

CHAPTER 184

EVALUATION OF AIR-SEA INTERFACE HEAT FLUX

ROBERT L. STREET¹, M.ASCE
A. WOODRUFF MILLER, JR.²

ABSTRACT

A theoretical analysis, laboratory experiments, and the routine availability of infrared remote sensors for boat, airplane or satellite use are combined to provide a simple method for evaluating the total heat flux through the air-water interface under nighttime field conditions in which the water surface is rough; all indications are that the technique can be extended to other conditions as well. To evaluate this flux all that is required is the local windspeed, the water surface temperature, a subsurface temperature, and the character of the sea surface. Conversely, the theory can also be applied to predict temperature differences across the interface if the heat transfer can be otherwise established.

INTRODUCTION

Rejection to the atmosphere is the ultimate fate of the heat released by thermal power plants. The probable development of off-shore nuclear plants and the existing coastal power plants make necessary an understanding of and a means of evaluating the effects of ocean waves and the wind on the total heat transfer and/or temperature gradients across the air-sea interface. This transfer is important to local ocean thermal conditions and as an input to the development of local climatology.

The key to estimating interface transfer is knowledge of the temperature structure in the aqueous boundary layer just beneath the air-

¹Professor, Department of Civil Engineering, Stanford University, Stanford, California 94305, U.S.A.

²Assistant Professor, Department of Civil Engineering, Brigham Young University, Provo, Utah 84602, U.S.A.

water interface. Saunders (1967) said, "For some decades it has been known that the surface is generally cooler than the subsurface water with the major temperature variation concentrated in the uppermost millimeters." He also noted, "When used from a near surface position, radiometers afford a unique tool for determining the magnitude of this surface anomaly, and ... it seems likely that they will become indispensable in field experiments where the transfer of heat, moisture and momentum are measured." Following this Hasse (1971) considered the heat flow through the surface. Disregarding initially the short wave radiation due to the sun, he developed a theoretical analysis, based on the diffusivity relationship of Reichardt, for the vertical transfer of heat in the aqueous boundary layer that is required to balance the heat loss (or gain) at the surface through evaporation, sensible heat transfer and effective back radiation. He then provided a theoretical relationship including the influence of the sun.

Using the insight provided by Saunders and Hasse, we have extended the latter's theoretical analysis, used our own laboratory experiments, and presumed the routine availability of infrared remote sensors for boat, airplane or satellite use to provide a simple method for evaluating the total heat flux through the air-water interface under *nighttime* field conditions in which the water surface is rough; all indications are that the technique can be extended to other conditions as well. To evaluate this flux all that is required is the local windspeed, the water surface temperature, a subsurface temperature, and the character of the sea surface. Conversely, the theory can also be applied to predict temperature differences across the interface if the heat transfer can be otherwise established.

The work of Scarpace and Green (1973) at Wisconsin gives a good illustration of the applicability of such a theory. Airborne thermal imagery (infrared radiometry) was used on thermal plumes. The imagery was supported by simultaneous data taken from boats. They reported:

The scanner is virtually impossible to calibrate from the air. It also gives no information on subsurface temperatures. Thus an intensive ground truth effort is essential to providing both accurate data and supporting information.

While the plane was overhead (sometimes shortly before and after), two or three surface craft measured underwater temperatures with Whitney resistance thermometers, while [a 40 ft R/V boat] made several passes over a predetermined course marked by 20 buoys. The PRT-5 [radiometer] was mounted on the bow.

Application of the method discussed below would either yield the heat transfer from the Scarpace and Green data or, combined with meteorological and wave data, obviate the need for either the subsurface or the PRT-5 data as ground truth.

BASIC THEORY

As mentioned above both Saunders and Hasse developed theories for the heat transfer and temperature difference relationship in the aqueous boundary layer. Saunders' theory was applicable across the so-called thermal sublayer where molecular effects are dominant, while Hasse's theory applies across a specific depth of the layer (as does ours). Interestingly the attempts to verify Saunders' theory have been made through experiments more relevant to Hasse's and our theories, leading to a diversity of results for Saunders' transfer coefficient λ .

Hasse (1971) suggests treating the flux Q_T in the aqueous surface layer in differential form as (Fig. 1)

$$Q_T = -\rho_w c_{p_w} K \frac{\partial T}{\partial z} \quad (1)$$

where $K = K(z)$ is the *effective* thermal diffusivity, T is the water temperature, ρ_w is the density of water, c_{p_w} is the heat capacity of water, and z is measured positive downward from the surface. If $\partial T/\partial x$ is essentially zero, and we speak of a steady mean flow so $\partial T/\partial t \equiv 0$, then

$$Q_T = -\rho_w c_{p_w} K \frac{dT}{dz} \quad (2)$$

In the surface layer Q_T is constant below the level of back radiation and in the absence of solar (incoming) radiation. We ignore the temperature effects on ρ_w and c_{p_w} across the thermal layers. Then,

$$\frac{dT}{dz} = -\left(\frac{Q_T}{\rho_w c_{p_w}}\right) \frac{1}{K(z)} \quad (3)$$

The key parameters of the problem are summarized as follows (see Fig. 1):

- h = mean surface roughness (wave) height
- k' = Kármán constant = 0.40
- Pr = molecular Prandtl number = ν_w/κ_w
- $U_{*w} = (\tau_o/\rho_w)^{\frac{1}{2}}$ = friction velocity in water
- z_b = depth of "bulk" water temperature measurement
- z_s = radiometer optical depth ($z_s \approx 140 \mu\text{m}$ for our experiments, while $z_s \approx 20 \mu\text{m}$ for a PRT-5)

- δ_t = thermal sublayer thickness
 δ_T = thermal layer matching thickness (i.e., the point at which the temperature profile becomes logarithmic)
 δ_v = viscous sublayer thickness
 κ_w = thermal diffusivity of water
 ν_w = kinematic viscosity of water
 τ_o = surface shear stress

We also define a set of nondimensional (Reynolds) numbers:

$$\begin{aligned}
 h^+ &= U_{*w} h / \nu_w; & z_b^+ &= U_{*w} z_b / \nu_w; & z_s^+ &= U_{*w} z_s / \nu_w; \\
 \delta_T^+ &= U_{*w} \delta_T / \nu_w; & \delta_t^+ &= U_{*w} \delta_t / \nu_w; & \delta_v^+ &= U_{*w} \delta_v / \nu_w
 \end{aligned}
 \tag{4}$$

Yaglom and Kader (1974) discussed the flow over and heat transfer from rough surfaces. They defined the functional form of the eddy viscosities and eddy diffusivities in the surface layers; we employ their formulations here. We also let $\Delta T_t = T_s - T_b$ where T_s and T_b are the "surface" and bulk water temperatures and the heat transfer coefficient becomes

$$\frac{1}{C_H} = \frac{\rho_w c_p U_{*w} \Delta T_t}{Q_T}
 \tag{5}$$

The immediate objective of our work was to obtain an explicit relationship between $1/C_H$ and the parameters of the flow. Our goals then were *first* to use our results from a laboratory experiment employing an infrared radiometer to relate δ_v^+ to h^+ for rough flows and *second* to employ this relation to predict $1/C_H$ for any rough flow. The experiments were run in the Stanford Wind, Water-Wave Research facility (Fig. 2).

The key to the analysis is prescription of $K(z)$. We have developed expressions based on Yaglom and Kader's analysis for fully rough wall flows, i.e., for $h^+ > 100$. Thus,

$$\epsilon_H = a_H' \nu_w (h^+)^{-3/2} (z^+)^3
 \tag{6}$$

$$\epsilon_M = a_M' \nu_w (h^+)^{-3/2} (z^+)^3
 \tag{7}$$

near the rough wall, where ϵ_H and ϵ_M are the eddy diffusivities for heat and momentum, respectively, and a_H' and a_M' are supposed constant for any given boundary geometry.

For the layers shown in Fig. 1 we must have

$$\text{a. at } z^+ = \delta_t^+ : \epsilon_H = a_H^+ \nu_w (h^+)^{-3/2} \quad (\delta_t^+)^3 = \kappa \quad (8)$$

$$\text{b. at } z^+ = \delta_v^+ : \epsilon_M = a_M^+ \nu_w (h^+)^{-3/2} \quad (\delta_v^+)^3 = \nu \quad (9)$$

$$\text{c. at } z^+ = \delta_T^+ : \epsilon_H = a_H^+ \nu_w (h^+)^{-3/2} \quad (\delta_T^+)^3 = k^+ \nu \delta_T^+ \quad (10)$$

Thus we require $\epsilon_H = \kappa$ at the edge of the thermal sublayer, $\epsilon_M = \nu$ at the edge of the viscous sublayer, and $\epsilon_H = k^+ U_{*w} z$ as a matching condition so the eddy diffusivity given by the cubic estimate matches that of the logarithmic layer (cf., Sec. 3.2, Kader and Yaglom, 1972). Using Eqs. (8-10) and other arguments (see, Street and Miller, 1976), we were able to conclude, first, that the classic relation

$$\delta_t^+ / \delta_v^+ = Pr^{-1/3} \quad (11)$$

is still valid, and, second, that

$$\delta_v^+ = k^{-1/3} \delta_T^{+2/3} \quad (12)$$

or

$$\delta_T^+ = 0.63 (\delta_v^+)^{1.5} \quad (12)$$

It follows that, if $K(z) = \kappa + \epsilon_H$, then

$$\text{a. in } z_s \leq z \leq \delta_T^+ : K(z) = \kappa [1 + (z^+ / \delta_T^+)^3] \quad (13a)$$

$$\text{b. in } \delta_T^+ \leq z \leq z_b : K(z) = \kappa (1 + 0.4 Pr z^+) \quad (13b)$$

Therefore, our basic equation (Eq. 3) can be integrated to give

$$\int_{T_s}^{T_b} dT = \left(\frac{-Q_T}{\rho_w c_p} \right) \int_{z_s}^{z_b} \frac{dz}{K(z)} \quad (14)$$

which leads to

$$\frac{1}{C_H} = \frac{\text{Pr}\delta_t^+}{3} \left[\frac{1}{2} \ln \left\{ \frac{(\delta_t^{+2} - \delta_t^+ z_s^+ + z_s^{+2})(\delta_t^+ + \delta_T^+)^2}{(\delta_t^{+2} - \delta_t^+ \delta_T^+ + \delta_T^{+2})(\delta_t^+ + z_s^+)^2} \right\} + (3)^{\frac{1}{2}} \left\{ \tan^{-1} \left(\frac{2\delta_T^+ - \delta_t^+}{(3)^{\frac{1}{2}}\delta_t^+} \right) - \tan^{-1} \left(\frac{2z_s^+ - \delta_t^+}{(3)^{\frac{1}{2}}\delta_t^+} \right) \right\} \right] \quad (15)$$

$$+ 2.5 \ln \left\{ \frac{1 + 0.4 \text{Pr}z_b^+}{1 + 0.4 \text{Pr}\delta_T^+} \right\}$$

EVALUATION OF RELATION BETWEEN δ_v^+ AND h^+

The data of Miller, et al. (1975) were used, together with Eq. (15), by Street and Miller (1976) to obtain an estimate of the relation between δ_v^+ and h^+ . Indeed Yaglom and Kader (1974) hypothesize (in effect) that

$$\delta_v^+ = a^+ (h^+)^{\frac{1}{2}} \quad (16)$$

Thus, we sought the constant a^+ for rough flows ($h^+ > 100$).

To establish a^+ we used 23 cases of wind-generated waves in the Stanford laboratory facility (Fig. 2) with water to air temperature differences of about 2.5, 5.0 and 7.5°C at fetches of 9.5 and 14.5 m (Miller, et al., 1975). The water surface temperatures were measured to $\pm 0.01^\circ\text{C}$ with an infrared radiometer employing an indium antimonide detector with a wave length range of 1.5 to 5.5 μm and an optical depth z_s of about 140 μm . (Extracting the surface temperature involved special calibration and computational techniques and consideration of the wave-length dependent, water and air, optical properties within the detector band width; details can be found in Miller, et al., 1975.) The thermal layer temperature difference ΔT_t was defined as the bulk water temperature T_b measured 100 mm below the mean interface minus the average temperature at $z = z_s = 140 \mu\text{m}$. The parameters used in calculating surface temperature and wave height were monitored continuously. Other parameters, such as mean free-stream air velocity, temperature and humidity, were obtained also. They formed the basis for calculating the total heat transfer and surface roughness Reynolds numbers h^+ through use of data collected by previous investigators in the Stanford channel.

From our experiments we derived the key result that $a^+ = 0.37$ so

$$\delta_v^+ = 0.37(h^+)^{\frac{1}{2}} \quad (17)$$

Thus, Yaglom and Kader's conjecture for rough solid surfaces applies as well to mobile surfaces and

$$\delta_v^+ = 0[(h^+)^{\frac{1}{2}}] \quad (18)$$

We also have from Eq. (11)

$$\delta_t^+ = 0.37(\text{Pr}^{-1/3})(h^+)^{1/2} \quad (19)$$

FIELD USE

The statistical distribution of wave heights in both laboratory and field is known to be a Rayleigh distribution. In addition, the similarity theory of Yaglom and Kader (1974) which underlies Eqs. (17) and (19) should not be scale dependent, a^+ being supposed to vary only with surface shape. Accordingly, we believe that Eqs. (17) and (19) should be valid in the field as well as in the laboratory, except perhaps in the presence of long-wave swell which should be subtracted before computing h^+ for the remaining sea (defined here as freshly generated waves).

Given that δ_v^+ and δ_t^+ are known via Eqs. (17) and (19) as functions of h^+ and Pr , we may use Eq. 15 with a minimum of measurements to obtain C_H . When C_H is known we can predict Q_T by measuring U_{*w} , the sea surface physical characteristics, the surface temperature (with an infrared radiometer to obtain T_s and, hence, ρ_w and c_{pw}) and the bulk water temperature T_b (in any convenient manner).

One possible approach to obtain U_{*w} and the necessary sea surface characteristics is as follows. Hsu (1974) shows that the dynamic roughness z_o of a water surface can be expressed as

$$z_o = \frac{1}{2} \frac{H}{(C/U_{*a})^2} \quad (20)$$

where H is the dominant wave height, C is its phase velocity and U_{*a} is the friction velocity in the atmospheric constant-flux layer. Using the results of Kondo, et al. (1973), Kondo (1975) determines that the mean surface roughness height h is given by

$$h = 30 z_o \quad (21)$$

for completely rough flow above the sea (established to occur when $U_{10} \geq 8 \text{ m}\cdot\text{s}^{-1}$ where U_{10} is the windspeed at 10 metres elevation). Of course, h can be measured directly also. Kondo (1975) also shows that the drag coefficient

$$C_D = (U_{*a}/U_{10})^2 = 10^{-3}(1.2 + 0.025 U_{10}) \quad (22)$$

for $U_{10} \geq 8 \text{ m}\cdot\text{s}^{-1}$. Applying the assumption of stress continuity across the interface yields

$$U_{*w} = (\rho_a/\rho_w)^{\frac{1}{2}} U_{*a} \quad (23)$$

where ρ_a is the air density. Given U_{10} and the temperatures in air and water, U_{*w} is found by the combination of Eqs. (23) and (22).

Hasse (1971), Saunders (1967), and others have followed the concept of stress continuity across the interface, according to Eq. (23). Wu₂ (1975) points out clearly that the fraction of the momentum flux $\rho_a U_{*a}^2$ from the air which goes to drift currents ranges from about 0.6 to 0.8 in the rough flow regime, the remaining flux going to wave generation. In a rough boundary case where eddying is driven by the drift current boundary layer flow and the random water wave motion together it is not entirely clear whether one should use U_{*w} or some fraction thereof as the reference velocity (the fraction ranging apparently from $\sqrt{0.6} \approx 0.8$ to $\sqrt{0.8} \approx 0.9$). We feel that, pending further evidence, the use of U_{*w} is best.

Thus, Eqs. (15), (17), (19), (21-23), and (12) are the needed set for prediction of the heat flux and temperature difference relationships for a given wind and sea state. Given basic physical data one can either predict ΔT_t from known Q_T or predict Q_T from known ΔT_t .

AN APPLICATION

An example illustrates the variation of parameters. Assume that T_b is 20°C so ν_w , ρ_w , c_{pw} and $Pr = 7.1$ are known. Take $T_{air} = 15^\circ\text{C}$ so ρ_a is known and assume $U_{*a} = 0.381$ (equivalent to $U_{10} = 10 \text{ m/s}$, for example). We employ a PRT-5 radiometer so $z_s = 20 \mu\text{m}$ and measure the bulk temperature at one of 3 depths, viz., $z_b = \delta_T$, $z_b = 1 \text{ cm}$ and $z_b = 10 \text{ cm}$.

Under these conditions $h^+ = (1.3 \times 10^4)h$ so $h^+ > 100$ when $h > 3/4 \text{ cm}$. One finds, for example,

$$z_s^+ = 0.27 (\delta_v^+ \geq 3.7, \delta_t^+ \geq 1.9 \text{ and}$$

$$\delta_T^+ \geq 4.5 \text{ for } h^+ \geq 100)$$

$$z_b^+ = \delta_T^+, 133 \text{ or } 1330$$

In Fig. 3 we show δ_v^+ and δ_t^+ as functions of h . As expected $\delta_t^+ \approx \frac{1}{2} \delta_v^+$ and both are proportional to \sqrt{h} . Figure 4 shows the inverse transfer coefficient C_H^{-1} as a function of h and z_b . As δ_T represents the point of transition to the fully turbulent, logarithmic temperature profile, $z_b = \delta_T$ corresponds to neglecting any temperature variation between δ_T and the actual measurement point for bulk temperature. As Saunders' theory neglects even the variation between $z = \delta_t$ and $z = \delta_T$, it is clear that comparison of laboratory or field measurements to Saunders' theory is moot because we cannot measure at $z = \delta_t$ which lies between the surface roughness. From this figure we see that neglecting the temperature variation in the logarithmic zone leads to an error of between 10 and 50 percent for $z_b = 10$ cm.

The final figure (Fig. 5) shows ΔT_t , i.e., the surface temperature depression for a typical field heat transfer value with all the previous assumptions in effect. Clearly a 1 to 2°C depression over 10 cm is worth considering.

The next step in our efforts will be to carry out the straightforward extension of our theory to daytime (significant solar radiation) conditions and to transition rough and smooth flow conditions.

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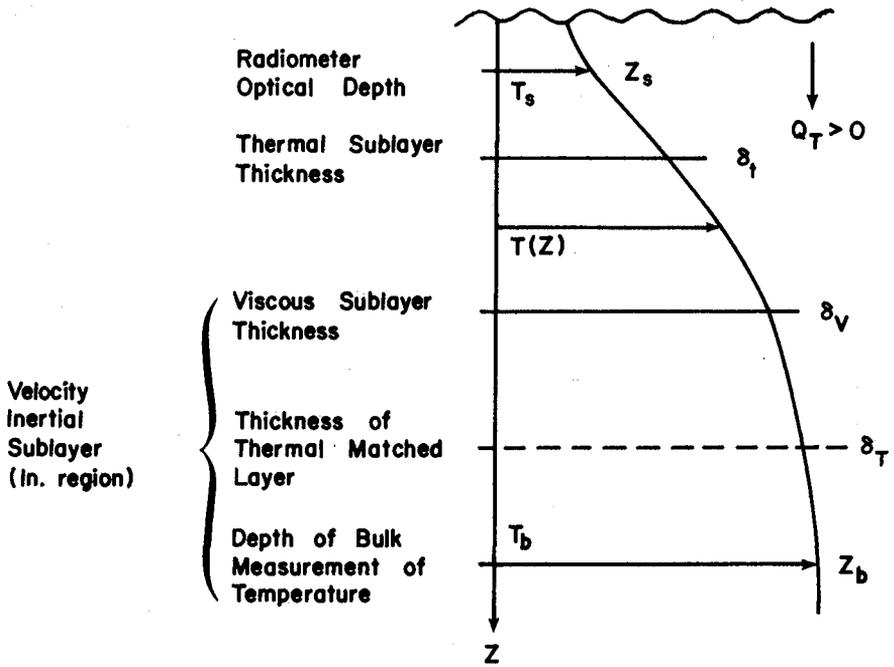


FIG. 1. SCHEMATIC OF AQUEOUS SURFACE LAYER.

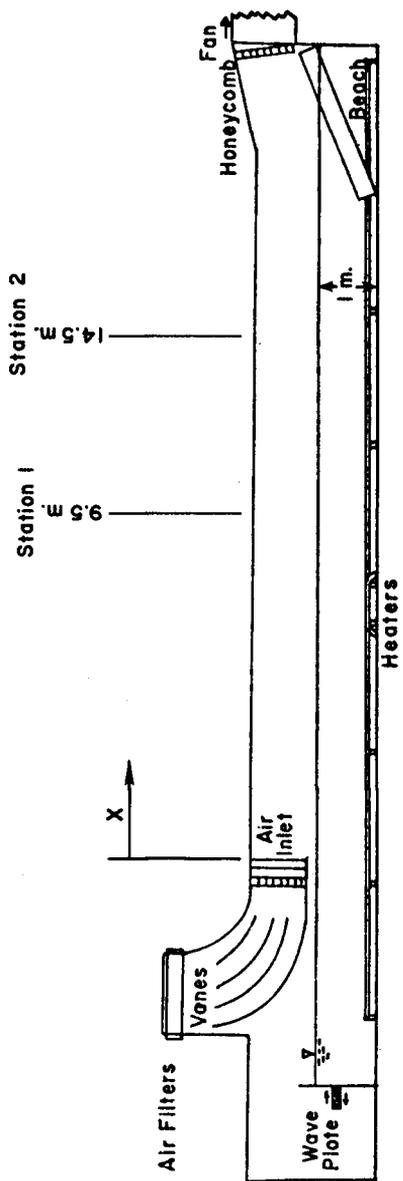


FIG. 2. SCHEMATIC OF STANFORD WIND, WATER-WAVE RESEARCH FACILITY.

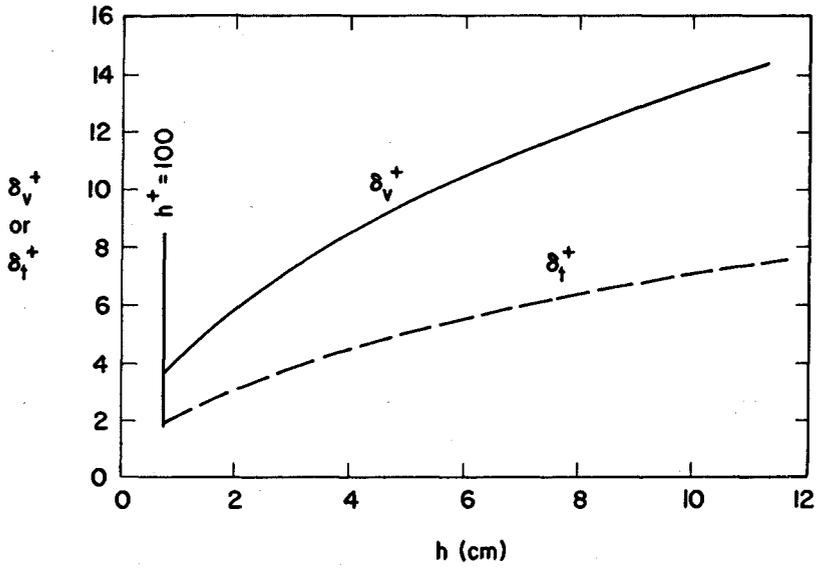


FIG. 3. VARIATION OF DIMENSIONLESS THERMAL AND VISCOUS SUBLAYER THICKNESS WITH MEAN WAVE HEIGHT FOR FULLY ROUGH BOUNDARY.

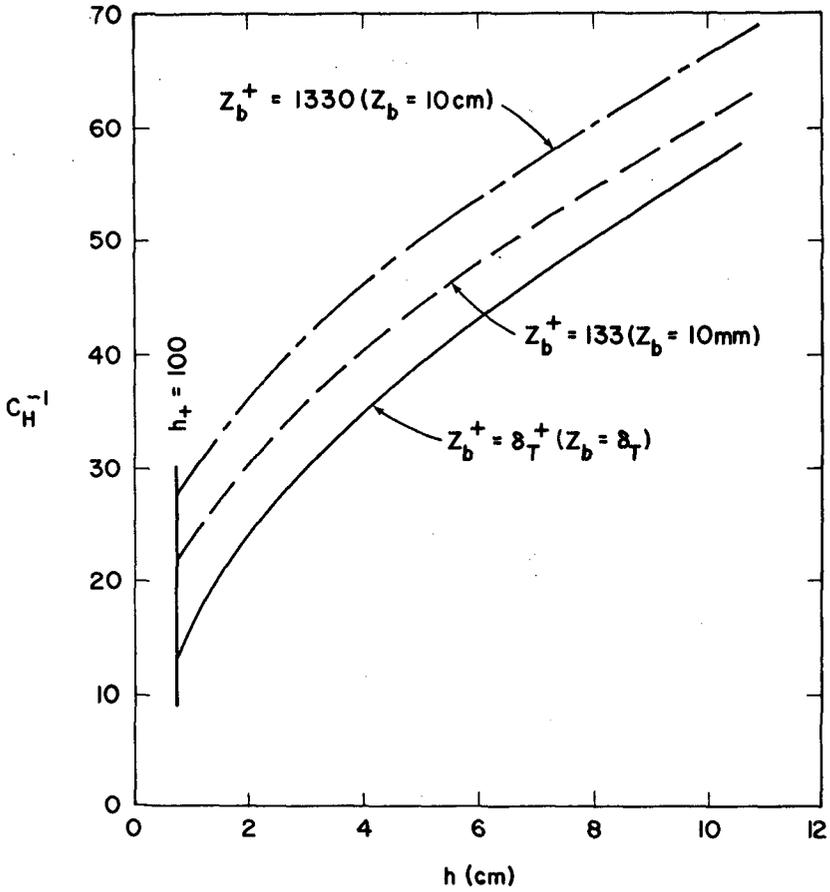


FIG. 4. VARIATION OF THE INVERSE OF THE TOTAL HEAT TRANSFER COEFFICIENT WITH MEAN WAVE HEIGHT FOR FULLY ROUGH BOUNDARY.

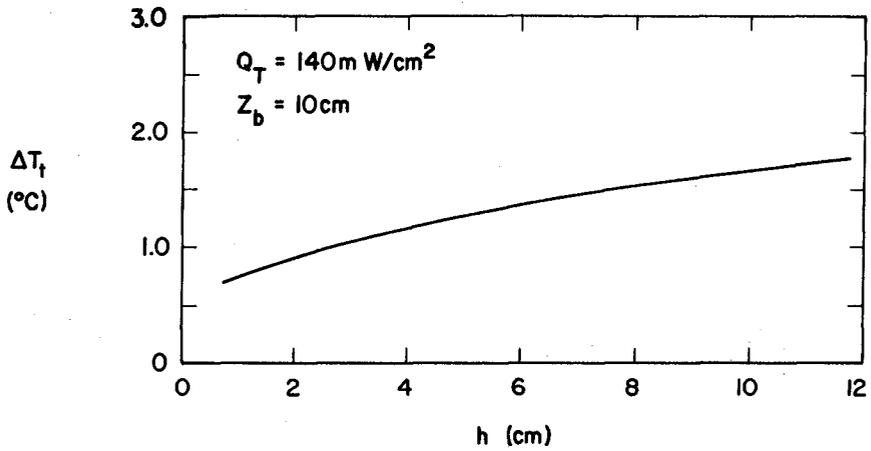


FIG. 5. VARIATION OF BULK-WATER TO SURFACE-WATER TEMPERATURE DIFFERENCE WITH MEAN WAVE HEIGHT FOR SPECIFIED Q_T AND Z_B AND A FULLY ROUGH SURFACE.