Estimating Overwater Convective Boundary Layer Height from Routine Meteorological Measurements for Diffusion Applications at Sea

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ABSTRACT

On the basis of hourly measurements of wind and air and sea surface temperatures for at least 6 yr at three buoy stations in the eastern Gulf of Mexico, the onset of the free convection regime, which coincides with the commencement of stability class C (for slightly unstable conditions in the Pasquill stability classification) at approximately $R_b = -0.03$, $-Z/L = 0.4$, and $-Z_i/L = 5$, is verified over the ocean, where R_b is the bulk Richardson number, Z (= 10 m) is the height above the sea, L is the Monin–Obukhov stability length, and Z_i is the height of the convective boundary layer (CBL). Datasets for the CBL are analyzed in the context of the boundary layer physics of Garratt. It is found that Z_i is linearly proportional to the surface buoyancy flux—that is, $(\overline{w'\theta_v'})_0$, where w is the vertical velocity and θ_v is the virtual potential temperature. For operational diffusion applications, a statistical formula is proposed—that is, $Z_i = 369 + 6004(\overline{w'\theta_v'})_0$. A method to compute this buoyancy flux from routine meteorological measurements is also provided.

1. Introduction

In certain oceanic regions the sea surface temperature during most times of the year is higher than the air temperature, making the overlying atmosphere unstable and convective. The eastern Gulf of Mexico is such a place (Fig. 1). Also shown in the figure is the stability parameter Z/L (where Z is the height above the sea surface, and L is the Monin–Obukhov length) and its classification (see Hsu 1992). The datasets were from the National Data Buoy Center (1990), and the stability computation was based on Eq. (1) (see Hsu 1992). Climatologically, this region is unstable all year round because Z/L is negative. Detailed analyses indicate that the most unstable region is located along the continental shelf break near Desoto Canyon at buoy station 42009. Note that hourly measurements of air temperature and wind at all three buoys (42003, 42007, and 42009) were at $Z = 10$ m. The data periods were from 1976 through 1988 for buoy 42003, 1981–88 for 42007, and 1980–86 for 42009.

Figure 1 shows that all values of Z/L are negative, however, if the value is small and within 0.4, the mechanical turbulence will dominate, making the vertical temperature distribution in the atmospheric boundary layer or the subcloud layer more adiabatic or in neutral stability class D condition according to Pasquill's stability classification (see Pasquill 1961), for example, at buoy stations 42003 and 42007 year round. On the other hand, when $-Z/L > 0.4$, heat convection will dominate. Thus, stability class C [for slightly unstable conditions, see Pasquill (1961)] exists in January, July, August, and December at the shelf break. Under these conditions, the CBL is prevailing. Therefore, the CBL height must be related more to the surface heat flux such that the mixing height determination for stability class D may not be applicable (see Venkatram 1978). Based on this reason, we are proposing a formula to estimate the overwater CBL height, which can be applied to offshore diffusion studies such as those being conducted in the Breton Island area in the northeastern Gulf of Mexico (Hsu 1995a). The proposed equation should be useful to other marine regions having similar CBL conditions.

2. Methods

To ensure that the unstable condition is also convective [i.e., for overwater stability classes C and B (for moderately unstable conditions)], one needs to find the onset of the free convective regime (see Wyngaard et al. 1971). This is done as follows.

From routine meteorological measurements at sea, the stability parameter Z/L is difficult to compute. A similar stability parameter, the bulk Richardson number R_b, is much easier to use. The generic relationship between Z/L and R_b is (Hsu 1989)

$$\frac{Z}{L} = \kappa C_T C_d^{-3/2} R_b, \tag{1}$$

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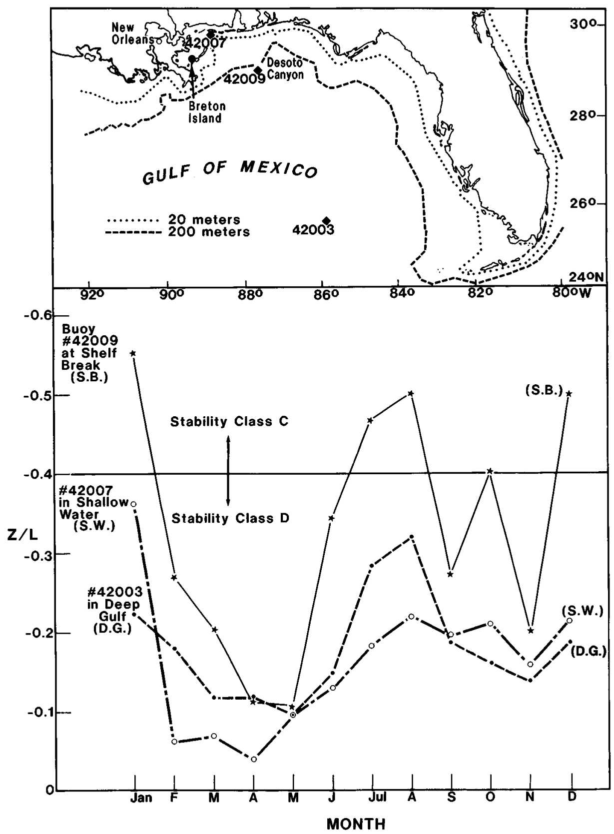

FIG. 1. The study area and its monthly variation of stability characteristics.

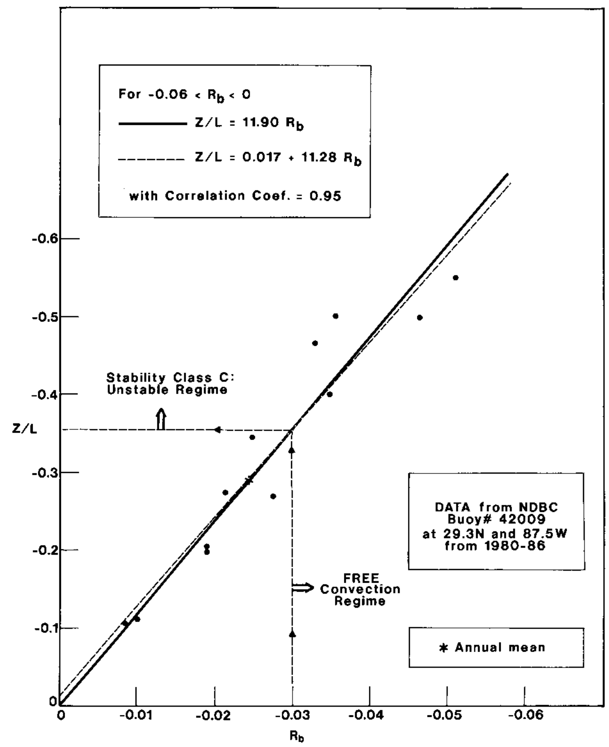

FIG. 2. A verification of the onset of the free convection regime and the commencement of stability class C over the ocean (see Fig. 1).

where $\kappa = 0.4$ is the von Kármán constant, C_T is the sensible heat flux coefficient, C_d is the wind-stress drag coefficient, and from Hsu (1992)

$$R_b = \frac{gZ(T_{\mathrm{air}} - T_{\mathrm{sea}})}{U_Z^2(T_{\mathrm{sea}} + 273)}, \tag{2}$$

where $g = 9.8$ m s^{-2} is the gravitational acceleration; T_{air} and T_{sea} are the air and sea surface temperatures, respectively; and U_Z (m s^{-1}) is the wind speed at height $Z = 10$ m above the mean sea surface.

Since buoy 42009 shown in Fig. 1 has both stability classes D and C, routine meteorological measurements of wind and air and sea temperatures were employed to compute values of R_b based on Eq. (2). For brevity, only monthly results are shown. Corresponding monthly values of Z/L are obtained from Eq. (1) where the value $C_T = 1.10 \times 10^{-3}$ for the unstable condition is adopted from Smith (1980). The proper C_d formulation for the continental shelf of the Gulf of Mexico is based on Hsu (1995b).

The data analysis is presented in Fig. 2. It is very interesting to note that stability class C is in the free-convection regime based on $R_b = -0.03$ from Priestly (1959, 47) and on $-Z/L$ between 0.3 and 0.4 from Wyngaard et al. (1971, Fig. 1). The consequence of the free convection to initiate the sea-breeze circulation over the coastal zone has been presented elsewhere (see Hsu 1973).

The growth rate of CBL height Z_i with time t, that is, $\partial Z_i/\partial t$, according to Garratt [1992, Eq. (6.18), 155 and Eq. (1.4), 10] is

$$\frac{\partial Z_i}{\partial t} = \frac{(1 + 2\beta)(\overline{w'\theta_v'})_0}{\gamma_\theta Z_i}, \tag{3}$$

where β is the ratio of entrainment and surface heat flux, $(\overline{w'\theta_v'})_0$ is related to the virtual heat flux at the surface or simply the surface buoyancy flux in which w is the vertical wind component and θ_v is the virtual potential temperature, and γ_θ is the potential temperature gradient above Z_i.

Experiments were conducted to estimate the overwater value of $\partial Z_i/\partial t$. The results shown in Fig. 3 were based on rawinsondings obtained during the summer of 1993 from the offshore platform GB 236A located at approximately 27.8°N and 93.1°W in the Gulf of Mexico (for more detail, see Hsu 1995a). It can be seen from Fig. 3 that between 0700 LT and 1300 LT the value of Z_i changed from approximately 669 to 659 m, a decrease of 10 m in 6 h, whereas between 1300 and 1900 LT it changed from 659 to 676 m, an increase of 17 m in 6 h. Therefore, we can say the $\partial Z_i/\partial t$ is small and may be treated as a slowly varying parameter. Similarly, if the external forcing due to synoptic-scale conditions does not change appreciably, we may also treat both β and

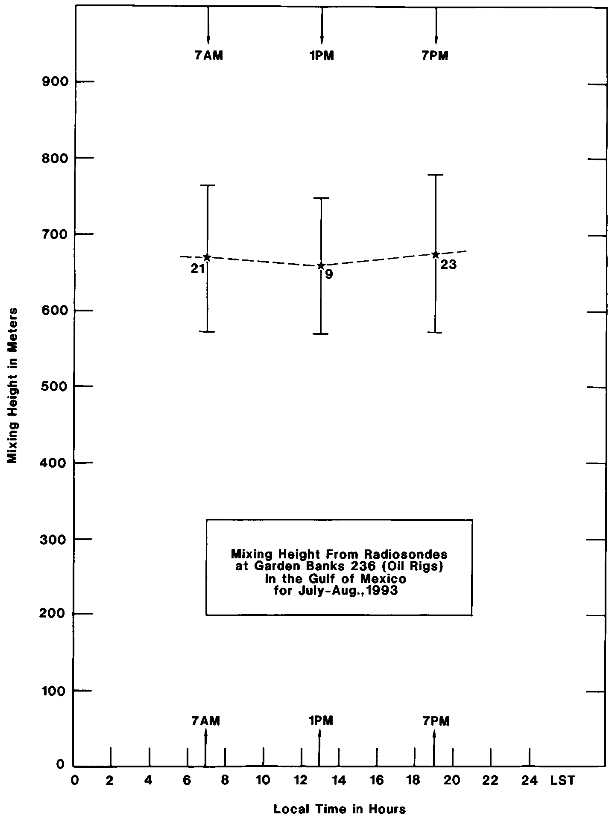

FIG. 3. Overwater variation of the mixing height between 0700 and 1900 LT over the Gulf of Mexico.

γ_θ as slowly changing variables as compared to Z_i and $(\overline{w'\theta'_v})_0$, we then have from Eq. (3)

$$Z_i = \frac{(1 + 2\beta)(\overline{w'\theta'_v})_0}{\gamma_\theta(\partial Z_i/\partial t)},$$

or from an operational point of view, we may assume that

$$Z_i = A + B(\overline{w'\theta'_v})_0, \tag{4}$$

where constants A and B need to be determined statistically from field experiments. Note that the value A is needed meteorologically because when $(\overline{w'\theta'_v})_0$ is getting smaller, Z_i should be approaching the boundary layer height for neutral conditions.

Since most measurements of Z_i over the ocean were obtained by research aircraft, values of Z/L in the surface layer were not available. On the basis of previous discussions we have established that the free convective regime begins in stability class C (see Fig. 2). On the other hand, according to Briggs (1988, 68) the dividing point between classes D and C is $-Z_i/L = 5$. For these reasons we have sythesized aircraft measurements made under CBL conditions over the ocean with $-Z_i/L \geq 5$ in which both values of Z_i and L were readily available as in Wyngaard et al. (1978) and Chou et al. (1986).

3. Results

Figure 4 shows our results. If one accepts the high correlation coefficient of 0.85 from two different geo-

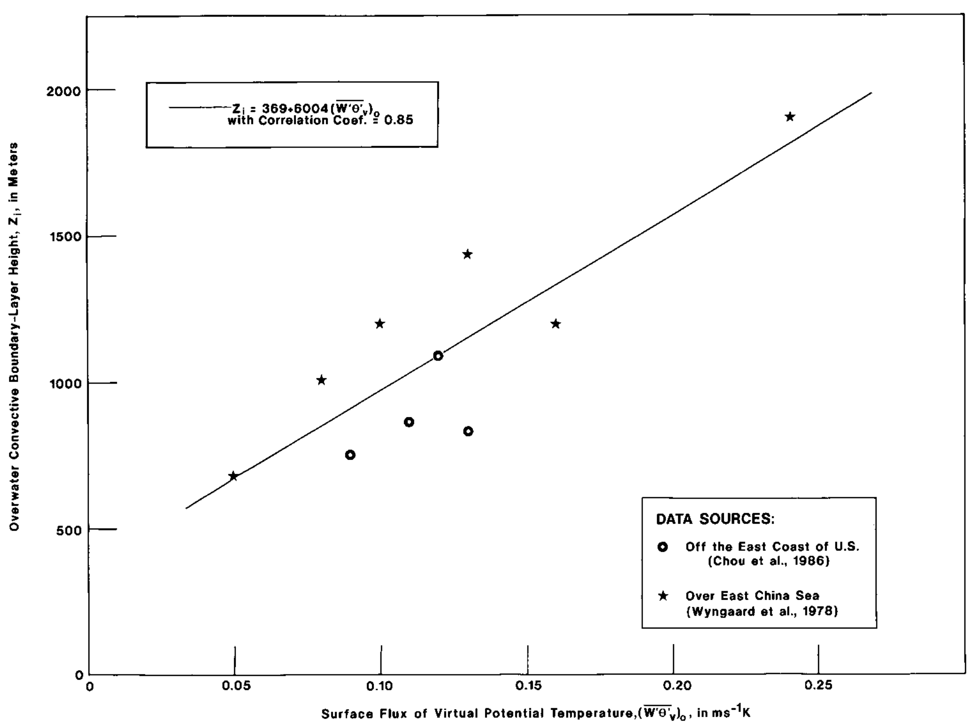

FIG. 4. A statistical analysis of Z_i vs $(\overline{w'\theta'_v})_0$ according to Eq. (5).

graphic regions, the following equation should be useful for overwater diffusion estimates.

$$Z_i = 369 + 6004(\overline{w'\theta_v'})_0, \tag{5}$$

where Z_i is in meters and $(\overline{w'\theta_v'})_0$ is in meters per second kelvin.

For operational applications, the buoyancy flux at the surface may be computed as follows: on the basis of Wyngaard et al. [1978, Eq. (14)], Panofsky and Dutton [1984, 132, Eq. (6)], and Hsu [1988, 112, Eq. (6.42)]

$$(\overline{w'\theta_v'})_0 = C_T U_{10}(T_{\mathrm{sea}} - T_{\mathrm{air}})\left(1 + \frac{0.07}{B}\right), \tag{6}$$

where C_T, T_{sea}, and T_{air} have been defined previously, U_{10} is the wind speed at 10 m, and B is the Bowen ratio from Panofsky and Dutton [1984, 132, Eq. (8)] that

$$B = \frac{T_{\mathrm{air}} - T_{\mathrm{sea}} + 0.01\Delta Z}{2500(q_{\mathrm{air}} - q_{\mathrm{sea}})}, \tag{7}$$

where ΔZ (m) is the height interval between T_{sea} and T_{air} (°C or K), and q's are the specific humidities (g g^{-1}).

Also, from Hsu (1988, 18–21),

$$q_{\mathrm{air}} = 0.62\frac{e_{\mathrm{air}}}{P} \tag{8}$$

and

$$q_{\mathrm{sea}} = 0.62\frac{e_{\mathrm{sea}}}{P}, \tag{9}$$

in which

$$e_{\mathrm{air}} = 6.1078 \times 10^{[7.5T_{\mathrm{dew}}/(237.2+T_{\mathrm{dew}})]} \tag{10}$$

and

$$e_{\mathrm{sea}} = 6.1078 \times 10^{[7.5T_{\mathrm{sea}}/(237.3+T_{\mathrm{sea}})]}, \tag{11}$$

where T_{dew} (°C) is the dewpoint temperature.

4. Conclusions

Several conclusions may be drawn from this study:

- Based on hourly measurements of wind and air and sea surface temperatures for at least 6 yr from three buoy stations in the eastern Gulf of Mexico, it is shown that unstable conditions prevail year round in this region. Furthermore, the free convective regime also exists along the continental shelf break.
- Three criteria normally used to classify the onset of the free convection on land indicated that they are also consistent with each other over the ocean. These three stability parameters are $R_b = -0.03$, $-Z/L = 0.4$, and $-Z_i/L = 5$.
- The onset of the free convection regime is found to coincide with the commencement of stability class C.
- According to the theory of Garratt, the height of the CBL, Z_i, should be linearly proportional to $(\overline{w'\theta_v'})_0$, which is the surface buoyancy flux. This is verified and for operational overwater diffusion applications for CBL, a statistical formula [Eq. (5)] is proposed.
- For operational use, a method to compute $(\overline{w'\theta_v'})_0$ from routine meteorological observations is also provided.

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