

Climatology and Trends of U.S. Surface Humidity and Temperature

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(Manuscript received 22 December 1997, in final form 13 April 1998)

ABSTRACT

Climatological annual and seasonal dewpoint, specific humidity, and relative humidity maps for the United States are presented using hourly data from 188 first-order weather stations for the period 1961–90. Separate climatologies were calculated for daytime (three observations per day between 0800 and 1600 LST), nighttime (three observations per day between 2000 and 0400 LST), and the full day (eight observations per day, every 3 h).

With extended datasets for the period 1961–95, trends in these same variables and temperature are calculated for each of 170 stations and for eight regions of the country. The data show increases in specific humidity of several percent per decade, and increases in dewpoint of several tenths of a degree per decade, over most of the country in winter, spring, and summer. Nighttime humidity trends are larger than daytime trends. The specific humidity increases are consistent with upward temperature trends. The upward temperature and humidity trends are also consistent with upward trends in apparent temperature, a measure of human comfort based on temperature and humidity. Relative humidity trends are weaker than the specific humidity trends, but they do show evidence of increases, especially in winter and spring.

The possibility that the detected trends may be artifacts of changes in instrumentation was examined, but several lines of reasoning suggest that they are not. Anthropogenic water vapor produced from fossil fuel consumption, both locally and globally, is too small a source to explain the observed trends.

1. Introduction

Surface specific humidity (q) and relative humidity (U) regulate evaporation and transpiration processes and so have obvious connections to both hydrological and surface energy budgets. However, despite current interest in quantifying and modeling these budgets (e.g., Roads et al. 1994), there is a lack of observational datasets describing the recent surface climatology of humidity over the United States. The two main purposes of this paper are to present climatological, seasonal surface humidity distributions over the United States for the recent climatic normal period 1961–90 and to evaluate trends in several humidity variables and temperature over the period 1961–95.

Climatological atlases for the United States have, in the past, presented maps and tables of selected humidity variables for various time periods. Kincer (1922) presented maps of relative humidity, wet-bulb depression, and vapor pressure for the period 1876–80. Visher (1954) gives relative humidity maps for 1899–1938 and maps of vapor pressure and wet-bulb depression for an unspecified data period. The *Weather Atlas of the United*

States (U.S. Department of Commerce 1968) includes monthly mean dewpoint maps and tables for 1946–65. The publication was reissued in 1993 with the title *Climatic Atlas of the United States* (U.S. Department of Commerce 1993) and with the same dewpoint maps. The surface dewpoint (T_d) studies of A. V. Dodd provide monthly maps (Dodd 1965a) and the seasonal variations of the diurnal cycle (Dodd 1965b) based on monthly T_d data for the United States from the 1950s, but other humidity variables are not included. Recent work by Robinson (1998) updates the U.S. surface T_d climatology for the period 1961–90 and discusses aspects of monthly and diurnal variability.

Three recent global humidity climatologies—two using radiosonde data (Peixoto and Oort 1996; Ross and Elliott 1996a), and one combining radiosonde and satellite-derived humidity observations (Randel et al. 1996)—document the distribution and variability of humidity in the free troposphere. Of these, only Peixoto and Oort (1996) deal explicitly with the surface, and then only for relative humidity. Of these recent studies, the longest data record is 21 yr (Ross and Elliott 1996a). In section 3 we present climatological maps of surface T_d , U , and q for the United States based on 30 yr of hourly data.

Among the expected climate changes due to increases in greenhouse gas concentrations is an increase in tro-

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ospheric water vapor concentrations (Kattenberg et al. 1995). Previous studies based on upper-air data have shown increases in tropospheric precipitable water over a variety of locations including, in particular, the tropical Pacific (Gutzler 1992, 1996), North America (Ross and Elliott 1996b), China (Zhai and Eskridge 1997), and some other regions since the early 1970s (Gaffen et al. 1992a). Over Europe, surface data for 1961–90 show increases in vapor pressure in all seasons, but with low statistical significance (Schönwiese et al. 1994; Schönwiese and Rapp 1997). Brazel and Balling (1986) found no trend in T_d , but decreasing U , in Phoenix, Arizona, during 1896–1984, which they suggest may be related to changes in local land use patterns. Surface T_d data from 15 U.S. stations for the period 1948–91 show increases at six western stations and little change at nine eastern stations (Knappenberger et al. 1996). In section 4, we use a much larger network of U.S. stations to estimate trends in surface U , T_d , q , temperature (T), and apparent temperature (T_a), a measure of human comfort developed by Steadman (1979, 1984) and used by the National Weather Service (NWS) as a summertime heat index.

2. Data

The source of data for this study is the National Solar Radiation Data Base (NSRDB; NREL 1992; Maxwell et al. 1995), which includes hourly meteorological and solar radiation data for 239 stations in the (50) United States and territories for 1961–90, and updated through 1995 (T. Ross, National Climatic Data Center, 1997, personal communication). These are surface airways observations from the NWS first-order stations. We employ the hourly pressure (p), T , and U data and compute T_d , q , vapor pressure (e), and T_a . For T_a we ignore the effects of wind and radiation and employ Steadman's (1984) regression equation

$$T_a = -1.3 + 0.92T + 2.2e,$$

where T is in Celsius and e is in kPa.

Beyond the NSRDB quality control (NREL 1992), we rejected all interpolated data and observations that were not physically reasonable. Data were accepted within the following ranges: for temperature and dewpoint, -70° to 60°C (with the provision that $T_d \leq T$); for relative humidity, 0%–100%; and for pressure, 600–1100 hPa. From 1965 to 1981 the meteorological data were saved only for every third hour, namely, 0300, 0600, 0900, . . . , 2400 UTC, which leads to somewhat different sampling of the diurnal cycle in different time zones. However, this diminished sampling has negligible impact on climatological monthly means. RIJC found that climatological monthly dewpoint values based on 24 and 8 observations per day differ by less than 0.1°C , based on more than 10 yr of data for which 24 hourly observations were available at 222 stations. Thus, for temporal consistency we used only eight ob-

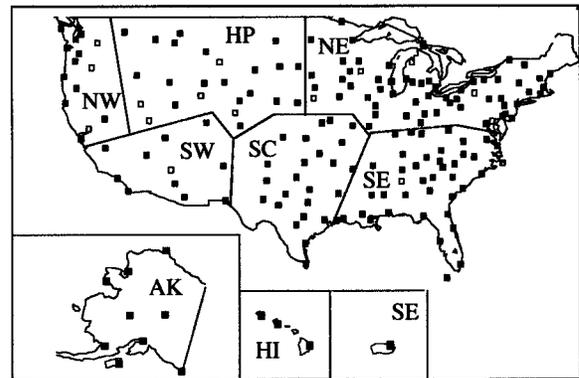


FIG. 1. Stations used in this study and the eight geographic regions into which they were placed. Open squares indicate stations where data were used only for the 1961–90 climatological analysis.

servations per day for the full period of record. After rejecting stations with significant data gaps, we retained 188 stations for the analysis of climatology for the period 1961–90 and 170 stations for the trends analysis for 1961–95, as shown in Fig. 1 and Table 1. Each of the station records is at least 98% complete, and the overall dataset has only 0.2% missing values.

Figure 1 shows the distribution of the stations and their placement in eight geographic regions. The regions are based on the climate zones identified by Fovell and Fovell (1993), who used cluster analysis of monthly surface air temperature and precipitation data for the conterminous United States to identify regions with coherent seasonal and interannual variability. We created new regions for Alaska and Hawaii, placed Puerto Rico in the Southeast region, and modified the boundaries of Fovell and Fovell's (1993) regions. The modifications involved aggregating some regions to avoid having regions with only a handful of stations, and shifting the boundaries of regions to maximize the correlation between the time series of monthly mean q for each station and the regional-average time series. The resulting eight regions, shown in Fig. 1 and listed in Table 2, are spatially uninterrupted and contain stations with coherent q variability. For the purposes of this study, it is not particularly important that the regions are not equal in size or number of stations, but that the seasonal and interannual variability of humidity is comparable at the stations within each region.

According to station history information, obtained from the National Climatic Data Center, and as summarized by Elliott (1995), different measurement systems have been employed at NWS sites during the period of interest. In the early 1960s, dry- and wet-bulb temperatures were measured manually using mercury thermometers and sling psychrometers (S. Yarkin, NWS, 1995, personal communication). Lithium chloride hygrometers, measuring T and T_d , were introduced in the early 1960s and remained in operation more than 20 yr. In the middle to late 1980s, the hygromer-

TABLE 1. Stations used in this analysis. Asterisks indicate stations included in the climatological analysis for 1961–90 but not in the trends analysis for 1961–95.

State	Cities
Alabama	Birmingham,* Huntsville, Mobile, Montgomery
Alaska	Anchorage, Barrow, Fairbanks, King Salmon, Kodiak, Kotzebue, McGrath, Nome, Yakutat
Arizona	Phoenix, Prescott,* Tucson
Arkansas	Fort Smith, Little Rock
California	Arcata,* Fresno, Los Angeles, Sacramento, San Diego, San Francisco
Colorado	Boulder,* Colorado Springs, Grand Junction
Connecticut	Hartford
Delaware	Wilmington
Florida	Daytona Beach, Jacksonville, Key West, Miami, Tallahassee, Tampa
Georgia	Athens, Atlanta, Macon, Savannah
Hawaii	Hilo, Honolulu, Lihue
Idaho	Boise, Pocatello
Illinois	Chicago, Moline, Peoria, Rockford, Springfield
Indiana	Evansville, Fort Wayne, Indianapolis
Iowa	Des Moines, Mason City, Sioux City, Waterloo
Kansas	Dodge City, Topeka, Wichita
Kentucky	Covington, Lexington, Louisville
Louisiana	Baton Rouge, Lake Charles, New Orleans, Shreveport
Maine	Portland
Maryland	Baltimore*
Massachusetts	Boston
Michigan	Detroit, Flint, Grand Rapids, Lansing, Muskegon, Sault Ste. Marie, Traverse City*
Minnesota	Duluth, International Falls, Minneapolis, Rochester
Mississippi	Jackson, Meridian
Missouri	Columbia, Kansas City, Springfield, St. Louis
Montana	Billings, Great Falls, Helena, Missoula
Nebraska	Grand Island, North Platte, Omaha,* Scottsbluff
Nevada	Elko,* Ely,* Las Vegas, Reno
New Hampshire	Concord
New Jersey	Atlantic City
New Mexico	Albuquerque
New York	Albany, Buffalo, Massena, New York City,* Rochester, Syracuse
North Carolina	Asheville, Cape Hatteras, Charlotte, Greensboro, Raleigh, Wilmington
North Dakota	Bismarck, Fargo
Ohio	Akron, Cleveland, Columbus, Dayton, Toledo, Youngstown
Oklahoma	Oklahoma City, Tulsa
Oregon	Astoria, Medford, North Bend,* Portland, Redmond,* Salem
Pennsylvania	Allentown, Bradford,* Erie, Philadelphia, Pittsburgh, Wilkes-Barre, Williamsport
Puerto Rico	San Juan
Rhode Island	Providence
South Carolina	Charleston, Columbia, Greenville–Spartanburg
South Dakota	Huron, Pierre, Rapid City, Sioux Falls
Tennessee	Bristol, Chattanooga, Knoxville, Memphis, Nashville
Texas	Abilene, Amarillo, Austin, Brownsville, Corpus Christi, El Paso, Fort Worth, Houston, Lubbock, Midland, Port Arthur, San Angelo, San Antonio, Waco, Wichita Falls
Utah	Cedar City, Salt Lake City
Vermont	Burlington
Virginia	Norfolk, Richmond, Roanoke, Sterling
Washington	Olympia, Quillayute, Seattle, Spokane, Yakima*

TABLE 1. (Continued)

State	Cities
West Virginia	Charleston
Wisconsin	Eau Claire, Green Bay, La Crosse,* Madison, Milwaukee
Wyoming	Casper, Cheyenne, Lander, Rock Springs,* Sheridan*

rometer model HO-83 was introduced. It includes a bead temperature sensor and a chilled mirror, held at the temperature at which a thin film of condensation is maintained. A second bead measures the mirror temperature as T_d (NWS 1994). The Automatic Surface Observing System (ASOS), introduced into the U.S. network from 1987 to 1997 (and continuing), includes the HO-83 sensors for T and T_d . A modification of the HO-83 system was introduced in ASOS systems beginning in 1991 (Jones and Young 1995). We defer discussion of the potential impact of these changes to section 5a.

3. Climatology

For each station, we calculated monthly and seasonal [December–February (DJF), etc.; Trenberth (1983)] mean T , T_d , T_a , q , and U , using data for the full day (eight observations per day), for daytime only (three observations between 0800 and 1600 LST), and for nighttime only (three observations between 2000 and 0400 LST). In this section we present maps depicting the 30-yr (1961–90) average climatological fields of selected humidity variables. The presentation for each variable shows the four seasonal mean maps and an annual mean map. Note that the scales for the color codes for each season are the same, except that the highest or lowest range of values may have a different maximum or minimum value for each season. A sixth map shows the annual range (the difference between the maximum and minimum 30-yr-mean monthly values) and the month of the maximum, depicted in vector form, where the range is indicated by color and the month of maximum (from January to December) is shown by direction

TABLE 2. Regions of the United States used in this study.

Region name	Abbreviation	Region number from Fovell and Fovell (1993)	No. of stations	
			1961–90	1961–95
Northeast	NE	2, 3	61	55
Southeast	SE	1, 7	42	40
High plains	HP	4, 10	24	19
Southwest	SW	6, 9	11	10
South-central	SC	5, 8	24	24
Northwest	NW	11, 12, 13, 14	14	10
Alaska	AK	None	9	9
Hawaii	HI	None	3	3
Total			188	170

of a line segment with respect to the numerals on a clock (from 1 to 12).

a. Specific humidity

The climatological distribution of specific humidity, q , the mass of water vapor per unit mass of moist air, is depicted in Fig. 2, including seasonal (Figs. 2a–d) and annual (Fig. 2e) maps of q (g/kg), based on data for the full day (eight observations per day). Over the conterminous United States, the patterns are comparable for each season and show maximum q values in Florida and around the Gulf of Mexico, with diminishing q with increasing latitude. In the Southeast and the West, coastal stations are moister than inland stations at comparable latitude, and stations in the eastern half of the country have q values about twice those at interior western stations. The annual map compares extremely well with the global presentation of Peixoto and Oort (1992, their Fig. 12.3a), based on twice-daily radiosonde data for 1963–73.

The tropical sites have the largest annual-mean q (13 to 15 g kg⁻¹) but June–August (JJA) values of 17 g kg⁻¹ along the Gulf Coast are national seasonal maxima. For most seasons, Alaskan stations have the lowest q , but in JJA the interior West has some values of 6 g kg⁻¹ that are comparable to those in Alaska (Fig. 2c). Daytime and nighttime q climatologies (not shown) are very similar because the diurnal cycle of q is small. Day–night differences in annual mean values average 3.7% of the full-day values.

As has been noted for the free troposphere (Gaffen et al. 1997), surface q is higher in fall than in spring. At most stations, the month of maximum q is July or August (Fig. 2f). The seasonal variation at West Coast and tropical stations (in Hawaii and at San Juan, Puerto Rico) is small compared with the rest of the country, consistent with the results from radiosonde observations (Gaffen et al. 1992b). The average annual range of q (Fig. 2f) over the United States is about 8.7 g kg⁻¹, is much larger in the eastern United States than in the west, and varies from 2.2 g kg⁻¹ at Honolulu, Hawaii, to 12.1 g kg⁻¹ at Topeka, Kansas.

b. Dewpoint

Because dewpoint temperature, T_d , is a function of q and atmospheric pressure, the spatial patterns of T_d are comparable to those of q . But because some readers may find dewpoint information more useful than q data, we present climatological fields of T_d (°C) in Fig. 3, where the data are for the full day. Like q , the average diurnal cycle of T_d is small. Seasonal mean daytime T_d differs from nighttime values by about 0.5 K, on average. The annual range of T_d (Fig. 3f) has a different structure from that of q , most notably where mean q and T_d are small (e.g., in Alaska). At those stations, the annual range of q is small because the relatively low

summertime T values limit maximum q values, but the annual range of T_d is large because T_d varies seasonally with T , which has a marked seasonal cycle.

The patterns are very similar to the average monthly T_d maps (°F) for the contiguous United States presented by Dodd (1965a), using data from about 200 stations from the 1950s. We have not attempted a detailed comparison to determine changes over time. The annual mean map compares favorably with that presented by Robinson (1998). At most locations, the values appear to differ by less than 1.0 K. The only exception is in the region of the Rocky Mountains, where our data show mean annual T_d values of -4° to -5°C , and Robinson's map shows no values this low, possibly due to differences in station networks.

c. Relative humidity

In contrast with q and T_d , relative humidity (U) has significant diurnal variability; therefore, we present maps of climatological daytime (Fig. 4) and nighttime (Fig. 5) U values, in percent. As noted above, daytime and nighttime are characterized using three observations spanning a 7-h period, corresponding to different local times in different time zones. Therefore, the period around sunrise and sunset is not included in this analysis and there is a slightly different sampling of the diurnal cycle at different stations. Examination of the hourly data shows that diurnal maxima of U occur around 0500 LST, or just before dawn, when T is lowest. Nevertheless, the relative smoothness of the spatial patterns, and the distinct difference between the daytime and nighttime averages, suggests that our sampling captures some of the main variability in U . Annual- and spatial-average daytime U for the United States is 59%, with a standard deviation (over all stations) of 9%. Nighttime observations average 75%, with a standard deviation of 8%.

Salient features of the U maps are the low values in the desert Southwest and the high values along the Northwest and Alaskan coasts. Over most of the country, the seasonal variability of U is minimal. The annual range of U is usually less than 20% (either day or night) but is higher in the Southwest and high plains regions, with typical values between 30% and 40%. In western states (day and night) and for stations north of 40°N (day), U maxima tend to occur in February and March, while summertime maxima are typical for the southeastern United States in day and throughout the eastern United States at night. We note, however, that the seasonal “cycle” of U is a misleading notion, as the climatological values often show multiple monthly relative maxima and, at some stations, have steplike changes in spring and fall. The details of the seasonality and diurnal variability of U will be the topic of a separate investigation.

Peixoto and Oort (1996) present maps of U at 850 mb for this region based on radiosonde data, but these are based on 0000 and 1200 UTC observations com-

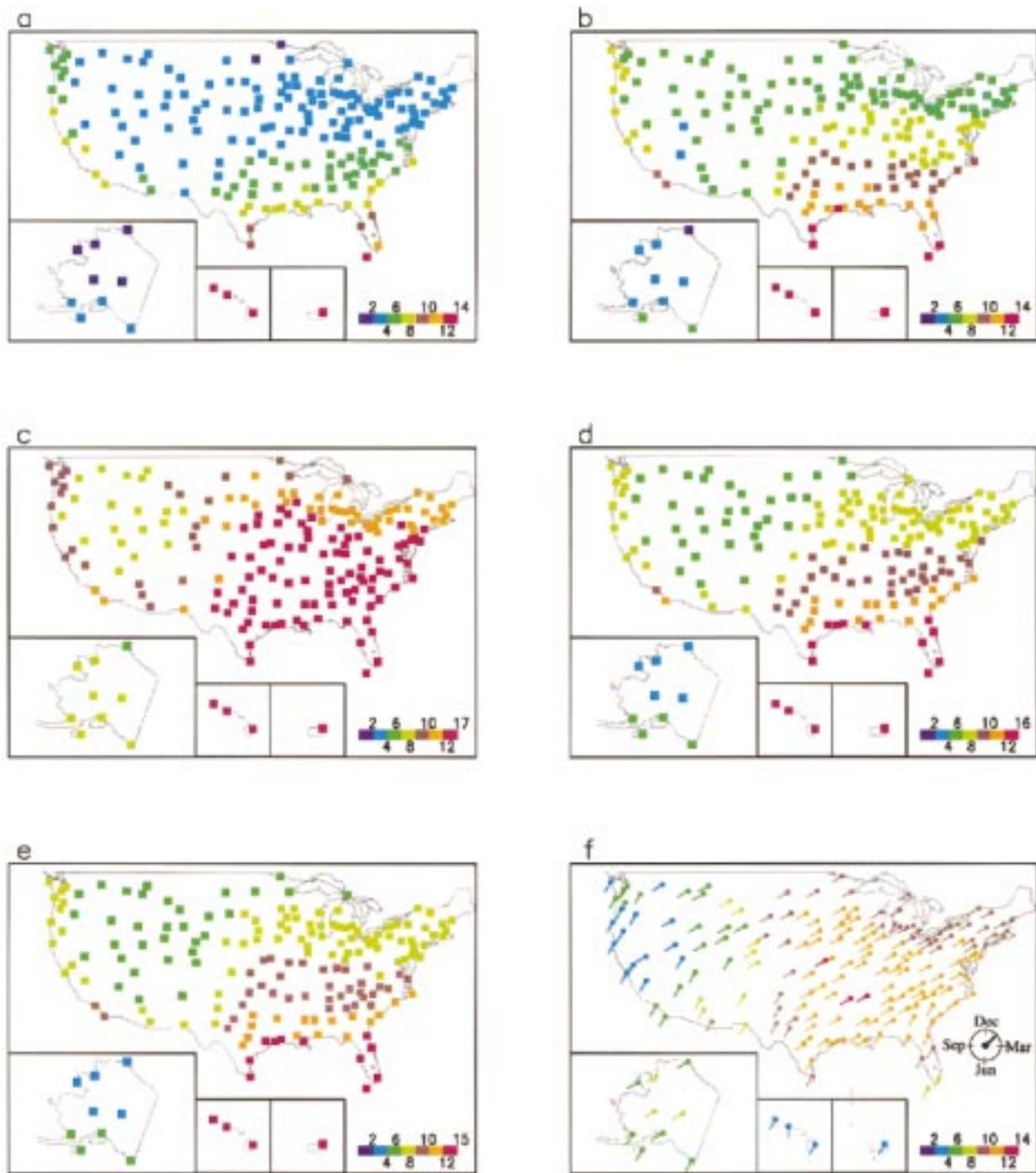


FIG. 2. Climatological specific humidity (g kg^{-1}) for (a) winter, (b) spring, (c) summer, (d) fall, and (e) annual means for 1961–90 based on data for the full day. The color bars are the same in (a)–(e), except for the range of the maximum values. The range of the annual cycle, indicated by color, and the month of maximum specific humidity, indicated by the clock direction of the line segment, are shown in (f).

binned, so a direct comparison with Figs. 4 and 5 is not feasible. Consistent with Peixoto and Oort (1996), we find no significant spatial correlation among seasonal mean values of U and either q or T .

On the other hand, as illustrated in Fig. 6, it is q and

T that are well correlated. At most stations, seasonal mean T is a reasonably good predictor of seasonal mean q . The solid lines in the figure show values of q as a function of T for different values of U and for $p = 1016$ hPa. Most of the station data cluster around the $U =$

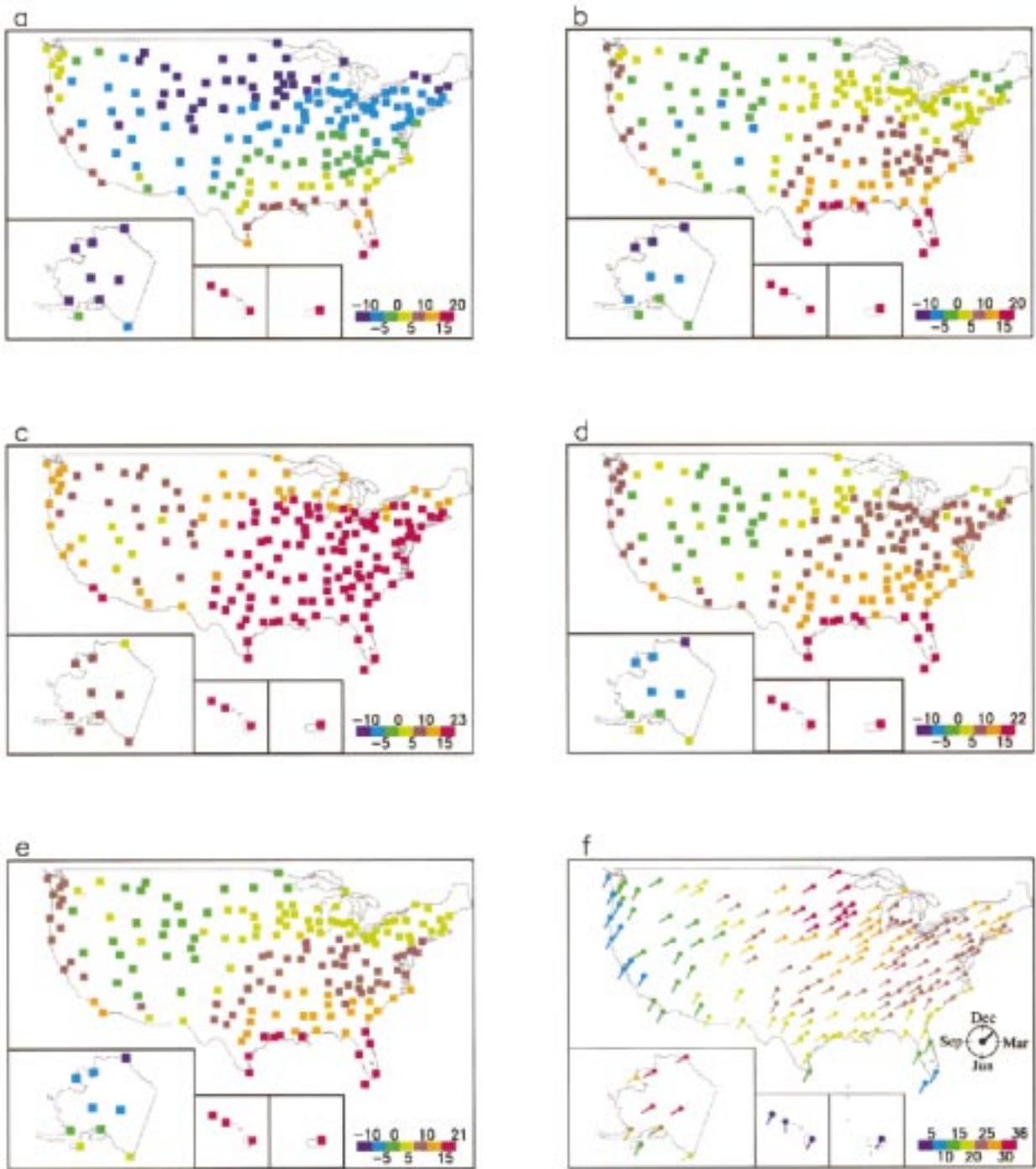


FIG. 3. As in Fig. 2 but for dewpoint ($^{\circ}\text{C}$).

60% or $U = 80\%$ lines, consistent with Fig. 4, which shows daytime U values generally within this range. The outliers fall into two categories. One is stations in the desert Southwest and high plains, where q and U are lower, particularly in summer. The other group of outliers is the Alaskan stations in winter, whose data fall above the $U = 100\%$ line because the actual pressure

is considerably less than 1016 hPa. The actual U values at these stations are closer to 70%.

4. Trends and interannual variations

Seasonal mean values of surface T_d , T , U , and q for daytime, nighttime, and full-day conditions form the



FIG. 4. As in Fig. 2 but for daytime relative humidity (%). The color bars are the same in (a)–(e), except for the range of the minimum values.

basis of 35-yr time series (1961–95) at 170 stations. Using these time series, we estimated seasonal trends using the nonparametric method of the median of pairwise slopes regression (e.g., Lanzante 1996). This method is resistant to outliers in the time series and robust to nonnormal data distributions (Wilks 1995). It in-

volves computing the slopes of lines connecting each pair of points in the time series and finding the median value of those slopes. This method avoids placing undue emphasis on large anomalies near the ends of time series. Two-tailed *t* tests of the robust and resistant Spearman rank-order correlation coefficient (between the hu-

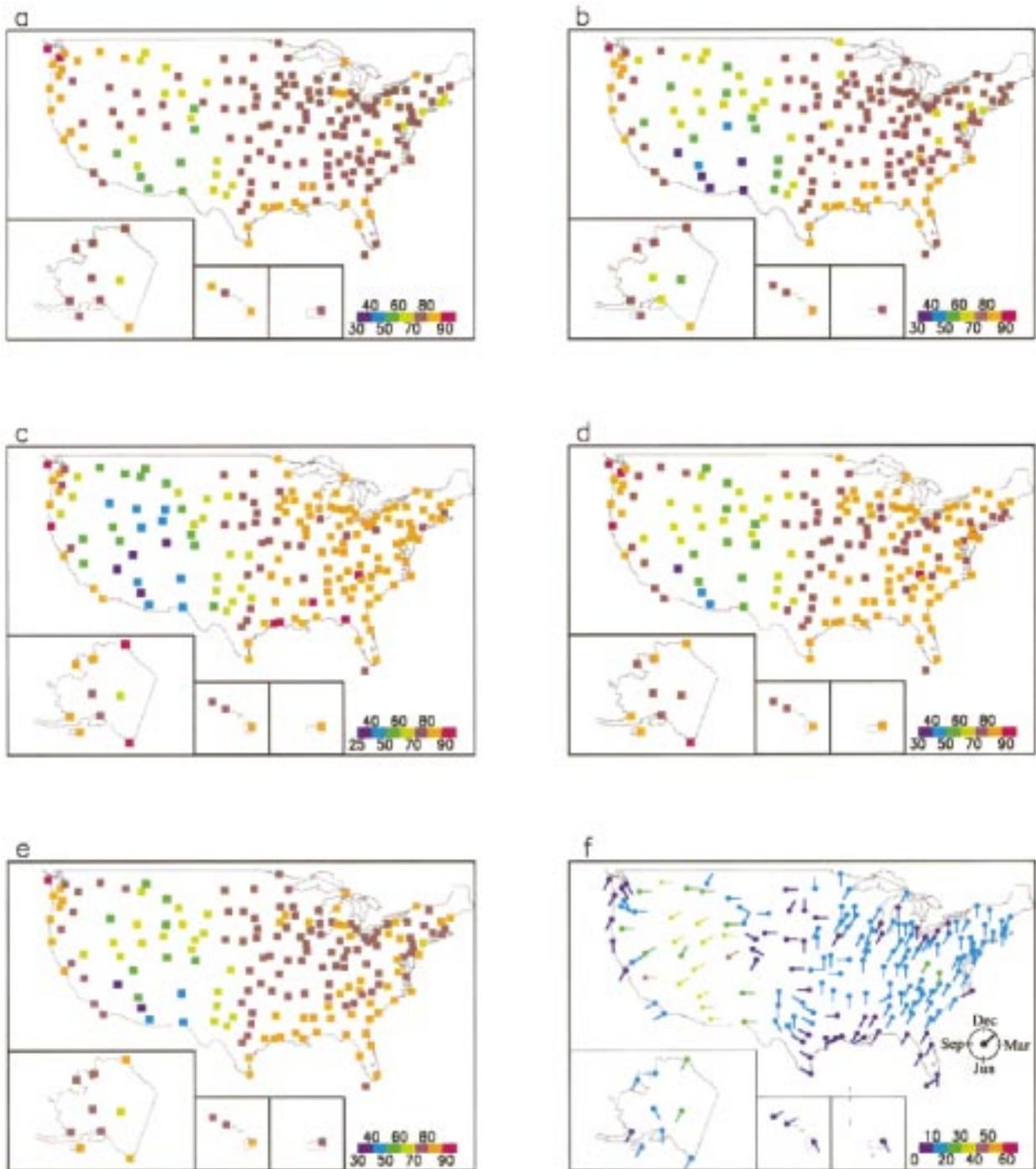


FIG. 5. As in Fig. 4 but for nighttime relative humidity (%).

midity variable and time) were used to evaluate the statistical significance of the trends.

Trends were also computed for each of the eight regions (Table 2 and Fig. 1). Regional time series were constructed by averaging the time series of seasonal values for each station within the region. This method

is predicated on the notion that the variability of all the stations' time series in a given region are similar. Recall that the regions were defined based on the correlation of time series of monthly q data. In many instances the regionally averaged time series had trends comparable in sign and magnitude to those at individual stations

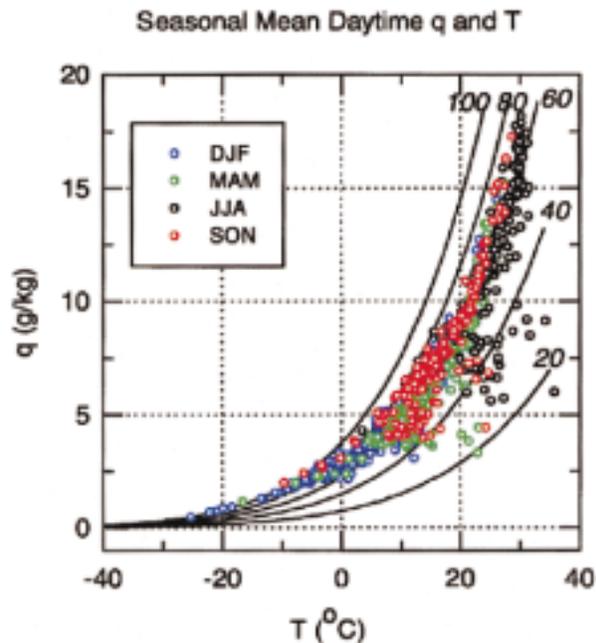


FIG. 6. The relationship between seasonal mean specific humidity and temperature at individual U.S. stations, based on daytime observations, shown by open circles. The smooth curves are the calculated relations, for relative humidity of 20%, 40%, 60%, 80%, and 100%, and pressure of 1016 hPa.

within the region, but the statistical significance was higher, presumably because of the reduction of noise variance associated with the averaging process.

a. Specific humidity

Figure 7 shows seasonal trends in regionally averaged q , normalized by the mean seasonal values and expressed in percent per decade. The statistical significance is indicated by the color of the bars: black for trends significant at the 0.01 level, gray for those significant at the 0.05 level, and white for the remainder. For each region and season, trends for daytime (leftmost bar), the full day (middle bar), and nighttime (rightmost bar) are shown.

The trends are overwhelmingly positive, except for small negative trends in Hawaii in winter and spring and in the Northeast and south-central regions in fall. The largest and most significant trends are in winter in Alaska, where q increased 9% per decade. In almost all cases, the nighttime trends exceed the daytime trends, both in a percentage sense (Fig. 7) and in an absolute sense (not shown).

The seasonal trend results are fairly consistent with trends in tropospheric (surface to 500 mb) column water vapor, W , over the United States for the period 1973–93 (Ross and Elliott 1996b; Elliott and Angell 1997). Increases in W over the western United States in spring, over the eastern and central United States in summer, and over Hawaii in summer and fall agree with these

trends in surface q . However, the large increase in surface q over Alaska in winter was not matched in the winter W trends. This result for water vapor is consistent with previous studies showing increases in surface T in the Arctic (Chapman and Walsh 1993) that are not reflected in the free troposphere (Kahl et al. 1993). Strong and frequent Arctic boundary layer inversions may explain this decoupling of the surface and free-tropospheric trends in both temperature and water vapor.

Schönwiese and Rapp (1997) found surface vapor pressure increases over most of Europe of less than 0.5 mb during 1961–90. Based on the mean values they present, we estimate an overall European trend of about 1% decade⁻¹. Gutzler (1996) found an increase of about 3% decade⁻¹ in average 1000-hPa q from four radiosonde stations in the western tropical Pacific during 1973–93. Trends at 700 and 300 mb were larger in a percentage sense, but not as statistically significant. These upward surface q trends in the United States as well as other parts of the world, combined with observed upward W trends over North America (Ross and Elliott 1996b) and China (Zhai and Eskridge 1997), suggest a widespread moistening of the lower troposphere over the past several decades.

Figure 8 shows seasonal anomaly q time series (based on monthly anomalies using full-day data), along with the standard deviation of the station anomalies about the regional mean, and linear regression relations for each of the eight regions. Clearly, there is variability not attributable to the linear trends, which only account for between 5% and 20% of the total variance. The seasonal trends (Fig. 7) explain about 25%–50% of the total variance for those trends significant at the 0.01 confidence level, and about 10%–15% of the total for those trends significant at only the 0.05 level. The actual trends in these time series may have a more complex form than the linear model examined here. Possible other explanations for the additional, nonseasonal, variability will be explored in section 5b.

b. Dewpoint

As expected, T_d trends exhibit the same general spatial and temporal patterns as q trends. Consistent with the q trends, nighttime T_d increases generally exceed the daytime increases. To demonstrate the spatial pattern of the humidity trends, Fig. 9 shows the sign and statistical significance level of station trends in T_d for each season based on full-day data. The trends are generally positive and are strongest in Alaska in winter (with a maximum trend of 2.4 K decade⁻¹), in the western half of the country in spring (maximum of 1.5 K decade⁻¹), in the eastern half of the country in summer (maximum of 0.7 K decade⁻¹), and at the southernmost stations in fall (maximum of 0.6 K decade⁻¹).

The field significance (Wilks 1995) of these trends was determined using the resampling methods of Livezey and Chen (1983). At the 0.05 confidence level, we

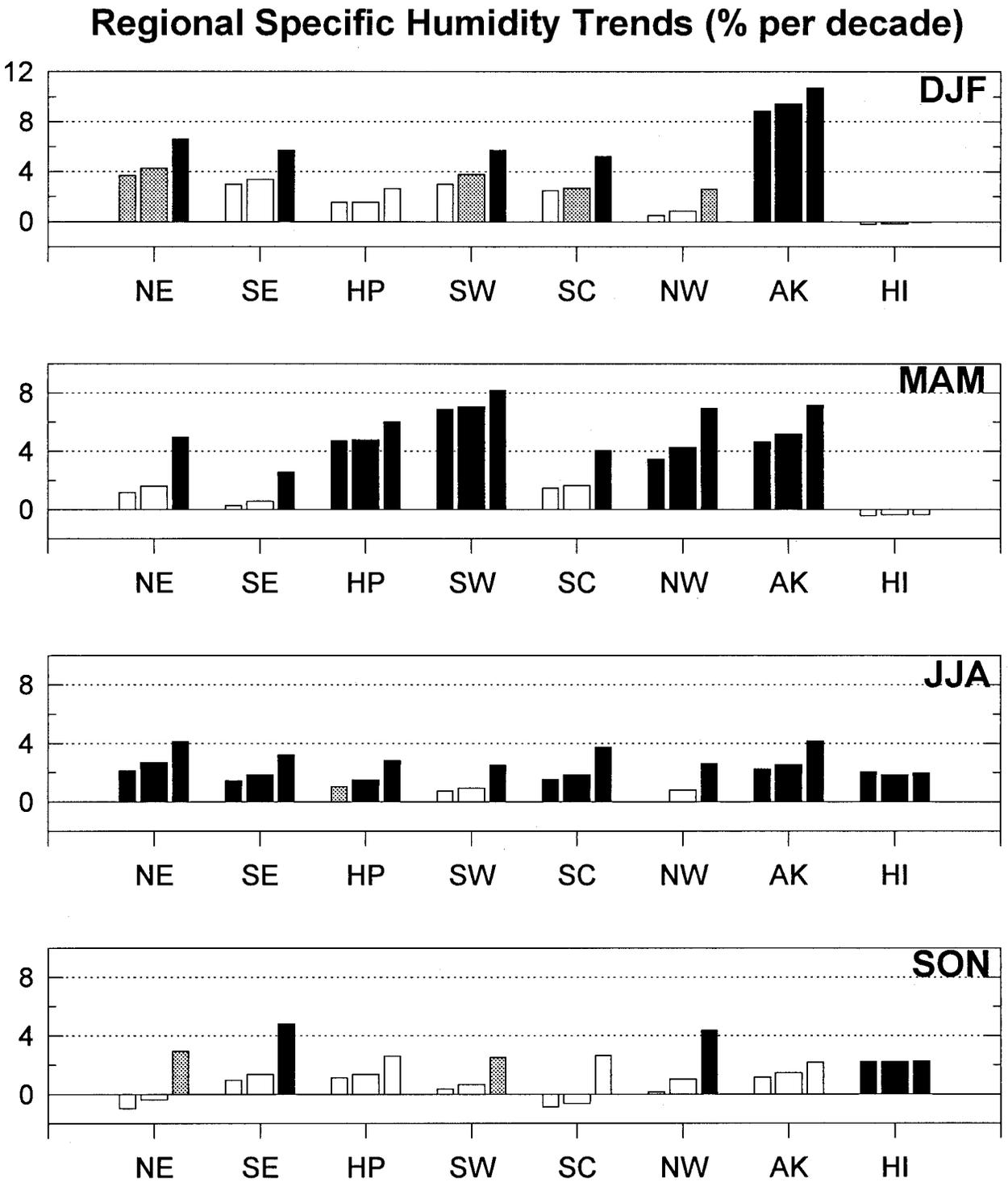


FIG. 7. Regional specific humidity trends for 1961–95 for daytime (left bar), full-day (wider middle bar), and nighttime (right bar) data. Seasonal trends have been divided by seasonal mean values to obtain trends in percent per decade. Trends significant at the 0.05 confidence level are shown in gray and those significant at the 0.01 level are shown in black.

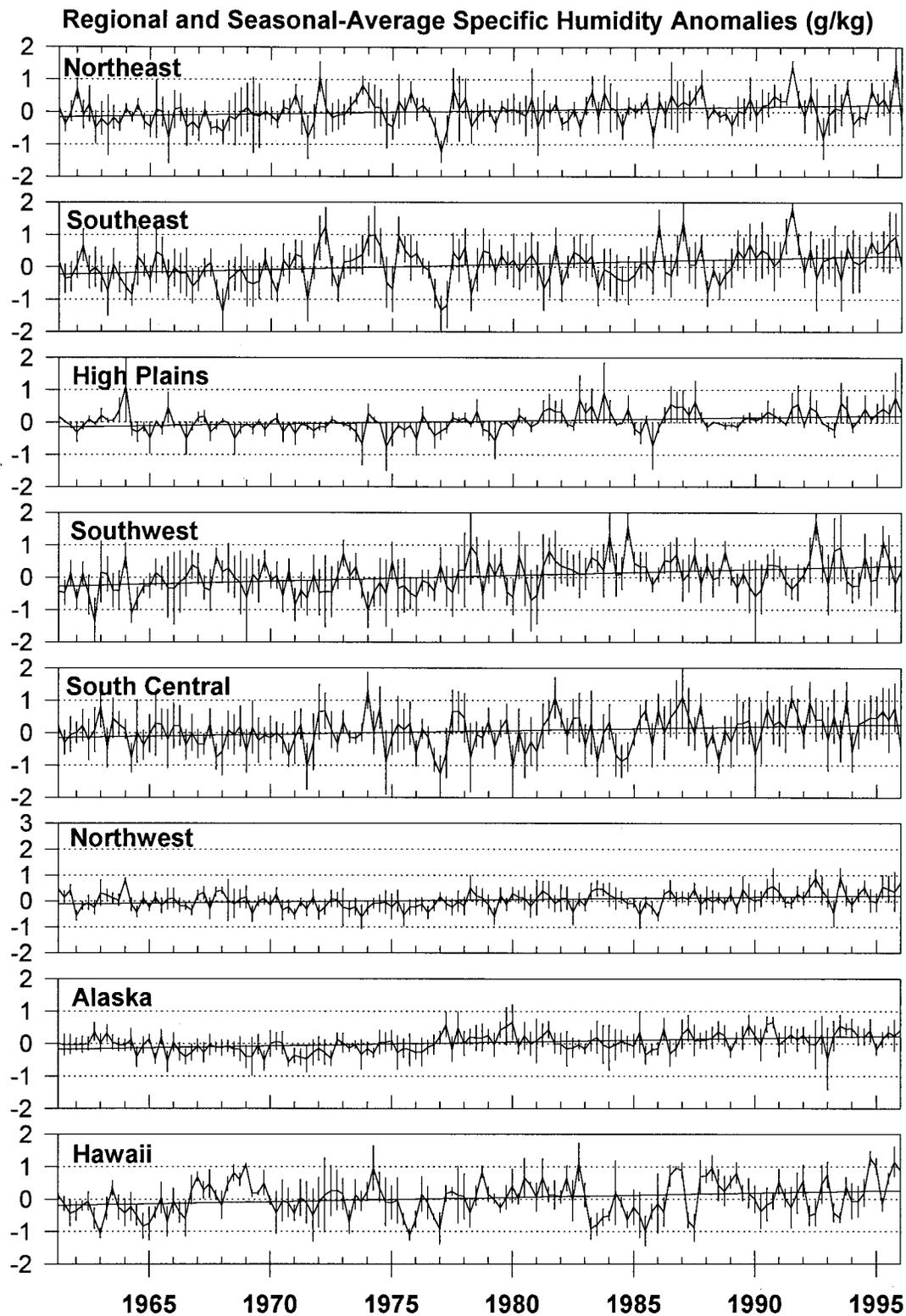


FIG. 8. Time series of seasonal anomalies of specific humidity (g kg^{-1}) for each of eight regions, ± 1 standard deviation of the station values about the regional means, and linear regression trend lines. Tick marks indicate the winter season at the start of the calendar year.



FIG. 9. Sign and statistical significance of trends, b , in dewpoint (K decade^{-1}) for (a) winter, (b) spring, (c) summer, and (d) fall based on full-day data. Statistical significance, indicated by the two-sided probability that the rank-order correlation between dewpoint and time is not different from zero, is shown for the 0.05 and 0.01 confidence levels.

are able to reject the hypothesis that the winter, spring, and summer trends shown in Figs. 9a–c are not significantly different from zero, due to possible spatial correlation among station time series. For fall (Fig. 9d), the number of significant station trends is just equal to the number required to reject this hypothesis at the 0.05 level. The spatial consistency of the station trends generally supports the grouping of stations into regions (Fig. 1) for trend computations. Two exceptions, where strong trends extend over a relatively small portion of the region, are the Northeast in spring and the Southeast in autumn.

Knappenberger et al. (1996) presented trends in T_d , for each hour of the day, for 15 U.S. stations for 1945–91, although most of the records had significant data gaps. They found increases in T_d for both day and night in all seasons except winter for the six-station western region and little change in the nine-station eastern region. The dissimilarity between those results and the trends shown in Fig. 9 could be due to the different data periods or the better data density and quality in the current study.

c. Temperature and apparent temperature

Regional temperature trends (Fig. 10) indicate warming of several tenths of a degree per decade over most of the United States in winter, spring, and summer. Fall shows less significant warming in most regions, and weak cooling in the Northeast and south-central regions. A comparison of daytime and nighttime T trends (Fig. 10a) shows some tendency for greater nighttime warming in fall and winter, but in spring and summer, the results are mixed. These results are somewhat at odds with the findings of asymmetric trends (with greater nighttime than daytime warming in all seasons, but especially in summer and fall) of Karl et al. (1993), who examined daily maximum and minimum T for the period 1951–90. Easterling et al. (1997) also report decreasing diurnal temperature range for the United States during 1950–93. However, we have examined trends over the period 1948–95 for 113 of the 170 stations in our network, and for that period we find stronger evidence for asymmetric trends, consistent with these previous studies. Nevertheless, it is important to note two substantial

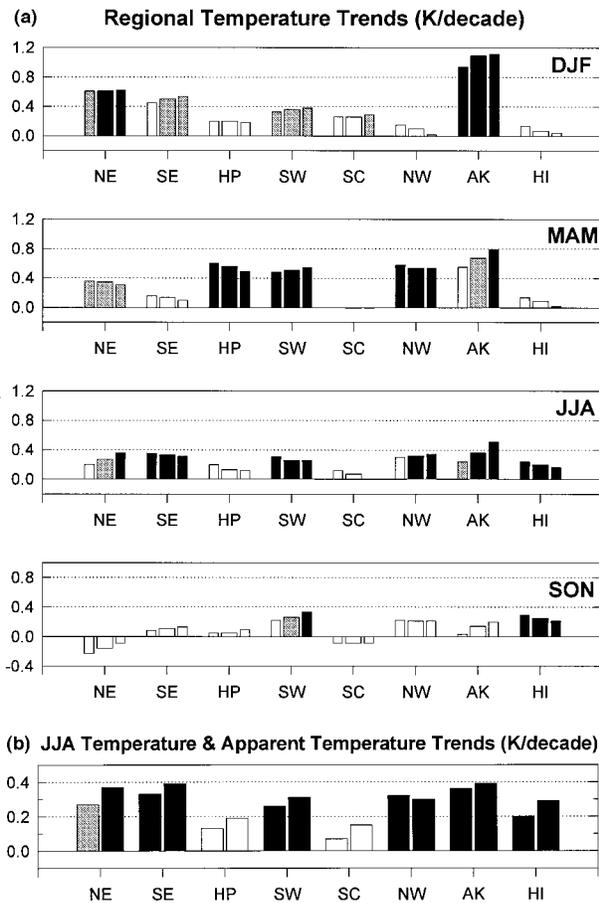


FIG. 10. (a) Regional temperature trends ($K \text{ decade}^{-1}$) for 1961–95 based on daytime (left bar), full-day (middle bar), and nighttime (right bar) data. (b) Regional trends in summertime temperature (left bar) and apparent temperature (right bar, $K \text{ decade}^{-1}$) based on full-day data.

differences between this study and those of Karl et al. (1993) and Easterling et al. (1997). First, we are not using minimum and maximum temperature data but rather daytime and nighttime average temperature, and our nighttime values probably do not include the minimum T , which tends to occur near dawn. Second, our data come from NWS first-order stations, whereas the other investigations use data from cooperative observing stations. In section 4a(1), we discuss possible biases due to instrument changes at the first-order stations.

These T increases are accompanied by increases in q (Fig. 9); however, the q trends are more often significant at the 0.01 level, especially at night. We also examined trends in apparent temperature T_a . As mentioned earlier, T_a is a measure of human heat stress in sultry weather. The summertime trends are shown to the right of the T trends in Fig. 10b. As would be expected from the joint q and T increases, the T_a trends tend to be slightly larger than the T trends. The largest difference is in the Northeast, where the T_a trend exceeds the T trend by $0.1^\circ \text{ decade}^{-1}$.

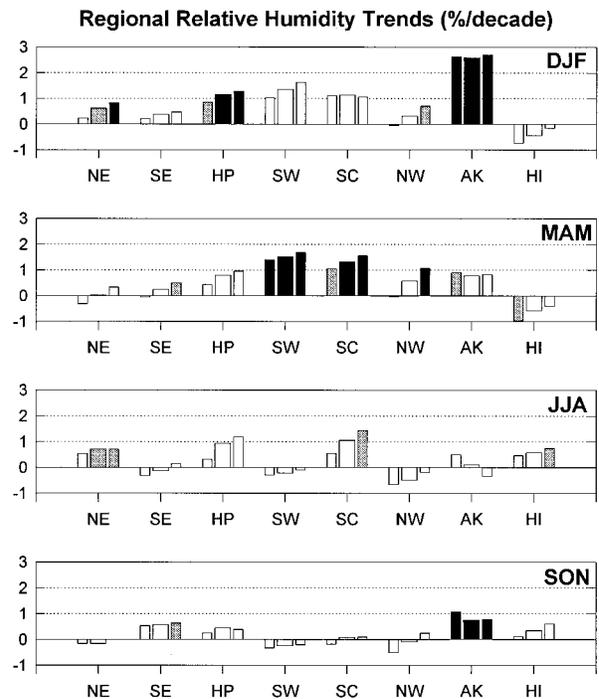


FIG. 11. Regional relative humidity trends ($\% U \text{ decade}^{-1}$) for 1961–95 based on daytime (left bar), full-day (middle bar), and nighttime (right bar) data.

Increases in T_a have implications for human comfort and heat-related illness and mortality, particularly if there have been changes in extreme values of T_a . In a separate study we have examined in more detail the climatology and trends in apparent temperature as well as the frequency of occurrence of high T_a values, and the duration of periods of high T_a , as an indicator of heat waves (Gaffen and Ross 1998).

d. Relative humidity

Trends in U at individual stations tend not to be significant at the 95% level and do not show the strong spatial consistency of the trends in q , T_a , and T . Consequently, the regional trends (Fig. 11) tend to be small and not statistically significant. However, there appears to be evidence for increases in U , especially at night, in winter and spring over most of the nation, with the most striking increase in Alaska and the high plains in winter, and in the Southwest and south-central regions in spring.

These surface trends differ in both seasonality and spatial pattern from tropospheric trends presented by Elliott and Angell (1997), who analyzed radiosonde data (at 0000 and 1200 UTC) for 1973–93 for the conterminous United States. They found seasonal 850-mb U trends were mostly not significant at the 5% level, except for an increase in summer of $2.8\% \text{ decade}^{-1}$ for the United States. On annual average, the only signif-

icant trends were in their south-central (2.1% decade⁻¹) and Southeast (1.6% decade⁻¹) regions, which do not have the same boundaries as the corresponding regions in this study. The differences in these results could be due to different stations and regions, different record lengths (21 vs 35 yr), different times of observation (twice daily vs eight times daily), or different instrumentation (radiosonde vs surface instruments). On the other hand, both studies find stronger evidence for increases in the water vapor content of the atmosphere, in this case q and in Elliott and Angell (1997) precipitable water, than for increases in U .

5. Discussion

a. Temporal homogeneity of station data

The trends presented here may not be representative of true regional humidity changes if the station data contain temporal inhomogeneities. In this section we consider two potential influences on the quality and continuity of the data. The first is abrupt changes in data biases due to changes in instrumentation or station locations. The second involves gradual developments both globally and near observing stations.

1) POSSIBLE ABRUPT CHANGES

Changes in observing methods and station moves could introduce abrupt changes in the time series. Station histories obtained from the National Climatic Data Center indicated that about 20% of the stations moved more than 1 mi during the period of this analysis, although some of the station history information was ambiguous. We visually examined time series of monthly anomalies of q , U , T , and T_d for those stations and found no evidence of abrupt changes associated with the moves.

Two major changes in U.S. surface observing methods could have influenced the data. These are the change from sling psychrometers to hygrothermometers in the early 1960s and the change to dewpoint hygrothermometers (model HO-83) in the mid-1980s (Elliott 1995). The more recent introduction of ASOS technology did not involve a sensor change. However the HO-83 system was modified starting in 1991, to reduce a warm bias in both daytime and nighttime observations (Jones and Young 1995). Station history information gave some, but not all, of the dates on which these changes were made. Karl et al. (1995) suggest that the change to the HO-83 hygrothermometers in the 1980s may have lead to spurious increases of 0.5°C in daily maximum temperature and “maybe” 0.1°C in daily minimum temperature, but that “the exact bias of any one instrument is unknown.” To our knowledge, no comparable assessments have been made of the effect of these instrument changes on humidity measurements.

We attempted to assess the impact of these changes

in four ways. First, we visually examined monthly anomaly U , T , and T_d time series for about half of the stations in search of obvious data discontinuities. We explicitly looked for level shifts at the time of instrument changes. None was found.

Second, we compared monthly anomalies of T and T_d before and after the nominal instrument change dates of mid-1964 and mid-1985. These dates were chosen because exact dates were not known for all stations. In reality, the changes were implemented over the course of several years throughout the U.S. network. Values for the 4-yr periods preceding and following these dates were compared using two techniques. One was a parametric two-sided t test of the mean values (Jaruskova 1996), and the other was a nonparametric rank-order test for the difference in medians (Lanzante 1996). All 170 station time series of both T and T_d anomalies were tested. Both tests yielded identical results regarding the existence of a level shift in all time series tested. At the 99% confidence level, for the nominal 1964 change, the mean (and median) T anomaly values were not significantly different for 94% of the stations, when data for all hours of the day were used. For daytime and nighttime data, 97% and 87% of the stations, respectively, showed no significant difference. For the nominal 1985 change, the data for all hours, daytime, and nighttime showed no significant differences at 89%, 90%, and 91% of the stations, respectively. However, of the remaining 9%–11% of the stations, almost all showed higher T following 1985 than before. The T_d results were similar to the T results, with one exception. About 23% of the stations, in the eastern United States, Alaska, and Hawaii, had significantly higher nighttime T_d following 1985 than before. However, more detailed analysis showed the result to hold mainly for the warm season (April–September), and is likely due to the moist summers of 1986 and, to a lesser extent, 1989, with drier years in between. Thus the change-point test results are not likely to be indicative of an artificial level shift, but rather a climatic fluctuation. Furthermore, the modification of the HO-83 beginning in 1991 was to reduce a warm bias (Jones and Young 1995), so any warm bias should have been reduced in the final years of our records.

Third, we compared trends based on daytime and nighttime data. If a larger upward daytime than nighttime T shift accompanied the introduction of the HO-83 instruments, as was estimated by Karl et al. (1995), we would expect larger daytime T trends for day than for night. However, as mentioned above, the asymmetries in the daytime and nighttime trends were not large, but they were in the opposite sense of what we would expect if the result were due to the bias in the HO-83 sensor. From this we conclude that any bias due to instrument changes at the first-order stations is less than the differential warming signal we would expect from the cooperative station results of Karl et al. (1993), although some bias may have reduced the asymmetry of

the trends in our data. We note, however, that, in this analysis, the nighttime humidity trends are greater than the daytime ones, and the difference is more pronounced than for temperature. It is possible that the humidity trend asymmetry is a manifestation of larger actual temperature trend asymmetry than our T data show.

Fourth, we compared trends for the period of homogeneous data (1966–83) to those for 1961–90, which includes the two main instrument changes, and found no consistent pattern to the differences. Furthermore, the median of pointwise slope method of trend estimation (Lanzante 1996) is less influenced by data at the ends of the time series than linear regression. This suggests that the consistency of the trends for the longer period is not due to events (such as instrument changes) near the ends of the period.

On the basis of these four findings, we conclude that, in general, the instrument changes did not have significant effects on the time series. It is possible that subtle effects beyond our detection have some influence on the computed trends. However, in the absence of reference stations making well-calibrated, highly precise observations for a sustained period, the first-order weather station data provide the best record of surface-level moisture.

2) POSSIBLE GRADUAL CHANGES

Gradual changes in the surface humidity patterns due to anthropogenic sources of water vapor are more difficult to detect and quantify. In this section, we consider three possible effects: water vapor emissions associated with fossil fuel use in general, water vapor emissions associated with aviation, and irrigation.

The consumption of fossil fuels produces both carbon dioxide and water vapor as combustion products. This anthropogenic water vapor source could influence background water vapor levels. Based on carbon emissions data (Marland et al. 1994), we estimate global water vapor emission from fossil fuel consumption to be of order 10^{12} (in 1960) to 10^{13} (in 1990) kg yr^{-1} . Evaporation contributes of order 10^{17} kg yr^{-1} to the atmosphere (van der Leeden et al. 1990). Thus, on a global basis, evaporation from the surface far exceeds anthropogenic water vapor emissions. Given the fast cycling of water vapor in the atmosphere, it seems unlikely that the global anthropogenic source can account for the observed trends of several percent per decade.

Next we consider a more localized water vapor source that may affect our data. The vast majority of the stations we have examined are at airports, some of which may have expanded during the last three decades. It may be possible that increased air traffic, and consequent increased jet fuel combustion and water vapor emissions, could account for the observed trends. Using data on U.S. aircraft fuel consumption trends, water vapor emissions from fuel consumption, and the height profile of fuel consumption from Baughcum et al. (1996), we es-

timate that recent aircraft fuel consumption represents a source of water vapor to the planetary boundary layer over U.S. airports of order 10^2 kg s^{-1} . This is about three orders of magnitude smaller than typical horizontal advective fluxes. (The latter were estimated to be of order 10^5 kg s^{-1} , assuming air with specific humidity of 7 g kg^{-1} and winds blowing at 5 m s^{-1} through a planetary boundary layer 1 km in height over an airport of horizontal dimension 10 km .)

Locally, anthropogenic modification of the hydrological cycle may be more important. Within the conterminous United States, the U.S. Geological Survey has estimated that consumptive use of water in agricultural irrigation contributes 100 billion gallons of water per day to the atmosphere, compared with 2800 billion gallons per day from evaporation and transpiration from surface water bodies, land surface, and vegetation (van der Leeden et al. 1990). In dry regions during the growing season, the ratio of consumptive use to natural evaporative sources may be greater, and it is possible that long-term increases in evaporation from irrigated fields may be large enough to influence the surface trends at some stations. Other confounding influences may affect the trends presented here. However, the spatial consistency of the trends leads us to speculate that they are not primarily due to local phenomena but represent regional, indeed national, increases in near-surface specific humidity.

b. Surface humidity and atmospheric circulation

Changes in surface humidity, such as reported above, could be caused by, or at least linked to, changes in atmospheric circulation patterns. Changes in the long-wave patterns, dominant airmass types, or strength or position of climatological “centers of action” will have important influences on local humidity and temperature regimes. In the Southeast, for example, more frequent spring and summer incursions of maritime tropical air masses could explain the observed trends. The linear correlation between T and q anomalies, shown in Fig. 12, indicates that in the eastern United States, along the west coast, and in Alaska, q and T share about 50% common variance. In most of the West, and in Hawaii and Puerto Rico, variations in T explain less than one-fourth the variation in q .

To assess the influence of large-scale dynamics on interannual humidity variations, we examined indices of aspects of atmospheric circulation patterns: the Southern Oscillation index (Ropelewski and Jones 1987, extended), North Pacific index (Trenberth and Hurrell 1994), and the North Atlantic Oscillation index (Hurrell 1995). Using singular value decomposition and minimizing chi square, we tested whether monthly q anomalies for each of the eight regions, individually, could be expressed in terms of a linear combination of monthly values of these three indices. In all eight cases, evalu-

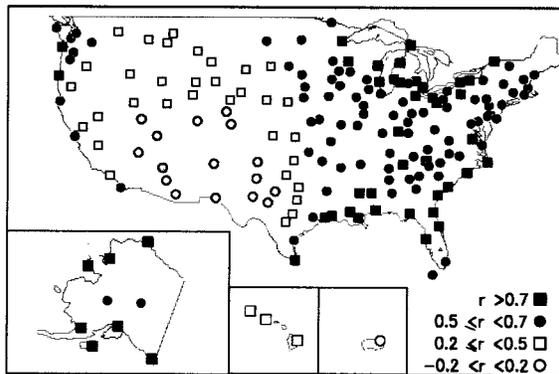


FIG. 12. Coefficient of linear correlation, r , between monthly anomalies of temperature and specific humidity. The values of $|r| < 0.2$ are generally not significant at the 0.05 confidence level.

ation of the chi-square probability function indicated the fit to be without statistical significance.

Similar results were obtained by linear regression of the regional monthly q anomalies, at lags of 0 to 12 months, with each index separately. Specific humidity anomalies in all regions are essentially uncorrelated with both the Southern Oscillation index and the North Atlantic Oscillation index for all lags tested. In the Hawaii, Northeast, Southeast, and Southwest regions there are small negative correlations with the North Pacific index that are significant at the 0.05 level or better. These results suggest that the interannual variation in near-surface q are not highly influenced by these large-scale dynamical features, but that when sea level pressure in the North Pacific region is above average, surface humidity in some parts of the United States may be below average.

c. Related trends

During the later years of our data period (1981–91), and especially in Alaska, Myneni et al. (1997) report a lengthening of the growing season, in association with springtime warming, earlier disappearance of snow, and an increase in the amplitude of the seasonal cycle of atmospheric carbon dioxide. For the period 1973–92, Groisman et al. (1994) found significant decreases in North American snow cover, associated with increasing surface T , especially in spring. Springtime warming, of both daily minimum and maximum T , was also detected by Karl et al. (1993) for the period 1951–90. Cooter and LeDuc (1996) have documented trends toward earlier dates of final spring frosts in New England during 1961–90. Angell (1998) showed a contraction of the 300-hPa circumpolar vortex in the Northern Hemisphere during 1963–96, particularly over North America, associated with tropospheric warming. Together these results suggest that springtime conditions are becoming more mild. The q , T , and T_d trends presented here are consistent with this interpretation.

Our findings of increased humidity at night could explain, at least in part, the diurnal temperature range trends found by others (Karl et al. 1993; Easterling et al. 1997). The enhanced greenhouse effect of water vapor at night may reduce nocturnal cooling and lead to increases in nighttime T , minimum T , or both. Increased nighttime U could also contribute to increased cloudiness, which, as suggested by Dai et al. (1997), could explain diurnal temperature range decreases.

6. Summary

We have computed climatological monthly and seasonal values of temperature, dewpoint, specific humidity, and relative humidity for 188 U.S. stations for the period 1961–90. Separate climatologies were calculated using daytime observations (0800–1600 LST), nighttime observations (2000–0400 LST), and eight observations representing the full day. Maps of seasonal and annual means, and the range and phase of the mean annual cycle are presented here. The monthly data used to create the maps are available from the authors.

The time series were extended using data for 1961–95 for 170 stations, and trends in these same variables were calculated for each station and for eight regions of the country. In general, we find increases of several percent per decade in specific humidity (and several tenths of a degree per decade in dew point) over most of the country in winter, spring, and summer, with larger trends at night than during the day. The specific humidity increases are consistent with upward temperature trends. Relative humidity trends are weaker than the specific humidity trends, but they do show evidence of increases, especially in winter and spring. The upward temperature and humidity trends are also consistent with upward trends in apparent temperature.

We have examined the possibility that the detected humidity trends may be artifacts of changes in instrumentation, but several lines of reasoning suggest that they are not. Fossil fuel burning involves emission of water vapor, but our estimates indicate that this source is too small to explain the observed trends. Locally, but probably not nationally, irrigation may be a contributing factor.

Acknowledgments. We thank Tom Ross of the National Climatic Data Center for supplying the meteorological data for this study. This work was supported, in part, jointly by the NOAA Office of Global Programs and the NASA Mission to Planet Earth.

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