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COMPARISON OF BULK AND DIRECT METHODS FOR ESTIMATING FLUXES AND STRUCTURE FUNCTION CONSTANTS OVER WATER

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COMPARISON OF BULK AND DIRECT METHODS FOR ESTIMATING FLUXES AND STRUCTURE FUNCTION CONSTANTS OVER WATER

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MANAGEMENT PERSPECTIVE

Lake studies often require estimates of wind stress, heat flux and evaporation which for general work by operational staff must be obtained from meteorological records. This paper compares estimates using available data with other direct measurements. The bulk methods presently employed and using available data, are reasonably reliable for surface stress, but greatly underestimate the heat flux. New relationships to improve the estimates of heat flux are proposed.

In other investigations by modellers and by engineers the details of the actual processes described in this paper are needed for:

atmospheric models for weather forecasts,
oceanic circulation and seasonal changes,
lake circulation and seasonal changes,
models of climatic variation and change,
computations of evaporation and heat loss in industrial cooling ponds and reservoirs.

This paper provides new information to all these subject areas.

T. Milne Dick
Chief
Hydraulics Division

PERSPECTIVE-GESTION

Lorsqu'on effectue des études sur les lacs, il faut souvent obtenir, à partir de relevés météorologiques, des estimations de la tension du vent, du flux thermique et de l'évaporation, aux fins du travail général du personnel des opérations. Dans le présent document, on compare ces estimations à l'aide des données offertes avec d'autres mesures directes. Les méthodes massales utilisées présentement et qui se servent des données disponibles sont assez sûres lorsqu'il s'agit de mesurer la tension superficielle, mais elles sous-évaluent grandement le flux thermique. On propose donc de nouveaux rapports pour améliorer les estimations du flux thermique.

Dans d'autres recherches effectuées par des modélisateurs et des ingénieurs, il est nécessaire d'obtenir les détails des procédés dont il est question dans ce document aux fins suivantes:

modèles atmosphériques aux fins des prévisions météorologiques,
circulation océanique et variations saisonnières,
circulation sur les lacs et variations saisonnières,
modèles de variation et de modification du climat,
calculs de l'évaporation et de la perte de chaleur dans les piscines et les réservoirs de stockage industriels.

Le présent document offre de nouveaux renseignements relatifs à tous les domaines susmentionnés.

Le Chef
T. Milne Dick
Division de l'hydraulique

RÉSUMÉ

Un ensemble d'observations du vent, de la température et de l'humitidé ont été effectuées au-dessus du lac Ontario, dans des conditions plutôt stables, pour établir des rapports flux-gradient et pour vérifier les méthodes massales employées pour calculer des constantes de tension, de flux thermique, d'évaporation et de fonction de structure. Bien que le rapport tension-gradient de vitesse ait correspondu à celui observé au-dessus du sol, le rapport flux thermique-gradient de température était très différent. La différence est attribuable à une transmission plus efficace de la chaleur au-dessus de l'eau et on postule que l'évaporation de goutelettes est le facteur qui contribue à cet accroissement. Lorsqu'on utilise les rapports flux-gradient nouvellement découverts dans les méthodes massales, la concordance des estimations massales et directes des flux et des constantes de fonction de structure est vraisemblable.

Comparison of Bulk and Direct Methods for Estimating Fluxes and Structure Function Constants over Water.

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Abstract

A set of observations of wind, temperature and moisture over Lake Ontario, under largely stable conditions, has been used to obtain flux-gradient relationships and to test bulk methods for computing stress, heat flux, evaporation and structure function constants. Although the stress-velocity gradient relationship agreed with that observed over land, the heat flux-temperature gradient was quite different. The difference implies an increase in heat transfer efficiency over water, and it is postulated that the agency for the increase in heat transfer efficiency is the evaporation of droplets. When these new found flux-gradient relationships are employed in the bulk methods the agreement between bulk and direct estimates of fluxes and structure function constants is reasonable.

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1. Introduction

During the International Field Year on the Great Lakes in 1972 measurements of turbulent fluxes and profiles were made from masts erected in Lake Ontario in water of depth 12 m. The masts were located 2.6 km north of the south shore of the lake and about 4 km west of the mouth of the Niagara River. Thus, the fetches varied from about 3 km for south winds to 240 km for NE winds. Profiles of wind and temperature were obtained automatically every ten minutes during four fortnightly periods (Donelan et al, 1974) and fluctuations of wind, temperature and humidity were recorded at selected times. The data covered a large range of stabilities from very stable air in May to unstable air in October, however most of the coincident flux and profile data were obtained under stable conditions.

The use of bulk aerodynamic methods for estimating fluxes is based on the Monin and Obukhov (1954) similarity theory, which states that the deviation of non-dimensional gradients, in the constant flux near-surface boundary layer, from their values under neutral conditions, depends only on $\zeta = z/L$; where z is the height above the interface and L is the Monin-Obukhov length. The functional forms of the various non-dimensional gradient parameters [e.g., $\phi_m(\zeta)$, $\phi_h(\zeta)$, $\phi_\varepsilon(\zeta)$] have been established from careful coincident measurements of fluxes and gradients over land. And, as the theory predicts, the measurements in general suggest that these flux-gradient relationships are universal functions of ζ only. Earlier attempts to establish these relationships (e.g., Dyer, 1965, 1967), while supportive of the theory, could not be considered definitive because some of the required fluxes were not measured directly but instead were inferred from the profiles. However, Businger et al. (1971) have made extensive measurements over wide ranges of ζ above flat grassland and their estimates of the flux-gradient relationships have been widely used to compute fluxes from bulk formulae over land and sea.

Where there have been coincident flux and gradient measurements over water the bulk formulae have been tested with somewhat indifferent results. Generally speaking the land-derived values of ϕ_m (momentum) and ϕ_q (moisture) appear to work quite well over water, but estimates of heat flux, based on land-derived ϕ_h values, differ considerably in many cases from the measured values. Table 1 summarizes the comparisons of direct and bulk estimates of boundary layer fluxes which have been made over water.

Of course, the bulk estimates of heat flux depend both on ϕ_h and z_{oT}, the roughness length for temperature. Poor agreement between direct and bulk flux estimates could mean that one or both of ϕ_h and z_{OT} are in error. However, with coincident direct flux and profile measurements ϕ_h may be determined independently of z_{OT}. In only two of the cases listed in Table 1 was this done for any of the non-dimensional gradients, ϕ_m, ϕ_h or ϕ_q.

In one case (Zank, 1981) coincident measurements of profiles and fluxes yielded estimates of ϕ_m in reasonable agreement with Businger et al (1971). In another (Zank, 1983) ϕ_h was tested in unstable conditions and found to be in general agreement with Businger et al (1971). However, the erratic behaviour of the bulk heat flux has been noted in measurements in the tropics with high moisture fluxes (Pond et al, 1971 and Paulson et al, 1972) and in temperate latitudes with relatively low moisture fluxes (Dunckel et al, 1974).

Phelps and Pond (1971) and Donelan and Miyake (1973) have noted large differences in the spectra of temperature and humidity in the marine boundary layer; while McBean (1971) found them very similar in the surface boundary layer over land. However, it has been argued (see, for example, Friehe and Schmitt, 1976) that at least some of the differences in marine boundary layer spectra of temperature and humidity are due to contamination of the temperature sensors by salt.

In our initial approach to the analysis of our lake Ontario data we set out to test the bulk formulae against direct flux observations, but we found that agreement was generally poor with the magnitude of the eddy heat flux consistently and substantially greater than the bulk estimates. Since we obtained concurrent profile and flux data we were able to derive ϕ_m (ζ) and ϕ_h (ζ) over water in much the same way as Businger et al (1971) did over land. The approach we have taken then, is to derive over water relationships for ϕ_m and ϕ_h under stable conditions, and to use them in bulk formulae for comparison with direct estimates of the fluxes of momentum, heat and moisture and the structure function constants of temperature and humidity. The large differences in ϕ_h from the land derived values call for an explanation and we suggest one, but are unable to prove it.

2. Theory

2.1 "Direct" measurements

Fluxes of momentum (stress, τ), heat (H) and moisture (E) could be derived directly from their definitions:

$$\tau = - \rho \overline{u'w'} \tag{1}$$

$$H = C_p \rho \overline{w'T'} \tag{2}$$

$$E = \rho \overline{w'q'} \tag{3}$$

where w is the vertical velocity, u the longitudinal velocity component, T potential temperature, q specific humidity, ρ air density, and C_p specific heat at constant pressure. Overbars denote time averages, and primes deviations from these averages.

In addition, $u_* = \sqrt{\frac{\tau}{\rho}}$ was determined from high-frequency fluctuations of the longitudinal velocity component by the dissipation method.

For high frequencies ($f > V/z$), the spectral density of the u-component $S_u(f)$ is given by:

$$fS_u(f) = 0.15\ \varepsilon^{2/3}\ (f/V)^{-2/3} \tag{4}$$

where V is the mean wind speed, ε is the rate of energy dissipation into heat, z is height, and the frequency f is measured in cycles per unit time.

Equation (4) permits the determination of ε from observations of velocity fluctuations. u_* is related to ε through:

$$u_* = (\varepsilon\ \kappa\ z/\phi_\varepsilon)^{\frac{1}{3}} \tag{5}$$

where κ is the von Kårmån constant, taken here as 0.4; ϕ_ε is a function of $\zeta = z/L$ where L is the Monin-Obukhov length, incorporating the buoyancy fluxes of both heat and moisture. The equations used for ϕ_ε were:

Unstable air: $\phi_\varepsilon = 1 - \zeta$

Stable air: $\phi_\varepsilon = [1 + 2.5\ (\zeta)^{0.6}]^{3/2}$ (6)

The Monin-Obukhov stability index, ζ, incorporates the buoyancy fluxes of both heat and moisture so that:

$$\zeta = \frac{z}{L} = -\frac{\kappa z}{u_*^3}\left(\frac{g}{T}\overline{w'T'} + 0.61\ g\ \overline{w'q'}\right)$$

In practice, ϕ_ε was never very different from unity, so that the resulting u_*'s were not sensitive to the actual estimates of L. Actually, we used the L-values determined by the bulk method (see Section 2.2).

The temperature structure function constant $C_T^{\ 2}$ was estimated from the spectrum of temperature fluctuations at high frequencies:

$$fS_T(f) = \frac{1}{13.6}\ C_T^{\ 2}\ \left(\frac{f}{V}\right)^{-2/3} \tag{7}$$

and the structure function constant $C_q^{\ 2}$ from an analogous expression for the spectral density of specific humidity.

2.2 Bulk Methods

In the surface layer, C_T^2 is assumed to obey Monin-Obukhov theory. Accordingly, it is given by:

$$C_T^2 = T_*^2 z^{-2/3} J(\zeta) \tag{8}$$

There, T_* is the scaling temperature $T_* = -\frac{\overline{w'T'}}{u_*}$ and $J(\zeta)$ is a presumably universal function which has been measured by Wyngaard et al. (1971). Thus, C_T^2 depends on the vertical fluxes of momentum, heat and moisture. We will therefore first discuss bulk methods for finding these fluxes and then derive C_T^2 and C_q^2 from equ. 8.

From the wind profile, the friction velocity is given by:

$$u_* = \frac{\kappa V}{\ln \frac{z}{z_o} - \psi(\zeta)} \tag{9}$$

z_o is the roughness length and ψ is a presumably universal function which is quite well known over land, and has been tabulated. It is negative in stable air and positive in unstable air. We will use our data to determine it for stable air over water.

For rough flow ($z_o u_*/\nu > 2$) the roughness length z_o is, on dimensional grounds, related to both the friction velocity, u_* and the wave age, c_p/V (where c_p is the phase speed of the waves at the peak of the spectrum and ν is the kinematic viscosity of the air) in the following form:

$$z_o = \alpha \frac{u_*^2}{g} \cdot G(c_p/V) \tag{10}$$

Which is a generalization of the Charnock (1955) relation due to Kitaigorodskii and Volkov (1965). In our data c_p/V is near full development and varies little (Table 3). Thus the Charnock relation should be an adequate representation of z_o.

At low wind speeds ($\lesssim$ 5 m/s) it is likely that the smooth flow condition will be more appropriate:

$$z_o = 0.13 \frac{\nu}{u_*} \tag{11}$$

Our data cover an intermediate range of wind speeds ($3 < V < 11$ m/s), where the observed linear wind speed dependence of Smith (1980) of the neutral drag coefficient, C_{DN} adequately bridges the regimes of (10) and (11).

$$C_{DN} \times 10^3 = 0.61 + 0.063\, V_N \tag{12}$$

The roughness length is related to the neutral drag coefficient C_{DN} by:

$$z_o = z \exp\left(- \kappa/\sqrt{C_{DN}}\right) \tag{13}$$

In order to evaluate the stress by equ. (9), we employ a process of successive approximations, which converges rapidly in practice. We first estimate $\overline{u'w'}$ for neutral conditions, and use the corresponding technique to infer $\overline{w'T'}$ (and $\overline{w'q'}$) from air-sea temperature difference and temperature roughness length for $\psi = 0$ (see below). Given $\overline{w'T'}$, $\overline{w'q'}$ and $\overline{u'w'}$, we compute L and hence $\psi(\zeta)$ and recompute $\overline{u'w'}$, $\overline{w'T'}$ and L. This procedure usually has to be repeated only twice to yield stable estimates of $\overline{u'w'}$.

For heat flux, Monin-Obukhov theory suggests:

$$\overline{w'T'} = \frac{\kappa^2 \,(T_s - T_z)V}{\phi_h(o)\,\left[\ln\frac{z}{z_{o_T}} - \psi_h(\zeta)\right]\left[\ln\frac{z}{z_o} - \psi(\zeta)\right]} \tag{14}$$

where T_z is the (potential) temperature at thermometer height and T_s is the surface temperature. Again, the function $\psi_h(\zeta)$ is well known over land. One difficulty with the use of (14) is that T_s is the "skin" temperature, whereas "bucket" temperatures are usually measured. Often, the bucket temperature is significantly warmer than the skin temperature due to evaporation at the surface.

Liu et al. (1979) have constructed a nomogram which permits estimates of differences between skin and bucket temperatures from wind speed and differences between air and bucket temperatures. We use the results of this nomogram in the numerical form suggested by Burk et al. (1979).

Another problem is the interpretation of z_{oT}. This is not necessarily only a function of ocean roughness, but a measure of the molecular transfer of heat through the surface. However the mixing length model of Donelan (1984) demonstrates that as the surface roughness increases so does the efficiency to heat and mass transfer so that z_{oT} and z_{oq} have the same general dependence on roughness Reynolds number ($R_* = z_o u_*/\nu$) as z_o. In the range of wind speeds of this experiment this is equivalent to:

$$\alpha_1 C_{HN} = C_{DN} = \alpha_2 C_{EN} \tag{15}$$

where C_{HN} and C_{EN} are the neutral equivalent heat and moisture transfer coefficients or Stanton and Dalton numbers. The values of α_1 and α_2 will be discussed below.

We then proceed to solve equ. (14) first for $\psi_h = 0$. This is combined as before with $-\overline{u'w'}$ computed for $\zeta = 0$, to form L. Hence the computation of $\overline{w'T'}$ is repeated several times leading to presumably improved estimates of L.

Although $\overline{w'q'}$ is not very important for C_T^2, bulk estimates of evaporation are important for other reasons. In analogy with equations (9) and (14), we then have

$$\overline{w'q'} = \frac{\kappa^2 (q_s - q) V}{\phi_q(o) \left[\ln \frac{z}{z_{oq}} - \psi_q(\zeta)\right] \left[\ln \frac{z}{z_o} - \psi(\zeta)\right]} \tag{16}$$

Here, q_s, the surface specific humidity, is evaluated from the surface skin temperature; ψ_q is equated with ψ_h.

Then, equ. (16) is solved by successive approximations, initially with $L = \infty$, and then with progressively more realistic L-values. In fact the iterative solutions of (9), (14) and (16) proceed simultaneously.

This whole procedure is analogous to that suggested by Burk et al. (1979) except that z_o, z_{oT} and z_{oq} are assumed to depend on the equivalent neutral 10 m wind.

2.3 Non-dimensional gradients.

The non -dimensional gradients of wind speed and temperature are defined as follows:

$$\phi_m = \frac{\kappa z}{u_*} \quad \frac{\partial V}{\partial z} \tag{17}$$

$$\phi_h = \frac{\kappa z}{T_*} \quad \frac{\partial T}{\partial z} \tag{18}$$

We have direct (eddy correlation) estimates of u_* and T_* but the gradients are approximated from differences of adjacent levels by:

$$\left[\frac{\partial A}{\partial z}\right]_{(z_1 z_2)^{1/2}} \doteqdot \frac{A_2 - A_1}{(z_1 z_2)^{1/2} \ln (z_2/z_1)} \tag{19}$$

Equation (19) is exact for a logarithmic profile but is slightly in error for diabatic profiles. The inexactness depends on the level spacing (i.e. the ratio z_2/z_1) and has been analysed in detail by Stearns (1970), who finds that for our case ($z_2/z_1 = 1.5$) the error is at most 1.5 %. This is substantially less than the average error in estimating A_* and so has been ignored.

3. The observations

Mean observations and fluctuations were measured on two separtate masts, 50 m apart. Fluctuations of wind components and temperature were made by a sonic anemometer-thermometer, manufactured by the Marine Instrument Company, Tokyo. It has a path length of 20 cm and a resolution of 2 cm/sec and 0.1°C. This instrument was mounted at 9.55 m above mean lake level. Temperatures were corrected for distortion due to moisture and wind fluctuations. Temperature fluctuations were also determined from a platinum wire thermometer manufactured by TSI, Inc., St. Paul, MN. Its frequency response was 5 khz. This instrument was originally mounted at 4.95 m height, and moved to 9.55 m after the June experiments. Unfortunately, the recording system of these temperature fluctuations was incompatible with the recording system used for the sonic anemometer-thermometer. Hence, the wire temperatures could not be directly correlated with vertical velocity to infer heat flux; but they were used to obtain C_T^2. Humidity fluctuations were measured by a Lyman-α humidiometer, manufactured by the Electromagnetic Research Corporation, College Park, MD. This instrument has a basic frequency response of 1500 Hz and was mounted at 9.55 m. To protect the instrument from rain it was enclosed in a horn which could be closed remotely by a solenoid-operated butterfly valve. Although we have no direct check of the reduction in frequency response caused by this arrangement, the roll-off of the -5/3 region of the spectrum suggests faithful response for scales larger than 30 cm. This is roughly comparable with the sonic anemometer-thermomemeter and is certainly adequate for direct estimates of the moisture flux made at several meters above the surface. Calibration changes, caused by etching of the water-sensitive windows, made it necessary to monitor the calibration of the Lyman-α humidiometer with a more stable device. We achieved this by comparing the low frequency spectral estimates with those obtained from a Dew Point Hygrometer. This check was made for each run and the calibration of the Lyman-α Humidiometer adjusted accordingly (Golus, 1983).

All turbulence data were recorded on a seven track Phillips FM-analog tape recorder and later digitized at 20 Hz for subsequent spectral analysis.

Mean wind speeds, temperatures and humidities were measured at logarithmically spaced heights of 12.1 m, 8.0 m, 5.2 m, 3.4 m and 2.1 m. Donelan et al. (1974) give detailed descriptions of the sensors.

Table 2 lists the runs, and their idiosyncrasies. Since the sonic instrument failed in the last runs (all in unstable air), no Reynolds fluxes are available for these cases. Most of the measured fluxes occurred in air with stable temperature gradients at 9.55 m. In contrast bulk estimates, and C_T^2 as well as C_q^2, were available for all runs.

4. Validation and corrections

Cup anemometers, as instruments for measuring the mean downwind component of the wind velocity, are subject to two main sources of error. The first of these arises because cup anemometers respond to the average magnitude of the resultant of the horizontal velocity components. Bernstein (1966) has analysed this error and shown that it is of the order of $\{1-\exp\left[-1/2\,(\sigma_v/U)^2\right]\}\times 100\%$. The variance of lateral velocity fluctuations, σ_v is given by Smith (1979), from measurements made at the same site during the same experiment; he finds that σ_v/U is about 0.05. The corresponding error is 0.12 % and is therefore negligible. The second error is caused by the tendency of cups to accelerate faster than they decelerate leading to an overestimate of the wind speed through "overspeeding ". The size of this error clearly depends on the design of the anemometer and on the gustiness of the wind. The latter is generally larger over land than water. By comparison with a sonic anemometer Izumi and Barad (1970) found 10 % overspeeding for the cups used over grassland by Businger et al. (1971). We have done the same here and found that the average overspeeding is 4 %. Therefore all cup measured wind speeds have been reduced by 4 %.

It has been argued (Friehe and Schmidt, 1976) that the anomalously large heat fluxes from eddy correlation measurements over the ocean are due, at least in part, to the appearance of "cold spikes" in the temperature data. These cold spikes are believed to be spurious and possibly due to the hygroscopic effect of salt particles adhering to the cold wire thermometers. Our temperature fluctuation measurements over fresh water with a sonic thermometer avoid these problems. However, it must be admitted that sonic thermometry is not free of problems. The signal-to-noise ratio is not large and substantial fluctuations of temperature are required before sonic thermometry is at all practical. Fortunately we have large air-water temperature differences (Table 3) and correspondingly high σ_T values. The heat flux measurements are not as plagued by low signal-to-noise ratios in the temperature since the noise is essentially uncorrelated with the vertical velocity signal. The apparent sonic temperature signal T_s must be corrected for fluctuations of humidity and of wind speed. The correction was given by Kaimal (1969) but unfortunately with a typographical error. The correct expression for the actual heat flux is

$$\overline{w'T'} = \overline{w'T_u'} - 0.51\ T\ \overline{w'q'} + \frac{4VT}{S^2}\ \overline{u'w'} \qquad (20)$$

where S is the speed of sound

The corrected ($\overline{w'T'}$) and uncorrected ($\overline{w'T_u'}$) heat fluxes are tabulated in Table 3, where it can seen that the corrections are typically 10 or 20 %, but, of course, can be arbitrarily large as neutrality is approached. All subsequent references to the heat flux refer to the corrected flux $\overline{w'T'}$.

5. Results

5.1 The non-dimensional wind gradient, ϕ_m (ζ)

Fig. 1 shows the non-dimensional gradient, ϕ_m versus the stability parameter ζ. In each case the gradient is derived from the average of the differences from the top three levels. The bottom two levels were not used since there is evidence (Donelan et al. 1974) that they are modified by the wave-induced Reynolds stress as predicted by Stewart (1961). The linear regression line is shown (dashed) but, since it is slightly above unity at $\zeta = 0$, we prefer to represent $\phi_m(\zeta)$ by:

$$\phi_m = 1 + 4\zeta \tag{21}$$

which is very close to the regression line ($\phi_m = 1.09 + 3.8\,\zeta$) and has the added merit that it implies no change in the von Karmân constant. Clearly the data are not precise enough to suggest any such change. The $\phi_m(\zeta)$ relationship found by Businger et al. (1971) is shown also (dotted line), and it is clearly but not greatly outside the range of our data. It is difficult to conceive of any reason why (above the zone of direct wave influence) ϕ_m should not be the same over land and water. However the flux measurements of Businger et al (1971) appear to be subject to some uncertainty (Wieringa, 1980) which may be sufficient to explain the difference between their results and ours.

The only other estimates of ϕ_m over water that we are aware of are due to Zank (1981), who found $\phi_m = 1 + 5.2\,\zeta$. However, his heat fluxes were bulk estimates which, as we shall demonstrate, are distinctly low when land-derived ϕ_h values are used.

5.2 The non-dimensional temperature gradient, $\phi_h(\zeta)$

The observed values of ϕ_h are plotted against ζ in Fig. 2. The temperature gradients were rather small and in only five cases were the profiles smooth enough to yield reasonable gradients. Each profile of five levels yielded four differences using only adjacent pairs. The corresponding ϕ_h values are shown with the top differences coded differently (closed circles) from the bottom (open circles).

The points are rather scattered and a straight line regression has the extremely low neutral value $\phi_h(o) = 0.1$. We have chosen instead to represent it by the quadratic in ζ shown (solid line):

$$\phi_h = 0.5 + 0.5\,\zeta + \zeta^2 \tag{22}$$

For comparison, the result of Businger et al. (1971) is also indicated (dotted line). It is highly unlikely that the enormous difference between these results could be due to error in either dataset. The remaining possibility is that ϕ_h is not a universal function of ζ. In the introduction (and Table 1) we drew attention to the generally unsatisfactory agreement between bulk and direct estimates of sensible heat flux over water. The bulk estimates are generally low by up to a factor of 2 (Table 1). This is just the error which would occur if ϕ_h is really about 1/2 that assumed (Fig. 2).

What agency could bring about such a large difference between temperature gradients over land and water? The four most likely possibilities are: (a) the presence of waves affects the transfer of heat and momentum differently; (b) the moisture laden atmosphere enhances radiative transfer; (c) buoyancy effects alter the simple flux-gradient concepts which are the basis for ϕ_h; (d) small droplets of spray in changing phase absorb substantial amounts of heat and are thereby an additional source of temperature fluctuations.

The first of these is unlikely to be the cause in our data since the waves are near full development and so the interaction with the wind near the spectral peak is weak. The shorter waves interact more strongly with the wind but the disturbance is negligible above one half wavelength and would not effect the highest levels of the temperature profile.

The radiative effects on a relatively opaque atmosphere have been analysed by Coantic and Seguin (1971) and Townsend (1958) from different points of view. Coantic and Seguin have calculated the mean radiative flux in a moisture laden boundary layer. According to their calculations the radiative flux would be less than 20 % of the sensible heat flux under the conditions of our observations. But, more importantly, the mean radiative flux does not alter the relationship between $\overline{w'T'}$ and $\partial T/\partial z$, i.e. it has no effect on the estimates of ϕ_h. Townsend (1958), on the other hand, looked at the effect of radiation on altering the rate of destruction of temperature fluctuations. He showed that radiative heat transfer always acts to reduce the magnitude of the convective heat transfer in a given velocity field and mean temperature gradient. This is just the opposite effect to that observed here and therefore its adjustments to the heat flux are clearly masked by another much stronger effect.

Priestley and Swinbank (1947) were the first to question the validity of an eddy conductivity approach to heat flux in the atmosphere. They argued that buoyancy forces would produce a "convective " heat flux which is always directed upwards and thus for positive temperature gradients is opposed to the down-gradient eddy flux. Deardorff (1966, 1972) has analysed this effect further and, by making plausible assumptions regarding the form of the pressure-strain term in the heat equation, has shown that the usual eddy conductivity expression for the heat flux must be modified to include a buoyancy term related to the ratio of the variances of temperature and vertical velocity. Under weakly stable conditions this buoyancy term can lead to upward directed "counter-gradient" heat flux. Warhaft (1976) has extended this analysis to include the effects of buoyancy due to moisture fluctuations and has shown that the correlation of temperature and moisture fluctuations enters the problem. According to Warhaft the complete expression for the heat flux is:

$$\overline{w'T'} = -K_h \left[\frac{\partial T}{\partial z} - \frac{1}{2} \frac{g}{\sigma_w^2} \left(\frac{\sigma_T^2}{T} + 0.61\ \overline{T'q'} \right) \right] \tag{23}$$

The factor 1/2 depends on the assumptions made regarding the pressure strain term and it may be different from 1/2 but, nonetheless, of order one. Although the $\overline{T'q'}$ term may be negative, it is under most circumstances swamped by the σ_T^2 term; so that for a given temperature gradient the heat flux is reduced in magnitude for stable stratification and increased for unstable stratification. The observations suggest a general increase in magnitude whatever the sign of the gradient. It would appear that known buoyancy effects are unable to account for the observed behaviour of the heat flux over water.

The correspondence between local gradients and fluxes also fails when the larger eddies associated with the fluxes encounter gradients in the planetary boundary layer which are very different from the surface gradients. Some of the consequences of this are discussed by Wyngaard and Brost (1984). The effect of large scale gradients is to alter the flux-local-gradient relation for the large eddies. However, one of the characteristics of temperature spectra over water is the enhanced high frequency variance (see, for example, Phelps and Pond, 1971) and this must be due to another cause.

The idea that substantial heat exchange through droplet evaporation may occur in the marine boundary layer has been advanced before. Various ideas and relevant measurements have been ably summarized in a review paper by Wu (1979). Wu's calculations of spray evaporation suggest enhancements of the evaporative flux of only 1 % or so at these wind speeds. Since the heat flux is generally 10 % or more of the latent heat flux (in these data they are about equal) and most of the droplet evaporation will occur near the surface it does not appear that heat flux divergence caused by spray evaporation can have much effect on ϕ_h several metres above the surface. However, the evaporating droplets leave a trail of cold air and are in effect a source of temperature fluctuations quite separate from those induced by eddy mixing of the mean temperature gradient. Townsend (1958) has demonstrated that radiative heat exchange reduces the variance of temperature fluctuations, thereby reducing $|\overline{w'T'}|$ or ϕ_h for a given $\partial T/\partial z$. The evaporating droplets increase the variance of temperature fluctuations and therefore have the opposite effect on ϕ_h. Further experimental work needs to be done to attempt to sort out the reasons for the differences between land and sea versions of ϕ_h. It seems to us that exploring the role of droplet evaporation in increasing temperature variance is a promising place to start.

Dunckel et al (1974) have examined the correlation between different frequency intervals of the heat flux cospectrum and the product $V(T_s - T_z)$ and found that only a central region of the cospectrum , amounting to 25% of the total flux, is correlated with the bulk estimator $V\Delta T$. Unfortunately they did not report a similar test for the moisture flux cospectrum so it is difficult to know how extraordinary this is. However, on the face of it, it would seem to augur poorly for the reliability of the bulk technique for heat flux over water. At the same time, the 53 % of the total heat flux on the high end of the cospectrum and uncorrelated with $V\Delta T$ may arise through the generation of temperature variance independently of the shear production. This is approximately the part of the spectrum of temperature fluctuations shown by Phelps and Pond (1971) to be anomalously high and it may well be that the agency for the production of temperature variance in this region is the evaporation of droplets in the marine boundary layer.

Of course, the effect of droplet evaporation on inducing moisture fluctuations is very much smaller because of the large latent heat of vapourization of water.

5.3 Friction velocity

In comparing bulk and direct fluxes the best agreement was achieved with $\alpha_1 = 0.6$ and $\alpha_2 = 1.0$ in (15), and with the ϕ functions given by (21) and (22) under stable conditions. For unstable stratification we used the ϕ_m given by Businger et al (1971) and their $\phi_h \times 0.5/0.74$ to agree with (22) at $\zeta = 0$. In all cases $\phi_q = \phi_h$.

Fig. 3 compares bulk and direct estimates of friction velocity. Three of the unstable "direct" values were obtained from profiles when no flux data were available. Generally the agreement is not bad. Some idea of the importance of the stability corrections to the bulk estimates can be gleaned from Fig. 4, which shows the very strong stability dependence of the uncorrected drag coefficient, u_*^2/V_{10}^2. The curve has been computed from (9) and (21) using the neutral drag coefficient appropriate to the average wind speed for all the runs (6.4 m/s). While the curve runs through the centre of the data, the points at low values of ζ are generally higher than those at high stability. Since strong stability is negatively correlated with wind speed this could well be an effect of wind speed on z_o. An attempt to test Charnock's relation produced rather poor correlation between measured z_o and u_*^2/g. Some of the reason for the scatter can be seen in Table 3 in which the roughness Reynolds number ($R_* = z_o u_*/\nu$) is indicated. Some of the values are less than the value appropriate to smooth flow ($R_* = 0.137$). These, cases are, in an aerodynamic sense, "smoother" than a completely smooth rigid surface. This is not, in principle , possible with a completely passive surface acting as an absorber of wind momentum. Some of these ultra-smooth cases have R_* values only a factor of 5 or less beneath the smooth limit and this could arise through an error in the estimate of z_o. (Remember that the process of estimating z_o involves a substantial correction for stability). However, in three cases (indicated by +) the R_* values are several orders of magnitude below the smooth limit. In these cases the fetch available is substantially greater than the fetch required for full development. (Both these fetches were computed by the method of Donelan (1980) as outlined in Bishop (1983)). In the other cases the ratio of fetches is near or less than one, indicating that the wave field is not or just fully developed. Full development is the term applied to a wave field, which under the influence of a steady wind, does not undergo any further perceptible development. Observations suggest that this occurs when the ratio of wind speed to phase speed of the peak waves $V/c_p = 0.83$. At this stage the net wind input of energy to the waves is entirely consumed locally in wave dissipation.

If such waves propagate from a region of higher wind to one of lower wind they will be overdeveloped and the largest waves will outrun the wind, and, by returning momentum to the wind, reduce the drag on the surface. This is known to happen and has been documented by Davidson (1974) among others. In our view it is very likely that this effect is one of the causes for scatter in estimates of oceanic drag coefficients, for which there is regrettably often no concomitant wave data. We have indicated in Table 3 the values of V/c_p for each case calculated from the fetch limited formulae of Donelan (1980). Values less than 0.83 signify over-development and it can be seen that the ultra-smooth R_* values are associated with the lowest V/c_p values. Strictly speaking the fetch-limited formulae are only valid at $V/c_p \geq 0.83$, but we have extended them here only to indicate, by their low V/c_p values, which cases are thoroughly developed and may be most likely to return momentum to the wind should the wind drop even slightly.

In our bulk calculations of friction velocity we have ignored effects associated with the state of wave development and this is evidently part of the reason for the scatter in Fig. 3. Methods for incorporating such effects are given by Kitaigorodskii and Volkov (1965) and by Donelan (1982).

However, these methods are strictly only applicable to fully rough flow ($R_* > 2.2$) and therefore have not been applied to our data (Table 3).

In Fig. 5, we compare friction velocities estimated by the dissipation method with the direct (eddy correlation) estimates. The agreement is good but not appreciably better than the bulk method (Fig. 3). Generally the dissipation method tends to overestimate u_* at low u_* and underestimate at high u_*. Again wind speed and $|\zeta|$ are negatively correlated and this behaviour may be due to an incorrect choice of $\phi_\varepsilon(\zeta)$.

5.4. Sensible heat flux

Fig. 6 compares the heat flux estimates by the bulk method with the direct measurements. The general agreement is quite good except for a few cases with generally small heat flux when the direct estimates were negative but the bulk estimates were positive. For some of these cases temperature profiles were available and these showed that the air was slightly stable around 10 m, where $\overline{w'T'}$ was measured, in agreement with the negative sign of $\overline{w'T'}$. But the surface temperature was higher than the temperature at 10 m leading to upward heat flux by the bulk method. It is likely that such profiles are produced when the air has passed over colder water and formed a stable temperature profile. On approaching the tower where the water was warmer an internal boundary layer develops with an unstable profile near the surface. While we have no record of the temperature distribution around the tower, such patchiness in the surface temperature can be caused by the development of the thermal bar in the Spring (Rodgers, 1965) or by the meandering of the plume of the Niagara River. If we had extrapolated the stable upper portion of the temperature profile, to obtain the upstream surface temperature appropriate to the negative heat flux above the internal boundary layer, then the bulk fluxes would also have been negative and in good agreement with the measured $\overline{w'T'}$.

5.5 Moisture flux

Fig. 7 compares bulk and direct moisture flux $\overline{w'q'}$. While there is considerable scatter, no particular bias is evident. In these bulk estimates we have used $\phi_q = \phi_h$ rather than $\phi_q = \phi_m$ on the grounds that moisture and heat are dynamically more similar than moisture and momentum in their transfer properties. However, the peculiarities of temperature spectra over water are not evident in the moisture spectra (Phelps and Pond, 1971) and our favoured droplet evaporation mechanism for the enhancement of heat flux is not likely to be nearly so efficacious for moisture flux. Clearly, careful measurements must be made to determine ϕ_q directly from flux and profile data. Unfortunately, our humidity profiles were not sufficiently accurate to be useful in this regard.

5.6 The temperature structure function constant, C_T^2

C_T^2 (equation (8)) contains a non-dimensional function of stability $J(\zeta)$. Panofsky and Dutton (1983) have shown that

$$J\ (\zeta) = 3\ \kappa^{-2/3}\ \phi_h \phi_\varepsilon^{-1/3} \tag{24}$$

which may be evaluated from the non-dimensional gradients already quoted.

In Fig. 8 we compare C_T^2 computed from the cold wire temperature spectra and C_T^2 obtained from bulk methods. Reasonable agreement is achieved for the stable points and for those unstable points in which bulk and direct heat fluxes agree in sign. Where there is disagreement in sign the bulk estimates of ζ will have the opposite sign to the local (direct) ζ and thus equation (8) will be inadequate. The neutral cases show generally poor agreement, which is not surprising since the gradient production of temperature variance will be small and so the other source, unrelated to temperature gradient, will be dominant and cause the spectral estimates to exceed the bulk.

5.7 The moisture structure function constant, C_q^2.

In Fig. 9 we compare bulk estimates of C_q^2 with those determined from the moisture spectra. There is substantial scatter but generally agreement is not bad and there is no obvious bias.

6. Summary

Our purpose in this paper has been to attempt to evaluate the bulk aerodynamic methods for the calculation of fluxes and structure function constants over water. However, we found that the non-dimensional temperature gradients established over land led to substantial underestimates of the heat flux and hence the stability. With our concurrent profile and flux data we were able to establish the non-dimensional gradients of wind and temperature. The former appears to be consistent with over-land observations, while the latter is quite different. The difference amounts to a much larger efficiency of heat transfer over water than land. We discussed various agents for bringing about such an increase in efficiency including: effects of waves; radiative transfer, buoyancy; droplet evaporation. All things considered, droplet evaporation is the most likely cause of the increase in heat transfer efficiency over water compared to that experienced over land.

Using the newly derived expressions for the non-dimensional gradients of wind and temperature (from flux and profile data) we compared the bulk estimates of fluxes and structure function constants with the direct measurements (from mean and turbulence data) and found the bulk methods generally acceptable but only if the over-water non-dimensional gradients are used.

We have offered the hypothesis that the increase in temperature variance caused by droplet evaporation leads to the observed increase in efficiency of heat transfer in the boundary layer over water compared to that over land. The question is far from settled and clearly needs some well designed experimentation to point the way for further theoretical development.

Table 1. Summary of previous over-water bulk-profile-direct comparisons of fluxes.

Reference	Location	Fluxes			Comments
		Bulk	Profiles	Direct	
1. Smith and Anderson (1984)	North Atlantic from Sable Island Nova Scotia			τ, H, E	ϕ_{TT} and ϕ_{qq} are alike, but ϕ_{WT} peaks at higher freqs. than ϕ_{wq} or ϕ_{uw}.
2. Zank, S (1983)	Tropical Atlantic 9°N (GATE)		τ, H	τ, H	Good agreement between profiles and direct. ϕ_h in reasonable agreement with land values for $\zeta<o$. Buoy data.
3. Zank, I (1981)	North Sea		τ	τ	$\phi_m=1+5.2\zeta$; $o<\zeta<o.45$, but ζ obtained from bulk H which would tend to underestimate ζ according to this work.
4. Francey and Garrett (1979)	East China Sea (AMTEX' 75)	τ,H,E		τ,H,E	C_D and C_H are fcn $(\bar{U})$, $C_E \neq$ fcn $(\bar{U})$. Measurements made from an island. At higher wind speeds C_H is anomalously large.
5. Hasse et al (1978)	Tropical Atlantic 9°N (GATE)	τ,H,E	τ,H,E		Good agreement between profile fluxes and bulk especially τ and H.
6. Friehe and Schmitt (1976)	North Pacific (NORPAX) 35°N	H,E		H,E	A review with some new data. Heat flux anomalously high on several occasions. These are associated with spurious cold spikes in temperature.
7. Dunckel et al. (1974)	Tropical Atlantic 7°N (ATEX)	τ,H,E	τ,H,E	τ,H,E	Good agreement between profile and bulk methods and bulk and direct except for heat flux which is 40% higher by direct than bulk methods. Poor correlation between bulk and direct heat flux
8. Kruspe (1974)	German Bight North Sea			H,E	Peak of temperature spectra at considerably higher frequencies than humidity spectra. However, heat flux and moisture flux cospectra were similar.
9. Müller-Glewe and Hinzpeter (1974)	Baltic Sea	H		H	Good agreement between direct and bulk methods.
10. Smith (1974)	Lake Ontario (IFYGL)	τ,H,E		τ,H,E	Peaks of temperature spectra and heat flux cospectra at considerably higher frequencies than humidity and moisture flux.
11. Paulson et al. (1972)	Tropical Atlantic (BOMEX)		τ,H,E	τ,H,E	Good agreement for stress and evaporation but direct heat fluxes are twice profile estimates.
12. Phelps and Pond (1971)	Tropical Atlantic (BOMEX) and San Diego			τ,H,E	High frequencies accentuated in temperature spectra over humidity spectra. This is more pro-

Table 1 (cont.)

Reference	Location	Fluxes			Comments
		Bulk	Profiles	Direct	
					nounced in Bomex than in San Diego data. In both cases heat flux cospectra peak at significantly higher frequencies than do moisture flux cospectra.
13. Pond et al. (1971)	Tropical Atlantic (BOMEX) and San Diego	τ,H,E		τ,H,E	Bulk and direct fluxes agree well except Bomex heat fluxes in which the direct are twice as great as the bulk.
14. Hicks and Dyer (1970)	Bass Strait, Australia	τ,H		τ,H	Good agreement for stress between bulk and direct methods. Poor agreement for heat flux but no appreciable bias.

Run	1972 Date	Duration (min)	Data Code see Table 2a	Run	1972 Date	Duration (min)	Data Code see Table 1a
3	May 22	24	A	23	June 16	46	A
4	22	26	A	25	21	32	A
4-1	22	26	A	26	22	11	A
5	24	30	A	27	24	63	A
6	24	31	A	30	Aug. 25	90	B
6-1	24	18	A	31	31	82	B
7	24	24	A	34	Oct. 13	40	C
8	25	45	A	34-1	13	40	C
9	25	13	A	35	13	50	C
10	25	53	A				
11	25	27	A				
12	25	51	A				
13	25	43	A				
14	25	44	A				
15	26	18	A				
16	26	61	A				
18	26	91	A				

Table 2. Date and duration of each run used for analysis, and data codes defined in table 2a.

Data Code	Fast response Sensor Height 4.95 m	9.55 m	Profile Information Available
A	cold wire	ultrasonic, LAH	yes
B		ultrasonic, LAH cold wire	yes
C		LAH cold wire	yes

Table 2a. Explanation of code used in table 2
LAH: Lyman-α Humidiometer.

Table 3. Summary of bulk conditions of data, heat flux corrections and effects of approach to full development.

Run	V_{10} cm s^{-1}	ΔT K	Δ_q $g.Kg^{-1}$	$-\overline{W'T'_u}$ cm s^{-1}K	$-\overline{W'T'}$ cm s^{-1}K	u_* cm s^{-1}	$C_D x10^3$ -	R_* -	Fetch Ratio -	V/σ_p -
3	539	0.25	-1.2	0.97	1.12	13.2	0.60	0.093	0.35	1.05
4	633	0.88	-1.3	2.05	2.25	14.4	0.52	0.14	0.30	1.09
4-1	636	1.90	-1.1	1.85	1.99	11.9	0.35	0.025	0.35	1.05
5	445	-.99	-2.0	1.20	1.43	15.4	1.20	7.5	0.87	0.85
6	484	-.63	-2.0	1.14	1.36	15.7	1.05	3.0	1.08	0.82
6-1	505	-.26	-1.9	1.97	2.21	16.6	1.08	8.0	1.02	0.83
7	446	1.88	-1.1	0.40	0.42	5.1	0.13	0.0006^{+}	1.27	0.79
8	1070	2.48	-1.5	3.57	4.96	31.9	0.89	0.79	0.22	1.17
9	1080	3.38	-1.6	6.36	8.16	35.9	1.11	4.1	0.22	1.18
10	977	3.70	-1.4	4.75	5.89	29.7	0.92	1.4	0.26	1.13
11	868	4.11	-1.4	4.75	5.57	26.5	0.93	2.0	0.34	1.07
12	819	3.58	-1.6	3.40	3.98	21.9	0.72	0.38	0.39	1.03
13	764	3.15	-1.3	3.55	4.06	21.5	0.79	0.93	0.43	1.01
14	576	2.10	-0.9	1.53	1.69	12.6	0.48	0.088	0.76	0.89
15	575	2.64	-1.1	2.51	2.75	15.0	0.68	1.7	0.76	0.88
16	532	1.21	-1.0	1.19	1.37	12.8	0.58	0.16	0.89	0.85
18	491	1.05	-0.8	1.00	1.17	13.6	0.77	0.57	1.05	0.82
23	409	1.06	-1.5	0.37	0.47	9.3	0.52	0.023	0.31	1.08
25	360	0.87	-0.8	0.14	0.15	5.6	0.24	0.00003^{+}	1.34	0.77
26	612	0.07	-	0.78	1.17	22.4	1.34	5.2	0.15	1.27
27	465	-.77	-0.6	-1.39	- 1.15	13.1	0.79	0.026	0.55	0.95
30	525	0.11	-	0.76	1.87	12.3	0.55	0.022	0.20	1.20
31	453	-.24	-3.4	0.17	0.65	8.8	0.38	0.0002^{+}	1.23	0.79
34	826	-4.46	-4.7							
34-1	804	-4.60	-4.7							
35	742	-5.43	-5.0							

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Figure captions

1. The non-dimensional wind gradient ϕ_m versus measured stability, ζ. (---) regression line; (——) relationship used; (....) Businger et al. (1971).

2. The non-dimensional temperature gradient ϕ_h versus measured stability, ζ. (——) relationship used; (.....) Businger et al. (1971).

3. Comparison of bulk versus direct estimates of friction velocity, u_*. Closed circles - stable; open circles - neutral (i.e. bulk estimates of stability, ζ_B, less than 0·2 in magnitude); open triangles - unstable and abscissal values are direct estimates; closed triangles - unstable and abscissal values are profile estimates. Where the symbol is accompanied by a horizonal tick the bulk and direct estimates of heat flux differ in sign.

4. The diabatic drag coefficient, u_*^2/V_{10}^2 versus measured stability, ζ. (See caption to Fig. 3 for symbolic code)

5. Comparison of dissipation versus direct estimates of friction velocity, u_*. (Symbols as for Fig. 3 except the direct measure of stability, ζ was used instead of the bulk, ζ_B).

6. Comparison of bulk versus direct estimates of heat flux, H. (Symbols as for Fig. 3).

7. Comparison of bulk versus direct estimates of moisture flux, E. (Symbols as for Fig. 3).

8. Comparison of bulk versus direct estimates of the temperature structure function constant, C_T^2. (Symbols as for Fig. 3).

9. Comparison of bulk versus direct estimates of the moisture structure function constant, C_q^2. (Symbols as for Fig. 3).

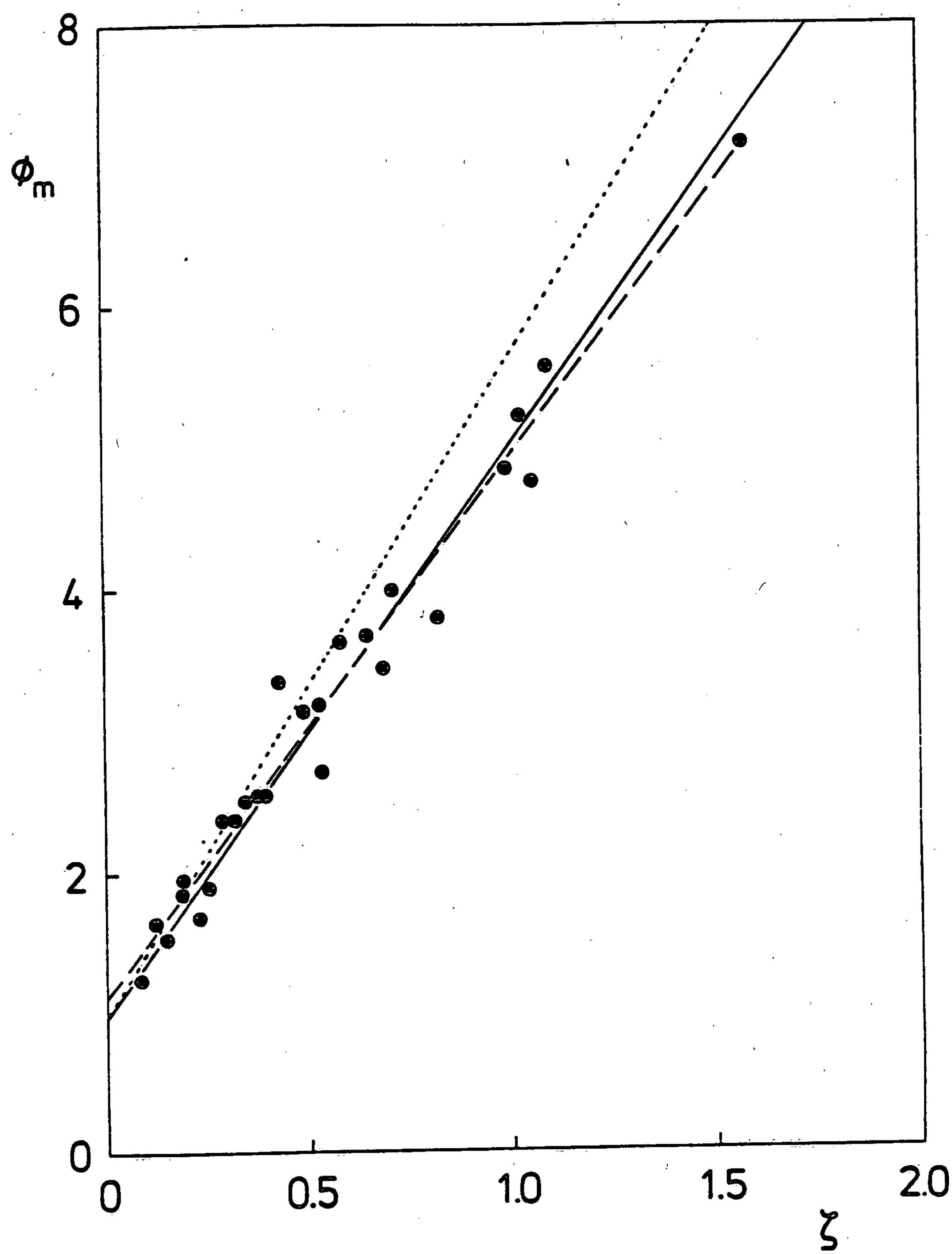


FIG. 1

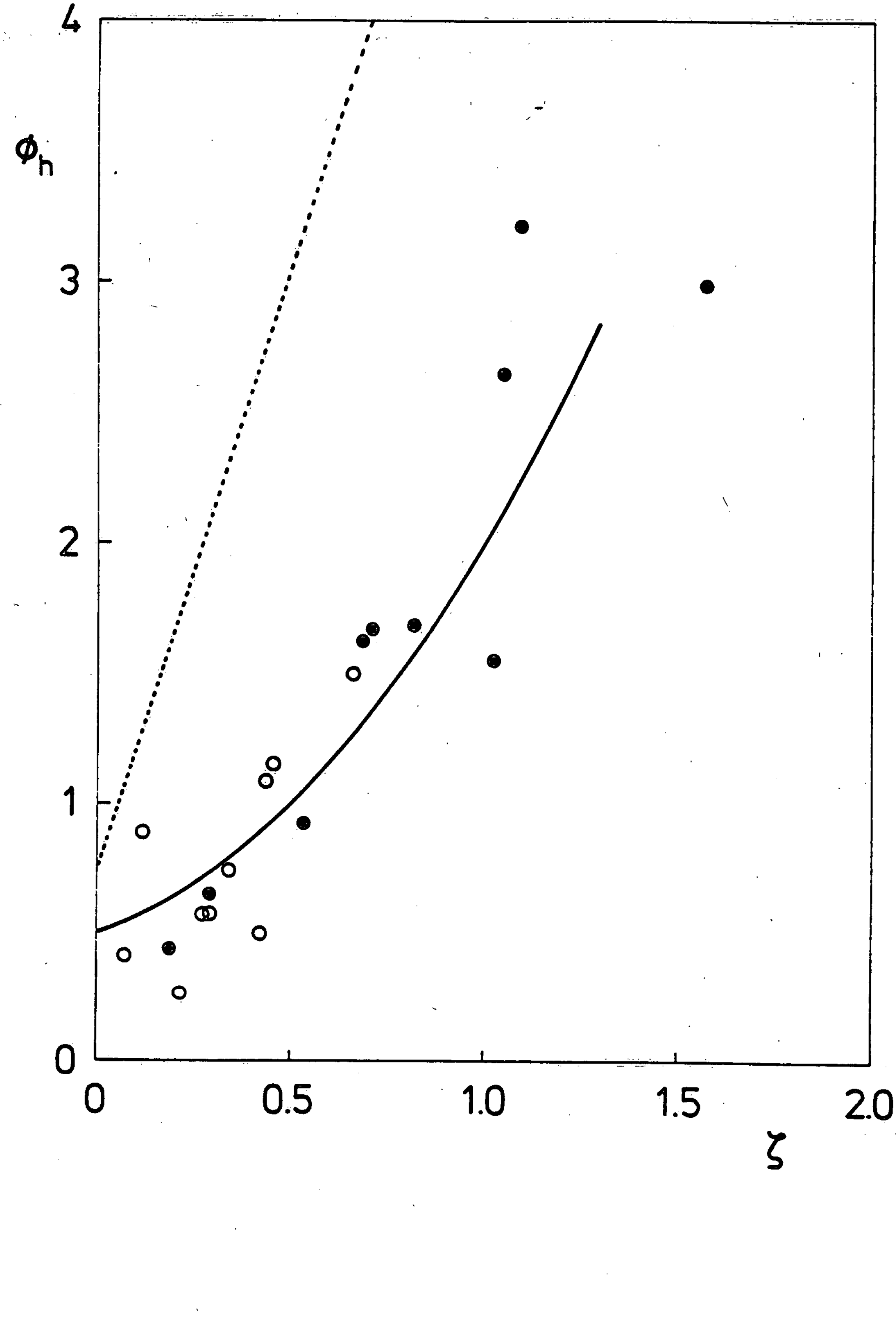


FIG. 2

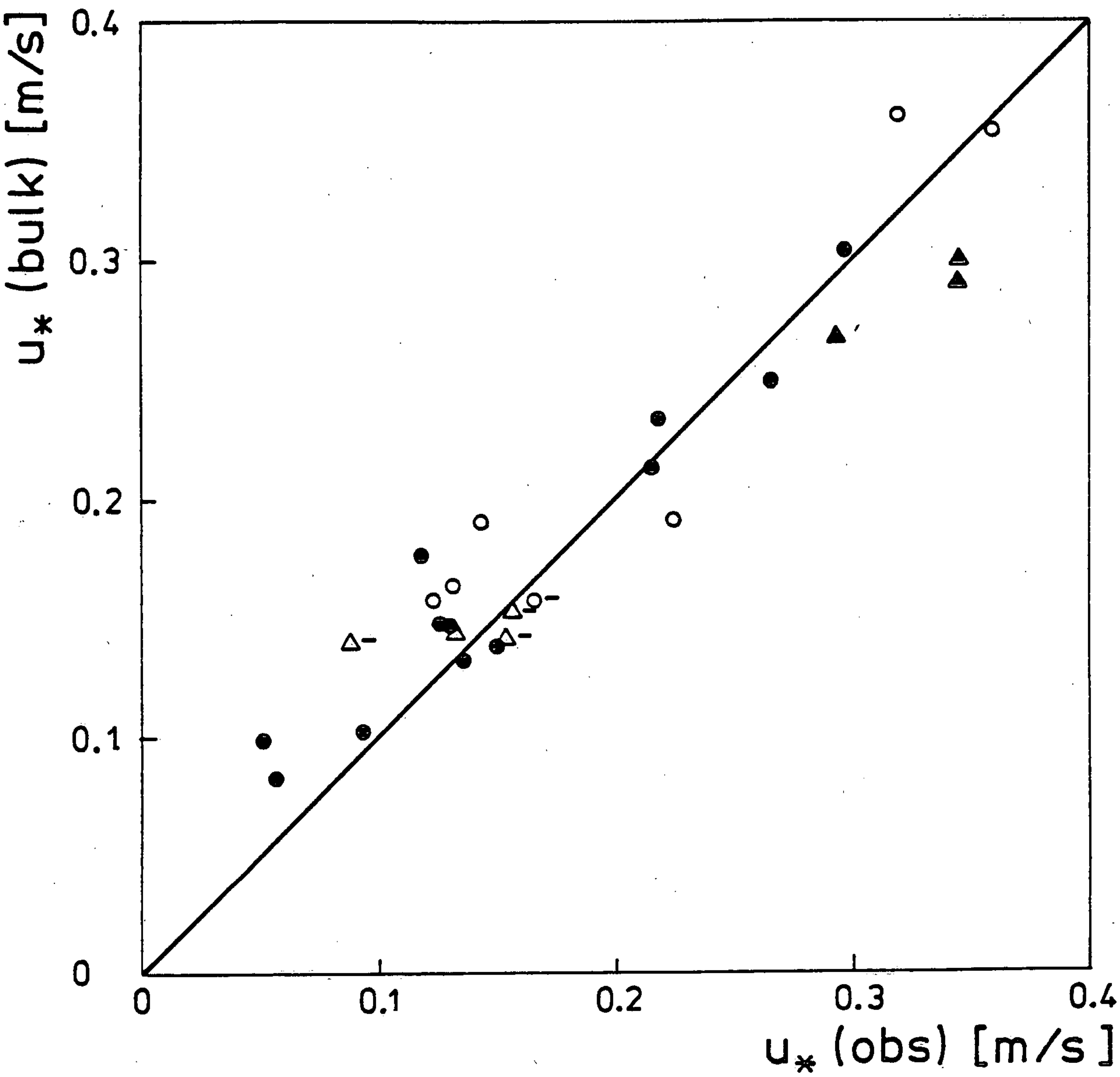


FIG. 3

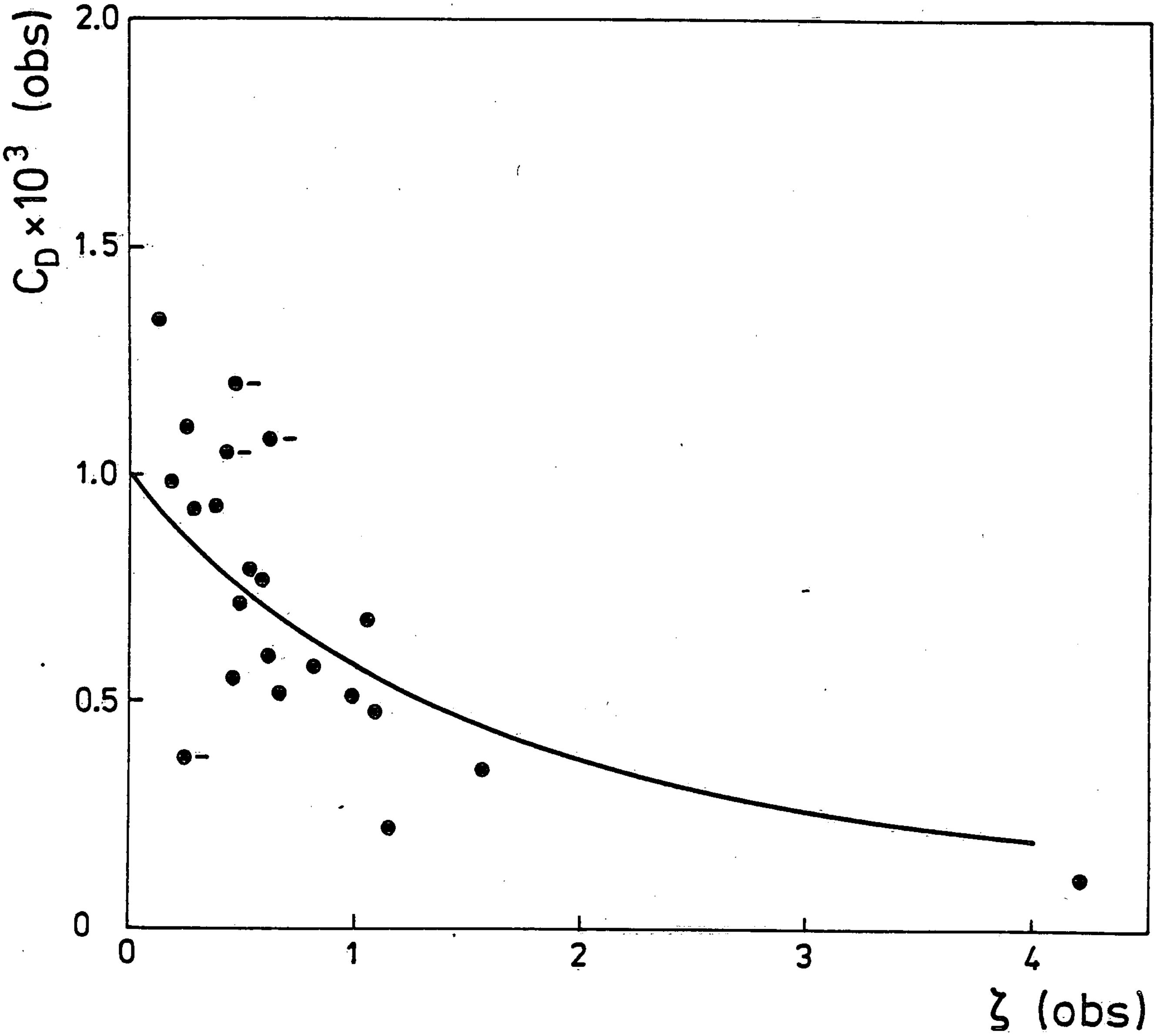


FIG. 4

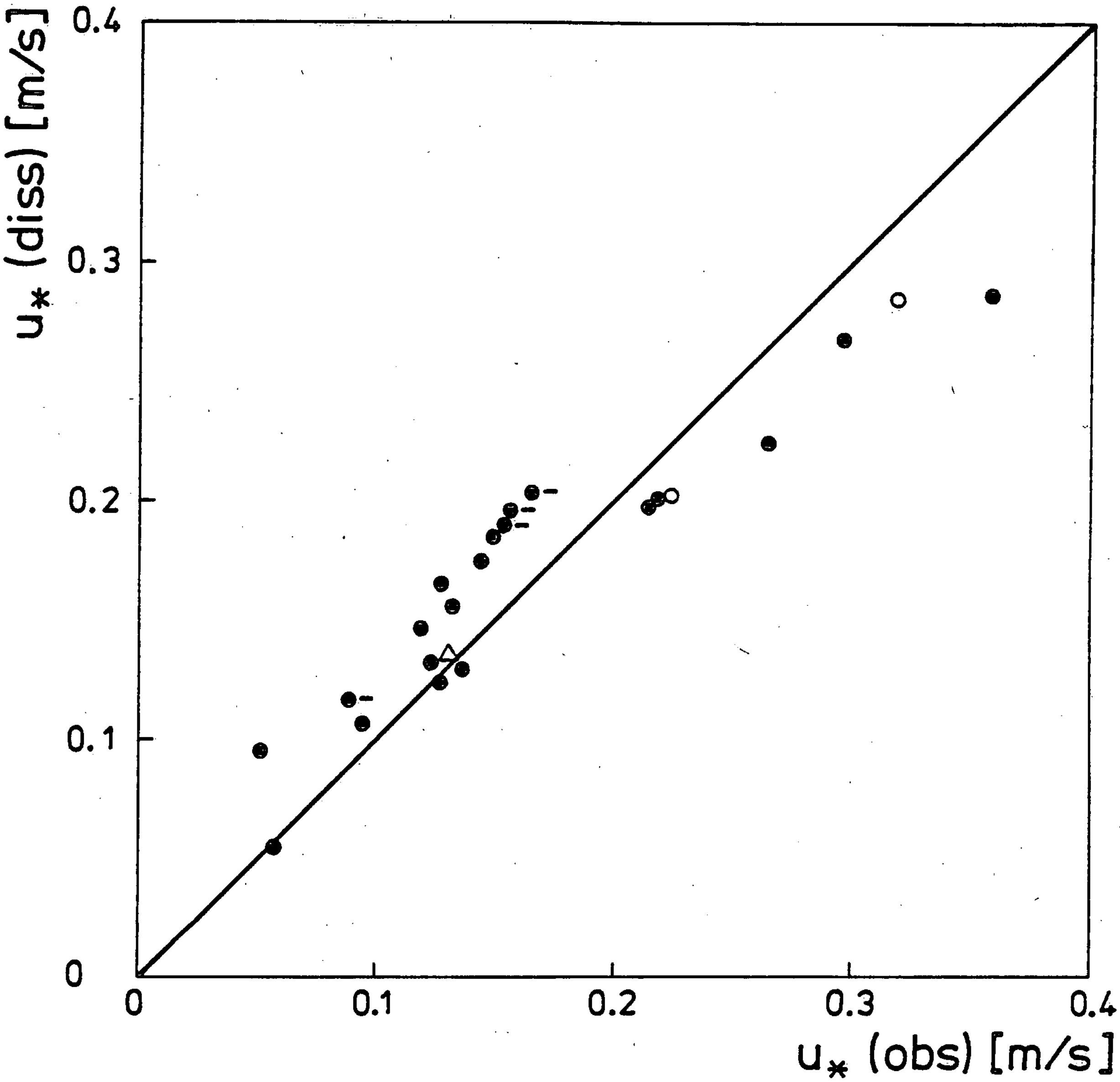


FIG. 5

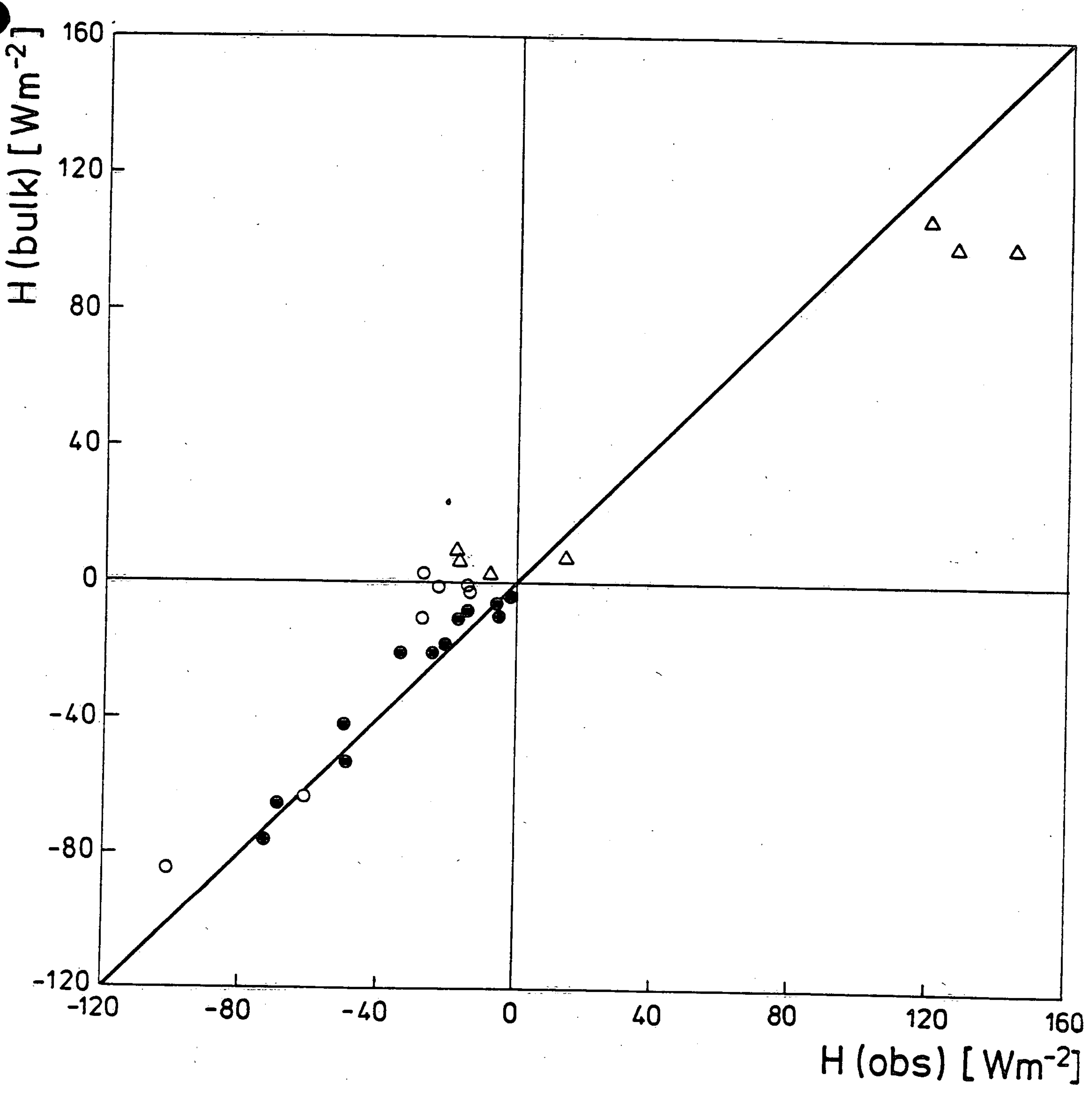

160
120
80
40
0
-40
-80
-120
H (bulk) [Wm^-2]
-120
-80
-40
0
40
80
120
160
H (obs) [Wm^-2]

FIG. 6.

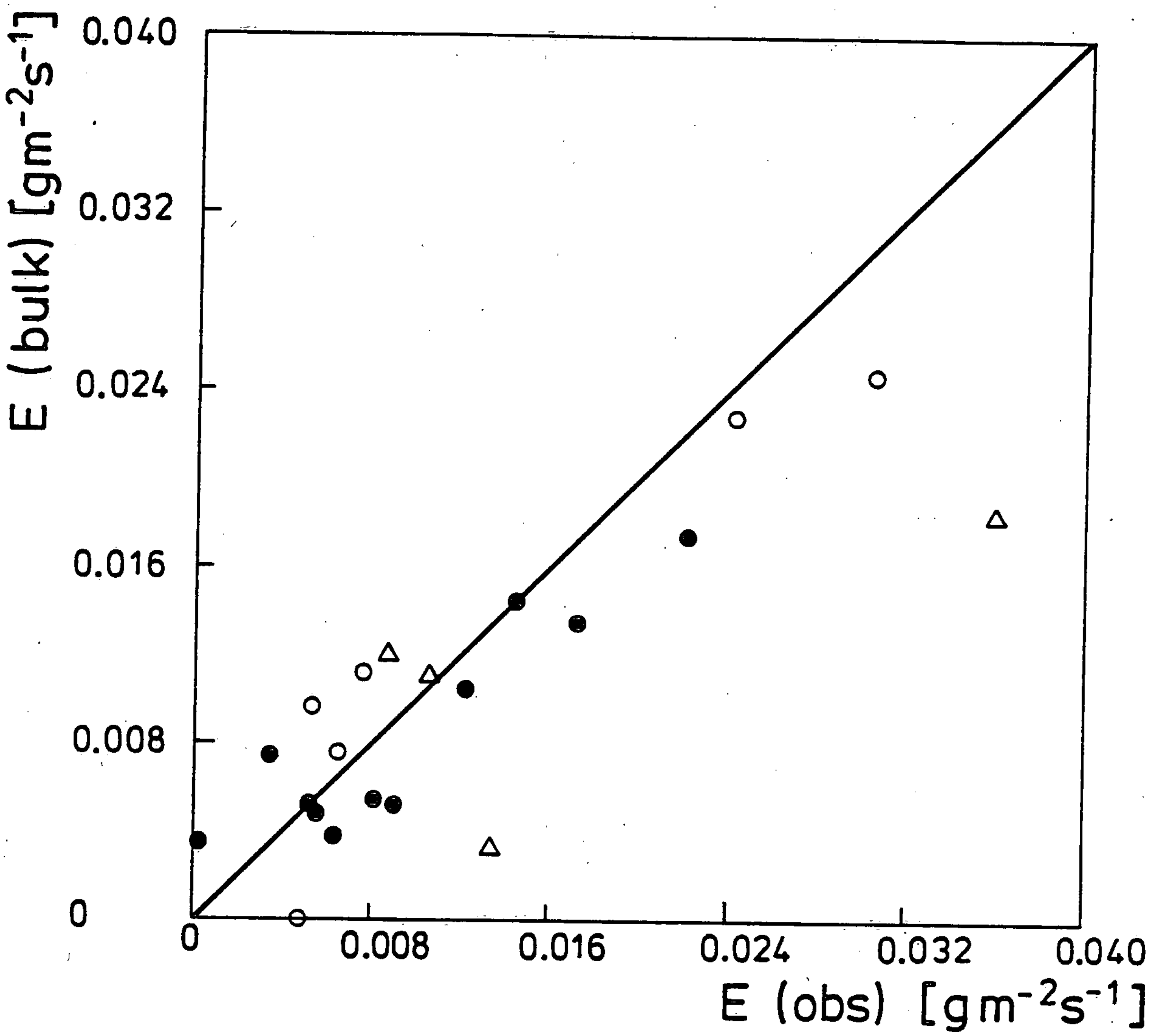
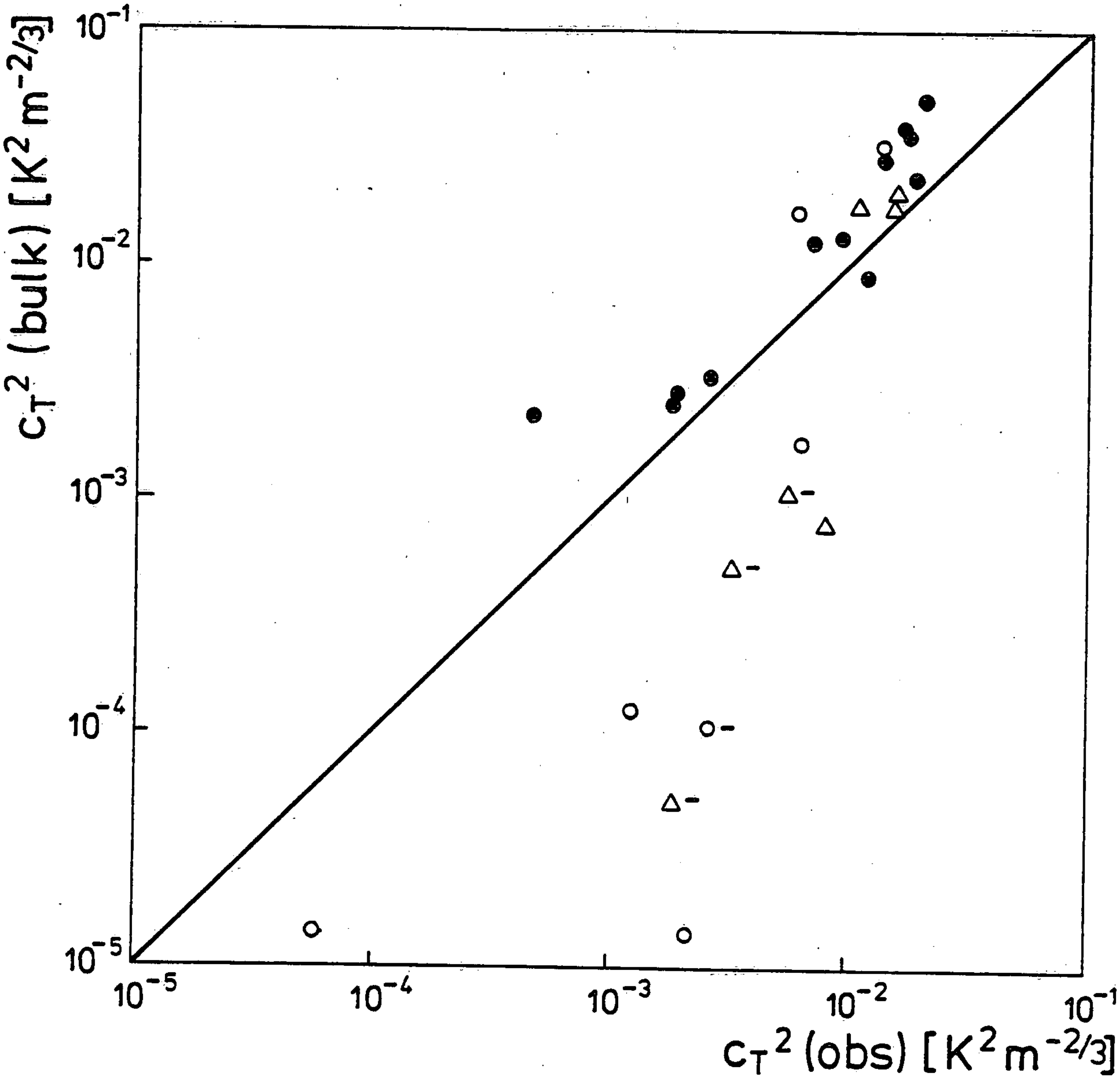
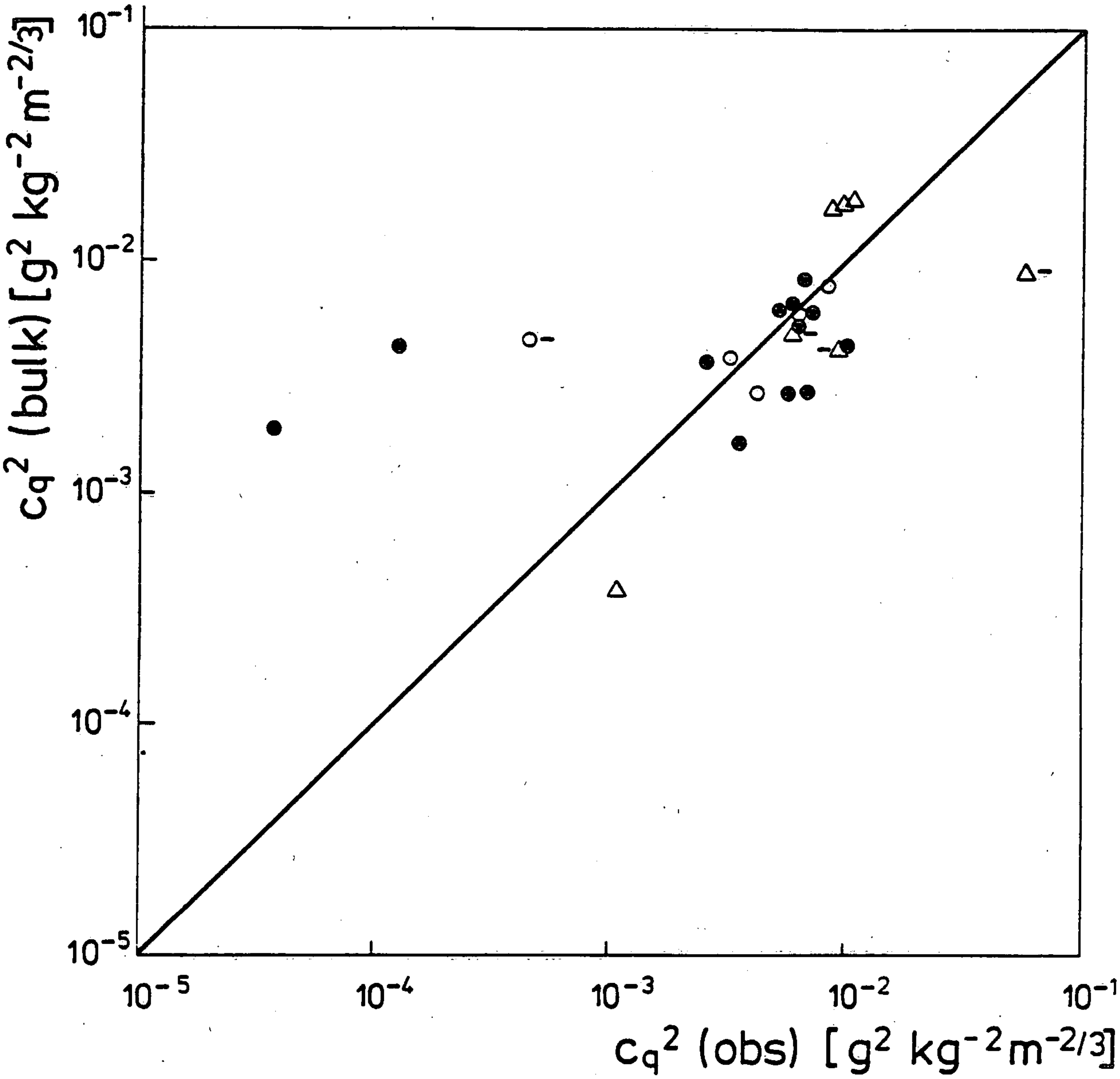

E (bulk) [gm-2s-1]
E (obs) [gm-2s-1]
0.040
0.032
0.024
0.016
0.008
0
0
0.008
0.016
0.024
0.032
0.040

FIG. 7



c_T^2 (bulk) [$K^2 m^{-2/3}$]
10^{-1}
10^{-2}
10^{-3}
10^{-4}
10^{-5}
10^{-5}
10^{-4}
10^{-3}
10^{-2}
10^{-1}
c_T^2 (obs) [$K^2 m^{-2/3}$]

FIG. 8



c_q^2 (bulk) [$g^2\ kg^{-2}\ m^{-2/3}$]
10^{-1}
10^{-2}
10^{-3}
10^{-4}
10^{-5}
10^{-5}
10^{-4}
10^{-3}
10^{-2}
10^{-1}
c_q^2 (obs) [$g^2\ kg^{-2}\ m^{-2/3}$]

FIG. 9.

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