THE DEPENDENCE OF THE MEASURED COOL SKIN OF THE OCEAN ON WIND STRESS AND TOTAL HEAT FLUX

HARTMUT GRASSL
Max-Planck-Institut für Meteorologie, Hamburg, F.R.G.

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Abstract. The temperature drop ΔT between the ocean surface and the 5-cm depth was recorded during GATE, Phase III. With measured values of the total heat flux Q and an assumption about the thickness of the viscous boundary layer of the ocean, the wind-speed dependence of the factor of proportionality between ΔT and Q is determined. This factor depends on the deviations of the thickness of the conductive layer from the thickness of the viscous layer and possibly partially on the wind stress. A further assumption about the thickness of the conductive layer leads to a wind-speed dependence of the ratio between total wind stress and its wave supporting part of it. This ratio increases from a value 1.5 at 1 m s^{-1} to 9 at 10 m s^{-1}, which is in agreement with existing estimates.

1. Introduction

The sea surface is usually cooler than the water immediately below. In recent years, a thin viscous boundary layer has been postulated and measured by different authors (Saunders, 1967; McAlister and McLeish, 1969; Clauss *et al.*, 1970; Hasse, 1971; Paulson and Parker, 1972; Katsaros, 1975). This thin surface layer, which is up to one millimetre in thickness in most circumstances, can be detected by the measurement of temperature gradient across the layer. This temperature gradient is necessary to maintain the energy flux from the ocean to the atmosphere or in some rare occasions in the opposite direction. The above-mentioned authors have found a temperature gradient at wind velocities of up to 10 m s^{-1}. At higher velocities, the destruction of the viscous layer by white capping is so dominant, that the recovery time of about 10 sec (Clauss *et al.*, 1970) is insufficient to rebuild the viscous layer between white caps. Since the wind velocity of large parts of the oceans is below 10 m s^{-1}, the viscous surface layer of the ocean is a basic phenomenon of water surfaces on the planet earth.

Together with the viscous layer, there should exist a conductive layer with a somewhat smaller thickness, but of the same order of magnitude. In this conductive layer, heat transport is only possible by molecular conduction. Knowing the temperature drop from surface to subsurface water and the thickness of the conductive layer, the total heat flux from the ocean to the atmosphere could be determined. This heat flux is the main energy input to the atmospheric circulation, because a high percentage of the incoming solar radiation is absorbed in the upper few metres of the ocean. This energy is given back to the atmosphere at the same or at a different location (after

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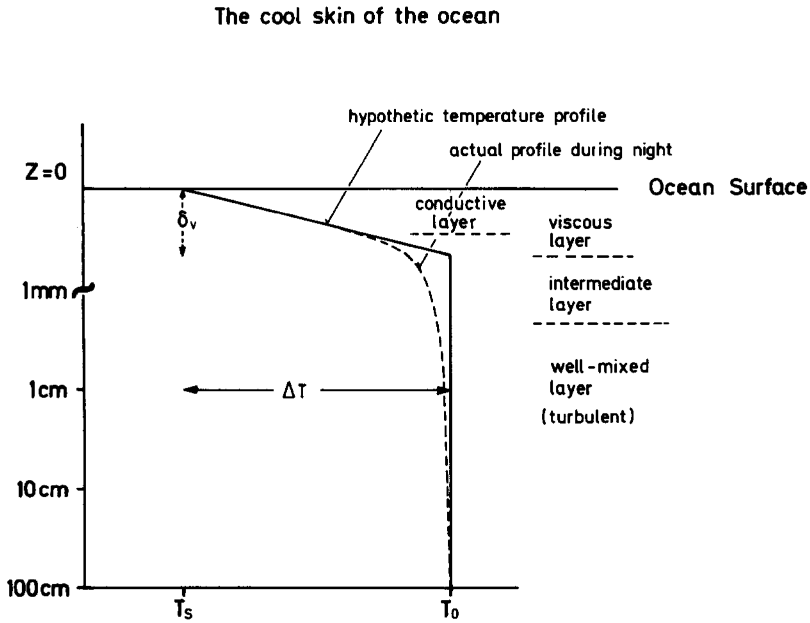

Fig. 1. Schematic representation of the near surface region of the ocean together with basic equations.

transport in ocean currents) in the form of turbulent heat fluxes and by the net long-wave radiative flux at the ocean surface. Therefore a large part of the energy driving the atmospheric circulation has to cross the viscous boundary of the ocean.

The total heat flux Q is the sum of the net radiative flux q_R, the turbulent flux of sensible heat q_s and the turbulent flux of latent heat q_L. The notations for and the magnitudes of the different parameters in the near surface layer of the ocean are shown schematically in Figure 1.

Fortunately we were able to measure all relevant parameters during GATE (GARP Atlantic Tropical Experiment), Phase III, August to September 1974, on board the German research vessel *Planet*. Together with all standard meteorological observations and the temperature difference between the surface and the 5-cm depth, we attempted to confirm already existing models or to reveal some new characteristics of the surface layer. With the experimental setup used, no gradient in the postulated conductive layer could be measured. Therefore we have had to assume a dependence of the layer thickness on wind stress. With this assumption and, for example, a knowledge of the total heat flux, we can perhaps show the ratio of total wind stress to that part of the stress necessary to maintain the wave field of the ocean. After some theoretical remarks in Section 2 and a short description of the experimental setup in Section 3, some results are given in Section 4. An additional Section 5 is devoted to the above-mentioned ratio of total wind stress to wave field supporting stress.

2. The Viscous Boundary Layer of the Ocean

Within the postulated conductive layer, the mean total heat flux Q should be proportional to the temperature gradient $\mathrm{d}T/\mathrm{d}z$.

$$Q = k\frac{\mathrm{d}T}{\mathrm{d}z} \tag{1}$$

where k is thermal conductivity and z is vertical coordinate with $z=0$ at the surface. Due to the strongly varying stresses and waves, the gradient $\mathrm{d}T/\mathrm{d}z$ will also vary strongly below the interface. Therefore the gradient used represents a time mean. The measurement should be made within the conductive layer to avoid distortion of the simple law in Equation (1) by the net long-wave radiation originating in the uppermost micrometres of the ocean. Since $\mathrm{d}T/\mathrm{d}z$ can only be determined by a very sophisticated experimental setup (McAlister and McLeish, 1969), another simpler technological approach was used during our own measurements.

We can change Equation (1) to

$$Q = k\frac{\Delta T}{\delta_c}, \tag{2}$$

now introducing the total temperature difference ΔT between surface and subsurface waters and the thickness δ_c of the conductive layer. If we were able to determine at least the relative magnitude of δ_c from the basic meteorological parameters, we could determine the total heat flux from Equation (2).

Dimensional arguments led Saunders (1967) to postulate a dependence of the thickness δ_v of the viscous boundary layer on the so-called bulk parameters of the form

$$\delta_v \sim \nu \bigg/ \sqrt{\frac{\tau}{\rho_w}} \tag{3}$$

where ν is kinematic viscosity, τ is wind stress, and ρ_w is density of sea water.

This relation should hold for wind speeds from 2–10 m s^{-1}. Using the so-called bulk parameterisation of the wind stress,

$$\tau = \rho_L \cdot c_D \cdot \bar{u}_{10}^2 \tag{4}$$

and assuming a fixed value δ_c/δ_v, the total heat flux Q could be determined, if the total wind stress were responsible for the thickness of the viscous layer. ρ_L in Equation (4) is the air density, c_D the drag coefficient and $\bar{u}_{10}$ the mean wind speed at 10 m above the surface. The combination of Equations (2) and (3) after a rearrangement leads to

$$\Delta T = \lambda \frac{Q \cdot \nu}{k\sqrt{\tau/\rho_W}} \tag{5}$$

with a factor λ containing deviations of the thickness δ_c of the viscous layer from the thickness δ_v of the conductive layer and including possible differences between the total wind stress τ and that part τ_w necessary to maintain the wave field. From Equation (5) one could, if all quantities besides λ were known, derive a dependence of τ/τ_w on wind velocity assuming a known ratio δ_c/δ_v. There have been efforts to determine this ratio τ/τ_w (Dobson, 1971; Hasselmann *et al.*, 1973). The ratio δ_c/δ_v can only be given for a rigid boundary, where it is

$$\delta_c/\delta_v = (\nu/\kappa)^{-1/3} \tag{6}$$

where κ is the thermal diffusivity. The Prandtl number $\mathrm{Pr} = \nu/\kappa$ varies from 6–13 in liquid water at ambient temperatures and therefore the ratio δ_c/δ_v is approximately $\frac{1}{2}$, indicating that the conductive layer is thinner than the viscous layer, if the analogy to a rigid boundary can be accepted.

Equation (5) is only valid during night-time, since the top millimetre of the ocean strongly absorbs incoming solar radiation in the daytime.

The results shown in Sections 4 and 5, however, are restricted to conditions with no or only small solar radiation, to avoid further complications. The influence of slicks on the cool skin are not considered here, since measurements were taken on the open ocean far from land in a water body with only small amounts of living matter. Additionally, most natural slicks are believed to be monomolecular. Therefore their influence can be disregarded (see Barger and Garrett, 1968). Conditions with free convection dominating, i.e., with no shear in the surface layer caused by vanishing ocean currents and nearly vanishing wind speeds, did not occur because there were drift currents of $\frac{1}{2}$ m s^{-1} and because conditions were not really calm.

3. Experimental Setup and Data Sampling

The temperature difference between the ocean surface and the lower boundary of the viscous or conductive layer is, under most conditions, of the order of a tenth of a degree, thus imposing a difficult measuring problem. Trying to find a mean temperature gradient according to Equation (1) from the temperature at two different depths means that temperature measurements must be made with an accuracy of approximately a hundredth of a degree. In this paper, an experimental setup has been used, based on Equation (2). We have measured the total temperature difference between the surface and the interior of the ocean, and then have used assumptions about the depth of the layer – as stated by Equations (3)–(5) – or about the total heat flux. From a research vessel, it is not possible to measure continuously the temperature at a depth of 1 mm. So we decided to take the value at 5 cm and assume a well-mixed layer everywhere below the viscous boundary layer, an assumption which is generally accepted and which is confirmed by measurements (Chang and Wagner, 1975). The surface temperature was determined by an Infrared-Filter-Radiometer (Barnes PRT 5). This radiometer was calibrated with a well-stirred temperature controlled sea water reservoir during a 1-min interval. The temperature of the calibration reservoir, together with the temperature at 5-cm depth – measured with the same type of resistance thermometer – could be determined to an accuracy of 0.03 K throughout the entire phase III of GATE. The stability of the calibration curve before and after phase III of GATE has already been published in Grassl and Hinzpeter (1975).

At station 27 of GATE, Phase III, digitized records on punched tape were continuously obtained. One record consisted of a 1-min mean of ocean surface radiation and a 1-min mean of radiation from the calibration water reservoir. For the evaluation reported in this paper, hourly mean values had to be generated, since some of the basic meteorological data were sampled at hourly intervals or derived as hourly mean values (for example, net radiation).

4. Basic Results

Before presenting results of the magnitude of the cool skin, we should look at the consistency of the data. The calibration curves already published (Grassl and Hinzpeter, 1975) show possible errors up to 0.03 K for the differences between the resistance thermometers during the entire phase III of GATE. In Figure 2 a set of measured temperatures shows the stability of the subsurface temperature and the stronger variations in the surface temperature. All later comparisons have been made with hourly mean values since most other meteorological parameters were available in this form only. The mean value of ΔT was 0.2 K, if all situations with solar radiation exceeding 200 W m^{-2} are disregarded. According to Equation (5), the assessment of the factor λ is only feasible if the total heat flux Q and the total wind stress τ are known. The hourly mean values of Q and τ were determined following the bulk parameterisation in

$$Q = q_L + q_S + q_R = \bar{u}_{10} \cdot \Delta l \cdot C_1 + \bar{u}_{10} \cdot \Delta T_{S,L} \cdot C_2 + (F_\downarrow - F_\uparrow) \tag{7}$$

where Δl = water-vapour pressure difference between surface and 10 m and $\Delta T_{S,L}$ = temperature difference between the ocean surface and air at 10 m.

The coefficients of proportionality c_1, c_2, c_D were taken from simultaneous measurements of the turbulent heat fluxes q_L and q_S by Müller-Glewe and Hinzpeter (1975) using the eddy-correlation technique. The values chosen are $c_1 = 1.45 \times 10^{-3}$, $c_2 = 1.3 \times 10^{-3}$, $c_D = 1.3 \times 10^{-3}$. The net long-wave radiation q_R was determined from our measurements of the sky emission $F_\downarrow$ and the emission of the ocean $F_\uparrow$. The latter value was calculated from the measured ocean surface temperature; and the incomplete blackness of the ocean surface was accounted for by an emissivity $\varepsilon = 0.96$. The dependence of λ on $\bar{u}_{10}$ is shown in Figure 3 as a line with crosses ($\times$). The line with closed circles (●) shows the results for a different approach to the bulk parameterisation of the wind stress τ. The parameter c_D no longer was held constant for the entire wind speed interval from 1–11 m s^{-1}. A dependence on mean wind speed $\bar{u}_{10}$ of the form $10^3 \times c_D = 0.63 + 0.066 \times \bar{u}_{10}$, as proposed by Smith and Banke (1975), was used. Both parameterisations in Figure 3 lead to a strong decrease in λ towards low wind

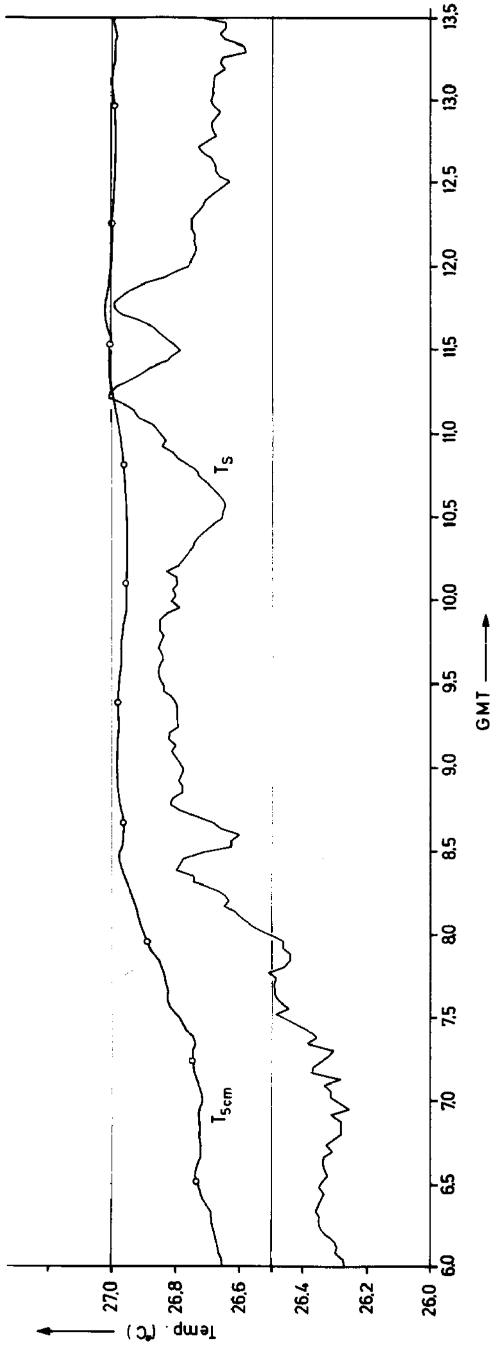

Fig. 2. 10-min-running mean of the ocean surface temperature T_s and temperature T_{5cm} at 5-cm depth on September 14, 1974, at Position 27 of GATE, Phase III, as measured from the German research vessel *Planet*.

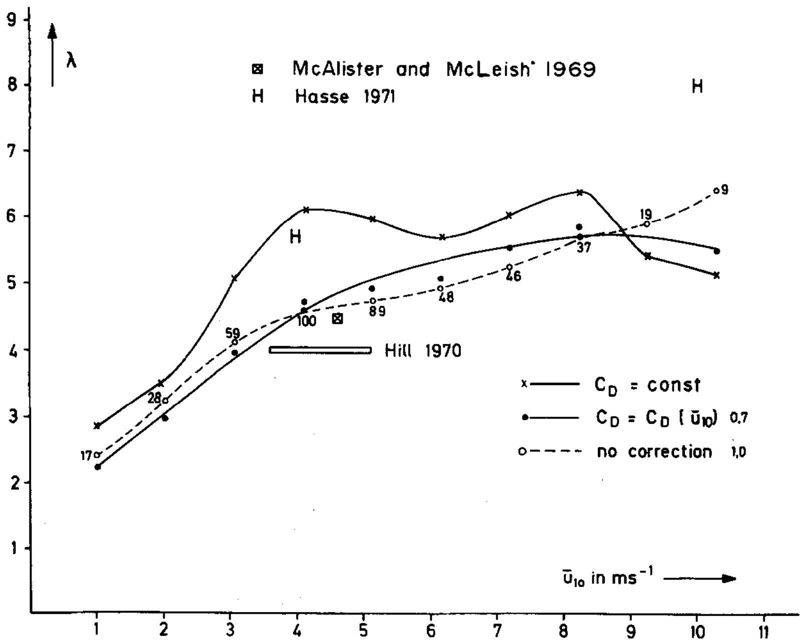

Fig. 3. The dependence of the factor λ on mean wind speed $\bar{u}_{10}$ at 10 m above the ground. Values presented in or derived from different publications are included. The line with crosses is valid for a bulk parameterisation of the wind stress with a constant $c_D \times 10^3 \times c_D = 0.63 + 0.066$ for the line with circles. The dashed line ignores the correction needed because of the mainly leeward field of view of the radiometer. Numbers on the dashed line give the numbers of hourly mean values.

speed and are in contradiction to the findings of Hasse (1971) and the model of Saunders (1967). Since the radiometric measurements of the surface temperature were taken 6–8 m off the port side of the ship, which was mainly drifting, the radiometer was frequently looking to the leeward with a decreased wind speed in the radiometer's field of view. This reduction in wind speed was accounted for by multiplying all measurements of wind speed during drifting periods by a factor 0.7. If this reduction is omitted, the dashed line in Figure 3 would result. The differences are astonishingly small.

One could attribute the strong decrease of λ with low wind speeds to a continuous change to free convection, which is governed by quite different laws. However, this free convective regime is only applicable if the drift current vanishes. This never occurred during Phase III of Gate at position 27. Therefore at least the values for 2, 3 and 4 m s^{-1} wind speeds, all showing a decreasing λ with decreasing $\bar{u}_{10}$, are free of a convective regime in the boundary layer.

We recall that λ depended on the component of wind stress used for wave generation and on the rather uncertain ratio δ_c/δ_v of the depth of the conductive viscous boundary layer. The strong variation of λ shows that quite different parts of τ are used for wave generation at different wind speeds. In Section 5 we will try to find an explanation, if we are courageous enough to introduce a numerical value for δ_c/δ_v.

The comparison of the λ-values found with those sampled by Paulson and Parker (1972) gives a new impression of the variations inexplicable up to now. The agreement with $\lambda = 4.5$ at 4.5 m s^{-1} wind speed deduced from McAlister and McLeish (1969) by Paulson and Parker (1972), is excellent. The value $\lambda = 8$ deduced from Hasse's compilation by Paulson should be corrected, if compared to the line with closed circles in Figure 3, because Hasse has used $c_D = 1.21$, which would be obtained at 10 m s^{-1} if a c_D as given by Smith and Banke (1975) had been chosen. At 4 m s^{-1}, the new λ would be 5.7 as shown in Figure 3. Now Hasse's values are still somewhat higher but the shape of the curve agrees very well. The comparison with the value $\lambda = 4$ given by Hill (1970) is somewhat arbitrary, since we did not know exactly the wind speeds in his measurements. No comparison to data given from the laboratory at very small fetch without generating waves was made, since from Paulson's discussion it is obvious that the absence of waves strongly influences ΔT and waves were always present at all measurements reported by Hasse and in this paper.

If we combine Equations (5) and (3) and solve for the depth of the viscous layer δ_v

$$\delta_v = \frac{\Delta T \cdot k}{\lambda \cdot Q} \tag{8}$$

we can now – with the knowledge of the λ-values – give a table of depths of the viscous boundary for various wind speeds. These values together with the mean values of Q, ΔT and λ for wind-speed intervals of 2 m s^{-1} appear in Table I. The temperature variation of δ_v is mainly determined by the temperature variation of the kinematic viscosity ν appearing in Equation (3). The depth δ_v decreases with increasing temperature at a rate of about 3 to 4% per degree.

TABLE I

Depth of the viscous boundary layer δ_v, total heat flux Q, temperature drop ΔT and λ as a function of mean wind speed $\bar{u}_{10}$

$\bar{u}_{10}$ (m s^{-1})	Q (in W m^{-2})	λ	ΔT (in K)	δ_v (in mm)
1	76.3	2.23	0.174	0.62
2	93.0	2.97	0.187	0.41
3	110.0	3.96	0.213	0.28
4	125.6	4.75	0.213	0.21
5	134.1	4.92	0.196	0.18
6	151.7	5.06	0.180	0.14
7	174.7	5.53	0.189	0.12
8	187.8	5.84	0.192	0.10
9	207.5	5.42	0.167	0.09
10	234.7	5.48	0.170	0.08

5. The Ratio of Total Wind Stress to the Wave Field Maintaining Wind Stress

Several authors (Stewart, 1961; Dobson, 1971; Hasselmann *et al.*, 1973) have argued that only a part of the total wind stress at the surface of the ocean is necessary to maintain the wave field. The factor λ shown in Figure 3 was introduced in Equation (5) to account for the unknown thickness δ_c of the conductive layer and the possible partial contribution of the wind stress to wave generation. If we can find a reliable value as mentioned in Section 4, the remaining variation of λ with $\bar{u}_{10}$ can be interpreted as being due to variations of τ/τ_w. This ratio is displayed in Figure 4 as a function of $\bar{u}_{10}$ using a value $\delta_v/\delta_c = 1.82 = \mathrm{Pr}^{1/3}$, if we accept this relation known to be valid for a rigid boundary. For comparison, the highest and lowest values for the ratio τ/τ_w as given by Hasselmann *et al.* (1973) and Dobson (1971) for quite different fetches are included. Unfortunately we are not able to give a conversion of wind speed to fetch to compare directly to results from Dobson, Hasselmann *et al.* and Stewart.

The following estimates of the smallest and highest values of the dimensionless fetch $\tilde{x} = gx/\bar{u}_{10}^2$ for Phase III of GATE are given for our data and for those published by Hasselmann *et al.* (1973). For the smallest observed wind system of an isolated cumulus cloud at a wind speed of 3 m s^{-1}, $\tilde{x} \simeq 500$; and for a well-developed wind field, $\tilde{x} \simeq 10^5$. If we compare these numbers with the corresponding τ_w as given by

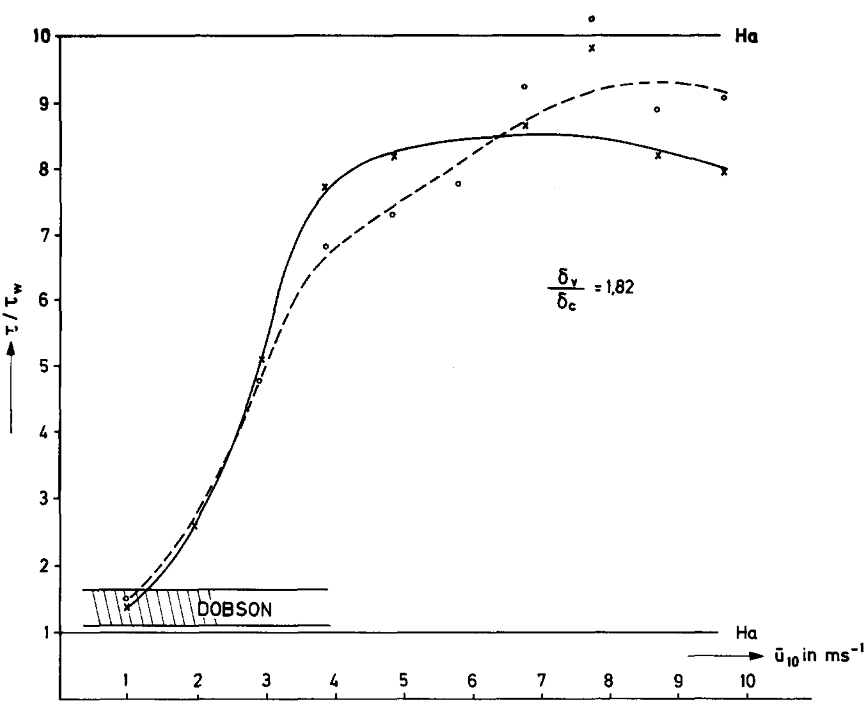

Fig. 4. Total wind stress to wave supporting stress τ/τ_w dependence on the mean wind speed $\bar{u}_{10}$ at 10 m above the ground using $c_D = (0.63 + 0.066\ u_{10})\ 10^{-3}$. The dashed line was generated by omission of all data with $\Delta T \geq 0.4$°C. Some results from the literature are included.

Hasselmann *et al.* (1973), our results are really bracketed by their minimum and maximum values. The strong slope of the factor τ/τ_w suggests nearly complete usage of the total stress τ for wave generation at low wind speeds; at low dimensionless fetch $\tilde{x}$, because of the small geometrical fetch x, there is only a 10% usage of τ at $\tilde{x} \simeq 10^5$, in good agreement with Hasselmann *et al.* (1973).

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