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Transport of Water Vapor by Eddy Diffusion in Continental Air Masses Flowing Over Coastal Waters Sept. 23, 1947 to Nov. 16, 1948

by
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Abstract

Measurements of the transport of water vapor in the turbulent layer of the atmosphere have been made from 12 series of airplane soundings. The airplane, equipped with a psychrograph which recorded dry- and wet-bulb temperatures, made ascents in the air over the mainland and at several positions downwind over the coastal waters of southeastern Massachusetts. The measured accumulation of the water vapor in the homogeneous layer was used to compute the net upward flux of vapor per unit time and area. Values of this transport in the non-steady state are given and are shown to vary with the wind speed. Under the assumption that the flow is the net result of the turbulent motions of the air, values of the coefficient of mass exchange have been computed from the classical diffusion equation. The magnitude of this coefficient increases with height, decreases with distance from shore, and depends on the wind speed and on the stability of the air.

I Introduction

The work described in this report was conducted by the Woods Hole Oceanographic Institution with the support of the Office of Naval Research under Contract N6onr-277, T. O. II, Sec. A, Item (2); The study of the transport of heat and water vapor by eddy diffusion. Many of the plans and preparations for the program of observations were made under Contract Nobs-2083.

1. Vertical Transport of Water Vapor in the Atmosphere

Determinations of the magnitude of the vertical transport of properties in the atmosphere have received little attention in the study of the eddy diffusion process. This phase of the problem has been slighted in favor of work leading to the understanding of the physics of diffusion. In the present work, the measurement of the transport of water vapor from a large body of water to a cooler atmosphere has been the main task. Similar studies are being made with other thermal conditions of the air and water.

The measured transport of water vapor has been interpreted, in accordance with present theory, as the result of the random vertical motions of the air as it flows over the earth's surface. When parcels of air move from one level to another, they carry amounts of water vapor characteristic of the level with them to the new level. If this process continues it results in a net transport of water vapor from the regions of high moisture content to regions of low content. From this simple statement of the process operating in the atmosphere it is apparent that (1) the rate of transport will vary greatly with existing turbulence conditions, and (2) direct observation of the motions of the transporting air or of rates of transport are impossible.

Indirect methods of determining the rates of evaporation or transport of water from land and water surfaces have been devised and used by numerous workers since Dalton (1802) established the relation between evaporation and the difference in vapor pressure between the air in contact with the water and the air at a higher level. Others have used atmometers which give values of the evaporation of water into the air. Unfortunately these values are not identical with the transport actually occurring in the atmosphere because these instruments do not approximate atmospheric conditions sufficiently. Sverdrup (1936, 1937), Montgomery (1940), Thornthwaite and Holzman (1942), and others have used moisture and wind gradients in conjunction with theoretical turbulence relations, such as are summarized by Dryden (1943), to obtain expressions for the evaporation or transport in the atmosphere. The problem has been studied also by observing modifications of air masses. Lettau (1937) obtained observations from a free balloon. Radiosonde data was studied by Petterssen (1940). Burke (1945) studied modification from weather maps, radiosonde, and ship reports, developing charts for use as forecasting aids.

In the present work air masses were followed by an airplane equipped with a psychrograph. Dry- and wet-bulb temperatures were recorded alternately each five seconds while the plane made spiral ascents. From these soundings, accumulations of water vapor in the air were obtained and rates of transport were determined. These rates of transport are the primary object of this research, and can be used for forecasting rates of modification of continental air as it moves out over the coastal waters. In applying them it must be noted that they are true only for (1) continental air blowing over warmer water, (2) for short distances (20-40 km) offshore and (3) when few or no cumulus clouds are present.

2. The Operational Use of the Homogeneous Layer of the Atmosphere

The water vapor that evaporates from the land or water surface of the earth passes first into a laminar boundary layer about 1 mm thick. Molecular diffusion alone transports vapor in this layer for no turbulence exists at this level. Above this skin layer, a turbulent boundary layer extends upwards about 100 m, characterised by a logarithmic wind distribution and transport by eddy diffusion. Montgomery (1940) describes these layers in his study of evaporation.

The top of the boundary layer merges into the overlying homogeneous layer, whose characteristics have been studied by many workers (Rossby, Montgomery, 1935), (Bunker, Haurwitz, Malkus, Stommel, 1949). This homogeneous layer is a natural division of the atmosphere, usually 500 to 1500 m thick, bounded on the top by the main body of the atmosphere. The stratification of the air into this natural division is accomplished in the following manner by the turbulence of the air. The turbulence causes rapid mixing of the air thus producing a nearly dry-adiabatic temperature lapse rate, 10^{-6}°C cm^{-1}, and a small mixing ratio gradient, -10^{-9} cm^{-1}. If, as is usually the case, the main atmosphere is stable, then mixing will extend upwards to a height depending on the energy available for mixing and the energy required to mix the stable air. A small inversion will be produced at the boundary of the homogeneous layer and the air above it. The inversion is stable and the turbulence and mixing is damped, thus partially isolating the homogeneous layer from the main air mass. This characteristic of small transport between the air of the homogeneous layer and the air above it has been used in this work for determining the flow of water vapor into the atmosphere. The assumption has been made that a negligible amount of water vapor is mixed through the inversion so that the accumulation of vapor as measured by the series of soundings is the total mixed into the layer during the time of observation. Thus nature has provided a convenient box with a lid which collects and stores the water vapor as soon as it evaporates from the sea.

The data collected has been examined for soundings that could be used as a check on the assumption of negligible transport through the inversion. The soundings of the series taken on Sept. 22, 1948 can be used for this purpose. They extended 1000 m above the top of the homogeneous layer into the region where the air mass was very stable. In the height range from 600 m to 1400 m, the potential temperature lapse rate was $+ 1.1 \times 10^{-4}$°C cm^{-1}. The lapse rate in the thin (about 50 m) layer of boundary air between the homogeneous layer and the air above it was at least twice this. An inspection of the mixing ratio curve (see Appendix, Fig. 22) shows that the air gained moisture in the homogeneous layer while it lost moisture in the air above the layer. This can only mean that the transport of water vapor through the 1400 m surface to the air above was much greater than the upward transport through the 600 m surface. A computation of the flow of water vapor necessary to account for the depletion of vapor in the 600-1400 m region leads to a value of 4.7×10^{-6} gm cm^{-2} sec^{-1}. This value, which is about 1/4 the average transport in the homogeneous layer, is representative of the stable air aloft and is large in comparison to the transport through the boundary inversion. A tentative value of 10^{-6} gm cm^{-2} sec^{-1} may be assigned to this transport, pending further evidence. A more complete discussion of the Sept. 22 soundings will be found in the appendix.

The decision to use the homogeneous layer as a meteorological tool in the study of water vapor flux by eddy diffusion was made after a brief study of airplane soundings taken during the war years by the M.I.T. Radiation Laboratory. These soundings, made available by Dr. R. B. Montgomery, were being studied for the same purpose as the present investigation. However, no reliable values were found since the soundings extended to only 1000 ft while water vapor was transported to higher levels. This limitation prevented the determination of the total amount of water vapor transported and stored in the atmosphere in a given time interval. An approximate value could be found by using the lowest inversion shown by radiosonde observations taken in the region. One fault of the radiosonde made it unsuitable for this work. Its great speed of ascent and infrequency of signal transmission could cause its failure to record weak inversions. The values determined from the study of the M.I.T. data have never been published for they were considered too uncertain. Now that the values from the present work are complete, a comparison of the values of the transport and of the coefficient of turbulent mass exchange show the two sets to be nearly the same. For similar synoptic conditions the M.I.T. data for 8 series yielded values between 1.4 to 13 x 10^{-6} gm cm^{-2} sec^{-1} for the transport and between 80 and 680 gm cm^{-1} sec^{-1} for the exchange coefficient, while the present work shows values, in the lowest layer, ranging from 3.7 to 91 x 10^{-6} gm cm^{-2} sec^{-1} and 105 to 5600 gm cm^{-1} sec^{-1}.

3. The Eddy Diffusion Equation

The computed values of the vertical flow of water vapor and mixing ratio gradients have been placed in the eddy diffusion equation,

$$E = - A \, \partial w / \partial z, \tag{1}$$

to obtain values of A, the coefficient of turbulent mass exchange. The value of A as determined from the observations includes any influence of convection that may be present in the atmosphere. Since both convection currents associated with cumulus clouds and "convective turbulence" as defined by Priestly and Swinbank (1947) will modify the value of A characteristic of eddy diffusion alone, it is logical to define A by equation (1), rather than by the classical derivation[1] based on eddy turbulence alone. Through this change, A is made to describe the state of mixing of the atmosphere as it actually exists, regardless of the process creating the turbulence or mass exchange. It should not be inferred from this change in the definition of A that eddy diffusion is no longer considered the predominating process of diffusion in the homogeneous layer. At the present time, the relative

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[1] See Haurwitz, 1941, pp. 217-218.

contribution of each process cannot be evaluated.

The observations were taken on days when few or no cumulus clouds were present in the areas studied. This selection of working days reduces the effect of convection currents associated with cumulus activity on the computed value of the transport of water vapor.

When using this equation it is essential to realize that its application depends on averages taken over long time-intervals. In this manner the random nature of the turbulent eddy motion of the air can be treated mathematically and the transport of properties by the air can be expressed in terms of gradients. The manner in which the present observations were taken does not satisfy the condition of a long time-interval. For the complete satisfaction of the conditions of the equation, several soundings should have been made at each point along the path of the air. This was operationally impossible. However, gradients may be determined by averaging values of the mixing ratio over a large height range of a single sounding. The gradient can have an intelligible meaning only when averaged over heights large in comparison with the turbulent eddies responsible for the transport of the vapor. Attempts to calculate gradients and exchange coefficients for small height ranges from single soundings will usually be unsuccessful since the procedure conflicts with the large-scale gusty nature of turbulent air. Such a method might work well in a stable air with small-scale turbulence.

To find significant values of the gradient, a selection of height ranges was made according to the appearance of the mixing ratio curve with height. It was noted that near the surface and up to 100-300 m there was a large gradient, while a smaller gradient existed from this level to the top of the homogeneous layer. Thus, mixing ratio values in the height ranges 15 to 100-300 m and from 100-300 m to the top of the layer were used in averaging gradients. It was assumed that each of these natural divisions was large in comparison with the eddies that produced them and was suitable for averaging. On averaging the values, the method of the least squares was used to determine the slope of a straight line fitted to the observations. This method was adopted not because it was expected to lead to values of great accuracy, but rather to give an impartial value to the gradients which are difficult to measure graphically.

Before applying equation (1) to the observations, mention should be made of a condition found in the atmosphere that does not satisfy the conditions required by the equation. Attempts to choose days without convective activity were made, but some cumulus clouds occasionally developed in the sounding areas. Some of these clouds may have pierced the inversion allowing unknown amounts of moist air to leave the homogeneous layer. This gave erroneous values of E to be entered into the diffusion equation.

II Equipment and Operational Procedure

1. Equipment

The major instrument used in this investigation is the airplane psychrograph designed and built by the Radiation Laboratory of the Massachusetts Institute of Technology. The entire program of observation was planned around the use of this instrument. The psychrograph will not be described here since it has been explained adequately by Katz (1947). It is sufficient to say here that it records alternately wet- and dry-bulb temperatures with a probable error of $\pm$ 0.01°C. The thermal lag of the resistor elements used for this study is 4 seconds with a ventilation speed of 100 knots. The housing containing the temperature sensitive elements was mounted on the wing strut of a Stinson Voyager. The amplifier, recorder, batteries and vibrator were placed in the back seat of the 4-place plane (see Figs 1 and 2). Mr. Kenneth McCasland operated the instrument and made altitude and airspeed notes while Mr. Marvin Odum piloted the plane.

A similar psychrograph was made for use with a captive balloon from a small boat. This instrument is described by Anderson (1946), (see Figs 3 and 4). Two amplifiers and recorders were installed so that dry and wet-bulb temperatures could be recorded simultaneously. The 1000 gm neoprene captive balloon had sufficient lift to carry the housing and connecting wire to 100 m. Only one sounding taken by this instrument is presented here for on many days electrical leakage occurred through the insulation of the wire rendering the values obtained uncertain.

FIG. 1 - PSYCHROGRAPH HOUSING BEING MOUNTED ON WING STRUT.

FIG.2- AMPLIFIER AND RECORDER FITTED INTO PLANE.

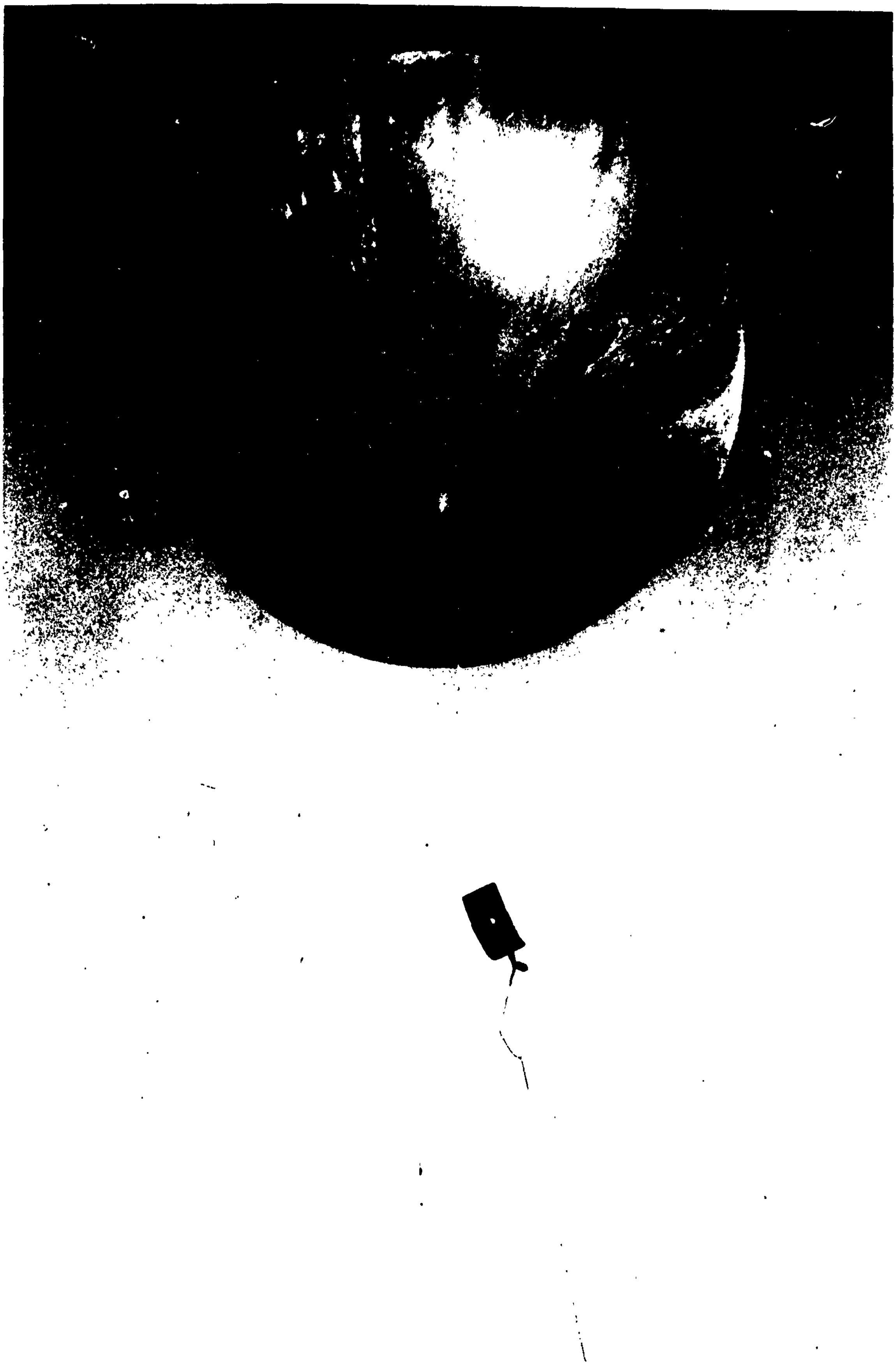


FIG. 3 - CAPTIVE BALLOON AND PSYCHROGRAPH HOUSING.

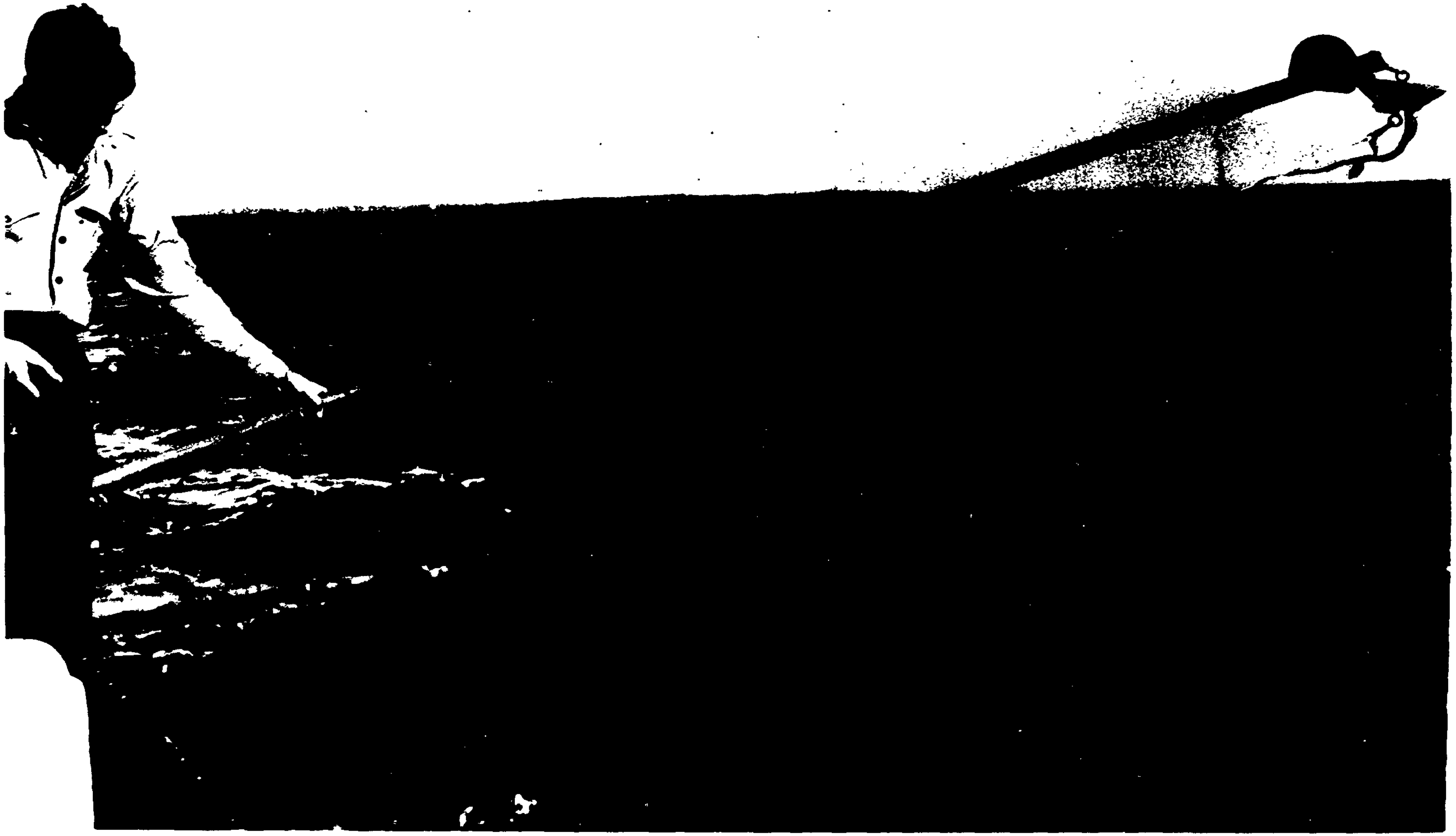


FIG. 4 - PSYCHROGRAPH OBTAINING GRADIENTS CLOSE TO THE WATER SURFACE.

Pilot balloon observations were made in the usual mannor as prescribed by the U. S. Weather Bureau (1942), except that 30-second time intervals were used.

Water temperatures were taken either by a dip bucket thermometer or by a bathythermograph. This instrument records water temperature against depth on a smoked glass slide. Information concerning the bathythermograph may be found in a Preliminary Instruction Book (1945).

The only other instruments used by the project are; (1) a sling psychrometer, (2) the plane's altimeter, (3) the plane's air speed indicator, (4) an anemometer, and (5) a mecurial barometer.

2. Operational Procedure

The first step in the day's work, once it has been decided that the day is suitable for observations, is to determine the present wind and check with the wind forecast as broadcast by the U. S. Weather Bureau. The winds aloft are computed from the pilot balloon observation taken early in the morning. From these and any changes in wind directions mentioned in the forecast, a probable trajectory of air moving offshore is determined. A distance interval downstream between soundings is decided upon so that measurements will be made in nearly the same air mass, thus reducing any advection effects that may be present. A series of observation points is located that satisfies the trajectory of the air, the proper spacing of the sounding, and that allows the single-engined plane to circle near an island or lightship for safety.

The observer proceeds to the airport, installs the psychrograph equipment in the plane, receives any last minute changes in the course, and takes off. Upon arriving at the selected spot overland a 500 yard radius spiral ascent is started from about a 200 foot altitude. The pilot attempts to maintain a 70 knot air speed and a 200 foot per minute climb. The observer makes altitude check marks on the recorder paper and keeps a log of altitude, air speed, roughness, cloud amounts, etc. The ascent is continued until it is noticed, either from the psychrograph record or from the sudden entry into the smooth air above the turbulent layer, that the inversion at the top of the homogeneous layer is reached. The climb is continued another few hundred feet before leaving the location for the next sounding area. Here the same procedure is repeated except that the plane is flown much closer to the water than to the land surface. Each series of observations consists of two to four soundings. Figure 5 is a photograph of a U. S. Coast and Geodetic Survey chart showing an area frequently used for soundings. Wire springs of the proper diameter were cut at the correct height to depict the 500 yard radius spiral made by the airplane as it climbed to the top of the homogeneous layer.

The second observer either remained on the Institution dock making wind aloft and water temperature observations or proceeded to the sounding area with a boat equipped with a balloon psychrograph and bathythermograph. With these instruments, water temperatures were obtained from the bottom of the bay or sound to the surface, and wet- and dry-bulb temperatures were obtained from 1 meter above the surface to 100 m. To assure ease of operating and freedom from salt spray the balloon soundings were made running downwind.

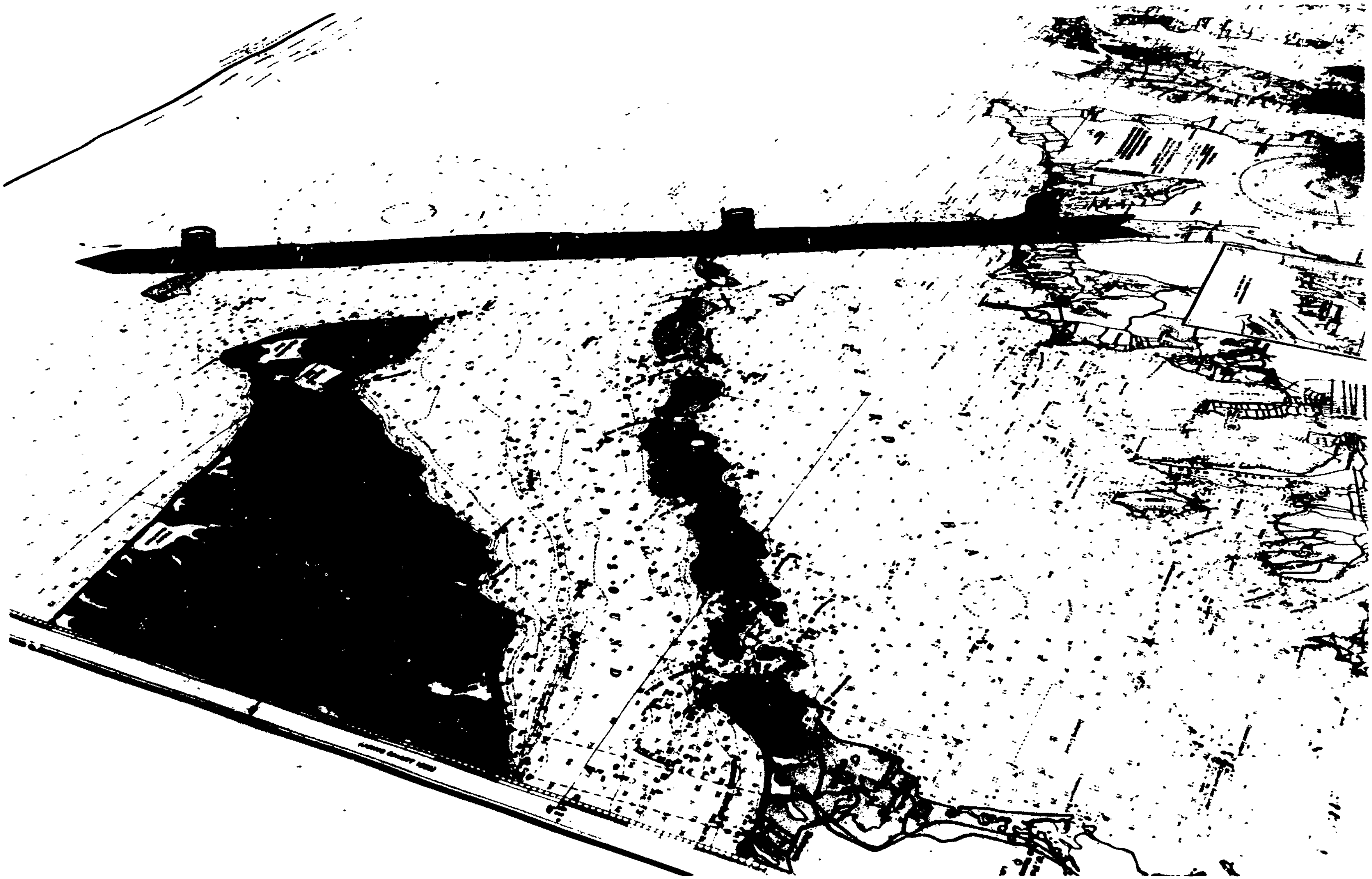


FIG.5- PHOTOGRAPH OF CHART WITH SPRINGS REPRESENTING THE HEIGHT, DIAMETER, AND RELATIVE POSITIONS OF THE SPIRALS MADE BY THE SOUNDING AIRPLANE.

THE ARROW INDICATES THE AVERAGE WIND AND SERVES AS A DISTANCE SCALE.

III Reduction of Data

1. Airplane Soundings

Methods of reducing the airplane observations were devised for determining rapidly the various elements desired and at the same time assuring that no fictious gradients or errors were introduced by approximations. Speed of reduction was essential for over 600 computations are required for the average sounding. A list of the observed values from which the computations are made is as follows; (1) the milliammeter reading of the dry and wet bulb resistances, (2) the indicated air speed, and (3) the altimeter reading.

a) Dry- and Wet-Bulb Temperatures

The milliammeter readings were reduced to dry- and wet-bulb temperatures by use of tables made up from the laboratory calibrations of the temperature elements. Corrections had to be applied to these values because of the heating caused by compression of the air ahead of the moving elements. These were determined from the following equation given by AAF Weather Service (1945);

$$\Delta T = \alpha_d \left(\frac{v}{100}\right)^2$$

Here α_d is a factor determined experimentally for the particular psychrograph. Its value is 0.9. The symbol v denotes the true air speed of the plane in miles per hour. The true air speed was found from the indicated air speed and the altitude by use of an aircraft navigational plotting board.

b) Pressure

The atmospheric pressure at each height, as indicated by the plane's pressure altimeter, was found from the surface pressure and the pressure height relation of the standard atmosphere.

c) Mixing Ratio

The mixing ratio, w, of each point was computed from the following form of the psychrometric equation;

$$w = w_s - (T_d - T_w)\,\frac{c_{pd} + wc_{pv}}{L}\,.$$

In this equation w_s is the saturation mixing ratio at the given wet-bulb temperature and pressure, T_d is the dry-bulb temperature, T_w the wet-bulb temperature, c_{pd} and c_{pv} the specific heats at constant pressure of dry air and water vapor respectively, L the heat of vaporization of water. The saturation mixing ratio was read from a graph giving the relation between wet-bulb temperature, pressure and mixing ratio. For a frozen wick, values of the saturation mixing ratio over ice were computed from the Smithsonian Meteorological Tables, Fifth Revised Edition (1939). The heat of sublimation was used in place of the heat of vaporization.

d) Equivalent, Virtual, and Potential Temperatures

Nomograms were used to compute these values. Potential temperatures are computed to the base 1000 mb.

e) Heights

True height in meters were computed from the virtual temperature for a few selected pressures and all other heights were found graphically from the resulting pressure height relation.

f) Total Amount of Water Vapor

The total amount of water vapor contained in the air column was found by multiplying the average mixing ratio for a small pressure range (2 mb below 1000 ft, 15 mb above 1000 ft) by the pressure difference divided by the constant of gravity. By always totaling the water vapor between the same altimeter heights one set of values for $\Delta p/g$ could be used for multiplication by w.

2. Captive-Balloon Soundings

a) Temperatures and Mixing Ratios

Milliammeter readings were reduced to temperature in the same manner as for the airplane soundings, except that dynamic corrections were not required. Mixing ratios were computed similarly.

b) Heights

Heights were read from markings on the wire holding the balloon. Corrections were made for the wire angle.

IV Analysis of Observations

1. Water Vapor Flux

The flow of water vapor through a unit surface of one square centimeter at a given height was found by totaling the water vapor in the column of air as it left the mainland and computing the accumulation of water vapor above the surface as the air moved over the water. As mentioned previously the water vapor was totaled to the top of the homogeneous layer where, according to the assumption, a negligible amount mixed through the inversion into the dry air aloft. Once the increase of water vapor was determined, the rate of flow could be found from the time the air had been over the water.

The pilot balloon data was used for this computation. The wind speed and direction averaged over the entire homogeneous layer was considered best for this work. In the case of stable air masses the turbulence is weak so that the air aloft moves faster than, and does not mix with, the lower air thus causing the phenomenon of shearing stratification as described by Craig (1946). In less pronounced cases of stability and in neutral and unstable air masses this stratification cannot develop because of the rapid mixing of the air. The characteristic frictional shearing of the wind with height is present, but this does not imply that a particular parcel of air travels long distances at one height. Rather, through the mixing process of eddy diffusion, it will travel at many heights in the homogeneous layer. For this reason it is correct to use the mean value of the velocities of the air at different heights to determine the trajectories of the air. Thus a parcel of air in the upper part of the layer has travelled over water about the same length of time as a parcel of air nearer the bottom, although their instantaneous velocities will differ greatly.

In the reduction of these observations and the computation of flux and austausch values, this principle is used in cases where there is evidence of sufficient mixing. This allows treatment of the homogeneous layer as a single body moving with a given velocity although its many component parts move with different velocities.

The flux of water vapor has been found in most cases for two surfaces, one E_0, at the bottom of the layer and another E_h, 100-300 m higher. Values of E, the flow through these surfaces, are presented in Table I. The units are gm cm^{-2} sec^{-1} x 10^{-6}.

Table I

Date	Time EST	Distance Offshore	E_0	E_h	
Sept. 23, 1947	1354-1515	1-13 km	7.3	-	
Oct. 1, 1947	1112-1311	1- 8 km	24.0	17.0	
		8-16	91.0*	40.0*	
Nov. 9, 1947	1520-1644	0-16	19.0	13.0	
Nov. 10, 1947	1337-1432	0- 8	28.0	14.0	
Nov. 13, 1947	1450-1602	0- 8	41.5	-	
Nov. 17, 1947	1405-1557	0- 3	23.0	18.0	
		3-10	2.0*	-2.2*	
		10-17	13.0	7.6	
Sept. 1, 1948	1232-1410	0-12	44.0*	43.0*	
		12-32	9.3*	-2.5*	
		0-32	22.0	14.0	
Sept. 15, 1948	1345-1552	0-12	13.6*	4.2*	
		0-32	10.3*	3.2*	
Sept. 16, 1948	1155-1346	0-43	3.7	1.7	E_s*
Sept. 22, 1948	1105-1305	0-12	-	6.0	
		0-37	-	5.0	
		12-37	-	-	4.7
Oct. 13, 1948	1130-1315	0-30	6.0*	1.5*	
Nov. 16, 1948	1145-1301	0-24	3.8	1.4	

* See discussion in appendix

The significance of the large range in the computed values of the flow can be clarified by plotting values of E against the wind speed observed on the same days. The values plotted in Figure 6 are the average of the wind for the entire homogeneous layer as tabulated in Table V of the Appendix. The graph shows that a relation exists between the flow and the wind speed. The scatter of the points is small, if the Oct. 13, 1948 value is omitted. The relative position of this point emphasizes the effect of stability, for, in all cases save this, the water is of the same temperature or up to 9°C warmer than the air blowing over it. On Oct. 13 the water was 2°C cooler than the air.

Over the mainland the air was very bumpy and turbulent, but as it moved over the cool water the turbulence subsided to what was described by the airplane observer as very smooth air. It is usual for the turbulence of the air to decrease over the water even when the water is warmer than the air, but this case shows that the magnitude of the decrease is much greater when the water is cooler.

Thus if the Oct. 13 value is omitted on the grounds that it represents a different stability condition of the air, the remaining points define a close relation between the vertical flow and wind speed.

From Figure 6, values of E can be obtained that are suitable for making forecasts or estimates of water vapor transport. Table II gives recommended values for regions between 0-30 km offshore and for water temperatures greater than the air temperature.

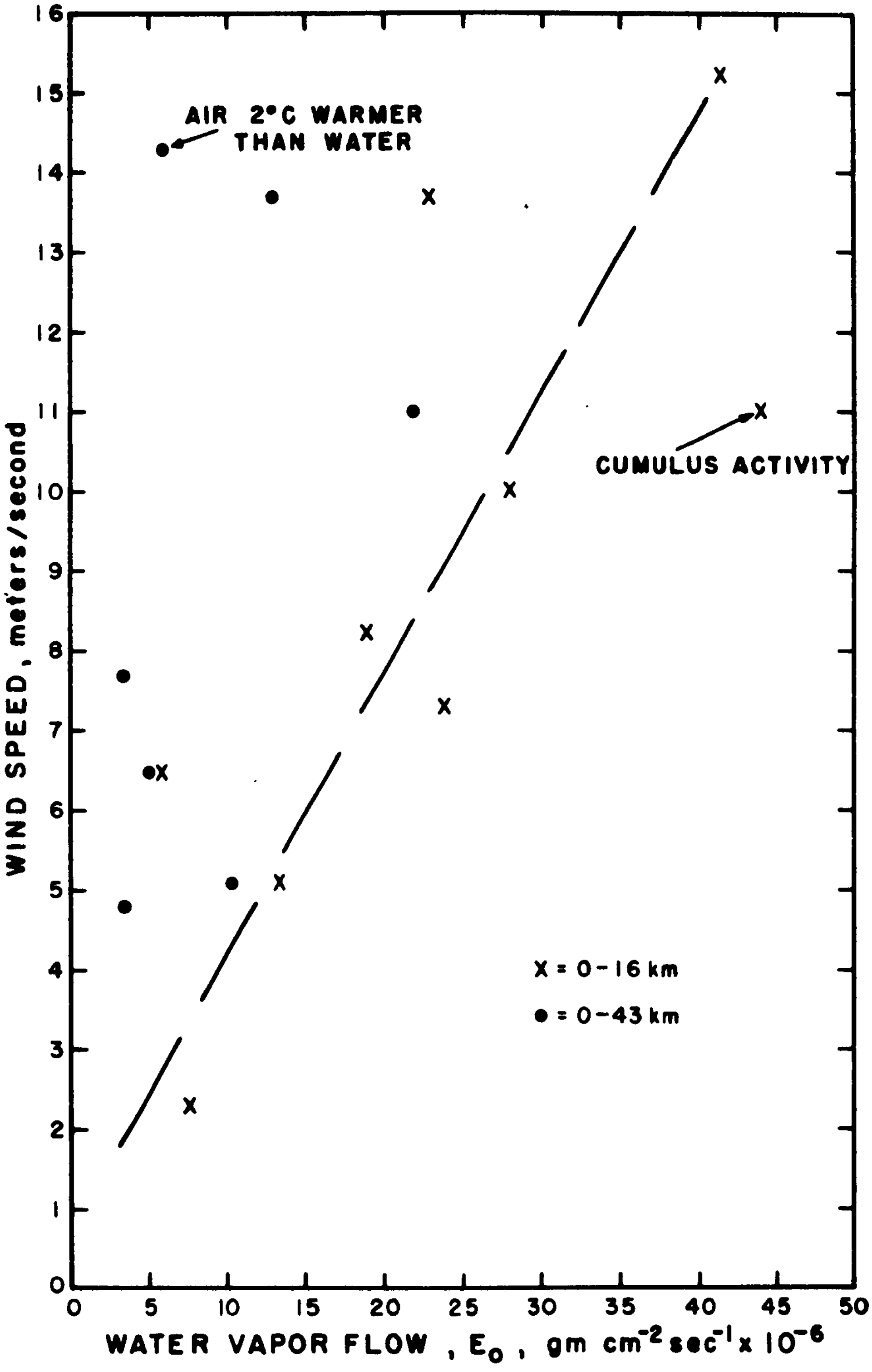

AIR 2°C WARMER
THAN WATER
CUMULUS ACTIVITY
X = 0-16 km
● = 0-43 km
WIND SPEED, meters/second
WATER VAPOR FLOW, E_o, gm cm^{-2} sec^{-1} x 10^{-6}
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
0
5
10
15
20
25
30
35
40
45
50

FIG.6

Table II

Wind Speed Average for Homogeneous Layer m sec^{-1}	Water Vapor Transport gm cm^{-2} sec^{-1}	$E_0/(w_s - w)$ gm cm^{-2} sec^{-1}
5	1.0×10^{-5}	1×10^{-3}
10	2.5×10^{-5}	4×10^{-3}
15	4.5×10^{-5}	7×10^{-3}

The scatter of the plotted points in Figure 6 can be reduced by eliminating the Dalton (1802) effect by the same method that Montgomery (1940) employed. Here the water vapor flux at the surface has been divided by the difference between w_s, the saturation mixing ratio of air immediately in contact with the water, and w, the average mixing ratio of the air of the homogeneous layer before passing over the water. In Figure 6A, this quotient is plotted against the wind speed of the homogeneous layer. Only the nine points representing values obtained between 0-16 km offshore are plotted. A close relation between the quotient of the flux and mixing ratio difference and the wind speed is indicated by the small scatter of the points.

Values of the quotient are given for three values of the wind speed in the third column of Table II. This column should be used for forecasts and estimates of transport if the values of the water temperature and mixing ratio of the air are known.

2. Mixing Ratio Gradients and the Coefficient of Turbulent Mass Exchange

The flow of water vapor upward through the homogeneous layer has

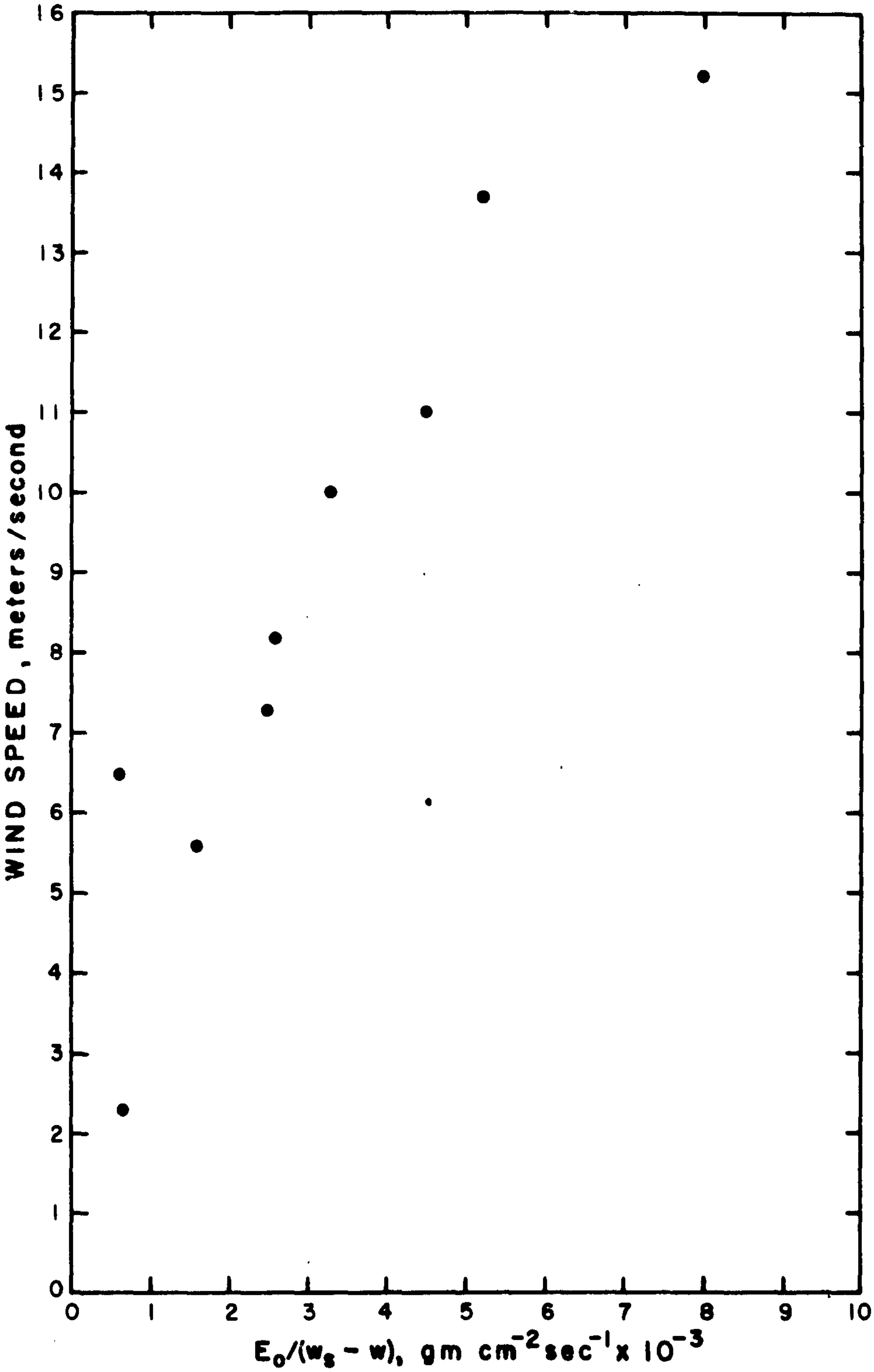

WIND SPEED, meters/second
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
0
1
2
3
4
5
6
7
8
9
10
E₀/(w_s − w), gm cm⁻² sec⁻¹ x 10⁻³

FIG.6A

been attributed to random vertical motions of the air and can thus be analyzed by the diffusion equation (1). The value of the coefficient of turbulent mass exchange obtained from the present data is characteristic of the unstable conditions existing at the time of the observations.

As mentioned in the introduction, the conditions required for the complete validity of the equation do not obtain in the atmosphere, and also the observations do not represent the necessary time average. Nevertheless, by taking averages over large height ranges and evaluating the coefficient at only two heights, the errors due to these unavoidable inconsistencies in the choice of data may be reduced.

On evaluating the exchange coefficient from equation (1), a standard system of determination of gradients and corresponding flux surfaces was used. Two gradients were determined from the slope of a straight line by the method of least squares, one for the region close to the water surface up to 100-300 m, the other from the 100-300 m to the top of the layer. The flux through the surface at the bottom of the atmosphere was matched with the gradient for the lowest level. The equation then yields a value characteristic of lowest layer but not associated with any exact height. Likewise the flux through the 300 m surface was matched with the gradient for the main body of the homogeneous layer. This value of the coefficient is considered to apply to the middle section of the homogeneous layer.

Table III presents the gradients of the mixing ratios as found by the least squares method. The value given is the average of the two soundings involved in the computation of A. Gradients are expressed in

Table III

Coefficient of Turbulent Mass Exchange

Date	$\left(\frac{\partial w}{\partial z}\right)_0$ $cm^{-1} \times 10^{-9}$	$\left(\frac{\partial w}{\partial z}\right)_h$ $cm^{-1} \times 10^{-9}$	A_0 $gm\ cm^{-1}\ sec^{-1}$	A_h $gm\ cm^{-1}\ sec^{-1}$		Distance Offshore km
Sept. 23, 1947	-	-5.6	-	1300		1-13
Oct. 1, 1947	-67	-69	360	2500		1- 8
	-6.7	-7.6	1300	5300		8-16
Nov. 9, 1947	-8.5	-2.3	2200	5700		0-16
Nov. 10, 1947	-5.0	-1.7	5600	8200		0- 8
Nov. 13, 1947	-	-2.7	-	15400		0- 8
Nov. 17, 1947	-4.4	-2.1	5200	8600		0- 3
	-12	-3.0	170	-		3-10
	-12	-4.5	1100	1700		10-17
Sept. 1, 1948	-14	-2.8	3100	15000		0-12
	-36	-7.0	260	330		12-32
	-37	-7.0	600	2000		0-32
Sept. 15, 1948	-19	-5.1	720	825		0-12
	-27	-5.5	380	580		0-32
Sept. 16, 1948	-28	-3.9	130	435	A_s*	0-43
Sept. 22, 1948	-8.0	-4.0	-	1100		0-12
	-17	-4.0	-	1300		0-37
					200	12-37
Oct. 13, 1948	-58	-0.16	105	9400		0-30
Nov. 16, 1948	-12	-1.8	320	780		0-24

* See discussion in Appendix

cm^{-1} with the factor 10^{-9} omitted. Exchange coefficients are expressed in gm cm^{-1} sec^{-1}, while the distance offshore is expressed in kilometers. A_0 designates the value determined for the region 0 to 100-300 m, while A_h represents the value characteristic of the region 100-300 m to the top of the layer. The one value labelled A_s is a value found for the stable air above the homogeneous layer.

Several facts are immediately obvious from the table. One is that the exchange coefficient is greater in the main body of the layer than it is closer to the surface. This is shown clearly by noting that the average value in the lower level (A_0 column) is 1400 gm cm^{-1} sec^{-1}, while at the higher level, the average of the A_h column is 4450 gm cm^{-1} sec^{-1}. The decrease in the value of the coefficient with distance from shore is shown best by averaging values of A_h obtained on days for which data was collected at several increasing distances from shore. The average of the values close to the shore is 5605 gm cm^{-1} sec^{-1} while the average value farther out is 2220 gm cm^{-1} sec^{-1}. From this we infer that the turbulence of the air decreases as it flows over the water regardless of the fact that the water temperature is higher than the air temperature.

An important point brought out by the table is the large values of the coefficient of turbulent mass exchange. Prior to this report and the work of Huss and Portman (1948) the coefficient was measured under stable conditions and values between 10 and 500 gm cm^{-1} sec^{-1} were assumed characteristic of most conditions encountered in the atmosphere. This work shows that values lying between 1000 and 10000 gm cm^{-1} sec^{-1} are characteristic of the unstable case.

Values of the exchange coefficient determined by other workers in the field have been collected in Table IV.

Table IV

Coefficients of Turbulent Mass Exchange and Eddy Viscosity

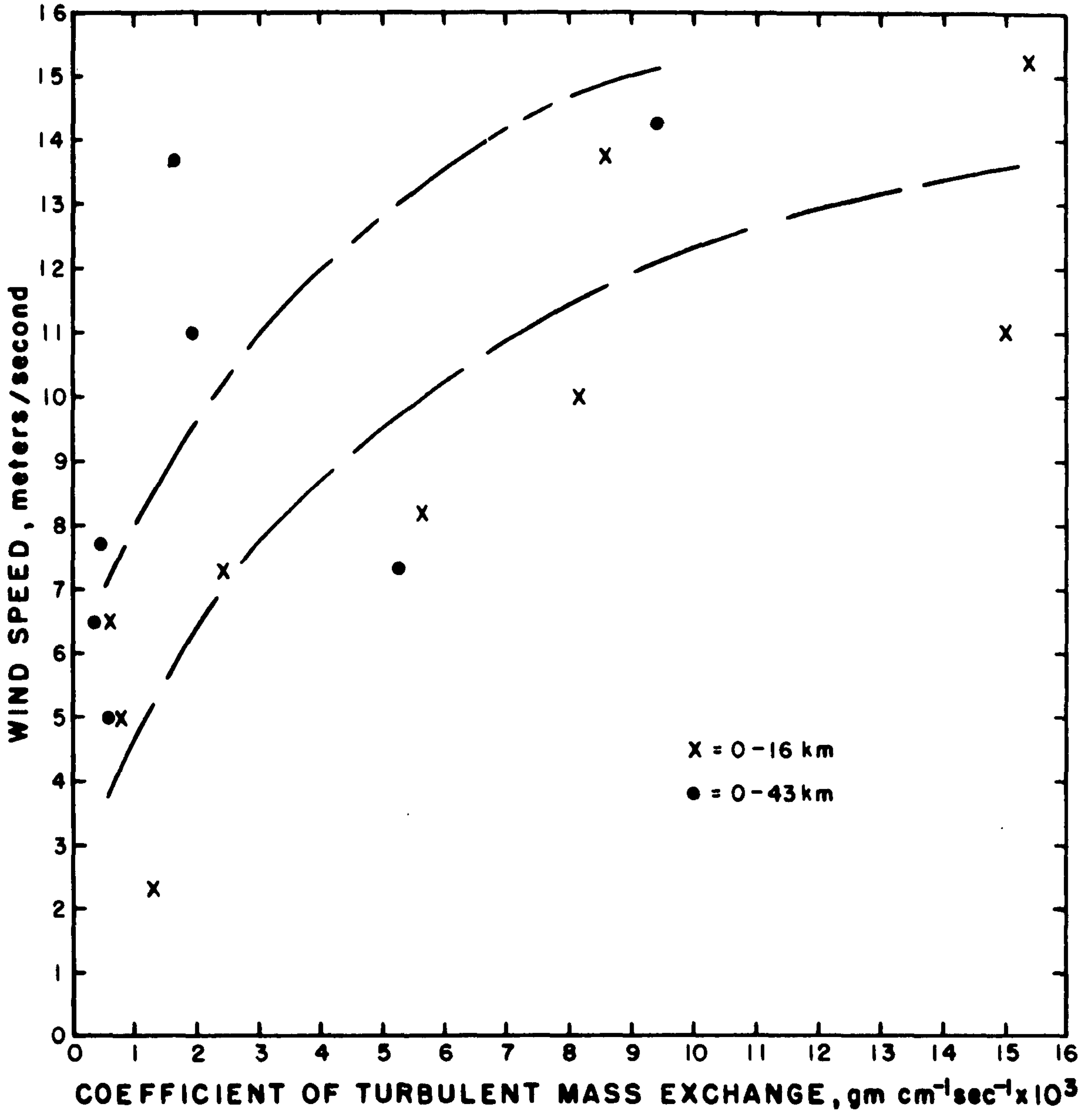
$gm\ cm^{-1}\ sec^{-1}$

Very Stable	Slightly Stable	Unstable
17 Lettau	400 Lettau	164-17700 Huss
	70-500 Mildner	130-15400 Bunker
	26-300 Mildner	
	105-825 Bunker	

This tabulation shows the variation of the coefficient with the stability of the air and compares values computed by the different authors. The coefficient of turbulent mass exchange is considered to be identical with eddy viscosity and the austausch coefficient. Values are expressed in $gm\ cm^{-1}\ sec^{-1}$. The table is divided into three columns representing the stability conditions existing at the time of the experiment. The heading "very stable" is used to denote the existence of an inversion. A value of the temperature lapse rate lying between the isothermal and the dry adiabatic lapse rate is classified as "slightly stable." By "unstable" it is meant that the lapse rate was super-adiabatic or that the water temperature was higher than the air temperature. The name following the value entered in the table indicates the author responsible for the determination. Lettau (1937) made his observations during two free balloon excursions. The value given is the exchange coefficient computed from equation (1). Mildner (1932) used pilot balloon observations to compute values of eddy viscosity. Huss and Portman (1948) computed austausch coefficients from wind observations after Ertel (1930). Bunker denotes values obtained in the present research.

The important role of stability in controlling turbulence is obvious from the table which shows that the ratio of values in different stability classes is about 25.

Since the source of energy for turbulence and eddy diffusion in the atmosphere is the wind shear and the kinetic energy of the wind, a relation might be expected between turbulent mass exchange and wind speed. Figure 7 shows that such a relation does exist. Here values of the coefficient obtained close to the shore are plotted against the average wind speed for the entire layer.



WIND SPEED, meters/second
COEFFICIENT OF TURBULENT MASS EXCHANGE, gm cm⁻¹sec⁻¹x10³
X = 0-16 km
● = 0-43 km

FIG.7

V Appendix

1. Supplementary Observations and Weather Conditions during the Observation Periods

Table V has been compiled to present supplementary observations taken during the days selected for the plane psychrograph observations. The values entered here are averages for the period of observation and are not necessarily characteristic of the entire day. The column headed pressure gives the atmospheric pressure as measured with a mecurial barometer. The column labeled Sky Conditions describes the amount and type of low cloud present in the sounding areas at the time of observation. Data was taken from the airplane observer's notes. The wind direction and speed presented in the table is the average for the homogeneous layer as determined for pilot balloon runs and surface data. Directions are in degrees while speeds are expressed in meters per seconds. Water temperatures were obtained as follows; (1) a dip bucket thermometer used either at the dock in Woods Hole Harbor or from a boat in or near the sounding area, (2) a bathythermograph in the sounding area, and (3) dip bucket temperature at the eastern entrance of the Cape Cod Canal, taken and made available by Mr. William Winsor, Deputy Sheriff of Barnstable Co. The saturation mixing ratio is given in the last column. It was computed from the water temperature and the surface pressure.

2. Graphs of Soundings and Discussion of Data

In Figures 8-23, potential temperature and mixing ratios are plotted against computed heights expressed in meters. Four moisture cross sections and one upper air map are included in the series of

Table V

Supplementary Observations

Date	Atmospheric Pressure mb	Sky Conditions	Wind Direction degrees	Wind Speed m/s	Water Temperature °C	Saturation Mixing Ratio gm/kg
Sept. 23, 1947	1026	No low cloud	310	2.3	19.9	14.5
Oct. 1, 1947	1025	No low cloud	315	7.3	16.6	11.7
Nov. 9, 1947	1007	5/10 stratus	290	8.2	12.2	9.9
Nov. 10, 1947	1020	No low cloud	305	10.0	12.2	8.8
Nov. 13, 1947	1013	5/10 cumulus	305	15.2	10.8	8.0
Nov. 17, 1947	1011	No low cloud	310	13.7	8.6	7.0
Sept. 1, 1948	1019	2/10 cumulus	340	11.0	21.6	15.2
Sept, 15, 1948	1025	5/10 over mainland	280	4.8*	17.8	12.8
		nil over water	320	6.4**		
Sept. 16, 1948	1032	3/10 cumulus	360	7.7	18.0	12.7
Sept. 22, 1948	1008	No low cloud	317	6.5	18.9	13.8
Oct. 13, 1948	1012	No low cloud	246	14.3	15.0	10.7
Nov. 16, 1948	1029	No low cloud	005	4.8	11.4	8.3

* 0-700 m

** 700-2100 m

figures. Large dots are used in plotting the potential temperatures. Usually only one sounding of the series is plotted, but if a significant change occurs in the temperature, additional points of another sounding are plotted with a large X. The water temperature, reduced adiabatically as though it were an air temperature to 1000 mb, is indicated by an arrow and θ_s. Mixing ratios are plotted with the following symbols; θ for the first sounding taken over the land, • for the second sounding of the series, and an x for the last sounding. In most cases all values of the mixing ratio have been plotted. Omissions have been made only to relieve crowding.

The inversion at the top of the homogeneous layer has been indicated by a horizontal line, although it is easily distinguishable by the decrease in mixing ratio and the increase in potential temperature.

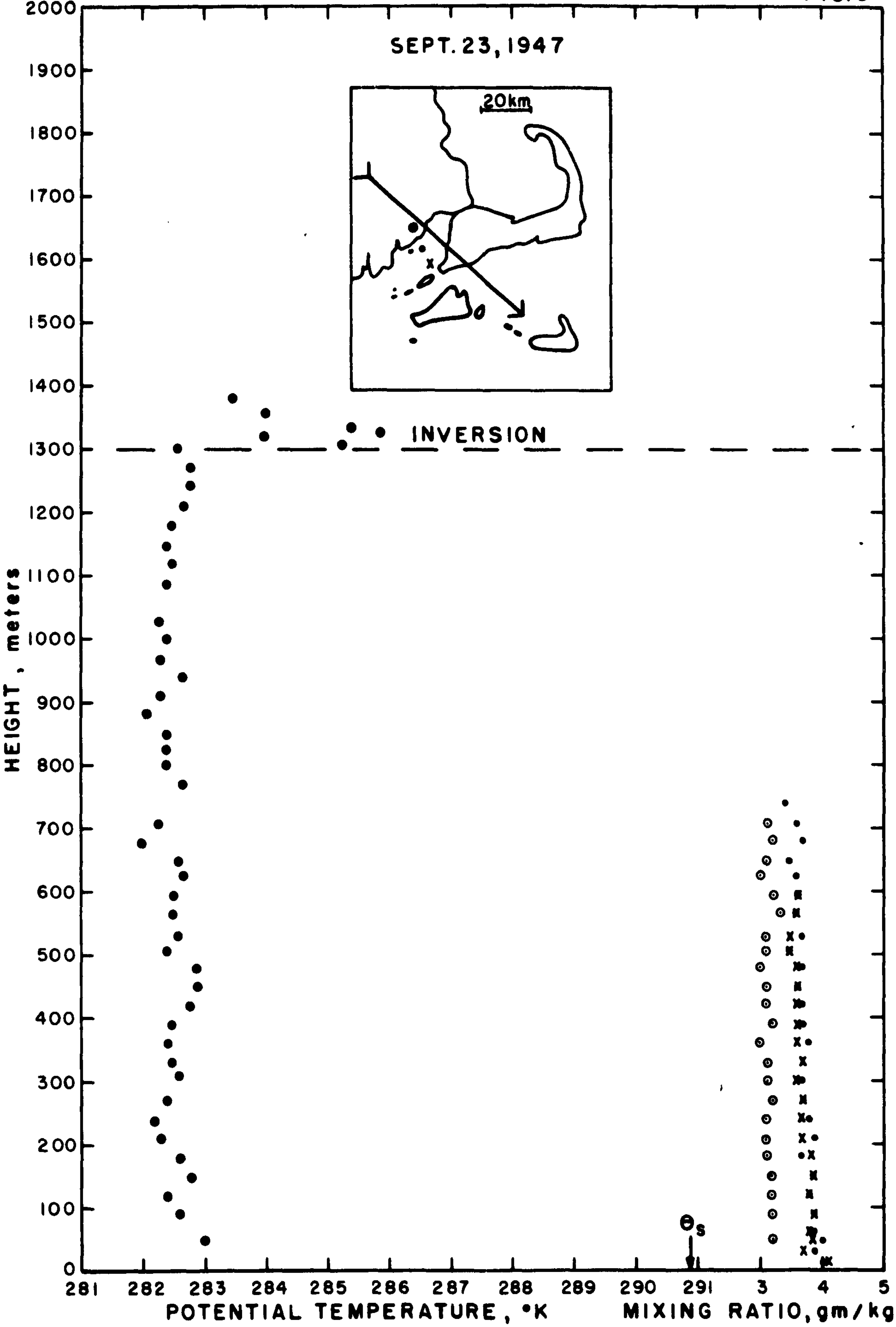
The map of southeastern Massachusetts is inserted to show the location of the sounding areas and the wind direction. The symbols indicating sounding areas correspond to the ones used in plotting mixing ratios.

A discussion of the days' soundings accompanies the graphs.

Sept. 23, 1947

This series, the first of all the observations, gives mixing ratio values to only 700 meters since the resistance of the wet-bulb temperature element exceeded the range of the amplifier at this height. The dry-bulb temperature continued on scale until the top of the homogeneous layer was reached. The computation of accumulation of moisture was

FIG. 8


SEPT. 23, 1947
20km
INVERSION
HEIGHT, meters
2000
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
281 282 283 284 285 286 287 288 289 290 291 3 4 5
Θ_s
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

Sept. 23,1947

Top of Homogeneous Layer

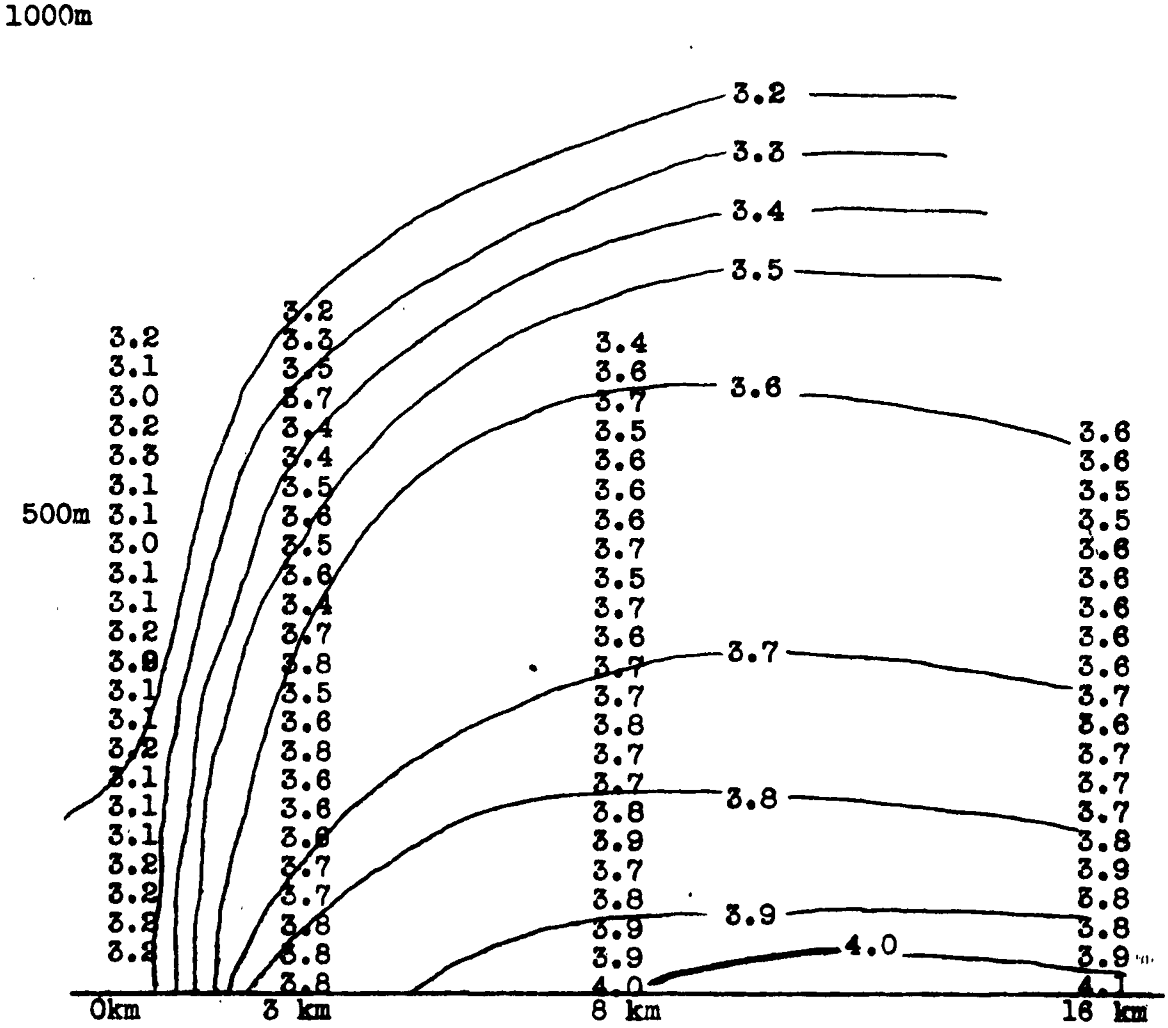


Figure 9

made by assuming values of the mixing ratio between the 700 m level and the top of the homogeneous layer at 1200 m. This series has the peculiarity that a large increase of moisture occurred between the first and second soundings taken 3 km apart. For computations of water vapor flow, the average of these two soundings were used as the initial point.

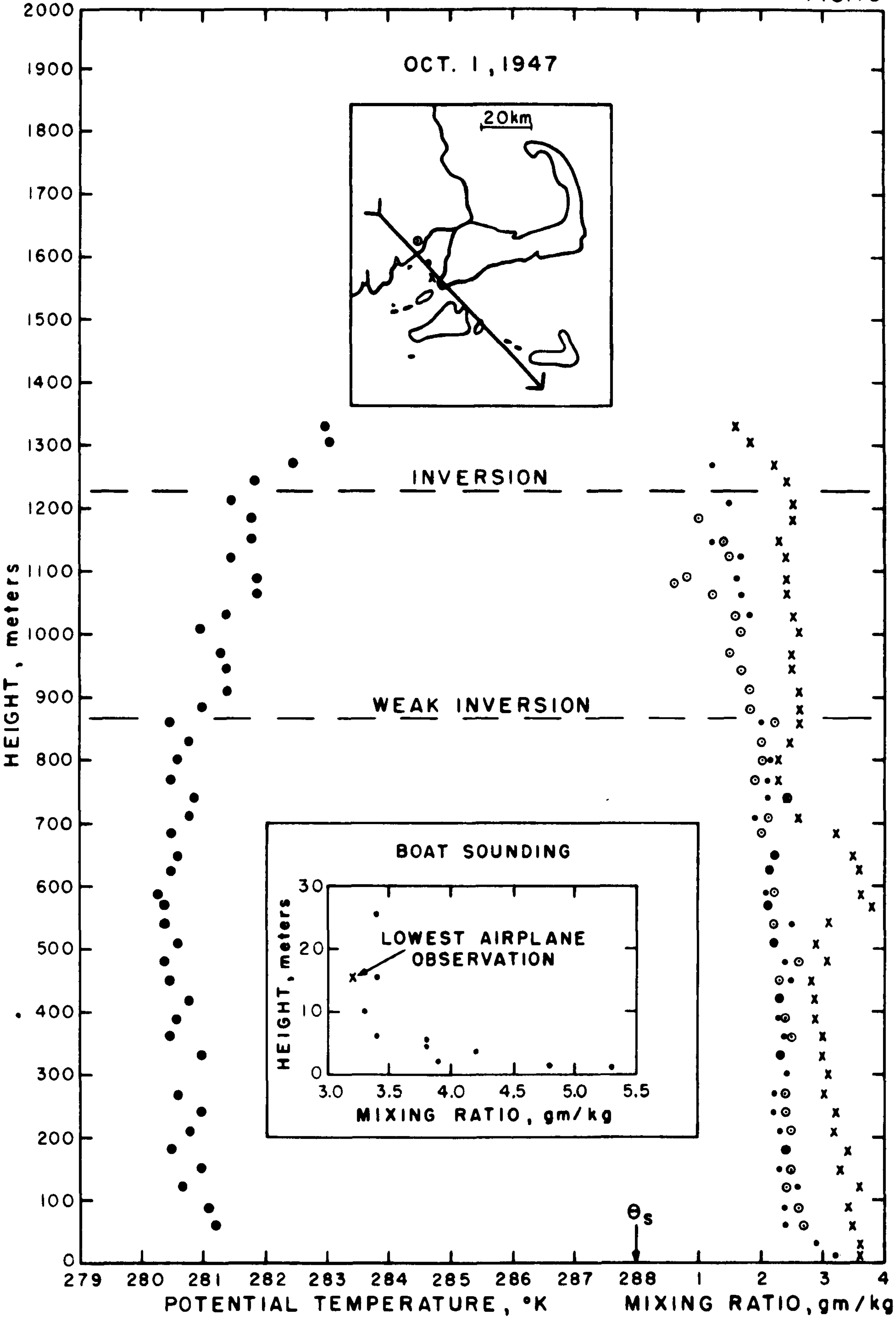
A mixing ratio cross section has been prepared to show the rapid increase near the windward shore. Values of the mixing ratio have been entered at the proper height and distance from the initial sounding. Values of equal water vapor content have been connected by solid lines. The height of the homogeneous layer is indicated by short horizontal lines.

Oct. 1, 1947

The four soundings of this series show two peculiarities worthy of note. One is that the second and third soundings obtained in the main body of the homogeneous layer have a smaller mixing ratio than the sounding taken overland. Convective action might account for this although no cumulus clouds were present in the area. The second phenomenon noticed is a large mass of air centered at 600 m with a very high mixing ratio. The value of 3.8 gm/kg is 0.2 higher than the value measured at 15 m. A cross section drawing is given to show these occurrences.

Values of the flow of water vapor were found by using the average of the first two soundings for the initial point. The flow for the 8-16 km distance is abnormally high because of the existence of the moist bubble of air just mentioned.

FIG. 10



OCT. 1, 1947
20km
INVERSION
WEAK INVERSION
BOAT SOUNDING
LOWEST AIRPLANE OBSERVATION
HEIGHT, meters
MIXING RATIO, gm/kg
θs
HEIGHT, meters
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

Oct. 1,1947

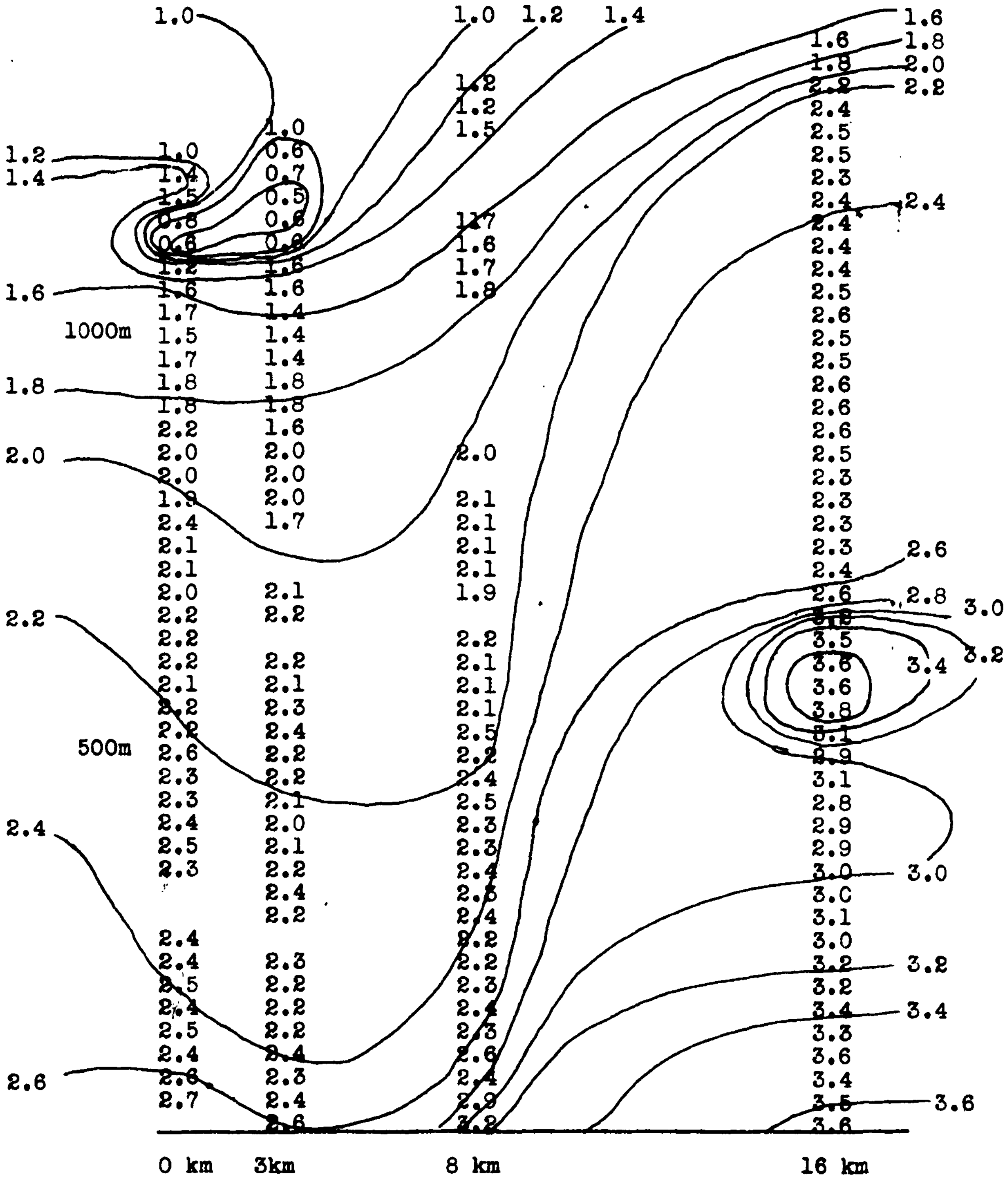


Figure 11

The captive balloon soundings obtained are presented in the insert of Figure 10 for the instruments were working perfectly on this day and the measurements are reliable. These soundings are used in conjunction with the airplane soundings to determine values of the exchange coefficient at frequent height intervals. Earlier in this report, it was pointed out that values could not be accurately determined for small height ranges. Values for this day are computed at five heights to show the magnitude of the variation. The accuracy of these values is not great for it is limited by the eddy nature of the atmosphere. Table VI gives the observed gradients determined graphically. Table VII gives the flow of water vapor through the surfaces and the values of the exchange coefficient.

Table VI

Mixing Ratio Gradients

$cm^{-1} \times 10^{-9}$

Heights m	Sounding #1	Sounding #2	Sounding #3	Sounding #4
1-5	-	-5800	-4800	-2600
5-10	-	-	-28	-28
50-150	-32	-10	-7	-12
300-700	-7	-3	-6	-9
700-1000	-17	-16	-9	-4

Table VII

Water Vapor Flow

gm cm^{-2} sec^{-1} x 10^{-5}

	Sounding #1 to #3	Sounding #3 to #4	Sounding #1 to #4
0	2.4	11	6.7
90	2.0	10	6.2
510	1.8	6.9	4.4
1005	1.5	2.5	2.0

Coefficient of Turbulent Mass Exchange

gm cm^{-1} sec^{-1}

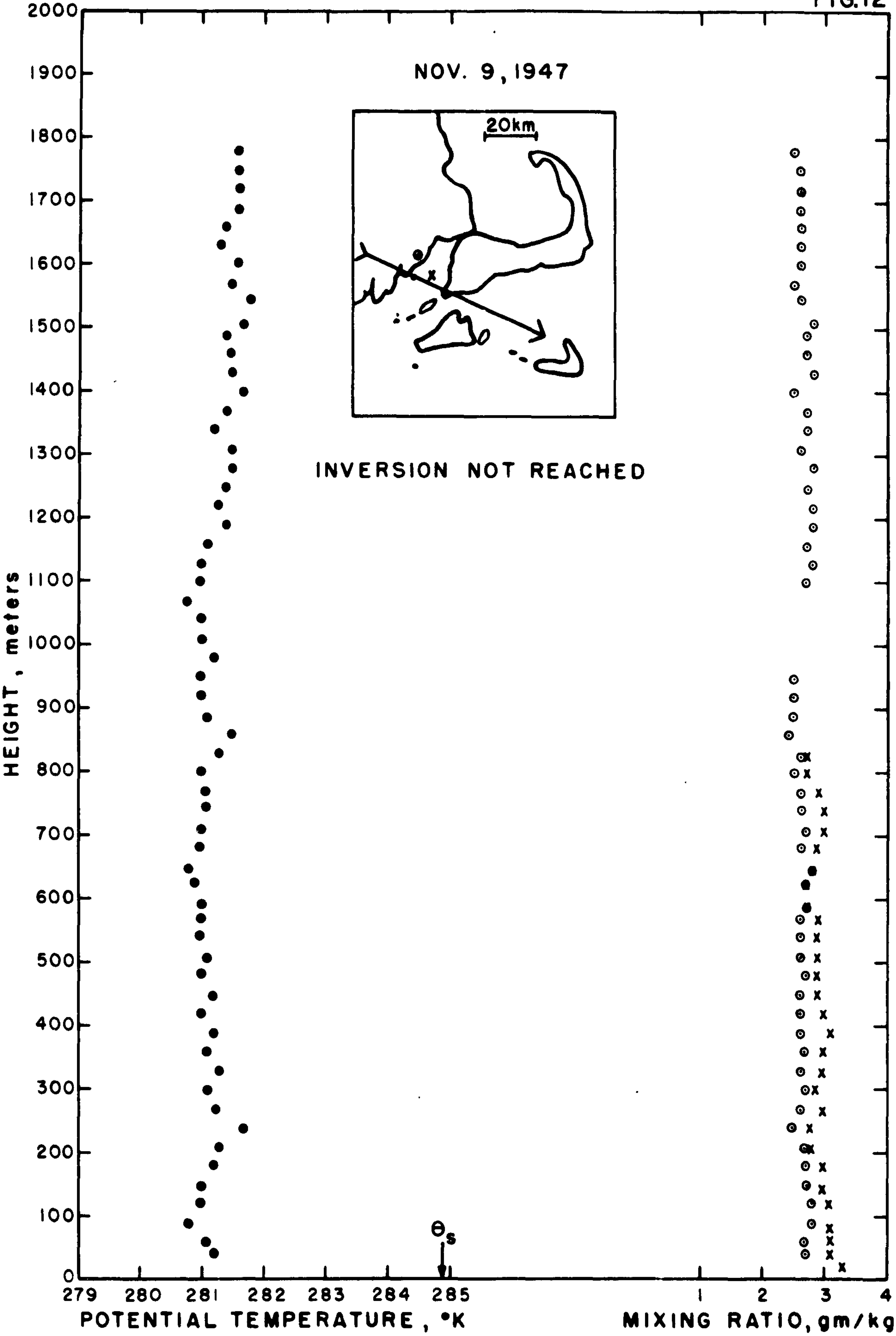
1-5	4.5	30	15
5-10	850	3900	2400
90	1500	10500	5400
510	3400	9200	6900
1005	1400	3800	2300

Nov. 9, 1947

The air mass was colder this day causing the wet bulb to freeze at - 3.0°C. Values of the mixing ratio were determined by using saturation values over ice and the heat of sublimation of ice in the psychrometric equation. The sounding in the middle of the bay, where the airplane descended had to be omitted for the frozen wick began thawing at some undetermined height.

The inversion at the top of the homogeneous layer was not reached at 6000 feet, but little error can be introduced by totaling moisture to this level. 5/10 coverage of stratus clouds occurred at times in the sounding area above the plane. The clouds had little or no vertical development and probably did not pierce the inversion and allow the moisture to mix into the upper air.

FIG. 12


NOV. 9, 1947
20km
INVERSION NOT REACHED
HEIGHT, meters
0
100
200
300
400
500
600
700
800
900
1000
1100
1200
1300
1400
1500
1600
1700
1800
1900
2000
279 280 281 282 283 284 285
Θ_s
1 2 3 4
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

Nov. 10, 1947

The wet bulb froze on this day permitting only the ascents to be used in the reduction of mixing ratio values. The height of the homogeneous layer had lowered so that the sounding extended up through the inversion.

Nov. 13, 1947

The top of the homogeneous layer was not reached on this day. 5/10 cumulus clouds were observed, creating the possibility that the determined value of the flow is too low due to leakage of moisture through the inversion.

Nov. 17, 1947

The wet bulb froze again on this day, but the sounding technique had been changed to making observations only during the ascent of the plane. This procedure allowed the wet bulb to thaw during the lower part of the descent, and to be ready for observations after a brief run at the lowest level. The inversion was reached and penetrated by the plane during the first sounding only.

A cross section of the region is included to explain the negative value of transport given in Table I. It is seen that the values increased greatly between the first and second sounding, while the values dropped in the 400-600 m region between the second and third soundings. In spite of this there has occurred a large increase in the water vapor between the first and third soundings. This example serves as a reminder of the eddy nature of the air, and that fine detail cannot be derived from a series of single ascents.

FIG. 13

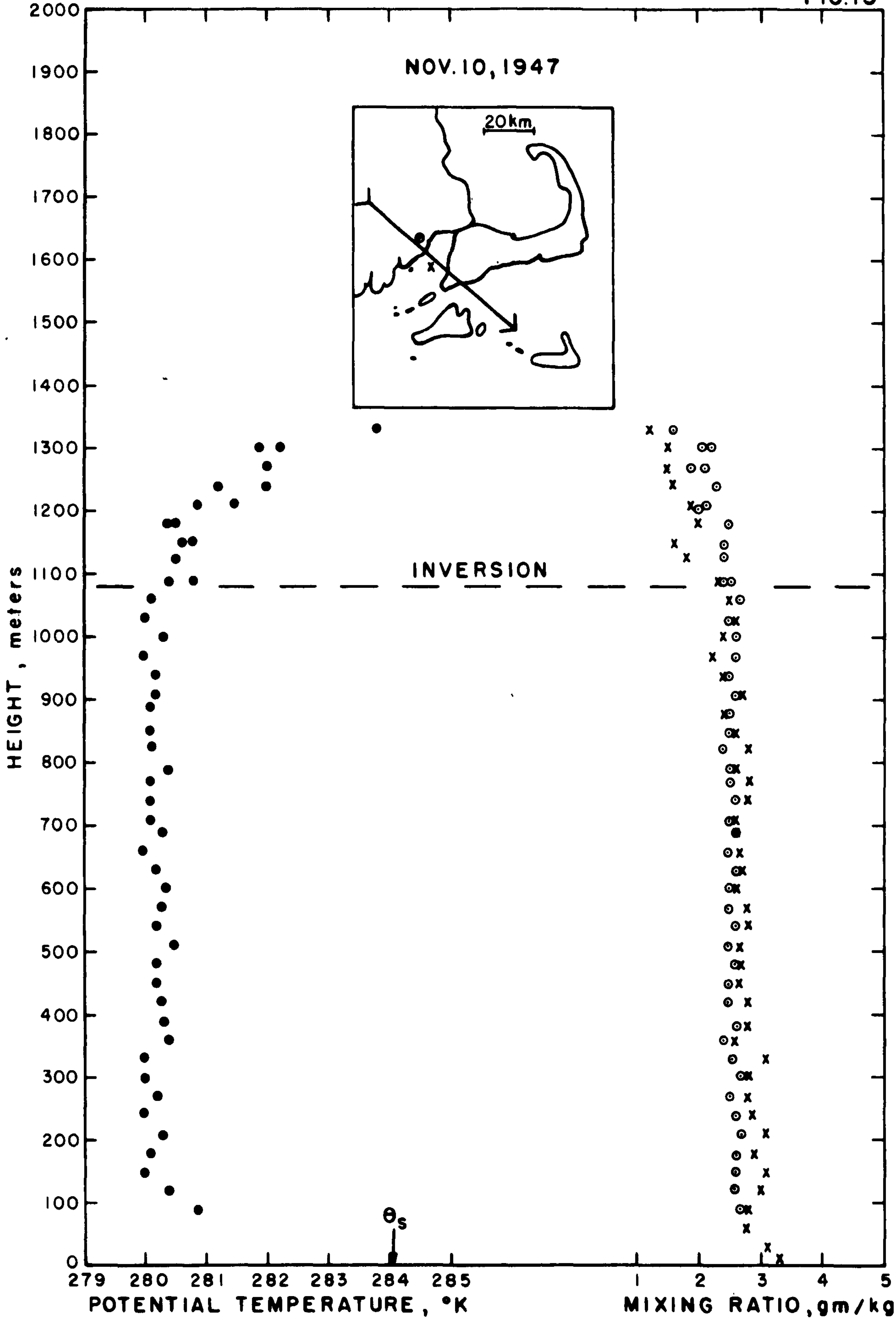
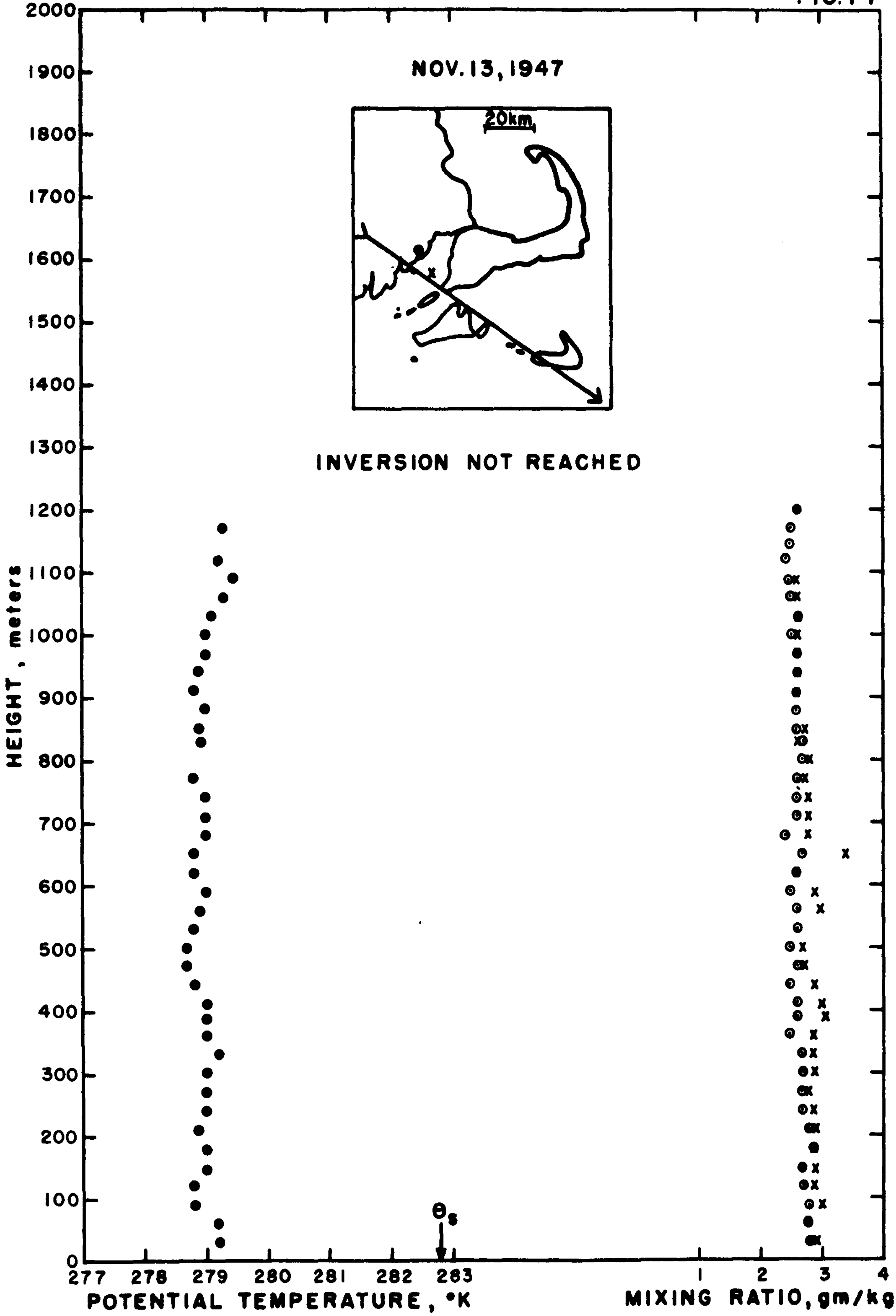
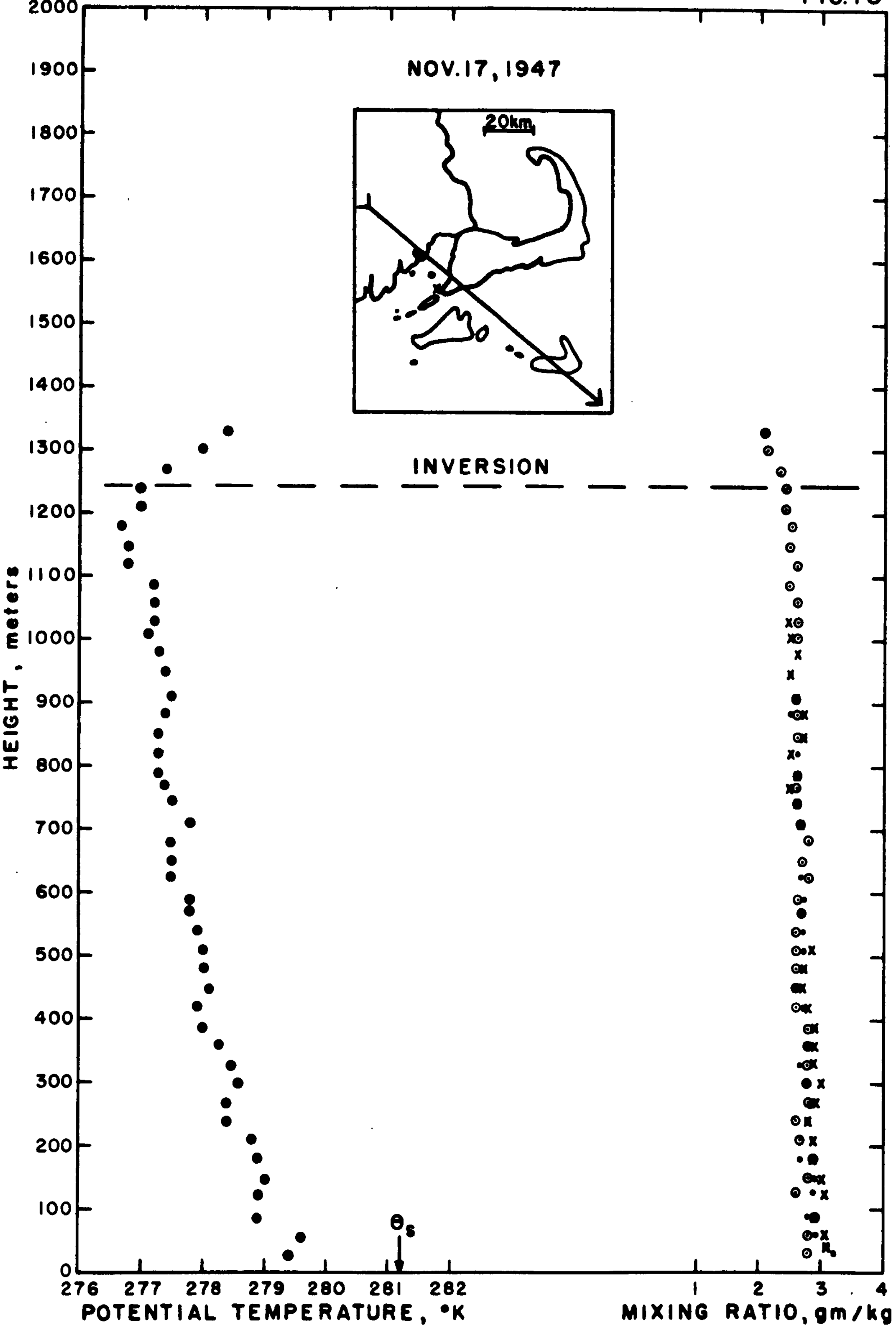

NOV. 10, 1947
20 km
INVERSION
HEIGHT, meters
0
100
200
300
400
500
600
700
800
900
1000
1100
1200
1300
1400
1500
1600
1700
1800
1900
2000
θ_s
279 280 281 282 283 284 285
1 2 3 4 5
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

FIG. 14


NOV. 13, 1947
20km
INVERSION NOT REACHED
HEIGHT, meters
0
100
200
300
400
500
600
700
800
900
1000
1100
1200
1300
1400
1500
1600
1700
1800
1900
2000
277
278
279
280
281
282
283
θ_s
POTENTIAL TEMPERATURE, °K
1
2
3
4
MIXING RATIO, gm/kg


NOV. 17, 1947
20km
INVERSION
HEIGHT, meters
2000
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
276 277 278 279 280 281 282
1 2 3 4
θ_s
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

Nov. 17,1947

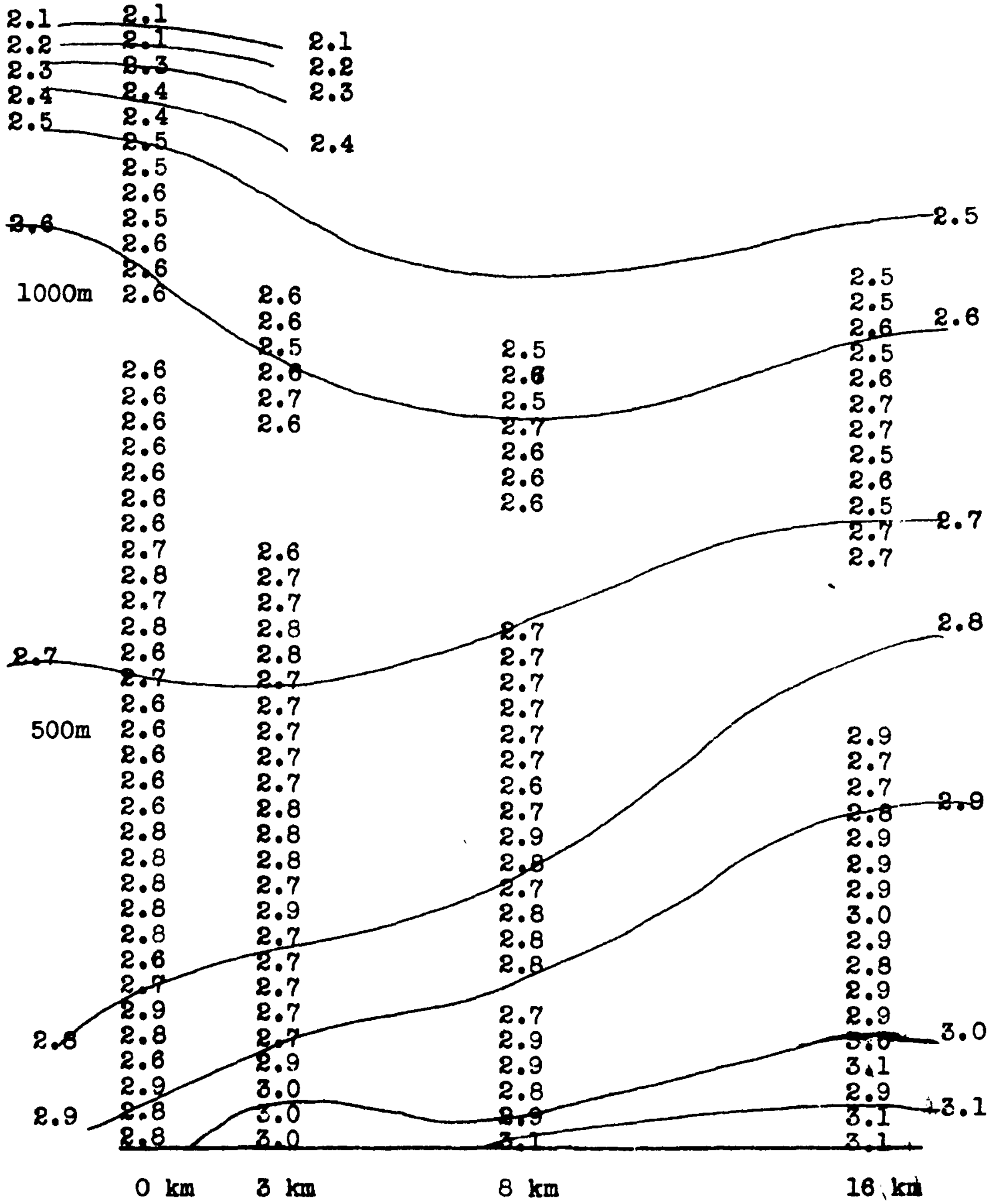


Figure 16

Sept. 1, 1948

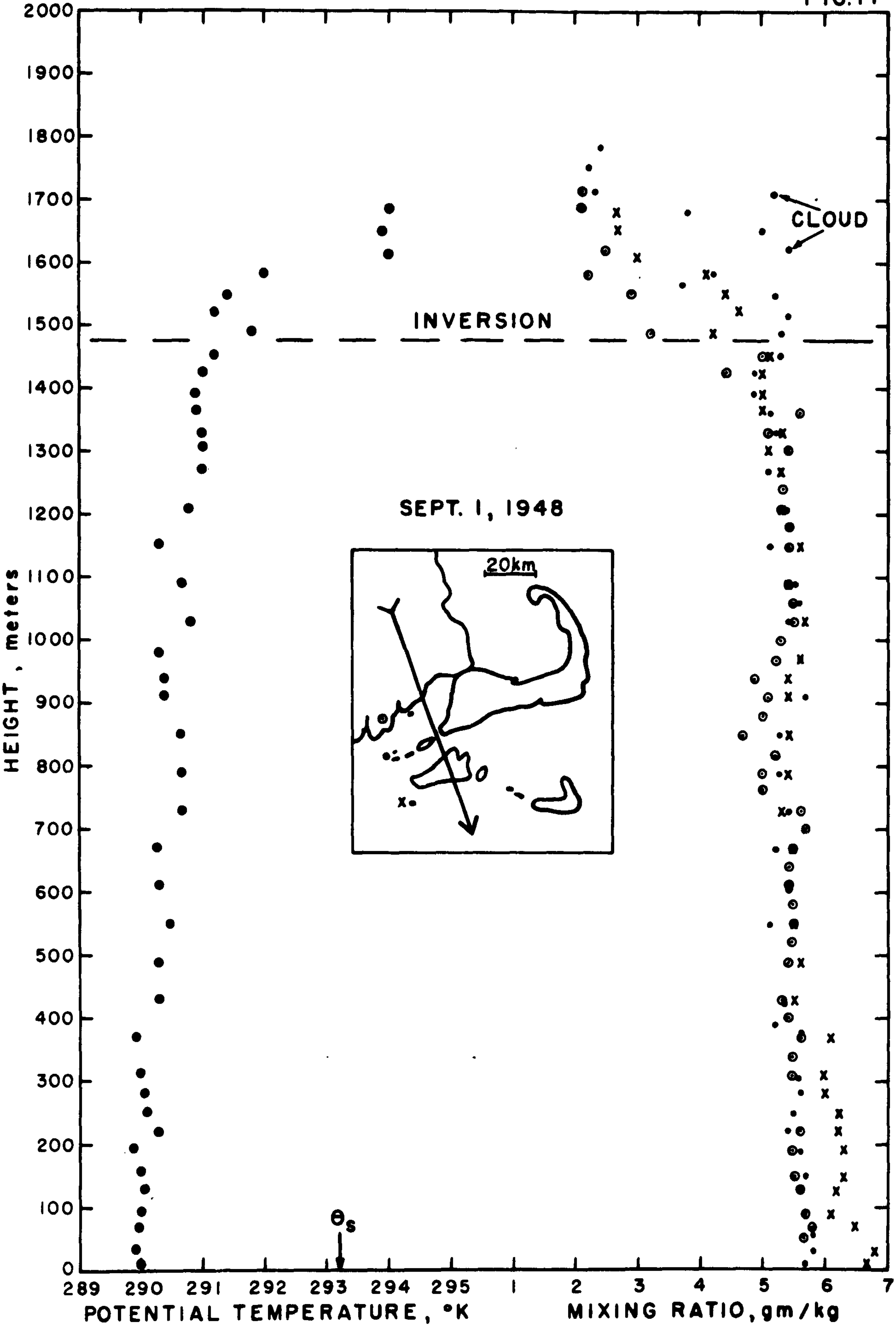
The direction of the sounding path was within 5° of the average wind determined for the homogeneous layer. 2/10 coverage of cumulus clouds were observed in the area. A cumulus cloud with small vertical structure developed within the spiral of the second sounding. Two runs were made through the cloud yielding mixing ratio values. The appearance of the cloud and the accompanying convective currents in the ascent spiral caused the moisture distribution to change radically from the normal. This led to uncertain values of the transport of water vapor, and explains the large values occurring between the first and second soundings and the negative transport between the second and third. For this reason the values obtained from the first and third are much more representative of transport by eddy diffusion.

The graph of the mixing ratio of the second sounding shows a separation of the values into two groups above the top of the homogeneous layer. One follows the decreasing curve of the first and third soundings while the other branch continues with the high values of the homogeneous layer up to the 1600-1700 m level where the cloud existed. A cross section of the air has been drawn to show the effect of the convective activity on the moisture distribution. The increase of moisture with distance travelled over the water and the cumulus clouds are shown clearly.

Sept. 15, 1948

The usual procedure of computing water vapor flows and exchange coefficients had to be completely modified to analyze the soundings of

FIG. 17



HEIGHT, meters
2000
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
INVERSION
CLOUD
SEPT. 1, 1948
20km
θ_s
289
290
291
292
293
294
295
1
2
3
4
5
6
7
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

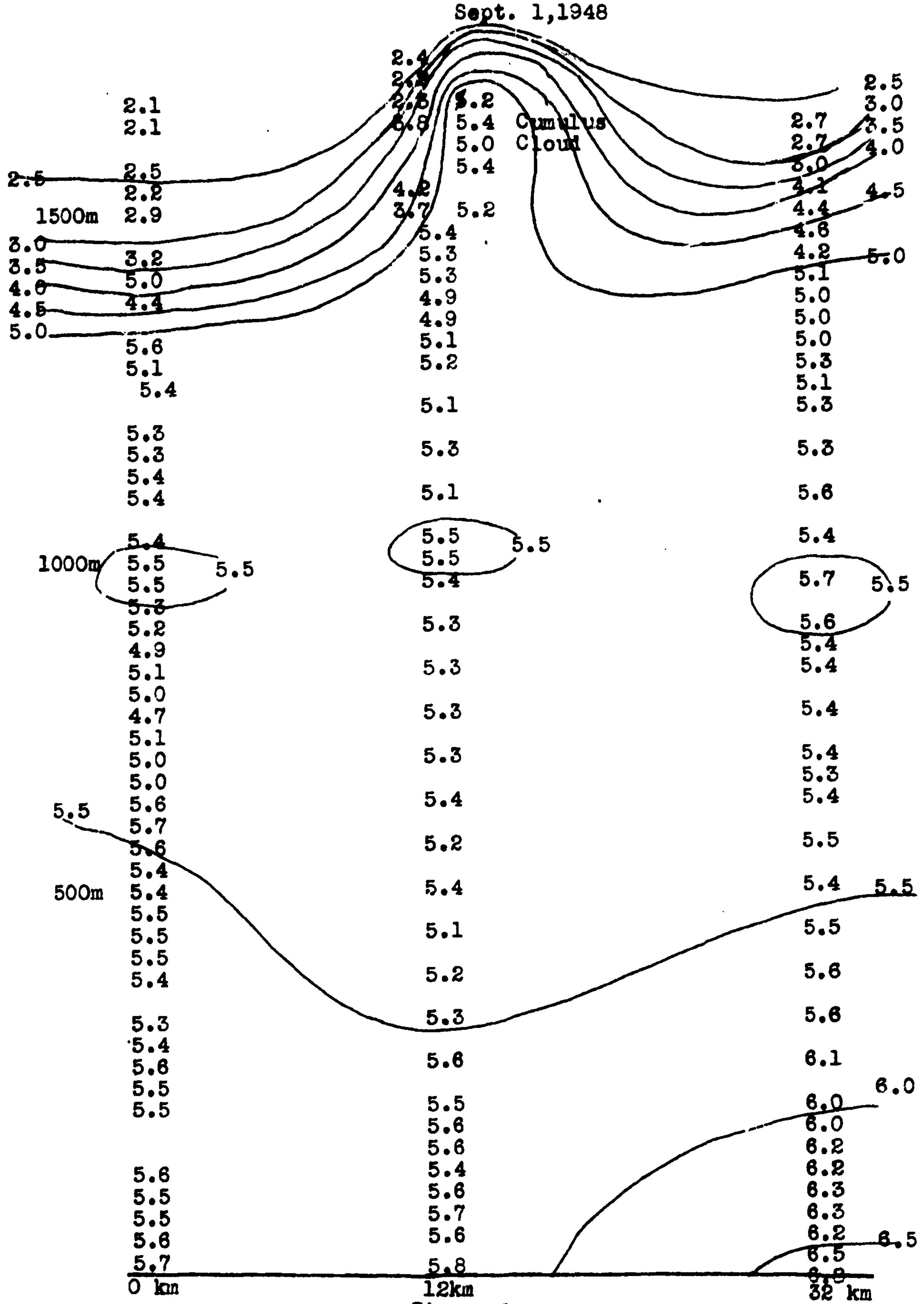


Figure 18

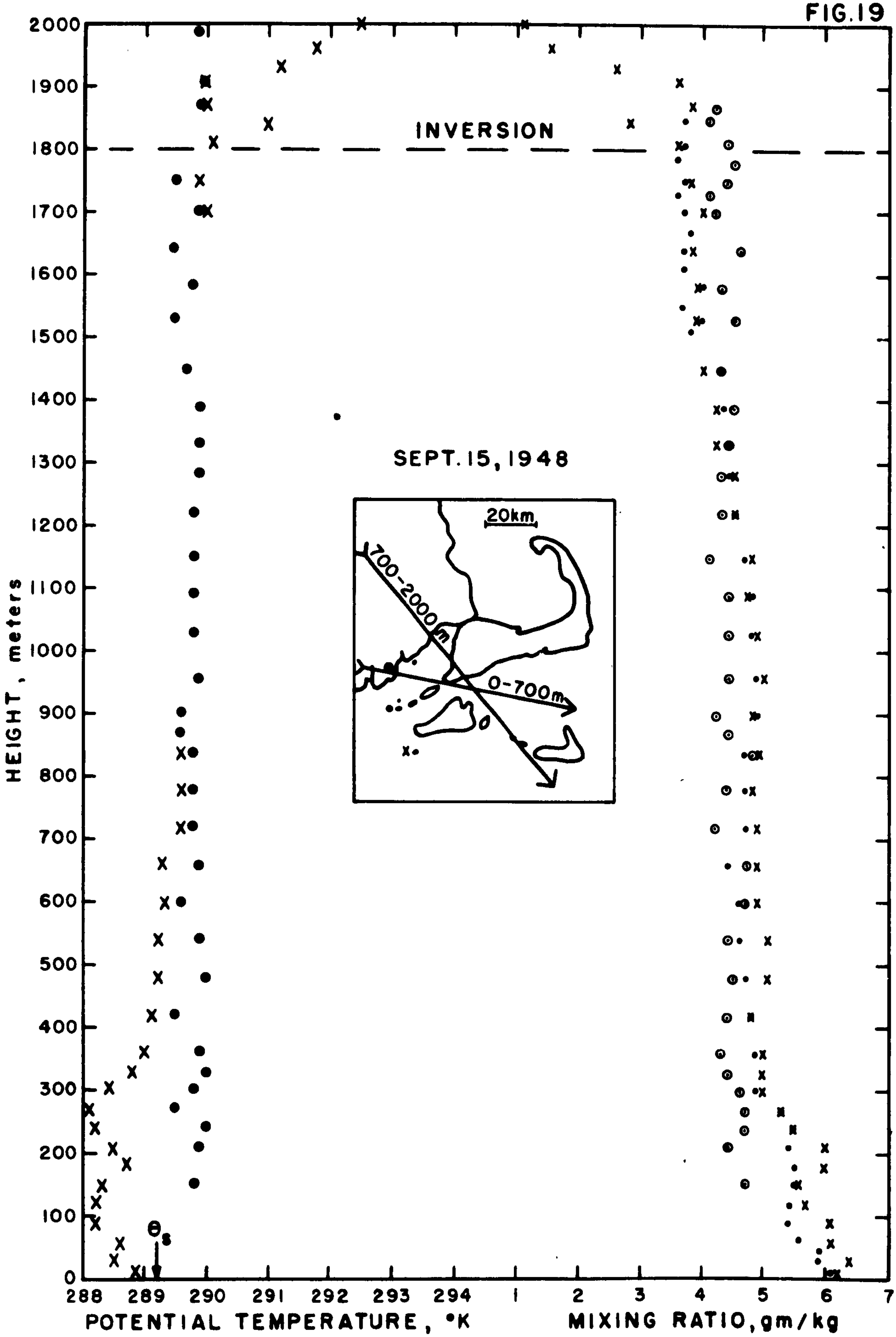
this day. The wind structure was not characteristic of the homogeneous layer for the layer was divided into two streams flowing in different directions. Pilot balloon observations showed a layer from 0-700 m having a 280° 4.8 m/s wind, and a layer from 700-2100 m having a 320° 6.4 m/s wind. The wind of the lower layer was variable, showing variations from 40° 0.6 m/s to 330° 3.0 m/s in two hours at Woods Hole. In determining the average wind of the lower layer, great weight was assigned to readings taken from a boat near the base of the second sounding at the time that the airplane was making the second sounding.

The procedure adopted to derive values of flow for this case is to (1) compute the flow into the top stream of air through the 700 m surface in the usual manner, (2) compute from this flow the amount of vapor that leaves the lower layer during its passage from land to the next sounding area, (3) add this loss of vapor to the accumulation found in the lower layer to determine the flow of vapor into the layer from the water surface below it. Figure 20 shows the two currents of air drawn in green and red and the geography of the region. Values of the exchange coefficient are found for the two layers.

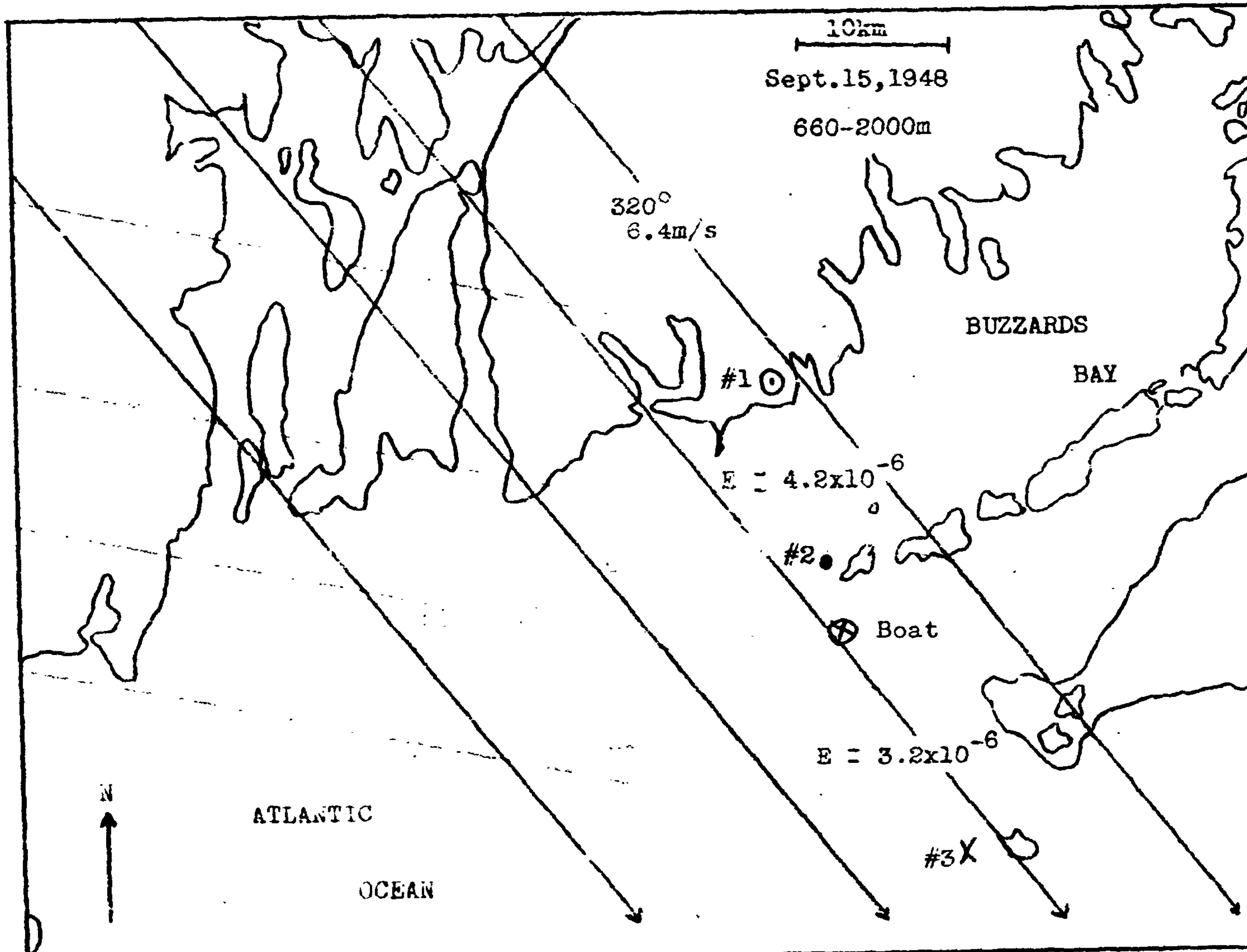
Sept. 16, 1948

The average wind was 360° while the orientation of the sounding areas was 350°. The air blew south over the waters of Buzzards Bay and passed over parts of the Elizabeth Islands before being measured again. The values of transport and diffusion are modified somewhat by the passage over the islands and are not strictly typical of transport from water surfaces.

FIG. 19



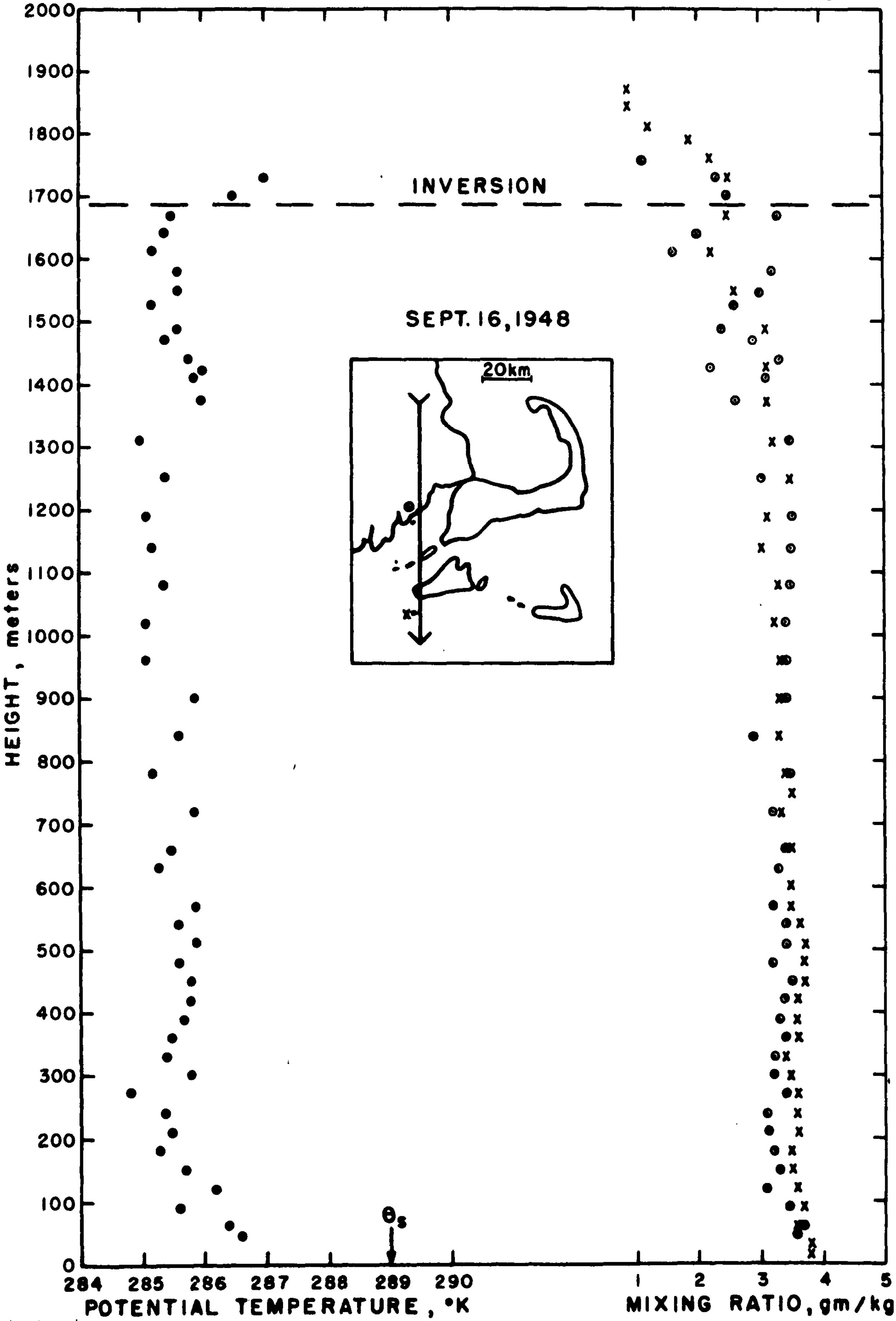
INVERSION
SEPT. 15, 1948
20km
700-2000m
0-700m
HEIGHT, meters
2000
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
Θ_s
288 289 290 291 292 293 294
1 2 3 4 5 6 7
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg



10km
Sept.15,1948
660-2000m
320°
6.4m/s
BUZZARDS
BAY
#1
E = 4.2x10^-6
#2
Boat
E = 3.2x10^-6
#3
N
ATLANTIC
OCEAN

Figure 20

FIG. 21



INVERSION
SEPT. 16, 1948
20km
HEIGHT, meters
2000
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
θ_s
284
285
286
287
288
289
290
POTENTIAL TEMPERATURE, °K
1
2
3
4
5
MIXING RATIO, gm/kg

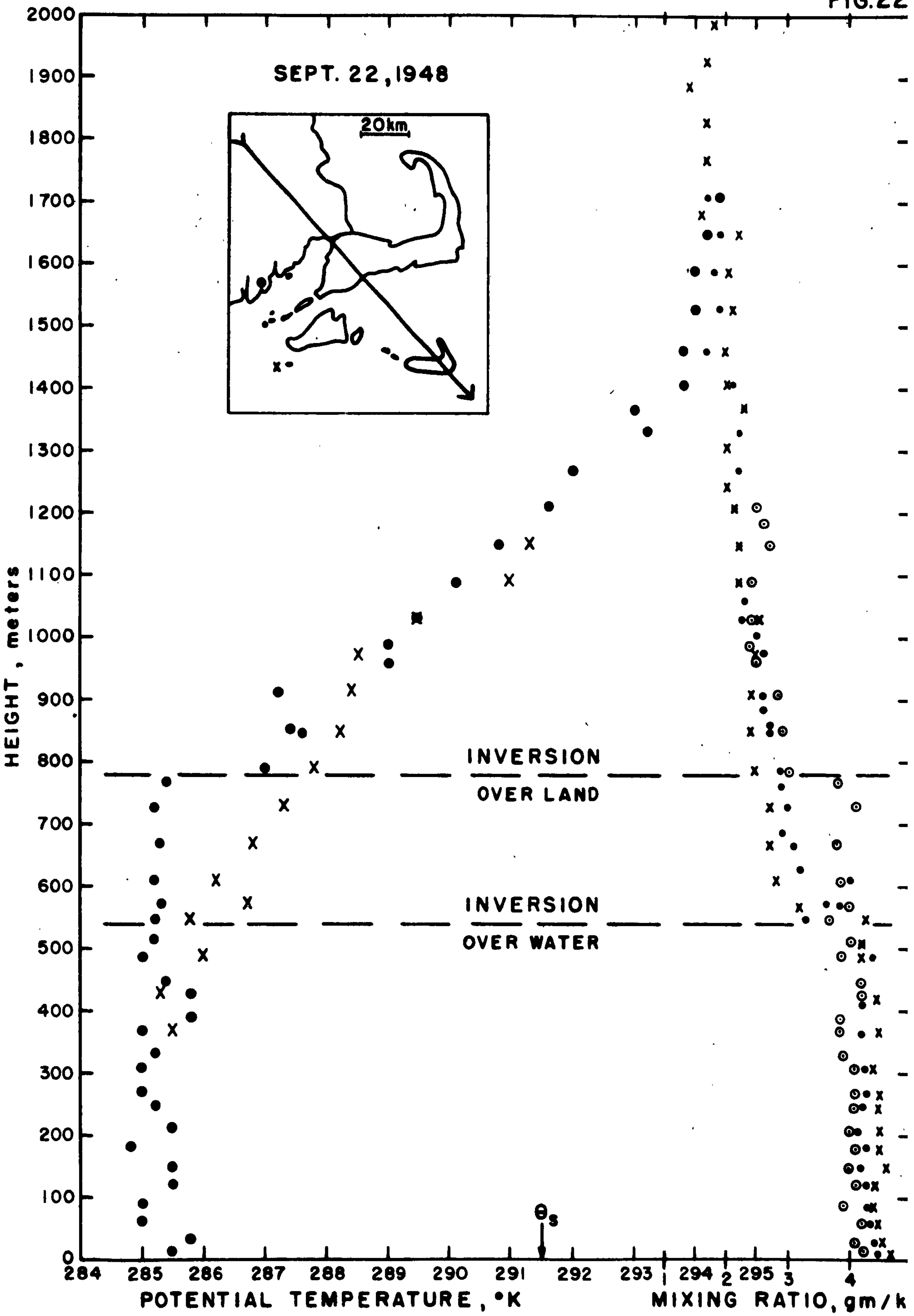
Sept. 22, 1948

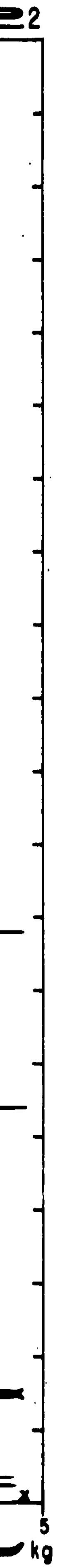
This series has been mentioned briefly in an earlier part of the report for it offered data that has been used to evaluate the mixing of water vapor through the inversion above the homogeneous layer. This computation could be made because the air above 600 m was isothermal to 2000 m. It was noted that this stable air lost water vapor as it travelled over the water. This would only mean that the vapor was mixed into the upper air and that the mixing in the stable air was many times greater than the mixing through the strong inversion below. Thus any measurement of the transport of vapor in the stable air must be many times larger than by the transport through the inversion. The computation was made by measuring the depletion of moisture in the 600 m-1400 m region and assuming that all of this amount mixed upward through the 1400 m surface. This leads to a value of 4.7×10^{-6} gm cm^{-2} sec^{-1}, corresponding to an exchange coefficient of 200 gm cm^{-1} sec^{-1}. Since the potential temperature lapse rate was less in the stable layer than in the thin inversion layer, the transport through the inversion must be several times less. Under the assumption that the transport is about five times less in the boundary inversion, its value will be 1×10^{-6} gm cm^{-2} sec^{-1}. The moisture gradient has been measured by inspection to be $- 18 \times 10^{-8}$ cm^{-1}. This leads to an exchange coefficient of 6 gm cm^{-1} sec^{-1}.

The computed values of the transport and exchange in the homogeneous layer were obtained in the usual manner.

The value of A for the stable region may be identified with the "residual turbulence" discussed by Rossby and Montgomery (1935). The

FIG. 22


SEPT. 22, 1948
20 km
INVERSION
OVER LAND
INVERSION
OVER WATER
HEIGHT, meters
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/k
θ_s
0
100
200
300
400
500
600
700
800
900
1000
1100
1200
1300
1400
1500
1600
1700
1800
1900
2000
284
285
286
287
288
289
290
291
292
293
294
295
2
3
4



value of 50 gm cm^{-1} sec^{-1} was assumed by them, after Ri
be a reasonable value. The value of 200 determined her
agreement considering the different manner of computati

Oct. 13, 1948

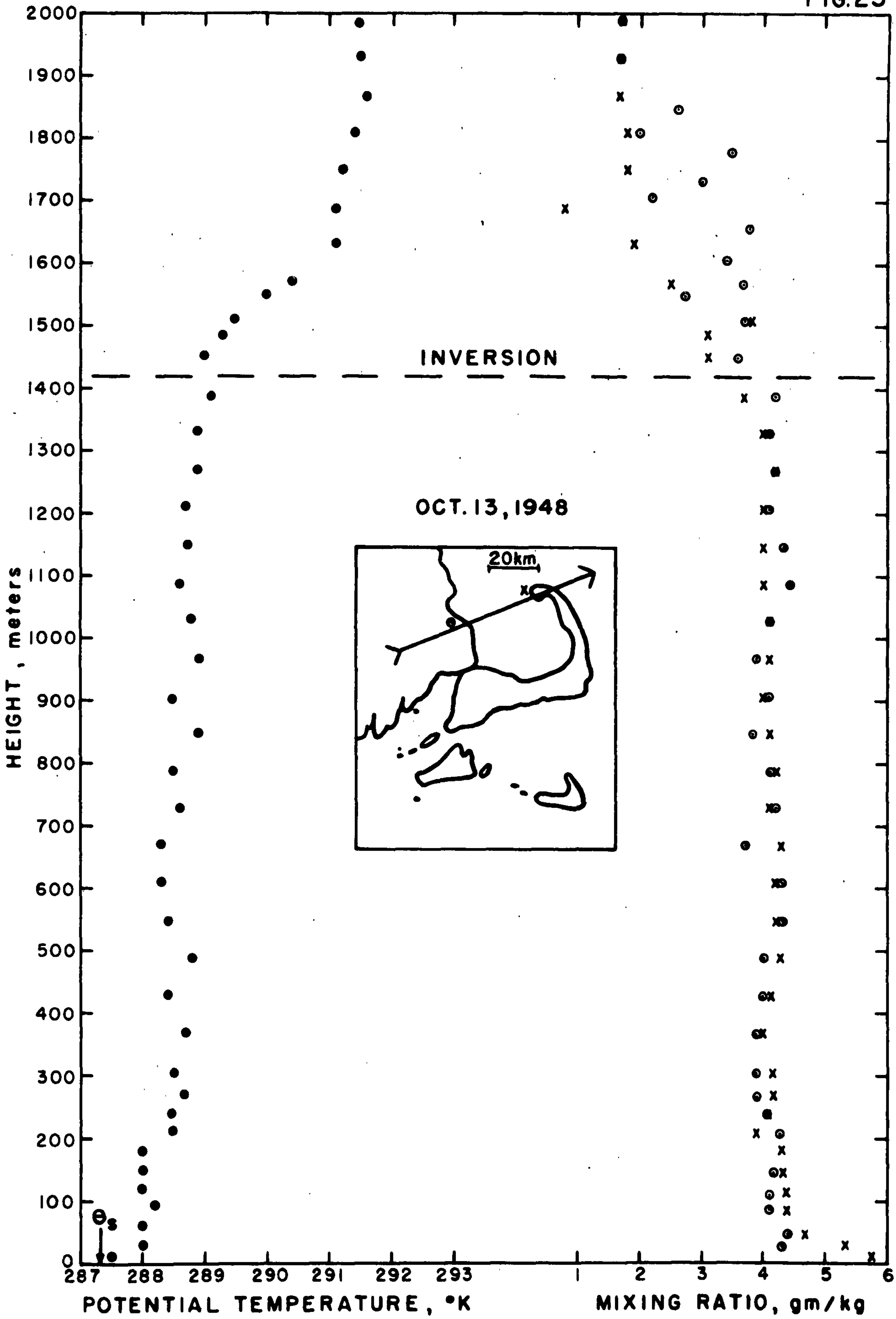
A sounding area was selected in Cape Cod Bay to ta
of a coastline nearly perpendicular to the 245° wind.
sounding was taken at Plymouth, the other at Race Point
distance. Thermal conditions of this series were diffe
others in that the air was 2°C warmer than the water.
the development of an inversion over the water with a c
decrease in turbulence and transport of water vapor. T
of transport is shown in Figure 6.

The computations of the exchange coefficient for t
from equation (1) give a value of 103 gm cm^{-1} sec^{-1}, wh
ent and reasonable. The high value of 9400 for the upp
the homogeneous layer offers an interesting puzzle. Th
consistent with the overland values for the wind speed
question that arises is whether the turbulence can cont
rate over cooler water for a distance of 30 km. The fa
the solution difficult to analyze is that the mixing ra
(6.7×10^{-10} cm^{-1}) overland was positive. If this were
the average as being a chance value depending on the ex
large parcel of moist air at 1100 m, then the value of
reduced to 1500 gm cm^{-1} sec^{-1}, a value that seems more
the conditions.

Nov. 16, 1948

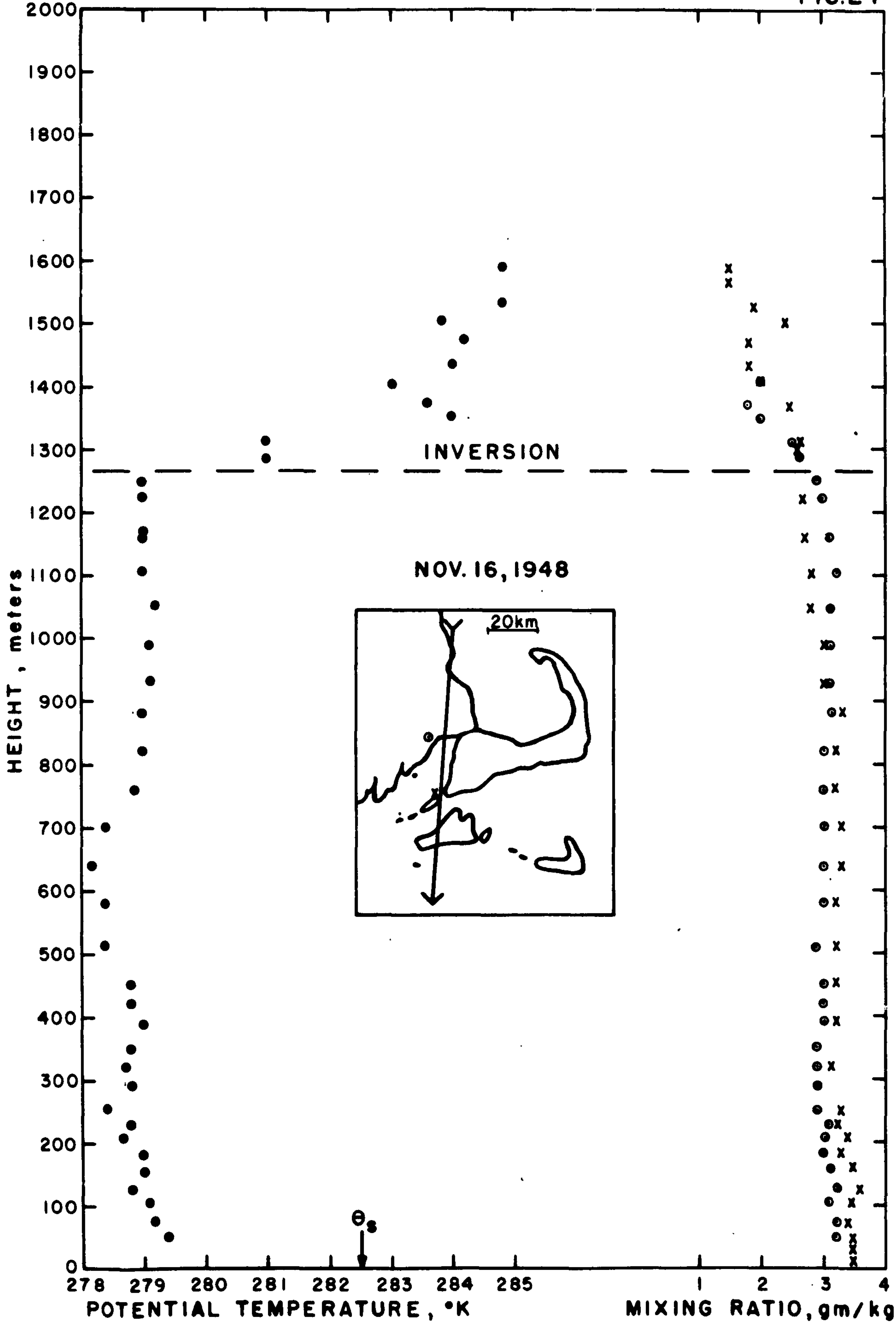
This is a typical case of modification of air by i
water. The wind was steady and nothing occurred to com

FIG. 23



2000
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
HEIGHT, meters
INVERSION
OCT. 13, 1948
20km
θ_s
287
288
289
290
291
292
293
1
2
3
4
5
6
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

FIG. 24



HEIGHT, meters
2000
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
INVERSION
NOV. 16, 1948
20km
θ_s
278 279 280 281 282 283 284 285
1 2 3 4
POTENTIAL TEMPERATURE, °K
MIXING RATIO, gm/kg

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