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LAKE EFFECTS ON AIR POLLUTION DISPERSION

Lars E. Olsson

Fred V. Brock
Project Director

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ABSTRACT

LAKE EFFECTS ON AIR POLLUTION DISPERSION

by

Lars Erik Olsson

Chairman: Aksel C. Wiin-Nielsen

Local wind regimes induced by a lake or a shoreline have a major influence on air pollution dispersion. Pressure differences due to differential heating of the air, e.g. those due to differences in surface characteristics between land and lake, are the driving forces of lake and land breeze circulations. Differences in roughness between land and lake surfaces will cause wind shear and aerodynamic downwash effects at a shoreline. Stability changes in the air result from differences in surface temperature and roughness between land and lake, e.g. when warm unstable air moves out over a cool lake a temperature inversion will develop near the surface, thus creating poor dispersion and trapping of pollutants. Fumigation will occur as this stable layer is broken up when the air moves across a shoreline and advances inland. A number of these lake influences have been examined in detail on two separate occasions.

The first observational study was carried out near the eastern shore of Lake Michigan on 25 June, 1965 and produced the most extensive measurements yet made of pure lake and land breeze circulations. Almost ideal anticyclonic conditions with clear skies and light winds prevailed.

In the second investigation constant level balloons, tetroons, were used to study air trajectories in lake breeze circulations near the highly urbanized and industrialized southwestern shore of Lake Michigan on 12 - 13 August, 1967. The tetroons were tracked by two optical theodolites and in one case a tetroon was observed to describe a complete lake breeze circulation loop. Extensive measurements of temperatures, humidities, aerosol concentrations, and winds were made both at the surface and aloft and photographs and time lapse movies of clouds, haze, and smoke were taken.

Based on these observational studies and an extensive literature review, the role of local wind systems near a large lake in dispersing atmospheric pollutants is explored in depth.

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1. INTRODUCTION

The availability of water is a necessity for almost all of man's activities. Thus, most population centers as well as most industrial complexes are located at or near large sources of water. The water body might be an ocean, a lake, or a river. We know that most human activities produce waste of one kind or another and that this waste has to be disposed of. Much of the waste is transported away by trucks or ships. It is dumped as fill or it is burned in incinerators, sometimes far away from its source of origin. Much is, however, released directly into the water or air. It is thus essential to have good estimates of how much pollution the water and the air can disperse; that is: transport away, diffuse, and finally eliminate. Some lakes, such as Lake Erie, are already over polluted, partly due to a lack of knowledge of the lake's capacity to disperse pollution.

Many bodies of water, such as lakes, rivers, and many ocean bays can be looked upon as almost closed systems, with very little dispersion across their boundaries. The problem of defining the dispersion patterns in the atmosphere is more complicated, because there are fewer constraints on air motions. It is thus difficult to draw many direct parallels between the dispersion of air and water pollution in general. There are, however, situations where an air mass may be trapped either physically or dynamically for several weeks; one location where such trapping causes severe air pollution problems is the Los Angeles Basin. Rivers and water currents may transport pollution many miles, and sometimes in high concentrations. The same is true for an air current under certain meteorological conditions. While it is easy to predict where a river will carry the pollutants it is much more difficult to predict the trajectory of pollution released into the atmosphere.

1.1 PURPOSE OF THE STUDY.

The objective of this study is to investigate the various effects a lake and a lakeshore have on the atmosphere's capability to disperse air pollution. Thus, the most important effects to be considered are air parcel trajectories and atmospheric stabilities.

Due to the differences in physical characteristics between soil and water, horizontal temperature, moisture, and pressure gradients normally exist across a lake shore. In situations when the synoptic pressure gradients are weak, local circulations, such as lake and land breezes are induced due to differential heating and cooling of lake and land surfaces. Under strong gradient wind conditions these temperature differentials as well as the discontinuity in surface roughness at the shore induce perturbations and shears in the gradient flow.

When an air mass is advected across a shoreline, its thermal and dynamic stability is likely to change in response to the changed characteristics of the underlying surface, e.g. strong surface inversions, formed in an air mass moving over a smooth and cool lake,would be destroyed as the air moves in over a rough and warm land surface.

A thorough search in the literature reveals that, although several observational studies of shoreline and lake meteorology have been made, only a few have been related to air pollution dispersion. A comprehensive summary of these earlier reported studies is presented in the next section.

As the most severe air pollution incidents normally occur under stagnant anticyclonic conditions, and as these synoptic conditions also promote development of land and lake breeze circulations, special emphasis is placed upon these circulations. Two observational studies of land and lake breeze circulations made on the shores of Lake Michigan are presented in detail in Chapters 2 and 3.

1.2 LITERATURE REVIEW.

Weather, especially local weather, has long been observed by man. That topographical discontinuities give rise to local weather anomalies have long been recognized and a large number of observational studies of such effects have been reported in the literature. However, most observations have been made near the surface, while only a few intensive measurements of how topographical transitions change atmospheric characteristics aloft have been made.

Many investigations have been devoted to land and sea breezes, while relatively few studies of lake breezes or other effects induced by lakes and shorelines have been reported. Even though air pollution has concerned man for hundreds of years, it is only in the last few decades that studies of air pollution dispersion have been made.

Local wind systems, such as land and sea breeze circulations and various slope winds, have been described in several text books, e.g. Hann (1907), Brunt (1939), Hewson and Longley (1944), Defant (1951), and Haltiner and Martin (1957). A survey of sea breeze studies was made by Wexler (1946), while Munn and Richards (1964) presented a survey of lake breeze studies. A bibliography on land and sea breezes was prepared by Baralt and Brown (1965). Meteorological aspects of air pollution dispersion have been discussed by Hewson (1951, 1960), Hewson and Olsson (1967), and Slade (1968).

The following literature review is based on a survey of more than 500 published papers and technical reports, and the main emphasis is placed on findings that are of importance in evaluation of air pollution dispersion in a lake region.

Plutarch (1st-2nd cent. A.D.) makes a reference to the Athenian general, Themistocles, who planned to use the onset of the sea breeze to capture the Persian fleet in the Bay of Salamis. Early descriptions of sea breezes, mainly at lower latitudes, were made by Halley (1686), Dampier (1705), Forster (1778), and Seignette (1782). The earliest known description of a lake breeze on the North American continent is due to Ellicott (1799), who reported on observations made on the southeastern shore of Lake Erie.

Several interesting findings and conclusions were reported during the 19th century. Martens (1829) and Schübler (1829) studied lake breezes on Lake Garda and Lake Constance, respectively, Coffin (1852) described local winds of the northern hemisphere, among them also the lake breeze at Lake Erie, and Rykatschew (1888) reported on sea breezes on the shores of the Caspian Sea. Tissandier (1870), Eloy and Lhoste (1883), and Toulon (1894) all described how the different directions of the air currents along a coast were used for "round trips" by balloon fliers. Laughton (1873) concluded that land and sea breezes are driven by water vapor differentials and that they have nothing to do with temperature differences. Measurements of the intensity of the sea breeze and of the water vapor pressure over land were reported by Smock (1888). He showed that the sea breeze is strongest in spring, when the water vapor pressure is weakest. In the same journal that contained Laughton's report, Stow (1873) described observations of the sea breeze at the coast of Yorkshire, England. He found, that of 82 clear days, 66% had a sea breeze, while only 33% had a land breeze. He also observed the largest average temperature difference across the shore to be 8°C. Blandford (1877) proved that a pressure difference does exist between land and sea breeze regimes even though it is small. By observing anchored balloons on Coney Island, Sherman (1880) found the average depth of the sea breeze to be 150-200 m. He also found a return flow layer aloft, which was 240-400 m thick. Köppen (1879, 1884) discussed gradient wind effects on sea breezes, and Forel (1892) described how land and sea winds at Lake Geneva are influenced by Mountain and valley winds. Forel stated that the land-wind starts over land and progresses against the sea-wind, while the sea-wind starts over the lake and has to overcome the land-wind. Daily wind speeds and directions at the water intake crib in Lake Michigan near Chicago, Ill. were studied by Hazen (1883, 1893). He found that the lake breeze lasted from 1100 to 2200, local time. Marzelle (1895) referred to the experience of a navy captain to prove the absence of land and sea breezes on the steep and high coast of Chile.

During the first half of the 20th century many observational studies of land and sea breezes were carried out in the tropics. Van Bemmelen (1922), Ramanathan (1931) and Dixit and Nicholson (1964) found that land and sea breezes occur almost every day in the summer months, while they appear on about 30% of the days during the rest of the year. They also reported on observed return currents above the tropical sea breeze. Braak (1925) presented a general description of local winds, that were observed on a mountainous island in the tropics. He found that with an offshore gradient wind a band of cumulus sometimes formed 5 km off shore and then moved inland with the sea breeze. He also noted that the down-valley wind at night was stronger than the up-valley day current. Leopold (1949) observed sea breezes on Hawaii and outlined their influences on local microclimate. Diurnal variations of winds up to 2,500 m have been studied by the aid of 5 years monthly means from Bombay and Poona in India by Parthasarathy and Narayanan (1952). They showed that the sea breeze correlates with high surface pressure over land in the morning and low pressure over land in the afternoon. Donn, Millie, and Briliant (1956) found that the dominant atmospheric pressure oscillations recorded by a microbarovariograph at Guantanamo Bay, Cuba, were caused by perturbations from internal gravity waves on the sea breeze interface aloft. Rao (1955) observed the average speed of inland penetration of the sea breeze front near Madras, India, to be 1-4 m sec^{-1} (3-7 mph). He noted that the front accelerated as it moved inland and that this acceleration was independent of wind speed, but varied depending on the hour of onset of the sea breeze, being lowest during the epoch of maximum temperature. Measurements reported by Berlage (1936) indicate that there are marked jumps in the electrical potential gradient associated with the tropical sea breeze front.

Shoreline circulations in the mid-latitudes have been intensively studied along the coasts in Europe, East Asia, and in the last few decades also in the United States. Sea breeze studies along the coasts of the Baltic Sea have been reported by Kaiser (1906, 1907), Grenander (1912), Šokalškij (1912), and Frank (1931, 1940, 1948). Average maximum temperature differences between land and sea, on sea breeze days, were observed to vary from 1.6°C to 10.9°C. Kaiser observed the maximum average surface pressure over water to be 1 mb higher than over land on sea breeze days and approximately 0.7 mb lower than over land during land breeze conditions. Koschmieder (1935, 1936, 1941), Koschmieder and Hornickel (1942), Hornickel (1942), and Möller (1942) reported on sea breeze circulation studies at Danzig. They found good correlations between changes of wind speed and directions, temperature drops, and humidity increases as the sea breeze front progressed inland. They also noted that the frontal characteristics were most pronounced under offshore gradient wind conditions and that the front penetrated inland in a pulsating manner and had a slope near the leading edge of 1:14.

On the basis of observational data from the Dutch coast of the North Sea, Balkema (1950) concluded, that insulation is more important for the occurrence of a sea breeze than the low level temperature difference between air over land and over sea. Batty (1921) reported, that balloons showed a descending air current off shore associated with a sea breeze on the southern coast of England. In analyzing pibal observations taken at Felixstowe, England, during the period 1924-33, Sutcliffe (1937) estimated the average depth of the sea breeze to be 500 m, but found no evidence of a return flow layer aloft. Pedlow (1952) reported on measurements made near the southern coast of England and found higher vapor density, steeper vapor lapse rates, higher temperature by day, and steeper temperature lapse in onshore winds than in offshore winds. Peppler (1920, 1929) discussed shoreline winds along the Dutch coast and Bleeker (1936) noted the effect of the Coriolis acceleration in wind measurements made at Hoek van Holland. He found, that in the late afternoon, the sea breeze blew parallel to the coast. In observing nocturnal winds near the coast in southern England, Lawrence (1954) noted the absence of katabatic winds over slopes less than 1 in 100 or 150. He found that these nocturnal winds occurred in moderate, but not strong inversions and that their course was markedly influenced by surface characteristics. That differences in surface friction between land and water causes vertical compensation currents was observed by Georgii (1923).

Shoreline winds along the coasts of the Black Sea were observed by Vasenko (1932) and Moltschanoff (1952). Vorontsov (1937) described the development of sea breezes in the Sochi region of the Caucasus on the basis of wind observations made by means of pilot balloons and standard level balloons.

Reichsmarineamt (1905) presented a summary of sea breezes around the Mediterranean and noted that they are not very predictable, since they are influenced by mountain and gradient winds and constantly change direction with the sun. Measurements of extents and rates of propagation of land and sea breezes in the Gulf of Lions were reported by Moye (1922). He found evidence of a return current above the sea breeze. Koch (1952) noted that the sea breeze near Rome penetrated 20 km inland in the winter and more than 60 km inland in the summer. Based on several years of analyzed surface data from coastal stations in northwestern Italy Bossolasco (1952) found that there are close relations between the characteristics of coastal sea currents and the characteristics of the sea breeze. He concluded that the strength of the sea breeze, whether on a daily or seasonal basis, arises either from cooling of coastal water or heating of littoral, or on the balance between the two factors.

Neumann (1951) found the frequency of nocturnal thunderstorms, within coastal areas in the eastern part of the Mediterranean, to be related to the convergence and divergence of the land and sea breeze circulations and that the curvature of the coast determine the convergence and divergence of the field of sea and land breezes. Observations of the sea breeze at Athens, Greece, have been reported by An der Lan (1949) and Karapiperis (1952, 1953). The depth of the sea breeze flow layer was observed to be about 500 m, and maximum mid-afternoon speeds 5 m sec^{-1} in the winter and 12 m sec^{-1} in the summer. Karapiperis also discussed moisture variations observed in the sea breeze regime, and An der Lan described characteristics of local katabatic winds on the Greek and Turkish coasts. Ward (1954) noted, that the sea breeze observed at Gibraltar, rarely extends above 300 m (1000 ft), often exceeds 7 m sec^{-1} (15 kt), and is associated with a stable temperature distribution up to 900 mb.

Shoreline winds along the coasts of Japan have been observed and reported by Fukuda (1953), Mano (1953), and Kikuchi (1954). They found land and sea breezes to be the dominant local winds, except in the winter. Mano also analyzed combined land-sea breeze and mountain-valley wind effects. A brief review of sea breezes in Australia was presented by Clarke (1955).

Rexboard (1954) described the sea breeze at Boston on the basis of surface data, while Fisher (1960) and Frizzola and Fisher (1963) observed the vertical structure of sea breezes on the coast of New England. They found that the maximum onshore velocity occurred in the lowest 300 m in the afternoon and rarely exceeded 7 m sec^{-1}. The sea breeze front was most pronounced under offshore gradient wind conditions and they noted that the leading edge of the front was quite steep, 1:20 to 1:100. Fisher and Frizzola also concluded that the depth of the sea breeze layer was less than 1000 m and that this depth was greater when onshore gradient flow existed than with offshore gradient flow. The Coriolis effect was evident and a return flow layer could be detected, although it was obscured by the overriding gradient flow. They observed a low level jet above the coast near sunset, which they thought to be due to a reduction of frictional drag, and thus associated with the onset of nocturnal cooling. The effects of coastal characteristics and various topographical features on sea breeze circulations were also discussed. Eddy (1966) found evidence of diurnal wind variations, associated with the sea breeze on the Texas coast at altitudes as high as 5000 m. He noted that land breezes at times extended 15 km off shore.

The various effects which meteorological factors, especially sea and land breezes, have on air pollution dispersion in southern and central California have been observed and discussed by several investigators, e.g. Beer and Leopold (1947), Kauper (1960), Edinger and Helvey (1961), Frenzel (1962), Edinger (1963), and Fosberg and Schroeder (1966). Topographical features and synoptic conditions were found to have a major influence on the local winds as well as on the atmosphere's capability to disperse air pollution. Fosberg and Schroeder found evidence of a return flow layer above the sea breeze. The topographical influences were also stressed by Staley (1957, 1959) and Lowry (1963), who observed winds and temperatures near the Pacific coast in the northwestern United States. Staley (1957) analyzed average sea level pressures for July, 1950 and found diurnal variations exceeding 2 mb near the coast. Maximum onshore pressure gradient was found in late afternoon and maximum off shore gradient in early morning.

Angell and Pack (1960, 1962, 1963) have reported on tetroon flights over the mid-Atlantic coast of the United States. They found the Coriolis effect to be pronounced and strong cross isobaric,antitriptic flow to occur in a sea breeze regime. They also found evidence of changes in wind direction and speed due to changes in atmospheric stability and surface roughness. The diffusive capacity of the atmosphere was observed to be better on non sea breeze days, than on sea breeze days. Similar findings were made during tetroon flights in the Los Angeles Basin, Pack and Angell (1963) and Angell et al (1966). Tetroon trajectories proved the existence of diurnal recirculation, i.e. air was carried off shore by the land breeze and again back on shore by the sea breeze. Angell and Pack (1962, 1966) measured the period of vertical oscillations in the atmosphere and found it to be a function of lapse rate and in agreement with the Brunt-Väisälä formulation. However, terrain features in the Los Angeles Basin, strongly influenced the circulations in that region. Hass et al (1967) reported on tetroon flights over New York City. Vertical oscillations were observed over the city with a generally high trajectory point over the Manhattan Island and low points over the surrounding rivers. Reversals, from land to sea breezes, off shore were successfully observed during several of these referenced tetroon studies.

Peterson (1966) found that long range, low-level tetroon trajectories could be predicted more accurately from observed surface winds than from geostrophic winds, and Druyan (1968) noted that such predictions were more difficult over urban areas due to the variability in thermal structure.

Wallington (1959, 1965) and Simpson (1965) observed average upward velocities of 250 cm sec^{-1} and occasional updrafts of about 800 cm sec^{-1} in sea breeze fronts in southern England. The fronts, sometimes less than 250 m wide, were occasionally well marked by "curtain of clouds".

Moisture, clouds, and showers associated with sea breeze fronts have been observed by several investigators. Tinn (1922) noted especially the cloud formation as an indication of updrafts, and Malcus et al (1951) studied the detailed structure of a cumulus cloud veil formed on a sea breeze front. Craig et al (1945) observed what has been considered a typical cross section of moisture in the sea breeze circulation. A dry tongue aloft, stretching landward at about 200 m over the water was associated with the return flow aloft and the subsidence off shore. Atlas et al (1953) and Atlas (1960) studied the sea breeze by means of radar. They found a sharp discontinuity in moisture along the frontal surface and estimated the rate of progression of the front to be 4 m sec^{-1} (9 mph). They observed a sharp lapse in vapor pressure near the surface in the sea breeze air, which was believed to be due to increased convection by passage of air from over cool water to the warmer coastal water.

Byers and Rodebush (1948) found the low-level horizontal convergence caused by the double sea breeze effect on the Florida peninsula to be the most rational explanation of the high frequency of thunderstorms over the interior of the peninsula.

Neuberger (1936) defined a "pseudo-sea breeze" to explain the high nuclei count near the ocean, which was observed as the wind veered toward land. Before the wind increased, pollution, that was carried off shore by the land breeze "piled up" and caused this high nuclei count. This "pseudo-sea breeeze" turned into a "pure sea breeze" as the wind speed increased. This phenomenon was also studied by Zenker (1954).

On the basis of 20 years actionometric observations, Sakali (1955) showed that the 24-hr total radiation may increase 15-20% under periods with land and sea breeze circulations. He found, that while the nocturnal totals of radiation balance increased perceptibly, the daily totals remained practically unchanged, since the diminution of effective, and increase in diffuse radiation was compensated by the decreased direct radiation.

Kotsev (1956) reported that increased general air ionization and a preponderance of negative ionizations are observed under pronounced sea breeze conditions. He found, that besides the Lenard effect, the circular character of the sea breeze circulation could explain this fact.

During the first half of the 20th century several observational studies of lake breezes were reported by European investigators. Weickmann (1921) observed descending currents over large lakes. Land and lake breezes at Lake Zürich and their interaction with local mountain and valley winds were studied by Frey (1926) and Gutersohn (1938). Kleinschmidt (1922) discussed the lake breeze at Lake Constance, while Kopfmüller (1922, 1923, 1924, 1927) reported on detailed observational studies of wind, temperature, humidity, and pressure variations associated with the land and lake breeze around that lake. He found, that, on the average, the lake breeze was 140 m deep with maximum onshore wind speeds of 2-3 m sec^{-1} 50 m above the surface and that there existed a 400 m thick return current aloft. Kopfmüller also studied the formation of land and lake breezes and found that the front penetrated 7-10 km inland. He noted that cloud formation was associated with the lake breeze front and that, at times, the breeze was homogeneous along the shore. It was also observed, that as the land breeze converged toward the center of the lake, accompanying rising motion caused stratus to form at a height of about 200 m over the lake. This occurred frequently in the spring and fall, but seldom in the summer. Peppler (1926, 1932) discussed temperature variations at Lake Constance during cold offshore winds and reported on lake breezes that occurred in March, while the land areas were snow covered. Rönicke (1962) studied the vertical structures of land and lake breezes at Lake Constance by means of photographing drifting smoke columns. His findings are in good agreement with those reported by Kopfmüller. The lake breeze along the shores of Lake Müritz has been described by Kuhn (1942).

Amplitudes and periodicities of local winds on the shores of Lake Balaton, under anticyclonic synoptic conditions have been studied by Czelnai (1955). Lake breeze circulations at that lake have also been observed by Peczely (1961, 1962). He noted that the depth of the onshore flow layer was approximately 500 m and estimated the height of the circulation to be 2000 m. Based on 5 years data, he also found that the lake breeze developed 15% of the time and that the front penetrated more than 5 km in 60% of these cases and more than 10 km inland in 12% of these cases.

Vorontsov (1958) has presented an extensive study of the lake breeze over Lake Ladoga. He reported that the depth of the onshore flow layer varied between 200 m and 900 m and that maximum winds were observed at 0.15 - 0.20 of that height. Vorontsov also observed average temperature drops of 1°C and relative humidity increases of 9% associated with the passage of the lake breeze front at a location near the shore. The effects of local topography and prevailing weather on a lake breeze circulation at a large fresh-water reservoir has been studied by Chestnaia (1962).

Temperature variations associated with the Lake breeze from the Dead Sea were studied by Ashbel (1934). Green (1935) discussed the relation between prevailing wind patterns and on-and off-shore winds on the south shore of Lake Victoria in East Africa. Mukade (1953) reported on meteorological observations made at Lake Inawashiro in Japan. He noted the influence of topographical features on land and lake breezes and that at 1000, local time, prevailing winds on the opposing shores of the lake showed opposite directions. He also observed some relationship between the wind and the surface temperature of the lake and that wind shifts occurred at 0730 and 1930, local time. An important influence of coastal topography on the shoreline circulations was also noted by Yamashita (1953), who studied winds over Lake Suwa.

Except for the referenced studies by Ellicot (1799) and Hazen (1883, 1893), only Miller (1939) has reported on lake breeze observations in the Great Lakes region before World War II. Miller found that, although Lake Michigan's breeze on the Door Peninsula in Wisconsin only entended 1.5 km over land, there was an observable lake effect extending 100 km inland. Landsberg (1958) reported on pibal observations made in the Chicago area during World War II. He noted that the strongest wind was found near the surface, that the lake breeze usually was less than 500 m thick, and that the lake air penetrated about 3 km inland.

Criteria for forecasting the occurrences of lake breezes and their effects on the visibility at Chicago's Midway Airport were discussed by Hall (1954). He stated that, if the gradient wind is less than 7 m sec^{-1} with an along shore or offshore direction, the cloudiness is less than 6 tenths, and the air temperature is more than $5^{\circ}C$ above the lake water temperature one could expect the lake breeze front to reach the airport, 12 km inland, in the afternoon. He noted that, especially on days with a general wind opposing the lake breeze, the smoke filled convergence zone is very narrow and its leading edge is very sharply defined. Visibility in this zone was occasionally observed to be less than 2 miles, which was considerable less than the visibility observed deeper into the body of the lake breeze circulation. He also indicated the occurrence of a lake high developing on lake breeze days.

Hewson et al (1960, 1961, 1963) and Bierly and Hewson (1963) found the natural ventilation on the western shore of Lake Erie and on the eastern shore of Lake Michigan to be above average. They found lake and land breezes to occur on more than one third of the days in the summer half of the year. Shallow lake breeze inversions were observed near the shore in the onshore flow. The depths of these inversions were usually less than 50 m and they were found to be destroyed within 3 km inland from the lake shore. Biggs and Graves (1962) developed a lake breeze index, based on the balance of forces that distinguish lake breeze days from non-lake breeze days. They found the occurrence of a lake breeze to be dependent on the magnitude of the gradient wind and the maximum temperature differential across the lake shore. Prolonged low level inversions, lasting for more than 24 hrs, along the western shore of Lake Erie were studied by Graves (1962). He found these inversions to occur most frequently in November and December and least frequently in August and September. Graves noted that warm air advection over cooled land areas favored occurrences of this type of shoreline inversions.

Munn and Richards (1964, 1967) and Richards (1964) studied shoreline circulations on the eastern shore of

Lake Huron. They observed onshore winds to occur between 1000 and 1700, local time, on more than two thirds of the days in spring and summer. Under lake breeze conditions, the depth of the stable onshore flow layer was normally less than 200 m and they noted that a new internal boundary layer developed over land in response to land heating. The depth of this superadiabatic layer varied between 30 m and 60 m at 300 m from the lake. The lake breeze was often observed as a 180° wind shift and a $3^{\circ}C$ temperature drop near the lake shore at about 0900, local time. Occasionally the lake breeze was observed to penetrate more than 19 km inland. Munn and Richards noted intense modification of the lake air as it passed over land. They found that this modification sometimes was so strong that the lake breeze disappeared completely, permitting a new lake breeze front to form and move inland. They also observed a controlling influence of the geostrophic wind, and effects due to changes in surface roughness, upon these shoreline circulations.

Observations of well developed lake breeze circulations reported by Moroz (1965, 1967) were made near Grand Haven on the eastern shore of Lake Michigan under predominantly anticyclonic conditions. He found that the lake breeze, in general, had a depth of less than 600 m and that a maximum onshore velocity of more than 4 m sec^{-1} occurred within 250 m of the surface immediately inland from the shore. Furthermore, the return flow layer aloft was more than 1500 m thick and a maximum offshore velocity exceeding 3 m sec^{-1} occurred in the afternoon at 1200 m at the shore. The depth of the offshore flow, the land breeze, seemed to exceed 1500 m at

0900 EST on 23 July 1964, and would have obscured the onset of the return flow in the late morning; there was, however, an increase in the offshore velocity at 900 m at the shore between 0900 EST and 1100 EST. The lake breeze onshore component was first observed at the shore at 1100 EST. At 0745 EST on 10 July 1963, Moroz observed a weak land breeze in the lowest 200 m, 5 km off shore, with indications of an overriding return flow in the land breeze circulation. Later that day, at 1710 EST, he observed a 2 m sec^{-1} lake breeze in the lowest 200 m and an overriding return flow at more than 16 km off shore. The effects of the Coriolis acceleration were clearly evident both in the onshore and in the offshore component of the lake breeze circulation. Moroz found the lake breeze front to penetrate inland in a pulsating manner and that the lake air was modified after short trajectories over land.

Observations reported by Strong (1968) were made on the western shore of Lake Michigan under essentially non-zero gradient wind conditions. Strong used single theodolite tracking of pibals in determining horizontal wind fields aloft. He calculated horizontal divergence and was thus able to evaluate vertical velocities. Upward velocities of more than 60 cm sec^{-1} were estimated 2.2 km inland and downward velocities exceeding 60 cm sec^{-1} were found between 800 m and 1000 m at 2.2 km off shore. At the surface, the first indications of a lake breeze were noticed along the shoreline, while lakeward these indications occurred aloft. Strong found the most dramatic shoreline effects to occur in warm offshore gradient wind situations and noted that with gradient winds exceeding 5 m sec^{-1} a Great Lake is of sufficient size to be modeled as an ocean. He observed intense overwater inversions, in the lowest 300 m, induced by the cold lake. Pronounced subsidence was observed above these inversions and shallow mesoscale high pressure systems were found to develop within these inversions.

Bellaire (1965) found the most intense modification of warm air moving over a cold lake to take place within 15-20 km from the shore. He observed the development of an intense inversion in the lowest 100 m and noted pronounced subsidence aloft. Strong and Bellaire (1965) noted, that under strong overwater inversion conditions, the sea surface could be smooth although the wind speed at 18 m exceeded 5 m sec^{-1}.

Richards et al (1966) analyzed 5 years of data and found the wind speed over the lake to be greater than over the upwind land areas in unstable overwater conditions, while it was less during overwater inversion conditions. This change was most pronounced in low wind speed ranges.

Lyons (1966, 1968) reported on lake breezes in the southern basin of Lake Michigan. He evaluated satellite cloud photos and compared them with cloud and smoke photos taken from the surface and from an airplane. A ring of towering cumulus associated with the lake breeze front was observed around the lake, suggesting along-shore homogeneity in the lake breeze circulation around the lake. The tops of these convective clouds were observed to be carried toward the lake in the return flow above the lake breeze and to dissipate near the lake shore. Lyons suggested that this dissipation was due to an increased stability in the lower layer, the lake breeze layer, over the lake and near the lake shore, thus a decreased convection coupled with a gravity type wave motion causing intense subsidence in the return flow layer. Subsidence over the lake was observed as smoke was carried off shore by the return flow aloft and was also evident from the lack of clouds over the lake itself. Lyons observed lake effects in cloud layers 5000 m aloft. He also reported on meteorological observations made in the Chicago, Ill., area. He found that lake breezes occasionally penetrated more than 60 km inland and that the depth of the lake breeze was approximately 300 m. At one time on 7 June 1963, the air temperature inland was more than 15°C higher than at the shore. The onshore wind component of the lake breeze was in general less than 5 m sec^{-1}. It should be noted that the urban heat island effect and increased surface roughness has to be considered in the evaluation of these observations.

Bierly (1966, 1968) has described characteristic discontinuities induced by a lake breeze. Based on observational data from shorelines along the Great Lakes he concluded, that three strata of air and thermally and dynamically induced internal boundary layers can be defined in lake breeze systems. He noted that wind shifts, temperature drops, and increases in humidity were associated with the passage of a lake breeze front, but also that strong modifications took place in the lake air in short trajectories and that the front normally penetrated inland in a pulsating manner.

Lansing (1965) observed average temperatures over land to be more than 1°C higher than over the lake during summer months. He found average maximum temperatures to be about 4°C higher. Lansing also noted that cloud cover and precipitation over the lake was less than over adjacent land areas. Pearson (1958) studied radar-observed precipitation echoes in the region around southern Lake Michigan. He noted that Lake Michigan, in general, discouraged the formation of air mass showers, while smaller lakes and urban areas caused no noticeable change in the form of the echoes. Average precipitation over Lake Michigan has been found to be up to 10% less than over adjacent up-wind land areas according to Chagnon (1961, 1967), Stout and Wilk (1962), and Williams (1964). Pettersen and Calabrese (1959), Thomas (1964), and McVehil and Peace (1965) have found that the Great Lakes have a marked effect on snowfall. Average solar radiation has been found to be considerably higher over the lakes than over adjacent land areas during the summer half of the year, Richards and Loewen (1965).

Moroz and Hewson (1966) observed the interaction of a lake breeze and a thunderstorm on the eastern shore of Lake Michigan. High frequency of thunderstorms and hail at LaPorte, Indiana have been associated with the lake breeze convergence zone and pollution from the industrial area along the southern shore of Lake Michigan, Chagnon (1968).

Slade (1962, 1966) and Van der Hoven (1967) discussed differences in dispersion rates over land and water surfaces. Slade observed that, on the average, wind direction variations decreased 50%, while wind speeds increased 25% as air traveled 10 km (7 mi) across Chesapeake Bay. The ratio between axial plume concentration over the water as compared to over the land was estimated to be 1.5 on the average. When the water surface was more than 4°C (7°F) colder than upwind land surface this ratio exceeded 2.5. Only at nights, when the water was more than 1°C (2°F) warmer than the air over land was dispersion over the water observed to be better than over land. Högström (1964) studied atmospheric diffusion by means of smoke puff photography on the western coast of the Baltic Sea. He found that air trajectories over land closely followed topographic features while dynamic downwash was evident in offshore winds, at least under neutral and unstable atmospheric conditions. Högström also discussed transitional diffusion and observed changes in wind direction shear near the shoreline. Moroz and Koczkur (1967) found that plume rise and dispersion, within a lake breeze regime on the shore of Lake Ontario, was less than what would normally be predicted on the basis of meteorological observations made at the surface.

Hewson (1945) reported on observed fumigation processes in a valley. Smoke plume behavior under certain restrictive meteorological conditions were described by e.g. Munn (1959), Bierly and Hewson (1962). They noted that fumigation can be caused by spatial discontinuities in the turbulence field, e.g. when a smoke plume passes from over rural areas in over a city at night or when its trajectory crosses a shoreline.

Leighly (1947) and Nyberg and Raab (1956) reported on measurements of temperature profiles normal to coast lines under onshore wind conditions. They found that strong modification of the air occurred in short overland trajectories. Herkhof (1969) measured horizontal temperature profiles near the eastern shore of Lake Michigan from a car. He found that the lake influence extended more than 10 km inland under onshore gradient wind conditions. Herkhof observed marked temperature discontinuities associated with lake breeze fronts within 2 km from the lake shore. As these fronts penetrated further inland, the temperature gradients across them became weaker due to strong modification of the lake air.

Several investigators have reported on observed urban heat island effects. Sundborg (1950), Summers (1965), and Daniels (1965) observed air temperatures at night in down town areas to be several degrees higher than in surrounding rural areas. DeMarrais (1961) and Munn and Stewart (1967) found that nocturnal surface inversions occurred less frequently in urban than in rural areas. Duckworth and Sandberg (1954) found that, during nights, when the gradient wind speed exceeded 6 m sec^{-1} (13 kt), there was no detectable urban heat island effect in Palo Alto, California. This critical wind speed was found to be 10 m sec^{-1} (22 kt) in London, England, by Chandler

(1960). Duckworth and Sandberg (1954) and Bornstein (1968) found that nocturnal inversions aloft occurred frequently over cities, while surface inversions were much less frequent in urban than in rural areas. Bornstein noted that the average height of the base of these elevated inversions over New York City was 310 m, almost identical to the average level at which the urban-rural temperature difference became zero. He also found that the average intensity of the heat island over New York was 1.6°C below 25 m and that the heat island occasionally extended well over 500 m. Chandler (1960, 1962), Pooler (1963), and Davidson (1967) reported on observed "nocturnal city winds" induced by the urban heat island.

Hosler (1961) found that inversions based below 160 m (500 ft) may be expected to occur about 20% to 30% of the time in any season in the Great Lakes area. Especially high frequencies of inversions, particularly during daytime hours, were observed in coastal area where the waters were cool and the frequency sea breezes was high. Hosler noted that locations only a few miles inland from a coast exhibited a continental-type frequency of low-level stability. Holzworth (1964) estimated the monthly mean maximum mixing depth to be less than 500 m in the winter and about 1000 m in the summer in the Great Lakes area, and below the averages for the contiguous U.S. He noted that the temperatures of the water surfaces have strong influences on the mixing depths.

Moses and Boyner (1967) prepared a 15 year climatological summary for Argonne, 35 km southwest of Chicago, Ill. They found that, besides a high frequency of southwesterly winds, normally the gradient wind in the region, there was also a high frequency of northeasterly winds, lake breeze winds, especially during the summer months. They noted that, on the average, the relative pressure in Argonne was lower than at the airports in Chicago (ORD and MDW) in the morning, ehile the pressure gradient was reversed in the afternoon. Moses et al (1967) found that dew point inversions occurred 41% of the time in the winter and at about 21% of the time in the summer and fall at Argonne.

Early theoretical treatments of land and sea breeze circulations were made by e.g. Jeffreys (1922), Bjerknes et al (1933), Schmidt (1947), Haurwitz (1947), and Defant (1951). Pearce (1955) developed an initial numerical sea breeze model and showed that differential heating across a shoreline produced velocity, temperature, and pressure distributions that were in general agreement with observations. Fisher (1961) modified Pearce's model and introduced surface friction and turbulent transport of heat and momentum. Estoque (1961, 1962) treated the sea breeze as a local perturbation on the general circulation flow. By specifying certain boundary layer conditions and by including non-linear advective terms, Estoque obtained results that describe the observed features of sea breeze circulations under various synoptic wind conditions. Modifications of the boundary layer have been made by Estoque and Bhumralkar (1968). McPherson (1968) extended Estoque's model into three dimensions and inverstigated the models response to variations in some of the modelling assumptions. Geisler and Bretherton (1969) used a linear theory to study the initiation of a sea breeze circulation.

Moroz (1965, 1967) developed a numerical lake breeze model based on Estoque's sea breeze model. By introducing a boundary in the middle of the lake he found the model to produce results, which were in good agreement with observed features of a lake breeze circulation. Wilson (1967) studied the response of Moroz's model to various heating functions and found that the model was unable to correctly describe a land breeze circulation.

Several theoretical investigations of variations in the boundary layer, in response to changes in the characteristics of the underlying surface, have been reported, e.g. Elliot (1958) and Panofsky and Townsend (1964). Modelling of regional air pollution dispersion have been described and discussed by e.g. Turner (1964), Miller and Holzworth (1967), Calvert (1967), Hilst et al (1967), and Slade (1967).

2. CLASSICAL LAND AND LAKE BREEZE CIRCULATION: AN OBSERVATIONAL STUDY

Since 1963 the Department of Meteorology and Oceanography at the University of Michigan has been conducting observational studies of the mesoscale wind systems and related temperature and moisture structures near the southern basin of Lake Michigan. Intensified studies have been conducted mainly during spring and early summer, when the daytime temperature difference between land and water is at a maximum and well defined lake breeze circulations occur. One particularly well documented lake breeze, which occurred on the eastern shore of Lake Michigan on 25 June, 1965, is presented in this chapter. A summary of these observations has been reported and a complete tabulation of observed meteorological variables prepared, Olsson et al (1968 a,b).

2.1 LOCATION OF STUDY AND OBSERVATIONAL PROGRAM.

Observations were made at a site on the eastern shore of Lake Michigan near Grand Haven, Michigan. The relative location of the site and the observational stations of primary concern are indicated in Figure 1. The coordinate system used throughout this report is such that the x-axis coincides with "the observation line" along Michigan highway 45 (M-45) and is positive to the east, the y-axis coincides with the coast south of origin (the intersection of M-45 and the shoreline) and is positive to the north. The altitude of the ground surface is everywhere set to be zero and the z-axis is positive upward.

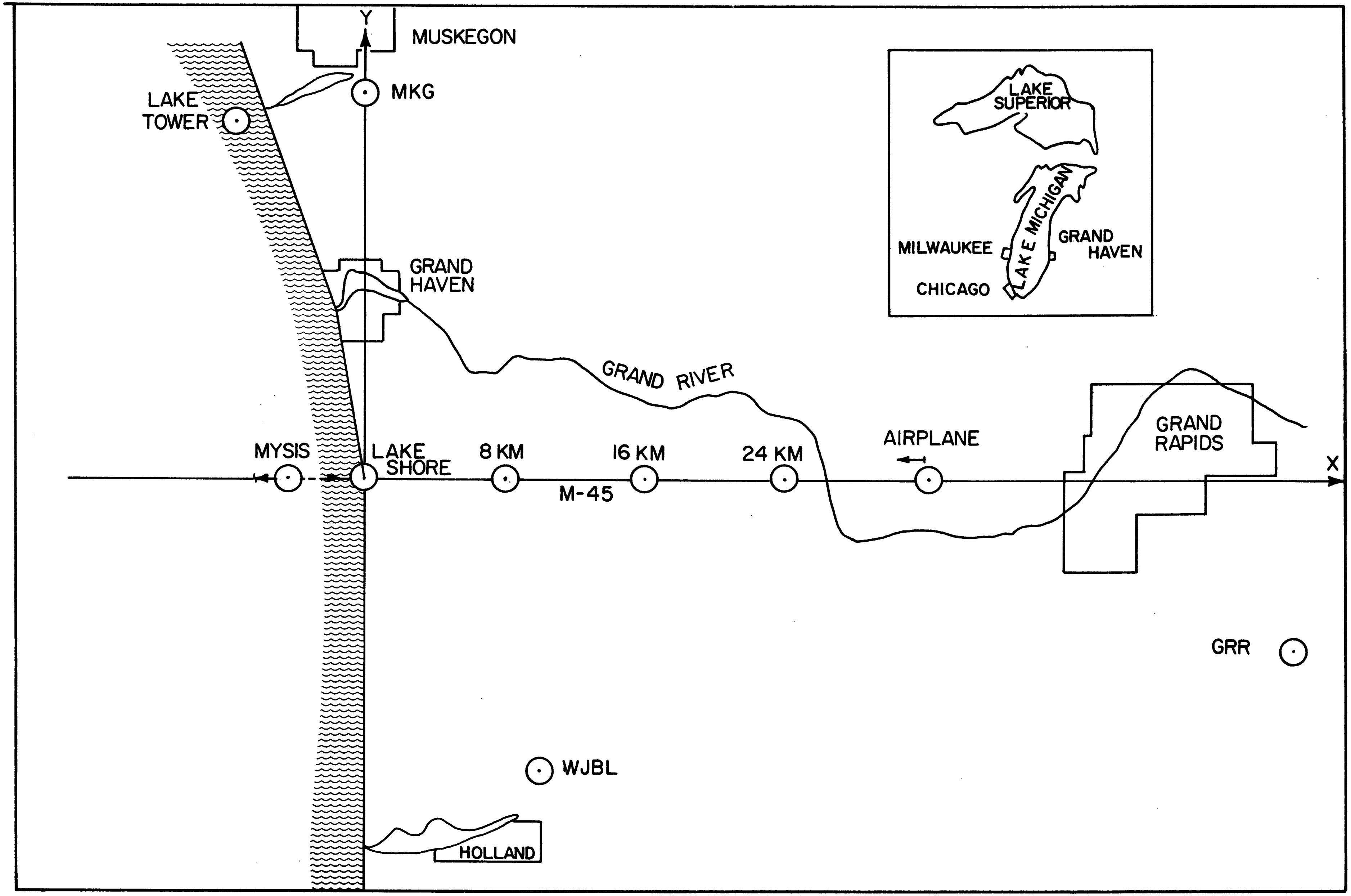

FIGURE 1. Location of observation stations on the eastern shore of Lake Michigan. Insert shows the location of the line of observations relative to Lake Michigan.

The site selected offers several distinct advantages. The lake is long and narrow (approximately 110 km wide) with its axis running nearly north-south, the curvature of the shoreline is small and does not strongly modify the flow patterns, and along-shore homogeneity can for all practical purposes be assumed in the neighborhood of the observation line. Sand dunes run parallel to the sand beach, with heights of up to 50 meters above mean lake level. The dunes are in general confined to within 1 km from the shore, and are partly covered with trees. Two or three km inland from the lake the land becomes flat and is uniformly developed for agriculture. The slope of the land from Grand Rapids toward the lake shore is small (approximately 1/1000) and it is assumed that no slope-winds develop.

A short description of the stations and their instrumentation follows:

Lake shore station (0 km) was located on the lawn at the Grand Rapids pumping station, in a 200 m wide break in the sand dune chain. A USWB-type instrument shelter was located on a lawn 50 m from shore, 15 m above mean lake water level, and 20 m lakeward (west) of the 10 m high pumping station, Figure 2. Temperature and relative humidity were recorded on a Bendix Friez Model 594 recording hygrothermograph in the shelter. Due to a failure in the recording device, no surface winds were obtained for 25 June. The winds used as representative for this station are averaged from observations obtained from nearby U.S. Coast guard stations. Winds aloft were measured by either singel or double theodolite tracking of pilot balloons. Figure 3

FIGURE 2. Lake shore station. An USWB-type instrument shelter at the foot of the 12.2 m meterological tower, which supports wind sensors and a radio telephone antenna.

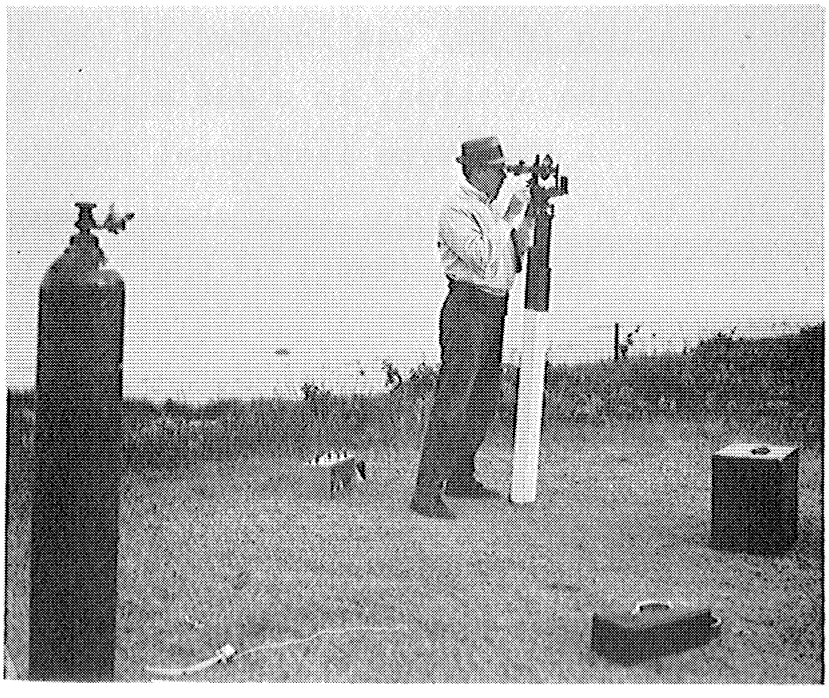

FIGURE 3. The north theodolite site at the Lake shore station viewed from the south-east. Lake Michigan in the background.

shows a theodolite crew at work at the lake shore station. The theodolites were located on the beach, south of the station, with a 305 m (1000 ft) north - south baseline. A Belfort Recording Pyrheliometer was located on the roof of the pumping station, approximately 10 m above the ground.

8 km station was located on a farm in flat terrain. A USWB-type instrument shelter, located approximately 100 m east of the farm buildings, housed a Bendix Friez Model 594 hygrothermograph. Close by the shelter, on the 12.2 m level of a 24.4 guyed steel tower was located a Science Associates #418 wind direction sensor and a Science Associates #402 anemometer, Figure 4. Wind speed and direction were recorded on an Esterline-Angus 20 pen recorder, Figure 5. Two theodolites were located on an open field to the south of the station with a 305 m north-south baseline, Figure 6.

16 km station was located on the lawn of a field station of the Michigan Department of Conservation. The lawn was shielded from open fields by hedges to the north and west, and by a small built-up area to the southeast. A USWB-type instrument shelter with a Bendix Friez Model 594 recording hygrothermograph and a 12.2 m guyed steel tower with a top mounted Science Associates #402 3-cup anemometer were located on the lawn. Wind speed and direction were recorded on an Esterline-Angus 20 pen event recorder. A station for single theodolite tracking of pibals was located on an open field south of the conservation station.

24 km station, with a Bendix Friez Model 594 recording hygrothermograph in an USWB-type instrument shelter, was located with open fields towards south and with wooded

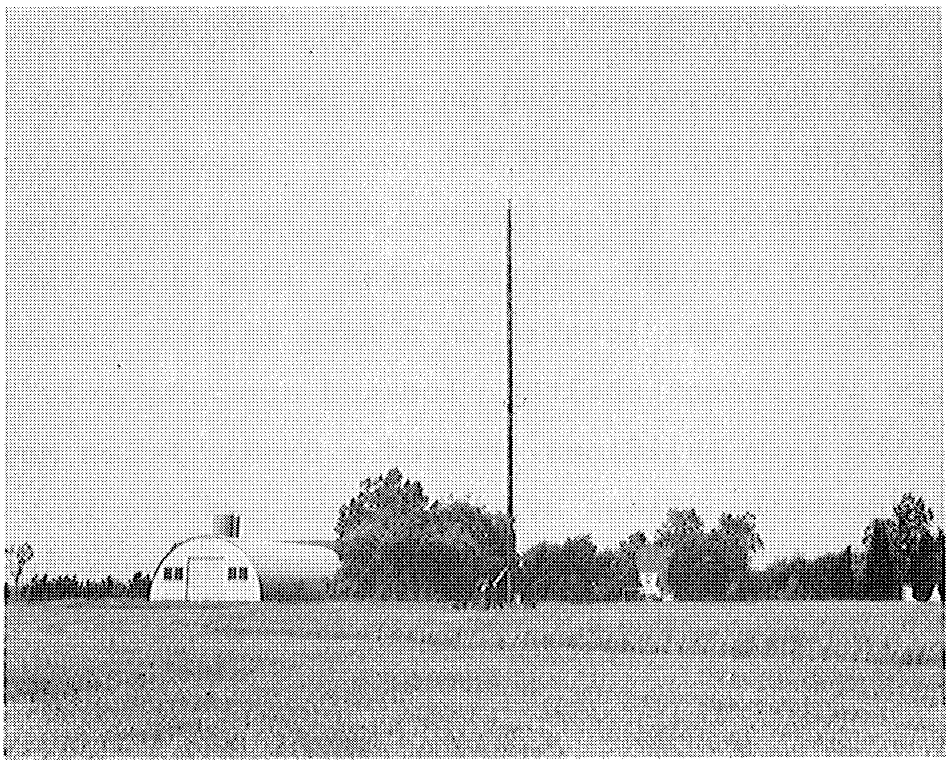

FIGURE 4. The meteorological tower at the 8 km station viewed from the east.

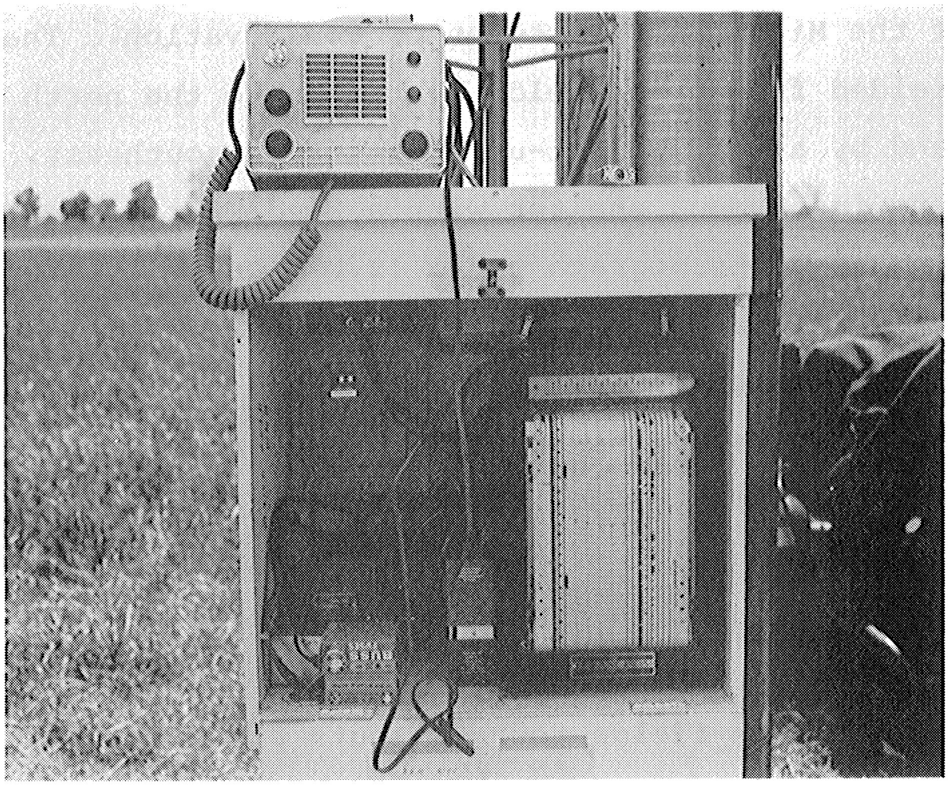

FIGURE 5. An Esterline-Angus 20 pen event recorder for wind speed and direction recording and a Messenger II, 5 watt Citizen's Band radio telephone.

areas toward north.

WJBL station is a radio station located north of Holland, Michigan; x = 10 km and y = -17 km. The sensors were mounted on a 91.4 m high radio transmitting tower located on an open field, Figure 7. Copper constantan thermojunctions in Thornthwaite shields at the 2.4, 4.9, 9.8, 19.5, 39.0, and 78.0 m levels sensed air temperatures and copper constantan thermojunctions at 0.01, 0.10, and 1.00 m depths in the ground sensed soil temperatures. These temperatures were recorded on a Honeywell - Brown multipoint recorder. Electric Speed Indicator 3-cup anemometers and wind vanes type F-420-C were mounted at the 19.5, 39.0, and 78.0 m levels. Wind speeds and directions were recorded on Esterline-Angus 0-1 ma dual strip-chart recorders.

Airplanes, type Cessna 172, flew trajectories in a vertical plane along the observation line. They were equipped with Friez aerometeorographs for temperature and relative humidity recording, Figure 8, and with Yellow Spring Instruments Corporation thermistors, mounted in radiation shields, for instantaneous readings of air temperature and wet bulb temperature, Figure 9.

MYSIS is a fifty foot, steel hulled research vessel belonging to the Great Lakes Research Division of the Institute of Science and Technology, The University of Michigan, Figure 10. She operated mainly 1.6 km to 6.4 km off shore along the observation line. Bow and beam winds with respect to the vessel were measured with an Electric Speed Indicator 3-cup anemometer and wind vane type F-240-C mounted on the top of the ship's mast, 12.1 m above the

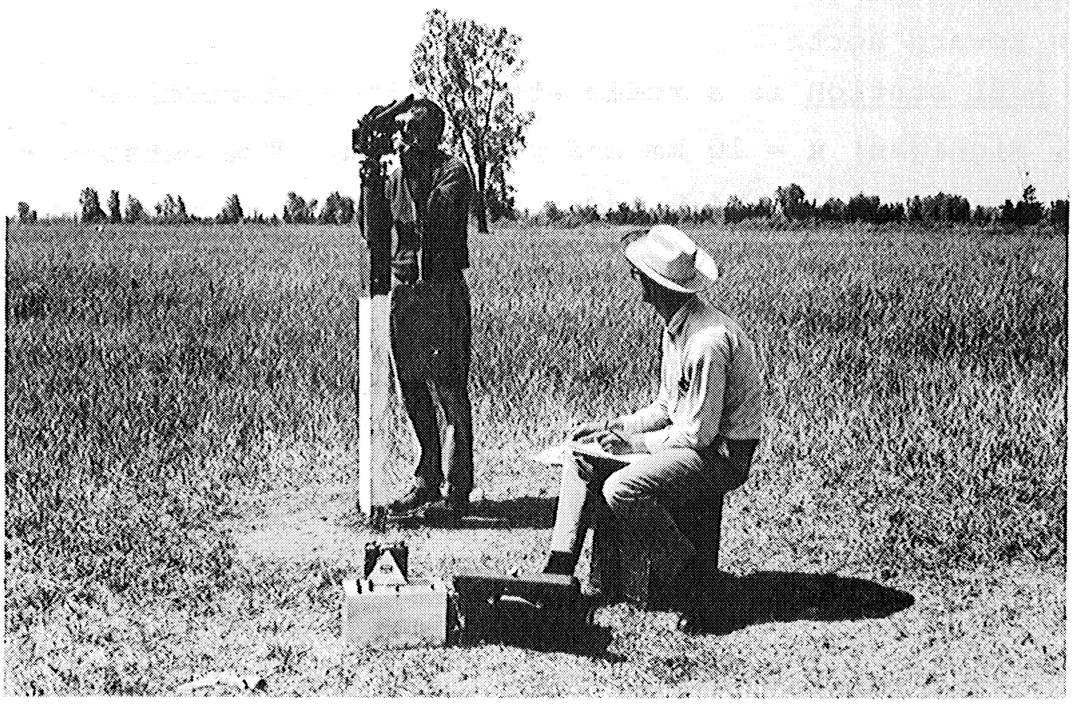

FIGURE 6. The north theodolite site at the 8 km station viewed from the west.

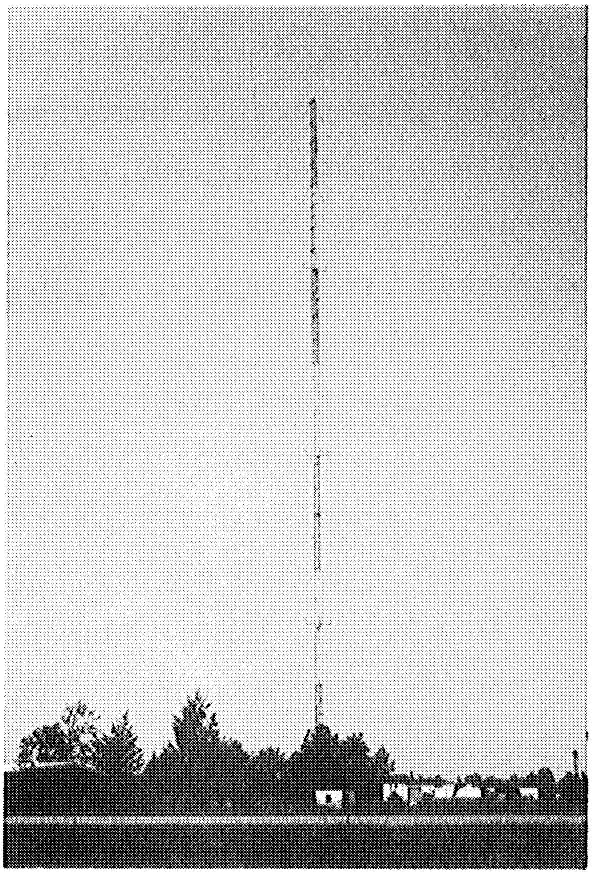

FIGURE 7. The WJBL, 91.4 m high, radio transmitting tower supporting wind and temperature sensors. The recorders are housed in the shelter at the foot of the tower.

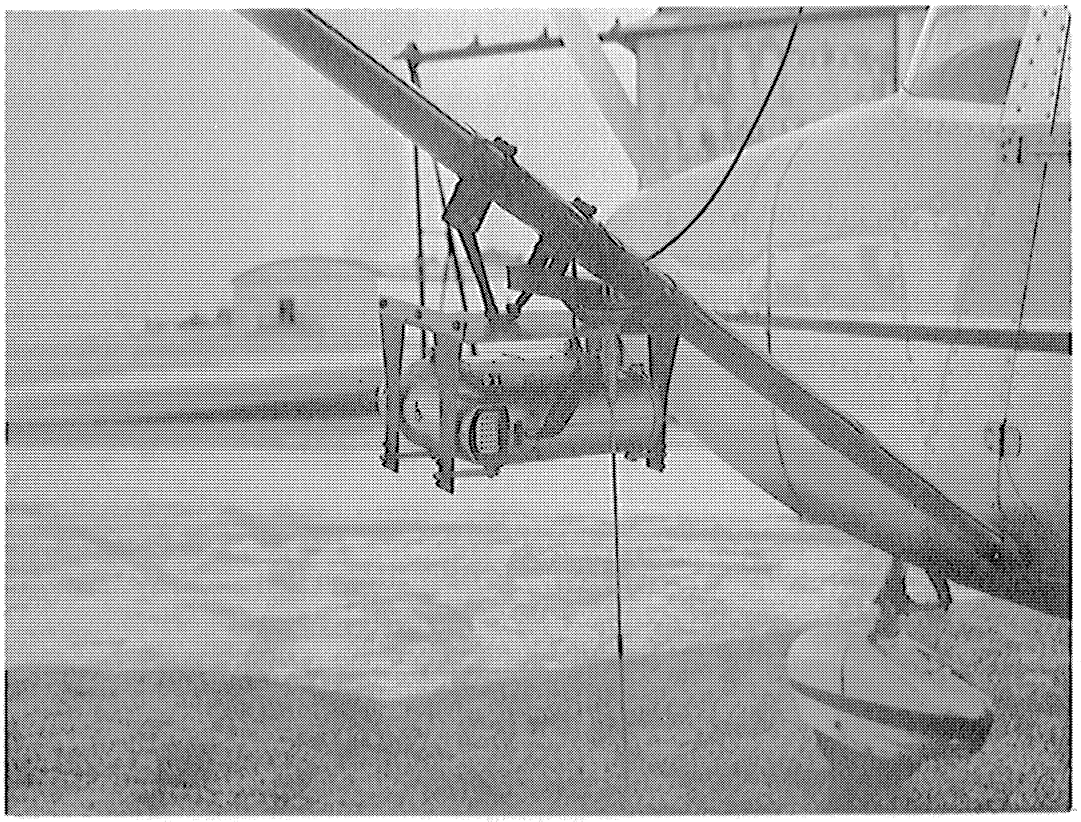

FIGURE 8. A Friez aerometeorograph, mounted on left wing strut of a Cessna 172 aircraft.

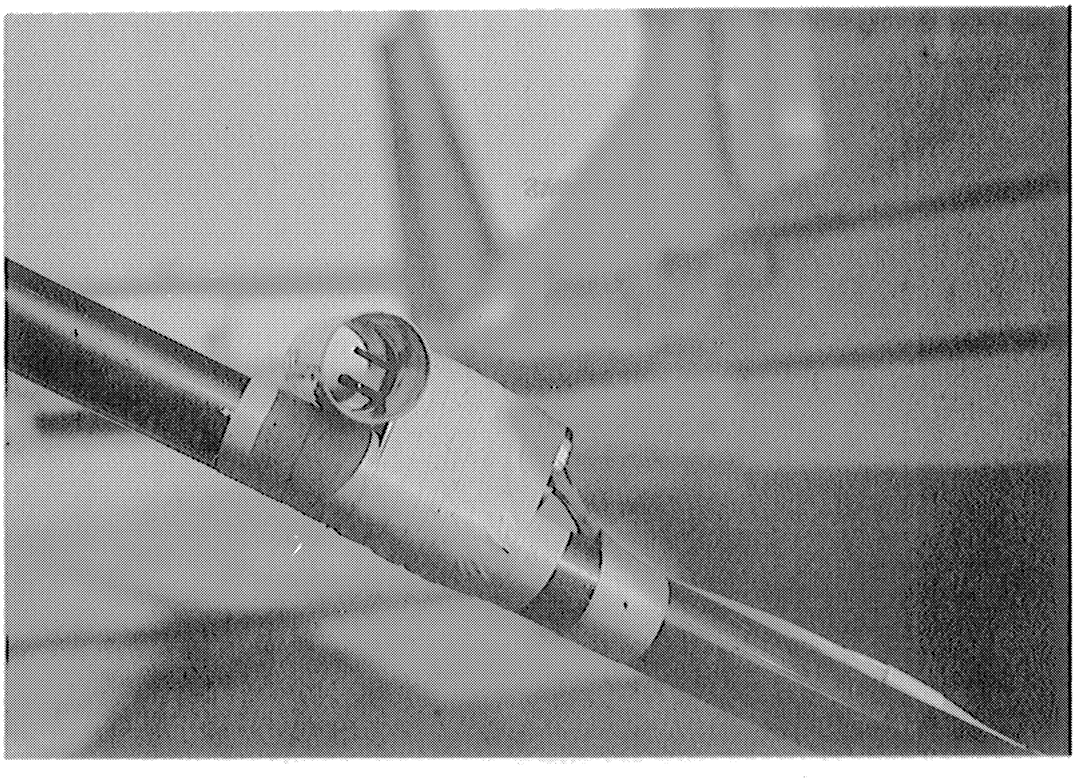

FIGURE 9. Yellow Spring Instrument Corporation Thermistors, for dry and wetbulb temperature sensing, in radiation shield mounted on right wing strut of a Cessna 172 aircraft.

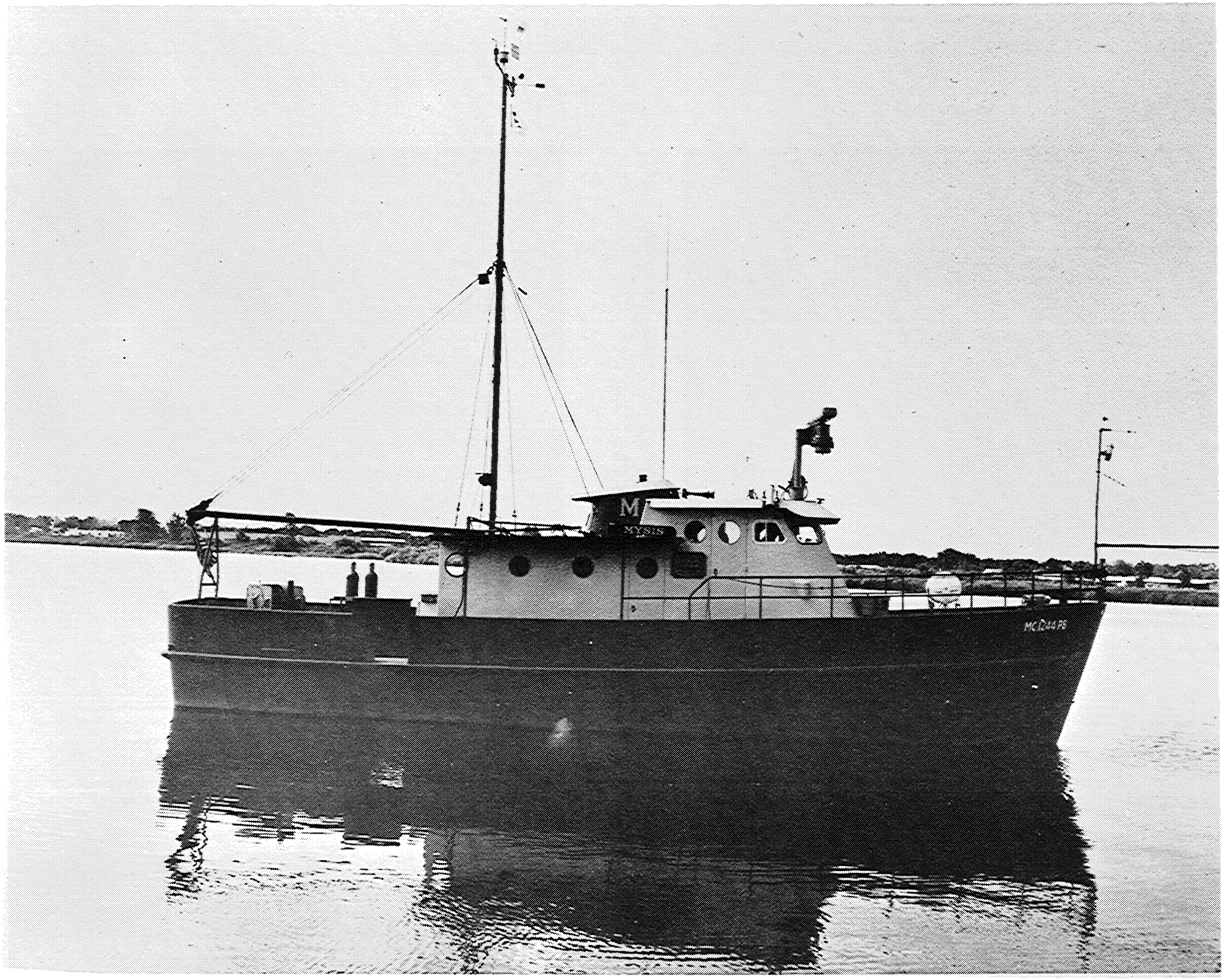

FIGURE 10. The R/V MYSIS equipped with temperature, humidity, radiation and wind sensors.

water surface. Air temperature was measured with a Rosemont Engineering platinum resistance thermometer system with sensors mounted in Thornthwaite-type radiation shields 11.8 m above the water on the mast and at 5.1 m above the water on the bowsprit. Water temperature was sensed at 1.2 m under the water surface. Humidity was sensed with a Honeywell lithium chloride dew probe on a bow mounting at 5.0 m above the water. Solar radiation was sensed by an Eppley solar radiation sensor mounted on the roof of the wheelhouse. Ship speed and heading were set manually, as was the date, while a digital clock produced time signals with one minute resolution. All instruments onboard the Mysis were recorded on an Information Instruments Inc. data logger model 641. This data logger converts analog voltage signals to digital form and records them on punched paper tape. Sampling time was in the order of milliseconds, with each sensor sampled and recorded once per minute. Releases of pilot balloons, for either shipboard tracking with marine theodolite or shoreline tracking with standard theodolites, were made.

All instruments on the previously mentioned stations were checked frequently and corrected, e.g. the hygrothermographs were checked once a day against sling psychrometers. Wind direction and speed sensors were mounted on towers, in accordance with recommendations set forth by Gill et al (1967), to give an accuracy of ± 5% in wind speed and ± 5° in wind directions. Soo and Elder (1967) found that errors in wind speed measurements from well exposed anemometers on a ship would be about ± 5%. Stevenson (1964) noted that a ship sometimes exerted a strong influence on surrounding air and water temperatures.

Communication between the different observational sites was provided by three Messenger II, and one Hallicrafters transistorized, 5 watt, Citizen's Band radio telephones. At the locations for double theodolite tracking, field telephones and buzzers were used between the two theodolites for coordination and timing.

Smoke plumes were generated near the lake shore and onboard the *Mysis* by means of several York-Hession oil-fog generators. The plumes were photographed both from the air and from the surface in an attempt to estimate diffusion characteristics in transitional states.

MKG is a USWB station located on the eastern side of Muskegon airport, 6 km inland from the shoreline, at coordinates x = 0, and y = 21 km.

GRR is a USWB station located at Grand Rapids airport at x = 53 km and y = -10 km.

Both MKG and GRR make routine observations on the hour.

Lake Tower is a research tower belonging to the Great Lakes Research Division and it is located at x = -7 km and y = 22 km, 1.8 km off shore, Elder (1964). Wind speeds and directions are sensed by an Aerovane mounted 15.0 m above the water and recorded on a data logger.

Auxiliary observations made at USWB-staions, U.S. Coast Guard stations, and onboard ships on the Great Lakes have been used in describing the synoptic situation.

Most of the data presented in this chapter have been abstracted and analyzed at the Department of Meteorology and Oceanography of the University of Michigan, and are punched on IBM cards. Some of the data are stored on magnetic tape. The data have also been presented in tabular form, Olsson et al (1968 b).

Winds aloft were determined by means of theodolite tracked, 30 gram, pilot balloons. A summary of these observations is presented in Table I. The balloons were inflated with hydrogen at the land stations giving an ascent rate of 183 m min^{-1} (600 ft min^{-1}) and onboard the ship with helium (for safety reasons) to have an ascent rate of 180 m min^{-1}. Theodolite observations were made at 1/2 minute intervals to permit analysis for winds through thin layers in order that details of flow changes with height could be discerned. Barnett and Clarkson (1965) have found that 20 second reading intervals give greatest accuracies in determination of balloon movements by means of double theodolites, but noted that the accuracies were only slightly decreased when longer time intervals were used. Analyses of the winds aloft observations were made on a digital computer, using an evaluation technique described by Biggs (1962) for cases of double theodolite tracking. No weighting functions were used to smooth the data, although some smoothing is introduced by using the conventional technique of averaging over two layers to evaluate the wind at the midpoint. As can be seen from Table 1 some of the balloon flights were analyzed, both as double and as single theodolite runs in order to determine errors made by assuming a constant ascent rate.

TABLE 1. A summary of available winds aloft observations for 25 June, 1965. "D" indicates that double theodolite tracking and analyses were used and "S" that single theodolite analyses were made. Superscript 1 indicates "balloon released from Mysis", from distance off shore as given in km in the Mysis column. Numbers in brackets are duration of tracks in minutes.

Time	Lakeshore	8 KM	16 KM	MYSIS
0600	D (20.0)			
0630	S^1(13.0)	S (20.0)		-1.6*)
0700		S (20.0)		*) Not tracked from MYSIS
0730		S (20.0)		
0800	D (20.0)	S (20.0)		
0830	D (20.0)	S (20.0)		
0900	S^1(20.0)	S (20.0)		-1.6 (12.5)
0930	D (20.0)	S (20.0)		
1000	D (20.0)	S (20.0)		
1030	D (20.0)	S (20.0)		
1100	D (13.0)	D (20.0)		-6.4 (12.5)
1130	D (19.5)	D (20.0)		
1200	D^1(20.0)	D (20.0)		-1.3 (12.5)
1230	S (20.0)	S (20.0)		
1300	S (20.0)	S (20.0)		
1330	D (20.0)	D (20.0)		
1347				-6.4 (12.5)
1400	D (20.0)	D (18.0)		
1430	D (20.0)	D (17.0)		
1500	S (20.0)	D (20.0)		
1530	S (20.0)	D (15.0)		
1600	S (20.0)	D (6.0) S (20.0)		
1630	S (20.0)	D (15.0)		
1700	S (20.0)	D (15.0)	S (15.0)	
1730	S (20.0)		S (15.0)	
1800	D (9.0) S (20.0)	D (15.0)	S (7.0)	
1830	D (20.0)	D (11.0)	S (8.5)	
1900	D (20.0)	D (15.0)	S (14.5)	
1930			S (15.0)	
1945	D (20.0)			
2000		S (15.0)		
2015	S (8.5)			
2030	D (6.5) S (15.5)			

These comparisons showed that the inherent error of wind analyses using single theodolite observations in very few cases exceeded ± 1 m $\sec^{-1}$ for land based theodolites and only in cases of balloon elevations over 1500 meters for the shipboard theodolite. In general, the accuracy estimated by Frizzola and Fisher (1963) and by Moroz (1965) of ± 2 m $\sec^{-1}$ for single theodolite observations in sea and lake breezes holds for the results presented here.

2.2 PREVAILING METEOROLOGICAL CONDITIONS, 25 JUNE, 1965.

On 25 June, 1965 the Lake Michigan basin was under the influence of a cold surface high, with a very weak pressure gradient over the entire Great Lakes region, Figure 11. During the day "mesoscale lake highs" developed over Lake Michigan and Lake Huron. In the late afternoon these highs had reached intensities of 3 - 4 millibars, Figures 12 - 14. Surface winds around Lake Michigan were light and variable in the morning hours. In the afternoon the surface winds reported from stations close to the shore, i.e. Muskegon, Chicago, Milwaukee, and Escanaba indicated lake effects around the entire lake. Similarly surface winds from stations around the other Great Lakes show clear evidence of lake effects. The winds aloft were weak and variable as a high pressure ridge moved in over the area, Figures 15 - 20 and lower insert on Figure 11. No clouds were sighted near the eastern shore of the lake at any time during the day. The solar radiation measured at the Lake shore is presented in Figure 21. The total intergrated solar radiation for 25 June, 1965 was 672 cal cm^{-2} day^{-1}, which is

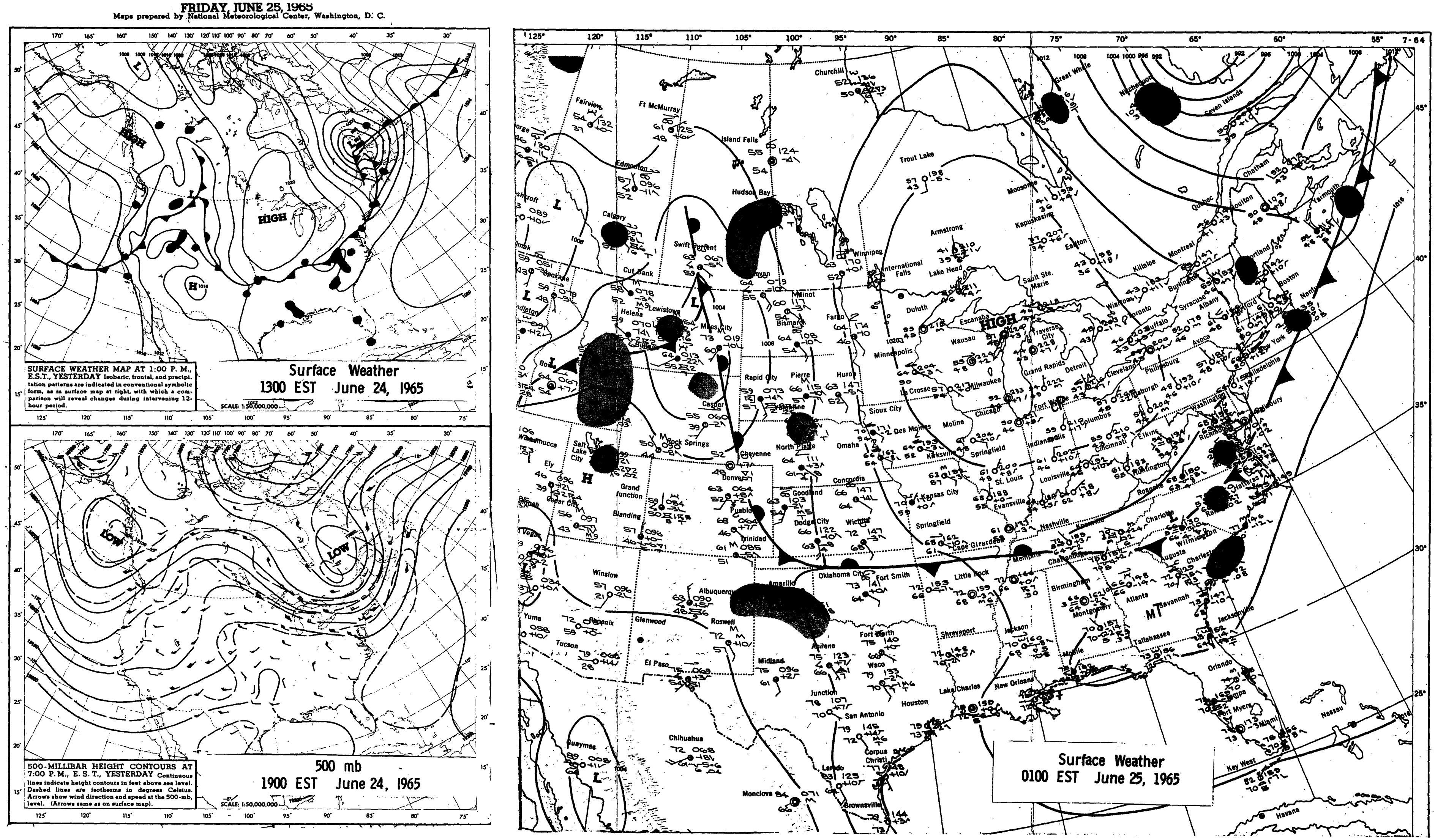

FIGURE 11. U.S. Weather Bureau Daily Weather Map at 0100 EST, 25 June, 1965. Upper insert shows the sea level isobars at 1300 EST, 24 June and the lower insert shows the height contours for the 500 mb surface at 1900 EST, 24 June.

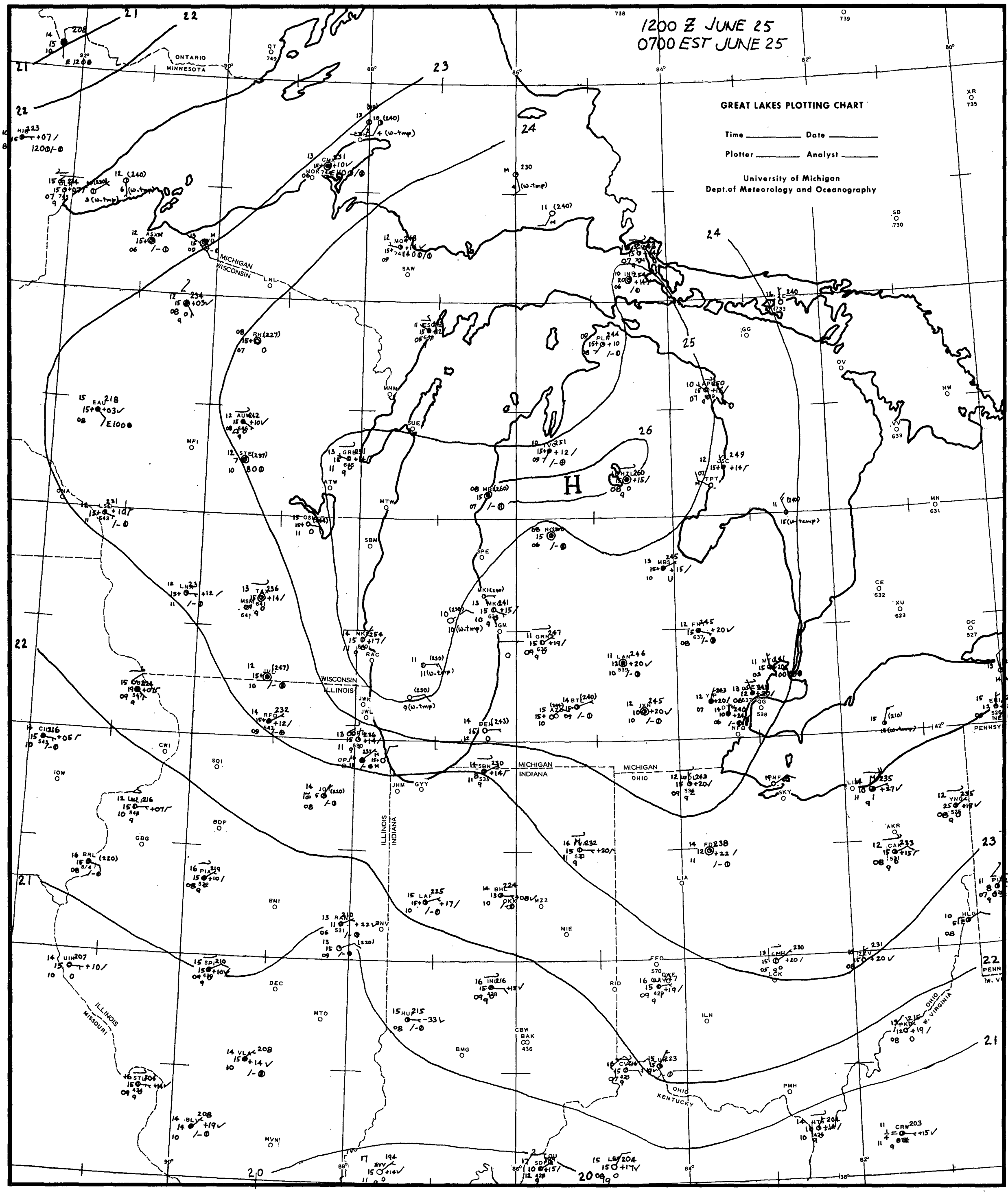

FIGURE 12. Mesoscale surface weather map and station weather at 0700 EST, 25 June, 1965.

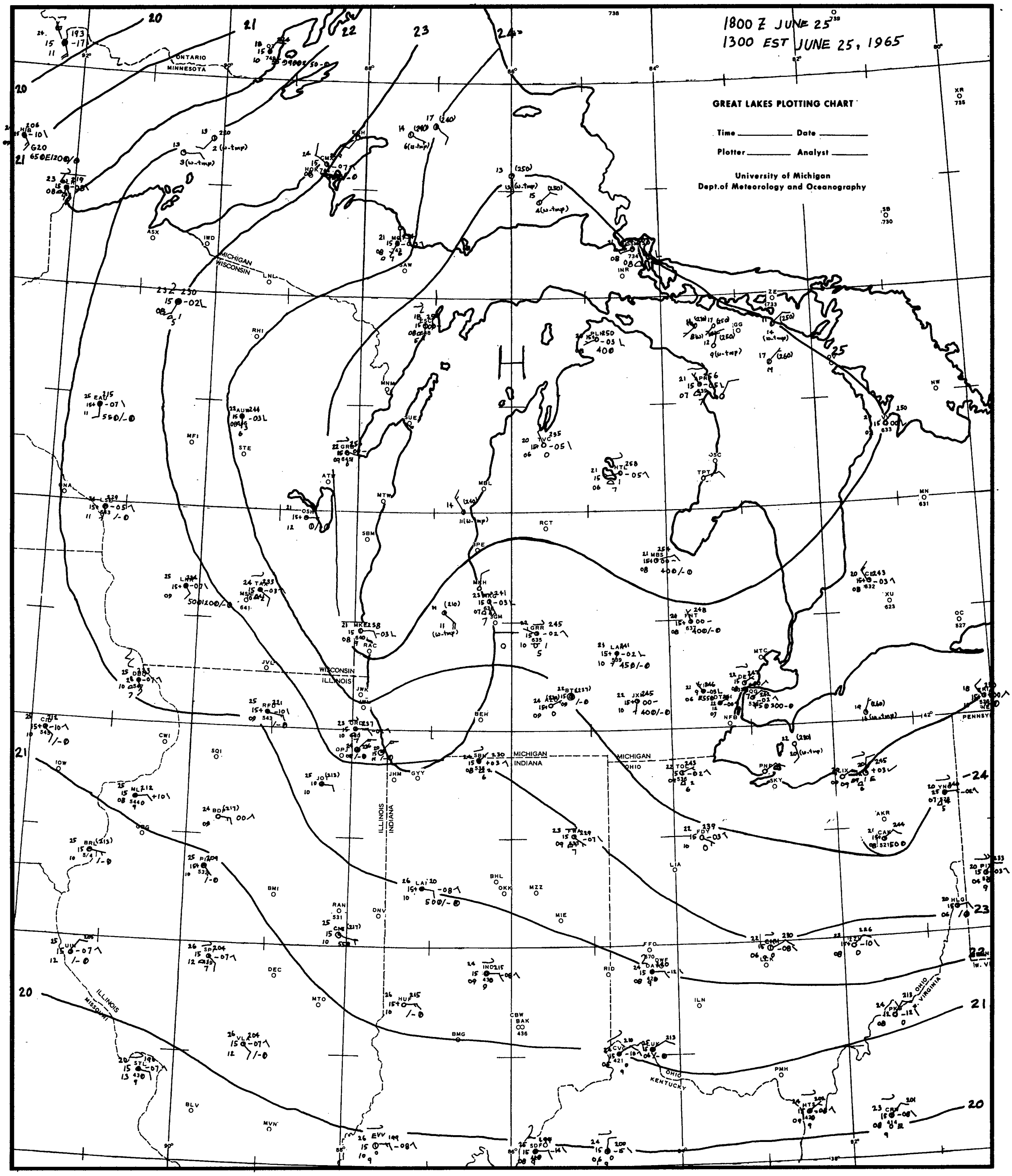

FIGURE 13. Mesoscale surface weather map and station weather at 1300 EST, 25 June, 1965.

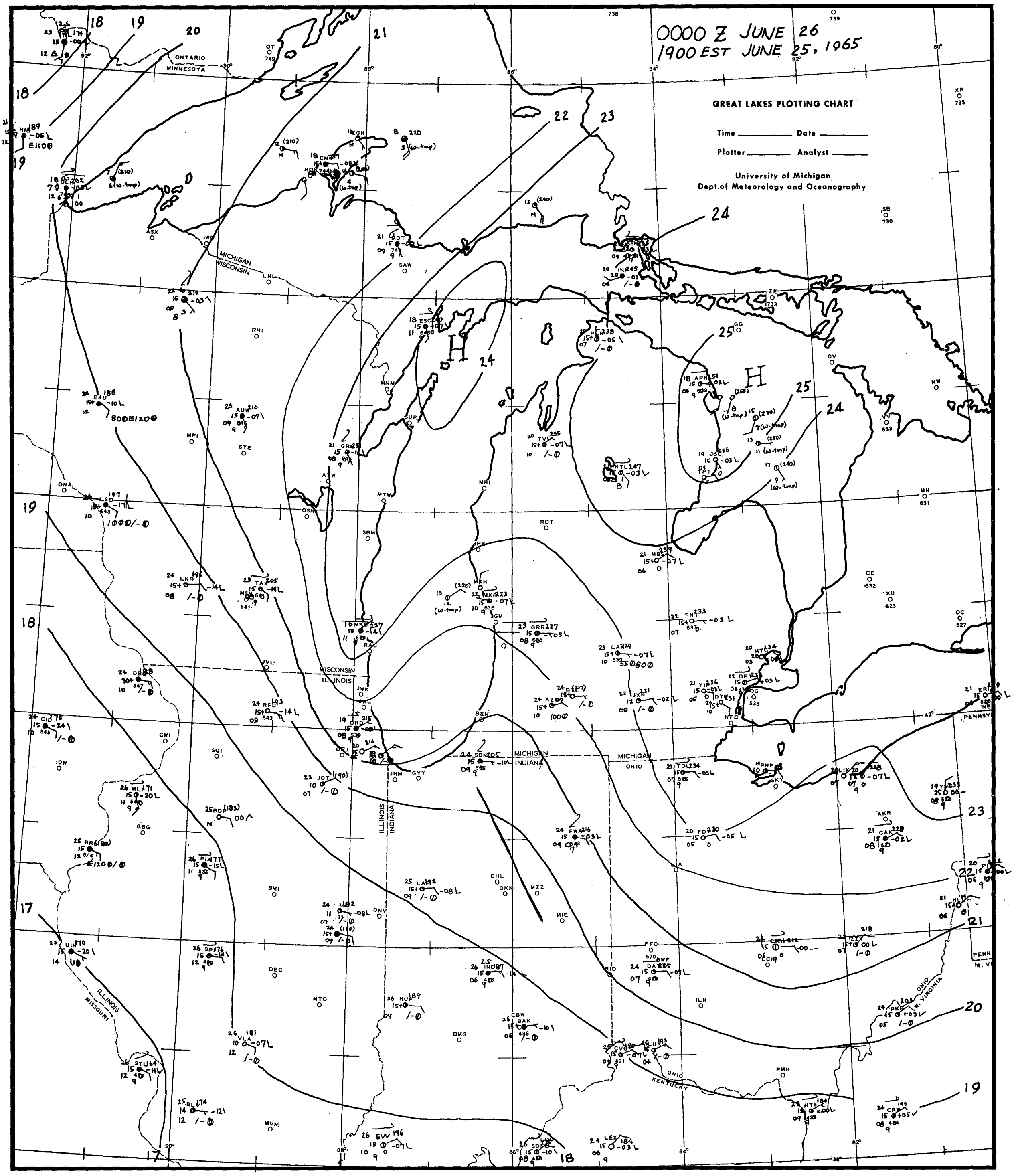

FIGURE 14. Mesoscale surface weather map and station weather at 1900 EST, 25 June 1965.

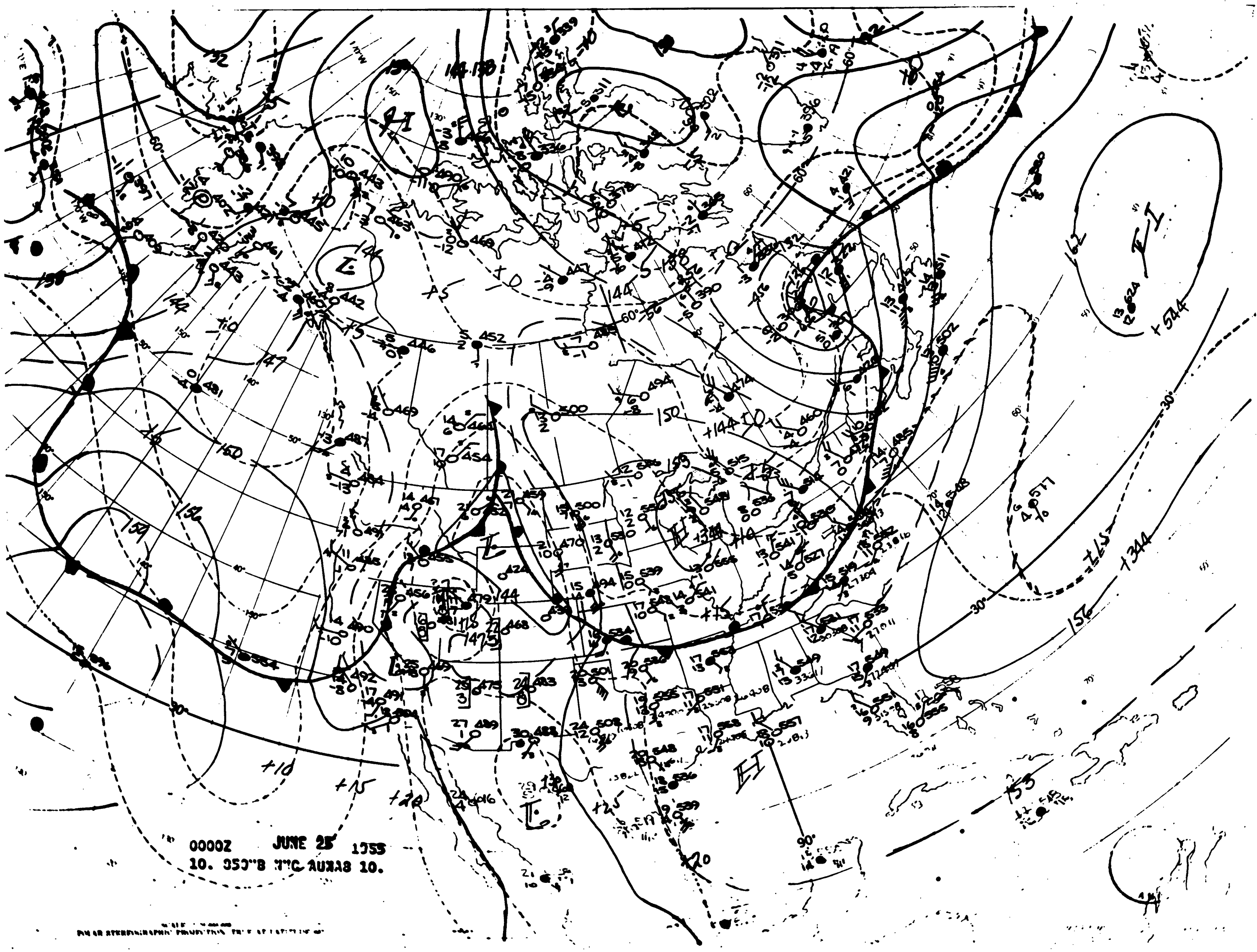

FIGURE 15. Height contours for the 850 mb surface at 1900 EST, 24 June, 1965.

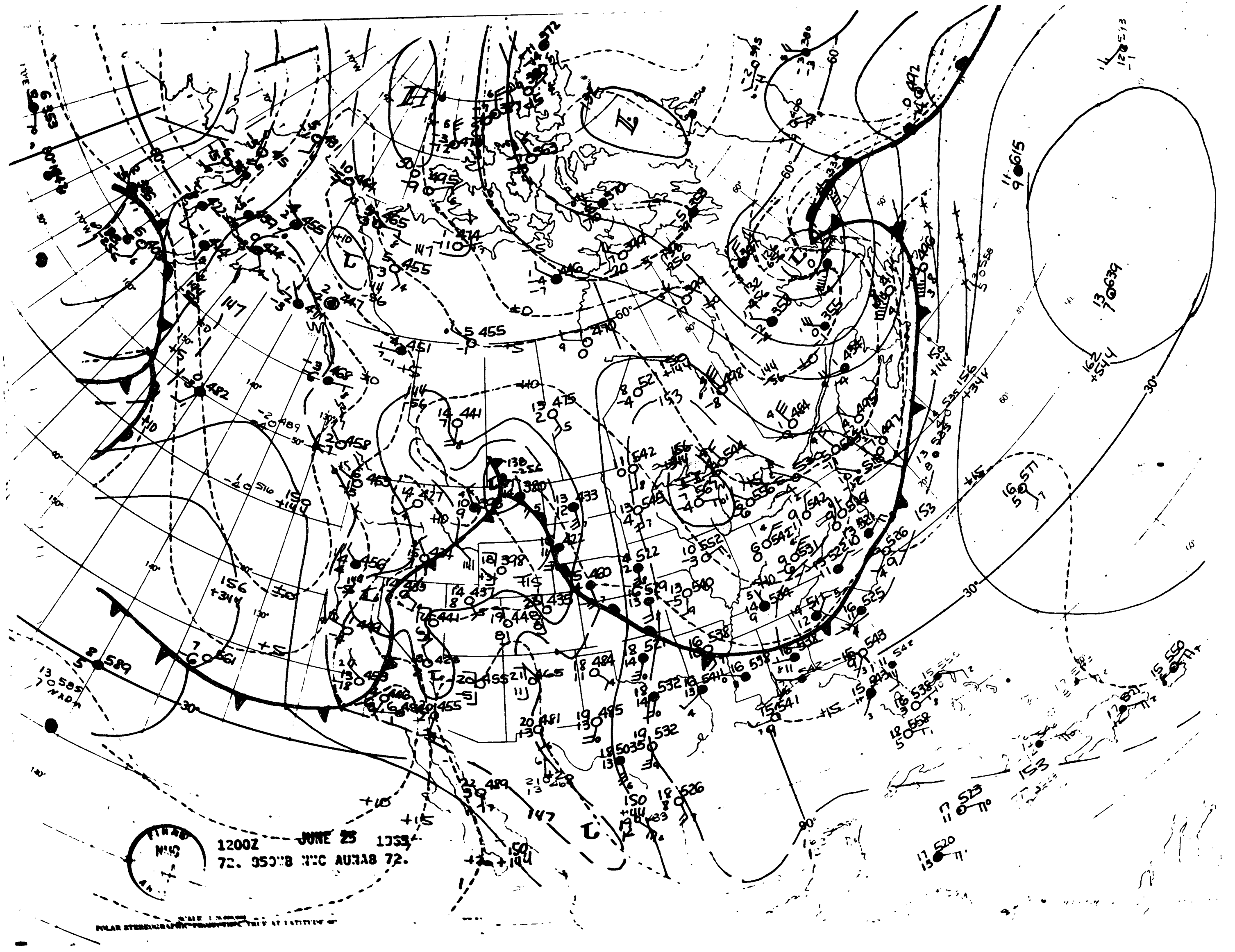

FIGURE 16. Height contours for the 850 mb surface at 0700 EST, 25 June, 1965.

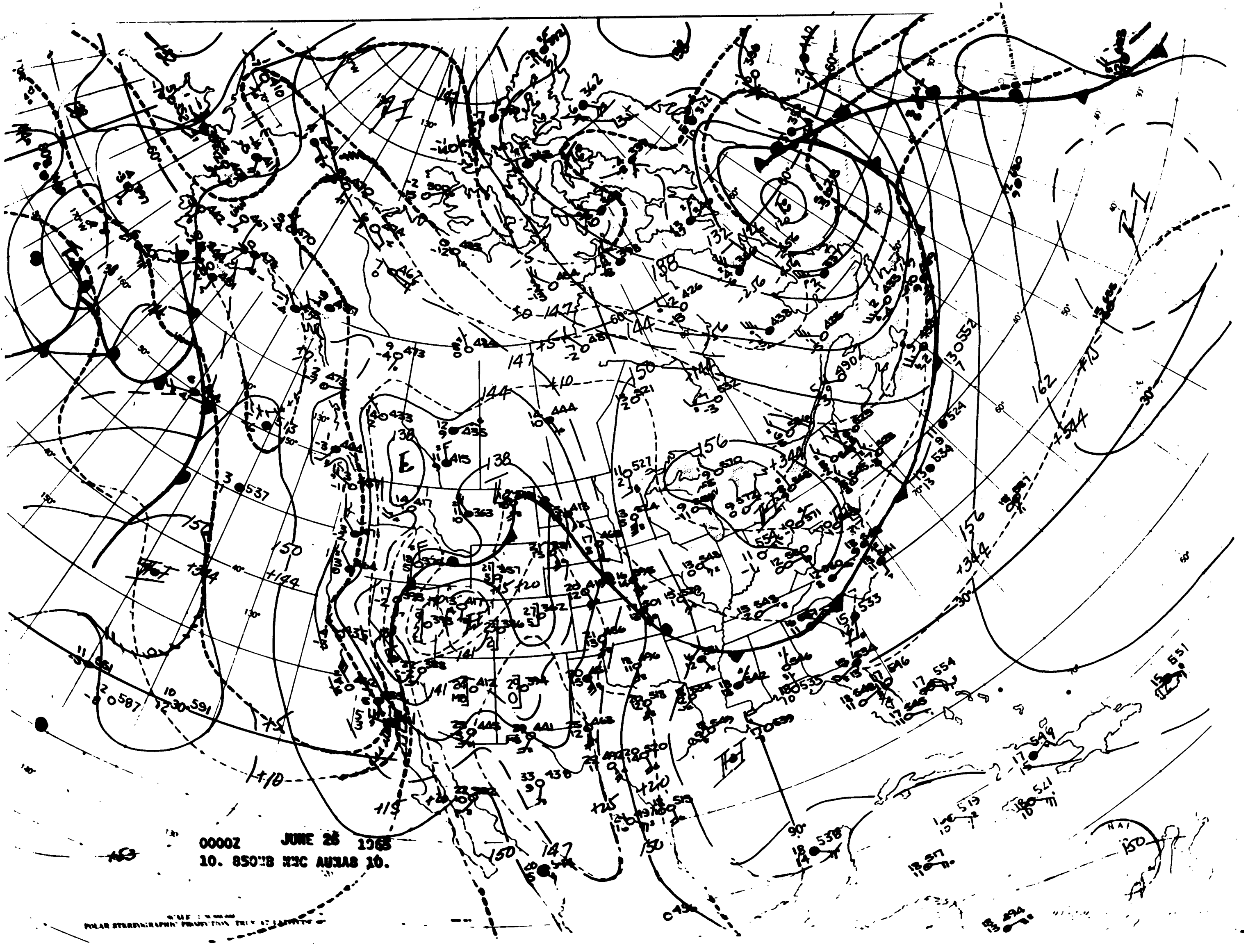

FIGURE 17. Height contours for the 850 mb surface at 1900 EST, 25 June, 1965.

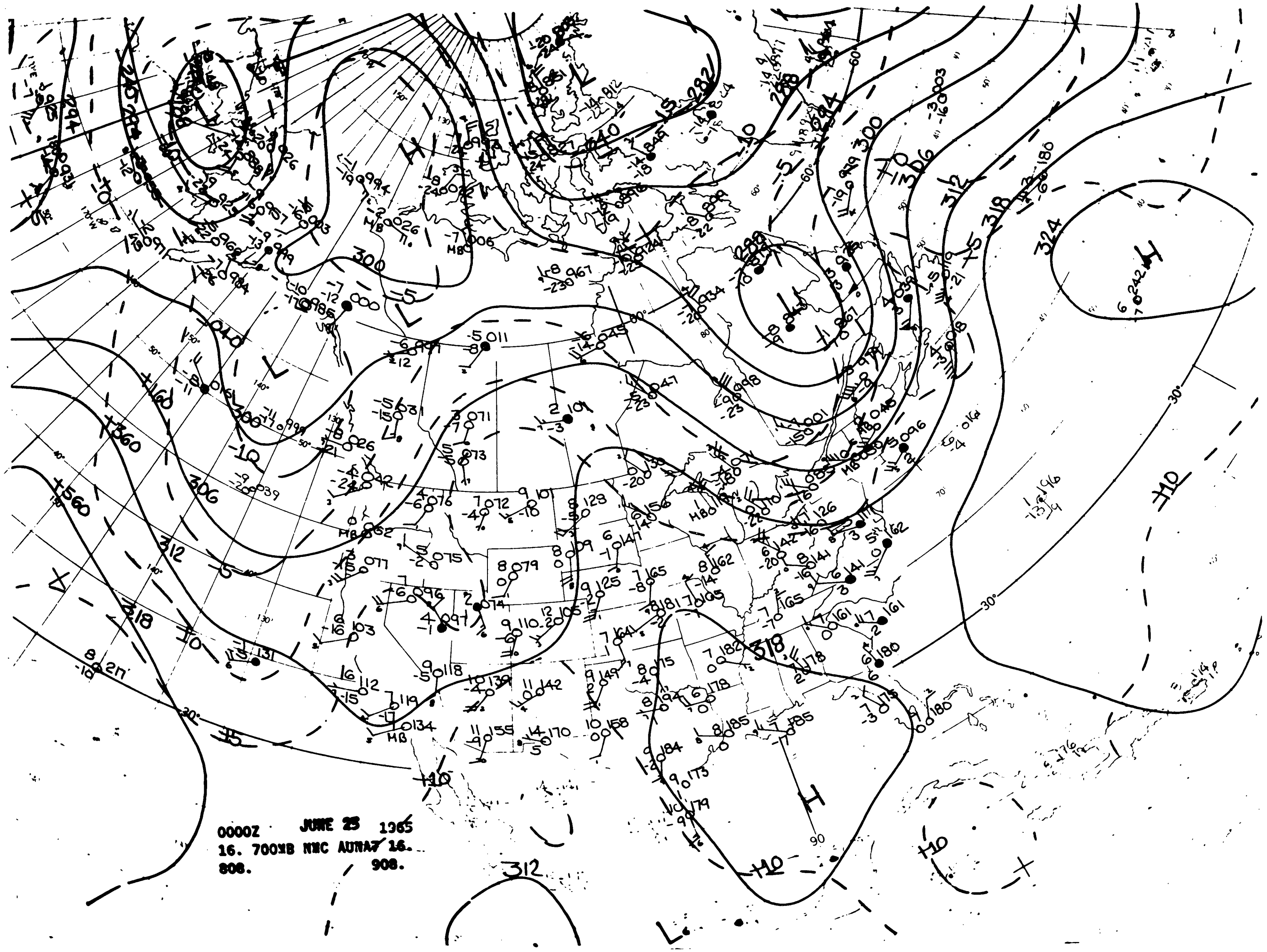

FIGURE 18. Height contours for the 700 mb surface at 1900 EST 24, June, 1965.

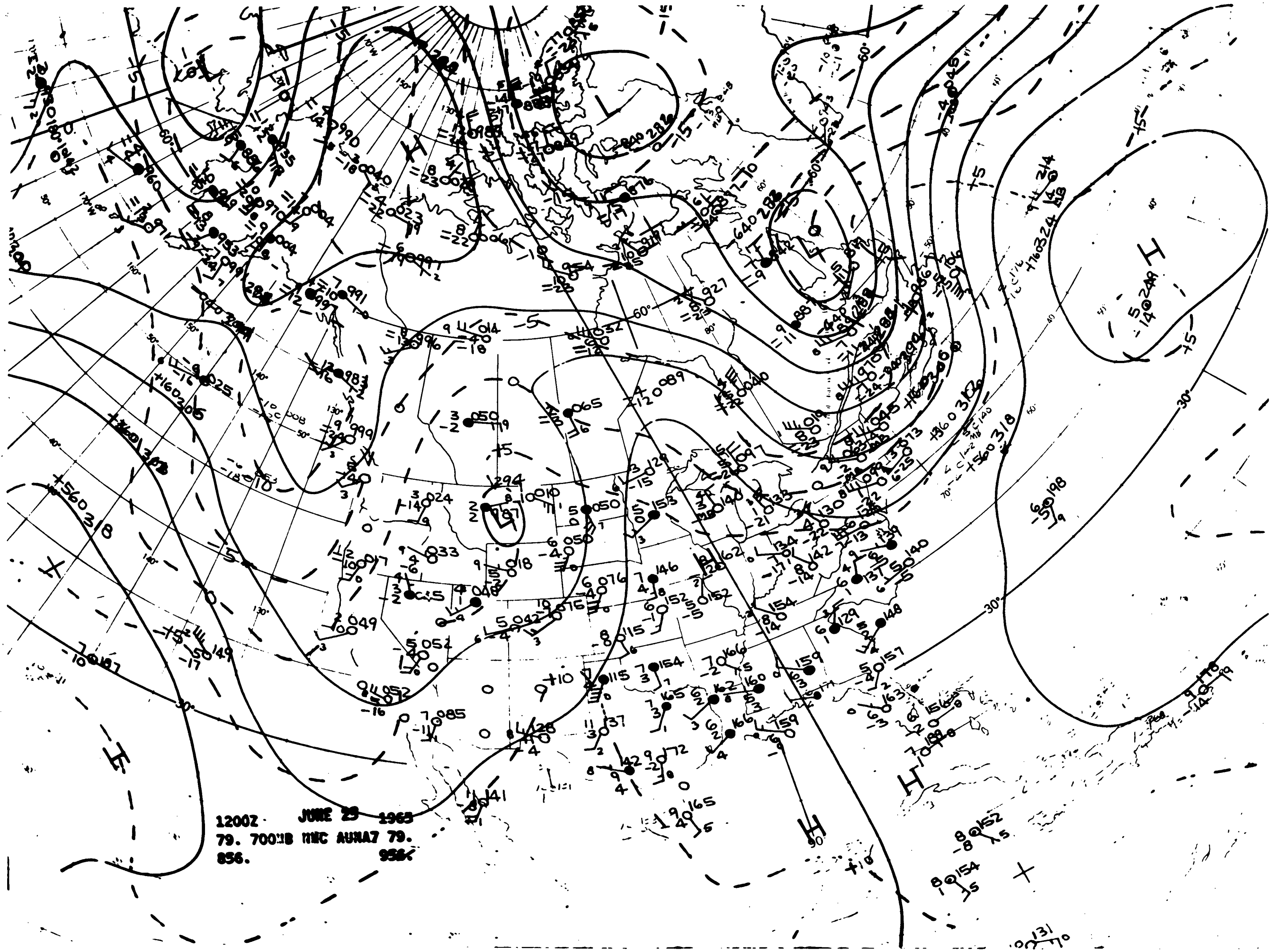

FIGURE 19. Height contours for the 700 mb surface at 0700 EST, 25 June, 1965.

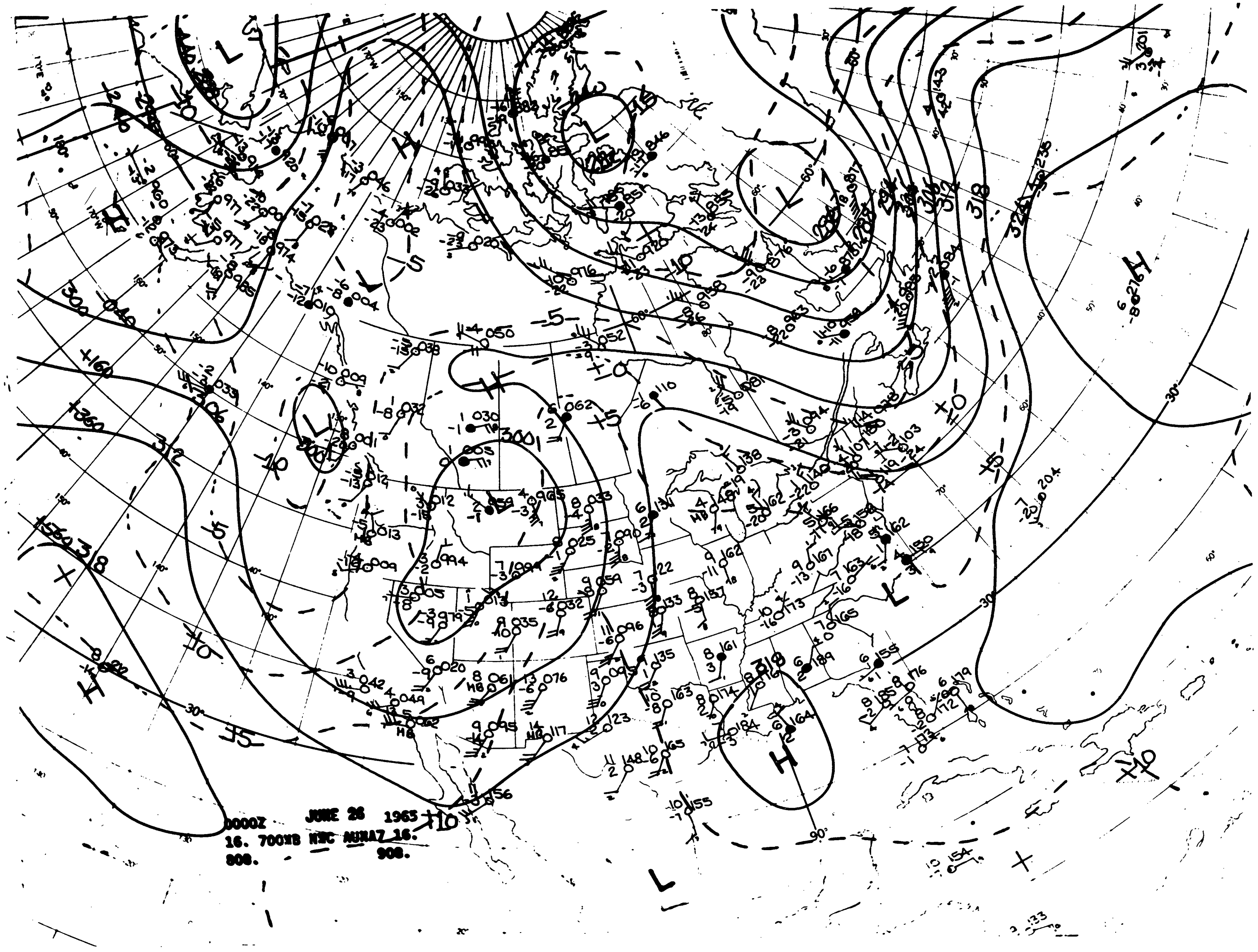

FIGURE 20. Height contours for the 700 mb surface at 1900 EST, 25 June, 1965.

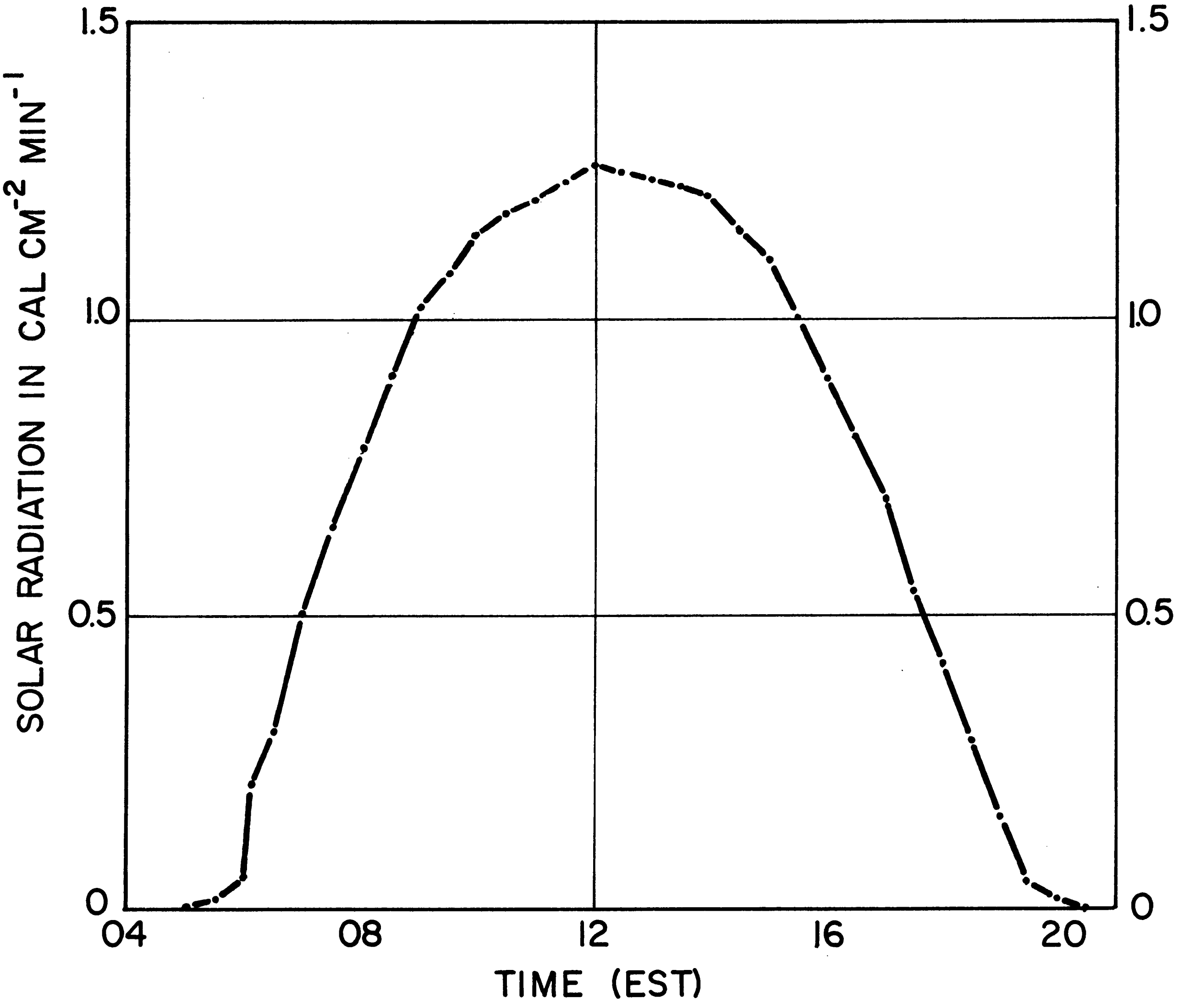

FIGURE 21. Solar radiation recorded at the Lake shore station on 25 June, 1965.

in good agreement with values given by CRREL (1964). The sun rose at 0506 EST and set at 2029 EST.

2.3 OBSERVED TEMPERATURES, HUMIDITIES, WINDS, AND SMOKE PLUMES.

This section presents observed temporal and spatial variations of temperature, moisture, and wind near the eastern shore of Lake Michigan on 25 June, 1965. The data, upon which this presentation is based, are tabulated and analyzed in detail in a report by Olsson et al (1968 b).

2.3.1 Temperature and Moisture.

The lapse rates measured in the lowest 80 m at WJBL at various times are presented in Figure 22, together with the temperature profiles in the first meter of the soil. While the diurnal temperature fluctuation at 1 m depth was less then 1.2°C, the difference between maximum and minimum temperatures at the surface (1 cm below the surface) was 26.0°C. At all times the air temperature recorded at the 2.4 m level was lower than those recorded at the surface. The problem with correct measurement of the surface temperature is well recognized. The top soil is "penetrated by", and rapidly responds to solar heating, while the response to nocturnal cooling is slow and retarded by heat flux through the soil. The nocturnal inversion was broken at sunrise. At 1300 EST a superadiabatic layer was present up to approximately 40 m. At 1600 EST the effect of the incoming lake breeze was seen in the lowering of the temperature in the lowest layers. The return to a nocturnal inversion condition after sunset was observed.

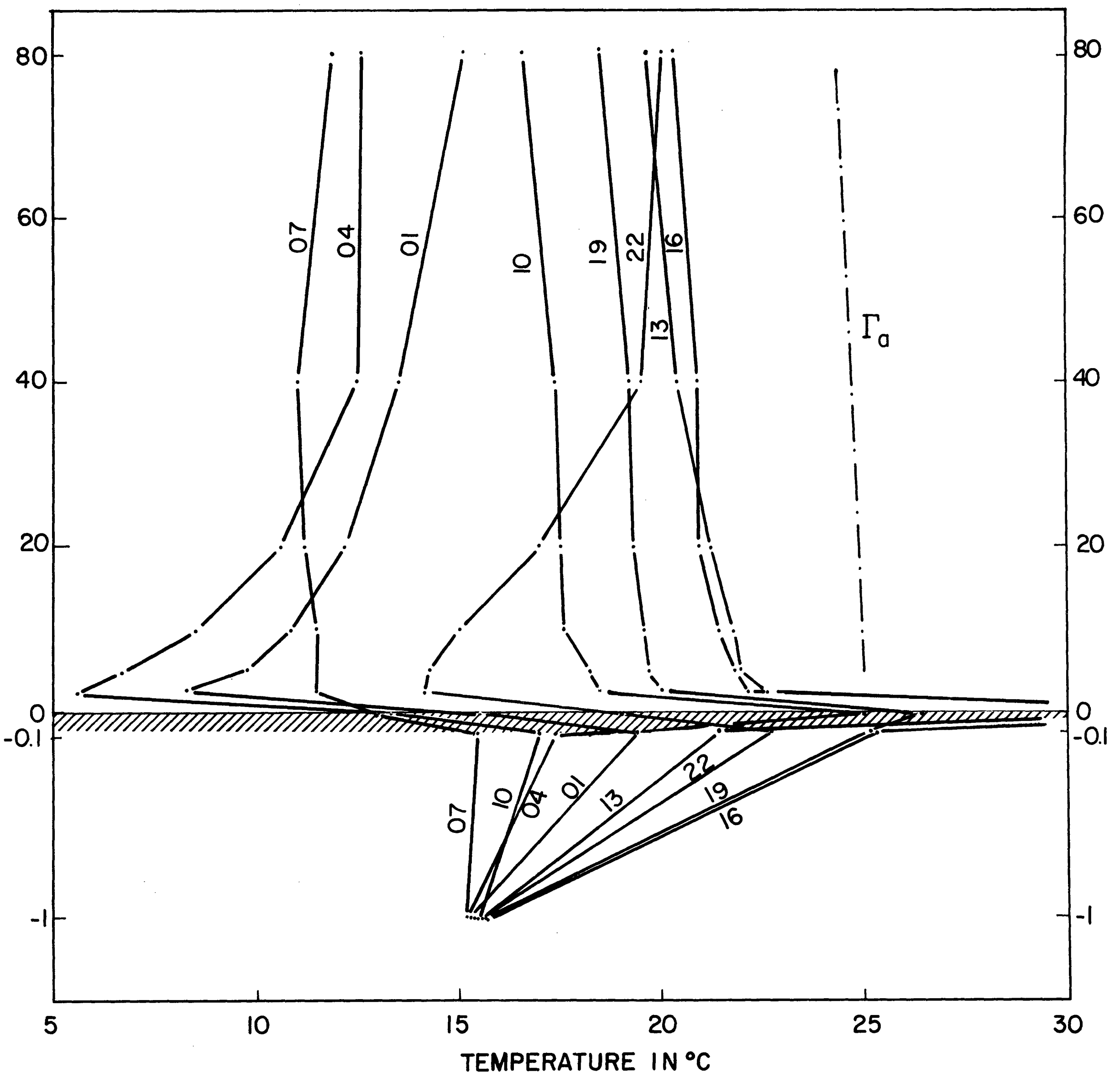

FIGURE 22. Lapse rates and soil temperatures as a function of time at WJBL tower on 25 June, 1965. Times indicated in Figure are EST.

Lapse rates in the layer between 150 m and 2135 m were measured from airplanes both over land and lake, Figure 23.

The airplanes ascended over the lake at $x = -32$ km, $y = 0$ km and descended over land at $x = 32$ km, $y = 0$ km. While the air over land assumed a near adiabatic lapse rate below 1500 m in the early afternoon, the air over the lake remained relatively stable in response to the cold surface water and off shore subsidence in the lake breeze circulation.

Vertical temperature cross sections along the observation line are presented in Figures 24-26. In order that the details near the surface could be included a logarithmic height scale was used. While there is a marked heating of the lower air layer over land, there is relatively little temperature change in that air layer over the lake. It should be noted that a horizontal temperature gradient across the shoreline of approximately 1.2°C km^{-1} (6°C in 5 km) had developed in the early afternoon.

Figure 27 presents temperatures as a function of time at various stations and at various distances from shore, as well as water temperatures recorded onboard the Mysis. The "heating curve" for the 24 km station agrees within ± 1°C with the one observed at GRR, 53 km inland, indicating that the lake breeze never penetrated as far as 24 km. The "heating curve" measured at the 2.4 m level at WJBL, 10 km inland, is in close agreement with the one observed at the 8 km station. Local envi-

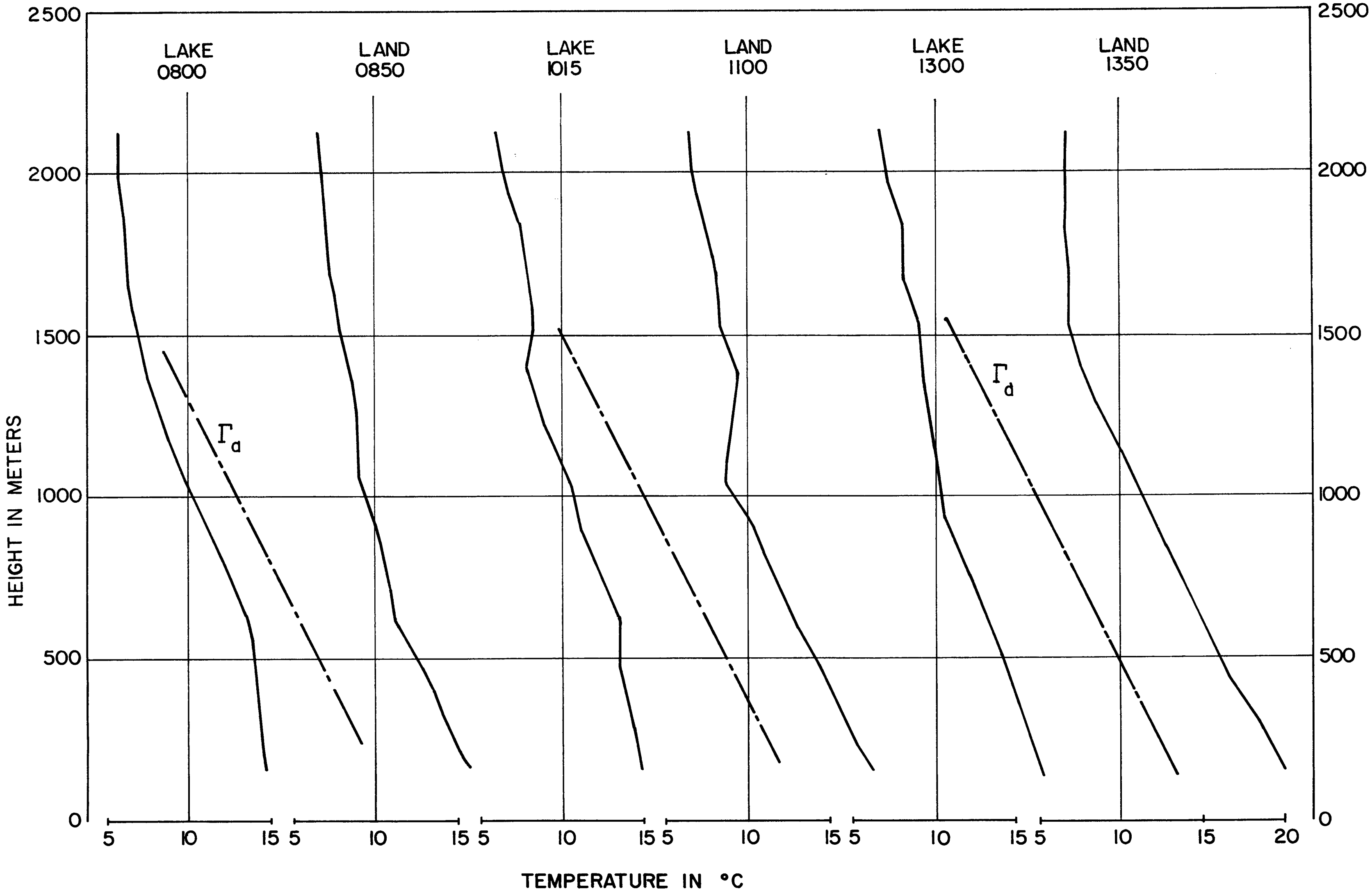

FIGURE 23. Lapse rates over lake and over land at various times on 25 June, 1965. The airplanes started ascent over the lake and descent over land at times indicated.

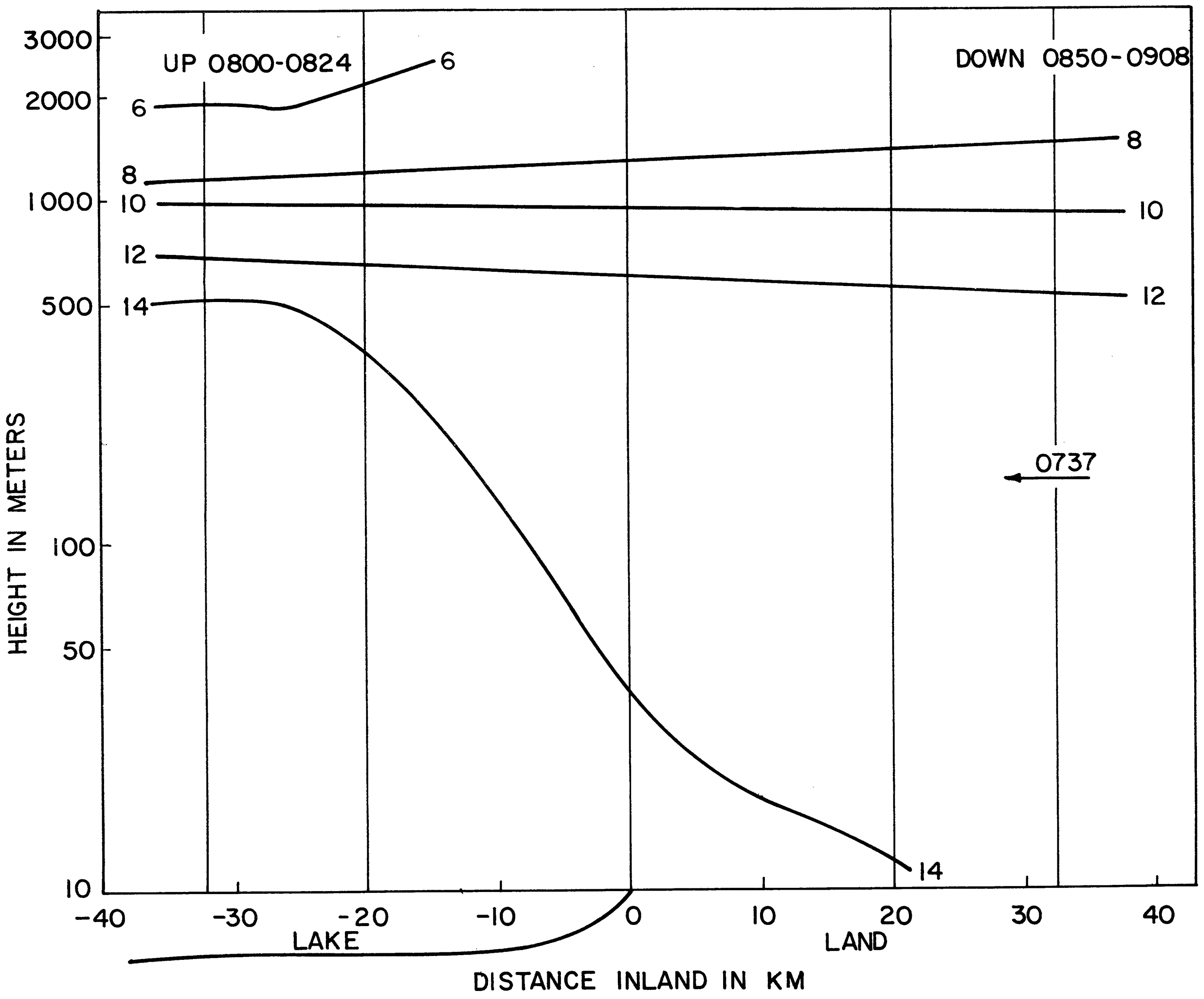

FIGURE 24. Analyzed vertical temperature field in early morning, 0737-0908 EST, 25 June, 1965. Note that a log-scale is used for the ordinate to allow surface and tower data to be illustrated as well as data obtained from further aloft by aircraft.

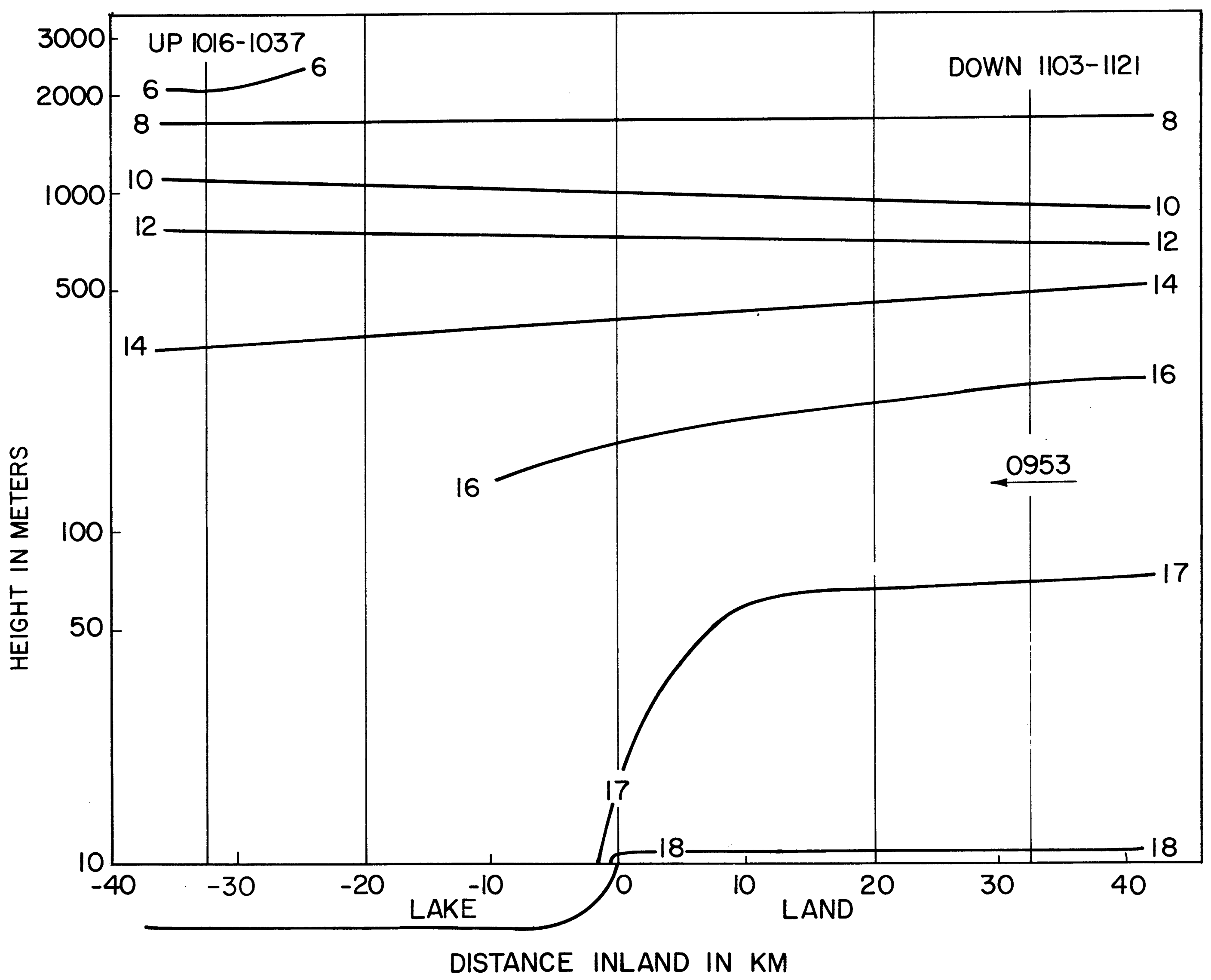

FIGURE 25. Analyzed vertical temperature field in late morning, 0953-1121 EST, 25 June, 1965. Note that a log-scale is used for the ordinate to allow surface and tower data to be illustrated as well as data obtained from further aloft by aircraft.

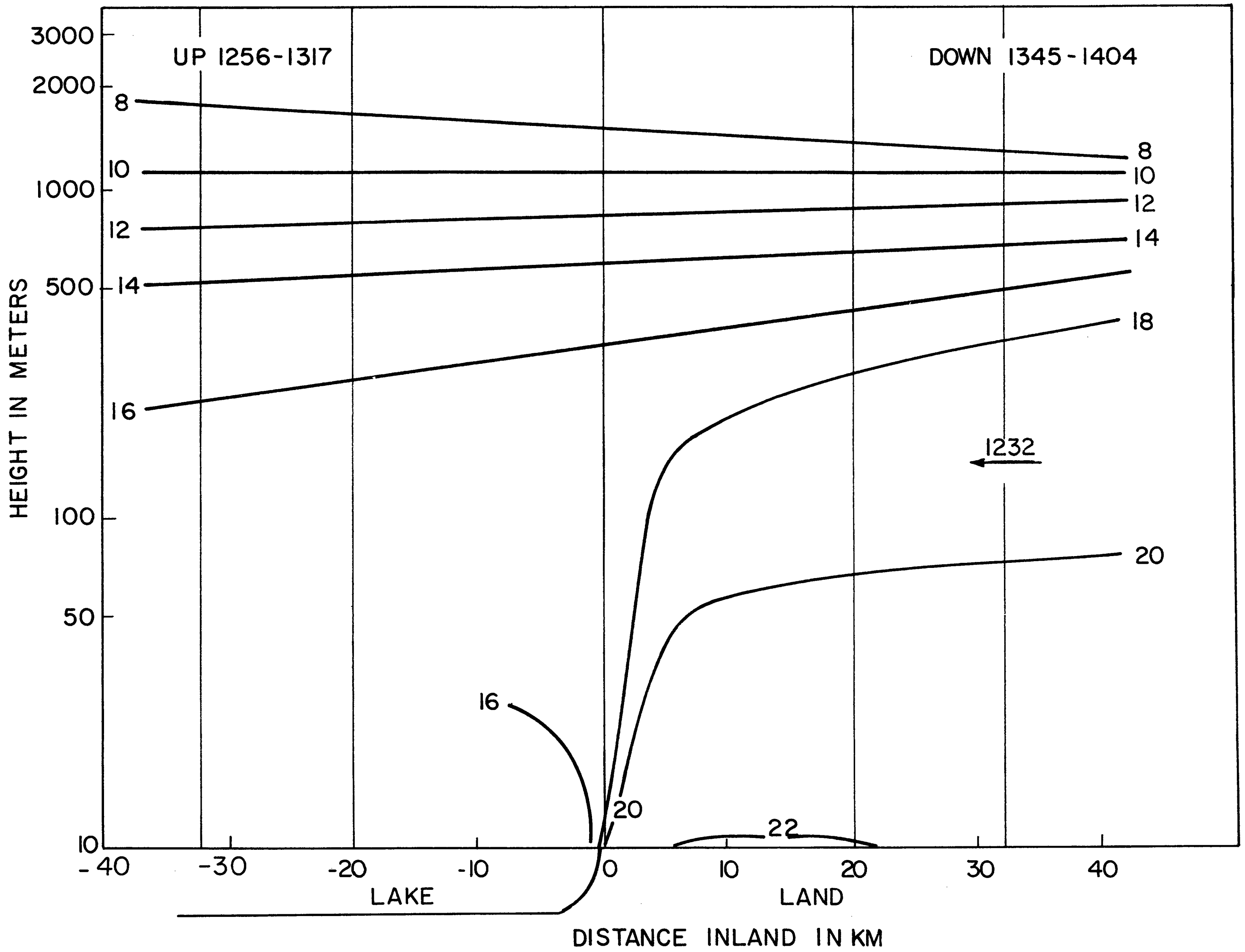

FIGURE 26. Analyzed vertical temperature field in early afternoon, 1232-1404 EST, 25 June, 1965. Note that a log-scale is used for the ordinate to allow surface and tower data to be illustrated as well as data obtained from further aloft by aircraft.

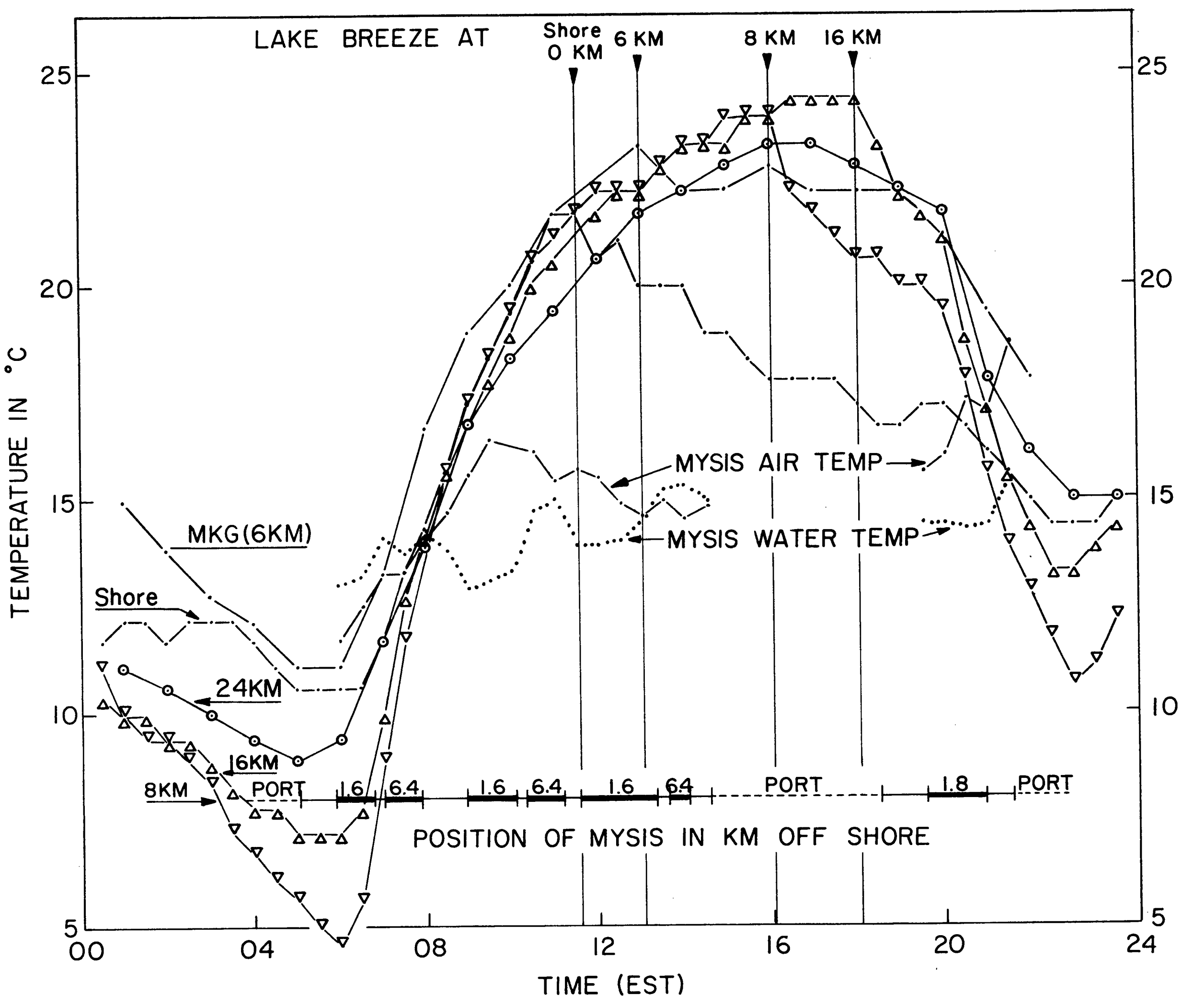

FIGURE 27. Temperature as a function of time at various stations and various distances from shore on 25 June, 1965. Included are air and water temperatures measured from the MYSIS at indicated distances off shore. Air temperatures onboard the MYSIS were sensed 11.8 m above the lake surface while the water temperatures were sensed 1.2 m below the lake surface.

ronmental differences seemed to effect the night-time values of temperature at the stations; however, the daily heating before lake breeze penetration was similar at all stations. When the cold lake air was brought inland by the lake breeze and reached a station, a definite drop in temperature was observed. The times at which the lake breeze reached the various stations are indicated with vertical lines in the figure. Changes in the air temperature over the lake indicated, that before approximately 0900 EST air was flowing from land, reaching at least 6 km off shore. In the evening the onset of a land breeze was noted off shore, as an increase in air temperature. Comparison of air and water temperatures recorded onboard the <u>Mysis</u> with air temperatures recorded elsewhere suggested the existence of a dome of cold and stable air near the surface over the lake. It should be noted that the water temperature was varying with time of day and distance from shore, the "day heating" observed being approximately 1.5°C both at 1.6 km and 6.4 km off shore, while the water was on the average 1 - 2°C warmer at 6.4 km than at 1.6 km off shore.

Although the difficulties in obtaining accurate measurements of moisture parameters are well recognized, some analyses of the moisture structure was attempted. The dew point temperatures are presented in Figure 28 as a function of time at the various stations. As the colder and more humid lake air moved inland with the lake breeze, a sudden increase in dew point temperature was observed at a station. The time of lake breeze passage at a station determined from "dew point increase" was in ex-

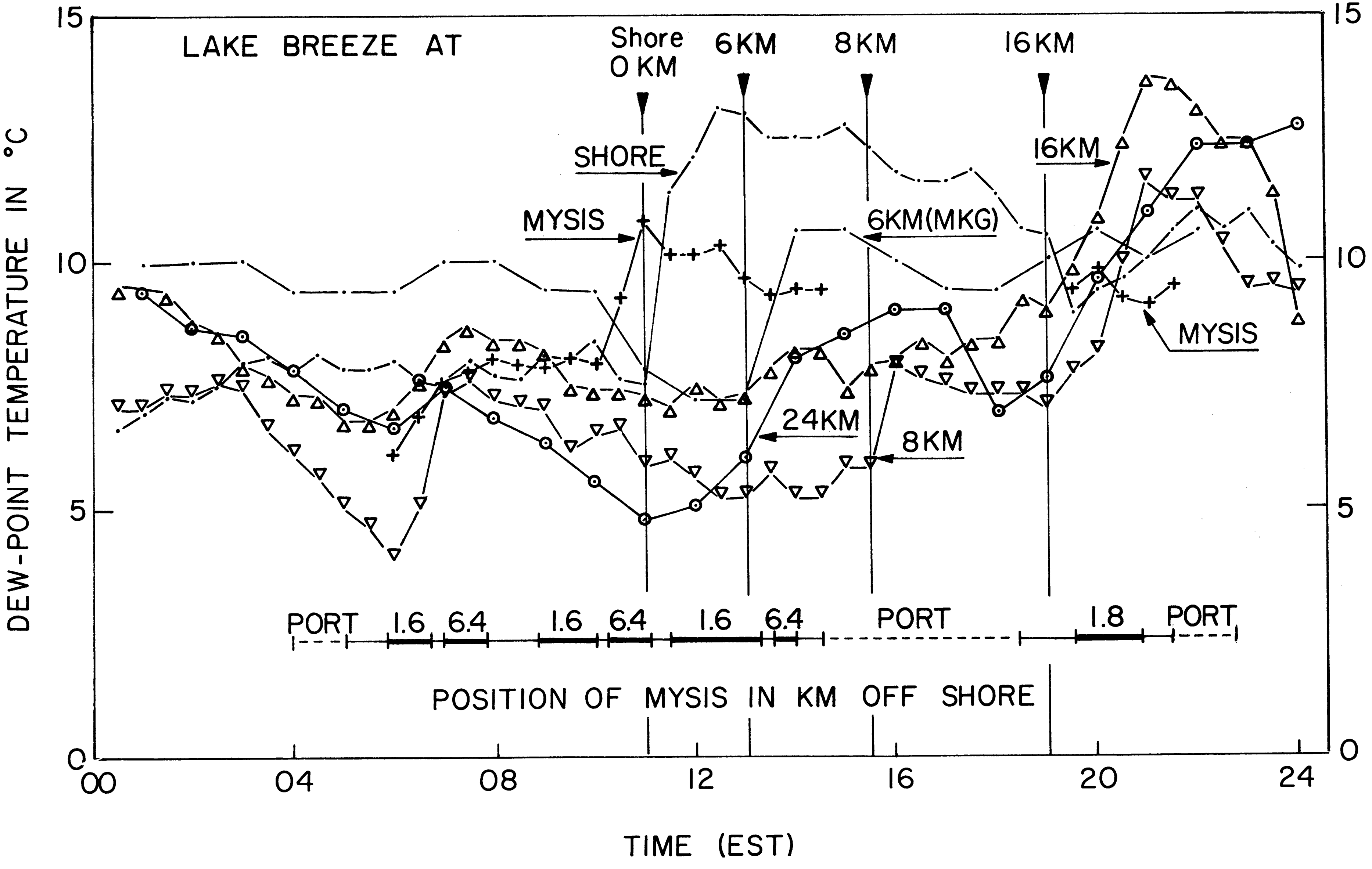

FIGURE 28. Dew Point temperature as a function of time at various stations and various distances from shore on 25 June, 1965. Included are dew point temperatures measured from the MYSIS at indicated distances off shore. The sensor onboard the MYSIS was located 5.0 m above the lake surface.

cellent agreement with the time determined on the basis of "temperature drop". Dry bulb temperatures and, to a lesser degree, dew point temperatures observed at the inland stations and compared with observations at the lake shore and onboard the Mysis indicated that intense modification of the lake air occurred in short trajectories over land.

Due to a lack of confidence regarding the time constant and accuracy of the airplane moisture measurements, no vertical cross sections of moisture structure are presented.

2.3.2 The Land and Lake Breeze Circulations.

The extensive measurements of the land and lake breeze circulations that were made along the observation line ($y = 0$) and elsewhere in the neighborhood of that line, are presented in this section.

Isotachs for the U (across-shore) and V (alongshore) components are analyzed for 0830, 1100, 1400, 1700, and 1900 EST and presented in Figures 29-38. In order that the details near the surface could be included, a logarithmic height scale was used. Vertical profiles of the U-component at various times and at the various stations are presented in Figure 39.

Figure 40 presents temporal variations of the wind in the land, lake, and land breeze circulations at various stations and at three different levels. The maximum lake breeze, an onshore wind, and the maximum land breeze, an offshore wind were observed, at about 200 m. The 1050 m level was chosen to be representative of the

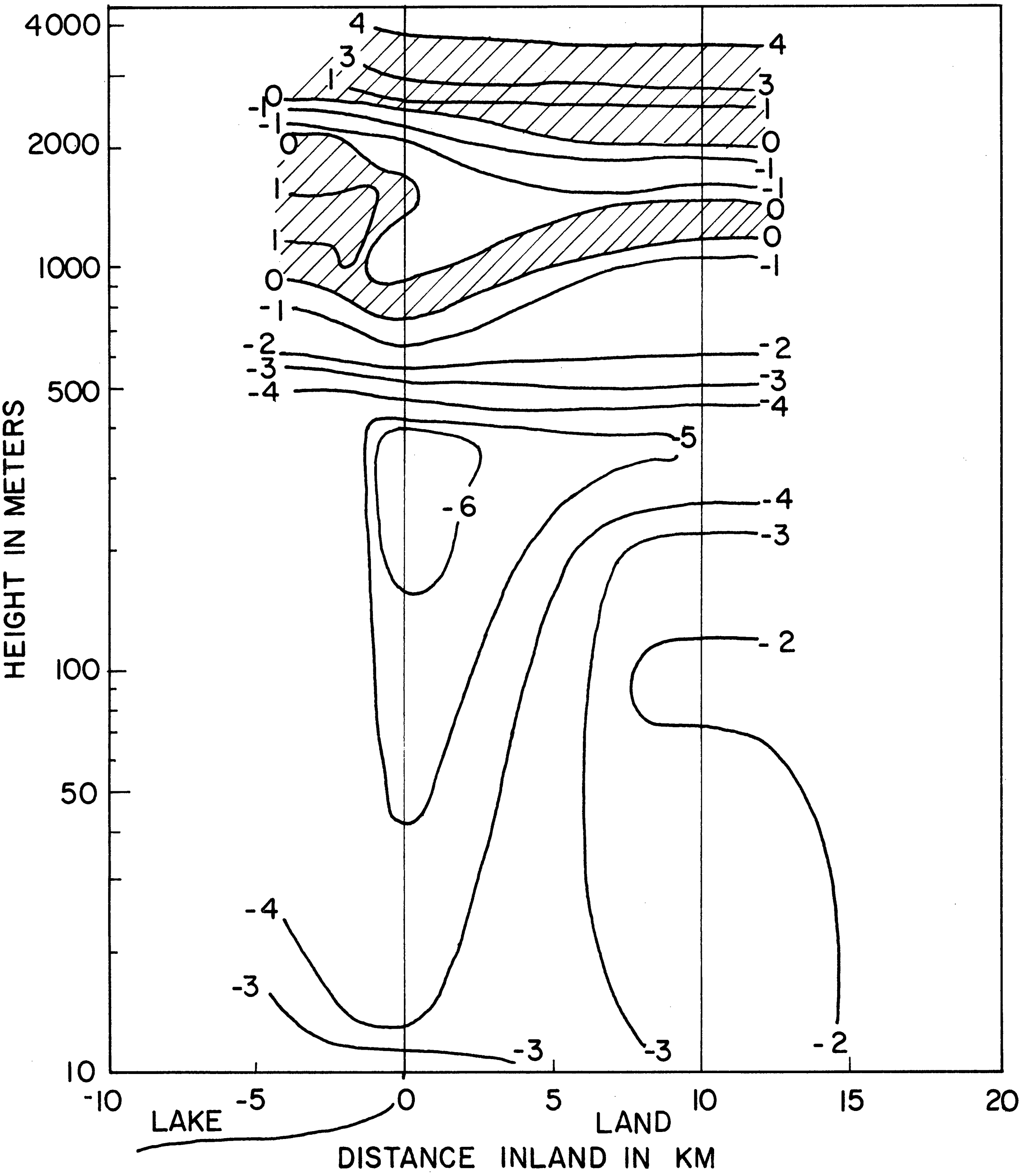

FIGURE 29. Isotachs for the across-shore component (U) at 0830 EST, 25 June, 1965. Positive values, hatched areas, indicate westerly winds (onshore) and negative areas indicate easterly winds (offshore). Wind speeds are in m sec^{-1}.

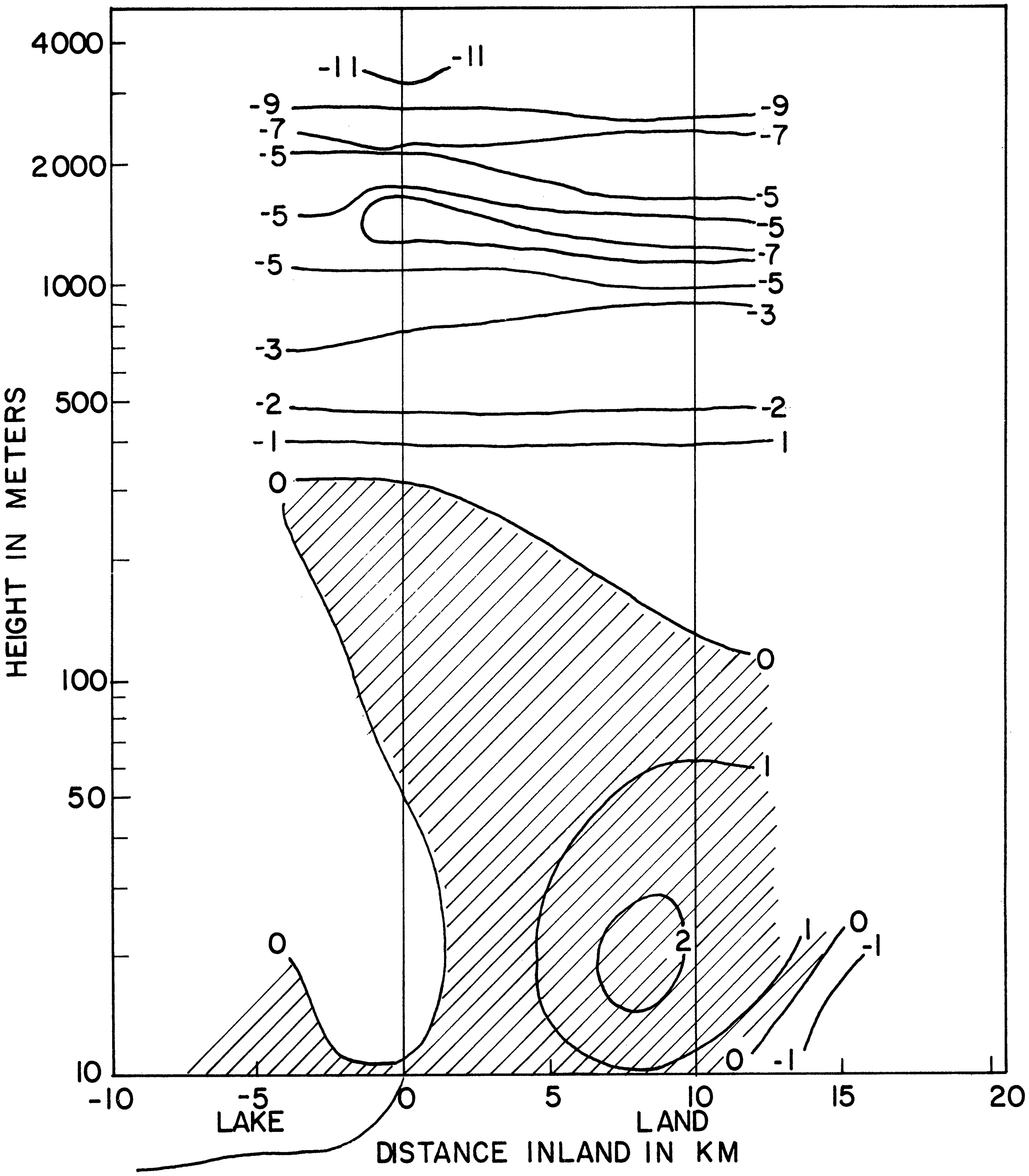

FIGURE 30. Isotachs for the along-shore component (V) at 0830 EST, 25 June, 1965. Positive values, hatched areas, indicate southerly winds and negative values indicate northerly winds. Wind speeds are in m sec^{-1}.

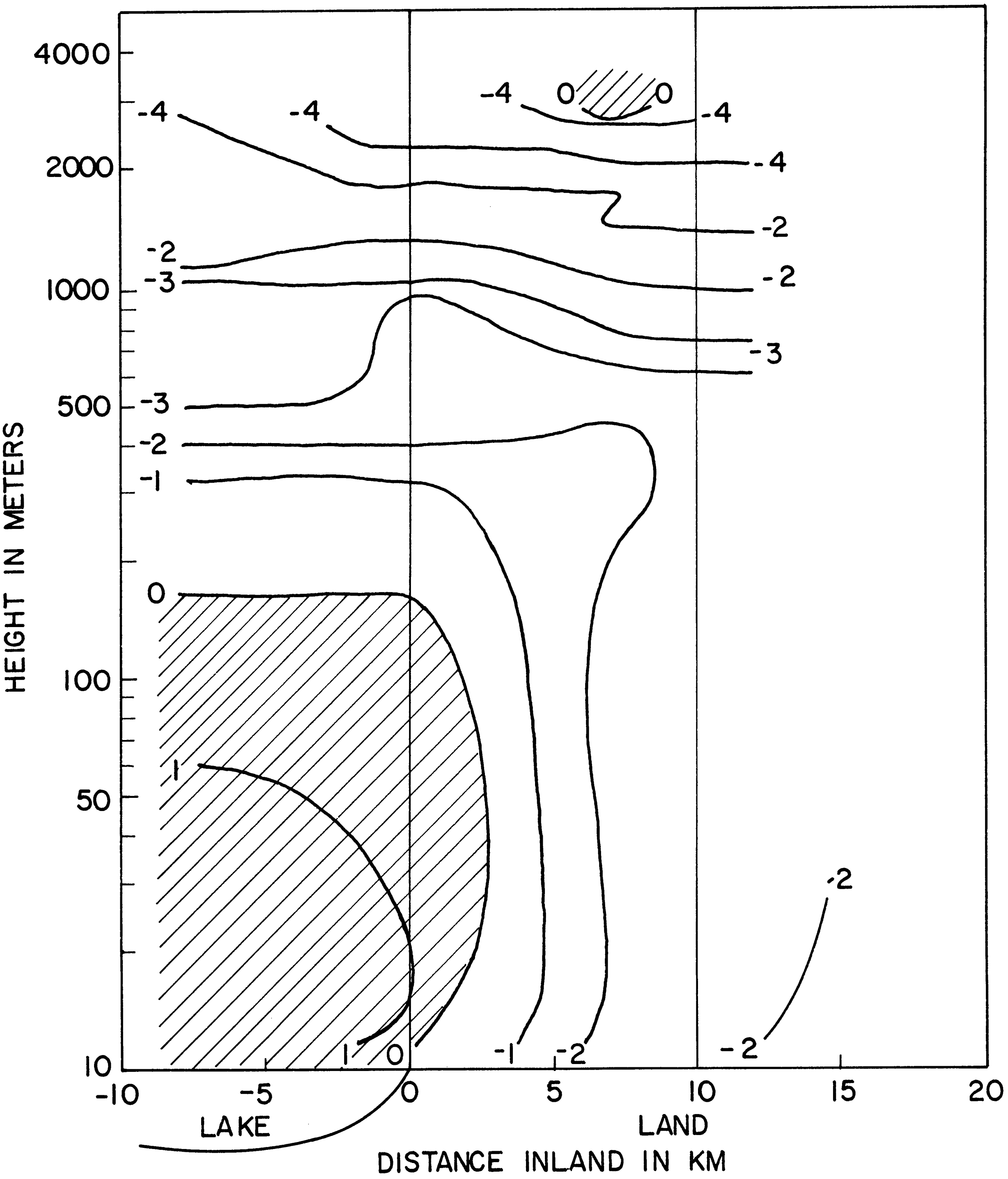

FIGURE 31. Isotachs for the across-shore component (U) at 1100 EST, 25 June, 1965. Positive values, hatched areas, indicate westerly winds (onshore) and negative values indicate easterly winds (offshore). Wind speeds are in m sec^{-1}.

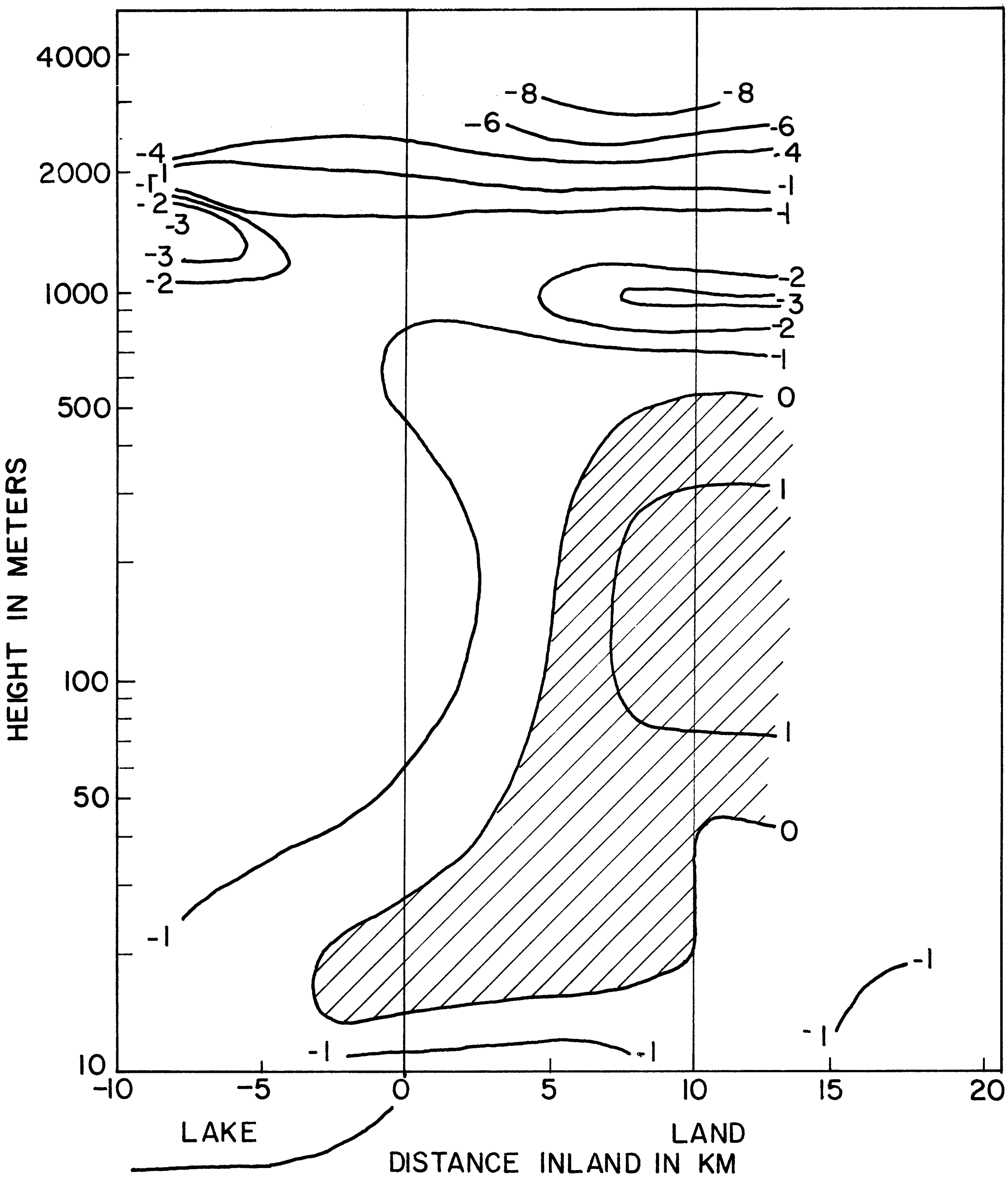

FIGURE 32. Isotachs for the along-shore component (V) at 1100 EST, 25 June, 1965. Positive values, hatched areas, indicate southerly winds and negative values indicates northerly winds. Wind speeds are in m sec^{-1}.

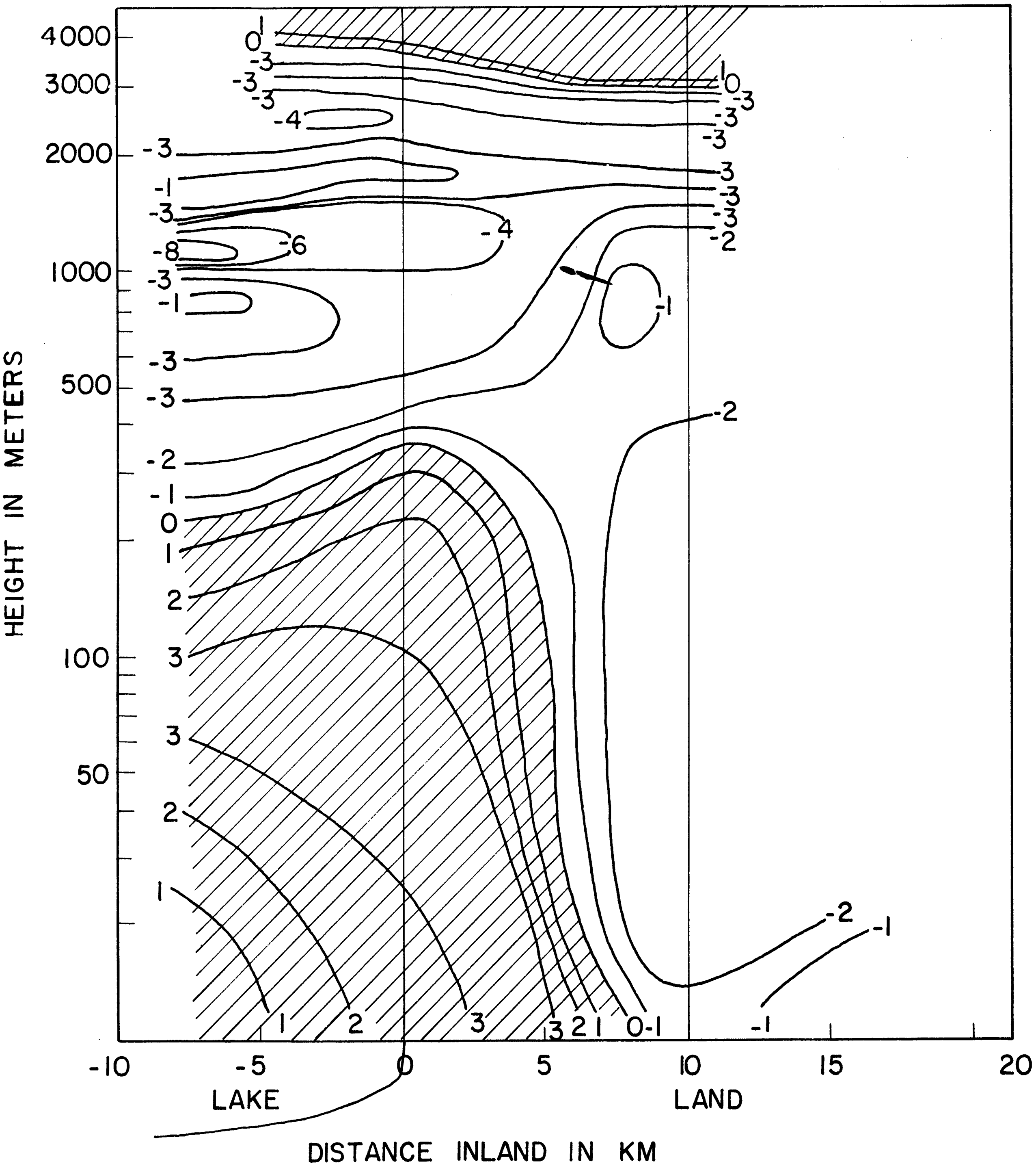

FIGURE 33. Isotachs for across-shore component (U) at 1400 EST, 25 June, 1965. Positive values, hatched areas, indicate westerly winds (onshore) and negative values indicate easterly winds (offshore). Wind speeds are in m sec^{-1}.

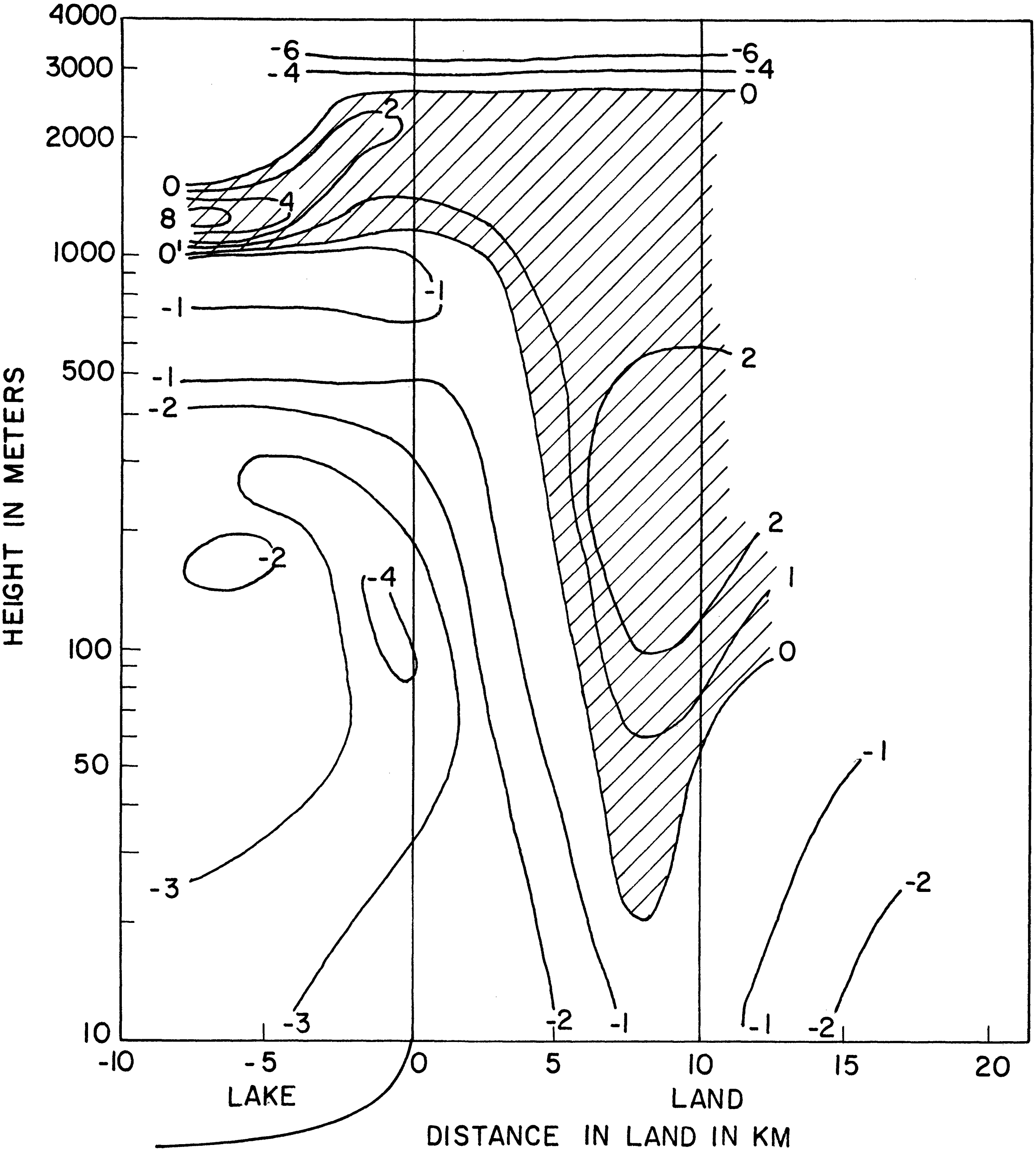

FIGURE 34. Isotachs for the along-shore component (V) at 1400 EST, 25 June, 1965. Positive values, hatched areas, indicate southerly winds and negative values indicate northerly winds. Wind speeds are in m sec^{-1}.

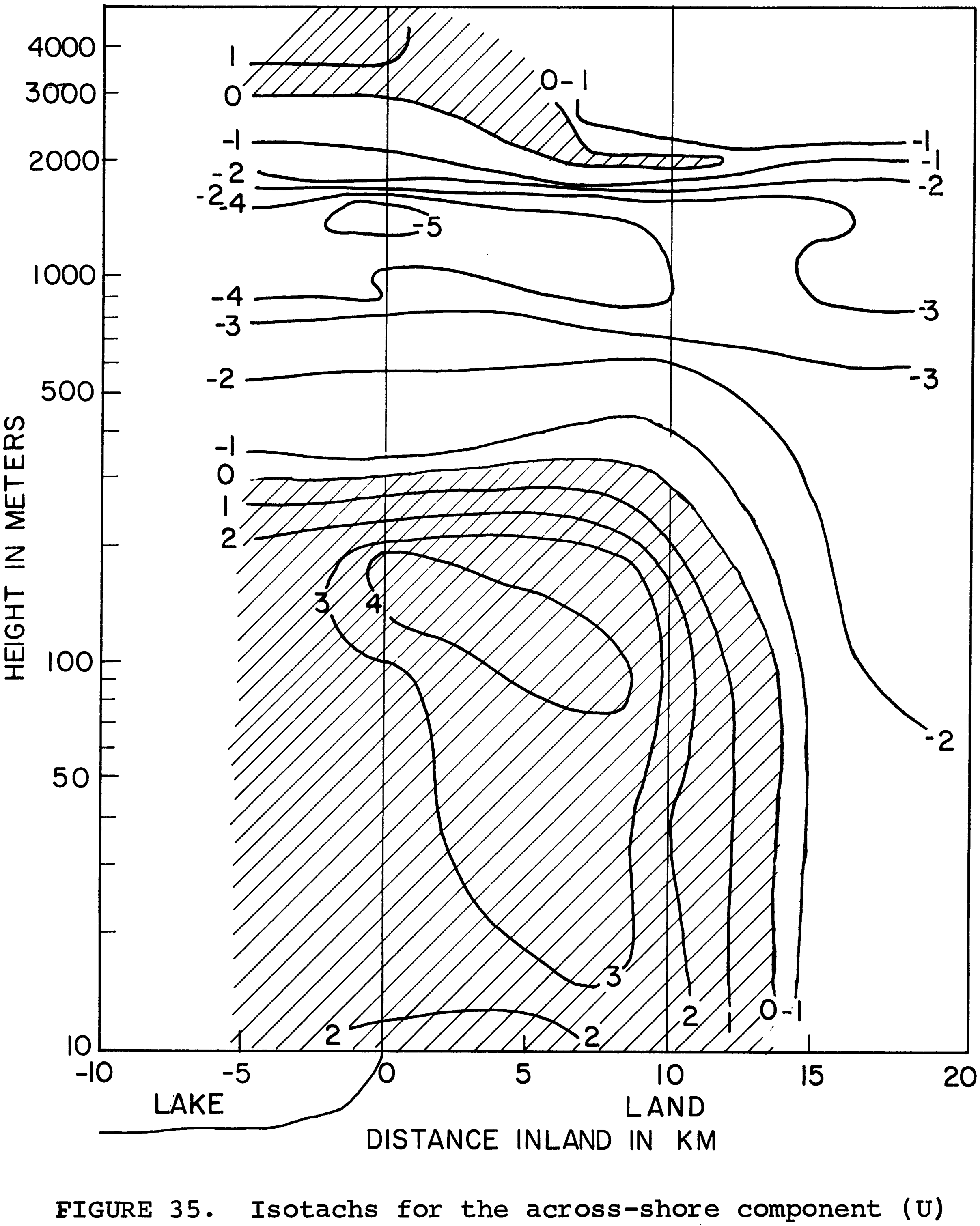

FIGURE 35. Isotachs for the across-shore component (U) at 1700 EST, 25 June, 1965. Positive values, hatched areas, indicate westerly winds (onshore) and negative values indicate easterly winds (offshore). Wind speeds are in m sec^{-1}.

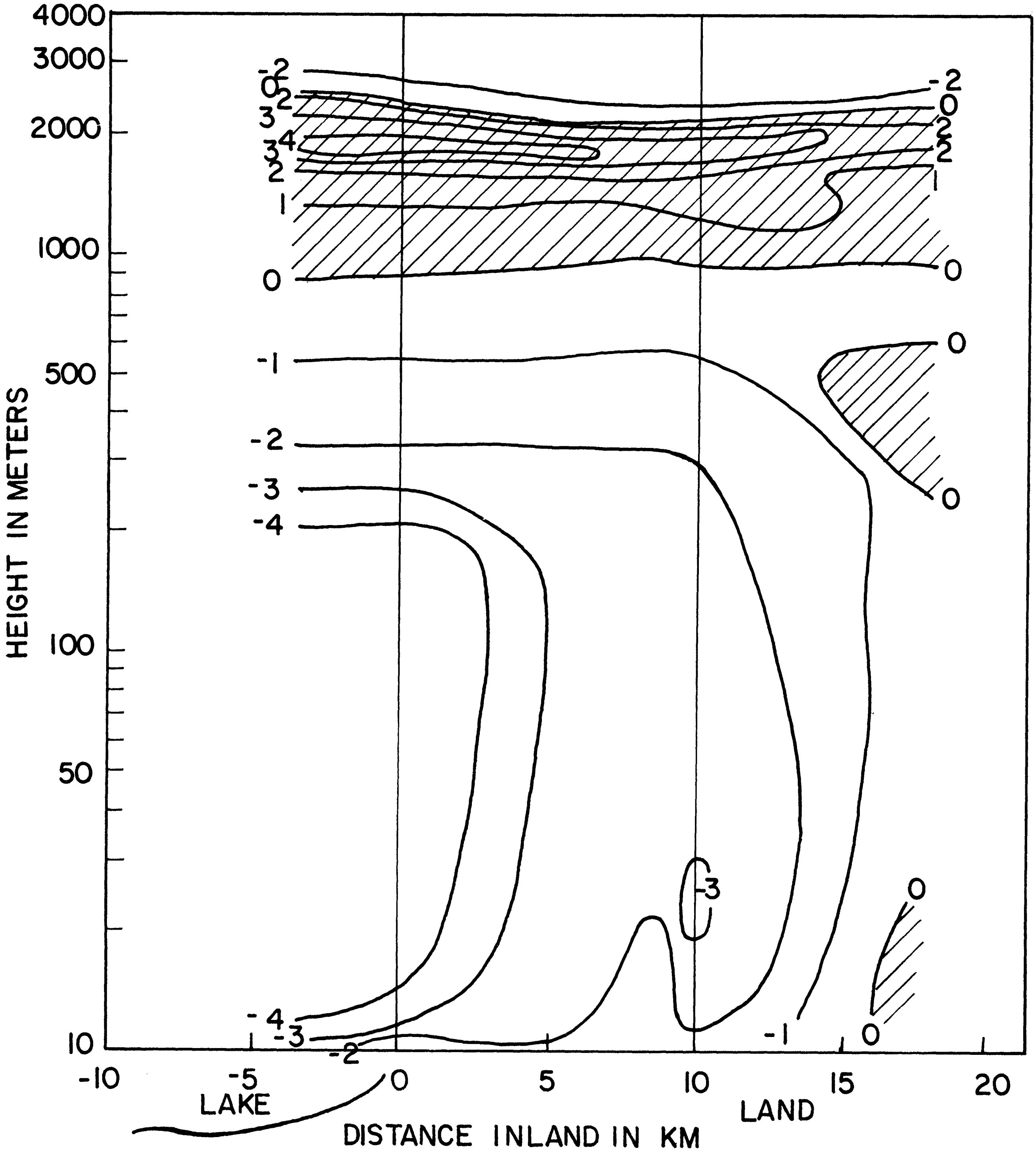

FIGURE 36. Isotachs for the along-shore component (V) at 1700 EST, 25 June, 1965. Positive values, hatched areas, indicate southerly winds and negative values indicate northerly winds. Wind speeds are in m sec^{-1}.

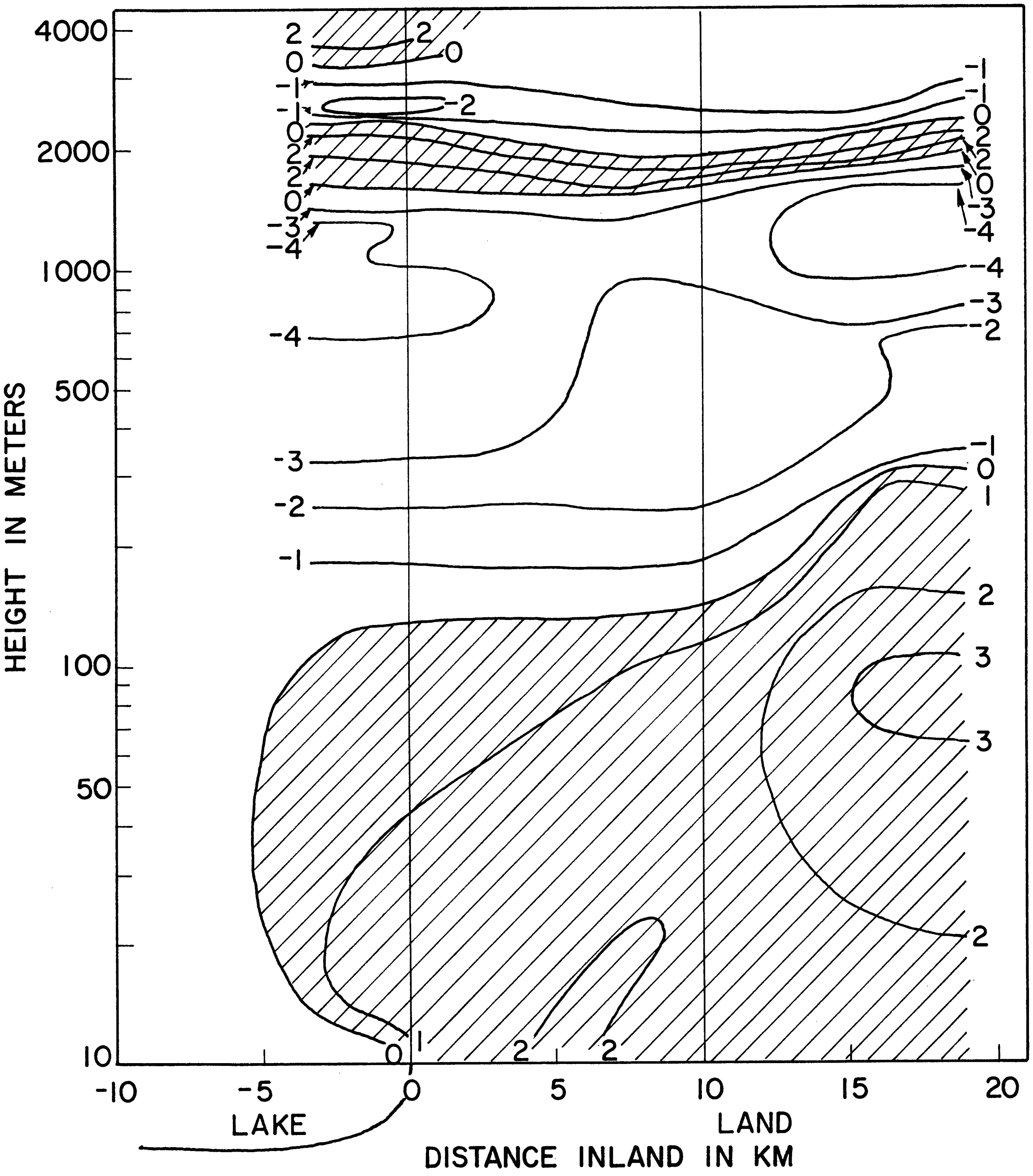

FIGURE 37. Isotachs for the across-shore component (U) at 1900 EST, 25 June, 1965. Positive values, hatched areas, indicate westerly winds (onshore) and negative values indicate easterly winds (offshore). Wind speeds are in m sec^{-1}.

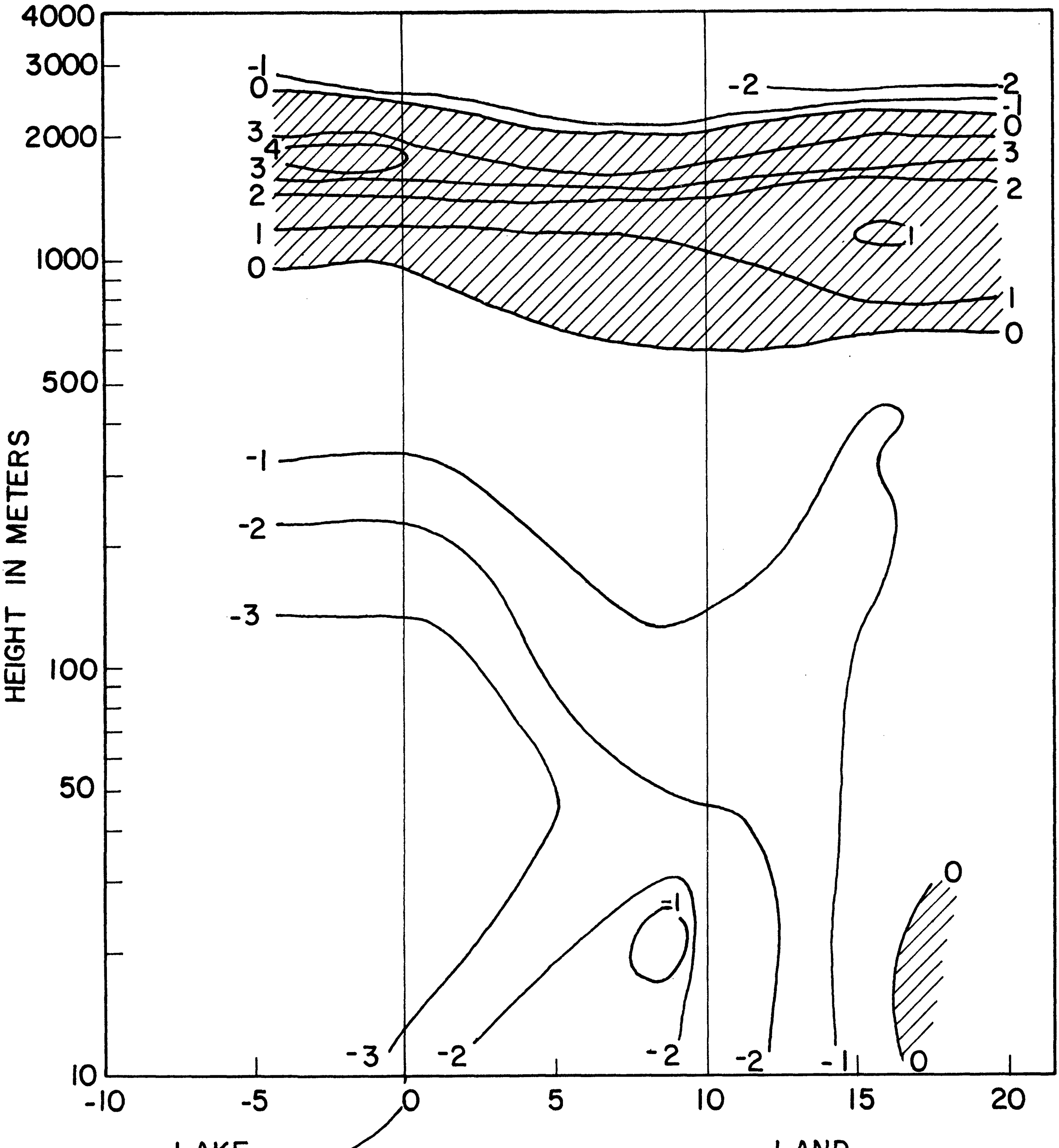

FIGURE 38. Isotachs for the along-shore component (V) at 1900 EST, 25 June, 1965. Positive values, hatched areas, indicate southerly winds and negative values indicate northerly winds. Wind speeds are in m sec^{-1}.

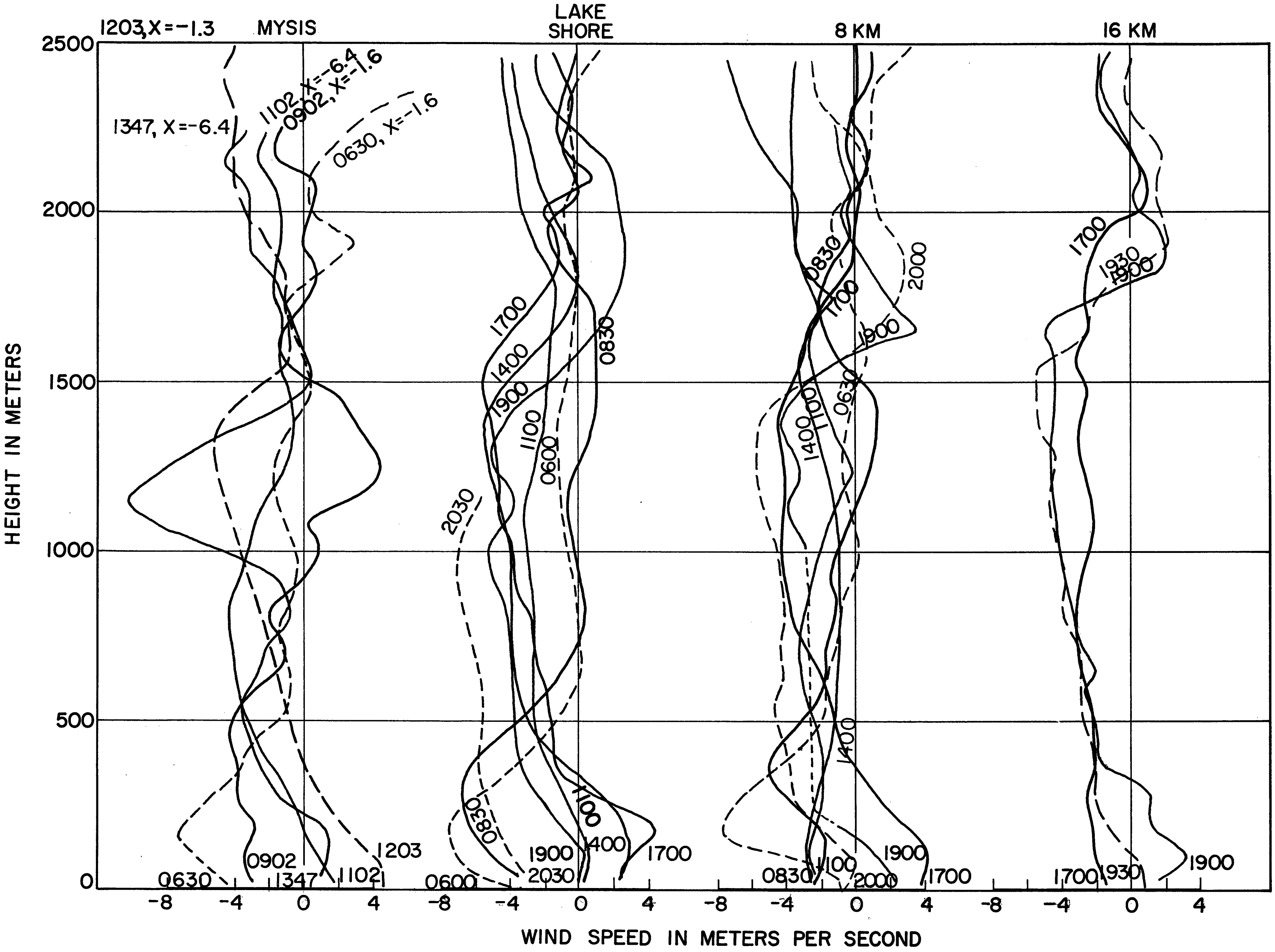

FIGURE 39. Across shore wind component (U) as a function of height at various times and stations, 25 June, 1965. Times given in figure are EST and x-values given for MYSIS indicate the distance off shore in km.

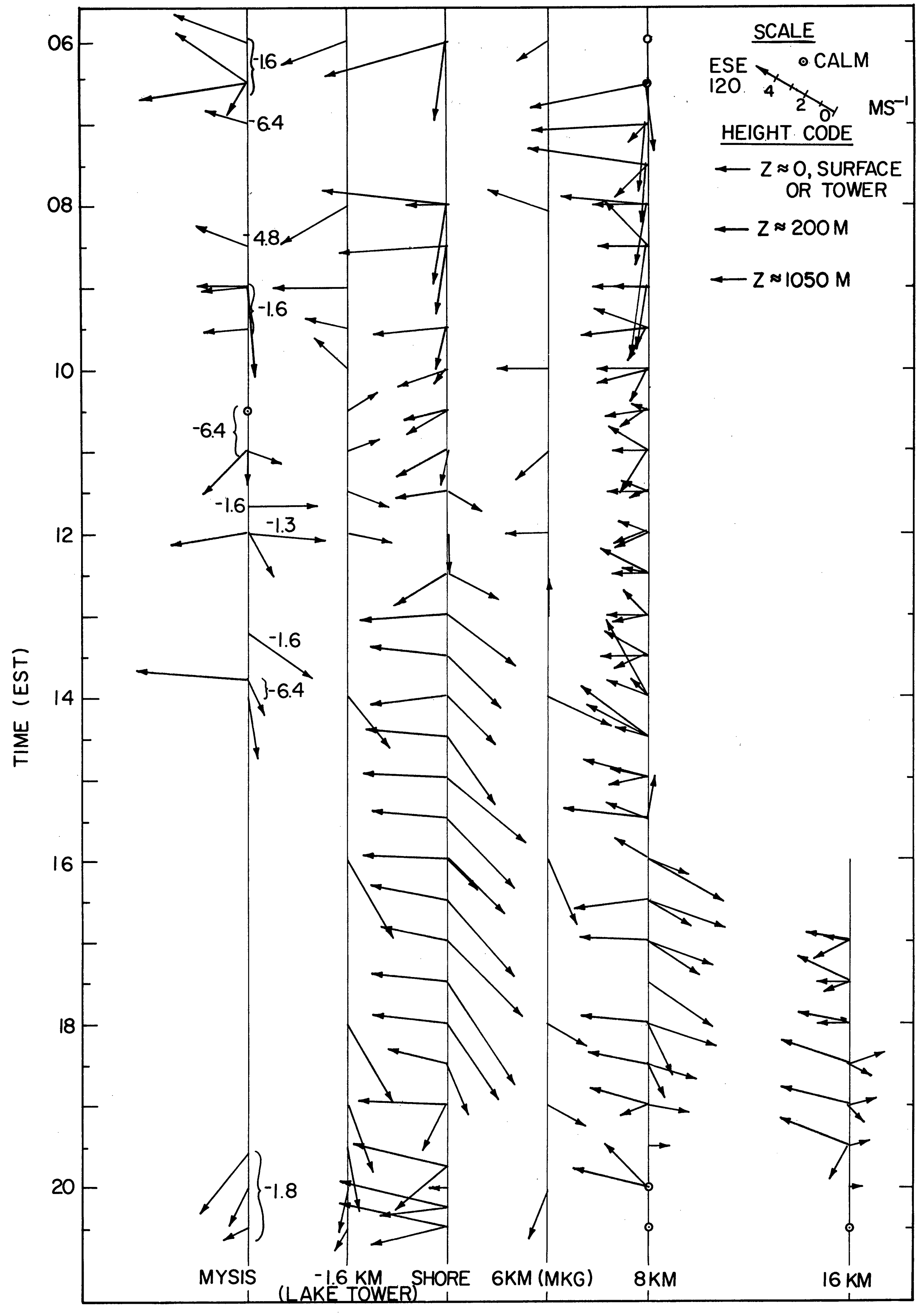

FIGURE 40. Wind vectors at the surface, 200 m and 1050 m aloft at various times and at various stations, 25 June, 1965. Number beside MYSIS' vector indicate distance off shore in km.

return flow, the offshore wind aloft in the lake breeze circulation. There was some indication of a return flow associated with the land breeze circulation, in this case an onshore wind aloft with a maximum in a layer above 1200 m; thus not truly represented in Figure 40.

As can be seen from Table 1, winds aloft data are available for every 30 minutes near the shore between 0600 EST and 2030 EST. A total of 62 pibal measurements were made and half of these observations were made by double theodolite technique and analyzed for vertical velocities as well as for horizontal winds.

The low level land breeze observed at all stations in the morning was reversed to a lake breeze at various times during the day and then again reversed to a land breeze in the early evening. The times for reversal to a lake breeze were in close agreement with the times at which the lake breeze could be detected at the respective stations, on the basis of changes in dry bulb and dew point temperatures, Table 2.

In the morning a strong offshore wind, exceeding 6 m sec^{-1}, was observed at about 200 m, both over land and lake, while the surface winds over land in general were light. As the stable land breeze air crossed the shore it subsided due to the change in surface friction. Downward velocities of more than 20 cm sec^{-1} were measured immediately off shore. The land breeze flow layer was more than 1000 m deep and extended more than 6 km off shore. A weak onshore flow layer aloft, in which a general ascent was observed, indicated the exsistence of a complete land

TABLE 2. Local times for lake breeze frontal passage at various stations as detected in wind direction change, temperature drop, and dew point rise near the surface on 25 June, 1965. The time selected is the time when a change began to take place. The values in the brackets are the changes that were observed in the next half hour in degrees and ^{o}C. For the lake shore the wind change at 200 m is given and for WJBL the wind change at 20 m is given. For MKG and WJBL half the subsequent one hour changes are given.

STATION: X-distance:	MYSIS -6.4 km	LAKE TOWER -1.6 km	LAKE SHORE 0 km	MKG 6 km	8 KM 8 km	WJBL 10 km	16 KM 16 km
WIND DIR.	1030(200)	1000(130)	1100(90)	1300(65)	1530(180)	1600(110)	1830(160)
TEMPERATURE	1030(0.7)	-	1130(1.1)	1300(0.6)	1600(1.7)	1630(0.9)	1800(1.1)
DEW POINT	1030(1.6)	-	1100(3.9)	1300(1.7)	1530(2.2)	-	1800(0,9)

breeze circulation cell. An onshore flow aloft in the evening, associated with the onset of the land breeze, was also observed.

Smoke plumes released near and off shore in the early morning hours indicated that the land breeze air was very stable, Figures 41 and 42. While, at more than 1 km off shore the water surface temperature was more than 1°C warmer than the deck height air temperature, there was a weak inversion in a layer between 5 m and 12 m. Strong wind speed and direction shears were depicted by the plumes. While the wind speed at 12 m above the lake surface exceeded 5 m sec^{-1}, there was only a few irregular streaks of ripples on the lake surface.

Strong updrafts, exceeding 200 cm sec^{-1} near the shore and a change towards more offshore wind direction in a layer between 500 m and 2500 m occurred during the mid-morning hours. The onset of the lake breeze was first observed 1.6 km off shore at 1000 EST. Half an hour later a sudden wind shift was observed at 6.4 km off shore. A smoke plume released from the Mysis depicted a sudden change in stability, the onset and increase of an onshore wind near the surface, and a return to more stable conditions, Figures 43 and 44.

Between 1100 EST and 1130 EST the lake breeze front moved across the shore. The front penetrated inland at a rate of 0.6 - 1.0 m sec^{-1}. Based on winds aloft measurements, the slope of the leading edge of the front was estimated to be less than 1:20. Qualitative observations, i.e. inspection of traces of recorded temperatures and winds, indicated

FIGURE 41. Aerial photograph of smoke plume released in stable land breeze air 1 km off shore ($x = 1$ km, $y = 0$) in early morning on 25 June, 1965. Note also plume at the shore.

FIGURE 42. Same as Figure 41, but end-view of plume.

FIGURE 43. Aerial photograph of smoke plume released at about 1000 EST, 6.4 km off shore ($x = -6.4$ km, $y = 0$) on 25 June, 1965. Note the vertical penetration of the plume and the vertical shearing of the plume.

FIGURE 44. Same as Figure 43, but 5 minutes later, when the onsetting lake breeze bends the plume toward shore. Note suppression of the plume and the "shearing off" of clouds from the plume by the offshore flow aloft.

that the front progressed inland in a series of pulses with periods less than 2 hours.

The lake breeze flow layer was in general stable over the lake, with inversion or isothermal conditions observed in the lowest 12 m. Mechanical turbulence induced by the rough lake surface could possibly explain observed fumigation off shore, Figure 45. At about 1 km off shore the lake breeze air started to ascend. Over land the lake breeze flow layer became moderately unstable and general upward velocities of about 50 cm sec^{-1} were measured. Near the lake breeze front the flow was very unstable, with vertical velocity variations between + 100 cm sec^{-1} and -100 cm sec^{-1} and pronounced low level convergence toward the front. Although observational errors gave some extreme values, measured updrafts in the front of 400 cm sec^{-1} appeared to be real.

The depth of the lake breeze flow layer grew from about 200 m before noon to more than 400 m, near the lake breeze front, in the afternoon. In late afternoon the wind speed at about 100 m immediately inland from the shore, exceeded 7 m sec^{-1} with a strong along-shore component, indicating the effect of the Coriolis acceleration. At 1830 EST, the wind was blowing parallel to shore near the lake shore, while the lake breeze front just passed the station 16 km inland.

Above the lake breeze layer a more than 1500 m thick, relatively stable, return flow layer was observed. Wind speeds in that layer reached approximately 5 m sec^{-1} in the afternoon and a slight veering in the direction was observed.

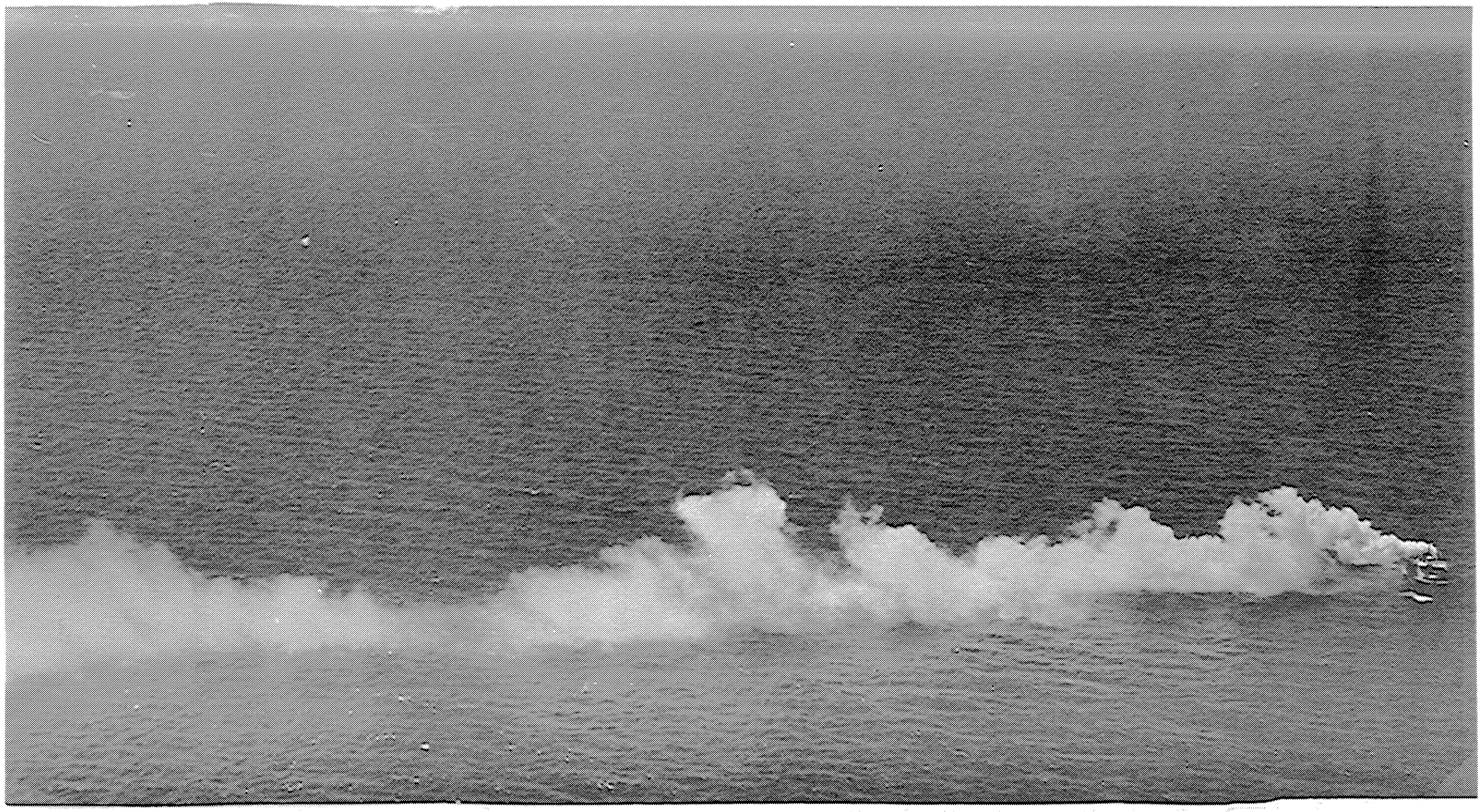

FIGURE 45. Aerial photograph of smoke plume released in stable lake breeze air 1 km off shore (x = 1 km, y = 0) in early afternoon on 25 June, 1965. Note the suppresion and fumigation of the smoke as it is carried toward shore by the lake breeze.

Marked vertical convergence occurred above the lake breeze front. A general subsidence, with downward velocities of less than 30 cm sec^{-1} was observed near the shore and over land, in the evening.

2.4 SUMMARY.

The spatial and temporal variations of temperature, moisture, and wind near the eastern shore of Lake Michigan have been observed and described for 25 June, 1965. During this day a weak northerly gradient flow prevailed and a well developed lake breeze circulation occurred. In summary the observations showed:

1. The effect of differential heating between land and lake surface led to a strong horizontal temperature gradient near the surface exceeding $1.2^{o}C\ km^{-1}$ ($6^{o}C/5$ km) across the shore in the early afternoon;
2. The temperature of the air over land became equal to the temperature of the air over the lake and also to the surface water temperature between 0700 EST and 0800 EST and again between 2000 EST and 2100 EST;
3. Air over the lake remained relatively stable througout the day in the lowest 600 m. Intense modification of this air in short trajectories over land was observed. Intense vertical mixing led to adiabatic and even superadiabatic conditions over land in the afternoon. A strong nocturnal surface inversion was observed over land.

Isothermal conditions in the lowest 80 m occurred at about 0730 EST and 2030 EST;

4. An offshore flow aloft, the return flow, was observed at the shore in a layer between 500 m and 2500 m between 0800 EST and 1100 EST. This observation was the first indication of a lake breeze circulation;
5. The onset of an onshore flow, the lake breeze, was observed near the surface, both over the lake and at the shore between 1000 EST and 1100 EST;
6. Strong vertical upward motion exceeding 200 cm sec^{-1} coupled with low level horizontal convergence, occurred near the lake breeze front, the lakeward sloping surface between the advancing cooler air off the lake, and the heated air over land;
7. Air temperature drops of more than $1^{o}C$, dew point increases exceeding $3^{o}C$, and reversal of surface winds occurred within one hour, as the lake breeze front passed a station near the shore. These changes were less pronounced the further inland the front penetrated;
8. The slope of the leading edge of the lake breeze front was on the average less than 1:20. The front progressed inland, in a series of pulses, at an average rate of 0.8 m sec^{-1}. The front penetrated approximately 20 km inland;
9. The maximum depth of the lake breeze flow layer was approximately 400 m near the lake breeze front in the late afternoon;

10. A maximum lake breeze of more than 7 m sec^{-1}, with an onshore wind component of about 4 m sec^{-1}, occurred in a layer 100-200 m aloft, immediately inland from the shore in the late afternoon;
11. Upward motions in general less than 50 cm sec^{-1}, were observed in the lake breeze flow layer over land. Downward motion was observed in that layer 1.2 km off shore;
12. A relatively stable, more than 1500 m thick, return flow layer was observed aloft. Offshore winds, exceeding 5 m sec^{-1} were measured above 1200 m in the afternoon;
13. The vertical velocity varied from +70 cm sec^{-1} to -70 cm sec^{-1} in the return flow layer. Several vertical convergence layers were observed, with the most pronounced at about 700 m near the lake breeze front;
14. The effect of the Coriolis acceleration was most pronounced in the onshore flow at about 200 m, but was also noted in the return flow aloft. In the afternoon the return flow was from the east-southeast and sandwiched between a northwesterly low level onshore flow and a weak northerly gradient flow above 2500 m;
15. Although no definitive criteria were found to determine the depth of penetration of the lake breeze circulation into the atmosphere, there were indications of influence in the flow pattern up to and above 4000 m in the afternoon;

16. A low level offshore flow, the land breeze, was observed in the morning and again in the evening;

17. The observed land breeze flow layer was on the average 1000 m thick and extended more than 6 km off shore. A maximum offshore wind component, exceeding 6 m sec^{-1}, was observed both over land and over the lake at 200 m between 0600 EST and 0700 EST;

18. There was evidence of a return flow associated with the land breeze in a layer berween 1000 m and 2400 m;

19. The effect of the Coriolis acceleration was observed in the land breeze circulation and was most pronounced in the lower layer in the early morning;

20. Shoreline downwash and upwash effects, caused by changes in roughness between land and lake surfaces were apparent, i.e. the air in the land breeze layer descended as it moved across the shore and out over the lake, while the air in the lake breeze layer ascended at the shoreline, before any thermal instability was expected to have been induced in the layer;

21. Temporal and spacial variations of the water temperature in response to solar heating and dynamic air-sea interaction exceeded 3°C;

22. A mesoscale "lake high" developed during the day in response to the low level divergence and the subsidence over the lake and reached maximum intensity of more than 4 millibars in the early evening;

23. Surface wind data from stations around the lake suggested that the lake effect was homogeneous along the lake shore.

3. URBAN AREA LAKE BREEZE: AN OBSERVATIONAL TRAJECTORY STUDY

During the summer of 1967 several constant level balloons were released near the shores of the southern basin of Lake Michigan. The emphasis in this chapter will be placed on three balloon flights made over the southwestern shore of the lake on two days, August 12 and 13, with weak synoptic pressure gradient and well defined lake breeze circulations.

On these two days winds aloft were measured by conventional single theodolite tracking of pibals and a dense network of surface stations reported winds, temperatures, humidities, and SO_2-concentrations. A large number of photographs were taken both from the surface and from an aircraft. On 13 August measurements of dew point temperatures and aerosol concentrations aloft were made, as an aircraft flew several traverses across the shoreline. The various measurements and observations are presented and correlated in the following sections.

3.1 THE OBSERVATIONAL PROGRAM.

3.1.1 The Tetroons.

The usefulness of super pressured constant volume balloons in atmospheric research has been discussed in several papers, e.g. Angell and Pack (1960), Angell (1961, 1962), Pack and Angell (1963), and Booker and Cooper (1965). The unique value of these balloons lies in their ability to seek pre-determined density (isosteric) surfaces. In the absence of vertical winds they will remain at such a surface throughout the course of the flights and, insofar as such surfaces are horizontal, the mean floating levels will be at constant altitudes. The balloons used in the program reported on here, have a tetrahedron like shape and are commonly called tetroons, Figure 46. They are made of 2-mil Mylar, have a gore length of about 1.5 m and a volume of close to 1 m^3. Tetroons of this type are used by the Environmental Science Service Administration (ESSA) and are available commercially. Operational procedures and characteristics for constant volume balloons have been described by Booker and Cooper (1965) and Booth (1965).

The tetroons were inflated with compressed air and helium and balanced to float at an altitude of approximately 300 m. Required "free lift" and "surface super pressure" were determined from known surface temperature and surface pressure, assumed flight level temperature, and a chart provided by ESSA's Air Resources Laboratory. It should be noted that it is difficult to predict the state of the atmosphere in a transitional zone, such as the one near a shore, and thus also difficult to pre-determine a flight level. However, as our main objective in this study was to investigate the lake breeze circulation near the shore and as double theodolite tracking was employed this difficulty was of minor concern.

To place the tetroon at the approximate flight level as soon as possible after release from the surface a tow and release arrangement recommended by ESSA was used. One, or more, slightly over-inflated 30 gram pilot balloons were used to tow the tetroon to flight altitude, Figure 46. Separation of the tetroon from the towing balloon was accomplished by means of a dynamite fuse. The fuse was ignited upon release

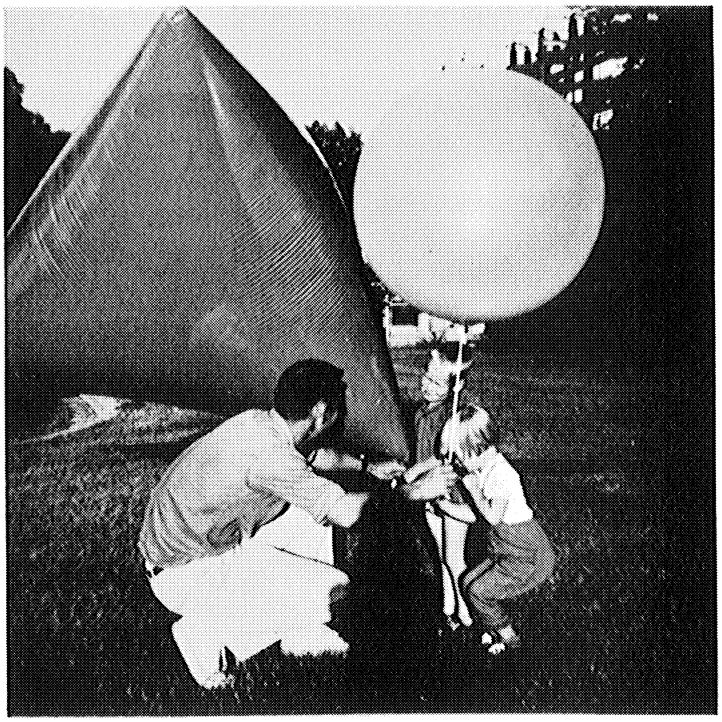

FIGURE 46. Tetroon ready for release.

from the surface and melted a mylar strip. Knowing the burning rate of the fuse and approximate towing speed it was possible, by choosing a proper length of the fuse, to place the tetroon very close to desired flight level within a few minutes after release.

Due to its large surface area and low inertia, a tetroon is expected to respond almost instantaneously to the medium and large scale turbulent air movements, at least in the horizontal. In the vertical the drag, exerted on the tetroon by up or down drafts, has to balance a combination of gravitational and buoyancy forces, which vary in space and time. Angell and Pack (1961) used tetroons for studying vertical air motions in desert terrain and reported balloon oscillations as large as 3000 m (10,000 ft.) in the vertical.

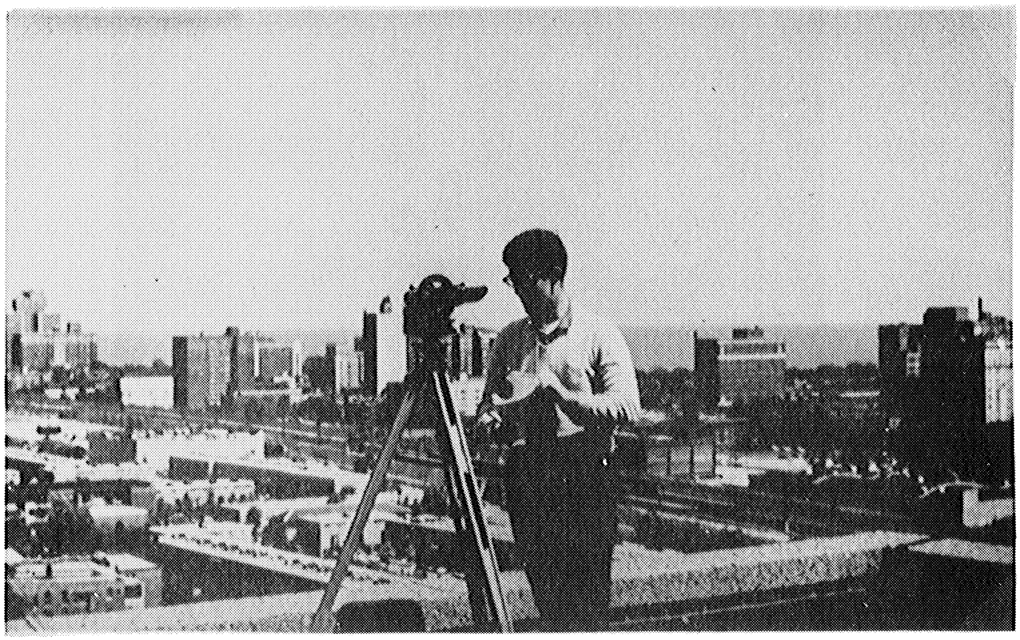

FIGURE 47. Theodolite site CLS on the University of Chicago Campus.

As the vertical oscillations encountered in this study, in general, were less than 1000 m it was assumed that the 3-dimensional tetroon trajectories were good approximations to the 3-dimensional air parcel movements.

Two optical theodolites, located on the roofs of 55 m high buildings near the Univ. of Chicago Campus, were used in tracking the tetroons, Figure 47. The chosen direction of the baseline proved to be good for two of the three runs made; however, the 680 m baseline was too short to give the desired accuracy over the full length of these two flights. During the trackings, elevation and azimuth angles were read to within one tenth of one degree every 30 seconds. In order to eliminate disturbances from small scale turbulence and possible inaccuracies in the theodolite readings, a three point smoother was employed in the trajectory analyses. All analyses were made on a digital computer using a scheme described by Biggs (1962). When the tetroon crossed the baseline or its extension, or traveled along it, the horizontal balloon-positions were calculated from the two elevation angles. Otherwise, these two angles were used to calculate the balloon altitude twice. The smoothed average of the two heights was used to describe the trajectory, while the difference between these two calculated heights gave an indication of the goodness of the specific observation.

3.1.2 Location of the Study and Auxiliary Observation Program.

The tetroons were released from the shore of Lake Michigan at a location near the University of Chicago Campus, approximately 7 km south of downtown Chicago, Illinois, Figure 48. The lake shore runs from SSE to NNW and can, for the purpose of this analysis, be considered a straight line. Topographically the land is slightly rolling, with only a small slope (on the average less than 1:1000) toward the lakeshore, and it was assumed that no slope winds developed. The area inland from the shore was moderately "built up" with two to ten story apartment or office buildings and there were no heavy industries within several kilometers of the University of Chicago Campus. Heavy traffic was noted on two expressways, which run parallel to the shore, 2 - 5 km inland and on several major streets closer to the shore. Heavy industry was located near Lake Calumet, 10 - 15 km south and along the shore of Lake Michigan, 5 - 20 km southeast of the release site. An aerial photo of the area is presented in Figure 49.

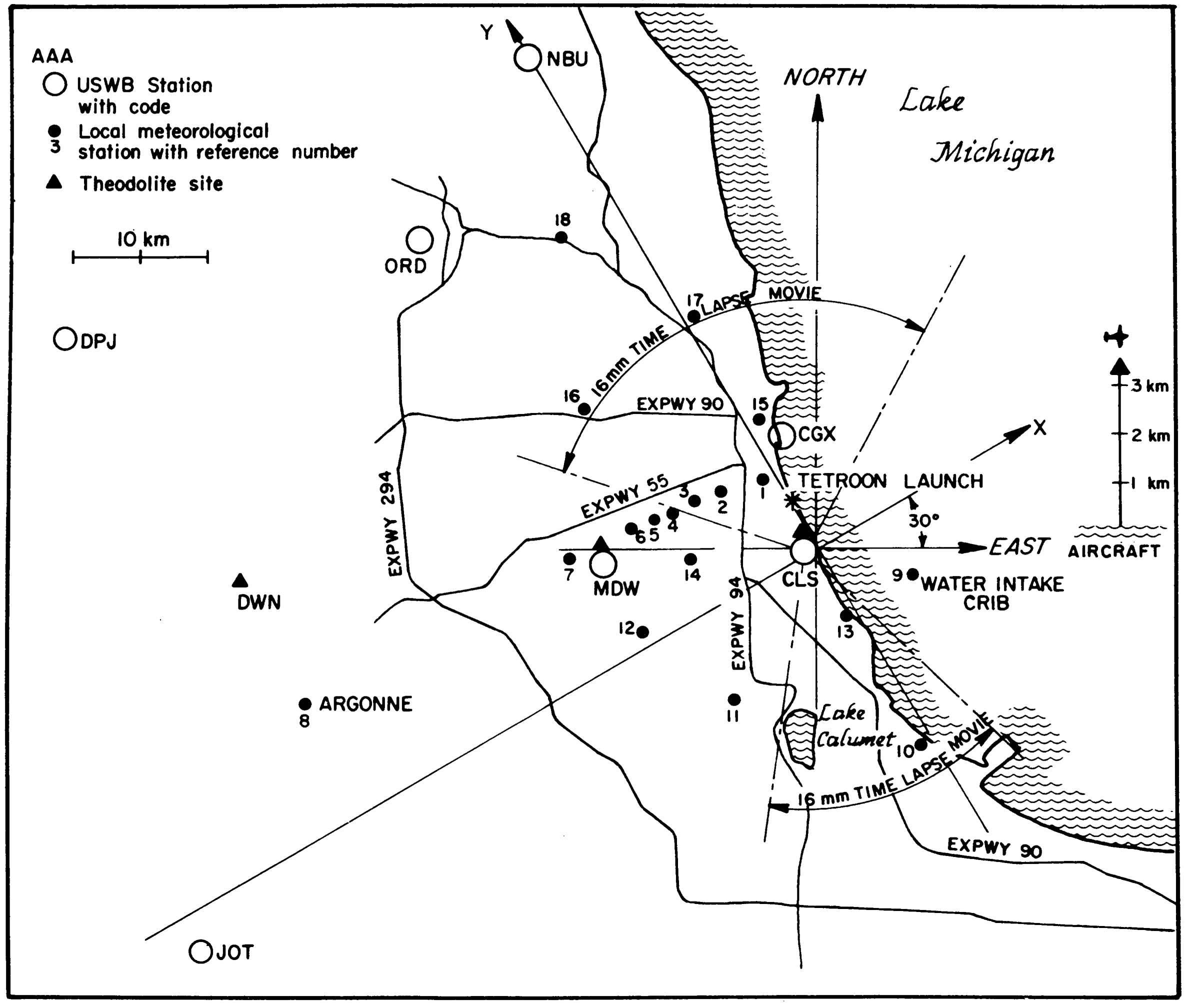

FIGURE 48. Location of the observation stations in the Chicago, Illinois area on the southwestern shore of Lake Michigan.

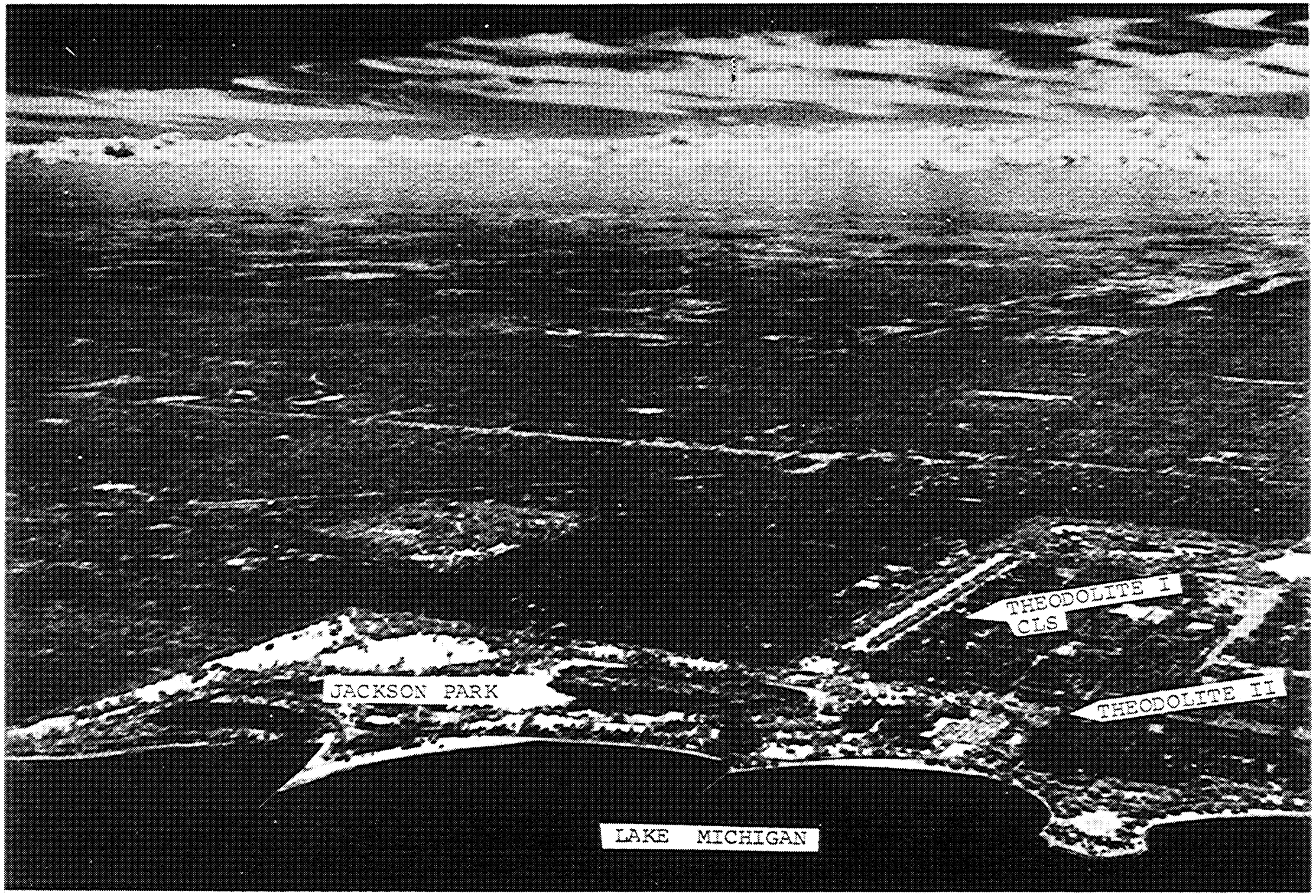

FIGURE 49. Aerial infrared photo of the Chicago, Illinois area taken from 3000 m at 1400 CST, 15 July, 1967.

One major reason for selecting this location was the abundance of supporting meteorological observations available in the region. As is indicated in Figure 48 and Table 3, temperatures, humidities, and winds were measured both at the surface and aloft at several stations, lake water temperatures were recorded 6 km off shore, and SO_2-concentrations were measured at several surface stations. Aerosol concentrations were measured from a Beech Queen Air plane belonging to the National Center for Atmospheric Research (NCAR). The aircraft was equipped with an automated data acquisition system, which has been described in detail by Langer et al (1968).

Pibal observations at MDW and DWN were made by local high school students. U. S. Weather Bureau stations and military airports reported hourly observations. 5 minute averages of winds and SO_2-concentrations were automatically recorded and telemetered to a centralized data acquisition system at least every hour. Temperatures and relative humidities were recorded on hygrothermographs at several local high schools. Moses and Bogner (1967) have described meteorological instrumentation and data acquisition system at Argonne.

In addition to these quantitative data, qualitative information was provided from time lapse movies and photographs of smoke and clouds, taken both from the surface and from the aircraft.

TABLE 3. Meteorological stations in the Chicago, Illinois area. Coordinate system and station locations are marked on map in Figure 48. Perpendicular distances to shore are given in brackets where those distances differ from the X-coordinate. Location coordinates are given in km.

STN	Name of Station	Location		Parameter			Remark
		X	Y	Temp	Hum	Wind	
CLS	USWB, Radar						Pibal, SO_2
	U. of Chicago	- 1	0				Tetroon
MDW	Midway Airport	-13(14)					USWB, Pibal
DWN	Downers Grove	-33(37)	17				Pibal
CGX	Miegs Airfield	2 (0)	7	x	x	x	USWB.
ORD	O'Hare Airport	-13(22)	31	x	x	x	USWB.
NBU	Glenview Airport	0 (9)	38	x	x	x	Military
DPJ	DuPage Airport	-28(37)	37	x	x	x	USWB.
JOT	Joliet Airport	-58	- 1	x	x	x	USWB.
1	Williams	- 1 (2)	7	x	x		*)
2	Armour	- 3 (4)	7	x	x		*)
3	Greene	- 5 (6)	7	x	x		*)
4	Shields	- 7 (8)	7	x	x		*)
5	Gunsualus	- 8 (9)	7	x	x		*)
6	Edwards	-10(11)	7	x	x		*)
7	Hale	-14(15)	7	x	x		*)
8	Argonne	-35(36)	8	x	x	x	
9	Water intake	6	- 5			x	Water Temp.
10		- 1	-16			x	*)
11		-10	- 6			x	*), SO_2
12		-13	1			x	*), SO_2
13		- 1 (0)	- 5			x	*)
14		- 7	3			x	*), SO_2
15		1 (1)	10			x	*), SO_2
16		- 9(13)	16			x	*), SO_2
17		1(13)	18			x	*), SO_2
18		- 4(11)	27			x	*), SO_2

*) Station run by the Department of Air Pollution Control in Chicago, Illinois.

3.2 PREVAILING METEOROLOGICAL CONDITIONS, 12 AND 13 AUGUST, 1967.

Between August 11 and 14, 1967 a stagnant surface high and a weak 500 mb trough dominated the large scale weather in the entire Great Lakes Basin, Figures 50 and 51. The pressure gradients both at the surface and aloft, and thus also the gradient winds, were very weak in the region. The winds were in general from NNE and on neither day did they exceed 15 m $\sec^{-1}$ at the 500 mb level. Winds and temperatures around the Lake Michigan basin at 0600 CST, 12 August are presented in Figure 52. Land breezes were, at that time, observed near the shore around the entire lake and the area shown in the figure was enclosed by a 3 mb isobar. Lake surface isotherms, included in Figure 52, showed uniform lake surface temperatures in the southwestern portion of the lake while there was a strong gradient along the eastern shore. Note also the apparent urban heat island effects on air temperatures near Chicago, Illinois and Muskegon, Michigan.

An ESSA III satellite photograph, taken 1410 CST, 12 August, Figure 53, showed a frontal zone moving east over the Appalachian mountains and a cloud free Great Lakes basin. The Great Lakes basin portion of the ESSA III photograph is presented in Figure 54 with superimposed weather station plots and an analysis of the wind field. A lake breeze effect was observed at that time around all the Great Lakes and the locations of the cloud bands were in good agreement with the locations of the lake breeze fronts as estimated in the analyses of the winds.

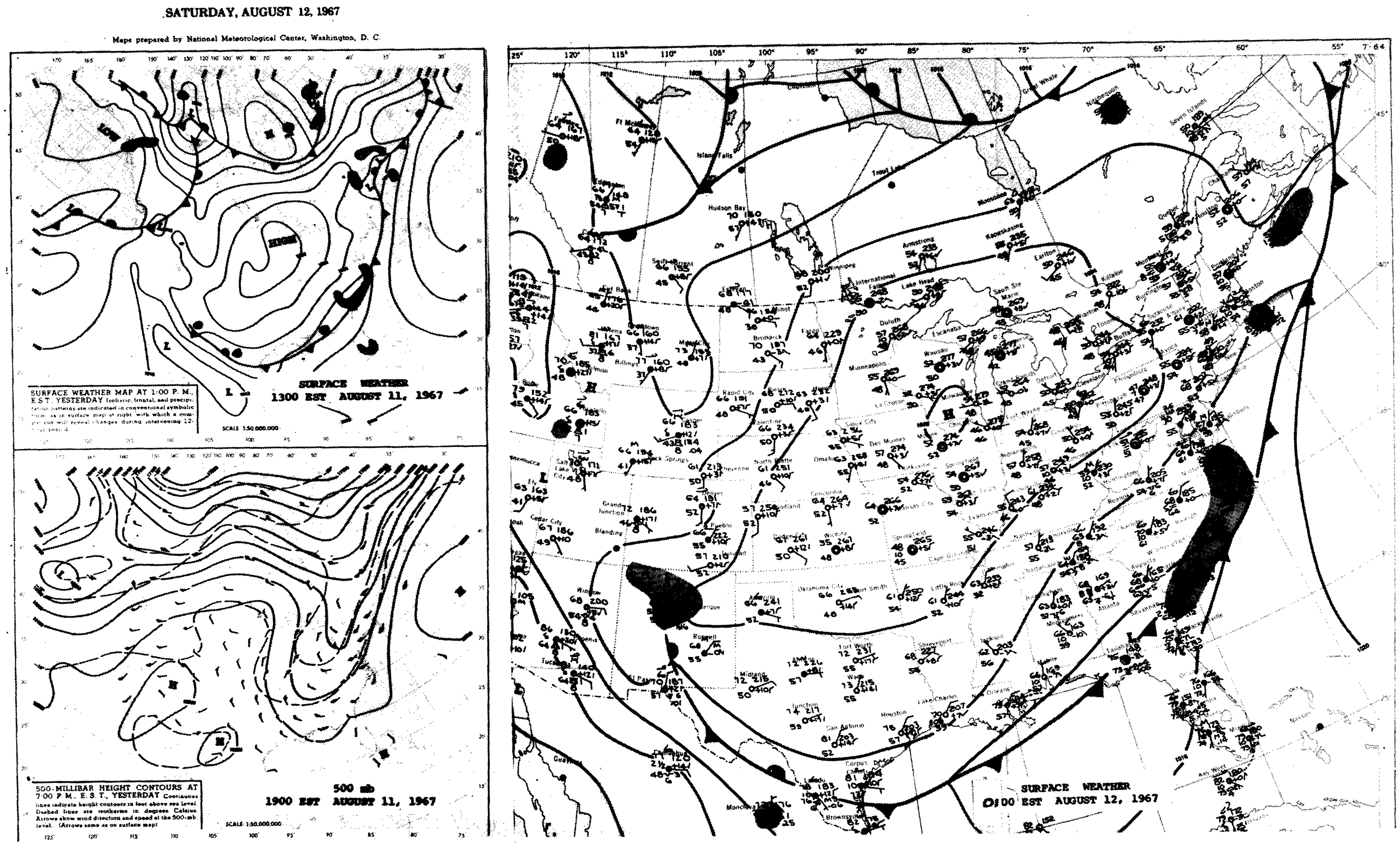

FIGURE 50. U.S. Weather Bureau Daily Weather Map for 0100 EST, 12 August, 1967. Upper insert shows the sea level isobars at 1300 EST, 11 August and the lower insert shows the height contours for the 500 mb surface at 1900 EST, 11 August.

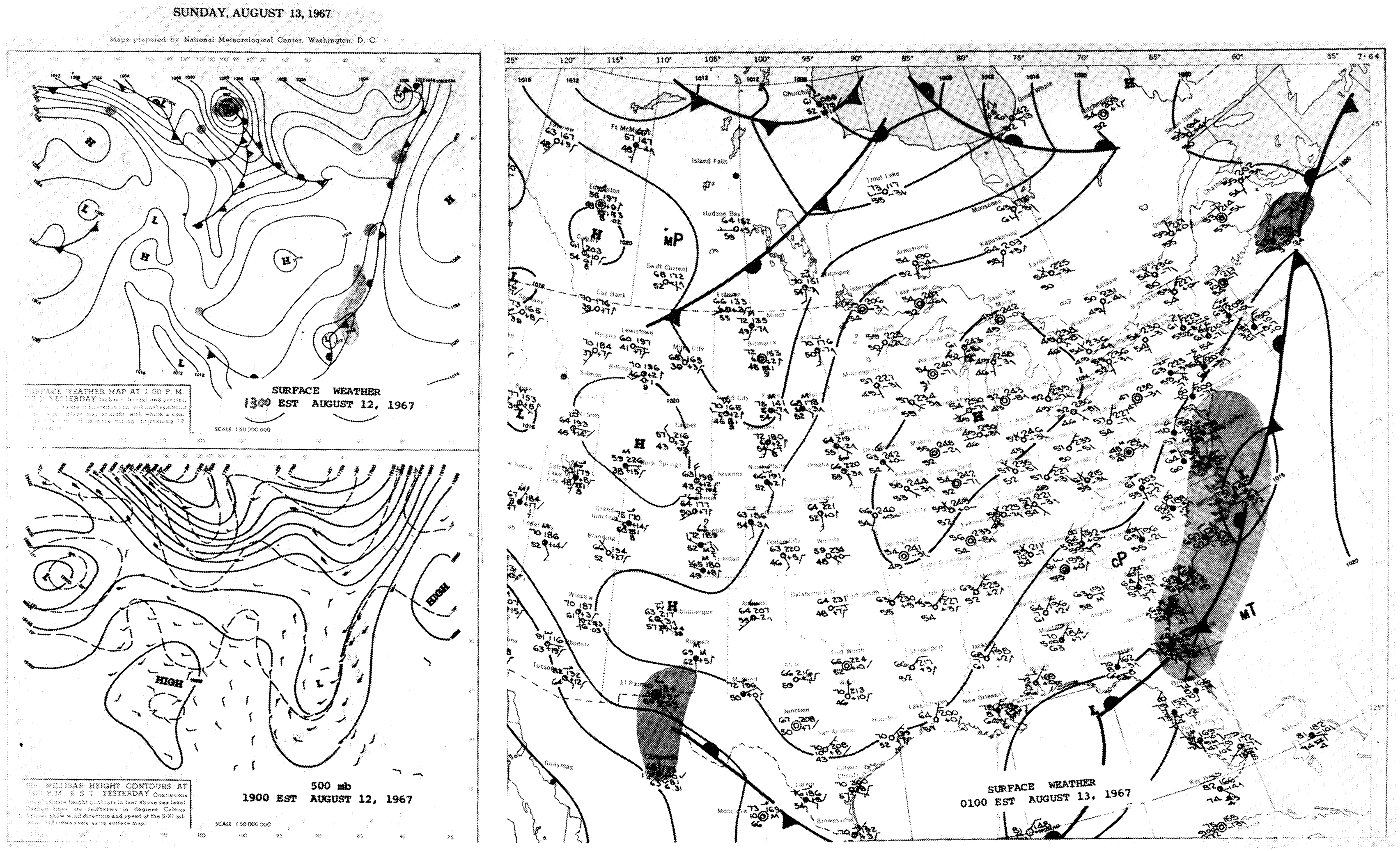

FIGURE 51. U.S. Weather Bureau Daily Weather Map for 0100 EST, 13 August, 1967. Upper insert shows the sea level isobars at 1300 EST, 12 August and the lower insert shows the height contours for the 500 mb surface at 1900 EST, 12 August.

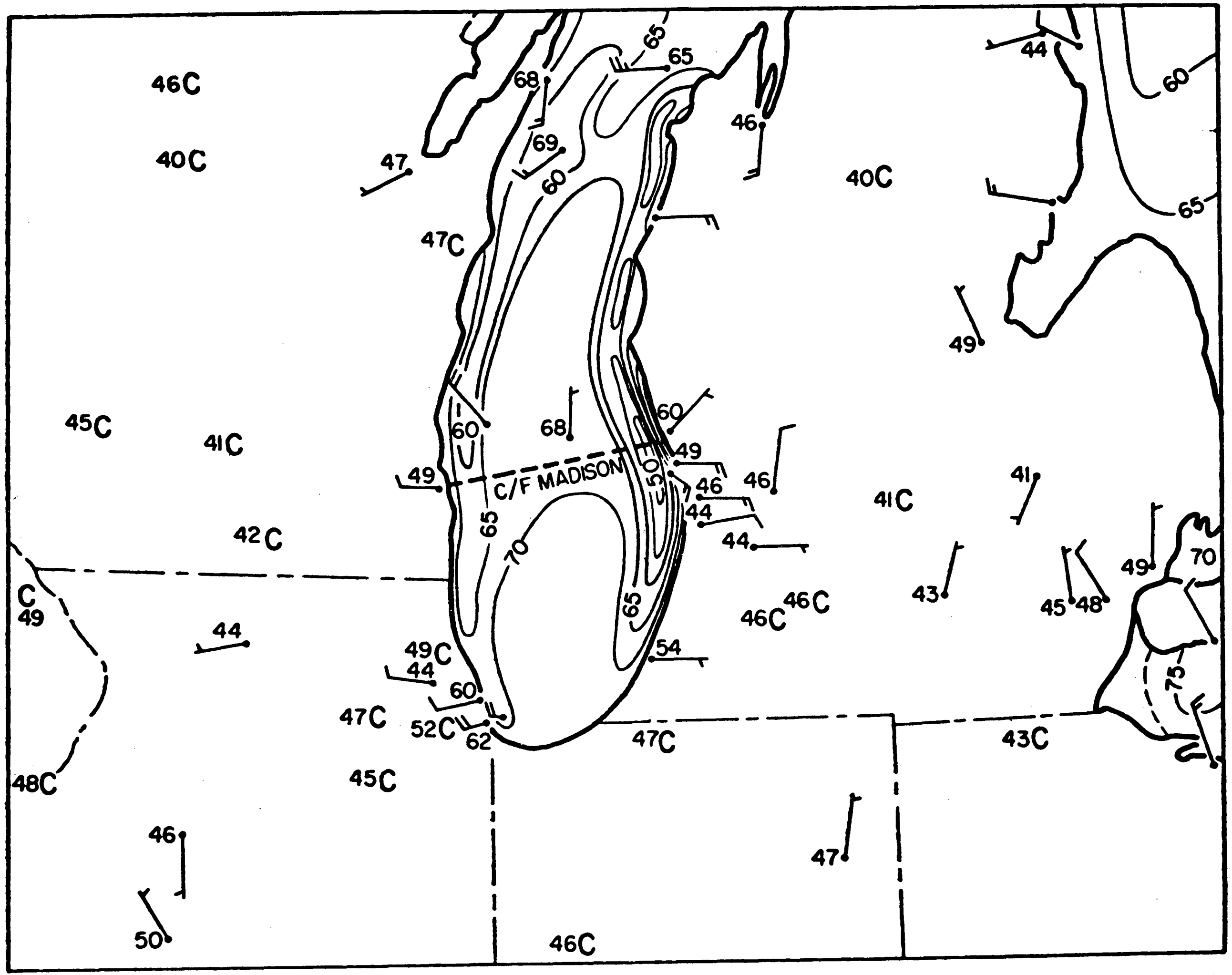

FIGURE 52. Mesoscale surface winds and air and water temperatures in the Lake Michigan Basin at 0600 CST, 12 August, 1967.

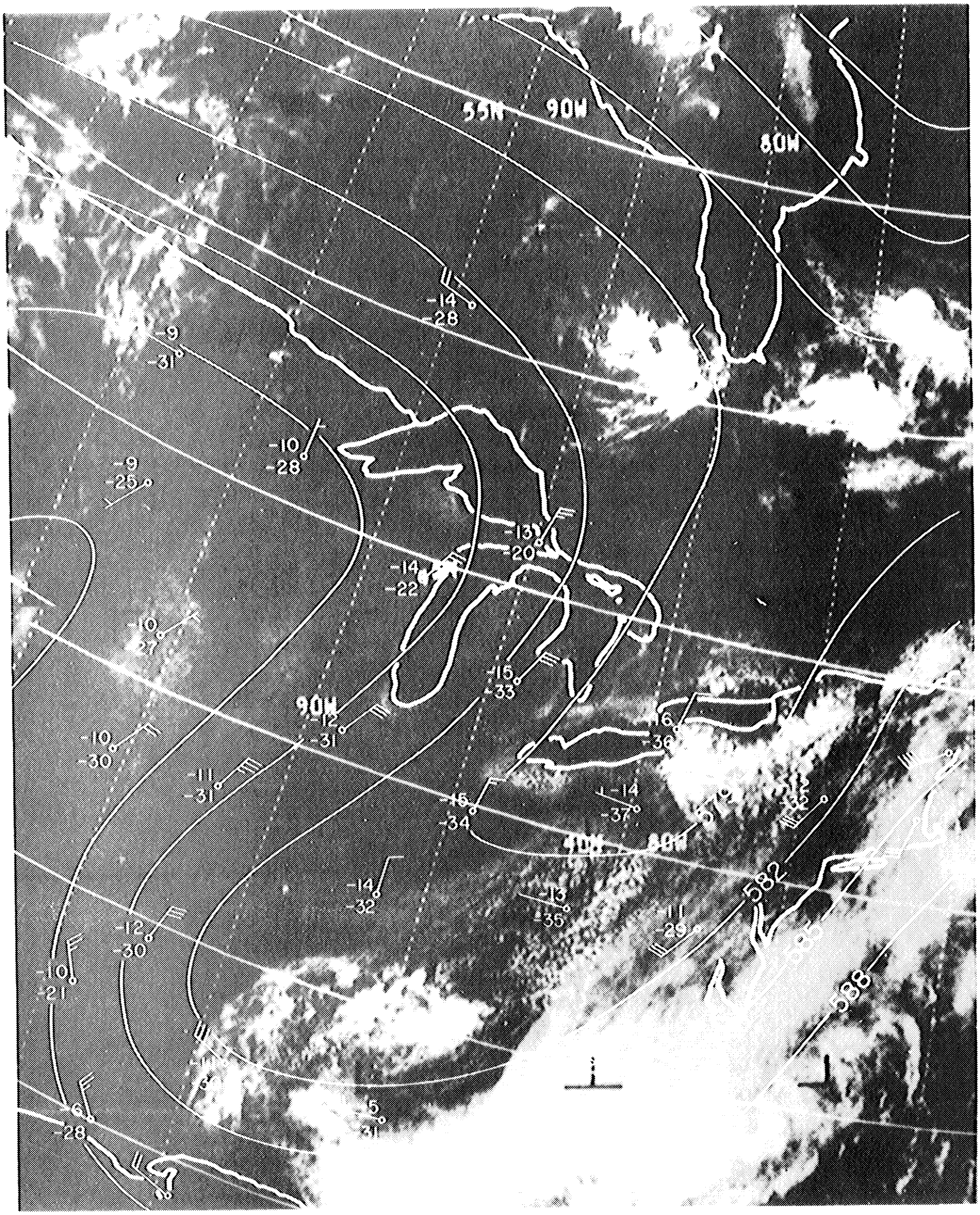

FIGURE 53. ESSA III photograph taken 1410 CST, 12 August 1967, with superimposed 500 mb height contours, winds, and air and dewpoint temperatures.

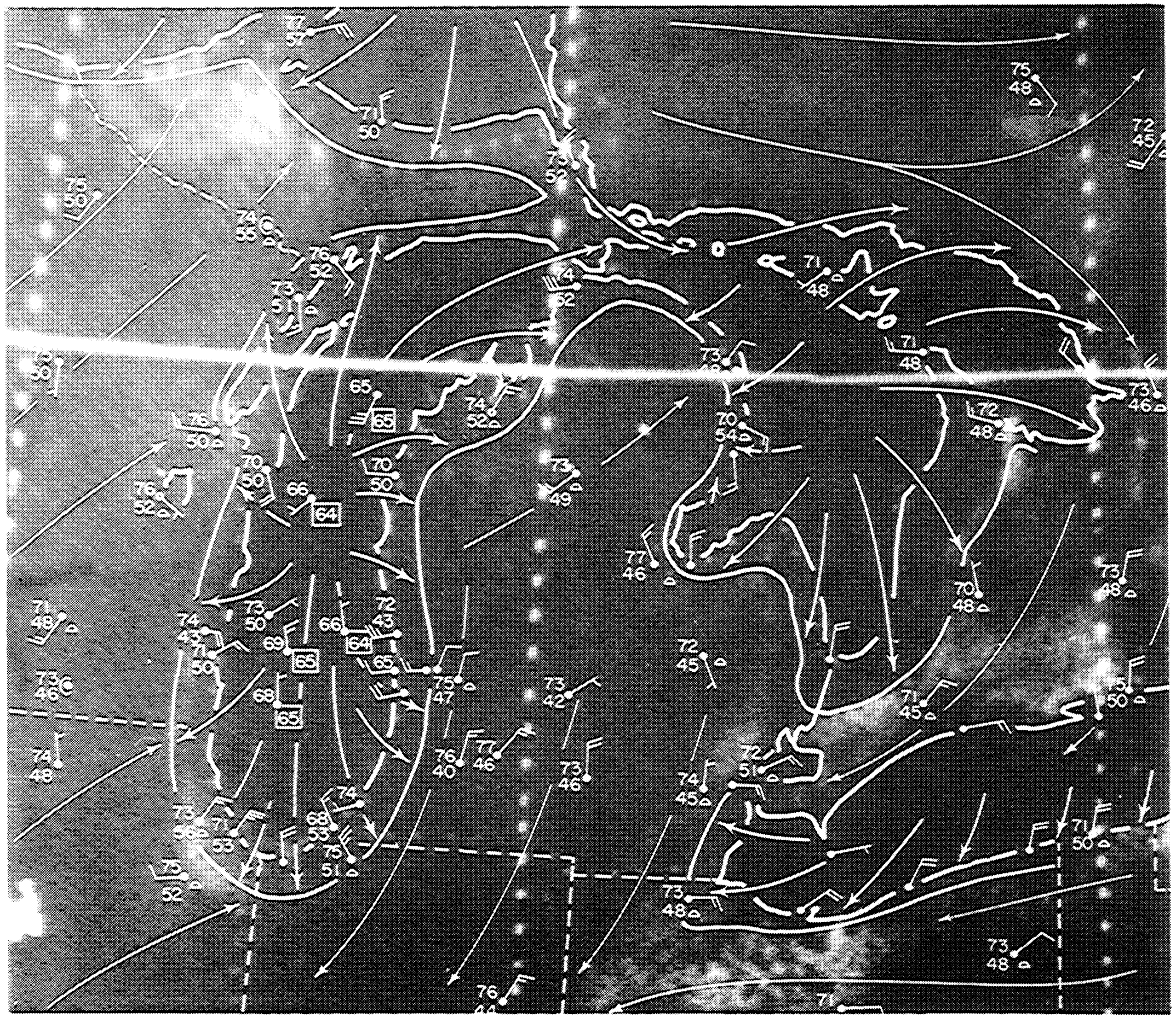

FIGURE 54. The Great Lakes basin portion of an ESSA III photograph taken 1410 CST, 12 August 1967, with superimposed surface wind analysis and station weather. The lake shores and estimated locations of the lake breeze fronts are indicated.

Inspections of satellite photos taken in the afternoons, 12 and 13 August showed that there were great similarities in large scale cloud patterns on the two days; however, the lake breeze clouds around the southwestern portion of Lake Michigan were not so apparent on the 13th, as they were thinner and more scattered.

Data from the University of Michigan meteorological towers on the eastern shore of Lake Michigan showed that nocturnal land breezes were reversed to lake breezes on both days. On 12 August the wind shifted from ESE to WNW between 0800 CST and 0830 CST near the shore. At 0930 CST the lake breeze front had reached 8 km, at 1130 CST 16 km, and at 1400 CST 24 km inland. A maximum lake breeze wind of 5 m $\sec^{-1}$ was observed at the shore, as well as at the stations 8 km and 16 km inland in the late afternoon and the effect of the Coriolis acceleration was apparent. Although the synoptic situation on 13 August was quite similar to the one on 12 August, the transition from land to lake breeze was very different. On 13 August the land breeze was strongest some hours before sunrise. Between 0500 CST and 0800 CST, the wind near the shore was light and variable. A westerly windshift and an increase in wind speed occurred between 0700 CST and 0930 CST at all stations from the shore to 24 km inland. After this "sudden start" the lake breeze developed relatively fast over land and reached a maximum speed of 5 m $\sec^{-1}$ both 8 km and 16 km inland. At the shore, however, the maximum strength of the lake breeze was 2 m $\sec^{-1}$, occuring more than one hour after the inland maximum.

On 12 August R/V Inland Seas of the Great Lakes Research Division at the University of Michigan was stationed approximately 160 km north of Chicago, Illinois, and 30 km off the western shore of Lake Michigan. Before 1000 CST the wind at her masthead, 12 m above the lake surface, was NNE and decreasing. Between 1000 CST and 1400 CST the wind was light and variable at the location of the ship. Between 1400 CST and 1600 CST, the ship was on her way to port. Within approximately 15 km from the shore the wind was SE at 3 m sec^{-1}.

On 13 August R/V Inland Seas crossed the lake from west to east approximately 150 km north of Chicago, Illinois. Close to the western shore, at 0600 CST, the wind was from the west at 3 m sec^{-1}. Around noon the winds over the center portion of the lake were in general from the south at 0 - 2 m sec^{-1}. Close to the eastern shore, at 1800 CST, the wind was WNW at more than 1 m sec^{-1}. Hourly observations, recorded in a logbook, showed that at all times, on both days, the lake surface was smooth and that the air was relatively stable in a layer between 5 m and 12 m above the lake surface. However, in all observed cases, the deck height air temperature, approximately 5 m above the lake surface, was more than 0.9°C lower than the lake surface temperature. In the afternoon on 12 August, this temperature difference exceeded 4°C for more than 2 hours. For 23 hourly observations at sea, on 12 and 13 August, the average temperature lapse rate in the lowest 5 m was 0.4°C m^{-1}.

3.3 OBSERVED TEMPERATURES, HUMIDITIES, WINDS, AND POLLUTANTS.

This section presents the various parameters measured near the southwestern shore of Lake Michigan on 12 and 13 August, 1967. Station references as well as the coordinate system used throughout this chapter are depicted in Figure 48 and listed in Table 3. Surface data, winds aloft as measured by conventional methods, and the analysed tetroon runs are presented and comparisons are made between winds derived from the tetroon trajectories and winds measured by conventional methods. Analyses of the dew point temperatures and aerosol concentrations, measured during the across shore airplane traverses, and the analysed photographs of smoke and haze are also presented.

3.3.1 Surface Temperatures, Humidities, SO_2-Concentrations, and Winds.

Temperatures and dew points obtained from hygrothermograph records from stations at various distances inland are presented in Figure 55. The lake effect was observed both as a leveling off or decrease in the dry bulb temperature and as an increase in the dew point temperature. At the stations closest to the shore the time of arrival of the lake breeze front could be determined to within ± 15 minutes from the records. However, these time-estimates became increasingly uncertain further inland, due to modification of the lake air as it passed over land. The pulsating penetration of the lake breeze front, shadows from the clouds associated with the front, and turbulence in and behind the convergence zone caused variability in the recorded temperatures and humidities. It is, however,

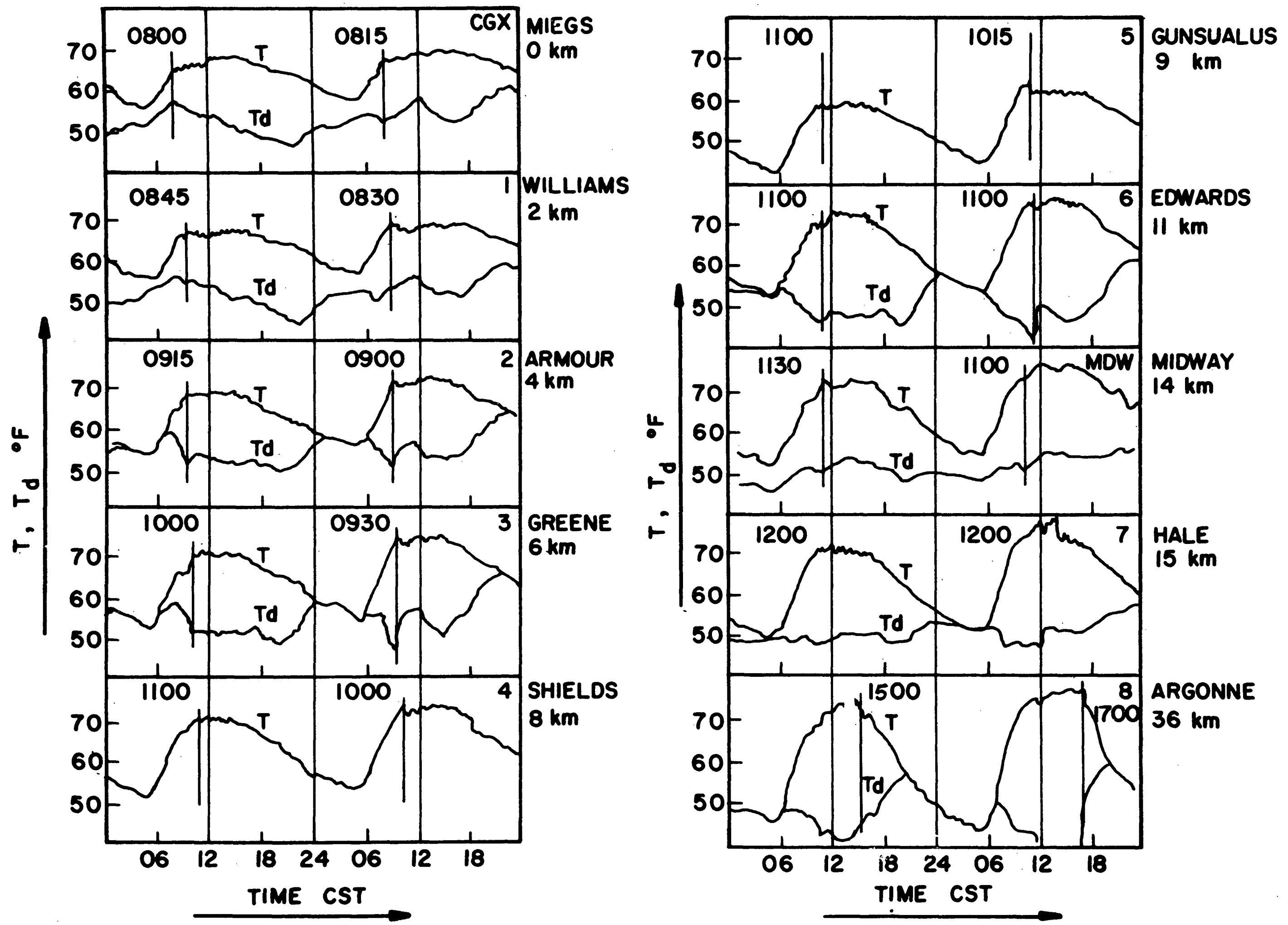

FIGURE 55. Temperatures and dew point temperatures as a function of time at several stations at various distances inland, 12 and 13 August, 1967. Locations of the stations are specified in Figure 48 and Table 3. Estimated local times for lake breeze frontal passages are indicated.

evident from Figure 55 that the lake breeze reached stations successively further inland both on 12 and 13 August.

Hourly analyses of the local surface winds on 12 and 13 August are presented in Figures 56 - 82, where measured SO_2-concentrations also are plotted. The location of the lake breeze front was determined on the basis of wind shifts and streamline convergence. The lake breeze fronts progressed inland at an average rate of 1.1 m sec^{-1}, in close agreement with the rate estimated from surface temperature and humidity changes near $Y = 0$. The progression was, however, slower to the north and faster to the south. There was also a slight difference in progression characteristics between the 12th and the 13th of August.

Unfortunately no detailed information regarding the measured SO_2-concentrations was available to the author during the preparation of this report. The plotted SO_2-concentrations were 5 minute averages, automatically recorded every hour and the observed changes, in these concentrations, seems to indicate that this gaseous pollutant might be an excellent tracer in a lake-land breeze regime.

3.3.2 Vertical Wind Profiles.

30 gram pilot balloons, tracked by single theodolites were released on the hour from Midway Airport (MDW) and Downers Grove (DWN), while from the University of Chicago Campus (CLS), such releases were made only before and after tetroon flights on 12 and 13 August, 1967. The times of release and the length of each track is given in Table 4. Theodolite observations were made at 1/2 minute intervals. A three point smoother was used as the analyses of the tracks

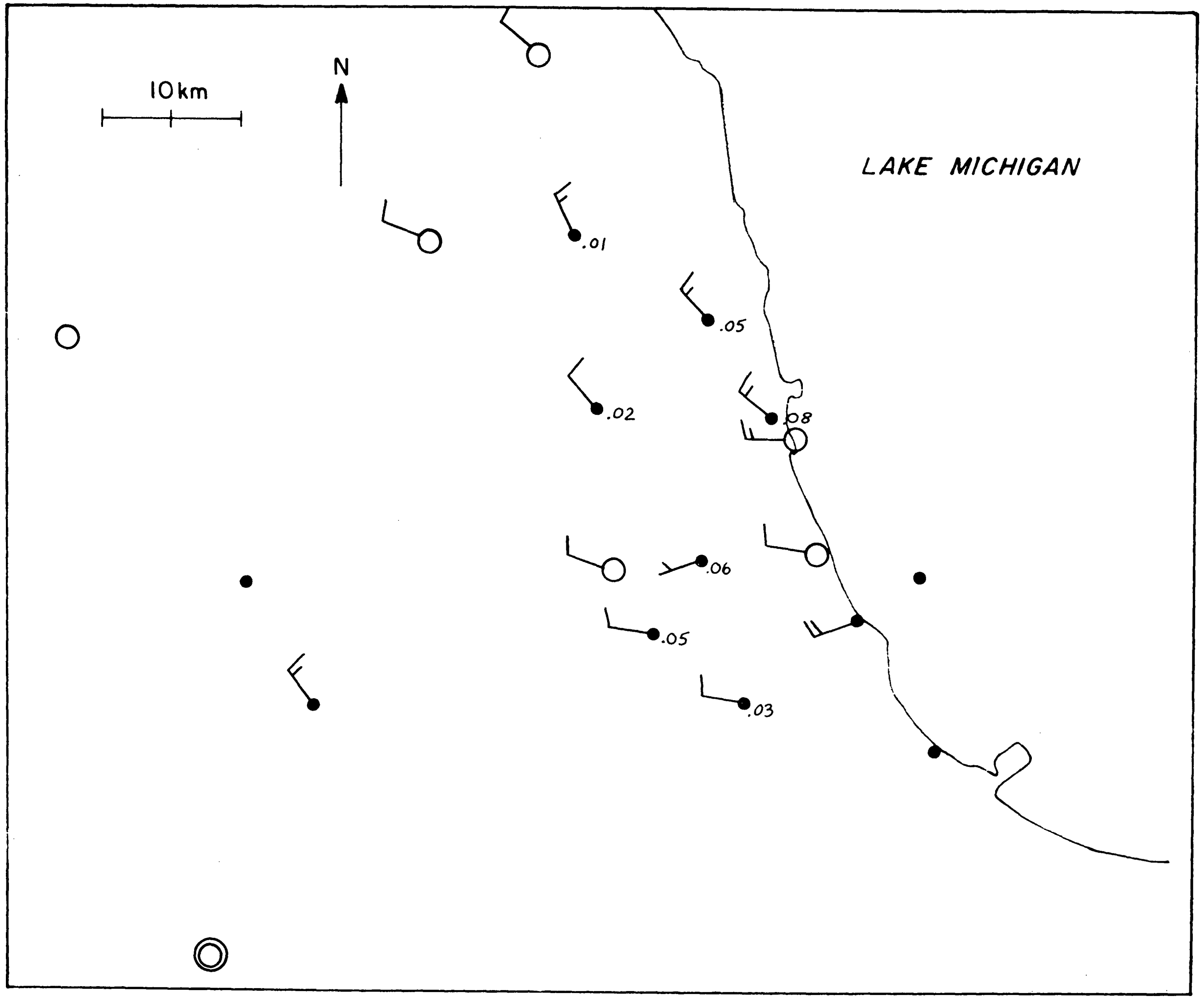

FIGURE 56. Winds and SO_2-concentrations at the surface in the Chicago, Illinois area at 0600 CST, 12 August, 1967. Circle around station indicates calm; no barb, 0.2-0.5; half barb, 0.5-1.5; full barb, 1.5-2.5 m sec^{-1} wind speed, etc. SO_2-concentrations are given in ppm.

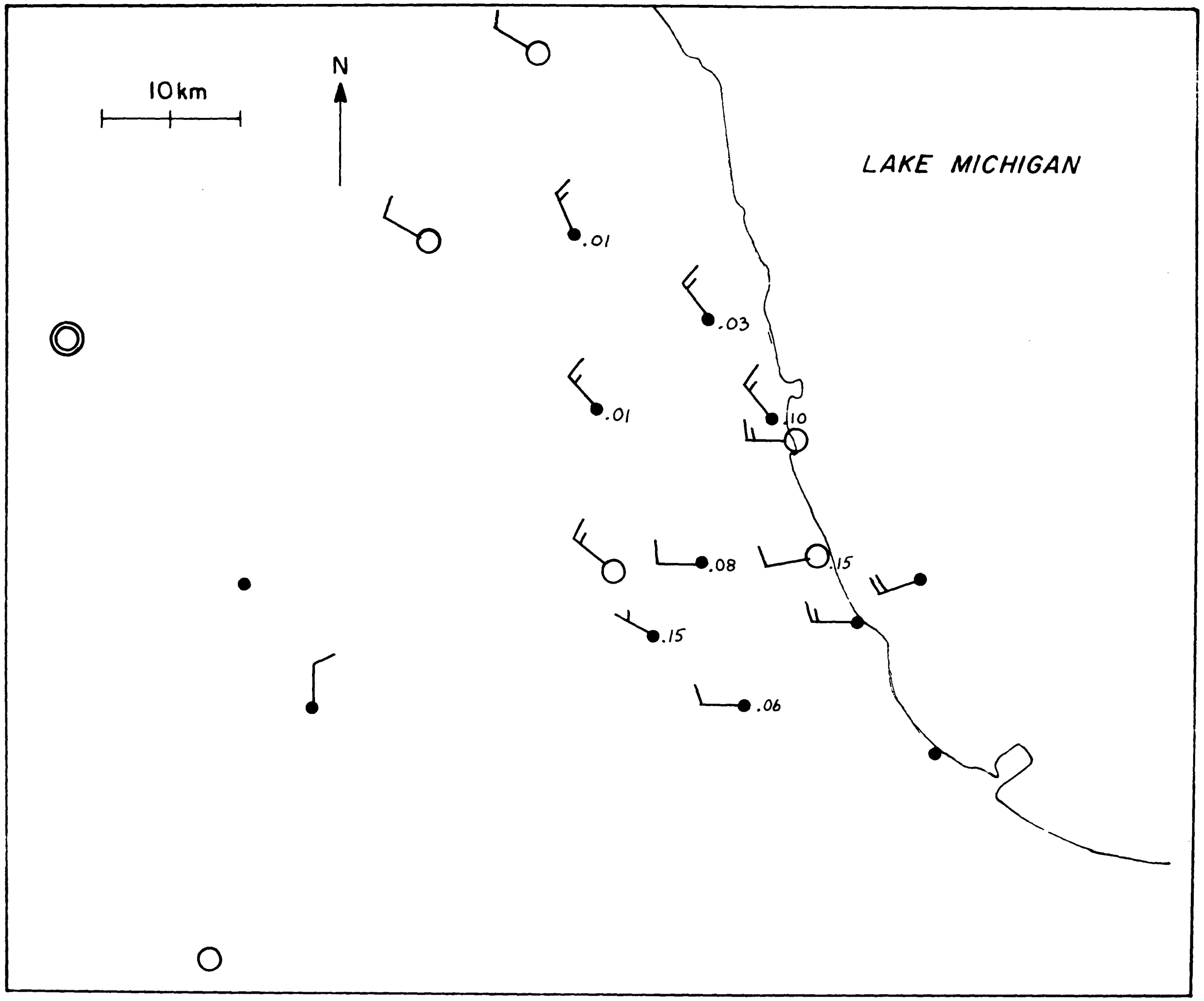

FIGURE 57. Same as Figure 56 except for 0700 CST, 12 August, 1967.

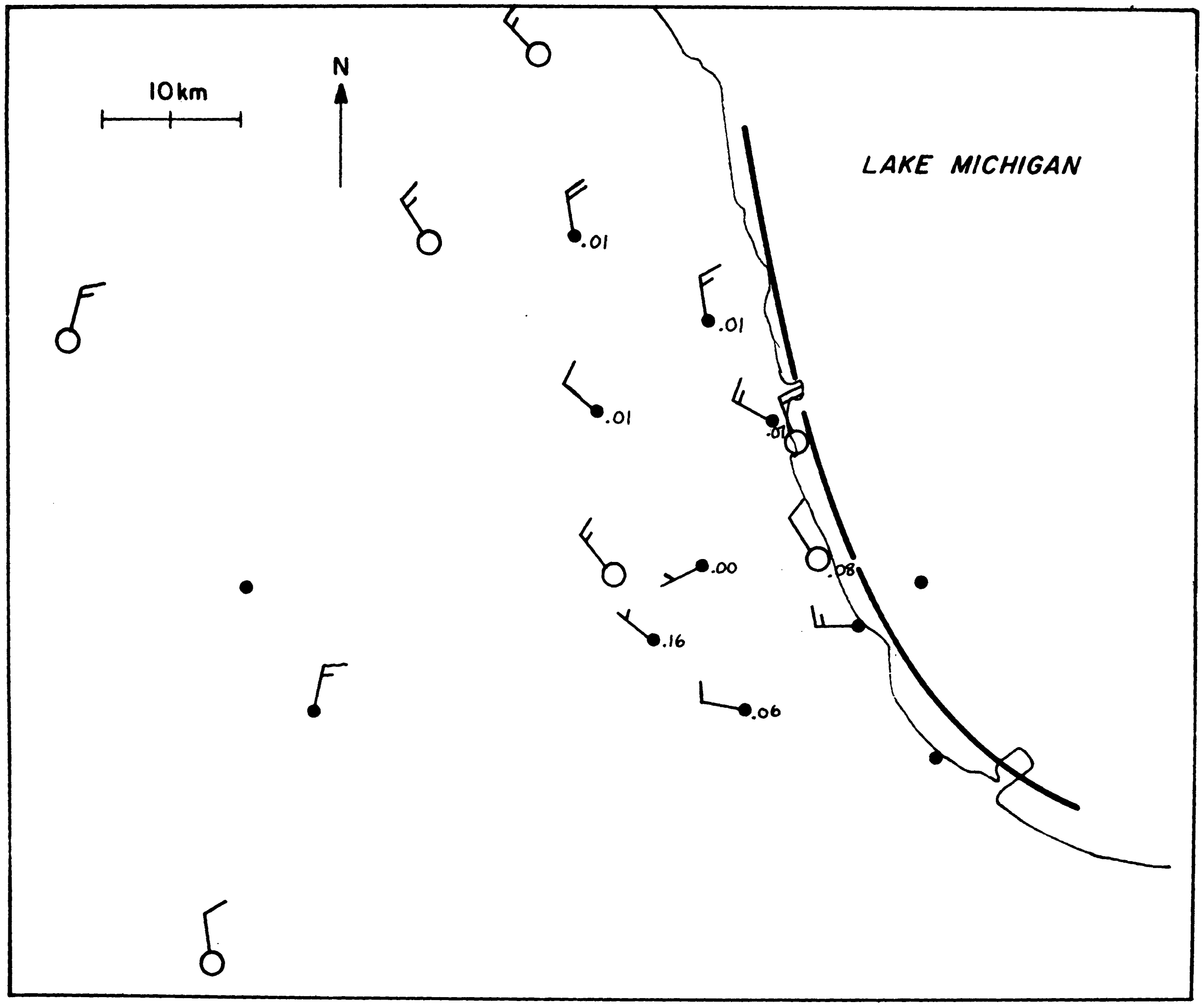

FIGURE 58. Same as Figure 56 except for 0800 CST, 12 August, 1967.

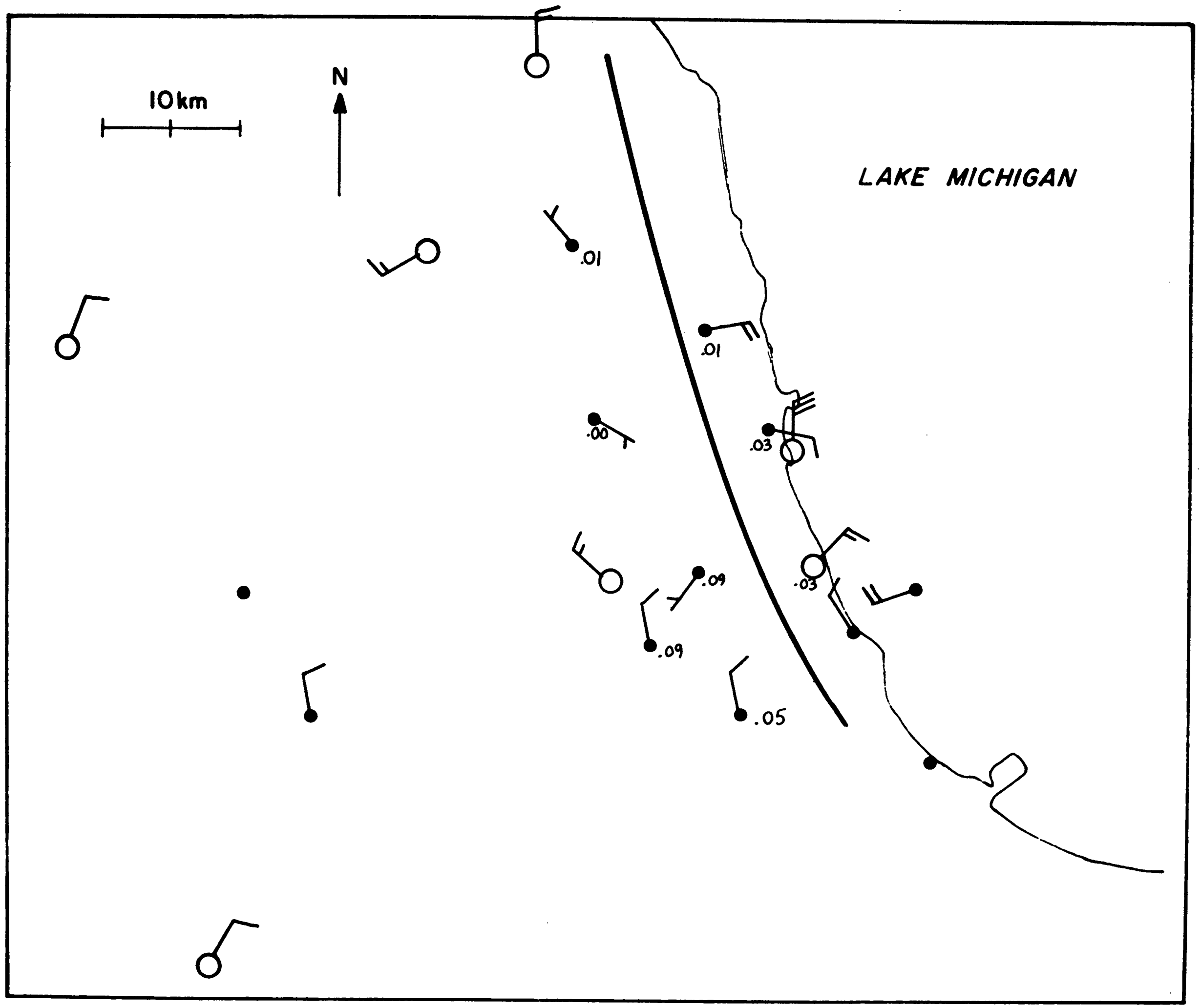

FIGURE 59. Same as Figure 56 except for 0900 CST, 12 August, 1967.

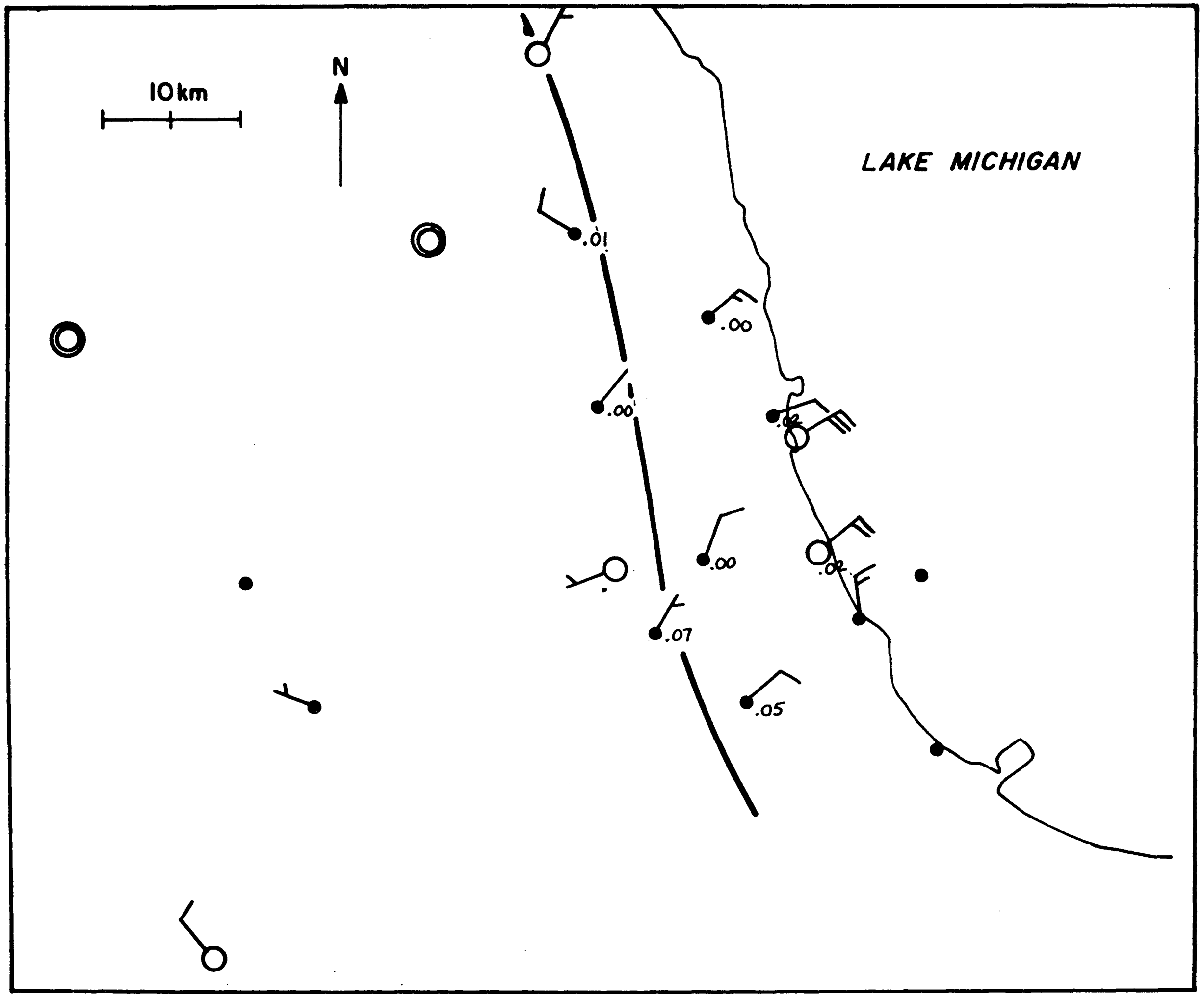

FIGURE 60. Same as Figure 56 except for 1000 CST, 12 August, 1967.

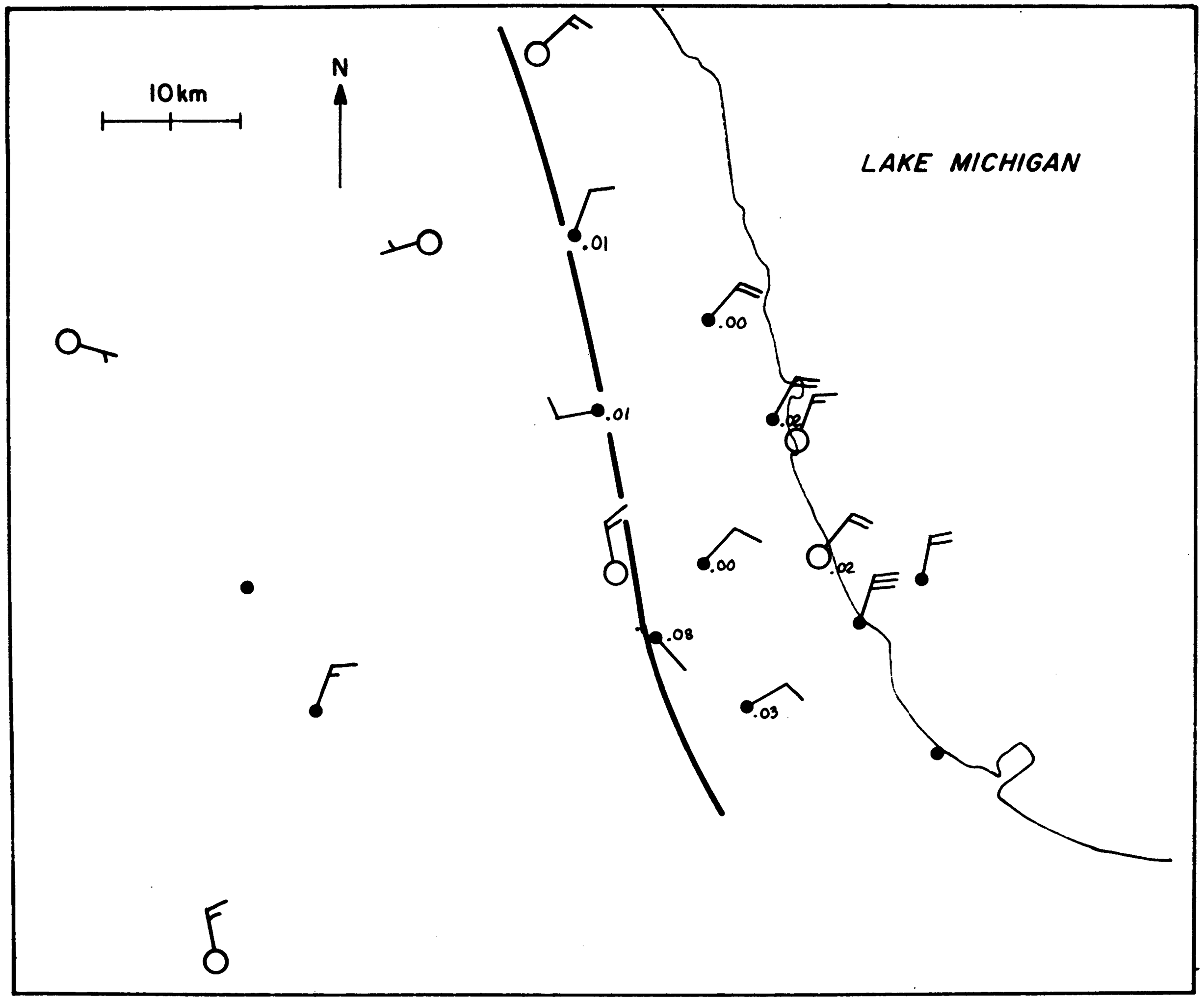

FIGURE 61. Same as Figure 56 except for 1100 CST, 12 August, 1967.

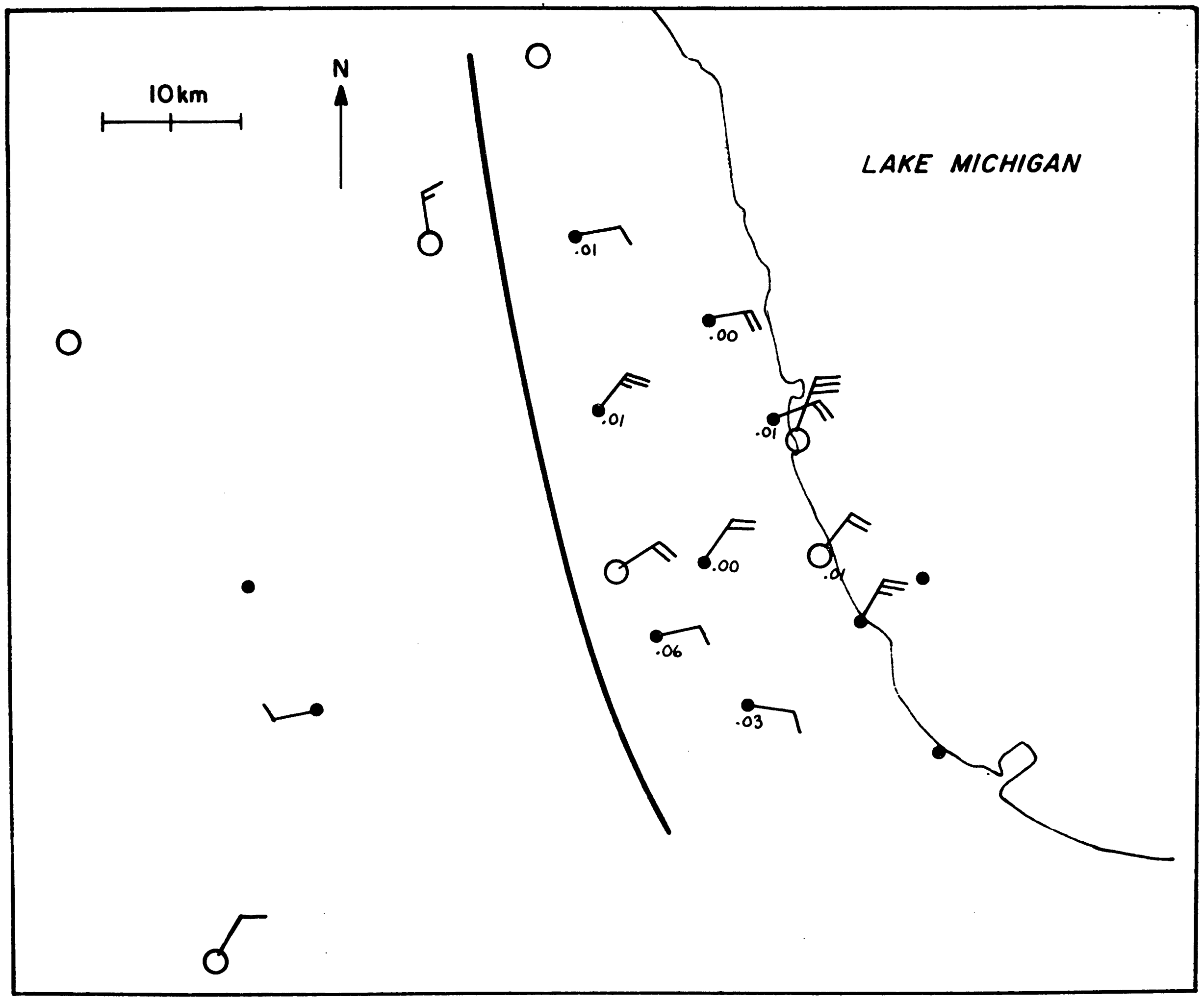

FIGURE 62. Same as Figure 56 except for 1200 CST, 12 August, 1967.

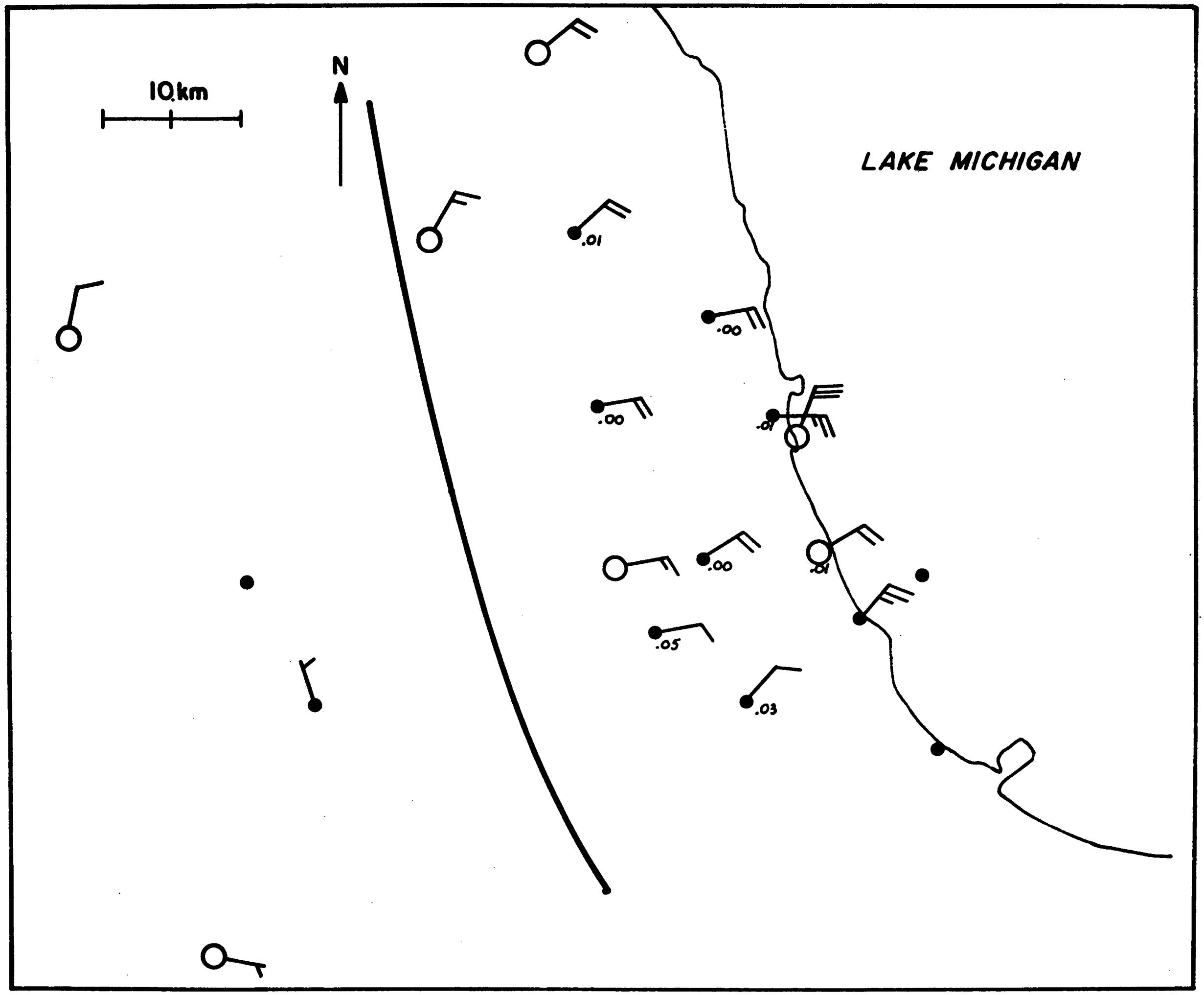

FIGURE 63. Same as Figure 56 except for 1300 CST, 12 August, 1967.

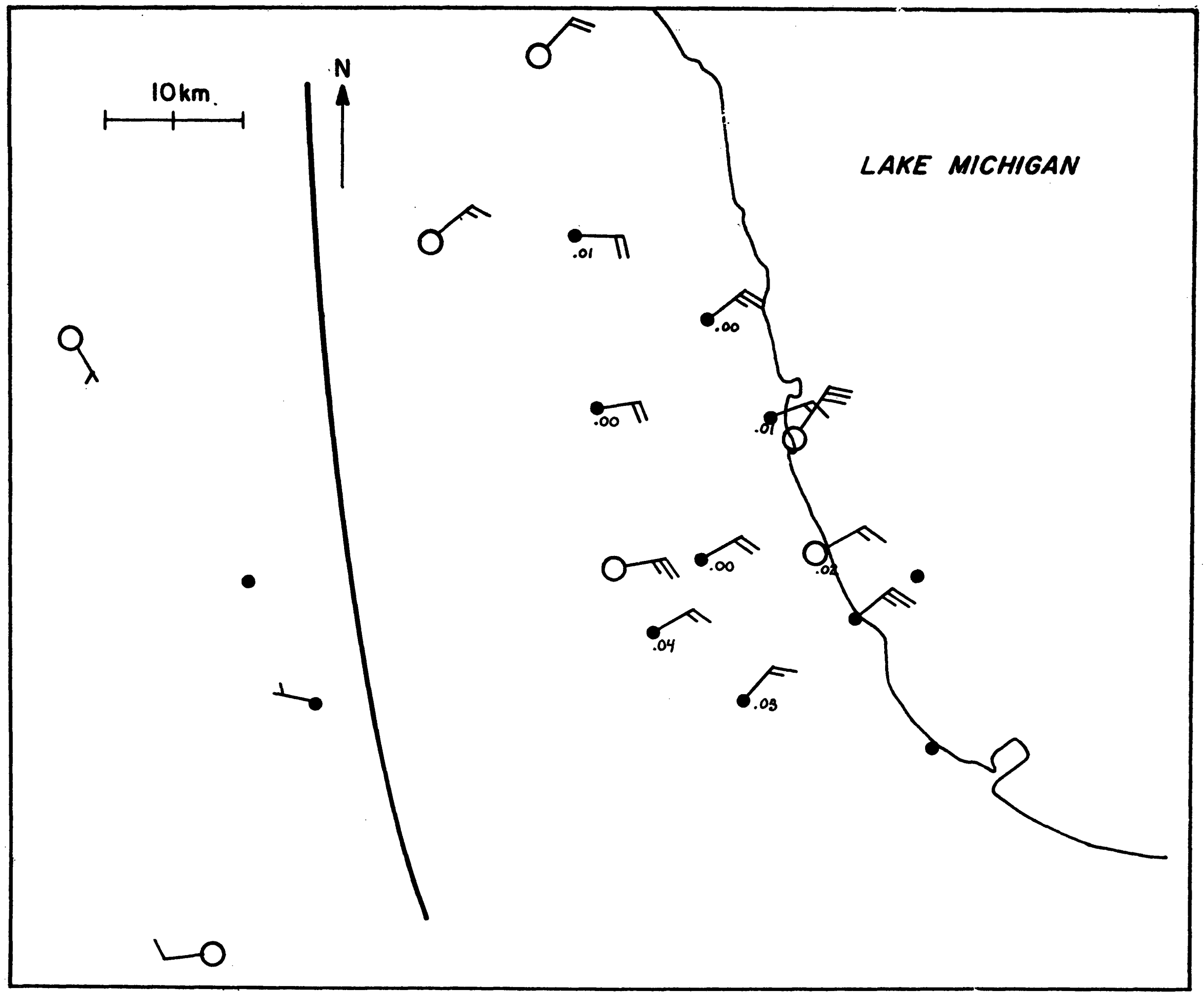

FIGURE 64. Same as Figure 56 except for 1400 CST, 12 August, 1967.

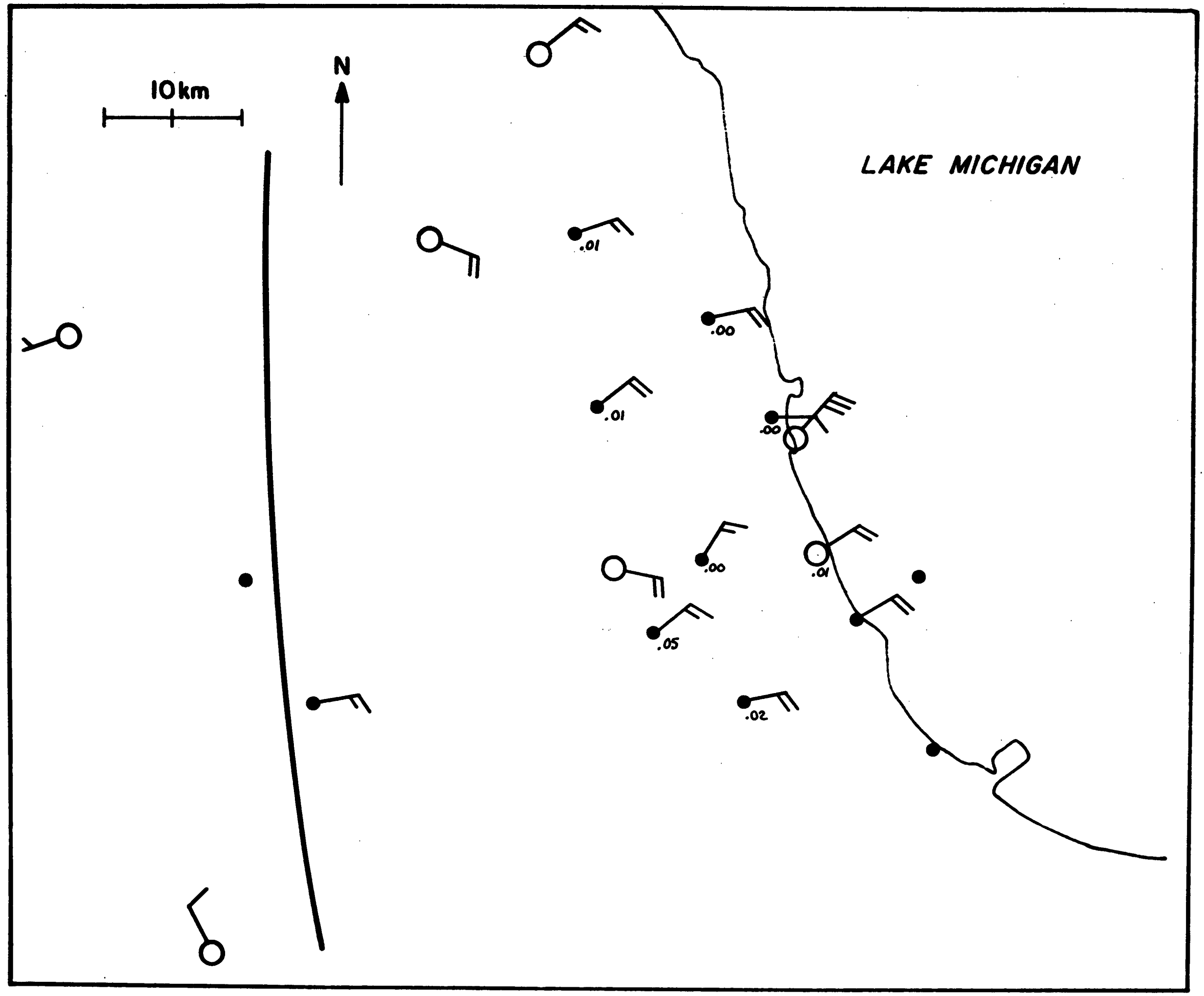

FIGURE 65. Same as Figure 56 except for 1500 CST, 12 August, 1967.

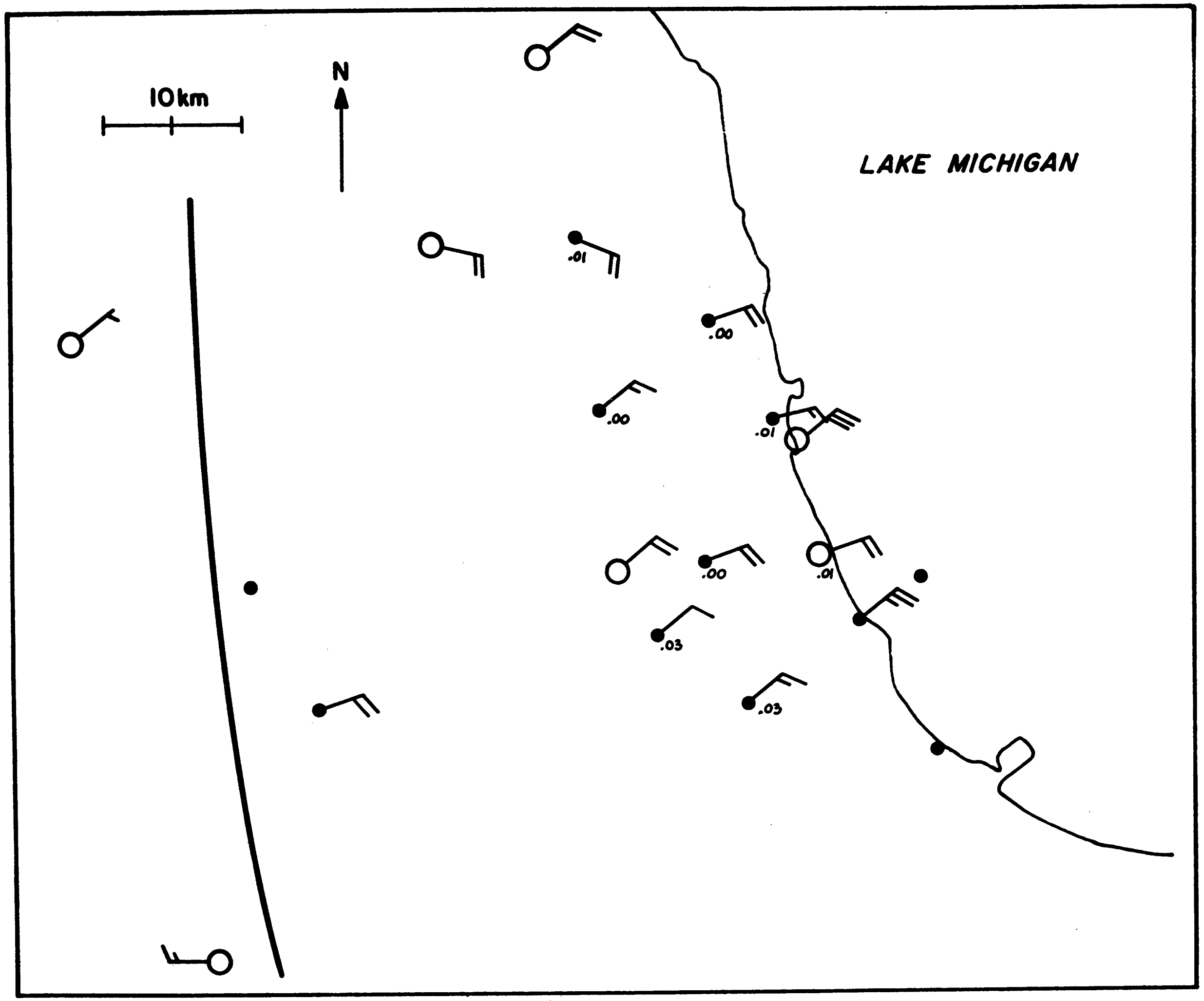

FIGURE 66. Same as Figure 56 except for 1600 CST, 12 August, 1967.

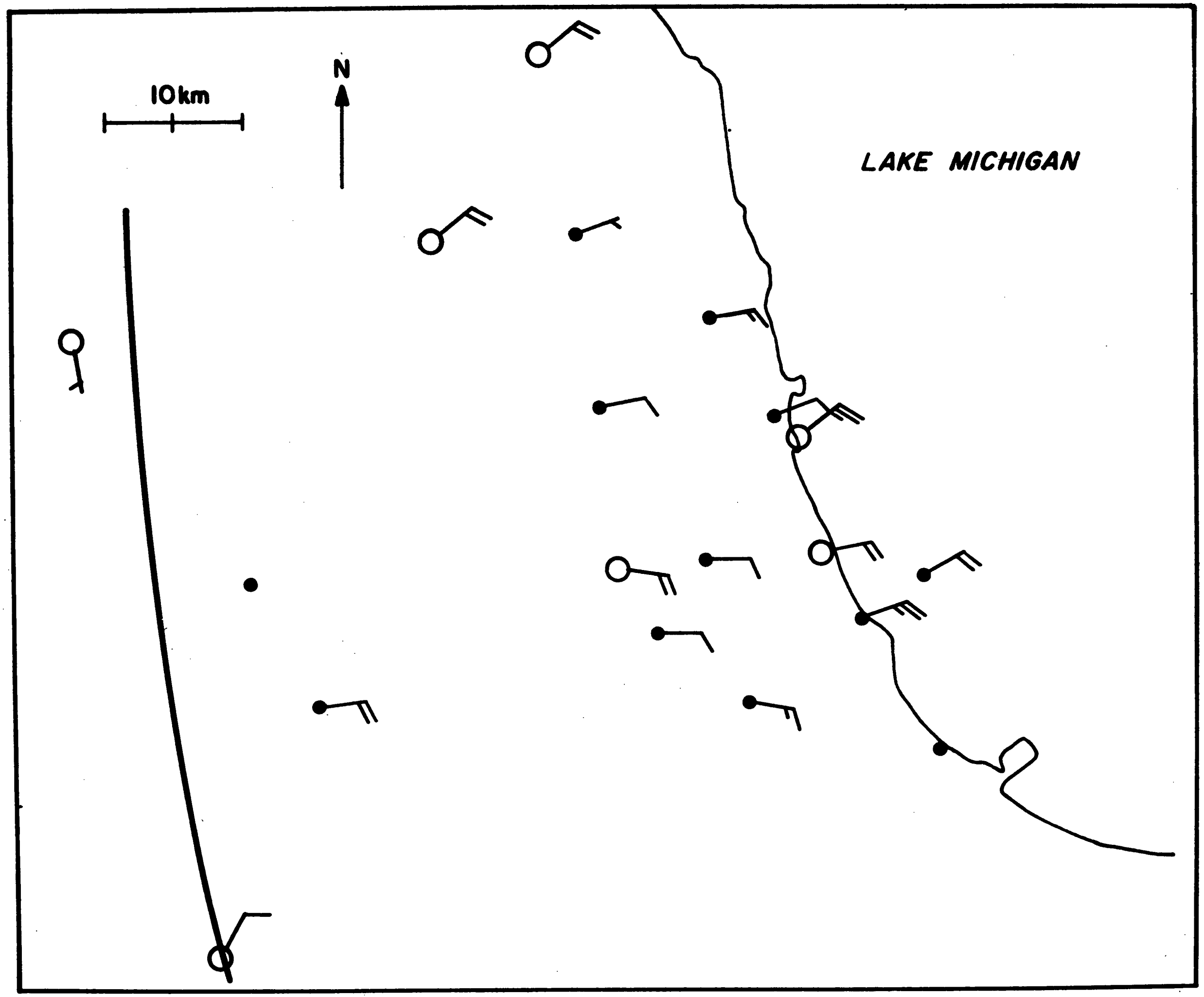

FIGURE 67. Same as Figure 56 except for 1700 CST, 12 August, 1967.

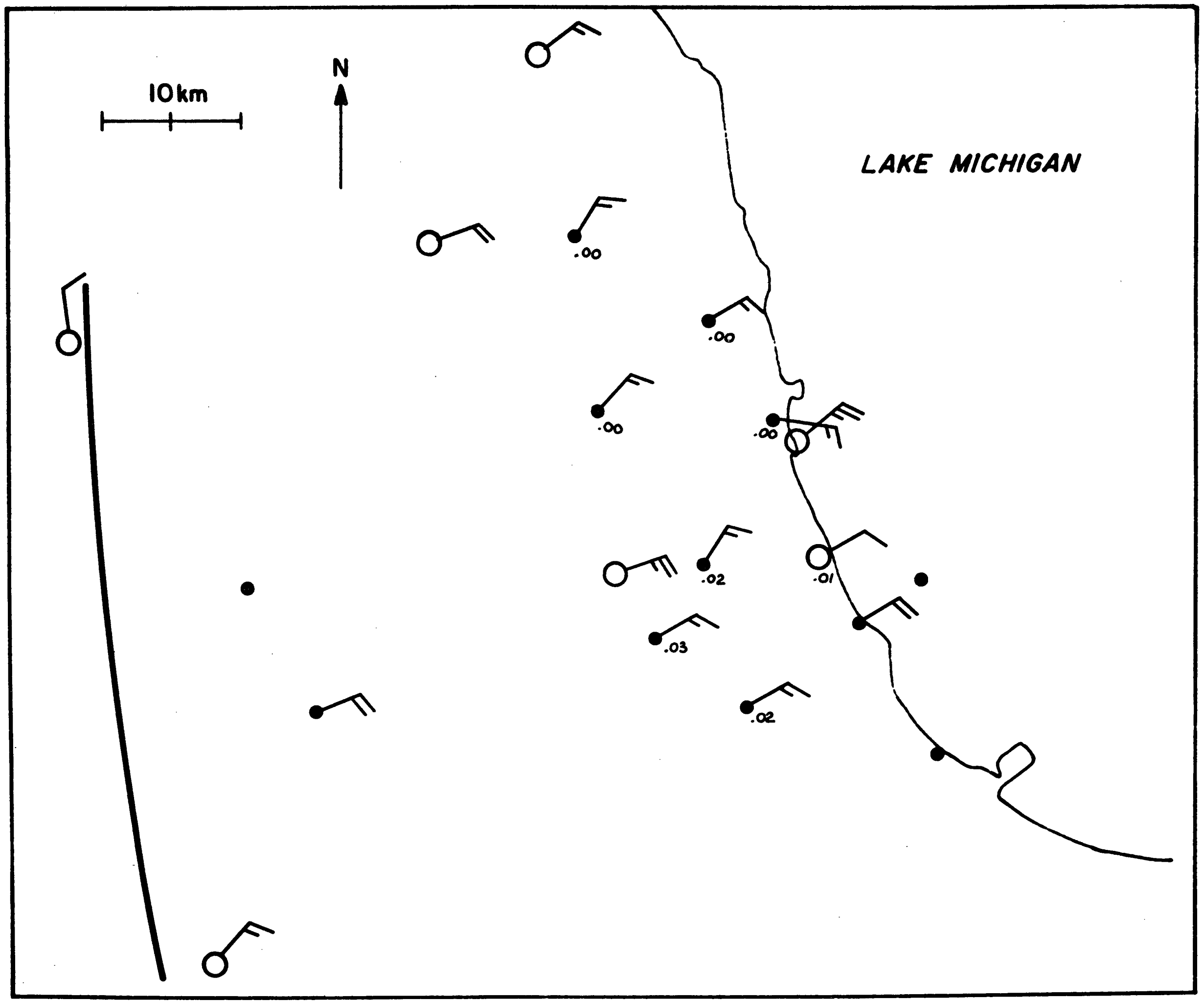

FIGURE 68. Same as Figure 56 except for 1800 CST, 12 August, 1967.

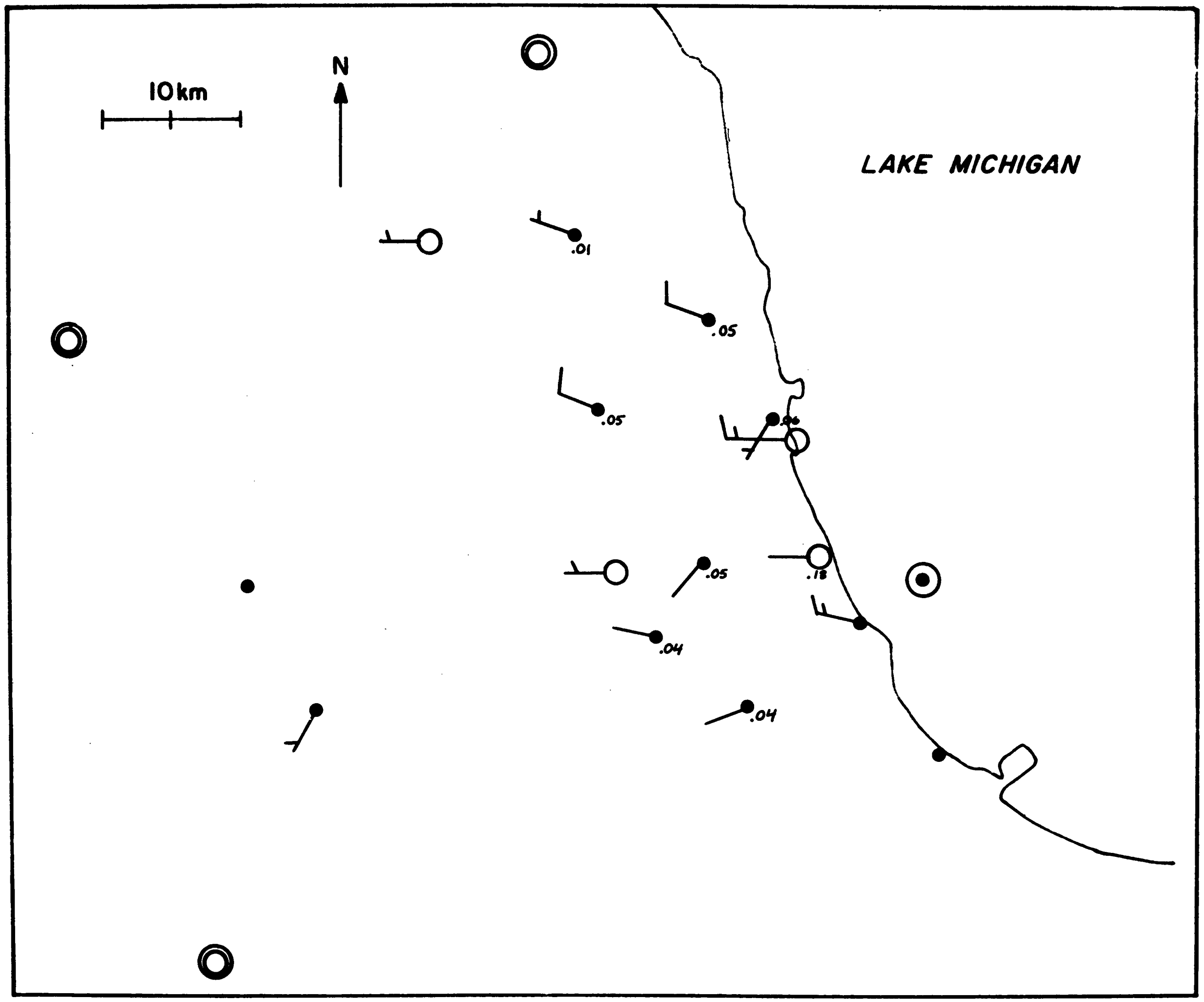

FIGURE 69. Same as Figure 56 except for 0500 CST, 13 August, 1967.

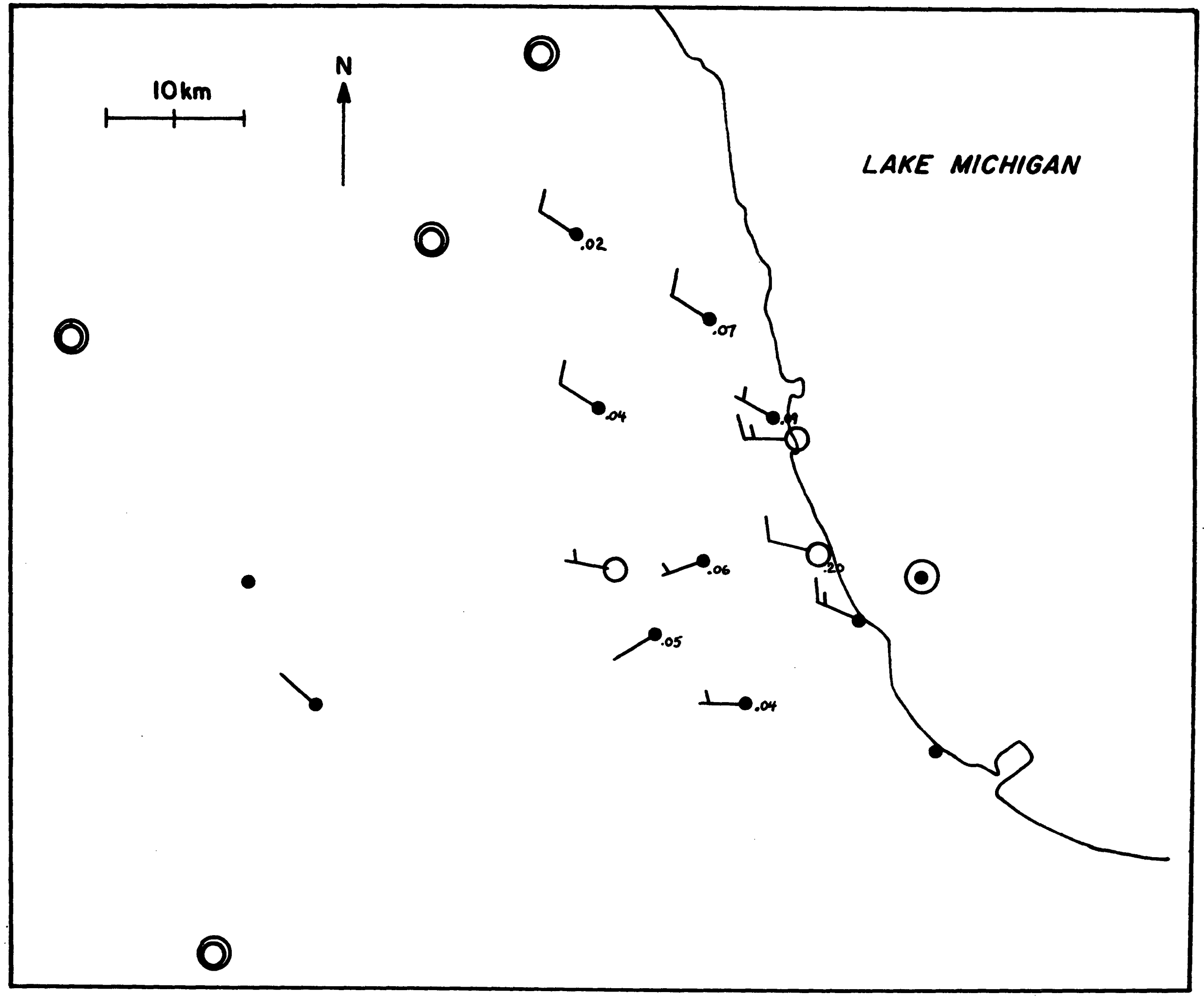

FIGURE 70. Same as Figure 56 except for 0600 CST, 13 August, 1967.

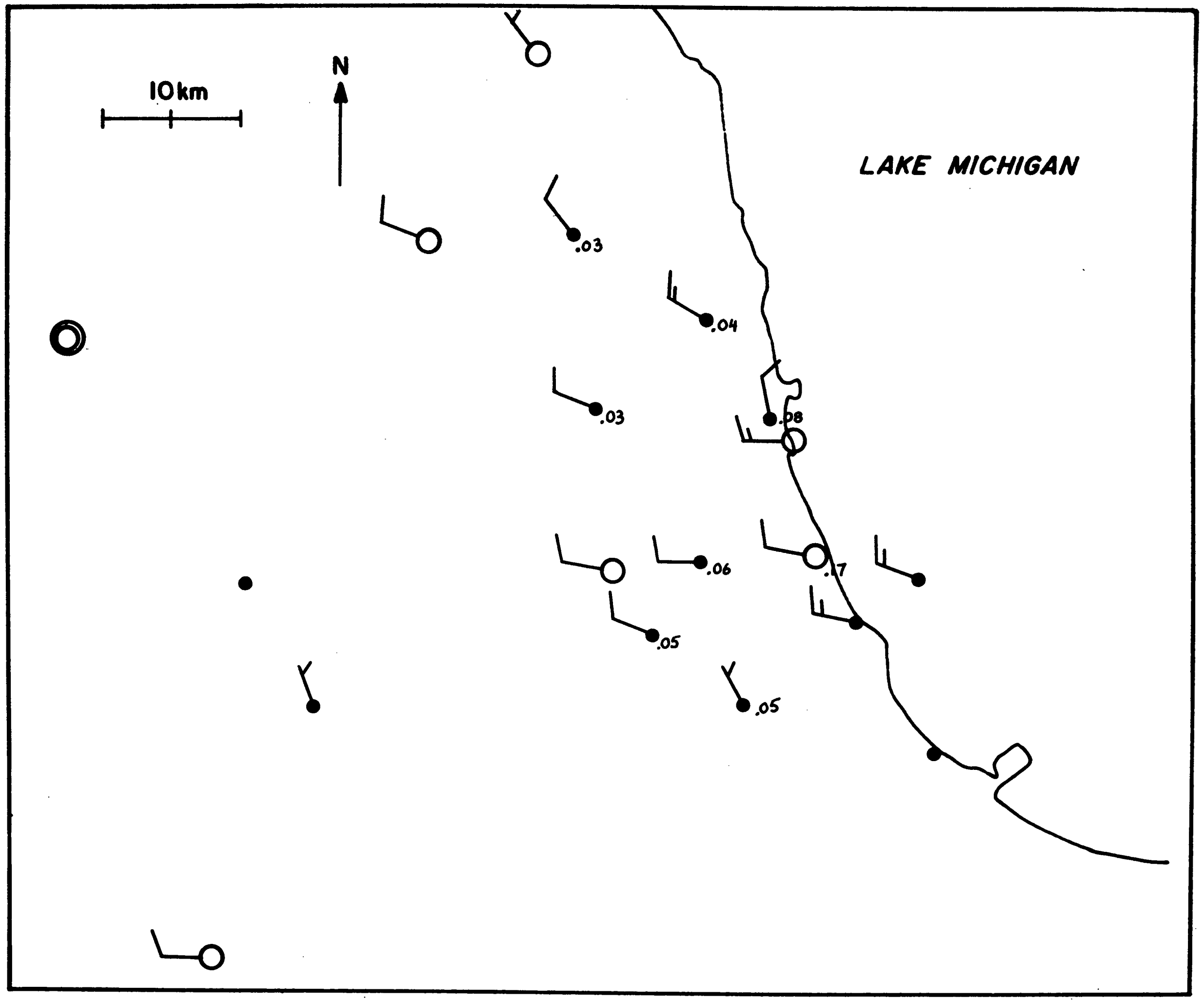

FIGURE 71. Same as Figure 56 except for 0700 CST, 13 August, 1967.

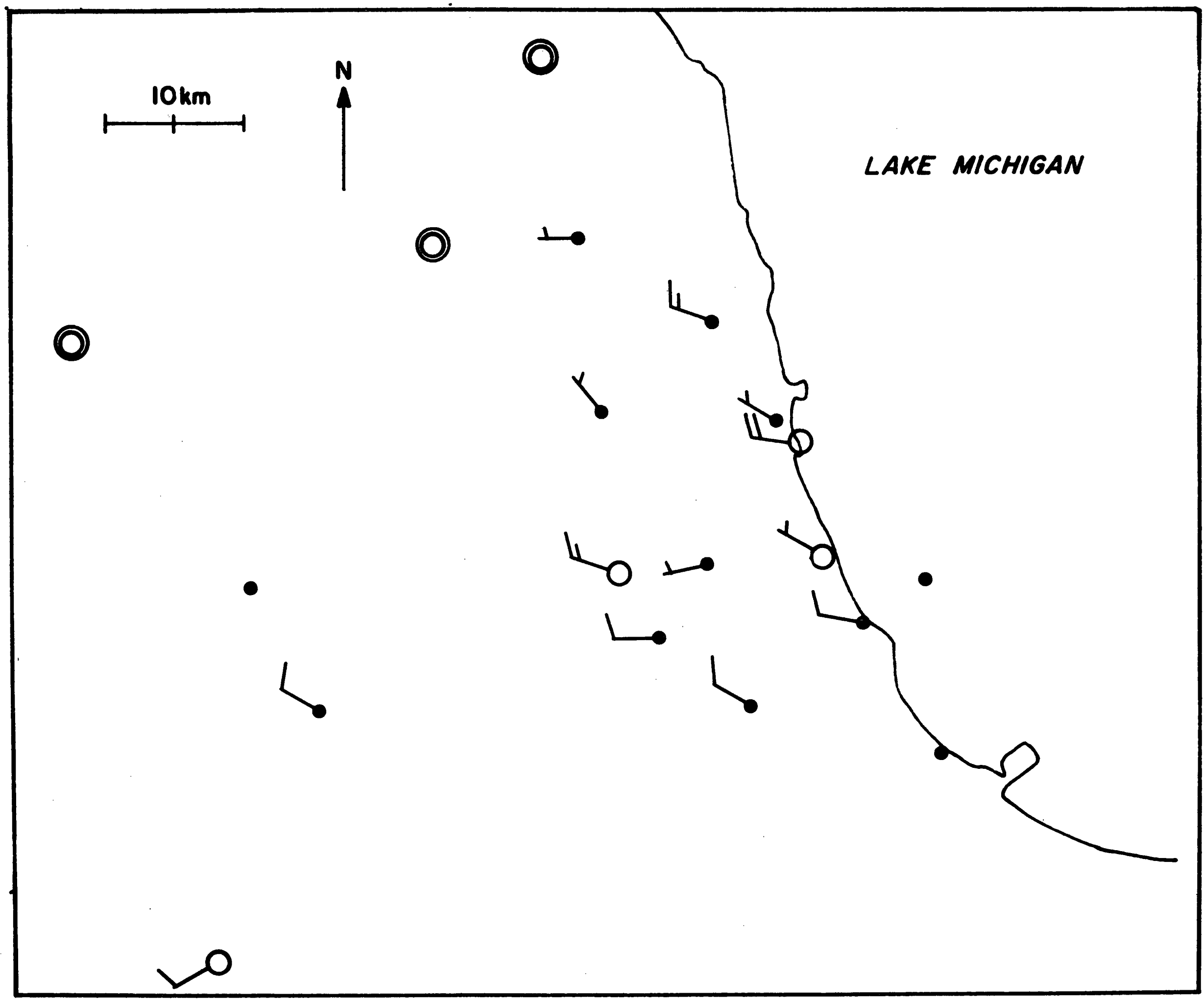

FIGURE 72. Same as Figure 56 except for 0800 CST, 13 August, 1967.

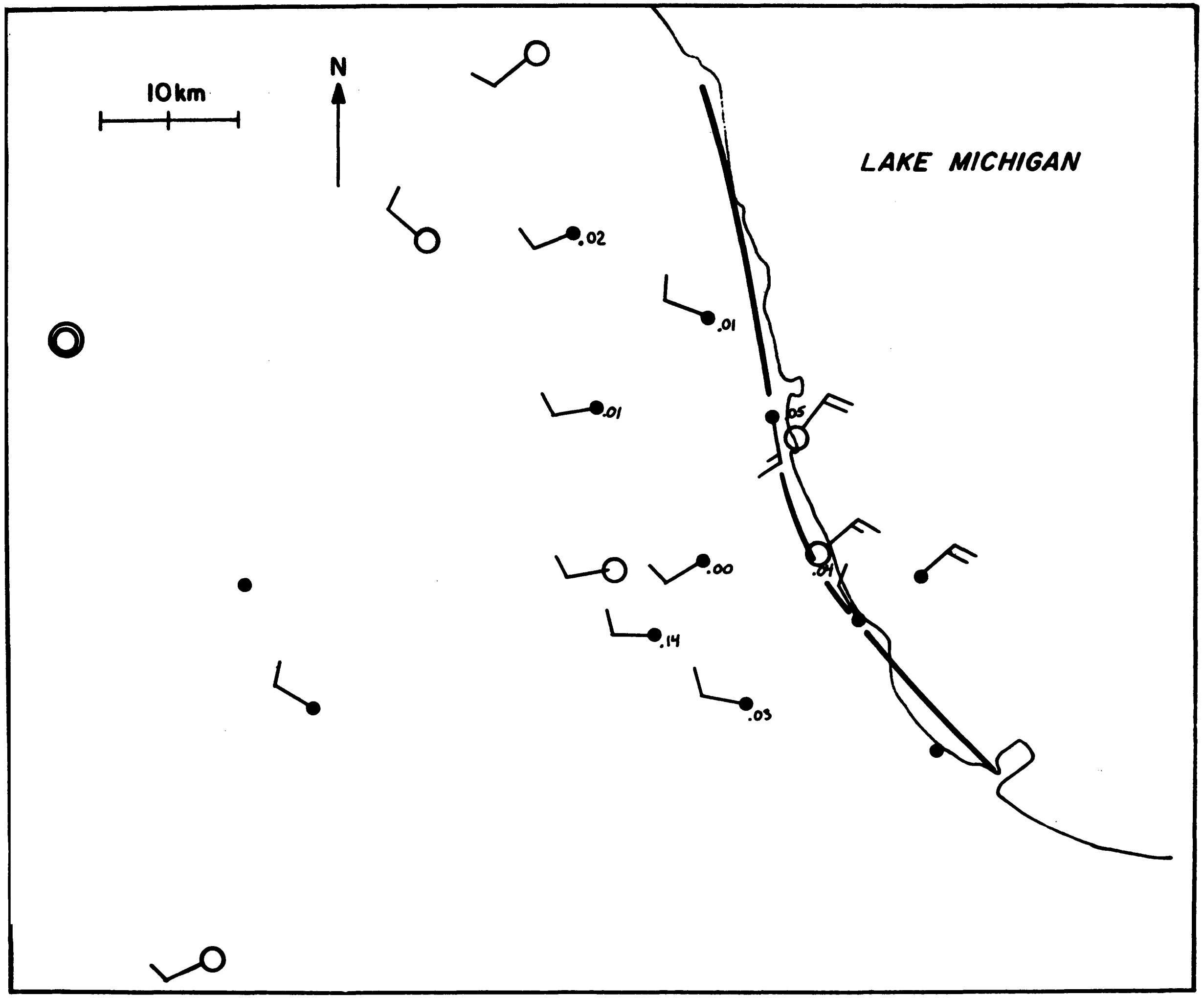

FIGURE 73. Same as Figure 56 except for 0900 CST, 13 August, 1967.

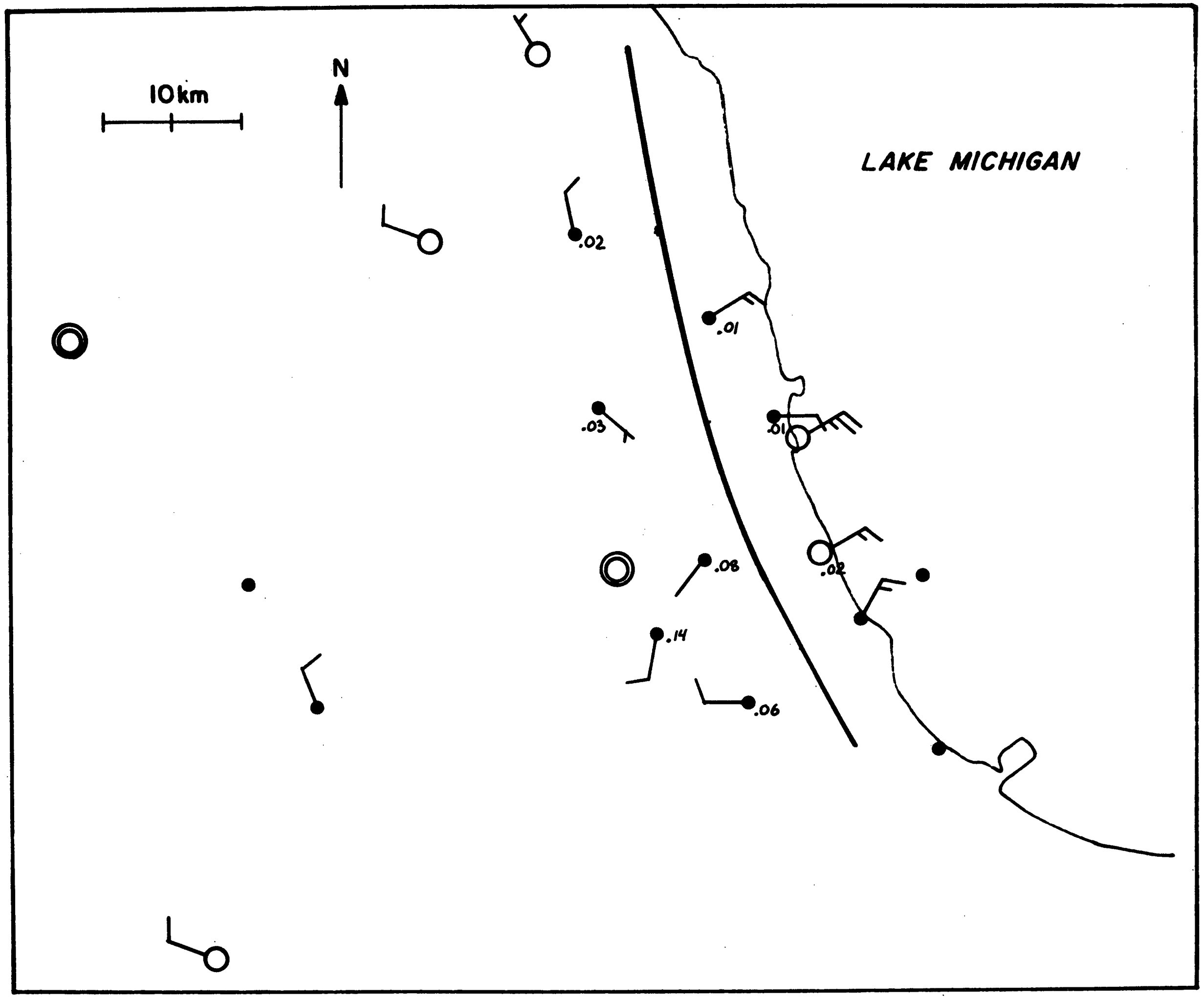

FIGURE 74. Same as Figure 56 except for 1000 CST, 13 August, 1967.

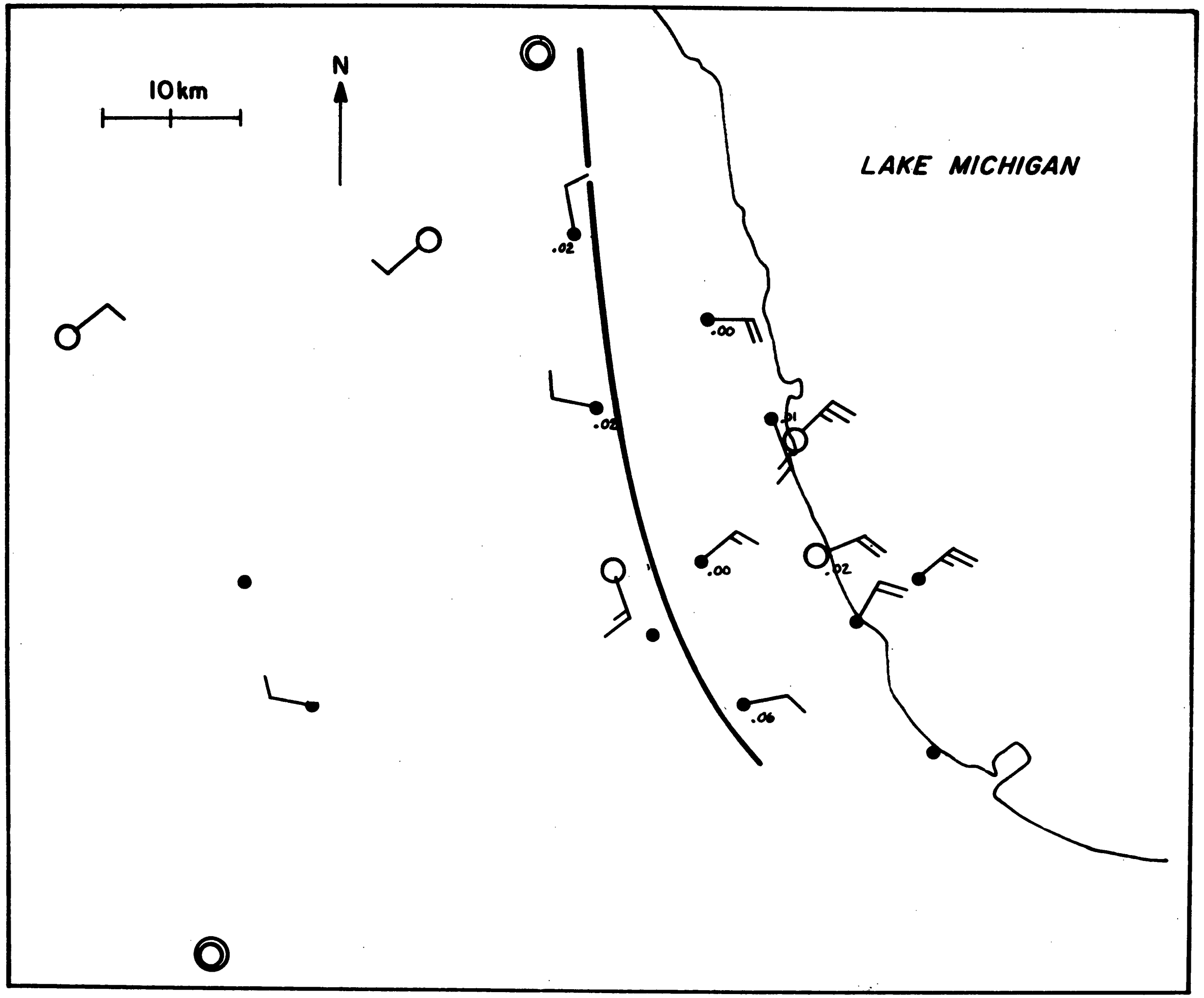

FIGURE 75. Same as Figure 56 except for 1100 CST, 13 August, 1967.

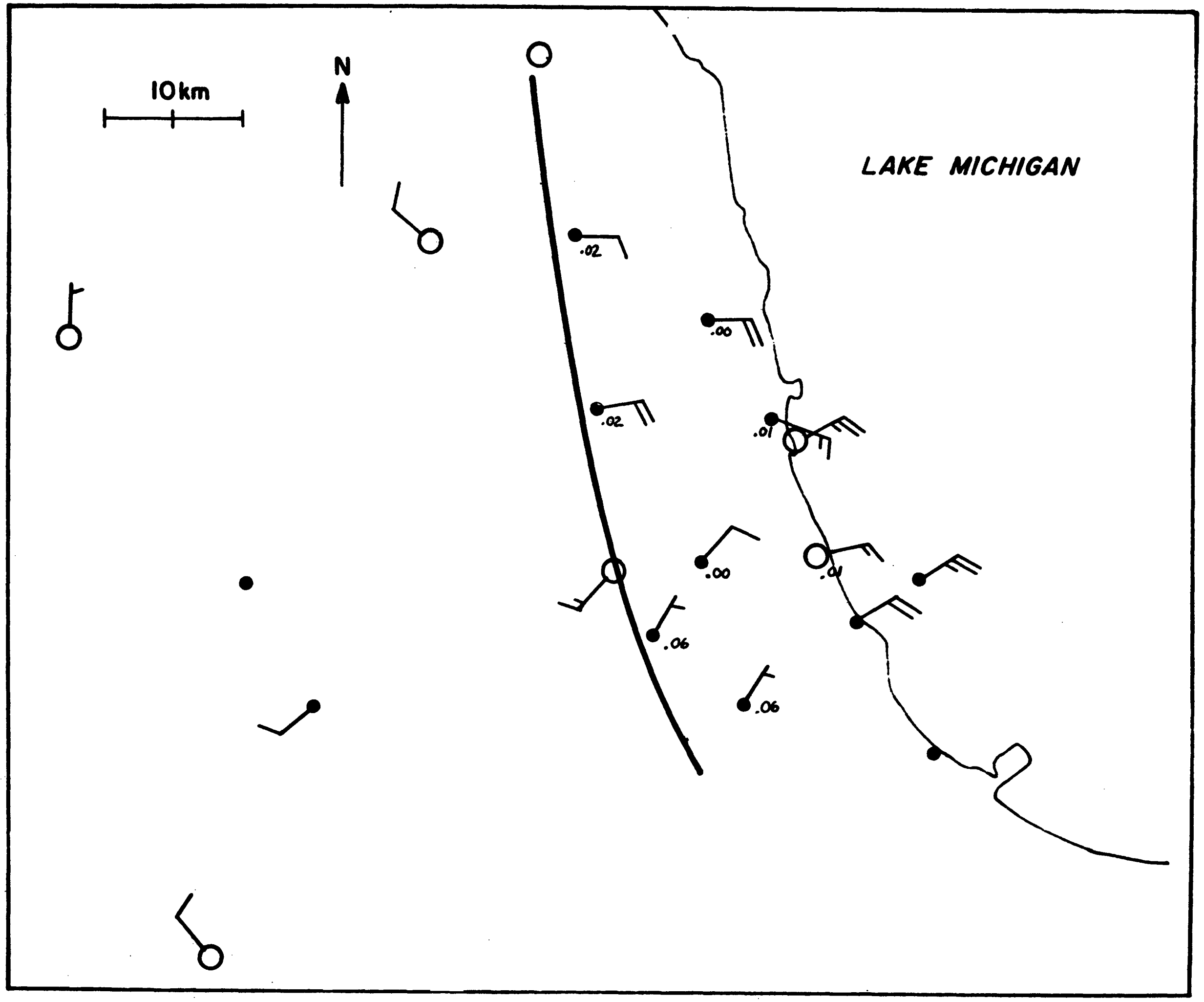

FIGURE 76. Same as Figure 56 except for 1200 CST, 13 August, 1967.

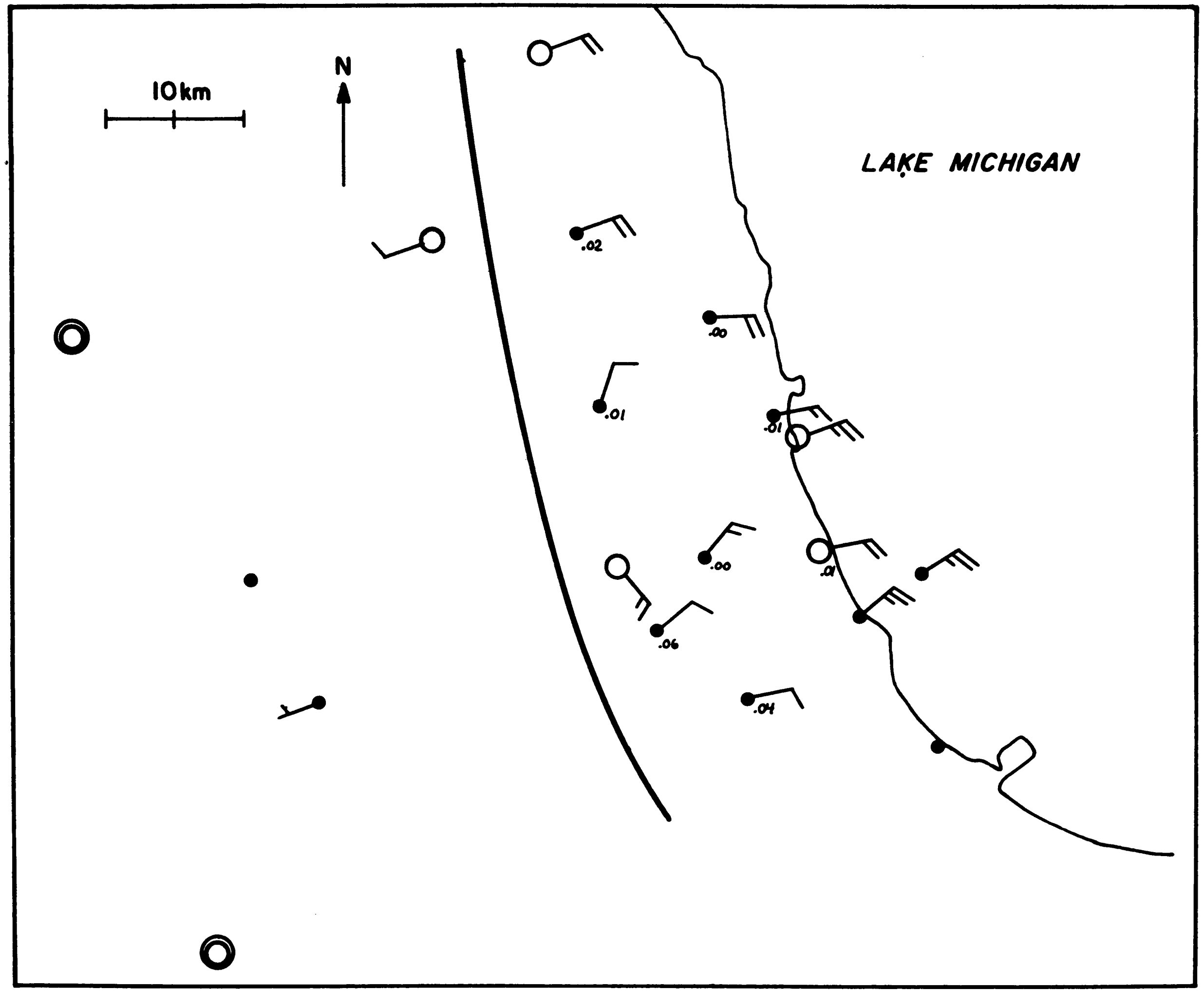

FIGURE 77. Same as Figure 56 except for 1300 CST, 13 August, 1967.

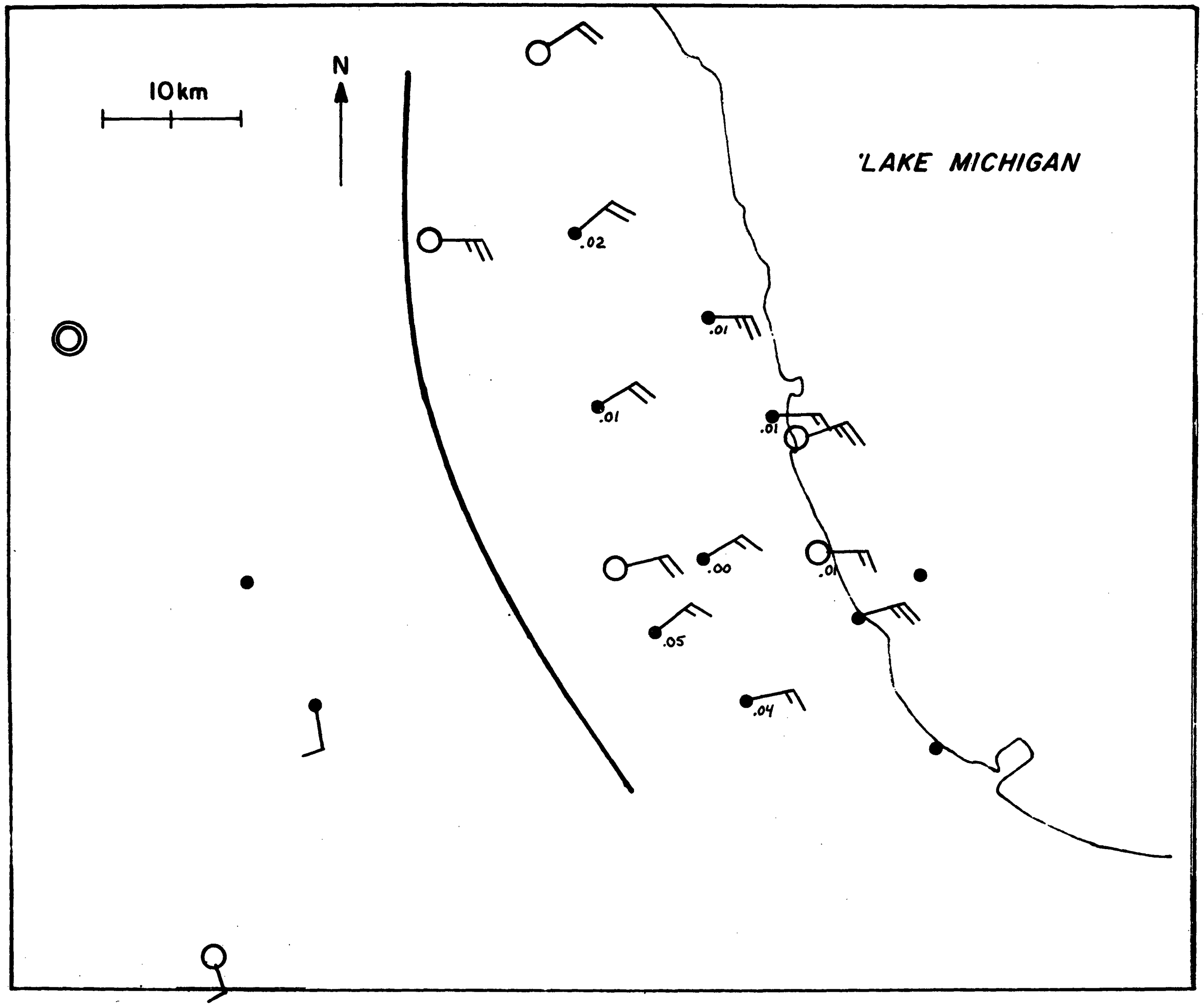

FIGURE 78. Same as Figure 56 except for 1400 CST, 13 August, 1967.

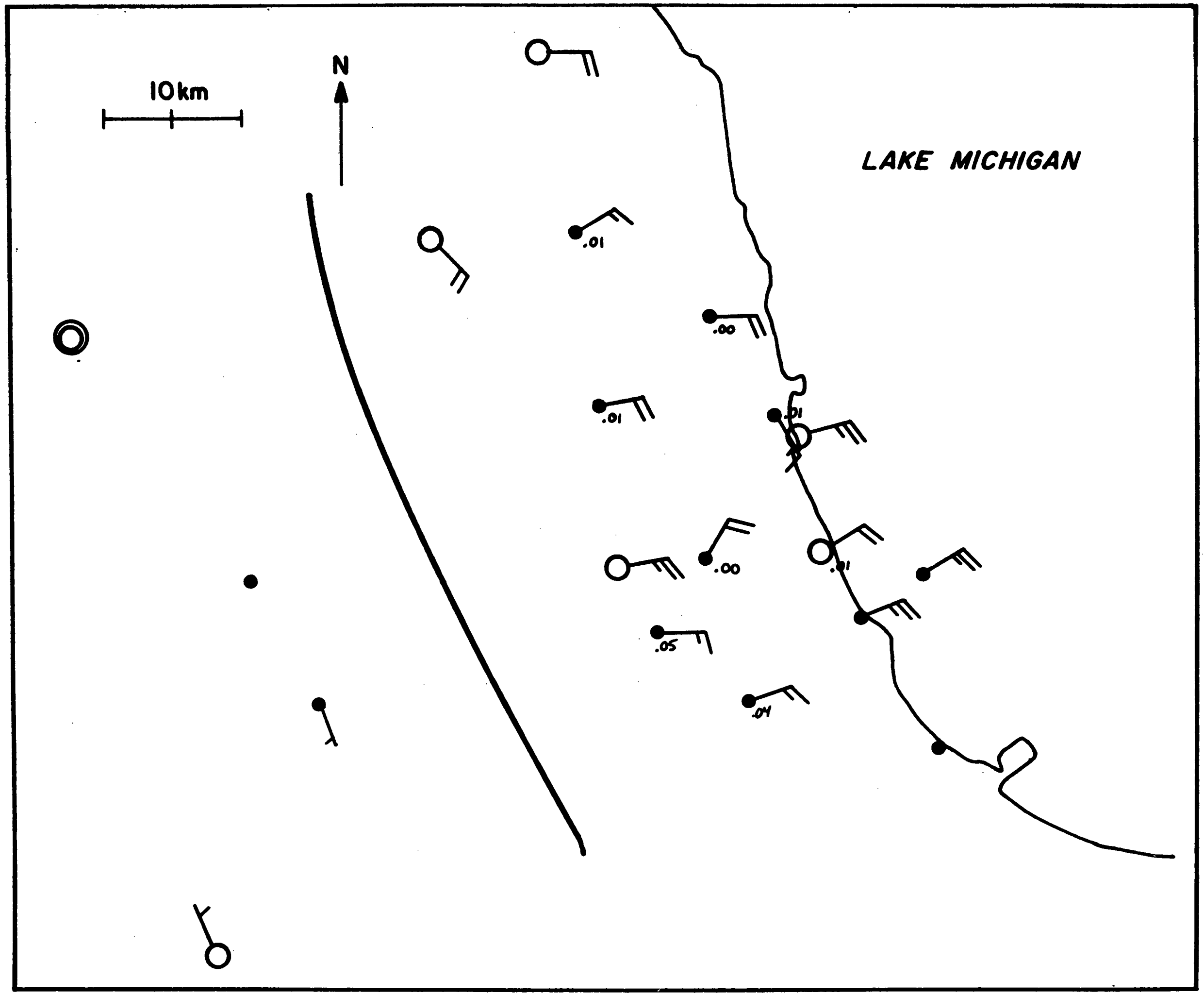

FIGURE 79. Same as Figure 56 except for 1500 CST, 13 August, 1967.

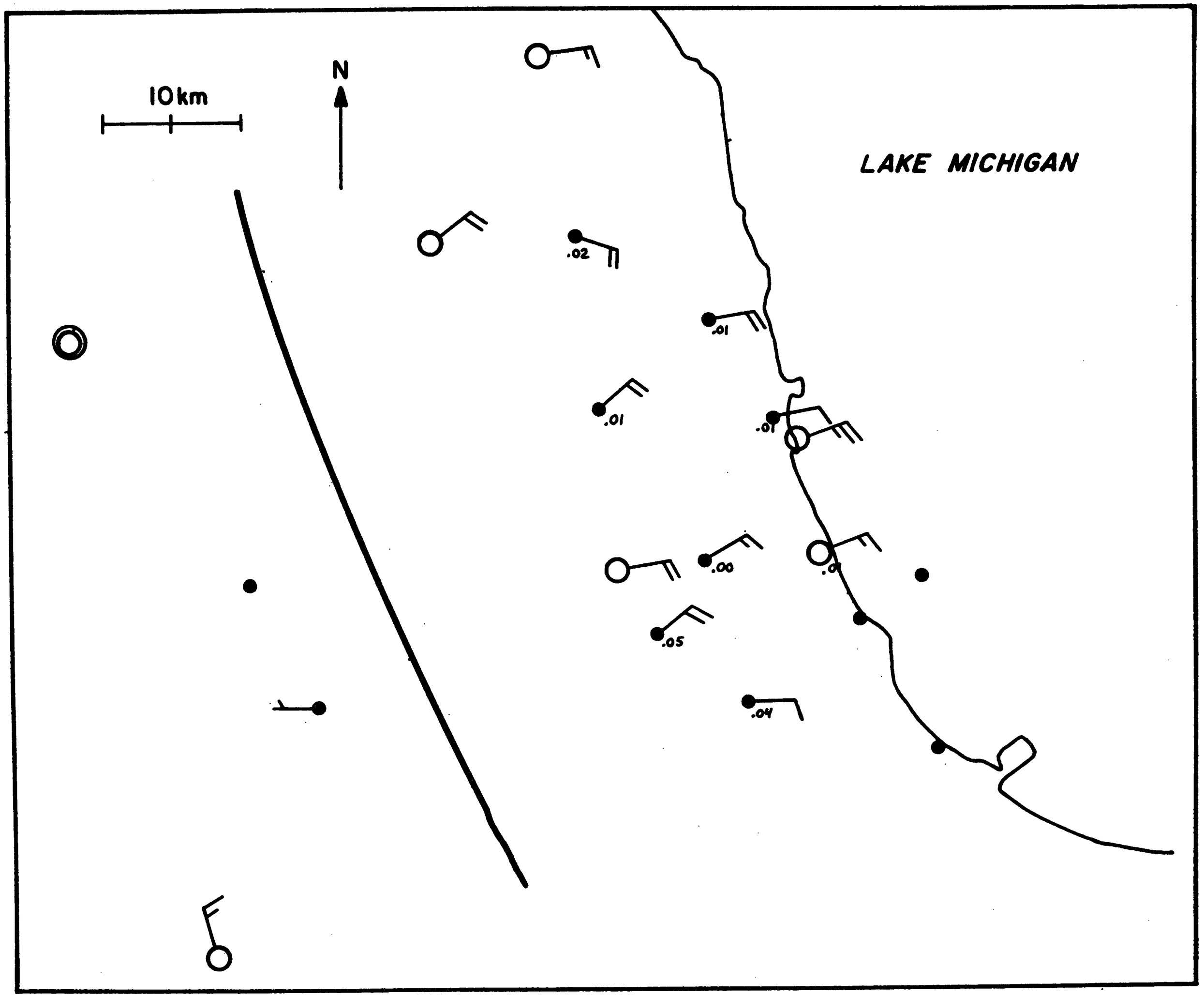

FIGURE 80. Same as Figure 56 except for 1600 CST, 13 August, 1967.

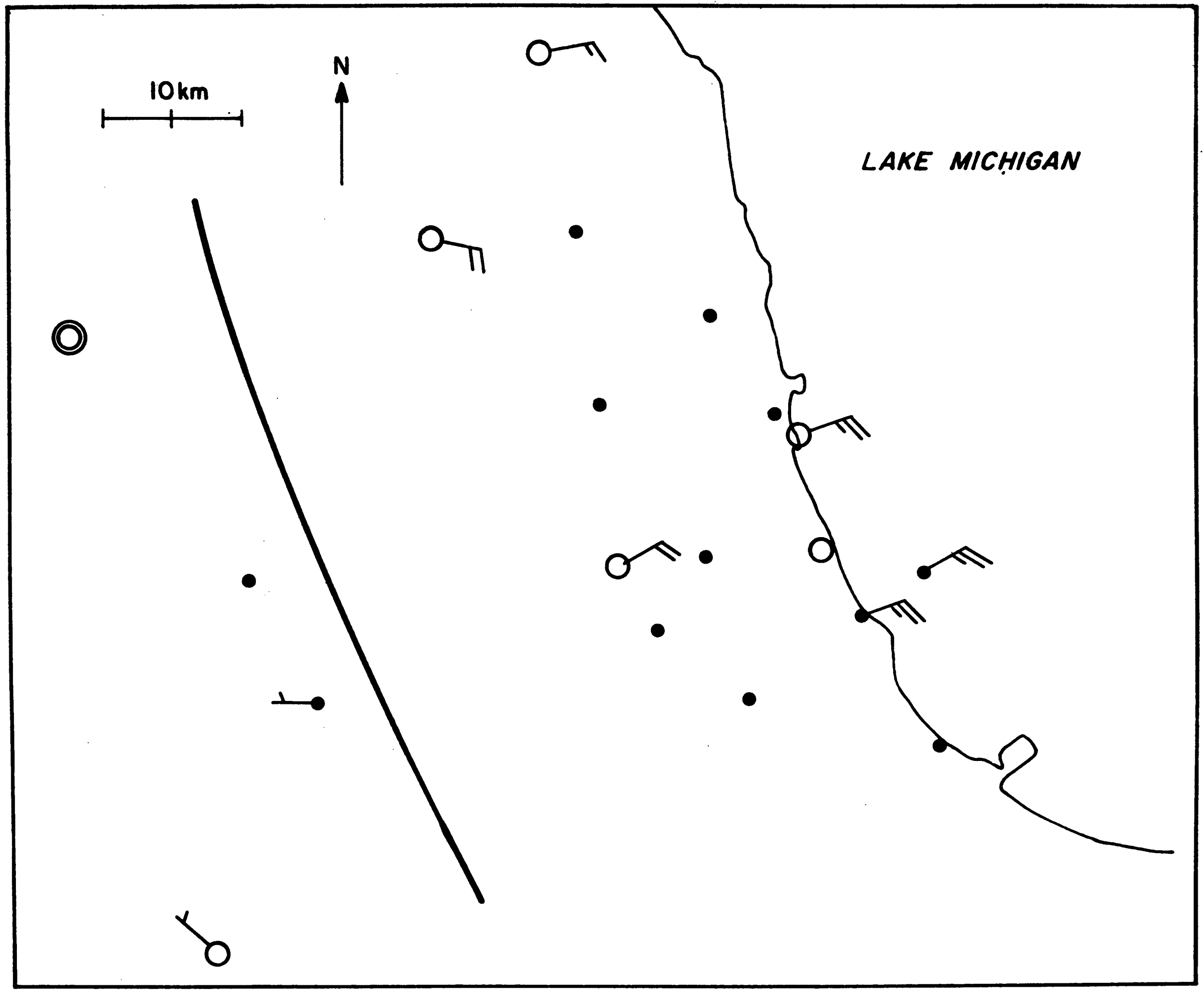

FIGURE 81. Same as Figure 56 except for 1700 CST, 13 August, 1967.

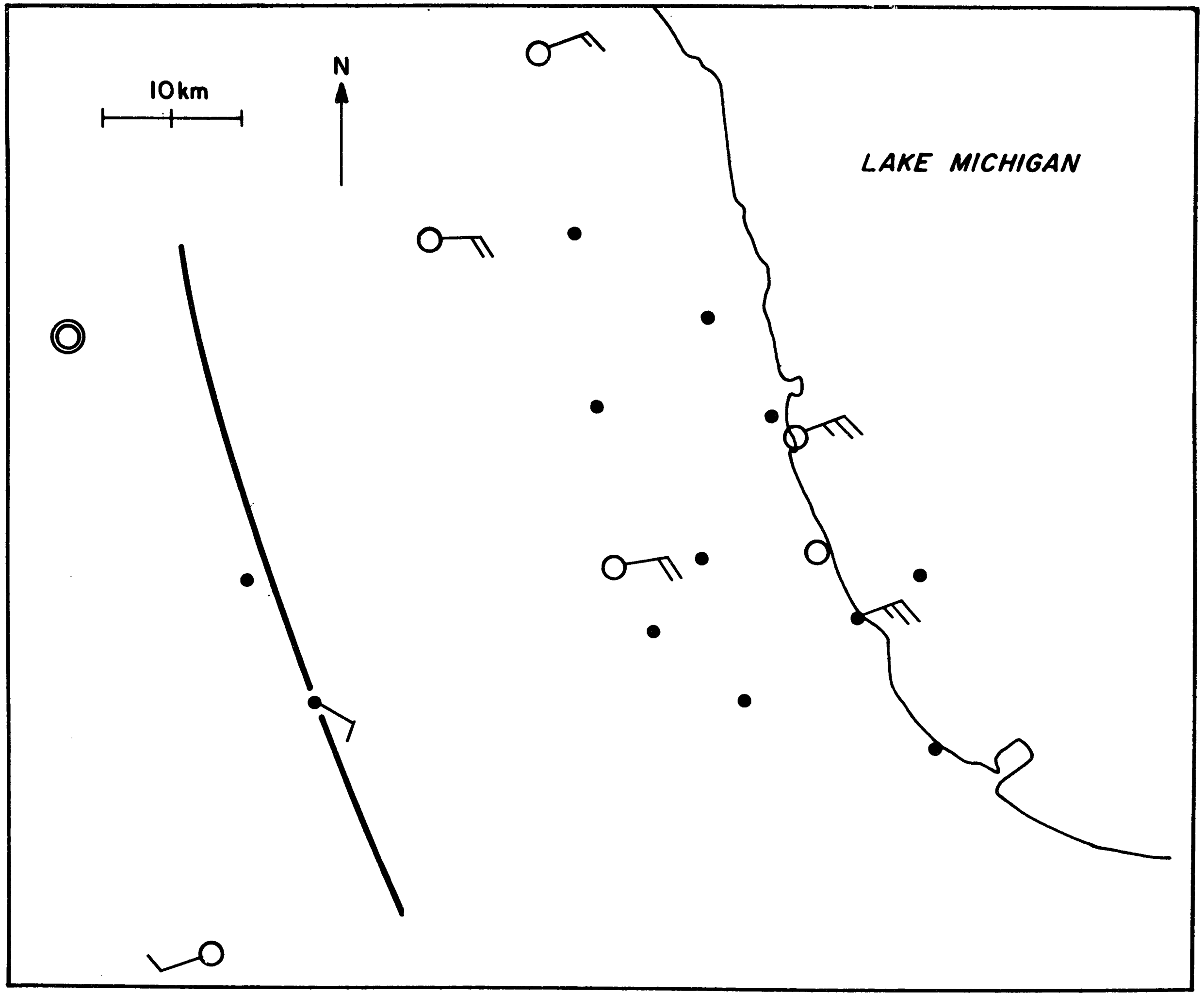

FIGURE 82. Same as Figure 56 except for 1800 CST, 13 August, 1967.

TABLE 4. A summary of available winds aloft observations on 12 and 13 August, 1967. Numbers given in the table are durations of tracks in minutes. The locations of the stations and the coordinate system are depicted in Figure 48.

Station: X/Y in km: DATE	Release time (CST)	CLS -1/0	MDW -13/6	DWN -36/18
8/12	0705	5.0		
	0733	5.5		
	0745	18.0		
	0800			8.5
	0820	19.5		
	0845	10.0		
	0900		19.5	7.0
	1000		8.5	19.5
	1100		4.5	19.5
	1200		6.5	10.0
	1300		3.5	18.5
	1400		3.5	6.0
	1450	8.0		
	1500			2.5
	1530	25.0		
8/13	0800			2.5
	0810	8.0		
	0900		10.5	19.5
	1000		8.0	19.5
	1030	5.0		
	1045	6.5		
	1100		7.5	19.5
	1200		5.0	19.5
	1300		6.5	19.5
	1400		5.5	19.5
	1500			19.5
	1505	17.0		

were made on a digital computor. The vertical profiles of the across-shore wind component, U, and the along-shore wind component, V, are presented in Figures 83-88.

On both days, a layer of weak onshore flow, was observed, above an offshore surface land breeze, in the early morning, at least near the shore. If not a pure return flow in the land breeze circulation, this onshore flow was probably a "city wind" above the conductive inversion layer and thus associated with the observed strong urban heat island over Chicago.

Although the depth of the lake breeze circulation cell was smaller on the 13th than on the 12th, several characteristics of the flow patterns were similar. Offshore flows aloft were observed before the surface lake breezes and the circulations intensified, as the lake breeze fronts penetrated inland. In the mid-afternoons, near the shore, onshore lake breezes of more than 5 m sec^{-1} were measured at approximately 200 m aloft with offshore return flows in a layer above the lake breezes.

3.3.3 Tetroon Trajectories.

The analysed tetroon trajectories, describing the flights made on 12 and 13 August, 1967, are presented in Figures 89-94. Average horizontal and vertical winds, derived from the trajectories, are presented in Table 5. Except for the regions near and in the convergence zones, there seemed to be relatively little turbulence, as the instantaneous winds were close to the mean winds at all times. Vertical up and downdrafts in excess of 100 cm sec^{-1} were measured in the convergence zone and subsidence velocity over the lake of 70 cm sec^{-1} was observed on 12 August.

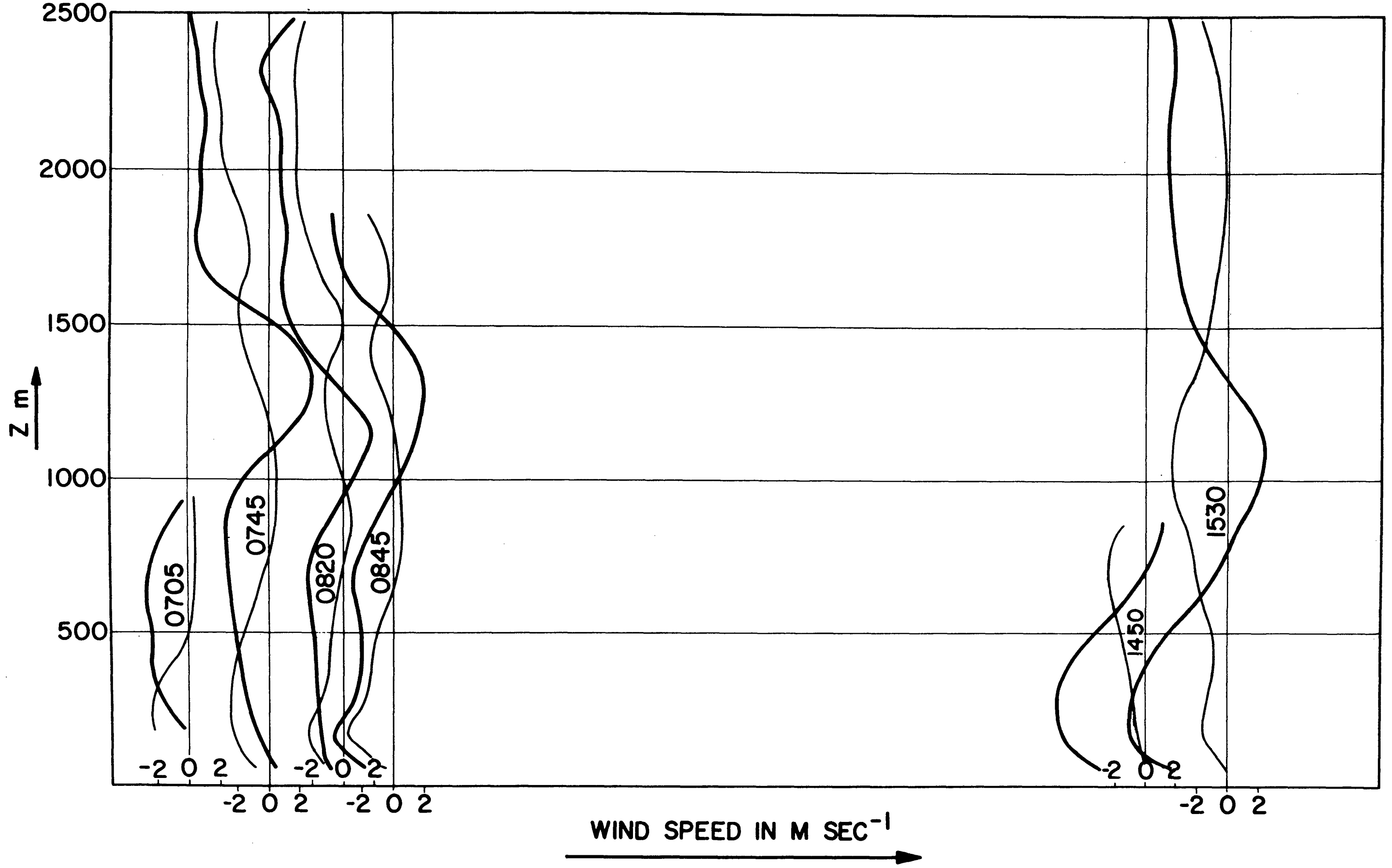

FIGURE 83. Winds aloft at the University of Chicago Campus (CLS) on 12 August, 1967. Heavy lines are profiles of the across shore components, U, positive off shore along the positive x-axis and the thin lines are profiles of the along shore components, V, positive toward NNW along the positive y-axis. Indicated times for pibal releases are in CST.

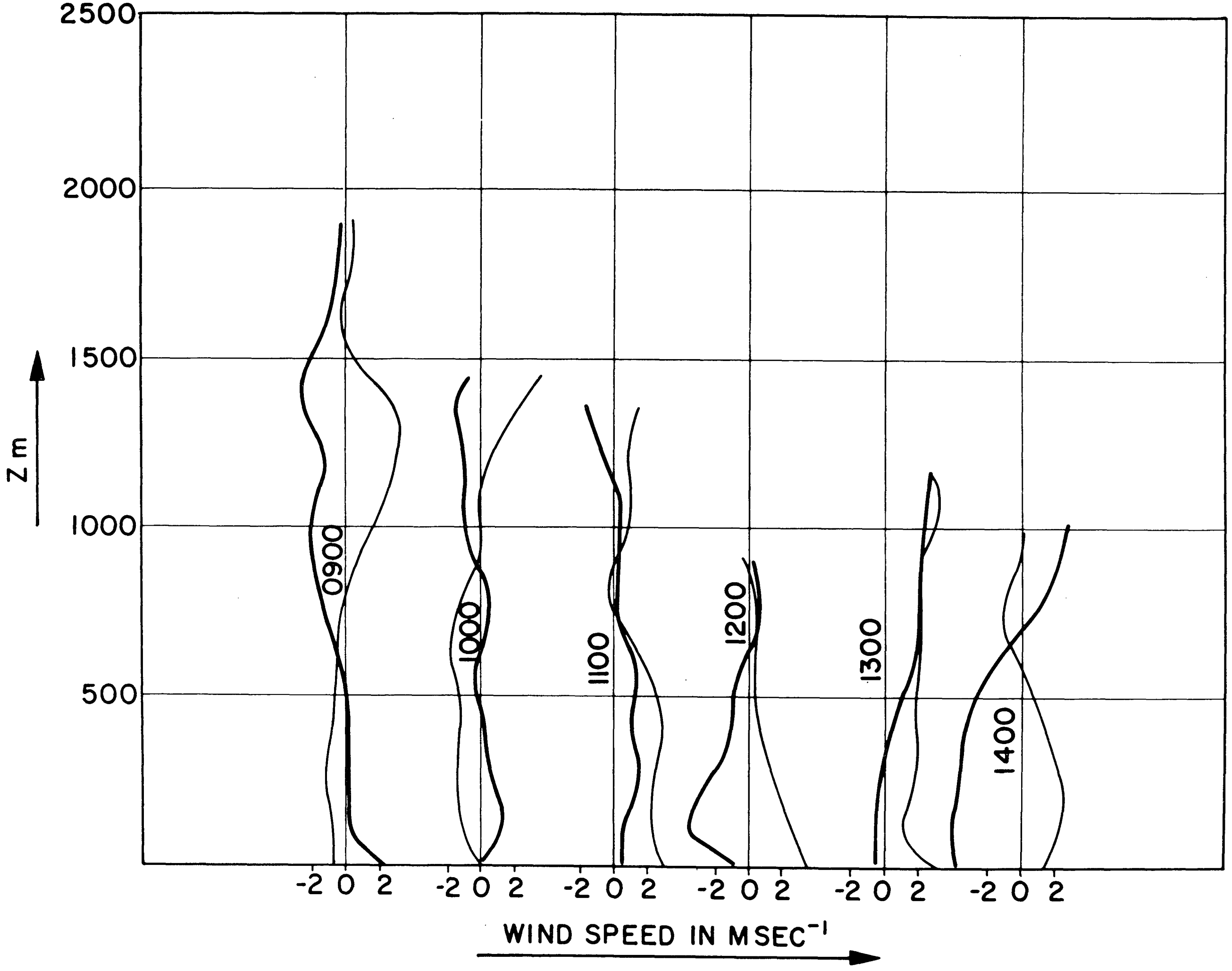

FIGURE 84. Winds aloft at Midway (MDW) on 12 August, 1967. Conventions as in Figure 83.

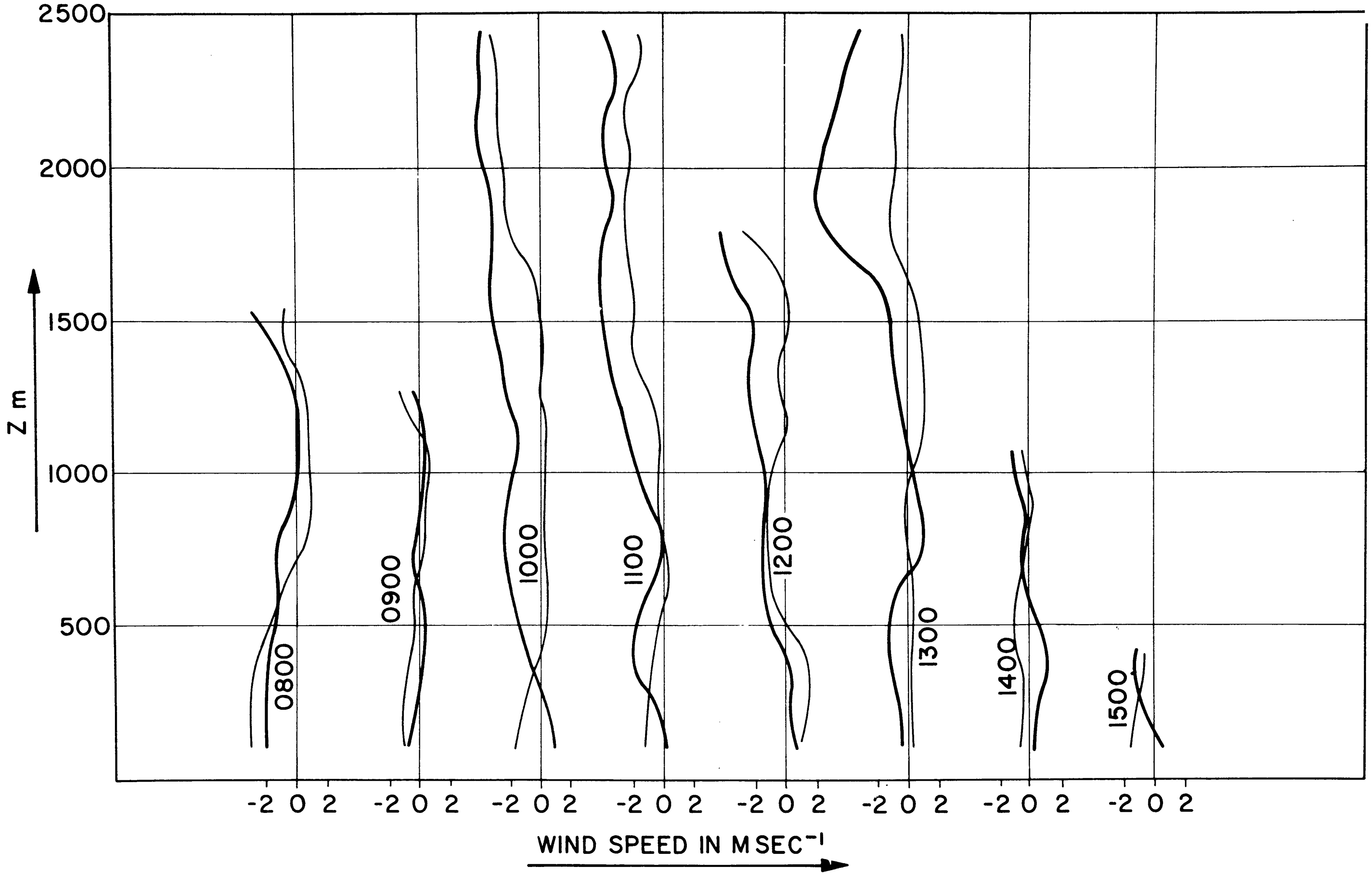

FIGURE 85. Winds aloft at Downers Grove (DWN) on 12 August, 1967. Conventions as in Figure 83.

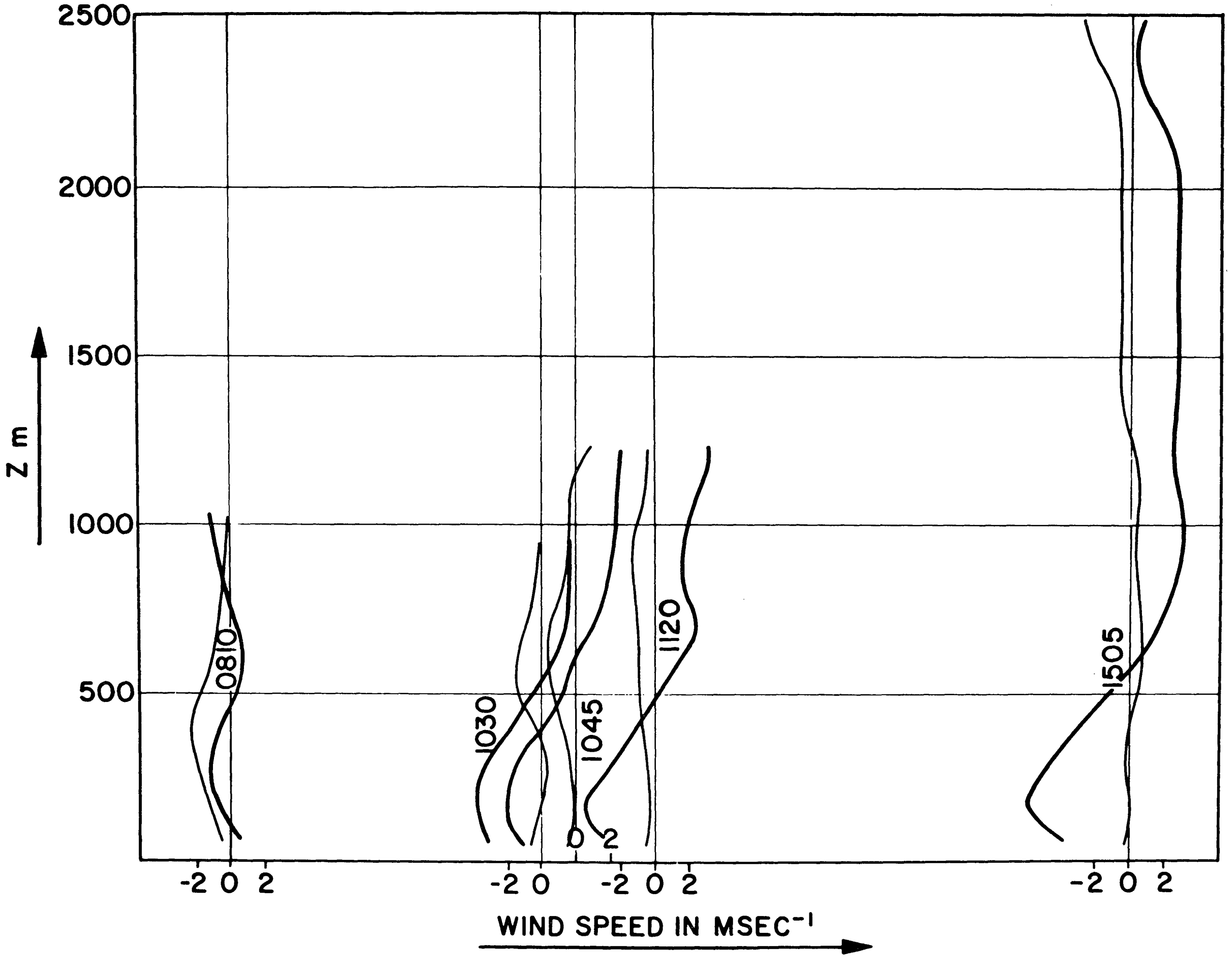

FIGURE 86. Winds aloft at the University of Chicago Campus (CLS) on 13 Aguust, 1967. Conventions as in Figure 83.

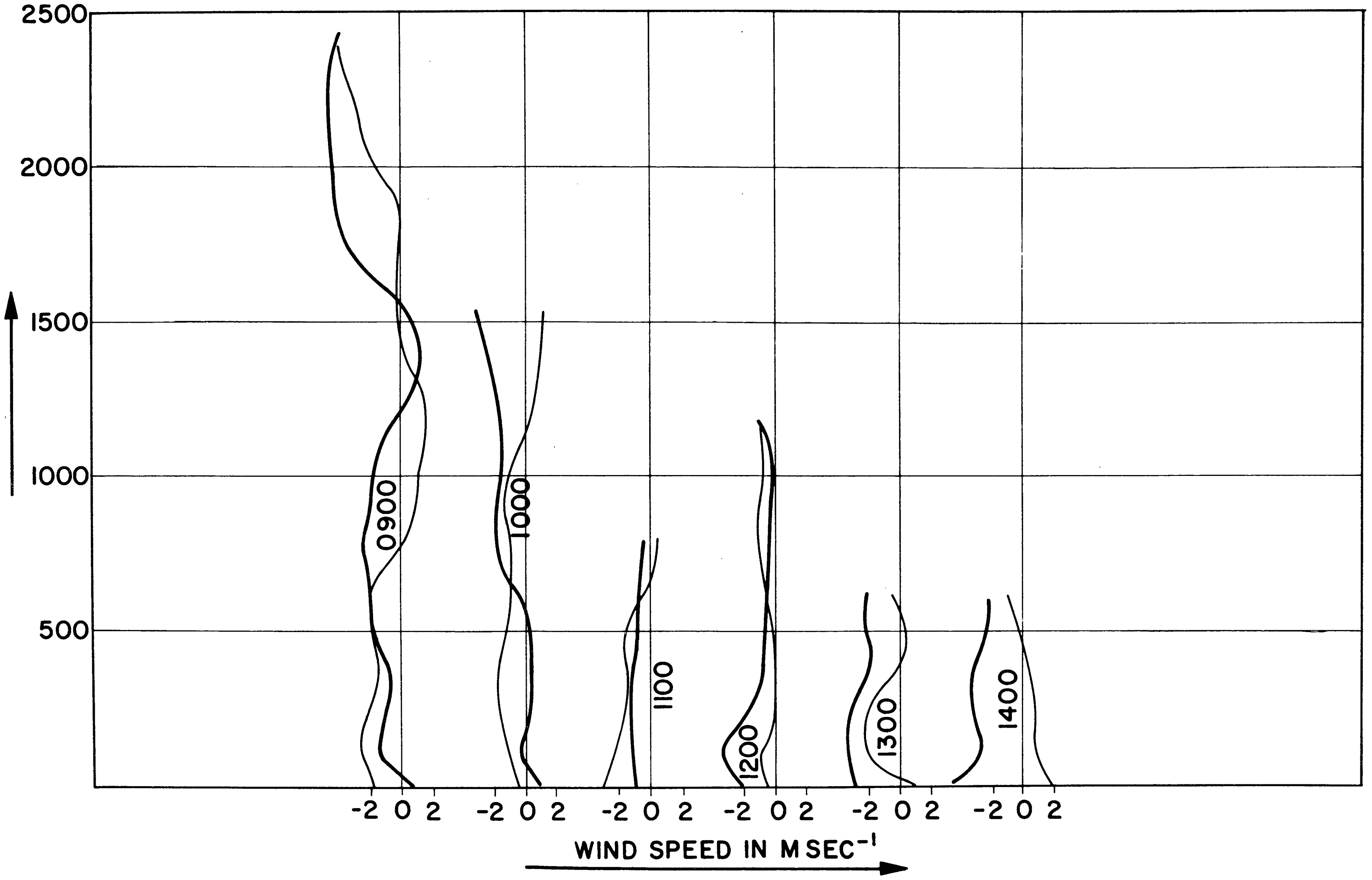

FIGURE 87. Winds aloft at Midway (MDW) on 13 August, 1967. Conventions as in Figure 83.

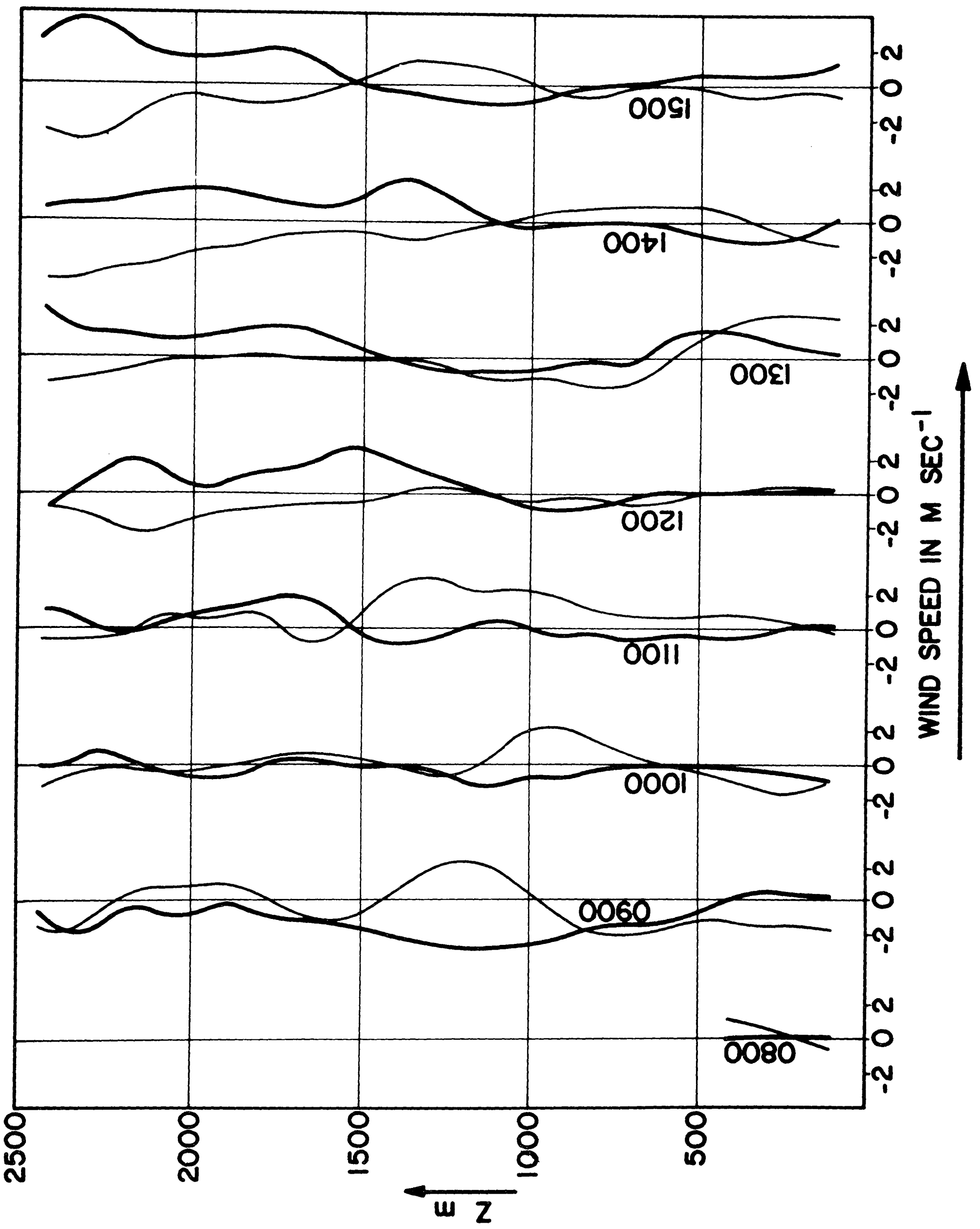

FIGURE 88. Winds aloft at Downers Grove (DWN) on 13 August, 1967. Conventions as in Figure 83.

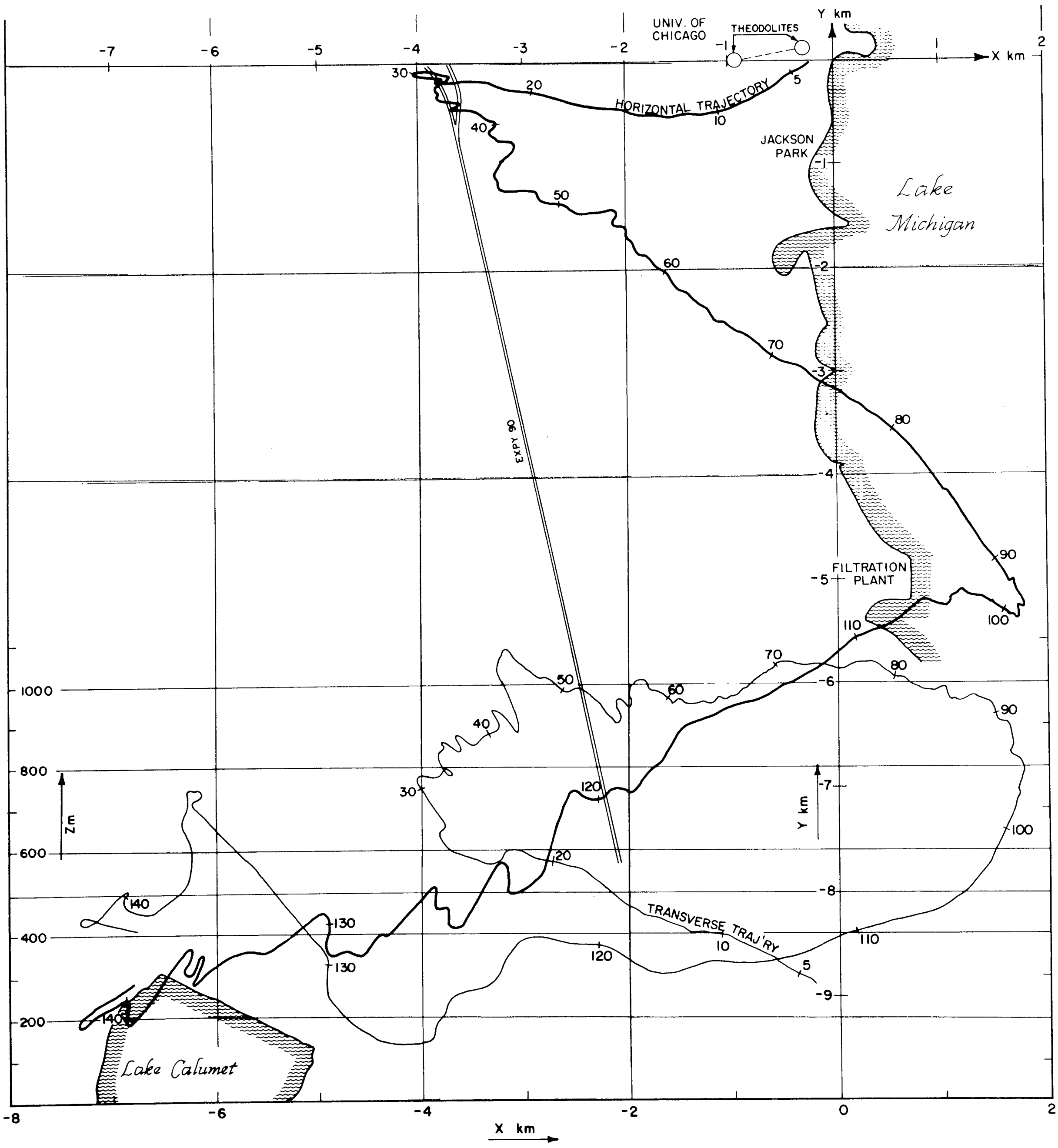

FIGURE 89. Horizontal and transverse trajectories of tetroon flight over Chicago, Illinois. The tetroon was launched at 0900 CST, 12 August, 1967. Elapsed flight times in minutes are indicated along the trajectories.

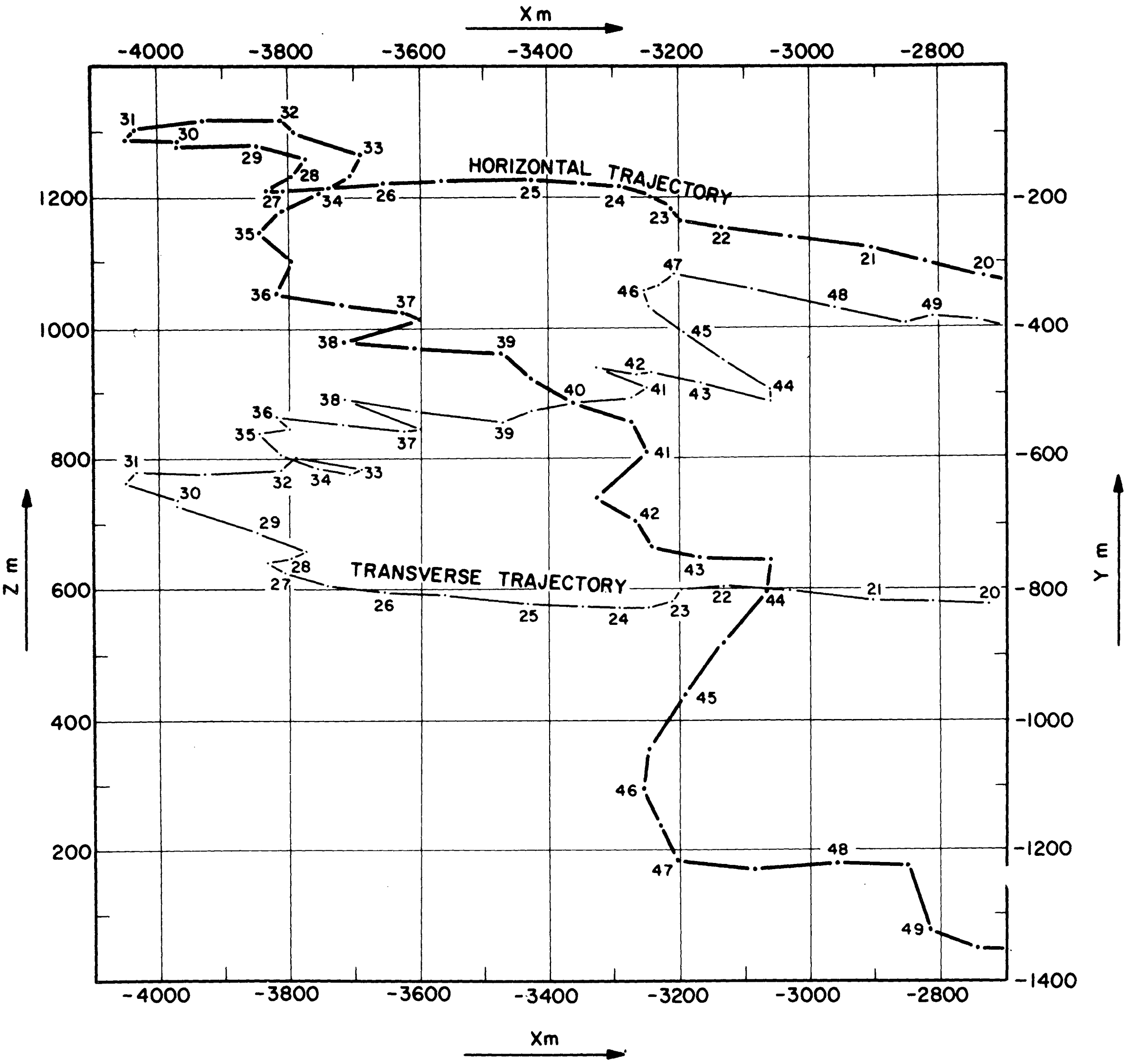

FIGURE 90. Horizontal and transverse trajectories of tetroon flight through the lake breeze convergence zone. The tetroon was launched at 0900 CST, 12 August, 1967 and its complete trajectory depicted in Figure 89. Tetroon positions are indicated every 1/2 minute and elapsed flight times in minutes are given.

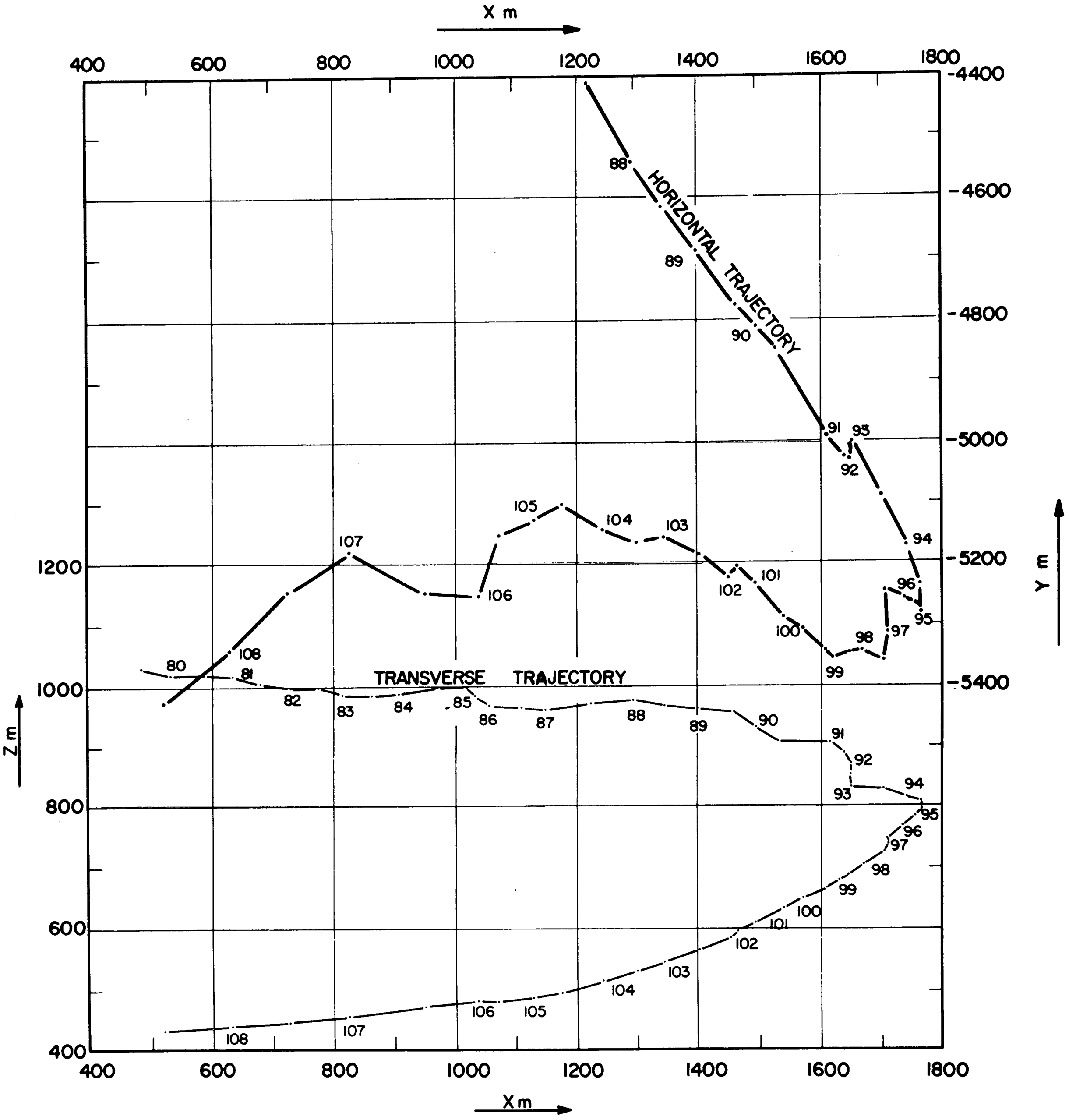

FIGURE 91. Horizontal and transverse trajectories of tetroon flight through off shore subsidence regime of the lake breeze circulation. The tetroon was launched at 0900 CST, 12 August, 1967 and its complete trajectory depicted in Figure 89. Tetroon positions are indicated every 1/2 minute and elapsed flight times in minutes are given.

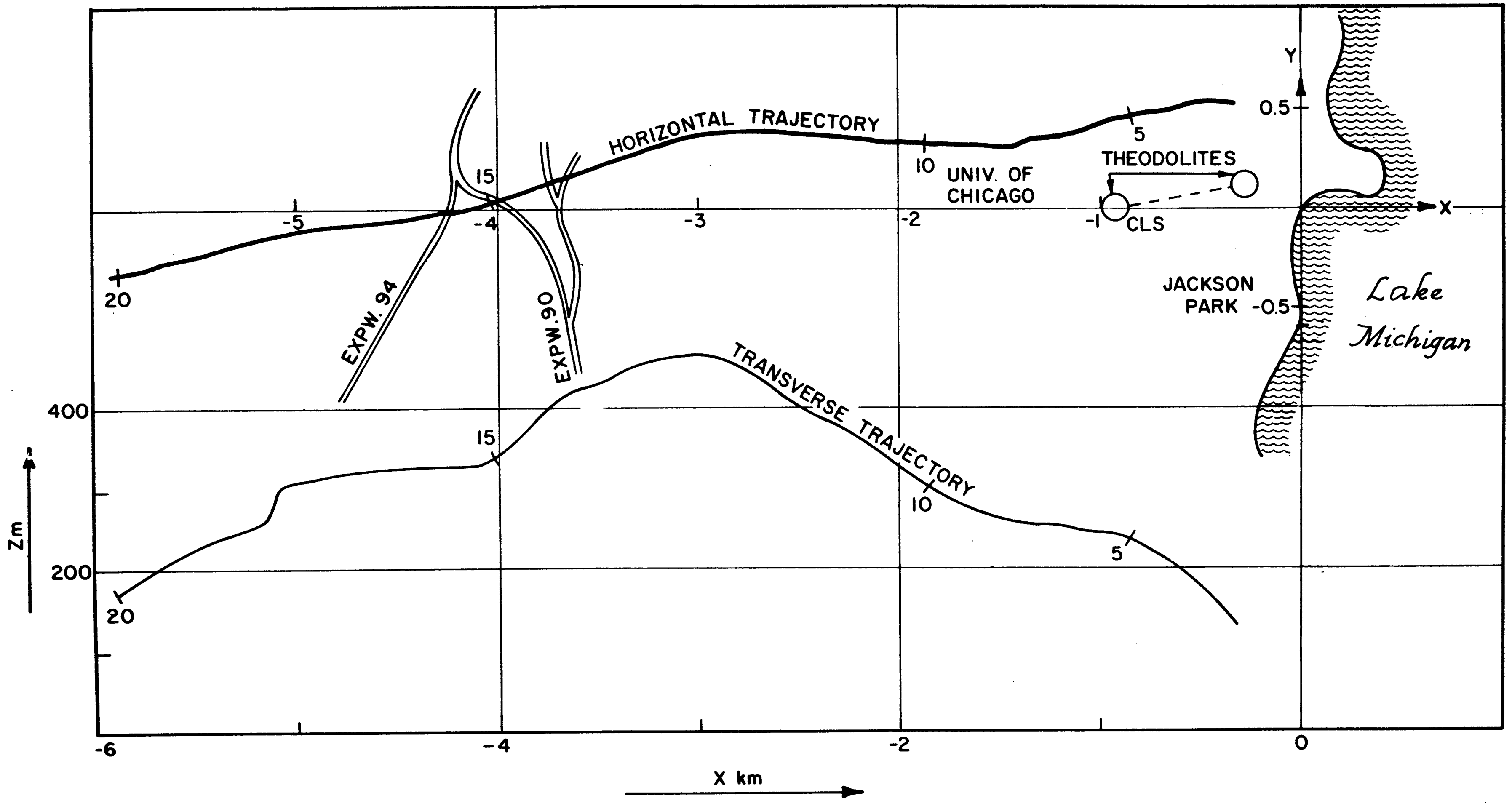

FIGURE 92. Horizontal and transverse trajectories of tetroon flight over Chicago, Illinois. The tetroon was launched at 1700 CST, 12 August, 1967. Elapsed flight times in minutes are indicated along the trajectories.

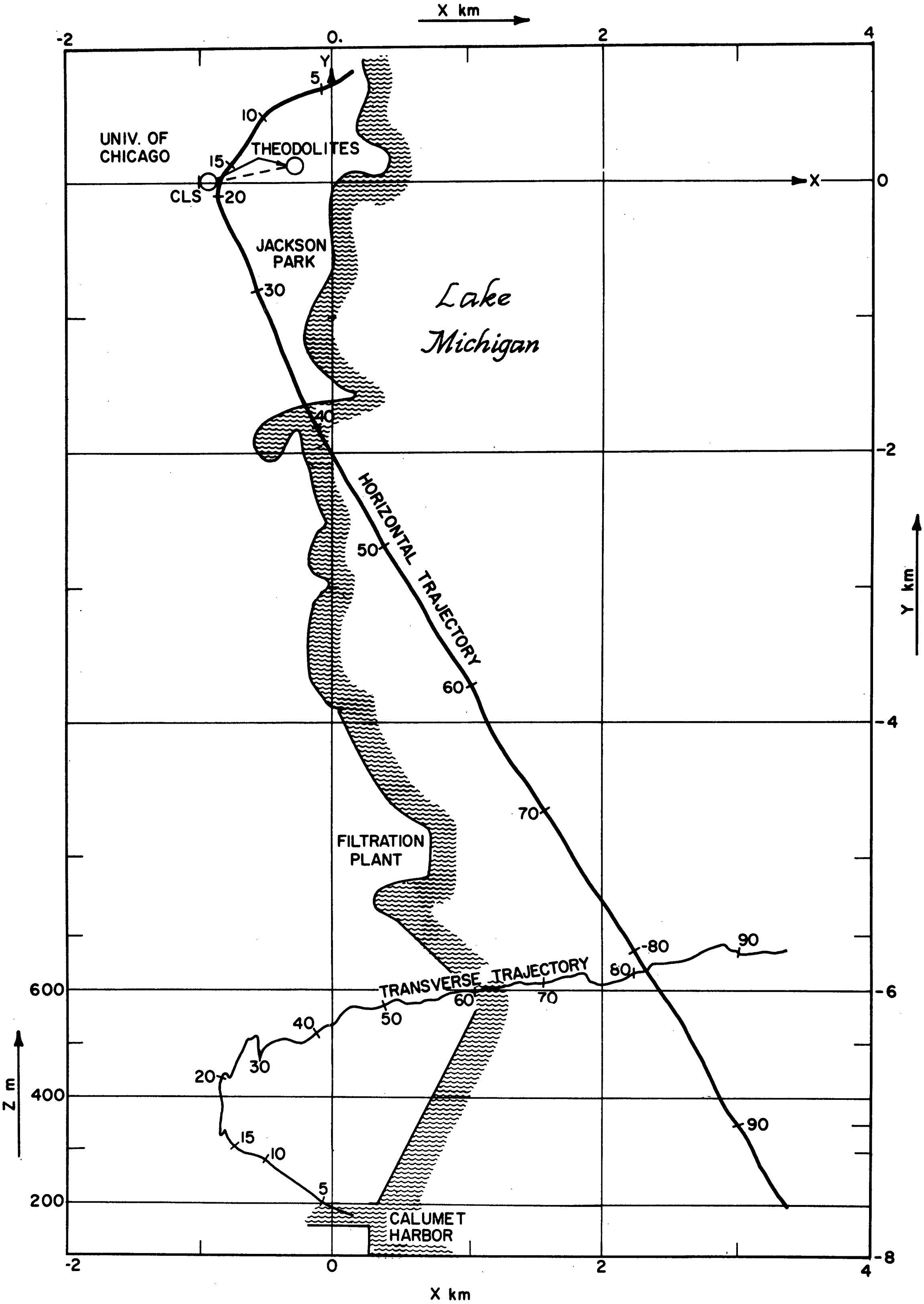

FIGURE 93. Horizontal and transverse trajectories of tetroon flight over Chicago, Illinois. The tetroon was launched at 0845 CST, 13 August, 1967. Elapsed flight times in minutes are indicated along the trajectories.

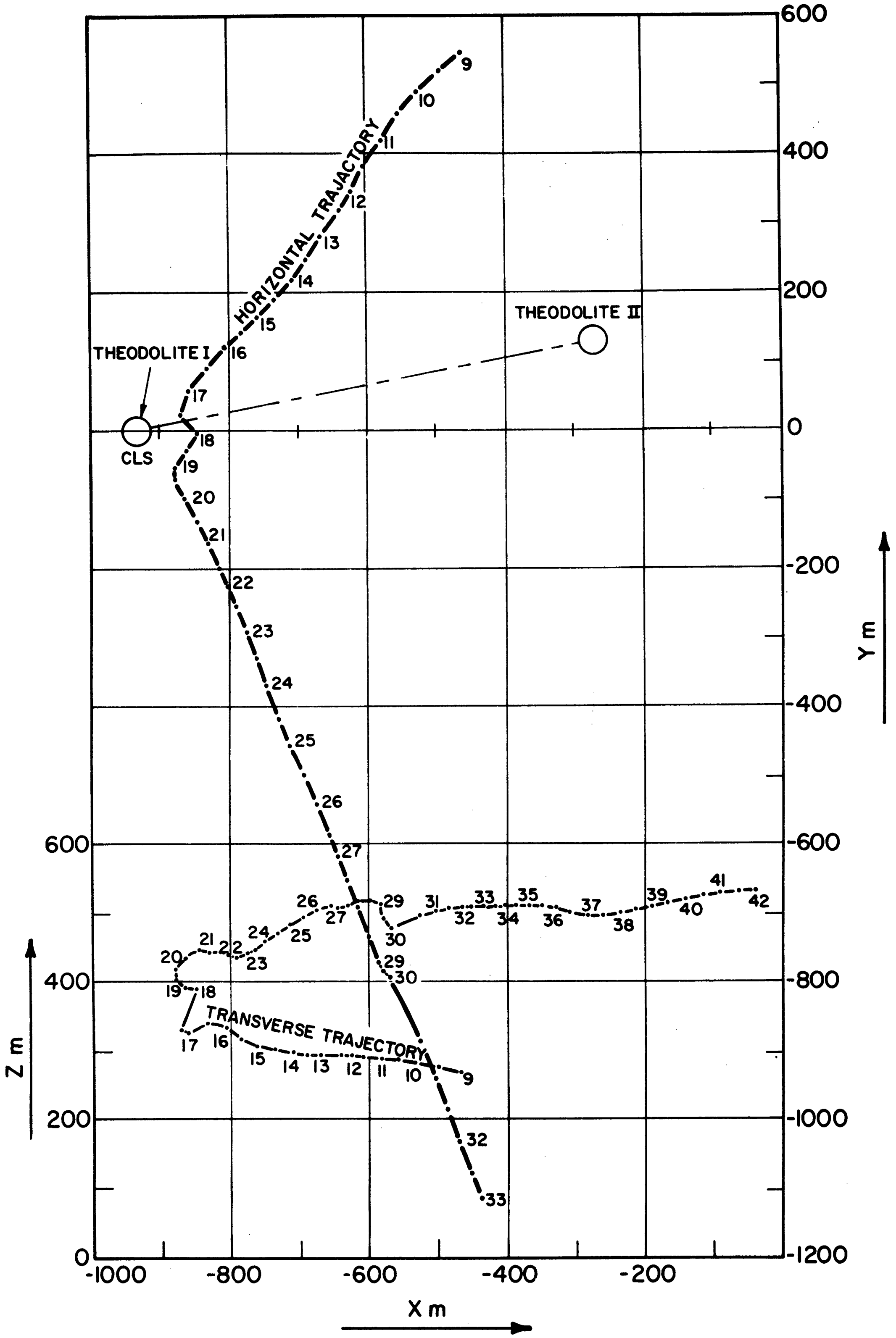

FIGURE 94. Horizontal and transverse trajectories of tetroon flight near the lake breeze front. The tetroon was launched at 0845 CST, 13 August, 1967 and its complete trajectory depicted in Figure 93. Tetroon positions are indicated every 1/2 minute and elapsed flight times in minutes are given.

TABLE 5. Average wind speeds and directions derived from trajectories of tetroon flights made near Chicago, Illinois on 12 and 13 August, 1967. Averages are derived for selected portions of the flights. The trajectories are shown in Figures 89-94.

Launch time	Time period (CST)	Altitudes in period (meters)	Wind speeds ($m\ sec^{-1}$)			Wind directions (degrees)	Remarks
			U	V	W		
0900, 12 Aug.	0905 - 0915	300 - 450	-2.5	-0.6	0.25	045	Lake breeze
	0920 - 0945	530 - 1000			0.31	Variable	Convergence zone
	0950 - 1020	1000 - 1000	1.7	-1.2	0.00	275	Return flow
	1030 - 1050	960 - 400			-0.47	Anticyclonic shift	Subsidence over lake
	1050 - 1120	400 - 500	-4.0	-1.9	0.06	035	Lake breeze
1700, 12 Aug.	1705 - 1712	240 - 460	-5.5	-0.3	0.52	055	Lake breeze
	1705 - 1720	240 - 170	-5.7	-0.7	-0.08	050	Lake breeze
0845, 13 Aug.	0850 - 0855	200 - 285	-1.5	-0.7	0.28	035	Lake breeze
	0850 - 0900	200 - 305	-1.1	-0.9	0.17	020	
	0900 - 0910	305 - 480			0.29		Near Convergence zone
	0910 - 0925	480 - 520	0.6	-1.5	0.04	307	Return flow over land
	0925 - 1015	520 - 670	1.1	-1.7	0.02	299	Return flow over lake

The accuracy of the presented trajectory is a function of the accuracy of the individual theodolite readings, the tetroon's position relative to the baseline direction, and of the distance between theodolites and tetroon. When the tetroon was within 2 km of either theodolite and the difference in the azimuth or elevation readings from the two theodolites was greater than 5 degrees the balloon position was accurately determined to within ± 10 m in the horizontal and ± 2 m in the vertical. The accuracy decreased as the tetroon moved further away from the theodolites.

The tetroon released at 0900 CST, 12 August was observed to follow more than one complete lake breeze circulation loop, Figure 89. The Coriolis effect was apparent both in the low-level onshore lake breeze and in the offshore return flow aloft as the trajectory is curved to the right at both levels. The tetroon's turbulent motion in the lake breeze convergence zone is depicted in detail in Figure 90. The tetroon was caught for more than 30 minutes in this, less than 1 km wide, zone. The wind variations observed in this frontal zone were in agreement with the observed pulsating nature of the frontal penetration. As the tetroon moved off shore its altitude was above its expected neutral level, indicating that the vertical velocities would be under-estimated for that portion of the trajectory. Over the lake the tetroon described an open-anticyclonic loop as it subsided, Figure 91. As the tetroon again moved inland, it was more than 6 km away from the tracking theodolites, but still easy to follow. The tracking was terminated as the tetroon was lost in smoke and haze near Lake Calumet, more than 10 km away from the theodolites.

The tetroon released at 1700 CST, 12 August was tracked for more than one hour, but its trajectory after approximately 20 minutes was close to the extended baseline and no accurate analysis was possible, Figure 92. The recorded azimuth angles from both theodolites were, however, slowly increasing, thus indicating a Coriolis effect, while the recorded elevation angles indicated that the tetroon oscillated in the vertical.

The neutral level for the tetroon released at 0845 CST, 13 August seems to have been at more than 500 m, Figure 93. It ascended rather smoothly near the lake breeze front, Figure 94 and was carried off shore by the return flow. This tetroon was tracked by a single theodolite for an additional 1/2 hr before being lost in smoke and haze after a total tracking time of more than 2 hrs. The tetroon was then more than 10 km away from the tracking site. The trajectory in the return flow was characterised by a broken straight line, the wind direction over the lake being approximately 8 degrees less than over land.

3.3.4 Comparison of Tetroon and Pibal Winds.

The observed lake breeze circulations are summarized in Figures 95 and 96. Surface winds and winds aloft are those observed from MDW and DWN on the hours indicated in the figure, while the winds aloft given at CLS are those observed closest to each respective hour. Representative winds derived from the tetroon trajectories are included in the plots.

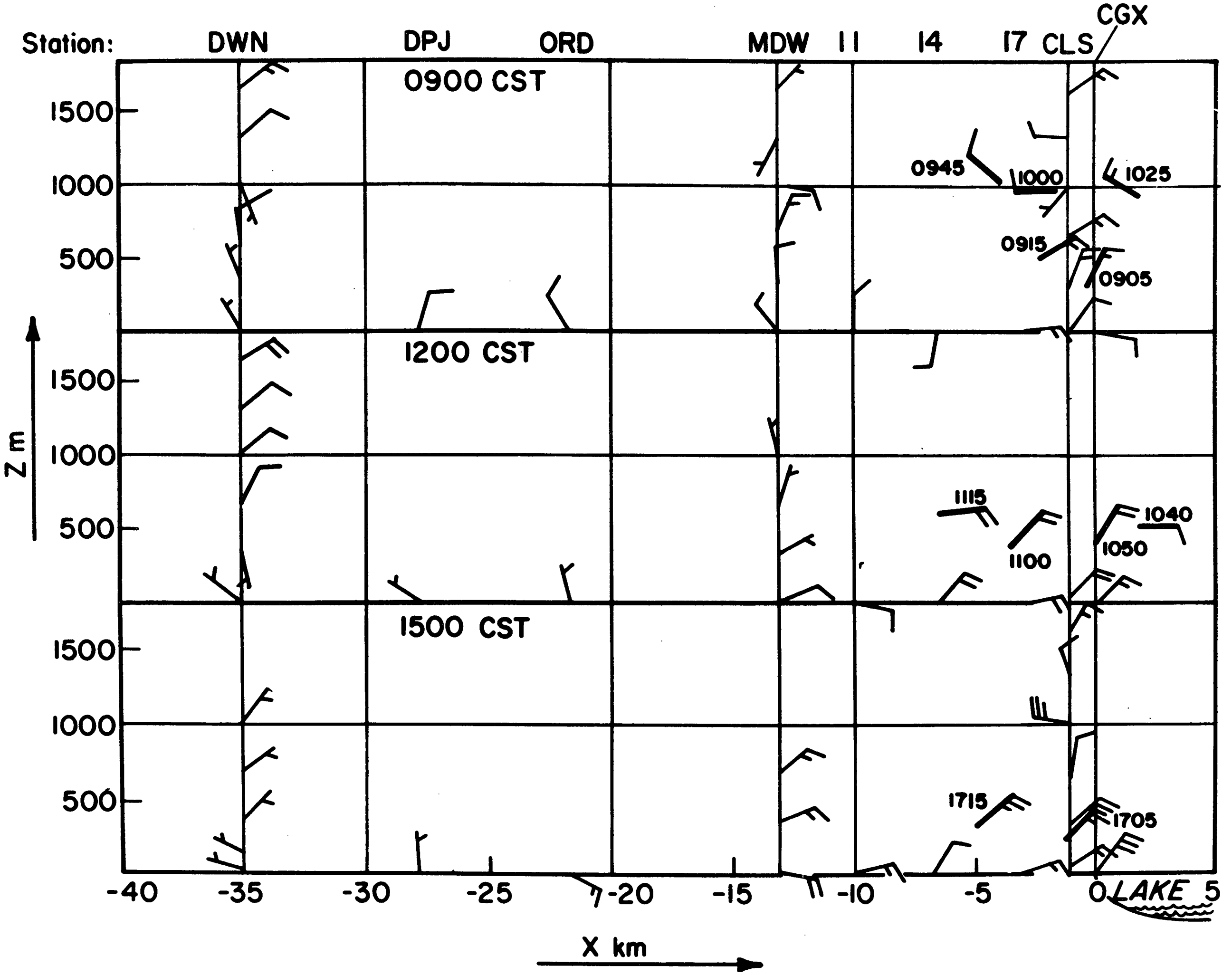

FIGURE 95. Vertical cross sections of winds at 0900, 1200, and 1500 CST on 12 August, 1967. Station codes and numbers refer to Figure 48 and Table 3. Circle indicates calm, no barb, 0.2-0.5; half barb, 0.5-1.5; full barb, 1.5-2.5 m sec^{-1} wind etc. Winds derived from the tetroon trajectories are indicated with heavy stem and CST-times, when these winds were measured are indicated.

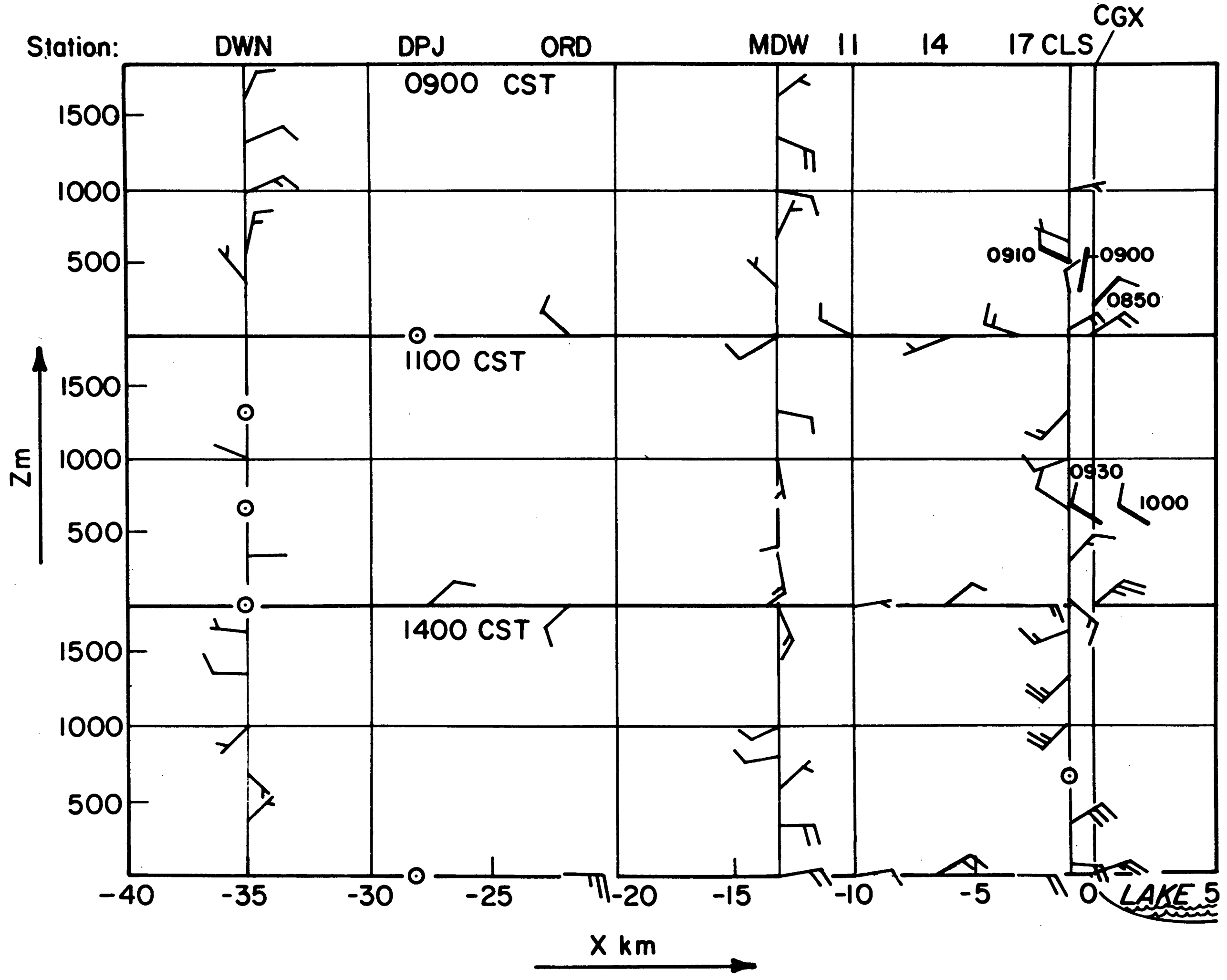

FIGURE 96. Vertical cross section of winds at 0900, 1100, and 1400 CST on 13 August, 1967. Conventions as in Figure 95.

In all cases the wind speeds derived from the tetroon trajectories are in close agreement with those observed during the pibal runs. Taking into account the right turning of the wind in response to the Coriolis acceleration and the time lags between the different wind observations the measured wind directions are also in good agreement.

3.3.5 Dew Point Temperatures and Aerosol Concentrations Aloft.

On 13 August the NCAR airplane flew three traverses across the lake shore. The flight path was parallel to the X-axis at Y = 7 km, thus passing over CGX and MDW. Several passes were made across the shore, at altitude increments of 150 m or 300 m up to 1500 m. The airplane's ground speed was 60 m sec^{-1} (120 kt). During these flights dew point temperatures and aerosol concentrations were successfully measured by means of a frost point hygrometer and a Royco aerosol counter, respective. The automated instrument system has been described in detail by Langer et al (1968). The specified accuracies were, ± 0.25°C in dew point temperature at 30°C and 1% in aerosol concentrations, while specified time constants were 2.0 sec and 0.25 sec, respective. Unfortunately the air temperature measurements were bad due to instrument failure and had to be discarded. Smoothed analyses of the dew point temperatures are presented in Figure 97. Spectral analyses of the observed aerosol concentrations were made and the spectral ranges, 0.5 - 3.0 microns and 3.0 - 9.0 microns are presented in Figures 98 and 99, respective. The horizontal and vertical scales in these figures are identical to those used in Figures 95 and 96.

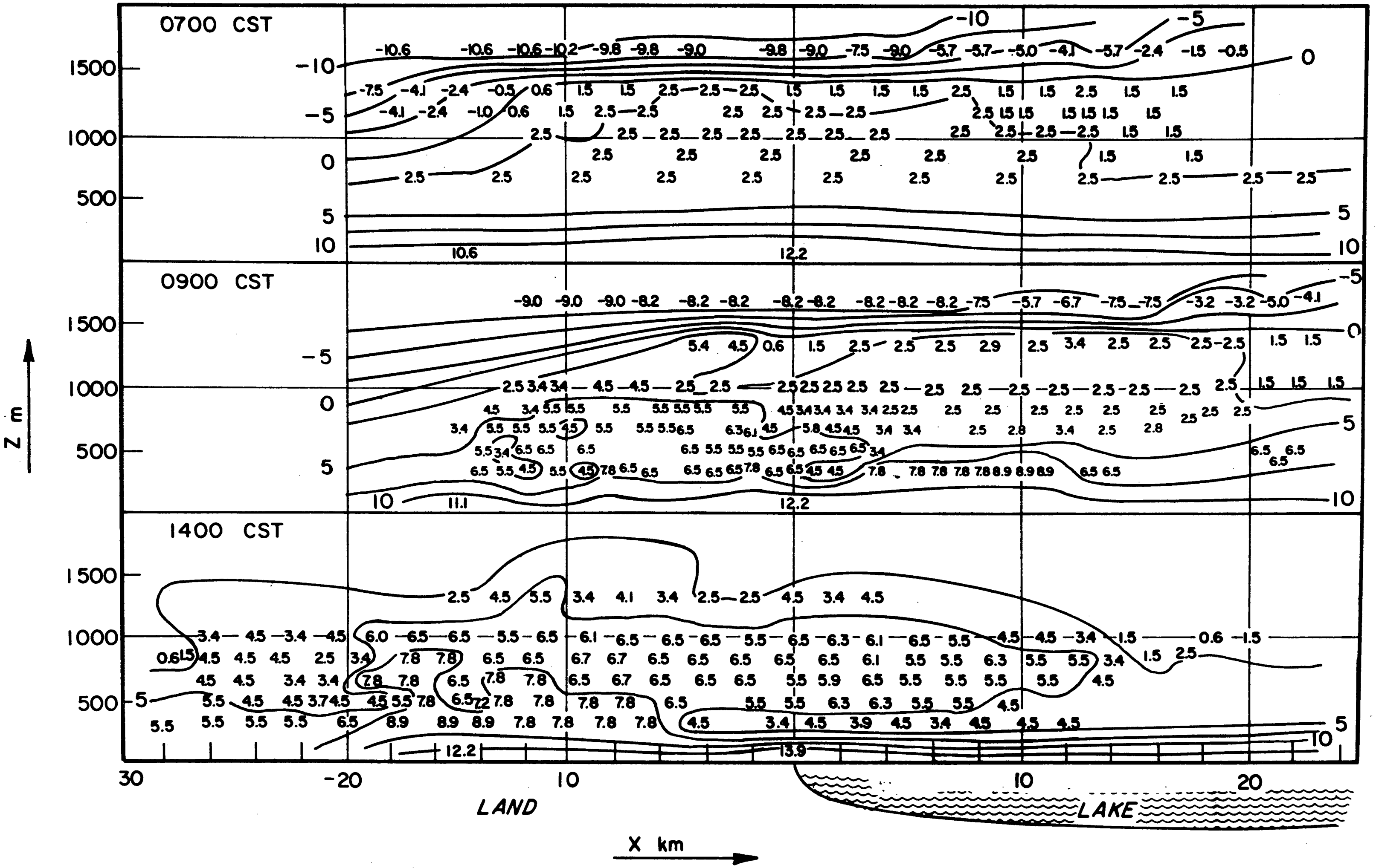

FIGURE 97. Analyzed vertical dew point temperature fields measured during three successive flights on 13 August, 1967. The flight paths were parallel to the X-axis at y = 7 km, see map in Figure 48. Dew point temperatures are given in °C.

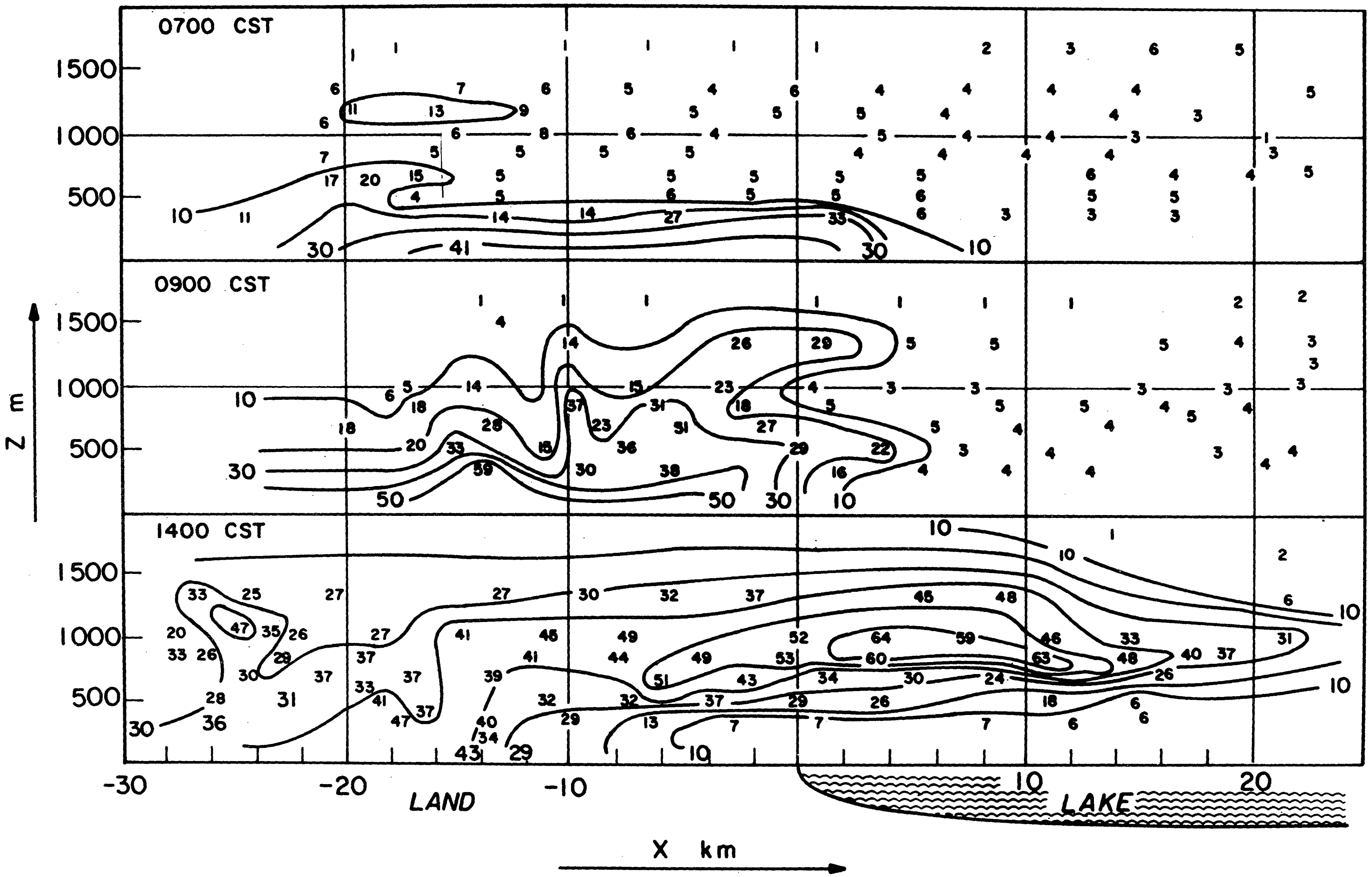

FIGURE 98. Analysed vertical fields of 0.5-3.0 microns aerosol concentrations measured during three successive flights on 13 August, 1967. The flight paths were parallel to the X-axis at $y = 7$ km, see map in Figure 48. Concentrations given are x 10^6 particles per cubic meter.

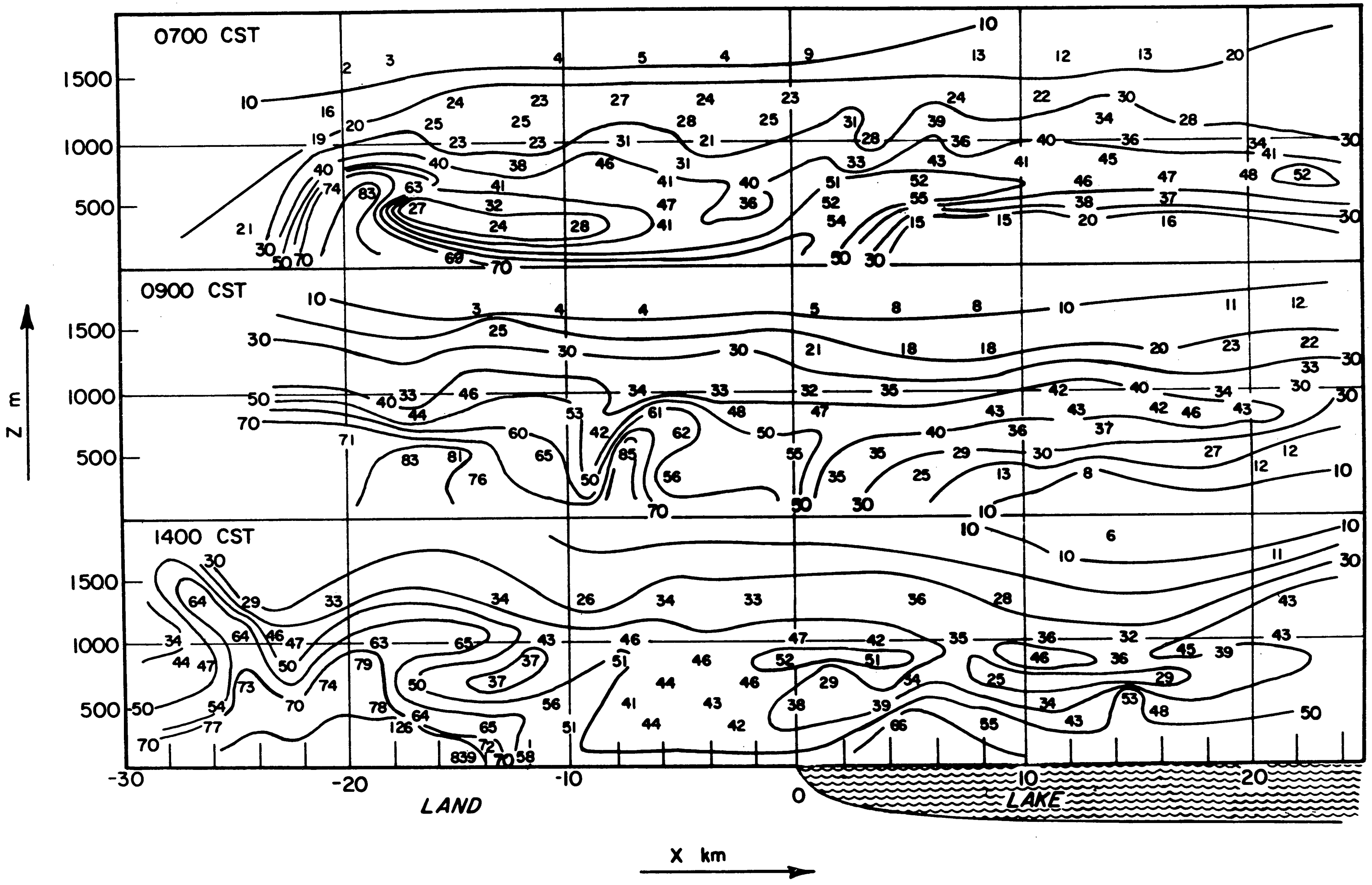

FIGURE 99. Same as Figure 98 except for 3.0-9.0 microns. Concentrations given are x 10^3 particles per cubic meter.

The first flight was made between 0631 CST and 0836 CST. As the winds were light and variable and the atmosphere relatively stable during this period no serious distortion of the sampled data field, due to drifting, was expected. By moving the sampling locations in accordance with the wind fields, the data fields sampled during flights between 0843 - 1009 CST and 1358 - 1450 CST were adjusted to correspond to 0900 CST and 1400 CST, respectively.

Many interesting features are depicted by the analyses. The urban heat island effect is seen in the dome-like shape of the lower isodrosotherms in the early morning. The land breeze had carried low level aerosols off shore during the night and early morning. The onshore flow layer, observed in the early morning at approximately 1000 m, Figure 86 - 88, brought in clean air aloft from off shore and could possibly also explain the strong subsidence, seen in the dew point temperature analysis inland from the urban heat island, Figure 97.

At 0900 CST the lake breeze circulation was initiated near the shore. Convective thermals were breaking through the stable inversion layer, bringing the moist and polluted air up into the return flow layer. Weak subsidence off shore, brought clean and dry air down to the onshore, lake breeze layer, Figures 97 and 98. At 1400 CST the return flow had carried polluted air more than 15 km off shore. Strong subsidence occured further out over the lake and also inland from the lake breeze front, Figure 97.

Average pollution, measured as total suspended particulates, has been found to be 170 micrograms m^{-3}, while average rural background has been estimated to 30 micrograms m^{-3}, Kenline (1968).

3.3.6 Photographic and Visual Observation of Clouds, Haze, and Smoke.

Several photographs and time lapse movies were taken of smoke, haze, and cloud patterns during 12 and 13 August. Combined with visual observations, made both from the various manned surface stations and from the aircraft, these photos reveal some interesting characteristics of the observed circulations.

On both days the surface layer, below 500 m was heavily polluted from several kilometers off shore to more than 10 km inland. During the early morning, before the onset of natural convection, columns of smoke were seen rising through the smoke layer above intense local heat sources. Over some of these columns, stationary cumulus clouds formed. One such cloud was photographed above a steel mill at 0630 CST, 12 August, Figure 100, and again at 0910 CST, after the passage of the lake breeze front, Figure 101.

As the lake breeze front started to move inland after 0800 CST, the smoke was observed to move toward the convergence zone forming an almost opaque wall. While the lakeward side of this 1 to 2 km wide wall was almost vertical and very distinct, the inland side was more diffuse. When the smoke had risen to the height of the return flow layer, it was carried off shore, leaving an area of clear air below it. A photograph taken at 0900 CST, 13 August, Figure 102,

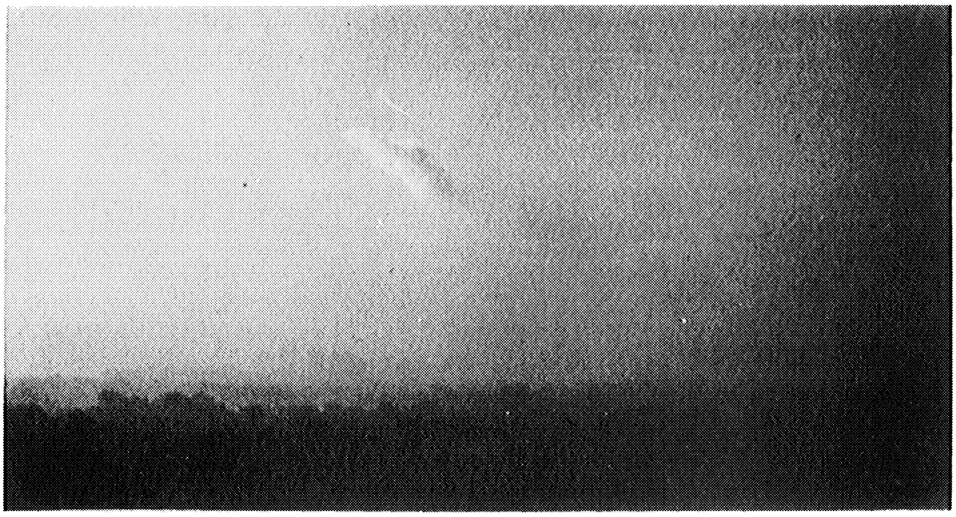

FIGURE 100. Stationary cumulus formed above a steel mill. Photograph taken toward SSE from CLS at 0630 CST, 12 August, 1967.

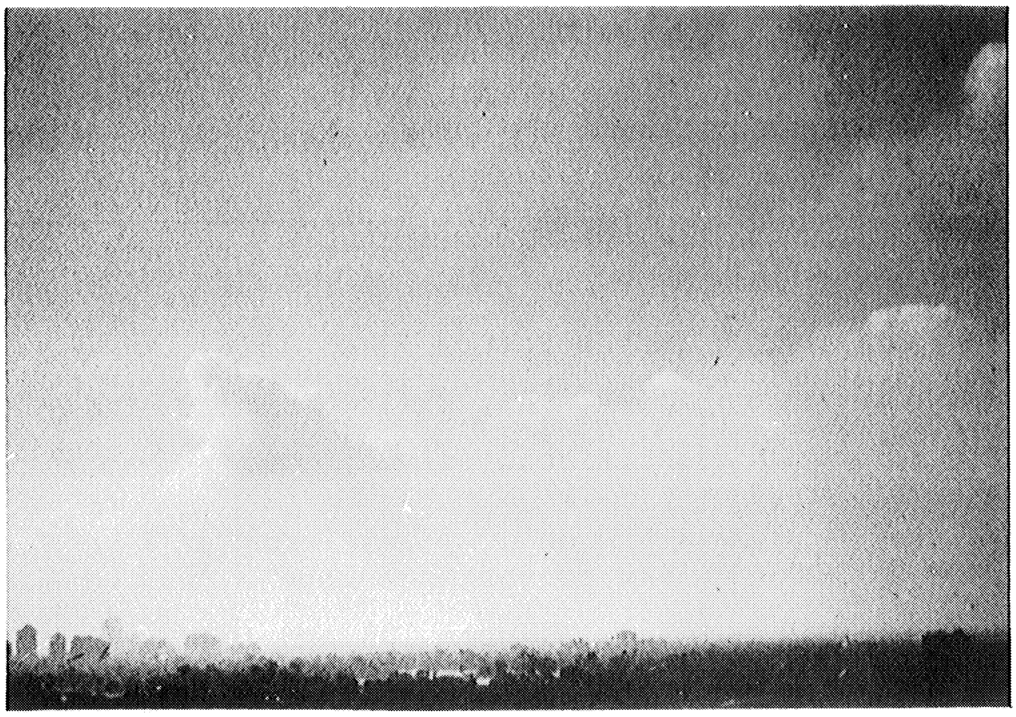

FIGURE 101. Stationary cumulus formed above a steel mill, left and cumulus associated with the lake breeze convergence zone, right. Photograph taken toward SSE from CLS at 0910 CST, 12 August, 1967.

FIGURE 102. Photograph taken toward NNW during midmorning on 13 August, 1967. Smoke caught in the return flow layer is carried off shore above the undercutting clean lake breeze air.

shows how smoke from near the downtown Chicago area had risen and was caught in the return flow aloft. In mid-afternoon the return flow layer, filled with red-borwn smoke, was clearly identifiable far out over the lake, Figure 103.

Above the lake breeze convergence zone a band of cumulus formed in the mid-morning, Figures 103 and 104. The cloud bank, deliniating the lake breeze front along the shores around the southern basin of Lake Michigan can be noted, as well as the stationary cumulus closer to shore, associated with the local heat sources. Late in the afternoon, when the lake breeze front had progressed far inland, the smoke

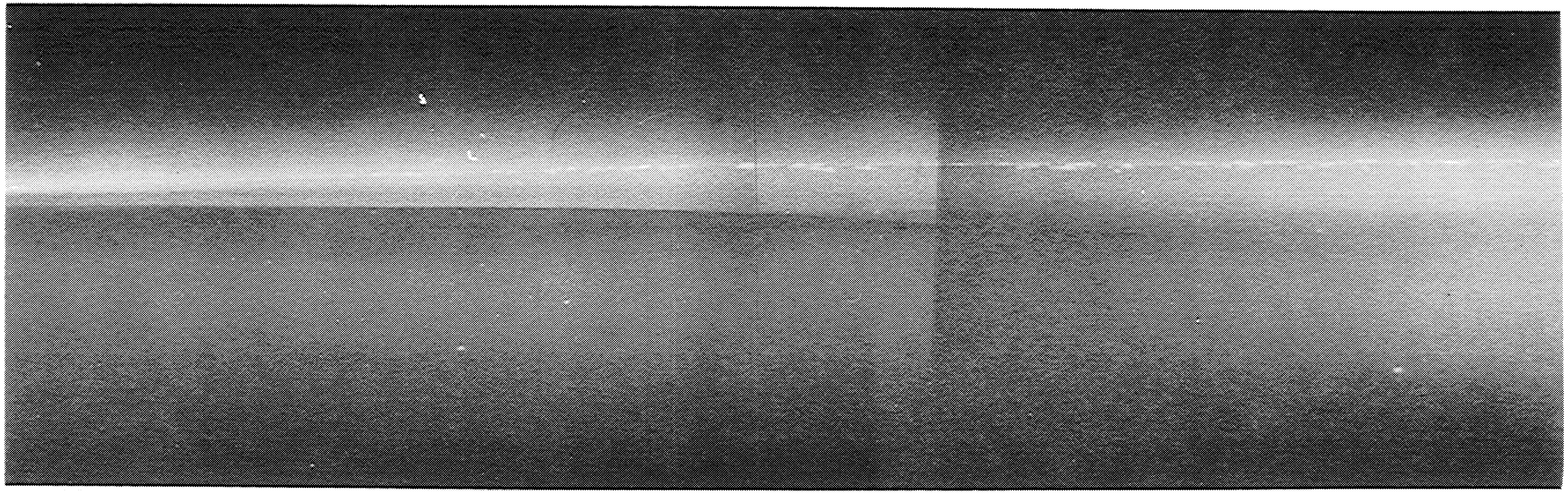

FIGURE 103. Partial panoramic infrared photograph taken toward SSW from an aircraft, at 2000 m and 20 km east of the University of Chigaco Campus (CLS) at 1415 CST, 13 August, 1967. The smoke filled return flow layer in the foreground and the band of lake breeze cumulus in the background.

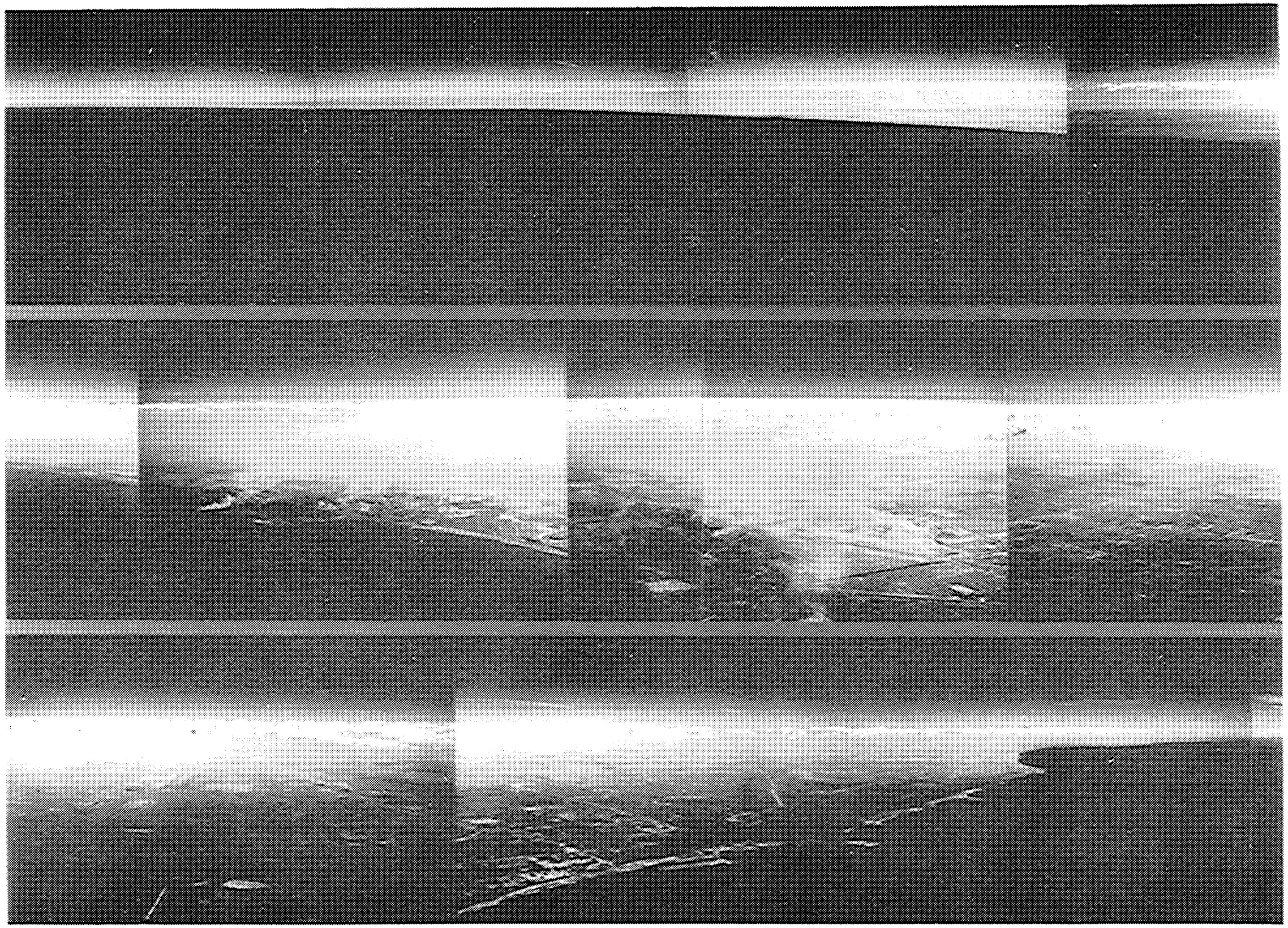

FIGURE 104. 360 degree panoramic infrared photograph taken from an aircraft, at 2000 m and 11 km east of the University of Chicago Campus (CLS) at 1400 CST, 12 August, 1967.

FIGURE 105. Photograph of rising smoke column taken toward NE from an aircraft at 1000 m and 5 km south of Downers Grove (DWN) at 1030 CST, 13 August, 1967.

density in the convergence zone had decreased and the frontal cloud bank disappeared. The fronts were, however, still clearly identifiable from the air, 25 km inland at 1500 CST on both days. The almost vertical rise of a smoke column, Figure 105, indicated that the wind was light and variable inland from the lake breeze front.

The inferred motions from the smoke and haze observations were in good agreement with the meteorological measurements of the circulations. Furthermore there was a good correspondence between measured aerosol concentrations and observed opacity of smoke and haze.

3.4 SUMMARY.

The spatial and temporal variations of temperature, moisture, wind, and pollutants near Chicago, Illinois have been observed and described for 12 and 13 August, 1967. During these days a weak NNE gradient flow prevailed and well developed lake breeze circulations occurred.

The usefulness of tetroons as mesoscale air parcel tracers was reaffirmed and the transport of aerosols in the lake breeze circulation documented. In summary the observations showed:

1. The effect of day time heating and nocturnal cooling led to large diurnal air temperature variations, exceeding 15°C over land, while the water surface temperature remained approximately constant;
2. The stability of the air over the lake varied from unstable to very stable in the layer between 5-12 m above the surface. However, the lake surface was at all times warmer than the air at 5 m. Intense modification of this air in short trajectories over land was observed;
3. Strong nocturnal urban heat islands, with more than 5°C higher temperatures in the city than in the surrounding rural areas, were observed in the early mornings;
4. An offshore flow aloft, the return flow, was observed at the shore before 0800 CST on both days. The onset of this return flow was the first indication of a lake breeze circulation;

5. The onset of an onshore flow, the lake breeze, was observed within 150 m of the surface, near the shore, between 0800 CST and 0900 CST on both days;
6. The average upward velocity near the lake breeze front was 30 cm sec^{-1}, where occasional updrafts exceeding 200 cm sec^{-1} were observed. Horizontal convergence near the surface, toward the front, and subsidence on the landward side of the front were noted;
7. Abrupt changes in air and dew point temperatures, reversals of surface wind directions, and drastic imporvements in visibility occurred within one hour, as the lake breeze front passed a station near the shore. These changes were less pronounced at stations further inland;
8. The lake breeze front progressed inland in a series of pulses, at an average rate of 1.1 m sec^{-1}. The front penetrated on both days more than 35 km inland;
9. The depth of the lake breeze layer varied between 700 m and 1000 m on 12 August and between 400 m and 600 m on 13 August;
10. A lake breeze of more than 6 m sec^{-1}, with an onshore component exceeding 5 m sec^{-1}occurred below 300 m, near the shore in mid-afternoon on both days;
11. Upward motion, in general less than 30 cm sec^{-1}, was observed in the lake breeze flow layer over land;
12. General subsidence was evident over the the lake, and downward velocities in excess of 40 cm sec^{-1}

were measured 1.5 km off shore on 12 August. A tetroon, descending over the lake, traced an anticylonic loop;

13. The return flow layers aloft were more than 500 m thick, and offshore winds in excess of 3 m sec^{-1} were observed in those layers;
14. The effect of the Coriolis acceleration was seen both in the lower lake breeze flow and in the return flow aloft;
15. Moisture and aerosols were transported by the lake breeze circulation. Aircraft measurements, photographs, and visual observations showed that moisture (measured as dew point temperature) and aerosols, especially in the 0.5 - 3.0 micron size range, were lifted by the convective updrafts near the lake breeze front and carried out over the lake by the return flow. Convective clouds were observed to form over the lake breeze front and also above local heat sources;
16. Surface land breezes and strong urban heat islands, centered around down-town Chicago and the industrial areas around the southwestern shore of Lake Michigan, were observed in the early mornings on both days;
17. Meteorological measurements from around the Great Lakes and aerial photographs taken from an aircraft and a satellite suggested that the lake effect was homogenous along the lake shore.

4. LAKE AND SHORELINE EFFECTS ON AIR POLLUTION DISPERSION

This chapter presents a generalized description of the effects a lake and a lake shore have on the atmosphere's capability to disperse pollution. The presentation is largely based on the findings made in the two observational studies reported on in Chapters 2 and 3 of this report, but draws in many cases also upon the experience of other investigators. As an extensive literature review is presented in Chapter 1, only a few direct references will be made in the following sections.

The effects that will be discussed have a length scale ranging from 1 km to 100 km, and thus include both convective and mesoscale systems. Holland (1967) has described the difference between displacement and dispersion of atmospheric pollution by relating length and time scales. Diurnal differential heating and cooling between land and lake surfaces are the cause of the predominant shoreline effects. Thus, according to Holland, the dispersion is dominated by the mesoscale motion systems, while displacement is dominated by the synoptic motion systems. When a lake region is under the influence of a stagnant anticyclone, displacement becomes small, while shoreline circulations normally become strong. In order to correctly predict air pollution dispersion near a lake shore it is thus essential to have a thorough understanding of these shoreline circulations.

When the displacement motion is strong, and an air mass moves over a lake or a shoreline, mesoscale effects will cause changes in the dispersion characteristics of the air.

The difficulties of constructing realistic mathematical models of these mesoscale motion systems and effects are partly due to complex interactions with convective and turbulent motion systems. In order to appreciate the complexity of the dispersion climate in a lake region one must first understand the various influences at work. Several effects can be described as isolated phenomena, although they rarely, if ever, have been observed in their "pure state" in the atmosphere. Any actual analysis of the dispersion pattern in a specific lake region requires a synthesis of these various influences.

It should be noted that, although the following descriptions and discussions will center around effects induced by lakes and lake shores, many of the characteristics presented would apply equally well to effects along ocean shores.

4.1 LAKE AND LAND BREEZE CIRCULATIONS.

In the middle latitudes lake breezes occur on between 30% and 60% of the days in spring and summer. They also occur, although less frequently, in the fall and winter. Land breezes are most frequent in fall and early winter, while they are expected to be least frequent in the spring.

The driving force in lake and land breeze circulations are horizontal pressure gradients across a shoreline. These gradients are primarily caused by differential heating and cooling between air over land and air over the adjacent lake.

Several reasons for the more rapid heating of a land surface during daytime and the more rapid cooling of a land surface during nighttime as compared with a lake surface have been given e.g. Hewson et al (1960). The ratio of the specific heats of sand, rock, and typical soil to that of water is about 1 to 5.5, while the ratio of their specific gravities is about 3.7 to 1. The net result is that the ratio of the thermal capacities, per unit volume, of typical land surface materials to that of water is approximately 1 to 1.5. However, the much greater thermal conductivity of typical land surface materials compared to that of water tends to neutralize this residual difference. Greater penetration of the sun's rays into the water than into the soil, differences in surface reflectivity as well as differences in water vapor pressure above the surfaces, and thus latent heating could explain differences in heating of the two surfaces, but these effects have been shown to be small and probably insignificant in most cases.

The most significant difference between the two surfaces is that turbulent mixing and transport, due to waves and currents in a lake, disperses heat energy, gained or lost by the lake surface, rather rapidly downward, while no such dispersion mechanism exist in a land mass.

The diurnal variation of the lake surface temperature in large and deep lakes is normally less than 1°C. Seasonal variations of lake water temperatures are strongly dependent on geographical location and physical characteristics of the lake. In the Great Lakes the surface water temperature ranges from 0°C in late winter and early spring to more than 20°C in late summer. When a lake becomes covered with ice and snow the temperature contrast between land and lake normally becomes negligible and the lake effect concept irrelevant.

The effect of daytime radiational heating may increase the land surface temperature more than 10°C above, while nocturnal radiational cooling may lower that temperature by more than 10°C below the diurnal average. These variations are strongly dependent on the characteristics of the land surface, e.g. type of soil, vegetation, snow cover, etc. Urban heat islands near the shore will modify the temperature differential across the shore. Anticyclonic synoptic conditions, with clear skies and light gradient winds, favor the development of strong temperature differential across a shore.

Assuming that isobaric and isothermal surfaces are horizontal in a lake shore region in the early morning, the following idealized description of the life cycle of a lake breeze circulation can be given. As the land surface is heated more rapidly then the lake surface, the air in a layer close to the ground becomes warmer than the corresponding layer over the lake, mainly due to convective heat transfer. Due to the warming, the air near the surface expands and becomes less dense, thus lifting the air column over land relative to the air column over the lake. The surface pressures over the land and over the lake are still the same, as no advection of mass has taken place. However, at some level aloft, this lifting creates a pressure gradient from land toward the lake. This pressure gradient initiates horizontal air motion toward the lake. This offshore flow aloft is usually called the "return flow" in the lake breeze circulation.

Due to the advection of mass from the air column over land to the air column over the lake, surface pressure decreases over land and increases over the lake. This surface pressure gradient initiates an onshore flow near the surface, the lake breeze.

When the heating of the land surface increases, the convection over land becomes stronger. Strong upward motion over land is compensated by subsidence over the lake, thus closing the lake breeze circulation cell. As the circulation intensifies, its horizontal extent increases and horizontal air trajectories become longer. While the initial across-shore flow was antitriptic, the later mature flow is deflected in response to the Coriolis acceleration and the wind veers toward a more along-shore direction. A mature lake breeze circulation is depicted in Figure 106a and the effect of the Coriolis acceleration shown in Figure 107.

In the early evening, when the radiational balance reverses, i.e. outgoing long wave radiation becomes stronger than the incoming solar radiation, the lake breeze circulation enters its final stage. As the driving temperature differential ceases, shortly after sunset, the lake breeze dies.

A similar reasoning in reverse can explain the life cycle of a land breeze circulation. At night the land surface is cooled more rapidly than the lake surface. However, as this surface cooling tends to stabilize the air near the surface, there is less exchange of heat through the surface air layer, i.e. no correspondence to daytime thermal convection. Thus the sinking of the air column over land, due to shrinking and increase in density of the air in the layer near the ground, is less pronounced and the development of the circulation slower. On the basis of this reasoning

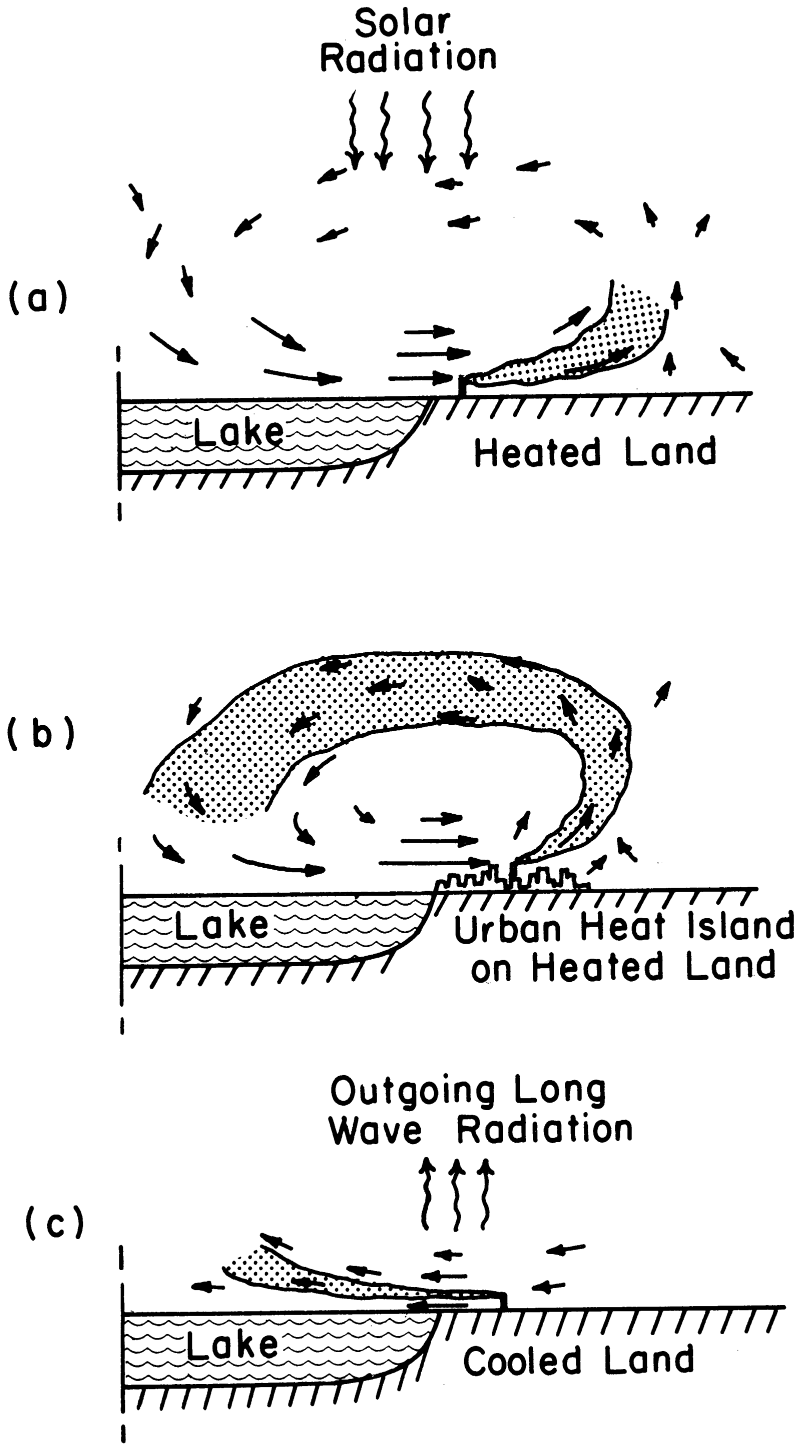

FIGURE 106. Thermal circulations near a shoreline. (a) lake breeze, driven by differential radiational heating. (b) lake breeze reinforced by an urban heat island. (c) land breeze, driven by differential cooling.

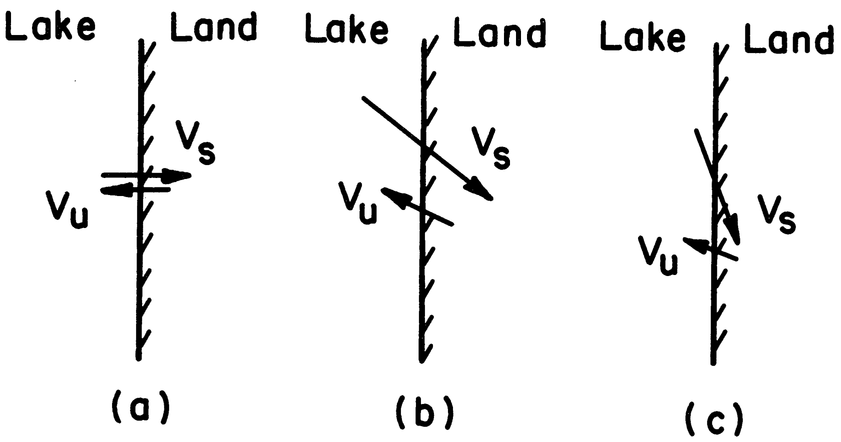

FIGURE 107. Plan view of lake breeze winds indicating the effect of the Coriolis acceleration. V_S = lower lake breeze wind, V_U = return flow wind aloft. (a) Initial stage-morning. (b) Mature stage - afternoon. (c) Final stage-late afternoon or early evening.

one would expect the land breeze circulation to be less intense than the lake breeze circulation.

In the following sections some of the observed features of these shoreline circulations will be discussed as well as how they are influenced by shoreline characteristics and varying synoptic conditions.

4.1.1 Classical Case.

Almost ideal lake and land breeze circulations developed on the eastern shore of Lake Michigan on 25 June, 1965. The measurements of these circulations have been presented in Sections 2.3.2 and 2.4 of this report. Different phases and various features of almost ideal circulations have also been observed and reported by other investigators and reviewed in Section 1.2. On the basis of these observational studies the following conclusions can be drawn:

- In order for a land or a lake breeze circulation to develop there has to be a temperature differential across the shore. Lake breezes occur when the land surface is warmer than the lake surface, land breezes when the land surface is cooler than the lake surface. The stronger this differential is the more intense the circulation and the greater its horizontal and vertical extent becomes;
- The onset of a lake breeze circulation is first observed as an offshore flow aloft, the return flow, one to two hours after sunrise on clear days. One to two hours later an onshore flow, the lake breeze, occurs within a couple of kilometers off shore;
- The leading edge of the lake breeze flow has the characteristics of a cold front. This front has a slope of

less than 1:20 near the surface and progresses inland at a rate of approximately 1 m sec^{-1};

- The front is most pronounced near the shore, before the cool and moist lake air behind the front has been modified by the land surface. In intense circulations, air temperatures may drop more than 5°C and dew point temperatures rise more than 3°C as the front passes a station. The lake breeze front may in such cases penetrate more than 30 km inland and the lake breeze extend more than 10 km off shore;
- Strong horizontal convergence in the frontal zone is coupled with updrafts in the zone and subsidence within one kilometer ahead of (inland from) the front. The width of the convergence zone may be less than 500 m. Average updrafts of more than 50 cm sec^{-1} and maxima exceeding 500 cm sec^{-1} occur in intense circulations. Frequently a line of cumulus clouds form above the front;
- The average depth of the lake breeze flow layer varies between 300 m and 700 m. Wind maxima in that layer are located within 100 m to 300 m of the surface, near the shore and occur in the late afternoon. The lake breeze wind speed may exceed 7 m sec^{-1};
- The return flow layer in the lake breeze circulation may be more than 1000 m thick and winds in that layer may exceed 5 m sec^{-1};
- Subsidence occurs over the lake and intensifies the inversion in the upper portion of the lake breeze flow

layer. The lower portion of that layer is stable over the lake due to the coolness of the water surface;

- A thermally induced, superadiabatic internal boundary layer develops in the lake breeze flow layer over land;
- The onset of a land breeze circulation is first observed aloft as an onshore flow, the return flow, near sunset on clear days. Within a few hours after sunset an offshore flow, the land breeze, occurs within 100 m of the surface;
- The land breeze flow layer is in general stable, with a strong, surface based inversion over land and may be more than 500 m deep. Wind maxima in that layer are located within 100 m of the surface and occur near sunrise. The land breeze wind may exceed 5 m sec^{-1};
- The return flow layer in the land breeze circulation may be more than 1000 m thick and winds in that layer may exceed 3 m sec^{-1};
- Horizontal convergence coupled with weak ascending air motion occur over the lake and may cause a layer of low stratus clouds to form over the lake. Subsidence occurs over land;
- The effect of the Coriolis acceleration causes the lake breeze to become almost parallel to shore in the late afternoon. This effect on the land breeze and the return flows is less pronounced;
- Lake and land breeze circulations occur homogeneously along the shores of a lake.

4.1.2 Changes due to Gradient Winds.

The characteristics of lake and land breeze circulations are strongly modified by gradient winds. Observational studies have shown that:

- No complete lake or land breeze circulations occur in cases when the gradient winds are strong, approximately more than 5 m sec^{-1} for large bodies of water, such as the Great Lakes. However, perturbations will be induced in the gradient flow due to across shore temperature differentials;
- The circulations might develop fully of partially in light gradient flow, e.g. the return flow aloft might be masked. The resulting winds can best be described as a superposition of the two wind fields;
- The onset of the lake breeze is delayed, and the lake breeze might form several kilometers off shore in the case of offshore gradient wind conditions or when the early morning land breeze is strong. An offshore gradient wind also causes the frontal characteristics to be more pronounced, the lake breeze flow layer to be shallower (occasionally less than 200 m even in intense circulations), the inland progression to be uneven, and the inland penetration to be less.

4.1.3 Changes due to Shoreline Geography and Topography.

The curvature of a lakeshore determines the location and strength of the convergence and divergence zones, and thus also the intensities of lake and land breeze circulations. For example, near a concave coast, lake breezes are divergent and land breezes convergent, while on a peninsula or an island the reverse holds true. (Note that in the southern basin of Lake Michigan the lake breeze is divergent while the return flow in the lake breeze circulation is convergent. This explains the high aerosol concentrations observed aloft over the lake, Section 3.3.5.) The curvature and the heat storage capacity are related to the physical dimensions of a lake. Although thermal circulations would develop near any lake shore, across which a temperature differential exists, they are less likely to occur around small and shallow lakes. Circulations around a circular lake would be different from those along the shores of a long and narrow lake.

The topography of the shoreline influences the circulations in several ways, e.g.:

- Lake and land breeze circulations are not likely to occur along steep and high coasts;
- Slope and valley wind circulations are driven by pressure gradients caused by differential heating of different surfaces (or parts) of a slope or a valley. A simplified depiction of these winds near a lakeshore is presented in Figure 108. These winds will tend to intensify lake and land breeze circulations. However, they are not expected to be significant, if the slope of the land surface is less than 1:150;
- A rough land surface will induce mechanical turbulence and thus aid in building up an internal boundary layer in the lake breeze flow layer over land. It will also, due to friction, strongly modify the wind-profile in the

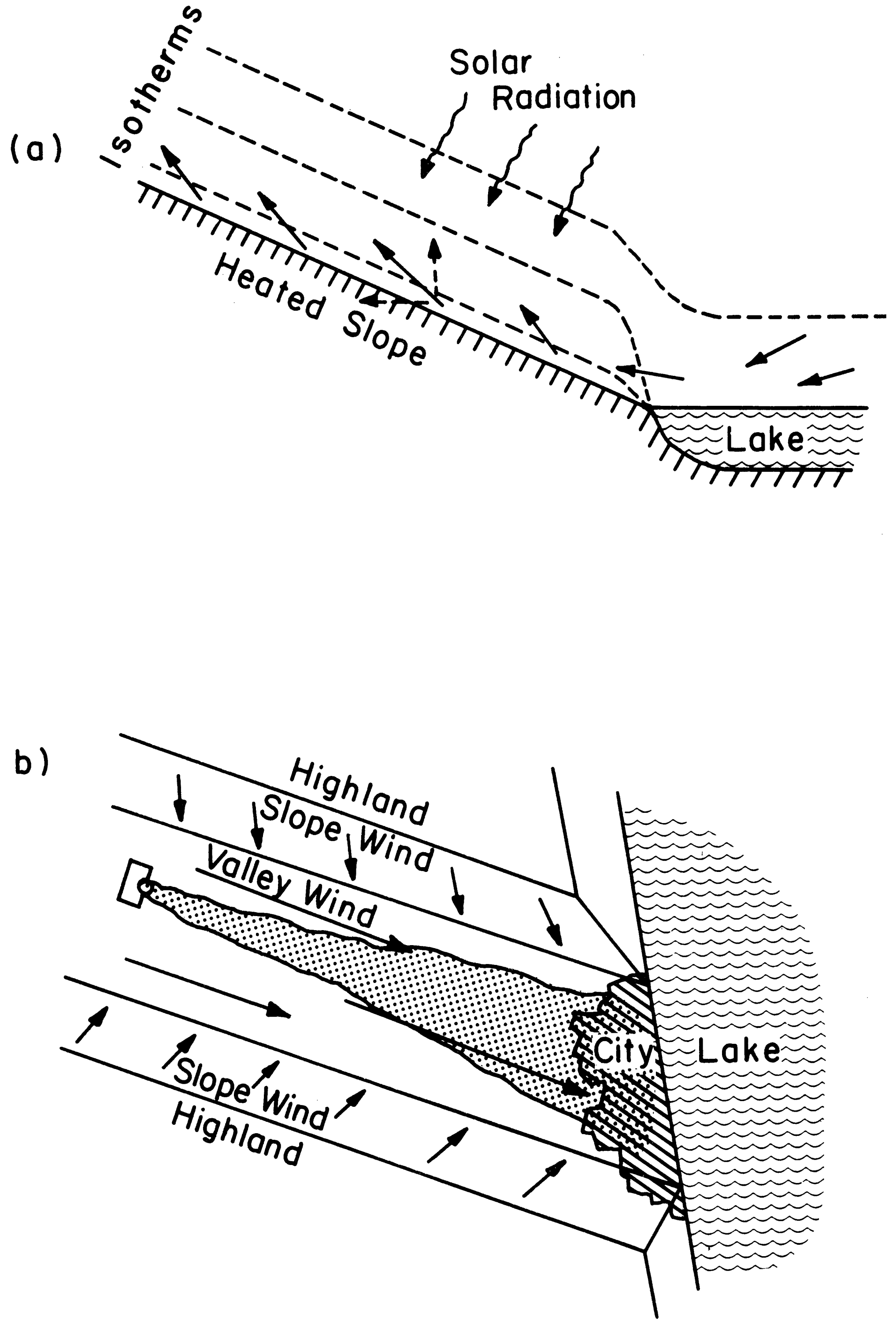

FIGURE 108. Slope and valley winds. (a) Upslope wind during the day. (b) Downslope winds reinforce a down-valley wind at night. Fumigation may occur as the stable air in the valley reaches the urban heat island and moves out over the warmer lake surface.

stable land breeze flow layer over land. Occasionally, when nocturnal cooling sets in and induces a surface based inversion over land, low-level jets form (at about 100 m above the surface) due to the decrease in downward momentum flux.

4.1.4 Effects due to Urban Heat Islands near the Shore.

From the standpoint of forecasting air pollution dispersion, the effects induced by an urban heat island near the shore is important, as these heat islands normally are associated with industrial and other sources of pollutants. A study of lake and land breeze circulations near Chicago, Illinois, on the southwestern shore of Lake Michigan on 12 and 13 August, 1967 has been presented in Chapter 3 of this report.

The urban heat island is caused by an excess of heat stored in massive buildings and pavements and by heat released in urban and industrial activities. The heat island is most pronounced at night, especially under anticyclonic synoptic conditions when surrounding rural areas are cool. Its size and strength depends on the physical dimensions and characteristics of the urban area. In a large city the air temperature is on the average between 1°C and 2°C higher than in adjacent rural areas, but the difference may exceed 10°C. While surface based inversions are frequent in rural areas at night, elevated inversions with their bases at 200 m to 500 m aloft and above an unstable air mass at the surface, are most frequent in urban areas. Thermal circulations, similar to lake and land breeze circulations, may be caused by these heat islands.

These "urban circulations" will cause lake breeze circulations to start earlier in the morning and add to their intensity as shown in Figure 106. However, they counteract the development of land breeze circulations. Figure 109 shows haze and pollution trapped beneath the elevated urban inversion and carried off shore by the land breeze and also pollution breaking through the inversion and carried off shore by the "early" return flow in the lake breeze circulation. About two hours later the return flow in the lake breeze circulation has carried pollution far out over the lake, leaving only a shallow layer of clear air near the lake surface, Figure 110. Note, that as infrared photography was used, the moist surface layer of the lake breeze flow shows up in the picture.

4.2 AIR FLOW CHANGES NEAR A LAKE SHORE.

When air flows across a lake shore, changes in the characteristics of the underlying surface induce shear-effects in the flow layer near the ground. The effects, which wind speed and wind direction shears will have on a smoke cloud or plume, are depicted in Figure 111. As the mixing by turbulent diffusion between clean air and a cloud is most active at the surface that separates them, strong shear effects will aid in dispersing the cloud.

4.2.1 Shoreline Downwash and Upwash Effects.

Downwash and upwash effects near a lake shore are basically caused by the difference in roughness between land and lake surfaces. As indicated in Figure 112, the wind velocity profile will change in an air mass that moves across a

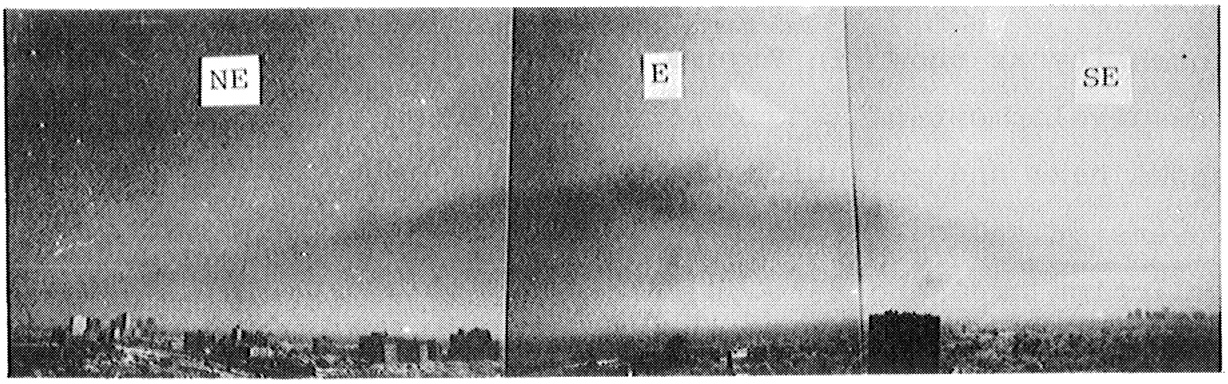

FIGURE 109. Partial panoramic infrared photograph taken toward SE from the University of Chicago Campus at 0950 CST on 4 August, 1966. A strong nocturnal land breeze had carried pollution far out over the lake. Pollution lifted by convection over land is carried off shore aloft by the return flow. Note that the density and thickness of the smoke and haze layer is reflected in the brightness in the picture. Clear sky is dark.

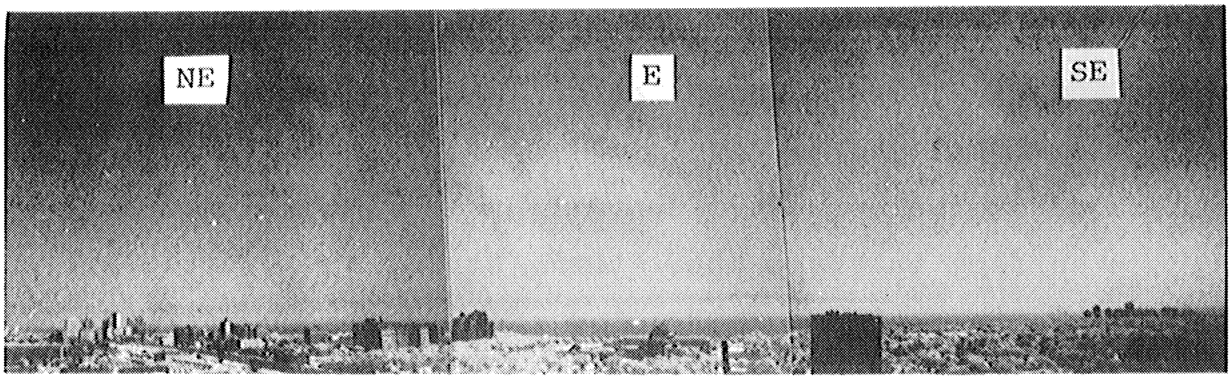

FIGURE 110. Same as Figure 109 except at 1135 CST. Note that a clear air zone is still detectable over the lake.

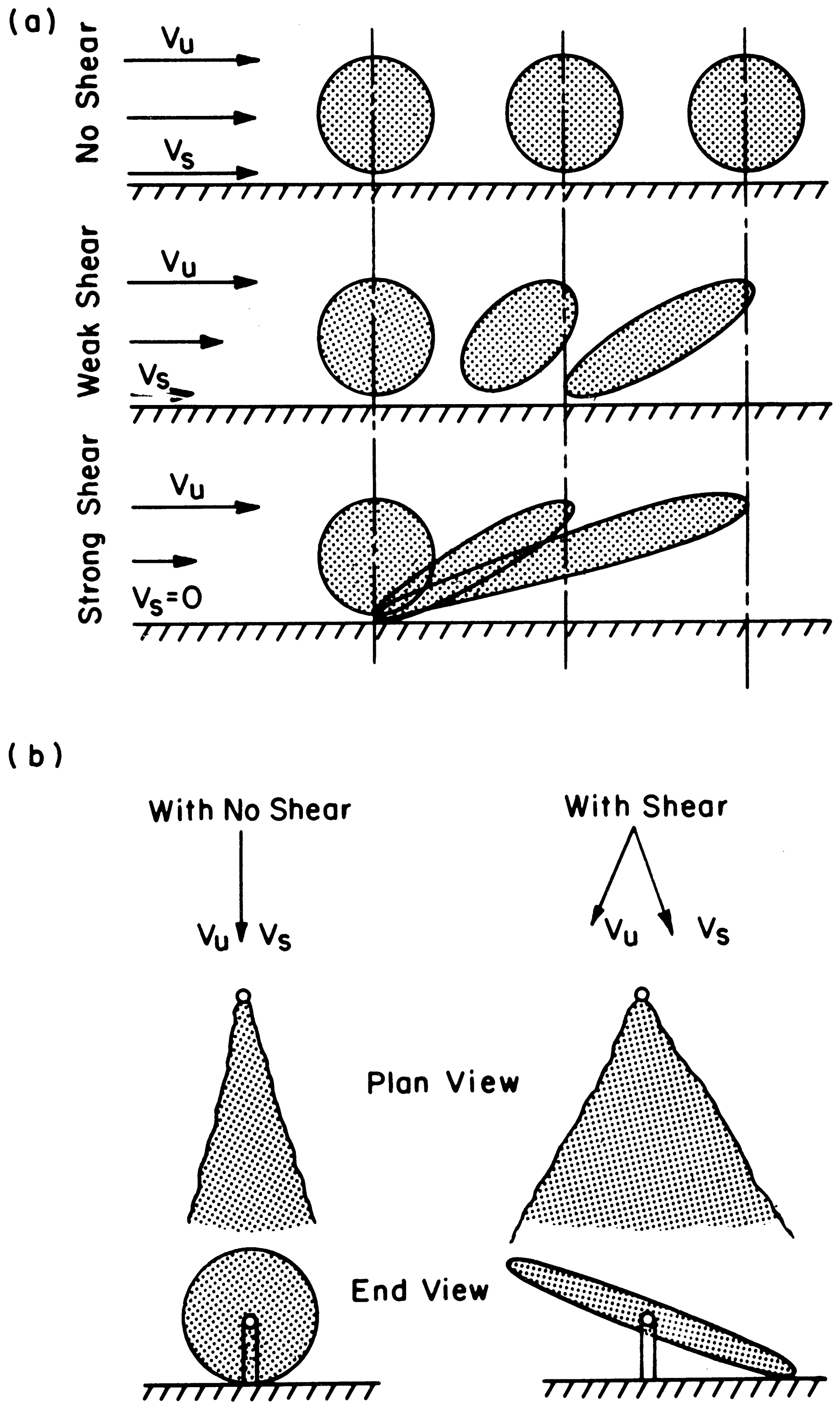

FIGURE 111. Deformation of a cloud due to (a) wind speed shear at three consecutive times, and (b) due to wind direction shear. V_s = surface wind, V_u = upper wind.

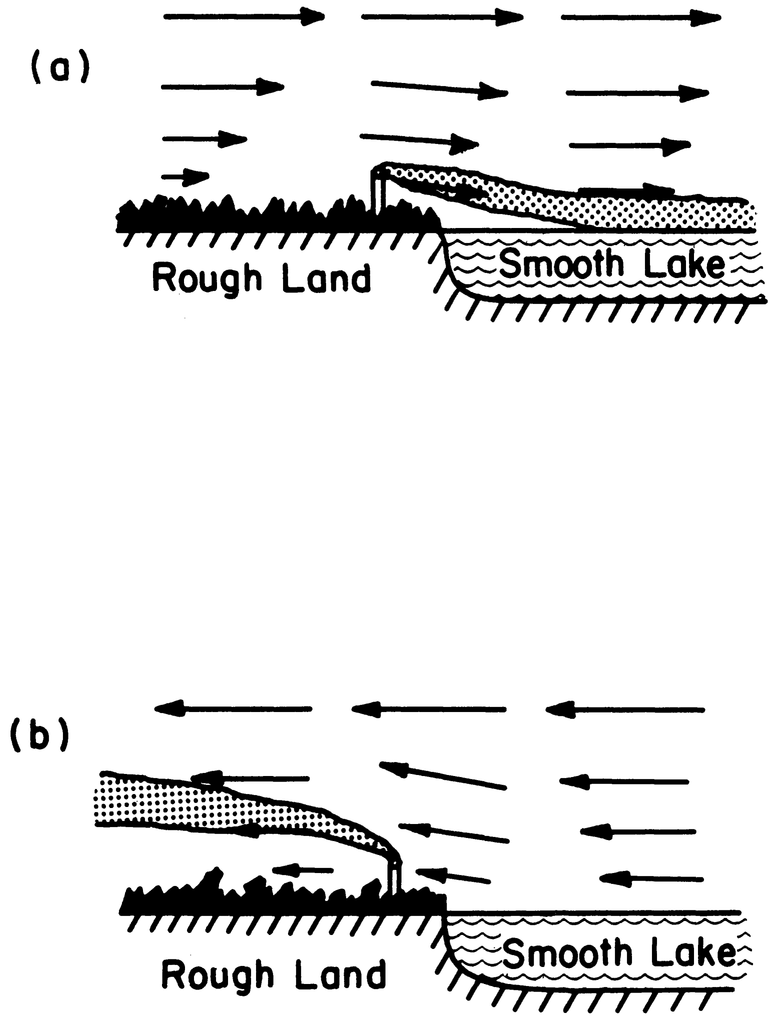

FIGURE 112. Changes in the airflow near a lake shore due to (a) dynamic downwash with offshore wind, and (b) dynamic upwash with onshore wind.

shoreline, causing compensating vertical motions near the shore. A high shoreline would, in addition to this frictional effect, induce dynamic downwash due to a wake effect with offshore winds.

While the upwash effect in general would be favorable from the standpoint of air pollution dispersion, downwash effects force pollution down toward the lake surface, beneath or into stable layers, that frequently persist over a lake. In land breeze flow, on the eastern shore of Lake Michigan, these downward velocities have been observed to exceed 15 cm sec^{-1} within 1 km off shore.

4.2.2. Wind Direction Changes Near a Lake Shore.

Wind direction changes near a lake shore are caused both by the difference in roughness between land and lake surfaces, and by the temperature differential that might exist across a lake shore.

Near the surface the flow is more or less antitriptic, due to surface friction. Through the frictional boundary layer, the wind normally veers with height and becomes geostrophic above that layer. The degree of veering, i.e. the degree of wind direction shear, depends on the surface roughness (surface friction), the stability of the air in the boundary layer, and the depth of the boundary layer. In general, the rougher the surface is, the more antitriptic the flow near the surface becomes; and the shallower and more stable the boundary layer is the stronger the wind direction shear.

A horizontal temperature gradient at the surface will influence the pressure aloft, thus changing the direction of the isobars aloft and hence the direction of the geostrophic

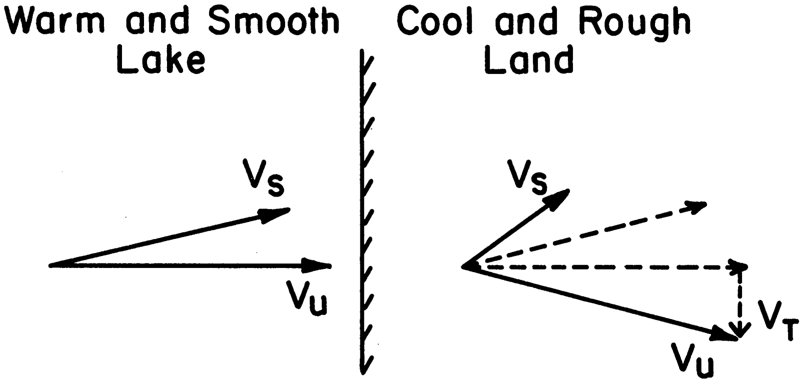

FIGURE 113. Wind direction change with height near a lake shore due to dynamic and thermal effects. V_s = surface wind, V_u = upper wind, V_t = thermal wind.

wind. Figure 113 indicates how an increase in both wind direction and wind speed shears would occur as air moves from over a warm smooth lake, across a shore, and in over a cold rough land area.

4.3 AIR MASS STABILITY CHANGES OVER A LAKE REGION.

When an air mass moves across a lake region its stability properties will change in response to differences in the character of the underlying surface. Advection inversions will develop in the surface layer of an air mass that moves from warm land areas out over a cool lake. As this air mass again moves inland an unstable internal boundary layer will develop in a manner similar to what has been described for lake breeze circulations. Pollution may be trapped in this stable air layer over the lake and transported, across large lakes in high concentrations for 100 km or more. Fumigation will occur during the breakup of this stable layer, when the air again moves inland. These phenomena are depicted in Figure 114, and a photograph of smoke plumes trapped in a stable air layer off shore is presented in Figure 115.

Stabilizing effects on an air mass moving across a lake region can also be caused by large scale topographical features and differences in surface roughness, which might induce downwash effects and hence subsidence inversions. If the air mass, that moves off shore, is warm and moist and the lake surface cool, strong inversion layers may develop aloft due to a release of latent heat.

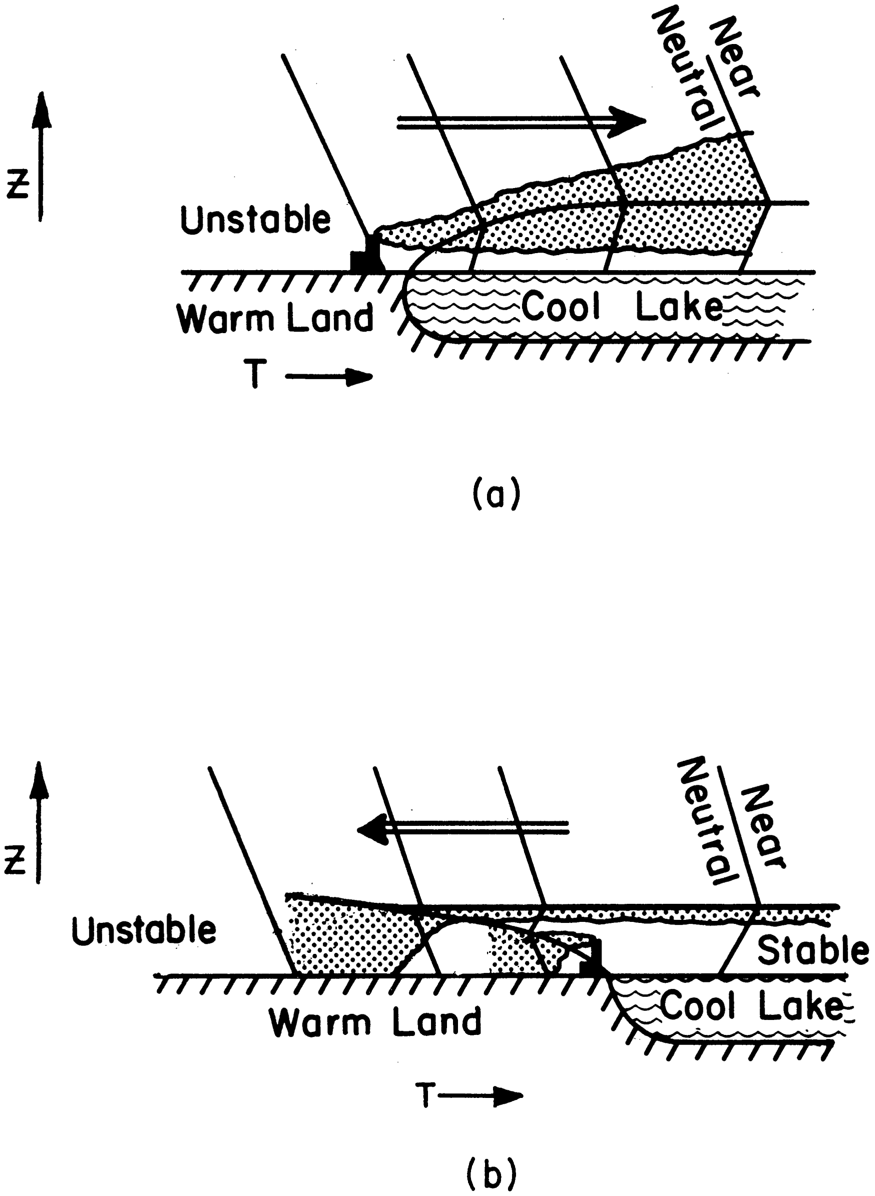

FIGURE 114. Variation of lapse rate and stability in an air mass as it moves across a lake. (a) Development of a stable inversion layer in warm air moving out over a cool lake. (b) Fumigation occuring as the stable layer breaks up over warm land.

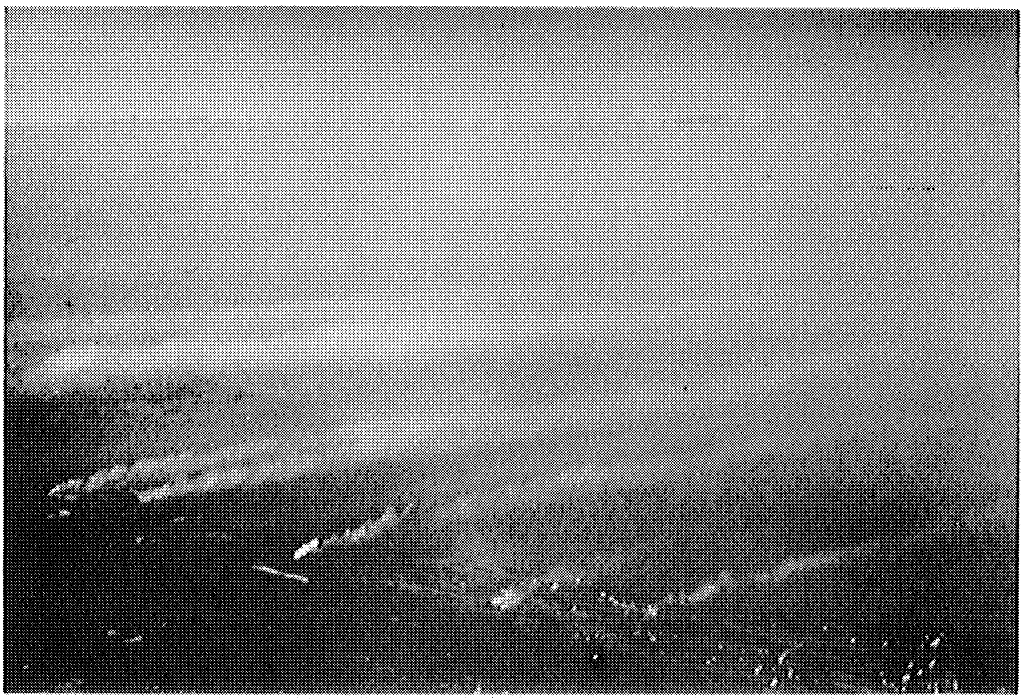

FIGURE 115. Photograph taken toward NE from an aircraft at 2500 m and SW of Gary, Indiana at 0530 CST on 15 July, 1968. Smoke plumes from mills near the shore of Lake Michigan are carried off shore by a 5 m sec^{-1} SW gradient flow.

5. CONCLUSIONS AND SUGGESTIONS FOR FURTHER RESEARCH

The presence of a lake or lakes in a region has to be considered in evaluation of the atmospheric dispersion climate. It has been shown how a lake alters the various meteorological parameters that are of prime importance in such evaluations. The predominant effects are those due to the differences of surface characteristics between land and lake. These differences cause local thermal circulations, perturbations on the large scale flow, and changes in atmospheric stabilities. Topographical features associated with lakes further complicate flow patterns and stability changes in a lake region. Any actual analyses of the dispersion patterns near a specific lake or shoreline require a synthesis of these various influences and in many cases special investigations may be required to determine the predominant factor involved.

This study has attempted to summarize our present knowledge of the various effects a lake and a shoreline would have on air pollution dispersion. While certain aspects of these effects have been fairly well documented in the literature, others have so far received little attention. It seems appropriate to point out some investigations and some definitive experiments that are needed and could be accomplished within the framework of the state of the art today:

- Climatologies of various dispersion conditions are needed for guidance of regional planning and disaster control. Based on available synoptic data, the frequency of air mass stagnation and related mixing heights, as well as the frequencies of occurrence of land and lake breezes should be determined. For regions with a dense network of continuously reporting meteorological stations, e.g. the region around the southern basin of Lake Michigan, such climatologies could be made very detailed and would thus constitute a vital tool for prediction of air pollution dispersion potentials in the region;

- Several specific aspects of lake effects could be studied by means of constant level balloons, tetroons, tracked by double theodolite techniques or by radar. For example, by placing two tetroons simultaneously at two different altitudes, wind shear influences as well as down- and upwash effects could be investigated, and by release of clusters of tetroons and the subsequent determination of their relative separation, estimates of Lagrangian diffusion characteristics could be obtained;
- Much data from intensive observational studies of land and lake breeze circulations are available and should be analyzed, in order that the characteristics of these circulations would be further discerned;
- Based on available observational data, improvements of existing, and developments of new theoretical models of various lake effects, e.g. land and lake breeze circulations, growth of internal boundary layers, etc., could be made;

- Finally the search for accurate prediction models of air pollution dispersion should continue. One approach that might prove useful, especially around medium size lakes, is depicted in Figure 116. When a lake region is under the influence of stagnant anticyclonic conditions for at least several days, the main air motions are those due to the differential heating between land and lake surfaces. Pollution released into the air near the lake shore under such conditions would essentially be confined within a "hypothetical box" with maximum dimensions determined by prevailing maximum daily mixing height, H, and the inland penetration of the lake breeze front. This "box model" approach has proved to be useful in estimating accumulations of air pollutants in a topograpically restricted basin and gives at least an estimate of anticipated average concentrations of various pollutants in such a region.

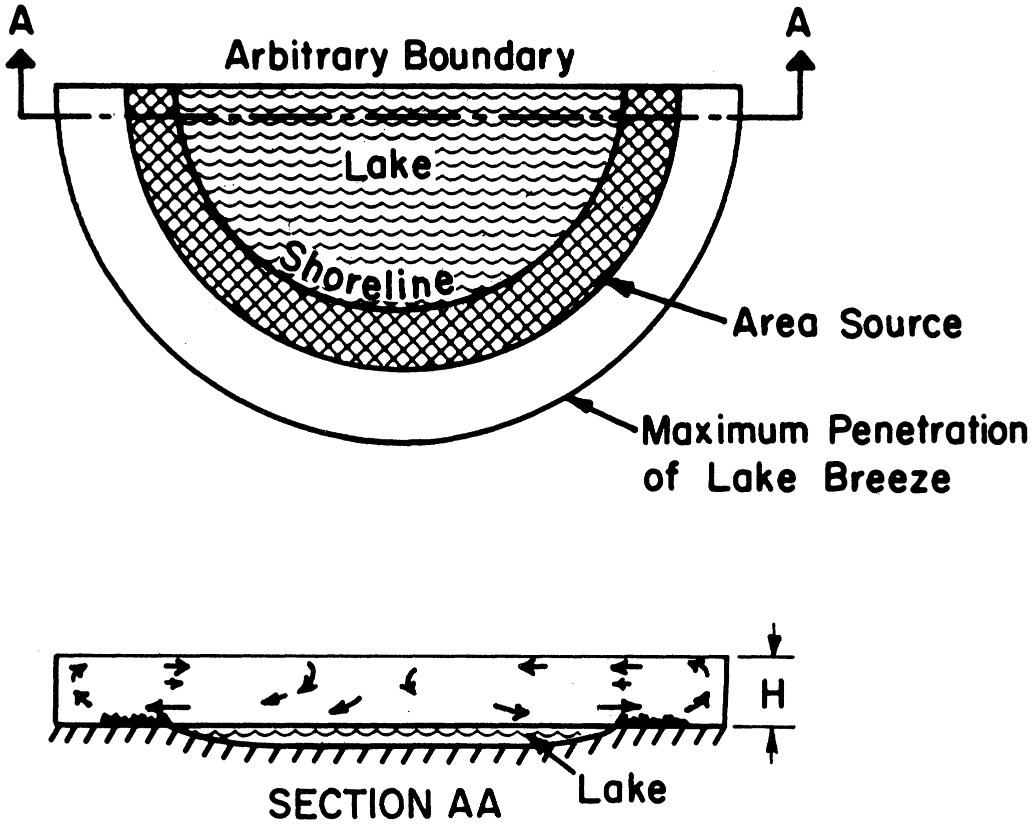

FIGURE 116. Simplified sketch of mixing volume, or hypothetical box, in which pollution from the source area bordering the lake will be dispersed, under zero gradient wind conditions. H = depth of mixing layer. In section A - A the lake breeze circulation is indicated.

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