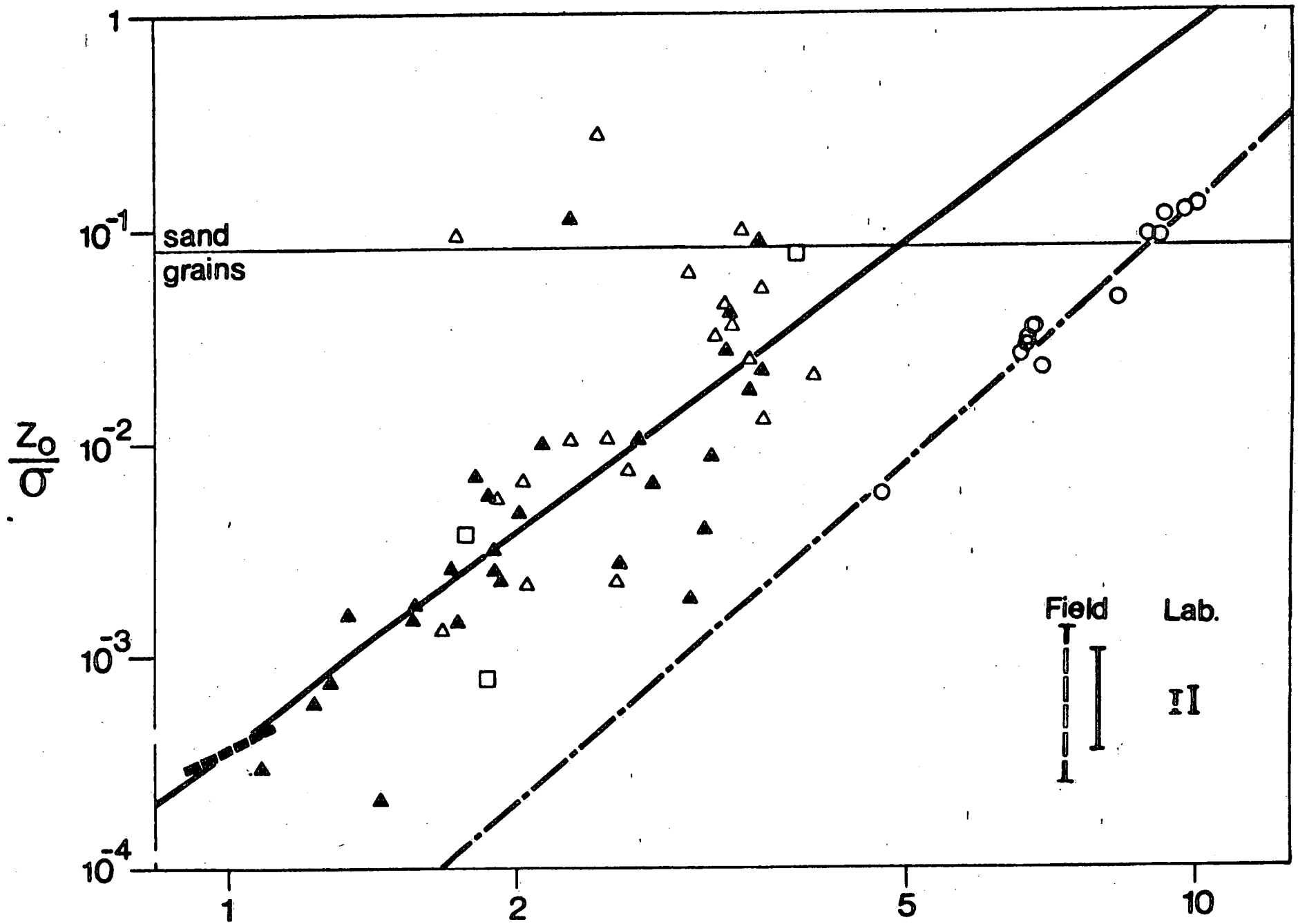


FIG. 13

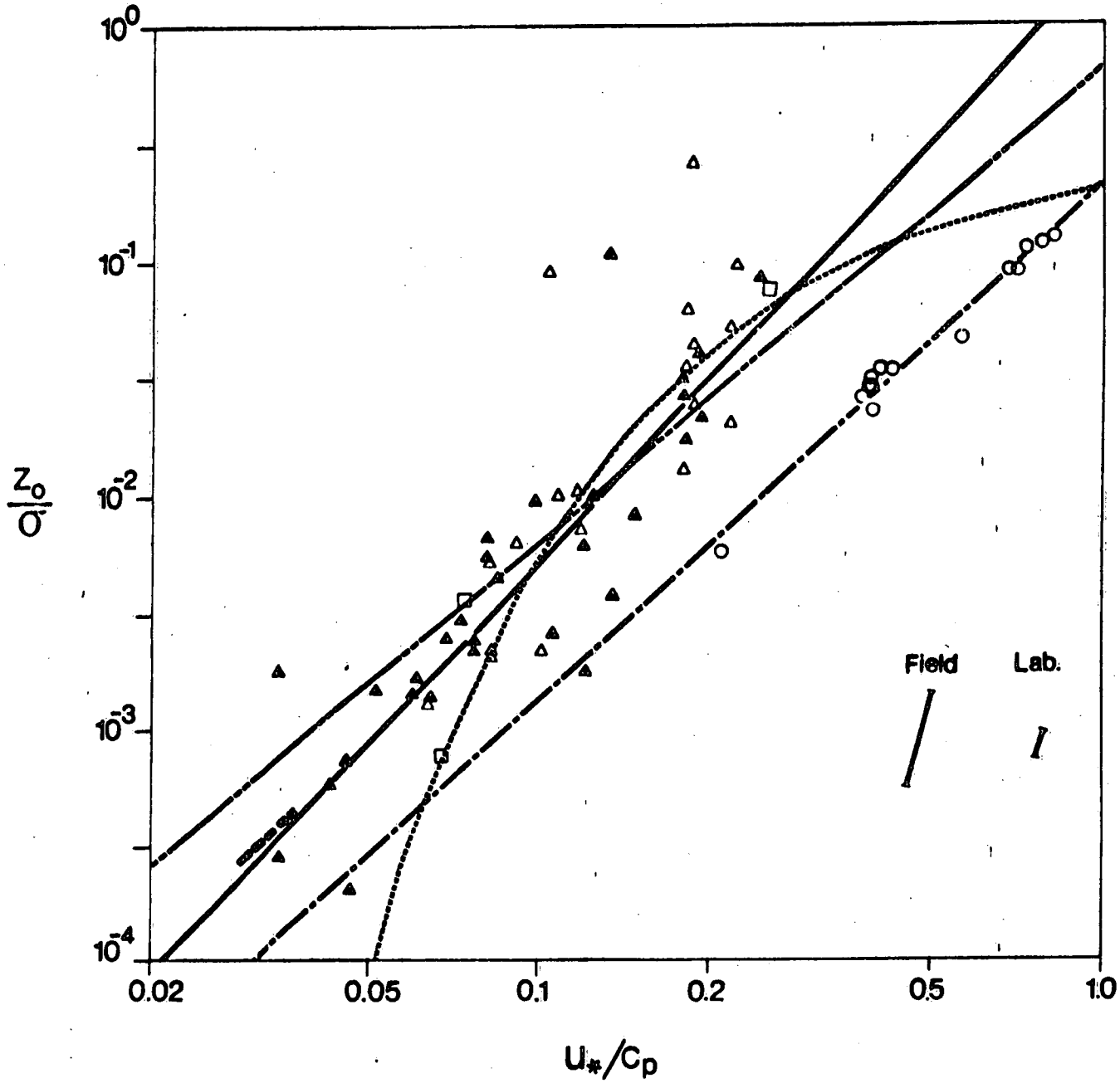


FIG. 14

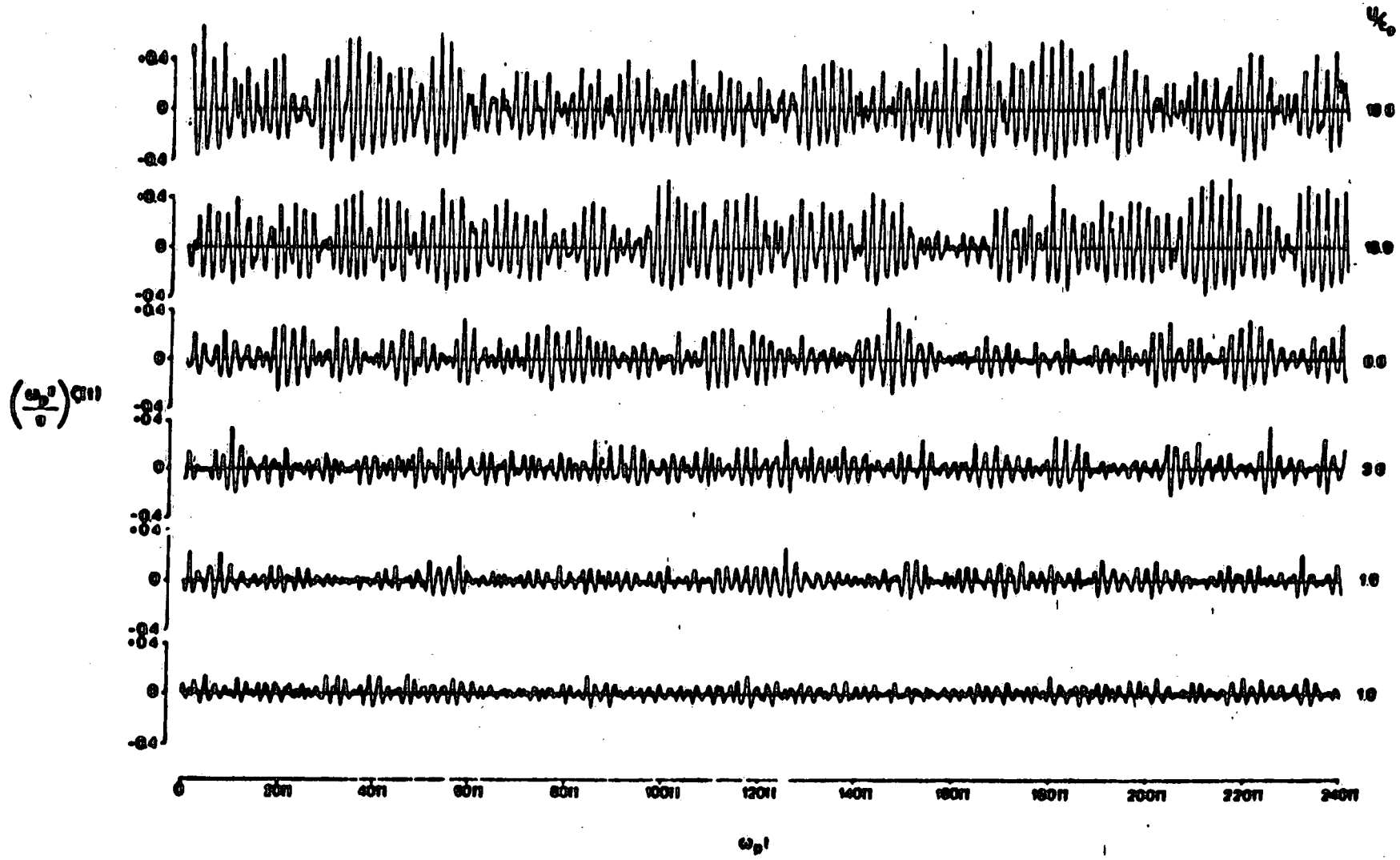

$\left(\frac{\omega_p^2}{\omega}\right)^{C(t)}$
$\omega_p t$
0
20π
40π
60π
80π
100π
120π
140π
160π
180π
200π
220π
240π
0.4
0
-0.4
10.0
8.0
6.0
3.0
1.0
1.0

FIG. 15.

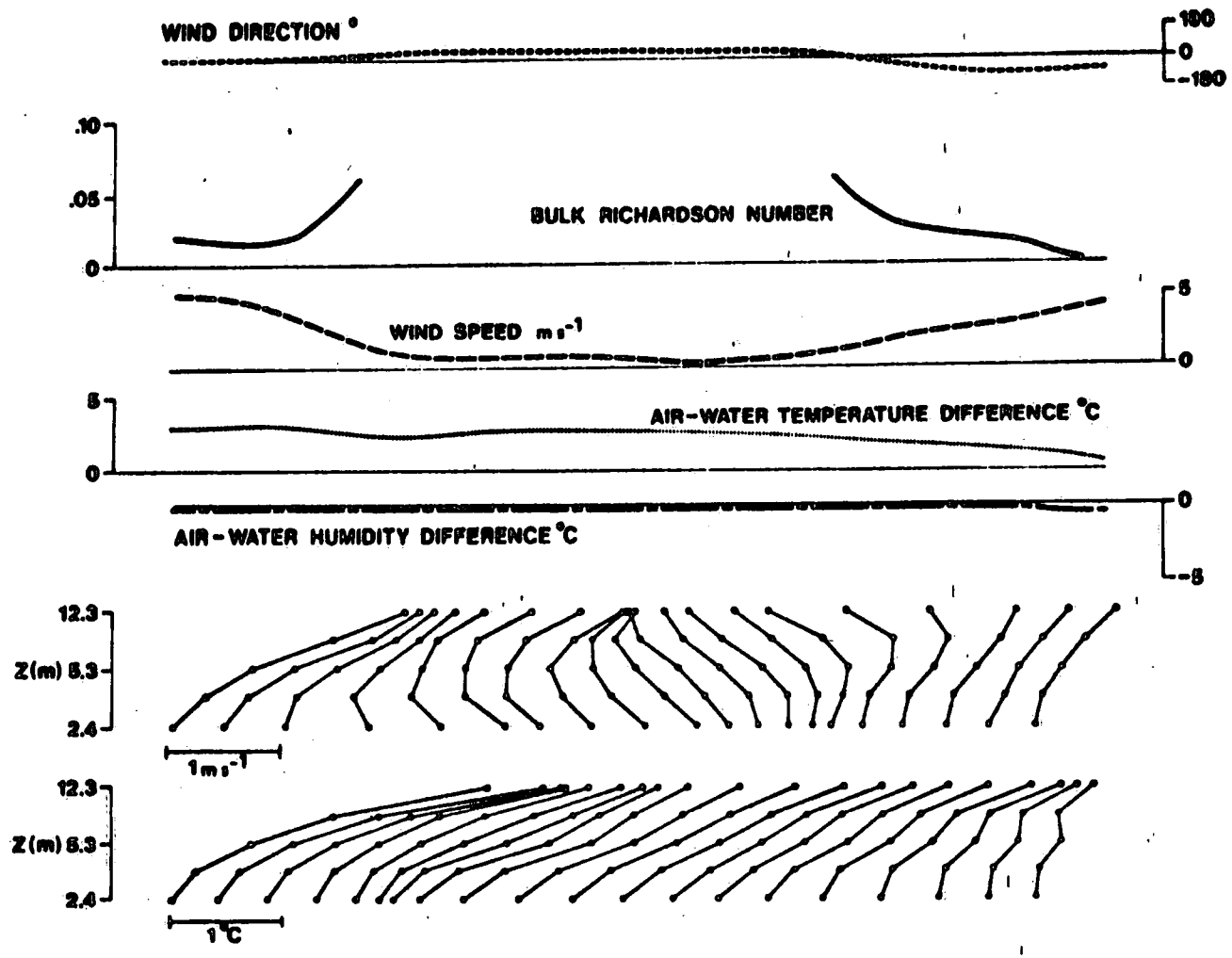

WIND DIRECTION °
180
0
-180
.10
.05
0
BULK RICHARDSON NUMBER
WIND SPEED m s-1
5
0
5
0
AIR-WATER TEMPERATURE DIFFERENCE °C
AIR-WATER HUMIDITY DIFFERENCE °C
0
-5
12.3
Z(m) 5.3
2.4
1 m s-1
12.3
Z(m) 5.3
2.4
1 °C

FIG. 16.

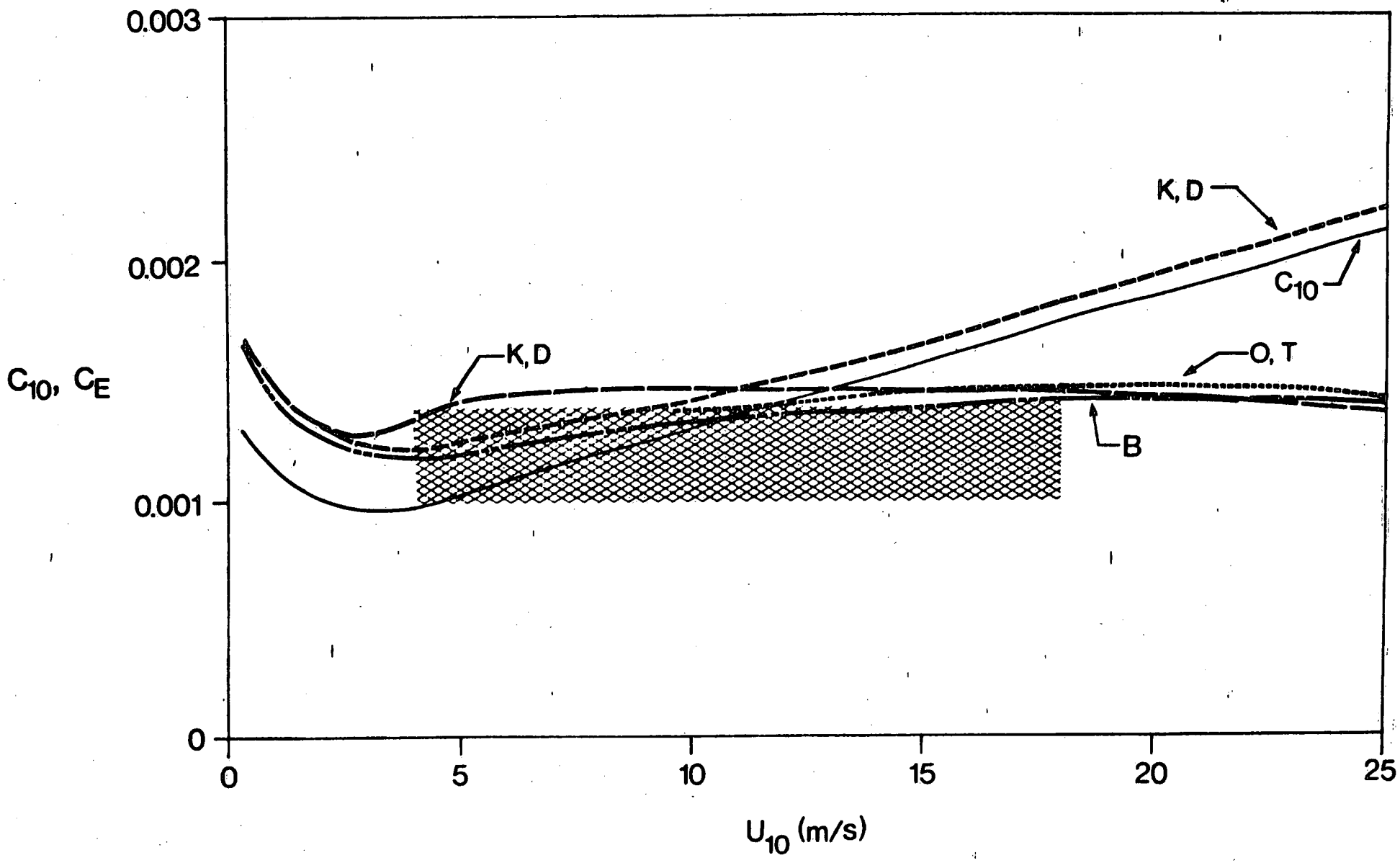


FIG. 17