Tropospheric Water Vapor Climatology and Trends over North America: 1973–93

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ABSTRACT

Here 21 years of radiosonde observations from stations in the Western Hemisphere north of the equator were analyzed for trends in tropospheric water vapor. Mean fields of precipitable water and relative humidity at several levels are shown. Annual trends of surface–500 mb precipitable water were generally increasing over this region except over northeastern Canada. When trends were expressed as a percentage of the climatological mean at each station, the trends south of ~45°N represent a linear rate of increase of 3%–7% decade^{-1}. Trends in the upper portion of this layer, 700–500 mb, were as large or larger than those of the middle (850–700 mb) or lower layer and were consistent in sign.

Annual trends in dewpoint generally agree in sign with trends in temperature. However, the dewpoint trends tended to be larger than those of temperature. This was consistent with the annual increases found in relative humidity over this period. Relative humidity increased except in Canada, Alaska, and a few stations in western mountainous areas. Largest percentage increases of relative humidity were in the Tropics.

Seasonal trends of precipitable water varied spatially more than the annual trends and fewer were statistically significant. More stations had significant trends in summer than in other seasons and these were located over the central and eastern United States and the Tropics. Spring trends were largest over the western United States, while the largest winter trends were along the Gulf Coast. The one area where significant water vapor increases were found in all four seasons was the Caribbean.

1. Introduction

Water vapor plays a major role in the dynamics of atmospheric circulation as well as in radiation exchange within the atmosphere. The IPCC (1990) report stated that the general circulation models considered there all showed an increase in specific humidity along with the warming predicted in equilibrium CO_2 doubling simulations, and this increased moisture would further increase the warming. The characterization of the present water vapor distribution and its interannual changes is thus an important undertaking. The contribution of this study is the presentation of the mean water vapor distribution and any trends over North America during the period 1973–93.

The radiosonde network is the source of the longest record of humidity throughout the troposphere, but the data, particularly in early years, are affected by measurement inaccuracies. Although humidity sensors continue to improve, problems remain, especially in very high or very low humidities and at all humidities under very cold conditions (Elliott and Gaffen 1991). Spatially, the radiosonde network is most dense over the most populated Northern Hemisphere land masses.

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Several early studies produced estimates of the climatological water vapor distribution. For example, Crutcher and Meserve (1970) used Monthly Climatic Data of the World for the period 1950–64 to generate climatological maps of Northern Hemisphere temperature and dewpoint at various atmospheric levels. Other early estimates of water vapor in the atmosphere focused on the total column water, sometimes called the precipitable water. Climatologies of precipitable water over the United States were prepared by Reitan (1960) using data from 1946 to 1956 and Lott (1976) using data from 1946 to 1972. Bannon and Steele (1960) estimated precipitable water in several layers for nearly the entire globe using data from 1951 to 1955.

More recently, Oort (1983) published global distributions of specific humidity based on radiosonde data from 1963 to 1973. This gave a more detailed description of the vertical variability as 11 vertical levels were presented. Peixoto and Oort (1992) summarized the general features of the water vapor distribution using precipitable water, specific humidity, and relative humidity, again based on the Oort (1983) data.

Interannual variability of water vapor was considered by Angell et al. (1984), who examined records from two United States radiosonde stations. One of their findings was that the separation of instrument and procedural changes from climate changes had to be addressed. Other studies also estimated trends in tropospheric water vapor from radiosonde observations.

Hense et al. (1988) reported an upward trend of 700–500-mb precipitable water and relative humidity over the western Pacific using monthly data from 1965 to 1986. Gutzler (1992) also found increases above the western tropical Pacific during 1974–88. Elliott et al. (1991), using a limited number of stations distributed globally, found an increase in the surface–500-mb precipitable water in the equatorial region but also some significant increases over the western United States and slight decreases over northern Canada. Gaffen et al. (1991, 1992) also found increases, particularly in the Tropics, using a somewhat greater number of stations. A recent study of stratospheric water vapor trends (Oltmans and Hofmann 1995) found statistically significant increases between 16 and 24 km above Boulder, Colorado, of 0.4%–0.8% yr^{-1}.

This study differs from the previous work in several ways. It represents a more recent and longer time period (1973–93) than most previous climatologies. Earlier studies were based on data from older and less reliable instruments. A considerably denser network of stations is treated here and the issue of data homogeneity is emphasized, although each of the trend studies listed above gave some consideration to the problem of data inhomogeneity. This study is limited to the North American region because it is an area of relatively homogeneous observations and one for which records of changes are available (Gaffen 1993; 1995, personal communication). The historical information on changes was combined with visual inspection to determine the final set of stations for analysis.

The details of the data processing, including quality control, to generate time series of monthly anomalies are described in section 2. In section 3 long-term annual and seasonal mean fields of precipitable water and relative humidity are presented. Section 4 presents the annual and seasonal trends of the quantities. In sections 5 and 6 the results are summarized, shortcomings discussed, and plans for further study outlined.

2. Data processing

The observations used here were obtained from the National Center for Atmospheric Research and comprise the routine radiosonde observations made by national weather services in the region and transmitted over the Global Telecommunication Service (GTS). For this study, North America is defined as the area north of the equator and ranging from 180°–50°W. It includes all of North America (excluding Greenland), Central America, South American stations north of the equator, and some stations in the central Pacific. In a few cases, where stations were moved less than 50 km but their identification number changed during the 1973–93 period, the data were treated as a continuous record.

Pressures, temperatures, and dewpoint depressions were extracted from each sounding at all reported levels (i.e., mandatory and significant levels) from the surface to 300 mb. Data extraction was terminated at 300 mb because of the poor performance of radiosonde humidity sensors at cold temperatures (in fact, observations above 500 mb may be adequate only in the Tropics). Observations at 0000 and 1200 UTC were treated separately.

Soundings transmitted over the GTS can and do contain clearly erroneous values and the application of quality control checks was necessary. The first step in the quality control procedure was the development of a climatology for temperature checking using data from all available soundings from 1973 to 1991. To avoid contamination by grossly erroneous values, the climatology was iteratively calculated. In the first pass through the data, gross errors in mandatory level temperatures were screened from the monthly means by rejecting temperatures falling outside a range of ±30°C of the zonal mean monthly temperatures of Oort (1983). Monthly means from the first pass were then used to screen the temperatures in a second pass by rejecting temperatures beyond $\pm 5\sigma$ of the first pass means, where the standard deviation is based on the daily values for a particular month. The monthly means from the second pass determined the "temperature-checking climatology." The reduction in the range of allowed temperatures between the two data passes was to reduce the impact of data errors on the temperature-checking climatology. Finally, each reported mandatory temperature was required to be within $\pm 4\sigma$ of the monthly mean of the temperature-checking climatology for that station. Temperatures at significant levels were excluded if they were colder than the lowest acceptable temperature for the higher mandatory level or warmer than the highest acceptable temperature at the lower mandatory level.

Because reported dewpoint depression values of a given sounding must be within the range of 0°–49°C, dewpoint depression checking was limited to insuring that each value was within the possible range. Surface pressures were required to be within a range appropriate to the station elevation (adjusted hydrostatically from a range of 950–1060 mb at sea level). For a sounding to be retained in the final dataset, it had to have a valid surface pressure as well as valid temperatures and dewpoints at the surface, 850 mb (if above the surface), and 700 mb. For example, a sounding with one missing or erroneous value of temperature or dewpoint at the surface resulted in the exclusion of the entire sounding from further processing. Most stations had fewer than 5% of soundings rejected by any of these restrictions. At 16 stations, (in northernmost Canada, Mexico, and the Caribbean) between 10% and 16% were rejected. Above 700 mb, a missing or erroneous value of temperature or dewpoint resulted in truncation of the sounding below that level. Thus, if a sounding was missing the dewpoint value at 500 mb, the sounding data up to 700 mb were still retained but data at 500 mb and above were eliminated. Significant levels with missing or erroneous values were ignored in the computation of precipitable water but otherwise did not cause truncation of the sounding. Note that the final temperature climatology was based only on observations that had valid humidity data, so there could be slight differences in mean temperatures from those based on all valid temperature measurements. However, because the percentage of rejections was small, this should be a minor difference.

In two situations the reported dewpoint depression values were adjusted, both arising because of United States reporting practices in effect until October 1993. If the ambient temperature was less than −40°C, United States stations reported the dewpoint depression as "missing," regardless of what the humidity sensor indicated. Canada, using the same humidity sensor, has reported the observation as received since about 1983, at least down to −65°C. To reduce a bias toward warm temperatures, missing dewpoint depressions at temperatures below −40°C at the 500-, 400-, and 300-mb levels were replaced with values corresponding to a relative humidity of 50% at 500 mb, 40% at 400 mb, and 35% at 300 mb. Significant levels with temperatures below −40°C were ignored. These relative humidity values were the median values of several years of Canadian observations when the temperature was below −40°C at those levels. The adjustments were most common in Alaska where ~8%, 45%, and 95% of soundings were adjusted at 500, 400, and 300 mb, respectively. This means that virtually all the humidity values at 300 mb and one-half at 400 mb in Alaska were reported as missing and so there is really little information there. This is one reason little is said here about water vapor above 500 mb.

The second case corresponds to conditions when the measured relative humidity was less than 20%. The U.S. practice, which was followed by most other countries in the region, was to transmit a dewpoint depression of 30°C in these conditions. Because a dewpoint depression of 30°C is quite dry (e.g., at a temperature of 0°C it corresponds to a relative humidity of about 8%), this value can lead to an underestimate of the true relative humidity. The median value of relative humidities less than 20% from several years of Canadian data was 16%, so transmitted dewpoint depression values of 30°C were replaced by values corresponding to a relative humidity of 16% for those stations following this practice. At 850 mb, the maximum percentage of soundings adjusted was ~25%, while at 700 mb and above it sometimes exceeded 50% in the subtropics.

Data from the quality-controlled soundings were then used to calculate dewpoints and relative and specific humidity values at the surface, 850-, 700-, 500-, 400-, and 300-mb levels and the precipitable water between these levels. The formulas used in those calculations are detailed in the appendix. Station monthly mean values of each variable at each observation time were calculated if at least ten observations during the month were available, otherwise that month was considered "missing." The distribution of observations within a month was not checked. Anomalies were computed as deviations of the individual monthly means from the averages of monthly data over the period 1976–90. This 15-yr period was selected because some stations lacked data before 1976 and others closed after 1990. This shortened period meant that the mean values of each station reflected the same time period.

Information about historical changes in radiosondes or observing practice of the countries in this region (Gaffen 1993) then guided the final choice of stations for inclusion. Stations were excluded if a known date of an instrumentation change appeared to coincide with an abrupt jump in the time series of 700- and 500-mb dewpoints. For example, four Canadian stations were eliminated because such a jump was visually apparent at the time when the radiosonde changed from a VIZ sonde to a Väisälä sonde. Time series from 17 U.S. stations that changed from VIZ sondes to SDD sondes were similarly excluded since there were apparent discontinuities in these time series as well. Of five stations for which we have no station history information, two were eliminated after visual inspection of the time series and the other three retained.

Known changes were not, in themselves, sufficient to reject a station record. For example, in 1980–81, 67 U.S. stations changed to a new version of the VIZ hygristor and a modified algorithm to compute RH. Information from the National Weather Service on limited comparison testing of the new and old hygristor (Blackmore, 1995 personal communication) indicated that humidities might average 2%–4% higher with the new sensor. Visual inspection of the time series from a subset of these stations was inconclusive; some stations showed a possible small increase near the date of the change, while others did not. Because of the inconsistency of the possible change at the stations, all stations that included this change were retained. However, the implications of an effect from this change will be further discussed in section 5.

To ensure that the time series at all stations are roughly similar in length, only stations with at least 220 months out of a possible 252 were included. There were 93 stations at 0000 UTC and 100 stations at 1200 UTC, satisfying all the conditions. The station locations are indicated by the filled squares in Fig. 1. Stations that were excluded because of instrumentation changes are indicated by the open squares. The excluded stations in the western United States changed from VIZ to SDD sondes.

3. Climatology

Before discussing the trends, some of the spatial variation of the mean fields of the variables is presented here. These fields are of interest in themselves and pro-

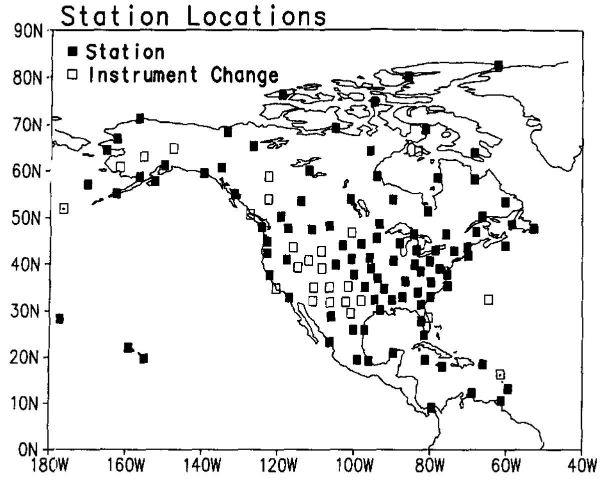

FIG. 1. Radiosonde stations used in this study are denoted by a filled square. Those stations with an open square were excluded from the analysis because of a change in instrumentation during the 1973–93 period.

vide a context for the subsequent trend analysis. This climatology is based on 15–yr means for the period 1976–90. At the pressure levels presented, there was no substantial difference between the 0000 and 1200 UTC climatologies. The 1200 UTC results are shown because there were a few more stations reporting at that time.

The reliability of humidity measurements becomes an important consideration at very cold temperatures and low humidities. The annual mean position of the −30°C isotherm was used as an estimate of the region of reliable measurements. Roughly speaking, the −30°C annual mean isotherm was at 68°N at 500 mb and 45°N at 400 mb. None of these stations had annual mean 300-mb temperatures warmer than −30°C. For these reasons the following discussion is confined to data from 500 mb and below. This layer contains more than 90% of the total precipitable water. The percentage of water vapor in the 500–300 mb layer was 4%–5% of the surface–500 mb annual total except in mountainous regions where the percentage can be slightly larger but total precipitable water is small.

The temperature determines the upper bound of the water vapor content, so there is a strong similarity in the large-scale patterns of temperature and the moisture variables except relative humidity. Temperature, dewpoint, specific humidity, and precipitable water are largest in low latitudes and generally decrease poleward. Precipitable water and relative humidity will be discussed below in greater detail. Annual mean values of specific humidity at the surface range from 18 g kg^{-1} along the north coast of South America to 1 g kg^{-1} in the Canadian Arctic. At 700 mb the values range from greater than 5 g kg^{-1} near the equator to less than 1 g kg^{-1} in northernmost Canada. At 300 mb values range from 0.2 g kg^{-1} in the low latitudes to less than 0.05 g kg^{-1} over most of Canada and Alaska.

a. Precipitable water

The annual mean surface–500 mb precipitable water (W) for the North American region is presented in Fig. 2. The largest values, 30–40 mm, occur south of 30°N and generally decrease northward to values less than 10 mm north of 60°N. Stations in the south-central and southeastern United States have values between 20 and 30 mm, but those in the remainder of the United States are in the range 15–20 mm. Values over high elevations in the western interior are slightly lower. The sharp gradient over Mexico is a plausible feature of the elevation contrast between coastal stations and those in the mountainous interior. The broad features are quite similar to those shown by Wittmeyer and Vonder Haar (1994) based on satellite observations during 1983–89.

The annual cycle of water vapor reflects the cycles of temperature and of circulation patterns. Figure 3 shows the mean winter (DJF) and summer (JJA) fields of the surface–500 mb W. While the values are everywhere higher in summer than in winter, the largest winter to summer increases (>20 mm) occur over the United States east of 100°W. The large annual range in the eastern United States is consistent with water vapor transport studies (e.g., Peixoto and Oort 1983) that show a deep northward flux from the Gulf of Mexico in summer and southward transport of cold, dry polar air masses from Canada in winter. In contrast, the increase of W from winter to summer is usually less than ~10 mm at low latitudes, the west coast of the United States, and northernmost Canada.

To examine the vertical distribution of W, we calculated the percent of the total water vapor below 500 mb that was in each of the three sublayers: surface–

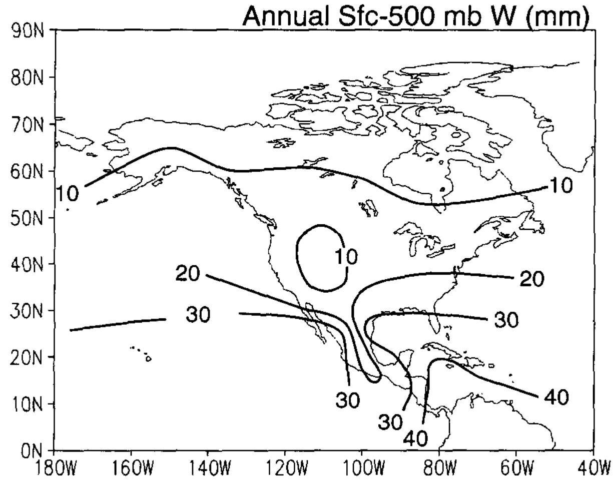

FIG. 2. Annual mean surface–500-mb precipitable water (mm).

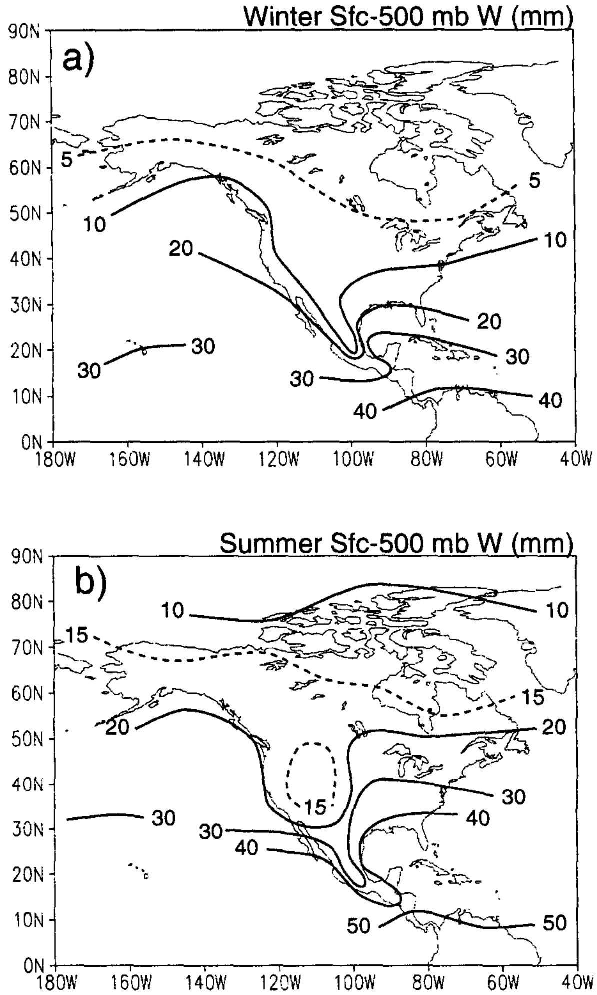

FIG. 3. Seasonal mean surface–500-mb precipitable water (mm) for (a) winter and (b) summer.

850 mb, 850–700 mb, and 700–500 mb. Figure 4a shows the percent of the annual mean water vapor in the surface–500 mb layer that is below 850 mb for those stations whose surface elevation is less than 200 m. (This confines consideration to coastal and island stations and to the lowlands of interior Canada.) The range is between 45% and 65%, with the lowest percentages at interior continental stations and values greater than 60% found at the central Pacific locations. Serreze et al. (1995) reported that 80% of the total water vapor was below 700 mb for Arctic stations north of 65°N in agreement with these results.

There is less geographic variation of the precipitable water fraction contained in the 700–500 mb layer (Fig. 4b), but the latitudinal gradient is reversed. A similar reversal also is seen in the 850–700 mb layer (not shown). In the middle layer, the higher-latitude values are 30%–35%, while the low-latitude and West Coast regions are 5%–10% lower. The relatively dry upper regions of the low latitudes show the influence of the descending branch of the Hadley cell.

There is little seasonal variation of the fractions in the lowest two layers except in Canada and interior Alaska. Large seasonal variations occur there; nearly 50% of the water vapor is below 850 mb in summer but less than 40% is there in winter. In fact, in winter in northeastern Canada the fraction between 850 mb and 700 mb is nearly the same as in the surface–850 mb layer; about 35%–40%. This probably reflects the lack of a low-level moisture source in winter when the ground and adjacent water surfaces are frozen. In contrast, Serreze et al. (1995) reported little seasonal variation of the percentage of W below 700 mb for Arctic stations north of 65°N. However, they averaged over all longitudes and the discrepancy with these results may indicate seasonal differences between Eurasia and North America.

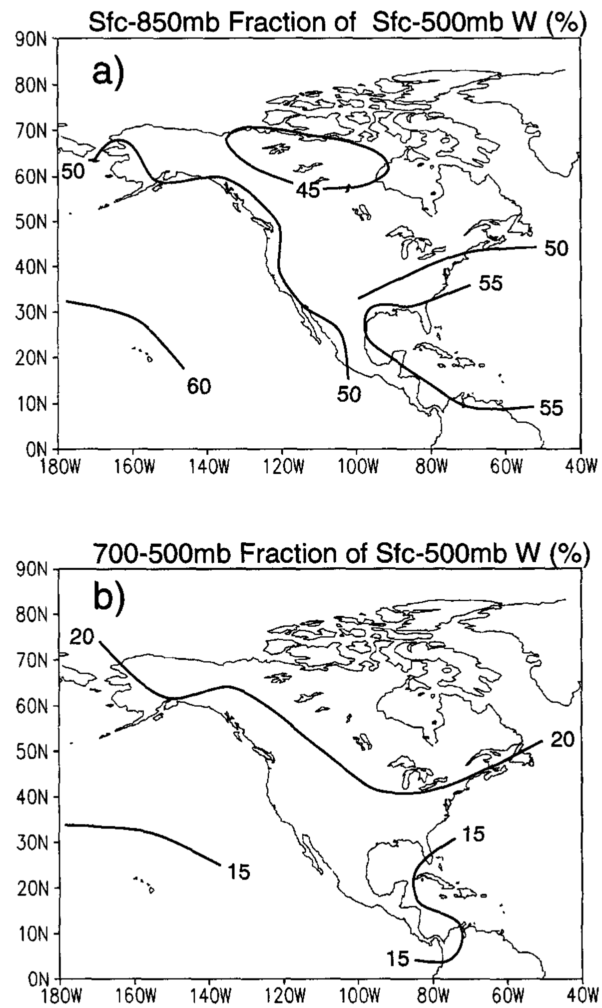

FIG. 4. The percentage of precipitable water in the (a) surface–850 mb layer and (b) the 700–500 mb layer relative to the surface–500 mb layer.

In the 700–500 mb layer the seasonal variation of the water vapor fraction differs between the Caribbean region (higher in summer than winter by ~4%) and eastern North America north of 30°N (lower in summer than winter by ~5%). The smaller summer percentage over the eastern United States compared with winter is consistent with earlier analysis by Reitan (1960). There is less seasonal variation along the West Coast, even as far north as Alaska and along the Aleutians.

b. Relative humidity

The annual mean relative humidity (RH) at 850, 700, and 500 mb is shown in Fig. 5. There is a general decrease of RH with height. At 850 mb the minimum values of RH are over the United States, while RH increases to the north and south. This is in broad agreement with Peixoto and Oort (1992). Note the very low values along the California coast and the high values over Hawaii at 850 mb. Above 850 mb the highest values are in the north, particularly in northwest Canada and interior Alaska. At 700 mb the Caribbean and United States have roughly comparable values, except for coastal California and Hawaii, which are dry. At 500 mb the same pattern is apparent but the range of values is reduced particularly in the western half of the domain.

The dry values along the California coast at all three levels and over the central Pacific stations at and above 700 mb reflect the subsidence associated with the subtropical anticyclone and the Hadley circulation. To a lesser extent, this also is seen over the Caribbean at 500 mb where relative humidities drop to ~30%. The difference in the height at which California, Hawaii, and the Caribbean become dry is consistent with the increase in the height of the top of the trade wind inversion westward from continental coasts (Hastenrath 1985).

The patterns of relative humidity in the four seasons (not shown) appear quite similar to the annual patterns with a few exceptions. At 850 mb the California coast has distinctly lower RH in summer (~25%) than at other seasons (~40%), while summer is the most moist season in the eastern United States (~70% in

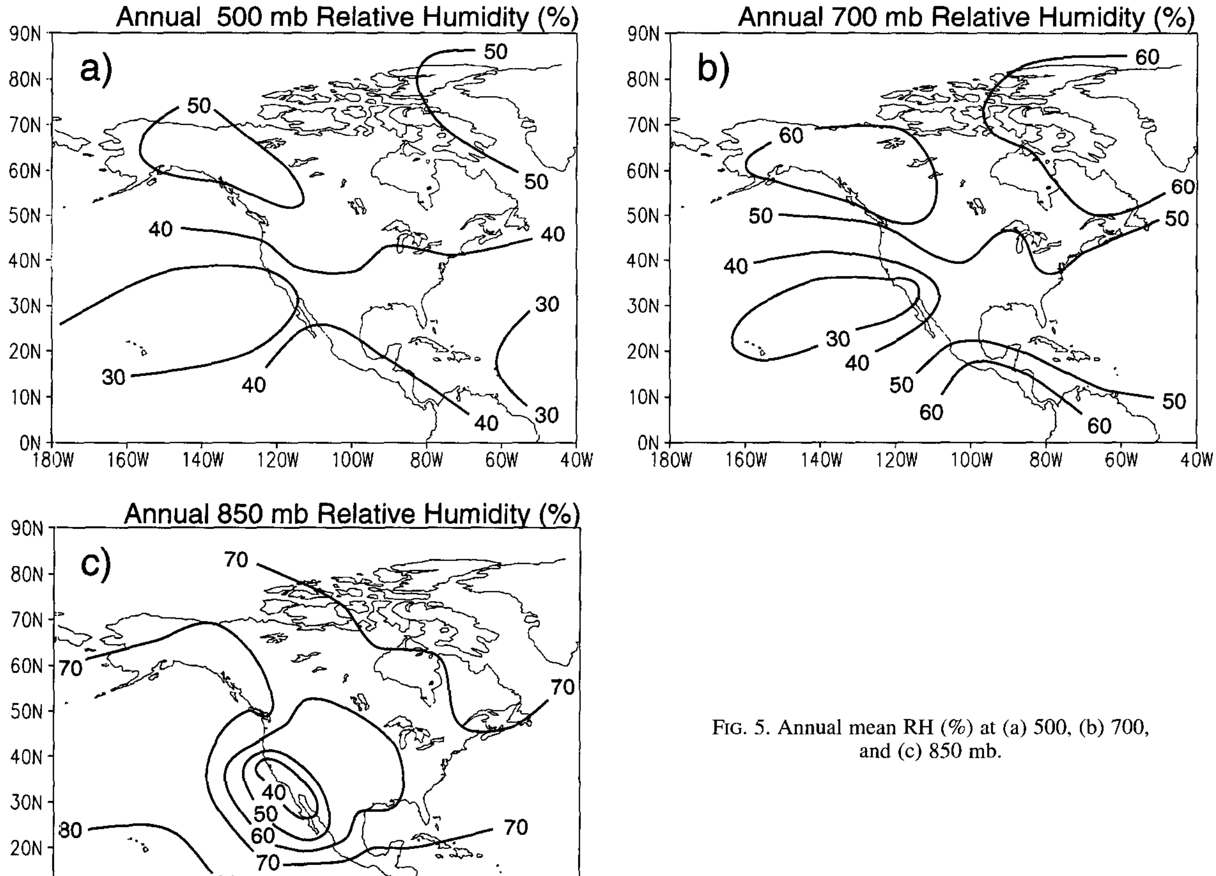

FIG. 5. Annual mean RH (%) at (a) 500, (b) 700, and (c) 850 mb.

summer compared with ~55% in winter). A similar seasonal variation appeared at 700 mb. Above the Caribbean stations at 700 and 500 mb, winter is relatively dry (e.g., 35% at 700 mb), while autumn is moist (e.g., 55% at 700 mb), consistent with their distinct rainy season in late summer and dry season in winter (Hastenrath 1985).

4. Trends

To examine the data for any existing trends in this 21-yr record, time series of the monthly anomalies of each variable were formed. These anomalies were deviations from monthly mean values calculated for the period 1976–90 as described above. The use of monthly anomalies eliminated the seasonal cycle from the data and was important for the computation of seasonal and annual averages when one month was missing. Trends at both observation times were examined, but the discussion is primarily based on the results for 1200 UTC. The differences between the trends at the two times were largest at the surface, where diurnal variability is largest. However, the differences were not systematic and do not affect the general conclusions. Annual trends were computed using the average of the 12 monthly anomalies (January–December), while seasonal trends were based on the average of the three monthly anomalies of each season. At least two months were required for each seasonal anomaly and at least 10 months were required for each annual anomaly.

Two methods were used to evaluate the trends in the anomaly time series. The Spearman test, a nonparametric test based on ranks (WMO 1966), tests for randomness against the alternative of a trend. As a nonparametric test, there is no assumption of the functional form of the trend, linear or otherwise. Furthermore, a Gaussian data distribution is not assumed. The level of confidence that a trend is present was determined by comparing the test statistic with a standard t distribution using $n - 2$ degrees of freedom, where n was the number of years or seasons of data (at most 21). The Spearman test statistic formula is given in the appendix. The sign of the test statistic indicated the trend direction. Thus, the Spearman test determined trend existence and trend sign but did not indicate a rate of change. In this study, values that rejected the null hypothesis of randomness at the 95% confidence level were considered statistically significant.

As an indication of the rate of change over this data period, the slope of the linear regression line was computed. The linear rate of change was useful for quantitative comparisons among those stations where the Spearman test indicated a trend was present. The standard error of the linear regression slope was also computed as an estimate of the uncertainty in the slopes. Seasonal and regional differences in the trends that were more than twice the standard error were assumed to represent actual differences. Trends were calculated for precipitable water, relative humidity, temperature, and dewpoint temperature at all stations and all levels or layers.

a. Annual trends

Magnitudes of linear trends of precipitable water in the surface–500 mb layer are shown in Fig. 6a; those statistically significant by the Spearman test are outlined by black boxes. The trends for this layer are nearly all positive (increasing W) except for some weak decreases in northern and eastern Canada. The largest values are found over the Caribbean, representing a linear rate of increase in W that was greater than 2 mm decade^{-1}. There is considerable spatial coher-

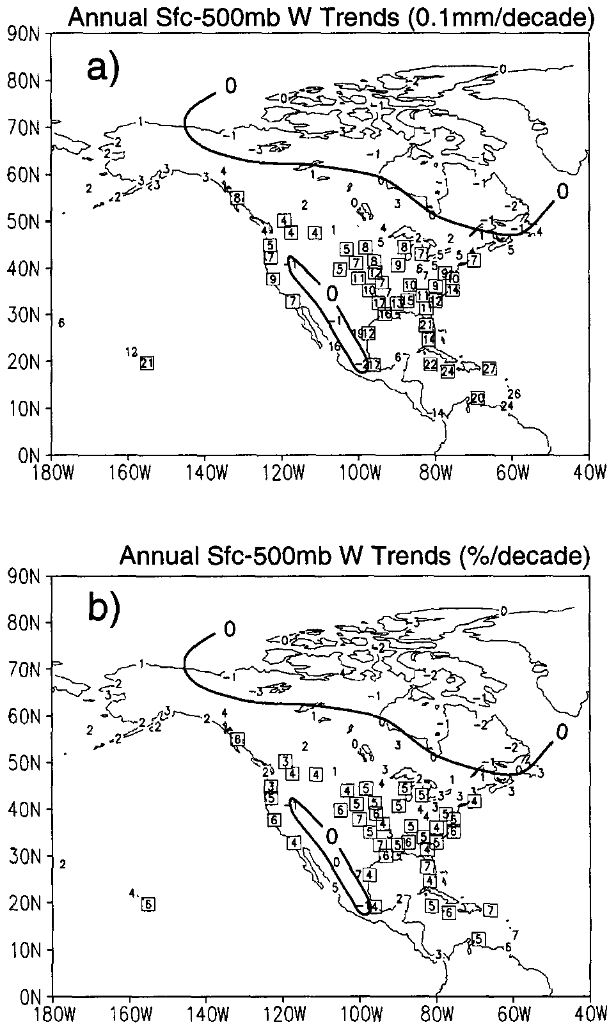

FIG. 6. Annual trends of surface–500 mb precipitable water expressed (a) in 0.1 mm decade^{-1} and (b) as a percentage of the 1976–90 mean value (% decade^{-1}). Printed values were rounded to the nearest percent. Trends outlined with a black square exceed the 95% confidence level.

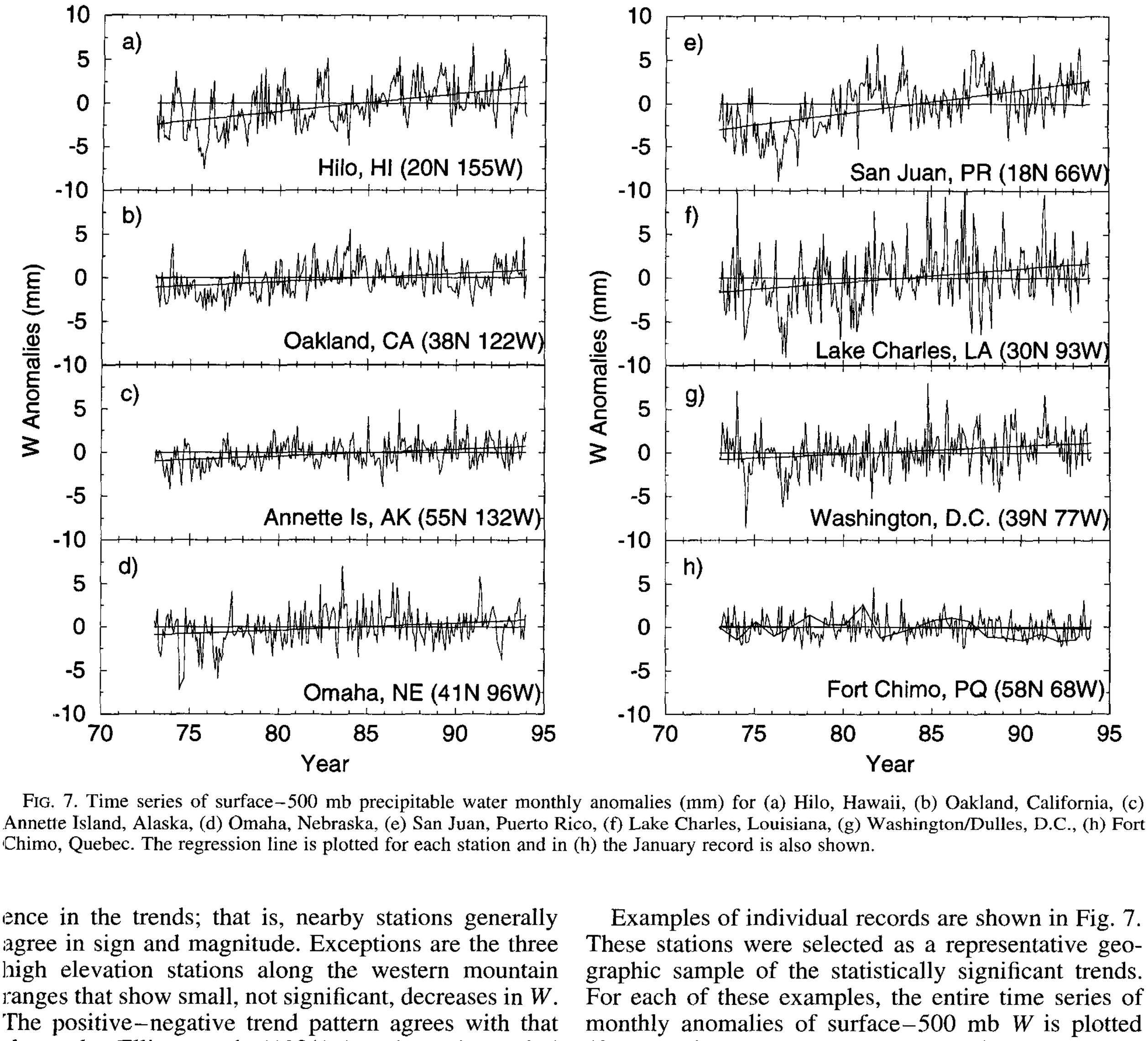

FIG. 7. Time series of surface–500 mb precipitable water monthly anomalies (mm) for (a) Hilo, Hawaii, (b) Oakland, California, (c) Annette Island, Alaska, (d) Omaha, Nebraska, (e) San Juan, Puerto Rico, (f) Lake Charles, Louisiana, (g) Washington/Dulles, D.C., (h) Fort Chimo, Quebec. The regression line is plotted for each station and in (h) the January record is also shown.

ence in the trends; that is, nearby stations generally agree in sign and magnitude. Exceptions are the three high elevation stations along the western mountain ranges that show small, not significant, decreases in W. The positive–negative trend pattern agrees with that shown by Elliott et al. (1991) based on the period 1973–88 (they did not report trend magnitudes). The geographic distribution of trend magnitudes is consistent with that shown by Gaffen et al. (1992) based on the period 1973–90.

Figure 6b presents the same trends as Fig. 6a with the values expressed as a percentage of the long term annual mean of each station. This normalization reduced the latitudinal gradient but there was still a tendency for a south to north decrease in percentage change. Most trends of magnitude greater than 4% exceed the 95% confidence level. No decreasing trend was significant. South of 45°N, W generally increased at a rate of 3%–7% decade^{-1}, with corresponding standard errors of the linear slopes of 1%–2%. The water vapor increases over the central Pacific and Caribbean are similar to the 6% decade^{-1} increase of W found by Gutzler (1992) at four western Pacific tropical stations for the period 1974–88.

Examples of individual records are shown in Fig. 7. These stations were selected as a representative geographic sample of the statistically significant trends. For each of these examples, the entire time series of monthly anomalies of surface–500 mb W is plotted (252 months, at most) along with the linear regression line. Seven of the stations showed significant increases for the period. Fort Chimo (Fig. 7h) is included as an example of a decreasing trend, although the trend was significant only in winter (January's record is also indicated).

The general increase in water vapor over most of North America in this period was not necessarily steady. Some records show that much of the increase occurred between about 1977 and the mid 1980s. San Juan (Fig 7e) illustrates the variability most noticeably. The periods of positive anomalies in 1983, 1987, and 1992/93 coincide with positive ENSO episodes. The rise in some records (e.g., Figs. 7a and 7b) beginning about 1977 may well be a manifestation of the 1977 climate transition (e.g., Trenberth 1990). Gaffen et al. (1991) also found a change in water vapor at this time in the Tropics as well as over North America.

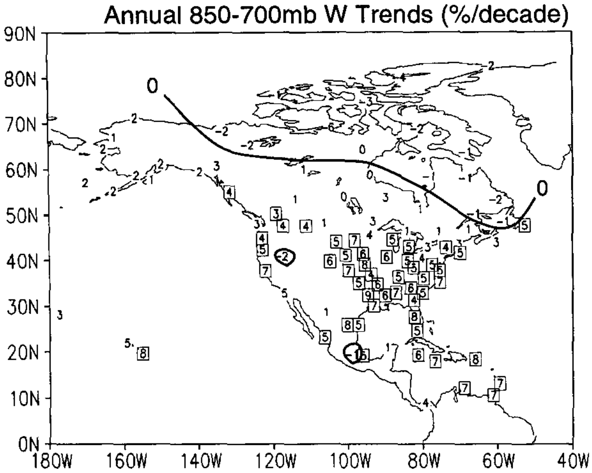

FIG. 8. Annual trends of precipitable water as a percentage of the 1976–90 mean in the 850–700-mb layer. Trends outlined with a black square exceed the 95% confidence level.

The W trends in the surface–500 mb layer are dominated by changes in the lower portion, which contains most of the water vapor in the column. Because of the importance for radiation flux of even small absolute changes of water vapor aloft (e.g., Shine and Sinha 1991), trends were calculated for three sublayers: a lower (surface–850 mb), middle (850–700 mb), and an upper layer (700–500 mb). The trends for the middle layer are shown in Fig. 8 as a percentage of the appropriate annual mean.

All three layers show a similar pattern; generally higher positive trends at low latitudes that decrease toward the north and become negative in northern and eastern Canada, similar to the total layer trends (Fig. 6b). Percentage increases in W are of equal or greater size at the middle and upper layers as in the lowest layer. At all levels, the positive values south of ~50°N are nearly all significant. Percentage trends in low latitudes and over the south-central and eastern United States are smallest near the surface and increase with height through the middle and upper layers. Over the Caribbean, the difference between the percentage trends of the lower and upper layer is 3%–6%.

Precipitable water trends at 0000 UTC showed the same general pattern and magnitudes. For the surface–500-mb layer, the differences between the 0000 and 1200 UTC trends are nearly all less than 0.1 mm decade^{-1}. The largest differences are at lower latitudes, especially the three central Pacific stations where the trends at 0000 UTC are about 0.5 mm decade^{-1} larger than those at 1200 UTC. Most of the difference probably reflects differences near the surface. For example, the difference in dewpoint trends at the surface are mostly less than 0.3°C decade^{-1}, while the corresponding 700-mb differences are generally less than 0.1°C decade^{-1}.

Trends in dewpoint permit direct comparison with temperature trends at different vertical levels. Trends in 700-mb temperature and dewpoint are shown in Figs. 9a and 9b. The standard error of the linear slopes was about 0.2°C for temperature and 0.3°C for dewpoint. In general there were more stations where trends of T_d are significant than of T. The positive–negative pattern of the temperature trends is very similar to that of W (Fig. 6b). The warming–cooling pattern is also similar to the surface temperature trends discussed by Jones (1988). The corresponding dewpoint trends usually have the same sign as the temperature trends but are generally larger when positive. This is especially true over the Caribbean where trends in temperature are less than 0.5°C decade^{-1}, while dewpoint trends are greater than 1.5°C decade^{-1}. Western Canada is the only area where positive temperature trends are comparable or greater than those of dewpoint over this 21-yr period.

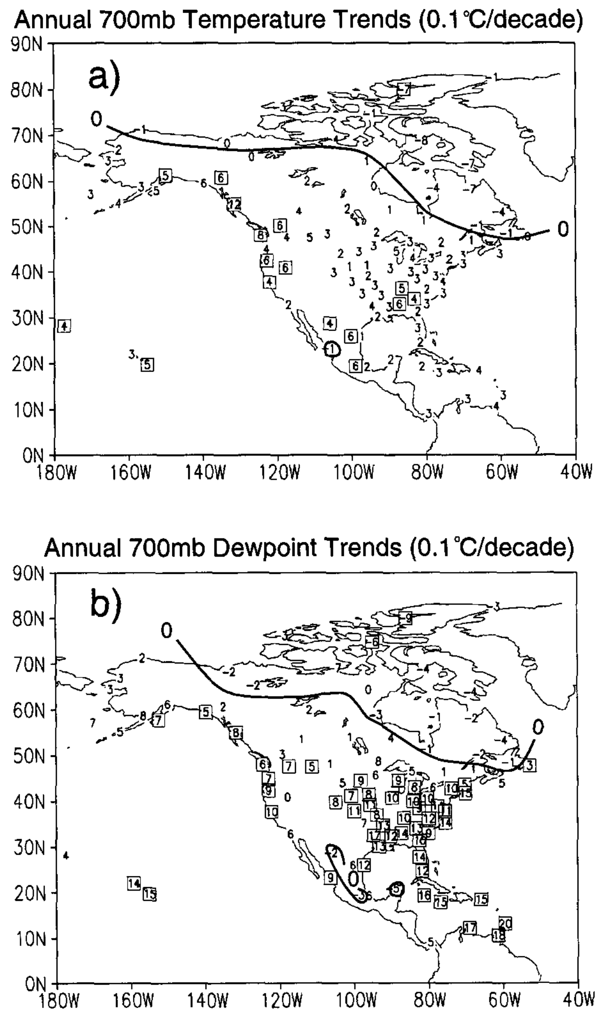

FIG. 9. Annual trends of (a) 700-mb temperature (0.1°C decade^{-1}) and (b) 700-mb dewpoint (0.1°C decade^{-1}). Trends outlined with a black square exceed the 95% confidence level.

TABLE 1. Number of stations where annual trends of T and T_d are both positive, both negative or of different sign. Last row is the number of stations where $dT/dt > dT_d/dt$ ($dT/dt < dT_d/dt$) when both trends are positive.

Annual trends	Surface	850 mb	700 mb	500 mb
dT/dt and $dT_d/dt > 0$	67	76	79	81
dT/dt and $dT_d/dt < 0$	11	14	13	8
$dT/dt > 0$ and $dT_d/dt < 0$	9	6	6	7
$dT/dt < 0$ and $dT_d/dt > 0$	13	4	2	3
$dT/dt > dT_d/dt$ ($dT/dt < dT_d/dt$)	16 (51)	11 (63)	15 (64)	10 (69)

These observations of the sign and magnitude of the temperature and dewpoint trends at 700 mb also hold at 850 and 500 mb. Table 1 categorizes the number of stations at each pressure level according to the sign of the T and T_d trends; both positive, both negative, or of different sign. At each level, most stations have positive trends of both T and T_d. The set of stations with positive trends of T and T_d are further subdivided into those where temperature trends are larger than dewpoint trends and the reverse. At most stations, the dewpoint trends are larger than the temperature trends; a situation that cannot continue indefinitely.

Relative humidity trends at 700 mb (Fig. 10) are generally positive in the low latitudes, decreasing at higher latitudes and primarily negative at northernmost latitudes. The trends in low latitudes and in the eastern United States were mostly significant. Similar patterns occur at 850 and 500 mb. At most of those stations the trends were largest at 700 mb and smallest at 850 mb. The pattern resembles the patterns of W and T_d but there are more negative trends. At some of these stations, temperature and dewpoint trends were both positive but temperature trends equaled or exceeded the dewpoint trends. At 850 and 700 mb, only a few stations had significant negative trends. The largest negative trends occurred at 500 mb, especially over interior and northern Canada, and more stations had significant negative trends.

Annual 700mb Relative Humidity Trends (0.1%/decade)

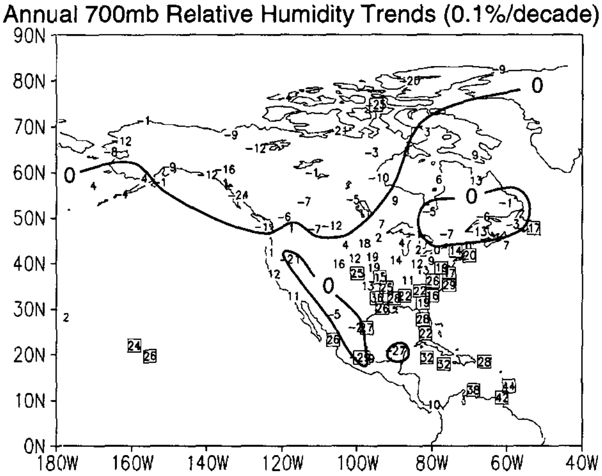

FIG. 10. Annual trends of relative humidity (0.1% decade^{-1}) at 700 mb. Trends outlined with a black square exceed the 95% confidence level.

These findings that relative humidity tended to increase over this time period are of interest because some earlier climate models (e.g., Manabe and Wetherald 1967) assumed that relative humidity would remain fairly constant under global warming conditions. Furthermore, most GCMs simulate an approximately invariant RH in doubled CO_2 experiments (Mitchell and Ingram 1992). The changes in water vapor in this period were greater than those required to maintain constant relative humidity.

b. Seasonal trends

To examine the contribution of each season to the annual trends, trends were calculated for the individual seasons (DJF, MAM, etc.). As expected, these seasonal trends were more variable in magnitude and sign than the annual trends and fewer stations had significant seasonal trends.

The seasonal variation of the surface–500 mb W trends are shown in Figs. 11a–d. Over the Caribbean moistening was generally evident in each season where trends were significant at all stations in winter, spring, and autumn and at most stations in summer. Over the entire North American region, summer was the season having the most widespread significant increases whereas autumn had the least. The drying over northern and eastern Canada derives primarily from the winter and spring seasons.

The seasonal moistening and drying regions were also generally regions of warming and cooling, respectively. To illustrate this consistency, changes in temperature and dewpoint at all levels are summarized in Table 2. The agreement in sign of trends of T and T_d among stations was highest during winter and least in summer. However, even for the season of lowest agreement, summer trends at 850 mb, 64% of the stations still showed consistent signs. Where both the dewpoint and temperature trends were positive, seasonal dewpoint trends were usually larger than those of temperature but less systematically so than the annual trends. Over the Caribbean and the southeastern United States dewpoint trends were consistently larger than temperature trends in all seasons.

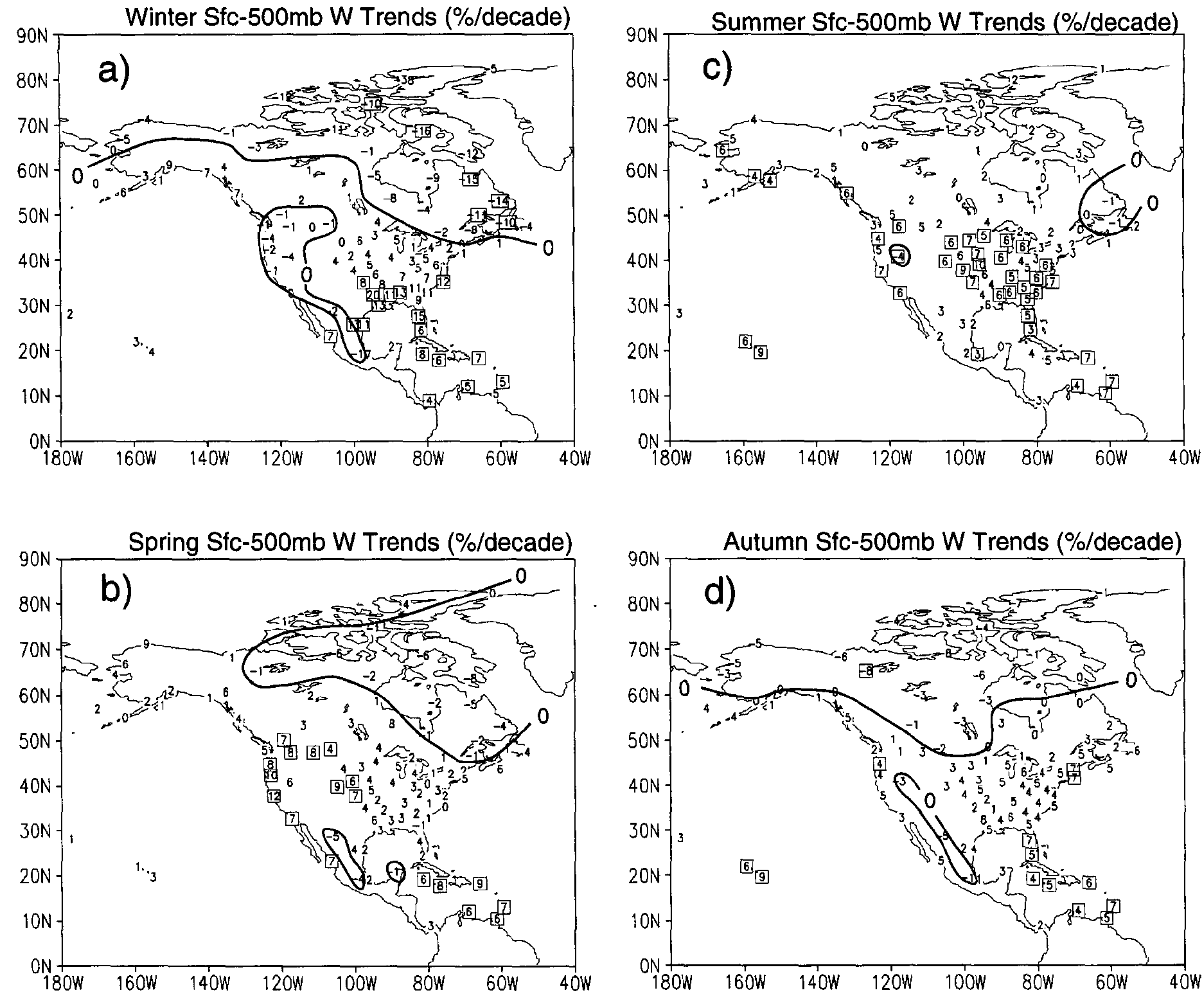

FIG. 11. Seasonal trends of surface–500 mb precipitable water (% decade^{-1}) for (a) winter, (b) spring, (c) summer, and (d) autumn. Trends outlined with a black square exceed the 95% confidence level.

5. Discussion

To detect any real climatological changes requires, first, a set of homogeneous data. North American observations come close to satisfying this criterion. Most stations have used similar humidity sensors over the period analyzed and the historical changes are well documented. This historical information was used to eliminate those stations that switched sensors during the period. Nevertheless, there were changes affecting the remaining stations so some uncertainty remains. In particular, a modification of the carbon hygristor and its associated algorithm in 1980–81 affected 67 of the United States stations used here. Visual inspection did not reveal a consistent effect, so if one exists, it does not dominate the record. To test further the sensitivity of the trends to this change, the trends of 700 mb T_d were computed using only data from 1982–93. This is clearly a short period of record, but the trends over the eastern half of the United States were about the same as those of the longer period. In the western United States the trends were near zero in the shorter period (but see below).

TABLE 2. Number of stations where seasonal trends of T and T_d agree (disagree) in sign.

Seasonal trends	Surface	850 mb	700 mb	500 mb
Winter	84 (13)	80 (15)	83 (14)	78 (15)
Spring	79 (19)	72 (25)	67 (32)	65 (32)
Summer	70 (30)	63 (35)	69 (31)	80 (20)
Autumn	70 (30)	78 (20)	79 (21)	74 (24)

Possible impacts on the trends from the two data adjustments that were made under cold ($T < -40°C$) or dry conditions (RH < 20%) were also tested. The large negative trends over Canada at 500 mb might have been an artifact of the data adjustments made to missing dewpoints at $T < -40°C$. Canada changed its cutoff criteria from $-40°C$ to $-65°C$ around 1983, so there is a potential discontinuity. Lowering the cutoff would produce more data at colder and drier conditions, which would suggest (an artificial) drying of the atmosphere (Elliott and Gaffen 1991). However, when the trends were recomputed with early Canadian data at $T < -40°C$ accepted as "missing" (rather than adjusted), the negative trends were slightly larger than those with adjustment of the early observations. Thus the drying in northern Canada does not appear to be a result of the adjustment.

Under the very dry conditions of the second data adjustment, the relative humidity is known to be below 20% and so the effect of the adjustment of the reported 30 dewpoint depression to a relative humidity of 16% generally leads to an increase in relative humidity (over the RH value corresponding to a dewpoint depression of 30). If the adjustment occurred more frequently at the latter portion of the time series it could result in a fictitious increasing trend. To test this, a time series of the number of dewpoint depressions reported as 30 each month was created for tropical regions (where the percentage of adjusted soundings can be greater than 50%). The trend of this time series was clearly downwards at nearly every station indicating that RH < 20% has been occurring less frequently with time. This supports the positive trends of water vapor at tropical stations.

The sensitivity of the trends to the 1977 climate transition (Trenberth 1990) was checked by computing annual trends of 700 mb T_d using data from 1978–93 (16 years). The trends in the western United States and west coast of Canada were much reduced in size and were no longer significant. However, trends at stations farther east were qualitatively consistent with the trends for the entire period. This suggests that the climate transition affected water vapor primarily in western North America. It also suggests that the post 1981 reduction in the rate of moistening of the western stations, discussed above, probably reflects this real climate change rather than the effect of the sensor modification in 1981.

Another question is how well do radiosonde observations reveal the true values of atmospheric humidity. Several qualitative speculations can be made.

The time lag of the humidity sensor is greater than that of the temperature sensor on the radiosonde so the reported humidity should be appropriate to a slightly lower elevation than the temperature (e.g., Williams and Acheson 1976). This represents a tendency to overestimate the humidity. That slower sensors give higher humidities at upper levels of the atmosphere has been shown by Soden and Lanzante (1996), who compared sondes of different types with satellite water vapor observations. Furthermore, to adjust the U.S. practice of reporting low humidity as a dewpoint depression of 30°C, the median value of such observations from Canada was used. However, Canada apparently uses an algorithm that does not allow relative humidities much less than about 14%. Although our adjustment permits comparability of U.S. and Canadian data, the result is likely to bias these low humidity observations too high. Both these factors suggest a high bias in the calculations.

On the other hand, there are also questions at very high humidities in U.S. data. Humidity reports greater than 96% are almost never recorded, even in clouds, indicating there is a low bias of the data at high values. Wade (1994) gives a good discussion of many of these problems, including their origins. While one can say qualitatively what their effects would be, their quantitative impact on climatological data remains to be investigated.

6. Summary and conclusions

This study has presented evidence for trends in water vapor over North America using radiosonde data from the period 1973–93. The climatological fields of temperature, dewpoint, relative humidity, and precipitable water from which the anomalies were generated are qualitatively consistent with previously published averages using data from earlier periods.

Total precipitable water, estimated from the surface–500-mb value, increased over much of North America. The changes in precipitable water over this period represent a linear rate ranging from greater than 2.0 mm decade^{-1} in the Caribbean to near zero or slightly decreasing at northern latitudes. When the trend was expressed as a percentage of the mean at each station, the trends south of ~45°N increased at a rate of ~3%–7% decade^{-1}. North of ~45°N, values were smaller and of mixed sign. Percentage trends in the upper portion of this layer, 700–500 mb, were as large or larger than those of the middle (850–700 mb) or lower layer.

Trends in dewpoint generally had the same sign as trends in temperature but tended to be larger and more stations had significant dewpoint trends than significant temperature trends. This was consistent with the increases found in relative humidity, which generally increased over this period except in Canada, Alaska, and a few stations in western mountainous areas. Largest relative humidity increases were in the Tropics and the trends decreased northward and eventually changed sign.

Fewer stations had significant seasonal trends compared with the annual trends and the areas where the trends were significant varied by season. More stations had significant trends in summer than in other seasons and these were located over the eastern United States and the Tropics. Spring trends were significant over the western United States and Canada, while the winter significant trends were primarily along the Gulf Coast. The one area where significant water vapor increases were found in all four seasons was the Caribbean.

Future analysis will consider the effects of ENSO on the water vapor record, as well as explore the relationships between water vapor and cloudiness or precipitation. Changes in circulation patterns including those associated with ENSO episodes may also play a role in the pattern of increases we report. For example, the southeastern United States was shown by Ropelewski and Halpert (1987) to receive more winter precipitation during ENSO periods. Since the latter half of our 21-yr period tended to have stronger and more frequent ENSO episodes than the first half, the increases in precipitable water and relative humidity shown in this study for that area may partially reflect the timing of ENSO episodes.

To expand the study to the global radiosonde network will involve more analysis of instrumentation changes because of the number of countries involved and the large body of historical information on instrumentation and procedural changes. However, it will lead to a better picture of the long-term variability of tropospheric water vapor.

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APPENDIX

Conversion and Trend Equations Used in This Study

The reported humidity data were the dewpoint depressions. After obtaining the dewpoint, the saturation vapor pressure e_s was calculated using the reported temperature as

$$e_s = f\,6.1121 \exp\left[\frac{(18.729 - T/227.3)T}{T + 257.87}\right], \quad \text{(A1)}$$

where T is in degrees Celsius. The ambient vapor pressure e was calculated using T_d in place of T in A1. The "enhancement factor" f is a correction to adjust the pressure of pure vapor at saturation over water for the mixture of air and vapor. This factor is approximated by

$$f = 1.0007 + 3.46p(10^{-6}), \quad \text{(A2)}$$

where p is the pressure. Equations (A1) and (A2) are from Buck (1981). The relative humidity is taken as

$$\text{RH} = 100\frac{e}{e_s}. \quad \text{(A3)}$$

The specific humidity, q, is given by

$$q = \frac{0.622e}{p - 0.378e}. \quad \text{(A4)}$$

The precipitable water (W) in a layer was evaluated as the summation over all sublayers defined by the reported mandatory and significant levels within that layer:

$$W = (1/g)\int q dp = \sum (1/g)\left(\frac{q_i + q_{i+1}}{2}\right)(p_i - p_{i+1}). \quad \text{(A5)}$$

With appropriate choices of units for g, p, and q, W has the units of kilograms per squared meter. When divided by the density of water, taken as 1000 kg m^{-3}, W is expressed as millimeters of liquid water.

The Spearman test-statistic (R) for trend is based on differences between the rank of each datum and its temporal position in the time series and is computed as

$$R = 1 - \frac{6\sum D_i^2}{n(n^2 - 1)}, \quad \text{(A6)}$$

where D_i is the difference between the rank and position of each datum and n is the number of differences.

REFERENCES

Angell, J. K., W. P. Elliott, and M. E. Smith, 1984: Tropospheric humidity variations at Brownsville, Texas and Great Falls, Montana, 1958–80. *J. Climate Appl. Meteor.*, **9**, 1286–1295.

Bannon, J. K., and L. P. Steele, 1960: Average water-vapour content of the air. Geophysical Memoirs 102, U.K. Meteorological Office, London, 37 pp.

Buck, A. L., 1981: New equations for computing vapor pressure and enhancement factor. *J. Appl. Meteor.*, **20**, 1527–1532.

Crutcher, H. L., and J. M. Meserve, 1970: Selected level heights, temperatures and dew points for the Northern Hemisphere. Naval Weather Service Command, NAVAIR 50-1C-52, 406 pp. [Available from U.S. Govt. Printing Office, Washington, DC 20402.]

Elliott, W. P., and D. J. Gaffen, 1991: On the utility of radiosonde humidity archives for climate studies. *Bull. Amer. Meteor. Soc.*, **72**, 1507–1520.

——, M. E. Smith, and J. K. Angell, 1991: On monitoring tropospheric water vapor changes using radiosonde data. *Greenhouse-Gas-Induced Climate Change: A Critical Appraisal of Simulations and Observations,* M. Schlesinger, Ed., Elsevier, 311–328.

Gaffen, D. J., 1993: Historical changes in radiosonde instruments and practices. Instruments and Observing Methods, Report # 50, WMO/TD-541, World Meteorological Organization, Geneva, 123 pp. [Available from World Meteorological Organization, Case Postale 2300, CH-1211 Geneva, Switzerland.]

——, T. P. Barnett, and W. P. Elliott, 1991: Space and time scales of global tropospheric moisture. *J. Climate,* **4**, 989–1008.

——, W. P. Elliott, and A. Robock, 1992: Relationships between tropospheric water vapor and surface temperature as observed by radiosondes. *Geophys. Res. Lett.*, **19**, 1839–1842.

Gutzler, D., 1992: Climatic variability of temperature and humidity over the tropical western Pacific. *Geophys. Res. Lett.*, **19**, 1595–1598.

Hastenrath, S., 1985: *Climate and Circulation of the Tropics.* D. Reidel, 455 pp.

Hense, A., P. Krahe, and H. Flohn, 1988: Recent fluctuations of tropospheric temperature and water vapor content in the tropics. *Meteor. Atmos. Phys.,* **38,** 215–227.

IPCC, 1990: *Climate Change: The IPCC Scientific Assessment.* J. T. Houghton, G. J. Jenkins, and J. J. Ephraums, Eds., Cambridge University Press, 365 pp.

Jones, P. D., 1988: Hemispheric surface air temperature variations: Recent trends and an update to 1987. *J. Climate,* **1,** 654–660.

Lott, G. A., 1976: Precipitable water over the United States, Vol. 1. NOAA Tech. Rep. NWS 20, 173 pp. [Available from U.S. Govt. Printing Office, Washington, DC 20402.]

Manabe, S., and R. T. Wetherald, 1967: Thermal equilibrium of the atmosphere with a given distribution of relative humidity. *J. Atmos. Sci.,* **24,** 241–259.

Mitchell, J. F., and W. J. Ingram, 1992: Carbon dioxide and climate: Mechanisms of changes in cloud. *J. Climate,* **5,** 5–21.

Oltmans, S. J., and D. J. Hofmann, 1995: Increase in lower-stratospheric water vapor at a mid-latitude Northern Hemisphere site from 1981 to 1994. *Nature,* **374,** 146–149.

Oort, A. H., 1983: Global atmospheric circulation statistics, 1958–1973. NOAA Prof. Pap. 14, 180 pp. [Available from U.S. Govt. Printing Office, Washington, DC 20402.]

Peixoto, J. P., and A. H. Oort, 1983: The atmospheric branch of the hydrological cycle and climate. *Variations in the Global Water Budget,* A. Street-Perrott, M. Beran, and R. Ratcliffe, Eds., D. Reidel, 5–65.

——, and ——, 1992: *Physics of Climate.* American Institute of Physics, 520 pp.

Reitan, C. H., 1960: Distribution of precipitable water vapor over the continental United States. *Bull. Amer. Meteor. Soc.,* **41,** 79–87.

Ropelewski, C. F., and M. S. Halpert, 1987: Global and regional scale precipitation patterns associated with the El Nino/Southern Oscillation. *Mon. Wea. Rev.,* **115,** 1606–1626.

Serreze, M. C., R. G. Barry, and J. E. Walsh, 1995: Atmospheric water vapor characteristics at 70°N. *J. Climate,* **8,** 719–731.

Shine, K. P., and A. Sinha, 1991: Sensitivity of the Earth's climate to height-dependent changes in the water vapour mixing ratio. *Nature,* **354,** 382–384.

Soden, B., and J. Lanzante, 1996: Satellite and radiosonde climatologies of upper tropospheric water vapor. *J. Climate,* **9,** 1235–1250.

Trenberth, K., 1990: Recent observed interdecadal climate changes in the Northern Hemisphere. *Bull. Amer. Meteor. Soc.,* **71,** 988–993.

Wade, C. G., 1994: An evaluation of problems affecting the measurement of low relative humidity on the United States radiosonde. *J. Atmos. Oceanic Technol.,* **11,** 687–700.

Williams, S. L., and D. T. Acheson, 1976: Thermal time constants of U.S. radiosonde sensors used in GATE. NOAA Tech. Memo. EDS CEDDA-7, 16 pp. [Available from U.S. Govt. Printing Office, Washington, DC 20402.]

Wittmeyer, I. L., and T. H. Vonder Haar, 1994: Analysis of the global ISCCP TOVS water vapor climatology. *J. Climate,* **7,** 325–333.

WMO, 1966: Climatic change. Technical Note No. 79, World Meteorological Organization, Geneva, 79 pp. [Available from World Meteorological Organization, Case Postale 2300, CH-1211 Geneva, Switzerland.]