Small Lake Daytime Breezes: Some Observational and Conceptual Evaluations

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ABSTRACT

The diversity of small lakes' (size < 50 km) configurations, sizes, surrounding terrain, and land use combined with relative sparsity of observations complicates the observational evaluation of the lake breezes (LB) that are induced by these lakes. In the present article observational data obtained from available documents, data archives, and special projects were surveyed to suggest characterization of the LB features. The observational survey was complemented by conceptual evaluations. A preliminary generalization of the LB intensity and inland penetration in relation to the surrounding land use was inferred. The conceptual evaluation suggested that for a given lake width the prime factor affecting the LB intensity is the magnitude of the surface sensible heat flux over the surrounding land. Cooling related to the lake water temperature was indicated to have usually a secondary effect on the LB intensity for small lakes. Surface observations implied that the onshore penetration of the LB by the early afternoon hours is typically less than the characteristic width of the lake. Lower atmosphere observations indicated that the vertical extent of the LB may reach several hundred meters. Implications of the observed LB features in support of characterization of the real-world vegetation breeze are discussed.

1. Introduction

About 1% of the global continental area consists of lakes. The vast majority of these lakes have a typical width of less than 50 km (henceforth termed small lakes). For lakes exceeding this width, such as the Great Lakes of North America, the lake breeze (hereafter LB) intensity tends to resemble that of the sea breeze. Van der Leeden et al. (1990) list 170 natural freshwater lakes in the United States (excluding Alaska) of area 100–2500 km^2, whereas thousands of smaller lakes exist. Within the context of LB some bays as well as wide rivers provide under certain constraints the equivalent of small lakes. In midlatitudes and lower latitudes, a considerable number of the small lakes are man-made and commonly reflect broadening of rivers. Fells and Keller (1973) estimate the global area of all water surfaces created by man at > 300 000 km^2, whereas those with area larger than 100 km^2 accounted for ~110 000 km^2 of this area. They list 26 man-made lakes that were in operation or were in construction by 1973 with areas larger than 1000 km^2 but less than 2500 km^2.

During the warm season the influence of the LB is important for a variety of reasons, including biometeorological heat stress, air pollution dispersion, local cloud formation, water management, and recreational activity. Therefore evaluation of LB features should be of interest to meteorologists as well as to various interdisciplinary scientists. Additionally, in recent years there has been a growing interest in the vegetation breeze, which is a thermally forced circulation between vegetated areas and surrounding dry areas (e.g., Segal and Arritt 1992). In many cases, particularly when irrigated crop areas are considered, the characteristic size of the vegetated area is similar to that of a small lake. Observational evaluations of the vegetation breeze are quite limited; thus, some insight into their intensity can be inferred through observations of the lake breezes.

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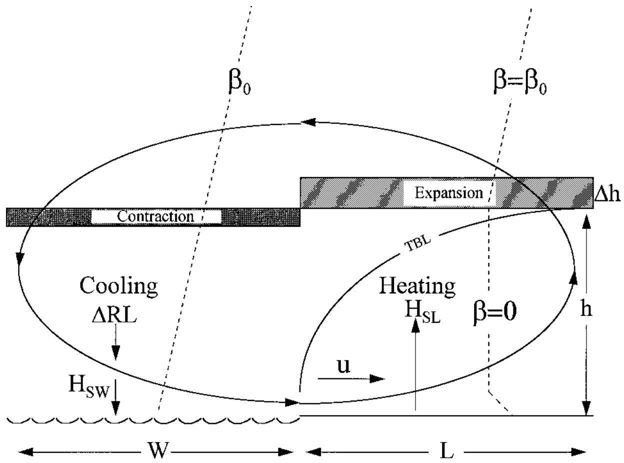

FIG. 1. Schematic illustration of features related to the scaling evaluation of the LB circulation (see the text for the notation). The air thermal expansion and contraction are magnified for purpose of illustration.

Unlike sea breezes, whose characteristics are widely documented (e.g. Atkinson 1981; Pielke 1984), there has not been an attempt for a systematic observational characterization of the small lakes' LB. This is partly because there have been fewer studies of lake breezes than of sea breezes. Additionally, the variety of lake sizes and shoreline configurations is a complicating factor in a generalization of LB characteristics. Further complication in the evaluation of the LB emerges from the variability of the large-scale meteorological characteristics that affect the LB. Climatic anomalies are likely to result in alteration in lake size, water temperature, and surrounding land use, consequently modifying the LB characteristics. In some geographical locations overuse of lake water, particularly during continuous drought years, may affect the lake's size and the characteristics of the related LB.

Several early observational studies of LBs were surveyed by Yoshino (1975). He presented two cases of small lakes of width of about 4 km (Lake Suwa, Japan) and 10 km (Lake Constance, Switzerland). The LB reached 2 m s^{-1} in the first case and somewhat greater in the second. However, these lakes are confined by nearby mountains, so that thermally induced upslope flows are likely to interfere to some extent with the LB. In later years, additional observational studies of LB were reported, as surveyed in section 3 of this article.

The objectives of the present article are twofold: (i) to provide conceptual evaluation of the characteristics of the LB, which includes scaling and illustrative numerical model simulations results, and (ii) to survey observational results of available documented studies as well as to analyze data from several recent projects in order to provide an initial step in the characterization of LBs induced by small lakes. We focus mainly on the warm season; however, some evaluations also are provided for other periods of the year. The evaluation primarily focuses on surface characteristics of the LB, and to a limited extent includes some characteristics of the vertical structure. Conceptual evaluations of small lakes' LB intensity are provided in section 2. The observational survey is given in section 3, followed by a discussion in section 4.

2. Conceptual evaluations of the surface LB speed

a. Scaling

1) SCALE ANALYSIS

Here we provide a simplified scaling in which the LB characteristic speed is estimated for midlatitude summertime conditions. (Figure 1 presents an illustration in support of the scaling.) The LB circulation is forced by conversion of potential energy into kinetic energy through mass redistribution. The potential energy available for this conversion is contributed by expansion of the onshore boundary layer due to diabatic heating, and to a lesser extent by contraction of the offshore air mass due to diabatic cooling (e.g., Pielke 1984). We consider a slab symmetric lake (i.e., an elongated lake with 2D symmetry) with no background wind. Thus, symmetric LB circulation cells should develop at the two shores.

Over the lake, on the average, stable stratification with potential temperature vertical gradient, β_o, is generated due to LB compensating subsidence in addition to near-surface diabatic cooling. The cooling is produced by two processes: (i) downward surface sensible heat flux, H_{sW} (assuming the air is warmer than the lake surface temperature) and (ii) cooling of the lower atmosphere forced by radiative longwave flux divergence, ΔR_L. In the following discussion we will refer to the combined effect of (i) and (ii) as the "effective" surface thermal flux. Using the nomograms in Kondo (1975), it is estimated that for typical LB speed and air–

surface water temperature difference, $|H_{sW}|$ < 20 W m^{-2}. Inferring from Ye et al. (1989), the radiative cooling is likely to be comparable to the cooling by sensible heat flux divergence. Thus the effective surface thermal flux over the lake is typically < 50 W m^{-2}, compared with typical values of onshore sensible heat flux, H_{sL}, of 200–450 W m^{-2} (e.g., Pielke 1984; Garratt 1992; Mahrt et al. 1994; Doran et al. 1995). Thus the magnitude of the potential energy gain of the onshore air column due to heating is noticeably larger than the potential energy drop offshore due to cooling. The cooling effect is likely to be greater for deeper lakes and in northern latitudes as these circumstances contribute to reduction in the summer water temperature. Also, the relative importance of cooling over the lake would increase with lake size because air parcels would reside over the lake for a longer period.

With the development of the inland LB a thermal boundary layer (TBL), in which the thermal stratification is unstable or neutral, is generated reaching depth h at onshore distance L. The direct LB flow is approximately confined to a layer depth similar to h. Here we evaluate the gain in potential energy due to expansion of the TBL by heating, and then approximate the LB characteristic flow induced through conversion of the gained potential energy into kinetic energy. Considering expansion Δh due to heating of a column of original height h, the net gain of potential energy per unit area, ΔE_p, is

$$\Delta E_p = \int_o^{h+\Delta h} \left(\rho + \Delta\rho\right) gzdz - \int_o^{h} \rho gzdz, \qquad (1)$$

where Δp is the characteristic decrease in the air density and g is the gravity acceleration. We adopt the approximation

$$-\Delta\rho/\rho \cong \Delta\theta/\theta \cong \Delta h/h, \qquad (2)$$

where ρ and θ are the background air density and potential temperature, respectively, and $\Delta\theta$ is the characteristic change in θ within the air column due to heating. Using (2) in evaluating the rhs of (1) yields $\Delta E_p = 0.5\rho g(\Delta\theta/\theta)h^2$, so that for onshore distance L the net gain of potential energy per unit transverse length is $0.5\rho g(\Delta\theta/\theta)h^2L$. In the equilibrium stage following redistribution of mass, only a fraction, $W/(L + W)$, of the gained potential energy would be converted to kinetic energy within a domain of horizontal size $(L + W)$ and characteristic height h (where W is the offshore extent of the LB circulation; W cannot exceed half the width of the lake). Therefore, neglecting frictional losses in the mass redistribution (and therefore overestimating somewhat the resulting characteristic LB speed u), the following balance equation should be fulfilled:

$$0.5\rho g(\Delta\theta/\theta)h^2 W\,[L/(W+L)] = 0.5\rho u^2(W+L)h. \qquad (3)$$

In order to simplify the derivation of u from Eq. (3), these common approximations (e.g., Garratt 1992) are used:

$$\Delta\theta = 0.5h\beta_o, \qquad (4)$$

and

$$h = \left[\frac{2C_\theta \int_{t_W}^{t_L} H_{s_L}\,dt}{\rho C_p \beta_o}\right]^{1/2}, \qquad (5)$$

where $C_\theta (\cong 1.2)$ is an entrainment coefficient used to parameterize the heat flux at the top of the TBL, β_o is the average vertical gradient of potential temperature over the lake, and C_p is the atmospheric specific heat at constant pressure. The time at which a column of air initiates its onshore advection by the LB is denoted by t_w and the time by which the column reaches distance L onshore is denoted t_L. Substituting (4)–(5) into (3), using as a scaling approximation $L = u(t_L - t_w)$, and assuming that the onshore average surface sensible heat flux during the period $(t_L - t_w)$ is H_{sL}, $\hat{H}_{sL}$ results in

$$u = \left[1.2\frac{g\hat{H}_{s_L}}{\theta\rho C_p}\frac{W}{\left(1+\frac{W}{L}\right)^2}\right]^{1/3}, \qquad (6)$$

where u is the characteristic LB speed. The expression (6) is similar to that derived using a different scaling approach by Mahrt et al. (1994; 2488) for vegetation breezes; however, it includes refinement to account for the impact of mass redistribution on the flow. To a first approximation the cooling over the lake can be accounted for in the scaling by taking $\hat{H}_{sL}$ in (6) as the *difference* between the sensible heat flux over the land and the effective surface thermal flux over the lake.

In the following evaluations (6) is applied under various constraints on L to estimate the typical range of u, in the absence of background flow. First we focus on situations in which the LB reaches its peak due to the merging of both shores' circulations at the center of the lake; thus $W = \tilde{W}$, where $\tilde{W}$ is the lake half-width. Assuming first that $L = W$, $\hat{H}_{sL} = 300$ W m^{-2}, $\rho = 1.2$ kg m^{-3}, and $\theta = 300$ K, then for $\tilde{W} = 1$ km the corresponding characteristic LB speed is $u = 1.4$ m s^{-1}. Holding all other parameters unchanged, for $\tilde{W} = 10$ km, $u = 2.9$ m s^{-1}, and for $\tilde{W} = 25$ km, $u = 4$ m s^{-1}. Increasing $\hat{H}_{sL}$ by 50% (to 450 W m^{-2}) would lead to intensification of u by about 15%, whereas decreasing $\hat{H}_{sL}$ by 50% (150 W m^{-2}) results in a decrease of about 26% in u. The scaling implies also that, for given W, (i) when $W/L >> 1$ then $u \to 0$, since in the absence of background flow the LB cannot be confined to the lake only; and (ii) when $W/L << 1$ then u increases by ~60% compared to its value for $L = W$.

2) Modeling estimations

Only a few numerical modeling studies of small lakes' breezes have been reported. Neumann and Mahrer (1975) simulated the LB of circular lake of diameter of 56 km, whereas Physick (1976) reports simulation of elongated lake of 52-km width. Both studies considered summer conditions and dry land situation. In the first study the LB onshore flow reached in the early afternoon a peak speed of 6 m s^{-1} (and near surface speed of 4 m s^{-1}), a depth of ~400 m, and an inland penetration distance of about half the lake width. In the second study the peak LB reached ~7 m s^{-1}, whereas the depth and inland penetration were almost doubled compared with the first study. Partially, the differences between the two studies are attributed to the shore curvature effect: in the circular lake the available potential energy per unit shore length is lower.

We provide additional insight into the LB characteristics in support of the scaling presented in the previous subsection through numerical model simulations of a 2D slab-symmetric lake. The simulations were performed utilizing a nonhydrostatic version of the model described in Arritt (1989) for standard atmosphere initial conditions, in absence of large-scale flow, and under midlatitude summertime conditions with noon land surface Bowen ratio of about 1 (with a corresponding $H_{sL} \cong 200$ W m^{-2}). The simulations commenced around sunrise (at 0600 LT).

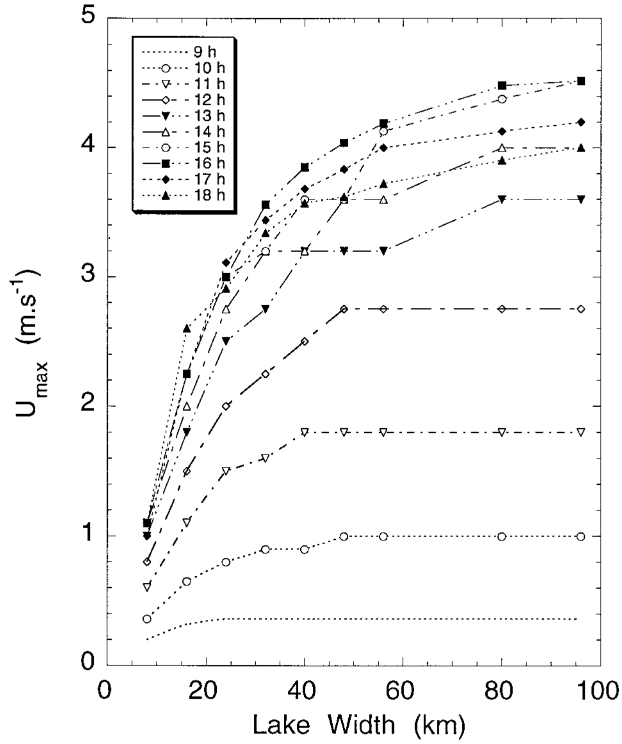

Fig. 2. Simulated maximum u component (u_{max}) of the LB (in the simulated vertical cross section) at various hours (as indicated) for 2D lakes of widths of 8, 16, 24, 32, 40, 48, 56, 80, and 95 km (indicated by the symbols on the curves).

Figure 2 provides an illustrative model evaluation of the relationship between the peak LB u component u_{max} and the lake width in support of the scaling presented in the previous section. The simulations were carried out for lakes with widths of 8, 16, 24, 32, 40, 48, 56, 80, and 95 km (the last three widths exceed our definition of "small" lakes but were included for comparison purposes). The simulated values of the LB speed support the scaled values presented in the previous subsection. For lake width exceeding about 80 km, a sea-breeze situation is approached and any additional increase in the lake width has little further impact on u_{max}. For given latitude, following Pearson (1973), the inland extent of the sea-breeze circulation is constrained by the Rossby radius of deformation L_D. Utilizing the scale analysis in section 2a(1) approximately $L_D \alpha H_{sL}^{1/2}$. Doubling the value of H_{sL} (as would be appropriate for dryland case; Bowen ratio >> 1) would increase by about 40% the inland extent of the LB when it approaches a sea-breeze situation. It is suggested that the corresponding increase of the lake width at which the circulation approaches a sea breeze would be similar. In the presented simulations, the daytime maximum LB speed u_{max} is dependent on lake width, $\bar{W}$ in km and, in reference to the daytime peak lake breeze speed for an 8-km wide lake, u_{8max} is approximated by a logarithmic fit, $u_{max} = 1.25u_{8max}\ln(\bar{W}/2.96)$. Finally, at 1400 LT for all the simulated small lakes the characteristics depth of the onshore LB was less than 600 m, and the inland penetration of the LB was less than half the width of the lakes (not shown).

b. Background flows

Based on conceptual considerations it can be suggested that even in the absence of background flow, the LB would not be unambiguously detectable when the LB intensity is of similar magnitude as the horizontal flow associated with convective boundary layer (CBL) large eddies. Over dry surfaces the typical midday large eddy flow speed is in the range 1–2 m s^{-1} for midlatitudes in the warm season. Alternatively, following the analysis in section a(1), using the free convection velocity scale w^* as a typical large eddy flow speed, Eq. (6) can be approximated as $u = w^*(\tilde{W}/h)^{1/3}$. Therefore the lake half-width $\tilde{W}$ should be at least larger than the CBL depth h in order to observe an onshore LB signal. This conclusion is in agreement with Hadfield et al. (1991), who utilized large eddy simulations to evaluate small-scale thermally forced circulations in the CBL. It should be noted that observational studies have indicated clearing of fairweather cumulus clouds over rivers of width greater than 1 km (e.g., over tributaries of the Chesapeake Bay; Scofield and Weiss 1977) and small width lakes in Oklahoma (Rabin et al. 1990), which may be attributed in part to weak LB divergence. However, no wind observations are available to confirm (or refute) a detectable LB in these cases. In the presence of background flow, divergence along a river or small lake may be detectable only as a perturbation in the background flow. Background flows are likely to obscure the identification of the LB most greatly for small lakes surrounded by wet surfaces. Following Segal and Arritt (1992) and Doran et al. (1995), we infer that for lakes of width 10 km, for example, the LB directed onshore (in the windward shore) is likely to be suppressed for a background wind at least 3 m s^{-1} for a lake surrounded by relatively moist land and 5 m s^{-1} for lakes in dry surroundings. In such cases the LB signal may be detected as a perturbation in the background flow. Typically, in this situation the LB circulation is shifted offshore, resulting in the disappearance of the branch of the LB flow on the windward shore. With the intensification of the background flow, coupling of the background flow with the LB is revealed as directional and speed perturbations of the background flow (e.g., Physick 1976; Segal and Pielke 1985; Arritt et al. 1996).

c. Impact of the surrounding land use

Many small lakes in the midlatitudes are located within vegetated areas, such as grassland, crops, or forested areas. Generally, when these canopies are mature (typically early in the warm season), peak seasonal transpiration occurs and thus some suppression of the LB is expected. Also for this type of land use, the characteristics of the rainy season are likely to affect evapotranspiration patterns and consequently the LB intensity. For example, during drought years the LB in these locations is likely to intensify due to an increase in H_s caused by dry soil and reduced evapotranspiration. In a reversed situation, a less detectable LB should be associated with an above-normal rainy season, which results in increased surface wetness as well as cloudiness and corresponding decrease in H_s in the lake surroundings. Short periods associated with these dry or wet surface conditions would cause transient modifications of the LB characteristics. In agricultural areas, variation of evapotranspiration related to the various stages in the growing season should affect the LB. Evapotranspiration of forests is most significantly affected by the rainfall frequency and the air temperature.

The impact of the evapotranspiration in the lakes' surroundings on the LB intensity can be inferred from some of the observations that are presented later. For example, as will be shown, the LB was less pronounced for Lake Okeechobee (which is surrounded mostly by swampland or irrigated crops) compared with that of similarly sized lakes like the Dead Sea (which is in desert or semiarid regions).

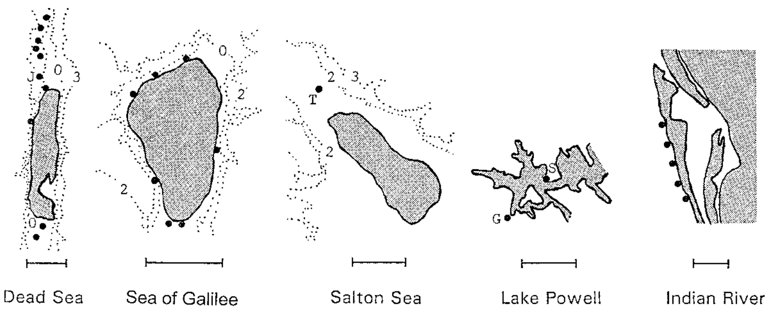

FIG. 3. Map illustration of several lakes for which LB observations are reported in the literature (reproduced from the studies cited in the text). The lake areas are shaded; relevant observational sites are indicated by dark circles and terrain contours are in hundreds of meters. The distance scale is 20 km except for 10 km for the Sea of Galilee.

d. Topographical effects

Many small lakes are located in valley basins (especially man-made lakes). In these cases coupling of the LB with daytime thermally induced upslope flow produces the local observed wind. When synoptic flow contributes to the observed flow, this additional complication must be accounted for. Segal et al. (1983) found that the wind speed of the coupled LB and upslope flow is similar to the upslope flow generated in the absence of a lake. Thus, even if the LB contribution to offshore flow is small (i.e., while assuming the same lake and environmental conditions, except for flat terrain), an apparent LB may be suggested that from a dynamical perspective is primarily attributable to thermally induced upslope flow. In order to resolve the LB in lakes located in valleys, observational sites located along the valley axis (i.e., in low-altitude locations of the valley) where thermally induced upslope flow is mild should be examined. In relatively narrow valleys, channeling of the LB is likely to enhance somewhat its intensity.

3. Observations

a. The surface LB based on reported studies

Lake-breeze characteristics based on available documentation were used in the following observational evaluations (focusing on clear sky and synoptically unperturbed days). The lakes' configurations and the locations of the observational sites are illustrated in Fig. 3 (reproduced from the original papers).

The Dead Sea in the Jordan Rift Valley is a terminal lake that is elongated and about 15 km wide, confined by elevated terrain to the west and the east. Conceptually an LB-dominated flow should be detected along the central axis of the valley where topographical gradients are small. In the summer, the synoptic and mesoscale flow features in this semiarid area are highly persistent thus enabling evaluations of the LB in monthly base averages. The signal of the LB along the northern shore of the Dead Sea was observed by Ashbel (1939) to be very noticeable following 0900 LT by the onset of southerly flow and drop of the shelter temperature. The monthly averaged flow direction and speed for the area in August are reported in Bitan (1977). By noon the LB is evident at sites extending to about 20 km from the northern shore with average flow of 4 m s^{-1}. However, in this direction the lake's cross section is ~80 km, which should enhance the LB in these locations given the valley channeling effect on the flow. Penetration of the Mediterranean sea breeze (SB) to the Dead Sea area in the afternoon hours masks the Dead Sea LB. The shallow southern portion of the lake vanished during the 1970s due to diversion of water from the Jordan River, which is the main water supply to the lake. Prior to this change, based on Ashbel (1939) and Bitan (1977), the lake breeze was observed in the southern sites, although with lower intensity than in the northern sites. Both studies attributed this difference to the somewhat higher surface water temperature in the southern portion of the lake. However, based on the conceptual evaluation in section 2, it is suggested that the semiseparation of the southern portion from the rest of the lake effectively corresponds to a smaller width lake than the northern sections, which is an additional factor in reducing the historical southern Dead Sea LB.

The roughly oval-shaped Sea of Galilee (Lake Kinneret) in the northern portion of the Jordan Rift Valley is surrounded by elevated terrain except for the southern (vegetated) and northern sections. Following Bitan (1981), onshore flow with speed of 4 m s^{-1} is observed on the average for August in the first half of the day (prior to the penetration of the Mediterranean SB to the area) at sites located at the bottom of the western and eastern slopes within a distance of less than 1 km from the shore.

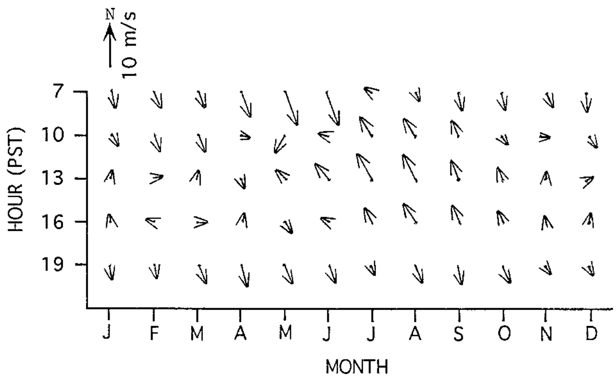

FIG. 4. The annual daytime pattern of the flow at Thermal (T), California, north of Salton Lake (see Fig. 3 for a schematic map illustration). Flow with southerly component implies an LB.

These flows reflect coupling of LB and the thermal upslope flows (e.g., Alpert et al. 1982).

A noticeable LB diumal cycle is observed for the northern Salton Sea area in the semiarid Imperial Valley of southern California. The area is surrounded by elevated terrain to the west and east; however, the valley is wide enough to be only marginally affected by daytime thermal upslope flows at Thermal, California (denoted by *T* in Fig. 3), which is located in the middle of the valley about 15 km northwest from the shore. Monthly averaged wind velocities for the daytime hours reproduced from Hayes et al. (1984), presented in Fig. 4, show for most of the year a daytime southerly flow reflecting an LB. It is initiated by 1000 PST in the summer and later in the rest of the year, while its cessation in the summer is later in the day. The peak LB speed in the summer is about 6 m s^{-1} (in July) and it should be enhanced because of the orientation of the lake major axis toward Thermal and associated valley channeling. It is worth noting that this strong signal of LB is observed even though much of the northern lake area is cultivated. It might be suggested, therefore, that the vegetated area by its suppression of sensible heat flux provides a virtual extension of the lake area.

Balling and Sutherland (1988) and Sutherland and Ostakup (1989) examined LBs at the shore of Lake Powell [at Glen Canyon (G) and Sand Hills (S) in Fig. 3]. The lake is located along the Colorado River and is surrounded by elevated terrain. The lake is elongated with large coves (> 5 km size) at the location of the measurements. The LB signal was observed during the winter and the summer at Glen Canyon and can be distinguished even though topographical thermal and dynamical effects may be present.

The Indian River is an elongated estuary with a maximum width of 5 km parallel to the Atlantic coast of central Florida. Zhong and Takle (1992) and Laird et al. (1995) present a single summer day observation of the daytime flow in this area. At several sites 1–1.5 km from the Indian River's western shoreline, a temporary weak flow attributed to an LB-like effect was detected prior to the penetration of the sea breeze from the Atlantic Ocean to the area.

Finally, Oliveira and Fitzjarrald (1993) inferred the existence of a river breeze based on surface and lower atmosphere observations close to the confluence of the Solimoes and Negro rivers in the Amazon rain forest, where the combined width of the two rivers is ~25 km. The river breeze signal was inferred near the surface and within the lower atmosphere for onshore distances as large as the combined river width. Negligible Coriolis force in this latitude should enhance the onshore penetration of the LB signal. Indirect support for river breezes induced by large rivers in Brazil is provided by Curtim et al. (1995). Monthly composites of satellite data indicated cloud clearing surrounding such rivers.

b. Surface LB observed in recent special projects

Several lakes' LBs were evaluated from the results of special field projects designed to characterize the LB, as well as from further analysis of data (Table 1). Schematic illustrations of the lakes and observational sites are presented in Fig. 5. Table 1 provides a summary of the analysis that was based on observations of surface wind and, in

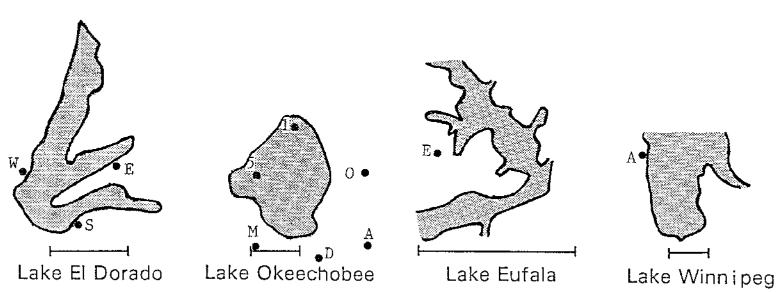

FIG. 5. Map illustration of several lakes evaluated for their LB in the present study and the location of related meteorological stations. The lake's area is shaded. The distance scale is 20 km except for 3 km for Lake El Dorado.

TABLE 1. Summary of LB observations for four lakes: Lake El Dorado (observation period: 23 February–8 April, 1995); Lake Eufala (observation period: 17 June–8 September 1994); Lake Okeechobee (observation period 21 May–22 September 1995); and Lake Winnipeg (observation period: 1 March–15 May, 1993–1994). See Fig. 5 for map illustration of the lakes. Landscape notation: D: dry surface, V: vegetated, S: bare soil.

Lake	Onshore distance (km)	# of fair weather days	# of onshore flow days	Land-scape	# of days with LB	Period of the day (LST)	Possible range of LB speed[a] ($m\ s^{-1}$)
Lake El Dorado, KS, South Site (S)	0.2	13	7	D	7	1000–1800	2–4[a]
Lake El Dorado, KS, East Site (E)	0.5	15	5	D	5	1000–1700	2–5[a]
Lake El Dorado, KS, West Site (W)	0.2	16	6	D	4	1000–1700	2–3[a]
Lake Eufala, OK (E)	3	26	10	V	9	1200–1700	1–3[a]
Lake Okeechobee, FL, Osce.[b] (O)	15	43	19	V	19	1100–1500	2–4
Lake Okeechobee, FL, Atla. (A)	25	52	20	V	19	1100–1500	2–4
Lake Okeechobee, FL, Mott. (M)	12	52	28	V	21	1100–1600	2–4
Lake Okeechobee, FL, Duda.[c] (D)	11	29	6	V	6	1100–1500	2–4
Lake Winnipeg, Canada, Arnes (A)	0.3	—	—	S	10	1000–1600	2–5

[a]Coupling with supportive background flow may contribute to increase in the LB upper range wind speed.
[b]No data after 2 August.
[c]No data after 12 June.

some sites, cloudiness, temperature, dewpoint temperature, and solar radiation. The table provides information about the distance of the stations from the representative shore (occasionally the lakeshores include irregularities that are minor in their impact on the LB). Also provided are the number of fairweather days (i.e., background flow < 5 $m\ s^{-1}$ and little cloudiness) that are conducive to LB development. For this set of days the number of days in which the daytime flow direction was the same as that of the anticipated LB is indicated.

The LB situations were discriminated subjectively based on the consistent timing and direction of daytime onset and cessation of the onshore flow in various days. For fairweather days we expect onset of onshore flow in the mid- or late morning and its cessation later in the afternoon (a sharp indicator is an associated change to a persistent background onshore flow).

The authors deployed three surface stations near the shores of the narrow El Dorado reservoir, Kansas (maximum width ~3 km), during late winter–early spring 1995. The surface in the surroundings of the lake was relatively dry, while daily integrated solar irradiance values toward the end of the observation period reached about 75% of midsummer values (thus implying supportive conditions for LB). Several days with LB signal were detected with the wind speed range listed in Table 1 (again it is worth emphasizing that the upper wind speed range is likely to be obtained through supportive background flow). Evaluations were also carried out for Lake Eufala, Oklahoma, based on the Oklahoma Climatological Survey station at Eufala. For this lake, which is slightly wider than the El Dorado reservoir, the LB signal was inferred in 9 summer days during the afternoon hours.

Lake Okeechobee, Florida, and its LB were observed during a comprehensive field study (Lake Breeze Experiment, LABEX) that was conducted in summer 1995 (Arritt et al. 1996). A summary of the characteristics of the LB as observed at four onshore permanent stations located various distances from the shores suggests that a detectable LB is less pronounced than would be anticipated from this relatively large lake. Relatively low Bowen ratio in the surrounding swampy or irrigated land, the warm lake surface, frequent occurrence of afternoon local deep convection (which may be triggered by the lake breeze itself), and penetration of the sea breeze from the Atlantic Ocean are factors that would mask the Lake Okeechobee LB (see Fig. 7 below).

Using the routine observations for the period 1 March–15 May of the years 1993–94, the LB was examined at Arnes, Manitoba, Canada, which is located on the southwestern side of Lake Winnipeg (the lake width in this location is ~40 km). In 10 days of unperturbed weather conditions, LBs with speed of several meters per second were observed [see Segal and Kubesh (1996) for illustrative cases]. The most pronounced LBs in the southwestern section of Lake Winnipeg typically occur in May, with onshore penetration of ~5 km (B. Atkinson 1995, personal communication).

The possible period of the day in which the LB was observed in each of the presented lakes is indicated in Table 1. It should be noted that in some cases, an LB was observed for the whole stated period of the day, in other cases only in a portion of the period, and occasionally not continuously. The range of the observed maximum LB was 3–5 m s^{-1}. It is likely that in some cases with background flow, the LB was reflected only as a mild perturbation in this flow velocity, so that the upper range for the LB speed at each of the evaluated sites may overestimate the intensity of the LB.

c. Observed vertical characteristics of the LB

Atkinson (1981) and Pielke (1984) surveyed observational studies that reported on the depth of the direct sea breeze circulation. Typically the depth of this layer is several hundred meters with peak depths of about 1000 m. The depth of the direct LB flow should be smaller for small lakes and should reduce with the lake size. Several observational studies, which are surveyed in the following, provide direct and indirect observations of the depth of the direct LB circulation.

For the Dead Sea, upper air observations at 1100 LT at Jericho (distance of 12 km from the northern lakeshore; J in Fig. 3) indicated, for August, a 300–600-m-deep layer with flow from the lake (Bitan 1977). This relatively deep lake breeze contrasts with measurements from flights crossing Lake Newell (13- × 5-km oval-shaped lake that is surrounded by dry farm land and uncultivated land) in Alberta, Canada, where in the early afternoon lake–land temperature differences were observed for heights only up to about 60 m (Holmes 1973). Direct LB flow would be limited to this depth. Similarly a shallow LB was implied for Lake McConaughy, Nebraska, which has approximate dimensions 20 km × 5 km along the South Platte River in western Nebraska. On 23 August 1993 (around noon), flights by the University of Wyoming King Air crossed the lake at an altitude of 300 m in various locations, including its widest section. Suppressed turbulence over the lake and slight air temperature differences between the lake and the land were observed. However, only a perturbation was revealed in the background flow. We have performed numerical model simulations that suggest that a weak LB circulation may have been present at altitudes below the lowest level of aircraft measurement (not shown).

Sun et al. (1996) provide flight observations of three lakes of width 3–10 km located in the southern boreal forest in Canada (Halkett Lake, White Gull Lake, and Candle Lake) during summer 1993. Direct LB circulations extending several kilometers onshore were observed for the largest lake (Candle Lake) at altitudes of 35–150 m. Flow divergence was observed over the smaller lakes that implies the existence of LB effect.

In the LABEX project, aircraft and radiosonde measurements were carried out around noon during the period 7–14 July 1995 in the area of Lake Okeechobee (Arritt et al. 1996). Aircraft observations were in most cases above 300 m and did not detect a LB. In one case (9 July) when lower altitude flights were carried out, a LB layer of 250–

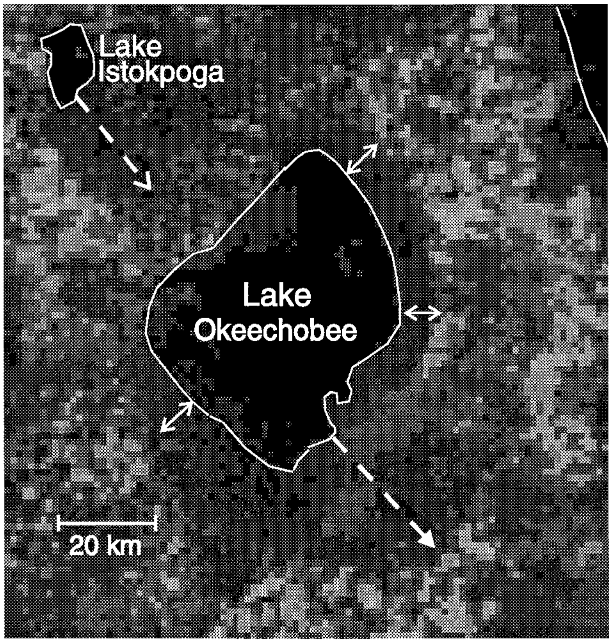

FIG. 6. GOES visible satellite image (pixel resolution 1 km × 1 km) of Lake Okeechobee on 1 July 1995 1300 EST. Double arrows adjacent to the lakeshore indicate the width of the cloud clearing, which implies the inland penetration of LB. Dashed arrows indicate cloud clearing supported by the background wind.

300 m depth was detected across the northwestern shore of the lake, contrasting strongly with an opposing background flow. A transition layer extended from the top of the LB layer to ~400 m above the surface. Winds in this transition layer were very light (≤ 0.5 m s^{-1}) and of indeterminate direction. Above the transition layer the background flow was reestablished.

d. LBs as implied by satellite imagery

Like sea-breeze situations, the area affected by an LB can be inferred from satellite visible imagery in areas affected by cumulus clouds or in a situation in which the LB front induces cumulus clouds. Current Geostationary Operational Environmental Satellite (GOES) satellite pixel resolution is 1 km; thus the LB onshore penetration should be at least a few kilometers in order to be resolved adequately. When background flow exists, an area of cloud clearing downwind is typical under appropriate environmental conditions (Rabin et al. 1990; Purdom 1991; Segal et al. 1996). This clear area is forced by LB subsidence and by the suppression of the CBL as air masses originating over the heated land are advected over the cooler water. The purely dynamical effect of the LB can be unambiguously established only for clearing under light background wind conditions and when clearing is observed upwind of the lake. Figure 6 provides an illustration of such a situation through a GOES visible image of Lake Okeechobee in the early afternoon. The background flow was weak northwesterly as evident from the extended clearing southeast of the lake (such clearing is evident also for the smaller Lake Istokpoga). However, a strip of at least ~8 km width of cloud-free area surrounded most of Lake Okeechobee, implying the extent of inland LB penetration. Figure 7 provides the diurnal variation of the surface winds for this day observed at two onshore sites and two offshore sites. At 1300 EST onshore flow indicating LB was observed in support of the satellite indication. The LB was observed for several hours at onshore sites (MOTT and OSCE) and for a shorter period at the offshore sites (#1 and #5).

4. Discussion and conclusions

The LB intensity is related to the pressure gradient in the lower atmosphere between the lake and

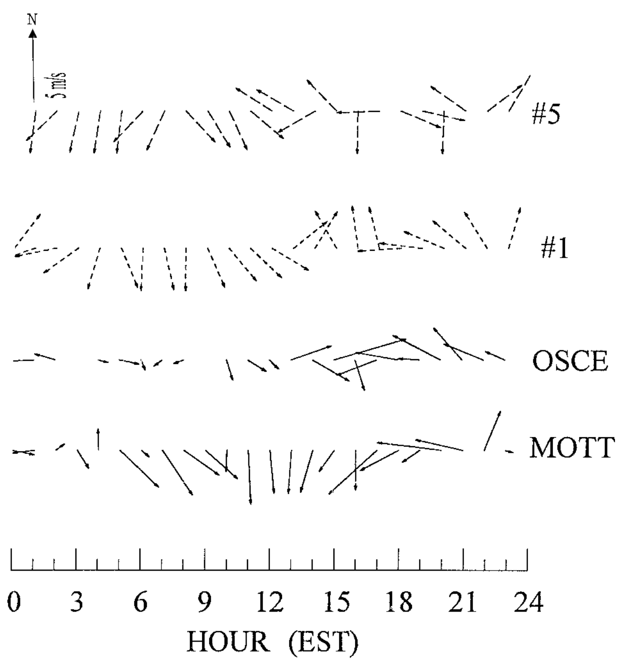

FIG. 7. Illustration of development of the lake breeze in Lake Okeechobee on 1 July 1995 at the onshore sites M (MOTT) and O (OSCE) and the offshore stations #1 and #5 (see Fig. 5 for site location). Dashed vectors indicate direction only.

the land. The magnitude of the inland sensible heat flux is a prime factor in establishing this gradient. Commonly, the cooling over a small lake (by sensible heat flux and longwave radiative flux divergence; both increase with drop in the lake water temperature) has only a secondary effect in contribution to the thermal contrast and therefore LB pressure forcing. The relative importance of the lake cooling effect as related to the surface water temperature would increase somewhat with the offshore extent of the LB circulation or when the offshore extent of the LB circulation is larger than the onshore one. Also, as the lake size increases, this cooling would further contribute to the LB intensity because of longer residence time for a given air mass over the lake prior to its onshore advection by the LB. Overall, assuming clear sky conditions the land use surrounding a small lake is of prime importance in determining the LB intensity for a given small lake because of its effect on the sensible heat flux. When a lake is surrounded by dense and highly transpiring vegetation that is juxtaposed by dry area, it is suggested that to some degree, in consideration of LB the vegetated area would provide a virtual extension of the lake.

The evaluations presented in this paper suggest that for lakes of characteristic width < 2 km the direct LB is likely to be undetectable along the lake shore. A mild background flow would result in distortion of the LB circulation. Even in the absence of background flow, the horizontal velocity component of the convective large eddies would mask the LB signal. On the other hand, when the lake width is > 80 km and the moisture availability of the surrounding land is moderate (Bowen ratio ~1), the LB intensity approaches that of the sea breeze, and any further increase in the lake width would not result in substantial intensification of the LB. The upper range for the surface LB intensity based on observations presented in the study reached about 6 m s^{-1}. Occasionally situations interpreted as strong LBs may in fact be attributable primarily to supportive background flow (i.e., onshore large-scale flow or daytime thermal upslope flows). Surface observations implied that the onshore penetration of the LB by early afternoon is typically less than the characteristic width of the lake. Lower atmosphere observations indicated that the LB signal can extend several hundred meters above the surface.

Lake breezes in dry geographical locations should be more frequent and pronounced compared with those in wetter areas for two main reasons: (i) in dry climates less perturbed weather is typical (e.g., infrequent passage of synoptic fronts and cyclones), and (ii) drier surfaces enhance the daytime surface sensible heat flux. Also, in dry locations the LB is likely to be more consistent from day to day, which simplifies its identification. In contrast, when lakes are affected frequently by synoptic perturbations, cloudiness, or rainfall, the character of the LB may differ from one day to another, and the number of days with clearly detectable LB may be low. Observations presented here imply that, indeed, in locations frequently affected by these conditions, the LB signal is obscured. It is worth noting that surface flow might be less revealing compared with measurements in the a layer of ≤ 50 m above ground as a lake breeze is expected to attain its peak wind speed above the surface.

The variety of lakes' shapes and the complexity of the coastlines makes it difficult to define their size as well as the correspondence between the land and water area and the LB. Distinction should be made between small coves and the major water area affecting the onshore LB. It may be difficult to determine whether vegetated swampland along the periphery of a lake should be treated as "modified" land or water (as in the case of Lake Okeechobee). Shrinkage of lakes as a result of the diversion of main water sources (e.g., the Dead Sea) may occur as the use of irrigation water increases. Occasionally, significant shrinkage of shallow lakes may occur because of drought conditions. Overall, the change in many small lakes' size and shoreline configuration as well as possible changes in land use may have considerable effects on the LB intensity of such lakes. Reasonable approximation to small lakes is likely for bays that are sufficiently distinct from larger water bodies. In these cases, the circulation associated with the bay can be examined as a distinct entity during the period of the day before the breeze from the large water body reaches the sites (e.g., Abbs 1986).

An implication of the present study is related to the characteristics of the "vegetation breeze" induced between a vegetated area and the surrounding dry area. Vegetation breezes have been simulated in numerous studies; however, the available observational results (e.g., Segal and Arritt 1992; Mahrt et al. 1994; Doran et al. 1995) have not indicated as strong circulations as implied by the simulations. In part this difference is the result of simulating idealized prescribed vegetation cover that is more extensive and uniform than in most real-world situations. Real-world irrigated areas (and in many situations other pertinent vegetated areas) are typically the size of small lakes. However, their suppression of sensible heat fluxes is in most cases less pronounced compared with water surfaces (due mostly to patches of bare dry soil or decrease in transpiration in response to environmental conditions). Thus the evaluations provided here for small lakes LB set an upper limit to the intensity vegetation breezes associated with vegetated areas of equivalent size and environmental conditions.

Finally, in order to establish an appropriate climatology of an individual small lake LB, several sites of observations in different onshore distances are needed at least during one warm season. Aircraft measurement would be especially useful in providing spatial characteristics of LB if flight altitude can be at least as low as 50 m.

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