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Stress, turbulence, and heat flow measurements over the Gulf of Maine and surrounding land.

by

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Stress, turbulence and heat flow measurements over the Gulf of Maine and surrounding land

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Abstract

This report presents turbulence, flux, temperature and water vapor data obtained from an airplane flying over the Gulf of Maine and adjacent shores. Measurements of the root-mean-square turbulent deviation velocities, shearing stresses, and heat flows were made at many heights and offshore distances in air masses moving from land to water. Stability effects on the turbulence and fluxes of heat and momentum have been observed over a wide range of conditions as air flowed over cooler or warmer surfaces.

The following conclusions have been drawn from a study of the data:

(1) The magnitudes of the root-mean-square turbulent deviation velocities, σ_w and σ_u, increases with height in the lowest 100 meters and then decreases slowly with height up to the inversion where the velocities drop to very low values.

(2) Shearing stresses were found to increase with height up to the 100 to 200 m level and then drop off with height. This height variation is in contrast to the generally accepted notion of a decrease of the stress from the surface to the geostrophic level. These observations confirm the findings of Scrase (1930) and others and demand an investigation of the acceleration of the air and the effects of thermal winds.

(3) Both the horizontal and vertical components of the turbulent wind are increased by hydrostatic instability and decreased by stability. The horizontal component is affected less than the vertical component by stability differences.

(4) The decrease in the turbulent velocities as air passes from land to cooler water is great and rapid while the increase in turbulent velocities as the air passes over warmer water is slight and slow.

(5) The observed heat fluxes also first increase and then decrease with height and usually become negative near the top of the mixed ground layer where the potential temperature gradient becomes strongly positive.

(6) The stability of the air above about 50 m is a very poor indicator of the temperature difference existing between the underlying surface and the air of the main mixed layer. Diffusion of heat downward from a layer of warm air above the ground layer frequently is the cause of a stable lapse rate regardless of the relative temperature of the surface below.

(7) No comparison of the coefficient of turbulent mass exchange for water vapor, heat flow and momentum could be made since the temperature gradient was stable even when heat was flowing upward, and no wind profiles were made over the water.

I Introduction

A. The relation of the present work to previous observational and theoretical studies

A vast number of observational studies of turbulence and diffusion in the atmosphere have been made in the past three quarters of a century. The Meteorological Abstracts and Bibliography (Kramer, et. al., 1953), lists 342 key papers on the subject, and many more have appeared since that date. A similarly large number of papers have been published (Rigby, et. al., 1953), on the theory of turbulence and turbulent diffusion. In spite of the great strides toward an understanding of the turbulent processes operating in the atmosphere which these papers represent, large gaps in our knowledge still remain. Most of the observational studies of atmospheric turbulence have been confined to the lower levels that can be reached by towers. Innumerable observations of wind and temperature gradients have been made to the higher regions of the free atmosphere, but only a few scattered observations of the turbulence have been made simultaneously at these levels. To fill this critical void of information, a system was developed to measure from an airplane in flight values of the horizontal and vertical components of the turbulent wind and variations of the air temperature. From a time series of these quantities the following means have been computed:

(1) root-mean-square horizontal turbulent component,

(2) r-m-s vertical turbulent component,

(3) shearing stress, from $-\bar{\rho}\overline{w'u'}$, and

(4) the heat flow, $c_p\bar{\rho}\overline{w'T'}$.

These values have been measured at many heights, from 15 m above the water to 2000 m, and many positions over land, a few km offshore and about 175 km offshore. Temperature and humidity measurements were made simultaneously so that stabilities, heat flows and water vapor flows could be measured. Winds were not measured except for the routine Weather Bureau observations; although winds would have been of great value had they been measured over the water.

The program of measurements was not developed around the idea of testing any one theory of turbulence of diffusion. It will be seen that data collected does rule out certain assumptions used frequently in turbulent theory and may act as a guide in suggesting new lines of approach and other assumptions on which to build a theory of turbulence. Spectra of the turbulent quantities have not been determined; hence much is left to be desired by the theoretician from the present study.

B. Meteorological quantities measured

The present study has been concentrated upon obtaining two types of data from flights with a PBY-6A amphibious aircraft. One type is the distribution of temperature and water vapor with height and distance over the waters of the Gulf of Maine. From this data the flux of moisture and heat can be obtained by computing the time rate of change of the water vapor and heat content of the air columns. The other type of observation is the recording of the rapid changes in the vertical acceleration, the airspeed, the attitude of the airplane and the temperature of the air. The instruments used for the measurement of these quantities has been described by Bunker and McCasland (1954). From these observations a time series of the rapid (1/5 second) variations of the vertical and horizontal components of the turbulent wind fluctuations and the temperature variations can be computed. The vertical velocities have been computed from the equation developed and described by Bunker (1955):

$$w' = 4.98 \frac{\Delta n}{\rho V} - \frac{8.9 \times 10^3 \Delta V_t}{\rho V^2} - \Delta \alpha_{att} V_t + \Sigma 1.1 \Delta n t_i + w_0$$

Where w' is the vertical turbulent component of the wind, Δn the vertical acceleration of the airplane, ρ the density of the air, V the velocity of the airplane, $\Delta\alpha_{att}$ the attitude of the airplane and w_o any original vertical velocity of the airplane. The constants appearing in the equation depend upon the aerodynamic characteristics of the PBY-6A. Deviations from 7 second averages were not taken as desirable in the above reference. The horizontal component fluctuations were found from the record of the airspeed assuming the ground velocity of the airplane to be constant or change linearly during the run. Since the completion of the present program, an additional horizontally mounted accelerometer has been attached to the airplane so that the assumption of a constant ground velocity is no longer necessary.

From these time series the meteorological quantities r-m-s vertical and horizontal deviation velocities, shearing stress and heat flows can be computed. Measurements of these quantities have been made at many heights from 150 m to 2000 m over the land and from 51 m to 2000 m over water and distances up to 175 km offshore.

C. Airplane instruments

The instruments installed on the PBY have been described in a report by Bunker and McCasland (1954); hence it will be stated here only that the outputs of a vertical accelerometer, an airspeed gauge, an attitude gyro, an altimeter and thermopile were recorded simultaneously on five galvanometers of a nine channel oscillograph. All of the fluctuation data was computed from these records.

Psychrometric data was recorded on a self-balancing potentiometer. The sensing heads were a calibrated thermistor and a lithium chloride strip mounted in the air-stream of the nose of the airplane.

D. Site of the experiment

The experiments to be described were carried out over the waters of the Gulf of Maine. This body of water is ideal for such work since it is close at hand, has a curved coastal line so that air from the SSW to N and reaching 42° 30' N, 68° 30' W has traveled over equal distances of water, and its temperature cycle and thermal structure are well known. In the summer and fall the water is usually cooler than the air from the land while in the winter it is usually much warmer, hence a wide variety of stability conditions occur.

Water temperatures were determined from mean temperature charts on file at the Institution adjusted for deviations of observed temperatures from the normal. The observed temperatures were taken from the Portland, Nantucket and Pollock Rip Lightships as well as at a few harbors along the coast. December, 1953, temperatures were found to be about 3°C above normal while January, 1954, was about 1.5°C above the normal. Water temperatures are expressed in this report as the potential temperature of an air parcel in contact with the water and having its temperature.

II Presentation of measurements

A. Presentation of observations taken on individual days

After the psychrograph and turbulence observations were reduced it became apparent that each day's character was different from all the others with respect to stability, cloud cover, wind speed, turbulence and transport. Hence the data pertaining to a given day are presented as a unit. A very brief description of the large scale synoptic weather pattern of the day is given at the beginning of each case presentation. The bulk of the data is given in a single graph to show the variation of the various quantities with height. The figures contain potential temperatures, heat fluxes, temperature deviations, mixing ratios, vertical and horizontal turbulent components of the wind, shearing stress, the wind profile, and the water temperatures. Notes concerning time, location, cloud height and amount, and precipitation have been added to the figure as needed.

Numerous quantities such as gradients, stabilities, fluxes, and mean winds have been measured from the graphs and presented in the text and in tables. To show both the height and offshore distance variation of σ_w, the observed values of σ_w have been plotted on cross sections. Lines of equal σ_w have been drawn for each 20 cm sec^{-1}. Stabilities of the air columns have been entered as well as differences of air and water temperatures.

Case I. October 20, 1953

Turbulent and convective heating of the air changed to cooling and greatly reduced turbulence as the air mass moved from land out over the cool waters of the Gulf of Maine. A ridge of high pressure west of the Gulf caused clear skies and a weak northwesterly flow out over the water. The land, heated by the solar radiation, was transferring its heat to the atmosphere and creating considerable convective turbulence. As the air flowed out over the water which was about 8°C cooler, the turbulence dropped to about 1/10 its previous level at 13 km offshore.

Three series of observations were taken on this day between 1200 and 1600 hrs. EST, the first over the land near Portland, Maine, the second 13 m offshore near the Portland Lightship, and the last 175 km offshore at about 42°N 68°W. Observations were made from 60 m to 1960 m above sea level. The sounding areas were chosen to lie along the air trajectory. The 2100 z pilot balloon runs show that the wind swung more northerly during the observation period so that the air trajectory cut across the observational line at an average of 30°.

Table Ia, has been compiled to show the average winds in the region at 1500 z and 2100 z. The values are averages of Portland, Maine, Boston, Massachusetts, and Nantucket, Massachusetts. From these data the average speed of the air was found to be 6.9 m sec^{-1}. Thus the air required

1900 seconds and 25000 seconds to travel to the 13 km and 175 km points respectively. These values are used in computations of heat and water vapor fluxes from the observed changes in the air columns. The temperatures and mixing ratios measured with the airplane psychrographs are plotted in Figure 1, for each of the sounding areas. From these plots of potential temperatures the stabilities have been computed for uniform portions of the three plots. The stabilities are tabulated as follows in Table Ib. The greatest stabilities observed over the Gulf are the result of the heat transfer from the air to the water which was about 14°C, 8° colder than the air. Mixing ratio gradients observed in the sounding areas are collected in Table Ic.

The heat flow computation from the net change in temperature of the column of air is greatly complicated by the change in wind speed and direction during the period. Using the winds presented in Table Ia, it is found that the air measured by the airplane in the center of the Gulf of Maine left the coast in the vicinity of Penobscot Bay around 0900 hrs, EST. Using the Old Town surface observations it is concluded that the air column started across the water about 3°C cooler than the air measured over Portland at 1200 hrs., the transit time to the center of the Gulf is estimated to be 3×10^4 seconds. With these values and the observed temperatures over the Gulf it is found that the heat flux was -0.2 m cal cm^{-2} sec^{-1}, with a gradient of 1×10^{-4} cm^{-1}, a coefficient of turbulent mass exchange of about 10 gm cm^{-1} sec^{-1} is found.

Table Id, presents the turbulent quantities and fluxes computed from the fluctuation data in the manner described in Section I. The following quantities are listed: (1) and identifying run number, (2) the height of the run, (3) σ_w the root-mean-square vertical velocity, (4) σ_u the r-m-s deviation horizontal velocity, (5) the shearing stress, (6) the r-m-s temperature deviations, and (7) the heat flux.

The data contained in Table Id have been plotted in Figure 1 together with the psychrograph, wind, and sea water temperature data. The various quantities for each of the sounding areas have been plotted against height as the ordinate and appropriate scale as the abscissa. From this graph many of the changes in the air mass produced by the flowing of the air over cooler water are clearly depicted. Most noteworthy are the cooling of the lowest 400 m of the air, the change of a large upward flow of heat to exceedingly small and negative flows, a decrease by one order of magnitude in the r-m-s vertical velocities, and a similarly large decrease in the shearing stress. These phenomena will be discussed in the following section in conjunction with the data from other days.

A cross section of the air from the Maine shore to the 175 km offshore sounding area is presented as Figure 2 to show the variation of the root-mean-square vertical velocities with height and distance. Most significant here is the rapidity and magnitude of the decrease in the turbulent velocities with offshore distance. Only a slight further decrease occurs between the 13 km area and the 175 km area.

Table Ia

Average Radio and Pilot Balloon Winds for Oct. 20, 1953

Height meters	1500 z Winds Direction °T	1500 z Winds Velocity m sec^{-1}	2100 z Winds Direction °T	2100 z Winds Velocity m sec^{-1}
sfc	310	6.6	010	5.3
300	320	6.6	020	6.1
600	330	7.7	360	6.7
900	330	8.3	350	7.0
1200	320	7.7	340	6.7
1500	290	7.2	330	6.5
1800	280	9.2	320	7.6
2100	290	10.2	310	8.3
2400	300	10.8	310	8.3

Table Ib

Stabilities of Air Regions

Portland, Maine Height Range m	Stability $cm^{-1} \times 10^8$	Portland Lightship Height Range m	Stability $cm^{-1} \times 10^8$	42° N 68° W Height Range m	Stability $cm^{-1} \times 10^8$
250 to 500	19	90 to 1000	5	70 to 400	36
500 to 1300	0			500 to 900	9
				900 to 1500	5

Table Ic

Mixing Ratio Gradients

Portland, Maine Height Range m	Gradient 10^8 cm^{-1}	Portland Lightship Height Range m	Gradient 10^8 cm^{-1}	42° N 68° W Height Range m	Gradient 10^8 cm^{-1}
250 to 1250	-1.4	90 to 940	-1.4	70 to 400	0.0
1275 to 1300	-24			400 to 470	-12
1300 to 1950	-1.5			470 to 1500	-0.8

Table Id

Turbulence and Flux Data

a. Portland, Maine, 1200 to 1240 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\,\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
462	225	109	86	7.3	0.10	1.5
463	300	128	60	0.9	0.18	5.5
464	600	66	49	0.3	0.06	-0.06
465	900	130	42	0.7	0.10	1.4
467	1200	43	40	-1.6	0.09	-0.02
468	1500	18	18	-0.2	0.07	0.1

b. Portland Lightship, 1410 to 1440 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\,\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
473	60	12	14	-0.00	0.10	-0.05
474	150	15	14	0.02	0.04	0.03
475	300	16	20	0.08	0.05	0.09
476	600	13	18	-0.15	0.04	0.08
478	900	15	16	0.02	0.03	0.06

c. 42° N 68° W, 1540 to 1615 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\,\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
479	60	13	16	0.07	0.03	0.06
481	150	12	8	0.03	0.03	-0.02
482	300	11	13	0.05	-	-
483	600	7	10	-0.03	-	-
484	900	13	23	0.11	0.03	0.00

Case II October 27, 1953

The surface flow on this day was controlled by a small high pressure cell located over the North Atlantic Ocean between Cape Cod and Bermuda. Thus air from southern New England passed northeastward over the cooler waters of the Gulf of Maine, losing its heat with a resulting condensation into a stratus undercast about 60 km offshore. The air was stable over the land since it had travelled long distances from the south and the remnants of a nocturnal inversion were still present. Fluxes cannot be calculated from the temperature changes in the column because of the development of the stratus undercast which was not penetrated by the airplane.

The soundings were made along a line that described the trajectory of the air within experimental verification. The first sounding was made over land north of Cape Cod, the second 29 km downwind over the water off Provincetown, Mass. and the third 200 km from the the first sounding at 43°N 69°W. Table IIa shows the Boston and Portland wind from 1500 z.

The average of these winds is 6.3 m sec^{-1}. The transit times to the second and third sounding areas are 4.6×10^3 sec. and 3.2×10^4 sec. The air measured over the Gulf thus left the land before sunrise and was already cool at the lowest layers from outward radiation.

The stabilities of the air mass have been measured from Figure 3 and entered in Table IIb. Gradients of the mixing ratio have been measured and entered in Table IIc. Table IId presents the fluctuation data obtained on this day.

Table IIa

Pilot Balloon and Rawin Winds at Boston and Portland, Oct. 27, 1953, 1500 z

Height meters	Boston Winds Direction °T	Boston Winds Velocity m sec^{-1}	Portland Winds Direction °T	Portland Winds Velocity m sec^{-1}
sfc	-	-	210	3.1
300	230	6.8	190	2.6
600	240	8.1	200	4.1
900	-	-	220	5.2
1200	-	-	230	7.2
1500	-	-	240	11.4

Table IIB

Stabilities of Air Regions

North of C. C. Canal Height Range m	Stability $cm^{-1} \times 10^8$	North of Provincetown Height Range m	Stability $cm^{-1} \times 10^8$	43°N 69°W Height Range m	Stability $cm^{-1} \times 10^8$
150 to 500	29	70 to 500	6	330 to 500	10
500 to 1000	3	500 to 1000	9	500 to 1000	8
1000 to 1500	21	1000 to 1500	21	1000 to 1500	14

Table IIc

Mixing Ratio Gradients

North of C. C. Canal Height Range m	Gradient $cm^{-1} \times 10^8$	North of Provincetown Height Range m	Gradient $cm^{-1} \times 10^8$	43°N 69°W Height Range m	Gradient $cm^{-1} \times 10^8$
150 to 1100	-3.9	70 to 500	-2.6	330 to 1100	-2.1
1100 to 1850	-0.9	500 to 1000	-5.5	1100 to 1260	-8.0
		1000 to 1850	-1.7	1260 to 1570	-2.6

Table IId

Turbulence and Flux data

a. North of Cape Cod Canal, 1040 to 1125 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal $cm^{-2}sec^{-1}$
486	180	27	27	-0.28	0.09	0.16
488	280	47	46	0.88	0.10	-0.52
489	460	9	10	-0.04	-	-
490	760	14	8	0.02	0.02	-0.17
491	1200	13	20	0.01	0.05	0.21

Table IId (continued)

Turbulence and Flux data

b. North of Provincetown, 1140 to 1225 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\rho \overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p \rho \overline{w'T'}$ mcal $cm^{-2} sec^{-1}$
498	50	10	25	-0.02	0.05	0.28
497	150	14	25	-0.18	0.03	-0.04
496	290	9	6	0.01	0.07	-0.00
495	580	12	16	0.08	0.05	-0.01
494	910	9	8	-0.00	0.05	-0.13
493	1530	10	21	-0.05	0.05	-0.03
492	1850	19	20	0.24	-	-
504	310	23	23	-0.22	0.11	-0.07
503	620	11	11	-0.03	-	-
502	910	6	12	-0.05	-	-
501	1200	14	10	-0.01	0.04	0.01
500	1560	13	17	-0.11	0.03	0.08

Figure 3 has been drawn presenting the psychrograph, wind and fluctuation data given in Table IId. It is seen that this situation is very similar to that of Case I, that is, the r-m-s velocities are very small as is also the stress and heat flux. Figure 4, the cross-section of the atmosphere shows the small values of the r-m-s vertical velocity and its small variation from land to sea.

Case III. December 7, 1953

The series of observations taken on this day are most remarkable for the large values of the r-m-s vertical velocities and shearing stresses observed over the land. A strong cold front passed the area around 1000 hrs EST, pushed by 20 to 30 m/sec winds. Equally remarkable is the manner in which the turbulence died off as the air blew out over the water which was almost exactly the same temperature as the air. The stress dropped from 21 dynes cm^{-2} to less than 1 dyne cm^{-2} during the 1/2 hours journey over the water. Two soundings were made on this day, one over land near Beverly, Massachusetts and the other over the water several kilometers north of Provincetown, Massachusetts. The two sounding areas lie along the air trajectory and are about 40 km apart. Table IIIa gives the Nantucket, Boston, and Portland 2100 z pibal observations. The Boston wind was used exclusively for determining transit times since the airplane observations were made so close to Boston. The air took 1600 sec to traverse the 40 km.

It will be noted that the instability over the land was greater than any other air mass studied during the series. Although no soil temperatures were obtained, it is obvious that the sun-heated land was warmer than the air. The ground layer must have been warmed to a considerable depth as the cold front was preceeded by a very warm period. The air retained a large part of its instability over the water even though the air and water were about equal in temperature. No fluxes of heat or water vapor could be determined on this day since the inversion at 1500 m was broken and the lower air mixed with the upper air.

Table IIIc presents the mixing ration gradients found in the two sounding areas, while Table IIId presents the fluctuation data.

Table IIIa

Pilot Balloon Observations, 2100 z

Height	Nantucket, Mass.		Boston, Mass.		Portland Maine	
m	°T	m sec^{-1}	°T	m sec^{-1}	°T	m sec^{-1}
sfc	280	8.3	-	-	270	8.9
300	280	15	280	25	280	17
600	290	16	280	25	290	21
900	290	19	280	24	290	22
1200	300	28	290	25	290	22
1500	300	32	300	27	290	22
1800	300	33	300	31	300	24

The stabilities of the various layers of the atmosphere over Beverly and the water are contained in Table IIIb.

Table IIIb

Stabilities of Air Regions

Beverly, Mass. Height Range m	Stability $cm^{-1} x 10^8$	North of Provincetown Height Range m	Stability $cm^{-1} x 10^8$
150 to 500	-10	300 to 500	-7
500 to 1300	2	500 to 1000	4
1300 to 1600	90	1000 to 1500	60

Table IIIc

Mixing Ratio Gradients

Beverly, Mass. Height Range m	Gradient $cm^{-1} x 10^8$	North of Provincetown Height Range m	Gradient $cm^{-1} x 10^8$
150 to 1430	-0.4	300 to 1100	-0.9
1430 to 1785	-0.4	1100 to 1600	-0.4

Table IIId

Turbulence and Fluctuation Data

a. Over Beverly, Massachusetts, 1415 to 1455 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\rho \overline{u'w'}$ dyne cm^{-2}	σ_T °C	$c_p \rho \overline{w'T'}$ mcal cm^{-2} sec^{-1}
522	150	127	130	8.7	0.05	0.03
523	300	197	173	21.3	0.06	-0.96
524a	600	90	92	3.9	0.06	-0.92
524b	900	33	47	-0.5	0.05	-0.08
530	1500	42	37	0.9	0.50	-2.95
531	1800	32	41	-0.1	0.19	-0.54

b. Downwind Flight from Plymouth, Massachusetts to North of Provincetown Massachusetts, 1525 to 1535 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\rho \overline{u'w'}$ dyne cm^{-2}	σ_T °C	$c_p \rho \overline{w'T'}$ mcal cm^{-2} sec^{-1}
534a	900	31	30	-0.30	0.07	-0.40
b	900	42	38	0.34	0.08	-0.38
c	900	74	90	-1.73	0.09	-0.37
d	900	57	38	0.49	0.06	0.49
e	900	69	39	-0.02	0.06	0.18
f	900	69	54	-1.47	0.06	-0.58
g	900	41	58	-0.07	0.06	0.00
h	900	32	39	-0.04	0.04	-0.01
i	900	28	69	0.06	0.07	-0.01

c. North of Provincetown, Massachusetts, 1540 to 1610 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\rho \overline{u'w'}$ dyne cm^{-2}	σ_T °C	$c_p \rho \overline{w'T'}$ mcal cm^{-2} sec^{-1}
535	300	43	54	-0.79	0.05	0.16
536	600	42	55	-0.35	0.04	0.06
537	900	24	29	0.03	0.10	-0.15
538	1200	30	38	-0.69	0.30	-0.52
539	1680	24	16	0.41	0.1	0.07

This data together with the psychrograph data has been plotted on a height diagram, Figure 5, and a root-mean-square vertical velocity cross section, Figure 6. The magnitude of the turbulence and the changes in turbulence on this day are brought out in the diagrams which have the same scales as the other diagrams in this paper.

Case IV. December 17, 1953

The air on this day was the coldest of the series and consequently had the greatest temperature difference, 18°C, with the sea surface. The flow of air was moderate and steady from the northwest under the influence of a high over the midwest of the United States and a low over the Atlantic Ocean northeast of Newfoundland. Over the land 8/10 of the sky was covered with cumulus clouds, bases 1050 m, tops 1800 m. As the air moved out over the water the cumulus merged into 10/10 sky coverage and snow showers were observed frequently in the center of the Gulf of Maine. The turbulence over the water on this day was the greatest observed during the series. Aircraft soundings were made over the land near Portland, Maine 11 km offshore and 150 km offshore in a direction exactly downwind. The 2100 z pilot balloon wind observations are presented in Table IVa. Using the Portland pibals we find an average wind of 7.9 m sec^{-1} from the surface to 1200 m, giving a transit time of 1400 sec to the second sounding area and 19000 sec to the third area. Stabilities of the air are presented in Table IVb.

Heat and water vapor fluxes have been computed but they are at best only crude approximations since much heat was liberated and water vapor lost through the process of snow showers. The value of the water vapor obtained between the land and the first offshore sounding is 29 x 10^{-6} gm cm^{-2} sec^{-1}. This value is reliable since no showers were observed in this area. One value of the coefficient of turbulent mass transport was computed to be 1200 gm cm^{-1} sec^{-1} although the process was primarily one of convection. The heat flux measured between the Lightship and 60° 30'W was 7.3 mcal cm^{-2} sec^{-1}. The observed mixing ratio gradients are presented in Table IVc. Table IVd presents the fluctuation data obtained

Table IVa

Pilot Balloon Winds 2100 z

Height m	Nantucket Direction °T	Nantucket Velocity m sec^{-1}	Boston Direction °T	Boston Velocity m sec^{-1}	Portland Direction °T	Portland Velocity m sec^{-1}
sfc	310	4.7	-	-	330	5.7
300	310	5.7	320	8.3	340	7.8
600	300	6.2	320	7.8	320	9.4
900	300	6.2	310	9.4	310	10.4
1200	-	-	290	9.9	290	6.2
1500	-	-	280	9.9	-	-
1800	-	-	290	8.3	-	-

Table IVb

Stabilities of Air Regions

Portland, Maine Height Range m	Portland, Maine Stability $cm^{-1} x 10^8$	Portland Lightship Height Range m	Portland Lightship Stability $cm^{-1} x 10^8$	42° 30'N 68° 30'W Height Range m	42° 30'N 68° 30'W Stability $cm^{-1} x 10^8$
150 to 1500	0.8	50 to 1200	0.4	150 to 350	-19
1500 to 2000	37			350 to 850	6

Table IVc

Mixing Ratio Gradients

Portland, Maine Height Range m	Portland, Maine Gradient $cm^{-1} x 10^8$	Portland Lightship Height Range m	Portland Lightship Gradient $cm^{-1} x 10^8$	42° 30'N 68° 30'N Height Range m	42° 30'N 68° 30'N Gradient $cm^{-1} x 10^8$
150 to 1700	-0.2	50 to 1200	-0.1	150 to 850	0.0
1700 to 1850	-2.0				
1850 to 1950	0.0				

Table IVd

Turbulence and Fluctuation Data

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{u'w'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2}sec^{-1}
a. Over Portland, Maine, 1340 to 1440 EST						
542	150	133	60	2.1	0.1	1.60
543	220	118	89	-2.6	0.11	2.2
545	530	72	43	1.4	0.03	-0.3
546	830	40	35	0.6	0.03	-0.08
547	1000	62	45	1.8	0.08	-0.01
548	1880	39	20	0.01	0.03	0.04
b. Over Portland Lightship, 1450 to 1525 EST						
554	50	91	111	-1.85	0.13	1.90
553	140	112	81	2.67	0.08	0.63
552	420	58	44	0.48	0.06	0.15
551	610	57	58	0.13	0.03	0.12
550	970	94	57	2.39	0.07	0.47
549	1215	74	55	2.36	0.06	-0.43
c. 42°30'N 68°30'W, 1600 to 1620 EST						
557	140	93	90	1.11	0.13	1.95
558	280	96	44	1.87	0.07	1.48
559	560	131	78	1.61	0.11	1.95

The above data is presented in graphic form as Figures 7 and 8

Case V. December 23, 1953

A cool air mass was blowing offshore from the northwest over the Gulf water which was about 6°C warmer than the air. No low clouds were present over the land with a few tenths sky coverage of flat top cumulus north of Provincetown and 5 to 8/10 coverage of flat top cumulus at 42°30'N, 68°30'W. The flatness of the cloud tops was caused by the stability of an intense inversion at 1200 m. Overland the potential temperature increased 6.5°C in 50 m.

The sounding areas were chosen in an east-west line starting at Beverly, Massachusetts. The second two soundings were made 56 km and 190 km offshore. The winds, which are presented in Table Va., actually came from a more northerly direction and showed a marked divergence. The average wind for the lowest 1200 m was found to be 11.6 m sec^{-1}, giving transit times of 4.8×10^3 sec and 11.6×10^3. The stabilities of the air layers are presented in Table Vb. The great stability of 480×10^{-8} cm^{-1} is noteworthy and indicative of strong subsidence.

Flows of heat were computed and found to be 4.7 mcal cm^{-2} sec^{-1} between land and 70°15'W and 6.9 mcal cm^{-2} sec^{-1}, between 70°15'W and 68°30'W through the surface layer. Water vapor content of the air was not measured on this day due to failure of the humidity strip circuit. The data obtained on this day has been plotted and presented as Figures 9 and 10.

Table Va

Pilot Balloon and Radio Winds for 1500 z

Height	Nantucket		Boston		Portland	
m	Direction °T	Velocity m sec^{-1}	Direction °T	Velocity m sec^{-1}	Direction °T	Velocity m sec^{-1}
sfc	320	8.3	-	-	250	7.8
300	300	14.1	280	12.0	270	10.4
600	290	14.1	290	12.0	290	10.9
900	270	14.1	300	13.0	290	9.4
1200	260	15.6	290	17.3	300	8.3
1500	260	21.4	270	22.4	290	9.4
1800	260	25.5	250	26.0	260	11.0

Table Vb

Stabilities of the Air Regions

Beverly, Mass.		42°30'N	70°15'W	42°30'N	68°30'W
Height Range m	Stabilities $cm^{-1} \times 10^8$	Height Range m	Stabilities $cm^{-1} \times 10^8$	Height Range m	Stabilities $cm^{-1} \times 10^8$
150 to 700	-1	40 to 400	-1	50 to 700	0
700 to 1200	6	400 to 1150	4	700 to 1600	45
1200 to 1250	480	1150 to 1200	480		-
1250 to 1477	45	1200 to 1425	45		-

Table Vc

Turbulence and Fluctuation Data

a. Over Beverly, Massachusetts, 1110 to 1145 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
563	150	123	111	2.49	0.07	0.62
564	285	51	52	0.42	0.04	0.00
565	570	45	37	0.26	0.03	-0.04
566	860	75	44	0.52	0.02	-0.47
567	1150	37	48	0.45	0.05	-0.06
568	1420	26	13	0.06	0.21	0.77

b. Sounding at 42°30'N, 70°15'W, 1155 to 1230 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
574	40	57	53	0.71	0.07	0.51
573	150	66	35	-0.14	0.09	0.91
572	280	100	64	1.47	0.05	0.01
571	570	108	51	1.34	0.04	0.05
570	1100	36	46	-0.43	0.41	-0.28
569	1420	14	13	0.001	0.08	0.01

c. Sounding at 42°30'N, 68°30'W, 1300 to 1325 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
580	50	54	46	-0.06	0.07	0.07
579	150	68	59	0.81	0.05	0.25
578	290	82	56	-0.36	0.06	0.47
577	580	87	39	2.55	0.06	0.09
576	840	39	40	0.43	0.10	-0.44
575	1620	12	15	-0.02	0.03	0.01

Case VI. January 5, 1954

This series of observations was made in a col between two coastal storms, one northeast of Newfoundland, the other over Virginia. Winds near the surface were light and generally from the northwest, while winds aloft were from the southwest bringing in a 10/10 cirrus overcast. The waters of Massachusetts Bay were about 6°C warmer than the air. Airplane observations were made along a 46 km line 150 m above the water from north of Provincetown, Massachusetts to Beverly, Massachusetts. There an ascent was made to 920 m, and then another sounding was made near the original point off Provincetown. The 1500 GCT pilot balloon and radio winds are presented in Table VIa. The average speed of the wind for the lower layers is 3.0 m sec^{-1} and transit time 1.5 x 10^{4} sec.

The stabilities of the air layers are given in Table VIb, while the mixing ratio gradients are presented in Table VIc. The heat flux was found to be 1.6 mcal cm^{-2} sec^{-1}, while the water vapor flux was found to be 2.5×10^{-6} gm cm^{-2} sec^{-1}. An austansch value of 800 gm cm^{-1} sec^{-1} was found. Turbulence and fluctuation data is contained in Table VId.

Table VIa

Pilot Balloon and Radio Winds, 1500 GCT

Height m	Nantucket Direction °T	Nantucket Velocity m sec^{-1}	Boston Direction °T	Boston Velocity m sec^{-1}	Portland Direction °T	Portland Velocity m sec^{-1}
sfc	300	2.3	-	-	320	1.8
300	260	5.5	280	3.2	330	2.3
600	250	8.2	260	5.9	280	1.4
900	250	10.0	270	8.7	250	2.7
1200	240	9.5	260	10.0	270	5.0
1500	240	9.5	250	10.9	270	7.2
1800	260	10.0	250	12.3	270	7.7

Table VIb

Stabilities of Air Regions

Beverly, Mass. Height Range m	Beverly, Mass. Stabilities $cm^{-1} \times 10^{8}$	North of Provincetown Height Range m	North of Provincetown Stabilities $cm^{-1} \times 10^{8}$
150 to 300	1	15 to 400	5
300 to 900	79	400 to 700	120
		700 to 900	37

Table VIc

Mixing Ratio Gradients

Beverly, Mass. Height Range m	Beverly, Mass. Gradient $cm^{-1} \times 10^{8}$	North of Provincetown Height Range m	North of Provincetown Gradient $cm^{-1} \times 10^{8}$
300 to 900	0.3	15 to 300	-0.3
		300 to 400	-10
		400 to 600	1.5
		600 to 920	0

Table VId

Turbulence and Fluctuation Data

a. Provincetown, Massachusetts to Beverly, Massachusetts Flight, 1105 to 1120 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2}sec^{-1}
583	150	53	44	-0.01	0.06	0.37
584	150	46	32	0.11	0.05	0.07
585	150	49	58	0.98	0.06	0.13
586	150	58	36	0.13	0.05	0.22
587	150	65	43	-0.05	0.05	0.60
588	150	62	44	1.37	0.06	0.28
589	180	52	36	0.33	0.04	0.19
590	230	50	45	0.08	0.06	0.20
b. Over Beverly, Massachusetts, 1120 to 1140 EST						
591	290	25	26	0.07	0.05	-0.01
592	595	16	11	-0.02	0.04	0.11
593	915	9	16	-0.003	0.03	-0.01
c. Over water north of Provincetown, 1205 to 1230 EST						
600	15	44	39	-0.18	0.09	0.68
599	45	50	39	0.33	0.07	0.51
598	95	39	26	0.42	0.05	0.15
597	305	33	33	0.18	0.08	-0.25
596	440	15	12	-0.06	0.04	0.01
595	610	16	21	0.10	0.10	-0.03
594	915	12	24	0.11	0.11	0.14

Case VII. January 7, 1954

A moderate north westerly wind was blowing offshore under the influence of a high over Louisiana and a low off Newfoundland. The water temperature was about 8°C warmer than the air as it left the land. As a result, convection started soon after the air left the coast and cumulus clouds increased until a complete overcast was present over the final sounding area. These soundings were made in a down-wind line, one over the land, one 18 km offshore, and the last, 190 km offshore. The observed 1500 z winds are presented in Table VIIa. Transit times for the air were 2.4×10^3 seconds and 2.4×10^4 seconds. The heat flux into the column was found to be 5.5 mcal cm^{-2} sec^{-1} between land and 18 km offshore and 3.7 mcal cm^{-2} sec^{-1} between 18 km and 190 km.

Table VIIa

Pilot Balloon and Radio Winds, 1500 z

Height m	Nantucket Direction °T	Nantucket Velocity m sec^{-1}	Boston Direction °T	Boston Velocity m sec^{-1}	Portland Direction °T	Portland Velocity m sec^{-1}
sfc	290	7.8	-	-	290	3.6
300	280	11.0	290	10.4	300	4.2
600	280	9.9	300	11.4	320	3.6
700	280	9.9	310	10.4	350	2.1
1200	280	12.0	300	9.4	330	2.1
1500	280	14.1	290	11.0	280	4.7
1800	280	13.5	280	12.0	270	8.3

Table VIIb

Stabilities of Air Regions

Newburyport, Massachusetts, 18 km offshore, 42° 30'N 68° 30'W

Height Run m	Stability cm^{-1}x 10^{8}	Height Run m	Stability cm^{-1}x 10^{8}	Height Run m	Stability cm^{-1}x 10^{8}
150 to 900	-0.2	50 to 400	-1	50 to 600	0.0
900 to 1400	52	400 to 1000	+6	600 to 750	10
		1000 to 1400	+65		

Table VIIc

Turbulence and Fluctuation Data

a. Over Newburyport, Massachusetts, 1100 to 1130 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
602	135	97	71	0.87	0.10	0.71
604	300	143	102	3.37	0.08	1.11
605	550	86	60	1.40	0.07	-0.26
606	820	61	49	-0.96	0.05	0.12
607	1130	37	28	0.12	0.32	1.44
608	1410	24	26	-0.02	0.08	-0.07

b. 18 km offshore, 1135 to 1200 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\bar{\rho}\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p\bar{\rho}\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
613	55	68	46	-0.11	0.11	1.28
612	140	95	49	0.78	0.06	0.39
611	260	76	58	-0.76	0.06	0.34
610	570	42	31	-0.59	0.04	0.05
609	890	67	56	-0.23	0.06	-0.08

Table VIIc (continued)

c. 42°30'N 68°30'W, 1300 to 1320 EST

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\rho\,\overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p \rho\,\overline{w'T'}$ mcal cm^{-2} sec^{-1}
616	55	69	68	1.84	0.05	0.46
617	140	79	58	1.15	0.10	1.63
618	270	58	35	0.41	0.05	0.19
619	550	98	55	1.44	0.06	0.99
620	740	46	32	-0.22	0.05	-0.40

Case VIII. January 29, 1954

On this day there was a strong northwest flow of cold air out over the Gulf as a result of an intense low pressure area northeast of Newfoundland. The air was subsiding and consequently clear; except for, at most, 2/10 sky coverage by cumulus clouds over the water. In spite of the small amount of cumulus clouds present, intense convection was going on as a result of the 13°C difference in water and air temperature.

One sounding was made over Portland, Maine. A down wind flight was made to 42°30'N, 68°30'W at 350 m above the sea. Six turbulence runs were made at five minute intervals during the flight. A regular sounding with turbulence runs were made at the terminus of the flight. Runs were obtained at 0, 31, 52, 73, 105, 145, and 175 km offshore.

An average wind speed of 10.2 m sec^{-1} is determined from the 1500 z winds which are reproduced here. Transit time of the air column to the final sounding area is 1.7 x 10^{4} seconds. The stabilities of the air column are presented in Table VIIIb. The observed mixing ratio gradient are tabulated in Table VIIIc. The heat and water vapor flows computed from the change in heat and water vapor content of the air column were found to be equal to 2.8 mcal cm^{-2} sec^{-1} and 2.3 x 10^{-6} gm cm^{-2} sec^{-1}, giving a coefficient of mass exchange of 770 gm cm^{-1} sec^{-1}.

Table VIIIa

Pilot Balloon and Radio Winds, 1500 z

Height	Nantucket		Boston		Portland	
m	Direction °T	Velocity m sec^{-1}	Direction °T	Velocity m sec^{-1}	Direction °T	Velocity m sec^{-1}
sfc	310	4.2	-	-	250	3.1
300	300	6.3	280	5.6	290	3.6
600	300	7.2	300	5.1	310	5.1
900	300	9.4	320	7.8	340	8.9
1200	310	10.3	320	10.3	350	12.5
1500	320	10.3	310	11.0	-	-
1800	320	13.9	310	11.0	330	15.6

Table VIIIb

Stabilities of Air Regions

Portland, Maine Height Range m	Stabilities $cm^{-1} \times 10^8$	42° 30'N Height Range m	68° 30'W Stabilities $cm^{-1} \times 10^8$
230 to 650	11	90 to 740	3
650 to 1220	55	740 to 1350	40

Table VIIIc

Mixing Ratio Gradients

Portland, Maine Height Range m	Gradients $cm^{-1} \times 10^8$	42° 30'N Height Range m	68° 30'W Gradients $cm^{-1} \times 10^8$
230 to 450	-0.4	75 to 730	-0.3
450 to 600	-7	730 to 1400	0
600 to 1200	+0.2		

Table VIIId

Turbulence and Fluctuation Data

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\rho \overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p \rho \overline{w'T'}$ mcal cm^{-2} sec^{-1}
a. Over Portland, Maine 1330 to 1400 EST						
635	235	86	29	0.76	0.10	0.82
636	370	42	49	1.07	0.07	-0.35
637	655	23	34	0.33	0.17	-0.76
638	925	14	12	-0.13	0.08	-0.11
639	1220	22	29	0.60	0.13	-0.57

Table VIIId (continued)

Run No.	Height m	σ_w cm sec^{-1}	σ_u cm sec^{-1}	$-\rho \overline{w'u'}$ dyne cm^{-2}	σ_T °C	$c_p \rho \overline{w'T'}$ mcal cm^{-2} sec^{-1}
b. Downwind flight from Portland, Maine to 42°30'N, 68°30'W, 1405 to 1440 EST.						
640	320	51	49	0.31	0.07	-0.35
641	315	122	50	1.23	0.09	0.98
642	340	60	42	0.12	0.07	-0.18
643	370	75	54	0.14	0.07	0.76
644	340	104	63	-1.73	0.07	0.67
645	360	79	37	1.24	0.04	0.47
c. 42°30'N, 68°30'W, 1445 to 1530 EST.						
646	85	92	78	3.88	0.14	1.94
647	185	108	59	-0.56	0.21	3.50
648	340	121	55	0.75	0.08	2.21
649	630	44	40	1.10	0.13	0.89
650	875	18	15	0.06	0.10	-0.29
651	1140	16	18	-0.24	0.05	0.06
652	1380	13	10	0.54	*	*

B. Mean σ_w maps.

Two maps have been prepared to show the geographical variation of the vertical component of the turbulence. The first map, Figure 17, shows the variation of σ_w at 300 m as it moves over water that is cooler than the air. Values of the turbulence over the land is usually high due to convection over the heated land. The turbulence drops off rapidly as the stability of the air is increased by the cool water. The air on December 7, 1953 almost was in equilibrium with the water.

The second map, Figure 18, was drawn up to show the range of values found when the water is warmer than the air. It is seen that the turbulence increased, or maintained its land values as the air was heated by the water and convection was established in the air.

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* temperature variations too small to measure

III. Average values and empirical relations

A. Average variation of σ_w and σ_u with height.

The average values of the root-mean-square vertical and horizontal deviation components of the turbulent wind has been found for several height ranges for both stable and unstable case over water and over land. The averages are presented in Table 1.

These averages show a general pattern with small values of the root-mean-square deviation velocities close to the ground or water increasing to a broad maximum in the 50 to 500 m range and a slow decrease as the temperature inversion or stable layer capping the ground layer is approached. The greater values of σ_w and σ_u over the land in comparison with the over water unstable case is interpreted to be the result of the greater roughness of the land; since wind and stability conditions were about the same in each case. In the case of stable air over the water the turbulent velocities are exceedingly small as a result of the great amount of turbulent energy required to displace the stable air and the relative smoothness of the water surface.

B. The ratio of σ_w to σ_u as a function of height and temperature contrast.

The present set of data lends itself to the study of the ratio of the vertical r-m-s velocity to the horizontal r-m-s deviation velocity. It is desired to see if this ratio varies with height above a surface or with temperature contrast between the surface and the air.

First, it is necessary to check the data to see if there are any systematic differences between σ_w and σ_u. As no tower comparisons of σ_u were made and no flight test can reveal any such difference, the only check possible is the determination of the over all value of the ratio of

Table 1

Averages of σ_w and σ_u Over Land and Water

Height Range m	Land $\sigma_{w'}$ cm sec^{-1}	Land $\sigma_{u'}$ cm sec^{-1}	Water Unstable $\sigma_{w'}$ cm sec^{-1}	Water Unstable $\sigma_{u'}$ cm sec^{-1}	Water Stable $\sigma_{w'}$ cm sec^{-1}	Water Stable $\sigma_{u'}$ cm sec^{-1}
0 to 50	-	-	47	39	-	-
50 to 100	-	-	72	64	13	16
100 to 200	103	81	70	58	12	8
200 to 500	93	69	70	46	17	18
500 to 1000	55	41	54	40	9	14
1000 to 2000	30	29	19	19	14	14
2000 up	39	20	-	-	-	-

Table 2

Shearing Stress Averages

Height Range m	Average Wind Land Stations m sec^{-1}	Land dy cm^{-2}	Water Stable dy cm^{-2}	Water Unstable dy cm^{-2}
0 to 50	6.5	-	-	0.26
50 to 100	-	-	0.07	0.95
100 to 200	-	3.70	0.03	1.47
200 to 500	7.9	3.73	0.14	0.62
500 to 1000	9.3	0.83	0.08	0.79
1000 to 2000	11.6	0.54	0.06	0.33
2000 up	-	0.01	-	-

Table 3

Heat Flux Averages Computed from $c_p \bar{\rho} \overline{w'T'}$

Height Range m	Land mcal cm^{-2}sec^{-1}	Water Stable mcal cm^{-2}sec^{-1}	Water Unstable mcal cm^{-2}sec^{-1}
0 to 50	-	0.28	0.68
50 to 100	0.51	-0.05	0.57
100 to 200	0.72	-0.01	1.05
200 to 500	0.66	0.01	1.01
500 to 1000	-0.07	0.02	0.28
1000 to 2000	-0.18	0.01	-0.08

$\sigma_w/\bar{u}$. The average over all heights for the unstable case is 1.25 while the average of all stable cases is 0.93. The average for both cases, giving equal weight to the two ratios is 1.09. As this ratio is nearly 1 and the ratios to be discussed in the next paragraph are greater than 2, it is concluded that systematic errors may be considered negligible.

The data has been studied by plotting ratios of σ_w/σ_u on a height-temperature difference diagram, Figure 19. When the ratios were plotted and contours drawn, a pattern emerged which shows the effect of the sea-air temperature contrast. The main feature of the pattern is a maximum of the ratio located roughly between 300 and 600 m above the surface and at greater than a 5°C excess of the water temperature above the air temperature. The maximum is the net result of the forces of buoyancy, viscosity, and turbulence and any theory attempting to describe the processes operating in an atmosphere with these stability conditions must be able to predict the observed maximum.

In other regions of the graph there is a wide scatter in the values of the ratio, but the mean value is roughly 1. Much of the scatter if produced by the exceedingly small r-m-s velocity values so that a difference of only 1 cm sec^{-1} may make a difference of 0.1 in the ratio. Above 1000 m much of the scatter is caused by varying heights of the ground layer inversion.

A 0.5 contour has been drawn in at the bottom of the graph, although there is only a weak suggestion that it may exist. Data taken in other studies such as tunnel and tower studies and one value of 0.5, obtained by the PBY over the Atlantic Ocean in an unstable atmosphere at 15 m, indicate such a decrease exists. Damping of the vertical component by the boundary surface is the cause of the decrease of the ratio.

C. Variation of the stress with height over Land and water.

Averages of the shearing stresses have been found and tabulated in Table 2.

A glance at this table will show that the stresses of the three groups differ widely from one another. The overland values are several times as large as the unstable over-water group. Since the degree of instability and the wind speed of these two groups are about the same, it is concluded that the roughness of the land has a large effect upon the shearing stress. The relation between the averages of the stable and unstable air masses is of particular interest because of the exceedingly small values of the stress in the stable case. The wind speed of the stable case is about 1/2 of the speed in the unstable case, yet the stress is much less than the 1/4 expected with such relative wind speeds. This small relative stress is attributed to the turbulence-damping ability of the stable air.

The variation of the stress with height is of great importance since it does not agree with the distribution observed in channels and generally accepted as describing the meteorological case. The accepted picture is that of a stress of zero at the height of the gradient wind increasing regularly to a maximum value at the boundary surface. The height distribution over the land may agree with this picture although no measurements were made within 100 m of the surface. Measurements made over the plains of Nebraska (to be published in a G.R.D. paper) indicate that the airplane-measured stresses reached a maximum at the lowest level measured, 18 m, in all but one series of observations. Perfect agreement with stress measuring equipment mounted in the ground was not attained, but no significant differences were obtained between the lowest airplane measurements and the ground equipment.

It should be noted that Scrases (1930), Cramer and Record (1953), observed an increase in the stress with height, while Panofsky (1956) observed a maximum of the stress at the 46 m or higher levels of the Brookhaven tower on all but one of his measurements. He suggested the maxima to be caused by the presence of trees upwind of the tower. Scrase's observations have been similarly explained. The presence of an increase over the Gulf of Maine indicates that the phenomena is of a much more general character and most likely is related to the thermal wind, variation of the turbulence intensity with height, and accelerations of the air. It seems most likely that the change in the sign of the volume frictional force, $\alpha \frac{\partial \tau}{\partial z}$, results in an acceleration of the air. If this is the case, a carefully planned observing program must be set up to measure each one of these factors. The equation that applies to the problem of an increasing stress with height is, in conventional vector notation,

$$\dot{V} = \alpha \nabla_H p + \nabla_H \frac{T'}{T^*} dz - 2 \Omega_z V + \alpha \frac{\partial \tau}{\partial z}$$

which accounts for the acceleration of a particle, changes in the pressure field due to heating along the trajectory, Coriolis force and the variation of the stress with height. Accelerations, downstream heating, and changes in the intensity of the turbulence are all considered important in the case since all of these factors are present in both Panofsky's studies at Brookhaven and the present studies over the Gulf of Maine.

D. Average values of the heat flux.

The values of the heat flux have been averaged for several height ranges over the land, and for the stable and unstable groups over the water. These averages are tabulated in Table 3.

Before interpreting the values in the table, let us state clearly what quantities are measured by the airplane technique. The heat flow computed from averaging the individual products of $w'T'$ is the net flux composed of a downgradient flow due to turbulent diffusion from the warm inverstion air and the upward flow produced by the convection of parcels heated by the water. Hence the total amount of heat entering the ground layer from both the top and bottom will be greater than the value found at any given level by the airplane measuring technique. The larger values of the heat flux measured from the temperature changes confirm this conclusion. From the small values measured in the two cases of great stability (Oct. 20 and Oct. 27) it is concluded that an appreciable amount of heat is transported by gusts with temperature deviations too small to be detected by the thermopile. The thermopile record was read to the nearest 0.1°C while the temperature variations of the smallest gusts were less than this value. Thus the heat temperature variations of the smallest gusts were less than this value. Thus the heat transported by the small gusts cannot be measured accurately by the present system, and the flux values due to diffusion will be systematically low. The fluxes due to large scale diffusion and convection will be recorded faithfully.

Returning to the table of fluxes, it is seen that the maximum flux recorded in the unstable column is 1 mcal cm^{-2} sec^{-1} yet the average flux for these cases as measured by the temperature change method is about 4 mcal $cm^{-2}sec^{-1}$. If both upward and downward fluxes were measured accurately throughout the spectrum of diffusion and convection, then we could conclude that 2.5 mcal cm^{-2} sec^{-1} entered the layer from the bottom and 1.5 mcal cm^{-2} sec^{-1} entered from the top. However, as we know the measurements are systematically low, we can only say that heat enters from the top and bottom but not specify the relative amounts.

The small negative values found in the stable cases show that technique gives the correct sign of the flux, but compared with the flux found from temperature changes, gives values at least one order of magnitude too small. The negative values found in the 1000 to 2000 m range for all three classes prove that there is an important heat flow downward from the warmer air aloft. Presumably these measured values are also too small by a factor of 10.

The increase in the flux with height is the lowest 100 m is puzzling if real, but may be the result of much of the heat flow being outside of the spectral range of the thermopile and airplane technique.

E. Stability and sea, air, and inversion potential temperature differences.

One of the unexpected results of the present study was the lack of a relation between the static stability of the air and the temperature difference between the air and the water. It was assumed that cool air flowing over warm water would display a negative stability. It was found that in the lowest 50 m to 300 m range negative stabilities are observed frequently, but not consistently, whenever air flows over warmer water. Above the 300 m level only one negative stability was observed. In the higher regions of the ground layer above the 300 m level the stabilities became positive and increased slowly up to the height of the inversion when they increased rapidly as the inversion is encountered.

Data on stabilities and temperatures have been collected in Table 4. Stabilities are given for each major region of the atmosphere studies. Heights are not given but may be obtained from Section II. Potential temperatures are tabulated for the water surface, for the main body of the ground layer at the point farthest from shore, and for the base of the warm air above the inversion. The last two columns give the temperature differences of the water and air, and the air and the inversion air.

From a comparison of the tabulated stabilities and temperature differences it is clear that while there is a weak relation between the temperature difference and stability of the lowest air many of the observations show no relation. For example, a stability of -7×10^{-8} cm^{-1} was found for December 7 where the temperature difference was only 0.5°C and a stability of 3×10^{-8} cm^{-1} was found on January 29 when the difference was 12°C. The case of December 7 may be explained as residual instability retained from the instability set up over the land. The rest of the cases are the result of great stability aloft and turbulence in the lower layers.

Here their turbulence is actively mixing downward the warm air of the inversion region as well as mixing upward the air warmed by the water. Hence an airplane flying as low as 50 m may completely miss recording the instability of the air although it will encounter numerous convective currents carrying heat upward. Thus the stability of the air as measured by the airplane frequently gives no clue to the nature of the underlying surface. The presence of warm convective currents does, however, prove to be a good indicator of a warm underlying surface. It is these convective currents that contribute the upward component of the heat flux in a stable atmosphere in accordance with Priestley's theory (1954). A confirmation of this theory was made by Bunker (1956), using data obtained by airplane on a Rhode Island to Bermuda flight.

The relation of stability of the upper region of the ground layer to the ground layer potential temperature minus inversion layer air potential temperature inconsistent in sign since all values are positive. However, the magnitude of the stability does not follow the magnitude of the temperature difference very well. This scatter is brought about by the variation in the magnitude of the turbulence in the various cases.

Table 4

Ground Layer Stabilities, Water, Ground Layer Air, and Inversion Layer Air Temperatures and Differences

Date	Stabilities $(d\theta/dz)/\theta \times 10^8$ cm^{-1}			Potential Temperatures °K			Temperature Differences	
	Land	Near-Shore	Offshore	Water	Ground Layer Air	Inversion Air	Sea minus Air	Air minus Inversion
Oct. 20, 1953	19	5	36	286	294	296	-8	-2
Oct. 27, 1953	29	6	10	285	293	297	-8	-4
Dec. 7, 1953	-10	-7 4	-	283	282.5	290	0.5	-7.5
Dec. 17, 1953	0.8	0.4	-19 6	283	267	269	16	-2
Dec. 23, 1953	-1 6	-1 4	0.0	282	277	288	5	-11
Jan. 5, 1954	1	5	-	278	272.5	283	5.5	-10.5
Jan. 7, 1954	-0.2	-1 6	0.0 10	281	275.5	280	5.5	-4.5
Jan. 29, 1954	11	-	3	278	266	275	12	-9

F. Relation between σ_w/u, σ_u/u and water minus air temperatures

Studies of turbulence by other workers have shown that the general amplitude of the atmosphere turbulence increases with increasing stability. The present data have been used to establish the relation between the water minus air temperatures and the root-mean-square vertical and horizontal turbulence deviation velocities. To demonstrate this relation, values of σ_w and σ_u, divided by the gradient wind, were plotted against the water-air temperature. The values observed at 300 m are presented in Figure 20. It is seen that a strong relation exists between the turbulence and the temperature differences. Both σ_w/u and σ_u/u increase with greater positive temperature differences, but σ_w/u increases much more rapidly; as might be expected from buoyancy considerations. Mean lines have been drawn through the points which curve upward, although this is only a hint from the data that the lines should not be straight.

A similar chart could be made for the overland cases if soil surface temperatures were available. Any plot based on stabilities of the air leads to a scatter of points. This is due to the poor relation discussed earlier between the observed stabilities of the air and the difference in temperatures between the air and the underlying surface.

G. Turbulence in atmospheres with nearly neutral stability.

On three occasions measurements of turbulence, stress, and heat flows were made in air masses that were of nearly neutral stability. One of these cases, Dec. 7, 1953, was measured 40 km offshore. Upwind over the land, the air was very unstable and turbulent. Another was observed over land on Dec. 17, 1953, while the third properly belongs to a different series of observations but will be reported upon here. The observations were made near Bermuda at a time of strong (26 m/sec) winds from the northwest. In Table 5 are listed values of σ_w and σ_w/u for the 300 m level. It is noted that the ratio is highest over land, but with considerable variation among the over water values. The relation between the Dec. 7 measurement and the Jan. 20, 1955, measurement is particularly significant since the air 40 km offshore had only 1/2 hour previously a σ_w of 197 cm sec^{-1} while the Jan. 20 must be considered an equilibrium state of turbulence since the air had been flowing at about the same velocity over about the same surface for 20 hours or better.

The only conclusion that can be drawn from such a small number of observations is that the turbulence over land is greater than over the water for a given wind speed.

H. Calculations based on the mixing length theory.

The classical mixing length theory of von Kármán provides a convenient basis for discussing observations of diffusion although its value is being diminished by the development of more basic theories of turbulent diffusion. Nevertheless, it may be profitable at this point to compare a few austausch values with root-mean-square vertical velocities and make an estimate of root-mean-square mixing lengths. Values of the coefficient of turbulent mass exchange and w are entered in Table 6. The coefficient values presented were computed from heat or water vapor flows and gradients on the only four days which yielded reliable results.

Too few values are available to establish whether a linear relation exists between the coefficient and σ_w or a square low exists, from the definitional equation

$$A = \bar{\rho}\,\overline{w'l'} = \rho r \sigma_w \sigma_l$$

it is seen that A would be a function of $\overline{w'^2}$ since $\sigma_{l'}$ might be expected to be proportional to σ_w.

A value of l' may be computed from this equation if r is assumed to be 0.8. Using A = 1200, and w' = 131, l' = 115 m. This value is consistent with the presence of connection in the air mass.

Table 5

Values of σ_w and σ_w/u at 300 m in atmosphere of neutral stability.

Date	Surface	σ_w cm sec^{-1}	σ_w/u x 10^2
Dec. 7, 1953	Water	43	1.7
Dec. 17, 1953	Land	118	13
Jan. 20, 1955	Water	111	5.8

Table 6

Values of the Coefficient of Turbulent mass exchange and σ_w.

Date	Coefficient	R-M-S Vertical Velocity cm sec^{-1}
Oct. 20, 1953	10	13
Dec. 17, 1953	1200	130
Jan. 5, 1954	800	60
Jan. 29, 1954	770	120

Using A = 10 and w = 13, ℓ' = 10 m. These values may be in error because of the assumptions made, but serve to give a rough idea of the vertical distance traveled by the gust before they lose their identity.

IV Summary of results

The observations obtained during this brief experiment have established the orders of magnitude of the turbulent velocities and fluxes associated with various combinations of stabilities, wind speed, and surface roughness. It was found that root-mean-square vertical velocities varies from a low value of 6 cm sec^{-1} obtained in a stable air mass over water to a maximum of 197 cm sec^{-1} in an unstable air mass over land. The root-mean-square horizontal deviation velocities have a similar range and pattern.

Both components of the turbulent deviation velocities are observed to increase with height up to a broad maximum in the 100 m to 500 m height range and to decrease slowly with height up to the inversion level. In the inversion layer the turbulent velocities drop to very low values. This variation with height need not be discussed in detail since this variation has been observed previously, although the range of values had not been established. The variation of the σw and σu with height emphasizes the errors that may result from using an assumption of constant eddy diffusion or viscosity with height in any theory. This danger has been recognized for many years and is the greatest fault of theories such as the Ekman wind spiral.

The ratio of the root-mean-square vertical velocities to the root-mean-square horizontal deviation velocities and its variation with height and stability of the air is a valuable bit of information for the study of turbulent flow near a surface or an inversion. It is seen that σu becomes relatively large close to the surface indicating a damping of the vertical component by the surface. In unstable air the buoyancy of the warmer air amplifies the vertical component of the wind so the $\sigma w/\sigma u$ becomes greater than 2.

Possibly the greatest contribution of the present work is the determination of the magnitude of the shearing stresses and the variation of the stress with surface roughness, wind speed, height and stability. The variation with stability is most pronounced, ranging over water surfaces from only a 0.01 dyne cm^{-2} to about 4 dyne cm^{-2}. This dependence of stress upon stability as well as wind speed forces new consideration of the energy balance of both the atmospheric circulation and the oceanic wind-driven circulations.

In addition to the stress supplied by the $\overline{w'u'}$ product, another stress due to the $\overline{u'u'}$ product should be mentioned. This stress is of the same order of magnitude as the $\overline{w'u'}$ stress but is of less importance because of the small resulting volume frictional force, $\frac{\partial(-\bar{\rho}\,\overline{u'u'})}{\partial x}$. This force is at least an order of magnitude smaller than $\frac{\partial(-\bar{\rho}\,\overline{u'w'})}{\partial z}$ since $-\bar{\rho}\,\overline{u'u'}$ changes appreciably in 10's of kilometers while $-\bar{\rho}\,\overline{w'u'}$ changes a similar amount in a single kilometer in the vertical.

As mentioned above, the gradient of the shearing stress in the vertical is large, and except for an increase in the lowest layer, is negative. Thus an appreciable retarding force acts upon the moving air mass. In the lowest layer where the gradient is positive an accelerating force is indicated. Although the causes and results of this positive gradient have not been determined, it seems clear that this gradient is not a steady state phenomenon and probably has a rather small effect upon the large scale circulation pattern. In certain regions where rapid changes of temperature and surface roughness occur, appreciable accelerations of the air may occur.

A comparison of the heat flows computed by the $c_p\overline{\rho w'T'}$ and the $c_p \frac{dp}{-g} \Delta T$ methods shows that the cross product technique gives a value that represents the total sensible heat flow as long as heat is being added to the air mass from one direction only. When heat is added from both top and bottom, the cross product gives the net flow of heat through any surface. The average net flow was found to be about 1 mcal cm^{-2} sec^{-1} from unstable air and much smaller positive and negative flows in the stable cases.

The downward heat flow observed in air masses flowing from land out over warmer water is the direct result of the degree of turbulence of the air and the stability of the air aloft. Because of the stability aloft and the turbulence, a stable lapse rate was established from the ground layer inversion down to a few hundred meters from the water surface. Thus it is seen that the observed stability of the ground layer has a very poor relation to the relative temperature of the sea and air. This phenomenon of stability is produced by the large scale subsidence of the air flowing from the northwest.

The effect of surface roughness upon turbulence was studied by comparing observations of σ_w/u obtained on days of similar stability. The study showed that the land produced from 2 to 6 times as great a value of σ_w/u as the water surface produced.

At the beginning of this program of observational studies it was planned to enter fluxes and gradients into the classical diffusion equation and obtain values of the coefficient of turbulent mass exchange for heat, water vapor, and momentum. It was soon found, however, due to the fluxes of heat against the gradient and the existence of increases of stress with height, that the diffusion equation failed to be any aid in the descriptions or understanding of the processes operating in the atmosphere. As no suitable general theory of turbulence in the atmosphere has been developed, we must, for the time being, cling closely to the details of the observed structure of convective panels and of turbulence.

V References

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Legends for Figures

Figure No. 1.	Psychrometric and Turbulence data observed	Oct. 20, 1953
" " " " 2.	Cross Section of R-M-S Vertical Velocity	Oct. 20, 1953
" " " " 3.	Psychrometric and Turbulence data	Oct. 27, 1953
" " " " 4.	Cross Section of R-M-S Vertical Velocity	Oct. 27, 1953
" " " " 5.	Psychrometric and Turbulence data	Dec. 7, 1953
" " " " 6.	Cross Section of R-M-S Vertical Velocity	Dec. 7, 1953
" " " " 7.	Psychrometric and Turbulence data	Dec. 17, 1953
" " " " 8.	Cross Section of R-M-S Vertical Velocity	Dec. 17, 1953
" " " " 9.	Psychrometric and Turbulence data	Dec. 23, 1953
" " " " 10.	Cross Section of R-M-S Vertical Velocity	Dec. 23, 1953
" " " " 11.	Psychrometric and Turbulence data	Jan. 5, 1954
" " " " 12.	Cross Section of R-M-S Vertical Velocity	Jan. 5, 1954
" " " " 13.	Psychrometric and Turbulence data	Jan. 7, 1954
" " " " 14.	Cross Section of R-M-S Vertical Velocity	Jan. 7, 1954
" " " " 15.	Psychrometric and Turbulence data	Jan. 29, 1954
" " " " 16.	Cross Section of R-M-S Vertical Velocity	Jan. 29, 1954
" " " " 17.	Map of Gulf of Maine showing variation of R-M-S Vertical Velocities when the air is warmer than the water.	
" " " " 18.	Map of Gulf of Maine showing variation of R-M-S Vertical Velocities when the air is colder than the water.	
" " " " 19.	Relation of σ_w/σ_u to the sea-air temperature difference.	
" " " " 20.	Relation of σ_w/u and σ_u/u sea-air temperature difference.	

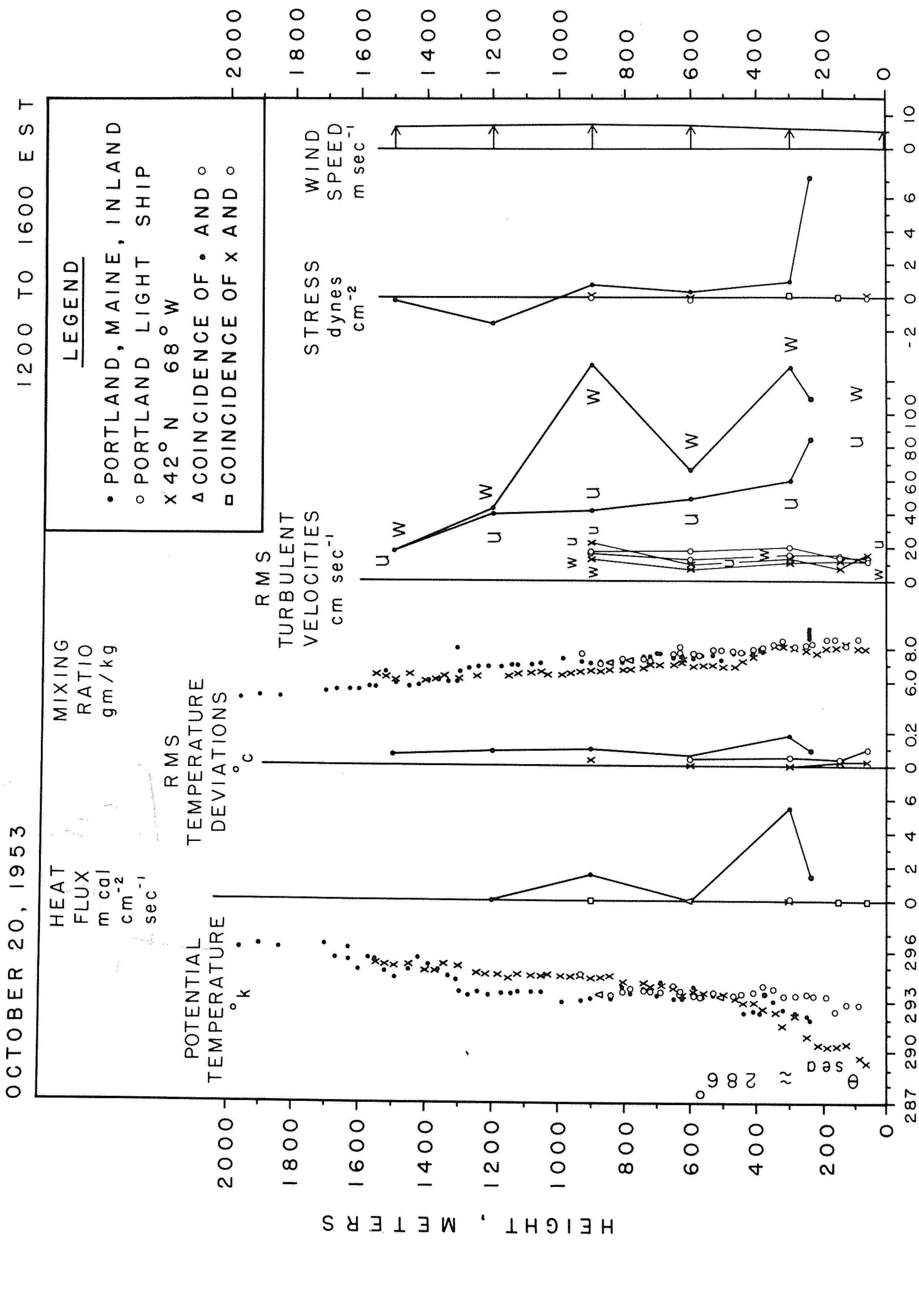

OCTOBER 20, 1953
1200 TO 1600 EST
LEGEND
• PORTLAND, MAINE, INLAND
○ PORTLAND LIGHT SHIP
X 42° N 68° W
Δ COINCIDENCE OF • AND ○
□ COINCIDENCE OF X AND ○
HEIGHT, METERS
POTENTIAL TEMPERATURE °K
HEAT FLUX m cal cm⁻² sec⁻¹
RMS TEMPERATURE DEVIATIONS °C
MIXING RATIO gm/kg
RMS TURBULENT VELOCITIES cm sec⁻¹
STRESS dynes cm⁻²
WIND SPEED m sec⁻¹
Θ sea ≈ 286

FIG. 1

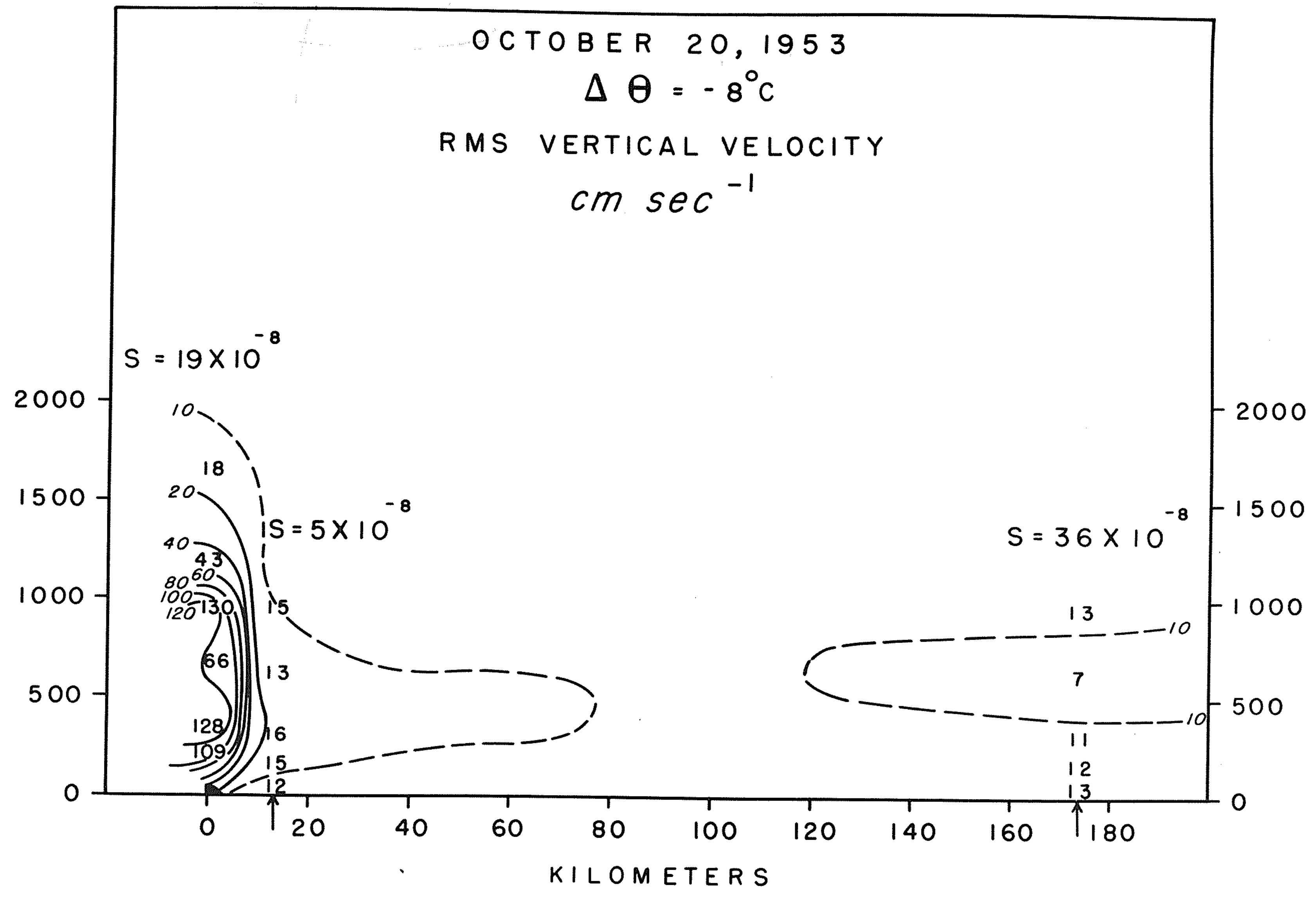

OCTOBER 20, 1953
Δ Θ = -8°C
RMS VERTICAL VELOCITY
cm sec⁻¹
S = 19 X 10⁻⁸
S = 5 X 10⁻⁸
S = 36 X 10⁻⁸
HEIGHT, METERS
KILOMETERS
0
500
1000
1500
2000
20
40
60
80
100
120
140
160
180
10
18
20
40
43
60
80
100
120
130
66
128
109
15
13
16
15
12
13
7
10
11
12
13

FIG. 2

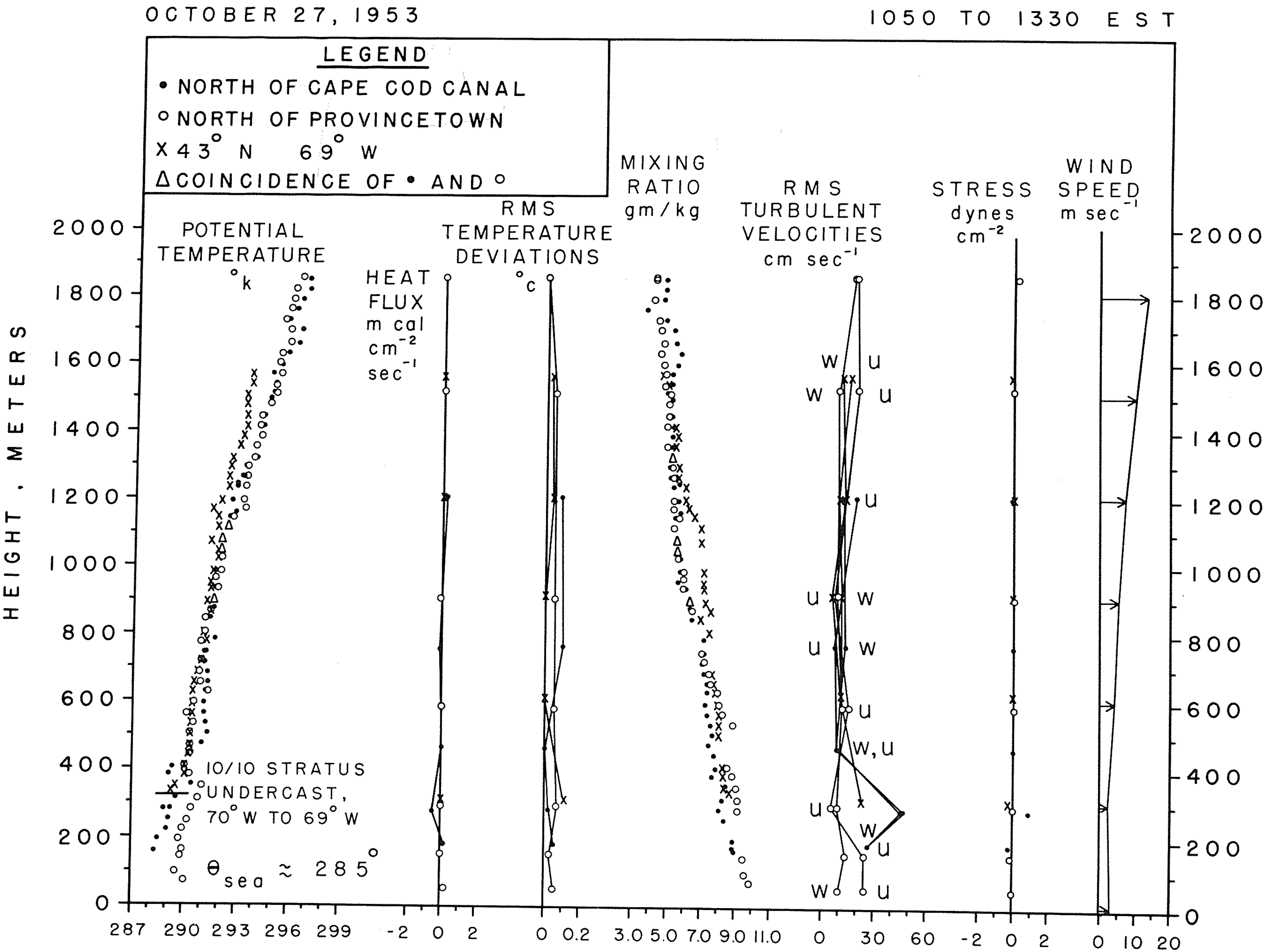

OCTOBER 27, 1953
1050 TO 1330 EST
LEGEND
• NORTH OF CAPE COD CANAL
○ NORTH OF PROVINCETOWN
X 43° N 69° W
Δ COINCIDENCE OF • AND ○
HEIGHT, METERS
POTENTIAL TEMPERATURE °k
HEAT FLUX m cal cm⁻² sec⁻¹
RMS TEMPERATURE DEVIATIONS °c
MIXING RATIO gm/kg
RMS TURBULENT VELOCITIES cm sec⁻¹
STRESS dynes cm⁻²
WIND SPEED m sec⁻¹
10/10 STRATUS UNDERCAST, 70° W TO 69° W
θsea ≈ 285°
0
200
400
600
800
1000
1200
1400
1600
1800
2000
287 290 293 296 299
-2 0 2
0 0.2
3.0 5.0 7.0 9.0 11.0
0 30 60
-2 0 2
0 10 20
W
U
W,U

FIG.3

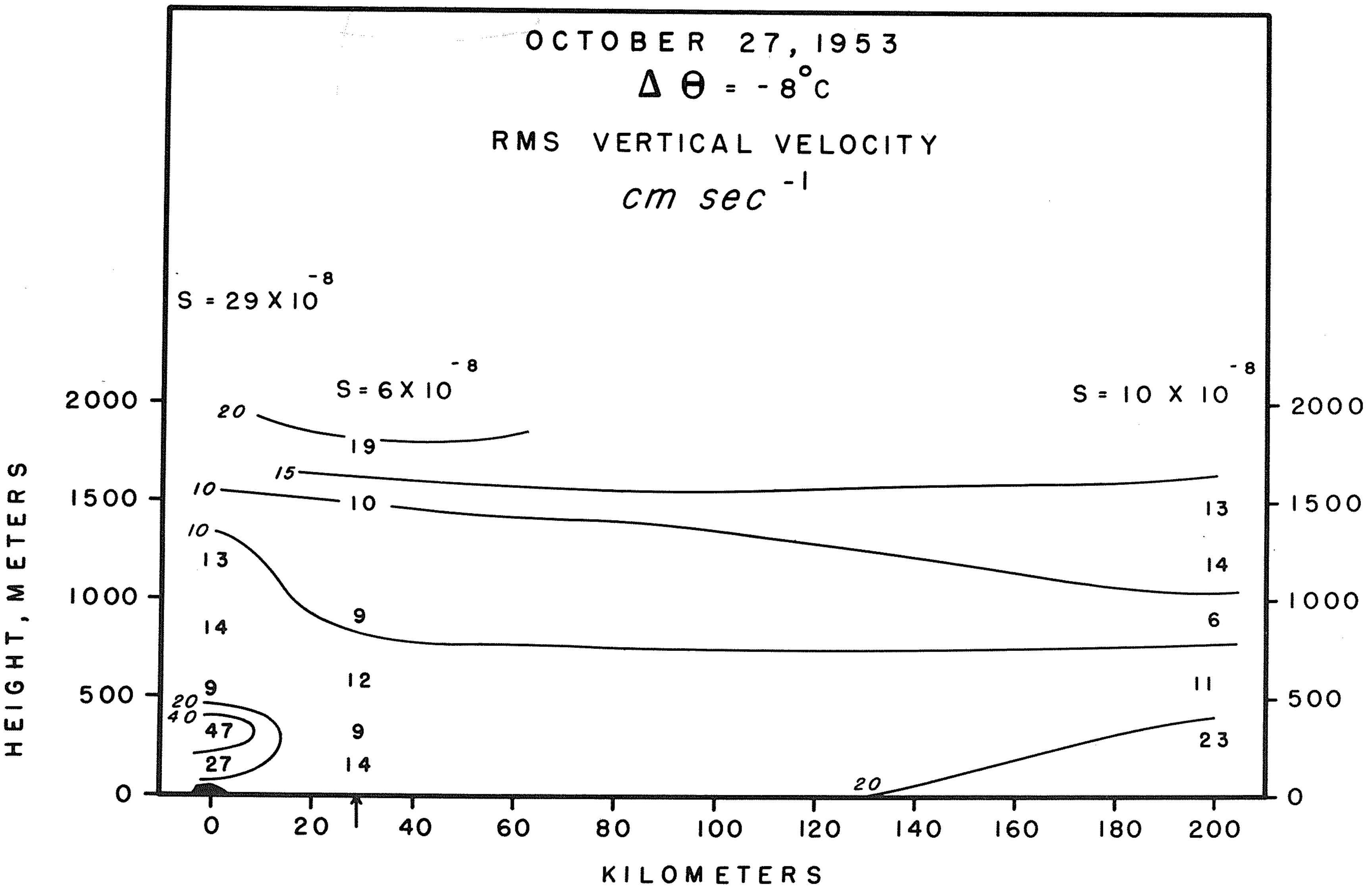

OCTOBER 27, 1953
$\Delta\Theta = -8°C$
RMS VERTICAL VELOCITY
$cm\ sec^{-1}$
$S = 29 \times 10^{-8}$
$S = 6 \times 10^{-8}$
$S = 10 \times 10^{-8}$
HEIGHT, METERS
KILOMETERS

FIG. 4

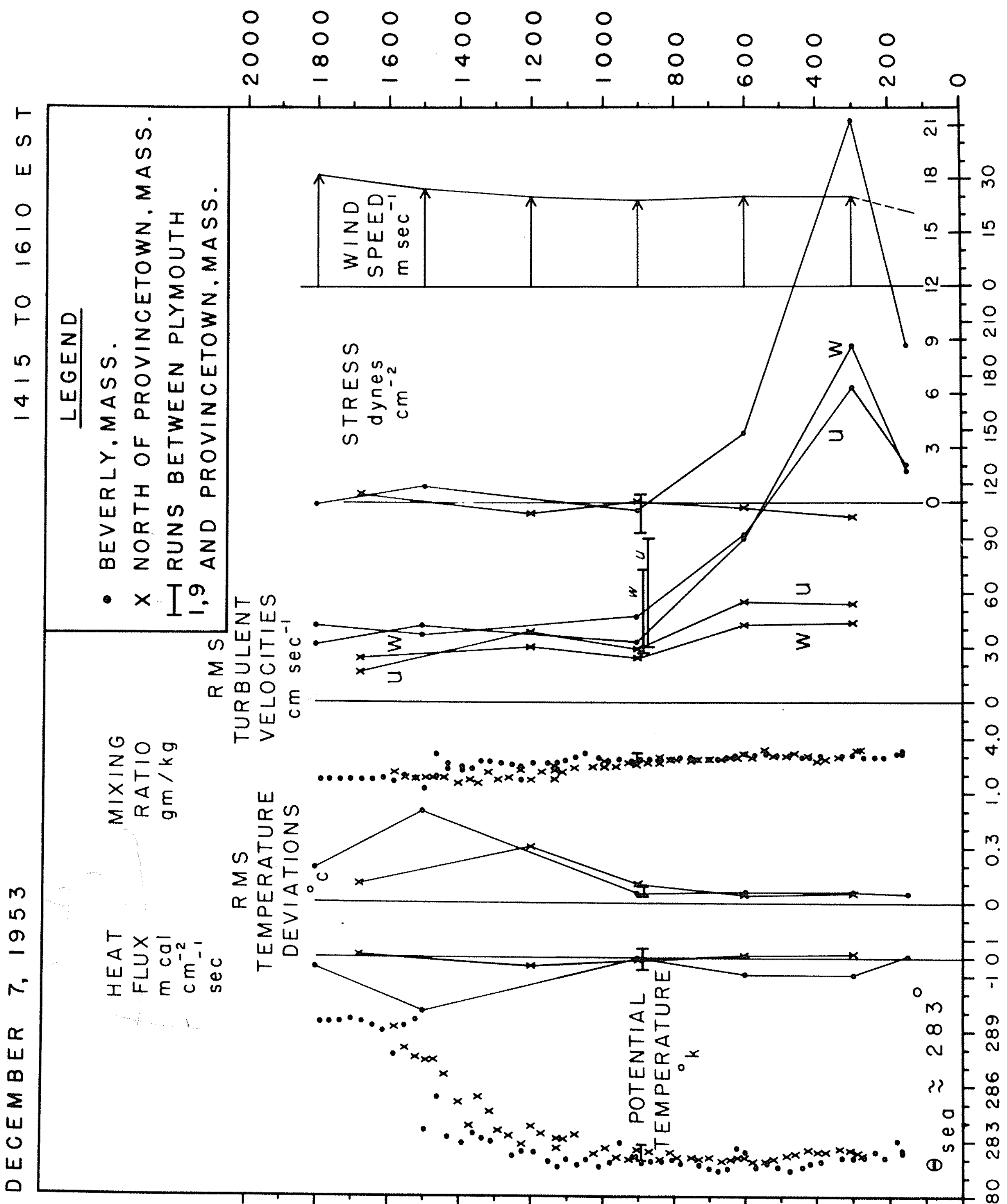
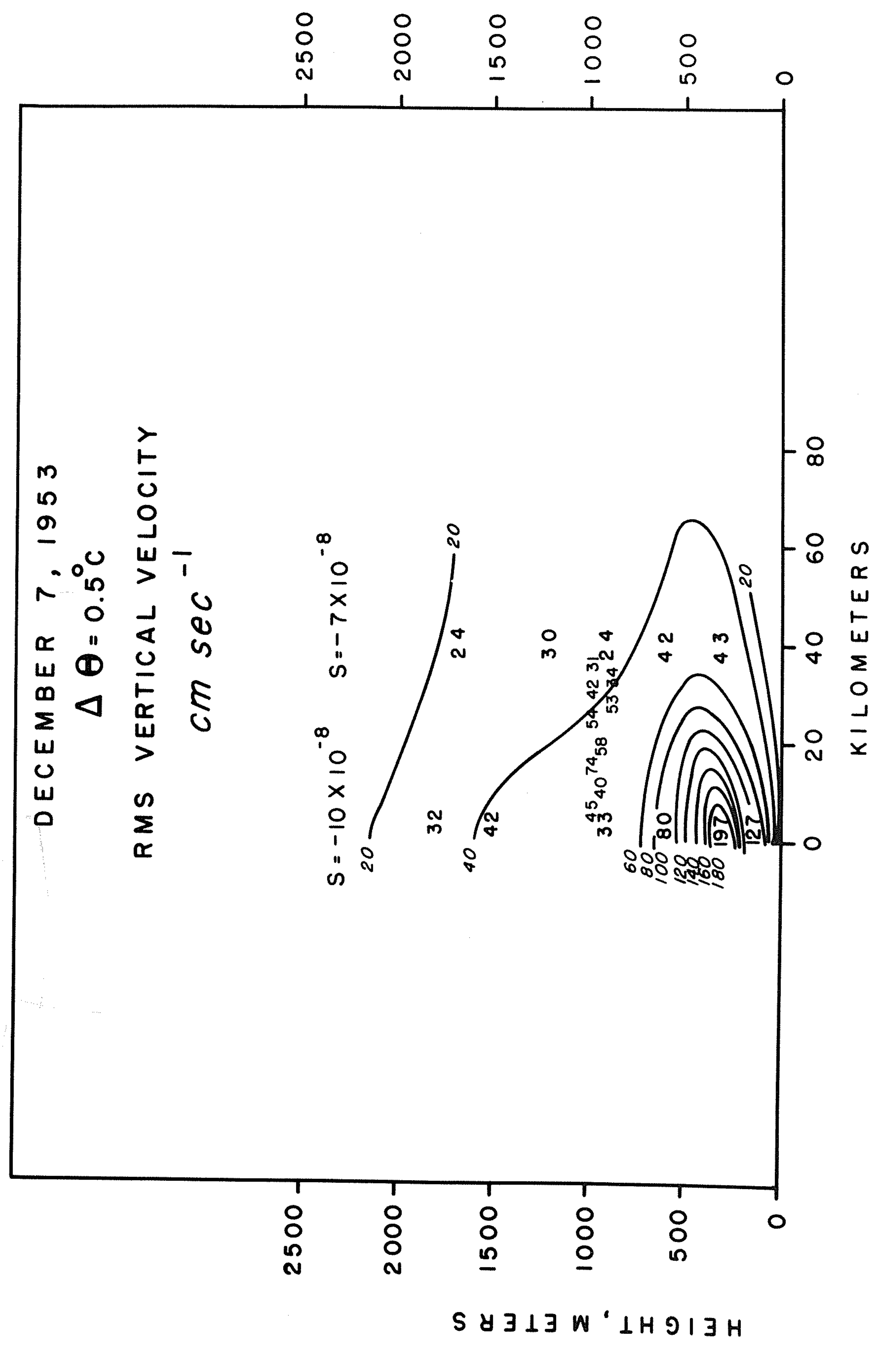

DECEMBER 7, 1953
1415 TO 1610 EST
LEGEND
• BEVERLY, MASS.
X NORTH OF PROVINCETOWN, MASS.
1,9 RUNS BETWEEN PLYMOUTH AND PROVINCETOWN, MASS.
HEAT FLUX m cal cm⁻² sec⁻¹
POTENTIAL TEMPERATURE °K
θ sea ≈ 283°
RMS TEMPERATURE DEVIATIONS °C
MIXING RATIO gm/kg
RMS TURBULENT VELOCITIES cm sec⁻¹
STRESS dynes cm⁻²
WIND SPEED m sec⁻¹
U
W
HEIGHT, METERS

FIG. 5



DECEMBER 7, 1953
Δ Θ = 0.5°C
RMS VERTICAL VELOCITY
cm sec⁻¹
S = -10 X 10⁻⁸
S = -7 X 10⁻⁸
KILOMETERS
HEIGHT, METERS

FIG. 6

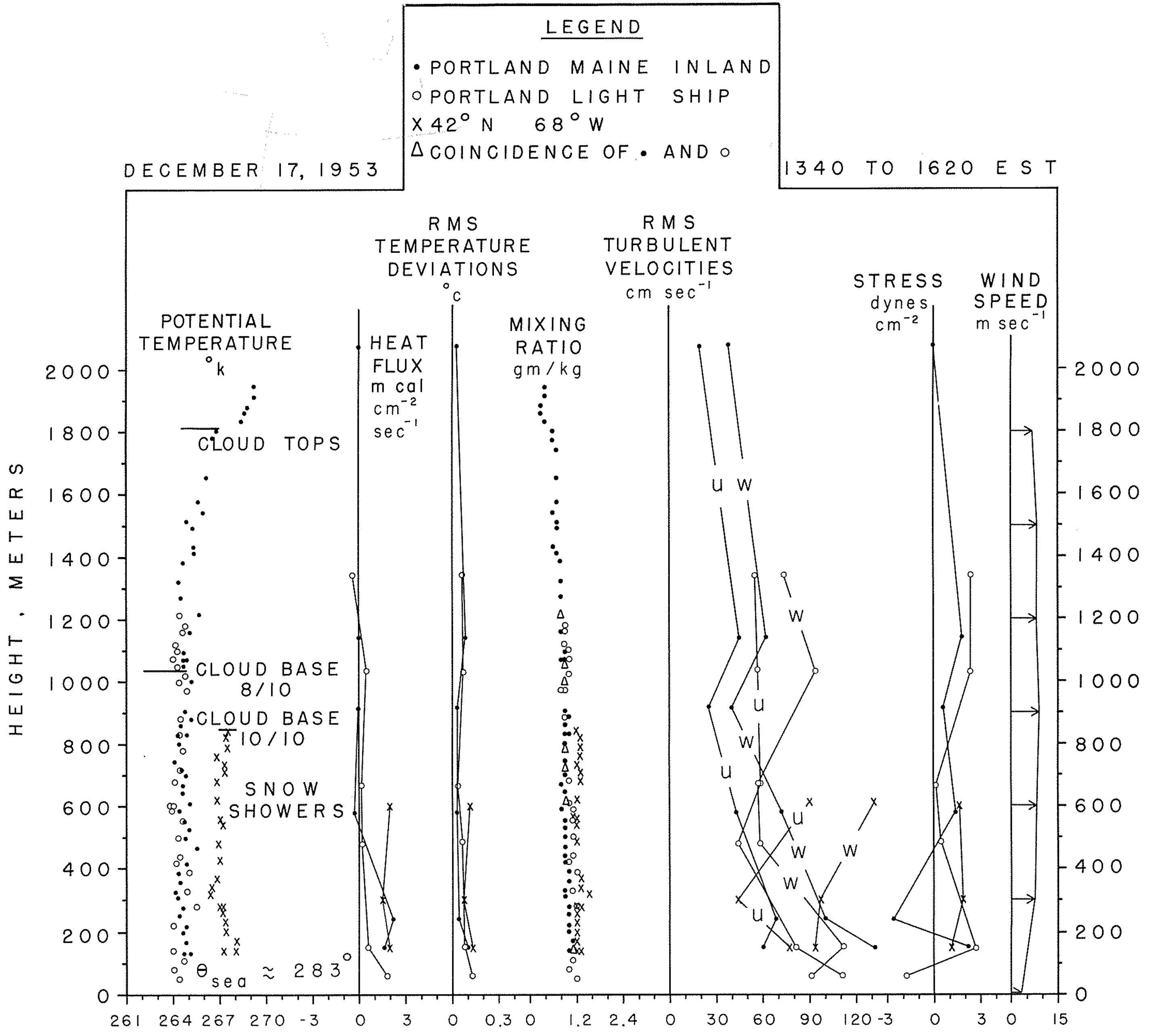
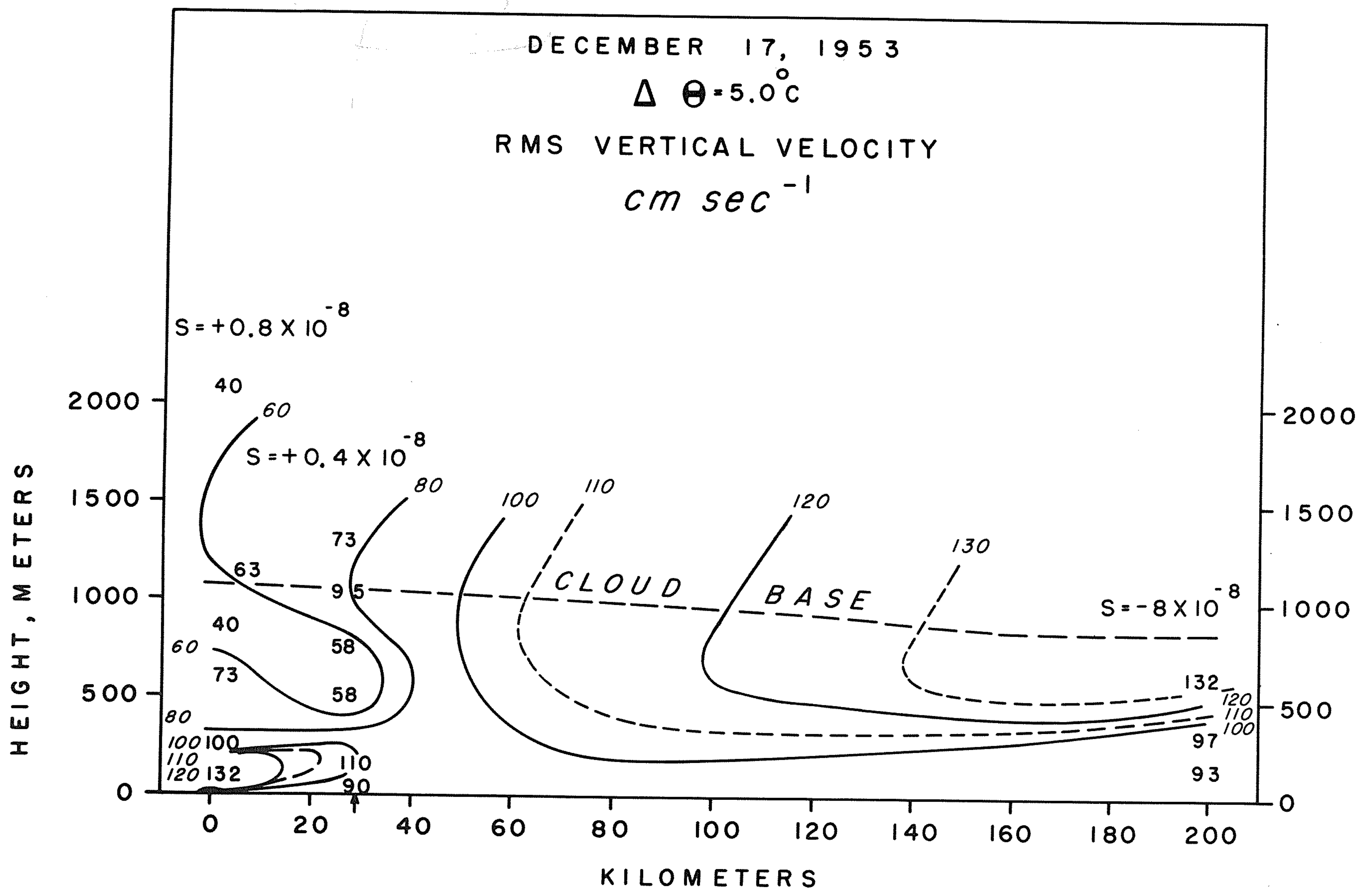

LEGEND
• PORTLAND MAINE INLAND
○ PORTLAND LIGHT SHIP
X 42° N 68° W
Δ COINCIDENCE OF • AND ○
DECEMBER 17, 1953
1340 TO 1620 EST
POTENTIAL TEMPERATURE °k
CLOUD TOPS
CLOUD BASE 8/10
CLOUD BASE 10/10
SNOW SHOWERS
θsea ≈ 283°
HEAT FLUX m cal cm⁻² sec⁻¹
RMS TEMPERATURE DEVIATIONS °c
MIXING RATIO gm/kg
RMS TURBULENT VELOCITIES cm sec⁻¹
STRESS dynes cm⁻²
WIND SPEED m sec⁻¹
HEIGHT, METERS

FIG.7



DECEMBER 17, 1953
Δ Θ = 5.0°C
RMS VERTICAL VELOCITY
cm sec⁻¹
S = +0.8 X 10⁻⁸
S = +0.4 X 10⁻⁸
S = -8 X 10⁻⁸
CLOUD BASE
HEIGHT, METERS
KILOMETERS

FIG.8

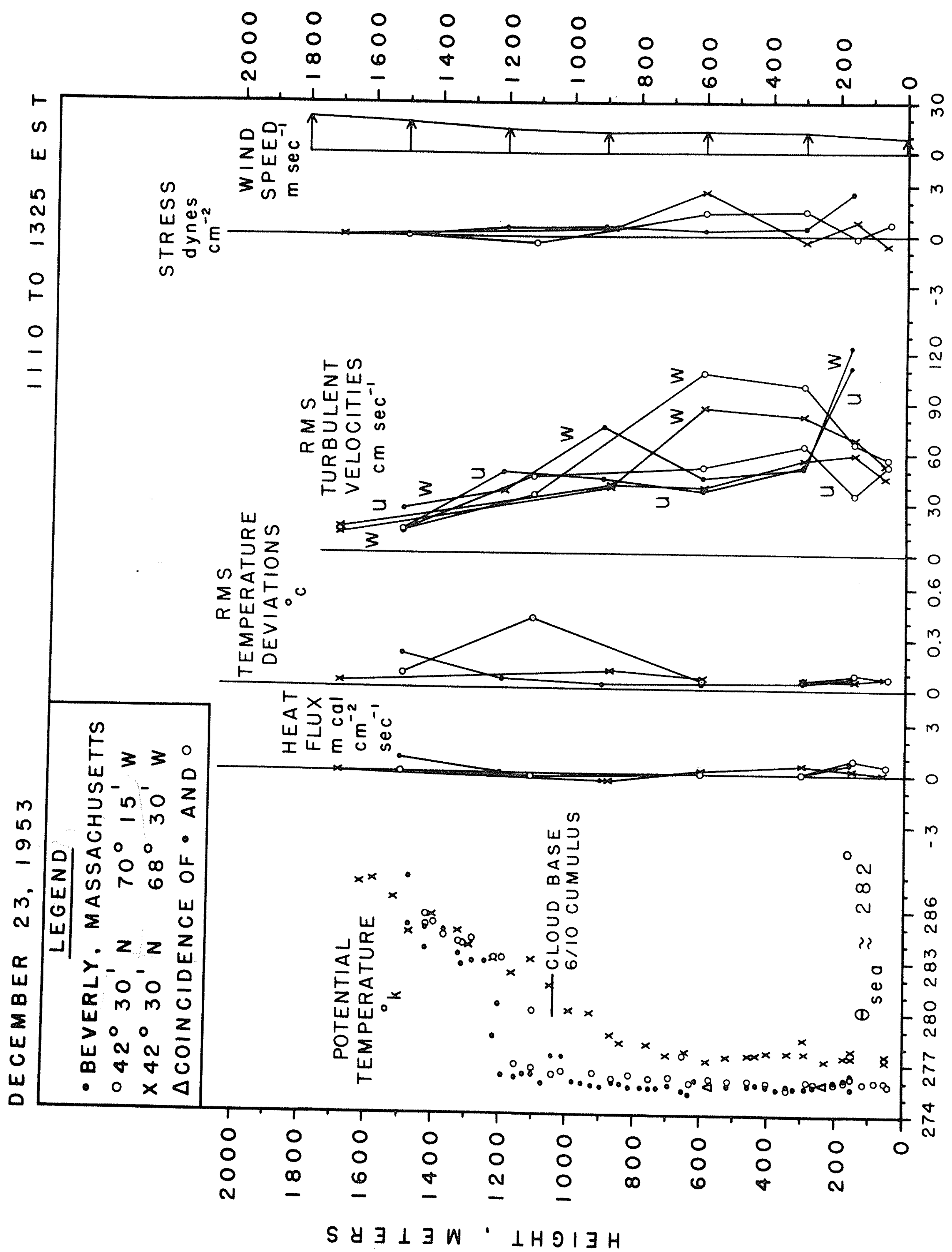

DECEMBER 23, 1953
1110 TO 1325 EST
LEGEND
• BEVERLY, MASSACHUSETTS
○ 42° 30' N 70° 15' W
X 42° 30' N 68° 30' W
Δ COINCIDENCE OF • AND ○
HEIGHT, METERS
POTENTIAL TEMPERATURE °K
CLOUD BASE 6/10 CUMULUS
$\Theta_{sea} \approx 282°$
HEAT FLUX m cal cm⁻² sec⁻¹
RMS TEMPERATURE DEVIATIONS °C
RMS TURBULENT VELOCITIES cm sec⁻¹
STRESS dynes cm⁻²
WIND SPEED m sec⁻¹

FIG.9

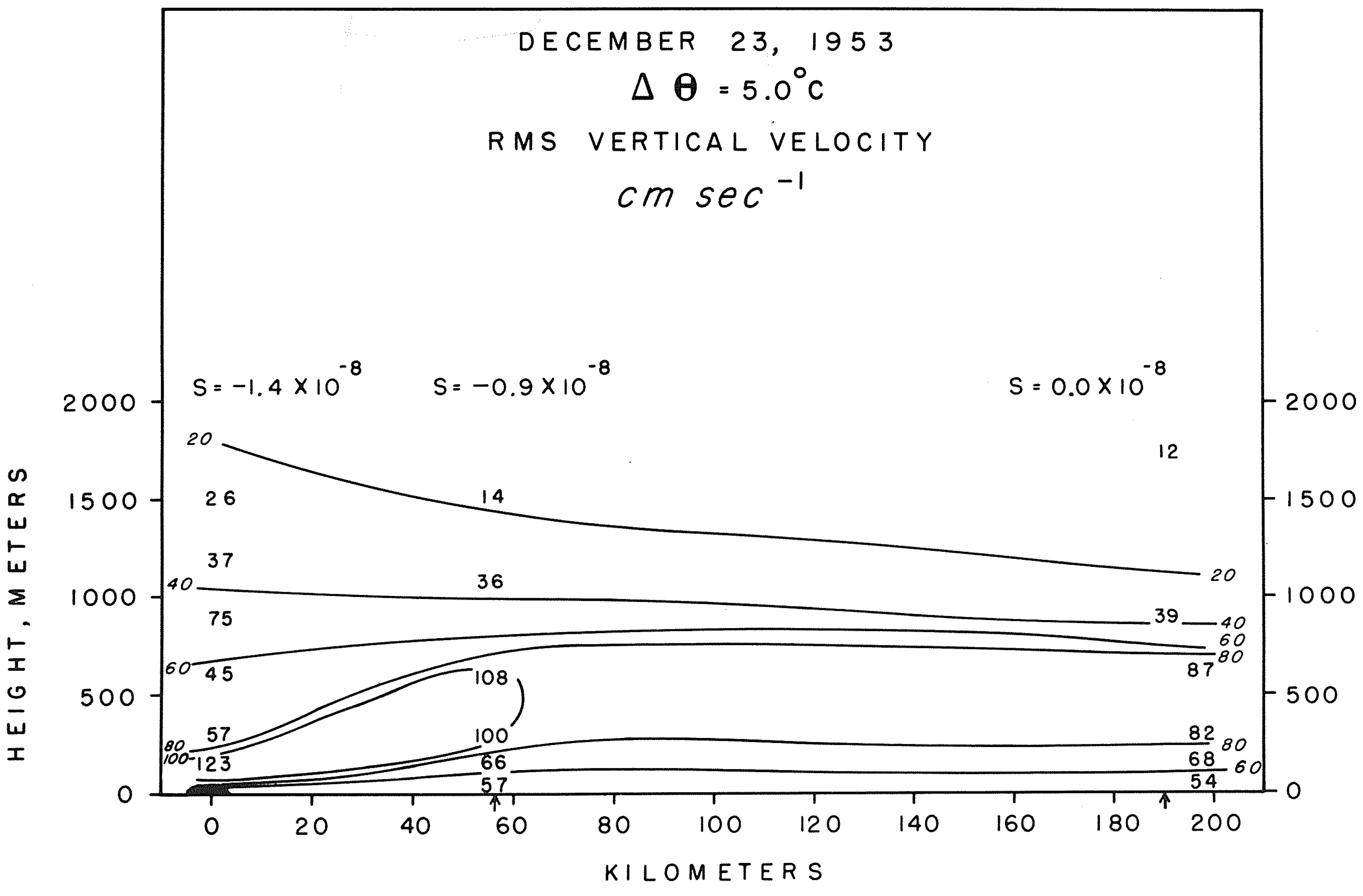
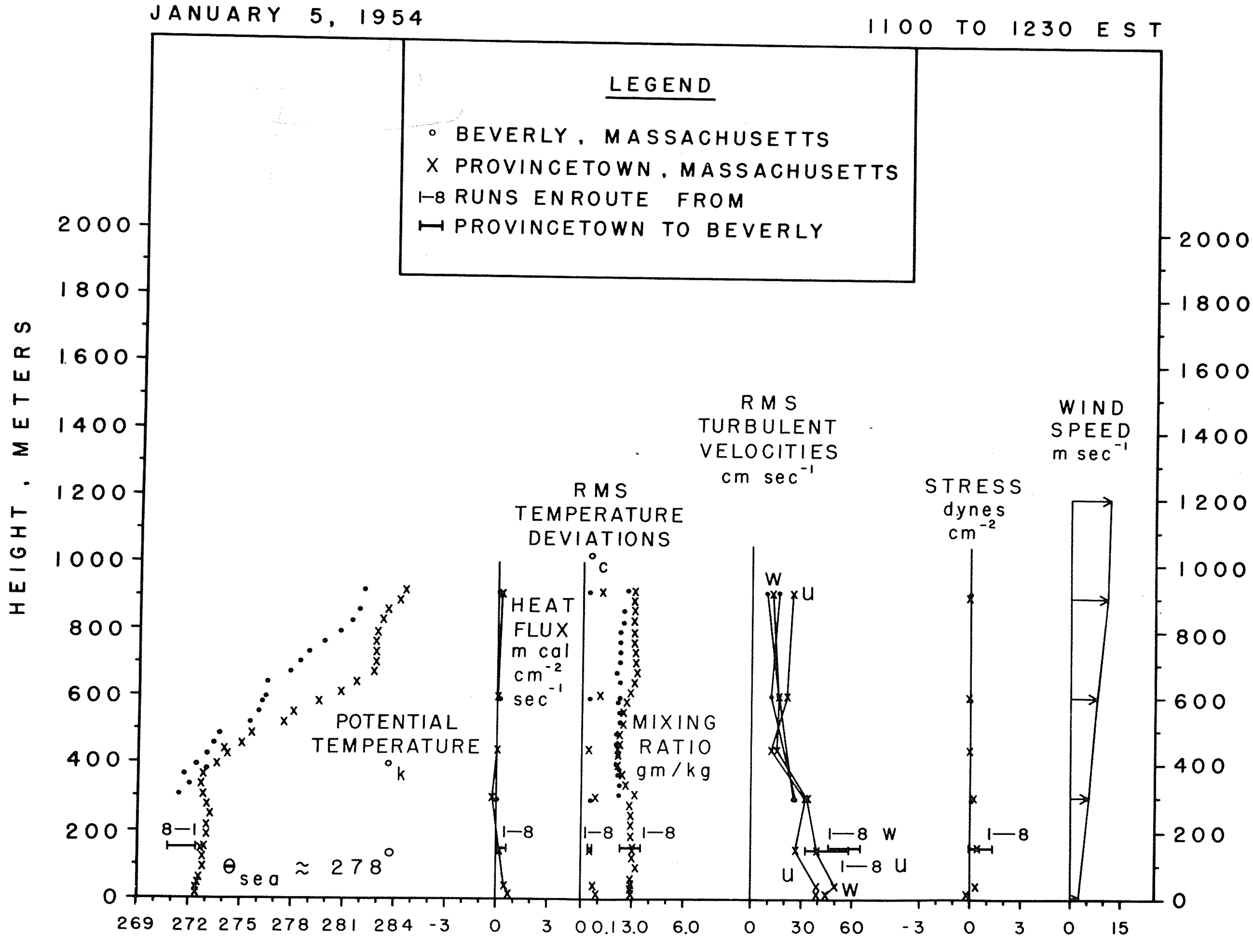

DECEMBER 23, 1953
Δ Θ = 5.0°C
RMS VERTICAL VELOCITY
cm sec⁻¹
S = -1.4 X 10⁻⁸
S = -0.9 X 10⁻⁸
S = 0.0 X 10⁻⁸
HEIGHT, METERS
KILOMETERS
0
20
40
60
80
100
120
140
160
180
200
500
1000
1500
2000
26
14
12
37
36
39
75
45
87
108
57
100
82
123
66
68
57
54

FIG. 10



JANUARY 5, 1954
1100 TO 1230 EST
LEGEND
° BEVERLY, MASSACHUSETTS
X PROVINCETOWN, MASSACHUSETTS
1-8 RUNS ENROUTE FROM
PROVINCETOWN TO BEVERLY
HEIGHT, METERS
POTENTIAL TEMPERATURE °k
θ sea ≈ 278°
HEAT FLUX m cal cm⁻² sec⁻¹
RMS TEMPERATURE DEVIATIONS °c
MIXING RATIO gm/kg
RMS TURBULENT VELOCITIES cm sec⁻¹
STRESS dynes cm⁻²
WIND SPEED m sec⁻¹

FIG.11

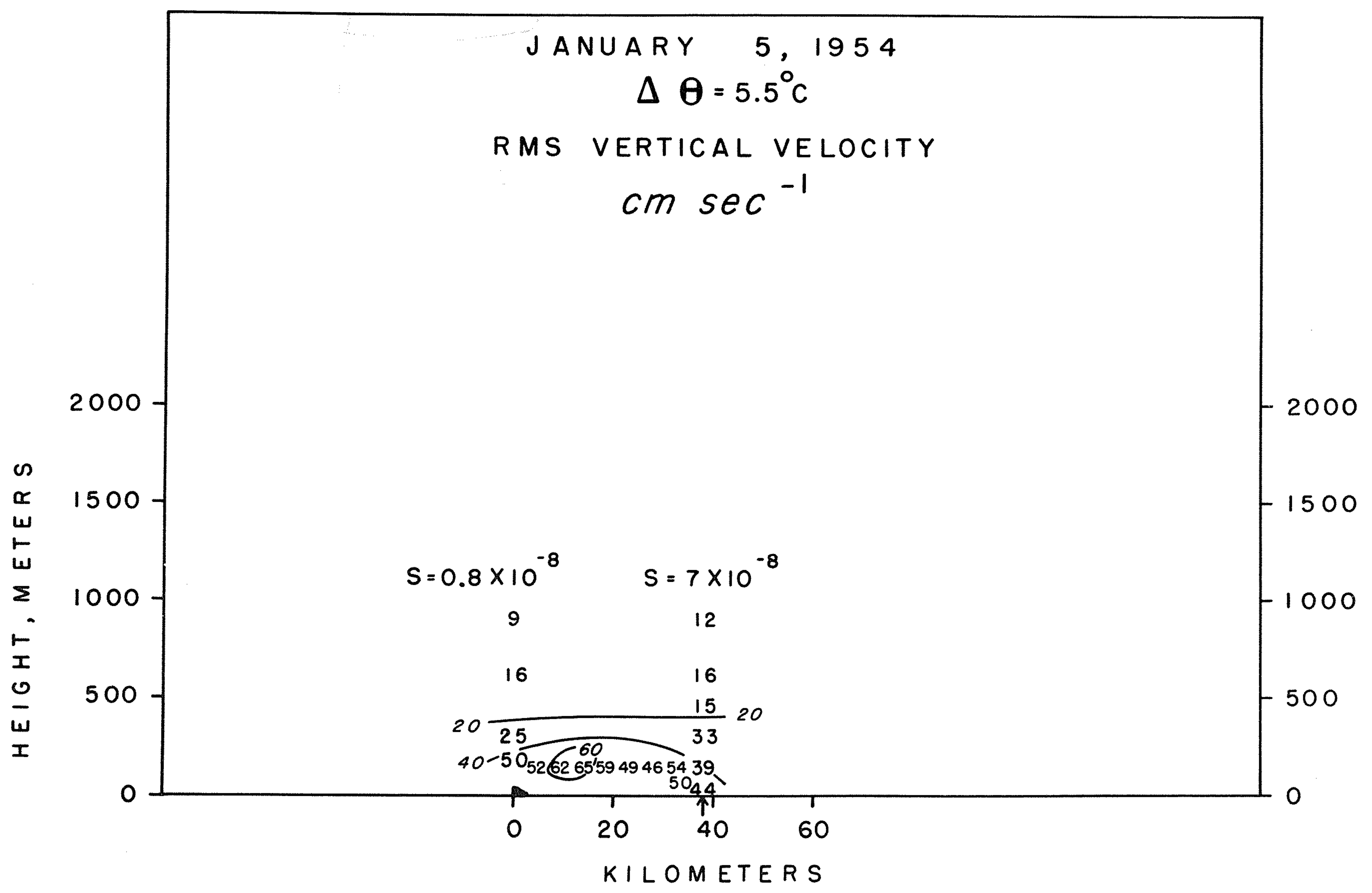

JANUARY 5, 1954
Δ θ = 5.5°C
RMS VERTICAL VELOCITY
cm sec⁻¹
S = 0.8 X 10⁻⁸
S = 7 X 10⁻⁸
9
12
16
16
15
20
20
25
33
40
50
60
52 62 65 59 49 46 54 39
50 44
2000
1500
1000
500
0
0
20
40
60
KILOMETERS
HEIGHT, METERS

FIG.12

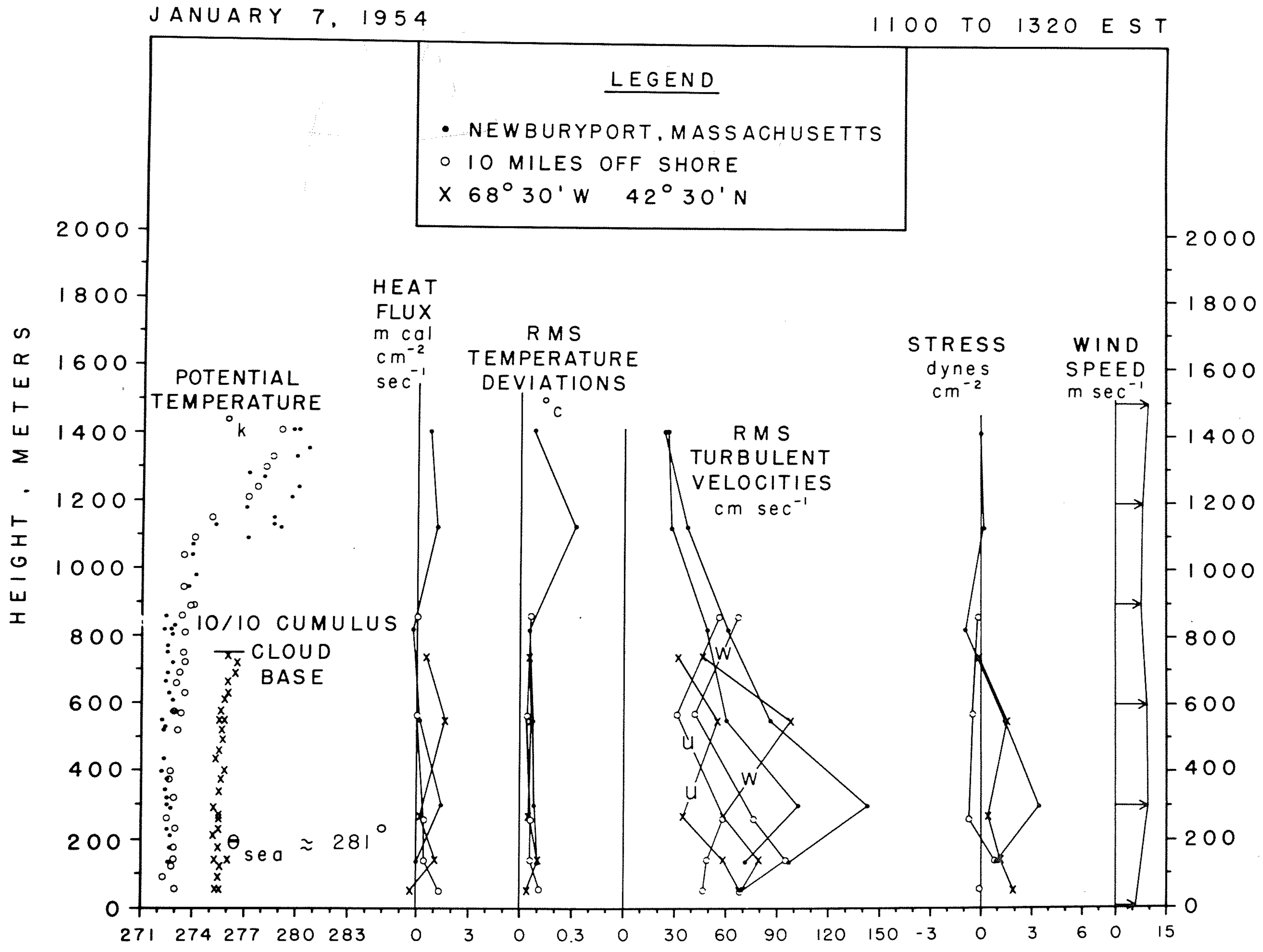

JANUARY 7, 1954
1100 TO 1320 EST
LEGEND
• NEWBURYPORT, MASSACHUSETTS
o 10 MILES OFF SHORE
X 68° 30' W 42° 30' N
HEIGHT, METERS
POTENTIAL TEMPERATURE °k
10/10 CUMULUS CLOUD BASE
θ sea ≈ 281°
HEAT FLUX m cal cm⁻² sec⁻¹
RMS TEMPERATURE DEVIATIONS °c
RMS TURBULENT VELOCITIES cm sec⁻¹
u
w
STRESS dynes cm⁻²
WIND SPEED m sec⁻¹
0
200
400
600
800
1000
1200
1400
1600
1800
2000
271 274 277 280 283
0 3
0 0.3
0 30 60 90 120 150
-3 0 3 6
0 15

FIG.13

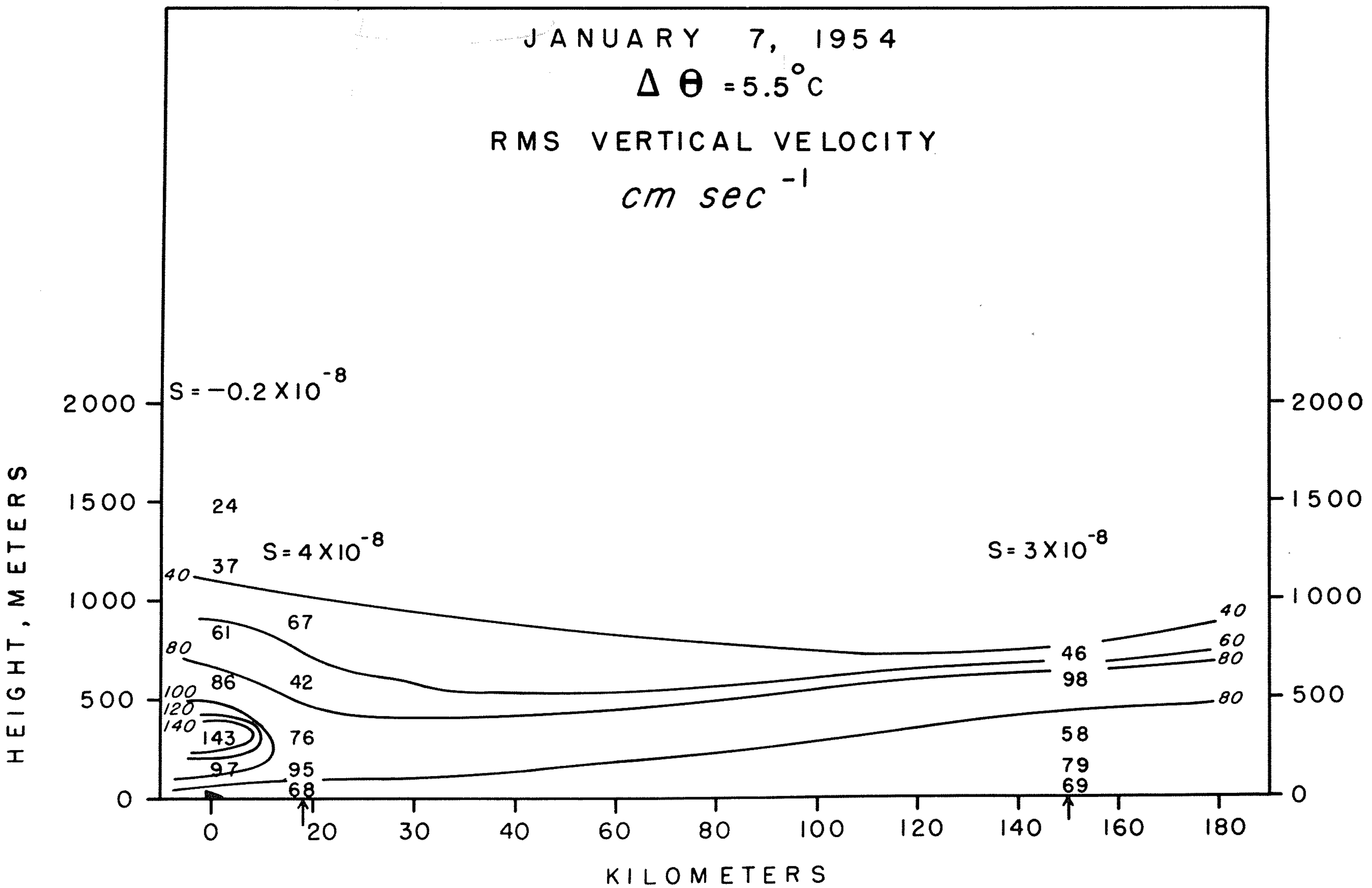
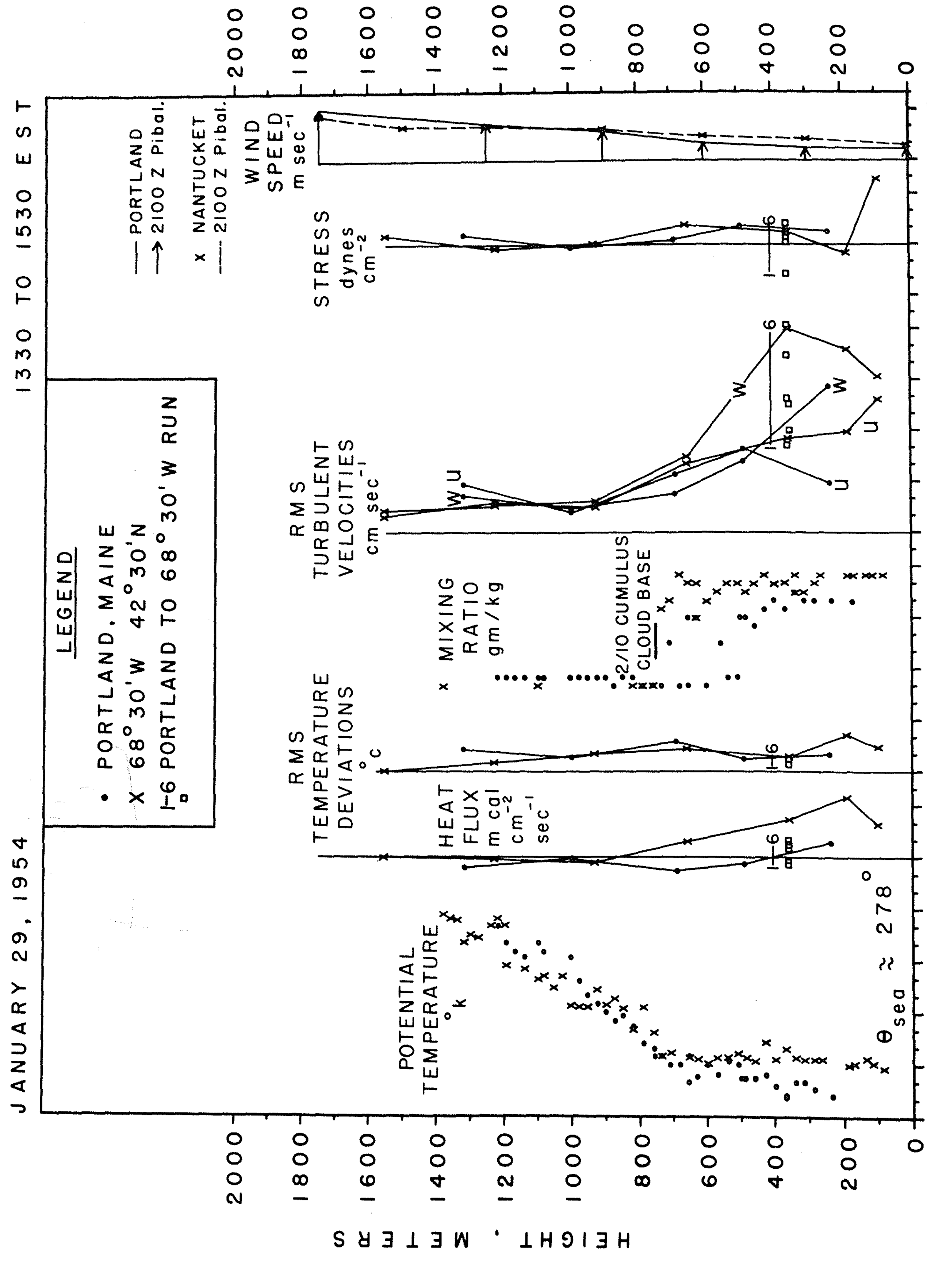

JANUARY 7, 1954
Δ Θ =5.5°C
RMS VERTICAL VELOCITY
cm sec -1
S = −0.2 X 10 -8
S = 4 X 10 -8
S = 3 X 10 -8
24
37
61
67
86
42
143
76
97
95
68
46
98
58
79
69
40
60
80
100
120
140
HEIGHT, METERS
2000
1500
1000
500
0
KILOMETERS
0
20
30
40
60
80
100
120
140
160
180

FIG. 14



JANUARY 29, 1954
1330 TO 1530 EST
LEGEND
• PORTLAND, MAINE
X 68° 30'W 42° 30'N
1-6 PORTLAND TO 68° 30'W RUN
— PORTLAND
→ 2100 Z Pibal.
x NANTUCKET
--- 2100 Z Pibal.
POTENTIAL TEMPERATURE °K
$\theta_{sea} \approx 278°$
HEAT FLUX m cal cm^{-2} sec^{-1}
RMS TEMPERATURE DEVIATIONS °C
MIXING RATIO gm/kg
2/10 CUMULUS CLOUD BASE
RMS TURBULENT VELOCITIES cm sec^{-1}
STRESS dynes cm^{-2}
WIND SPEED m sec^{-1}
HEIGHT, METERS

FIG.15

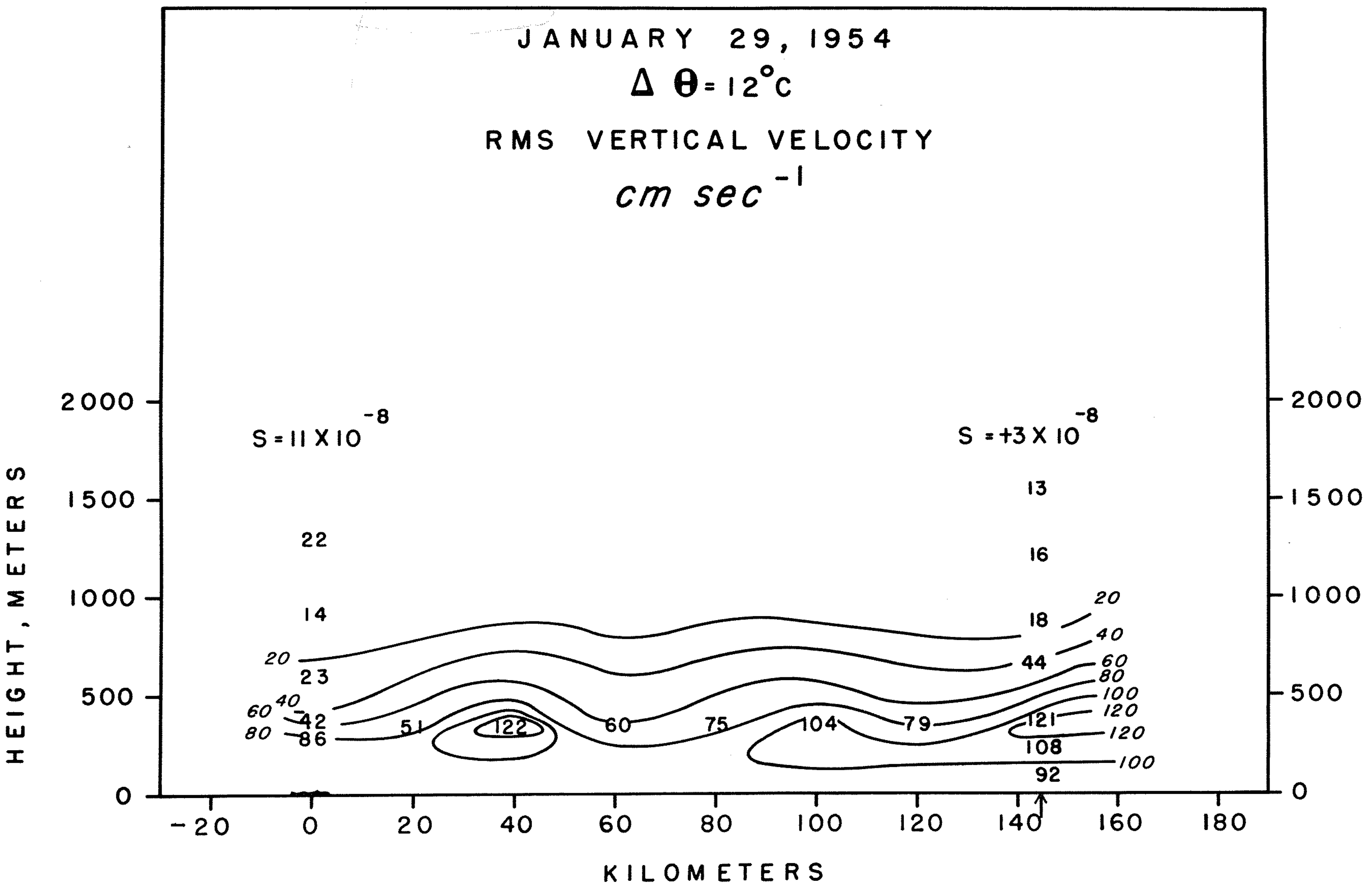
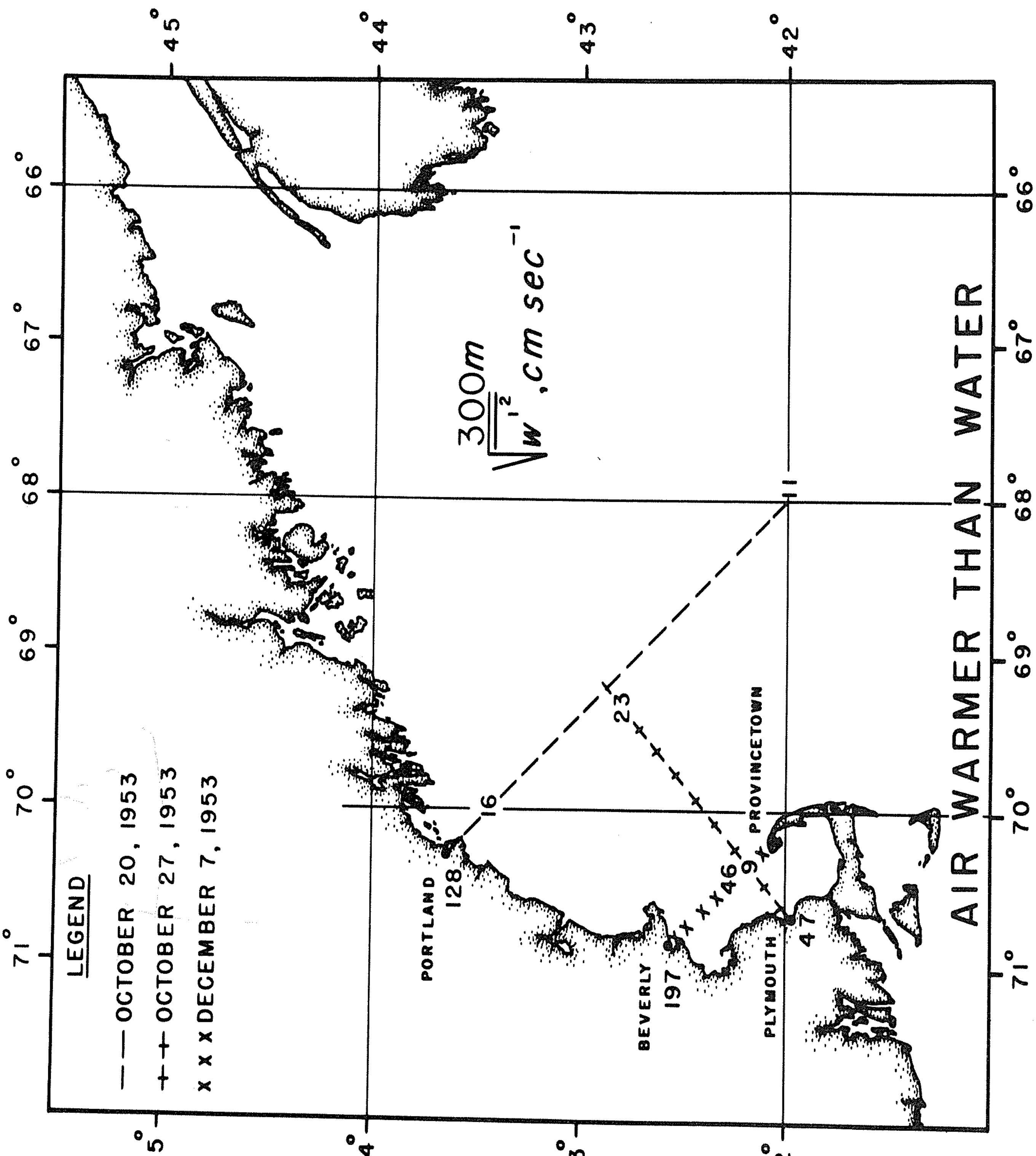

JANUARY 29, 1954
Δ θ = 12°C
RMS VERTICAL VELOCITY
cm sec⁻¹
S = 11 X 10⁻⁸
S = +3 X 10⁻⁸
HEIGHT, METERS
KILOMETERS
22
14
23
42
86
51
122
60
75
104
79
121
108
92
13
16
18
44
20
40
60
80
100
120
0
500
1000
1500
2000
-20
140
160
180

FIG.16



LEGEND
— — OCTOBER 20, 1953
+ + OCTOBER 27, 1953
X X X DECEMBER 7, 1953
$\frac{300m}{\sqrt{w'^2}}$, cm sec^{-1}
AIR WARMER THAN WATER
PORTLAND
BEVERLY
PLYMOUTH
PROVINCETOWN
128
16
23
11
197
46
9
47
45°
44°
43°
42°
71°
70°
69°
68°
67°
66°

FIG. 17

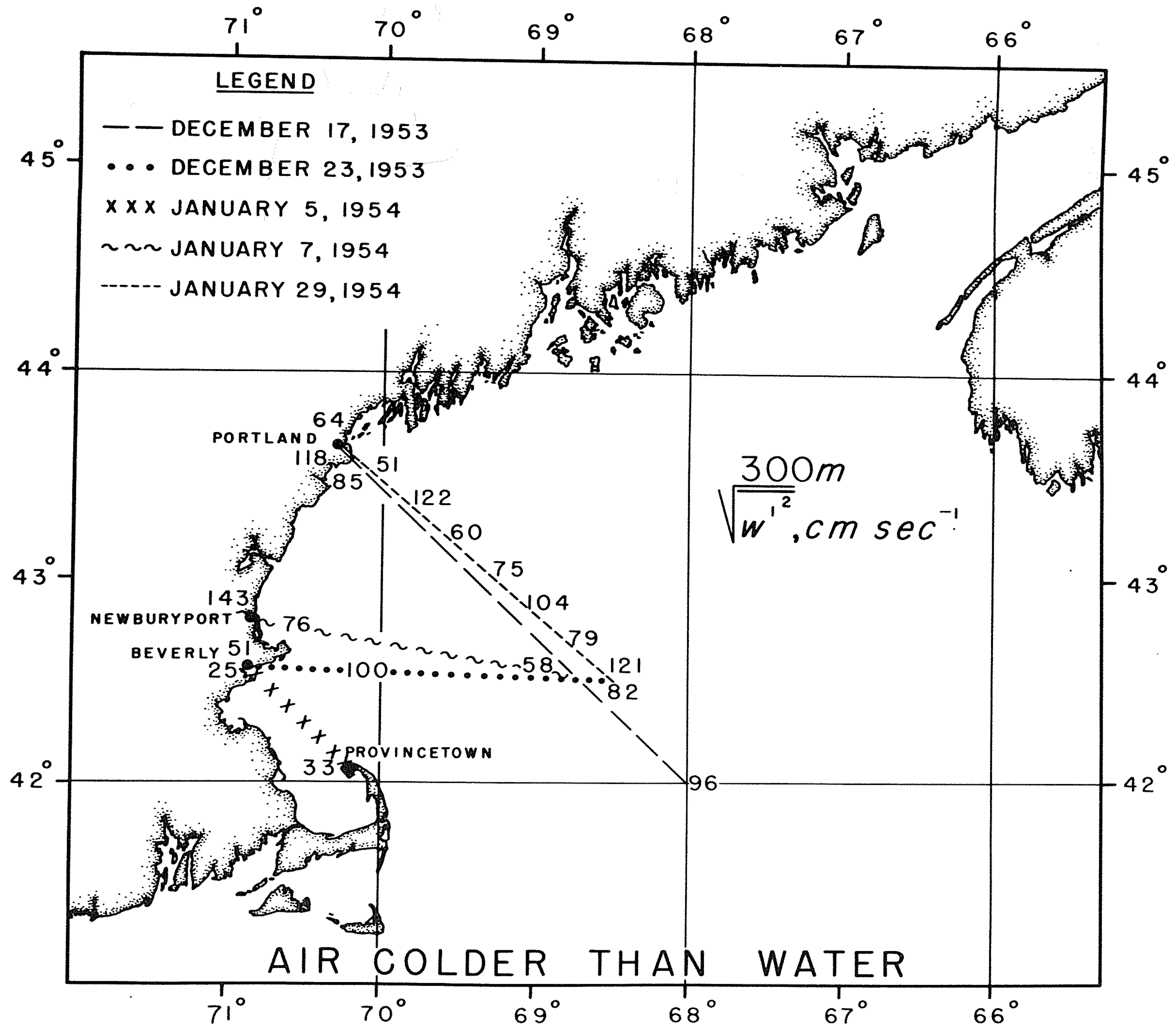

LEGEND
DECEMBER 17, 1953
DECEMBER 23, 1953
XXX JANUARY 5, 1954
~~~ JANUARY 7, 1954
JANUARY 29, 1954
300m
$\sqrt{\overline{w'^2}}$, cm sec^{-1}
AIR COLDER THAN WATER
PORTLAND
NEWBURYPORT
BEVERLY
PROVINCETOWN
64
118
51
85
122
60
75
104
79
121
82
96
143
76
58
51
25
100
33
71°
70°
69°
68°
67°
66°
45°
44°
43°
42°

FIG.18
~~~

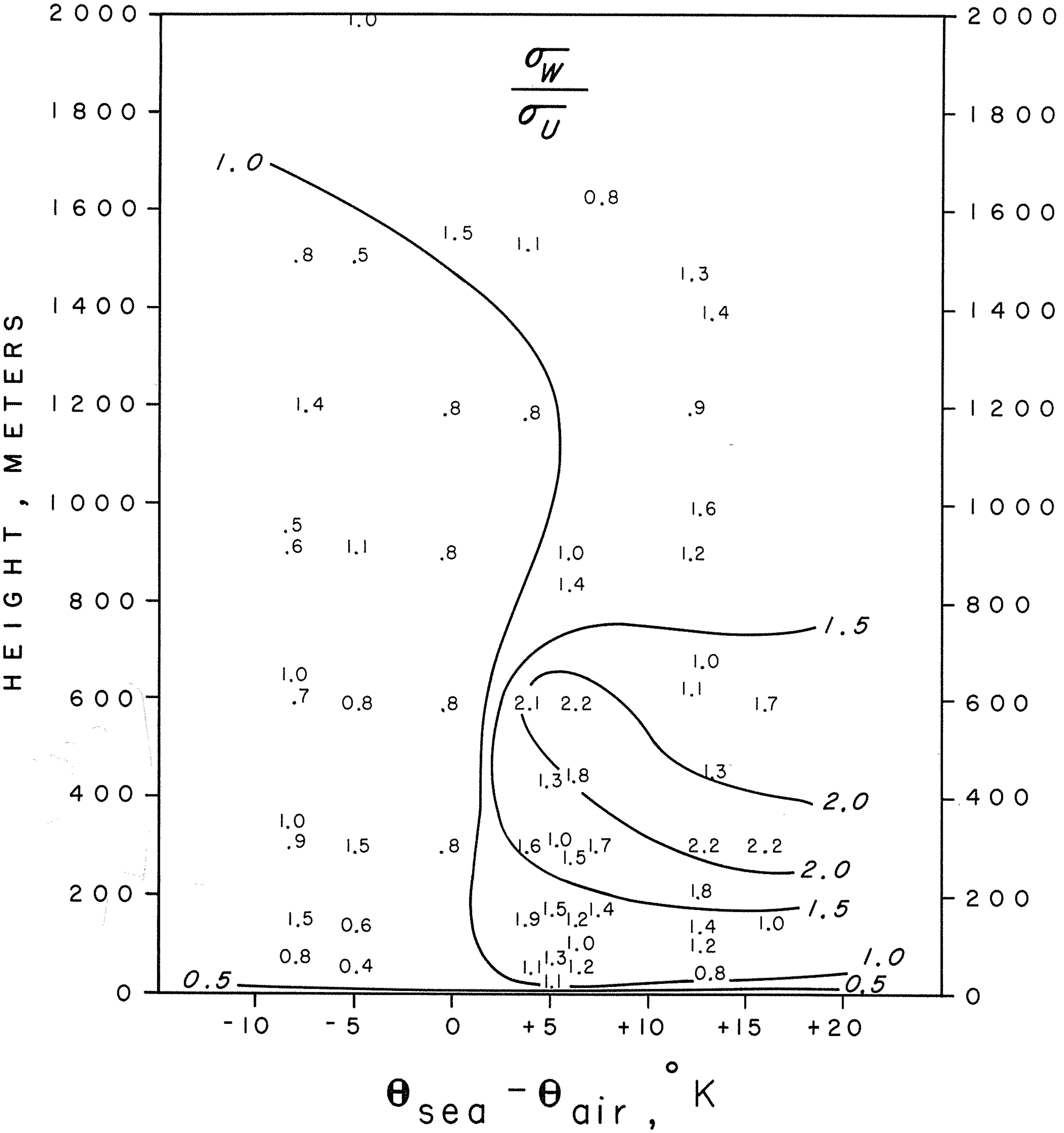
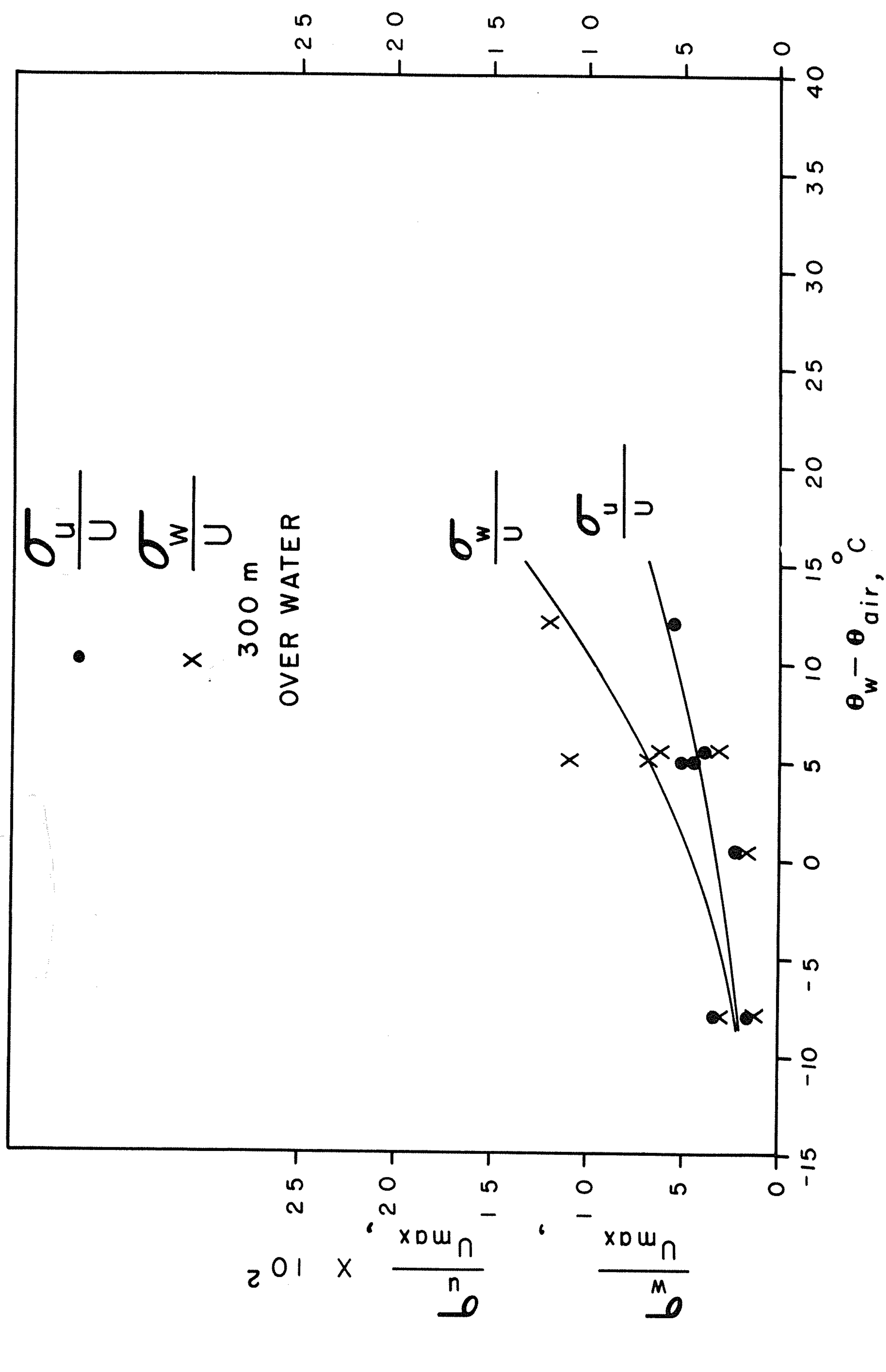


FIG. 19



$\frac{\sigma_u}{U}$
$\frac{\sigma_w}{U}$
300 m
OVER WATER
$\frac{\sigma_w}{U}$
$\frac{\sigma_u}{U}$
$\theta_w - \theta_{air}$, °C
$\frac{\sigma_u}{U_{max}}$, $\frac{\sigma_w}{U_{max}}$, X 10^2
-15
-10
-5
0
5
10
15
20
25
30
35
40
0
5
10
15
20
25

FIG. 20

24 August 1953

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