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# Effectiveness of Unstable Atmospheric Boundary Layer on Heat loss from Alexandria Eastern Harbor, Egypt

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**Abstract**

The heat storage of a water body regulates by the energy fluxes across the water surface. Sensible and latent heat fluxes are affected by atmospheric stability. So, the energy fluxes will affect the water balance. In this study two consecutive summer seasons time scales at Alexandria Eastern Harbor, Egypt was selected to examine the effect of atmospheric stability on the heat fluxes by estimating daily average of sensible and latent heat fluxes. Persistent atmospheric instability resulted in a 27.4%, 33.8% (2010, 2011), respectively increase in the mean heat loss by latent heat fluxes, relative to neutral stability conditions. Also, 291%, 220% (2010, 2011), respectively increase in the mean heat loss by sensible heat fluxes relative to neutral stability conditions. In Alexandria Eastern Harbor, the atmosphere could be unstable even with air temperatures higher than the water surface by about less than 1 °C.

***Key words:*** *Heat Flux, Latent heat, Sensible heat, Heat exchange coefficient, Alexandria Eastern Harbor.*

## 1. Introduction

Sensible and latent heat fluxes represent the turbulent exchange of heat between the ocean surface and overlying atmosphere (Larid and Kristovich, 2002).

The heat budget for the water body is controlled by interaction with the atmosphere. This interaction is affected by the stability of the atmospheric boundary layer (ABL) above the water surface (Verburg and Antenucci, 2010). When the Obukhov stability length is negative the atmosphere above the water surface is unstable and convective (Brutsaert, 1982). The difference between the temperature of air and water surface can be used as an indicator of stability (Derecki, 1981; Croley, 1989) with an unstable atmosphere typically associated with a water surface temperature that exceeds the air temperature, although this is not strictly correct as wind speed and humidity also need to be considered (Verburg and Antenucci, 2010). Heat loss from a water body by sensible and latent heat transfer is reduced when the atmosphere above it is stable and enhanced when the atmosphere is unstable (Brutsaert, 1982).

Lakes and may be harbors, as opposed to the ocean, generally extreme variability in (ABL) stratification (change from highly stable to highly unstable) due to the generally lower wind speeds and also due to the greater heating and cooling by the surrounding land (Katsaros, 1998).

The aim of this study is to examine the effect of atmospheric stability on latent and sensible heat fluxes during the available data in semi-arid region; heat fluxes across the water surface of Alexandria Eastern Harbor (AEH) were determined using aerodynamic methods. Latent and sensible heat fluxes were estimated on daily basis from wind speed, air temperature, surface water temperature, humidity, and air pressure.

## 2. Study area

Alexandria Eastern Harbor is a shallow, protected, semi-enclosed circular basin. It covers an area of about 2.8 $km^2$ and occupies the central part of the coast of Alexandria Figure (1). The Harbor is connected to the Mediterranean Sea through two openings: El-Silsila (northeast opening) and El-Boughaz (main central outlet). The seabed within the Harbor slopes gradually seawards, with an average depth of 5 m inside the Harbor and a maximum depth of 13 m at the eastern corner of El-Boughaz (El-Geziry *et al.*, 2007; Hussein and El-Geziry, 2014; Hussein and Mohamed, 2016).

Within the framework of collaboration between the Ocean Data Information Network of Africa (ODINAfrica) and the Global Sea Level Observing System (GLOSS), GLOSS-ODINAfrica tide gauge system has been installed (since November 2009) on a sea wall at NIOF head office within AEH Figure (1). The site is secure with adequate water depth of 1.5 m. The system provides near real time data access

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using satellite communications with digital data recording capability and new radar/pressure water level sensors. In addition to the pressure sensor, a temperature sensor does also exist (31.2128°N, 29.8854°E) to record the sea surface temperature.

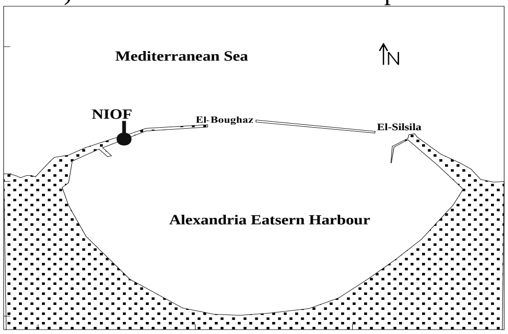



Fig. 1 Alexandria Eastern Harbour and the location of NIOF

## 3. Data Availability

The measurements of sea surface temperature ($T_o$, °C) and surface meteorological parameters: air temperature (T, °C), relative humidity ($R_H$, %), wind speed ($U_z$, m/s) and atmospheric pressure (p, mbar) were made for two consecutive summer seasons. The SST and Meteorological data, collected from 6/6/2010 to 30/9/2010 (phase-I), and from 1/7/2011 to 19/9/2011(phase-II) have been used to compute the latent and sensible heat fluxes. Meteorological data has been measured by Alexandria Nozha Air Port meteorological station.

## 4. The Transfer Method

There is a linear relationship between the fluxes of sensible and latent heat and the transfer coefficients, which vary with the local stability of (ABL), and affected by the humidity and gradients of temperature and wind velocity above the water surface. The algorithm which has been used here is similar to the types of algorithms that are typically used for the computation of ocean surface fluxes, so this manuscript follow the same procedure which has been described in details by (Verburg and Antenucci, 2010). Sensible heat (H) and latent heat (E) fluxes ($W/m^2$) are estimated with bulk aerodynamic methods,

$$H = \rho_a C_a C_h U_z (T_o - T) \quad (1)$$

$$E = \rho_a L_v C_e U_z (q_s - q_z) \quad (2)$$

$$\rho_a = 100 p / (R_a (T + 273.16)) \quad (3)$$

$$\Delta\theta = Tov - Tv \quad (4)$$

Where $C_a$ is the specific heat of air (1005 J $kg^{-1}$ $K^{-1}$), $C_h$ and $C_e$ are the transfer coefficients for sensible heat and latent heat, respectively, $L_v$ is the latent heat of vaporization (J $kg^{-1}$), $q_s$ is the specific humidity at saturation (kg/kg), $q_z$ is the specific humidity (kg/kg), $\rho_a$ is the air density ($kg/m^3$), and $R_a$ is the gas constant for moist air (J $kg^{-1}$ $K^{-1}$). Tov is the virtual temperature of saturated air at the water surface and Tv is the virtual air temperature.

## 5. Results and discussion

### 5.1 The Roughness Lengths

The average value of momentum roughness length Zo during (I) was 2.07 x $10^{-4}$ m (1.03 x $10^{-4}$ m to 3.30 x $10^{-4}$ m) and during (II) was 1.85 x $10^{-4}$ m (1.04 x $10^{-4}$ m to 3.27 x $10^{-4}$ m), both are increased with wind speed above (1.68 m/s) (Figure 2). The Zo and wind speed relationship is similar to the results represented by Renfrew et al. (2002) and Verburg and Antenucci, (2010). The average value of vapor roughness length Ze during (I) was 1.18 x $10^{-4}$ m (7.89 x $10^{-5}$ m to 1.74 x $10^{-4}$ m) and during (II) was 1.27 x $10^{-4}$ m (7.94 x $10^{-5}$ m to 1.72 x $10^{-4}$ m), both decreased with wind speed above (1.68 m/s) as shown in Figure 2. The Ze and wind speed relationship agreed with results of Katsaros (1998), Renfrew *et al*. (2002) and Verburg and Antenucci, (2010). Ze here was on average slightly higher due to the study area has high air temperatures.

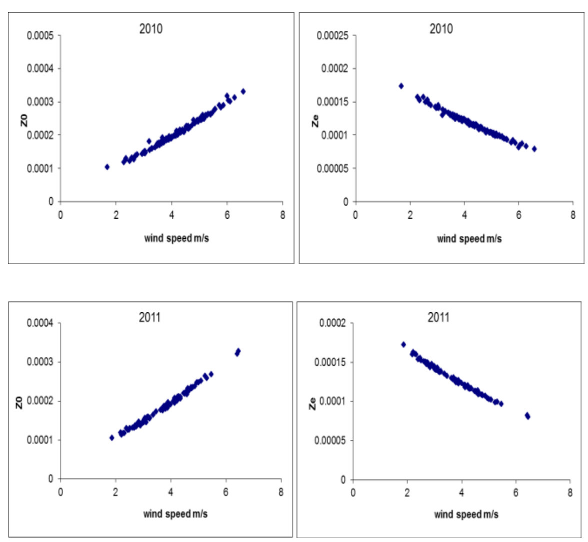



Fig. 2 The Roughness length for momentum (Zo) and vapor (Ze)

## 5.2 Atmospheric Stability Effectiveness

The average value of the neutral drag coefficient ($C_{dn}$) during phase-I was 1.44 x $10^{-3}$ ($C_{dn}$ =1.27 x $10^{-3}$ at $U_z$ =1.68 m/s and $C_{dn}$ = 1.58 x $10^{-3}$ at $U_z$ = 6.58 m/s) and during phase-II ($C_{dn}$) was 1.41 x $10^{-3}$ ($C_{dn}$ =1.28 x $10^{-3}$ at $U_z$ =1.87 m/s and $C_{dn}$ = 1.58 x $10^{-3}$ at $U_z$ = 6.46 m/s). Both are increased with wind speeds above 1.65 m/s as estimated by Verburg and Antenucci, (2010). The stability function (ψ) during both phases are increased the average value of drag coefficient ($C_d$) by 19.96 % to 1.73 x $10^{-3}$, and by 25.27 % to 1.77 x $10^{-3}$, respectively (Table 1).

$C_{en}$ (neutral transfer coefficient for heat) during (I) was on average 1.37 x $10^{-3}$ ($C_{en}$ = 1.34 x $10^{-3}$ at $U_z$ = 1.68 m/s and $C_{en}$ = 1.39 x $10^{-3}$ at $U_z$ = 6.58 m/s). It was increased between wind speeds 1.68 m/s to less than 5.50 m/s and slightly increased above 5.57 m/s to 6.58 m/s (Figure 3). $C_{en}$ during (II) was 1.36 x $10^{-3}$ ($C_{en}$ = 1.33 x $10^{-3}$ at $U_z$ = 1.87 m/s and $C_{en}$ = 1.39 x $10^{-3}$ at average $U_z$ = 6.46 m/s) and also it increased with wind speed (Figure 3).

The ABL at Eastern Harbor during (I and II) was unstable most of the time (stability parameter $\zeta < 0$ occurred 97.43 % and 100%, respectively), resulting in enhancement of transfer coefficients and fluxes of the heat. Table (1) represents the average transfer coefficient $C_e$ for the two phases which increased by stability functions with 28.5 % to 1.76 x $10^{-3}$ and 36.19 % to 1.86 x $10^{-3}$, respectively.

When (Δθ) was large and wind speeds were low the stability effect on the transfer coefficient was largest, both in the stable ABL (reduced transfer for E and H) and unstable ABL (enhanced transfer).

The effect of wind speed on the stability adjusted transfer coefficient is represented by (Figure 3), $C_e$ converging to $C_{en}$ with increasing the value of wind speed, as ζ decreased gradually to zero (Figure 4).

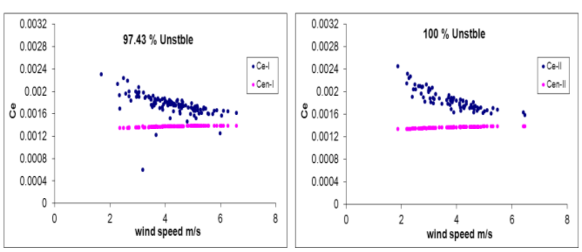



Fig. 3 The transfer coefficient $C_e$ depends on wind speed. The ABL is unstable when $C_e$ is higher than $C_{en}$ and stable when $C_e$ is lower than $C_{en}$.

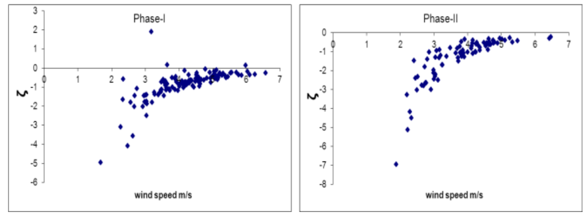
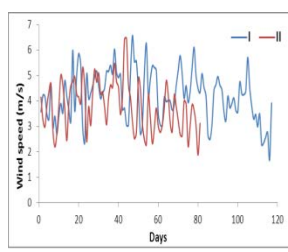



Fig. 4 Relationship between stability parameter (ζ) and wind speed

## 5.3 The Transfer Coefficients

Wind speed during (I) fluctuated between 1.68 to 6.58 m/s with an average (4.2 m/s +/- 0.98 SD) and during (II) it oscillated between 1.88 to 6.47 m/s with an average (3.8 m/s +/- 0.97 SD) (Figure 5). Relative humidity variation during (I) was between; 47% to 78% with an average (67% +/- 5 SD) and during (II) it was between; 61% to 79% with an average (70% +/- 3 SD) (Figure 6). Daily wind speed and Δθ for both phases were inversely correlated r = -0.18 and -0.48, respectively (Figure 7). Stability parameter ζ of the two phases depends on Δθ and wind speed. When wind speed was low │ζ│ was large (phase I: r = -0.76, phase II: r = -0.77) and │ζ│ was moderate when Δθ was high (phase I: r = 0.43, phase II: r = 0.75). The stability adjusted transfer coefficients closely tracked ζ of the two phases ($C_e$: r = 0.94, 0.96 and $C_d$: r = 0.68, 0.87). The atmosphere was more unstable and the mean transfer coefficient was higher in (II) mean ζ = -1.38 than in (I) mean ζ = -0.89. When wind speeds decreasing below about 3 m/s, both $C_e$ and $C_d$ increased.



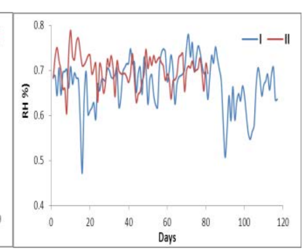




Fig. 5 The wind speed variations Fig. 6Relative humidity variations

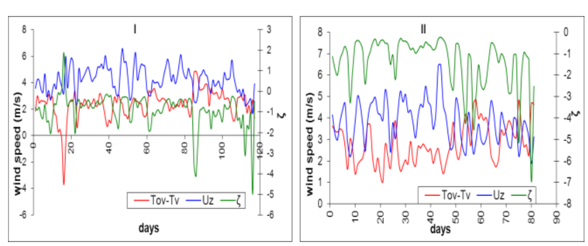



Fig. 7 Daily mean of wind speed, Δθ and ζ

## 5.4 Sensible and Latent Heat Fluxes

SST generally affected by the sensible (H) and latent heat (E) as shown in (Figure 8). While the daily mean $C_e$ during (I) was 28.53 % and during (II) was 36.19% higher than the mean $C_{en}$ (Figure 9), the daily mean E was enhanced during (I), by 27.4 % and during (II) by 33.89% relative to $E_n$ with an assumed neutral ABL (Table 1). This is because at low wind speed and small E the effect of atmospheric stability on $C_e$ was highest. Also, the daily mean H enhanced during (I), by 291 % and during (II) by 220% relative to $H_n$ with an assumed neutral ABL (Table 1).

During (I) the differences between estimated neutral and stability adjusted values of E and H were significant (paired t test, t = 24.45, 40.95, respectively, df = 116, p = 0) for the difference between E and $E_n$, and H and $H_n$. Also, during (II) the differences between estimated neutral and stability adjusted values of E and H were significant (paired t test, t = 33.17, 40.31, respectively, df = 80, p = 0) for the difference between E and $E_n$, and H and $H_n$.

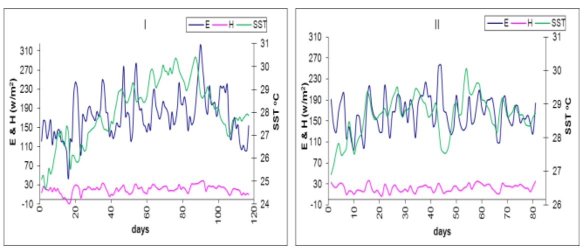



Fig. 8 Values of latent heat (H) and sensible heat (E) and SST

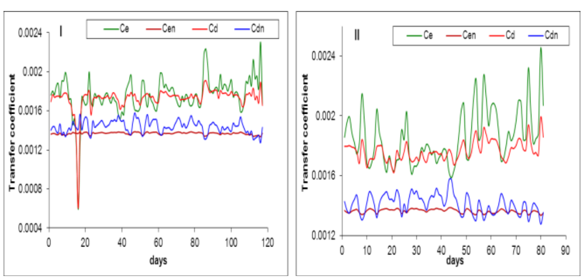



Fig. 9 Daily mean of transfer coefficients

Table 1. Comparison of daily mean neutral and stability dependent drag, transfer coefficient and heat fluxes

| | Phase-I | Phase-II |
|---|---|---|
| $C_{dn}$ | 1.44 | 1.41 |
| $C_{en}$ | 1.37 | 1.36 |
| $E_n$ | 137.5 | 122.7 |
| $H_n$ | 5.28 | 6.96 |
| $C_d$ | 1.73 | 1.77 |
| $C_e$ | 1.76 | 1.86 |
| E | 175.2 | 164.3 |
| H | 20.69 | 22.31 |
| Differences | | |
| $C_d$ | 19.9% | 25.27% |
| $C_e$ | 28.5% | 36.19% |
| E | 27.4% | 33.89% |
| H | 291% | 220% |

E and H units: $W/m^2$. $C_{dn}$ , $C_{en}$ , $C_d$ , $C_e$ are multiply by $10^{-3}$. Differences are relative to the values for the neutral case.

Taking into account the transfer coefficients from the literature ($C_e = C_h = 1.35 \times 10^{-3}$ (MacIntyre *et al.*, 2002), E for both phases was underestimated by 22.71% and 26.21 %, respectively relative to the results achieved here. On the other hand the high transfer coefficients value suggested in literature (e.g., $1.9 \times 10^{-3}$) (Imberger and Patterson, 1989), E for both phases was overestimated by 8.73 % and 3.84%, respectively relative to the results achieved here.

Relatively high unstable or stable conditions ($|\zeta|$ > 4.98 (Figure 4) occurred when wind speed was nearly low value (average 1.86 m/s). According to (equations 1 & 2) the latent and sensible heat fluxes are proportional to wind speed. So, the fluxes were small under highly ($|\zeta|$) value.

## 5.5 ABL Stability

SST was on average 0.81 °C and 1.24 °C during summer seasons of 2010 and 2011, respectively above air temperature in AEH. Mean temperature differences between SST and the air temperature were in positive values, allowing unstable conditions to develop (Figure 10).

According to Verburg and Antenucci, (2010) the atmosphere in the boundary layer is only unstable when the virtual temperature of air decreases with increasing height above the water surface. Figure (11) represents the relationship between (Tov-Tv) and $\zeta$ above AEH during the summer seasons of 2010 and 2011. In AEH and during (I) the atmosphere could be unstable even with air temperatures higher than the water surface by less than 1 °C (Figure 12).

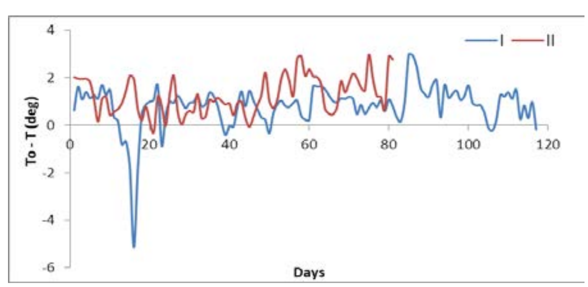


Fig. 10 Daily difference (SST-Air temperature)

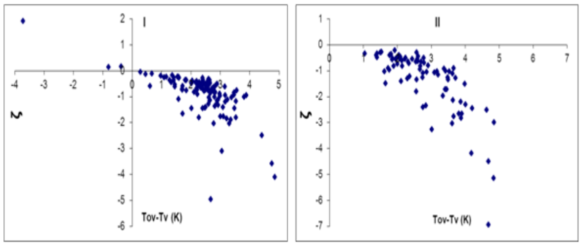



Fig. 11 Relationship between Δθ and ζ

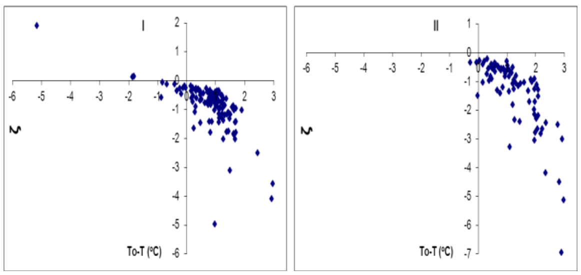



Fig. 12 Relationship between (SST – air temperature) and ζ

When Δθ was positive and then the atmosphere was unstable the ratio of the stability adjusted transfer coefficients to the neutral coefficients $C_e/C_{en}$, was > 1.

The change in the proportional adjustment of the coefficients with ζ was rapid for ζ close to zero (Figure 13) the ratio $C_e/C_{en}$ was larger for unstable conditions than $C_d/C_{dn}$ (Figure 13).

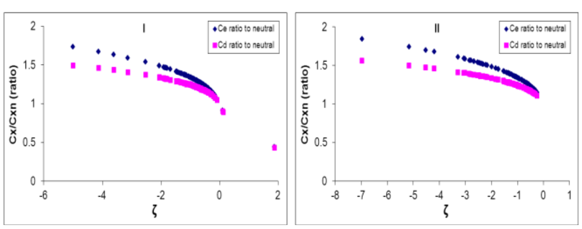



Fig. 13 Relationship between the stability parameter ζ and the ratio of the stability adjusted transfer coefficients $C_e$ and $C_d$ to respective neutral coefficient

## 6. Conclusion

Sensible (H) and latent (E) heat fluxes are affected by atmospheric stability. The Eastern Harbor atmosphere was almost always unstable during summer seasons of 2010 to 2011, in contrast to the atmosphere in other regions which are unstable in winter and stable in summer, resulting in enhancement of transfer coefficients and heat fluxes. The persistent unstable atmosphere is caused by a higher water surface temperature compared with air temperature, which is mainly the case Alexandria Eastern Harbor (AEH). In AEH, the atmosphere could be unstable even with air temperatures higher than the water surface by about less than 1 °C. The stability adjustment of the transfer coefficient was largest at the high stability parameter |ζ|, strong heat loss only occurred when ζ was closed to zero due to the effect of high wind speed. The effect of decreasing humidity and increasing $T_o$-T on Δθ would increase the incidence of unstable events. According to the results of paired t-test the differences between estimated neutral and stability adjusted values of E and H were significant.

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