Measurement of Evaporation From Lakes and Ponds in Florida

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INTRODUCTION

The Florida legislative mandate of 1972 charging water management districts with the responsibility of rationally allocating the state's water resources has resulted in increased pressure to assess water management techniques. In particular, due to the unique climatic conditions in Florida, relative to other states, the need for local assessment of methodologies for water resource management is quite critical. A key factor in any model of the hydrological cycle is the estimation of water loss by evaporation from lakes and ponds. Currently hydrological models often calculate evaporation in terms of corrections to evaporation pan data [see e.g. Holtan et.al.(1974)]. A more satisfactory approach would be to estimate evaporation rates in terms of the turbulent transport processes which govern the diffusion of water vapor in the atmospheric boundary-layer.

In addition to the importance of lakes in the hydrological cycle, the moderating effects of large bodies of water on surface temperatures caused by the advection of energy from lakes is also quite important. Florida growers have found through experience they can obtain some moderating effects on minimum temperatures which occur at night during freeze conditions by planting on the southern shores of lakes.

Thermal protection of crops by lakes will become increasingly important as fuel costs for freeze protection continue to rise. Unfortunately, many lakes in Florida are undergoing severe eutrophic processes either through natural conditions or conditions aggravated by man's use of water. The Florida Department of Environmental Regulation and the Florida Fish and Game Commission have recommended lake drawdown as a method of removing excess nutrients from water, germinating lake bottom vegetation, and solidifying

lake bottom sediments in studies of eutrophic lakes such as Lake Carlton and Apopka. However, many of the lakes are in need of renewal in key agricultural areas. Thus an assessment of lake elevation levels compatible with thermal protection needs in agriculture is critical. A better understanding of the turbulent mechanisms involved in the advection of heat from lakes is necessary in order to assess these needs.

We have sought in this report to evaluate conditions for which current techniques used to estimate turbulent transport would be effective. In this report turbulent diffusion is evaluated through the use of 1) the Bowen ratio energy method, 2) numerical simulationsof the atmosphere, 3) the eddy correlation technique and 4) bulk transfer coefficients. In Chapter I, instrumentation to measure the Bowen ratio (ratio of sensible to latent heat flux to the atmosphere) is discussed. The instrumentation was designed to be both flexible and rugged for practical field measurement of evaporation. After development, the instrumentation was adapted for use on an instrumented boat and was subsequently used to measure latent and sensible heat fluxes from Lake Apopka. These results along with a comparison with the results of other investigators are reported in Chapter II. In addition to the measurement of energy fluxes from Lake Apopka, thermal imagery indicated the downwind distance for which thermal effects due to the lake were significant. These observations suggested that the atmospheric model of Gutman (1974) (used to model the response of the urban heat island to surface roughness and heat addition) would adequately model the transport of energy from Lake Apopka. A comparison of the model results with measured data indicates that a 2-dimensional model may be successfully employed to predict air temperature maxima downwind of a lake the size of Apopka. In Chapter IV, direct measurements of evaporation on Orange Lake are reported. The eddy correlation technique was employed using

an ultra-sonic anemometer thermometer and a Lyman-alpha humidity sensor. Results were compared with bulk transfer coefficients, as well as a new relationship by Hicks (1975) which predicts fluxes solely on the basis of surface temperature. Finally in Chapter V, bulk transfer coefficients are used to predict the diurnal variation in the flux of latent heat from East Lake Tohopekeliga.

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CHAPTER I

A COMPARISON OF EDDY-CORRELATION AND BOWEN RATIO SENSORS IN THE MEASUREMENT OF EVAPORATION[1]

R.G. Bill, Jr., L.H. Allen, Jr., and J.F. Bartholic

Abstract

The feasibility of using the Brady array as a humidity sensor in the Bowen-ratio and eddy-correlation techniques was determined. Instrumentation for the Bowen-ratio included the Brady array and a 3-mil copper-constantan thermocouple which continuously traversed up and down 1.5 m over a bare soil surface. A comparison of the humidity fluctuation spectra of the Brady array and a Lyman-alpha indicated that the response of the Brady array was adequate for profiles of mean vapor pressure. Fifteen-minute profiles of temperature and vapor pressure were consistent with expected results for a constant flux layer. The variance spectra indicated that the Brady array was not suitable for eddy-correlation measurements.

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[1] Summary of paper presented at 13th American Meteorological Society Conference on Agriculture and Forest Meteorology, Purdue University, West Lafayette, Indiana, April 4-6, 1977.

1. Introduction

Two techniques useful for measurement of evaporation where mass balance methods are inaccurate or unfeasible are the Bowen-ratio and the eddy-correlation. The present study investigates the Brady array for use in these techniques.

As discussed by Sinclair et. al. (1974), the Bowen ratio, B, in finite difference form, is (notation standard):

$$B = (\rho C_p K_H \Delta t)/(\ell K_v \Delta \rho_v).$$

The chief difficulty of the Bowen-ratio method results from the small temperature and humidity gradients within the surface boundary layer. Either temperature and water vapor sensors must be closely matched to provide accurate profiles gradients, or a single sensor may be translated up and down. Another problem is the estimation of the transport coefficients, K_H and K_v. Typically $K_H/K_v = 1.0$ is used (Lemon, 1965).

In the eddy correlation technique, it is possible to measure sensible and latent heat flux density directly by correlating flow variables from a time series of digital samples. The samples in the time series must be taken frequently enough and over a long enough time period to include energy from the entire spectrum of turbulent eddy scales. McBean (1972) suggested that the frequency spectrum for atmospheric turbulence is about 0.005 to 10 Hz.

2. Bowen-Ratio Instrumentation

Instrumentation for the Bowen-ratio technique was developed using a 3-mil copper constantan thermocouple and a Brady array (Thunder Scientific

Corporation[2]). The thermocouple and Brady array, placed in a solar radiation shield, were mounted to the belt of a pulley system that continuously traversed up and down 1.5 m above the soil surface. Each cyclic traverse required 25 seconds.

Given the traverse speed, a response time of 1/3 second seemed adequate to measure mean profiles. The Brady array was mounted with a Lyman-alpha humidiometer (Electromagnetic Research Corp.) 2 m above the soil surface to determine the response characteristics. Miyake and McBean (1970) found the Lyman-alpha to be superior to a dewpoint hygrometer for flux density measurements. A variance spectrum of vapor density was calculated for our two humidity sensors using a Fast-Fourier transform algorithim.

In Figure I-1, variance spectra, Φ, of the fluctuations, ρ_v', and analog records of the humidity sensors are shown. The analog records indicate that much of the detailed picture of turbulence is cut off by the response of the Brady array. The variance spectra further indicate that the response of the Brady array is not adequate for measuring turbulent flow. Below 0.1 Hz, the variance spectrum is overestimated, while at higher frequencies, the variance spectrum is underestimated. Above 3 Hz, signal noise was encountered and no conclusions could be made.

The static calibration of the Brady array was checked over a relative humidity of 30 - 80% using an Atkins lithium chloride dewpoint hygrometer. The two sensors agreed within 5% relative humidity. However, large sudden

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changes of relative humidity (of the order of 30%) caused hysteresis problems. The overestimation of variance at low frequencies may result from this hysteresis. Thus, it appears that the Brady array is unsuitable for eddy-correlation studies. However, its response characteristics appear adequate for the measurement of mean humidity profiles.

In Figure I-2, profiles of vapor pressure and temperature from the Brady array and the thermocouple are presented. The profiles are from successive 15-min time periods between 1200 and 1300 EDT. The Brady array and thermocouple moved continuously up and down 1.5 m above a bare soil surface starting about 5 cm above the surface. Sensor outputs were digitalized approximately every 4 seconds. The position of the sensor was checked at the beginning of each cycle by using an event marker for the computer. With increasing time, both the temperature and vapor pressure increased.

Vapor pressure is plotted as a function of temperature in Figure I-3 for two profiles. Within the constant flux layer, the points should form a straight line. These profiles indicate the maximum and minimum deviations from linearity. Both curves, however, show extensive regions of linearity and are similar to results of Sinclair et. al. (1974) for the constant flux region above a corn crop. This result confirms both the adequacy of the response characteristics of the sensors and the sampling technique.

Conclusions

The Brady array has response characteristics adequate for a Bowen-ratio technique using moving sensors. Profiles of vapor pressure based upon 15-min averages appear to be statistically stable and consistent with results expected for a constant flux layer. However, the Brady array is not an optimal sensor for eddy-correlation studies as its response range is limited to about 3 Hz.

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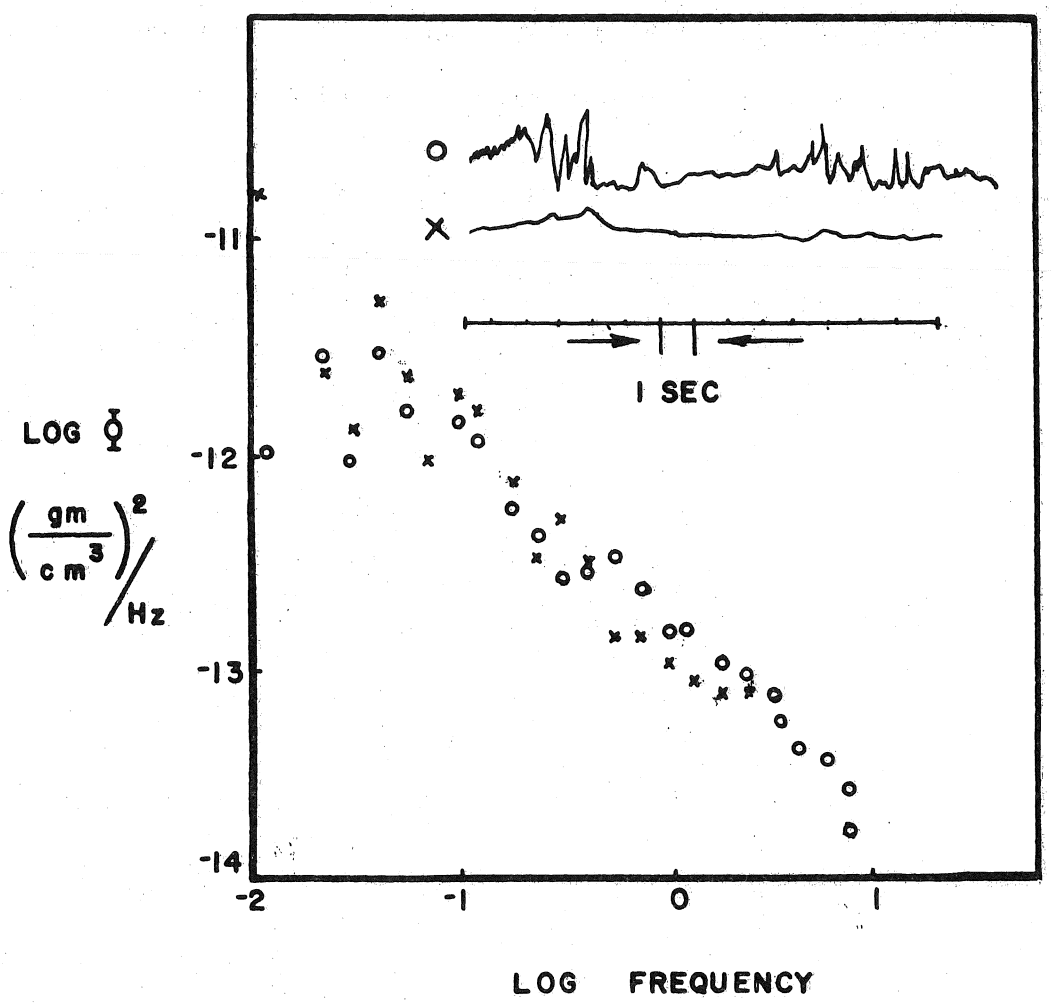

Figure I-1. Comparison of the variance spectra of the Brady array (x) and the Lyman-alpha (o).

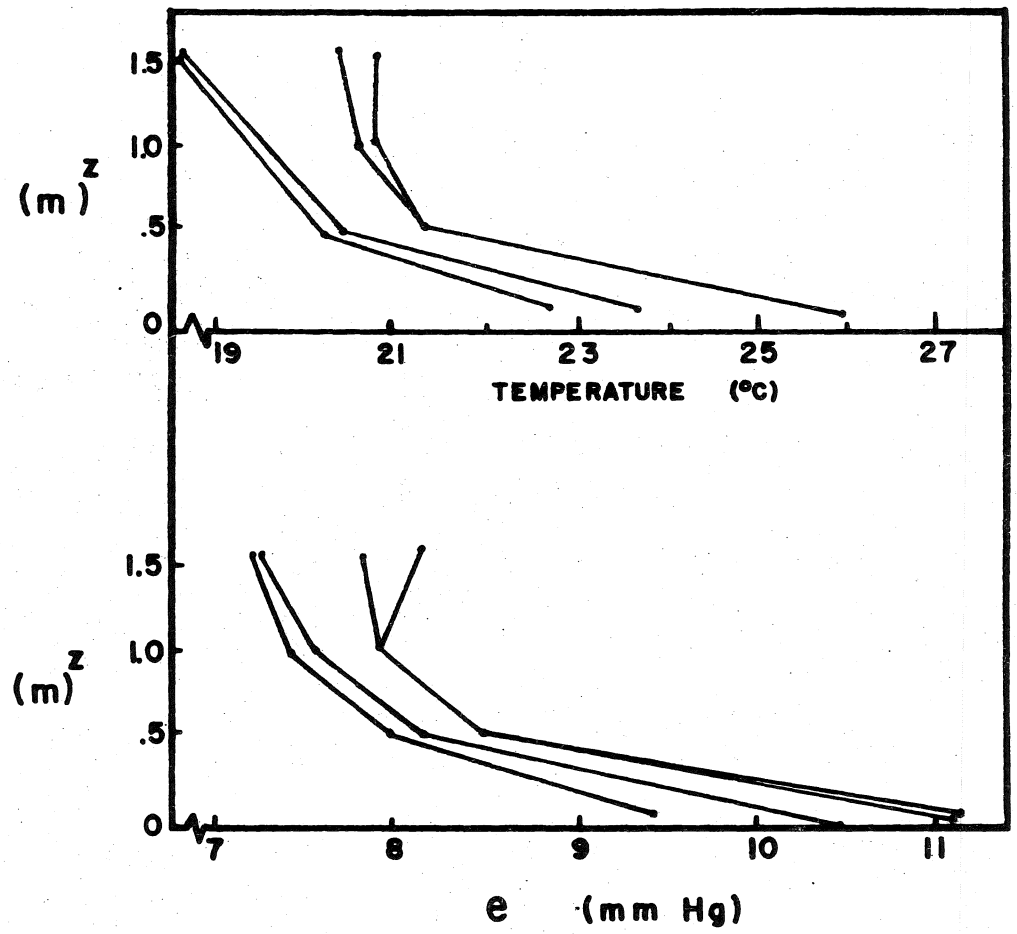

Figure I-2. Profiles of temperature, t, and vapor pressure, e, above the soil surface.

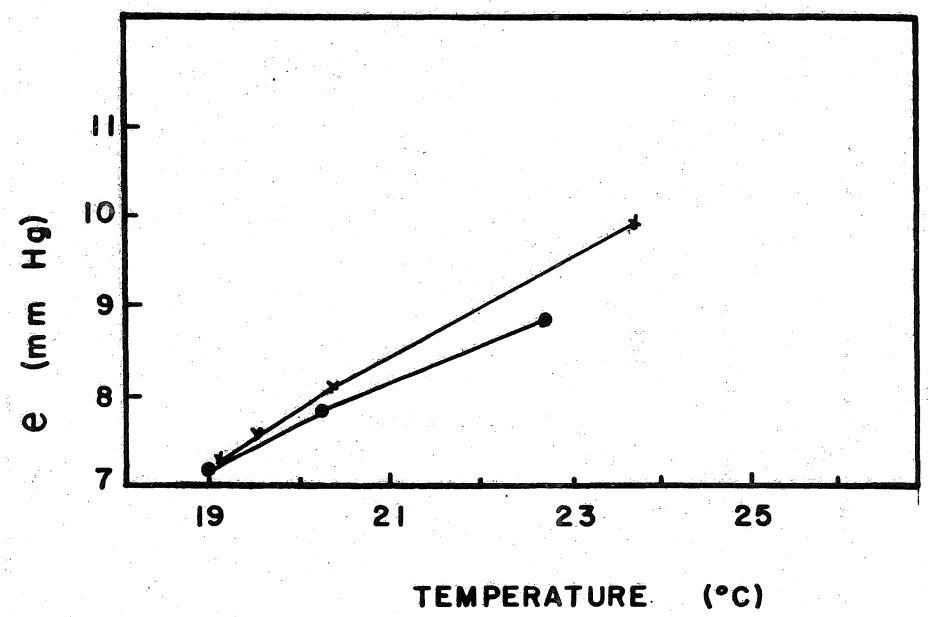

Figure I-3. Vapor pressure, e, as a function of temperature, t, above the soil surface.

CHAPTER II

Observations of the Convective Plume of a Lake Under Cold Air Advective Conditions

R.G. Bill, Jr., R.A. Sutherland, J.F. Bartholic, and E. Chen

Abstract

Moderating effects of Lake Apopka, Florida on downwind surface temperatures were evaluated under cold air advective conditions. Point temperature measurements north and south of the lake and data obtained from a thermal scanner flown at 1.6 km, indicate surface temperatures directly downwind may be higher than surrounding surface temperatures by as much as 5°C under conditions of moderate winds (∿4 m/sec). No substantial temperature effects were observed with surface wind speed less than 1 m/sec. Fluxes of sensible and latent heat from Lake Apopka were calculated from measurements of lake temperature, net radiation, relative humidity and air temperature above the lake. Bulk transfer coefficients and the Bowen ratio were calculated and found to be in agreement with reported data for non-advective conditions.

1. Introduction

Moderating effects of large bodies of water on surface temperatures of coastal regions have been noted qualitatively by climatologists for many years. Agriculturists have been aware of similar effects on a much smaller scale produced by the plumes of lakes and have used this information as an aid in the selection of growing sites for cultivars subject to cold damage. Florida citrus growers, for example, have found through experience that they can obtain some moderating effects on minimum temperatures which occur at night during freeze conditions when winds are predominately from the north by planting on the southern shores of lakes [Lawrence (1963), Bartholic and Sutherland (1975)]. These practical effects are well known but surface temperature data is quite sparse [see eg. Geiger (1965)]. No detailed quantitative data are available, however, concerning energy transport and flow conditions associated with such convective plumes.

Transport mechanisms include both turbulent and radiative transfer. Lakes, being warmer than the air and surrounding land, release sensible and latent heat under typical cold conditions following the passage of a cold front. Sensible and/or latent heat (depending upon the downwind dewpoint temperature) is transferred by turbulent diffusion to the ground as the moist buoyant plume is advected beyond the lake by the moving air mass. The radiation energy balance of the surface, downwind of the lake, may also be affected by advection of water vapor from the lake surface. That is, local changes in the vertical profiles of absolute humidity may change net radiation loss from the ground.

The situation described above is comparable to the effect that urban areas have on the surface boundary layer at night. A complete survey of the "urban heat island" effect is given by Gutman (1974). Energy is supplied in

the case of the urban heat island not only through the release of internal energy but also by waste heat added to the atmosphere by man's utilization of energy for industrial and residential purposes. These energy sources result in high air and surface temperatures in and downwind of cities as compared with surrounding rural areas. For example, Bornestein (1968) reported elevated temperatures of about $4^{o}C$ for Manhattan. Preston-White (1970) in a study of the urban heat island surrounding Durban, South Africa indicated the center of the heat island, defined in terms of temperature contours, may be displaced from the central business district by a sea breeze. Similar temperature contours downwind of the center of the urban heat island were predicted by Gutman's numerical model (1974).

Data are provided herein on surface temperatures, radiation, turbulent transport and wind flow over Lake Apopka. Data were taken on the night of 19 January, 1977, a night of severe cold conditions. It was possible to delineate flow conditions in which significant thermal effects could be expected to occur downwind of lakes and estimate the energy transport associated with these effects.

2. Description of the Site

Lake Apopka is located in the central portion of the Florida peninsula (see Figure II-1). The lake is approximately 13 km across and has a mean depth of 1.65 m. Large areas of citrus exist both south and east of the lake. These are low flat areas with sandy soils not normally well adapted for citrus due to their freeze susceptibility from poor air drainage. There is also a considerable area of citrus growth west of the lake, on land with relatively high elevation. The area immediately north of the lake is characterized by organic soils and is used for vegetable farming.

3. Experimental Methods

Measurements of temperature, humidity, wind speed and direction are needed in order to properly assess the thermal effects of Lake Apopka on the surrounding area. Furthermore, radiation and lake temperature data are also necessary to characterize the energy flux associated with the plume of the lake. Towers were erected to a height of 15 m at locations north and south of the lake on the "freeze night" of 19 January 1977 (see Figure II-1). Temperature profiles were sensed with 24-gauge copper-constantan thermocouples and recorded on a multipoint thermocouple potentiometer with an ice-point reference junction. Transects were run along Route 50 as shown in Figure II-3 measuring variations in temperature using an Atkins* thermistor unit to supplement these data. The time for a complete traverse south of the lake was less than 45 min.

Systematic temperature differences from location to location quite independent of any lake effect would be expected due to variations in land cover, soil type, and drainage. A NASA aircraft using an infra-red scanner (8 - 14μ) was used to determine temperature patterns south of the lake to avoid biasing conclusions with results from only a few point temperature samples. The aircraft flew a set of 6 flight lines south of the lake at 3 different times in an east-west direction. The aircraft flew at an altitude of 1.6 km scanning a total area approximately 10 x 17 km. Information from the scanner was recorded on analog tape in the aircraft. Data were subsequently processed at the Kennedy Space Center. The film was

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'false colored' with each color representing a specific temperature interval [Sutherland and Bartholic (1974)].

Measurements on Lake Apopka proper were acquired on a specially instrumented boat provided by NASA, Kennedy Space Center. The boat was located near the eastern shore as shown in Figure II-1, since prevailing winds for the night were west to northwesterly. Depth of the lake at this location was 1.7 m, as compared to an average lake depth of 1.65 m.

Instrumentation in the boat included a wet bulb-dry bulb psychrometer, a thermistor for measuring water temperature, a Swissteco S-1 net radiometer, and a cup anemometer and wind vane. Instrumentation to determine the Bowen ratio (B = H/LE) was mounted on a boom extending approximately 2 m beyond the bow of the boat to estimate the amount of energy released by the lake through sensible heat, H, and latent heat, LE. The Bowen ratio instrumentation consisted of a Brady array humidity sensing device and an 80μ copper-constantan thermocouple referenced to an ice-water junction. The temperature and humidity array were attached to the belt of a pulley system and continuously traversed up and down a distance of approximately 1.5 m. The bottom point of the traverse was approximately 0.4 m above the mean surface level of the lake. Time for a complete traverse was 40 seconds. Changes in relative humidity and temperature levels were recorded continuously on strip chart recorders. A microswitch at the top of the traverse triggered an event marker so that a record of the position of the array could be maintained. Differences of vapor pressure and temperature across the 1.5 m were calculated from instantaneous outputs of the Brady array and the thermocouple. These results along with measurements of net radiation and rate of change of lake temperature then allowed latent and sensible fluxes to be calculated.

4. Results

4.1 Temperature measurements and wind conditions

Hourly meteorological conditions are plotted from data obtained from the instrumented boat on Lake Apopka (Figure II-2). The data are for the time period of 0030 - 0530 EST of the morning of 20 January 1977. Figure II-2 indicates wind speed, air temperature, and specific humidity were relatively constant throughout most of the night. Wind direction varied gradually from the west at 0030 to the north at 0400. Turbulent transport of sensible heat was sufficient to keep the air mass over the lake above 0.5°C at wind speeds above approximately 4 m/sec. The mean 2 m air temperature obtained from the measured temperatures profiles north and south of the lake were -5.8°C and -3.8°C, respectively, during the same time period.

Mean wind decreased from 0400 to 0600 until it was below the stall speed of the cup anemometer (∿40 cm/sec). Reduction in wind speed occurred simultaneously with the reduction in air temperature (Figure II-2). This result is expected since a reduction in wind speed reduces the coefficient of turbulent diffusion [Saito (1967)]. A similar variation in specific humidity would be expected; however, no variation was detected due to the relative insensitivity of the wet bulb-dry bulb psychrometer used.

The source of energy for the transported sensible and latent heat is the change in internal energy of the lake. The lake temperature at midnight was 5.6°C. A vertical traverse of the lake indicated that the lake was isothermal to a depth of 1.5 m. Below this level, the lake was characterized by colloidal organic matter and the lake bottom was not well defined. Due to the high degree of mixing in the lake, no gradients in temperature were detectable in this layer and the temperature remained constant throughout the night. The temperature above a depth of 1.5 m changed at a mean rate of

0.14°C/hour over a period of 5 hours.

The effect of the sensible heat transferred from Lake Apopka on the air temperature downwind of the lake is dramatically shown in Figure II-3, a plot of air temperature measured along Fl. 50. Data are plotted for two time periods, 0218 to 0303 and 0548 to 0611.

Wind fluctuated between 4 and 5 m/sec during the time period 0218 to 0303 (see Figure II-2). A temperature maximum is readily apparent in Figure II-3, along the axis of the lake as viewed from the wind direction. A general pattern of decreasing temperature with distance from this axis is also discernable. The temperature maximum is generally greater than 1°C higher than other measured temperatures along Fl. 50.

No regular temperature pattern is apparent for the time period 0548 to 0611 EST when wind speed was well below 1 m/sec. Variations in temperature for this time period do not appear to be correlated in any way with location with respect to the lake. It can be presumed that the observed variations in temperature are due to differences in elevation and land use as well as to time intervals between the various measurements.

The contrast between the temperature patterns for the two time periods indicates the pronounced effect that wind speed has on downwind temperatures. Variations in wind speed not only affect the rate at which energy may be transported from the lake but also the rate at which this energy is advected. Furthermore, the water vapor content of the atmosphere could not have substantially changed during the time period between the two sets of temperature measurements, hence it appears that effects due to radiation are not as important as the turbulent processes.

A further indication of the correlation between downwind air temperature and position with respect to wind axis would be data indicating increased water

vapor content. No dewpoint data are available along Fl. 50 for the night of 19 - 20 January 1977. However, dewpoint data as well as temperature data are available for 0100 February 1977, a night of similar meteorological conditions (Figure II-4). Mean wind speed was 6 m/sec and the wind direction is as noted. Figure II-4 clearly shows a strong correlation between vapor content and position with respect to the wind direction axis. The outline of the plume as in Figure II-3 is quite visible.

Temperature profile data for 19 - 20 January, 1977 from the towers north and south of the lake are shown in Figure II-5 for the time periods 0130, 0330, and 0530. At 0100, air temperatures at a given elevation above the ground are warmer south of the lake particularly below 6 m. However, differences in temperature above 6 m are quite small. This is to be expected since the wind direction previous to this time period was westerly and the plume of the lake would not interact strongly with the air mass at the location of the southern tower. By 0330, the wind direction was northwesterly and air temperatures are consistently warmer by $1^{o}C$ or more south of the lake. Finally, at 0530, when the wind speed dropped below 0.4 m/sec, the differences in air temperature at elevations below 6 m has significantly decreased. The decrease in the air temperature in the air surface layer at the southern tower location is due to the decrease in energy advected from the lake and the subsequent drop in ground temperatures.

4.2 Data from the thermal scanner

It is now possible to interpret the more complex temperature patterns provided by the thermal scanner with the above background data in mind.

Temperature patterns south of Lake Apopka as established by the thermal scanner are shown in Figures II-6 and II-7. The color code associated with

these two figures is detailed in Table II-1. Data for Figures II-6 and II-7 were obtained, respectively, between midnight and 0100; and between 0500 and 0600; the morning of 20 January 1977.

A warm region southeast of Lake Apopka is quite apparent during the period of relatively vigorous winds (Figure II-6). Temperature ranged between -1.7 to 1.7^{o}C (red and yellow) along the southern lake shore. This contrasts markedly with temperatures east and west of the lake where temperatures -5.0^{o}C or lower (magenta and black) predominate. The temperature decreases southeast of the lake in the direction of the wind. However, the surface temperatures, typically -3.3^{o}C (blue and green), are still higher than the surface temperatures east and west of the lake where temperatures below -8.3^{o} occur. These temperature patterns established by the thermal scanner are quite consistent with the air temperatures shown in Figures II-2 and II-3. The surface temperature pattern south of the lake is somewhat complicated by the presence of smaller lakes but it appears that the pattern of elevated temperatures vanishes about 6 km southeast of Lake Apopka.

No correction was made for emissivity or for water vapor absorption in the atmosphere. Sutherland and Bartholic (1977) have shown that crop surfaces, such as the citrus area south of Lake Apopka, radiate approximately as black-body cavities and that the error due to emissivity is on the order of only 1°C. Occasional bare, or nearly bare fields will sometimes appear (erroneously) as cold spots due to the low emissivities of the sands in the area. These, however, occur nearly randomly throughout the entire data set and can be ignored when looking at large scale phenomena. The scanner used for the flight receives IR data in the 10-14μ "window" in water vapor absorption. Calculations based on the method of Greenfield and Kellogg (1959) using radiosonde data provided by the National Weather Service indicate that the error due to

atmospheric absorption is less than 0.5°C. Emissivity and atmospheric corrections are moreover of opposite sign, hence, the resultant error is quite small for the night of 19 - 20 January 1977.

The organized plume apparent in the surface temperature patterns shown in Figure II-6 is greatly diminished in Figure II-7. Regions below -8.3°C appear throughout the data set. The mean wind for this time period was below 1 m/sec. The rate at which sensible heat was advected from the lake and transferred to the ground was greatly decreased. Hence, the surface temperatures rapidly decreased through radiative cooling. Small variations in temperature such as those shown in Figure II-3 are visible; however, the predominate temperature range is less than -5.0°C. Only those regions quite close to the smaller lakes or to marsh lands in the region have maintained their elevated temperatures.

The wind direction was from the northeast on the night of 9 January 1976. Data for that night from a thermal scanner are shown in Figure II-8. The color code for this figure is shown in Table II-2. In contrast to the data for 20 January, 1977, the region of elevated temperature is shifted to the area southwest of the lake. These data further confirm that the thermal plume is responsible for the reported temperature effects.

The temperature patterns from the thermal scanner provide a large data base on which to develop models of advection. A comparison of existing models with the results of this study is now being undertaken and will be reported in the future.

4.3 Determination of energy fluxes from the lake

A final consideration is the energy balance of the lake that was associated with the temperature patterns described above. The energy balance

may be written as:

$$\frac{-du}{dt} = Rn + LE + H + S \qquad [1]$$

(outward flux from the water surface positive)

where Rn, LE, H, and S are respectively the flux densities of net radiation, latent heat, sensible heat, and heat flux from the bottom of the lake; $\frac{du}{dt}$ is the time rate of change of the internal energy per unit area of the lake. The rate of change of internal energy may be calculated from temperature measurements as below:

$$\frac{du}{dt} = \int_o^d \rho c \frac{d\theta}{dt} dz \qquad [2]$$

Lake temperature, θ, decreased at an average rate of 0.14°C/hour over a 5-hour period as noted above. The flux density associated with this change in temperature is 245 watts/m^2 (0.35 cal/cm^2-min) from the measured isothermal temperature profiles to a depth, d, or 150 cm.

Net radiation between midnight and 0500 was approximately constant and averaged 140 watts/m^2 (0.20 cal/cm^2-min). As discussed above, no temperature changes were detectable below 1.5 m. Due to the low thermal conductivity of organic soils [see e.g. Van Wijk (1966)], the flux from the soil was neglected. Hence, residual energy left for latent and sensible heat was 105 watts/m^2 (0.15 cal/cm^2-min).

Recently, Hicks (1975) suggested the following formula for the Bowen ratio of clean water surfaces:

$$B = (K/D) \left(\frac{\gamma}{\Delta}\right) \qquad [3]$$

where K and D are, respectively, the molecular diffusivities for heat and water vapor, γ is the psychrometric constant, and Δ is the slope of vapor pressure versus temperature curve. The ratio K/D is $\sim$0.84, from Hicks (1975). Priestley and Taylor (1972) derived an empirical relationship for the Bowen ratio which, following Hicks, may be expressed:

$$B = 0.8 \left(\frac{\gamma}{\Delta}\right) - 0.2 \qquad [4]$$

The measured Bowen ratio, $B = \frac{C_p \Delta t}{L \Delta q}$, (conventional notation) as well as, the predictions of Hicks (1975) and Priestley and Taylor (1972) are shown in Table II-2. The Bowen ratio was calculated over 20 minute periods from 0100-0220 from the instantaneous analog records of the Brady array and fine thermocouple. Constants have been evaluated at a temperature equal to the average of the water surface temperature and the 2 m air temperature. The Bowen ratio is intermediate to the results predicted by the 2 theories. No choice, however, may be made between the 2 due to the relatively large error resulting from vertical motion of the instrumented boat on the water surface.

4.4 Calculation of bulk transfer coefficients and comparison with previous results

Turbulent fluxes of heat and water vapor above water surface have frequently been evaluated in terms of bulk transfer coefficients of heats, D_H, and water vapor, D_W as in equations [5] and [6].

$$H = \rho\, C_p\, D_H\, U_z\, (T_o - T_z) \qquad [5]$$

(conventional notation)

$$LE = \rho\, L\, D_W\, U_z\, (q_o - q_z) \qquad [6]$$

Hicks (1975a) compiled a survey of existing data on such coefficients and suggested a formula for transfer coefficients taking into account the dependence upon atmospheric stability. Stability is characterized there by the parameter, z/L, where z is the distance above the water surface and L is the Monin-Obukhov length defined below in conventional notation [see e.g. Webb (1965)].

$$L = \frac{-U_*^3}{K_g H_m / \rho C_p \Theta} \qquad [7]$$

H_m, following Webb (1965) is the modified sensible heat flux taking into account the effect of buoyancy by the flux of water vapor; that is,

$$H_m \sim (H + .07\ LE)$$

The friction velocity, U_*, is estimated as $U_* \simeq .03\ U_{2m}$, as suggested by the comment of Hicks (1975a) that $U_* \approx U_o$ and $U_o \approx 0.03\ U_{10m}$. The following results were calculated at a 2 m reference from data taken on the instrumented boat using the measured Bowen ratio energy balance: $z/L = -0.6$, $D_H = 1.7 \times 10^{-3}$, $D_w = 1.3 \times 10^{-3}$. Hicks (1975a) estimates $D_H \sim 1.5 \times 10^{-3}$ for the stability conditions of this study. Hicks (1975a) suggests that $D_H = D_w$, but the data reported in the survey by Hicks (1975a) indicates D_w is typically slightly lower than D_H. Due to the large error estimate for the Bowen ratio, no firm conclusions concerning the formulation of Hicks (1975a) are possible. However, from an operational point of view, our data indicate that the formulation of Hicks (1975) for the transfer coefficient, D_H, appears to be an adequate description of turbulent transfer for the conditions of this study.

Conclusions

Observations of this study indicate surface temperatures downwind of lakes may be higher than surrounding surface temperatures by as much as 5°C under conditions of moderate winds. No substantial thermal effects are to be expected, however, under low wind conditions. Radiative transfer effects due to increased water vapor appear negligible as compared to turbulent transport processes. A comparison of measured sensible and latent heat fluxes indicates the formulations of Hicks (1975a, 1975) are adequate for predictions of the local Bowen ratio and the sensible heat transfer coefficient for the conditions of this study.

Acknowledgments

This work supported by the U.S. Department of Interior, Office of Water Research and Technology as authorized under the Water Resource Research Act; the Florida Department of Environmental Regulation, Office of Lake Restoration, and NASA, Meteorology and Climatology, Kennedy Space Center.

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Table II-1

Color code for Figures II-6 and II-7.

Color	Temperature Interval (^{o}C)
red	0.0 to 1.7
yellow	-1.7 to 0.0
green	-3.3 to -1.7
light blue	-5.0 to -3.3
dark blue	-6.7 to -5.0
magenta	-8.3 to -6.7
black	<-8.3

Table II-2

Color code for Figure II-8.

Color	Temperature Interval (^{o}C)
red	5.0 to 6.7
yellow	3.3 to 5.0
green	1.7 to 3.3
light blue	0.0 to 1.7
dark blue	-1.7 to 0.0
magenta	-3.3 to -1.7

Table II-3

Comparison of measured Bowen ratio with theoretical predictions.

B (Hicks) =	1.03
B (Measured) =	0.95 $\pm$ .15
B (Priestley - Taylor) =	0.80

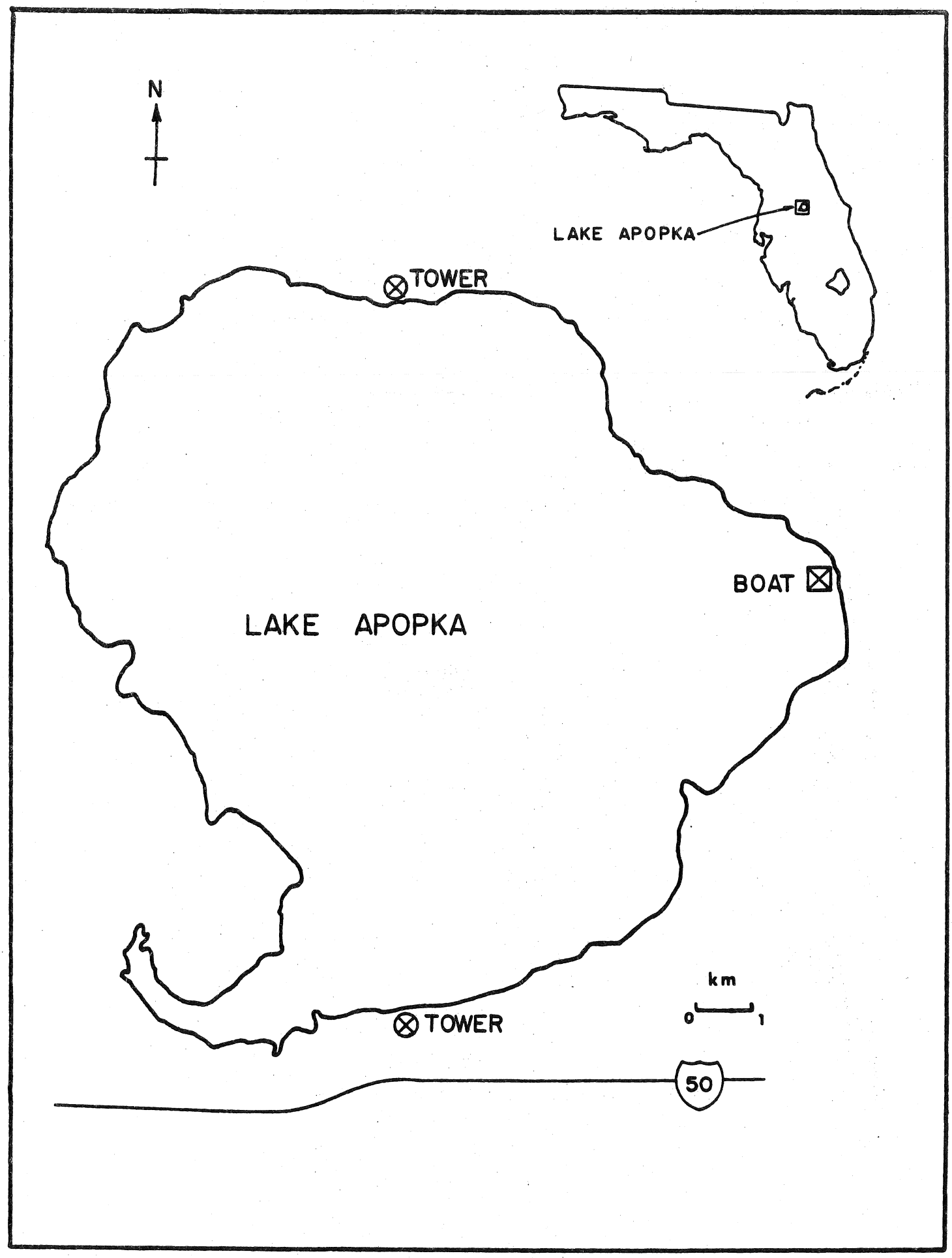

Fig. II-1. Location of Lake Apopka and position of measurement stations.

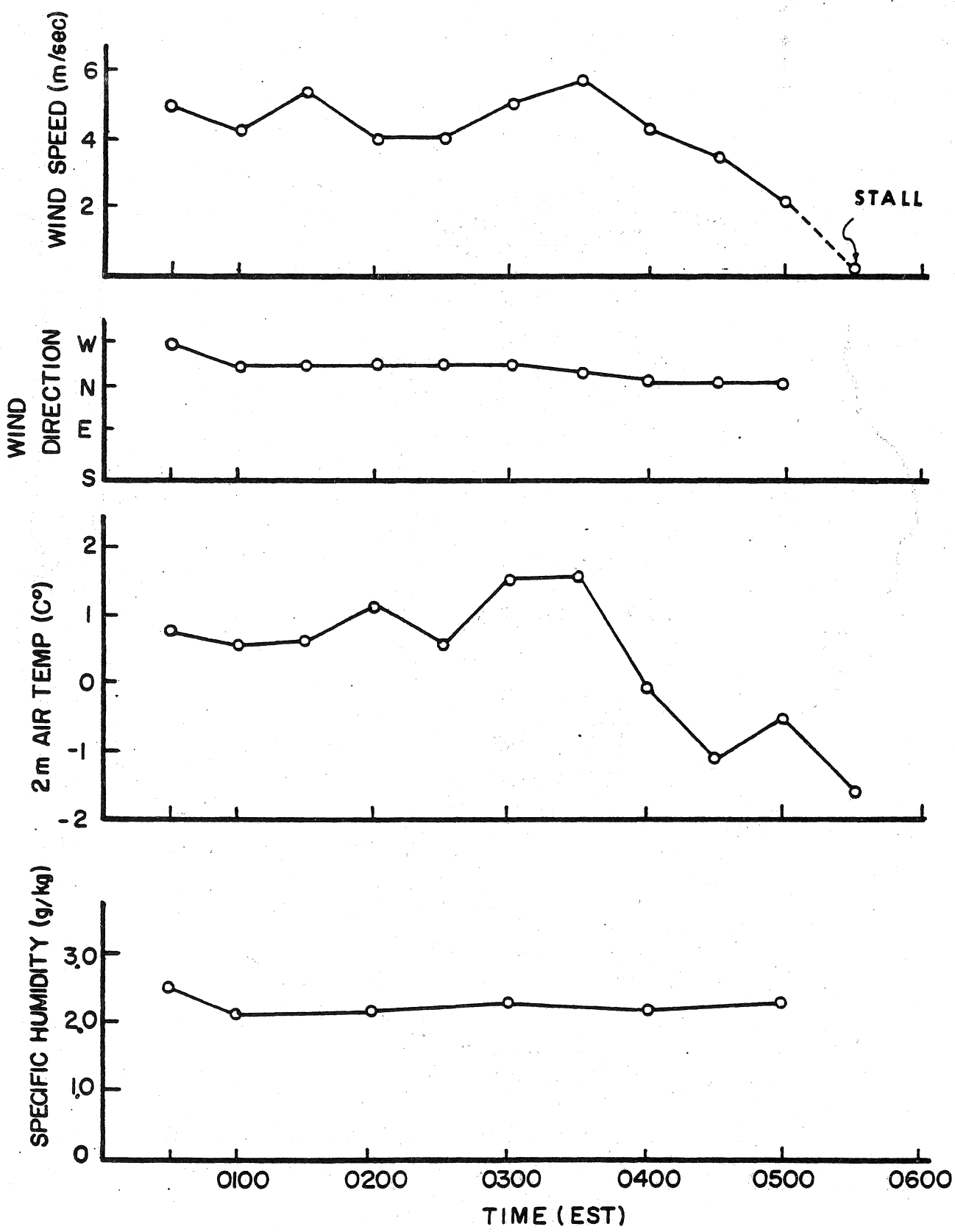

Fig. II-2. Mean meteorological data on Lake Apopka.

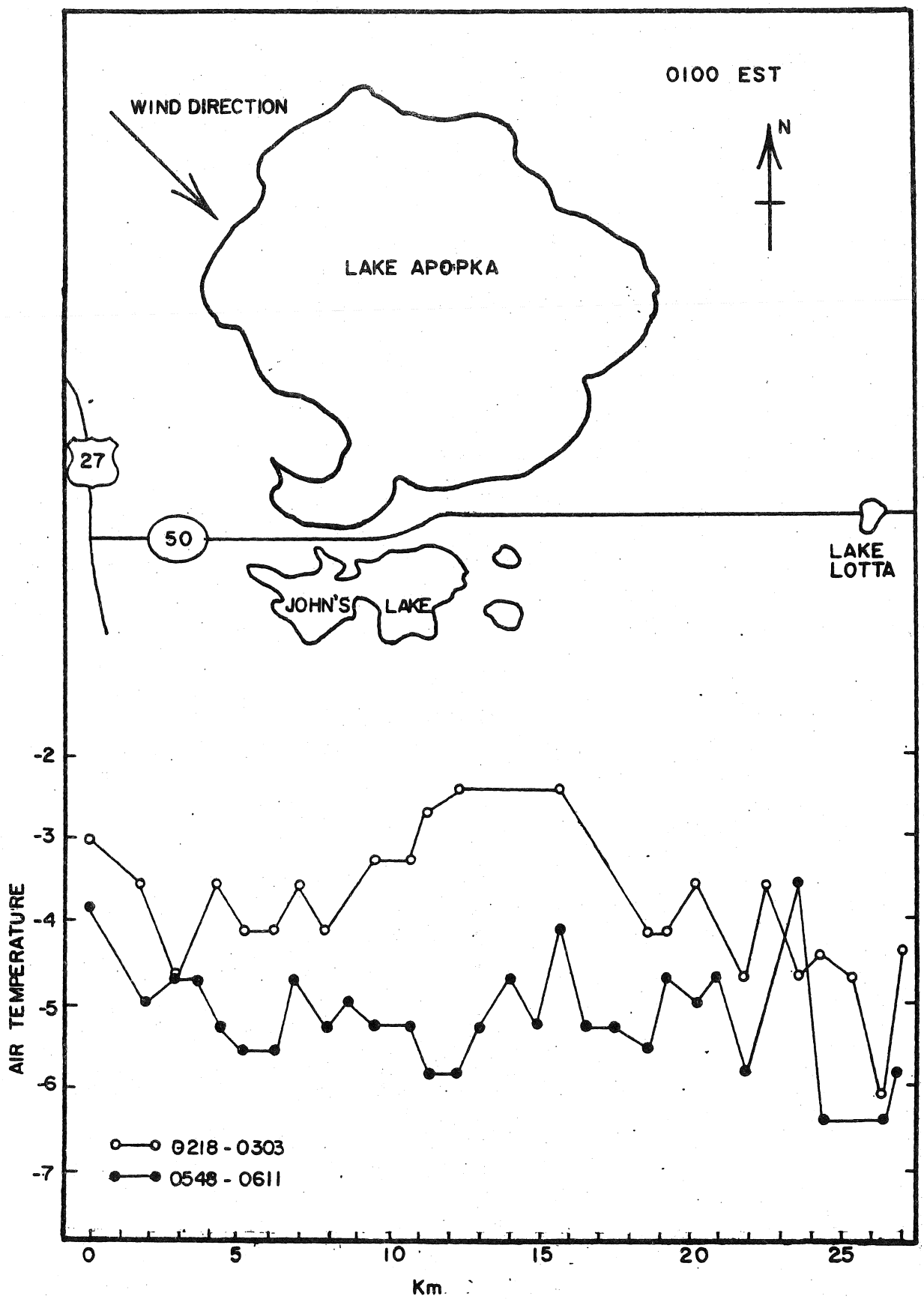

Fig. II-3. Air temperature along Fla. 50 at 02:18 to 03:03 and 05:43 to 06:11 EST, 20 Jan. 1977

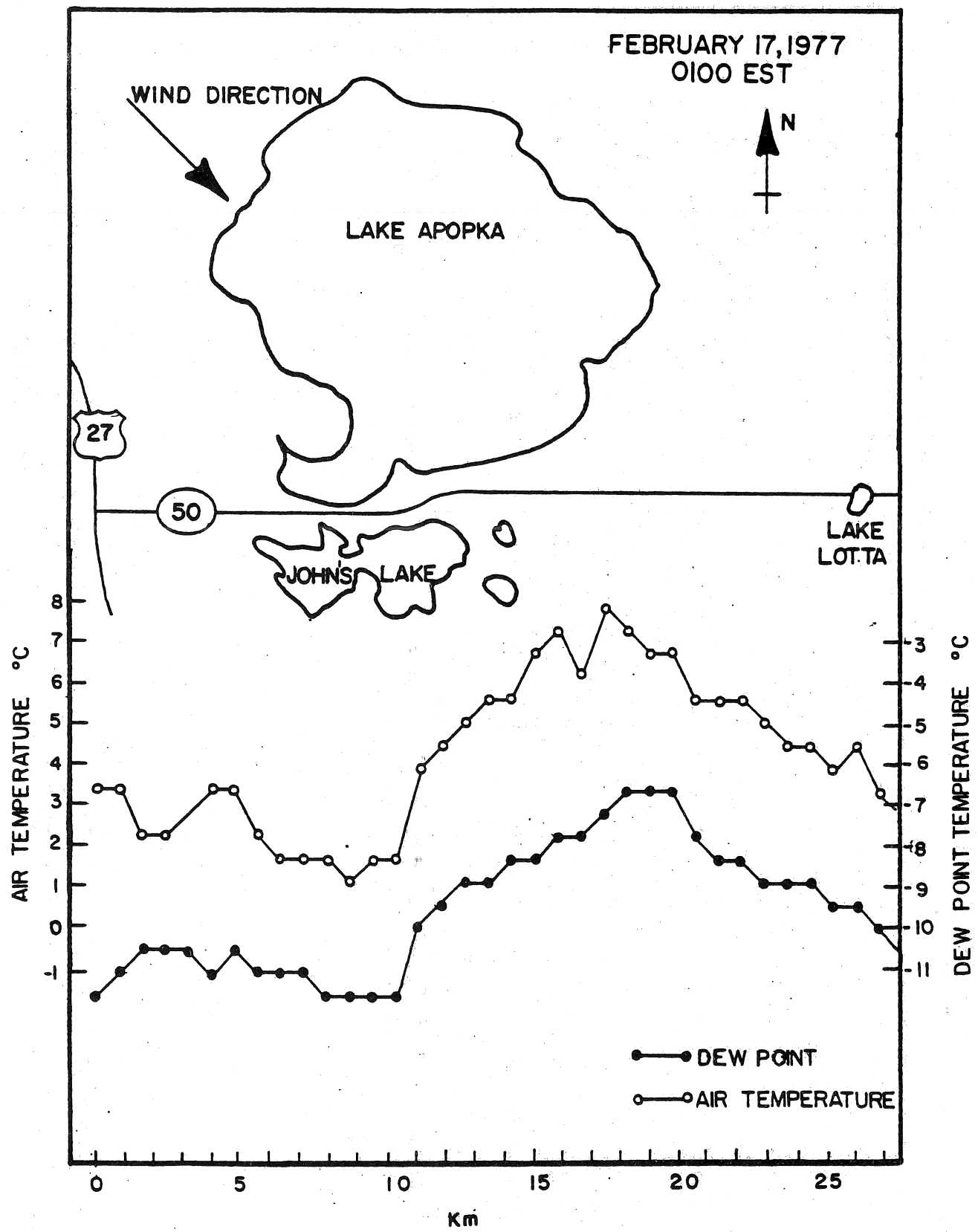

Fig. II-4Air temperature and dewpoint temperature along Fla. 50 01:00, 17, Feb. 1977

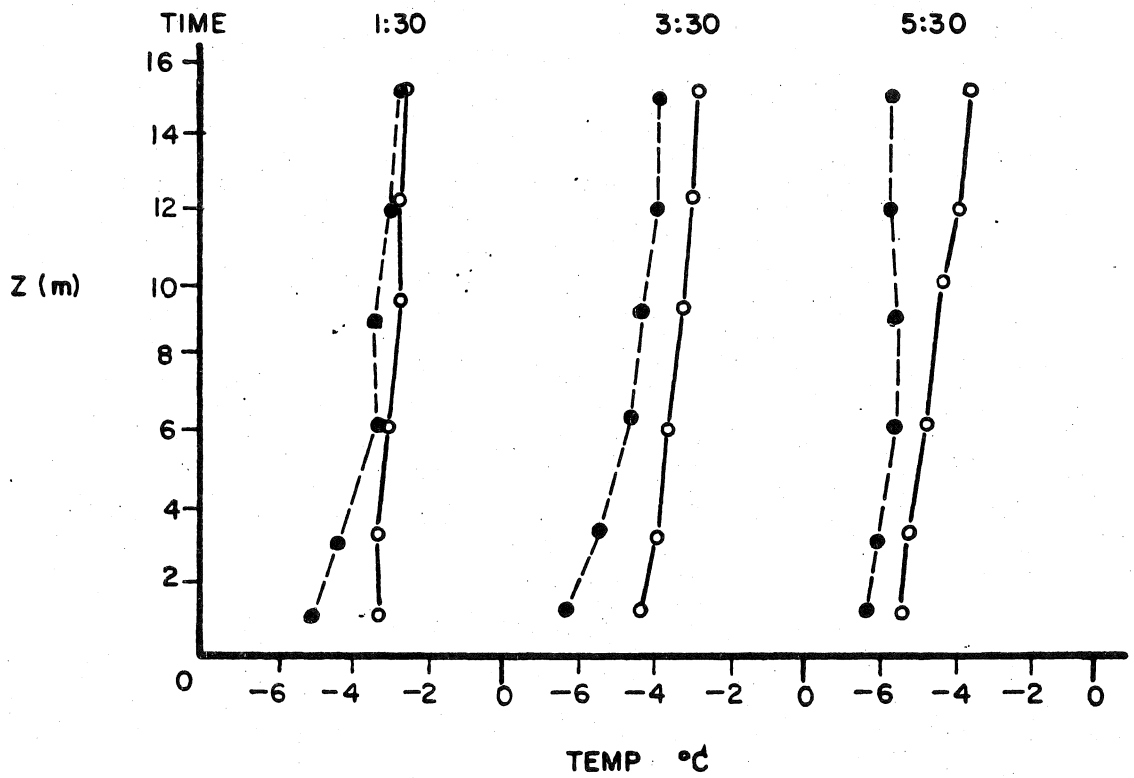

Fig. II-5 Profiles of mean air temperature from tower stations north and south of Lake Apopka. ;——o——south tower; ---●---north tower

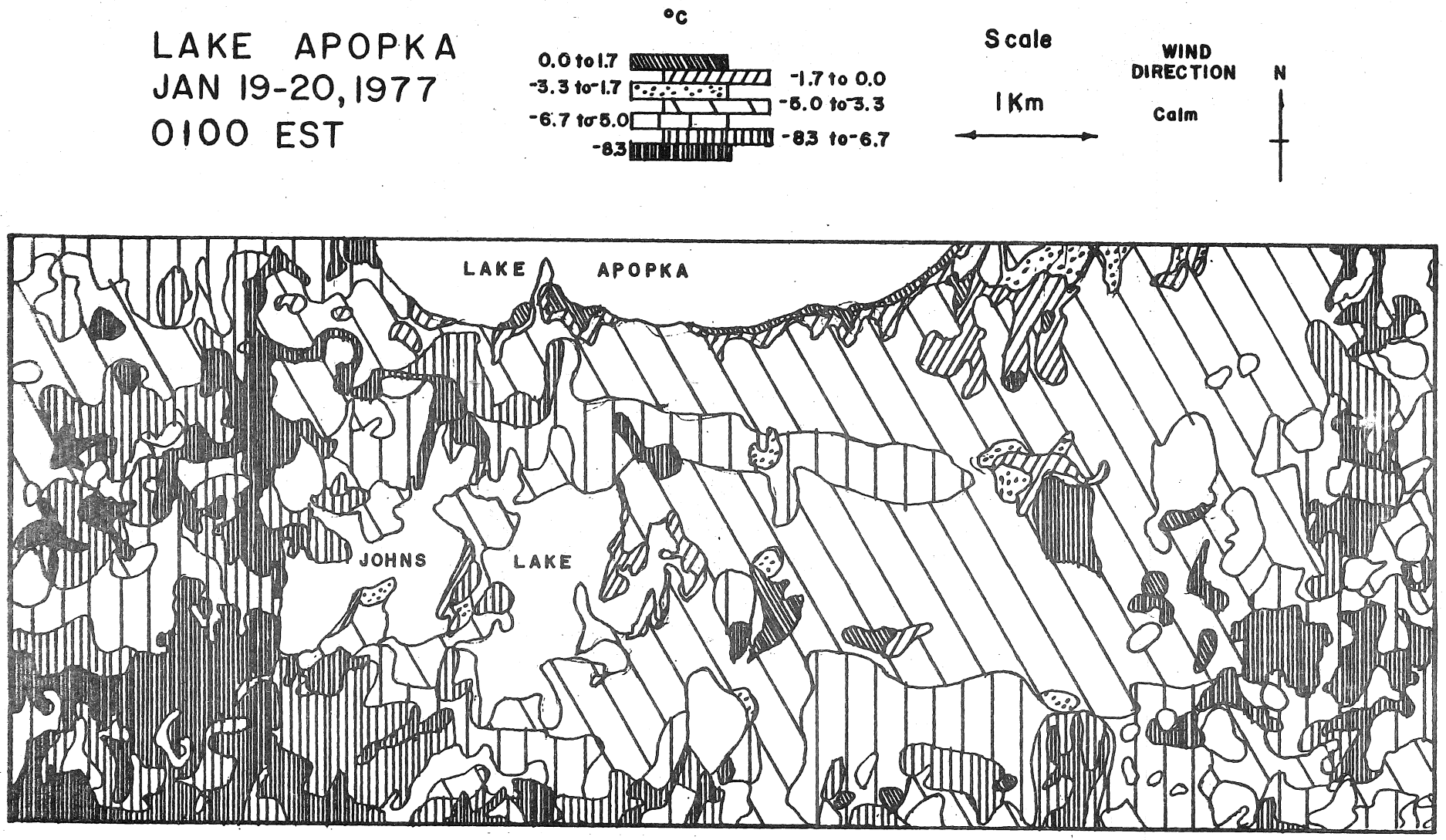

Figure II-6. Thermal Scanner Data for 0100, 20 January 1977.

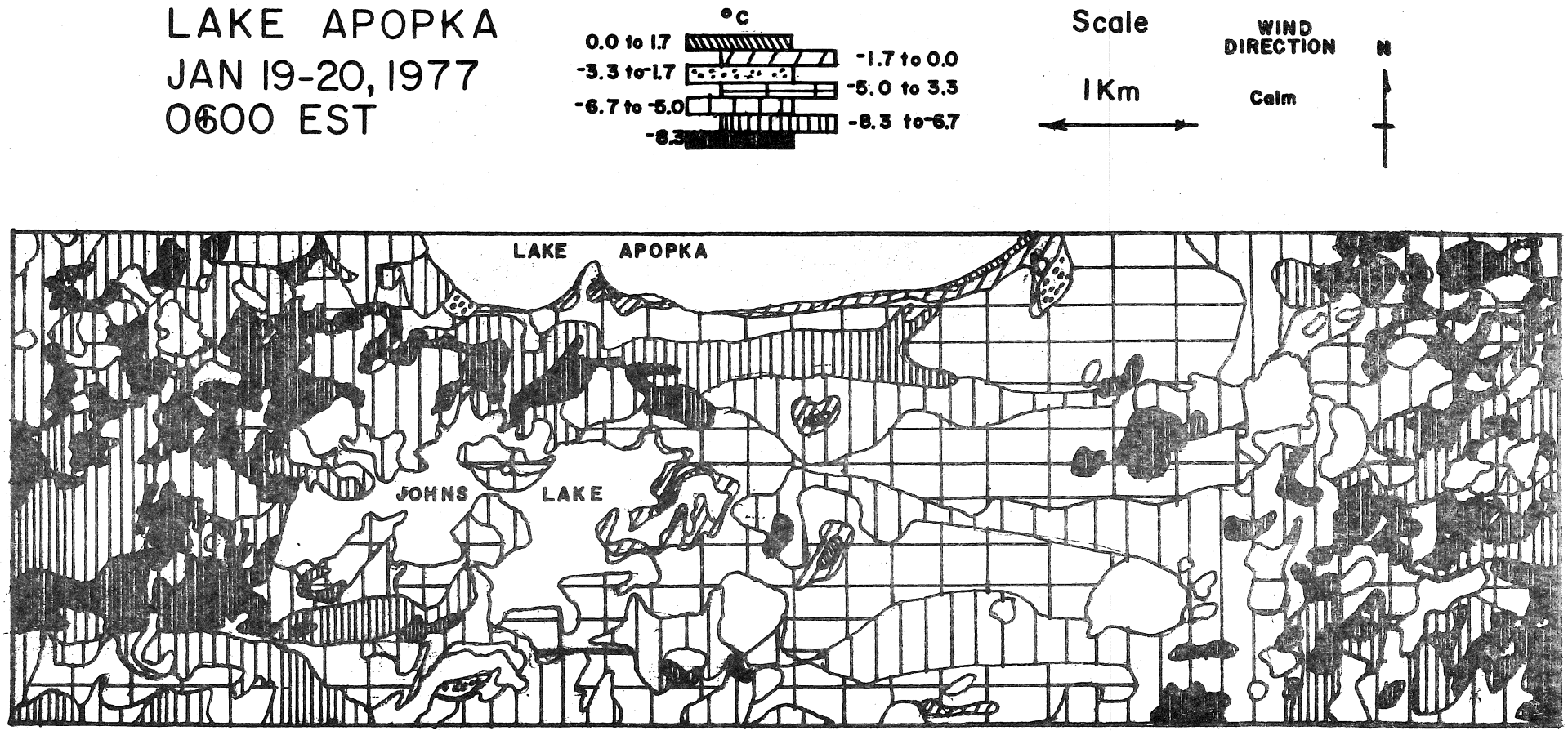

Figure II-7. Thermal Scanner Data for 0600, 20 January 1977.

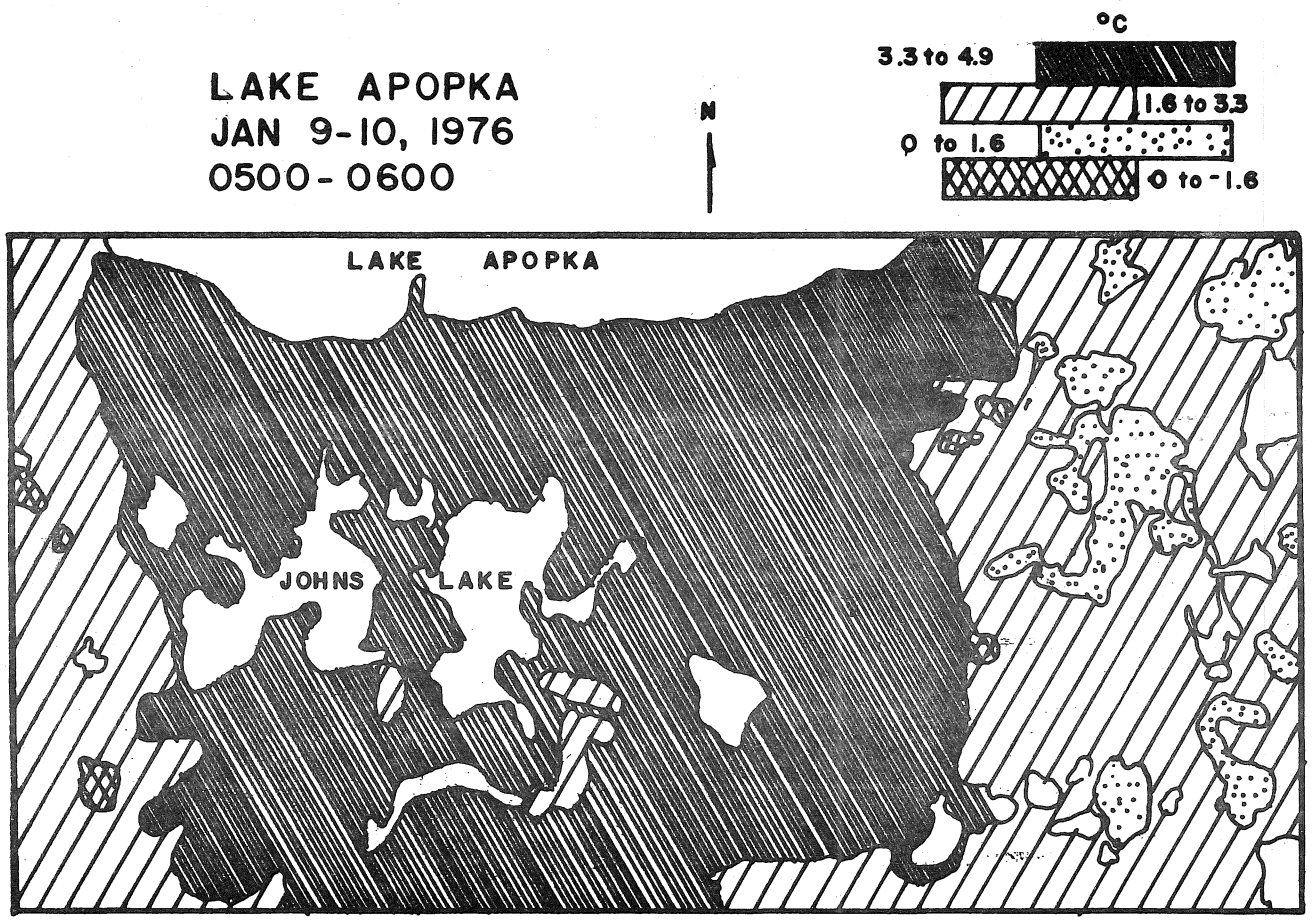

Figure II-8. Thermal Scanner
Data for 0500, 10 January 1976.

CHAPTER III

SIMULATING THE MODERATING EFFECT OF A LAKE ON DOWNWIND TEMPERATURES*

R.G. Bill, Jr., E. Chen, R.A. Sutherland, and J.F. Bartholic

Abstract

A steady-state, two dimensional numerical model is used to simulate air temperatures and humidity downwind of a lake. Thermal effects of the lake were modelled for the case of moderate and low surface winds under the cold air advective conditions that occur following the passage of a cold front. Surface temperatures were found to be in good agreement with reported observations. A comparison of model results with thermal imagery indicated the model successfully predicts the downwind distance for which thermal effects due to the lake are significant.

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Nomenclature

f, Coriolis parameter;

g, Acceleration of gravity;

K, Turbulent eddy coefficient;

M, Moisture availability parameter;

Q_{rad}, Net long wave absorbed radiation;

q, Specific humidity;

t, Time;

Ug, Geostrophic wind;

u, v, w, Horizontal, crosswind, and vertical velocity components;

x, y, z, Horizontal and vertical coordinates;

Greek Symbols

ζ, Vorticity;

θ, Potential temperature

1. Introduction

Numerical models have been used with increasing frequency to simulate complex microscale and mesoscale phenomena within the planetary boundary layer. One important parameter that has received little attention is the prediction of surface and lower air temperatures under advective conditions. The ability to model and predict such temperature patterns is of increasing importance to agriculture with increasing fuel costs for frost protection of crops. Detailed nocturnal temperature patterns downwind of a lake under cold air advective conditions were reported recently [Bill, et. al., 1978]. These patterns provided by a thermal scanner flown at 1.6 km, were complemented with other standard meteorological data. Such climatic observations, however, at best provide only a limited amount of information compared to the wide range of climate parameters possible. These considerations make development of realistic numerical models quite attractive.

A review of the literature indicates that the Gutman model (1974) used to study response of the urban boundary layer to heat addition and surface roughness incorporates the necessary transport mechanisms involved in the interaction between a warm lake and the passage of a cold front. Ideally, the model should be 3-dimensional and time dependent, such as that of Pielke (1974). However, the computational complexity of such a model makes a 2-dimensional steady state model more attractive. It is expected that steady state conditions of energy and momentum transport will be more rapidly approached at night when the energy budget consists of a balance of smaller fluxes, as compared with the daytime case. A complete annotated listing of the program and its results are given in Gutman (1974). Further discussions of the model are given by Gutman and Torrance (1975). Governing equations,

boundary conditions, and numerical procedures adapted for use here are briefly described below.

2. The Gutman Model

Gutman's model is a 2-dimensional, steady state finite difference procedure employing transition, constant flux and soil substrate layers to calculate the transport of momentum, heat and water vapor in the lower 1400 m of the planetary boundary layer. The conservation equations used in the transition layer are shown below:

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0 \qquad [1]$$

$$\frac{\partial \zeta y}{\partial t} + u \frac{\partial \zeta y}{\partial x} + w \frac{\partial \zeta y}{\partial z} = f \frac{\partial v}{\partial z} - \frac{g}{\Theta\infty} \frac{\partial \theta}{\partial x} + \frac{\partial^2}{\partial z^2} (K\zeta y) \qquad [2]$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + w \frac{\partial v}{\partial z} = -f(u - Ug) + \frac{\partial}{\partial z} \frac{(K\partial v)}{\partial z} \qquad [3]$$

$$\frac{\partial \theta}{\partial t} + u \frac{\partial \theta}{\partial x} + w \frac{\partial \theta}{\partial z} = \frac{\partial}{\partial z} \frac{(K\partial \theta)}{\partial z} + Q_{rad} \qquad [4]$$

$$\frac{\partial q}{\partial t} + u \frac{\partial q}{\partial x} + w \frac{\partial q}{\partial z} = \frac{\partial}{\partial z} \frac{(K\partial q)}{\partial z} \qquad [5]$$

(Note that conservation of momentum in the x direction is replaced by a conservation equation for the y component of vorticity, ζy. Net long wave radiation, Q_{rad}, is calculated by the method of Zdunkowski and Johnson (1965). The turbulent diffusion coefficient, K, assumed equal for momentum, heat and water vapor, is calculated as in Gutman, et. al. (1973). The stability-dependent form of K is similar to that of Blackadar (1962). As in Gutman (1974), only steady state solutions are of interest. The time dependent form is retained as a numerical iterative procedure to produce the steady state

result).

The model region and boundary conditions are shown schematically in Fig. III-1. The boundary conditions remain the same as those of Gutman and Torrance (1975) with the exception of the lake area where the lake is assumed to be a constant temperature surface. The lake temperature was observed to decrease at an average rate of 0.14°C per hour during the night simulated, 19 - 20 January 1977. A no-slip condition was imposed for computational simplicity instead of allowing for drift at the lake surface. Kraus (1972) has indicated surface drift is only approximately 5% of the 10-meter wind speed.

Specific humidity at the surface is set as a fraction of the saturation value associated with the temperature of the surface; i.e. $q = M \cdot q_{sat}$, with $0 \leq M \leq 1$. For example, $M \simeq 1$ with a freely transpiring crop surface. The upwind and downwind terrain to be modelled were a homogeneous dry organic soil surface and an extensive region of citrus groves, respectively. These conditions are representative of the northern and southern shores of Lake Apopka, Florida. Stomata of citrus are closed and evapotranspiration (ET) is negligible under nocturnal conditions. Hence, a value of $M = 0.5$ as indicated by Gutman (1974) was employed upwind and downwind of the lake. This value produced the desired effect of reducing ET to zero while maintaining reasonable humidity values. This point will be discussed further below. $M = 1.0$ on the lake surface.

Boundary conditions for specific humidity and temperature at the 1400 m level were taken from radiosonde data provided by the U.S. National Weather Service at Ruskin, Florida. A geostrophic wind condition was determined by trial and error to produce a wind velocity at the 5 m level in

approximate agreement with observations.

The upstream condition needed to implement the numerical procedure was calculated using the same series of equations [1 - 5], with the exclusion of derivatives in the downstream, x, direction. The kinematic condition of Gutman and Torrance (1975), $\partial^2/\partial x^2 = 0$ was imposed at the downstream end of the grid region.

Non-varying parameters used in this study are listed in TableIII-1Surface roughness parameters are from Chang (1968). Thermal diffusivities are from Van Wijk (1963).

A one second time stop was used in approaching a steady state solution. Such a small stop was needed to avoid numerical instabilities. Approximately 6000 iterations of the scheme were necessary before steady state was reached. See Gutman (1974) for discussion of stability of the numerical scheme.

3. Results

Gutman's model (1974) was employed using the physical parameters of Table III-1 for the case of moderate and low winds at the 5 m level. Computed 5 m winds on the leeward shore of the lake were respectively 3.94 and 1.1 m/sec for the 2 cases. These conditions were found by trial and error by imposing various wind speeds at the geostrophic level. The 2 wind conditions correspond to geostrophic winds of 10 m/sec and 4 m/sec.

In Fig. III-2, the computed 5 m air temperature is plotted as a function of the distance from the leeward shore of the lake. The upstream temperature at 5 m is computed to be $-3.6^{\circ}C$ for the moderate wind case. The temperature is $-4.4^{\circ}C$ in the low wind speed case. Hence, the maximum temperature increases at the 5 m height for the two cases are 2.1 and $1.1^{\circ}C$. The difference in

upstream air temperatures is due to the reduction of the turbulent eddy coefficient, K, with decreasing wind speed. K, at a height of 5 m, is 0.3 m^2/sec and 0.1 m^2/sec, respectively, above the organic soil surface for the 2 cases.

The 5 m air temperatures computed for a location 1.5 km from the leeward shore of the lake is compared with measured air temperatures south of Lake Apopka in Fig. III-2. Air temperatures at the 2 m level were measured along route 50 for 2 time periods, 0218 - 0303 and 0548 - 0611 EST on 20 January 1977 [Bill, et. al. (1978)]. The wind direction varied gradually on this night from the west at 0030 to the north at 0400. Wind speed was relatively constant during this period varying from 4 - 5 m/sec. Wind speed dropped below 1 m/sec after 0400 and was less than 0.4 m/sec at 0600. A temperature maximum is readily apparent in Fig. III-2 for the time period 0218 - 0303 along the axis of the lake as viewed from the wind direction. No regular temperature pattern is apparent, however, for the period 0548 - 0611. Citrus fruit damage occurs at approximately -2.2°C [Puffer and Turrel (1967)] with 95 percent tree damage occurring at temperatures as high as -4.4°C for some cultivars [Gerber and Hashemi (1965)], hence, the thermal protection afforded by the lake under moderate wind conditions is quite significant.

The location of the measured temperature maximum shown in Fig. III-3 is between 1.5 and 1.7 km from the shore of the lake as measured along the axis of the wind direction. The calculated temperature for the moderate wind case is within 0.1°C of the measured maximum. The computed temperature appears to be in reasonable agreement with measured temperatures in the low wind case, although the temperature pattern is more complicated.

Thermal imagery reported by Bill, et. al. (1978) indicated thermal

effects from the lake penetrated a distance of approximately 6 km under moderate wind conditions, while penetration of the plume was limited to approximately 1 km under low wind conditions. The computed temperature excess with respect to the upstream air temperature has vanished at a distance of 20 km and 3 km for the 2 wind conditions. The model predicts that air temperature drops off quite sharply at the edge of the lake and then asymptotically approaches the upstream temperature with increasing distance. The temperature excess drops to 10 percent of the maximum value at distances of 7 and 1.5 km from the edge of the lake for the 2 cases.

In order to make a more detailed comparison between model and surface radiative temperatures observed from the thermal scanner, the scanner signal was digitized from FM tapes of the data of 0100 and 0600, January 20, 1978. Due to the varying amounts of overlap between thermal scenes from successive flight paths of the aircraft and the resultant ambiguity in determining the precise position of the digital data in the thermal scene, only data from the first flight path was analyzed.

The spatial resolution of the data resulting from the digital conversion was approximately 10 m^2. Mean surface temperature in the central portion of the plume downwind of the lake were computed from blocks of data representing an area 1.0 x 0.25 km. The mean surface radiative temperatures are compared with results from the model in Fig. III-4. Agreement both in regard to trend and temperature level are quite good in the low wind case. Computed and measured temperatures agree within 0.4^{o}C. For the moderate wind case, the rate of decrease in measured temperatures with distance downwind of the lake is in good agreement with the model. The scanner data is systematically low, however, but only by 0.8^{o}C. This slight discrepancy may result from the use

of a steady state calculation to model conditions in which wind speed and direction changed with time.

The good agreement between measured and computed temperatures, as well as the favorable comparison with thermal imagery, indicates that a 2-dimensional model adequately predicts air temperature maxima downwind of a lake the size of Lake Apopka (approximately 13 km across). Moreover, a steady-state calculation appears adequate since the model successfully predicts the effect of a decrease in wind speed from 4 to 1 m/sec.

The ability of the model to compute air and surface temperatures must depend to a great extent upon the realism of the computed eddy coefficients at low levels. Computed coefficients at a height of 5 m for moderate and low wind conditions downwind of the lake varied between 1.2 to 0.57 m^2/sec and 0.32 to 0.2 m^2/sec, respectively. Diffusivities decreased monotonically with distance from the leeward shore of the lake in both cases. No data are presently available on such coefficients above citrus; however, Allen (1974) has measured eddy diffusivities at night above a maize canopy. At a height of 4.5 m, under low conditions (0.8 m/sec), the eddy diffusivity was approximately 0.1 m^2/sec. Diffusivities of the order of 1 m^2/sec were reported under unstable conditions when wind speed was 3.8 m/sec. Thus, the technique employed by Gutman (1974) appears to be in reasonable agreement with measured values.

Profiles from computed and measured air temperature directly upwind and downwind of the lake are shown in Fig. III-5. (Note grid locations used in the numerical model are indicated by x's). Air temperature measurements are from tower data reported by Bill, et. al. (1978). The upwind data is in reasonable agreement with the model profile and appears to indicate that the model pro-

vides an appropriate upwind temperature profile. As in the case of the surface data of Fig. III-4, the air temperature is systematically lower than the model temperature. Although the tower [as indicated by both the temperature data shown in Fig. III-3 and the thermal imagery reported in Bill, et. al. (1978)] was not in the central portion of the plume, the gradients in temperatures shown in Fig. III-3 suggest that a finer grid mesh in the vertical direction would improve the accuracy of the model. No other air temperature data is available for comparison.

Computed air temperatures are plotted in Fig. III-6 to a height of 1400 m for locations upstream and downstream of the lake for moderate and low wind speed conditions. The profiles are analagous to computed results reported by Gutman (1974) up and downstream of the urban "heat island". The temperature profile downwind of the lake converges with the upstream temperature profile at a height of approximately 200 m for the 2 cases. Profiles over the lake similarly converge at this height; hence, it may be inferred that the ceiling for the plume from the lake is approximately 200 m.

The increase in downwind air temperature seen in Fig. III-1 is associated with a similar increase in specific humidity due to the transport of water vapor from the lake. Specific humidity is plotted versus distance from the leeside of the lake in Fig. III-7. Upstream values for the 2 wind conditions were 1.43 and 1.31 g/kg. Patterns of decreasing specific humidity with increasing distance are quite consistent with the picture of transport revealed in Fig. III-3. The specific humidity excess vanishes at 20 and 3 km for the case of moderate and low wind speeds, as in the case of the 5 m air temperature.

Meteorological conditions similar to those of the 19 - 20 January existed on 17 February 1977. Bill, et. al. (1978) reported increases in

dewpoint of approximately $4^{o}C$ along Fla. route 50. Mean wind speeds were 6 m/sec. The increase in dewpoint corresponds to an increase in specific humidity of 0.96 g/kg for the conditions reported. This compares with a computed excess in the simulated case of 0.4 g/kg with wind speeds of 4 m/sec. Sufficient data necessary to run the model for the higher wind speed, however, are not available.

Results shown in Fig. III-7 are highly dependent on the choice of the moisture availability parameter. Choice of M = 0.5 produced the desired result of no evaporation or condensation downwind of the lake; i.e., computed gradients in specific humidity were positive and saturation values were not obtained. The limited data available suggest M = 0.5 is appropriate for the modelled conditions.

Conclusions

Computed air temperatures at the 5 m level are in good agreement with observed temperature maxima downwind of a lake and observed radiative surface temperatures in the central portion of the plume. The time constant of the transport processes associated with the meteorological conditions of this study are such that a steady state model may be employed for the cases of moderate and low surface winds. Furthermore, comparison of model results with measured data indicates that a 2-dimensional model may be successfully employed to predict air temperature maxima downwind of lakes the size of Lake Apopka.

Table III-1.

Parameters used in the study and height, z, at which they are applied.

Name	Symbol and Value	z(m)
Coriolis parameter	$f = 6.732 \times 10^{-5}$	
Soil Temperature	$T = 283.8^{o}$ K	-1
Lake Temperature	$T = 278.6^{o}$ K	0
Pressure	$P = 1.01 \times 10^{5}$ Nm^{-2}	0
Specific Humidity	$q = 0.35$ g/kg	1400
Upper Air Temperature	$\theta = 275^{o}$ K	1400
Thermal Diffusivity of Substrate		
Upwind	$Kg = 7.4 \times 10^{-7}$ m^2/s^{-1}	
Downwind	$= 1.2 \times 10^{-6}$	
Surface Roughness		
Upwind	$z_o = .01$ m	
Lake	$= .06$ m	
Downwind	$= .6$ m	

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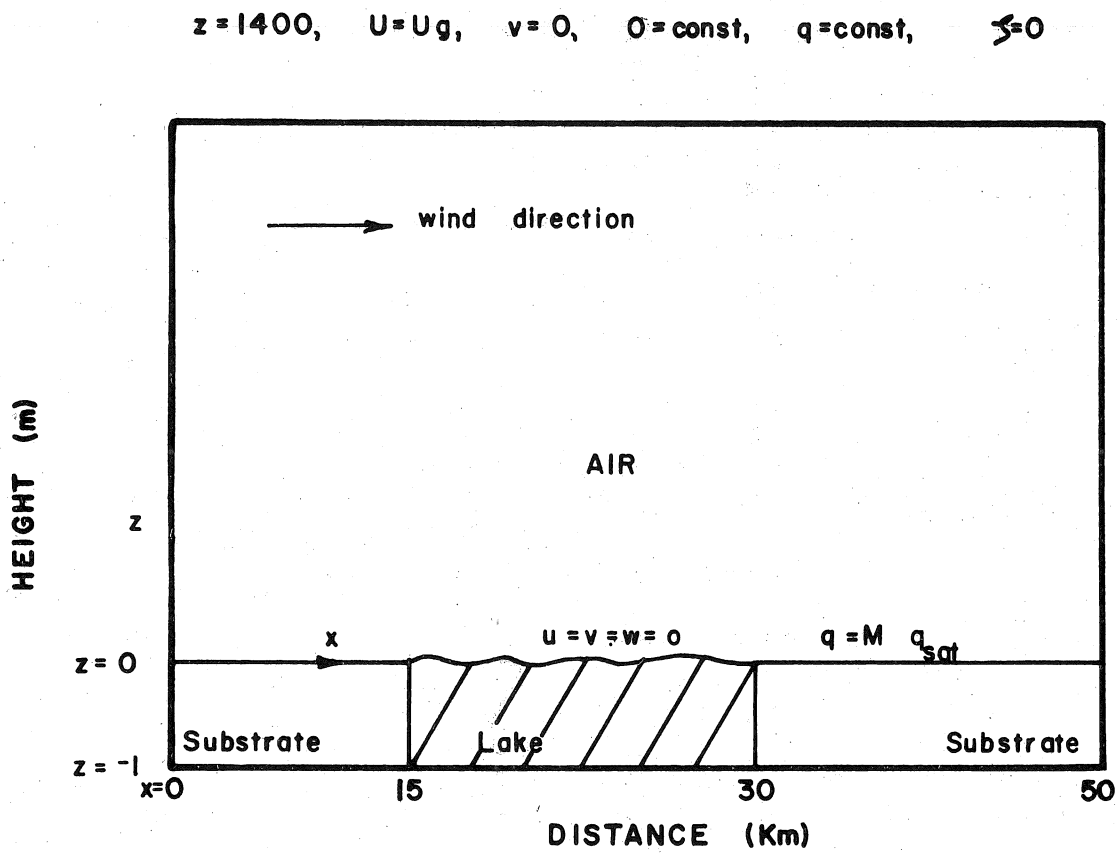

At $x = 0$, $\frac{\partial}{\partial x} = 0$

$w = 0$

Figure III-1. Schematic of computing region with boundary conditions.

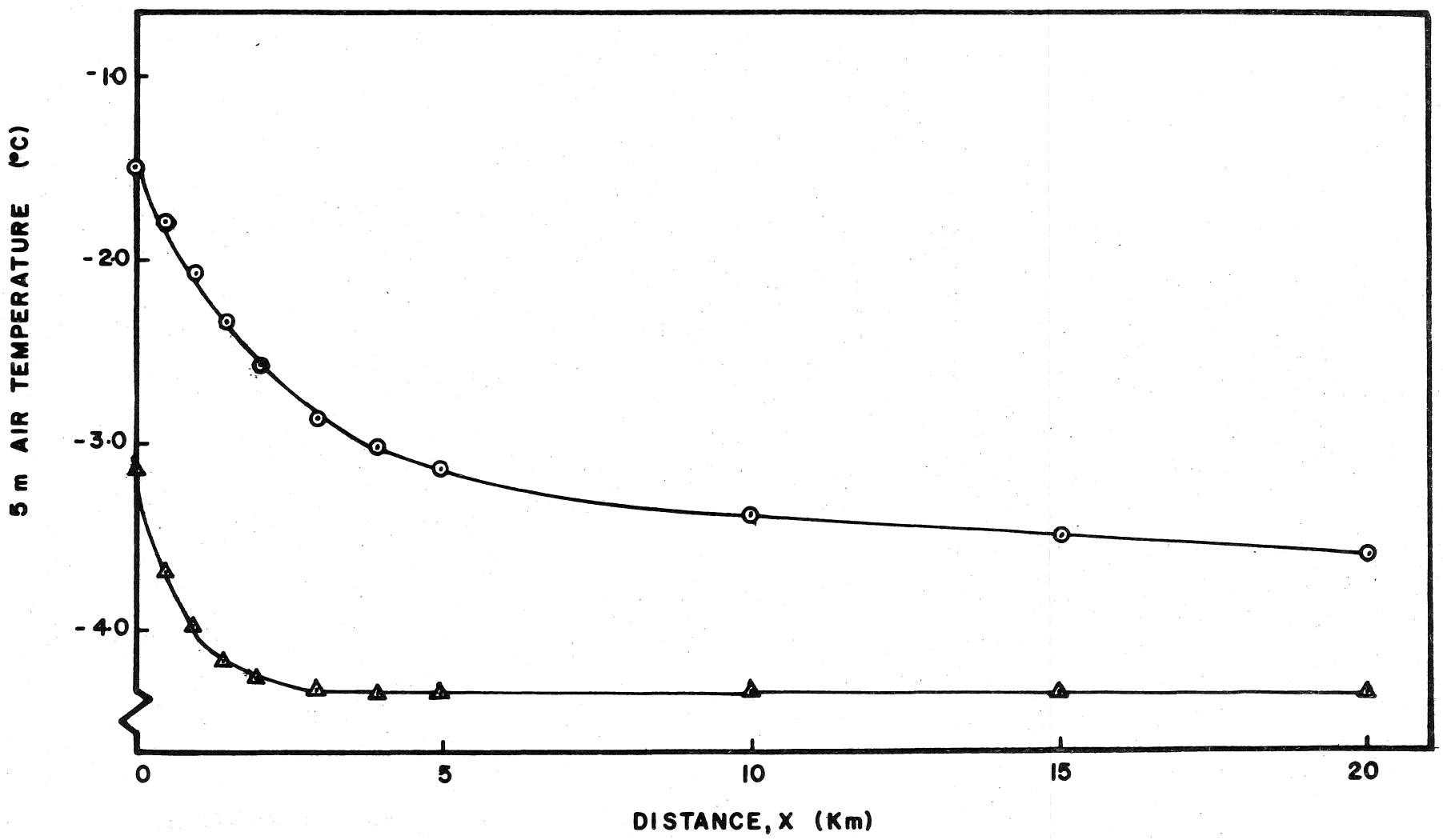

Figure III-2. Computed 5 m air temperature downwind of Lake Apopka: u = 4 m/sec, ⊙, u = 1 m/sec, Δ.

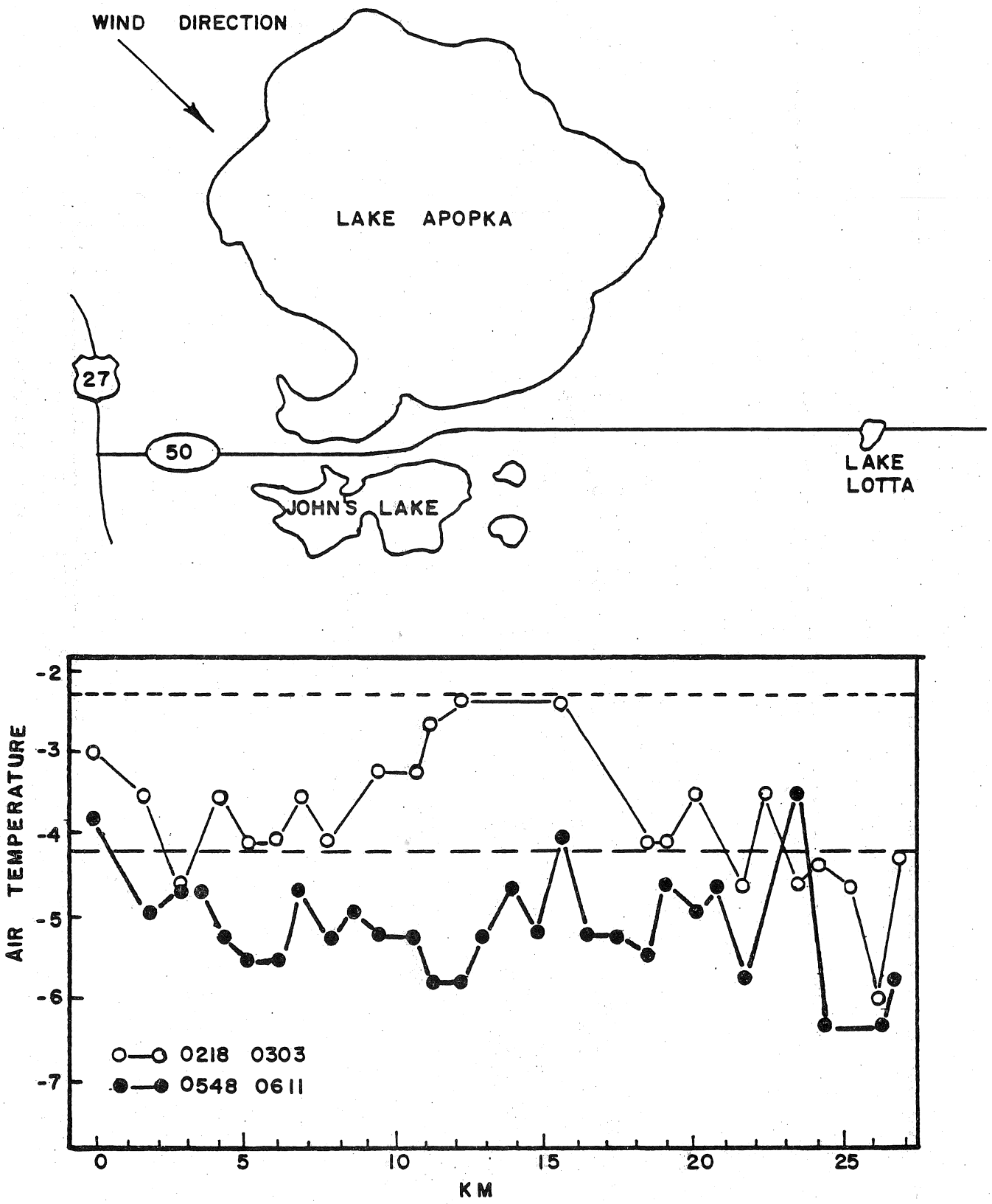

Figure III-3. Transects of a measured air temperature along route 50, 20 January 1977; computed temperature maximum, u=4m/sec, ----; u=1m/sec, — — —.

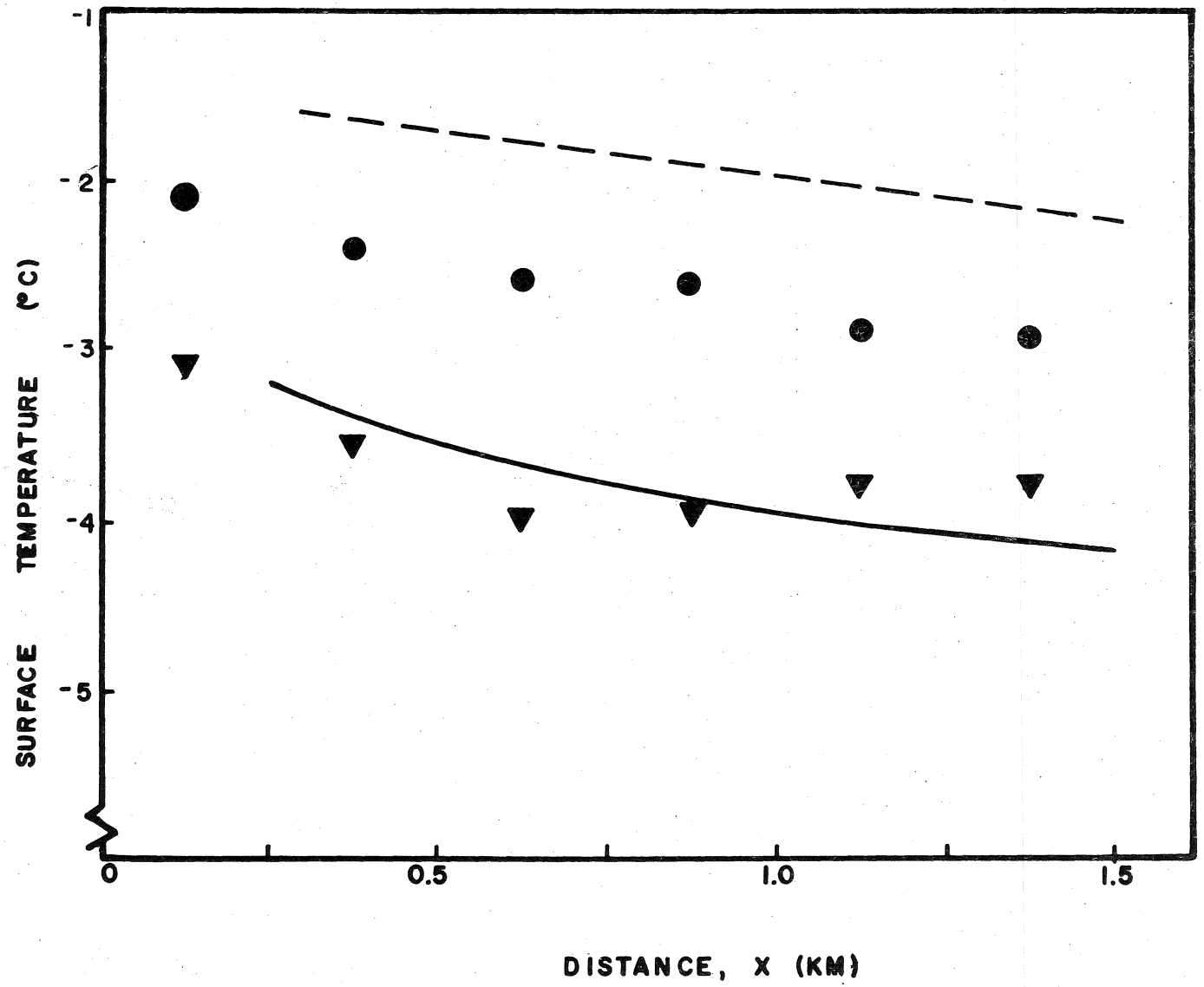

Figure III-4. Computed downwind surface temperatures vs. distance compared with mean surface radiative temperatures; u = 4 m/sec: model (----), data (●); u = 1 m/sec: model (____), data (▼).

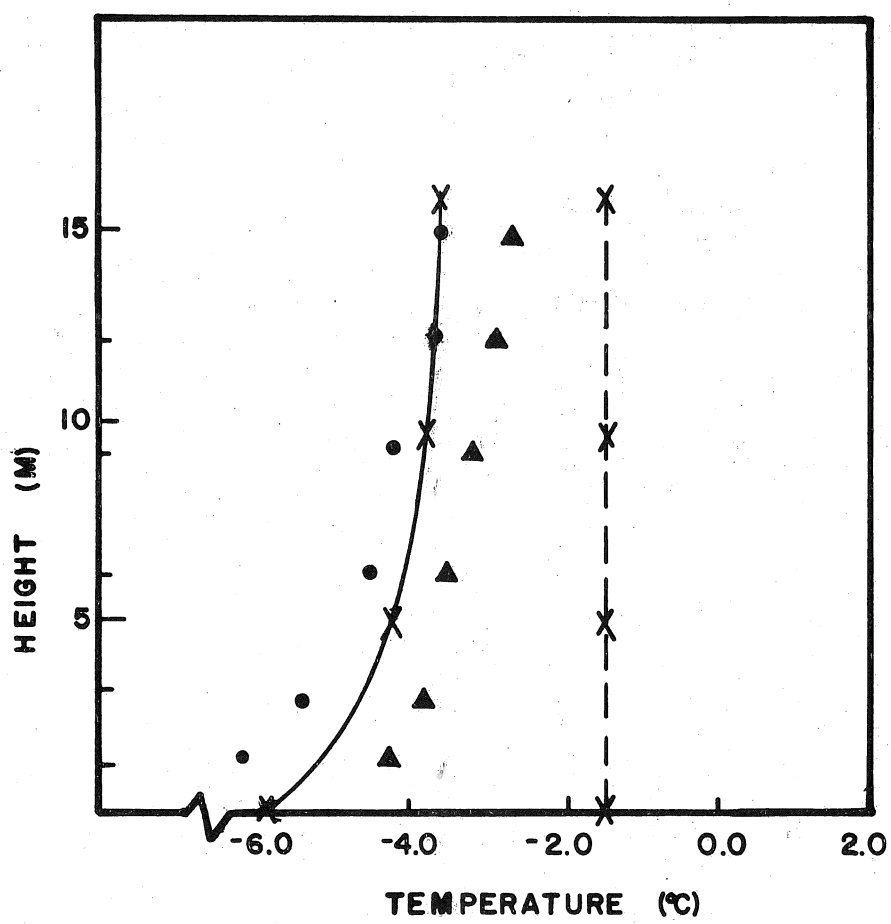

Figure III-5. Comparison of computed temperature profiles with tower data of Bill, et al. (1978); Upwind: model (--x--), data (▲); Downwind: model (—X—), data (●).

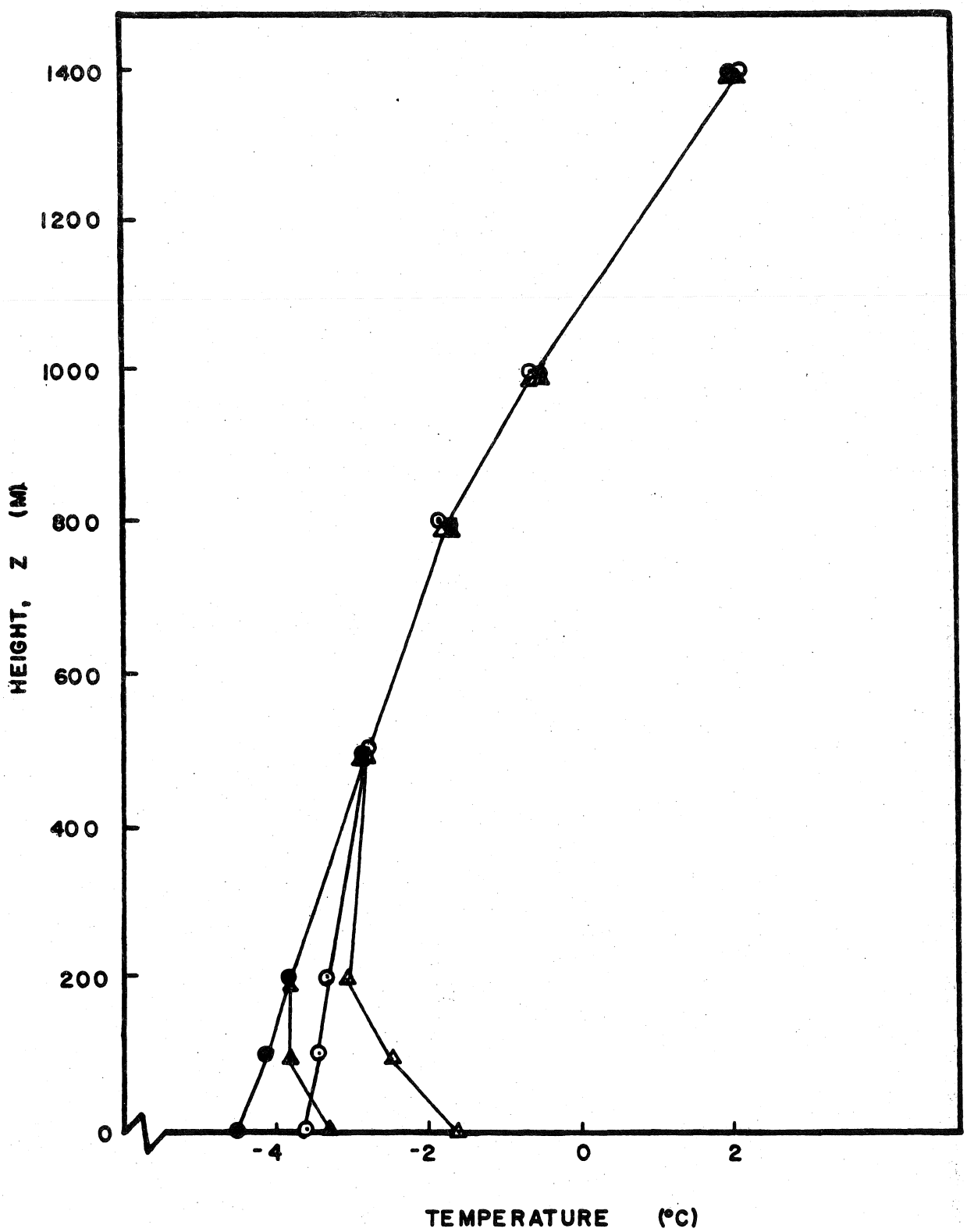

Figure III-6. Computed vertical temperature profiles;
u = 4 m/sec: upwind (⊙), downwind (▲);
u = 1 m/sec: upwind (●), downwind (▲).

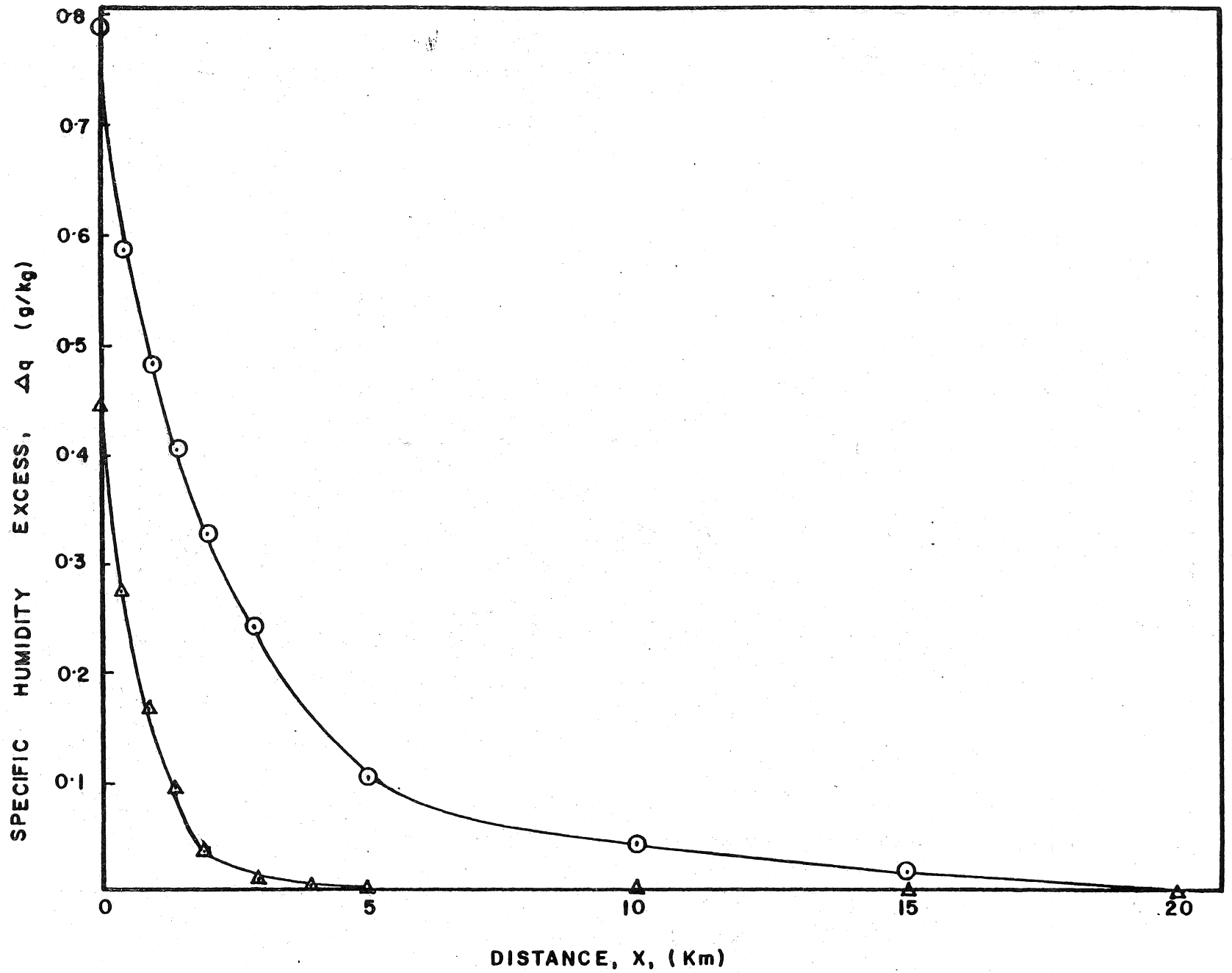

Figure III-7. Computed specific humidity excess downwind of Lake Apopka; u = 4 m/sec, O, u = 1m/sec, Δ.

CHAPTER IV

Predicting Fluxes of Latent and Sensible Heat of Lakes From Water Surface Temperatures

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and

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Abstract

Wind velocity, air temperature, and humidity fluctuations were measured over Orange Lake, Florida using an ultra-sonic anemometer-thermometer and Lyman-alpha humidity sensor. Latent and sensible heat fluxes were calculated from the data using the eddy-correlation technique. Lake temperature at a 2.5 cm depth and air temperature and wind speed at 10 m were measured in order to compare measured fluxes with predicted fluxes based upon reported bulk transfer coefficients and the formulation of Hicks (1975a) based solely upon water surface temperature. Measured fluxes were in good agreement with predictions based upon transfer coefficients. However, disagreement with predicted fluxes based solely upon water surface temperature indicate that such formulations are inadequate for short term flux evaluations under the meteorological conditions of this study.

1. Introduction

Considerable progress has been made in understanding the details of turbulent flow over bodies of water as the result of the correlation of flux measurements with aerodynamic similarity parameters [Brutsaert (1975), Merlivat (1978)]. Such relationships require detailed flow measurement of surface roughness and friction velocity. Due to the considerable scatter in data for the roughness parameter as a function of wind speed and atmospheric stability [see e.g. Kraus (1972)], such formulations are of limited value to hydrologists for the prediction of evaporation when only standard meteorological data such as wind speed, U, relative humidity, RH, and air, T_a, and lake temperatures, T_ℓ, are available.

Hicks (1975a) has suggested a long term relationship for sensible, H, and latent heat flux, LE, based solely upon water surface temperature. The supporting data, obtained from artificially heated cooling ponds, shows considerable scatter and no detailed discussion of the meteorological conditions of the study is given. The purpose of this study is to investigate the validity of the relationship of Hicks (1975a) for natural bodies of water. Latent and sensible heat fluxes were measured using the eddy-correlation technique over a period for which the lake, air temperature, and specific humidity were approximately constant. These conditions would appear to be favorable for the application of Hicks' formulation.

2. Basic Relations

The relationship of Hicks (1975) for sensible, H, and latent heat, LE, is given below:

$$H = 4\varepsilon\sigma T_\ell^3 B \;/\; (dB/dT_\ell) \qquad [1]$$

$$LE = 4\varepsilon\sigma T_\ell^3 \ / \ (dB/dT_\ell) \qquad [2]$$

Here ε is the water surface emissivity and σ is the Stefan-Boltzmann constant. In developing this relationship, the Bowen ratio, B = H/LE, is introduced and assumed also to be solely a function of the water surface temperature, T_ℓ (absolute). If the air above a water surface is saturated, then B may be expressed as the ratio of the psychrometric constant, γ, to the slope of the saturation specific humidity versus temperature curve S. Since the air over a water surface is normally unsaturated, Hicks (1975) modified the result by the ratio of the molecular diffusivity for heat, K, and water vapor, D.

$$B = \frac{K}{D} \left(\frac{\gamma}{S}\right) \qquad [3]$$

Priestley and Taylor (1972) also suggested that the Bowen ratio over a saturated surface is solely a function of surface temperature. Expressing their formulation in a manner similar to that of Hicks (1975), the Bowen ratio is:

$$B = 0.8 \left(\frac{\gamma}{S}\right) - .2 \qquad [4]$$

As in the case of [3], the temperature dependence of B is primarily in the S term.

With the temperature dependence of B prescribed, the formulae for H and LE, [1] and [2], may be evaluated for any given surface temperature. Hicks (1975a) evaluated flux and water temperature data from cooling ponds using both [3] and [4] for B. Scatter in the data precluded any judgment as to which expression for the Bowen ratio was most appropriate.

3. Description of Site and Instrumentation

A fixed data collection platform was installed on Orange Lake, Florida for eddy-correlation and meteorological measurements. Orange Lake, Florida is an "L" shaped lake in north-central Florida with legs approximately 6.5 km long and 3.5 km wide. At the time of the study, the average depth was approximately 3 m.

In Figure IV-1, the collection platform (2.4 x 5 m^2) and instrumentation are shown. The platform was located approximately 120 m from the west shore of the lake. During the period of data collection, the wind was from the northeast or east, providing a minimum upwind fetch of 3.5 km.

Wind speed was measured at the top of the 10 m tower (seen in Figure IV-1) with a RM Young* bivane anemometer. Air temperature at that level and at a water depth of 2.5 cm was measured with 24 gauge copper-constantan thermocouple connected to a multipoint recorder with ice-point reference junction. Eddy-correlation instrumentation extended 1.5 m in front of the platform at a height of 2.5 m. This equipment, seen in greater detail in Figure IV-2, consisted of a three dimensional ultra sonic-anemometer-thermometer (10 cm path, Kaijo Denko Co.) and a Lyman-alpha humidity sensor (Electromagnetic Research Corp.). The two horizontal wind sensing paths of the sonic anemometer cross at a 120° angle allowing wind variation of $\pm 45^{\circ}$ from center without interference effects from the wakes of transducers and supporting members [Mitsuta *et. al.* (1967)].

The Lyman-alpha sensor was calibrated by determining the response of

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*Mention of proprietary products is for the convenience of the reader and does not constitute endorsement or preferential treatment by the University of Florida.

the sensor to dry air (-57°C dewpoint) and mean humidity of ambient air. While the sonic anemometer was fixed and oriented approximately in the direction of the wind, the Lyman-alpha sensor was vane mounted to minimize the blockage of flow through the 1 cm path of the sensor.

Analog signals were processed by a Hewlett Packard 2100 computer with a 2313 Analog/Digital subsystem located on the shore of the lake. All signals were transmitted to the computer system through individually shielded wires.

Analog signals were digitized at a rate of 25 samples/second. The 3 cartesian components of the wind velocity were computed as in Mitsuta et. al. (1967). Fluctuations in water vapor density from the Lyman-alpha were computed as in Pond et. al. (1971). The Reynolds stress tensor and sensible and latent heat flux vectors were calculated over 8.7 minute time intervals. A coordinate transformation was then applied to rotate all tensor quantities and heat flux vectors to a coordinate system in which the vertical component of the wind, W, vanished and the horizontal wind velocity, U, was parallel to the mean wind direction [Pond et. al. (1971)]. A correction in the resultant sensible heat flux was made as in Smith (1974) since the sonic anemometer-thermometer senses the virtual temperature, not the dry-bulb temperature. The difference between the virtual and the dry-bulb temperature for the conditions of this study was 0.5°C or less.

Results

Meteorological conditions for the night of January 26-27, 1978 may be described briefly in terms of the half-hourly averages of the 10 m wind speed, 10 m air temperature and lake temperature (depth = 2.5 cm) shown in Figure IV-3. Early in the evening (2300-0000, EST), the mean wind speed was low (~2.5 m/sec)and air temperature decreased monotonically from 5.8 to 5.0^{0}C. At 0000, the mean wind speed began to increase and air temperature stabilized at approximately 4.4^{0}C. The wind speed was relatively steady from 0400 until 0530, averaging 9.7 m/sec. Mean wind speed then decreased reaching a value of 6.8 m/sec at 0630. This decrease was accompanied by a decrease in air temperature to 3.7^{0}C. During the time period 2300-0700 EST, the lake temperature decreased approximately linearly from 14.1^{0}C to 13^{0}C.

Eddy-Correlation Measurements

Flux densities, measured by the eddy-correlation instrumentation and the 10 m wind speed, U, are shown in Figure IV-4. The initial period on the time scale corresponds to 0437 EST. Data was collected beginning at 2300. However, data is reported here only for those cases in which there were no obvious errors in Reynolds stress values over 3 consecutive measurement periods. These cases correspond to observations in which the deviation in mean wind direction from the center axis of the sonic anemomoter was less than $\pm$ 10 degrees. The missing data point in the time series results from a voltage fluctuation causing a sampling error in the data acquisition system.

For the time period presented, the mean flux densities of latent and sensible heat (see Table I) were respectively 234 and 160 w/m^2. Standard deviations are respectively 42.8 and 21.7 w/m^2. During the first 60 minutes

when air temperature, specific humidity and wind speed were relatively constant, there did not appear to be a correlation between the small changes in wind speed and changes in the flux density levels. Note, during this time period, temperature as measured by the sonic thermometer and specific humidity measured by the Lyman-alpha sensor varied only from 4.2-4.4°C and 2.5-2.7 g/kg. However, after 60 minutes, when there was a consistant decline in wind speed, there did appear to be a correlation between decreasing mean wind speed and flux densities such as would be suggested by the bulk transfer coefficients, CD, CH, CE, that have been used to model respectively momentum, heat and vapor transport. These relations are shown below [see e.g. Francey and Garratt (1978)].

$$\tau/\rho = C_D U_{10}^2 \qquad [5]$$

$$H = cpC_H U_{10} \ (T_{10} - T_\ell) \qquad [6]$$

$$LE = \rho L \ C_E U_{10} \ (q_{10} - q_\ell) \qquad [7]$$

(Here, τ is the surface shear stress; ρ the air density; C_p, the specific heat capacity and L is the latent heat of vaporization).

Values of C_D, C_H, C_E, were computed and shown along statistics for the flux densities of sensible and latent heat. Since, the specific humidity was not available at the 10m height, the mean value measured by the Lyman-alpha was used in equation 7. There is considerable scatter in reported transfer coefficients. Frances and Garratt (1978) have recently reviewed these results and from their own eddy correlation measurements have reported $C_D = 1.7 \pm .5$, $C_H = 1.4 \pm .5$, $C_E = 1.8 \pm 0.3$. Both C_D and C_H are in good agreement with values reported here. Our results, however, indicate that $C_H \approx C_E$. However, the Lyman-alpha sensor is more appropriate for measuring fluctuation in humidity.

The flux densities predicted by Hicks (1975a) from equations 1 and 2 are also shown in Figure IV-4. At high wind speeds, Hicks (1975a) underestimates both H and LE. At the lower wind speeds, agreement is more reasonable.

Hicks (1975a) has suggested that the simple relationships of equations 1 and 2 may only be applicable over long time periods. Figure 3 suggests that these relationships may be interpreted as being valid for some average wind conditions in which conditions of air temperature and specific humidity are approximately constant.

In Figure IV-5, Bowen ratios (B=H/LE) are calculated and plotted from eddy correlation data. The mean and standard deviation are shown in Table I. For comparison, equation 3 is shown for the conditions of the study along with the expected result for saturated conditions; i.e. $B=\alpha/s$. The result of Priestley-Taylor (1972), equation 4, appears to be totally invalid for the conditions of this study since results of equations 3 and 4 differ by approximately 0.2. The data is scattered along the result, $B=\alpha/s$. Considering the dry conditions that were experienced, this result suggests that it is improper to evaluate thermal properties in equation 3 using water surface temperature since the temperature gradient is quite large. A common engineering practice for many heat transfer correlations is to evaluate properties at a "film temperature"; T_f i.e., a temperature intermediate between a surface temperature and the bulk fluid temperature [see e.g. Gebhart (1971)]. Results of this study and equation 3 are in agreement if $T_f = T_\ell - .3(T_\ell - T_{10})$.

Conclusions

Eddy correlation results indicate that the relationship of Hicks (1975a) is inadequate for evaluation of flux densities over water on a short time basis. The result may, however, be adequate for some wind conditions. The Bowen ratio formulation of Hicks (1975a) is valid if properties are evaluated at a "film temperature" intermediate to the water surface and 10 m air temperature.

Table I

SUMMARY OF EDDY CORRELATION RESULTS

	MEAN	STANDARD DEVIATION
H	- 160 w/m	43.0
LE	- 234 w/m	22.0
C_D	- 1.75×10^{-3}	0.26×10^{-3}
C_H	- 1.17×10^{-3}	0.13×10^{-3}
C_E	- 1.19×10^{-3}	0.13×10^{-3}
B	= H/LE .677	.0306

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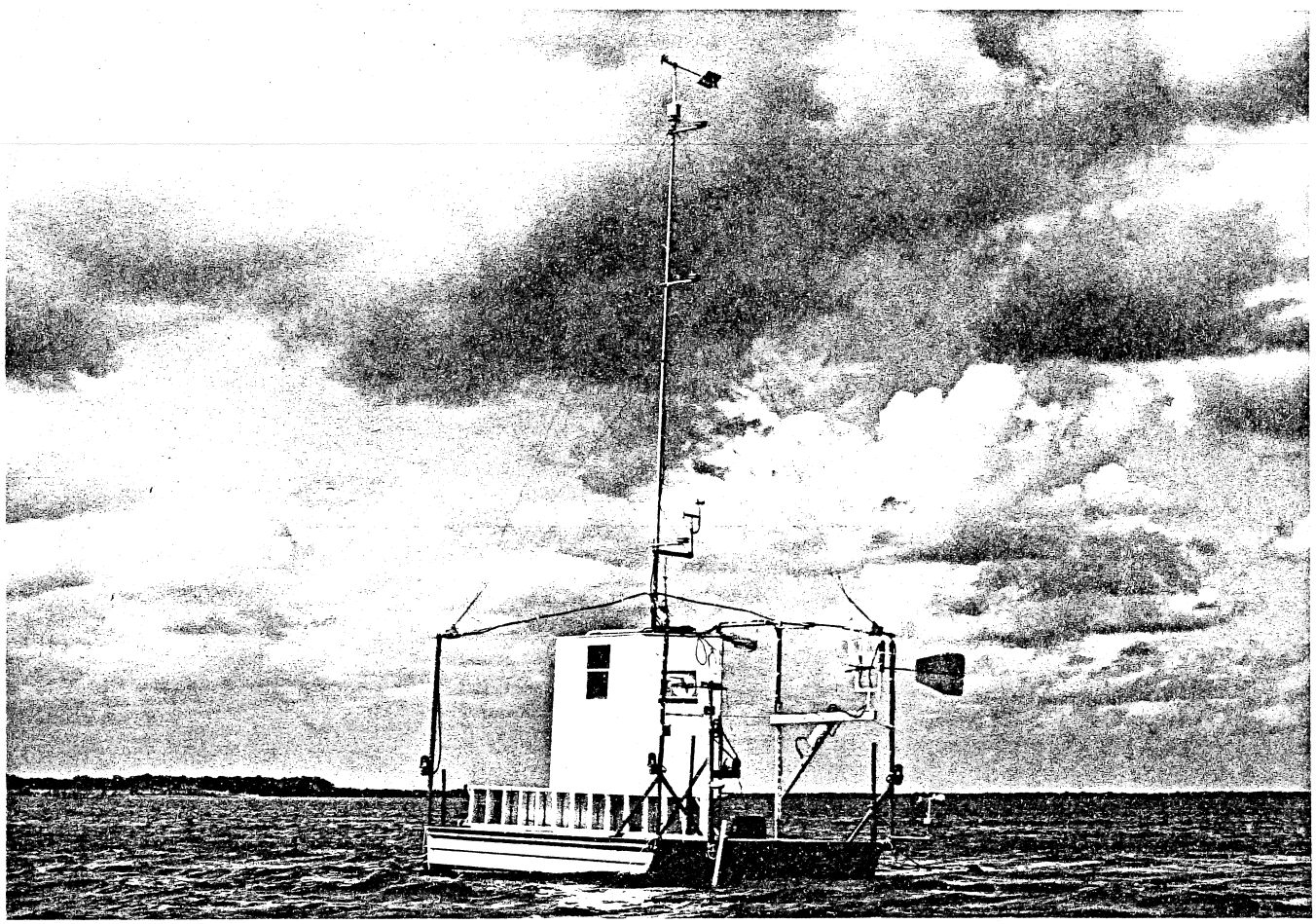

Figure IV-1. Instrumentation Platform on Orange Lake.

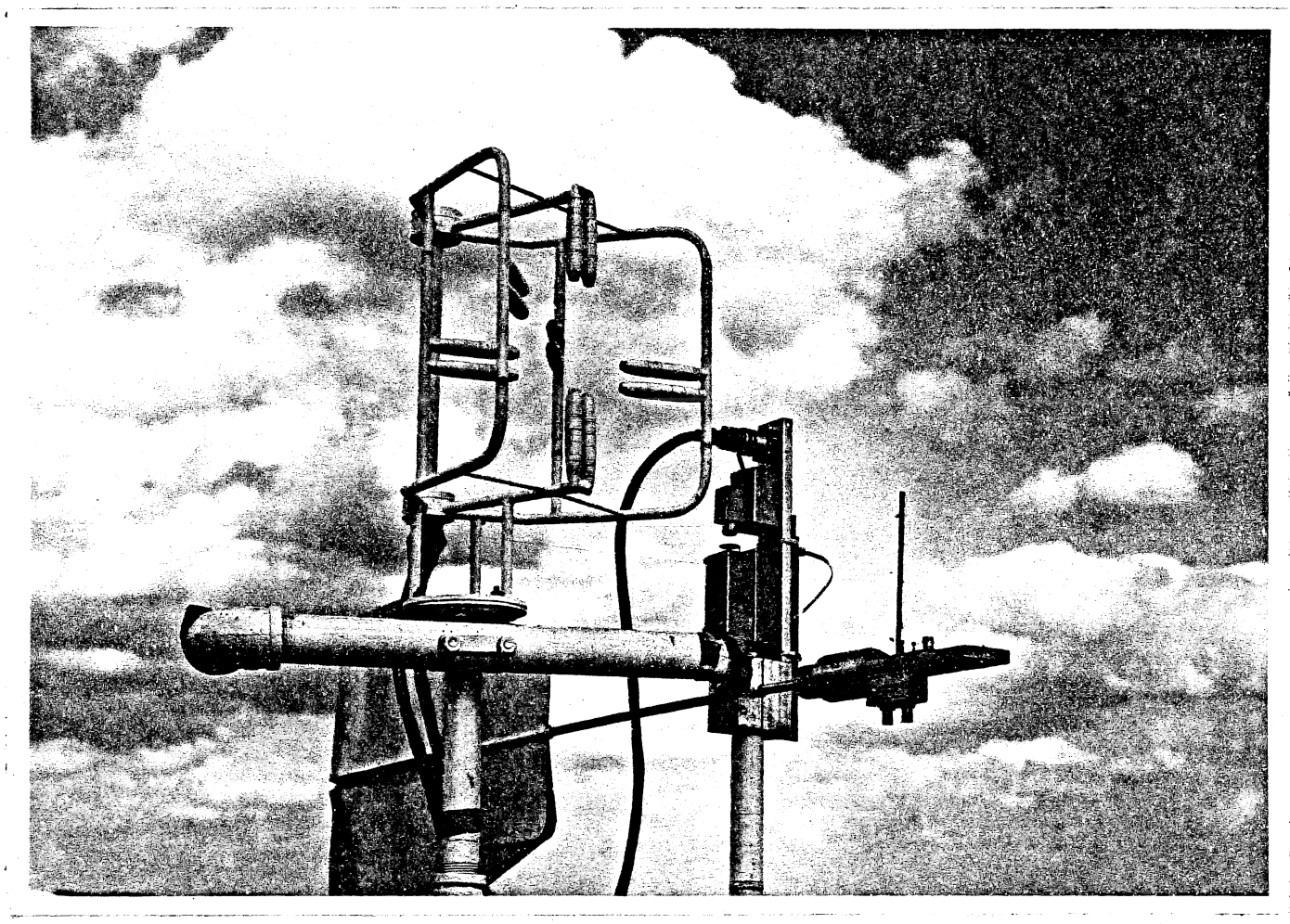

Figure IV-2. Eddy correlation instrumentation.

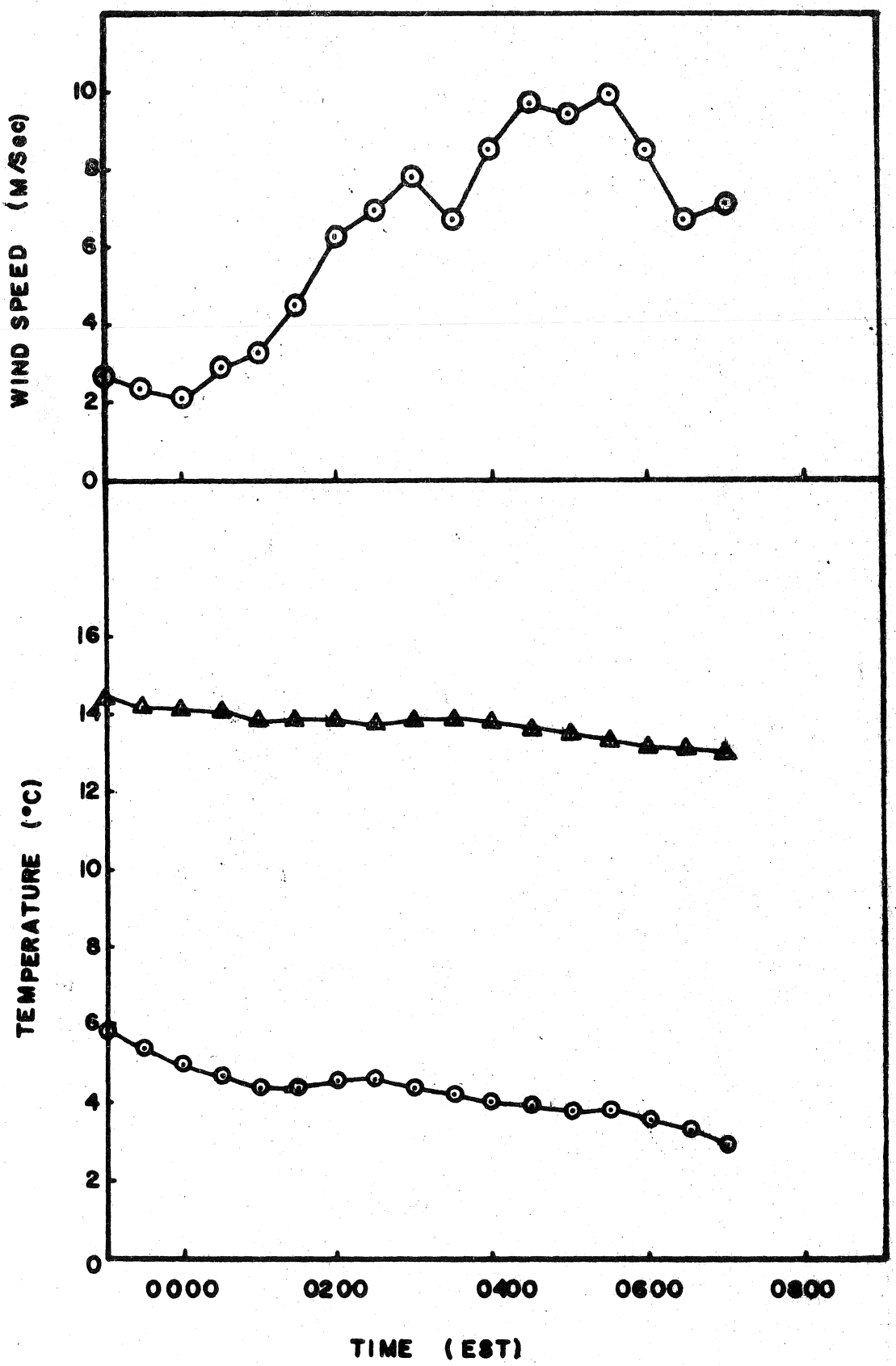

Figure IV-3. Meteorological data for January 25-26, 1978 on Orange Lake; 10m windspeed Θ, lake temperature Δ, 10m air temperature.

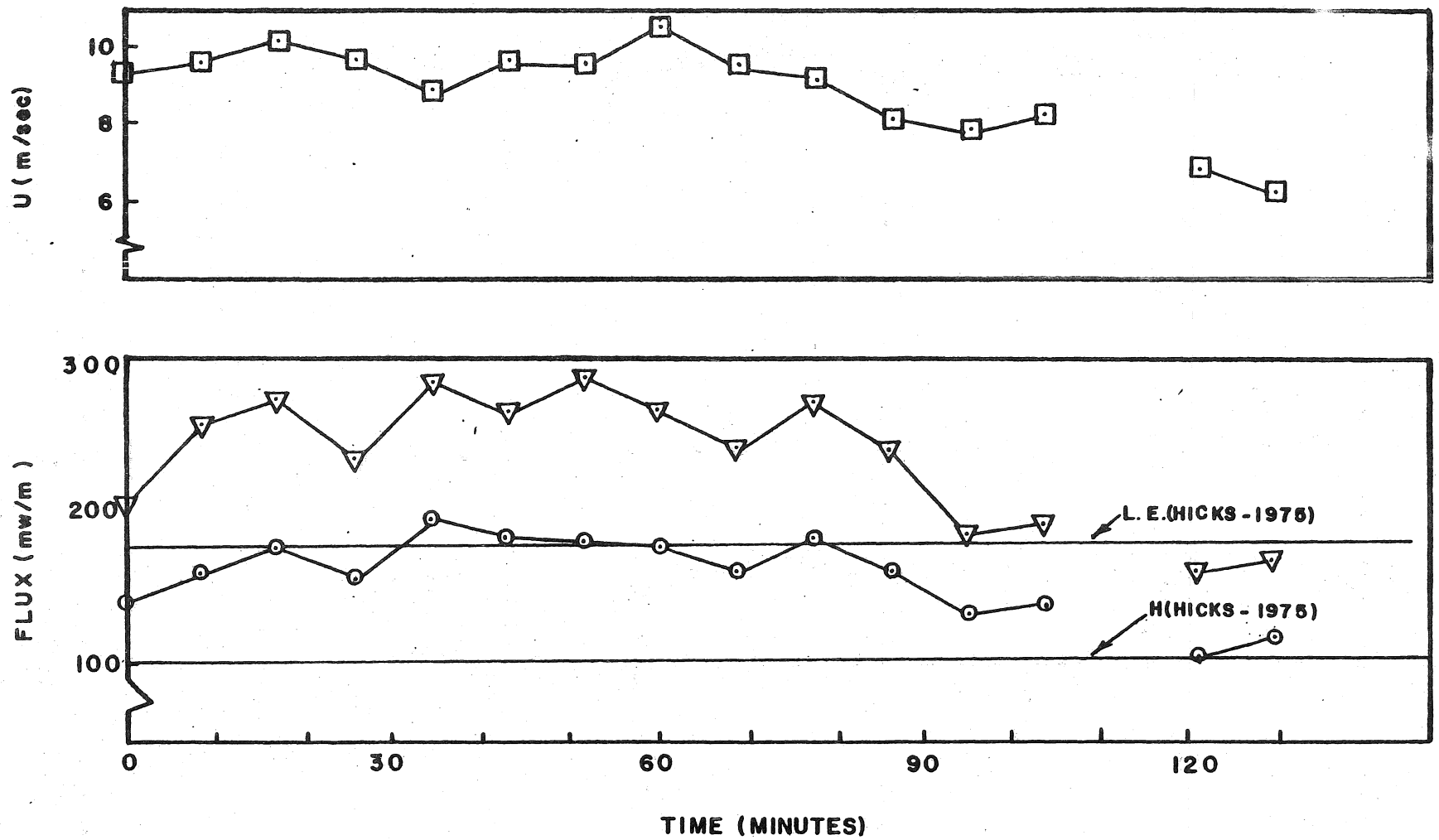

Figure IV-4. Comparison of flux densities of latent heat, ∇, and sensible heat, O, with the 10m wind speed.

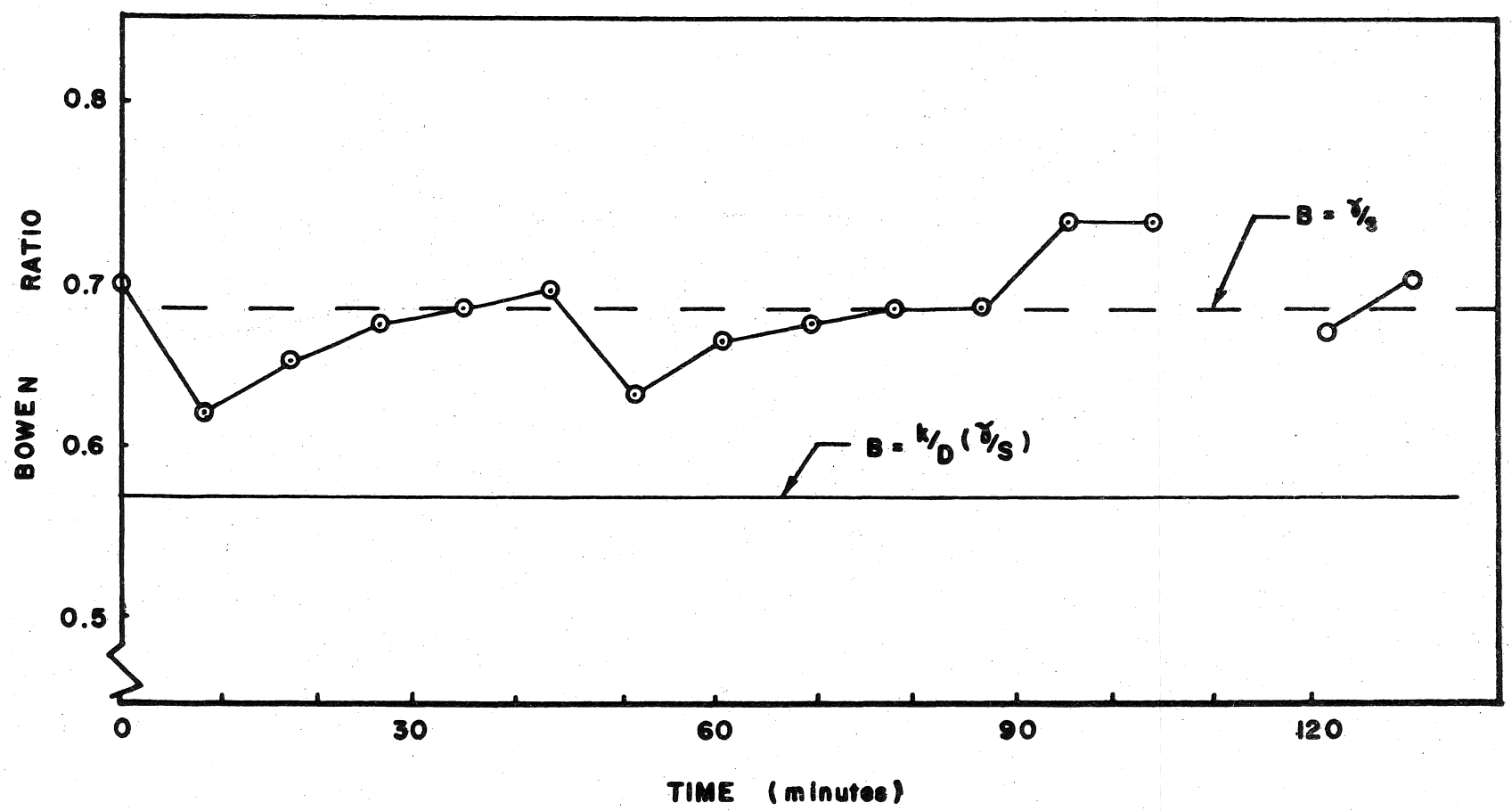

Figure IV-5. Bowen ratio measurement from eddy correlation results.

CHAPTER V

DIURNAL VARIATION OF LATENT HEAT FLUX ON EAST LAKE TOHOPEKALIGA

R.G. Bill, Jr.

Abstract

Dewpoint, wind speed, and lake temperature were measured on East Tohopekaliga, Florida during 24 hours on April 11-12, 1978. Latent heat flux densities were calculated using bulk transfer coefficients of Francey and Garratt (1978). Total flux for a 24 hour period was in agreement with a wind speed corrected formulation of Hicks (1975).

1. Introduction

Bulk transfer coefficients have been used with increasing frequency to calculate latent heat flux densities from bodies of water. For example, Francey and Garratt (1978) calculate the latent flux density as:

$$LE = \rho L U_{10} \; C_E (q_\ell - q_{10}) \qquad [1]$$

Here ρ is the air density; L, the latent heat of vaporization; U_{10} the mean wind speed at a height of 10m; q, the specific humidity at either the lake surface (subscript ℓ) or at a 10m height (subscript 10) and C_E is the bulk transfer coefficient for the transport of water vapor.

On April 11-12, 1978 data necessary to calculate latent heat flux densities from East Lake Tohopekaliga were obtained. East Lake Tohopekaliga, seen in Figure 1, is a circular lake in east-central Florida approximately 7.5 km in diameter. Elevations shown in Figure 1 are from the U.S. Geological Survey with a shoreline elevation of 56'. The approximate location for the observations of this study is also shown in Figure 1.

The instrumented boat, used for data collection, is shown in Figure 2. Wind speed and direction were measured at a height of 3m above the water surface. Air from this height was pumped to a dewpoint hygrometer to provide a measurement of the specific humidity of the air. Water temperature was measured at depths of oil, 0.3, 1.0, 2.0, 3.0m and 0.3m in the lake bottom with copper-constantan thermocouples connected to a multipoint recorder with an ice-point reference junction.

Results and Conclusions

Wind speeds for the 24 hour period beginning at 1530, April 11, 1978 are shown in Figure 3. Throughout this time period, the wind speed was moderate or low. The mean wind speed was 3.1 m/sec, with a maximum and minimum of 6.1 and 1.2 m/sec. During this period, wind was primarily from the south or southwest.

Water temperature profiles in the lake are shown in Figure 4. Throughout the 24 hour period substantial temperature gradients occurred in the lake, except at night between 0330 and 0630. (We note parenthetically, that such large gradients make measurement of changes in the internal energy of a lake subject to significant error. This makes the use of the Bowen ratio technique (Chapter II) impractical.) The mean water temperature at a 10cm depth was 26.1°C.

Dewpoint was relatively constant during the observation period. A minimum value of 15.6°C occurred at 1530. The maximum value of 20.6°C was maintained between 2000 and 0730. The minimum air temperature, 20.8°C, was reached at 0530.

From wind speed, dewpoint, and lake temperature at a depth of 10cm, latent heat flux densities were calculated using equation 1 with a transfer coefficienty $C_E = 1.8 \times 10^{-3}$, [Francey and Garratt (1978)]. These results are shown in Figure 5. Maximum and minimum flux

densities are 152 and 31.4 w/m^2. The average flux density for the total 24 hour period is 86.3 w/m^2.

Hicks (1975) has suggested a relationship for predicting evaporation rates based solely on the lake surface temperature (Chapter IV). For the mean lake temperature, Hicks relationship predicts the flux density to be 313 w/m^2 or a value 3.63 times higher than the mean value as calculated using transfer coefficients.

In Chapter IV, eddy correlation results, indicate that short term flux-densities are strongly affected by the wind speed. For example in the first 60 minutes of Figure IV-4 when the mean wind speed was 9.66 m/sec the mean flux density was 257 w/m^2. Later when the mean wind speed decreased to 7.3 m/sec, the flux density was 176.4 w/m^2. For the lower wind speed, the difference between the measured and predicted flux is less than 1%. This suggests the following wind correlation to the relationship of Hicks.

$$LE = LE_{Hicks} \left(\frac{U}{U_0}\right)^n \qquad [2]$$

where U_0 = 7.3 m/sec.

From the above data, n = 1.36. (Obviously, this result may not be extended to U = 0, since transport would still occur by free convection).

If the mean wind speed, and LE_{Hicks} based upon the 24 hour mean lake temperature are substituted into equation 2, then the predicted flux density is 96.8 w/m^2. The difference between the corrected prediction [2] and the calculated mean flux density is less than 12%. While equation [2] must be tested over a much greater range of climatic conditions, results from this chapter and Chapter IV indicate [2] is valid for $1 < U < 10$ m/sec.

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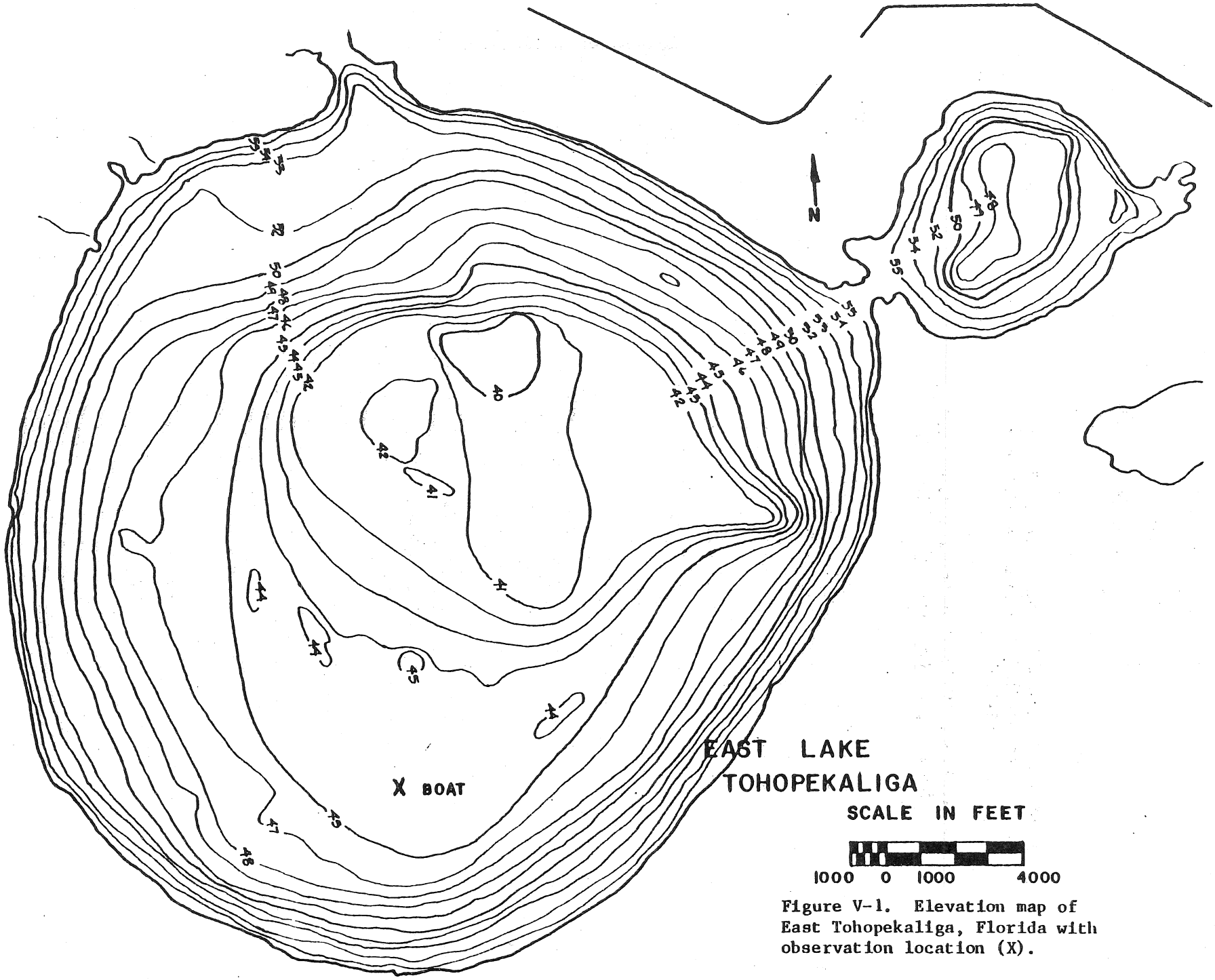

Figure V-1. Elevation map of East Tohopekaliga, Florida with observation location (X).

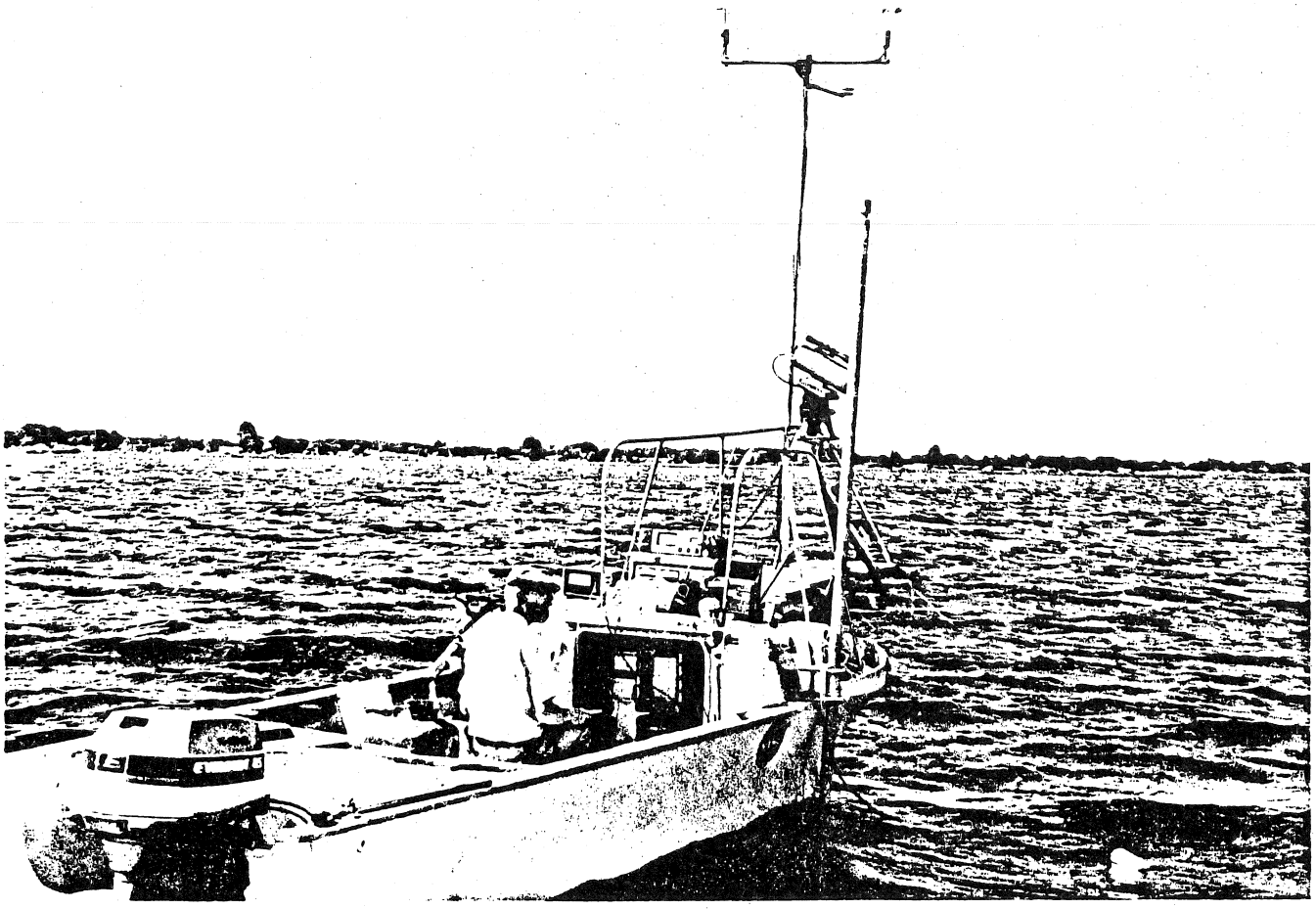

Figure V-2. Instrumented boat for meteorological data collection.

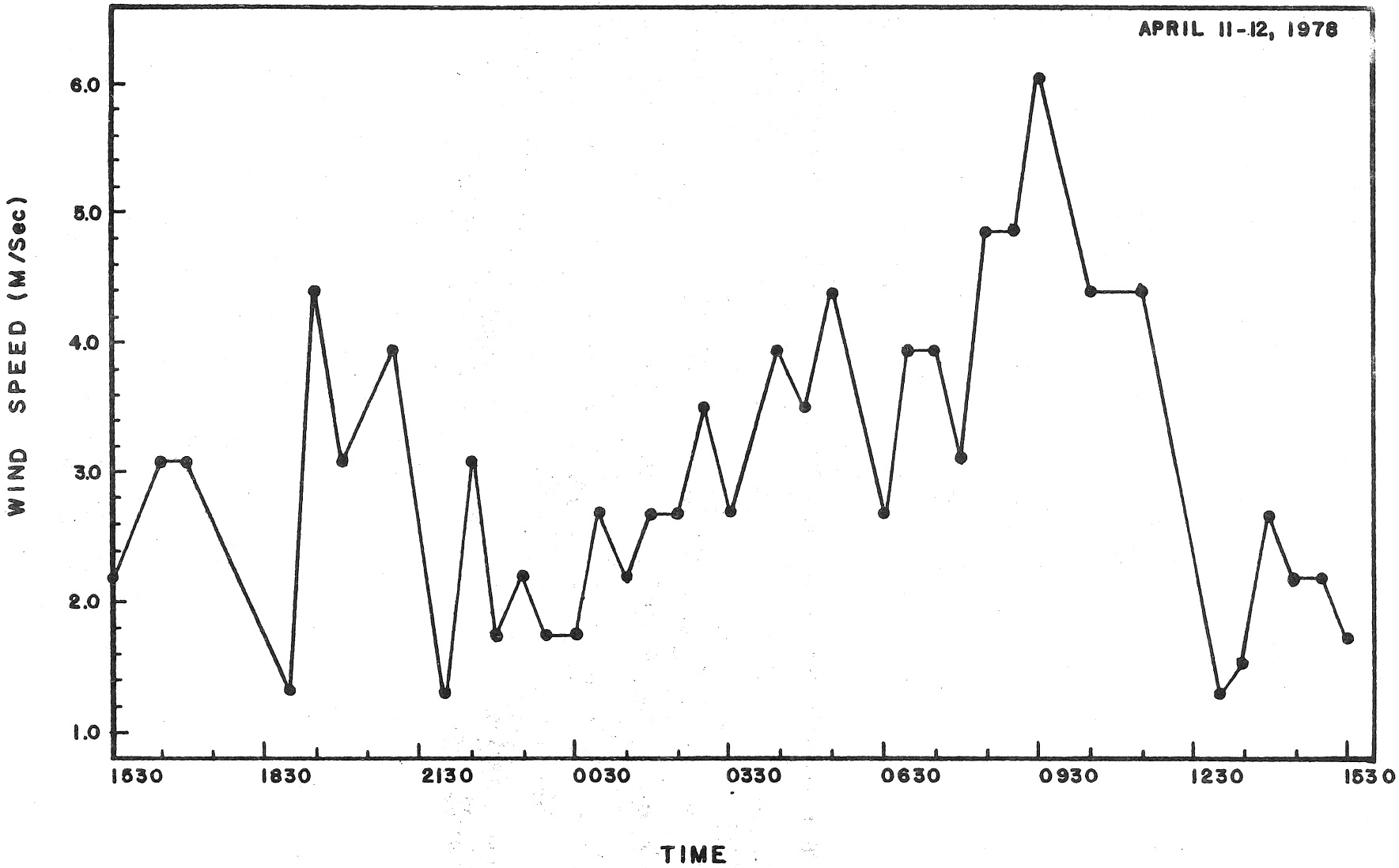

Figure V-3. Wind speed on East Lake Tohopokaliga, April 11-12, 1976.

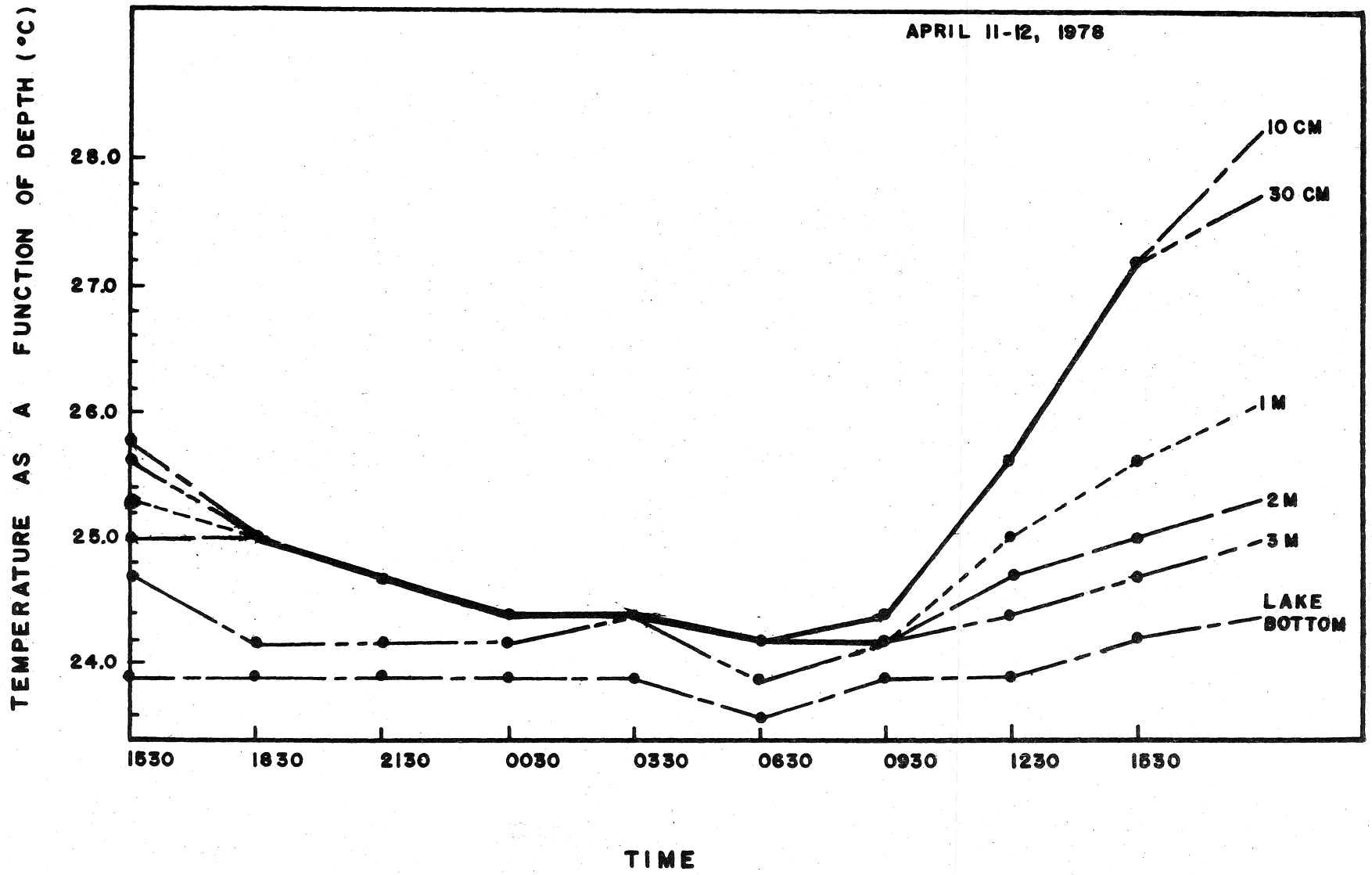

Figure V-4. Lake temperature at .10, .3, 1.0, 2.0, 3.0 m depth and lake bottom temperature.

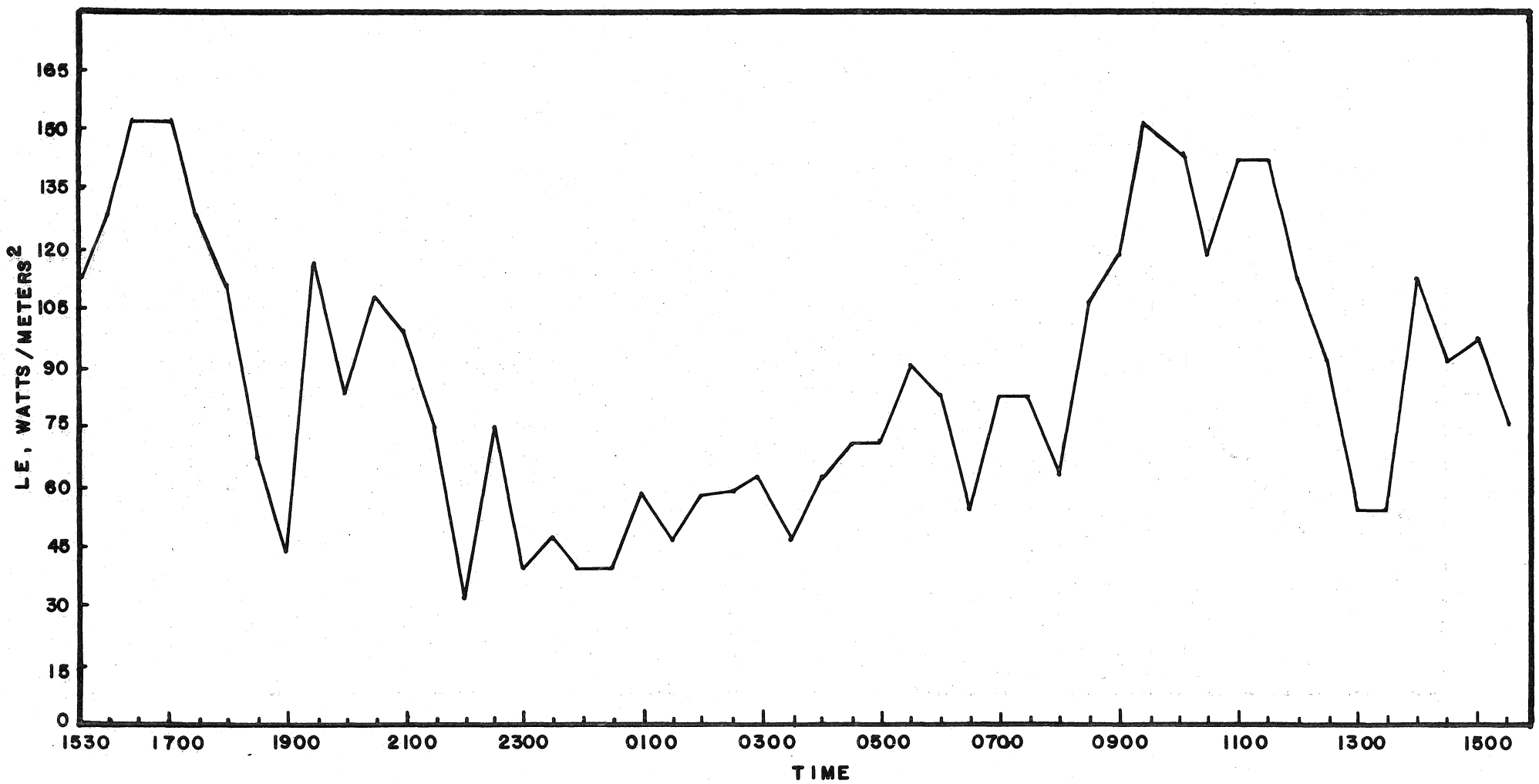

Figure V-5. Diurnal variation in latent heat flux density (w/m^2).