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THE MICROCLIMATE OF A LAKE IN A SUBARCTIC CLIMATIC REGION

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THE MICROCLIMATE OF A LAKE IN A SUBARCTIC CLIMATIC REGION

A

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By

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THE MICROCLIMATE OF A LAKE IN A SUBARCTIC CLIMATIC REGION

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ABSTRACT

The microclimate of Lake Minchumina, Alaska (65 km^2), situated in a subarctic, continental climatic region, was investigated in the late summer of 1979 and compared with similar studies of lake microclimates in temperate climatic zones. Measurements of the lake effect like land-lake breeze, local warming, air temperature moderation and high dew-point, showed almost no difference with its temperate zone counterparts. A heat balance calculation at the lake's surface indicated small net radiation but very large sensible and latent heat losses.

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I. INTRODUCTION

For centuries farmers have been aware of the effect of a body of water on the microclimate of its surroundings. This information has been taken advantage of for the selection of preferred growing sites which are less susceptible to frost damage than other areas.

The occurrence of a lake effect on its surroundings is due to the different characteristics of water and soil: water is mobile and has the highest specific heat capacity among materials, soil is immobile and has a lower specific heat capacity. The specific heat capacity of water is 1.00 cal/g^{o}C at 15^{o}C and normal atmospheric pressure. The specific heat capacity of dry soil ranges from 0.16 to 0.21 cal/g^{o}C (Munn, 1966) depending on the moisture content, porosity, etc.. Owing to the convection in water, heat can be transferred to other levels more quickly in a lake than in soil. Therefore, supported by the higher specific heat capacity of water, a lake is able to store and release a much higher amount of heat than soil.

Consider the consequences of heating a lake at an initial uniform temperature of 1^{o}C. At this temperature warmed water has the unusual property of becoming denser and therefore it sinks, mixing the lake thoroughly until temperatures at all levels reach 4^{o}C. Thereafter, the heated water surface is lighter than the water at deeper levels, and further vertical motions are suppressed. Thereafter, the heat transfer takes place by turbulent mixing action of wind-generated waves (in the top few meters) and slowly by conduction (throughout the fluid). In summer, therefore, a body of water tends to be divided into a relatively warm upper layer, called the epilimnion, and a deep, cold, and undisturbed region called the hypolimnion. The plane of separation is the thermocline, defined as the level of maximum rate of decrease in temperature. In fall, when the water surface cools to a temperature of 4°C, its density increases and it sinks. The cooling of the entire epilimnion proceeds rapidly. However, additional cooling below the temperature of maximum density brings a return of stable stratification, and subsequent heat losses to the air are reduced (Munn, 1966).

The different specific heat capacities of water and soil and the convection of water have an immediate effect on the temperature distribution over the lake and its surroundings. During the course of a day, the lake maintains its surface temperature at an almost constant value, whereas the temperature of the land surface varies over a wider range. These different temperature conditions over lake and land during night and day cause a local wind phenomenon, known as the lake-land breeze (Geiger, 1965).

But the land surface temperature variations are obviously controlled by the distance from the lake. Daily and annual temperature fluctuations have been found to be smaller in areas near the shoreline of the lake than inland. For example D'yakonov and Reteyum (1965) reported that the maximum cooling influence is observed in April and May for the Rybinsk Reservoir in the U.S.S.R. (57°5'N, 37°5'E, 500 km^2). Phillips and McCalloch (1971) found that the decrease in the diurnal air temperature varation for areas in the proximity of the Great Lakes is about $5^{o}C$. This is the combined effect of a drop in temperature during the daytime and the warming effect of 3^{o}- $4^{o}C$ at night. The former effect was pointed out by Eley et al. (1961), who studied the variation of the afternoon temperature near the shores of Lake Winnipeg (Manitoba, Canada). The latter effect was reported by D'yakonov and Reteyum (1965) for the reservoir mentioned above.

In autumn, when the lake surface is cooled but the water temperature is still above $4^{o}C$ and has therefore not reached its maximum density, thermal convection starts, releasing a great amount of heat which, moving with the prevailing air flow, can significantly increase the environmental temperature. This heat release of large lakes or reservoirs in autumn is a major factor in extending the growing season around them. Phillips and McCulloch (1971) reported the addition of 16 days to the growing season for the Great Lakes area. Furthermore, there is less fluctuation in the dates of annual frost occurrence on the shores of the Great Lakes. In the Soviet Union, similar effects - including the extension of the frost-free season by an average of 5 to 15 days - were reported by D'yakonov and Reteyum (1965) for the Rybinsk Reservoir.

The lake effect is a function first of the climatic condition in which the lake is situated, and second the morphometry of the lake itself (Ragotzkie, 1978). Large lakes (more than 1000 km^2) have been studied previously by many meteorologists, for example, Great Lakes (about 5000 km^2) by Phillips and McCulloch (1971). The effects of some small lakes with areas of about 10 to 200 km^2 have also been studied, for example, Lake Suwa, Japan (14 km^2) by Yoshino et al. (1970) and Lake Inawashiro, Japan (104 km^2) by Shitara (1967), and Lake Mendota by Dutton and Bryson (1962).

However, all of these studies mentioned were carried out for lakes in temperate climatic regions, none in subarctic or arctic regions. Since the lake effect is dependent upon the climatic conditions as mentioned above, it is very important to examine if a lake in a subarctic climatic region has the same effect or the magnitude of the effect is the same as for lakes in a temperate climatic region.

II. OBJECTIVES

Since there have been no studies of subarctic and arctic lakes, it would be of interest to know if they have the same effect on their surroundings as do lakes in temperate regions. A subarctic or arctic lake is assumed to have the following characteristics: (1) a location north of 60^0 latitude; (2) a large seasonal variation in the duration of available daylight; (3) a continuous winter ice cover. We have chosen Lake Minchumina, which is located in a subarctic continental climatic region at the center of Interior Alaska.

The main objectives of this study are to answer the following questions:

a) Is the size of Lake Minchumina large enough to produce a lake effect?

b) What are the characteristics of the lake effect in a subarctic region?

c) Does the marked seasonal change of the daily sunshine duration affect the lake and its surroundings?

d) Are the characteristics of the lake effect produced by a subarctic lake substantially different from those produced by a lake in a temperate climate?

In order to answer these questions, the geography of Lake Minchumina will be described, followed by the location of the study sites and their instrumentation (Chapter III). Then the general weather conditions of Interior Alaska during the summer and fall, 1979 will be discussed (Chapter IV A), because they represent the "external forcing function" for the observed lake's microclimate. In the remaining part of Chapter IV the results of wind, air temperature, dew-point temperature and heat budget data and their interpretations will be shown. In this Chapter IV B, the wind regime of Lake Minchumina will be discussed first because the lake's effect is expected to be observed during favorable conditions of the synoptic wind regime. Then the lake's influence on air temperature and dew-point temperature around the lake will be examined followed by the heat budget calculations which may give us a quantitative estimate of the thermal responses of the lake to its surrounding boundary layer. In the final chapter, the answers for all the questions proposed and the discussion of their implications will be attempted.

III. STUDY SITE AND INSTRUMENTATION

A Study Site

Lake Minchumina (63°53'N, 152°15'W) is located in Central Alaska (Fig. 1) 106 km north of Mt. McKinley (6187 m, the highest mountain on the North American Continent). The lake's area is 65 km^2, at an elevation of 196 m above mean sea level, with hills (150-300 m above the lake's water surface) to the north-west and flat land to the east and south (Fig. 2).

The heavily silt-laden Foraker River is the primary inlet. The river is a glacier-fed stream flowing from the Alaska Range into the south-east end of the lake. The silty water circulates throughout the lake except at the extreme western end. As a result, most of the water is gray. Deep Creek is the second largest inlet, estimated to have less than one tenth the flow amount of the Foraker River. The brown-stained but clear water from several other small streams entering the west end of the lake keeps the lake clear enough for sunlight penetration, resulting in lush growth of aquatic macrophytes which choke the west bay by early August. The Muddy River is the only outlet of Lake Minchumina, flowing out from the east end of the lake into the Kantishna River.

The shoreline is primarily rocky, except in the west end and south side where the banks are boggy. The rocky part of the shorelines is adjacent to hills that rise abruptly from the lake. There are also some sandy beaches at the tip of North Bay and some silt banks on East Bay.

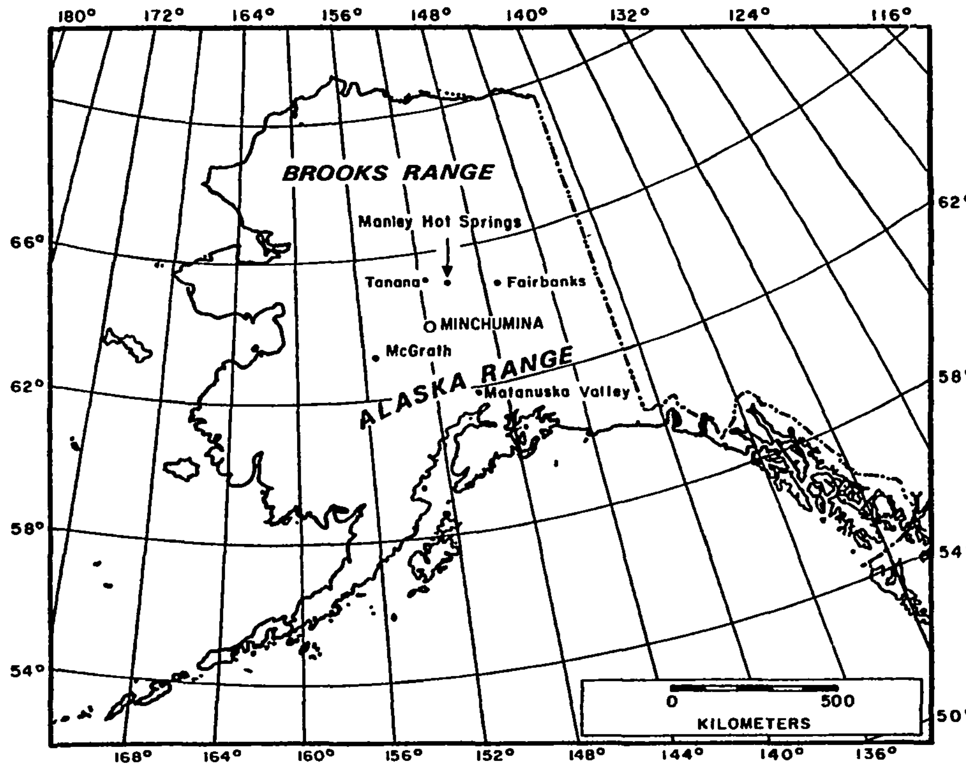


Fig. 1: The geographic setting of Lake Minchumina.

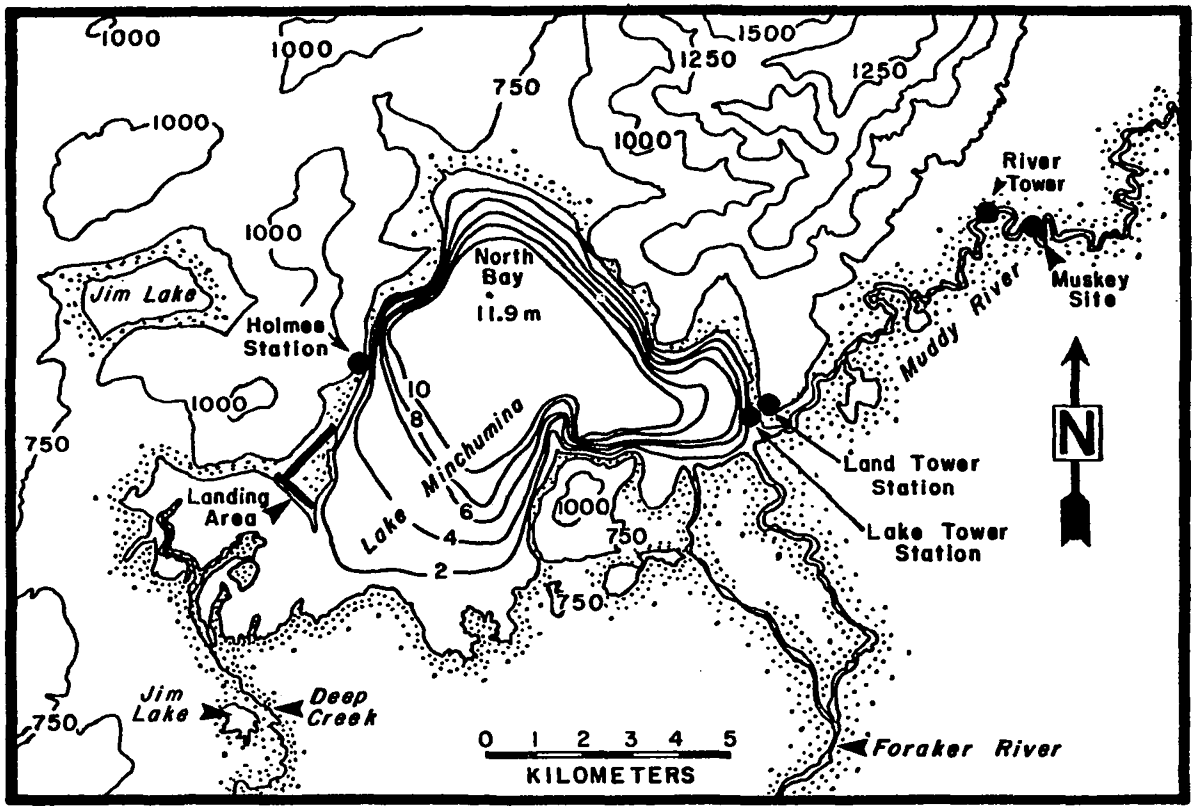


Fig. 2: Topography around Lake Minchumina (after U.S.G.S,) and the depth contours of the lake (after Van Hulle *et al*.). The altitudes are indicated by feet above MSL, the depths are in meters.

The lake's most prominent shoreline features are two gravel spits. One spit extends 0.5 km in a southerly direction from the point on which the airplane landing strip is located. The other spit extends 1 km north from Yutokh Hill (whose height is 150 m above the lake's water surface).

The lake is relatively shallow, with a maximum depth of 12 m. The depth contours are shown in Fig. 2. The bottom of the lake is silt-covered except at the extreme west end where organic material is the primary constituent. Near the hills, where the lake bottom almost uniformly drops off quickly, there is little aquatic vegetation because sunlight cannot reach through the deeper water.

The landscape south of the lake is flat with the exception of the Yutokh Hill near the south shore. In the eastern part of this plain many bogs and ponds are seen, which are the old river beds of the Foraker and Muddy Rivers.

The vegetation around the lake is variable, but dominated by broad leaf deciduous trees. Many pure stands of paper birch are present, especially on the northwest and northeast hills.

The reasons Lake Minchumina was chosen for this study are the following:

1) It is one of the largest lakes in Interior Alaska.

2) Accessibility by commercial flights was available.

3) Instruments could be ferried by a river boat via the Muddy River.

4) The population around Lake Minchumina (about 20) hardly disturbs the natural environment, and the lake is far away from any population centers (250 km from Fairbanks whose population is about 75,000).
5) There was a weather station at the airplane landing strip in operation for many years, and data from this indicated a lake effect (see Chapter IV c).

B Location of Observation Sites and Instrumentation

Five meteorological stations were established near Lake Minchumina (Fig. 2). Since the prevailing winds are westerlies it was decided to set up most of the stations at the eastern side of the lake at different distances from the lake's shore. This also had the advantage that the stations could be reached easily by boat on the Muddy River. Because of the limited number of instruments the number of stations was restricted to five, but these stations were sufficient in recording the lake effect.

The first, referred to as "Holmes Station", was set up 50 m away from the western shore, where wind speed, wind direction, air temperature, humidity and global radiation were measured from the beginning of May until the end of October 1979. A Woelfle-type anemometer, which records wind speed and direction on a mechanically wound recorder with pressure-sensitive paper, was used. A hygrothermograph for measuring air temperature and humidity, and a Robitzsch-type pyranograph for measuring global radiation were also installed. Holmes Station was expected to represent background conditions during the periods of the prevailing westerlies.

At the eastern shore two towers (6 m high) were constructed at the beginning of July 1979 - one in the lake, the other on land - to obtain vertical profiles of wind speed, air temperature and humidity (at the heights of 0.5, 1.0, 2.0, 4.0 m). At the same time temperature profiles in soil (depths: 0.05, 0.1, 0.2, 0.4, 0.5, 1 m) and water (depths: 0.1, 0.2, 0.5, 2 m) were measured near the land and lake tower, respectively. These data were recorded on the Grant CR50 data logger (Grant Instruments Ltd. Cambridge, England) which is battery-operated, allows operation for several weeks unattended and records on magnetic cassette tape. Besides these observations, the water level was measured. These stations are referred to as "Land Tower Station" and "Lake Tower Station".

The fourth station, "River Tower Station", was set up 6 km from the lake down the Muddy River, which drains Lake Minchumina. A tower was built to obtain vertical profiles of wind speed and air temperatures (at 0.5, 1.0, 2.0, 4.0 m). Soil temperature profiles were also measured. These data were recorded on a second CR50 Grant cassette tape data-logger.

"Muskeg Site", the fifth station, 1 km further down the Muddy River from the River Tower Station, included a hygrothermograph to measure air temperature at 1.2 m, a Woelfle-type anemometer for recording wind speed and direction.

In summary the five stations were expected to reflect the following conditions: Holmes Station was expected to be used as a reference for the climatic background, especially during periods of westerlies, Land Tower, Lake Tower, and Site 3 should be strongly influenced by the lake, whereas River Tower and Muskeg Site should show little or no effect due to the lake.

Table 1 summarizes the data taken at the individual stations and used in the study. The column "height" specifies the height at which the particular data were taken.

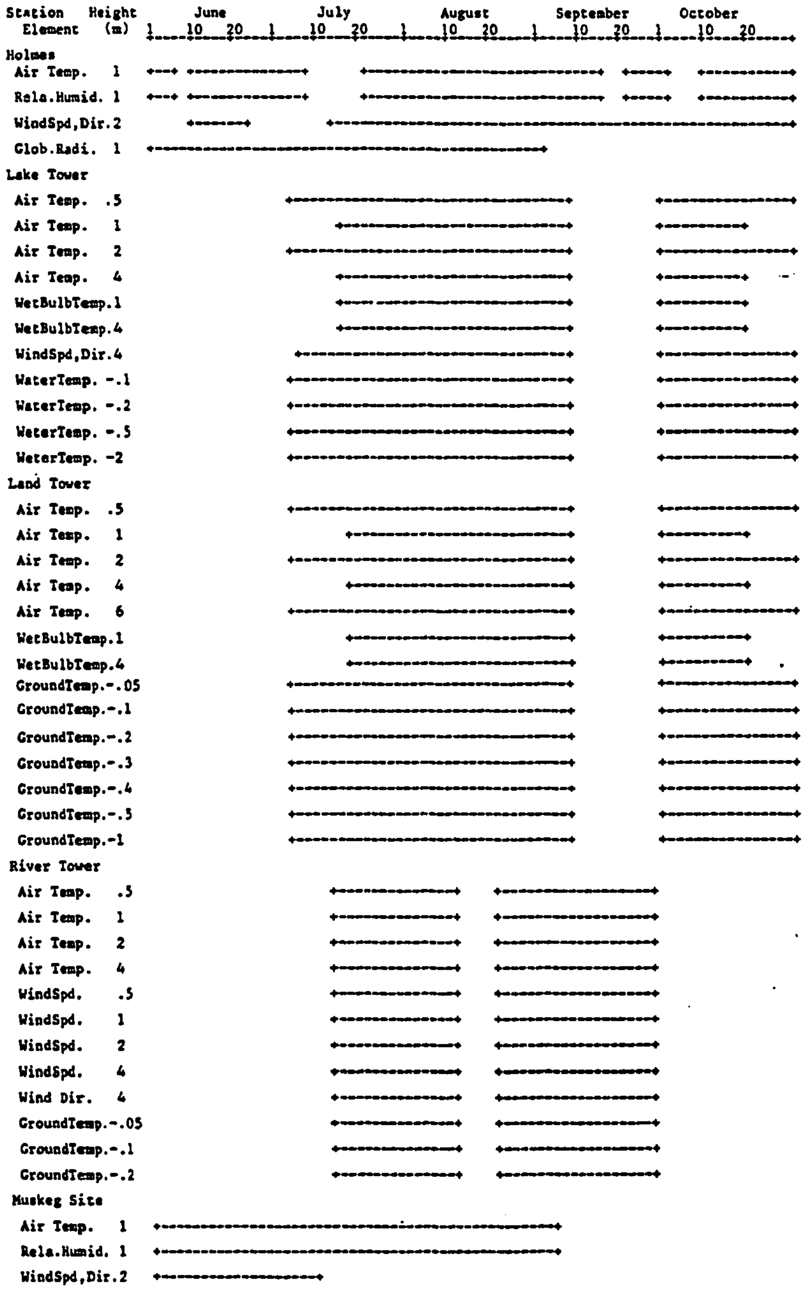


Table 1: The data taken at each station around Lake Minchumina in 1979.

IV. RESULTS AND THEIR INTERPRETATION

A General Overview of the Weather Conditions of Interior Alaska in Summer and Fall 1979

A lake's microclimate is defined as the prevailing weather condition in the lower boundary of the atmosphere which is modified by the underlying water body of a lake, and it is determined by the thermal characteristics of the lake as well as by the broad synoptic conditions to which the lake is exposed. A lake responds to its surrounding climate. From this point of view a lake can be considered as a climate recording instrument (Ragotzkie, 1978). Thus, besides the physical characteristics of Lake Minchumina, it is important to have knowledge about the general weather conditions of Interior Alaska during the summer and fall of 1979 when the experiment was conducted. If the lake is large enough, its microclimate can interact with the synoptic-scale climate. For example, Bruce and Clark (1966) found an enhanced development of thunderstorms over the surrounding terrain of large lakes. For Lake Minchumina this interaction is assumed to be negligible and we have not further investigated it.

The hours of possible sunlight in Fairbanks (65^{o}N) vary from 3 3/4 to 20 1/2 hours annually. On June 1, the sun rises at about 2:00 and sets at about 21:30 (local time, AST): the possible sunshine duration is about 19.5 hours: in midsummer the possible sunlight duration reaches 20.5 hours. At the end of October (when the measurements ended), it decreases to about 8 hours: the sun rises at about 7:00 (AST) and sets at about 15:00 (AST).

Two east-west aligned mountain ranges (Alaska Range and Brooks Range) divide Alaska into three distinctly different climatic zones: maritime, continental, and arctic. The interior of Alaska is dominated by continental climatic conditions, with great diurnal and annual temperature variations, low precipitation, a fairly low amount of cloudiness and low humidity. Yearly precipitation totals range from less than 20 to over 40 cm, with a distinct August maximum. Normal January temperatures are around -25°C, while July values are about 15°C. Summer temperatures have exceeded 30°C on occasion. Frost may occur during any month of the year, but is rare from June through mid-August (Bowling, 1979).

The local meteorological data and the deviations from the 30-year mean for two stations in the interior of Alaska in summer and fall of 1979 are given in Table 2. For a period of about 30 days, starting from June 10, the average temperatures for Fairbanks and McGrath were recorded below normal (based on the period 1930-60) and the cloud cover was above normal. These conditions made the month of June of 1979 in Interior Alaska a cold one. Precipitation in June was relatively low except in Fairbanks. Wind directions from June to August were mostly south to west with wind speeds of 2.0 to 3.0 m/sec on the average for both stations. At the beginning of July, moist air masses coming from the west caused high amounts of rainfall. Thus, July had the highest precipitation records during the summer at both stations, although the

	JUN	JUL	AUG	SEP	OCT
			Average air temp. (^{o}C)		
Fairbanks	14.1(-0.9)	16.3(+0.3)	15.8(+2.8)	8.1(+1.2)	0.2(+4.0)
McGrath	11.4(-1.8)	14.2(-0.3)	13.8(+1.9)	7.9(+1.4)	1.1(+4.8)
			Average dew-point temp. (^{o}C)		
Fairbanks	6.7	10.0	9.4	1.7	-5.0
McGrath	5.0	8.3	9.4	2.2	-2.2
			Relative humidity (%)		
Fairbanks	61*(-2)$^{!}$	66*(+1)$^{!}$	66*(-4)$^{!}$	64*(-4)$^{!}$	68*(-10)$^{!}$
McGrath	65*(±0)$^{!}$	68*(-3)$^{!}$	75*(-5)$^{!}$	67*(-13)$^{!}$	79*(-1)$^{!}$
			Precipitation (mm)		
Fairbanks	39(+3)	65(+16)	31(-25)	5(-23)	24(+5)
McGrath	28(-15)	51(-7)	47(-36)	46(-8)	23(-8)
			Resultant wind direction (^{o})		
Fairbanks	240	250	230	20	20
McGrath	190	290	190	40	50
			Average wind speed (m/s)		
Fairbanks	3.2(+0.3)$^{!}$	2.9(+0.2)$^{!}$	2.4(-0.1)$^{!}$	2.6(±0)	2.1(-0.3)$^{!}$
McGrath	2.8(±0)$^{!}$	2.4(-0.3)$^{!}$	2.1(-0.4)$^{!}$	2.6(+0.2)$^{!}$	2.2(+0.1)$^{!}$
			Cloud cover (1/10)		
Fairbanks	3.2(+0.9)$^{!}$	7.2(-0.4)$^{!}$	7.9(-0.2)$^{!}$	5.9(-1.8)$^{!}$	7.5(-0.4)$^{!}$
McGrath	8.0(+0.1)$^{!}$	7.4(-1.1)$^{!}$	7.6(-0.9)$^{!}$	6.1(-2.0)$^{!}$	8.0(-0.2)$^{!}$

Table 2: Local climatological data for 1979 of fairbanks and McGrath. () indicates the departure from normal (1930-1960). * denotes the normal was calculated by Searby(1968). ! denotes that the normal was calculated from the average air and dew-point Temperatures.

maximum precipitation usually occurs in August. But after July 10, sunny and dry days continued for about two weeks, and caused the summer's highest air temperature of 26.7°C on July 20 at McGrath, and of 27.8°C on July 24 at Fairbanks. Interior Alaska during July 1979 was close to a normal month except for the precipitation amounts at Fairbanks. From August to October, the air temperature was higher, precipitation and cloudiness were less than during the average year. In September, resultant wind directions changed from westerly to northeasterly. The air temperatures in October were markedly warmer (4-5°C) than in a normal year. Overall, Interior Alaska during summer and fall of 1979 was warmer and drier and less cloudy than during an average year, except in June which was slightly colder than normal.

B Results Observed at Lake Minchumina in Summer and Fall 1979

a) Wind Regime

A well-known feature of the lake's climate is its lake-land breeze. Therefore, the wind regime at Lake Minchumina shall be investigated first.

Figure 3 shows the numbers of cases of wind directions for all wind speeds at Holmes Station on the west side of the lake from May 16 - September 14, 1979 (data missing during the period June 27 - July 14). The data were taken on an hourly basis. The number of cases of wind directions shown in Fig. 3a has four modes: 90^{o}, 150^{o}, 240^{o}, 330^{o}. Although the prevailing wind direction is 240^{o}, the other three peaks are comparable in magnitude. Figures 3b and c are presented in order to determine the contributions of different wind speeds, and the time of the day to the distribution. Fig. 3b shows the numbers of cases for winds with speeds less than 0.5 m/sec for the daytime (9:00 - 20:00, AST) and the nighttime (21:00 - 8:00, AST). Fig. 3c shows the same analysis for wind speed higher than 0.5 m/sec. Thus the four modes of the distribution can be explained in the following way: mode I (90^{o}) is caused mainly by high speed winds ($>$ 0.5 m/sec) during the daytime, mode II (150^{o}) by low speed winds ($<$ 0.5 m/sec) during the daytime, mode III (240^{o}) by high speed winds during the daytime, and mode IV (330^{o}) by weak winds during the nighttime. Considering that Holmes Station is located at the west side of the lake, it is possible that the land

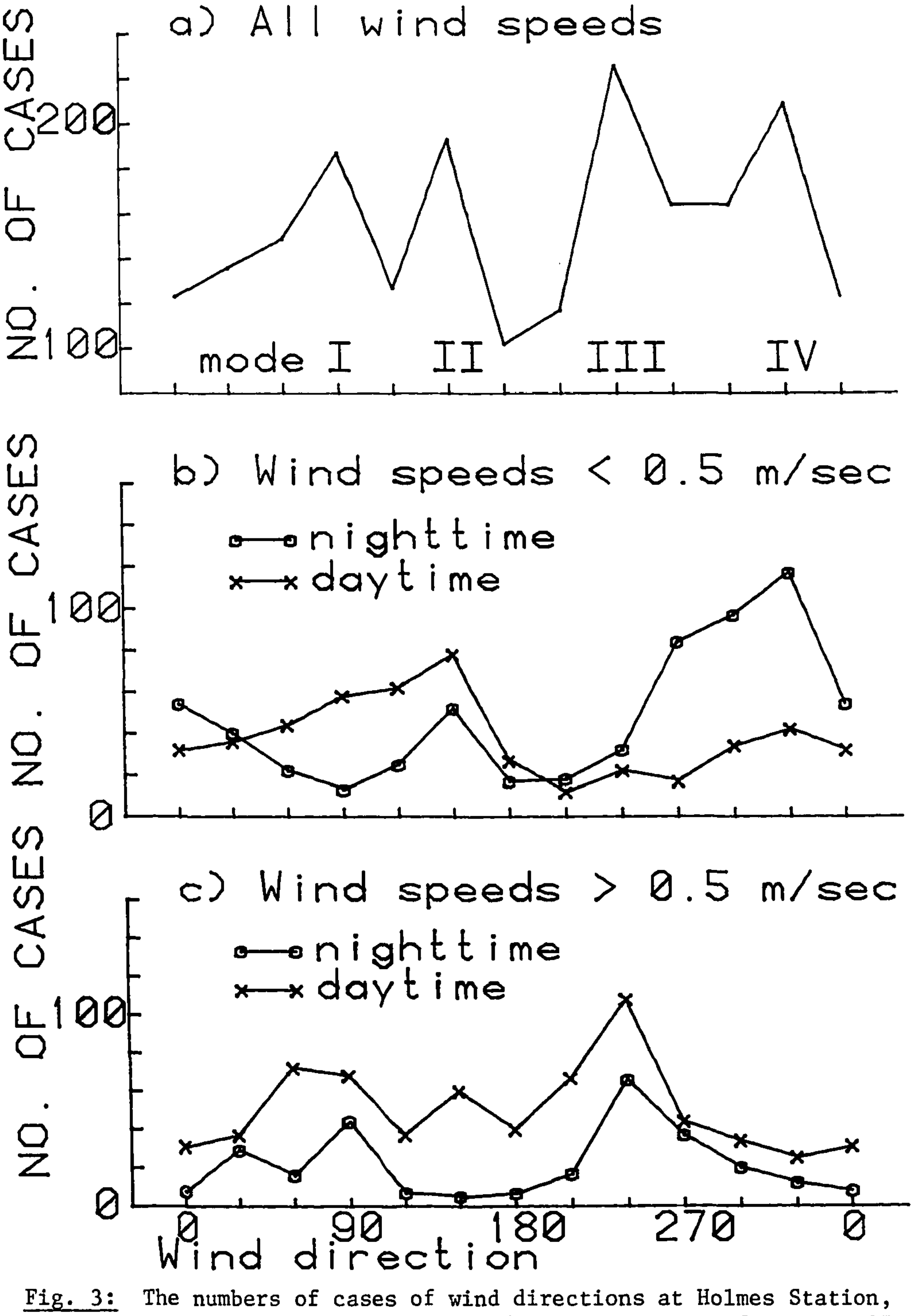


Fig. 3: The numbers of cases of wind directions at Holmes Station, May 16 - September 14, 1979 (Data are missing from June 27 - July 14.).

breeze (weak wind at night) contributes to the mode IV (there might be additional gravitational winds in mode IV because Holmes Station is located on a slope with a gradient of about 1/12) and the lake breeze (weak wind during daytime) to mode II. Modes I and III are found mainly for high speed winds (> 0.5 m/sec) which can be considered to be a part of the synoptic regime. If all low speed winds (< 0.5 m/sec) during the daytime from the directions 30° - 180° are considered to be a part of the lake breeze, the contribution of the frequency of the occurrence of the lake breeze amounts to 13% of all winds observed which is slightly less than the values (14% at 9h, 24% at noon) reported by Yoshino et al. (1970). If all low speed winds during the nighttime from the directions 210° - 360° are assumed to be a part of the land breeze its contribution amounts to 20% of all winds observed.

Figures 4a, b, and c show the same analysis as Figures 3a - c but for Lake Tower Station. The distribution of the numbers of cases shown in Fig. 4a does not have four modes like Fig. 3a, but has the highest peak in easterlies. From Figures 4b and c, can be seen that this peak is dominated by the weak easterlies during the nighttime which might be the land breeze. But in general Figures 4b and c are not as conclusive as Figures 3b and c.

Additional information can be obtained by comparing the wind regimes at Holmes Station and Lake Tower Station. Figures 5a - h show the corresponding distributions of wind directions observed at Lake Tower Station for a given wind direction at Holmes Station. The nighttime distribution is indicated by shaded bars; unshaded bars

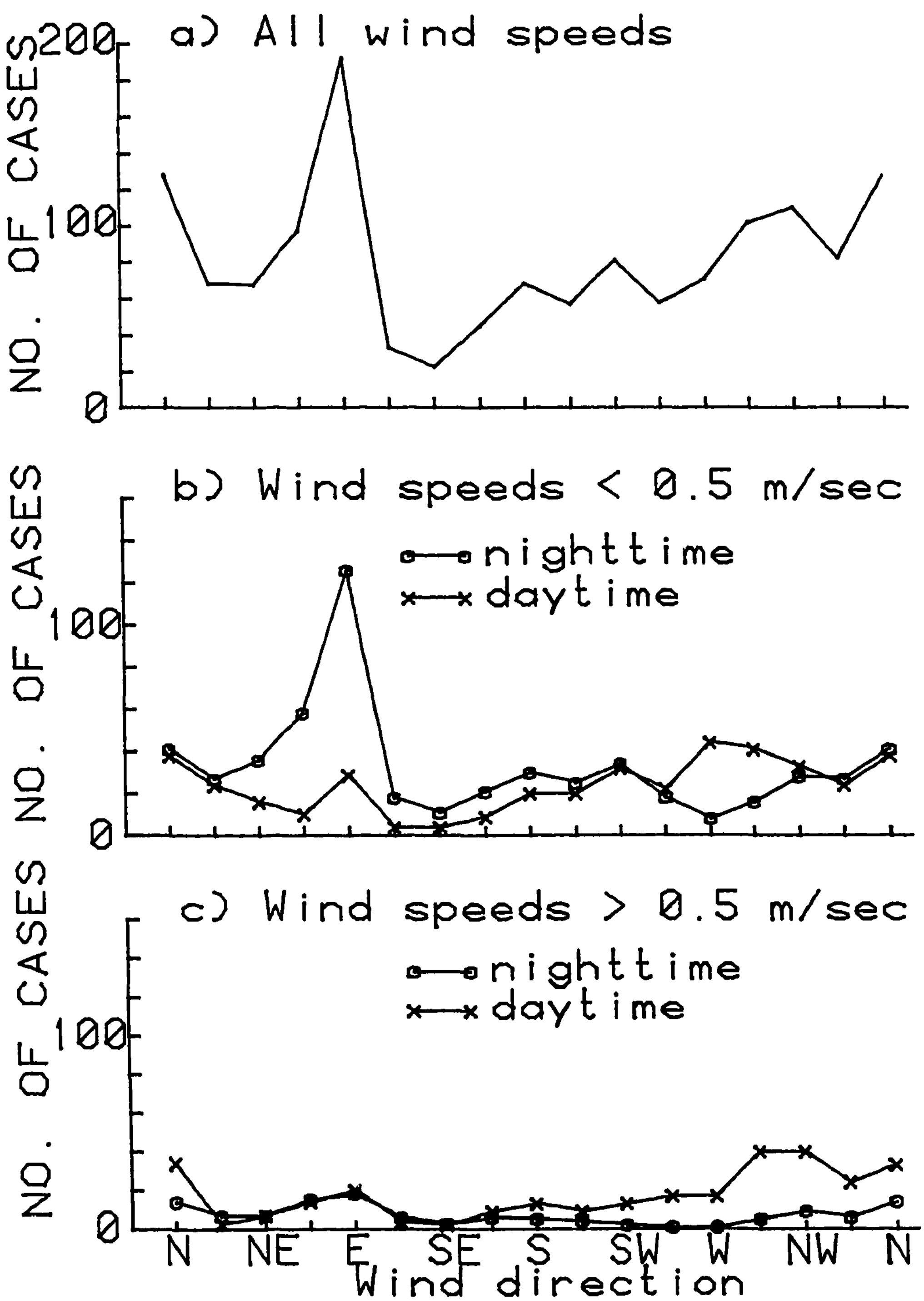


Fig. 4: The numbers of cases of wind directions at Lake Tower Station from July 6 to September 8 in 1979.

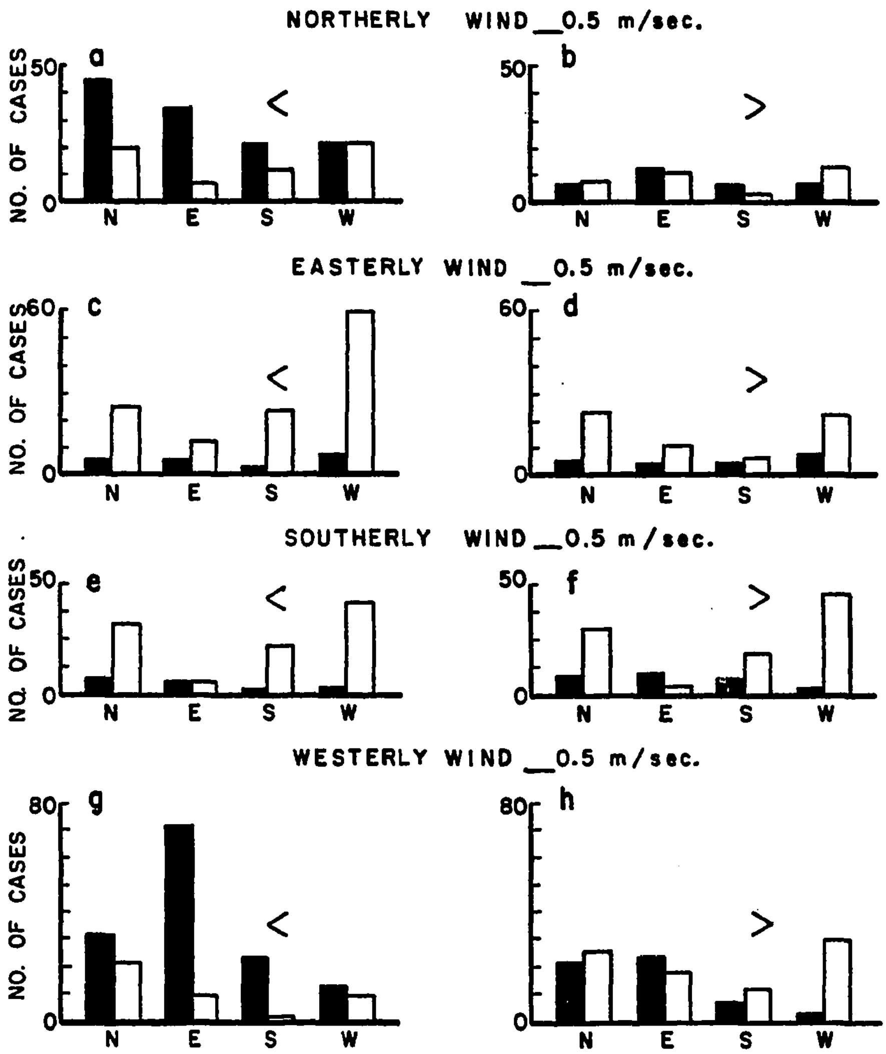


Fig. 5: The numbers of cases of wind directions at Lake Tower Station for given wind directions at Holmes Station.

represent the daytime conditions. These Figures show that there are complicated wind direction distributions around the lake; however, there are a few interesting results:

a) When there are weak (< 0.5 m/sec) easterlies (NE - SE) during the daytime (9:00 - 20:00 AST) at Holmes Station (Fig. 5c), westerlies (SW - NW) are dominant at Lake Tower Station. This agrees with a lake breeze effect causing easterlies at Holmes Station and westerlies at Lake Tower Station.

b) Similarly, when there are weak westerlies at Holmes Station (Fig. 5g) during the nighttime, a maximum of easterlies is found at Lake Tower Station in agreement with a land breeze.

c) When there are weak northerlies (NW - NE) at Holmes Station (Fig. 5a), the nighttime northerlies at Land Tower Station are also pronounced, a part of which might be explained by a land breeze effect.

d) For the weak southerlies (SE - SW) at Holmes Station (Fig. 5e), the daytime westerlies at Lake Tower Station are dominant, which is explainable also as a component of the lake breeze.

e) For all directions of strong winds (> 0.5 m/sec) at Holmes Station, which should represent the synoptic regime, there are not the corresponding wind directions found at Lake Tower Station except for strong west winds. This might be explained by the influence of the terrain around Lake Minchumina on the synoptic wind regime.

Figures 6a, b, and c show the distribution of the numbers of observations for given wind directions at River Tower Station during the period June 19 - September 24. It shows the highest peak for westerlies

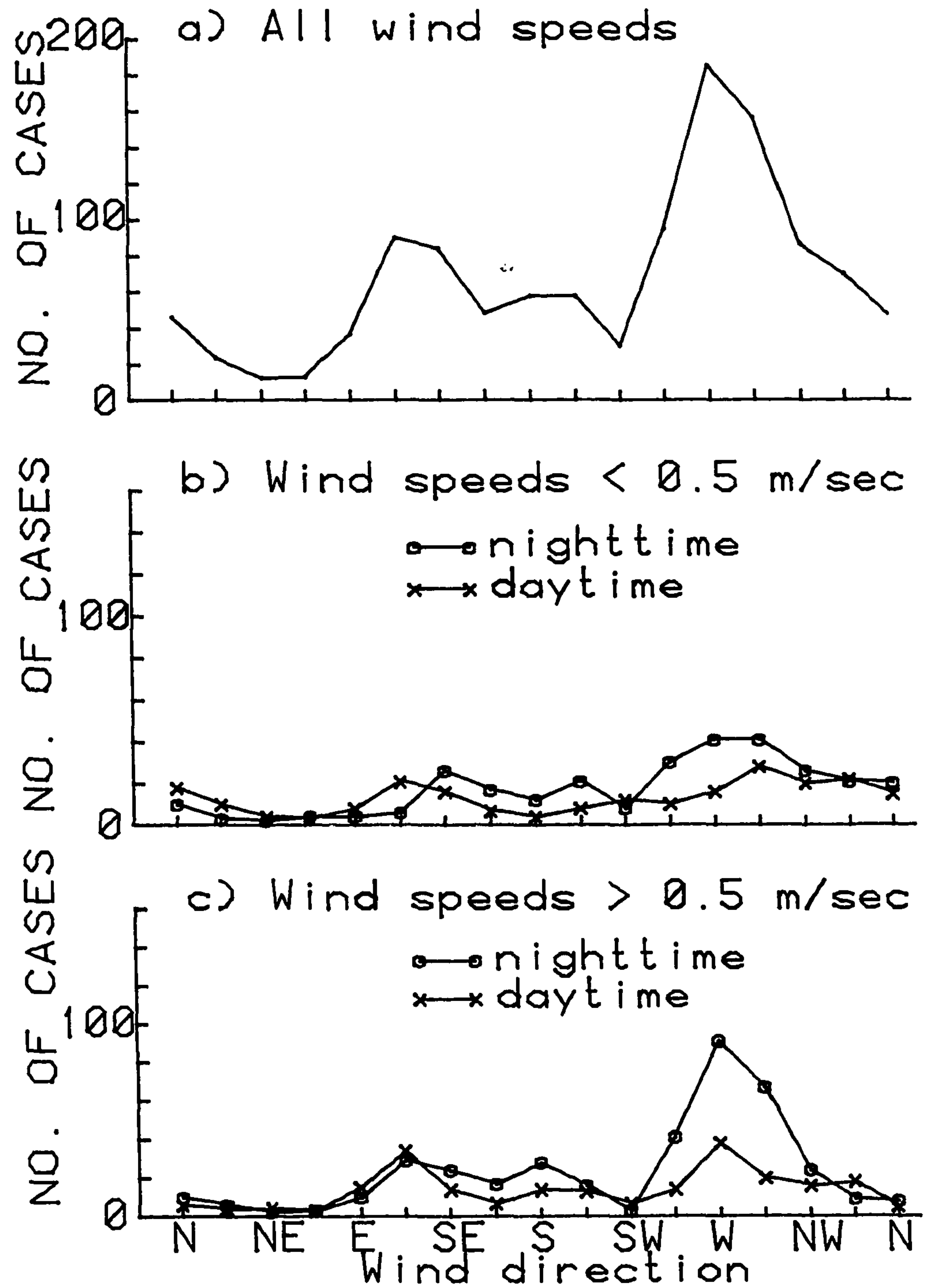


Fig. 6: The numbers of cases of wind directions at River Tower Station from July 19 to September 24 in 1979.

in Fig. 6a, and does not show any land-lake breeze in Figures 6b and c. A similarly shaped distribution for Muskeg Site is found. This means that there is no extension of the land-lake breeze 6 km away from the lake's shore.

Overall, the prevailing wind direction is west around the lake except that easterlies are dominant at Lake Tower Station. For the shore stations, the wind regime is very complicated. However, the land-lake breeze is clearly found.

b) Air Temperature

The lake effect can be most easily detected by its influence on the ambient air temperature. The moderating effect es expected to be more pronounced at sites close to the lake. First, the overall seasonal temperature changes at Holmes Station due to the synoptic conditions during the summer and fall 1979 will be investigated. Then, the lake effect will be shown by computing the temperature differences between stations close to and distant from the lake.

Seven days running averaged daily mean (arithmetic mean of daily maximum and minimum) air temperatures at Holmes Station (western shore of the lake) are shown in Fig. 7 which indicates the general trend of the air temperature at Lake Minchumina in summer and fall 1979. The actual water temperatures measured at Lake Tower Station are added to this figure. According to residents of Lake Minchumina, the break-up of the lake ice was in the middle of May. From June 10 to July 10 the air temperatures were lower than one would expect for June, but this observation agrees with the below normal air temperatures measured at Fairbanks and McGrath for June (see Chapter IV A). When the measurements of the water temperatures started (at the beginning of July), the lake's water temperature was already higher than the daily mean air temperature at Holmes Station. At the end of August the air temperature decreased rapidly. The lake ice formed around October 20 according to residents.

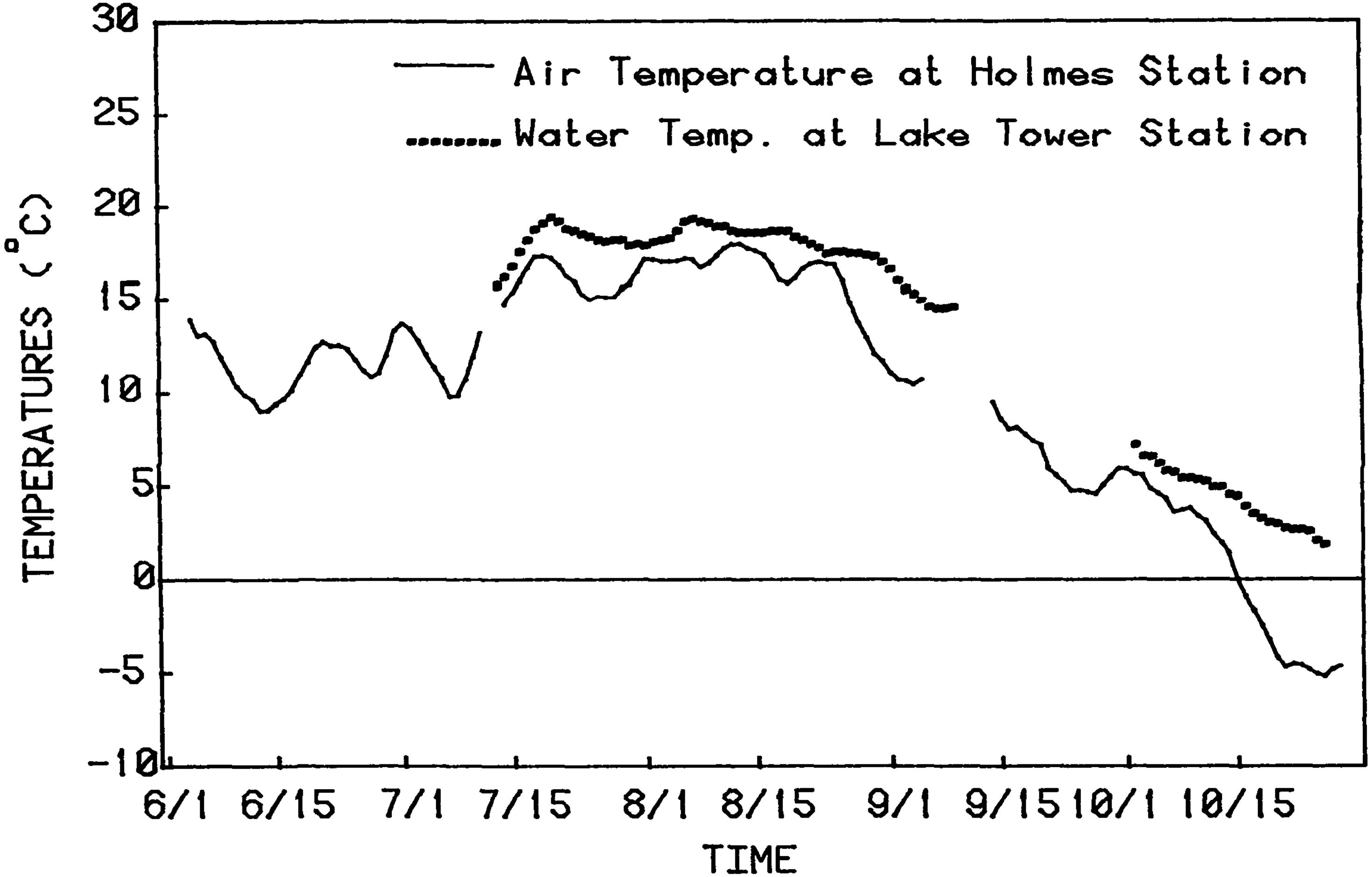


Fig. 7: Seven days running average daily mean air temperature at Holmes Station and water temperature at Lake Tower Station.

It is a well-known fact that air masses which travel across the lake are affected by the lake, thus their characteristics are different from those air masses that move over land. Therefore, the lake effect on the air temperature at the shore stations is assumed to depend upon the direction of the winds. Table 3 shows the mean air temperature differences between Holmes Station and Land Tower Station and between Holmes Station and Muskeg Site grouped for different wind directions during the period of July 14 - September 8, 1979. The values were calculated only for those cases when the wind directions at the three stations were the same. When there are easterly winds at all three stations, the air temperature at Land Tower Station is 0.8°C warmer than at Muskeg Site, and the temperature at Holmes Station is 2.4°C warmer than at Land Tower Station. It seems that the air mass has been warmed up during its travel from Muskeg Site to Land Tower Station, and then the air has been further warmed up by the lake water while traveling over it. In the case of westerlies at all three stations, the same tendency is observed: the air temperature at Land Tower Station is 1.4°C higher than at Holmes Station, and the temperature at Muskeg Site is 1.8°C lower than at Land Tower Station. For the northerlies and the southerlies, the temperature differences are not significant enough to be mentioned. The reason the temperature difference between Holmes Station and Lake Tower Station for easterlies is larger than for westerlies is not clear. However, it might be worthwhile mentioning that the easterlies at both stations are more frequently observed at the

	Land Tower Station	Muskeg Site
Northerly winds	+ 0.8	- 0.5
Easterly winds	- 2.8	- 3.6
Southerly winds	+ 0.4	- 0.3
Westerly winds	+1.2	- 0.6

Table 3: Mean air temperature differences with Holmes Station for a given wind direction.

end of August and the beginning of September than the earlier period of July and the beginning of August.

Daily mean air temperature differences between Land Tower Station and Muskeg Site and between Holmes Station and Muskeg Site are shown in Figures 8a and b, respectively. According to Fig. 8a, Land Tower Station is slightly warmer (by 1.2°C on the average) than Muskeg Site from the beginning of July to the middle of August. The difference increases towards the fall, and reaches its maximum of 6°C on September 5, 1979. Fig. 8b shows the same tendency as Fig. 8a except that Holmes Station is sometimes slightly cooler than Muskeg Site from the beginning of July to the beginning of August. From these two comparisons an average value of the differences is found for the period of the beginning of August to the beginning of September of about 3° - 4°C, which is comparable to the same value reported by Phillips and McCulloch (1971) for the proximity of the Great Lakes.

Figure 9 shows the daily mean air temperature differences between Land Tower Station and Holmes Station. From the beginning of July to the beginning of August Land Tower Station is warmer than Holmes Station, then the relationship reverses but the differences are smaller. The lower air temperatures at Holmes Station compared to Land Tower Station in the first period seem to explain the unexpected behavior of the graphs in Figures 8a and b for the first period. To explain why Land Tower Station is warmer than Holmes Station in the first period and cooler in the second period the ratio of number of occurrences of winds blowing from the lake to the total number of winds observed was

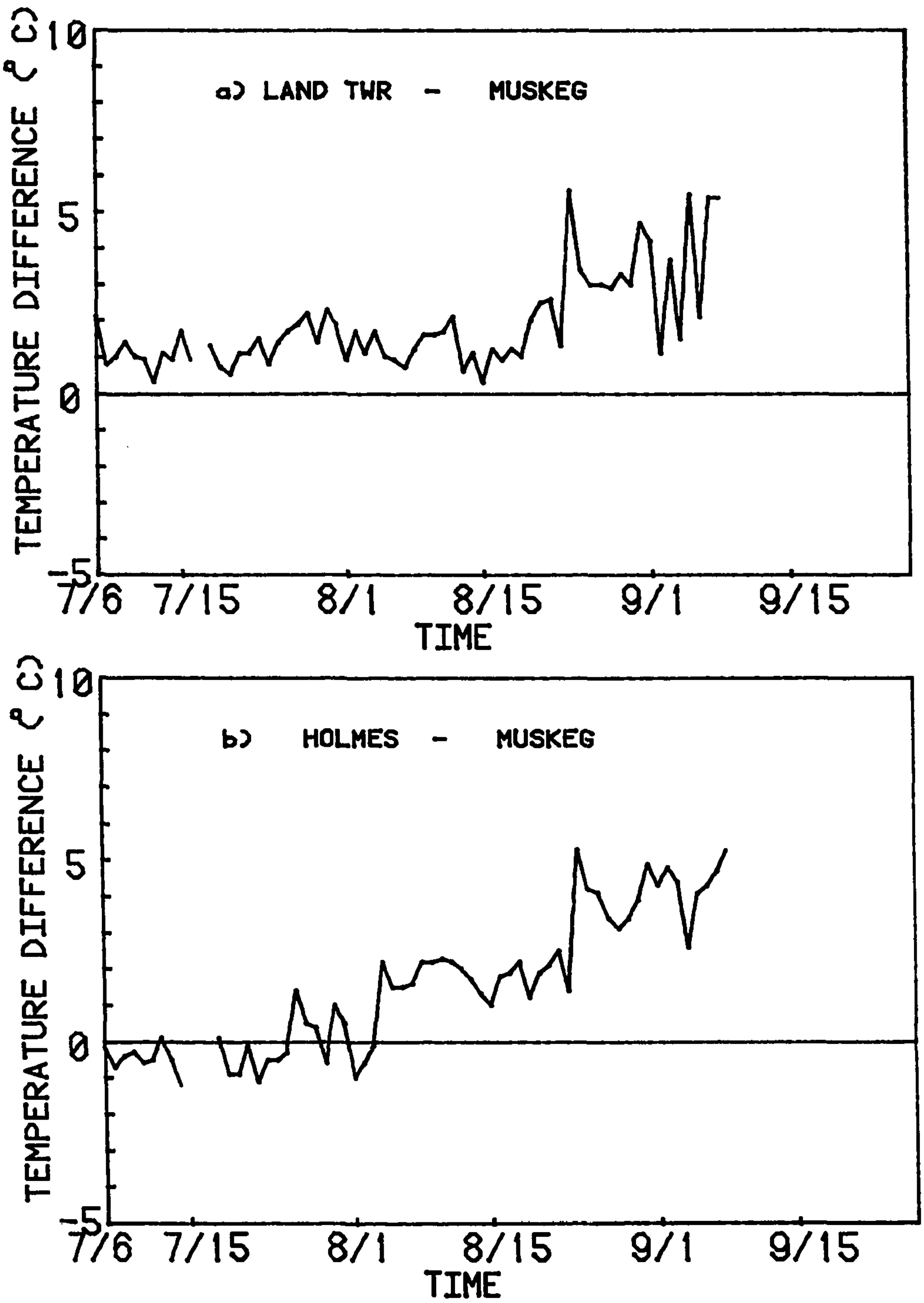


Fig. 8: The daily mean air temperature differences- a) between Land Tower Stationand Muskeg Site and b) between Holmes Station and Muskeg Site at Lake Minchumina in 1979.

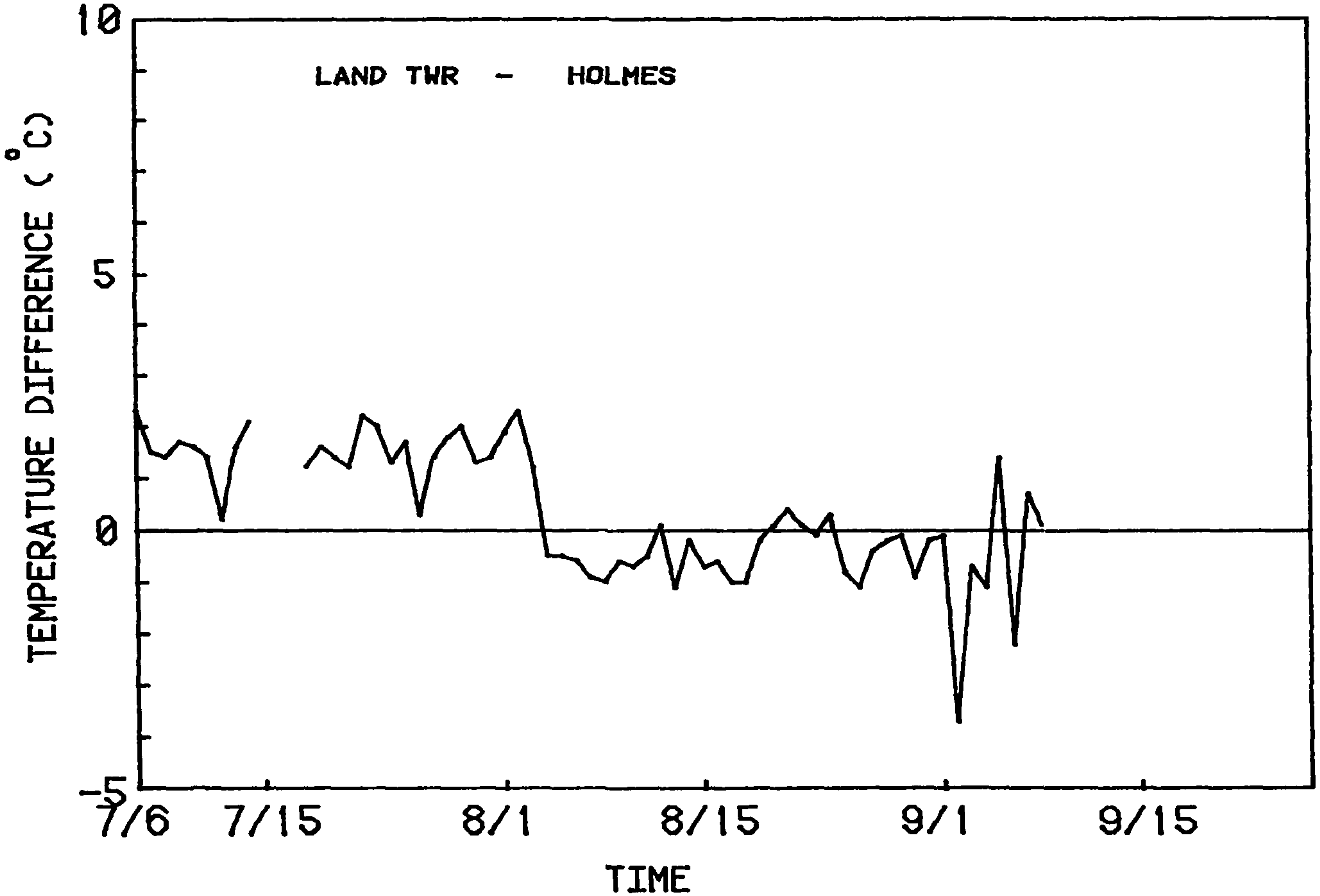


Fig. 9: The daily mean air temperature differences between Land Tower Station and Holmes Station at Lake Minchumina in 1979.

calculated for the two stations. The ratio is 35% at Holmes Station and 53% at Land Tower Station for the first period. Since the air mass travelling over the lake is warmer on the average than that moving only over land, the higher ratio of winds blowing from the lake causes a higher air temperature at Land Tower Station than at Holmes Station. For the second period the ratios increase to 46% at Holmes Station and decrease to 48% at Lake Tower Station. This might explain why the relationship reversed for the second period, but hardly explains why Holmes Station is slightly warmer than Land Tower Station in the second period for the same ratios of winds blowing from the lake.

The presence of a body of open water (lake or reservoir) reduces extremes in ambient air temperature due to its large specific heat capacity and mixing, which occurs in the water body. Figures 10a and b show the differences of the spread of diurnal air temperatures (T_{max} - T_{min}) between the lake shore stations (Land Tower and Holmes Station) and Muskeg Site. Both Figures show that the diurnal air temperature spread at Muskeg Site is much bigger than at the lake shore stations. This is considered to be the moderating effect of the lake on the air temperature.

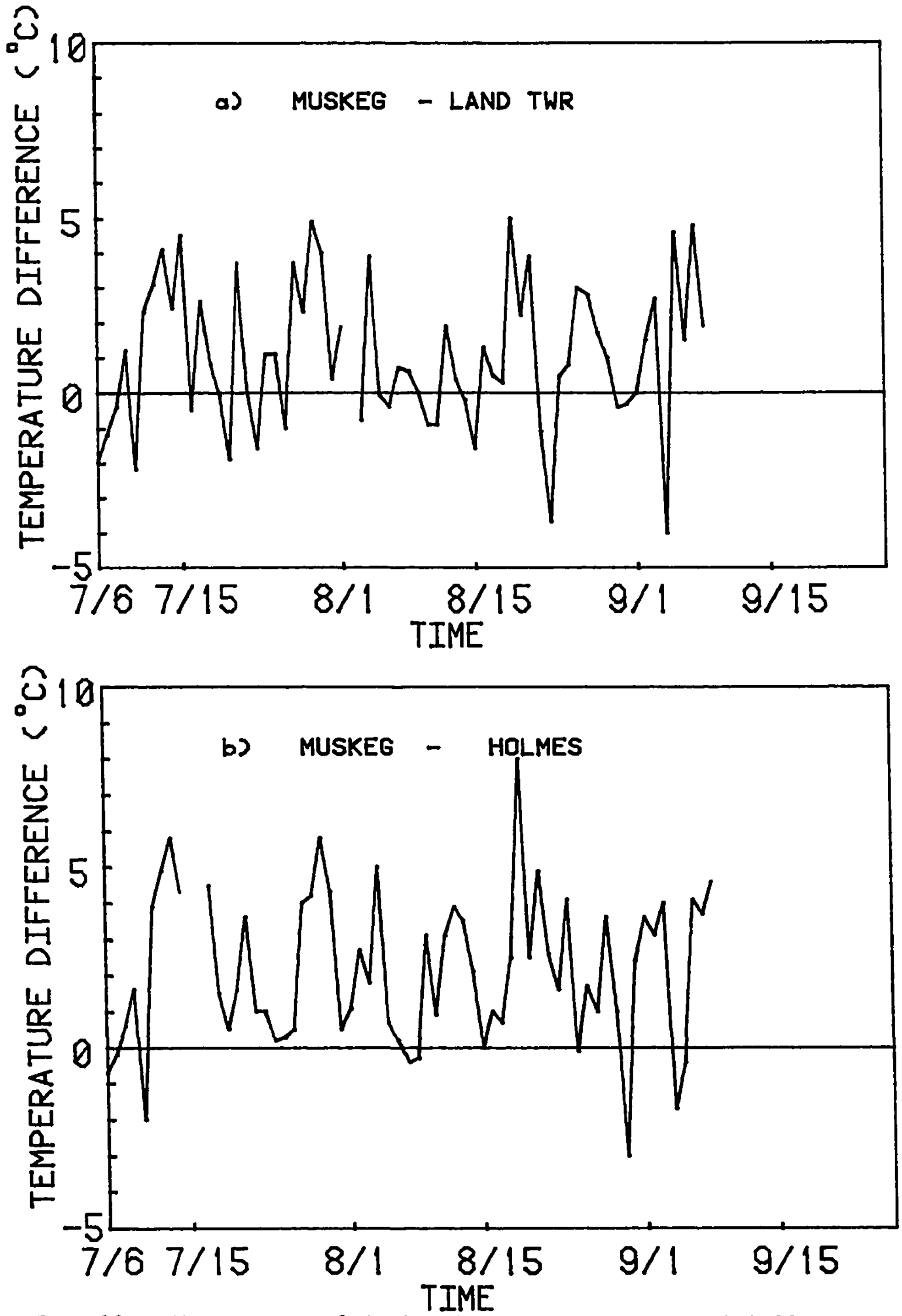


Fig. 10: Time series of daily air temperature spread differences a) between Muskeg Site and Land Tower Station and b) between Muskeg Site and Holmes Station.

c) Frost-free Season

The moderating effect of a lake on annual air temperature variation can easily be found in the length of the frost-free season. Table 4 shows the duration of the mean length of the growing season for some selected stations in Interior Alaska (after Searby, 1968). The location of these stations in relation to Lake Minchumina is shown in Fig. 2. Table 4 shows the last day in spring and the first day in fall of the occurrence of $0^{o}C$ in daily minimum air temperatures, the mean number of days between those dates, and the number of years of record. Caution is advisable in comparing these data with each other, because the average is based on different lengths of observations and different time periods which could not be found in the original paper. According to Table 4, Lake Minchumina (data obtained at the airport, see Fig. 2) has the longest frost-free season of the seven weather stations in Interior Alaska. Lake Minchumina has 117 frost-free days on the average, which compares favorably with 81 for Tanana, 104 for Fairbanks, 43 for Manley Hot Springs, and 106 for McGrath. Even the Matanuska Valley (located south of the Alaska Range), agricultural center of Alaska, has only 111 frost-free days and Big Delta 114. In fact, these last two stations in Table 4 have growing seasons that are shorter than Lake Minchumina's by averages of 6 and 3 days, respectively. This could be interpreted that Lake Minchumina causes some effects, which make the growing season longer than that of the other stations in Interior Alaska.

Station	occurrence of $0^{o}C$ last day in Spring	first day in Fall	frost-free period (days)	length of of obs. (years)
Lake Minchumina	May 20	Sept. 14	117	10
Tanana	May 30	Aug. 19	81	10
Fairbanks	May 20	Sept. 1	104	30
McGrath	May 23	Sept. 6	106	10
Manley Hot Springs	June 19	Aug. 1	43	9
Matanuska Valley	May 25	Sept. 13	111	28
Big Delta	May 16	Sept. 7	114	10

Table 4: Duration of the mean length of the growing season for some selected stations in Interior Alaska (after Seaby,1968).

In Spring 1979 the minimum temperature at Holmes Station dropped to the freezing point for the last time on May 18. In fall, on September 16, the minimum temperature fell again to the freezing point. Therefore, the frost-free period comprised 121 days, 4 days longer than the average which is shown in Table 4. In Fairbanks, the minimum air temperature reached the freezing point for the last time on May 19 and again on September 17. In 1979, Fairbanks had 121 frost-free days, 17 days more than the average of Table 4. McGrath recorded 124 frost-free days, 18 days longer than the average value. This longer frost-free season might have been a result of the warmer summer and fall of 1979. The lesser departure of the frost-free period at Holmes Station compared to Fairbanks and McGrath is in agreement with the results of Phillips and McCulloch (1971), who found that the annual variability in the dates of frost occurrence is less along a lake shore.

d) Dew-Point

In the literature, only a few papers discuss the moisture content in the air mass above and around a lake although there are many reports of evaporation from lakes. Naturally it is expected that the moisture content in the air mass (only the ice-free period is of interest) is normally higher over the lake and its closest surroundings than further inland. Richards and Fortin (1962) reported for the Great Lakes that this is true except for May and June, when lakes are colder than the surrounding land areas. D'yakonov and Reteyum (1965) reported a water-vapor pressure difference between the lake and the surrounding land area of up to 2 mbs. during the night.

Dew-point temperature is defined as the temperature to which moist air must be cooled during a process, in which pressure and mixing ratio remain constant, in order to become just saturated with respect to water. In other words, the dew-point temperature indicates the amount of water-vapor per unit volume of air mass.

In this section, the trend of the dew-point temperature for the period June 1 - October 15 at Holmes Station will be shown. Second, the dependency of the dew-point temperatures at Holmes Station and at Land Tower Station on the wind direction will be discussed. Then a comparison of dew-point temperatures between the lake shore stations and Muskeg Site will be made.

Figure 11 shows the average daily dew-point temperature and daily precipitation amount at Holmes Station from June 1 to October 14, 1979.

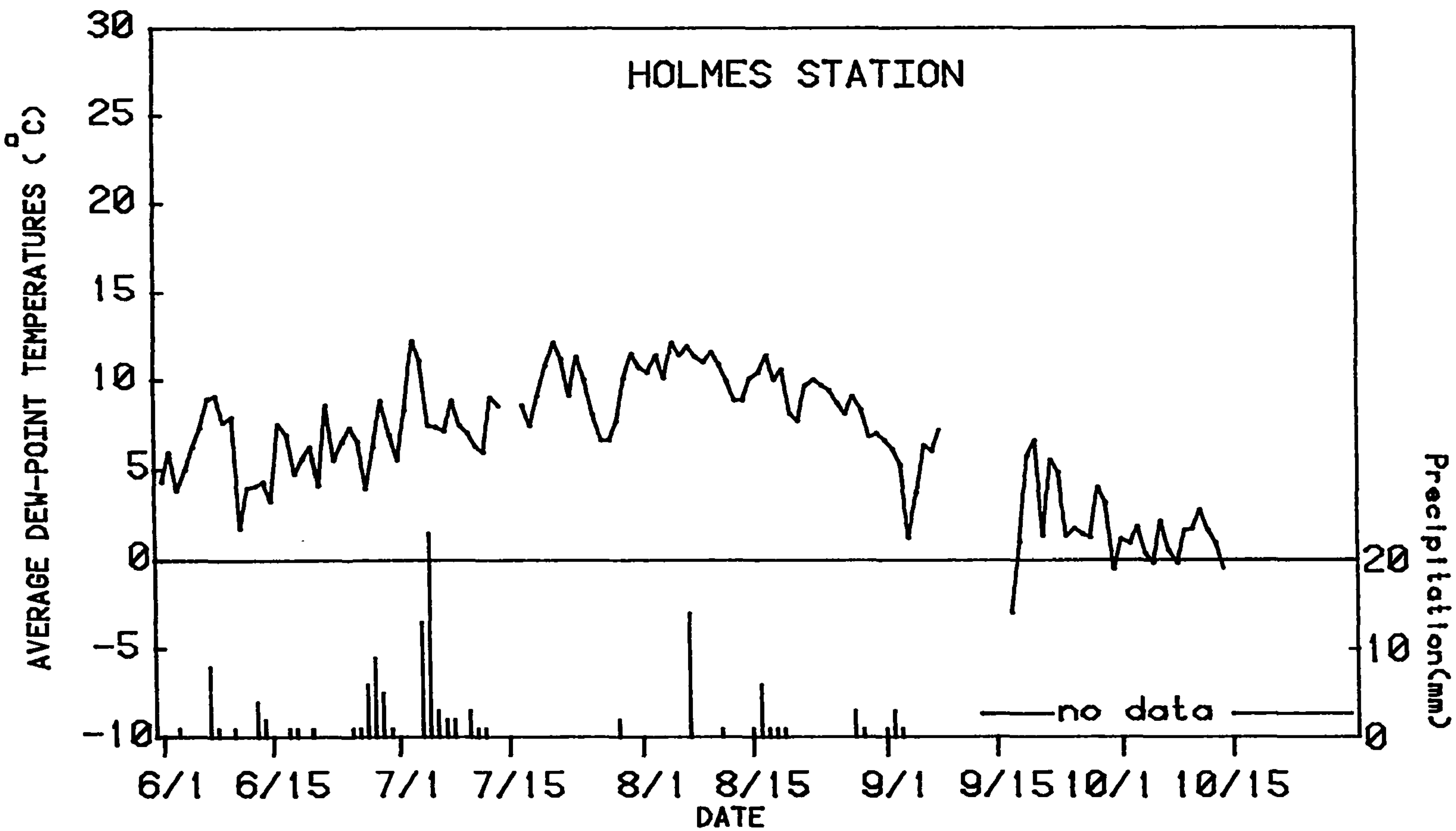


Fig. 11: The average daily dew-point temperature and daily precipitation amount at Holmes Station at Lake Minchumina in 1979.

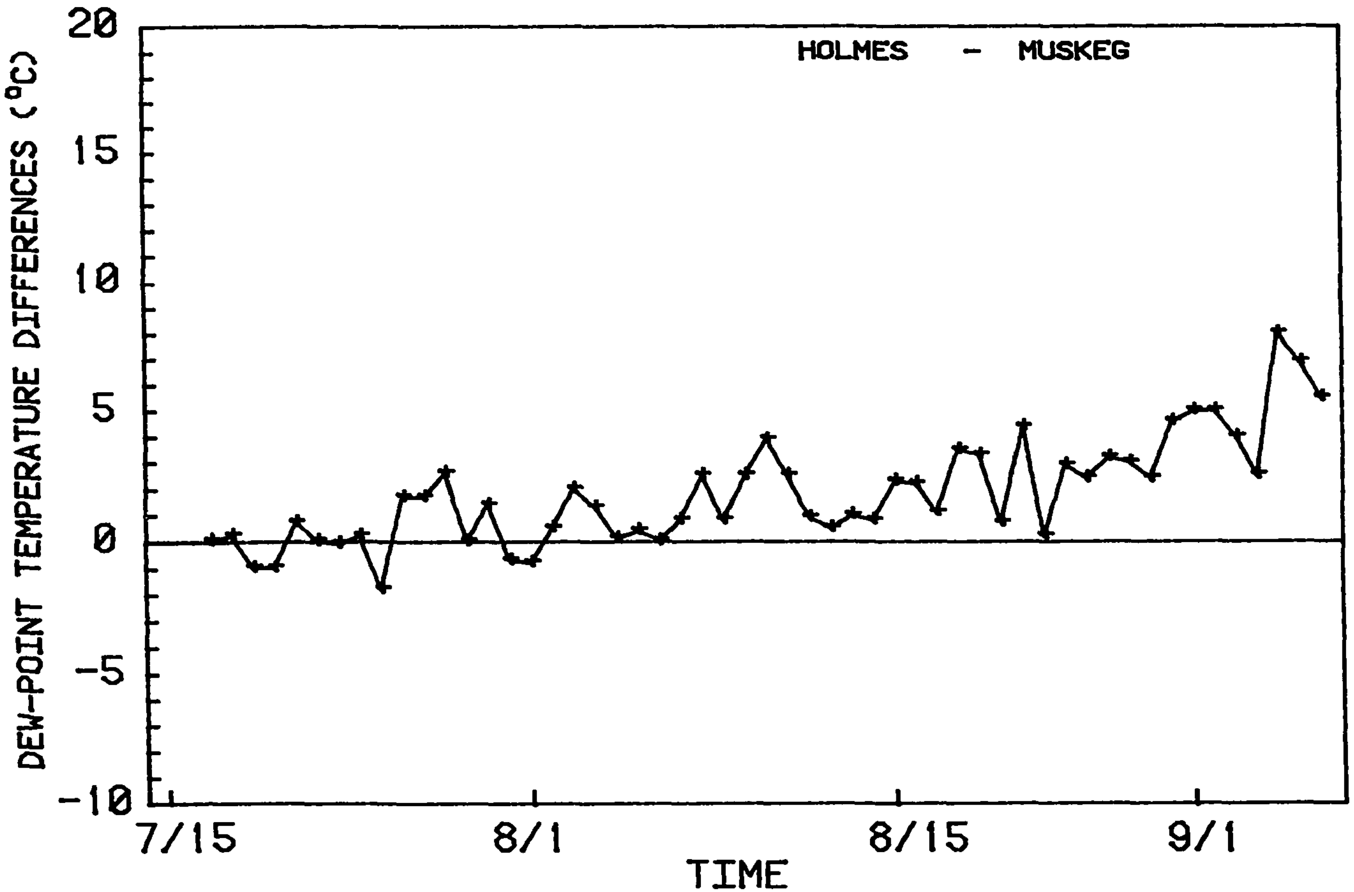
The general trend of the dew-point temperature shows low values at the beginning of June, increasing gradually, reaching its maximum at the middle of August. Each irregular high peak of the dew-point temperature seems to coincide with precipitation. Similar trends were observed between Land Tower Station (east side of the lake) and Muskeg Site (far inland).

The dependency of the dew-point temperatures on wind direction calculated is shown in Table 5. When there are westerly winds (SW - NW) at Holmes Station (west side of the lake) and Land Tower Station (east side of the lake), the water-vapor pressure difference between these two stations is 3.0 mbs. on the average and corresponding dew-point temperature difference is 3.3^{o}C. The air mass arriving at Lake Tower Station apparently has obtained moisture from the lake while travelling over the lake. The corresponding result is found for easterlies. The reason the dew-point temperature differences for easterlies are lower than for westerlies is that the easterly winds are mostly observed at the end of August and the beginning of September when the air temperatures are generally lower than those at the first part of the period.

Comparison of the dew-point temperatures between all stations around the lake were done. The daily average dew-point temperature differences between Holmes Station (shore) and Muskeg Site (inland) are shown in Fig. 12. In July the differences are almost zero. Then, they increase towards the fall, reaching the largest value of 8^{o}C on September 5, 1979. The same tendency is found in the comparison between

	Holmes Station	Land Tower Station	Differences
		W E S T E R L I E S	
Water vapor preeure (mb)	12.1	15.1	3.0
Dew-point temperature (°C)	9.8	13.1	3.3
		E A S T E R L I E S	
Water vapor pressure (mb)	12.3	10.9	1.4
Dew-point temperature (°C)	10.0	8.2	1.8

Table 5: The dependency of the dew-point temperature and the water vapor pressure on a given wind direction at Holmes and Land Tower Stations.



<u>Fig. 12:</u> The daily average dew-point temperature differences between Holmes Station and Muskeg Site at Lake Minchumina in 1979.

Land Tower Station and Muskeg Site. This higher moisture content in the air mass at the shore stations than at the stations further inland might be caused by the continuous evaporation from the lake (see Chapter IV B e). The almost equal dew-point temperatures found at Muskeg Site and Holmes Station during July might have occurred because the heavy rainfall at the end of June and the beginning of July (see Fig. 11) supplied an additional amount of moisture to the air mass at Muskeg Site.

The average dew-point temperatures from August 1 to September 7, when there is a detectable lake effect on the moisture content of the air mass, are 10°C at Holmes Station and 7.4°C at Muskeg Site. Its difference of 2.6°C is equivalent to 2 mbs. in water-vapor pressure, which is the same value reported by D'yakonov and Reteyum (1965).

e) Heat Budget

It is important for a better understanding of a lake's effects on its surroundings to calculate the heat budget components of the lake. As mentioned in the introduction, the thermal behavior of a lake is affected by the synoptic conditions and has influence on the lake's microclimate. The calculation of the heat balance components of the lake gives a quantitative estimate of the thermal behavior of the lake.

First, the methods for calculating the heat budget terms will be introduced, then their results and the expected errors in each term and the data will be disscussed. Third, the heat balance components of Lake Minchumina will be compared with those of some temperate lakes.

The heat balance of an unfrozen lake (Sellers, 1965) can simply be expressed as

$$Q_T = R_N + Q_S + Q_L + Q_{misc} \tag{1}$$

where Q_T = rate of change of heat stored in the lake,

R_N = net radiation,

Q_S = sensible heat exchange,

Q_L = latent heat of evaporation,

Q_{misc} = miscellaneous energy fluxes.

All rates will be expressed as cal cm^{-2} day^{-1}.

The equation states simply that a change in the lake's heat content is equal to the algebraic sum of radiation processes, evaporation, conduction to the atmosphere, and other terms, such as advection in the lake and heat conduction from the water to the ground at the bottom of the lake. Heating from chemical and biological processes is neglected. All terms calculated are listed in the Appendix A.

1) Rate of Change of Heat Stored in the Lake, Q_T

Heat gain or loss by a column of water of depth D is given by

$$Q_T = c\rho_w \int_{-D}^{0} \frac{\partial T}{\partial t}\, dz \qquad (2)$$

where c = specific heat capacity of water,

ρ_w = water density,

T = water temperature,

t = time,

dz = depth element.

The density and specific heat capacity are assumed with sufficient accuracy to be equal to unity (Ragotzkie, 1978).

2) Net Radiation

Net radiation is the algebraic sum of incoming net solar radiation and outgoing thermal emitted radiation,

$$R_N = (1 - \alpha)\, R_S + R_L \qquad (3)$$

where R_N = net radiation in cal cm^{-2} day^{-1},

R_S = incoming vertical flux of solar radiation in cal cm^{-2} day^{-1},

α = mean daily shortwave reflectance or albedo,

R_L = net longwave radiation emitted to the atmosphere in cal $cm^{-2}day^{-1}$.

The quantity $(1 - \alpha)$ R_S represents the net shortwave radiation values received at the water surface. Incoming solar radiation values were measured at Holmes Station (west side of the lake) by a Robitzsch-type pyranograph. This pyranograph was calibrated by the calibrated pyranometer at the roof of the Geophysical Institute, University of Alaska, in summer 1978. The error in daily sums of incoming solar radiation was estimated to be about 10%.

The shortwave albedo for an inland water is 5-10% (Smithonian Meteorological Tables). When skies are clear, the albedo of a water surface is 2-3% for solar altitudes greater than 45^o increasing to 40% at 5^o (Munn, 1964). In the middle of June, 3 hours out of the 19 hours of possible sunshine duration have a solar altitude greater than 45^o, and 30% of the daily total of the solar radiation was received during this period. At the end of July, the solar altitudes do not exceed 45^o. Considering these factors, a value for the albedo of 10% was chosen. The error of the value of the net shortwave radiation caused by this rough estimate is probably within 10%.

Net longwave radiation was estimated by using an equation based on a formula proposed by Jensen, Robb, and Franzoy (1970) (see Appendix B for derivation). The formula of the equation is

$$R_L = (1.31 \frac{R_S}{R_{SO}} - 0.31) R_{LO} \tag{4}$$

where R_S = daily solar radiation,

R_{SO} = daily solar radiation on a cloudless day,

R_{LO} = net longwave radiation on a cloudless day.

The daily solar radiation on a cloudless day, R_{SO}, was derived from the table computed by Bolsenga (1964). He calculated R_{SO} as a function of the precipitable water vapor content of the atmosphere and the dust attenuation of radiation. These are assumed to be 2.0 cm and 0.06, respectively.

The net longwave radiation R_{LO} on a cloudless day was estimated by using an equation based on work done by Jensen, Robb, and Franzoy (1970). The form used is

$$R_{LO} = (1 - \alpha_w) (a + b\sqrt{e_d}) \sigma \frac{T^4_{max} + T^4_{min}}{2} - R_W \tag{5}$$

where α_w = reflectivity of water for longwave radiation (3%),

e_d = saturation vapor pressure at mean dew-point temperature in mb,

σ = Stefan-Boltzmann constant in cal cm^{-2} day^{-1} ^{o}K^{-4},

a,b = constants,

T_{max} = daily maximum absolute air temperature in ^{o}K,

T_{min} = daily minimum absolute air temperature in ^{o}K,

R_W = longwave radiation emitted by a body of water in cal cm^{-2} day^{-1}.

Since the conditions in Alaska are different from those calculated for Idaho by Jensen, Robb, and Franzoy (1970), new constants were calculated (a = 0.62, b = 0.023) based on data taken during the summers of 1979 and 1978 at the Geophysical Institute, University of Alaska, Fairbanks (derivation see Appendix B).

Longwave radiation emitted from the lake, R_W, can be calculated from the Stefan-Boltzmann law for black body radiation and the emissivity of the water surface. Emissivity indicates the relative power of a surface to emit heat by radiation in comparison with the maximum possible intensity of a black body. The value for the emissivity of the water surface used in this study was 0.97 after Derecki (1976). The relationship for the emitted radiation is expressed by the equation

$$R_W = \varepsilon \sigma T_W^4 \qquad (6)$$

where ε = emissivity of the water surface (0.97),

T_W = water surface temperature in ^oK.

3) Sensible Heat Exchange Q_S and Latent Heat of Evaporation Q_L

The sensible heat exchange and the latent heat of evaporation are computed by using the gradient method (Webb, 1965). The transfer coefficients are measures of turbulent conductivity, representing the effectiveness of the turbulence in transfering a property down its gradient. The equations used here are:

$$Q_S = -c_p \rho K_H \frac{\partial \theta}{\partial z} \tag{7}$$

$$= -c_p \rho\, k_h^* u_* \frac{T_1 - T_W + G(z_1 - z_0)}{\ln(z_1/z_0)} \tag{8}$$

$$Q_L = -\rho L K_E \frac{\partial q}{\partial z} \tag{9}$$

$$= -\rho L k_H^* \; u_* \frac{q_1 - q_0}{\ln(z_1/z_0)} \tag{10}$$

$$u_* = \frac{k u_1}{\ln(z_1/z_0)} \tag{11}$$

$$k_H^* = f(R_i) \tag{12}$$

$$R_i = \frac{g}{\theta_a} \frac{\frac{\partial \theta}{\partial z}}{\left(\frac{\partial u}{\partial z}\right)^2} \tag{13}$$

where c_p = specific heat capacity of air at constant pressure,

ρ = air density,

K_H, K_E = eddy diffusivities of sensible and latent heat,

Θ, Θ_a = potential temperature, a denotes the average value

z, z_1 = height, 1 denotes a height of a sensor,

z_0 = roughness length (0.05),

k_H^* = modified von-Karmann constant for any lapse conditions,

u_* = friction velocity,

T_1 = air temperature at height z_1,

T_W = water surface temperature,

G = adiabatic lapse rate,

L = latent heat of evaporation,

q_1, q_0 = specific humidities at heights z_1, z_0,

u_1 = wind velocity at height z_1,

R_i = Richardson number,

g = gravitational acceleration.

The parameter k_H^* is a function of the Richardson number. The Richardson number was calculated according to equation (12), k_H^* was obtained from a table by Webb (1965).

The roughness length, z_0= 0.05 cm, was assumed to be constant, and its value was referred from Geiger (1965).

4) Miscellaneous Energy Fluxes, Q_{misc}

The heat flow from geothermal sources deep within the earth averages about 1µcal cm^{-2} sec^{-1}. On an annual basis this heat source adds about 30 cal cm^{-2} $year^{-1}$ to most lakes. This heat source is insignificant compared to Q_T, R_N, Q_L, Q_S and can be ignored in lake heat budget studies (Ragotzkie, 1978).

The heat flow through the bottom of the lake is dependent on seasonal changes of water temperature. Heat flows from the water to the sediments during the summer and early fall. The amount of heat involved in this seasonal exchange depends mainly on the seasonal temperature range of the water in contact with the bottom and secondarily on the thermal properties of the bottom sediments or underlying rock. In the summer this heat exchange is only a small percentage of the total heat content of the water and is generally ignored in the evaluation of a lake's summer heat budget (Ragotzkie, 1978).

The advective term in the heat balance of a lake describes changes in the heat content of the lake due to addition or removal of water which is at a temperature different from the average temperature of the lake. Myrup et al. (1979) estimated the average monthly advective term for Lake Tahoe (area of 499 km^2, mean depth of 313 m, latitude 39°N, California, U.S.A.). The maximum found is 9.6 cal cm^{-2} day^{-1} in August, the minimum is -4.1 cal $cm^{-2}day^{-1}$ in May. During summer months, these values are less than 5% of the rate of the heat storage. In this study the advective term was neglected but Foraker River, which carries cold water from the glaciers in the Alaska Range, certainly contributes to this term.

5) Results of the Heat Budget Calculations and Possible Errors

The semimonthly averages of the component of the heat balance of Lake Minchumina are shown in Table 6. This table shows the following results:

a) Net radiation, R_N, is decreasing continuously from July to the beginning of September.

b) Sensible heat fluxes are negative during the whole period. The maximum value is found for late August, the minimum value for the end of July.

c) Large negative latent heat fluxes are characteristic for the whole period. Maximum and minimum values coincide with those for the sensible heat fluxes.

d) In the second half of July the rate of heat storage in the lake is positive but very small. Then it becomes negative and shows a tremendous amount of heat loss in the second half of August, which is comparable with the net radiation and contributes to the latent and sensible heat fluxes.

e) From the heat budget data averaged over all periods more than the incoming net radiation is consumed by the latent heat flux, and the sensible heat flux is comparable to the rate of the heat storage loss.

f) Column #5 of Table 6 gives the rate of change of heat storage obtained as a residue from the quantities R_N, Q_S, Q_L. For a balanced heat budget these values should equal the corresponding values of Q_T calculated.

	R_N	Q_S	Q_L	Q_T	$R_N+Q_S+Q_L$
7/13 - 7/31/1979	214	-39	-157	5	18
8/1 - 8/16/1979	163	-44	-207	-16	-88
8/17 - 9/1/1979	114	-72	-208	-101	-166
7/13 - 9/1/1979	164	-52	-191	-41	-79

Table 6: Results of the heat budget components calculations for Lake Minchumina (cal cm^{-2} day^{-1}).

Errors in calculating the heat budget components are due to the inaccuracy of the collected data and due to the assumptions made for the calculations.

The error in the values of the global radiation is estimated to be within 10% due to the inaccuracy of the Robitzsch-type pyranograph found during the calibration mentioned above. The water temperature was usually measured beneath the water surface, and does not truly represent the actual water surface temperature. The depth of the temperature sensor differed from 5 cm to 20 cm depending on the change of the water level. A 10% error in the calculation of R_W could have been caused by an error of 7^oC in the water surface temperature. However, the error in the water surface temperature was estimated not to exceed 7^oC. The instrumental error in the dry-bulb, wet-bulb Psychrometer from which the value of q was calculated was described to be within 10% (manufacturer of the instrument).

There are many problems in the assumptions which are inevitable: The daily mean albedo to calculate R_S is assumed to be constant, 10%. The albedo is a function of the solar altitude and cloud conditions. The solar altitude at Lake Minchumina is low relative to the ones in the temperate climatic regions because of their latitudes. For Lake Huron (45^oN) Bolsenga (1964) used the value of 7% for July and 8% for August. Dutton and Bryson (1962) used 3% for Lake Mendota (34^oN). According to Fresnel's formula (Neumann and Pierson (1966)) for the albedo of a sea surface, it can be calculated using the average solar altitude and R_S/R_{SO} for a day. Using this formula, albedos within the ranges of 6.5

- 14% in July and 6.5 - 13% in August can be calculated for Lake Minchumina. The assumption of a constant 10% albedo lies within these ranges. The error in R_{SO}: the propriety of the values used for precipitable water vapor and dust attenuation, which are assumed to be constant for the entire period of the heat budget calculation, is beyond discussion because information on the actual data is unavailable. The constants (1.31 and -0.31) in the equation for R_L were estimated by using data measured at the roof of the Geophysical Institute, University of Alaska, Fairbanks. Jensen, Robb, and Franzoy (1970) used 1.35 for the coefficient of R_S/R_{SO} from the data taken at Davis, California. But there is no reasonable explanation that the ratio R_L/R_{LO} can be a linear function of R_S/R_{SO}, however, high correlation coefficients were found between them. R_{LO} consists of the Brunt-type equation and the Stefan-Bolzmann equation. Since the coefficients in the Brunt-type equation were obtained from data of only 9 clear days (Appendix B), the correlation coefficient was poor, 0.5. As mentioned above, the water surface temperature T_W is inaccurate and causes the error in the Stefan-Boltzmann equation. Overall, R_L is estimated to have a relative error within 10%. The calculation of Q_S, Q_L: the roughness parameter is assumed to be constant but varies with the wind velocity aloft and the characteristics of the surface, and ranges from 1×10^{-4} to 50 cm over water (Munn, 1966). The function of the Richardson number used for the estimation of the modified Karmann constant has not yet been firmly established. The sampling time for u, T_W, T_1, q_W, q_0 was one hour, which should have been less than one hour in order to eliminate the diurnal change from the change caused by the turbulent phenomenon. In order to calculate Q_T, the water temperature change for the depth of 3.5 m must be accounted for and it is assumed that there is no heat flow at a depth lower than 3.5 m, which might not be true but negligible. The location of the measurements taken for the heat budget calculation was at the east side of the lake, which might not be a representative area to show the average phenomenon of the lake but has at least the characteristics of the lake.

6) Comparison of Heat Budgets for Different Lakes

After having seen the main features of the heat budget estimates of Lake Minchumina the results of this study were compared with studies at lakes in the temperate climatic region. Table 7 shows a comparison for the heat budget components in July and August for Lake Minchumina, Lake Mendota (after Dutton and Bryson, 1962), Lake Tahoe (after Myrup et al., 1979) and Lake Huron (after Bolsenga, 1975).

a) The net radiation for Lake Minchumina is smallest among the lakes for both months. This can mainly be explained by the different latitude. Although the average sunshine duration is longer at Lake Minchumina, it does not compensate for the low solar angle which causes a high water surface albedo.

b) The values of Q_S and Q_T seem to be comparable in magnitude to the small lakes but not to the big ones.

	Latitude	Area (km^2)	Depth (m)	R_N	Q_S	Q_L	Q_T	R_N	Q_S	Q_L	Q_T
				JULY				AUGUST			
Lake Minchumina	64°N	65	12 (max.)	214	-39	-159	5	139	-58	-208	-59
Lake Mendota	43°n	90	?	363	65	-276	65	293	-10	-268	-10
Lake Tahoe	39°N	499	313 (aver-age)	457	-24	-173	308	387	-71	-253	204
Lake Huron	45°N	59,570	229 (max.)	443	-37	-103	368	302	-7	-80	125

Table 7: Comparison of heat budget components for different lakes: Lake Mendota (after Dutton and Bryson,1962), Lake Tahoe (after Myrup *et al.*, 1979), Lake Huron (after Bolsenga, 1979).

c) The values of Q_L for all lakes have the largest contribution to the heat sink. It seems that this term does not strongly depend on the size of the lakes and on the latitude in which they are located.

d) The values of Q_T in August are negative for the small lakes in contrast to the large lakes, which have large positive values.

V. SUMMARY AND DISCUSSION

In this paper the subarctic lake's microclimate was described and compared with studies which were done for lakes in temperate climates. Now the questions which were raised in the beginning of this study will be answered.

The size of Lake Minchumina is large enough to produce a lake effect under subarctic and continental climatic conditions showing the following characteristics:

1) The lake-land breeze is found clearly: the frequency of the occurrence of the lake breeze is 13% of all winds observed and that of the land breeze is 20% at Holmes Station. No lake-land breezes are found at the station 6 km away from the lake shore.

2) The warming effect of the lake is observed: the air mass is warmed up by 2.4°C when crossing the lake in a westward direction and by 1.4°C when moving eastwards. The lake shore stations are 3.4°C warmer than the station 6 km away from the lake. The longest frost-free period (117 days) is found at Lake Minchumina compared with other stations in Interior Alaska (10 years of data).

3) The moderating effect is detected in the comparisons of the spread of diurnal air temperatures between the lake shore stations and the station 6 km away from the lake. This effect is also found in the lesser departure of the frost-free period from the 10 years average of Holmes Station in 1979 compared with Fairbanks and McGrath.

4) The moisture content of the air mass is higher at the shore stations (2 mb) than at the stations further inland.

5) The heat budget components calculation shows that the rate of change of heat storage of the lake turns from its positive sign in July to the negative one in August and indicates a tremendeous amount of heat loss from the lake, which is almost comparable to the net radiation and is consumed by the latent and sensible heat fluxes.

Effects of the long sunshine duration which is characteristic for the high latitude region in summer are not clearly found for the lake effect. On the contrary, an effect of the low solar altitude on the small net radiation in the heat budget of the lake relative to that of lakes in temperate regions could be found. As a constant albedo of water for the calculation was used, the global radiation must have been weak. This low intensity of the global radiation might have caused the slightly lower value for the occurrence of the lake breeze at Holmes Station than the value reported by Yoshino et al. (1970).

Unfortunately the heat budget components further inland could not be measured at the same time as at the lake because of a failure of the recorder. It might have given some interesting information. To understand the complicated wind regime of a lake caused by the thermal response of the lake to the boundary layer a three dimensional observation is needed.

APPENDIX A

The Heat Budget Componenets for Lake Minchumina in 1979 (Ly/day).

DATE	R_N	Q_S	Q_L	Q_T	$R_N+Q_S+Q_L$
195	277	-8	-39	183	230
196	277	-59	-310	-168	-171
197	231	+6	-58	154	179
198	247	-12	-167	8	68
199	258	-39	-332	-136	-113
200	255	-53	-255	-113	-53
201	256	-4	-30	163	222
202	245	-13	-51	208	181
203	209	-41	-128	-74	40
204	258	-35	-101	76	122
205	152	-22	-102	136	28
206	164	-120	-298	-228	-254
207	124	-86	-216	-309	-478
208	180	-72	-166	92	-58
209	198	-65	-146	-145	-2
210	255	-6	-38	152	211
211	167	-36	-162	60	-31
212	136	-37	-222	28	-123
AVE	214	-39	-157	5	18
213	158	-40	-232	-15	-114
214	181	-57	-274	-114	-150
215	190	-32	-148	22	10
216	170	-24	-158	51	-12
217	187	-87	-320	-204	-220
218	157	-100	-334	-158	-277
219	159	-73	-273	-117	-187
220	157	-83	-247	-186	-173
221	165	-64	-184	-156	-83
222	165	-19	-126	87	20
223	166	-8	-78	-59	80
224	175	-1	-124	159	50
225	172	-15	-259	-191	-102
226	131	-11	-107	201	13
227	148	-24	-193	-166	-69
228	134	-60	-250	-38	-180
AVE	163	-44	-207	-16	-88

DATE	R_N	Q_S	Q_L	Q_T	$R_N+Q_S+Q_L$
229	141	-11	-124	-29	6
230	121	-31	-128	-141	-38
231	142	-83	-209	-82	-150
232	107	-140	-282	-150	-315
233	132	-54	-116	-50	-38
234	163	-10	-35	197	118
235	153	-31	-76	-16	46
236	177	-3	-111	-90	63
237	96	-13	-195	-81	-112
238	115	-85	-318	-144	-288
239	86	-106	-321	-192	-341
240	84	-154	-350	-330	-420
241	74	-145	-337	-174	-408
242	70	-84	-220	-156	-234
243	87	-83	-206	-126	-202
244	78	-125	-295	-53	-342
AVE	114	-72	-208	-101	-166
TOTAL AVE	164	-52	-191	-41	-79

APPENDIX B

Brunt (1932) found high correlation coefficients between $R/\sigma T_a{}^4$ and $\sqrt{e_d}$, where R is the incoming atmospheric radiation, σ is the Stefan-Boltzmann constant, T_a is the air temperature, and e_d is the saturation water-vapor pressure at the dew-point temperature. He also derived an approximation formula for the net longwave radiation, R_{LO}, on a clear day (Brunt, 1939).

$$R_{LO} = \sigma T_a^4 \ (1 - a - b \sqrt{e_d}) \qquad \text{(B-1)}$$

Jensen ,Robb, and Franzoy (1970) used the following equation for a clear day net longwave radiation by modifying the Brunt equation:

$$R_{LO} = (a + b\sqrt{e_d}) \ \sigma \frac{T^4_{max} + T^4_{min}}{2} - \varepsilon\sigma \frac{T^4_{max} + T^4_{min}}{2} \qquad \text{(B-2)}$$

where T_{max} and T_{min} are daily maximum and minimum absolute air temperatures in degrees Kelvin and ε is the emissivity of the surface.

In order to calculate the constants a and b, the equation (B-2) was rearranged,

$$-R_{LO} \ / \ \sigma \frac{T^4_{max} + T^4_{min}}{2} = A - b\sqrt{e_d} \qquad \text{(B-3)}$$

where $A = \varepsilon - a$, to give an equation of the form $y = ax + b$.

The left hand side term and e_d were calculated from the data measured on the roof of the Geophysical Institute, University of Alaska, Fairbanks, on clear days. There were 9 clear days during the summers of 1979 and 1980. The technique which was used is linear regression by the method of least squares. The result is shown in Fig. B-1. ε is assumed to be 0.98. From the graph the following results were obtained:

$$a = 0.62$$

$$b = 0.023$$

The correlation coefficient is 0.5. To improve the reliability of the estimates for a and b more cloudless day data are needed.

Jensen, Robb, and Franzoy (1970) estimated R_L using the following formula:

$$R_L = (c \frac{R_S}{R_{SO}} + d) R_{LO} \ . \tag{B-4}$$

The constants, c and d, they used were 1.35 and -0.35, respectively. The linear relationship between R_L/R_{LO} was calculated by the method of least squares using daily data measured at the roof of the Geophysical Institute, University of Alaska, Fairbanks, in the summer 1979. The estimates for the coefficients c and d are:

$$c = 1.31$$

$$d = -0.17$$

$$\text{correlation coefficient} = 0.77.$$

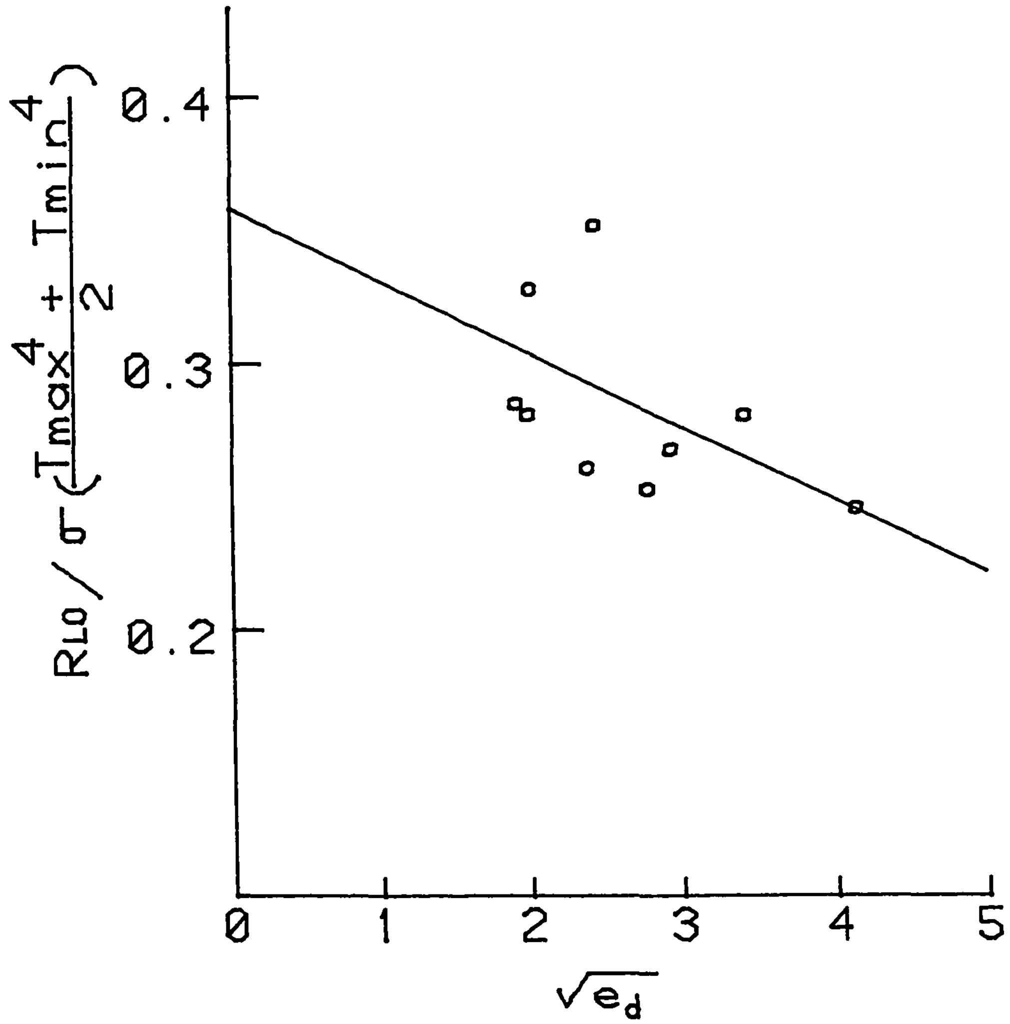


Fig. B-1: The result of the linear regression by the method of least squares for the equation (B-3).

But, when $R_S = R_{SO}$ viz. on a cloudless day, R_L should be equal to R_{LO}. To satisfy this requirement, the value of -0.31 for d was used instead of -0.17.

APPENDIX C

List of Symbols

a, b	constants
c	specific heat capacity of water
c_p	specific heat capacity of air at constant pressure
dz	depth or height increment
e_d	saturation vapor pressure at daily mean dew-point temperature in mb
G	adiabatic lapse rate
g	gravitational acceleration
K_H, K_Q	eddy diffusivities of sensible and latent heat, respectively
k_H^*	modified von-Karmann constant for any lapse conditions
L	latent heat of evaporation
q_0, q_1	specific humidities at heights z_1, z_0
Q_L	latent heat flux
Q_{misc}	miscelleneous energy fluxes
Q_T	rate of change of heat stored in the lake
R_i	Richardson number
R_L	net longwave radiation emitted to the atmosphere
R_{LO}	R_L on a cloudless day
R_N	net radiation
R_S	incoming vertical flux of solar radiation
R_{SO}	R_S on a cloudless day

R_W	longwave radiation emitted by a body of water
t	time
T	water temperature
T_{max}	daily maximum absolute air temperature
T_{min}	daily minimum absolute air temperature
T_1	air temperature at height z_1
T_W	water surface temperature
u_*	friction velocity
u_1	wind velocity at height z_1
z, z_1	heights, 1 denotes a height of a sensor
z_0	roughness length
α	mean daily shortwave reflectance or albedo
α_W	Reflectivity of water for longwave radiation
ε	emissivity of the water surface
Θ, Θ_a	potential temperature, a denotes the average value
ρ	air density
ρ_W	density of water
σ	Stefan-Boltzmann constant

REFERENCES

Bolsenga, S.J., Daily sums of global radiation for cloudless skies, U.S. Army C.R.R.E.L., Res. Rep. 160, Hanover, N.H., 1964.

Bowling, S.A., Alaska's weather and climate, Geophysical Institute Report, Univ. of Alaska; UAG R-269, pp 1-25.

Bruce, J.P. and R.H. Clark, Introduction to Hydropetrology, Pergamon Press Ltd., Oxford, 1966.

Brunt, D., Physical and Dynamical Meteorology, Pergamon Press Ltd., Cambridge, 1952.

Derecki, J.A., Lake Erie Terrestrial Radiation, Water Resour. Res., Vol. 12, NO. 5, 979-984, 1976.

Dutton, J.A. and R.A. Bryson, Heat Flux in Lake Mendota, Limnol. and Oceanogr. 7 (1), 80-97, 1962.

D'yakonov, K.N. and A.Yu. Reteyum, The local climate of the Rybinsk Reservoir, Sov. Geog.: Review and Translation, Vol. VI, No. 10, 40-52, 1965.

Eley, J., D. Keeley and T. Olien, Variation of afternoon temperatures in summer near the shores of Lake Winnipeg, Meteorological Branch, Canada Department of Transport, CIR-3559, TEC-379, 1961.

Geiger, R., The climate near the ground, Harvard University Press, Cambridge, Mass., 1965.

Jensen, M.E., D.C.N. Robb, and C.E. Franzoy, Scheduling irrigations using climate-crop-soil data, Journal of the Irrigation and Drainage Division, Proceedings of the American Society of Civil Engeneers, 25-37, 1970.

Munn, R.E., Descriptive Micrometeorology, Academic Press, New York, 1966.

Myrup, L.O., T.M. Powell, D.A. Godden, and C.R. Goldmann, CLimatological Estimate of the Average Monthly Energy and Water Budgets of Lake Tahoe, California-Nevada, Water Resour. Res. Vol. 15, No. 6, 1499-1508, 1979.

Neumann, G. and W.J. Pierson, Jr., Principles of Physical Oceanography, Prentice-Hall, INC., Englewood Cliffs, N.J., 1966.

Phillips, D.W. and J.A.W. McCulloch, The climate of the Great Lakes basin climatological studies No. 20, Atmospheric Environment Service, Environment Canada, 1971.

Ragotzkie, R.A., Heat Budgets of Lakes, Lakes-Chemistry, Geology, Physics, edited by Abraham Lerman, Springer Verlag, New York, 1-19, 1978.

Richards, T.L. and J.P. Fortin, An evaluation of the land-lake vapor pressure relationship for the Great Lakes, Publ. No. 9, Great Lakes Res. Div., Univ. of Michigan, pp 103-110, 1962.

Searby, H.W., CLimates of the States, Alaska, ESSA, Climatology of the United States No. 60-49, U.S. Weather Bureau, 23pp, 1968.

Sellers, W.D., Physical Climatology. The University of Chicago Press, Chicago pp272, 1965.

Shitara, H., Air temperature in summer on and around the Lake Inawashiro, Fukushima Pref., Sci. Rep. Tohoku Univ. (Geogr.), 16, 69-84, 1967.

Van Hulle et al., Annual Performance Report for Inventory and Cataloging. Federal Aid in Restoration Sport Fish Division, Vol. 16, 1975.

Webb, E.K., Aerial microclimate, Agr. meteoro., Meteorol. Monograph, Vol. 6, No. 28, 27-58, 1965.

Yoshino, M., K. Urushibata, M. Owada, Lake-breeze of Lake Suwa, Nagano Pref., Central Japan, (written in Japanese), Tenki, 17, 55-62, 1970.