Trends in Intense Precipitation in the Climate Record

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ABSTRACT

Observed changes in intense precipitation (e.g., the frequency of very heavy precipitation or the upper 0.3% of daily precipitation events) have been analyzed for over half of the land area of the globe. These changes have been linked to changes in intense precipitation for three transient climate model simulations, all with greenhouse gas concentrations increasing during the twentieth and twenty-first centuries and doubling in the later part of the twenty-first century. It was found that both the empirical evidence from the period of instrumental observations and model projections of a greenhouse-enriched atmosphere indicate an increasing probability of intense precipitation events for many extratropical regions including the United States. Although there can be ambiguity as to the impact of more frequent heavy precipitation events, the thresholds of the definitions of these events were raised here, such that they are likely to be disruptive. Unfortunately, reliable assertions of very heavy and extreme precipitation changes are possible only for regions with dense networks due to the small radius of correlation for many intense precipitation events.

1. Introduction

In this paper we present an overview of findings related to changes in very heavy or intense precipitation, changes that are often disruptive to the environment and the economy (Edwards and Owens 1991; Easterling et al. 2000a, b; Soil and Water Conservation Society 2003; more information available online at http://earthobservatory.nasa.gov/NaturalHazards/). We do not focus on particular environmental and/or agricultural thresholds (e.g., floods, landslides, erosion, etc.) to define intense precipitation levels but instead, we use event frequency thresholds.[1] In general, throughout this paper, we count the upper 0.3% of daily rainfall events. This can be equated to a return period of approximately one daily event in 3 to 5 yr for annual precipitation and approximately 10 to 20 yr for seasonal precipitation, depending on the probability of daily rain events for a given location. Regionally averaged frequencies of these events and their changes are calculated using long-term datasets.

The structure of this article is as follows. First, we

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[1] For a given location and season, we define a daily precipitation event as heavy when it falls into the upper 10% and/or 5% of all precipitation events; as very heavy when it falls into the upper 1% and/or 0.3% of precipitation events; and extreme when it falls into the upper 0.1% of all precipitation events. Therefore, for heavy and very heavy precipitation events, we always specify the specific percentile used to avoid ambiguity.

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Table 1. Trend characteristics in annual precipitation totals; in heavy (upper 5%), very heavy (upper 1%), and extreme (upper 0.1% of daily rain events) precipitation totals; and in the fraction of total precipitation occurring in heavy, very heavy, and extreme precipitation events over the contiguous United States, 1910–99. Asterisks (*) indicate trends that are statistically significant at the 0.05 or higher level.

	Annual precipitation			Contribution to annual totals		
		Linear trend			Relative change	
Precipitation	Mean value (mm)	Estimate [% (10 yr)$^{-1}$]	Variance (%)	Fraction	Estimate [% (10 yr)$^{-1}$]	Variance (%)
Total	750	0.6	5*	1.00		
Heavy	195	1.7	12*	0.26	1.0	20*
Very heavy	62	2.5	15*	0.08	1.9	17*
Extreme	12	3.3	11*	0.016	2.7	9*

briefly review previous work and results related to heavy and very heavy precipitation over the land and describe the global network of daily precipitation stations available for analysis. Next, we describe the methodology used in our study and present our most recent results for several countries over the globe, including the contiguous United States. Finally, changes in heavy precipitation from global climate change simulations with changes in greenhouse gases and other forcings from three different global climate models are compared with observed changes.

2. An overview of the past studies

Changes in heavy and extreme precipitation were first documented by Iwashima and Yamamoto (1993) who used the data from scores of stations in Japan and the United States. A more detailed assessment for heavy precipitation increases over the contiguous United States was published by Karl and Knight (1998). Using the data from ~200 long-term stations, Karl and Knight (1998) showed that the sums of the highest monthly daily precipitation events had statistically significant increasing linear trends over four of the nine regions studied over the period 1910–95. Nationwide, there was a 7% increase. At the same time, Karl and Knight (1998) demonstrated that the contribution of the upper 10% of precipitation events to annual totals has increased nationwide. Thus, two important aspects of heavy precipitation were changing, amount and frequency, but the analysis was for a relatively sparse network. When attempts were made to raise the threshold above the 10 percentile, it immediately became obvious that a denser network was required. Section 4b explains theoretical considerations behind this requirement.

Groisman et al. (1999a) assumed a simple and quite flexible three-parameter model for the distribution of daily precipitation totals. The model was tested and fitted to the data of eight countries (Canada, the United States, Mexico, Norway, Poland, the former USSR, China, and Australia), in order to study the sensitivity of the probability of heavy rainfall (P_{heavy}). This was done by varying the model parameters according to their observed temporal and spatial variations as well as by the time series analysis of variations in P_{heavy}. Groisman et al. (1999a) found a disproportionate change in precipitation intensity whenever the mean precipitation changed. This was also shown theoretically by Katz (1999).

These studies and an increasing number of model projections indicate that changes in intense precipitation are more likely as global temperature increases (e.g., Meehl et al. 2000; Cubash and Meehl 2001; Zwiers and Kharin 1998; Kharin and Zwiers 2000; Allen and Ingram 2002; Semenov and Bengtsson 2002). This has triggered a set of studies to determine the change in the probability of heavy precipitation over the world using all available daily data. Easterling et al. (2000c) summarized these efforts. Thereafter, a number of regional studies have now become available (e.g., Stone et al. 2000; Zhang et al. 2001; Frei and Schär 2001; Alpert et al. 2002; Frich et al. 2002; Roy and Balling 2004). Some basic tenets emerging from analyses include:

- To obtain statistically significant estimates, the characteristics of heavy precipitation should be areally averaged over a spatially homogeneous region. Otherwise, noise at the spatial scale of daily weather systems masks changes and makes them very difficult to detect (e.g., Frei and Schär 2001; Zhang et al. 2004).
- Whenever there are statistically significant regional changes in the rainy season, relative changes in heavy precipitation are of the same sign and are stronger than those of the mean. A search at various sites around the globe using our data holdings and results from others (e.g., Osborn et al. 2000; Tarhule and Woo 1998; Suppiah and Hennessy 1998; Zhai et al. 1999; Groisman et al. 2001) confirm this.
- This search also revealed several regions where mean precipitation does not noticeably change in the rainy season but heavy precipitation does change. In such cases, there was always an increase in heavy precipitation. Among these regions are Siberia, South Africa, northern Japan (Easterling et al. 2000c), and eastern Mediterranean (Alpert et al. 2002).

In the recent report by the U.S. Soil and Water Conservation Society (2003; Tables 1, 2, and 3), changes in

TABLE 2a. Trend characteristics in the number of days with heavy and very heavy precipitation over the contiguous United States, 1910–99 (percentile definition). Asterisks (*) indicate trends that are statistically significant at the 0.05 or higher level.

	Days with precipitation			Contribution to total days with precipitation above 1 mm		
		Linear trend			Relative change	
Events	Mean (days yr^{-1})	Estimate [% $(10\ \text{yr})^{-1}$]	Variance (%)	Fraction	Estimate [% $(10\ \text{yr})^{-1}$]	Variance (%)
Total days with precipitation above 1 mm	75	0.5	6*	1		
Heavy (upper 5% of precipitation events)	4.4	1.5	12*	0.06	1.0	11*
Very heavy (upper 1% of precipitation events)	0.88	2.2	14*	0.012	1.7	13*

nationwide (contiguous U.S.) annual precipitation with a partition of daily rainfall into heavy (above the 95 percentiles and/or above 50.8 mm), very heavy (above the 99 percentiles and/or above 101.6 mm), and extreme (above the 99.9 percentiles) events were assessed. Tables 1 and 2 showed that as the mean total precipitation and the number of rainy days over the conterminous U.S. increased, the heavy and very heavy precipitation increase was significantly greater, as was the proportion of the total precipitation and counts attributed to these events. Thorough analysis of the United States data indicated that practically the entire nationwide increase in heavy and very heavy precipitation occurred during the past three decades (Table 3).

Groisman et al. (2001, 2004) considered different definitions for heavy and very heavy precipitation and their changes during the past century for the contiguous United States. Groisman et al. (2004) raised the threshold definition of very heavy precipitation events to the upper 0.3% of daily precipitation events. This translates to an 80 mm day^{-1} threshold for daily precipitation in the major agricultural area of the midwestern United States and over 100 mm day^{-1} for the southern and southeastern parts of the United States (Table 4). When selected from a 12-month period, events above the upper 0.3% threshold have a return period of only once in approximately 3 to 5 yr. When selected among a 3-month season of daily events, the return period of very heavy precipitation events varies from 10 to 20 yr, depending upon the total frequency of days with measurable precipitation in the region. For these defined thresholds, Groisman et al. (2001, 2004) found statistically significant century-long trends in the frequency of very heavy precipitation events within three major regions (Fig. 1; the South, Midwest, and Upper Mississippi) of the central United States. These regions are particularly important because they cover most of the Mississippi River basin and most of the wheat and corn belts of the country. Their analysis showed that regionally and seasonally, changes in very heavy precipitation vary significantly, and the magnitude of the trends is most notable in the eastern two-thirds of the country, and primarily in the warm season when the most intense rainfall events typically occur. We extend this work as described in section 5g.

3. Data

Subdaily precipitation time series provide more information about precipitation intensity than daily totals (Trenberth et al. 2003). However, subdaily data are readily available in sufficient quantities only for the contiguous United States (Frederich et al. 1997; NCDC 1998), and homogeneity problems remain to be overcome prior to their use (Groisman et al. 1999b). Therefore, in this study we use daily total precipitation data sets compiled at the National Climatic Data Center (NCDC; Fig. 2; NCDC 2002). Precipitation information is available for virtually all of these stations. Several regions of the world (most of North America, East Asia, eastern Australia, eastern Brazil, India, South Africa, central Mexico, southern half of the former USSR, and Northern Europe) have a sufficiently dense network to make it feasible to study changes in very heavy precipitation. There are many more precipitation sta-

TABLE 2b. Trend characteristics in the number of days with heavy and very heavy precipitation over the contiguous United States, 1910–99 (absolute value definition). Asterisks (*) indicate trends that are statistically significant at the 0.05 or higher level.

	Days with precipitation			Contribution to total days with precipitation above 1 mm		
		Linear trend			Relative change	
Events	Mean (days yr^{-1})	Estimate [% $(10\ \text{yr})^{-1}$]	Variance (%)	Fraction	Estimate [% $(10\ \text{yr})^{-1}$]	Variance (%)
Total days with precipitation above 1 mm	75	0.5	6*	1		
Heavy (above 50.8 mm)	1.4	3.3	30*	0.02	2.8	33*
Very heavy (above 101.6 mm)	0.13	4.9	22*	0.002	4.4	21*

Table 3. Trends in share of total annual precipitation occurring in heavy, very heavy, and extreme daily precipitation events in the contiguous United States, 1910–70 vs 1970–99. Asterisks (*) indicate trends that are statistically significant at the 0.05 or higher level.

Precipitation	1910–70 Mean (mm)	1910–70 Linear trend Estimate [% (10 yr)$^{-1}$]	1910–70 Linear trend Variance (%)	1970–99 Mean (mm)	1970–99 Linear trend Estimate [% (10 yr)$^{-1}$]	1970–99 Linear trend Variance (%)
Total annual precipitation	737	−0.4	1	772	1.2	2
Heavy (upper 5% of precipitation events)	188	−0.1	0	208	4.6	12*
Very heavy (upper 1% of precipitation events)	59	0.9	1	67	7.2	15*
Extreme (upper 0.1% of precipitation events)	12	1.5	1	14	14.1	22*

tions in the world (Groisman and Legates 1995), but their daily data are not presently available. The appendix shows the regional availability of long-term stations used in this study. Daily precipitation data for three countries, the former USSR, Canada, and Australia, underwent a special preprocessing to restore the homogeneity of the time series. These countries were affected by changes in observational practices and instruments (Groisman and Rankova 2001; Groisman et al. 1999a; Groisman 2002).

4. Analysis method

The regional averaging technique employed throughout this paper is described in section 4a. Section 4b discusses the representativeness of the area-averaged time series used in this study. In most cases we present the actual time series, which allows the reader to judge the form of systematic trends revealed. We did not focus on the linearity of the changes and used (in additional to the linear trend assessment) a nonparametric test to check for a monotonic change of the time series. In a few cases, when significant nonlinearity was detected, we point it out explicitly. Once a statistically significant trend has been discovered, we characterize it by the mean rate of change. A linear trend estimate is an essential characteristic in this case. We tested the presence of systematic change in the time series using two standard methods: least squares regression (Draper and Smith 1966; Polyak 1996) and a nonparametric method based on Spearman rank order correlation (Kendall and Stuart 1967). We used two-tailed tests at the 0.05 or higher significance level. We tested for autocorrelation of the detrended time series of very heavy precipitation, but the residuals of the frequencies of heavy and very heavy precipitation events were never found to be autocorrelated.

a. Area-averaging routine

Meteorological stations are not uniformly distributed. Stations tend to cluster around major metropolitan areas and are sparse in mountainous terrain. Missing values are present in most of the records. Both factors had to be addressed to properly represent regional averages of the frequency and/or amount of very heavy precipitation derived from in situ observations. Area-averaged calculations presented in this paper all use the same method. First, we selected a reference period with the greatest availability of data to estimate the long-term mean values for each element and for each season. For most of the countries/regions, the period selected was 1961–90 but, for example, in the Nord-Este region of Brazil the reference period used

Table 4. Area-averaged annual and seasonal daily precipitation thresholds (in mm) for very heavy (upper 0.3%), and extreme (upper 0.1%) precipitation events in different regions of the contiguous United States. Note that at each specific location the threshold precipitation value may be different. The return period for such events varies depending upon the frequency of days with measurable precipitation and varies, for example, from 3 to 5 yr for annual and 10 to 20 yr for seasonal very heavy precipitation events. Region numbers correspond to those shown in Fig. 1

	Subregion	Area, 10^3 km^2	Annual 0.3%	Annual 0.1%	Winter 0.3%	Spring 0.3%	Summer 0.3%	Autumn 0.3%
1	Northwest	660	45	55	45	35	35	40
2	Missouri River basin	1240	50	65	20	45	60	45
3	Upper Mississippi	680	65	80	30	50	80	65
4	Northeast	480	65	80	50	55	75	70
5	California & Nevada	720	65	80	65	50	40	60
6	Southwest	1120	45	55	30	35	45	45
7	South	1500	105	130	65	95	100	110
8	Midwest	820	80	100	65	75	85	80
9	Southeast	780	105	130	85	100	100	110
	48-states average	8000	70	90	50	60	70	70

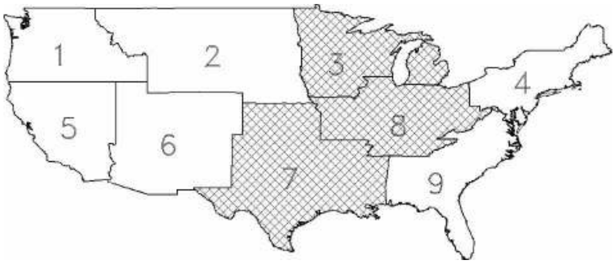

FIG. 1. Regions of the contiguous United States (hatched) where statistically significant annual increases in very heavy precipitation for the 1908–2002 period were reported by Groisman et al. (2004).

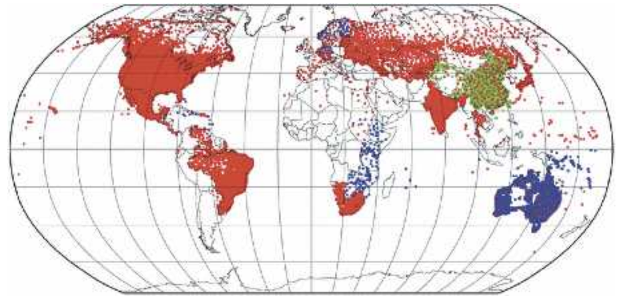

FIG. 2. Map of stations with daily precipitation available at the U.S. NCDC (as of 15 Jul 2003). A subset of ~32 000 is available through Global Daily Climatology Network (GDCN), Version 1.0 (NCDC 2002; red dots). Only typhoon-related precipitation data are available for most of stations from China (green dots).

was 1951–80 because for a significant number of stations the data ended in 1980. For each station, we determined the empirical distribution function and the set of upper threshold values (90, 95, 99, 99.7, and 99.9 percentiles of rainy day events) for daily precipitation during the reference period. Then, exceedences (or precipitation totals for some analyses) above the threshold values were totaled and the climatological mean was calculated for the threshold based on the reference period. For each region, season, year, and intense precipitation threshold, we calculated the anomalies from the long-term mean number of exceedences (or precipitation totals above the thresholds) for each station and then arithmetically averaged these anomalies within 1º × 1º grid cells. These anomalies were regionally averaged with the weights proportional to their area. Data from the large regions use the regional area weights to form a national average when those analyses are presented. The long-term mean values (normals) were area-averaged in a similar fashion and used to determine actual precipitation amounts. This approach emphasizes underrepresented parts of the region/country because a region, even with a relatively low percentage of grid cells with data will receive the full weight comparable to the region's area relative to other regions. It also allows the preservation of the regional time series unaffected by the changing availability of data with time.[2]

In some situations, time series of exceedences of very high climatological thresholds vary substantially between reference periods due to changing climate conditions (Zhang et al. 2005). Special experiments with varying reference periods were conducted to assure that the conclusions presented in this paper are not affected by this problem.

b. *Representativeness of regional estimates of very heavy precipitation frequency*

1) THEORETICAL FRAMEWORK

Each estimate of the area-averaged anomaly[3] of a variable, X, is based on a set of point (or grid cell) measurements of these anomalies, x_i, (i = 1, 2, . . . N), within a region with area S. The estimate should be representative of the regional quantity. Formally, this means that a linear combination of our point measurements, $X' = \Sigma\, w_i x_i$, should be as close to variable X as possible (w_i are the weights of averaging in approximation of X by X'). The mean square error, E^2, of the linear estimate of the area-averaged anomaly X over the region S using data (x_i) from N locations (or grid cells) is given by Kagan (1997) as

$$E_2 = \sigma_s^2 - 2\sum w_i \Omega_i + \sum\sum w_i w_j R_{ij} + \sum w_i^2 \delta_i^2, \tag{1}$$

where σ_s^2 is the variance of the variable X averaged over the region S, Ω_i is the covariance of x_i and X, R_{ij} is the covariance between x_i and x_j, and δ_i^2 is the variance of the error of measurement at location i. If the statistical structure of the x field is known, the error E^2 can be estimated for each set of sites (grid cells) inside the region with any selection of w_i. Our selection of weights w_i is described in section 4a.

[2] We used this area-averaging routine during the past decade for various climate variables (e.g., Groisman and Legates 1995; Karl and Knight 1998; Groisman et al. 2001, 2004) after extensive testing regarding the robustness of the algorithm. The results of its implementation are close to those based on area-averaging procedures built on optimal interpolation and optimal averaging with normalizing weights (Gandin and Kagan 1976; Kagan 1997). Optimal procedures (i.e., those deliver the minimal standard error of area averaging) are much more computationally extensive and preserve their optimal properties only when specifics of the statistical structure of meteorological field to be averaged are well known. This is not the case for many of the regions we analyzed and, thus, we opted not to use them.

[3] The use of anomalies guards against faulty trends due to station dropouts.

The spatial correlation function of the frequency of very heavy precipitation has never been estimated previously, although it is likely that the radius of correlation of these quantities is quite small. We calculated the spatial correlation function of the seasonal and annual frequencies of very heavy precipitation (our x fields) and approximated it in the form

$$r(\rho) = C_0 \exp(-\rho/\rho_0), \tag{2}$$

where ρ is the distance, ρ_0 is the radius of correlation, and C_0 is a constant less than or equal to 1. The term $(1 - C_0)$ is an estimate of the portion of the variance of the x field that is not spatially correlated. As a result, C_0 characterizes both microclimatic variability and errors in x measurements (i.e., δ^2). In our estimates for the contiguous United States, Australia, Brazil, South Africa, and Mexico, C_0 varies between 0.55 and 1.0.

The relative root-mean-square error, Z, of the mean anomaly of the x field over the region S that is approximated by x_i at N points (grid cells) evenly distributed over the region S (below we denote the area of the region with the same letter, S) can be described by

$$Z_S = C_v\{[(1 - C_0)/C_0 + 0.23(S/N)^{1/2}\rho_0^{-1}]/N\}^{1/2}, \tag{3}$$

where the spatial correlation function is approximated by Eq. (2), and C_v is the coefficient of variation of the x field (Kagan 1997). If the points/cells are not evenly distributed over the region, Z_S is increased by a factor influenced by the area-averaging routine, the parameters of spatial correlation function, and the measure of unevenness of the station distribution.

Following the area-averaging procedure described in section 4a, we first estimated the representativeness of gridcell area averaging. This step provided us estimates of the accuracy of the $1° \times 1°$ grid cell values for the average frequency of heavy and very heavy events. These accuracy estimates were then used for evaluation of the regional Z_S values.

2) Results

The application of Eq. (3) shows that for the annual frequency of very heavy precipitation events for a typical $1° \times 1°$ grid cell on fairly level terrain[4] (with a ρ_0 of ~30 km and C_v of ~0.3), with three, two, or one stations we cannot reduce our error below 10%, 15%, and 25%[5] and a similar assessment in mountainous $1° \times 1°$ grid cells[6] (with a ρ_0 of ~10 km and C_v of ~0.4) gives estimates of Z_S in the range of 25%, 35%, and 60%, respectively.[7]

The appendix Table A1 provides estimates of ρ_0 and C_0 for the frequency of heavy (H) and very heavy (VH) seasonal and annual precipitation (above the upper 10 and 0.3 percentiles, respectively) for several regions of the contiguous United States and two regions of the European part of the former USSR. In the latter, thunderstorm activity associated with very heavy precipitation is less spatially expansive compared to the former (i.e., it rarely manifests itself as a multicell event) and we rarely have more than one station per grid cell. Consequently, here we obtained estimates of C_0 below 0.5 for approximations of the spatial correlation function of the frequency of VH annual precipitation events between $1° \times 1°$ grid cells. Large values of seasonal and annual radius of correlation for frequency of heavy precipitation events of several hundred kilometers (up to 600 km in the northwestern United States in winter) assure the representativeness of area-averaged values of this quantity based on a point/gridcell network similar to that for mean seasonal/annual precipitation (cf. Czelnai et al. 1963; Huff and Changnon 1965; Kagan 1997). For VH events, further analysis was required.

In a region larger than a grid cell (e.g., the midwestern United States, which encompasses 82 $1° \times 1°$ grid cells, with nearly complete gridcell data coverage during the entire twentieth century Fig. A2), we obtained Z_S less than 2% throughout the twentieth century for the area-averaged annual VH frequency. In this region, Z values remain less than 3% even in the last decade of the nineteenth century. The opposite situation (among the regions considered in this paper) is evident in northwestern Russia between 60°N and the Arctic Circle. This region, with area ~10^6 km^2, does not have a complete (or nearly complete) $1° \times 1°$ grid cell coverage to start with (Fig. A3). Fifteen $1° \times 1°$ grid cells with data (usually a single station within a cell and the cells unevenly distributed over the region) result in a value of Z_S close to 15%. The term $C_v[(1 - C_0)C_0^{-1}N^{-1} + \ldots]^{1/2}$ is a major component in Eq. (3) for this region and it decreases slowly with increasing N. The above illustrates that the number of $1° \times 1°$ grid cells with valid station data is an important component in the accuracy of area averaging. Therefore, this quantity was used throughout the study to control the level of representativeness of our results for each region discussed below.

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[4] For example, the midwestern United States, European Russia, or Australia.

[5] We consider the Z_S estimates to be on a low side because of several assumptions [e.g., that the stations are distributed evenly over the grid cell, that the approximation of the covariance function, R, with the help of Eq. (2) is precise, etc.] in reality do not materialize and/or are only convenient approximations.

[6] For example, the southwestern or northwestern United States, Caucasus, or Mexico with large micrometeorological variability.

[7] The use of frequencies of intense precipitation with thresholds derived from the local distributions may not be very different at low and high elevations, because we are flexibly changing the definitions of heavy events. This alleviates to a certain extent the impact of the elevation-inhomogeneous distribution of stations that interferes with steep precipitation gradients. But, in the mountains, assumptions of implementation of Eq. (2) (in particular the isotropy of the spatial covariance function, R) are less reliable than on the plains (Gandin et al. 1976; Gandin 1988). This adds additional uncertainty and indicates that the Z_S estimates are to be on a low side.

3) A major concern

Scattered thunderstorms in the area, or strong gradients in precipitation totals and/or variances (e.g., due to elevation changes) may cause a fraction of the rain events to remain unnoticed and/or the estimates of the area total to be biased. For example, in the contiguous U.S. west of 105°W, the mean average elevation of the synoptic station network is about 500 m below the mean elevation of the surface and most of the cold season precipitation is orographically defined (Daly et al. 1994). To avoid biases associated with spatial inhomogeneity in the mean and variance of the field, we analyzed anomalies and used point-defined individual thresholds and counts above these thresholds. The use of counts (instead of actual values of precipitation) gave us more robust results even when we do not have a dense network. For example, if we have widespread heavy rain events that fall mostly in the mountains, while only their remnants show up in the valley where our station happens to be located, then we may still capture these events and our count of the heavy and/or very heavy events is still unbiased. Problems arise for small ρ_0 (e.g., scattered thunderstorms) and a relatively sparse network that cannot capture most of these events. Biases, however, are not a major concern in this situation. Let us assume that the network is so sparse (or ρ_0 is so small) that each event is counted only once (i.e., only at one station). Then the area-averaged count of the events according to our area-averaging routine will be equal to the count of events divided by N and remains the estimate of the probability of the event at a single station within the region. If we assume the statistical field to be isotropic with a spatial covariance function provided by Eq. (2), this will still be an unbiased estimate of the average probability of this type of event within the region. However, if N is small, the estimate will be of very low accuracy. This immediately would be noticed by the second term in Z_S [Eq. (3)], which would grow to very high values. When Z_S becomes greater or comparable to C_v, the practical importance of our area-averaged estimate based on the point measurements in that particular instance becomes low and the information carried by the measurements is not clearly seen beyond the noise level. Moreover, one can select an alternative network within the same region, which will provide alternative estimates that are independent (uncorrelated) with those produced by the original network. Thus, for the processing and analyses applied to this study, it is not biases but representativeness that is the key problem. Empirically, we observed the manifestation of this problem in the portion of the regionally averaged time series that is based on an insufficiently dense network. The interannual variability of this portion starts behaving badly. It become highly variable compared to the period with sufficiently dense network in the region (cf. Osborn and Hulme 1997) meaning that the variability of the time series was dominated by the random error of the estimation process.

5. Analyses for several regions in the world

a. European part of the former USSR

For this region, more than 700 long-term stations during the period 1936–97 are available for analyses of heavy and very heavy precipitation (Bulygina et al. 2000). The numbers of stations for the two regions under consideration shown in Fig. 3 are 70 and 633. In general, maximum precipitation in this area occurs during the warm season, with very heavy rainfall coming almost entirely from convective clouds (Sun et al. 2001). Note that approximately 95% of the daily precipitation events are less than 10 mm day^{-1}. Table 5 and Figure 3 summarize the results of trend analyses for these two regions. Both show a large increase of 10% to 15% in annual precipitation in the region for the study period although the century-long increase is smaller (e.g., Groisman 1991; Groisman and Rankova 2001). During the same period, the rates of increase in heavy precipitation, in very heavy precipitation, and in extreme rainfall were higher than for mean annual precipitation. The linear trend of the time series of heavy precipitation was statistically significant at the 0.01 level in both regions. In the southern region, trends in very heavy (upper 1% of rain events) and even in extreme precipitation are also statistically significant at the 0.05 level or above. The trends of very heavy precipitation in the north are not statistically significant, partly due to large sampling errors resulting from a small number of stations in that area (Figs. 1 and A3). The network here is adequate for capturing total precipitation and the upper 10% and 5% of precipitation events. However, when totaling the precipitation of rare very heavy rain events (that occur once per year or even less frequently), one needs a denser network to suppress the very high weather variability associated with these events. The Z estimates based on Eq. (3) show that the random errors of the area-averaged frequency of very heavy and extreme precipitation are approximately 4 times higher than in the southwestern part of the former USSR.

b. Northern Europe

Fennoscandia is very well covered by precipitation stations (Groisman and Legates 1995), but only a fraction of the daily data for this network is available publicly (Klein Tank et al. 2002), or for special research projects such as Arctic Climate Impact Assessment (ACIA 2004; Groisman et al. 2003). In the framework of ACIA, a study of contemporary climatic changes in high latitudes during the past 50 yr has been conducted (Groisman et al. 2003). To define heavy precipitation events in high latitudes, a special effort was made to

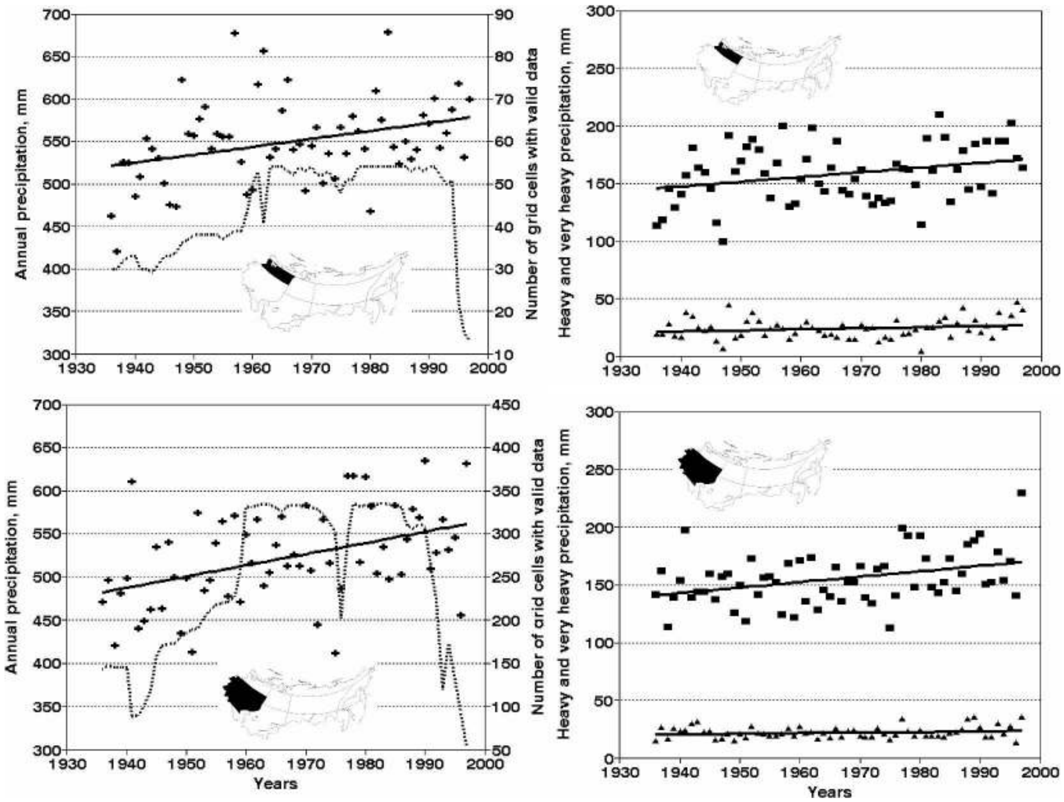


FIG. 3. Annual (pluses), heavy (upper 5%; squares), and very heavy (upper 0.3%; triangles) precipitation totals over the western half of the former USSR (regions of area averaging are shown darkened in the maps within the plots) and their linear trends (solid lines). Statistical significance of linear trends is provided in Table 5. The numbers of 1° × 1° grid cells with valid station data are shown by dotted lines.

separate and further consider only discernible precipitation events, which we have defined as those above 0.5 mm. The reason for this is that the median of daily precipitation events over most regions in high latitudes is close to or even less that 0.5 mm. This coupled with the frequently changing precision of measurements can interfere with our analyses, for example, in Canada and Norway (Groisman et al. 1999a). There were previous reports describing the total precipitation increase in northern Europe (Groisman 1991; Hanssen-Bauer et al.

TABLE 5. Trend characteristics of the annual precipitation for the western part of the former USSR over the period 1936–97. Trend values statistically significant at the 0.05 level or at the 0.01 level are marked with asterisks (*) and double asterisks (**), respectively

Precipitation	Totals mm	Thresholds mm	Linear trend and its variance % $(100\ \mathrm{yr})^{-1}$	Linear trend and its variance %
		North of European Russia (north of 60ºN)		
Total	560	0	17	11**
Heavy, 90 percentile	240	7	27	14**
95 percentile	160	10	26	9*
Very heavy, 99 percentile,	55	20	25	3
99.7 percentile	23	30	44	4
Extreme, 99.9 percentile	10	35	52	3
		European part of the former USSR south of 60ºN		
Total	540	0	24	18**
Heavy, 90 percentile	240	9	290	15**
95 percentile	160	13	30	14**
Very heavy, 99 percentile	55	25	40	15**
99.7 percentile	22	37	20	2
Extreme, 99.9 percentile	10	47	50	10*

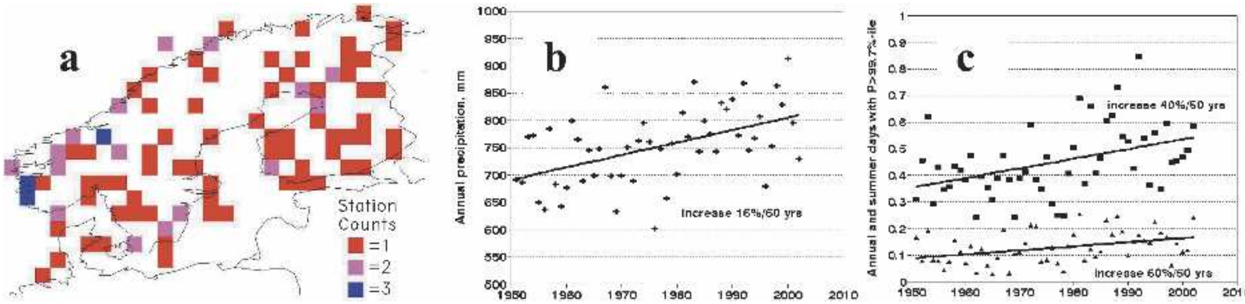


FIG. 4. (a) Data availability at the 88 stations over Fennoscandia generalized within the 1° × 1° grid cells, (b) annual totals, and (c) frequency of very heavy annual (squares) and summer (triangles) precipitation events during the 1951–2002 period. All linear trends (shown by solid lines) are statistically significant at the 0.01 level or above. All increases have occurred in the past 25 yr. The average regional upper 0.3% thresholds are 50 and 45 mm for summer and year, respectively.

1997; Heino et al. 1999; Førland and Hanssen-Bauer 2000; Folland and Karl 2001), but they all reported a smaller relative change compared to our results for changes in very heavy precipitation frequency both in summer (a season with the most intense precipitation) and throughout the year (Fig. 4).

c. *Pacific coast of northwestern North America*

This is the only large high-latitude region with both large annual precipitation totals and a sufficiently dense precipitation network available for our analyses. Figure 5 shows the time series of the frequency of heavy and very heavy precipitation in southern Alaska (south of 62°N) and British Columbia, Canada (south of 55°N). In both regions, precipitation increased during the period of record, but the double-digit increase in the frequency of heavy and very heavy precipitation is especially noteworthy (Table 6). Given the high thresholds for these events, these changes reflect an increasing societal and/or environmental threat in both areas.

d. *Southeastern and southwestern Australia*

Southeastern Australia is densely populated and thus is well covered by a long-term precipitation network (Lavery et al. 1997; Fig. A4). The southwestern tip of the continent has good station coverage since the mid-1910s. Precipitation occurs in southeastern Australia year round without particular peaks in the seasonal cycle while a winter (June–August) maximum is observed in the southwest. The regionally averaged thresholds for annual very heavy precipitation (upper 0.3% of daily events) are 82 and 53 mm, respectively. Figure 6 shows time series of both annual precipitation and the frequency of days with very heavy precipitation for these two regions. Precipitation totals increased by 16% $(100\ \mathrm{yr})^{-1}$ in the southeast and decreased by the same amount in the southwest during the period with sufficient data (1907–98 and 1913–98, respectively). The 52% $(100\ \mathrm{yr})^{-1}$ increase in frequency of very heavy precipitation in the southeast is statistically significant at the 0.05 level, while the 43% $(100\ \mathrm{yr})^{-1}$ decrease in the southwest is statistically significant only at the 0.10 level.

e. *South Africa*

We focused on the eastern part of the country, which is more humid and is mostly farmland. This area is thoroughly covered by a dense, long-term precipitation network (Fig. A5). The regionally averaged thresholds for very heavy precipitation in the upper 0.3% of daily events for annual and summer (December–January–

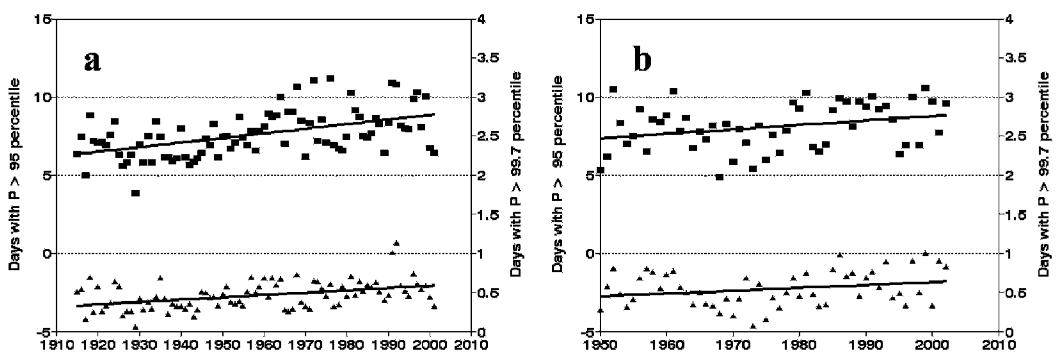


FIG. 5. Heavy and very heavy annual precipitation variations and linear trends along the northwestern coast of North America (a) British Columbia south of 55°N and (b) Alaska south of 62°N. Statistical significance of linear trends is provided in Table 6.

TABLE 6. Heavy and very heavy annual precipitation along the northwestern coast of North America. Values of mean annual precipitation, area-averaged thresholds for heavy (upper 5 percentile) and very heavy (upper 0.3%) are shown. Trend estimates in total precipitation and annual number of days above the two thresholds are also presented. Trend values statistically significant at the 0.05 level or at the 0.01 level are marked with asterisks (*) and double asterisks (**), respectively

	Total precipitation		95 percentile threshold		99.7 percentile threshold	
Region, period assessed	Mean mm	Trend % (50 yr)$^{-1}$	Value mm	Trend % (50 yr)$^{-1}$	Value mm	Trend % (50 yr)$^{-1}$
British Columbia, south of 55ºN 1910–2001	1 625	7.2**	26	16**	56	19**
Alaska, south of 62ºN 1950–2002	1 640	10.3*	28	18*	66	37

February, hereafter DJF) are 85 mm (return period 5 yr) and 90 mm (return period 10 yr), respectively. Figure 7 shows time series of both precipitation totals and frequency of very heavy precipitation for this region. While annual and summer precipitation totals did not change during the period with sufficient data (1906–97), there is an increase in the annual frequency of very heavy precipitation that is statistically significant at the 0.05 level. These results broadly correspond to those by Fauchereau et al. (2003). In addition, we noted a significant increase in very heavy precipitation during the last three decades (a feature that was also evident for the contiguous United States; cf. Groisman et al. 2004; section 5g).

f. Eastern Brazil and Uruguay

A sufficiently dense network for the past 70 yr is available for the eastern half of Brazil and Uruguay (Figs. 2 and A6). This became possible after the National Meteorological Service, in cooperation with the private company *Agência Nacional de Energia Elétrica* (ANEEL) made their national precipitation data publicly available online. An explosion of research resulted (Liebmann et al. 1998, 1999, 2001; Liebmann and Marengo 2001; Marengo et al. 2001; Carvalho et al. 2002).

Figure 8 shows time series of both annual precipitation and the frequency of very heavy precipitation for three regions of eastern Brazil and Uruguay. For these three regions, the regionally averaged thresholds for annual upper 0.3% of daily rainfall events are 100, 95, and 120 mm, respectively, and have return periods of 3 to 4 yr. Very high precipitation variability in a relatively dry climate of the Nord-Este is modulated by El Niño–La Niña events (Ropelewski and Halpert 1996). In the Nord-Este, we found a statistically significant increase in the annual frequency of very heavy precipitation events of 40% (100 yr)$^{-1}$, but all of the increase occurred during the first half of the twentieth century. In the subtropical part of Brazil there was a systematic increase of very heavy precipitation since the 1940s. In the northern subtropical region, where an extremely dense network of São Paulo state was used in our analysis (cf. Liebmann et al. 2001), we obtained an increase of 58% (100 yr)$^{-1}$, or by 34% for the period of record, statistically significant at the 0.01 significance level.

g. Central United States

In three large neighboring regions in the central United States (hatched in Fig. 1), statistically significant increasing trends in very heavy precipitation were documented for the period 1908–2002 (Groisman et al. 2004). Data availability restricted Groisman et al. (2004) in the nationwide analyses of the very heavy

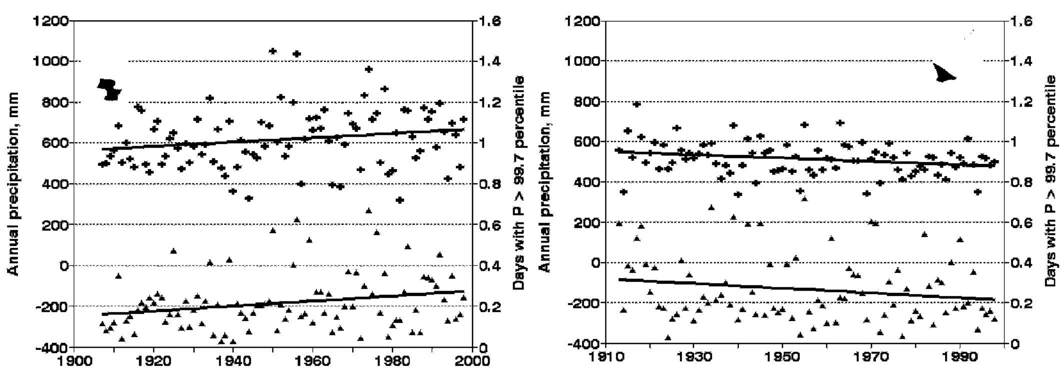


FIG. 6. Annual precipitation (pluses), frequency of very heavy precipitation (triangles), and their linear trends (solid lines) over southeastern and southwestern Australia (regions of area averaging are shown darkened in the maps within the plots). The linear trend estimates for annual precipitation decrease in southwestern Australia [−16% (100 yr)$^{-1}$] and increase in the number of days with very heavy precipitation in southeastern Australia [(62% (100 yr)$^{-1}$] are statistically significant at the 0.05 level. The linear trend estimates for annual precipitation increase in southeastern Australia [16% (100 yr)$^{-1}$] and decrease in the number of days with very heavy precipitation in southwestern Australia [(−45% (100 yr)$^{-1}$] are statistically significant at the 0.10 level.

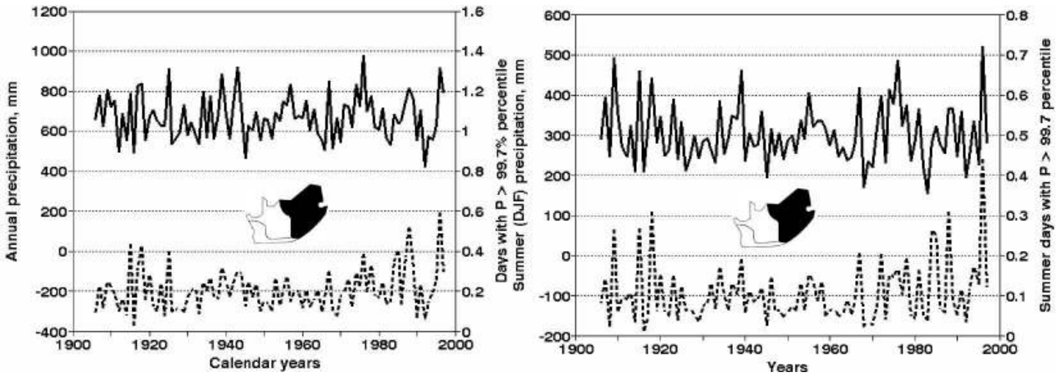


FIG. 7. Annual and summer (DJF) rainfall (solid lines) and frequency of very heavy rains (dashed lines) over the eastern half of South Africa (region of area averaging is shown darkened in the map). The linear trend estimates for an increase in the annual and summer frequency of very heavy precipitation [44% (100 yr)$^{-1}$ and 42% (100 yr)$^{-1}$] are statistically significant at the 0.05 and 0.10 levels, respectively.

precipitation prior to the 1908 starting year. This, however, is mostly due to a data deficiency in the western part of the country (Fig. A2). Here, we extend our analyses of very heavy precipitation back to 1893 (Fig. 9). In 1893, the analysis covers only 84 1° × 1° grid cells compared with more than 330 during the second half of the twentieth century. This deficiency adds some noise to the regionally averaged time series. Over the study period (1893–2002) the frequency of very heavy precipitation has increased by 20% (statistically significant at the 0.01 level). All of the increase has occurred during the last third of the century (Fig. 9). The longer the time series are, the better is our understanding of the variability of heavy and very heavy precipitation in the twentieth century. But, it is probably a paradox that so much effort was made to collect, quality control, pre-

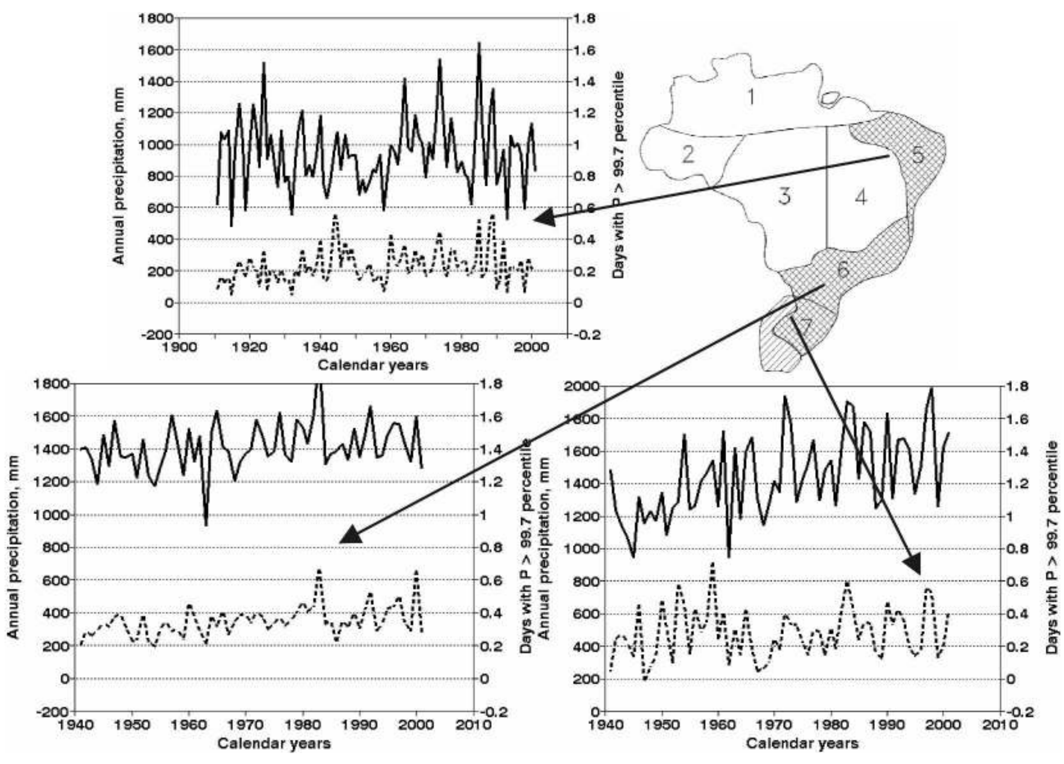


FIG. 8. Annual rainfall (solid lines) and frequency of very heavy rains (dashed lines) over three regions in Brazil, Uruguay, and adjacent Argentinean and Paraguay areas (regions are hatched in the map). Linear trend estimates for increases in the annual frequency of very heavy precipitation in the Nord-Este (1911–2001) and northern subtropics (1941–2001) are statistically significant at the 0.05 level or higher.

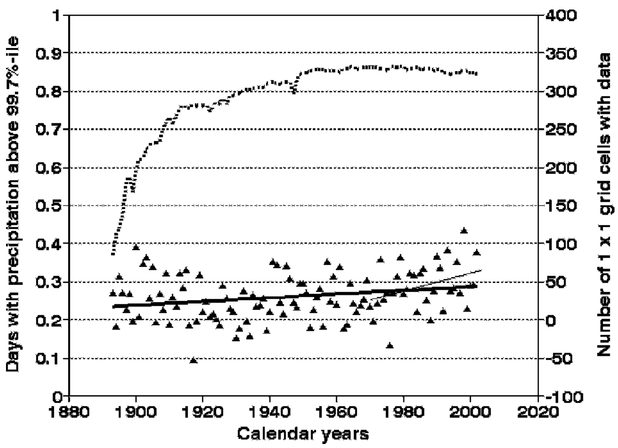


FIG. 9. Very heavy precipitation (upper 0.3% of daily rain events with return period of 4 yr) over regions of the central United States (hatched in Fig. 1) and their linear trends. Linear trends for the 1893–2002 and 1970–2002 periods (solid lines) are equal to 20% $(110\ \mathrm{yr})^{-1}$ and 26% $(30\ \mathrm{yr})^{-1}$, respectively, and are statistically significant at the 0.05 level or higher. Note that there was not any change in very heavy precipitation prior to 1970. The numbers of 1° × 1° grid cells with valid station data are shown by dotted line.

process, and analyze data for the full 110 yr, only to reveal that during the first 80 yr no systematic changes occurred in very heavy precipitation frequency (Table 3; Fig. 9). However, this emphasized that the change over the last 30 yr is unusual and is (at least for the contiguous United States, South Africa, and central Mexico; cf. Fig. 7; Fauchereau et al. 2003; and the next section) a relatively new phenomenon.

h. Central Mexico

Central Mexico is reasonably well covered with precipitation data for the past 60 yr (Figs. 10 and A7). The North American monsoon is the major cause of summer rainfall away from the coastal areas in the central part of the country. Figure 10 shows that during the past 30 years a substantial precipitation decrease has occurred over the Central Plateau of Mexico. The frequency of heavy precipitation events (those above the 25 to 35 mm thresholds in this region) generally followed the tendency of the mean totals (although these changes were statistically insignificant). However, the frequency of very heavy precipitation (above the upper 1% and 0.3% of the rain events or above 55 and 75 mm, respectively) increased during the same 30-yr-long period. The frequency of very heavy rain events (above the upper 0.3%) has increased substantially [by 110% $(30\ \mathrm{yr})^{-1}$]. Thus, while in the early 1970s the average return period of such events was approximately 12 yr, in the early 2000s it is estimated to be around 5 yr.

i. Summary of observed trends in heavy and very heavy precipitation

Figure 11 is a substantial update of the map from Easterling et al (2000c) where signs (+ and −) show the regions with changes in heavy precipitation found in our studies and in the studies of others that follow the pattern outlined above: changes in mean precipitation are less or insignificant while changes in heavy/very heavy precipitation are statistically significant. The shaded regions in this figure are results from our work, although others have studied some of these regions as well and reached similar conclusions, for example, Stone et al. (2000) for Canada, Iwashima and Yamamoto (1993) for Japan, Roy and Balling (2004) for India, Osborn et al. (2000) for the United Kingdom, Tarhule and Woo (1998) for Nigeria, Zhai et al. (1999, 2005) for China, Kunkel et al. (1999) and Kunkel (2003) for the United States,[8] Alpert et al. (2002) for eastern

[8] More recent work by Kunkel et al. (2003) found similar in-

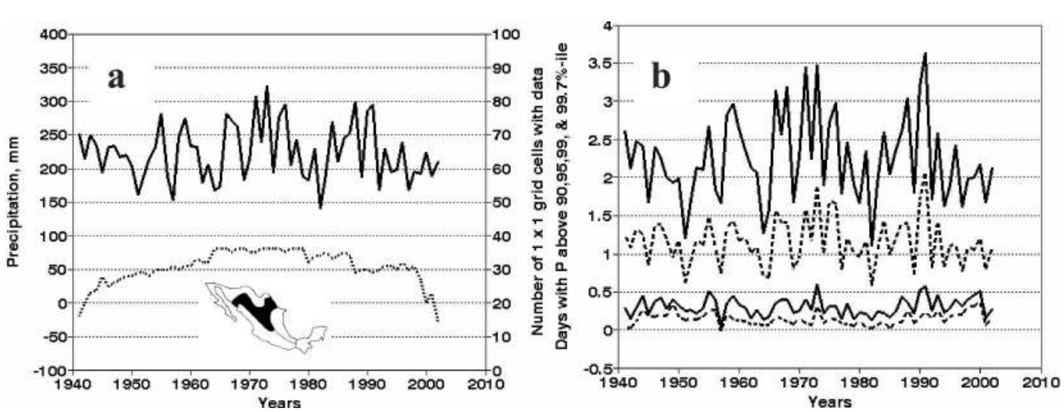


FIG. 10. Mexico, Central Plateau (region is shown darkened in the map). Summer precipitation and number of days with heavy (above 90 and 95 percentile) and very heavy (above 99 and 99.7 percentile) precipitation. Since 1970, more than a twofold increase in the frequency of days with very heavy precipitation (above 99.7 percentile) and a decrease in precipitation [by 20% $(30\ \mathrm{yr})^{-1}$] are statistically significant at the 0.05 level.

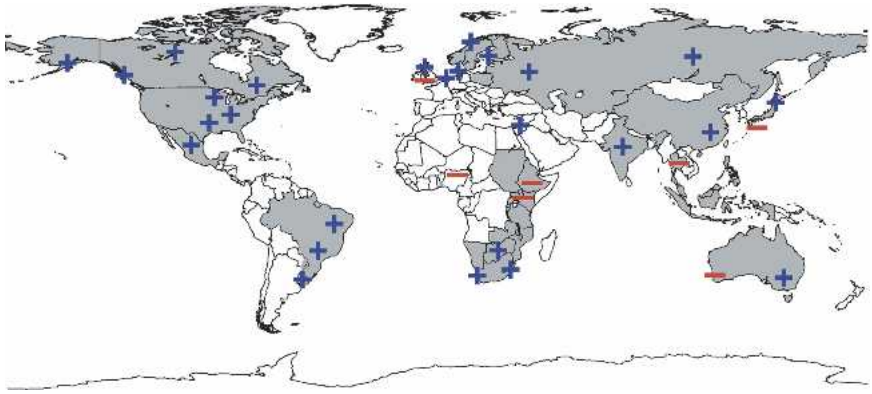


FIG. 11. Regions where disproportionate changes in heavy and very heavy precipitation during the past decades were documented compared to the change in the annual and/or seasonal precipitation (Easterling et al. 2000c, substantially updated). Thresholds used to define heavy and very heavy precipitation vary by season and region. However, changes in heavy precipitation frequencies are always higher than changes in precipitation totals and, in some regions, an increase in heavy and/or very heavy precipitation occurred while no change or even a decrease in precipitation totals was observed.

Mediterranean, and Haylock and Nicholls (2000) for Australia.

6. Model projections

What are the causes of all the observed trends in intense precipitation? Unfortunately, observed data and/or their analysis alone cannot provide answers to this question. These questions are better addressed by physical models of the global climate system. Precipitation is a variable that is more difficult to simulate than other meteorological variables (McAvaney et al. 2001).[9] Most global climate models project an increase in precipitation intensity with global warming (Schaer et al. 1996; Jones et al. 1997; Hennessy et al. 1997; Mason and Joubert 1997; Zwiers and Kharin 1998; Meehl et al. 2000; Kharin and Zwiers 2000; Cubash and Meehl 2001; Wilby and Wigley 2002; Allen and Ingram 2002; and Hegerl et al. 2003, 2004). Recently, Karl and Trenberth (2003) showed that even without change in total precipitation, there was an increase in the frequency of intense daily precipitation when comparing warmer with cooler climates. There are physical explanations for the reason that changes in heavy precipitation should be more pronounced than changes in precipitation totals in ongoing climatic change (Trenberth 1999; Groisman et al. 1999a; Bengtsson 2001; Allen and Ingram 2002; Trenberth et al. 2003). These explanations will be further discussed in section 7.

Climate models simulate a global-scale increase in mean precipitation due to increased greenhouse gases (Cubash and Meehl 2001). Changes in global mean annual precipitation from an ensemble of coupled model simulations driven by both natural (changes in volcanism and solar forcing) and anthropogenic forcing closely follow the observed trajectory (Allen and Ingram 2002). However, this similarity is caused only by the effect of volcanic eruptions on global precipitation and the anthropogenic signal cannot presently be detected (Lambert et al. 2004; Gillett et al. 2004). Furthermore, the spatial patterns of annual precipitation change are very model-dependent and poorly correlated between different models at the time of the CO_2 doubling (Cubash and Meehl 2001; Hegerl et al. 2004). However, simulated changes in heavy and very heavy precipitation tend to become more positive and stronger than in mean precipitation (Allen and Ingram 2002; Semenov and Bengtsson 2002). The spatial pattern of changes in heavy precipitation shows more regions of increase than the pattern of annual precipitation changes, and hence more similarity between models (Figs. 12–14). This finding of a more consistent pattern of increase in heavy precipitation in climate model simulations is supported by a comparison of two global circulation model (GCM) projections of the warming effects on precipitation associated with the doubling of CO_2 in the atmosphere. Hegerl et al. (2003, 2004) assessed projected changes at the time of CO_2 doubling, but here we highlight additional results from that experiment for North America. The pattern of the difference for annual precipitation and various indices of heavy and very heavy precipitation were analyzed between the "average climate" for the 20 yr 2040–60 in

creases for the United States during the twentieth century, but also found evidence that there was a period of increased heavy rainfall events during the 1890s over the western part of the country.

[9] However, reanalyses-based studies (e.g., Janowiak et al. 1998; Widmann and Bretherton 2000) showed that temporal variability in model-simulated precipitation can be quite realistic while the errors in climatological precipitation remain very large.

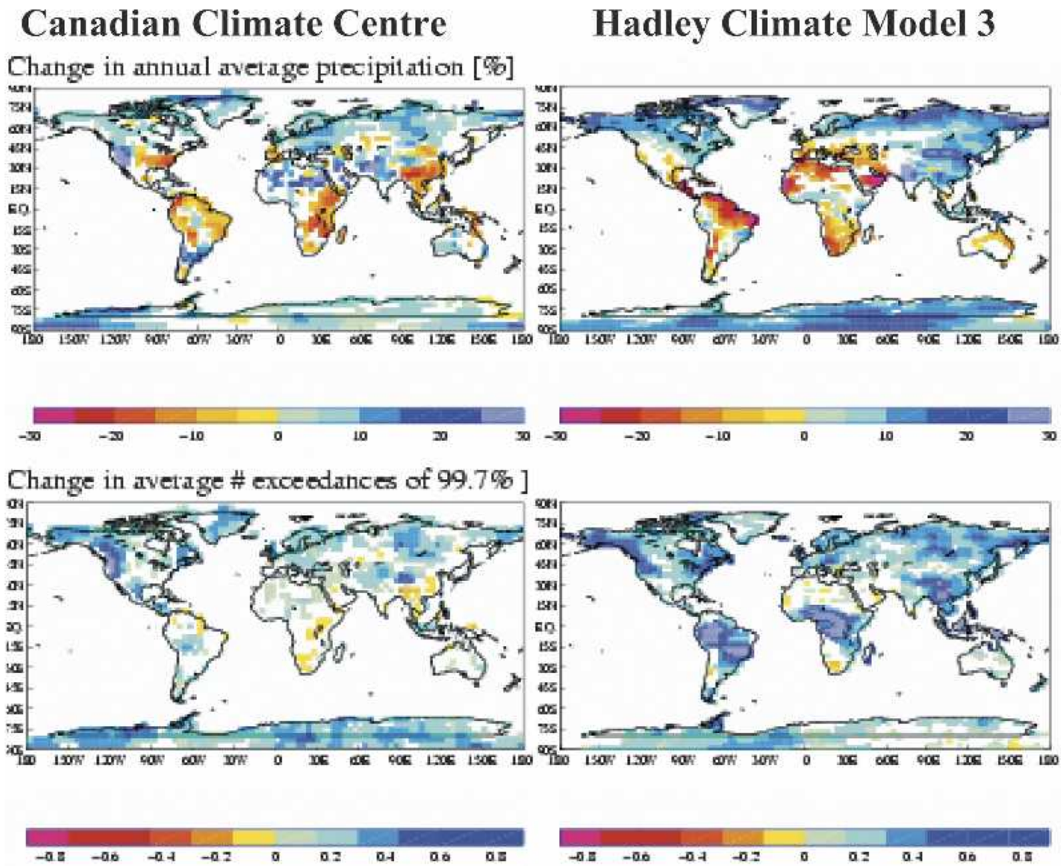


FIG. 12. (top) Model simulations of the effect of CO_2 doubling on annual precipitation (%) and (bottom) the average number of exeedences of the 99.7 percentile of precipitation distribution from the HadCM3 and CGCM1 models. The red end of the scale depicts decreases and the blue increases. Changes are only shown where they are significant relative to the control climate (from archive of Hegerl et al. 2003, 2004).

the $2 \times CO_2$ simulation (doubled CO_2 achieved in latter half of the twenty-first century), and conditions at the end of the twentieth century.[10] This was done for three-member ensemble simulations for each of two models:

- The Canadian Climate Centre model CGCM1, which is the first version of the coupled climate model from the Canadian Centre for Climate Modelling and Analysis (CCCma; Flato et al. 2000; Boer et al. 2000a, b). The atmospheric component of this model has a resolution of 3.75° latitude × 3.75° longitude.[11]
- The third cycle of the Hadley Centre climate model (Gordon et al. 2000; Pope et al. 2000; Johns et al. 2002; see also Stott et al. 2001). This model has the same longitudinal resolution and a higher latitudinal resolution of 2.5°.

While the historic change in atmospheric composition for non-greenhouse-gas anthropogenic forcing is implemented differently in both models, it is expected that by the middle of the twenty-first century, differences between both ensembles will be largely due to differences in the model response. These two coupled climate models show reasonably realistic five-day extreme rainfall compared to observational and/or reanalysis products (Kharin and Zwiers 2000; Hegerl et al. 2004). The errors and uncertainties are larger in the Tropics, where, despite substantial differences in climatological mean and extreme rainfall, both models simulate moderately similar changes in extreme rainfall, with global mean land precipitation changes in anthropogenic climate change simulations being reasonably similar between the models (Hegerl et al. 2004).

Figure 12 compares the change in annual total precipitation and in the number of exceedences of the 99.7 percentile in two climate models. The comparison re-

[10] For signal detection purposes, the use of the earlier period as reference period, which is less affected by an anthropogenic increase in greenhouse gases concentrations, would be preferable. However, the end of the twentieth century was selected as reference period in this study (as well as in Hegerl et al. 2004), because much more station data are available at that time. This facilitated the comparison between model simulated and observed trends.

[11] Changes in the second version of this model (CGCM2) were quite similar and, therefore, are not assessed here.

veals how inconsistent the pattern of total precipitation changes is compared to that of the change in the frequency of the wettest days in the model output records. Both models show a pronounced pattern of changes of opposite signs with a global mean precipitation increase of between 1% and 2% (Fig. 12; Hegerl et al. 2004), but these patterns correlate poorly between the models. When the change in the upper 0.3% of precipitation events is studied, both models show an increase in precipitation extremes over most regions of the globe, although regional details still differ (Fig. 12). When compared to observed changes in very heavy precipitation reported in this study (Fig. 11), both GCMs predict increases in very heavy precipitation in Siberia, Fennoscandia, central and eastern Europe, southeastern Australia, the northwestern coast of North America, northern Canada, and central Mexico in agreement with the sign of the observed trends. However, both GCMs predict a decrease in very heavy precipitation in South Africa contradicting the sign of the presently observed trends.

Figure 13 shows the "consensus" climate change pattern for the annual mean precipitation over North America. The pattern is based on the average of both model's simulations. Values are plotted only where the changes in both models are consistent (i.e., the changes in the ensemble simulations with both models at a grid box are not significantly different at the 90% level based on a Mann–Whitney rank test). This represents a pattern of precipitation change that is robust between both models. The difference is striking in this consensus intercomparison. Over most of the continent the change, often even the sign of change in total precipitation, is different between both models (Fig. 13). In a different way, the consensus change in precipitation on the wettest day per year and (to a lesser extent) in the wettest five consecutive days per year, precipitation increases everywhere over North America (Fig. 14). This increased similarity in the patterns of change should lead to a more robust signal in heavy rainfall compared to annual mean precipitation. A simple detection analysis using model data suggests that changes in extreme rainfall may therefore be more robustly detectable than changes in annual total rainfall (Hegerl et al. 2004). By the time of CO_2 doubling, annual total rainfall changes simulated in one model could not be reliably detected using fingerprints of climate change based on the other model, while changes in heavy rainfall could be confidently detected by that time between the model simulations. If this finding based on two models can be generalized, it suggests that observed changes of very heavy precipitation are easier to detect and attribute to global warming than changes in the mean annual or seasonal precipitation.

The results based on GCMs are more credible when simulated variability and trends roughly correspond to

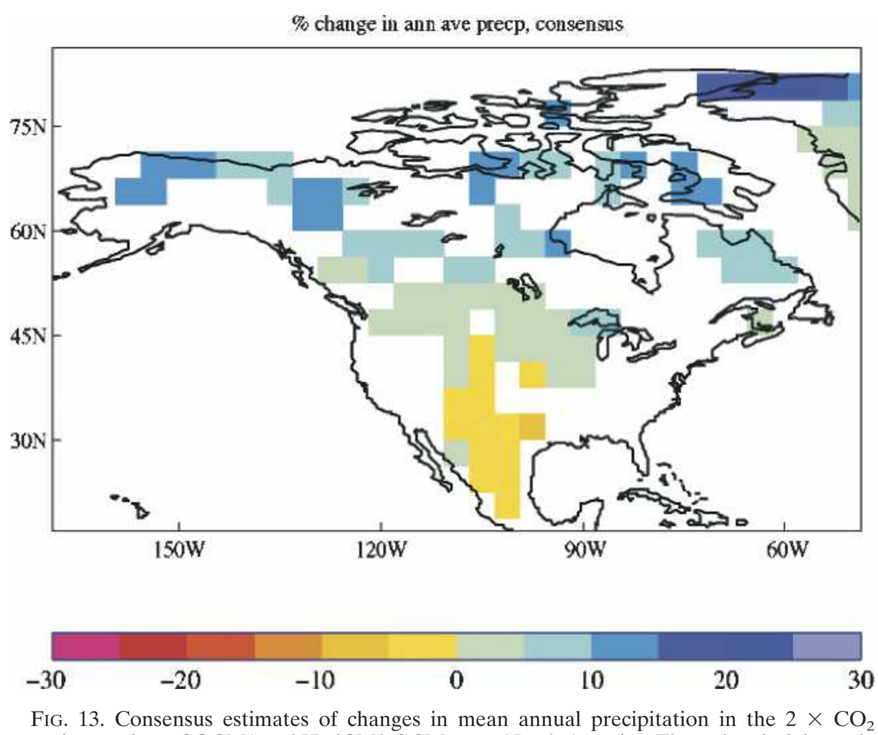


FIG. 13. Consensus estimates of changes in mean annual precipitation in the 2 × CO_2 experiments from CGCM1 and HadCM3 GCMs over North America. The red end of the scale depicts decreases and the blue increases. The pattern shows the average precipitation change between the models, it is only shown where the simulations with each model are consistent with the respective other model at the gridpoint level.

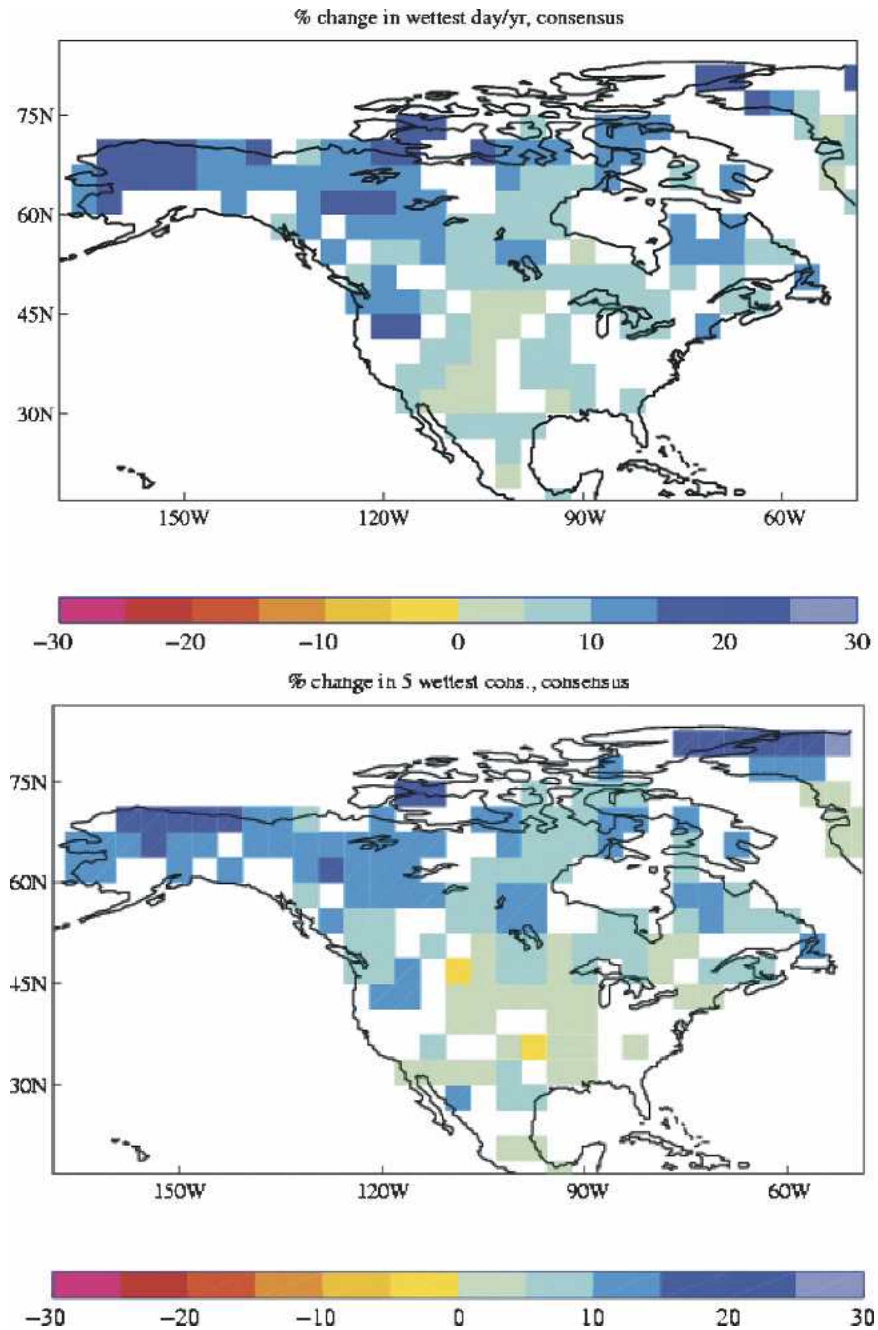


Fig. 14. (top) Consensus estimates of changes in the wettest day per year and (bottom) the wettest 5-day accumulation per year precipitation in the 2 × CO_2 experiments from CCC and HadCM3 GCMs (see Fig. 13 for details).

those observed during the past century. Unfortunately, the direct intercomparison of transient climate change in heavy precipitation as reproduced by the GCM and the observed change is not a frequent exercise and is hampered by differences in the scale represented by model data (smooth grid boxes) and observed station indices. One such example is presented and discussed below.

In their assessment of heavy rainfall changes in the transient greenhouse gas simulation using a coupled atmosphere–ocean global circulation model (ECHAM4/OPYC3 for 1900–2099), Semenov and Bengtsson (2002; their Fig. 4) looked at results of the simulated changes of the contribution of the upper 10% quantile of daily precipitation to the annual total precipitation over the contiguous United States for the twentieth century. These were compared with the empirical results of Karl and Knight (1998). A close resemblance of positive trends, mean values, and the amplitude of the interdecadal variability suggests that the observed changes are in part related to an increase in greenhouse gases. The only other region of the United States that was assessed by Semenov and Bengtsson (2002) was the northeastern quarter of the country, which roughly corresponds to the Northeast and Midwest regions used in the study conducted by Groisman et al. (2004).

Figures 15 and 16 show the changes in the number of days with heavy precipitation and wet days in the eastern United States from model results (Semenov and Bengtsson 2002) and observations (Groisman et al. 2004). Comparing observations in this region with changes in the number of days with heavy precipitation from this model, we did not expect that year-to-year variations in the GCM simulation and observations would coincide. Moreover, the model grid cells generate precipitation twice as frequently as observed in point observations. However, this comparison reveals a similar increase of approximately 10% over the twentieth century in both time series. This model projects a substantial increase (up to 40% to the end of the twenty-first century) of days with heavy precipitation. In spite of the above-mentioned similarities, these are obviously two very different estimates of climate change over the northeastern United States. For example, a century-long change during the twentieth century was shown during the first 70 yr in the model simulation, while the opposite is true in observations. This points to the need for ensemble assessments of future climate projections (Kattsov and Walsh 2000; Kharin and Zwiers 2002).

It is clear that if mean precipitation does not change appreciably compared with the highest part of the precipitation distribution, then the frequency of precipitation events will be affected. Indeed, in several regions of the world this appears to be the case. For instance, in South Africa, Siberia, eastern Mediterranean, central Mexico, and northern Japan, an increase only in heavy and/or very heavy precipitation is observed while total precipitation and/or the frequency of days with an ap-

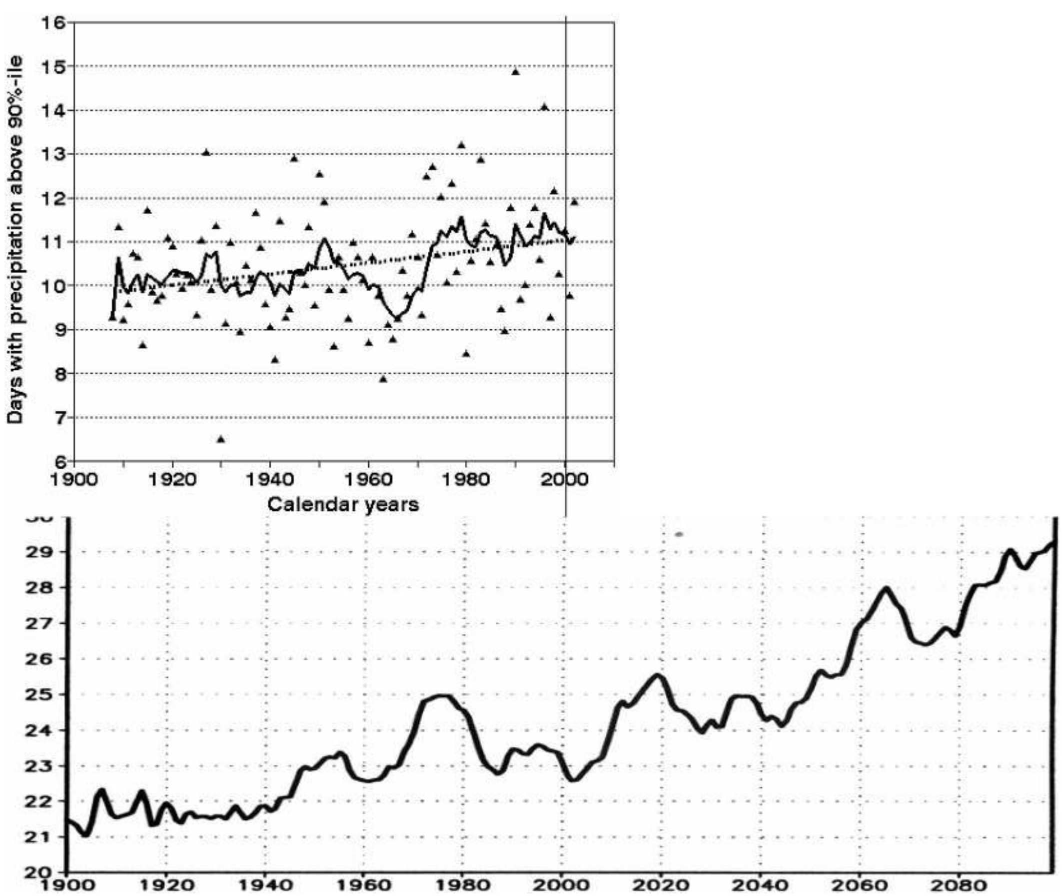
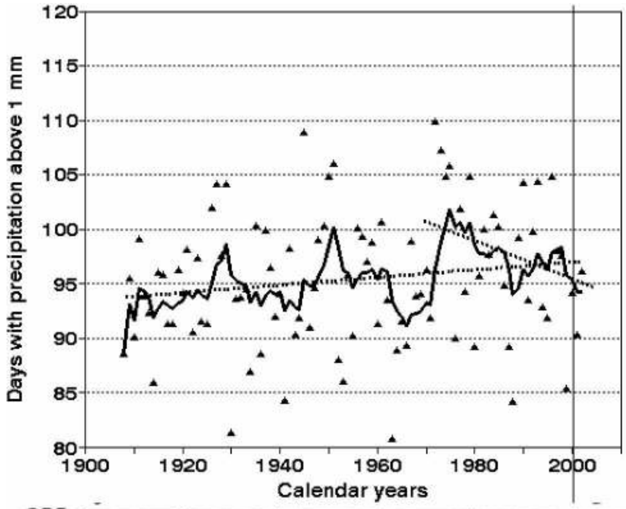


FIG. 15. Frequency of the upper 10% of rainy days over the northeastern quadrant of the contiguous United States. (top) Observations (Northeast and Midwest regions; regions 4 and 8 in Fig. 1) for 1908–2002. Annual values (triangles), 10-yr running mean values (solid line), and linear trend (dotted line) are shown. Linear trend for 1908–2002 [12% $(100\ \mathrm{yr})^{-1}$] is statistically significant at the 0.05 level. (bottom) ECHAM4 (35°–45°N; 75°–85°W; adapted from Semenov and Bengtsson 2002) 10-yr running mean values for 1900–2090.



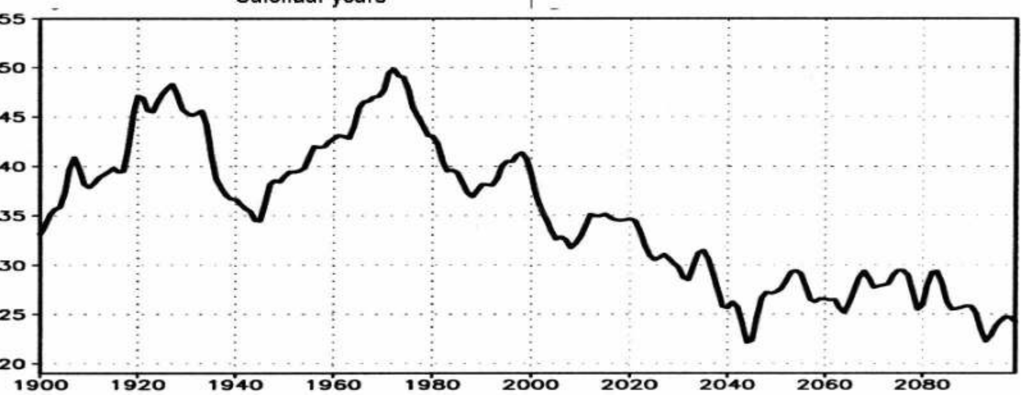


FIG. 16. Number of wet days over the northeastern quadrant of the contiguous United States. (top) Observations (Northeast and Midwest regions) for 1908–2002. Annual values (triangles), 10-yr running mean values (solid line), and linear trends (dotted lines) are shown. Linear trends for 1908–2002 [4% $(100\ \mathrm{yr})^{-1}$] and for 1972–2002 [−9% $(30\ \mathrm{yr})^{-1}$] are statistically significant at the 0.05 level. (bottom) ECHAM4 (35°–45°N; 75°–85°W; adapted from Semenov and Bengtsson 2002) 10-yr running mean values for 1900–2090. Both show a decrease during the past 30 yr.

preciable amount of precipitation (wet days) are not changing and/or are decreasing (Figs. 7 and 10; Sun and Groisman 2000; Easterling et al. 2000c; Alpert et al. 2002). The first indication that this feature might also be present in the contiguous United States is shown in Fig. 16 for the northeastern part of the country. Observations show that the annual number of wet days has increased during the past 100 yr, but during the last 30 yr (exactly at the time when most of increase in very heavy precipitation started) a decrease in the number of wet days was observed. This figure is presented to compare the observations with the variations of the regional number of wet days reproduced by the ECHAM4/OPYC3. We see a similarity of tendencies and a very strong decrease in the wet-day frequency projected for the twenty-first century by the model. However, once again the reliability of the model simulation is somewhat hampered by an exaggerated number of days with model-generated precipitation (on average annually, 240 of modeled days are wet, which is much higher than observed).

7. Discussion

A physical explanation for an increase in heavy precipitation with global warming is provided by Trenberth et al. (2003). An expanded body of evidence for the twentieth century presented in this paper appears to support this concept for very heavy precipitation as well. Global warming, which has been especially pronounced during the recent decades in extratropical land areas and in minimum temperatures (Karl et al. 1991; Folland and Karl 2001), is related to a reduction in spring snow cover extent (Brown 2000; Groisman et al. 1994, 2001), and thus to an earlier onset of spring- and summerlike weather conditions (Easterling 2002). Warming also relates to a higher water vapor content in the atmosphere (Douville et al. 2002; Trenberth et al. 2003), which has been documented in many regions of the world (Sun et al. 2000; Ross and Elliott 2001). This in turn results in an increase in the frequency of Cumulonimbus clouds [documented for the former USSR and the contiguous United States by Sun et al. (2001)], which is related to the general increase in thunderstorm activity [documented for most of the contiguous United States by Changnon (2001)]. This development can explain an observed widespread increase in very heavy precipitation in the extratropics. Furthermore, in humid regions an increase in summer minimum temperatures is related to an increase in the probability of severe convective weather (Dessens 1995) and is likely related to changes in the frequency of heavy and very heavy rain events. It is difficult to directly relate estimates of changes in very heavy precipitation with flooding (e.g., Groisman et al. 2001; Kunkel 2003). However, great floods have been found to be increasing in the twentieth century (Milly et al. 2002).

We are confident in our finding that very heavy precipitation has increased during the period of instrumental observations over most of the contiguous United States. Characteristics used to define very heavy precipitation (frequencies) are robust and have an advantage of being insensitive to scaling errors. Clearly, more work is needed, but the evidence is growing that the observed historical trends of increasing very heavy precipitation are linked to global warming. Simulated changes in intense precipitation and precipitation extremes are generally greater than in mean precipitation and are consistent among the models studied here. It is likely that changes in heavy precipitation will probably be more easily detected than changes in annual mean precipitation in the future climatic changes.

8. Summary and conclusions

An empirical assessment of observed changes of the characteristics of intense precipitation (mostly, the frequency of very heavy precipitation defined as the upper 0.3% of daily precipitation events), and analysis of the output of three GCM simulations with transiently increasing greenhouse gases during the twentieth and twenty-first centuries have been conducted for over half of the land area of the globe (Fig. 11). In summary, these are the major findings:

- Reliable assertions of very heavy and extreme precipitation changes are possible only for regions with dense networks due to a small radius of correlation for most of intense precipitation events.
- In the midlatitudes, there is a widespread increase in the frequency of very heavy precipitation during the past 50 to 100 yr.
- By raising the thresholds for the definition of very heavy precipitation and providing empirical evidence of changes in the frequency of these events, we can better provide a basis for impact assessments of the consequences of these changes, including landslides, floods, and soil erosion.
- Three model projections of a greenhouse-enriched atmosphere and the empirical evidence from the period of instrumental observations indicate an increas-

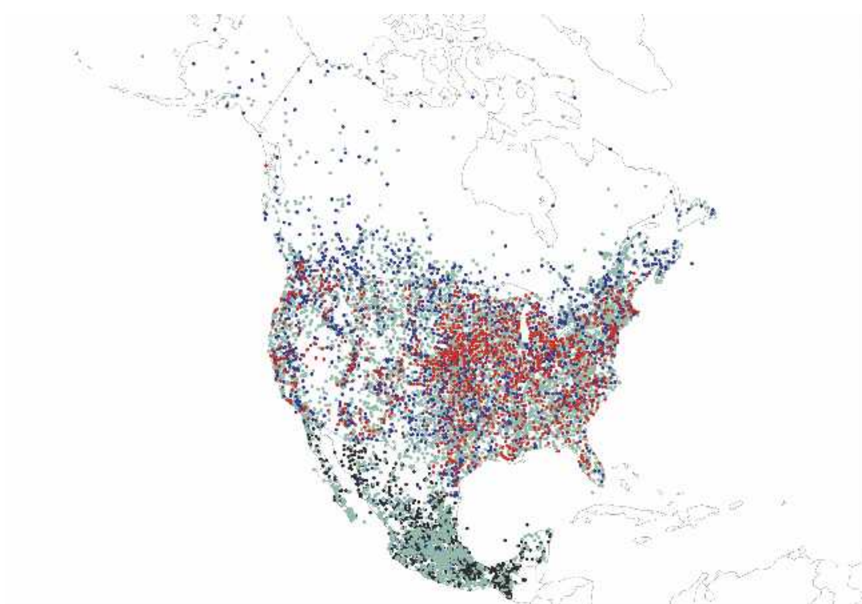


FIG. A1. Present coverage of North America south of 55°N with long-term (at least 25 yr of data) stations. Red and blue dots show stations with ~100 and 80 yr of data, respectively. Green dots indicate stations with at least 25 yr of data during the 1961–90 period. Black dots show additional long-term Mexican stations that are presently undergoing extensive quality control.

TABLE A1. Parameters of statistical structure of the fields of regional frequency of heavy (above the upper 10 percentile) and very heavy (above the upper 0.3 percentile) seasonal precipitation events over the contiguous United States and the European part of the former Soviet Union. Parameters of the spatial correlation function in Eq. (2) (ρ_0 and C_0) and variance coefficient, C_v, of the 1°-gridcell-averaged values of the frequencies are presented.

Region	Season	Events	ρ_0, km	C_0	C_v
USA, Northwest	Winter	Heavy	505	0.85	0.40
	Annual		325	0.85	0.20
	Winter	Very heavy	250	0.55	0.85
	Annual		160	0.65	0.45
USA, Southwest	Winter	Heavy	300	0.90	0.50
	Annual		255	0.90	0.20
	Winter	Very heavy	125	0.70	0.75
	Annual		95	0.65	0.30
USA, Midwest	Summer	Heavy	190	1.00	0.20
	Annual		300	1.00	0.15
	Summer	Very heavy	95	0.65	0.40
	Annual		110	1.00	0.30
USA, Southeast	Summer	Heavy	270	0.75	0.15
	Annual		420	0.85	0.15
	Summer	Very heavy	155	0.55	0.45
	Annual		155	0.85	0.40
European part of Russia, north of 60°N, south of 66.7°N	Summer	Heavy	265	0.75	0.20
	Annual		400	0.65	0.12
	Summer	Very heavy	135	0.30	0.55
	Annual		215	0.30	0.30
European part of the former USSR, south of 60°N	Summer	Heavy	275	0.70	0.15
	Annual		400	0.75	0.12
	Summer	Very heavy	200	0.15	0.25
	Annual		135	0.40	0.25

ing probability of heavy precipitation events for many extratropical regions including the United States.

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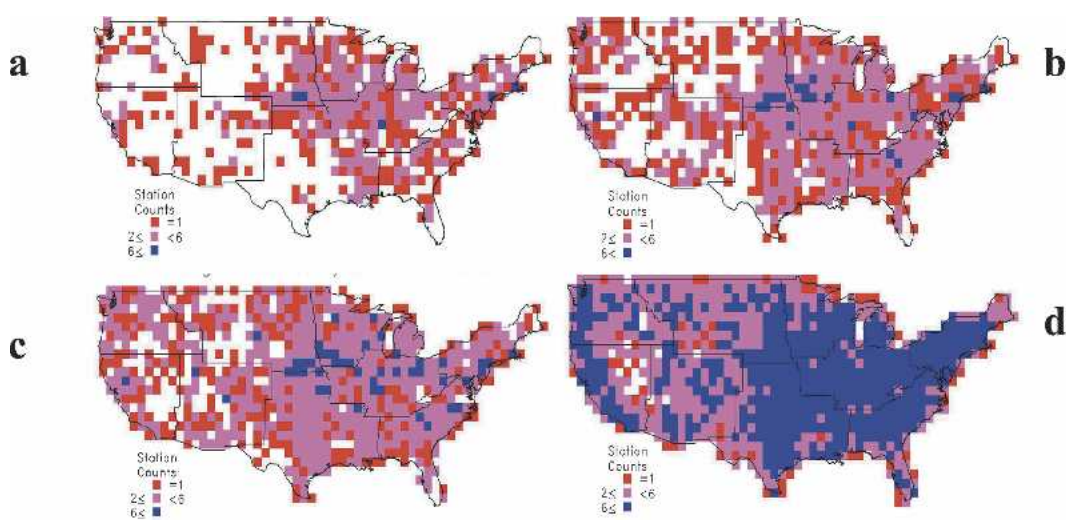

FIG. A2. Data availability over the contiguous U.S. generalized within the 1° × 1° grid cells for the periods of a near-constant network**.** The regional partition of the country used throughout this paper for area averaging is also shown. Counts of stations with more than 50% of daily precipitation data within each grid cell for the (a) 1891–1900, (b) 1901–10, (c) 1911–20, and (d) 1951–2002 periods, respectively.

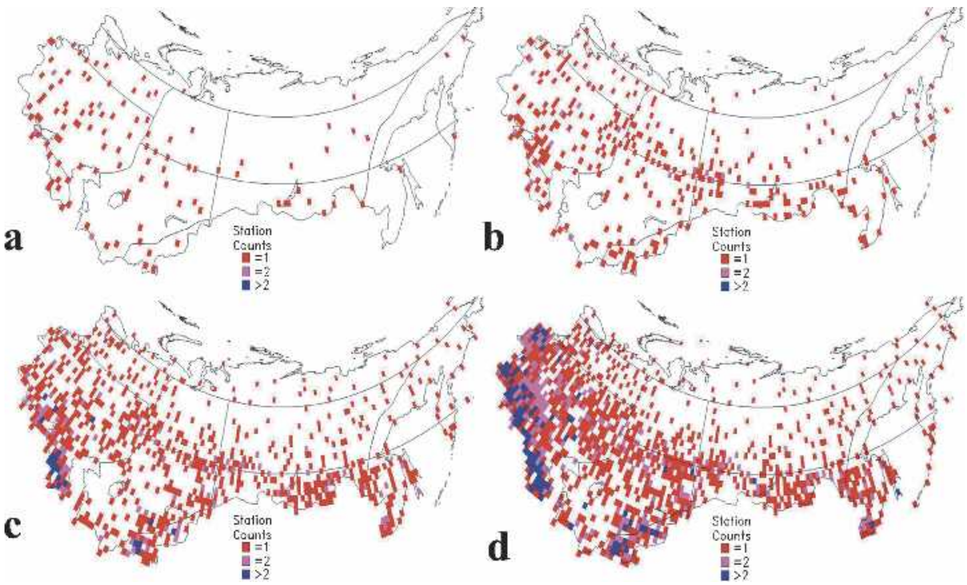

FIG. A3. Same as Fig. A2, but over the former USSR for the (a) 1901–10, (b) 1931–40, (c) 1941–50, and (d) 1951–96 periods, respectively.

APPENDIX

Availability of Long-Term Stations

Figure A1 shows the availability of long-term stations in North America used in this study. It can be noted that for most of the contiguous United States there is a sufficient number of century-long daily precipitation time series available. The following set of similar figures (Figs. A2–A7) provide counts of stations with precipitation data within each 1° × 1° grid cell for periods with relatively stable networks for regions used in this study. A station was considered present during the listed period when it had at least 50% of daily precipitation data within it. The figures also depict the regions used for area averaging for the contiguous United States, former USSR, Australia, South Africa, Brazil, and Mexico. Area averaging can be counterproductive and even misleading when changes of opposite signs have occurred within the region. Therefore, we first selected regions that can be considered as relatively climatologically homogeneous according to independent criteria and then applied the area-averaging routine. To preserve the comparability of results, the regional partition for the conterminous United States remains the same as used in Karl and Knight (1998) and Groisman et al. (2001, 2004). When no published climate regions could be found for a country (e.g., South Africa, Brazil, and Mexico), we created them using a combination of climate classifications from Trewartha [available on line at http://fp.arizona.edu/khirschboeck/climate/Trew.map.large.htm], plots of seasonal precipitation averages, and considerations of terrain, vegetation, latitude, and data availability.

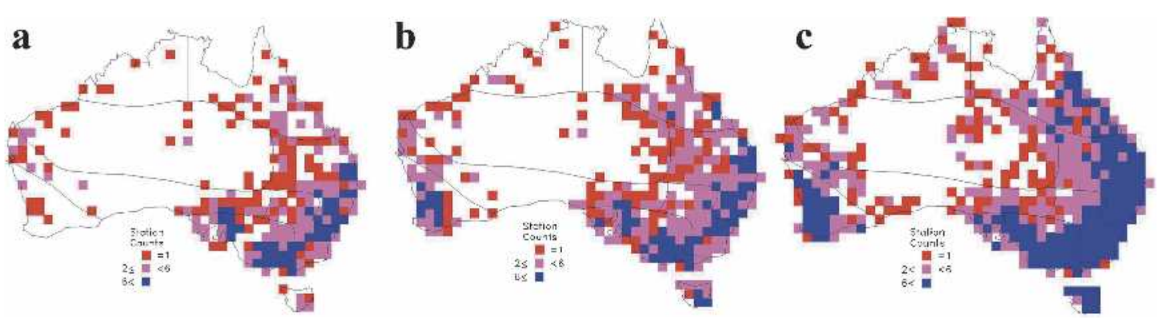

FIG. A4. Same as Fig. A2, but over Australia for the (a) 1901–10, (b) 1911–20, and (c) 1951–99 periods, respectively. When creating the climate regions for the continent, we primarily used the climate zones based upon rainfall published by the Australian Bureau of Meteorology [available online at http://www.bom.gov.au/lam/climate/levelthree/ausclim/zones.htm]. The regions were modified somewhat using plots of the seasonal precipitation cycle and data availability considerations.

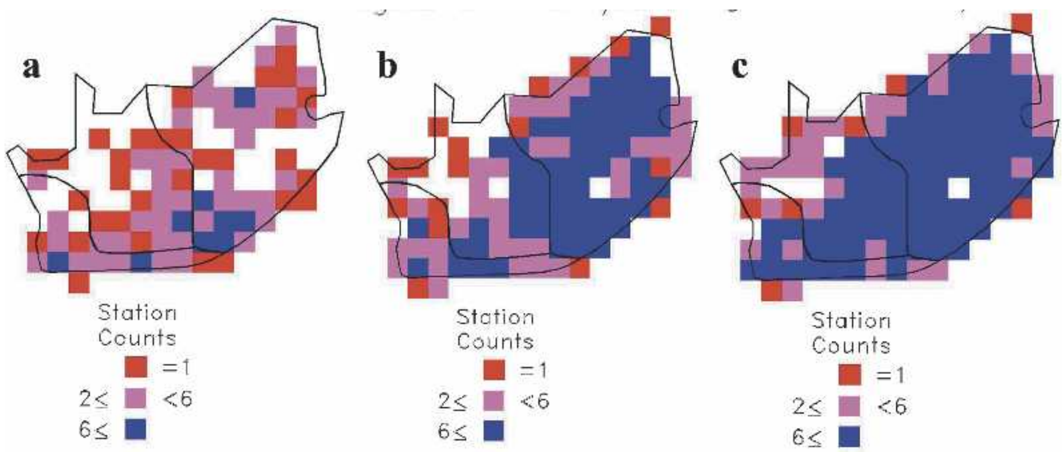

FIG. A5. Same as Fig. A2, but over South Africa for the (a) 1901–10, (b) 1921–30, and (c) 1931–97 periods, respectively. Three climatological regions were selected by taking into account the seasonal precipitation cycle. In the eastern half of the country the precipitation maximum is observed in austral summer (DJF) while along the Atlantic coast it is observed in winter. The desert region between the two has an annual precipitation of ~250 mm.

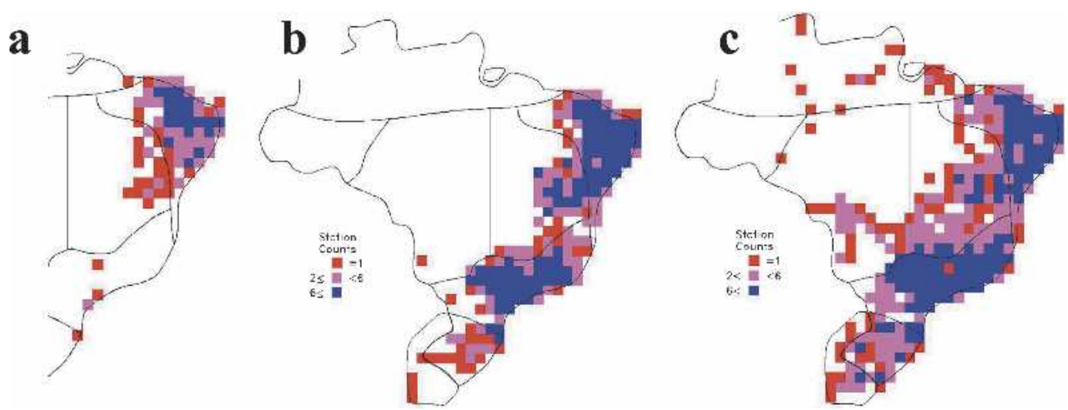

FIG. A6. Same as Fig. A2, but over Brazil for the (a) 1911–20, (b) 1941–50, and (c) 1951–2001 periods, respectively. The Nord-Este region has an autumn (March–May) precipitation maximum and two extratropical regions have a summer precipitation maximum (DJF). The southernmost region was expanded beyond the national borders to include Uruguay and adjacent small areas of Argentina and Paraguay, which were also well covered by the observations in our dataset.

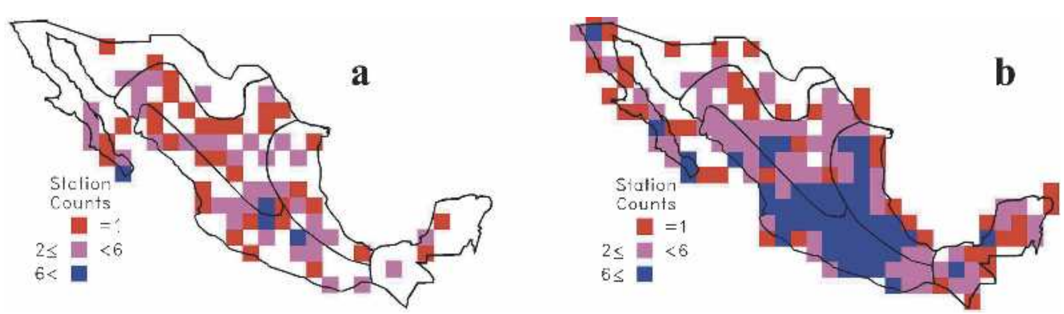

FIG. A7. Same as Fig. A2, but over Mexico for the (a) 1941–50 and (b) 1951–2002 periods.

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