Water Vapor Distribution in the Stratosphere and High Troposphere[1,2]

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(Manuscript received 27 June 1967, in revised form 6 November 1967)

ABSTRACT

Fifty-one soundings with balloon-borne frost-point hygrometers provided measurements to a height of 94,000 ft of the vertical distribution of water vapor over Trinidad, West Indies, Washington, D. C., and Thule, Greenland, during 1964 and 1965, the International Years of the Quiet Sun. The stratospheric measurements were obtained during the balloon descent with sample collection ahead of the instrument train to avoid extraneous moisture from the flight system.

The observed mixing ratios of the lower stratosphere to a height of 73,000 ft are for nearly all cases within the range of 1.2×10^{-6}–3.3×10^{-6}. Above 73,000 ft the range broadens to include higher mixing ratios, but the majority of cases fall within the same narrow range as for the lower stratosphere. The vertical distribution of mixing ratios for most purposes is better represented by the medians of the distributions by level than by the arithmetic means. The median vertical distribution of stratospheric mixing ratio to a height of 94,000 ft for Washington, D. C., and for Trinidad is within the mixing ratio range of 2×10^{-6}–3×10^{-6}.

A seasonal variation of mixing ratio for the low stratosphere is evident for both Washington, D. C., and Trinidad, with lowest mixing ratios in late winter and spring and highest in late summer and fall. The magnitude of the seasonal change decreases with height so that a seasonal reversal of the vertical gradient of mixing ratio for the lower stratosphere is also observed. A systematic latitudinal gradient of mixing ratio in the stratosphere is not discernible in the data.

1. Introduction

The first measurements of stratospheric water vapor were made by the British in 1953 using a manually operated frost-point hygrometer on an aircraft. The instrument and measurements are described by Dobson *et al.* (1946). The humidity was found to decrease rapidly with height above the tropopause and mixing ratios as low as 2×10^{-6} were observed. The aircraft soundings continued for many years thereafter and by the end of 1955, a total of 339 ascents had been made over England. The measurements from these flights, referred to as the MRF (Meteorological Research Flight) data, are summarized and discussed by Tucker (1957). The early aircraft ascents seldom exceeded 40,000 ft, and provided data only for the lowest stratospheric levels. In 1954 and 1955 the use of a Canberra aircraft extended the soundings to 50,000 ft. More than 60 soundings at the higher altitudes show a mixing ratio of 2×10^{-6} at the 150-mb level, and 1.9×10^{-6} at 125 mb. These two highest levels show the vertical distribution of mixing ratio approaching a constant value of about 2×10^{-6} in the stratosphere.

The first measurements to higher altitudes were made in 1949 by Barrett *et al.* (1950), using an automatically controlled frost-point hygrometer carried on a large plastic balloon. They describe three soundings which reached an altitude of 100,000 ft and show mixing ratios in the stratosphere which are an order of magnitude higher than the MRF measurements.

The next high altitude measurements came nine years later when Barclay *et al.* (1960) used a balloon-borne vapor trap to freeze out water vapor and carbon dioxide from an air sample collected at 27 km. They found a vapor-to-air mixing ratio at this level of 3.7×10^{-5}. Several more measurements over the next two years were consistent with the earlier observation, showing the mixing ratio at 27 km to be about 4×10^{-5} (Brown *et al.*, 1961).

In 1957 and 1958, Houghton and Seeley (1960), using an infrared solar spectrometer mounted on an aircraft, measured the atmospheric absorption in the 2.7-μ water-vapor band above an altitude of 45,000 ft. With the assumption of a uniform mixing ratio in the path between the aircraft and the sun, they obtained a mixing ratio for the stratosphere of 3×10^{-6}. The observations do not define a unique distribution, and Houghton cites a mixing ratio distribution of 2×10^{-6} for the region 150–31 mb and 40×10^{-6} above 31 mb as also consistent with the absorption data (Houghton and Seeley, 1961). The same equipment was flown in 1959 to obtain high resolution spectra in the 6.3-μ band (Houghton *et al.*, 1961). Houghton (1963) attempted to infer a vertical distribution for water vapor by examining the weak lines for which the pressure broadening effect is small,

[1] Field research supported by the Atmospheric Science Program, National Science Foundation, NSF Grant No. NSF-GP876.

[2] Instrument development supported by Naval Air Systems Command (AIR-540).

and the sensitivity to water vapor at low pressures is high. Assumed distributions were compared with observations between 30,000 and 45,000 ft, and the best fit was found for a low mixing ratio of 1.5×10^{-6} in the lower stratosphere and a much higher mixing ratio (up to 5×10^{-5}) above 25 mb.

Murcray *et al.* (1962) used a balloon-borne infrared solar spectrometer to determine the vertical distribution of water vapor to a height of 95,000 ft. Absorption in the 6.3-μ band measured at many levels allowed for the calculation of the average mixing ratio for 10,000-ft layers in the stratosphere. The vertical distribution of mixing ratio reported by Murcray *et al.* has a minimum of 7.5×10^{-6} in the lower stratosphere and increases to 7.5×10^{-4} above 95,000 ft.

The Japanese describe a frost-point radiosonde (Hayashi, 1961) which they used to measure the vertical distribution of water vapor over the Japanese Islands during the 2-yr period of the International Geophysical Years of 1959–1960. Over 150 soundings for this period (Japan Meteorological Agency, 1961, 1962) show a high water-vapor content for the stratosphere with an average mixing ratio of about 9×10^{-6} for the lower stratosphere increasing to 7×10^{-5} at 90,000 feet.

Mastenbrook and Dinger (1961) measured the vertical distribution of water vapor using a balloon-borne frost-point hygrometer which utilized the standard radiosonde for telemetry. Their sounding of April 1959 showed a high stratospheric mixing ratio in general agreement with Barrett *et al.* Two soundings in 1960 showed the lower stratosphere to be very dry, in agreement with the MRF measurements, and an increase in mixing ratio with height, agreeing closely with the vapor trap measurements at 90,000 ft.

By the early nineteen sixties the accumulated data supplied by all investigators suggested a model for the vertical distribution of water vapor in the stratosphere for which the mixing ratio reached a minimum in the low stratosphere and increased with altitude becoming an order of magnitude higher at 100,000 ft. A wide range in the data suggested considerable variability in the mixing ratio at all stratospheric levels. Inconsistencies were evident, however, in the comparison of observations from individual programs which cautioned against the modeling of the water distribution of the stratosphere at that time. One of the disturbing inconsistencies was the disparity between the two largest bodies of data, the MRF observations and those of the Japanese. In the lower stratosphere where the two sets of data can be compared, the MRF data show consistently low mixing ratios of about 2×10^{-6}, whereas the Japanese data show an average mixing ratio nearly an order of magnitude higher with considerable variability.

Mastenbrook (1965) considered the problem of water-vapor contamination from balloon-borne systems and concluded that water vapor evolved from the instrument package at stratospheric levels might well account for the higher and more variable concentrations of water vapor measured from balloon platforms. He described two soundings with a frost-point hygrometer in 1962 and 1963 for which great care was taken to minimize the extraneous moisture entering the sample. The instrument was an improved model of the hygrometer used earlier by Mastenbrook and Dinger (1960, 1961). Contamination from the surfaces of the sensing cavity and inlet duct was minimized through the use of stainless steel and a high volume flow rate. The air was sampled at the lowest level of the flight train during a balloon descent to avoid contamination from external sources. The two soundings showed stratospheric mixing ratios of 1×10^{-6}–4×10^{-6} for all stratospheric levels to 90,000 ft. They were the first observations to show a vertical distribution of mixing ratio for the stratosphere which is essentially constant, and consistent with the large number of MRF observations at the base of the stratosphere.

In 1964 and 1965, 51 descent balloon soundings with the frost-point hygrometer were made under the auspices of the IQSY (International Years of the Quiet Sun) (Mastenbrook, 1966a). The program called for monthly observations at more than one latitude during the 2-yr period. Twenty-five soundings were made at Washington, D. C. (39N), and 23 at Trinidad, West Indies (11N). Three soundings were made at Thule, Greenland (77N), during the summer of 1965 during a temporary cessation of operations at Trinidad. The water-vapor distribution in the stratosphere is here reexamined in the light of this new body of data.

2. Instrumentation

The frost-point hygrometer when used as a sounding instrument provides a vertical distribution of water vapor in terms of the frost-point temperature. From the known relation of frost-point temperature and partial pressure of water vapor, the concentration of water vapor is determined. The frost-point sonde used in the collection of the IQSY data is an improved model of the sonde used earlier and described elsewhere (Mastenbrook and Dinger, 1960). In operation, the instrument maintains a condensate on a polished mirror which is interposed between an induction heater and a heat sink of freely-boiling Freon 13. An optical detection system senses the scattering of light directed upon the mirror and controls the input to the heater to maintain the size of condensate essentially constant, thereby maintaining the mirror temperature at or about the frost-point temperature. A thermistor embedded in the surface of the mirror senses this temperature which is telemetered as a modulation of the standard radiosonde transmission.

The suitability of Freon 13 as the coolant at stratospheric levels has been determined by measuring the minimum mirror temperature in the absence of control heating for several stratospheric flights. Observed minima of −99C at 50,000 ft and −114C at 100,000 ft provide a reserve of cooling of 19C in this altitude range when the mirror is at a frost-point temperature corresponding to a mixing ratio of 2×10^{-6}.

Some investigators have questioned whether an optical frost-point hygrometer can detect ice collected at temperatures <-90C. Brewer *et al.* (1948) describe experiments in which ice collected on a mirror cooled below −90C was not detected until the mirror was warmed to a higher temperature. He concluded that the condensate formed as glossy ice and did not become visible until crystallization occurred at a higher temperature. The IQSY instrumentation was subjected to repeated laboratory tests over the range of stratospheric pressures when ice was collected at mirror temperatures between −90C and −110C. A white deposit was observed to form in all cases which was readily visible to the eye. Present sounding procedure, adopted after the IQSY program, includes a partial removal of condensate from the mirror at the 40-mb level during the balloon ascent by means of a 3-sec heating impulse. A small reduction of condensate size drops the temperature of the mirror to the heat sink temperature of −110C, at which temperature new condensate forms. Approximately 7 min is required to collect the amount of condensate necessary to restore control. The new deposit forms a finer grain structure which improves instrument response for the subsequent descent sounding. The procedure has demonstrated that ice collected in the stratosphere at mirror temperatures well below −90C is detected with this instrumentation.

The measurement of very low frost-point temperatures is limited by the response time of the instrument which depends not only on the sensitivity of the optical detection system but also upon temperature and optical characteristics of the condensate. Since the distribution and character of condensate differs for each flight, the time response cannot be uniquely defined as an instrument function but can only be given as an approximate figure. Typical response times to a change of frost-point temperature have been determined under simulated stratospheric conditions.

Step changes of humidity are not readily produced in the laboratory for the determination of response time. An alternative method was therefore employed consisting of a step change in the bias control to the mirror heater input of the hygrometer, to produce a step change of mirror temperature in a constant humidity environment. The return of mirror temperature to the frost-point temperature requires a minute change of condensate in a manner equivalent to the response of mirror temperature to a step change of frost-point temperature. The thermal response of the mirror at low frost points is many times more rapid than the optical response resulting from condensate change, so that the response time may be considered as determined by the

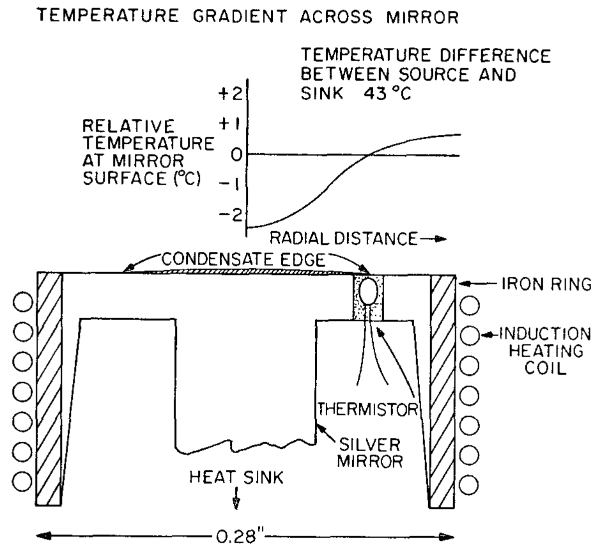

FIG. 1. A schematic representation of the frost-point hygrometer mirror and the radial departure of temperature from that measured by the thermistor.

efficiency of condensate collection or evolvement. The response time therefore increases with decreasing frost-point temperature and decreases with decreasing air pressure. The measured $(1-1/e)$ response time for a 3C offset to a lower temperature is 0.36 min at −85C frost point and 100 mb pressure, and 0.60 min at −90C frost point and 45 mb. The condensate size must also change in response to a change of heat sink temperature to maintain the same mirror temperature, since condensate size and brightness determine the heat input to the mirror. The same response considerations apply in this case as applied to the response due to frost-point change.

For an atmospheric sounding, it is the change of the difference between the heat-sink temperature and frost-point temperature which must be considered in determining the overall response. At stratospheric levels, the change of heat-sink temperature with height has been found to very closely parallel the frost-point temperature that corresponds to a constant mixing ratio. The response lag error is therefore minimal when sounding a stratosphere of near constant mixing ratio.

The largest single source of error inherent in the system results from the gradient of temperature across the face of the mirror. The condensate size is adjusted initially so that the condensate edge is at the same radial distance as the thermistor. A design feature of the instrument is a programmed clearing of the mirror at a mid-tropospheric level and a reforming of the condensate in the ice phase. Since no two condensates are identical in structure, brightness and distribution, it cannot be assumed that the controlling edge of the reformed condensate will be at the same radial distance as before. A partial instrument failure might also alter the condensate size without giving evidence of malfunction. The

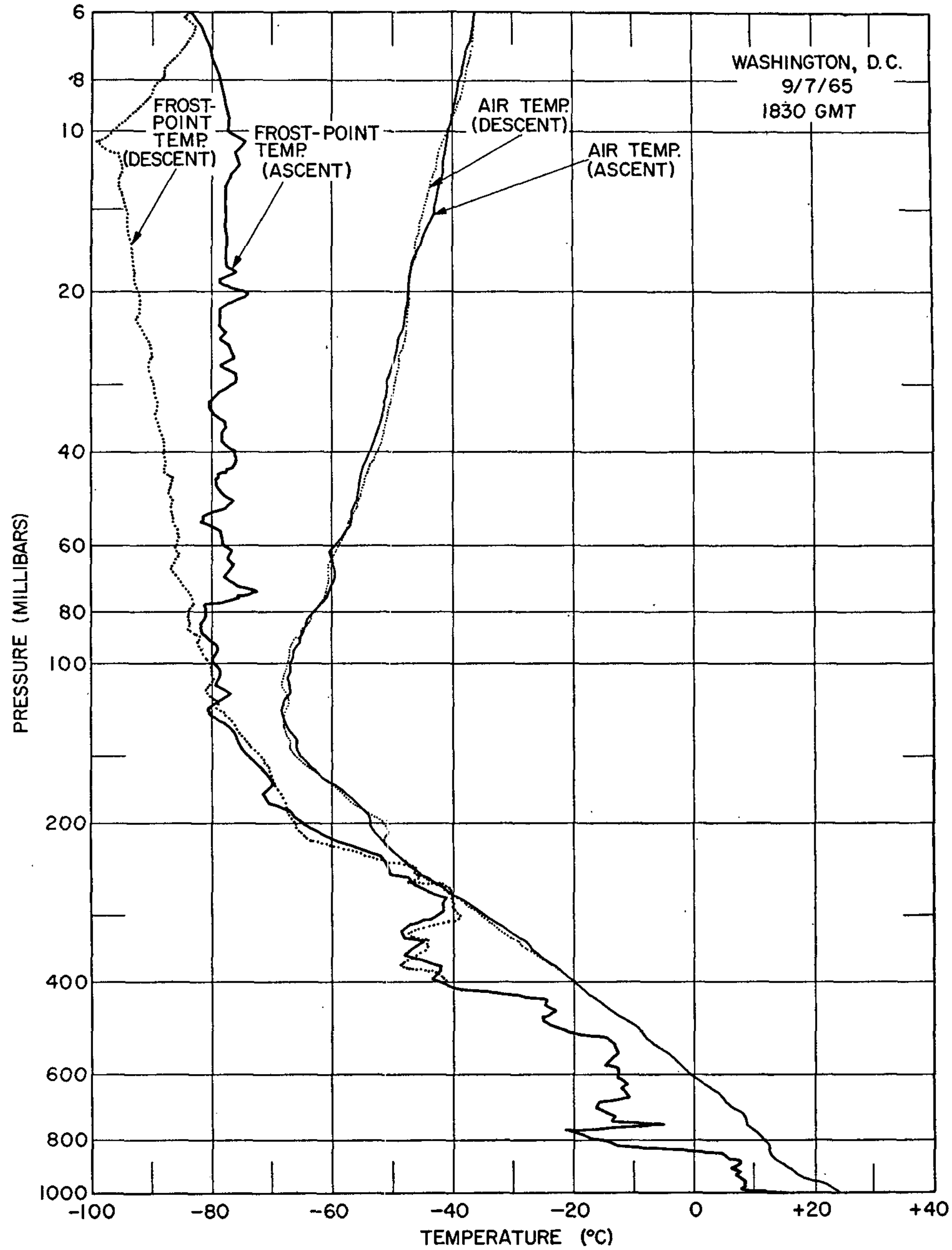

FIG. 2. A typical ascent-descent water-vapor sounding. The descent after turn-around adjustment is taken as the measure of stratospheric moisture.

gradient distribution has been examined (Fig. 1) and the limits of error determined as +2.5C and −0.7C. In the absence of an instrument malfunction, the gradient error will generally not exceed 1C.

The instrumental error of measurement at stratospheric levels is of less concern than the problem of extraneous water vapor in the sample. The design of the instrumentation and sounding procedures have been directed toward minimizing the level of extraneous moisture in the sample. The low frost points measured in the stratosphere indicate that these efforts have to a large degree been successful. The threat of water-vapor contamination, however, is ever present like a bad genie and may appear through a leak, a blower failure, a cloud, or in other obscure corners to give an anomalously high moisture indication for the stratosphere. Caution must be exercised in the interpretation of a single sounding for this reason, and it is considered good practice to seek verification of unusual features of a sounding with a second flight. This was not always possible, however, for the observations reported here.

The turn-around of the balloon, which initiates the

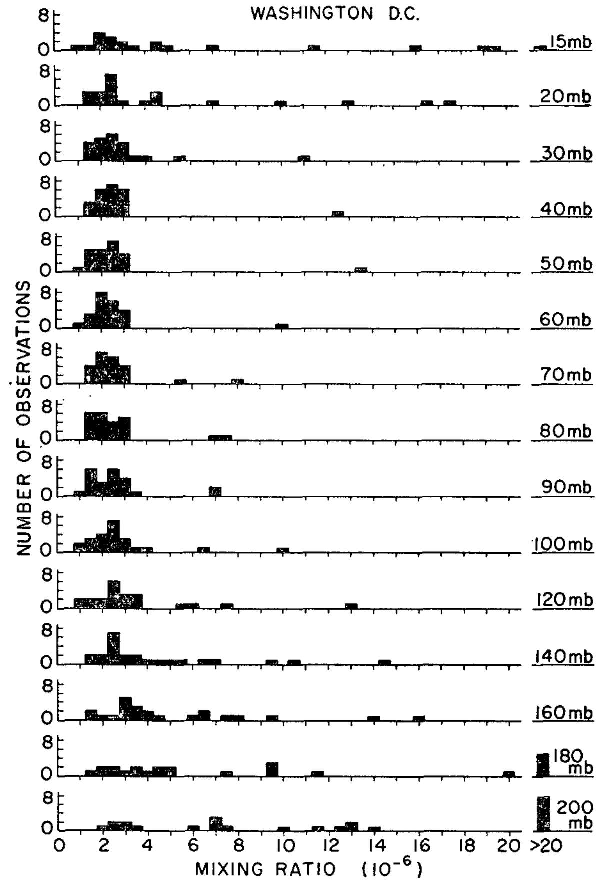

FIG. 3. Frequency distribution of mixing ratio for selected pressure levels over Washington, D. C., for the years 1964 and 1965.

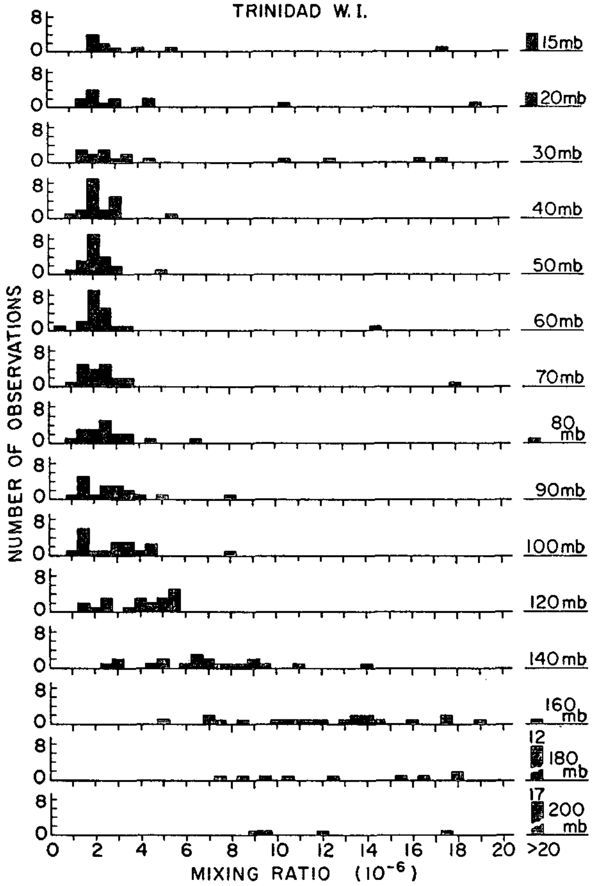

FIG. 4. Frequency distribution of mixing ratio for selected pressure levels over Trinidad, W. I., for the years 1964 and 1965.

descent flight from which the stratospheric observations are taken, is achieved by means of a balloon control system (Mastenbrook, 1966b) which valves gas from the balloon at a predetermined altitude. During the turn-around, the air sample changes from relatively moist air collected in the contaminated wake of the flight train to dryer air collected ahead of the flight train. Also associated with turn-around is a change in ventilation rate due to an added ram pressure on the inlet duct during descent which increases the flow rate. The location of the inlet duct beneath the instrument is for the purpose of favoring the descent flight for the sampling of the stratosphere. The transition from moist sample to dry sample occurs over a period of several minutes as the balloon accelerates to a descent velocity of about 1800 ft min^{-1} and the inlet duct moves ahead of the moisture emanating from the instrument package. The transition is characteristically accompanied by a decrease of the observed frost-point temperature as the contaminant contribution decreases. When the decrease in frost-point temperature occurs quickly so as to relate closely in time with balloon turn-around, data for this portion of the record are discarded. When the drop in observed frost-point temperature is not well defined, or protracted in time, the data are kept in order not to exclude observations which may reflect a real increase in moisture at the upper levels. A typical water-vapor sounding is that of 7 September 1965 (Fig. 2). The descent flight starting at the 12-mb level is considered indicative of the stratospheric water-vapor distribution.

3. Results and discussions

Frequency distributions of mixing ratio. The two years of IQSY observations at Washington, D. C., and Trinidad, W. I., are first examined as distributions of mixing ratio at selected stratospheric and high tropospheric levels. Data were extracted from the sounding record at half-minute intervals and the mixing ratios for the levels determined by averaging six datum levels, corresponding to a height interval of approximately 3000 ft. The distributions appear in Figs. 3 and 4.

An outstanding feature of the distributions for Washington, D. C., is the concentration of observations in the mixing ratio range of 1.2–3.3 parts per million for the levels 100 through 40 mb. The cases of substantially

TABLE 1. The median distribution of mixing ratio for the stratosphere and high troposphere during 1964 and 1965 at Washington, D. C., and Trinidad, W. I.

Level (mb)	Median mixing ratio ($\times 10^{-6}$)* Washington, D. C.		Trinidad, W. I.		Both	
15	3.0	(21)	3.3	(14)	3.2	(35)
20	2.6	(23)	3.0	(16)	2.7	(39)
25	2.5	(23)	2.8	(16)	2.6	(39)
30	2.5	(23)	2.8	(16)	2.6	(39)
40	2.4	(23)	2.1	(20)	2.3	(43)
50	2.3	(23)	2.1	(20)	2.2	(43)
60	2.2	(23)	2.1	(20)	2.2	(43)
70	2.3	(23)	2.3	(20)	2.3	(43)
80	2.2	(23)	2.5	(19)	2.4	(42)
90	2.4	(23)	2.6	(18)	2.4	(41)
100	2.4	(23)	2.8	(19)	2.5	(42)
120	2.7	(22)	4.3	(20)	3.2	(42)
140	3.1	(25)	6.8	(20)	5.1	(43)
160	3.7	(23)	13.0	(21)	8.0	(45)
180	5.0	(24)	24.0	(21)	11.0	(45)
200	11.5	(25)	33.0	(21)	24.2	(46)

* Number of cases in parentheses.

higher mixing ratio are so few as to be regarded as anomalous occurrences. Above the 40-mb level the distributions show more scatter toward higher mixing ratio but with a majority of cases in a narrow range about a value of 2.5×10^{-6}. The predominant observation of low mixing ratios at the higher levels is a characteristic not evident in the observations of the early 1960's and may well reflect the greater care taken to avoid water-vapor contamination in the later measurements. Below the 100-mb level and extending downward to the 200-mb level, the distributions broaden toward higher mixing ratio while retaining the feature of a concentration at the low end of the range.

The distribution for Trinidad does not differ substantially from the Washington, D. C., distribution for the intermediate stratospheric level of 50 and 60 mb. Progressing downward by levels to 120 mb, the distributions appear progressively more bimodal which will

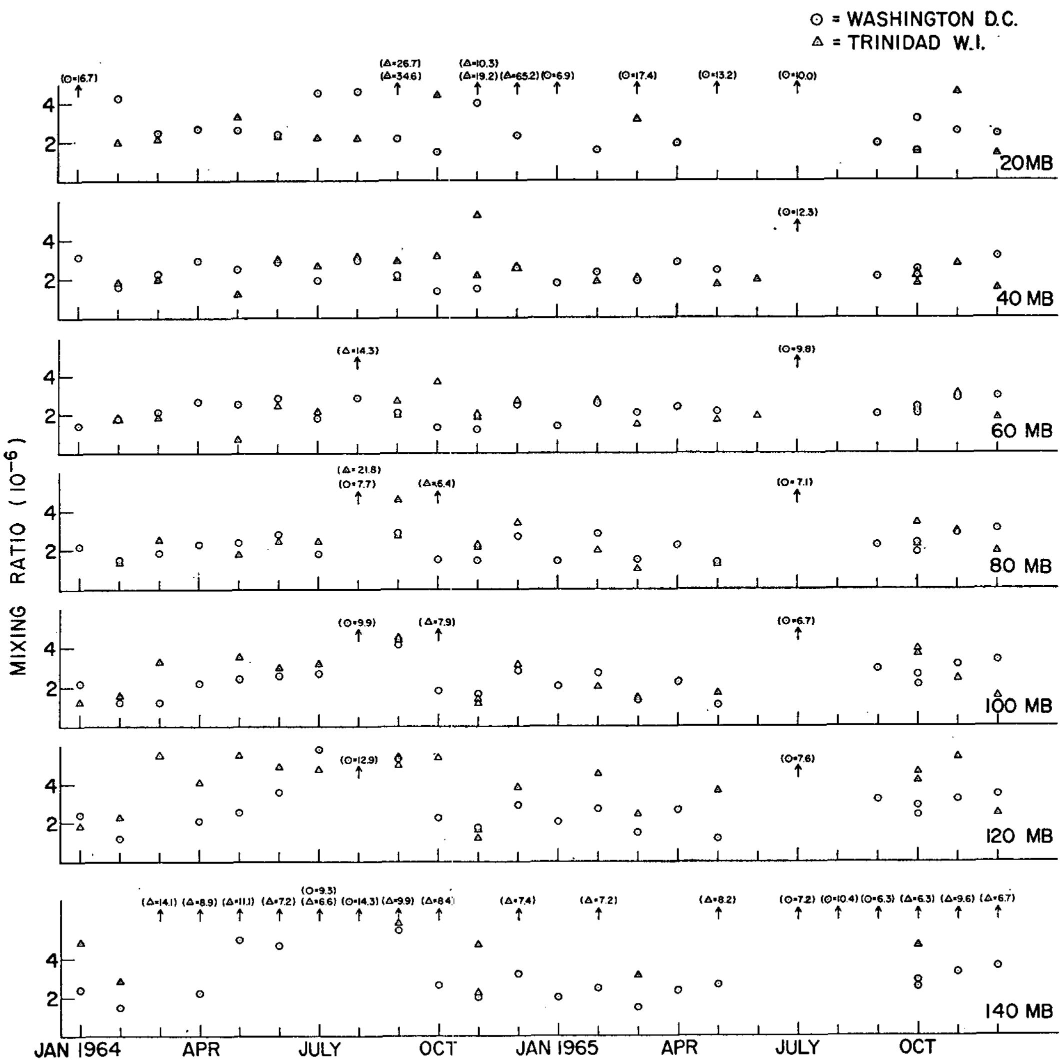

FIG. 5. Monthly mixing ratios for Washington, D. C., and Trinidad, W. I., for the years 1964 and 1965. Each symbol represents data for a selected level from a single sounding.

later be shown to be due to a seasonal variability. Most of the observations at 140 mb and below were within the troposphere and the mixing ratios are much higher than for corresponding levels at Washington. At the two highest levels, the more frequent occurrence of high mixing ratios may be the consequence of a humid environment at the launch site and a more frequent occurrence of high cirrus clouds, factors contributing to the contamination problem.

In Table 1, the vertical distribution of mixing ratio is presented in terms of the medians of the layer distributions. The medians were selected to represent the vertical distribution in preference to the arithmetic means for the reason that the medians more closely represent the main body of data. The arithmetic means, on the other hand, fall outside of the main body of data, giving a distribution which was seldom approximated in the two years of observations.

Temporal and latitudinal distribution of mixing ratio. In Fig. 5 the two years of observations appear as a time distribution of mixing ratio by selected levels. The 2-yr sequence of data may be examined for evidence of time and space variability, bearing in mind that the individual observations which make up the distribution may not constitute a good representation of the monthly means. Examining first the 140-mb level, the data clearly divide seasonally, with lowest mixing ratios in winter and highest in summer, with higher values at Trinidad for both seasons. At 120 mb the seasonal trend is the same, except that the minimum of mixing ratio appears somewhat later in the winter at Washington, D. C., than at Trinidad. Moving upward by successive layers, the amplitude of the seasonal variation decreases with height and is almost indiscernible at 40 mb. A persistent or seasonal mixing-ratio difference between the two stations at these levels is not apparent.

Corresponding features in the monthly distributions for the two stations suggest a scale of variability which can produce similar effects at the two stations within the space of a month. For example, the low mixing ratios for November 1964 at 100 and 120 mb, which is not consistent with the seasonal trend, is observed at both stations within a period of ten days. Another case in point is the abnormally high mixing ratio for the lower stratosphere of both stations during the months of August and September of 1964 (Fig. 6).

Vertical gradients. Since the amplitude of the seasonal oscillation decreases with height, a seasonal reversal in the vertical gradient of mixing ratio for the lower stratosphere might be expected. To examine the vertical gradient, mixing ratios were determined for two levels, 90 and 50 mb ,by averaging 12 datum levels corresponding to 6000 ft about each layer. The difference for the two levels provides a mean gradient for the 13,000-ft height interval (Fig. 7). A seasonal reversal appears in the Washington, D. C., distribution with a downward gradient in late winter and spring and an upward gradient in late summer and fall extending into early winter. The July 1965 gradient was derived from a sounding of anomalously high stratospheric moisture, which appears to result in an anomalous gradient as well. At Trinidad, a seasonal reversal of gradient is also evident; however, the gradients are generally of higher magnitude and the time distribution more variable.

The significance of the gradient pattern may be assessed in terms of the known sources of error. The gradi-

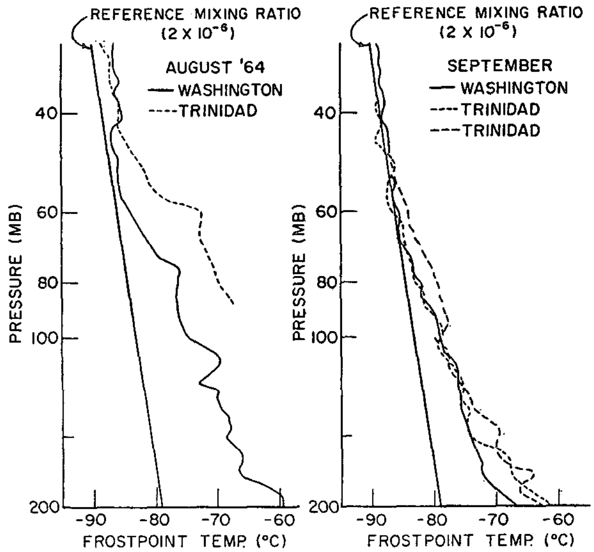

FIG. 6. Vertical distribution of frost-point temperature for the months of August and September at Washington, D. C., and Trinidad, W. I., showing abnormally high moisture levels for the low stratosphere for both stations.

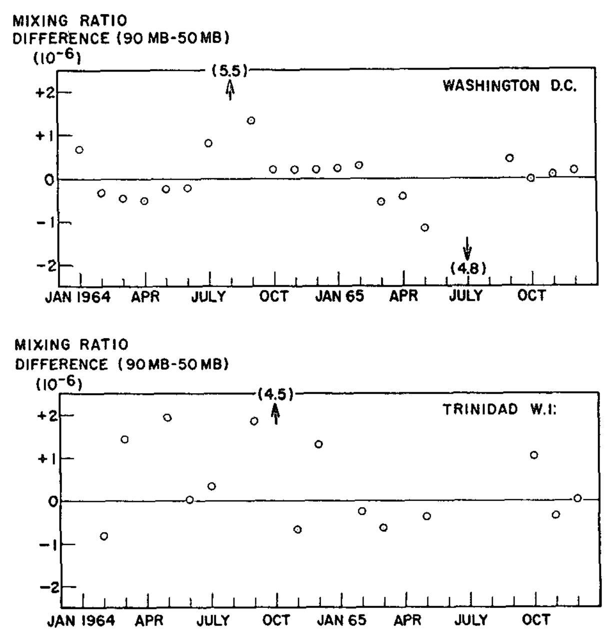

FIG. 7. The mean gradient of mixing ratio by months for the stratospheric layer 90–50 mb for Washington, D. C., and Trinidad, W. I., during 1964 and 1965.

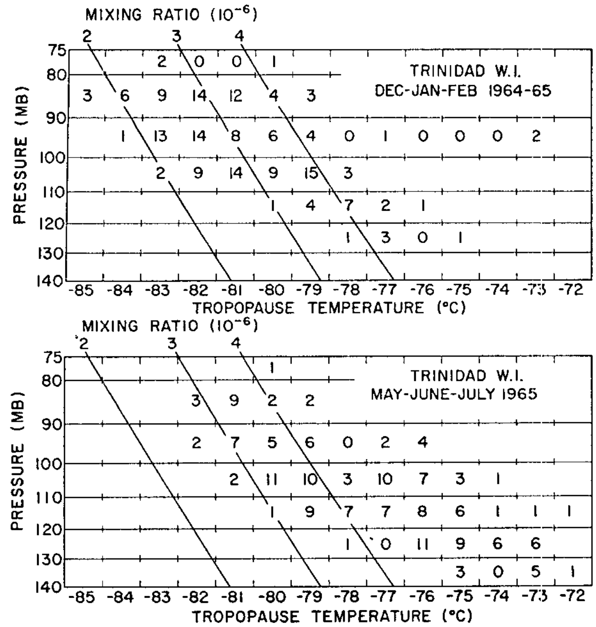

FIG. 8. The distribution of tropopause temperature and pressure at Trinidad, W. I., for three months in winter and three months in summer.

ent, determined as the difference between two means, is not affected by bias error. The error arising from the relative position of condensate and thermistor in the gradient field of the mirror remains essentially unchanged over short time intervals and may be ignored, providing no disruption of operation occurred as the instrument traversed the layers in question. Random errors of measurement and fine structure in the moisture profile appear in the deviations from the layer means. A standard deviation of the datum level mixing ratios from the layer means was determined for the eight Washington, D. C., soundings for which the mixing ratio difference between the two levels was less than 0.3×10^{-6} and the standard error of the difference was found to be 0.07×10^{-6}, which is approximately represented by the radius of the data circles of Fig. 7.

The question arises as to whether water-vapor contamination, affecting the two layers unequally and differing in magnitude seasonally, might account for the observed gradients and their seasonal variation. This question is examined in terms of the reasoned effect of water-vapor contamination upon the error of measurement. Since the soundings were made during the instrument descent with sampling ahead of the flight train, the concern is with the moisture evolved within the sensing cavity. The rate of evolvement can be expected to decrease as the instrument descends through the stratosphere, both because of the decreasing temperature of the stainless steel cavity and the depletion of the contamination. At the same time, the water-vapor density of the atmosphere, corresponding to a constant mixing-ratio distribution, increases with descent, amounting to an order of magnitude increase between the 10- and 100-mb levels. The error of mixing ratio due to the water-vapor contamination therefore decreases with decreasing altitude and produces a component in the observed vertical gradient of mixing ratio which is always directed downward, that is, a negative gradient. The error is proportional in magnitude to the amount of contamination. The mixing-ratio gradients determined for the lower stratosphere are positive in more than 50% of the cases. These positive gradients cannot reasonably be attributed to water-vapor contamination. Consider next a seasonal variation of the water-vapor contamination. Instruments prepared and launched at the Washington, D. C., site might be expected to exhibit more contamination at stratospheric levels in summer than winter because of higher absolute humidities at ground level in summer. This could result in a larger negative gradient error in summer than in winter. The observed gradient change is in the opposite sense, being negative in winter and spring and positive in summer and fall.

The seasonal reversal of gradient appears real on the basis of the error considerations. The reversal also is consistent with the seasonal pattern of mixing-ratio fluctuation for the low stratosphere which is best illustrated by the distributions at the three lowest levels of Fig. 5, which are below the 90-mb level used for the gradient analysis, but still within the stratosphere throughout the year at Washington, D. C. The greater variability of the gradient at Trinidad is not surprising, since the 90-mb level is frequently close to the tropopause and more directly affected by the fluctuation of the tropopause and the transport of moisture to high levels by cumulus clouds.

Seasonal change of temperature of the tropical tropopause. The seasonal height and temperature variation of the tropical tropopause is examined to determine if this variation could reasonably account for the seasonal change of mixing ratio. The distribution of tropopause temperature and pressure is shown (Fig. 8) for three winter and three summer months for Trinidad. The numbers represent the frequency of tropopause occurrence within the temperature and pressure intervals. The sloping lines are lines of constant saturation mixing ratio. Since the distributions are based upon only two soundings per day at fixed time intervals, it may be presumed that the daily minima of tropopause temperature would be represented by distributions somewhat displaced to lower temperatures. The minima of saturation mixing ratio which intersect the distributions are indicative of the ability of the cold region of the tropical tropopause to remove water vapor by precipitation to ice crystals.

Comparing winter and summer, it is seen that the winter fluctuation of the tropopause could lower the moisture content in the vicinity of the tropopause to less than two parts per million, while in the summer the removal capability is less by about one part per million.

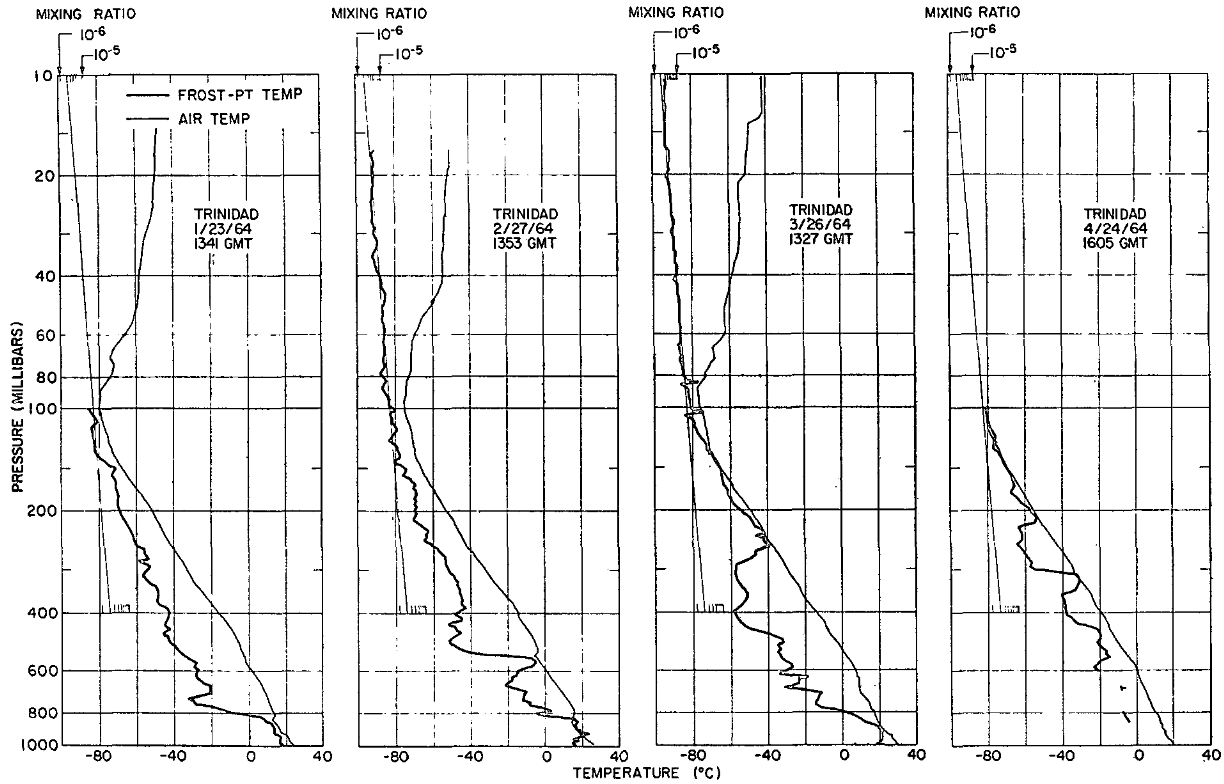

FIG. 9. Vertical distributions of frost-point temperature and air temperature for Trinidad, W. I. The 2×10^{-6} mixing-ratio curve provides convenient reference and appears on figures.

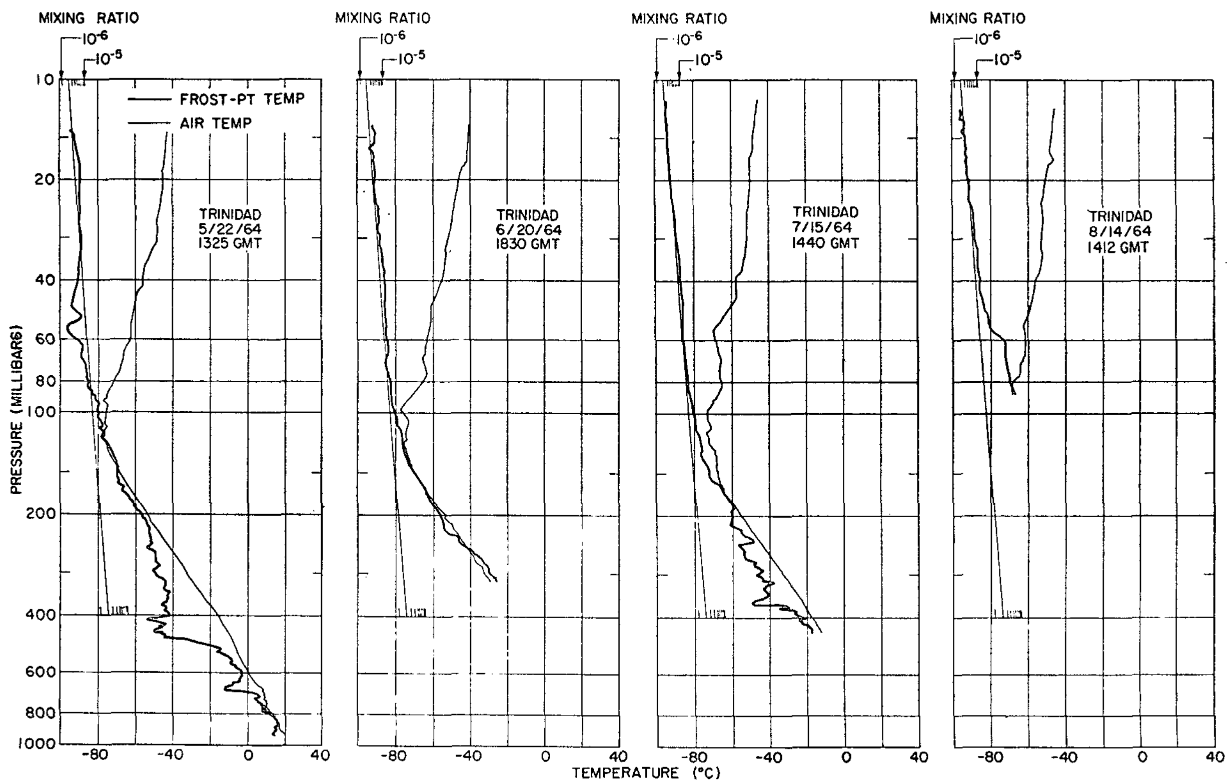

FIG. 10. Vertical distribution of frost-point temperature for Trinidad, W. I. (continued).

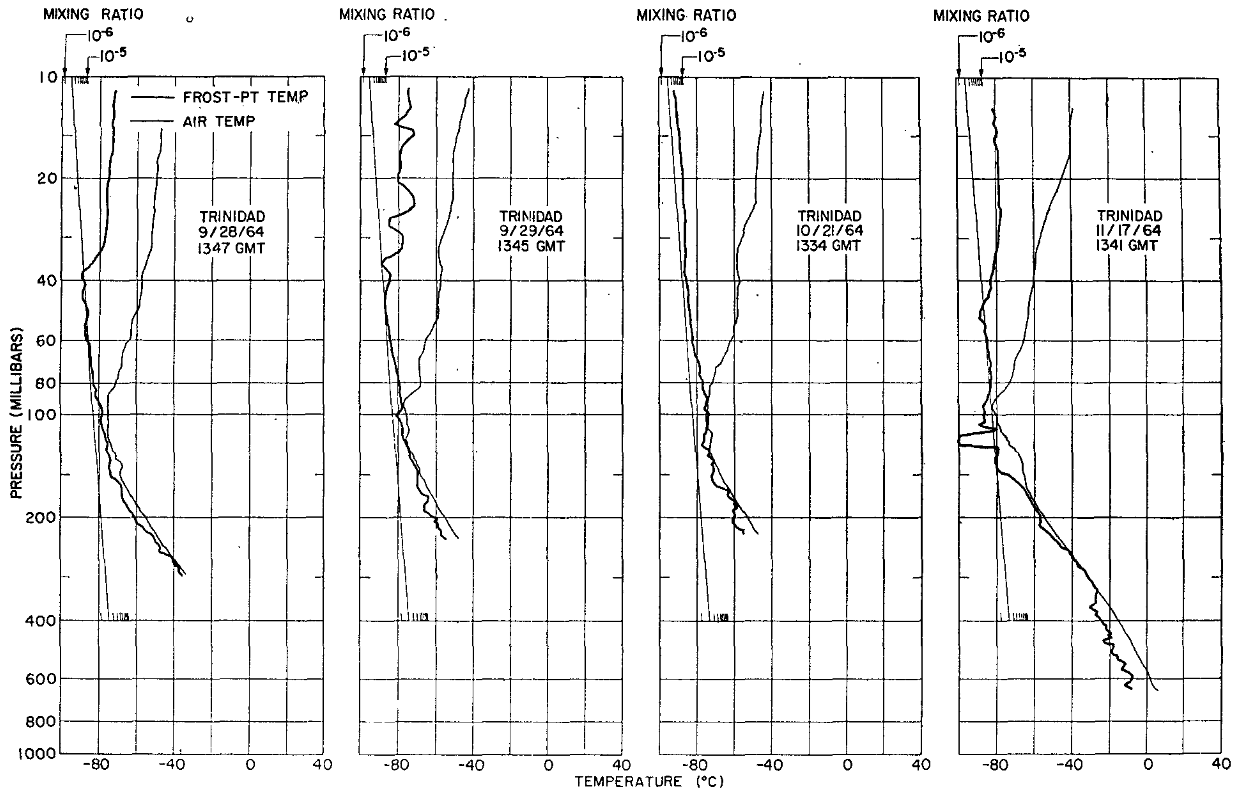

FIG. 11. Vertical distribution of frost-point temperature for Trinidad, W. I. (continued).

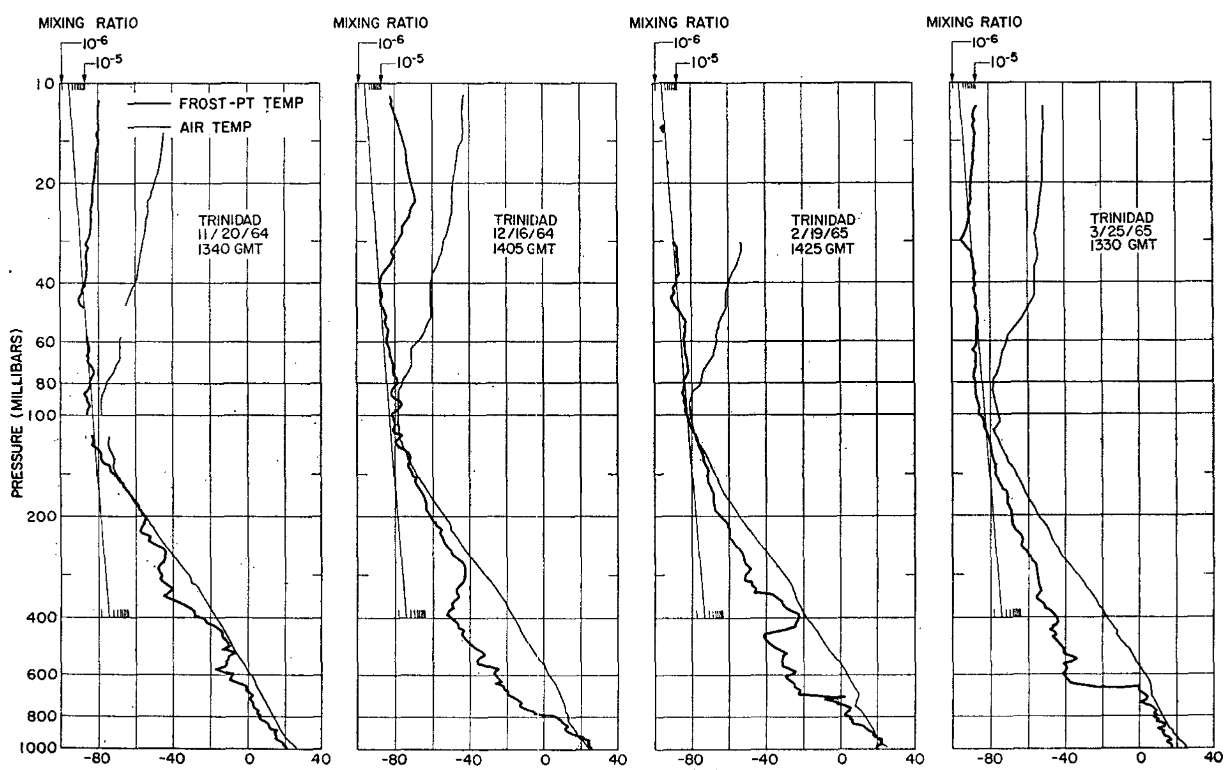

FIG. 12. Vertical distribution of frost-point temperature for Trinidad, W. I. (continued).

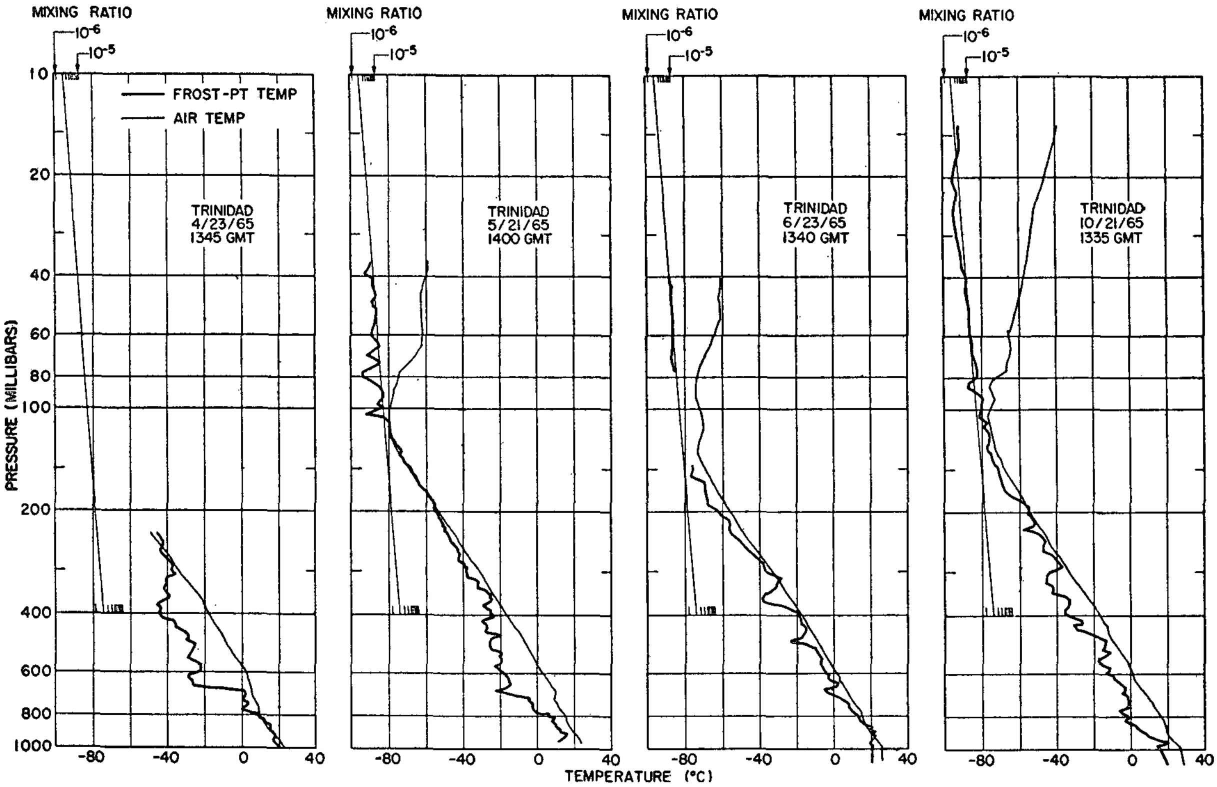

FIG. 13. Vertical distribution of frost-point temperature for Trinidad, W. I. (continued).

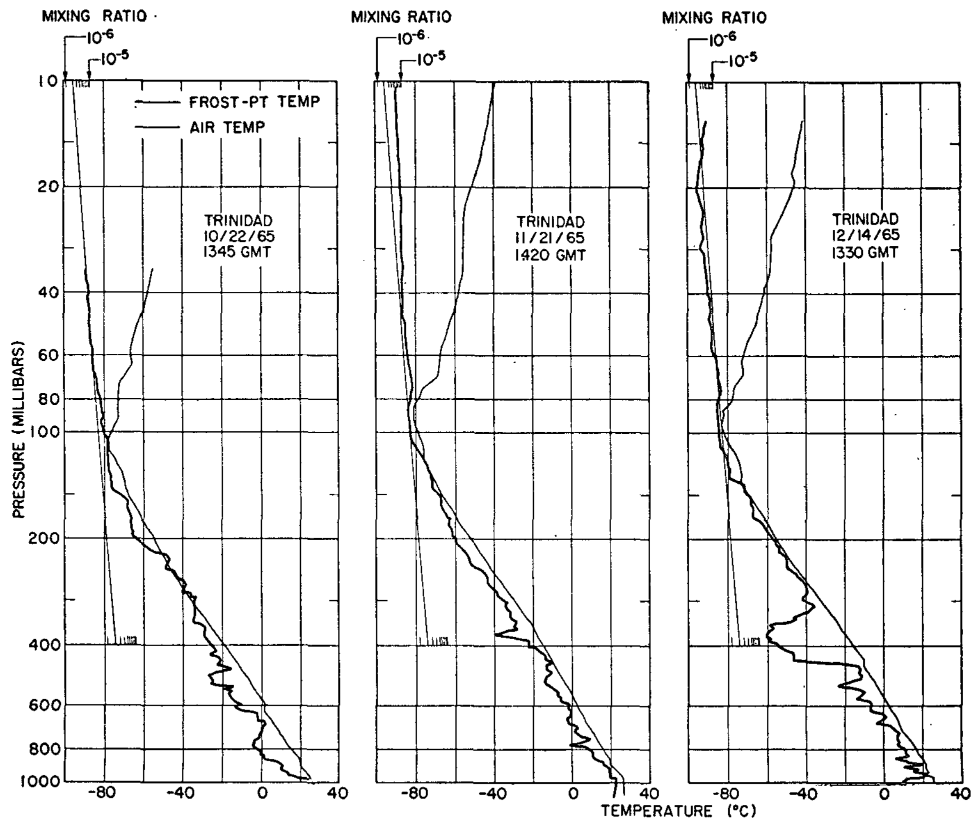

FIG. 14. Vertical distribution of frost-point temperature for Trinidad, W. I. (continued).

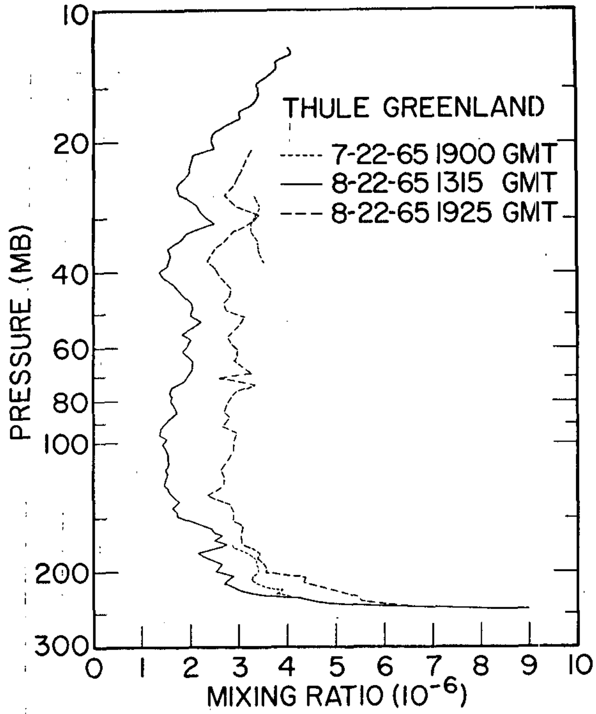

FIG. 15. Stratospheric mixing-ratio distributions over Thule, Greenland, during the summer of 1965.

The seasonal change in the moisture concentration at levels near the tropopause for both Washington, D. C., and Trinidad reflects this seasonal change in the removal capability of the tropical tropopause over Trinidad. At the 100-mb level (Fig. 5) 13 of 15 cases for which the mixing ratio was two parts per million or less, were observed in the winter half of the year. The similar patterns in the seasonal change in tropopause temperature at a tropical station and the seasonal change of mixing ratio for corresponding levels at a low and a middle latitude, is consistent with the hypothesis of stratospheric moisture control by the cold region of the tropical tropopause.

The role of temperature in the vicinity of the tropical tropopause in regulating the stratospheric moisture content is illustrated in the sequence of soundings at Trinidad (Figs. 9–14). The sequence starts in midwinter with a troposphere that is relatively dry. In February, a mixing-ratio minimum and inversion appears near 80 and 90 mb, the region of tropopause control. The 2×10^{-6} mixing-ratio line appears as a fine line on each chart for convenient reference. In March, the minimum is still evident above 80 mb, but the upper troposphere is now moist and the lowest stratospheric levels show an increase in moisture over February. In the course of summer the minimum of the lower stratosphere disappears and is replaced by a continuous upward gradient of mixing ratio. The moisture content of the lower stratosphere reaches a maximum by late summer. Through fall and into early winter the upper troposphere remains nearly saturated and the moisture content of the high troposphere and low stratosphere appears to be under the direct and nearly continuous control of local temperature. A new minimum appears near 100 mb in early winter which deepens and moves upward during winter, as the tropopause becomes colder and higher. The deepening minimum reverses the gradient of mixing ratio for the lower stratosphere, directing it downward. Sporadic data for spring and the absence of summer soundings break the continuity of data. The sequence resumes again in October when the lower stratosphere is again moist with an upward gradient of mixing ratio. The appearance of a new minimum near 100 mb in the early winter repeats the pattern of the previous year.

Water-vapor distribution over Thule, Greenland. In the months of July and August of 1965, three water-vapor soundings were made at Thule, Greenland (Fig. 15), providing the only data collected at high latitude in the course of the program. The tropopause levels and features are essentially the same for both months and the mixing-ratio transition from near saturation in the high troposphere to the low level of the stratosphere occurs within the few thousand feet of the tropopause inversion. The stratospheric mixing ratio is in the range of 1.5×10^{-6}–3.5×10^{-6} and is not significantly different from that of lower latitudes. The observations, though few in number, suggest that the stratospheric gradient of moisture between high and low latitudes in summer is very small. High latitude observations are not available for winter; however, many of the winter soundings at Washington, D. C., were north of the polar jet stream and well within the polar vortex circulation and may be considered to represent a high latitude distribution. These winter observations also do not indicate a higher moisture level for the high latitudes.

4. Summary and conclusions

Fifty-one water vapor soundings for Washington, D. C., and Trinidad, W. I., give a median vertical distribution of water vapor for the stratosphere to a height of 28 km which approximates a constant mixing ratio within the range of 2×10^{-6}–3×10^{-6}. These results obtained with a frost-point hygrometer agree well with the radiometer measurements over England of Williamson and Houghton (1965) which indicate stratospheric mixing ratios in the region of 3×10^{-6} at least to 25 km.

Calfee and Gates (1966) have calculated mixing ratios above 13.7 km over Florida using spectrographic absorption measurements from aircraft. Assuming the mixing ratio to be constant, they obtained an average mixing ratio of 3.2×10^{-6} for summer and 2.4×10^{-6} for winter, values which are consistent with the distributions reported here.

The vertical gradient in the lower stratosphere is found to reverse seasonally at both Washington, D. C., and Trinidad, W. I. The gradient is directed downward in late winter and spring and upward in late summer and fall. The seasonal change of gradient is almost entirely due to the seasonal change of moisture in the low stratosphere. The moisture content of the low stratosphere at both stations varies with the seasonal change of temperature of the tropical tropopause, supporting the hypothesis that the temperature of the tropical tropopause regulates the level of stratospheric moisture.

The two years of observations for low and middle latitudes and one summer of observations at high latitude did not provide evidence of a systematic latitudinal gradient of moisture within the stratosphere and suggest that this gradient, if it exists, must be of small magnitude.

Acknowledgments. The author wishes to express his gratitude to Messrs. Winton W. Werner and Robert E. Daniels for the capable and conscientious preparation of equipment and the conduct of field operations which contributed so greatly to the success of the program, and to Mr. Donald R. Purdy for computer programming and data processing. Special thanks are due the officers and men of the Naval Air Station, Trinidad, and the Thule Air Force Base, Greenland, for support services and facilities.

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