12.5. CHANGES IN PRECIPITATION DISTRIBUTION SPECTRA AND CONTEMPORARY WARMING OF THE EXTRATROPICS: IMPLICATIONS FOR INTENSE RAINFALL, DROUGHTS, AND POTENTIAL FOREST FIRE DANGER

Pavel Ya. Groisman*
UCAR Project Scientist at National Climatic Data Center, Asheville, North Carolina, USA
Richard W. Knight
University of North Carolina, Asheville, North Carolina, USA
David R. Easterling, David H. Levinson, Richard R. Heim, Jr., and Thomas R. Karl
National Climatic Data Center, Asheville, North Carolina, USA
Paul H. Whitfield
Meteorological Service of Canada, Vancouver, British Columbia, Canada
Gabriele C. Hegerl
Nicholas School of the Environment and Earth Sciences, Duke University, Durham, North Carolina, USA
Vyacheslav N. Razuvaev and Boris G. Sherstyukov
Russian Institute for Hydrometeorological Information, Obninsk, Russia
Jesse G. Enloe and Nina S. Stroumentova
STG, Incorporated, Asheville, North Carolina, USA

1. INTRODUCTION

Significant climatic changes over the high latitudes in the 20th century (Fig. 1) have been reflected in many atmospheric, oceanic, and terrestrial variables. Changes in surface air temperature, precipitation, growing season duration, and snow cover cause changes in numerous derived variables of economic, social and ecological interest, including the natural frequency of forest fires, droughts, frequency of intense precipitation and of wet days (Karl and Knight 1998; Folland and Karl 2001; Nicholson 2001; Frich et al. 2002; Heim 2002; Heim et al. 2003; Zhang et al. 2003; Groisman et al. 2003a,b, 2004, 2005; Zhai et al. 2004, etc.). In this paper, we (1) present the summary of our (and others) studies of changes in precipitation distribution by daily intensities, e.g., Fig. 2, Section 3; (2) present/summarize the evidence of changes in high latitudinal land areas that may imply drier summer conditions (Section 4); (3) present the current state of

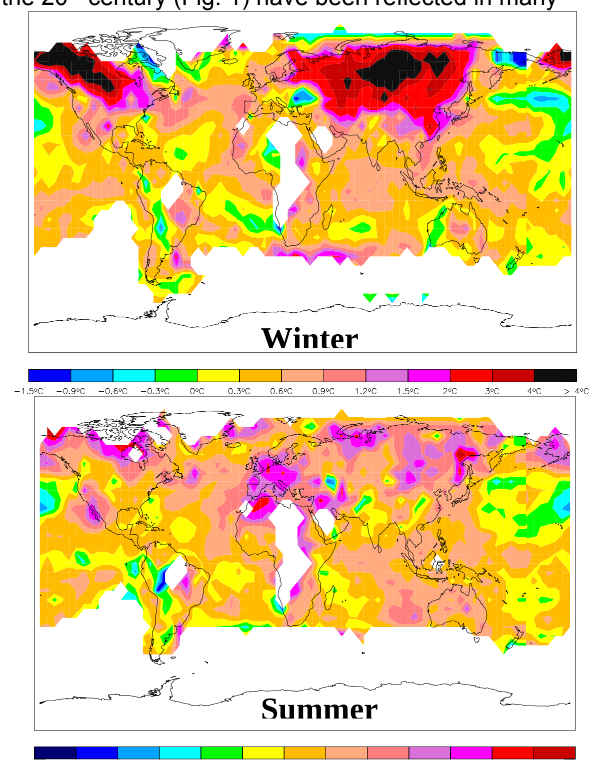

Figure 1. Mean seasonal temperature change 1965 to 2004 over the globe. Data source: (Jones and Moberg 2003, updated). Processed by the NCDC Global Climate at the Glance Mapping System.

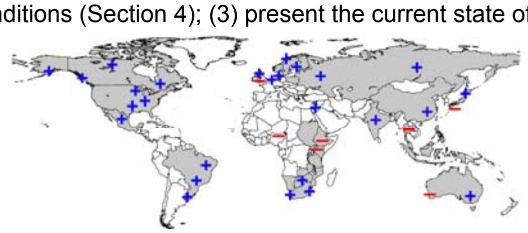

Figure 2. Regions where disproportionate changes in heavy and very heavy precipitation during the past decades were documented compared to the change in the annual and/or seasonal precipitation. Thresholds used to define "heavy" and "very heavy" precipitation vary by season and region. However, changes 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 (Groisman et al. 2005). This map summarizes our results (over shaded regions) as well as the results of others (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; Osborn et al. 2000; Tarhule and Woo 1998; Suppiah and Hennessy 1998; Zhai et al. 1999, 2004).

* *Corresponding Author Address*: Pavel Ya. Groisman. National Climatic Data Center, Federal Building, 151 Patton Avenue, Asheville, North Carolina,USA; e-mail: Pasha.Groisman@noaa.gov.

evidence of changes in drought indices, which in high latitudes correspond to characteristics of the potential forest fire danger (Sections 5 and 6); briefly discuss our findings (Section 7) and, in Appendix, (4) provide the rationale for the approach used to analyze changes in the frequency of rare events (such as very heavy precipitation) based on assessment of the statistical structure of the field of these events and representativeness of the network at hand.

2. DATA

Global daily precipitation data sets accumulated at NCDC (Fig.3) were used for analyses of changes in precipitation distribution spectra over the globe (Easterling et al. 2000; Groisman et al. 1999, 2005). For analyses of climatic changes in Northern Hemispheric extratropics (including extreme events and potential forest fire danger changes), we performed our

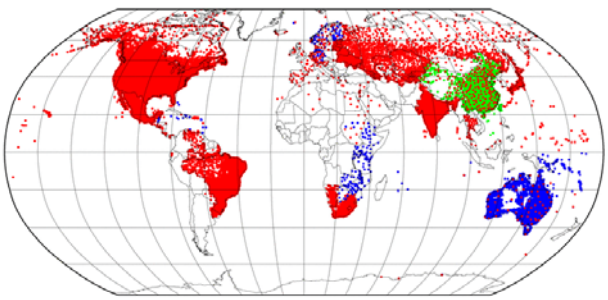

Figure 3. **Map of stations with daily precipitation available at the U.S. National Climatic Data Center (as of July 15, 2003). A subset of ~32,000 is available through Global Daily Climatology Network, GDCN (Gleason et al. 2002; red dots). Only typhoon-related precipitation data are available for most of stations from China (green dots).**

analysis using a subset of about 1500 stations north of 50ºN from the recently created Global Daily Climatology Network archive (GDCN, Gleason et al. 2002; Fig. 4), the U.S. cooperative network (more than 6000 long-term stations), and a subset of the Global Synoptic Data Network for the former USSR (the former USSR stations are practically the same as those shown in Fig. 4). When considering temperature variations over Canada we included a subset of 210 Canadian stations with homogenized temperature time series (Vincent and Gullett 1999). A priority was given to these high-quality data instead of the observations in the GDCN. Precipitation time series for the former USSR (fUSSR) and Canada were homogenized to account for changes in instrumentation and observational practice as described in Groisman and Rankova (2001), NCDC (1998), and Groisman (2002). Figs. 3 and 4 show the GDCN stations with precipitation daily data. Generally, the number of stations with daily precipitation *and* temperature data available in high latitudes is somewhat less than shown in Fig. 4.

3. PRECIPITATION DISTRIBUTION SPECTRA

We analyzed the changes in the precipitation distribution spectra for more than half of the land area of the globe (Groisman et al. 2005). These observed changes have been analyzed in relation to changes in intense precipitation for three transient Global Climate Model (GCM) simulations, each with greenhouse gas concentrations increasing during the 20th and 21st centuries and doubling during the later part of the 21st Century. Groisman et al. (2005) found that the empirical evidence from the period of instrumental observations and the model projections of a greenhouse-enriched atmosphere both indicate an increasing probability of intense precipitation events for many extra-tropical regions including the United States (Figs. 2, 5, 6 and 7).

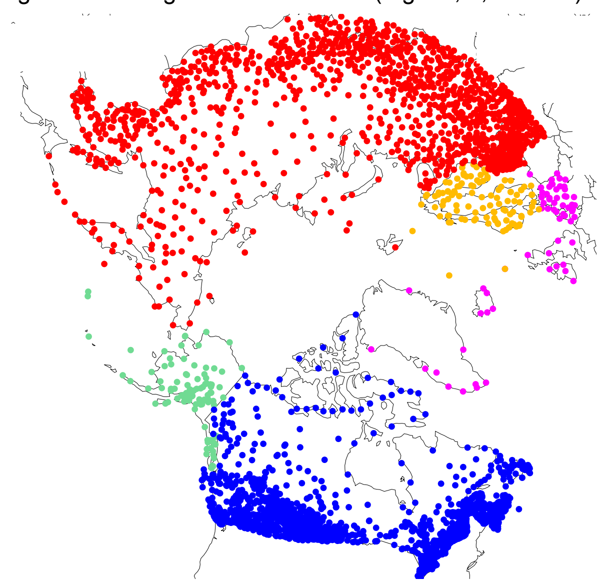

Figure 4. Map of GDCN stations with more than 25 years of valid daily data north of 50°N (plus southeastern Canada and Russian Far East).

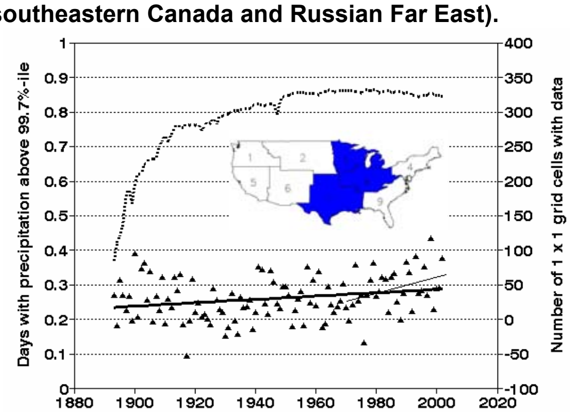

Figure 5. Very heavy precipitation (upper 0.3% of daily rain events with return period of 4 years) over regions of the Central United States (shaded in insert) and their linear trends. Linear trends for the 1893-2002 and 1970-2002 periods (solid lines) are equal to 20%/110 yrs and 26%/30 yrs respectively 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 1ºx1º grid cells with valid station data are shown by dotted line (Groisman et al. 2005).

Under “very heavy” precipitation events, we assume the upper 0.3% of daily rainfall events. This equates to a return period of approximately one daily event in 3 to 5 years for annual precipitation and approximately 10 to 20 years for seasonal precipitation, depending on the probability of daily rain events for a given location. For the contiguous U.S., the mean threshold for such daily events is ~70 mm day^{-1} but for southern and southeastern regions the thresholds used to define these events exceed 100 mm (Groisman et al. 2005). In several extratropical regions with a sufficiently dense meteorological network, we found significant increases in the frequency of very heavy precipitation events. These regions include the Central U.S., the northwestern coast of North America, southern Brazil, Fennoscandia, the East European Plain, South Africa,

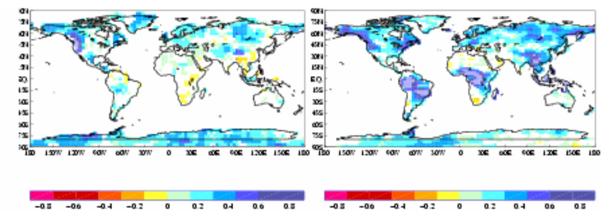

Figure 6. Model simulations of the effect of CO_2 doubling on the average number of exeedences of the 99.7 percentile of precipitation distribution from the Canadian Climate Centre (CCM1, left) and Hadley Climate Model (HAD3, right) models. The red end of the scale depicts decreases and the blue increases. Changes are only shown where they are significant relative to the control climate (Hegerl et al. 2004; Groisman et al. 2005).

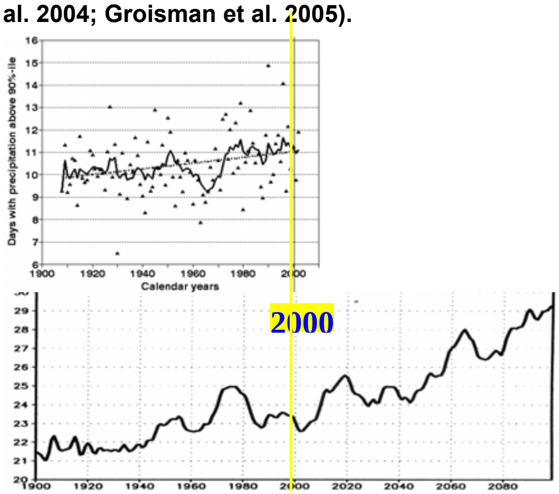

Figure 7. Frequency of the upper 10% of rainy days over the northeastern quadrant of the contiguous United States. Top. Observations (Northeast and Midwest regions) for 1908-2002. Annual values (triangles), 10 year running mean values (solid line), and linear trend (dotted line) are shown. Linear trend for 1908-2002 (12%/100 yrs) is statistically significant at the 0.05 level. Bottom. ECHAM4 global climate model run with transient CO_2 forcing (observed prior to 2000 and project for the 21st century; sector 35N-45N; 75W-85W; adapted from Semenov and Bengtsson 2002) 10 year running mean values for 1900-2090.

southeastern Australia, and Siberia. When century-long time series of intense precipitation were available (e.g., the United States and South Africa), we found that the increases in very heavy precipitation frequency are relatively new phenomena and occur mostly during the past few decades (cf., SWCS 2003).

Both empirical data and model projections for the CO_2-enriched world show that mean precipitation does not change appreciably compared with the highest part of the precipitation distribution. This implies that the frequency of precipitation events has been (will be) affected. Indeed, in several regions of the world this appears to be the case and rainy days are becoming less frequent. For instance, in South Africa, Siberia, the Eastern Mediterranean, Central Mexico, and northern Japan, an increase only in heavy precipitation is observed while total precipitation and/or the frequency of days with an appreciable amount of precipitation (wet days) are not changing and/or are decreasing (Easterling et al. 2000; Alpert et al. 2002; Fauchereau et al. 2003; Groisman et al. 2005).

The first indication that this feature might also be present in the contiguous United States is shown in Fig. 8 for the northeastern part of the country (cf., Sun and Groisman 2004). Observations show that the annual number of wet days has increased during the past 100 years, but during the last 30 years (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

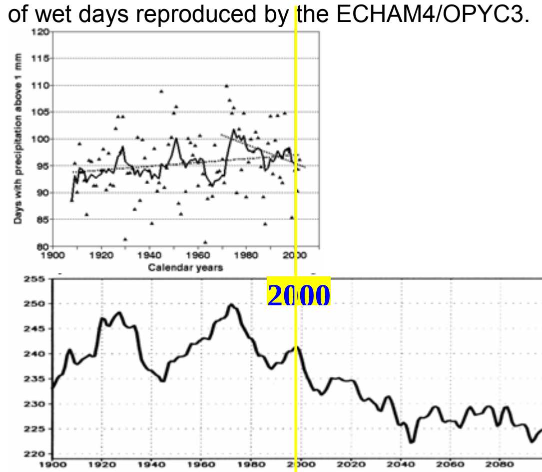

Figure 8. Number of wet days over the northeastern quadrant of the contiguous United States. Top. Observations (Northeast and Midwest regions) for 1908-2002. Annual values (triangles), 10 year running mean values (solid line), and linear trends (dotted line) are shown. Linear trends for 1908-2002 (4%/100 yrs) and for 1972-2002 (-9%/30 yrs) are statistically significant at the 0.05 level. Bottom. ECHAM4 global climate model run with transient CO_2 forcing (observed prior to 2000 and project for the 21st century; sector 35N-45N; 75W-85W; adapted from Semenov and Bengtsson 2002) 10 year running mean values for 1900-2090. Both show a decrease during the past 30 years.

see a similarity of tendencies and a very strong decrease in the wet-day frequency projected for the 21st century by the model. However, 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).

These two patterns may occur in the same regions where an increasing probability of intense precipitation with a decreasing frequency of rainy days makes changes in mean precipitation less notable than changes in precipitation distribution spectra thus also affecting the entire terrestrial hydrological cycle (runoff, soil moisture available for evapotranspiration, etc.).

4. CLIMATE CHANGES IN HIGH LATITUDES THAT MAY INFER DRIER SUMMER CONDITIONS

In the northern regions, observed increases in surface air temperatures (Fig. 1) have resulted in (a) an earlier snow cover retreat (Brown 2000; Groisman et al. 1994, 2003b, 2004), (b) an increase in the frequency of cold season thaws (e.g., Fig. 9), and (c) a significant expansion of the warm season (Easterling 2002; Groisman et al. 2003a; Pagano et al. 2004; Tabs. 1 and

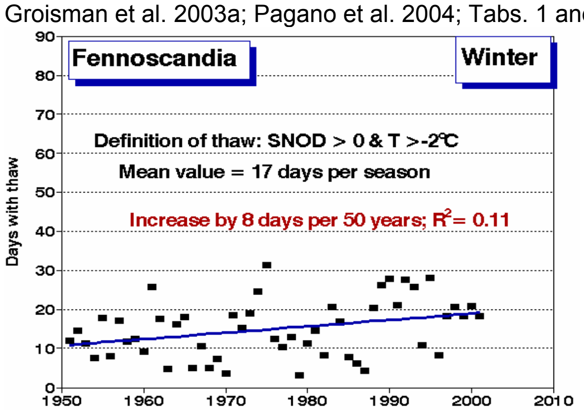

Figure 9. Changes of the winter frequency of days with thaw over Fennoscandia. Thaw-day is defined as a day with mean daily temperature above 2°C and the presence of snow on the ground.

2). The two digit increases in warm season characteristics (marked in **red** in Tabs. 1 and 2) during

Table 1. Statistically significant changes in temperature-derived characteristics over northwestern quadrant of North America [north of 49ºN west of 95ºW] during the past 50 years. Asterisks indicate that in Eastern Canada similar changes have been observed and they are statistically significant at the 0.05 level.

Characteristic	Trend, %/50 yrs Alaska	W. Canada
Heating-degree days	-8	-9
Degree-days below 0ºC	-17	-21
Degree-days above 15ºC	31	17*
Duration of the growing season (T> 10ºC)	19	8*
Frost-free period	6	9*

the past 50 years implies that there should be an additional water "demand" for evaporation and transpiration. It can be compensated by increasing precipitation. But, in high latitudes annual precipitation increased only by ~5% and there are large regions where it has not changed at all (e.g., Siberia; Sun and Groisman 2000; Groisman and Rankova 2001; Zhang et al. 2001; Groisman et al. 2003a). This coupled with changes in precipitation distribution spectra and an increase in Cb cloudiness (documented for the former USSR and the United States by Sun et al. 2001) has resulted in the possibility for a "summer dryness" scenario in several large interior regions of the continents (cf., recent forest fires and droughts in high latitudes such as Siberia, Central Asia, and several regions of the western part of North America). Specifically, in high latitudes we have:

Table 2. Statistically significant changes in temperature-derived over Northern Eurasia during the past 50 years (east of 30ºE, north of 50ºN).

Characteristic	Trend, %/50 yrs East of 80°E	West of 80°E
Heating-degree days	-6	-7
Degree-days below 0ºC	-12	-19
Degree-days above 15ºC	12	-
Duration of the growing season (T> 10ºC)	8	8
Frost-free period	10	-

- Up to a two-digit (%) increase in temperature-derived characteristics => evapotranspiration may increase;
- Earlier snowmelt and more frequent thaws => more cold season precipitation become unavailable in the warm season;
- Only a moderate increase in precipitation but an increase in thunderstorm activity => more warm season precipitation contributed to increased runoff;
- All the above => a possibility of drier summer conditions => an increase in forest fire danger.

These conditions were projected by some GCMs for the greenhouse gases enriched atmosphere and suggest reduced future water availability in high and mid-latitudes over the land (Folland and Karl 2001). The next Section describes our assessment of potential forest fire indices (that can be also considered as drought indices) in detail.

5. INDICES OF POTENTIAL FOREST FIRE DANGER

Each summer, forest and tundra in high latitudes suffer from numerous naturally caused fires that are difficult to fight due to their remote locations. Among meteorological variables that affect the potential fire danger are surface air temperature, soil moisture, humidity deficit, probability of lightning, and atmospheric stability (Condra 1944; Turner and Lawson 1978; Keetch and Byram 1968; Karl 1986). To characterize the level of potential fire danger, numerous indices have been suggested. Figure 11 shows one such index (Nesterov Index) based on the humidity deficit, temperature, and information about the days with precipitation below 3 mm day^{-1} (Sherstyukov 2002). Specifically, this index of fire danger, G, is calculated according to the formula:

$$G = \Sigma\ (T\ d), \qquad (1)$$

where T is the mid-day temperature, d is the dew point deficit ($d = T - T_d$, where T_d is the dew point), and totals are calculated for positive temperatures for a sequence of days with precipitation less than 3 mm. Rainfall above 3 mm resets G to zero. Time series with the number of days with thusly defined G above two different thresholds are shown in Figure 10 for the Moscow region, Russia. Days with G below 300 are the days without substantial forest fire danger. Days with G above 1000 are characterized as days with high forest fire danger. The figure shows an increase in potential forest fire danger in the region during the past 40 years. Index G and several other indices require sub-daily meteorological information which is not easy to access on a century time scale. Therefore, we used a simplified index suggested by Keetch and Byram in the late 1960s but for Canada and Russia tested two other indices based on sub-daily synoptic information: modified Nesterov and Zhdanko indices.

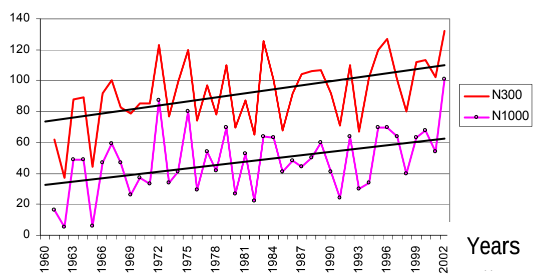

Figure 10. Interannual variability of the number of days with medium and above potential forest danger (G > 300) and high forest fire danger (G >1000) for the Moscow region (Sherstyukov 2002).

5.1. Keetch-Byram Drought Index

The Keetch-Byram Drought Index (KBDI; Keetch and Byram 1968) is a measure of meteorological conditions conducive to forest fires. It uses only daily maximum temperature and precipitation information and estimates soil moisture deficiency on a scale ranging from 0 to 800. Zero is the point of no moisture deficiency and corresponds to no danger of forest fire and 800 is the most severe drought that is possible and corresponds to extreme forest fire danger. The logic behind the index is that wet soil suppresses wild fires while dry soil organic matter enhances these fires and makes them difficult to control. Certainly, factors in addition to soil moisture influence the occurrence and behavior of forest fires. However, in the USA, experience over the years has established the close association of extremely difficult fire suppression with cumulative dryness, or drought. For these reasons, we have chosen the KBDI to study changes in forest fire potential. Below is a short summary of the algorithm of the KBDI evaluation:

Begin calculations at the point when soil is close to saturation (e.g., after snowmelt) and KBDI ≈ 0. Then

- Increase KBDI by an amount of evapotranspiration, E (a function of the current KBDI, T_{max} values, and mean annual P)
- Reduce KBDI by the amount of "net" rain, if any. In the sequence of humid days with non-zero precipitation, the "net" rain is the actual rainfall on the second day and days thereafter, but is reduced by 5 mm (0.2 inch) during the first day of the sequence.

In Fig.11, the KBDI index is presented for North American and North Eurasian stations with daily temperature and precipitation available for the past fifty years. This is a climatology of the index for the month of July (the long-term mean values). Note the difference in the ranges of the index between high and mid-latitudinal stations.

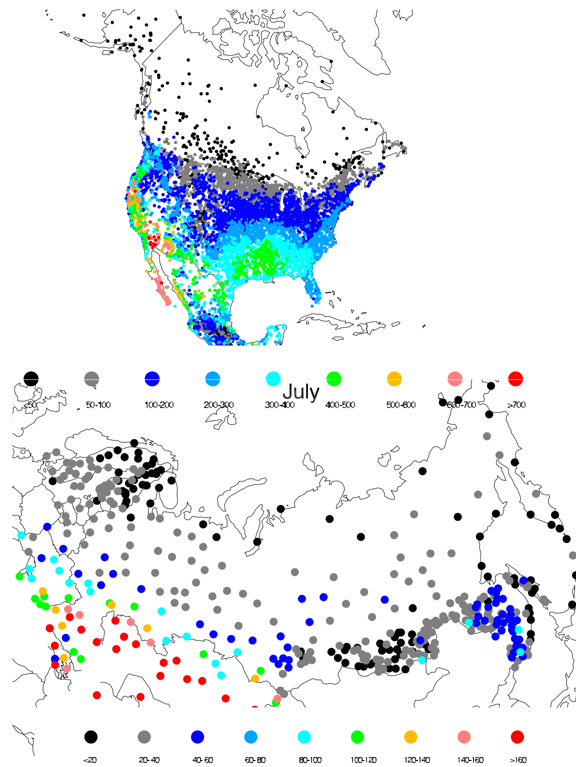

Figure 11. Climatology of July KBDI values for North America and Northern Eurasia. Note a different scale ranges used in these two plots.

The only regional adjustment in the KBDI computation depends on the long-term annual precipitation, P, that is assumed to be proportional to the total regional vegetation density (that adjusts itself to P) and thus controls E. Surface radiation balance is not used for this control which implies that the KBDI should be regionalized. The reason being that the actual vegetation density is a function of (at least) surface radiation balance (that can be associated with potential evapotranspiration) and precipitation (Fig. 12). In regionalizing the KBDI, we did not change the calculation scheme but instead, we abandoned a direct relationship between the index and soil moisture in the upper soil and duff layers. The authors of the index wrote: "For forest fire control, a useful concept of drought is one which treats it as a continuous quantity which can be described in numerical terms. The values would range from zero (soil and duff saturated with water) up to some maximum value which corresponds to an absence of available moisture in the soil and duff. The upper part of the scale does correspond to those conditions for which many definitions of drought require that the dryness or moisture deficiency be abnormal or unusual" (Keetch and Byram 1968). Therefore, we fully

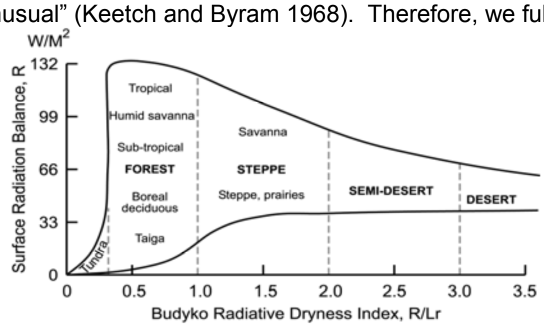

Figure 12. Geobotanical zones as a function of the surface radiation balance, R, and Budyko radiative dryness index (R/Lr). r is annual precipitation and L is specific latent heat of vaporization (Budyko 1971).

employ this peculiarity of KBDI by using the upper frequency part of this scale in our analyses to select the days with abnormally dry soil conditions, i.e., the days conducive to forest fire. Specifically, at each point for the entire period of record, we calculated KBDI and then estimated its distribution for the part of the year (season, month) when KBDI was above zero. We estimated its distribution for the reference period 1961-1990 (when a majority of stations have the data available). Upper percentiles of this distribution were defined at each station (upper 10th percentile, upper 1st percentile, etc). Then we calculated the number of days above these percentiles. Keep in mind that whatever the absolute values of KBDI actually are or whether forest fires do or do not actually occur in the region, we assume these days as potentially dangerous for forest fires (i.e., conducive for fire). The count of these days was performed for each season and year and area averaged** to present the time series of "regionally-averaged" conditions that are conducive to forest fire. In this approach, however, locations in swampy areas that would never be able to burn (at present climate conditions) are included in the analyses and participate in the area-averaging process as well as locations with dry soils that have a great chance to catch fire from natural and anthropogenic causes. For regionally-averaged characteristics based on all forest fire indices, we used normalization described above. Note that initially Groisman et al. (2004) used fixed KBDI thresholds for the contiguous U.S. (e.g., 200, 300, and 400). This was done only for narrowly defined small regions where we can expect the KBDI variations to be relatively small. The processing described above (with percentiles) is more universal and flexible.

5.2. Zhdanko and modified Nesterov indices.

Zhdanko (1965) suggested a recurrent index of potential forest fire danger during the warm snowfree period of the year similar to G (Eq. 4):

$$Zh(N) = (Zh(N-1) + d) \times K(N), \qquad (2)$$

where d is the dew point deficit and K(N) is a scale coefficient in a [0,1] that controls the index change when precipitation occurs on day N. This reduction factor is equal to 1 when no rainfall occurs, is equal to 0 when daily rainfall is above 20 mm, and gradually decreases between these thresholds (e.g., it is equal to 0.4 when daily rainfall is in the range of 3 to 5 mm). Thus, the differences between Eqs. 1 and 2 arise from a steeper temperature dependence used in G and the introduction of a more relaxed influence of rainfall on the forest fire index in Eq. 2 (Molchanov index assumes K(N)=0 for rainfall as low as 3 mm without accounting for the level of dry conditions prior to this rain event). In our further analyses we used a modified Nesterov Index (MNI) by introducing to its values a K(N) reduction factor similar to that used by Zhdanko in Eq. 2. A comparison of remaining differences between both indices (Zhdanko and MNI) shows that their application delivers similar results (Fig.13).

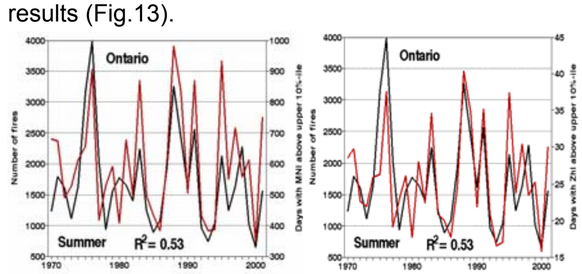

Figure 13. Comparison of Modified Nesterov (left) and Zhdanko Indices with the count of forest fires over Ontario Province in Canada. Both indices were normalized (i.e., number of days with indices above the upper 10th percentile were area-averaged over the province).

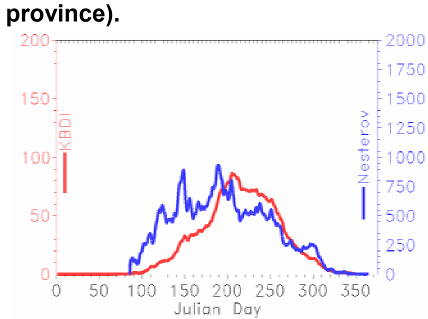

Figure 14. Seasonal distribution of KBDI and MNI area-averaged over the Ontario province.

** Area-averaging was performed as follows: First, station data anomalies from the long-term mean for the reference 1961-1990 period (or statistics) were averaged within each 1° x 1° gridcells. The gridcell values (those with at least 1 valid station value within) were then averaged further over the region with weights proportional to the gridcell latitude cosine. This approach does not claim to cover the entire region but focuses on the data-elucidated regions where people live and maintain meteorological observations for sufficiently long periods of time.

MNI and Zhdanko indices have climatology that differs noticeably from that of KBDI. For example, Fig. 14 shows that the atmospheric dryness (better described by the dew point deficit accumulated values) occurs earlier than dryness associated with soil and duff moisture deficit associated with KBDI. However, when normalized and area-averaged, both types of indices deliver similar conclusions about changes in potential forest fire danger. Whatever trends were in place, an increase of wet conditions (as over the Great Russian Plain; Fig. 15) or the progress of dry conditions (as over East Siberia; Fig. 16), normalized time series of various "fire" indices are reasonably well correlated and support conclusions about the regional tendencies of change in potential forest fire danger based on each of them.

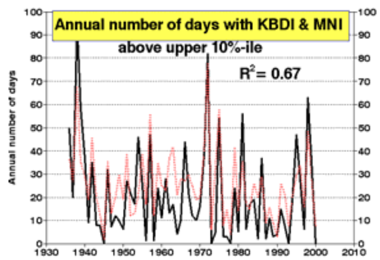

Figure 15. Annual number of days with KBDI and MNI above the upper 10th percentile of non-zero daily values area-averaged over the European part of the former USSR south of 60°N.

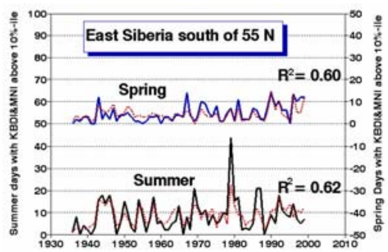

Figure 16. Spring and summer number of days with KBDI and MNI above the upper 10th percentile of non-zero daily values area-averaged over southern part of East Siberia (region around the Baikal Lake).

6. TIME SERIES OF DAYS WITH KBDI ABOVE SELECTED THRESHOLDS

6.1. Alaska south of the Polar Circle

In Alaska, we first compared the area-averaged number of days with summer KBDI above the 90th percentile with an observed number and area of forest fires around the state (Fig. 17). Correlation of the index with counts of fires improves with time. The peak of the index in 1957 corresponds to a widespread fire year (by area but not by counts; not shown). When analyzing Alaskan data, we separated the very humid southernmost part of the state south of 62°N. Thus, the following figure presents time series of the frequency of high KBDI values in Central and Southern Alaska separately. Since 1944, in the center of the state, and since 1937, in the south, we have a a sufficient number of stations to provide the area-averaging of the fire indices. After the fire outbreak in 1957/1958, a steady increase in frequency of spring and summer days with high KBDI has occurred. Fig. 18 does not include year 2004 when the fire area and fire counts were close to the infamously remembered year 1957.

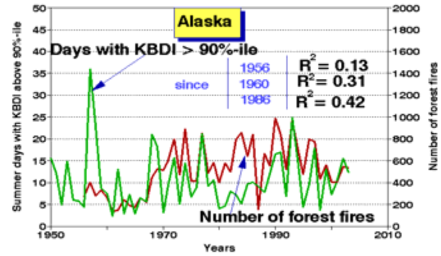

Figure 17. Frequency of summer days with KBDI above 90th percentile in Alaska and number of forest fires (the 1950-2003 period).

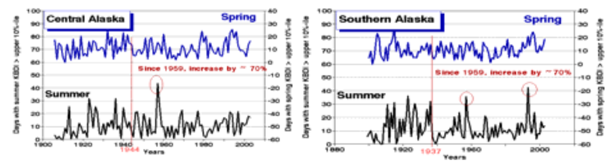

Figure 18. Frequency of high KBDI values in Central (between 62°N and 66.7°N) and Southern Alaska in spring (MAM) and summer (JJA) seasons.

6.2. Southern Canada

While assessing the trends in potential forest fire danger in Southern Canada (south of 55°N), Groisman et al. (2003b) did not find statistically significant trends (e.g., Fig. 13) but revealed a very high interannual variability in regional KBDI estimates. Keeping in mind significant temperature changes (Fig. 1; Tab. 2), forest regions of western Canada (especially, its northern part) should be prone to an increasing forest fire danger. We, however, did not see this in our data due to the sparse network and short time series available at our disposal (Fig. 11). Studies of changes of potential forest fire danger in the northwestern part of the country should be continued.

6.3. Siberia south of the polar circle

Siberia is a region that is prone to forest fires (by some estimates: ~10^7 ha/yr are burned) but the KBDI is always below 200 (Fig. 11). Most of Siberia is in the permafrost zone and, thus, has a shallow active soil layer. Low precipitation and rough terrain of Eastern Siberia (that assists runoff) leave soil above the permafrost relatively dry. Groisman et al. (2003b) shows that over the entire Siberia study area, the frequency of summer days with KBDI above the "non-

Figure 19. Five regions in Siberia where potential forest fire danger in spring and summer has increased in the 20th century. These increases are statistically significant at the 0.05 or above levels.

zero" 90th percentile has increased during the past century in the range of 70% to 115% (Figs. 19 and 20; Tab.3). It is worth noting that an earlier spring onset accompanied by a significant warming and an earlier snow cover retreat (Brown 2000; Groisman et al. 1994, 2003b) also cause an increasing KBDI in the spring season, especially in southern parts of Siberia and the Russian Far East during the past two decades (Fig. 20).

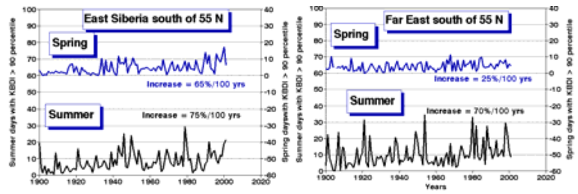

Figure 20. Summer and spring days with daily KBDI values above the upper 10th percentile of its seasonal distribution over southern part of East Siberia and Russian Far East (regions 9 and 11 in Fig. 19).

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 20th 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, 2003b), and thus to an earlier onset of spring- and summer-like 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 20th century (Milly et al. 2002). *Both model projections of a greenhouse-enriched atmosphere and the empirical evidence from the period of instrumental observations indicate an increasing probability of heavy precipitation events for many extratropical regions including the United States.*

During the past century, global temperature and particularly temperature in high latitudes has increased dramatically. This is associated with a retreat in spring snow cover, changes in the length of the growing season period and, therefore, an increase in the precipitation needed to be spent on evapotranspiration (if available). In high latitudes, most of the evapotranspiration occurs in a narrow interval of early summer. For example, according to Rauner (1972), up to half of the evapotranspiration in the forest-steppe zone of the European part of Russia and Western Siberia occurs in June-July. If the precipitation increase (which is also occurring in high latitudes during the past 50 years) does not match the "needs" of evapotranspiration, the summer dryness will increase and potential forest fire danger will also increase. It appears that this is the case for Alaska and Siberia south of the Arctic Circle. In Siberia, we do not observe a substantial increase in summer precipitation during the past 50 years (Gruza et al. 1999; Sun and Groisman 2000). Furthermore, a redistribution of precipitation (increase of heavy rainfall and decrease in days with precipitation) was found over the Asian part of Russia (Sun and Groisman 2000). Heavy rainfall (which is increasing in frequency in Siberia) usually comes in thunderstorms from *Cumulonimbus* clouds (Sun et al. 2001). The accompanying lightning produces an additional (and independent) factor that acts to increase the forest fire danger. This explains why, using meteorological information for the past century, *we found a significant (sometimes a twofold) increase in indices that characterize the weather conditions conducive to forest fires. The areas where this increase was statistically significant coincide with the areas of most significant warming during the past several decades in Alaska and Siberia south of the Arctic Circle.*

Table 3. Increase in mean regional frequency when summer KBDI exceeds the "non-zero" 90th percentile of its distribution during the reference period 1961-1990. All linear trend estimates are statistically significant at the 0.05 or above levels. Asterisk (*) marks a different period, for which trend has been estimated.

Regions south of 66.7° N	Period assessed	Frequency changes, %/100yr
West Siberia, north of 55° N	1900-2001	115
East Siberia, north of 55° N	1900-2001	85
Far East, north of 55° N	1936-2001	75*
East Siberia, south of 55° N	1900-2001	75
Far East, south of 55° N	1900-2001	70

8. APPENDIX. REPRESENTATIVENESS OF REGIONAL ESTIMATES OF VERY HEAVY PRECIPITATION FREQUENCY

Reliable assertions of trends in some of the characteristics of extremes (e.g., very heavy and extreme precipitation changes) are possible only for those regions with dense networks due to the restricted radius of correlation for these characteristics. Therefore, an assessment of the representativeness of results based on sparse networks has been a part of the analyses and is presented below.

8.1. Theoretical framework

Each estimate of the area-averaged anomaly 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

$$E^2 = \sigma_s^2 - 2\ \Sigma\ w_i\Omega_i + \Sigma\Sigma w_i w_j R_{ij} + \Sigma w_i^2 \delta_i^2 \quad , \qquad (3)$$

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 footnote **.

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

$$r(\rho) = C_0 \exp(-\rho/\rho_0), \qquad (4)$$

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 $(1\text{-}C_0)$ 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 "very heavy" precipitation over most of the regions with sufficiently dense network in Fig.3, 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 (here we use the area name, S) can be described by

$$Z_S = C_v \{[(1\text{-}C_0)/ C_0 + 0.23(S/N)^{1/2} \rho_0^{-1}]/N\}^{1/2}, \qquad (5)$$

where the spatial correlation function is approximated by Eq. 4, 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 our area-averaging procedure, we first estimated the representativeness of grid cell area-averaging. This step provided us estimates of the accuracy of the 1° x 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.

8.2. Results

The application of Eq. 5 shows that for the annual frequency of "very heavy" precipitation events for a "typical" 1° x 1° grid cell on fairy level terrain (e.g., the Midwestern United States, European Russia, or Australia with a ρ_0 of ~ 30 km and C_v of ~ 0.3), with 3, 2, or 1 stations we cannot reduce the error below 10%, 15%, and 25%. We consider the Z_S-estimates to be on the 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. 4 is precise, etc.) in reality do not materialize and/or are only convenient approximations. A similar assessment in mountainous 1°x1° grid cells (e.g., the southwestern or northwestern United States, Caucasus, or Mexico with large micrometeorological variability and 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. 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 corresponds with steep precipitation gradients. But, in the mountains, assumptions of implementation of Eq. 4

Table 4. Parameters of statistical structure of the fields of regional frequency of heavy (above the upper 10th percentile) and very heavy (above the upper 0.3th percentile) seasonal precipitation events over the contiguous U.S. Parameters of the spatial correlation function in Eq. 4 (ρ_0 and C_0) and variance coefficient, C_v, of the 1-degree-grid-cell-averaged values of the frequencies are presented.

Region	Season	Events	ρ_0, km	C_0	C_v
North-west	Winter Annual	Heavy	505 325	0.85 0.85	0.40 0.20
	Winter Annual	Very heavy	250 160	0.55 0.65	0.85 0.45
South-west	Winter Annual	Heavy	300 255	0.90 0.90	0.50 0.20
	Winter Annual	Very heavy	125 95	0.70 0.65	0.75 0.30
Mid-west	Summer Annual	Heavy	190 300	1.00 1.00	0.20 0.15
	Summer Annual	Very heavy	95 110	0.65 1.00	0.40 0.30
South-east	Summer Annual	Heavy	270 420	0.75 0.85	0.15 0.15
	Summer Annual	Very heavy	155 155	0.55 0.85	0.45 0.40

(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 the "low side".

Table 4 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 10th and 0.3th percentiles respectively) for several regions of the contiguous United States. 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° x 1° grid cells, with nearly complete grid cell data coverage during the entire 20th century Fig. 21), we obtained Z_S less than 2% throughout the 20th century for the area-averaged annual VH frequency. In this region, Z_S-values remain less than 3% even in the last decade of the 19th century.

8.3. Representativeness problem

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 meters 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 (a) we may still capture these events and (b) 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 1 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.4, 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. 5) 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 of 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 a 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. Fig. 21 provides counts of stations with precipitation data within each1° x 1° grid cell for periods

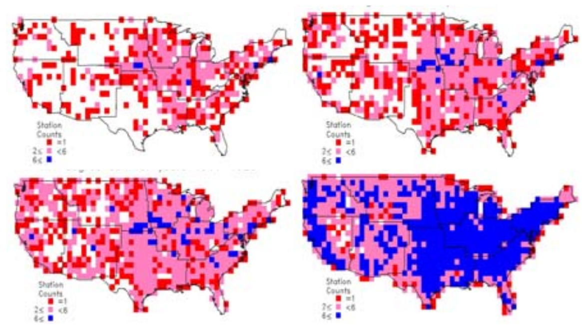

Figure 21. Data availability over the contiguous U.S. generalized within the 1º x 1º grid cells for the periods of a near-constant network. Counts of stations with more than 50% of daily precipitation data within each grid cell for the 1891-1900, 1901-1910, 1911-1920, & 1951-2002 periods respectively. Note a poor data coverage for the last decade of the 19th century over the western half of the country.

with relatively stable networks for the contiguous U.S. A station was considered present during the listed period when it had at least 50% of daily precipitation data within it. Fig. 21 shows that data availability is restricted in the nationwide analyses of the very heavy precipitation prior to the first decade of the 20th century. This, however, is mostly due to a data deficiency in the western part of the country and, therefore, allows analyses of the 110 year-long time series for the Central U.S. presented in Fig. 5.

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