Are Temperature and Precipitation Extremes Increasing over the U.S. High Plains?

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Received 4 April 2012; accepted 14 September 2012

ABSTRACT: Large-scale environmental, social, and economic impacts of recent weather and climate extremes are raising questions about whether the frequency and intensity of these extremes have been increasing. Here, the authors evaluate trends in climate extremes during the past half century using

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DOI: 10.1175/2012EI000454.1
the U.S. High Plains as a case study. A total of eight different extreme indices and the standardized precipitation index (SPI) were evaluated using daily maximum and minimum temperature and precipitation data from 207 stations and 0.25° gridded data. The 1958–2010 time period was selected to exclude the 1950s and 2011 droughts. Results show general consistency between the station data and gridded data. The annual extreme temperature range (ETR) decreased significantly ($p < 0.05$) in $\sim 54\%$ of the High Plains, with a spatial mean rate of $-0.7^\circ\text{C decade}^{-1}$. Decreases in ETR result primarily from increases in annual lowest temperature in $\sim 63\%$ of the stations at a mean rate of $-0.9^\circ\text{C decade}^{-1}$, whereas increases in annual highest temperature were much less. Approximately 43% of the stations showed increasing warm nights ($T_{\text{min}} > 90$) with a spatial mean rate of 0.5% decade$^{-1}$. Precipitation intensity generally did not vary significantly in most grid cells and stations. Significant decreasing trends in consecutive dry days (CDD) were restricted to 21% of the stations in the northern High Plains with a spatial mean of $-0.8$ days decade$^{-1}$. Areas experiencing severe dry periods (1-month SPI $< -1.5$) decreased over time from 8% to 4%. The number of dry months (SPI $< 0$) in each year also decreased. In summary, the ETR is decreasing and low temperatures are increasing. Precipitation extremes are generally not changing in the High Plains; however, high natural climatic variability in this semiarid region makes it difficult to assess climate extremes.

**KEYWORDS:** Climate extremes; High Plains; Precipitation; SPI; Drought

1. **Introduction**

Changes in extreme weather and climate have large-scale socioeconomic and environmental impacts. The 2011 drought in Texas and surrounding states was the most severe 1-yr drought in the instrumental record (1895–2011) and was associated with record high temperatures (40 days $> 100^\circ\text{F}$ in the central, western, north, and north-central parts of the state) (http://www.ncdc.noaa.gov/sotc/drought/2011/8). Reported economic losses totaled $7.6$ billion (Fannin 2012). The United States has been subjected to 133 weather/climate disasters since 1980 in which overall damages/costs exceeded $1$ billion, totaling $875$ billion (http://www.ncdc.noaa.gov/billions). However, comparison of trends in these societal impacts (economic losses and fatalities) with trends in atmospheric climate forcing suggests that increasing economic losses reflect rising societal vulnerability from demographics and lifestyle changes rather than weather/climate trends (Kunkel et al. 1999).

Weather and climate extremes are projected to increase with climate change. Large shifts in extreme temperature and precipitation can result from small shifts in mean values (Karl et al. 2008). U.S. average temperatures increased by $\sim 2^\circ\text{F}$ during the past 50 years (Karl et al. 2009); however, annual record high maximum temperatures exceeded record low minimum temperatures by a ratio of $\sim 2$ during the past decade (Meehl et al. 2009). Because rising temperatures increase atmospheric water holding capacity, rainfall intensities are projected to increase when it rains (Trenberth et al. 2003). While precipitation in the United States increased on average about 5% from 1950 to 2000, there was a disproportional increase in heavy precipitation events with up to 67% increase in the heaviest 1% of daily events in the northeastern United States from 1958 to 2007 (Karl et al. 2009). Increases in intensity may be offset by reductions in duration or frequency of events (Trenberth et al. 2003).
Various approaches have been used to describe extreme events. Frich et al. (Frich et al. 2002) described 10 different indices for quantifying variations in frequency and intensity of extreme events, 5 related to temperature and 5 related to precipitation. These indices include total number of frost days \( (T_d) \), intra-annual extreme temperature range (ETR), growing season length (GSL), heat wave duration index (HWDI), warm nights \( (T_{\min,90}) \), number of days with precipitation larger or equal to 10 mm day \(^{-1}\) (R10), maximum number of consecutive dry days (CDD), maximum 5-day precipitation total (R5d), simple daily intensity index (SDII), and fraction of annual total precipitation due to events exceeding the 95th percentile for the period 1961–90 (R95T).

The indices are calculated based on station data for analysis of observational records, and the results are often gridded to evaluate spatial variations in trends in extremes (e.g., Alexander et al. 2006). Gridded data can also be used instead of station data for analyses of observational trends and gridded data are always used for evaluation of model outputs. The climate extreme indices have been used in global analyses of extreme events based on observational data and models (e.g., Alexander et al. 2006; Frich et al. 2002; Tebaldi et al. 2007). Results based on 2223 temperature and 5948 precipitation stations globally from 1951 to 2003 showed a decrease in cold nights and an increase in warm nights over 70% of the land area and similar but lower increases in daily maximum temperatures than in daily minimum temperature (Alexander et al. 2006). Precipitation showed widespread increases, but precipitation changes are less spatially coherent. Some studies have focused on threshold exceedances for different event durations (e.g., 1 and 7 days) and different return periods (e.g., 1 and 5 years) (e.g., Kunkel et al. 1999). The analysis is often represented as percentile exceedances that also correspond to return period. Raw exceedance counts are sometimes shown as anomalies relative to the climatology for different periods (e.g., 1961–90, 1971–2000, and 1981–2010). Trends can be calculated directly using station data or using spatially averaged data. Linear trends are generally used based on least squared regression or Kendall slope method; the latter does not require any assumptions about statistical distribution. The beginning and ending periods are important for trend estimation. For example, trends calculated from 1950 to 2000 in the United States may be biased because much of the United States was subjected to extreme drought in the 1950s. The standardized precipitation index (SPI) is one of the commonly used indices for drought monitoring and can provide information on precipitation surplus or deficit relative to mean precipitation for a specific time period (McKee et al. 1993). This index can also be used as an indicator for extreme dry and wet conditions (e.g., SPI < -1.5 and SPI > 1.5).

The objective of this study was to assess trends in climate extremes using the U.S. High Plains (HP) as a case study. The analysis is based on station data (207 stations) for precipitation and temperature and gridded data (0.25° grid) for precipitation. The HP is a critical region because of the high agricultural productivity that is vulnerable to climate extremes. The HP has a large north–south extent, representing a temperature gradient, and there is also a large precipitation gradient mostly from west to east. The time period from 1958 to 2010 was selected for trend analysis to avoid the droughts in the 1950s and in 2011. The precipitation data allow comparison of use of station versus gridded data for trend analysis.
2. Methods and materials

2.1. U.S. High Plains

The U.S. HP (450,000 km²) is a subregion of the U.S. Great Plains, generally encompassing the western part of the Great Plains. The HP extends from southeast Wyoming and southwest South Dakota to northwest Texas. Elevation of the HP varies from ~350 m in the east to ~2400 m in the west. Mean annual precipitation is 520 mm (semihumid climate) and ranges from 338 mm in the northwest to 831 mm in the central-east HP [1958–2010; National Oceanic and Atmospheric Administration (NOAA)/Climate Prediction Center (CPC) 0.25° gridded daily data] (Figure 1a). There are no significant changes in long-term annual precipitation in 90% of the stations with only 10% of stations showing increasing trends at a rate of 26 mm decade⁻¹ (F statistic; p < 0.05) (Figure 1b). In general, atmospheric moisture from the Gulf of Mexico results in precipitation concentrated mostly in months from April to September. A grassland biome dominates this region.

The HP is one of the most productive agricultural regions in the United States (Scanlon et al. 2012). Agriculture, including livestock and crops (e.g., corn, wheat, and cotton), is the primary economic activity in the region, with agricultural products totaling $35 billion relative to a U.S. total of $300 billion in 2007. Agricultural productivity relies heavily on groundwater from the HP aquifer for
irrigation because of the dry climate. Irrigation has caused large-scale groundwater depletion (~340 km$^3$) in the HP aquifer from the 1950s to 2009 (McGuire 2009).

2.2. Methods

Four metrics for extreme temperature and four metrics for extreme precipitation proposed by Frich et al. (Frich et al. 1996; Frich et al. 2002) and widely used in the literature (e.g., Alexander et al. 2006; Tank and Konnen 2003) were selected to examine climate extremes over the HP. The temperature-related indices include the following: 1) ETR (°C), 2) HWDI (days), 3) number of days with exceeding 100°F ($T_{\text{max}100}$; days), and 4) $T_{\text{min}90}$ (%). The ETR was calculated from the difference between the highest daily maximum temperature ($T_h$) and the lowest daily minimum temperature for the same calendar year ($T_l$). The HWDI is defined as the maximum period of at least 5 consecutive days with daily maximum temperature ($T_{\text{max}}$) higher by at least 5°C than the climatological norm (1961–90). The $T_{\text{max}100}$ was calculated by counting the number of days with $T_{\text{max}} > 100°F$ in a year. The $T_{\text{min}90}$ was calculated by the percentage of times in a year when daily minimum temperature ($T_{\text{min}}$) is above the 90th percentile of the climatological distribution (1961–90) for that calendar year.

Rainfall-related indices include the following: 1) CDD (days; dry day is defined as $R_{\text{day}} < 1$ mm where $R_{\text{day}}$ is the daily precipitation), 2) maximum number of consecutive wet days (CWD; days; a wet day is defined here as $R_{\text{day}} \geq 1$ mm), 3) R95T (%), and 4) SDII (mm day$^{-1}$). CDD and CWD (days) were calculated from the maximum number of consecutive dry days and wet days in a year, respectively. Regarding the R95T, the 95th percentile for 1961–90 is the mean of the 95th percentiles for each year using all wet days (Frich et al. 2002). The R95T was then calculated from the amount of precipitation for the year due to events exceeding the 95th percentiles of the 1961–90 period divided by annual total precipitation. SDII was calculated from the annual total precipitation divided by the number of wet days in a year.

Precipitation is not normally distributed. Calculation of SPI involves transforming monthly precipitation into a standard normal distribution (mean of 0 and standard deviation of 1). In this study, the 1-month SPI was calculated from monthly precipitation accumulated from daily precipitation. The 1-month SPI map is similar to a map of the percent of normal precipitation for a month. Because the 1-month SPI reflects relatively short-term conditions, its application can be closely related to short-term soil moisture and crop stress, especially during the growing season (Guttman 1999; Quiring 2009). A positive SPI value indicates greater than median precipitation, whereas a negative value indicates less than median precipitation.

Ordinary least squares regression (OLSR) with 95% or 90% confidence intervals (both were tested but only 95% confidence intervals are shown in the figures) was used for trend analysis for the time series of climate extreme indices. The non-parametric Kendall’s tau-based slope estimator was also used and showed similar results as OLSR. Discussion of the results is based on OLSR with 95% confidence intervals.
2.3. Data sources

Daily maximum and minimum temperature data from 207 NOAA meteorological stations for 1958–2010 with less than 10%–20% missing data were used to calculate temperature indices within the HP (Figure 1). To maximize the number of stations used for analysis of $T_{\text{max}100}$, stations with 60% of the missing data concentrated in the shoulder seasons (January–March and October–December) were included in the analysis.

Daily precipitation data from the 207 NOAA stations and also from the NOAA/CPC $0.25^\circ \times 0.25^\circ$ daily U.S. unified precipitation data were used to calculate precipitation extremes on a station and grid basis. Station time series with less than 10% missing records annually were used to calculate precipitation extremes for that year. Precipitation extremes from stations and grids were compared to evaluate the two data sources. When calculating trends stations with at least 80% complete data for the study period were used. Trends were calculated for the period between 1958 and 2010 to exclude the 1950s and 2011 extreme droughts because such extreme droughts can impact trends (e.g., Easterling et al. 2000; Kunkel et al. 1996).

3. Results and discussion

3.1. Trends in extreme temperature indices from station data

ETR decreased significantly across ~54% of the stations (72 out of 134 stations) within the HP (Figure 2 and Table 1). The $T_i$ increased significantly in 63% of the stations (84 out of 134), with a slope of ~0.9°C decade$^{-1}$ for the 84 stations (Figure 3). However, increasing trends in $T_h$ were much lower than those in $T_i$, with some stations (17 stations) even showing statistically significant decreasing trends. For 69 stations with increasing trends in $T_i$ and $T_h$, the average ratio of the increasing trends of $T_i$ relative to $T_h$ is ~2. Therefore, decreasing ETR over the HP resulted mostly from the increase in $T_i$ and partly from the decrease in $T_h$.

HWDI did not show significant trends in most stations (73% of stations: 88 out of 121 stations) with significant trends restricted to the northeast of the northern High Plains (NHP) and east-central High Plains (CHP) at a mean rate of about ~2 days decade$^{-1}$. Increasing trends in $T_{\text{min}90}$ were found in 43% of stations (51 out of 120) indicating increases in warm nights with an overall rate of 0.5% decade$^{-1}$ for the 120 stations. There is no trend in $T_{\text{max}100}$ for 76% of the stations (54 out of 71 stations). A total of seven stations show statistically significant decreases in $T_{\text{max}100}$, six of which are in the NHP. Indices associated with low temperatures at daily and annual time scales (e.g., $T_{\text{min}90}$ and $T_i$) show more significant changes than those associated with high temperatures (e.g., $T_{\text{max}100}$ and $T_h$). This finding is consistent with that in Alexander et al. (Alexander et al. 2006).

3.2. Trends in spatial means of extreme precipitation indices from gridded data

There is no significant decreasing trend in CDD over the CHP and southern High Plains (SHP) ($F$ statistic, $p = 0.28$ CHP and $p = 0.78$ SHP). However, there is a
Figure 2. Trends in ETR (°C decade⁻¹), HWDI (days decade⁻¹), $T_{\text{max}100}$ (days decade⁻¹), and $T_{\text{min}90}$ (% decade⁻¹) within the High Plains for the period 1958–2010 using station temperature data. Filled blue and red dots denote statistically significant trends ($F$ statistic; $p < 0.05$). Open circles denote no statistically significant trend.
significant decreasing trend across the NHP at $-1$ day decade$^{-1}$ and the anomaly decreasing at $-3\%$ decade$^{-1}$ (Figure 4). The spatial mean of CWD over all sub-regions of the HP does not show any significant decreasing trend for the study period. CDD and CWD are not closely correlated, which means that a higher CDD value does not mean a lower value of CWD. CDD can be indicative of effects on vegetation and ecosystems, and is a potential drought indicator. A decrease in CDD would reflect a wetter climate if the change was due to more frequent wet days (Frich et al. 2002). Pryor et al. (Pryor et al. 2009) found that the number of precipitation days (i.e., precipitation exceeding 1.27 mm) increased significantly across the NHP over the twentieth century. For the NHP, the reduced CDD therefore indicates a tendency toward wetter conditions over the study period.

There is no trend in spatial means of R95T and SDII time series for the NHP and SHP, with only the CHP showing statistically significant increasing trends in the spatial mean and anomaly of R95T at rates of $\sim$1% decade$^{-1}$ and $\sim$3% decade$^{-1}$ and of SDII at rates of $\sim$0.2 mm m$^{-1}$ day$^{-1}$ decade$^{-1}$ and $\sim$2% decade$^{-1}$.

### 3.3. Spatial distributions of mean annual extreme precipitation indices from gridded data

Large mean annual CDD values are generally concentrated in the SHP and CHP (spatial mean of 45.9 and 42.6 days, respectively): that is, the region with a semiarid and arid climate (Figure 5). The NHP shows a generally homogeneous distribution of CDD of 36.5 days. Mean annual CWD across the HP is generally homogenous spatially, with a value of $\sim$5.3 days. SDII varies markedly between the western and eastern HP, which is generally consistent with the distribution of mean annual precipitation (with reference to Figure 1a), with mean annual precipitation being more spatially homogenous than SDII. Mean annual R95T is more spatially heterogeneous than the other three indices.
3.4. Trends in gridded and station time series of extreme precipitation indices

3.4.1. Gridded extreme precipitation indices

CDD does not show significant trends in 84% of grid cells (635 out of 754) within the HP with decreasing trends in the remaining ~16% of grid cells that are restricted mostly to the NHP (Figure 6). Significant decreasing trends in CDD in the west HP neighboring the Rocky Mountains were also found. There are only five grid cells in the NHP showing statistically significant decreasing trends in CWD. Grid cells with decreasing CDD in the northwest of the HP exhibited a statistically significant increasing trend in CWD.

There is no statistically significant trend in R95T and SDII over the NHP and SHP. Significantly increasing trends in R95T were restricted to 14% of the grid cells (30 out of 215) in the CHP (spatial mean of 0.7% decade\(^{-1}\)) and increasing trends in SDII were restricted to 25% of grid cells (54 out of 215) in the CHP (spatial mean 0.1 mm day\(^{-1}\) decade\(^{-1}\)). These results indicate that areas of increasing precipitation intensity are limited to some parts of the CHP.

3.4.2. Station extreme precipitation indices

There is general consistency between station-based and grid-based trends (Figures 6, 7). Data in the NHP show significantly decreasing trends in CDD in
Figure 4. (a),(c),(e),(g) Trends in time series and (b),(d),(f),(h) anomalies of CDD, CWD, R95T, and SDII across the northern, central, and southern High Plains for the period 1958–2010.
21% of stations (25 out of 121) (mean slope of $-0.8$ days decade$^{-1}$ across the 121 stations) relative to $\sim24\%$ of grid cells (mean slope of $-1.1$ days decade$^{-1}$). Approximately 12% of stations (7 out of 57) across the CHP show statistically increasing trends in R95T (mean slope of 0.8% decade$^{-1}$) relative to 14% for
Figure 6. Trends in CDD (days decade$^{-1}$), CWD (days decade$^{-1}$), R95T (% decade$^{-1}$), and SDII (mm day$^{-1}$ decade$^{-1}$) over the High Plains and adjacent areas for the period 1958–2010 using gridded precipitation data. Upward-pointing triangles denote statistically significant increasing trends, and downward-pointing triangles denote statistically significant decreasing trends ($F$ statistic; $p < 0.05$).
Figure 7. Trends in CDD (days decade$^{-1}$), CWD (days decade$^{-1}$), R95T (% decade$^{-1}$), and SDII (mm day$^{-1}$ decade$^{-1}$) within the High Plains for the period 1958–2010 using station precipitation data. Filled blue and red dots denote statistically significant trends ($F$ statistic; $p < 0.05$). Open circles denote no statistically significant trend.
gridded data (mean slope of 0.7% decade$^{-1}$), and $\sim$19% of stations (11 out of 57) across the CHP show statistically increasing trends in SDII (mean slope of 0.2 mm day$^{-1}$ decade$^{-1}$) relative to 25% for gridded data (mean slope of 0.1 mm day$^{-1}$ decade$^{-1}$).

### 3.5. Comparison of the results with other studies

Reduced ETR and a more pronounced increase in $T_I$ relative to $T_h$ over $\sim$50% of stations in the HP are consistent with previous studies with a general trend of increasing $T_I$ at a greater rate relative to $T_h$ globally (e.g., Alexander et al. 2006; Easterling et al. 1997; Jones et al. 1999; Russo and Sterl 2011) and regionally like Europe and China (e.g., Hu et al. 2010; Tank and Konnen 2003; Yan et al. 2002; Yao et al. 2008), Southeast Asia and the South Pacific (Manton et al. 2001), Canada (Vincent and Mekis 2006), South America (Vincent et al. 2005), and the United States (e.g., Cooter and Leduc 1995; DeGaetano and Allen 2002). This study also confirms that it might be more accurate to view the world as getting less cold instead of viewing it as getting hotter (Alexander et al. 2006), even though some regions show a more rapid increase in extreme maximum temperature than the increase in extreme minimum temperature: for example, the western Indian Ocean region (Vincent et al. 2011) and western central Africa, Guinea Conakry, and Zimbabwe (Aguilar et al. 2009). Changes in extreme temperatures are more spatially coherent than changes in extreme precipitation that exhibit strong local variability over the HP, which are also consistent with the findings in the previously mentioned studies.

A decreasing trend in CDD at a rate of about $-1$ day decade$^{-1}$ over the NHP appears to be consistent with the decreasing trend in CDD at rates ranging between $-4$ and $0$ days decade$^{-1}$ over most parts of the contiguous United States, as shown in Figure 6 of Alexander et al. (Alexander et al. 2006). In addition, an increasing trend in R95T at a rate of 1% decade$^{-1}$ and in SDII at a rate of 0.2 mm day$^{-1}$ decade$^{-1}$ over the CHP in this study seems also to fall within the ranges of decadal trends of R95T of 0%–1% decade$^{-1}$ and SDII of 0–0.5 mm day$^{-1}$ for this region of the HP also (Alexander et al. 2006). Trends in extreme precipitation indices are also consistent with the results in, for example, Frich et al. (Frich et al. 2002) and Tebaldi et al. (Tebaldi et al. 2007). In addition, magnitudes of variation in CDD and SDII of our studies are comparable with the counterparts from station data and simulations of an atmosphere-only GCM of trends of $-2$ to $0$ days decade$^{-1}$ and 0–0.2 mm decade$^{-1}$, respectively, for the HP as shown in Figure 2 of Kiktev et al. (Kiktev et al. 2003).

### 3.6. Trends in SPI over the High Plains

Spatial mean of the 1-month SPI generally increases over the study period (Figure 8). According to McKee et al. (McKee et al. 1993), a value of SPI $< -1.5$ indicates severe dry periods with $\sim 6.7\%$ of time in the long-term climate record. The percentage of areas that experienced severe dry periods over the U.S. HP was computed at a 5-yr interval from 1961 through 2010 (Figure 9). Apparently, there is a significant decreasing trend (at the 90% confidence interval) in severe dry areas
from ~8% to ~4% with the highest percentage occurring during 1976–80 and the lowest percentage during 1991–95 (Figure 9a). The percentage of area with floods (SPI > 1.0) increased (at the 95% confidence interval) from ~10% in 1961–65 and 1966–70 to ~20% in 1996–2000 and 2006–10 (Figure 9b).

The number of months that experienced dry conditions (SPI < 0) decreased (Figure 10). For the entire HP, the years 1974 and 1976 had 10 months with drier than normal conditions, whereas 1993, 1997, and 2007 had 2 drier than normal months. The decrease in percentage of severe dry areas and number of months with dry conditions indicates that the study region most likely is becoming wetter.

To examine which season has the largest change in SPI, the number of months with SPI < 0 based on seasons (each season has 3 months) was calculated. All four seasons exhibit large changes in the number of months with SPI < 0 (Figure 11). For instance, there are approximately 1.7 months per season with SPI < 0 from
1961 to 1985 and 1.1 months per season after 1985 (Table 2), reduced by 0.6 months per season. On average, spring had a slightly higher number (1.76 months) of months with SPI < 0 than other seasons during the period 1961–85. For both periods (1961–85 and 1986–2010), summer had slightly less months with dryer than normal conditions than other seasons. Comparison of the period 1961–85 with 1986–2010 indicates that the number of drier than normal months decreased by 0.3–0.5 months in each season.

### 3.7. Difficulties of estimating climate extremes

Estimating changes in extreme events is difficult because extreme events by their nature are limited in extent and evaluating changes can be constrained by temperature and precipitation network density. Climate variability is generally higher in more semiarid/arid regions, making it more difficult to estimate extremes in these regions. Focusing on the period from 1958 to 2010 increased the number of stations in this study relative to studies evaluating extremes over the past century. Previous studies have suggested that the spatial coherence of precipitation extremes is much less than that for temperature extremes (e.g., Pryor et al. 2009), requiring a higher density network to evaluate precipitation extremes. In the United States, the precipitation station density seems to be higher in the more humid regions in the east than in the more arid regions toward the west; however, the HP being in a semiarid region may require a higher network density to detect climate extremes. Gridding would also be expected to smooth data, reducing extremes; however, the general consistency between station and gridded results in this analysis suggests that gridded data may be valuable. There is also a problem with missing data, particularly for the temperature records; therefore, continuous records are very important. Most parts of the HP did not show significant trends in precipitation extremes in this study; however, this may be related, in part, to the
low station density and the high variability of precipitation extremes. Further studies on hydrological drought by examining streamflow extremes (e.g., Mishra et al. 2011b) and wet and dry spells in precipitation time series (e.g., Mishra et al. 2011a; Ozger et al. 2011) will be conducted for further understanding climate extremes and droughts over the High Plains.

4. Conclusions

Analysis of temperature and precipitation extreme indices in the HP from 1958 to 2010 shows that no changes in extremes throughout much of the HP. The intra-annual extreme temperature range (ETR) decreased significantly over 54% of the

Table 2. Statistics of the number of months with SPI < 0 for different seasons and different periods.

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Figure 11. Number of months with drier than normal conditions (SPI < 0) for (a) spring, (b) summer, (c) fall, and (d) winter.
HP stations (72 out of 134 stations; mean of $-0.7^\circ$C decade$^{-1}$). The decrease in ETR is attributed primarily to much higher increases in annual lowest temperatures ($T_L$) in 63% of the stations ($0.9^\circ$C decade$^{-1}$) relative to lower increases in annual highest temperatures ($T_H$) in $\sim$8% of stations ($0.4^\circ$C decade$^{-1}$). There has been a general increase in warm nights ($T_{min90}$) over 40% of the stations throughout the HP. Trends in extreme temperatures are spatially coherent similar to findings from other studies. These trends in ETR, $T_L$, and $T_H$ are consistent with published studies on ETR, $T_L$, and $T_H$ regionally and globally.

There is no trend in annual precipitation in 90% of the stations across the HP. In addition, there is no trend in extreme precipitation over most parts of the HP. Decreasing trends in CDD were restricted to 21% of stations in the NHP (mean of $-0.8$ days decade$^{-1}$). Significant changes in R95T were restricted to the CHP with 12% of stations showing increases and SDII in the CHP with 19% of stations showing increases. The spatial mean and anomaly in CDD over the NHP have been decreasing significantly at a rate of about $-1$ day decade$^{-1}$ and $-2.9$% decade$^{-1}$, respectively. HWDI did not show significant trends over the CHP and SHP but decreased over the NHP at a rate of about $-2$ days decade$^{-1}$. Trends in grided indices are generally consistent with station-based indices increasing confidence in the trends from this analysis.

Examination of short-term dry/wet conditions over the High Plains using the 1-month SPI indicates that there is no long-term trend in dry conditions, but the area experiencing excessive wetness appears to be increasing. Areas experiencing dry conditions (SPI $<-1.5$) have decreased from 8% to 4% of the entire High Plains. In the meantime, wet areas with SPI $>1.0$ increased from 10% to $>20$% of the total area. Furthermore, the number of dry months (SPI $<0$) in each year has decreased. The reduction in dry conditions was found in all four seasons from approximately 1.7 dry months per season during the period 1961–85 to 1.1 dry months per season after 1985.

In this semiarid region with high natural climate variability, it is difficult to quantify trends in climate extremes. Because of the greater spatial coherence of temperature data, trends in temperature extremes could be identified with the current network; however, lower spatial coherence of precipitation trends may require much higher station density than is currently available.

**Acknowledgments.** This study is financially supported by the NASA Project NNX09AN10G and the Bureau of Reclamation. Additional support was provided by the Jackson School of Geosciences at the University of Texas at Austin. We greatly appreciate the high quality station data provided by Dr. Kenneth Kunkel at the NOAA/National Climatic Data Center. Publication authorized by the Director, Bureau of Economic Geology, Jackson School of Geosciences, The University of Texas at Austin.

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