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SATELLITE & MESOMETEOROLOGY RESEARCH PROJECT

Department of the Geophysical Sciences
The University of Chicago

THE CONTROL OF SUMMERTIME CUMULI AND THUNDERSTORMS BY LAKE MICHIGAN DURING NON-LAKE BREEZE CONDITIONS

by

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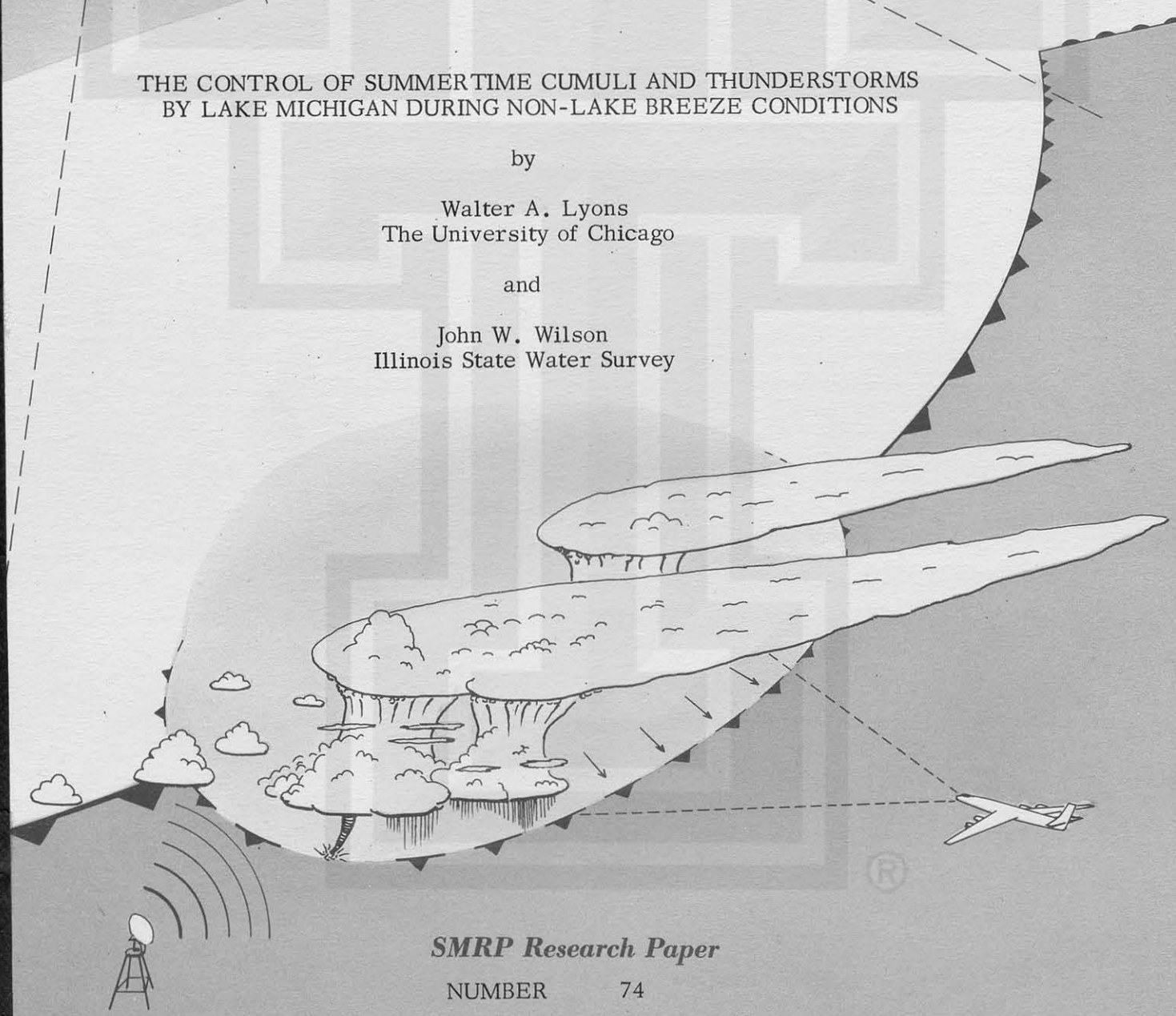

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The research reported in this paper has been partially sponsored by ESSA (MSL) under Grant Cwb WBG-34 (University of Chicago); U.S. Public Health Service Grant No. 5 ROl AP 00380-03 (University of Michigan); and, Atomic Energy Commission Grant No. AT(11-1)-1199 (Illinois State Water Survey, The University of Illinois).

THE CONTROL OF SUMMERTIME CUMULI AND THUNDERSTORMS BY LAKE MICHIGAN DURING NON-LAKE BREEZE CONDITIONS[1]

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ABSTRACT

This paper reports on a one day field study over Lake Michigan (26 July 1966), observing the response of convective nephsystems to the cold water surface during non-lake breeze conditions. As appears typical of these situations, large cumulus clouds dissipated after about 10 miles of over-water fetch with no cumulus reforming further out over the lake. Although all advected cumuli dissipated, a moderately intense squall line was able to cross the southern basin in the afternoon. Gain-stepped radar data indicate no decrease of echo strengths over water. Cloud photographs and surface data also established the existence of, (1) a band of shoreline convergence along the southeastern shore where the prevailing winds paralleled the coast, and (2) a 20-mile wide cumulus-free zone on the eastern shore, apparently marking a zone of subsidence and air-mass remodification.

A simple two-layer model was developed to compute the temperature field over the lake. The lowest layer (about 60 m deep) is defined as that where the over-land superadiabatic layer is replaced by an intense inversion through conduction of heat to the water surface. Various types of evidence lead to the conclusion that small-scale cumulus convection, conceived as continuous travelling plumes, originates within the superadiabatic layer, which is a necessary, though perhaps not sufficient, condition for such motions. Thunderstorms in squall lines and other systems, however, may often have their origins at much higher levels and thereby remain largely uninfluenced by a large lake.

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[1]The research reported in this paper has been partially sponsored by ESSA (MSL) under Grant Cwb WBG-34 (University of Chicago); U.S. Public Health Service Grant No. 5 RO1 AP 00380-03 (University of Michigan); and, Atomic Energy Commission Grant No. AT (11-1)-1199 (Illinois State Water Survey, The University of Illinois).

Above the conduction layer it is assumed that once over water, there is no upward convective transport of heat in an air column, while the surrounding air over land continues to heat. Ignoring compressional warming due to subsidence, assuming hourly increments of heating over the land and knowing the wind field from serial pibals on both shores, the temperature deficits in the subcloud layer over the lake were estimated. Integrating the hydrostatic pressure equation yields an excess pressure in the lake "mesohigh" of about 3.5 mb.

1. Introduction

The effect of topographic and boundary layer differences upon convective cloud formation is a relatively unexplored area of mesometeorology. Braham and Draginis (1960), and Fujita, et al., (1961), produced some of the first well-documented reports showing how the high level heat source of a mountain can control the local convective regime. Malkus and Stern (1953) demonstrated how a heated island in a relatively cold ocean can generate rows of cumuli and act very much like a "thermal equivalent mountain". Yet the reverse situation has been largely ignored. Given a relatively cold lake, surrounded by a heating land mass, what are the resultant effects upon convective clouds?

Among previous studies, Plank (1965) discussed summer cumulus populations over Florida and noted that in 19 flights made under varying synoptic situations, not once was there seen an active cumulus cloud over the 25-mile wide Lake Okeechobee. The same can be said from 15 Gemini photographs of southern Florida inspected by the authors. Moroz (1967) showed that during the warm season, when winds are sufficiently light, a lake breeze will develop over a large lake comparable in strength to the sea breeze.[2] The response of convective clouds to a lake breeze circulation has been fairly well established. Lyons (1967) showed with time-lapse films how the lake and the surrounding shore areas covered by inflowing, subsiding lake air, are completely free of cumuli and cumulonimbi. Convective clouds will either move along the convergence zone or dissipate rapidly while crossing it, depending on whether the large scale flow is parallel or normal to the lake breeze front. Stout and Wilk (1960), using a CPS-9 radar 100 miles south of Chicago, noted that some of 30 squall lines weakened while crossing Lake Michigan. In search of a specific mechanism which might cause such weakening, Lyons (1966) combined radar data and detailed mesoanalyses showing that a squall line can be rapidly dissipated when it intersects a strong

[2]For the two summers of 1966 and 1967, lake breeze conditions occurred along the western shore of Lake Michigan 35% of all days (mid-May through mid-August). A lake breeze is defined as a shift of greater than 30 deg at shoreline stations to onshore flow without a corresponding shift at stations inland.

lake breeze circulation. Thus the lake breeze appears to exercise a significant degree of control upon convection of all scales.

In this paper, by intensive analysis of data collected for a typical summer day (26 July 1966), we wish to examine the response of convective clouds to Lake Michigan during a period of warm advection when geostrophic winds are strong enough to prevent the development of a lake breeze on the western shore. Cole (1967) showed that along the western shore (specifically Milwaukee) there is an off-shore component to the wind 41% of the daylight hours in July. During 1966-67, abnormally cool and dry years, 29% of all summer days were characterized by cross-lake flow with a westerly component advecting cumuliform clouds over the colder water. Figure 1 illustrates the cloud pattern usually seen in satellite photographs obtained under these conditions.

This paper largely represents the results of a one-day field program involving four organizations. By simply coordinating their regular data collection schedules for the first time a relatively complete three-dimensional picture of events over the southern basin of the lake was obtained at virtually no additional cost. The Illinois State Water Survey C-45 aircraft, on a tritium sampling mission, took infrared and color cloud photographs and temperature soundings near and while crossing the lake. The University of Chicago (SMRP) took serial pibals and infrared panoramic pictures on the western shore, while the Mesoscale Winds Project of the University of Michigan took serial pibals and surface data along the eastern shore. Gain-stepped photographs at regular intervals were taken by the WSR-57 radar at the Chicago Weather Bureau Forecast Center, which also supplied the Nimbus II APT photograph (Fig. 2).

The paper is divided into four main sections. First, we will illustrate the complete dissipation of cumulus which are advected off the upwind shoreline. Second, the radar data will show how a squall line, associated with a 500-mb short wave trough, crossed the southern basin during late afternoon, and which, by contrast, showed no signs of dissipation whatever. Third, in light of a simple model for the temperature deficit pattern over a large cold lake, we will show how these observations are consistent with some recent studies on the origins of cumulus convection. Fourth, cloud photographs and surface data appear to verify the Estoque (1962) numerical sea breeze model, (1) along the southeastern shore, where a zone of shoreline convergence was maintained, and (2) on the eastern shore, where a distinct cloud-free zone was present for 20 miles inland.

The study will confirm one's intuitive guess that there should be less cumulus over cold water during warm air advection. The energy budgets of the Great Lakes currently are being investigated to answer questions of vital importance to natural resources

management. Therefore, knowing the degree to which convective cloudiness is suppressed by the lake is of great practical value. In spite of the expected land-lake differences in cloudiness, several authors (Rogers and Anderson, 1961; Bruce and Rogers, 1962; Derecki, 1964) have been forced to use pyrheliometric data from land stations in their energy budget and evaporation studies of the Great Lakes. As Bellaire (1965) points out, even a ten percent change in insolation over a lake would greatly affect the ratio of conduction/evaporation to radiation terms as factors in heat input into the water. Derecki (1964) noted that after air-water temperature difference, percentage of sunshine appeared to be the most important factor in determining Lake Erie's evaporation.

With the growth of the controversy on the effect upon lake water levels by the diversion of water by large cities, the question of land-lake precipitation differences is even of greater importance. Since the first estimates of Freeman (1926), estimates have ranged from zero difference (Brunk, 1962) to 21 percent less annual rainfall over Lake Michigan (Hunt, 1959). A recent report by Changnon (1967), based upon land and lake station climatological data, concludes that Lake Michigan receives 14 percent less rain in the summer than does the surrounding land. What we are attempting here is to further specify what mechanisms, if any, might be acting to decrease precipitation over the lake for various types of convective storms.

2. Observations on 26 July 1966

A. Cumulus Distribution Prior to Squall Line Passage.

On 26 July 1966, warm southwesterly geostrophic flow was present over southern Lake Michigan, with a stationary front north of Milwaukee. At 0600 CST, a 500-mb short wave trough over Minnesota was moving southeast at 30 kt, touching off numerous thundershowers in Iowa and Minnesota.

The C-45 flew from Champaign, Illinois to just north of Chicago between 0930 and 1125 CST. Infrared photographs (Fig. 2b) taken near the western shore of the lake at 1051 CST, 24 minutes prior to the Nimbus II APT photograph are shown in Fig. 2a. Cumulus formed about 0930 CST in the Chicago area and rapidly grew into the broken cover of swelling clouds shown in the ground panoramic in Fig. 2c. The aircraft pictures confirm that the lake was indeed free of cumuli, and that the clouds, some with tops as high as 10,000 feet, were advected only about 10 miles out over the water. Also visible in Fig. 2b is a line of buildups along the southeastern shoreline, with smaller cumuli present further north along the eastern shore. The flight continued east-northeastward across the lake and then north along the eastern shore from 1250 to 1345 CST.

Figure 3 is a view from 3,000 feet looking south along the shoreline towards Chicago at 1258 CST and portrays the large cumuli advecting over the lake. By 1304 the plane was 10 miles offshore at 6,300 feet. Figure 4 shows the remnants of the cumuli on top of the haze layer as they were dissipating. Figure 5, an infrared photograph taken from 9,500 feet looking northwest over mid-lake, strikingly shows the lake completely clear of clouds, with cumulus dissipating offshore and several small thundershowers south and southwest of Milwaukee which had formed ahead of the developing squall line.

By plotting a summary of the low-level cloud types with greatest vertical development (exclusive of cumulonimbus, which were mostly associated with the squall line) reported by the airway weather observations during the day (Fig. 6), one can see that the cloud-free region over the lake and along the downwind shore could be expected after noting that Muskegon (MKG) reported no convective clouds at all, and Benton Harbor (BEH) and Manistee (MBL) could only indicate buildups inland during the afternoon.

B. Passage of the Squall Line Across the Lake.

By mid-morning, surface reports and Chicago radar indicated that the thundershowers in western Illinois were beginning to organize into a definite squall line mesosystem. Figure 7 shows the 1100 CST mesoanalyses, roughly concurrent with the satellite photograph. During the next several hours, the squall line developed and moved east-northeastward, with echo tops reaching to 40,000 feet, and producing 50 kt surface wind gusts. By 1530 CST, the northern half of the squall line began pushing off the western shore. In the 1700 CST mesoanalysis (Fig. 8), radar indicated a solid line of thunderstorms of moderate intensity over mid-lake. No cumuli were as yet reported over the eastern half of the lake by surface or aircraft observations. The squall line continued into Michigan and Indiana and only after 1900 CST (Fig. 9) did the cells, both those which had and had not passed over the lake, begin to weaken rapidly.

Gain-stepped photographs were taken at roughly 25-minute intervals on the Chicago WSR-57 radar. After applying a rather crude range normalization scheme to all echoes within 100 miles, the areas covered by any echoes at full receiver sensitivity ($Z_e = 10^{-1}$ mm^6 /m^3 or less), by light showers (those areas within a $Z_e = 10$mm^6 /m^3 contour), and by moderate thunderstorms ($Z_e = 10^3$ mm^6 /m^3) were mapped (Fig. 10). The above indicated that many of the heavier echoes were actually located over the lake. While it cannot be determined if the precipitation would have been more intense had the lake not been present, this case hardly resembles the complete annihilation of a vigorous squall line which attempted to cross the lake during lake breeze conditions reported by Lyons (1966). Many other squall lines, such as the one producing the 1965 Palm Sunday Tornadoes (when the lake was 35F colder than the air), and many observed during the summer of 1967, have crossed the lake with little or no apparent modification.

3. Modeling Mesoscale Temperature and Pressure Fields over the Lake

In the above sections we noted that cumulus formation is markedly suppressed over the lake whereas the effect upon thunderstorm cells in a moderately intense squall line is so negligible as to be undetectable by radar. In order to arrive at a possible explanation for this apparently differential aspect of the lake on convection of different scales, we must first try to estimate what mesoscale perturbations in atmospheric structure are induced by the lake.

The following model is preliminary, but will at least serve to point out the major processes believed to be involved. We will break the subcloud layer up into two layers. The lowest, less than 100 m deep, is the superadiabatic layer over the heating land, or the region of significant conductive heat loss to the cold surface over the water. The upper layer, from about 100 m to cloud level, is one where temperature change is brought about mostly by convective heat transport from below and adiabatic expansion or compression due to vertical motions.

The basic structure of the lower layer has already been described in the wiresonde observations of Bellaire (1965), which show that an intense conduction inversion rapidly develops in about the lowest 60 m when warm air flows over the underlying cold surface. While no wiresonde data were available for the 26 July 1966 case, it appears that the equation for simple Fickian diffusion of heat to the surface gives a good estimate of the low-level temperature structure over the cold water. The temperature over water $T_W(z, s)$ at a given height z and fetch s is found from the equation

$$T_W(z,s) = (T_L - \alpha\Gamma z) - (T_L - T_W)\left\{1 - P\left[\frac{z\sqrt{\bar{u}}}{K_H s}\right]\right\} \qquad (1)$$

where P indicates a probability function, T_L the surface temperature over land, and $\alpha\Gamma$ the lapse rate in the lowest layers in terms of multiples of the dry adiabatic lapse rate Γ. T_L and α were obtained from the micrometeorological tower at the Argonne National Laboratory, 40 miles southwest of Chicago. T_W, the lake water surface temperature, was assumed to be constant in the solution of the original partial differential equation. Church (1945) indicated that by July water temperatures over the southern basin tend to become rather uniform in the horizontal. All available ship and intake crib temperature reports for 26 July 1966 show little variation and we will assume the surface water temperature of the lake south of Milwaukee to be uniform at 22C for purposes of calculation. As for $\bar{u}$, the mean surface boundary layer wind, the marked thermal stratification over water causes wind speeds near the surface to drop to well below values over land, with a streaming effect above the inversion surface. $\bar{u}$ was chosen to be 3 mps for the lowest 60 m, or 50 percent of the overland readings.

K_H , the eddy thermal diffusivity, can vary over several orders of magnitude within the boundary layer. Here, K_H was selected to be 5×10^2 cm^2 /sec, certainly several orders of magnitude less than that over the heating land. This will probably serve as a good average over water, inasmuch as Bellaire's observations give strong evidence that turbulent transport drops markedly after less than 20 miles of over-water fetch. Moroz (1967) used this value as a minimum near the lake surface in his lake breeze model.

By way of illustration, Fig. 11a represents the buildup of a conduction inversion over water using the above values of constant T_W and K_H . In essence, modification of a single column of air after it leaves shore is calculated. Figure 11b represents the field of computed temperature loss by conduction to the lake surface out to a fetch of 75 km at 1300 CST. It was assumed that air leaving the shore at 0600 CST had $T_L = T_W$ (land temperature and water temperature equal), with hourly increments of T_L and α . Thus, in the conduction layer, in order to completely specify the land-lake temperature deficit at a given time (here, 1300 CST), one subtracts the computed values in Fig. 11b from the temperature which the parcel had when it left shore, and then subtract this over-water temperature from the observed Argonne tower temperatures at 1300 CST. The final result is shown in Fig. 11c. Note that the conductive heat loss is only significant in about the lowest 60 m.

We will now discuss a simple model for temperature deficits over the lake in the layer from 60 m to the top of the mixed layer. In general, the major factor causing temperature increases in the lowest few thousand meters would be (1) heat transported upward from the surface layers by the mechanism of free convection, and (2) adiabatic compression resulting from subsidence caused by a diverging wind field over the lake. Our major assumption will be that once air moves from the heated land to the colder lake, virtually all convective motions terminate almost immediately. Thus, for any parcel of air above the conduction layer,

$$D\theta_W(z)/Dt = 0 \qquad (2)$$

where $\theta_W(z)$ is the potential temperature of an air parcel traveling over water. To this time, there is no study available which has estimated or measured the intensity or pattern of divergence over a large lake during non-lake breeze conditions. For this paper, as a first estimate, it is assumed that all air flow over the lake is horizontal, and therefore, any air parcel above 60 m would be denoted by,

$$DT_W(z)/Dt = 0. \qquad (3)$$

Expanding Equation 3 and expressing in natural coordinates, we find

$$\frac{\partial T_W(z,t)}{\partial t} = -u(z)\frac{\partial T_W(z,t)}{\partial s} \qquad (4)$$

that is, the local change in over water temperature $T_W(z)$ at any given level can be described purely in terms of horizontal temperature advection. The gradient of temperature over the lake can be easily estimated if one knows the wind speed over the lake and changing values of $T_L(z,t)$, the temperature at any level above the heating land of the air that advects past the shoreline.

Specifically, for 26 July 1966, due to the homogeneous character of the air mass surrounding the southern portion of the lake on this day, we can assume uniform heating of the triptosphere over the entire region surrounding the lake. The initial temperature sounding $T_i(z)$ at 0600 CST[3] is shown in Fig. 12. Due to lack of convective modification over the cold water surface, save for advection of warm air from land during the day, the sounding over the lake should remain as $T_i(z)$ all day. Since serial radiosondes are not available, the hour-by-hour heating in the lowest layers was estimated as shown in Fig. 12. Convective mixing reached the lowest condensation level (LCL) just before 1000 CST, and the adiabatic layer increased to about the 2,000 m level by 1300 CST.

The calculated land-lake temperature differences $\Delta T(z,t)$ at 1300 CST for several levels are shown in Fig. 13. These were obtained by the relationship

$$\Delta T(z,t) = T_L(z,t) - T_W(z,t) =$$

$$= \left\{+\int_{t=0600}^{t=1300}\frac{\partial T_L(z,t)}{\partial t}\,dt - T_i(z)\right\} - \left\{-\int_{t=0600}^{t=1300}u(z)\frac{\partial T_W(z,t)}{\partial s}\,dt - T_i(z)\right\} \qquad (5)$$

which was evaluated using one hour time steps and the six-hourly mean vector winds obtained from pibal data, the streamlines of which are shown in Fig. 13.

With the temperature field over the lake computed, by integrating the hydrostatic equation downwards from the level of negligible heating, assumed to be 2,000 m, the excess pressure over the lake can be estimated. The land-lake pressure difference at a given pressure level can be computed by

$$\Delta P_i = P_{W_i} - P_{L_i} = P_{W_{i+1}}\,EXP\left[\frac{g\Delta z}{R\,\overline{T_W(z)}}\right] - P_{L_{i+1}}\,EXP\left[\frac{g\Delta z}{R\,\overline{T_L(z)}}\right] \qquad (6)$$

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[3]The 0600 CST Dayton radiosonde report was used to estimate $T_i(z)$ because this was the only area station not in a frontal zone or having a rain shower during the day.

Here the equation was evaluated in 250-m steps. Figure 14 shows the field of excess pressure at the surface at 1300 CST, when the magnitude of the mesohigh was close to its maximum. The peak excess pressure of 3.5 to 4.0 mb near the eastern shore closely fits values measured under similar conditions by University of Michigan research vessels. The field of excess pressure is a function of time, wind field, and convective mixing. It generally builds during the morning hours but begins to decay in the afternoon when heated air from the western shore begins to arrive at the eastern shoreline. Less than 0.5 mb of the total excess pressure is due to the conductive heat loss to the lake, the remainder being the result of lack of convective heat transport over the lake while the surrounding land is heating.

This result clearly explains the frequent observation over the Great Lakes in summer that winds instead of veering over the water in response to less surface friction, often back by as much as 40 deg in adjusting to the excess pressure field (note the south-southeasterly winds over mid-lake in Fig. 1). Furthermore, this appears to be a smaller-scale negative analogue to the distortion of the pressure field by the heating effects of the Great Lakes during winter, as reported by Petterssen and Calabrese (1960).

The latent heat input into the atmosphere by the lake during winter unstable conditions is a major factor to be considered. However, during spring and early summer, rather than evaporation from the lake, condensation from the moist air onto the cold water is the common occurrence. On 26 July 1966, the high dewpoints in the air mass indicate that relatively little evaporation would occur on this day. Furthermore, any increase in water content of the air would probably be restricted to the conduction layer where the high thermal stability would greatly inhibit vertical transport.

4. Theoretical Implications of the Observations

A. Cumulus Dissipation Over a Cold Lake.

We now wish to discuss what the above observations appear to confirm about some recent researches on the nature and origins of cumulus-scale convection.

The 26 July 1966 and the 5 June 1966 cases (shown in Fig. 1) are typical of many others which have been observed. From inspection of such satellite pictures an interesting question soon arises. On 5 June 1966 moderate southwesterly flow prevailed over the Great Lakes. Cumulus began forming in the Chicago area around 1000 CST and soon showed a definite banded structure paralleling the 5,000-foot flow. The Nimbus II AVCS photograph taken at 1113 CST (Fig. 1) indicates that these clouds were advecting no more than 10 or 15 miles off-shore before dissipating, with no new cumulus forming over the lake. This situation remained

essentially steady state through the day, since the same cloud pattern was noted by one of the authors at 1430 CST while flying on a commercial flight across the lake. Cloud streets such as these have been attributed to the presence of mesoscale roll vortices operating through the depth of the adiabatic subcloud layer, Faller (1965). Byers (1963) is among those who have suggested that most smaller cumuli are the result of eddying motions such as a mesoscale horizontal roll vortex, crests in waves, or random turbulence, at the top of a deep adiabatic layer which lifts air a short distance to the condensation level. Therefore, cumuli usually would not mark the arrival of specific buoyant parcels from near ground level. As plausible as this idea seems, the cumulus pattern shown in Fig. 1 raises an intriguing question: if the above suggestion is correct, where there clearly appears to be an organized circulation in the subcloud layer sufficient to raise warm air to the condensation level, just what happens in less than 20 miles of transit over the lake to completely disrupt the cloud-producing mechanism? The analysis of the 26 July 1966 data will attempt to answer this query.

Extensive time-lapse cloud photography in the Chicago area during the summers of 1966 and 1967 have shown that in cases of offshore flow cumuli dissipate roughly in the time required for a cloud to complete its normal life cycle of growth and decay rather than with the extreme rapidity which is normally observed during lake breeze conditions, and attributed to intense subsidence and wave motions in the return flow region. It does not seem likely that in the offshore flow case there are "shoreline downwash" effects, which even if they exist, are strong enough to disrupt a vigorous roll vortex circulation producing cumulus scale convection. In fact, the streaming of the flow over the conduction inversion, plus some indications from Estoque's model, would suggest that there might be some low level convergence and slight upward motion for the first few miles offshore.

What the time-lapse films have indicated, however, is that many cumuli are that portion of a continuous traveling plume, originating within the surface superadiabatic layer, that penetrates the condensation level. On numerous mornings, the dense industrial pollution of the Chicago area is trapped below the nocturnal inversion. Plumes of smoke of more or less cylindrical shape, or even tapering with height, can be seen rising as heating progresses and the superadiabatic layer deepens beyond some critical depth. Such a plume is shown in Fig. 15 in a photograph taken from a plane flying near downtown Chicago. These plumes become numerous and more vigorous with continued heating, eventually penetrating to above the condensation level; that portion of the air in the plume rising through the lowest condensation level (LCL) is the visible cloud. These plumes, due to transport of the low-level horizontal momentum in a sheared atmosphere, stand into the wind at higher levels. The air

flows around and over the thermal, eroding portions of the cloud in the downwind direction. This produces the apparent "rotation" of the cumulus about a horizontal axis and accounts for the frequent observation of the "hard" side of the cumulus being on the upshear side.

On many occasions, such cumulus-topped smoke plumes have been seen moving out over the lake. The air that has already ascended out of the superadiabatic layer continues to rise in the plume, allowing apparent cloud growth for some short distance over the lake, but no new smoke columns are seen ascending from near the surface after several miles of over-water passage. It is to be emphasized that we are not stating that cumuli originate from any given "hot spot" on the ground, but rather from within the superadiabatic layer. The important fact is that plume formation is observed to cease after a few miles of travel over the lake at average wind speeds. This is just the distance computed above for conduction to eliminate the superadiabatic layer and replace it with the conduction inversion. Thus, by using the lake as a "natural laboratory" we see that cumulus formation is primarily suppressed by the elimination of the superadiabatic layer, a necessary, though not perhaps sufficient, condition for this scale of convection.

Both radar and aircraft data exist which lend further support to this suggestion. The powerful 10-cm radars operated by Johns Hopkins University at Wallops Island, Virginia, have upon numerous occasions photographed echoes apparently related to refractive index gradients in clear air thermals, (Hardy and Glover, 1966). Figure 16 is an RHI scan looking southwestward along the heated peninsula near Wallops Island showing a series of inverted cone-shaped returns rising from the ground level to about 5,000 feet, the base of an inversion. These are thought to mark refractive index gradients at the boundaries of thermals best described as continuing plumes resembling the smoke plume in Fig. 15. A PPI representation of the same phenomena in Fig. 17 reveals something even more interesting. With northwesterly flow at low levels, lines of hollow, more or less circular, clear air echoes about 500 - 1,000 m across, formed in the early morning with the onset of heating. It is evident that these echoes dissipate after less than 5 miles of travel over swamp and open water. Most likely the continuing plumes, and their associated refractive index gradients, are organized into lines by the typical roll vortex structure of the boundary layer. It is assumed that the plumes extend through a layer approximately 1,500 m deep and move with the wind. After a short passage over water the superadiabatic layer is destroyed, plume formation ceases, and the remaining buoyant air rises, halted finally by the inversion layer aloft. Assuming a mean vertical motion in the plume of about 1 mps and a wind speed of 5 mps, 4 miles is approximately the distance needed to evacuate all refractive index gradients from the lower levels after warm air flux onto the base of the plume ceases.

Over the last four years, work undertaken in Australia has produced from extensive aircraft data a model of cumulus scale convection identical in almost all qualitative aspects to the one postulated here from visual observations and deductions. Warner and Telford (1963, 1967) have demonstrated that most thermals, under conditions of light to moderate winds, are continuing plumes which almost certainly organize within the surface superadiabatic layer, 200 m deep or less. They rise out of the superadiabatic layer with a typical excess temperature of 1C and vertical velocity of 1 mps, and are 200- to 400 m or more across, are approximately cylindrical in shape, and show little increase in horizontal dimension as they rise. Inasmuch as they are rising into a sheared transporting environment carrying considerable low-level horizontal momentum, they will lean into the wind, causing flow around their sides and over the rising tops, accounting for the erosion of buoyant material in the downwind direction and the rolling motion discussed previously. These thermals appear during the day over any type of heated surface, but have never been observed over water colder than the air, though they have been seen over sea with a temperature warmer than the air when it left shore. So sensitive does the plume formation appear to be to changes in the superadiabatic layer that it has been observed that beneath the shadows cast by the first cumuli which form during the day, the low-level temperature pulsations, the identifying signature of the plumes, disappear almost immediately. The original cloud pattern can then be seen changing in readjustment to this temporary elimination of an area of superadiabatic lapse.

Thus, evidence from cloud photography, high power radar, and aircraft measurements indicates that convective plume formation ceases almost immediately upon the removal of the underlying superadiabatic layer. While subsidence over the lake would certainly make cumuli dissipate somewhat faster than normal, it appears, until definite data can show otherwise, to be only a second-order effect in preventing cloud formation over a large lake.

B. Thunderstorm Passage Over the Lake.

As mentioned above, the cells associated with the squall line that moved across the southern basin in late afternoon showed no sign of weakening on the Chicago radar. There have been other occasions when aircraft photographs and ground time-lapse photography have clearly shown small cumuli dissipating as they move over the lake--and yet, simultaneously, large thunderstorms both advecting and/or forming over the cold water.

The air-mass thundershower is generally considered to be the mature stage of normal daytime cumulus convection occurring under rather unstable conditions. Certainly, if no cumuli exist over the lake, then the true air-mass shower must be absent too. In fact, on 26 July 1966, some scattered showers, apparently of the air-mass type, did break out

around, but not over, the lake. Pearson (1958) has several graphic radar case studies of this effect. Changnon (1966) notes that from 1200-1800 CST, Muskegon, Michigan, in the average summer has less than 4.0 thunderstorms compared to an average of from 7.5 to 8.5 along the western shore. The 24-hour average though shows 17.0 at Muskegon and 21.0 at Chicago. These figures reflect probable suppression of air-mass types, but further indicate that many other thunderstorms and thunderstorm systems can cross the lake with little difficulty.

The point is that the originating levels of convection for thunderstorms of the non air-mass kind may be higher than the superadiabatic layer. Fujita and Bradbury (1966) discuss the concept of Best Lifted Index (BLI) which shows that often layers other than the surface provide the greatest potential instability, given the proper triggering mechanism. Thus the initiating motions of many convective storms, such as those associated with Positive Vorticity Advection (PVA) in 500-mb short wave troughs, frontal lifting, or convergence in low level nocturnal jet streams, can be many thousands of feet above the lake's surface. The lake then would be expected to exercise little control on their formation and movement.

6. Meteorological Regimes on the Southeastern and Eastern Shores

So far this paper has concentrated on the convective clouds which move off the windward shore of the lake during periods of warm, brisk flow having a westerly component. In order to complete the picture of the effect of the lake upon convective clouds during these conditions, we will now focus attention on the presence of, (1) a strip of shoreline convergence along the southeastern shore, and (2) a cloud-free zone along the eastern shore believed to mark a region of subsidence.

A. Shoreline Convergence Zone.

When west through southwest winds are strong enough to prevent a general lake breeze around Lake Michigan, there exists an area where the general flow is tangent to the shoreline. This is usually located somewhere between Gary, Indiana and Benton Harbor, Michigan. Estoque's (1962) model for parallel shore flow for a 5-mps geostrophic wind indicates that shoreline frictional convergence and the land-lake temperature gradient should combine to produce a definite convergence zone associated with strong vertical motions that moves inland about 10 miles by late afternoon.

On 26 July 1966, the C-45 approached the shoreline just north of Benton Harbor at 1326 CST and took the infrared photograph shown in Fig. 18. Using cloud shadows visible on the ground in color slides, the wall of towering cumuli paralleling the shore was found to be 4 to 5 miles inland. This line of cumuli near the shore was found only along the 40-mile

section of coast where the average surface wind streamlines (1100-1500 CST) were essentially parallel to the coast. Mean shoreline and inland winds indicate that there was about a 40 deg confluence of the streamlines in this zone. Using a smoothed isotach-streamline field, divergence was calculated, yielding values as high as -45×10^{-5} sec^{-1} . These values are comparable to those found along windshift zones in squall line mesosystems. The estimated divergence pattern is plotted in Fig. 19. The cloud line fits very closely the area bounded by an isopleth of $-30 \times 10^{-5} sec^{-1}$. Note that it is only over the extreme southeastern corner of the lake that surface winds veer over water. Warm air that has just left the heated land is overhead, and values of excess pressure are quite small (Fig. 10).

A most interesting feature further confirming the strip of shoreline convergence was supplied by the Chicago radar. At 1505 CST, exactly in the position of the towering cumuli shown in Fig. 18, several small thundershowers appeared. They moved along the convergence zone for about 45 minutes and then broke away into the general flow, dissipating near Grand Rapids about two hours later. Aside from the squall line, these were the only showers in southwest lower Michigan during the day. It would seem that the shoreline convergence was the significant factor in their formation. Similar patterns of cloud lines and thunderstorms have been seen on several occasions, although it must be remembered that the tangent point is found at different points along the shore with different geostrophic wind directions.

B. Air-Mass Modification Along the Eastern Shore.

Flying north parallel to the eastern shore, the C-45 took additional color and infrared photographs which indicate that a radically different cloud pattern existed north of the shoreline convergence region (Fig. 21). There was a distinct cumulus-free zone extending from the shoreline to about 20 miles inland at 1335 CST. Using the University of Michigan hygrothermograph data taken at points on the shoreline, 5, 10, and 15 miles inland, plus other data, the reheating of surface air can be seen in Fig. 20 to have been essentially completed within 10 to 15 miles off shore, indicating the rapid buildup of a new superadiabatic layer above the heated land. This was presumably the source region of the cumuli developing on the eastern shore. One could expect that the lakeward edge of the clouds moved closer to the shore during the morning and afternoon as heating increased. However, cloud observations made at the 10-mile station show that the cumulus boundary moved progressively further inland during the day. The reason for this can be found again by inspection of Estoque's model. North of the convergence zone, above the lowest layers where marked backing of the winds was occurring over the lake, the flow was essentially normal to the shoreline. Inland heating resulted in an acceleration of the flow in the sub-cloud layer, inducing subsidence in a strip

along the shore which increased in magnitude during the day. The increasing stability below the LCL due to the continuing subsidence, and the arrival of warmer air from the western shore after 1400 CST, presumably inhibited the growth of the plumes to the condensation level. An analysis of the serial pibals taken at the shoreline and 10-mile stations, using differential horizontal motions of balloons released concurrently, indicated that there was indeed a definite downward motion in the subcloud layer of 1 to 3 cm/sec, which intensified during the day.[4] Time-lapse film on the western shore under analogous conditions of easterly onshore flow have frequently shown cumulus forming about 5 miles inland during the morning with their formation zone pushing rapidly inland as heating, and presumably subsidence, increased.

The apparent discontinuity in surface winds along the shore resulted from downward mixing from about the 500-m level, where winds had backed only a few degrees while over the lake. This effect began about 2 miles downwind from the shore. Note the southerly flow along the shoreline and the stronger southwesterly winds at Muskegon and the 5-mile station. Pyrheliograph traces for shoreline (GHN) and Grand Rapids (GRR) show that no cumuli were present along the shore all day, while the cloud pattern 40 miles inland remained essentially unaffected by the lake.

7. Conclusions and Future Research

In essence, a large lake has been used as a natural laboratory for convection studies. By assuming a constant surface water temperature over the lower basin of Lake Michigan, the use of a simple heat diffusion equation, which closely models observed low-level temperature structures over the lake, indicates that the destruction of the surface superadiabatic layer and its replacement by a conduction inversion during periods of offshore flow occurs on exactly the same time scale as the observed cessation of cumulus convection. This provides further strong observational support for the conclusion of Warner and Telford (1967) that under most conditions convective plumes of cumulus scale organize nowhere else but in the superadiabatic layer.

The frequent appearance of thunderstorms over the cold lake during summer indicates that many larger convective clouds may be initiated at levels well above the superadiabatic layer.

An instrumented aircraft could easily test the proposed model for the thermal structure over a large cold lake by measuring $T_w(z, t)$. Assuming that there is no upward heat transport allows one to calculate the temperature field over the water with knowledge of the

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[4] Analysis performed by Mr. Pieter J. Feteris of the Illinois State Water Survey.

over-land sounding and Lagrangian air trajectories since DT/Dt = 0 above the conduction level, which is limited to 60-100 m above the water. Temperature soundings made by the C-45 at 1330 CST off Muskegon, Michigan, indicated that the subcloud layer was still essentially unaffected by daytime heating at this time and location. Further observation may be necessary to establish how serious an omission it may be to ignore adiabatic compressional warming. This, of course, will be a function of (1) the intensity of over-water subsidence, and (2) by what degree the sub-cloud lapse differs from the dry adiabatic.

The calculated excess pressure field over the lake explains the frequently observed south-southeast winds over the cold lake during periods of strong southwesterly geostrophic flow. A relatively simple computer model could show the time evolution of temperature deficit and excess pressure patterns under varying heating rates, wind patterns, and stabilities.

This case, as well as several other similar ones, will be useful in the construction of a meaningful cloud climatology of the southern basin of Lake Michigan using satellite cloud photographs, which do not always possess sufficient resolution to provide unambiguous identification of cloud types and amount.

Richards and Loewen (1965) attempted to directly evaluate summertime land-lake insolation differences by comparing four years of shipboard pyrheliometric data with simultaneous readings at three land stations. Their findings indicated monthly averages showing as much as 1.56 times more insolation over the lake. This value might be low, because the three land pyrheliometric sites were located quite close to the shore. If a lake breeze exists, according to Lyons (1967) the shoreline areas will normally be free of cumulus even though vigorous convection may be occurring further inland. Furthermore, as the current study shows, when there is no lake breeze, any station within 20 to 25 miles of the downwind shore may be within the cloud-free zone. Thus, since their "inland" cloudiness is influenced by the lake, land-lake insolation ratios may be somewhat higher than 1.56.

Daytime shoreline convergence zones probably exist frequently somewhere along the southern or southeastern shore of Lake Michigan under non-lake breeze conditions. Research currently being carried on by Changnon (1968) has found a marked increase in precipitation in the La Porte, Indiana area, which has become more pronounced with time over the last several decades. Inasmuch as La Porte is at the expected mean position for the shoreline convergence zones and directly downwind of the Gary, Indiana steel mill complex, known producers of vast quantities of ice and condensation nuclei, (Langer, 1968), another interesting area of research has been opened up.

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Our indebtedness is also acknowledged to Mr. Glenn E. Stout and Mr. Stanley A. Changnon, Jr. of the Illinois State Water Survey, and Prof. E. Wendell Hewson, the University of Michigan for their constructive criticisms and manuscript reviews.

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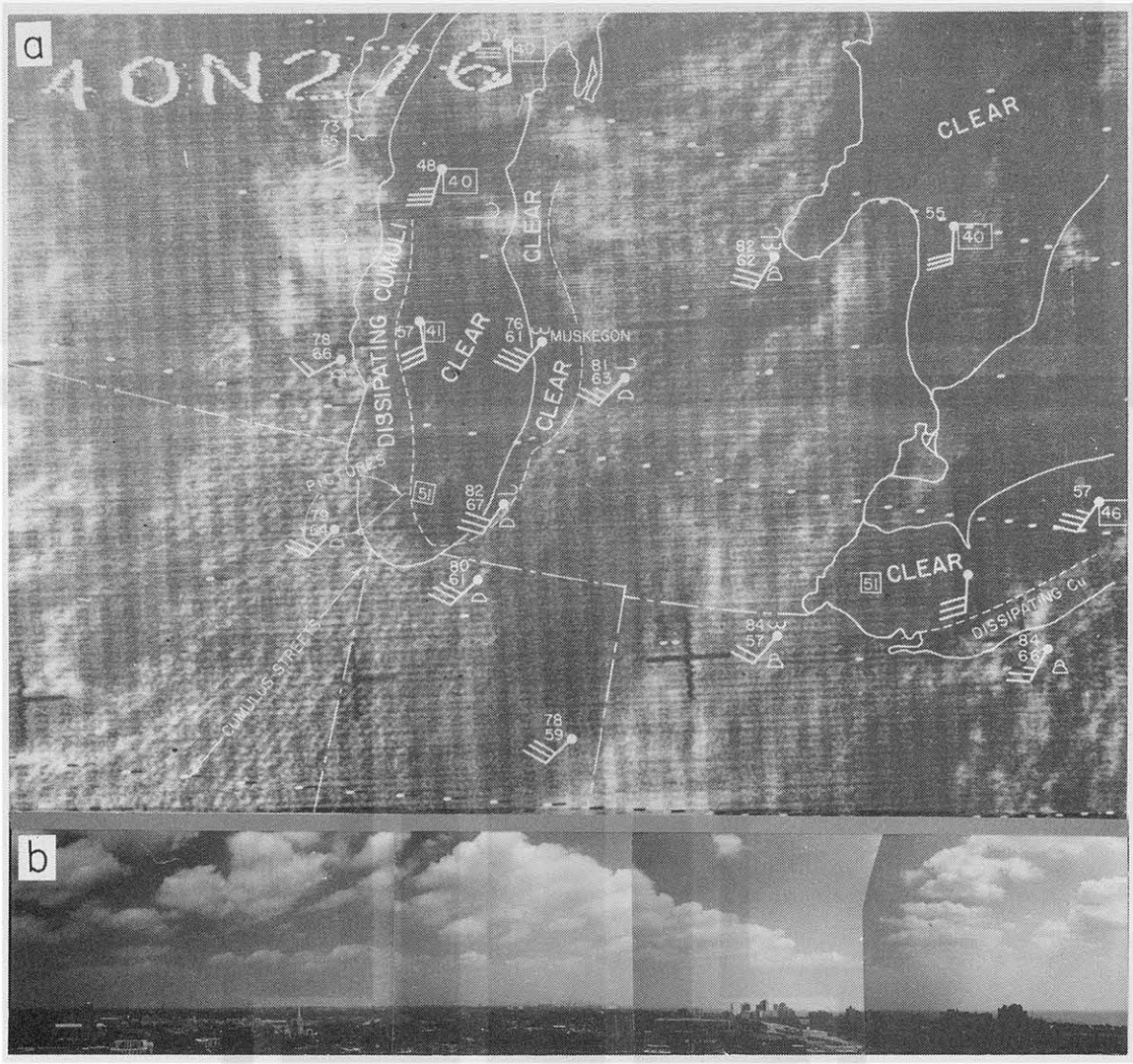

Fig. 1. (a) Nimbus-II AVCS photograph, Orbit 285, 1113 CST, 5 June 1966, showing cumulus dissipation over Great Lakes. Observed temperature, dewpoints, cloud types and wind arrows (one barb = 5 knots) are plotted for 1100 CST. Water temperatures are given in boxes. (b) Portion of panoramic photograph taken from 250-ft tower on Chicago lakefront at 1112 CST, 5 June 1966, showing cumuli, plainly organized into rows, moving northeastward (arrow) out over Lake Michigan.

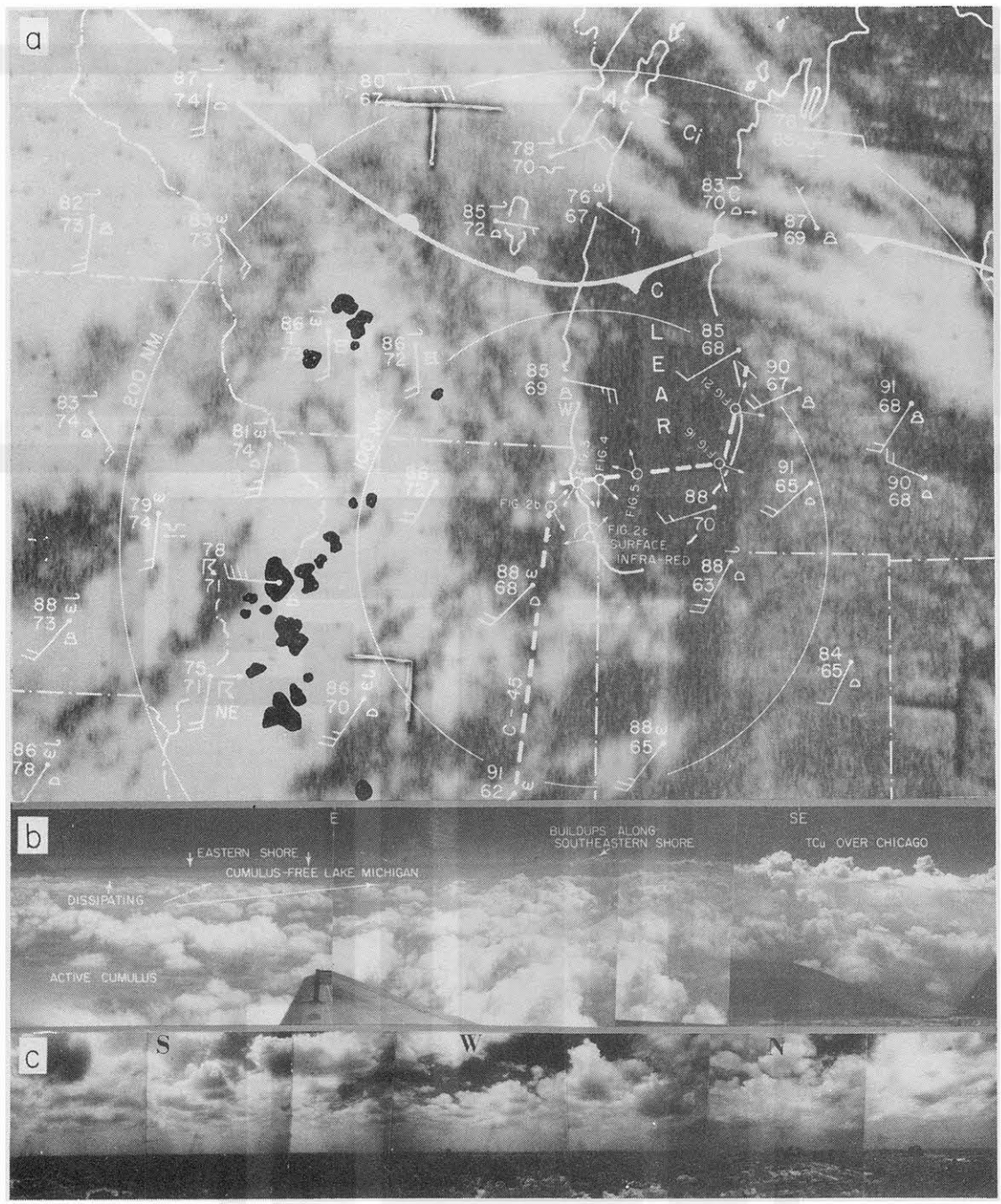

Fig. 2. (a) Portion of Nimbus-II APT photograph, Orbit 964, 1115 CST, 26 July 1966, showing cumulus-free southern basin of Lake Michigan. Airway data for 1100 CST represented as in Fig. 1(a). Radar echoes (solid black) at 1114 CST and 100-mile range markers from Chicago WSR-57 are superimposed. Path of C-45 flight also marked. (b) Infrared photomosaic taken at 1051 CST from C-45 at 8500 ft (all heights MSL) 10 miles north of O'Hare Airport showing the clear lake and towering cumuli along southeastern shore. (c) Portion of panoramic infrared photomosaic taken from 250-ft tower on Chicago lakefront at 1112 CST showing appearance of cumuli from below as they moved northeastward over the lake.

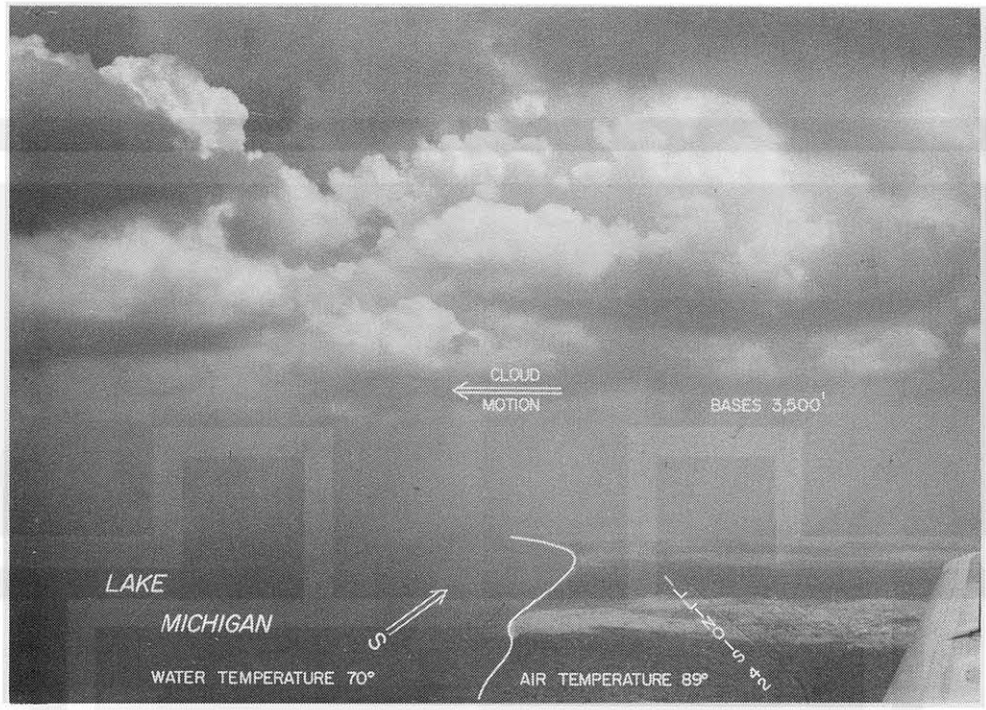

Fig. 3. Crossing shoreline at 3000 ft looking south towards Chicago at 1258 CST showing large cumuli advecting offshore.

Fig. 4. Ten miles offshore at 1304 CST, showing the remnant cumuli on the top of a dense haze and moisture layer (the triptopause) at 6300 ft.

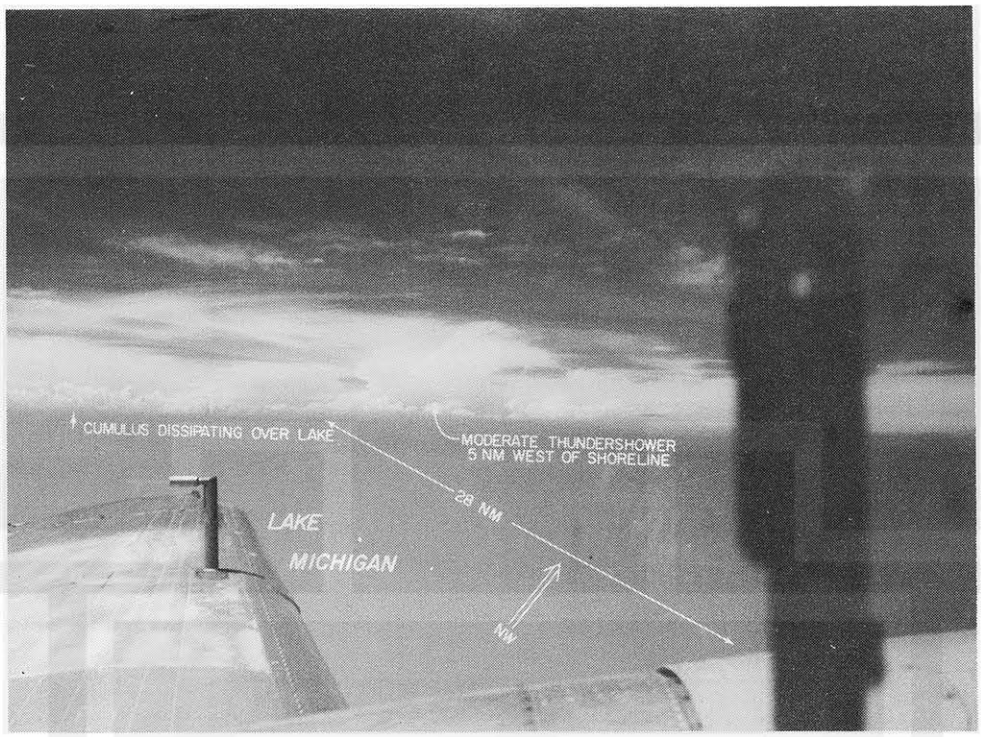

Fig. 5. At 9500 ft over midlake at 1314 CST showing completely cumulus-free lake, with zone of dissipation for about the first 10 miles offshore and several small thundershowers over land well ahead of the main squall line.

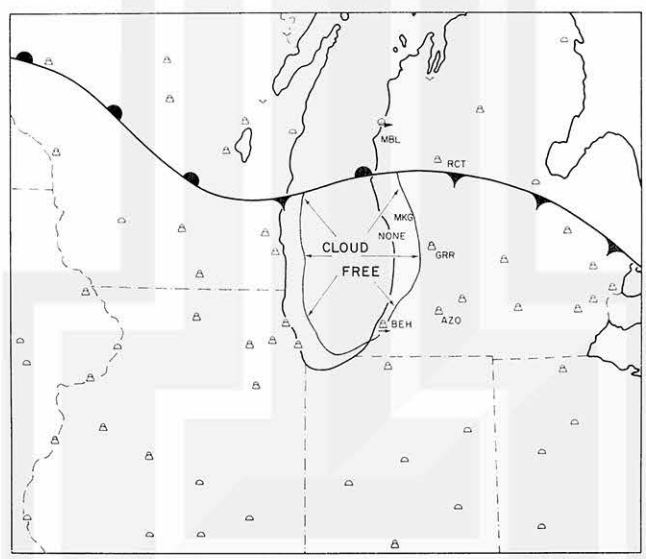

Fig. 6. Map showing type cloud with maximum vertical development (exclusive of cumulonimbi which were mostly associated with the squall line) observed at regular airway stations from 0800 to 2000 CST, 26 July 1966. Note that the only clouds visible along the eastern shoreline were inland.

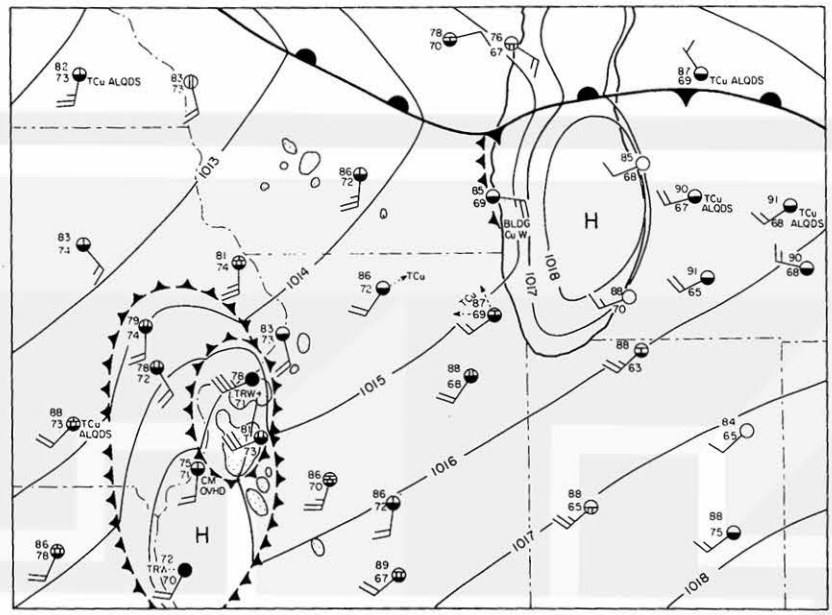

Fig. 7. Mesoanalysis at 1100 CST, 26 July 1966. Wind, dewpoint and remarks are plotted. One wind barb = 5 knots. The proportionate amounts of low clouds (black), middle clouds (double line) and high clouds (single line) are shown in the station circle. A stationary front lies across the northern section of the area. The larger, mature mesosystem in western Illinois contains the incipient stages of the squall line mesosystem which traveled across Illinois, Michigan and Indiana later in the day. Radar echoes from the Chicago WSR-57 are stippled.

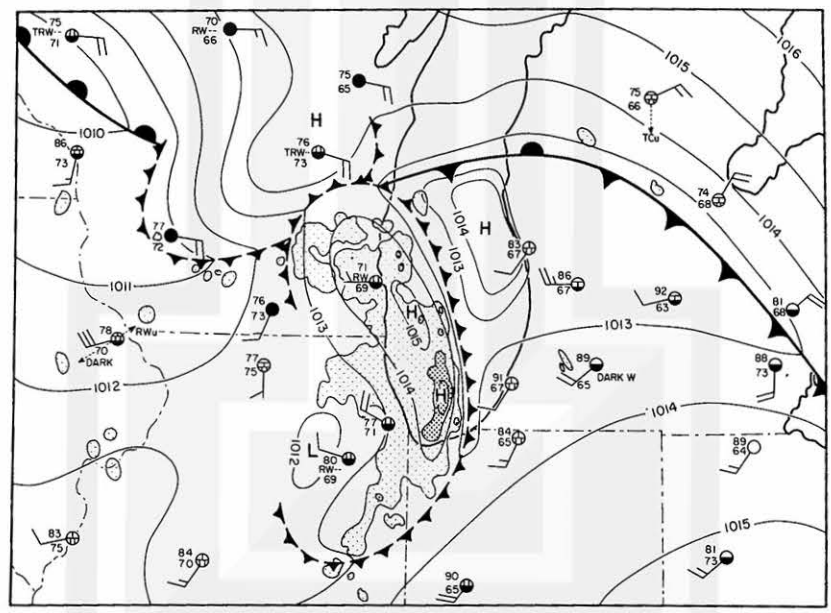

Fig. 8. Mature squall line mesosystem over mid-lake at 1700 CST. The lighter stipple indicates echoes visible at full receiver gain and darker stipple corresponds to an equivalent reflectivity factor of about $10^3 mm^6/m^3$ (moderate shower).

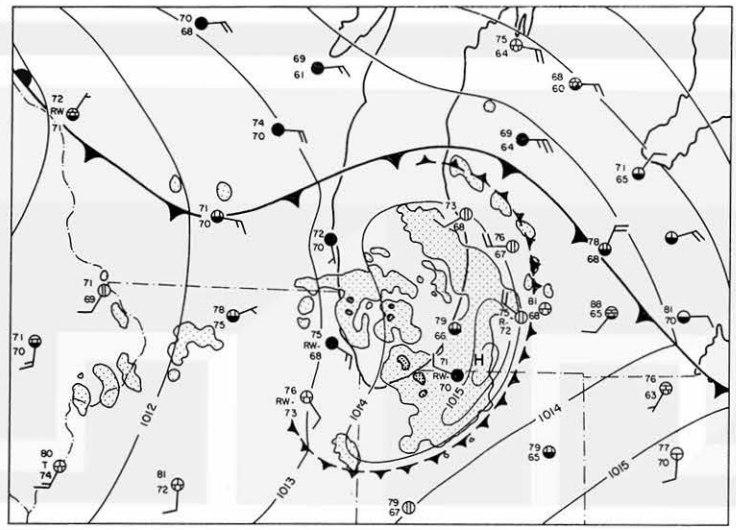

Fig. 9. The weakening squall line mesosystem at 2000 CST. A few moderate showers still exist, several over the lake.

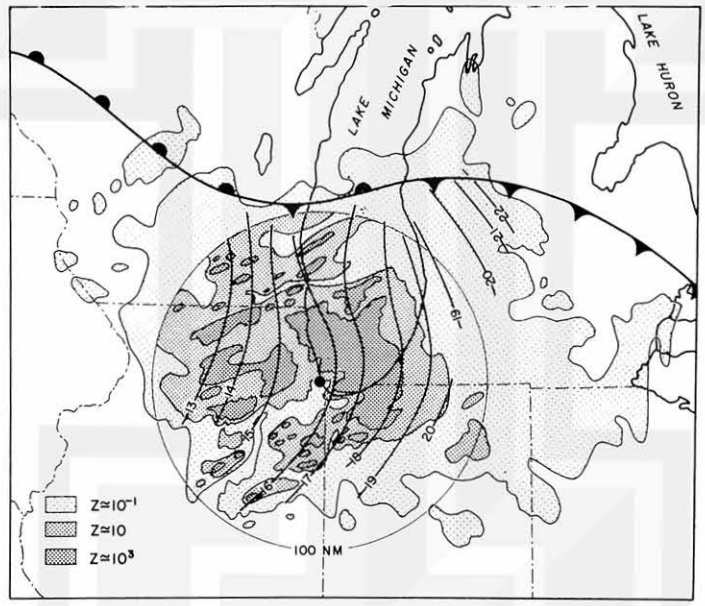

Fig. 10. Composite of strongest radar echo observed within 100 miles range of Chicago at 25-minute intervals from 1000 to 2200 CST, 26 July 1966. Darkest stipple indicates regions which were at any time covered by an equivalent radar reflectivity factor of 10^3 mm^6/m^3 or greater, medium stipple corresponds to a coverage by a Z of 10 to 10^3 mm^6/m^3, and the light stipple shows those areas (out to 250 miles) covered by any echo at all with receiver on full gain. The leading edge of the radar squall line is shown as a solid dark line.

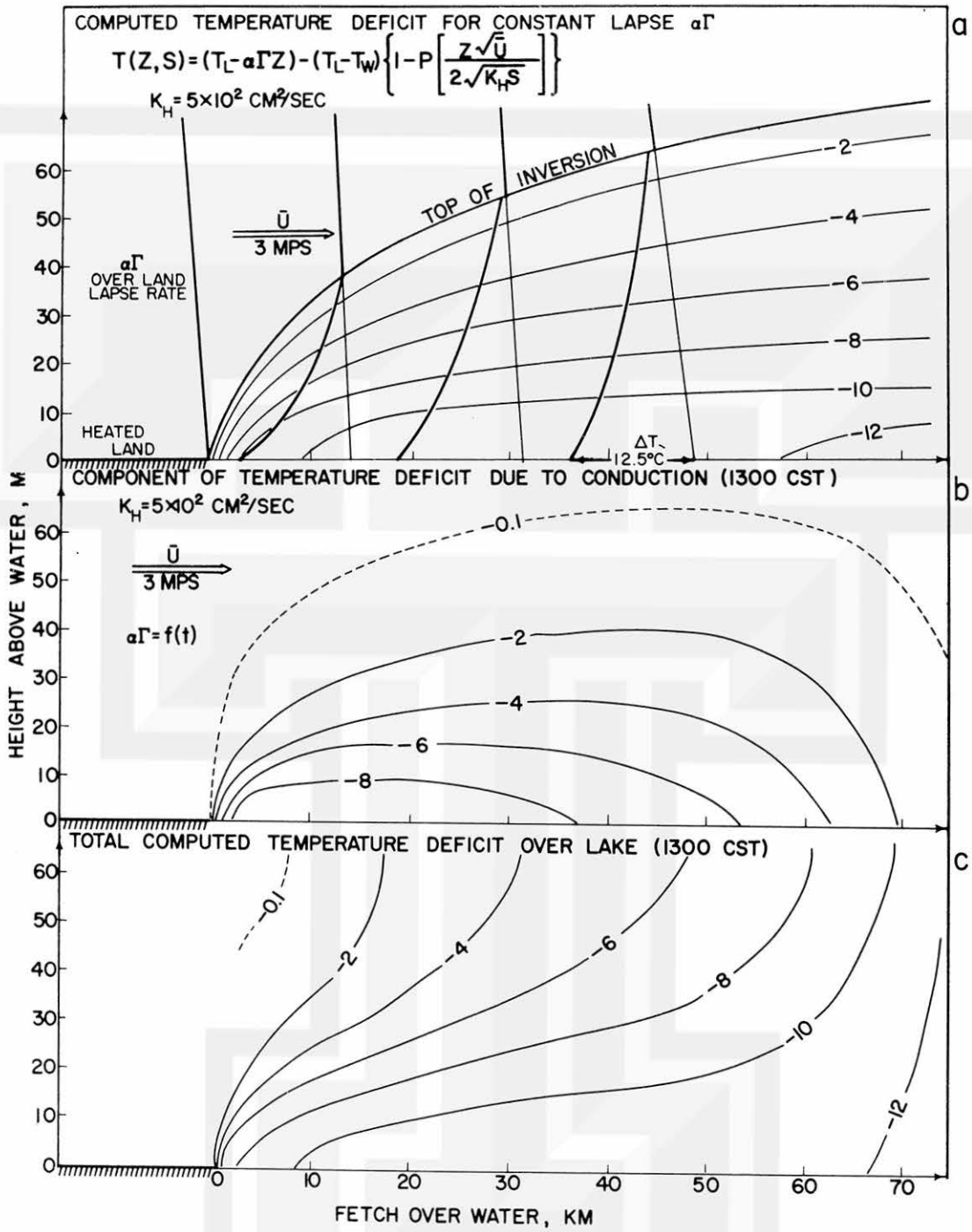

Fig. 11. (a) The computed temperature loss for a given column of air moving out over the cold lake surface with constant values of K_H and T_W assumed. The top of the inversion is taken to be the 0.5C isopleth. (b) The component of the land-lake temperature difference due solely to conduction of heat to the surface calculated by taking hourly increments of $\alpha\Gamma$ and T_L from 0600 to 1300 CST. (c) The total (actual) land-lake temperature difference at 1300 CST computed by subtracting the computed over-water temperatures from the actual temperatures measured over the heating land.

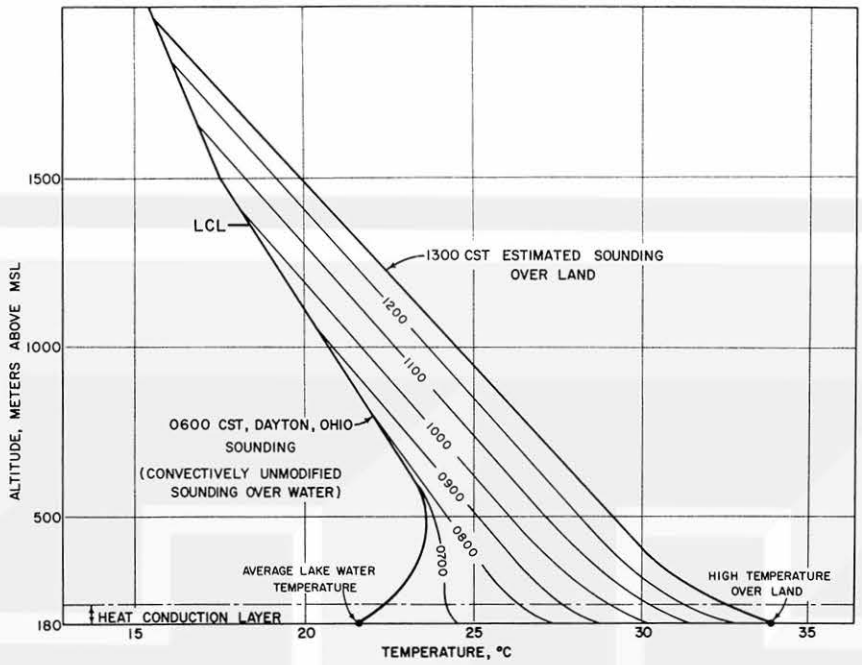

Fig. 12. Estimated temperature soundings over the heating land from 0600 to 1300 CST. The Dayton, Ohio 0600 CST sounding (Ti(z)) was most representative of the region during the morning hours and is assumed to be the sounding for those areas over water which had not been reached by advected warm air from the shore.

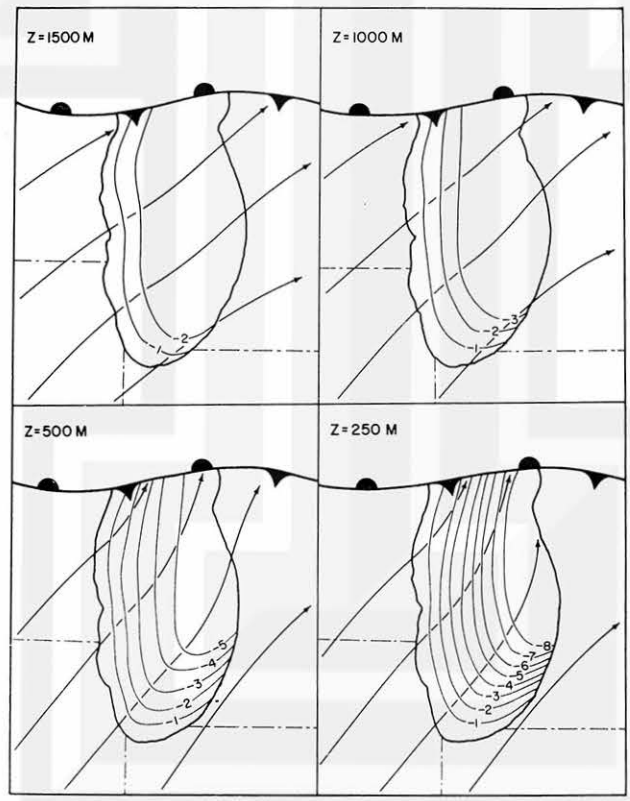

Fig. 13. Computed land-lake temperature differences at 1300 CST. Arrows indicate mean streamlines derived from serial pibal data used in computations. Winds above 100 m were almost constant at 10-12 knots.

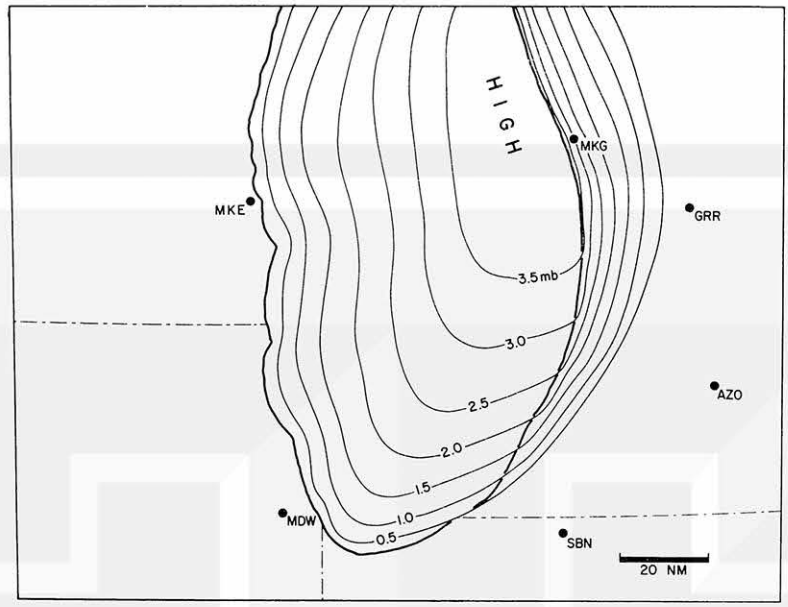

Fig. 14. Computed values of excess pressure over the southern basin at 1300 CST, approximately the time of maximum heating, and before warm air from the windward shore had reached the eastern shore.

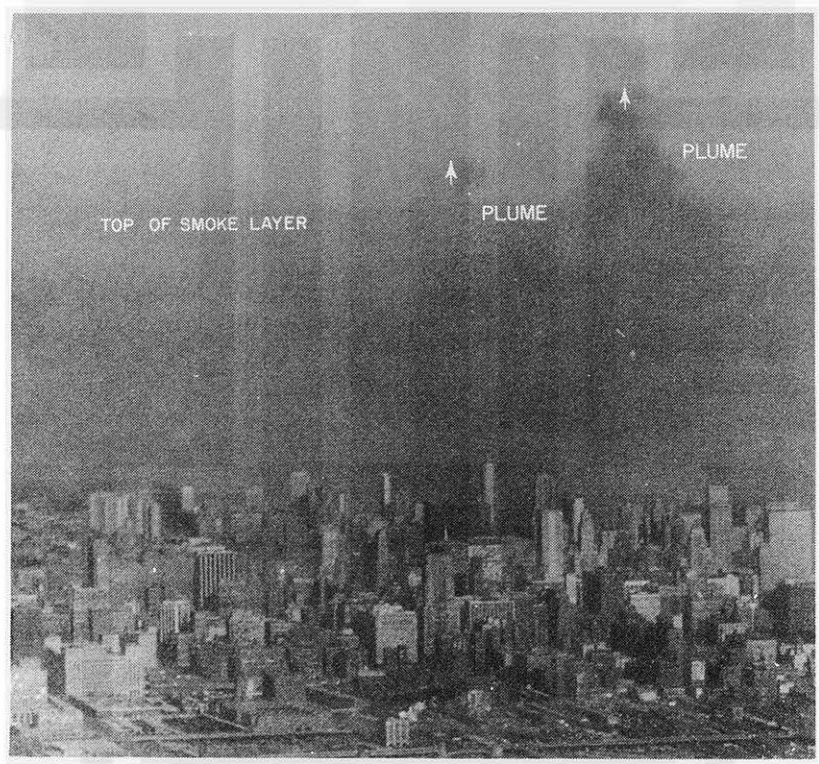

Fig. 15. Aircraft photograph showing typical plumes breaking through early morning smoke layer over Chicago's Loop. Often these plumes can be seen extending from near ground level into the bases of forming cumulus clouds (Photo courtesy of Art Shay).

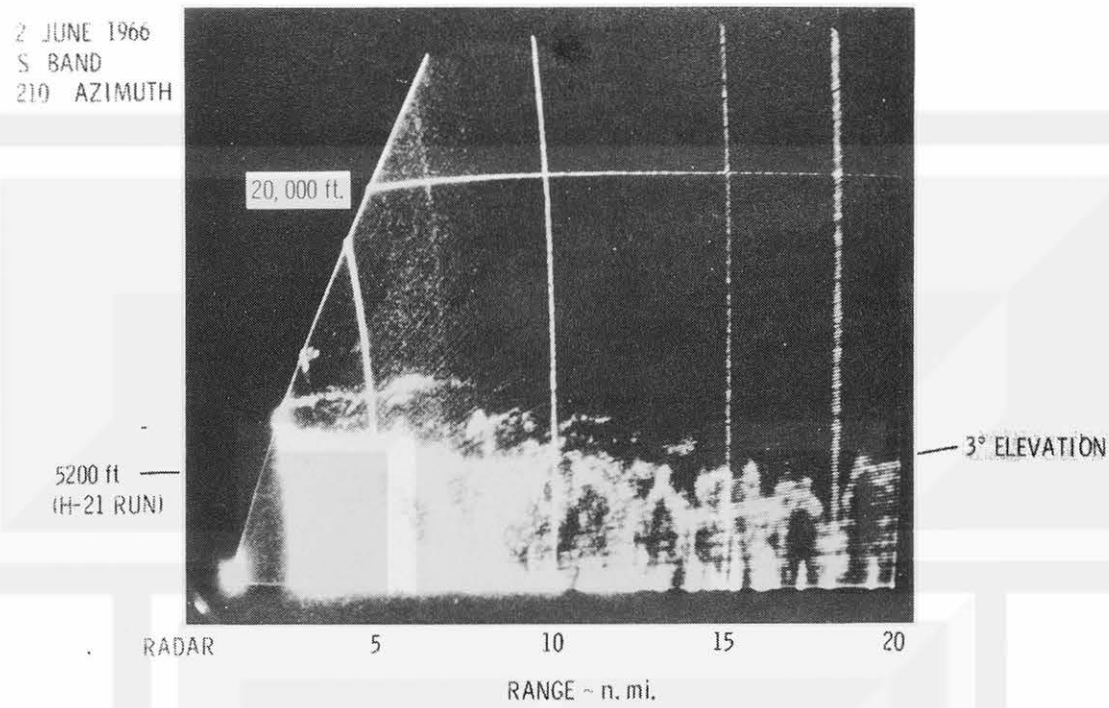

Fig. 16. RHI scan on Wallop's Island 10-cm radar showing echoes returned from clear air thermals over heated land on 2 June 1966. They appear to mark refractive index gradients at boundaries of traveling, continuing convective plumes rising from near the surface layers (Photo courtesy of Prof. David Atlas).

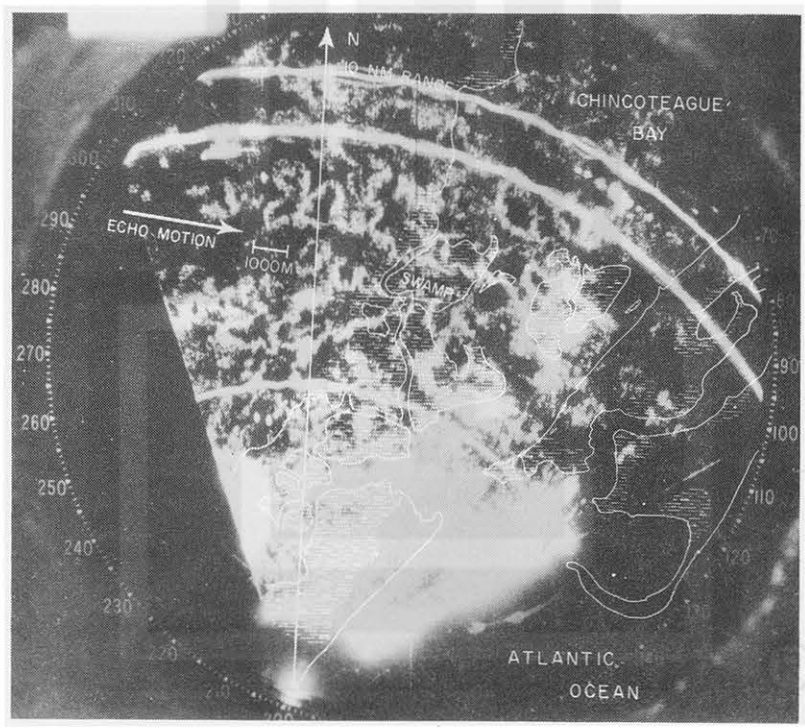

Fig. 17. PPI photograph from Wallop's Island radar on 2 June 1966 showing clear air thermals aligned in rows parallel to the low-level wind. Range markers are labelled at 5 and 10 n mi. The echoes can be seen to dissipate less than 5 miles after beginning passage over an area of swamps and open water (Photo courtesy of Prof. David Atlas).

Fig. 18. Infrared photograph of cloud line in convergence zone along southeastern shore taken from C-45 at 9500 ft at 1326 CST, 26 July 1966. Probable low-level streamlines are superimposed on picture.

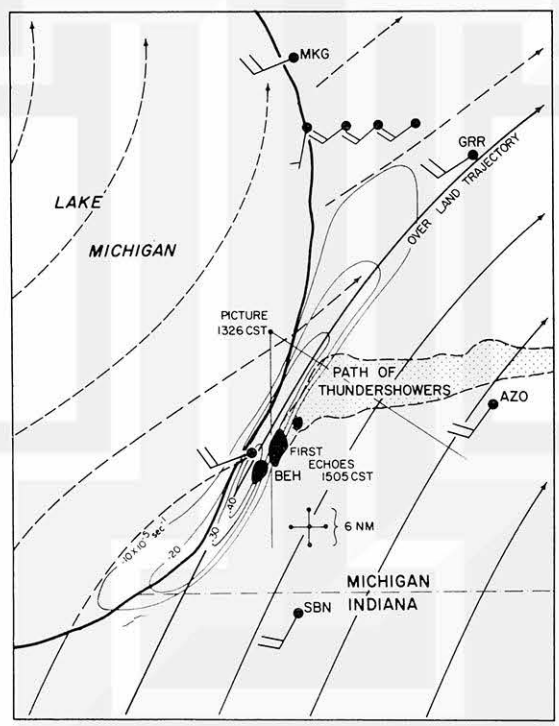

Fig. 19. Surface streamlines averaged from 1100 to 1500 CST. Isopleths of computed divergence are shown. The first radar echoes developing out of the cloud line shown in Fig. 18 are indicated by solid black areas, and their total area of coverage is stippled.

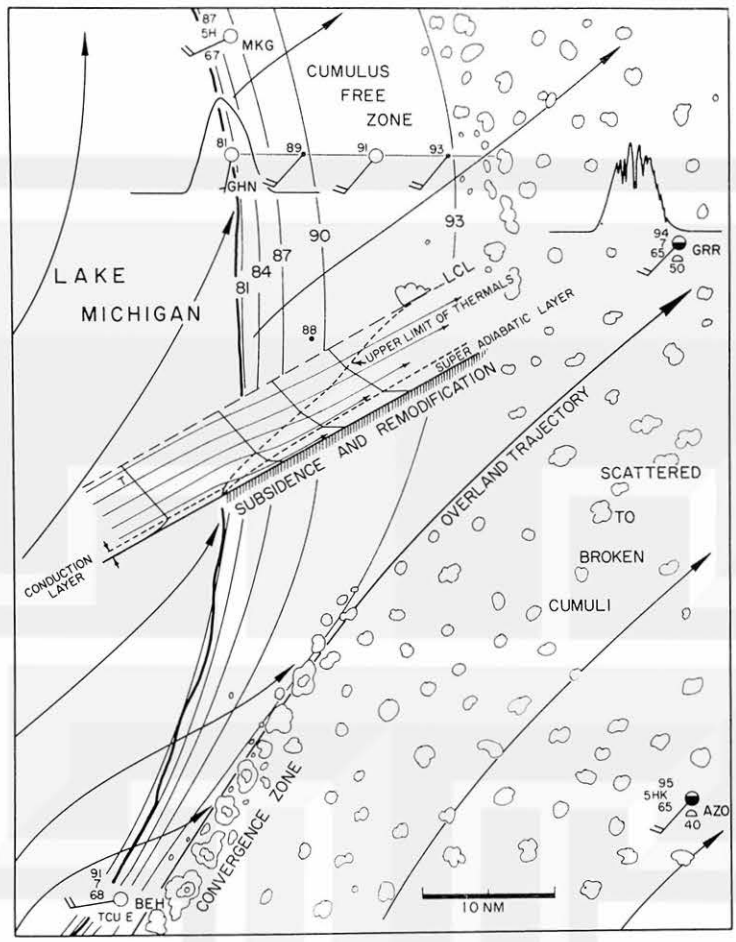

Fig. 20. Surface wind and cloud distribution along the southeastern and and eastern shore at 1400 CST. Surface isotherms are drawn every 3F. The inset is a vertical section showing the hypothesized subsidence in the onshore flow beneath the subcloud layer and the build-up of the superadiabatic boundary layer due to surface reheating. The first cumulus appear when the plumes are able to penetrate to the LCL. Shoreline (GHN) and inland (GRR) pyrheliograph traces show the distinctly different convective cloud regimes. The overland trajectory marks that surface flow which never passed over the lake.

Fig. 21. Photomosaic made from color photographs taken at 9500 ft above eastern shore about 35 miles south of Muskegon at 1335 CST. The thunderstorm anvils visible 80 to 90 n mi to the northeast were located by the Chicago radar.

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