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CHANGES IN THE PROBABILITY OF HEAVY PRECIPITATION: IMPORTANT INDICATORS OF CLIMATIC CHANGE

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Abstract. A simple statistical model of daily precipitation based on the gamma distribution is applied to summer (JJA in Northern Hemisphere, DJF in Southern Hemisphere) data from eight countries: Canada, the United States, Mexico, the former Soviet Union, China, Australia, Norway, and Poland. These constitute more than 40% of the global land mass, and more than 80% of the extratropical land area. It is shown that the shape parameter of this distribution remains relatively stable, while the scale parameter is most variable spatially and temporally. This implies that the changes in mean monthly precipitation totals tend to have the most influence on the heavy precipitation rates in these countries. Observations show that in each country under consideration (except China), mean summer precipitation has increased by at least 5% in the past century. In the USA, Norway, and Australia the frequency of summer precipitation events has also increased, but there is little evidence of such increases in any of the countries considered during the past fifty years. A scenario is considered, whereby mean summer precipitation increases by 5% with no change in the number of days with precipitation or the shape parameter. When applied in the statistical model, the probability of daily precipitation exceeding 25.4 mm (1 inch) in northern countries (Canada, Norway, Russia, and Poland) or 50.8 mm (2 inches) in mid-latitude countries (the USA, Mexico, China, and Australia) increases by about 20% (nearly four times the increase in mean). The contribution of heavy rains (above these thresholds) to the total 5% increase of precipitation is disproportionally high (up to 50%), while heavy rain usually constitutes a significantly smaller fraction of the precipitation events and totals in extratropical regions (but up to 40% in the tropics, e.g., in southern Mexico). Scenarios with moderate changes in the number of days with precipitation coupled with changes in the scale parameter were also investigated and found to produce smaller increases in heavy rainfall but still support the above conclusions. These scenarios give changes in heavy rainfall which are comparable to those observed and are consistent with the greenhouse-gas-induced increases in heavy precipitation simulated by some climate models for the next century. In regions with adequate data coverage such as the eastern two-thirds of contiguous United States, Norway, eastern Australia, and the European part of the former USSR, the statistical model helps to explain the disproportionate high changes in heavy precipitation which have been observed.

Keywords: heavy precipitation changes, North America, Eurasia, Australia

1. Introduction

Analyses of trends in mean precipitation during the past century reveal compelling evidence of the presence of trends over many regions of the world (Groisman and Legates, 1995; IPCC, 1996, 1998). In many countries (e.g., Russia, Norway, Sweden, Canada) the increase in precipitation was more pronounced in the cold season (about 10-15 %/100 yrs) than in the warm season (about 5%/100 yrs), but the absolute values of these changes are comparable because of the seasonal cycle of precipitation in most of the northern extratropics. In Poland and Australia, the century-long increase in precipitation was predominantly in the warm half of the year (Kozuchowski, 1985; Suppiah and Hennessy, 1998). In most parts of Norway the annual precipitation has increased by 8-14%/100yrs (Hanssen-Bauer et al., 1997), while the summer precipitation has increased less prominently (by 5-10%) and mostly in northern part of the country (Hanssen-Bauer, 1994).

Of particular interest, from both practical and theoretical considerations, are the analyses of precipitation change that reveal increases in extreme and very heavy precipitation from North America, Australia, and Japan. Karl et al. (1995) and Karl and Knight (1998) provide evidence for a statistically significant increase in extreme precipitation (greater than 50 mm per day) precipitation in the United States. Similarly for Australia, Suppiah and Hennessy (1996, 1998) show significant increases for the higher percentiles, e.g., the 90th and 95th percentiles. This was augmented by an increase in heavy-rain days in eastern Australia associated with East Coast cyclones reported by Hopkins and Holland (1997). Iwashima and Yamamoto (1993) analyzed daily precipitation data from 1890 to 1980 at 55 Japanese stations and found that more stations recorded their highest, 2nd highest or 3rd highest precipitation event in more recent decades. Thus, the frequency of years with extremely heavy daily precipitation is increasing at Japanese stations throughout the 20th century. Analysis of a small subset of 14 U.S. stations performed by Iwashima and Yamamoto (1993) suggests that this increase has occurred over the contiguous United States too. Tsonis (1996) shows that the variability of monthly precipitation totals over the United States, Europe, and Australia has also increased during the past 100 years. Beniston et al. (1994) concluded that “in a warmer global climate, precipitation in Alps would be generally reduced but the extreme precipitation events could be expected to increase significantly”. This empirical conclusion was supported later by the modeling assessment of Schaer et al. (1996). Generally, climate model simulations consistently project increases in global precipitation due to global warming stemming from increases in greenhouse gases, particularly for the mid and high latitudes (IPCC, 1990, 1996). An increase in heavy precipitation is also simulated by climate models (IPCC, 1996; Schaer et al., 1996; Jones et al., 1997; Hennessy et al., 1997).

We are interested in heavy precipitation during the three warmest (and often wettest) summer months, which coincide with the period of the primary growing season. In this paper heavy precipitation changes during summer are assessed in eight countries: Canada, the United States, Mexico, the former Soviet Union, China, Australia, Poland,

and Norway. We show that if the shape of the precipitation distribution (often well described by the gamma distribution) does not change as total precipitation increases, a disproportionate increase in heavy precipitation is expected.

2. Data Used

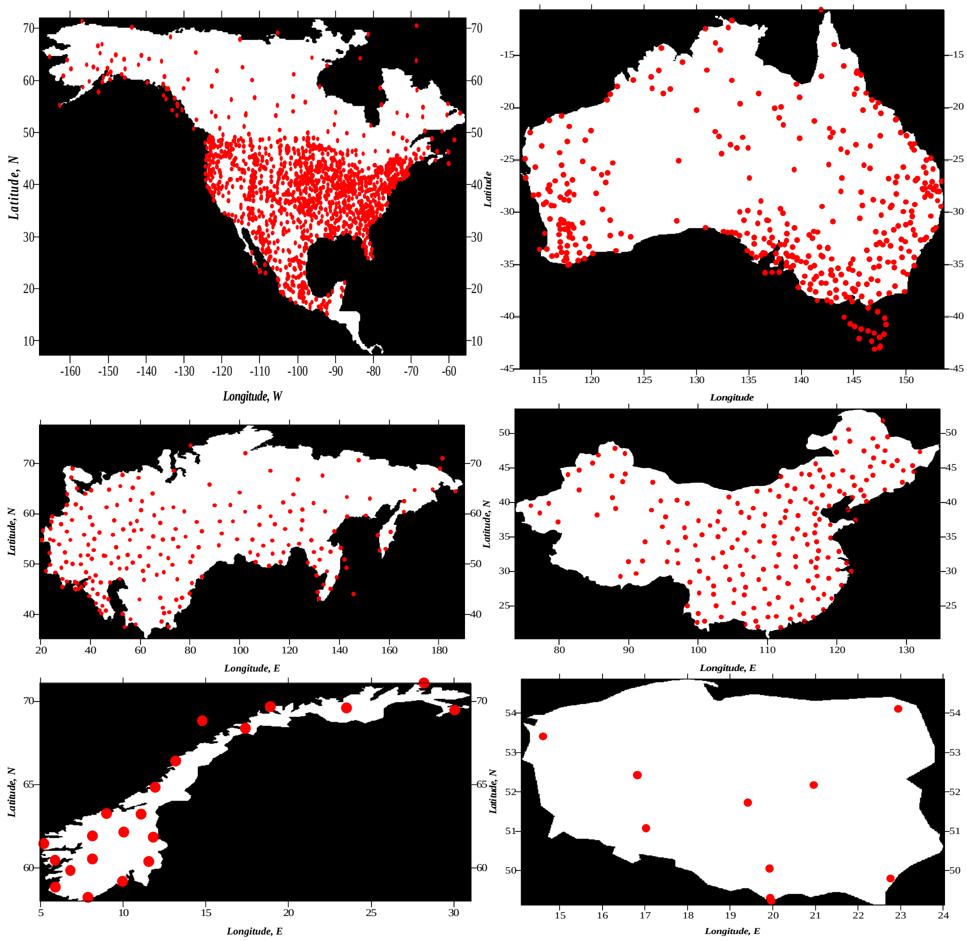

Figure 1. Maps of the stations with daily precipitation time series used in this study for North America (Canada, the United States, and Mexico), Australia, the former Soviet Union, People Republic of China (PRC), Norway, and Poland. Only the continental part of all these countries is shown. Several stations from adjacent islands were also used in the analyses. Note the different spatial scales in each map.

Daily precipitation data sets for eight countries were used in our analyses. For the former Soviet Union we used an archive of 223 stations of the international exchange available from the Carbon Dioxide Information Analysis Center (Razuvaev et al., 1993, updated) from the beginning of observations to 1994. For North America we employed a new daily precipitation data set accumulated at the National Climatic Data Center (Easterling, 1997; Easterling et al., 1998). Daily data from Canada (93 stations) and Mexico (202 stations) spanned the years 1900-1995 and 1950-1990 respectively. A subset of the highest quality stations from the U.S. Historical Climatology Network (HCN) of 134 stations with century-long daily precipitation time series (Hughes et al., 1992) comprised the data base for the contiguous United States, supplemented by an additional 53 stations to provide more representative spatial coverage. The U.S. time series, now updated through 1996, were previously used by Karl et al. (1995) and Karl and Knight (1998) in the analyses of extreme precipitation over the contiguous United States. Additionally, for mapping of precipitation distribution parameters only, we used 1060 HCN stations from the contiguous U.S. and 44 Alaskan stations spanning the years 1948-1995. Data for 198 Chinese stations of international exchange span the period from 1951 to 1994 (Baker et al., 1995). An extended high-quality historical precipitation data set for Australia comprises 379 stations from the beginning of observations (113 start as early as 1891) up to 1996 (Lavery et al., 1997). A subset of 13 century-long homogeneous daily precipitation time series was used for estimates of the precipitation extremes over Norway. Somewhat shorter homogeneous daily precipitation time series (40 to 60 years of data) from another 8 Norwegian stations were used mostly for mapping of precipitation distribution parameters. Data from ten first order stations well distributed over Poland (except the north-eastern) were available for the post-World War II period. The station networks are shown in Figure 1 and their pre-processing is described in Appendix 1.

3. Model of the Daily Precipitation Distribution

It is widely recognized that the distribution of daily precipitation totals, P, can be approximated by the gamma-distribution $\Gamma(\eta, \lambda)$ (Thom, 1951, 1958; Bagrov, 1965; Mooley, 1973; Crutcher et al., 1977, Buishand, 1978; Guttman et al., 1993) with the density function

$$p(\eta, \lambda, x) = \mathrm{const}(\eta, \lambda) * x^{\eta-1} * \exp(-\lambda x), \qquad (1)$$

when $x>0$, and zero when $x \leq 0$. For this family of distributions the η-parameter defines the shape of the distribution, while the λ-parameter characterizes the scale. The mean, μ, variance, σ^2, and the coefficient of variation C_v of this distribution are defined by these parameters:

$$\mu = \eta / \lambda; \qquad \sigma^2 = \eta / \lambda^2; \qquad C_v = \sigma/\mu, = 1/\mathrm{sqrt}(\eta) \qquad (2)$$

Note that C_v is only a function of the shape parameter.

Since it does not rain every day, a mixed distribution model is considered for daily precipitation totals. Under this model, it is assumed that the occurrence of daily precipitation events has a binary distribution with the probability of a single event P_{pr} and the distribution function of precipitation totals F(x) is expressed as:

$$F(x) = P(X \leq x) = (1-P_{pr}) + P_{pr} \int_0^x p(\eta, \lambda, t)\, dt \qquad (3).$$

The precipitation amount during this event is considered to have a gamma-distribution. For (3), we have three parameters: P_{pr}, η , and λ, where the density function (1) now characterizes a conditional distribution of daily precipitation. For this model, Eqs. 2 will be transformed into:

$$\mu = P_{pr}\eta/\lambda; \ \sigma^2 = (P_{pr})^2\, \eta\, /\lambda^2 \ ; \ C_v = \sigma/\mu, =1/\mathrm{sqrt}(\eta) \qquad (2').$$

We shall use this model throughout and apply it to daily precipitation totals over Eurasia, Australia, and America with the following simplification: a precipitation event is defined as a non-zero 24-hour total. Analysis of the weather duration tables from the United States primary meteorological network shows that this is not exactly the case. In the summer time, rainy days are composed from two rain events on average divided by a short no-rain period. But we do not have weather duration tables for most of the network data we are using for most of the period under consideration, i.e., we have no better choice.

Another simplification is that we are assuming independent daily precipitation events. In fact, the probability of having a summer day with precipitation after a rainy day is higher than after a day without precipitation and, similarly, the probability of a day without precipitation after a dry day is higher than after a wet day (Katz, 1977, Richardson, 1981). No efforts were made throughout this paper to address temporal correlation of precipitation behavior: grouping of dry and wet days into spells. This could adversely affect our assessment of the probability of heavy rains in the framework of the model (3), although the theoretical analysis by Katz (1998) indicates that temporal correlation is not crucial for estimates of the probability of extreme precipitation events. Therefore, we specifically tested the goodness of fit of model (3) for estimates of the probability of heavy rains. In the regions with a dense network of long-term homogeneous precipitation observations with a sufficient amount of precipitation events (Eastern United States, Eastern Australia, European Russia, Southern Norway) we calculated empirical estimates of the probability of "heavy" precipitation (i.e., above a given threshold) and compared them with calculations based on model (3). Figure 2 presents an example of such a comparison for the contiguous United States. It shows that, for the threshold under consideration (here, 50.8 mm), model (3) reasonably well reproduces the pattern of probability but sometimes underestimates its absolute values. We cannot test empirically the goodness of fit of our model in the regions with short or few homogeneous precipitation time series, "dry" regions, and for the probability to exceed higher thresholds (e.g. 150 mm) due to the lack of sufficient heavy rain events.

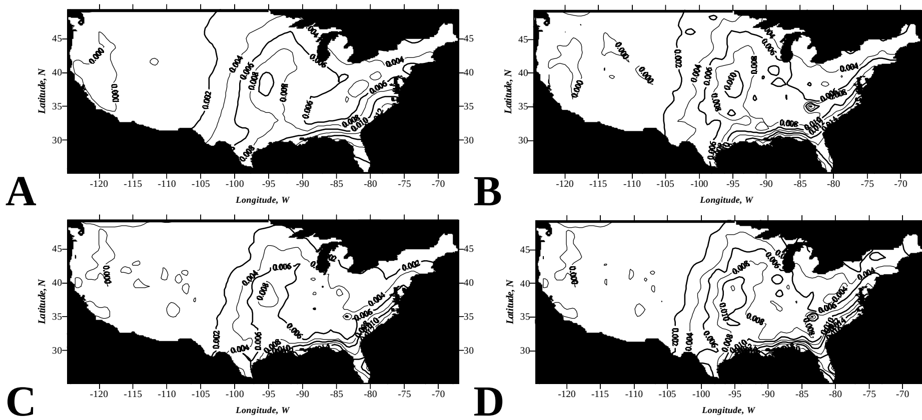

Figure 2. Probability of a summer day with precipitation exceeding 50.8 mm over the contiguous United States: (A) direct empirical estimates using the century-long homogeneous time series of 172 stations (Hughes et al., 1992, updated; Karl and Knight, 1998); (B) the same as (A) but using the time series of 1088 HCN stations for period 1950-1995 ; (C) calculations based on model (3) with parameters η, λ, and P_{pr} estimated from the data of 1088 HCN stations for period 1950-1995; and (D) the same as (C) but with the λ-parameter reduced by a factor of 1.05 to allow a 5% increase in mean daily precipitation.

The mean precipitation in model (3) is a product:

$$\mu = P_{pr}\eta/\lambda \qquad (4)$$

and its change can be a result of the contribution of all three parameters. We are interested in the changes in the probability of heavy rains that can accompany changes in mean precipitation. Therefore, we tested the sensitivity of this probability to changes in μ that are introduced by the variation of each of these three parameters. In the regions with mean daily summer precipitation above 1 mm day^{-1} for typical combinations of η, λ, and P_{pr}, the change in the probability of exceeding heavy precipitation thresholds with a change in μ was analyzed. The strongest changes in heavy precipitation probability occur when the changes in μ are associated with variation of scale parameter, λ, and the smallest changes occur when the changes in μ are associated with variations of P_{pr}. For example, the probability of exceeding a 50.8 mm day^{-1} threshold over the eastern two-thirds of

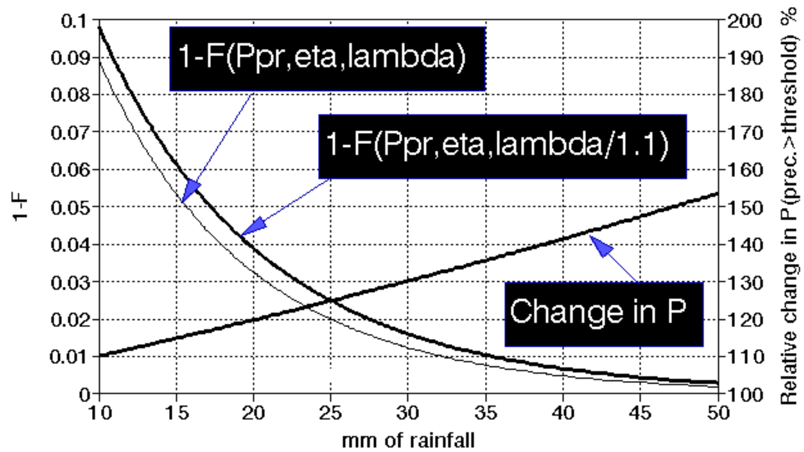

Figure 3 A. Tail of the distribution function [1-F(x, P_{pr}, η , λ)] that mimics July precipitation in Toronto, Ontario (P_{pr} =0.3; η = 0.76; λ= 0.09 mm^{-1}; μ = 2.5 mm day^{-1}) and the same function for a 10% increase in mean value, μ, of July precipitation assuming that P_{pr} and η do not change. The relative change in exceedance the x-threshold is also given as a function of the change in x.

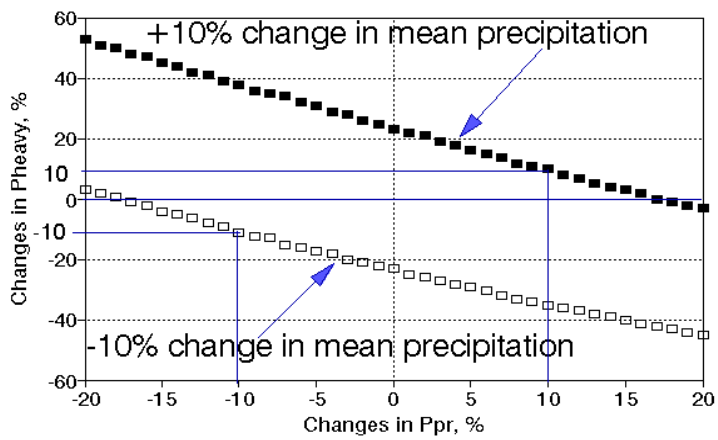

Figure 3 B. Changes in the probability of heavy rains (above 50.8 mm) in Guangzhou, PRC (P_{pr} =0.6; η = 0.56; λ= 0.04 mm^{-1}; μ = 8 mm day^{-1}) with a 10% increase/decrease in mean summer precipitation, μ, assuming that η does not change and the changes in μ are due to changes in P_{pr} and λ. Because $\mu = P_{pr}\eta/\lambda$, the changes in λ in these scenarios are a function of $\Delta\mu$ (= ±10%) and ΔP_{pr} and are not shown.

the contiguous United States with a 10% increase in μ changes by approximately 40%, 20%, and 10%, if this increase in μ is produced in Eq. 4 by an appropriate change in λ, η, or P_{pr} respectively. Obviously, an increase/decrease in P_{pr} produces a linear 1:1 increase/decrease in probability to exceed any given threshold. Changes in the two other parameters produce disproportionally high changes in the probability of extreme precipitation compared to the corresponding changes in mean precipitation.

Figure 3A further illustrates how a 10% increase in μ due to a change in the scale parameter increases the probability of daily precipitation above 25.4 mm (1 inch) from 0.018 to 0.023, assuming μ = 2.5 mm day^{-1}, η = 0.75, and P_{pr} = 0.3. This is a 25% increase in extreme precipitation occurrence compared to a 10% increase in the mean. This hypothetical example was selected to match the July daily precipitation distribution in Toronto, Ontario[*], and the precipitation increase documented over southern Canada by Groisman and Easterling (1994) (cf., also IPCC, 1996). Figure 3A also shows the effect of the threshold selection on the change in exceedance of this threshold with a 10% increase in mean precipitation due to a change in λ. In Figure 3B we sketch the changes in the probability, P_{heavy}, of summer daily precipitation above 50.8 mm in Guangzhou, PRC, when the mean precipitation, μ, changes by ±10% due to changes in P_{pr} and λ but without changes in η. It shows that depending upon the ratio of changes in these two parameters to the change in μ, P_{heavy}, can change

- with a higher than linear rate, when changes in P_{pr} are less than the changes in μ by absolute value;
- linearly, when changes in μ are solely due to changes in P_{pr} (fixed λ).
- with a lower than linear rate or inversely (in this example, when absolute values of ΔP_{pr} are above 17%), when changes in P_{pr} are higher than the changes in μ by absolute value.

Therefore, in constructing scenarios of a future climate change, we have to judge which of these three parameters will be responsible for the change in the mean precipitation. This will affect substantially the behavior of precipitation extremes in these scenarios and, in turn, will have important socio-economical and ecological consequences.

The above provides a rationale for our approach. We presume that for daily precipitation described by (3), the changes in λ, η, or P_{pr} which have occurred interannually and in the seasonal cycle during the past century as well as their spatial variability contain information about the stability of these parameters in moderate climate and weather variations. Then, using this information, we can apply a plausible scenario of the mean precipitation change and derive valuable

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[*] Empirical estimates of the probability of the daily precipitation total in July to exceed a threshold of 25.4 mm are equal to 0.025 for the Toronto International Airport and 0.020 for the downtown Toronto meteorological stations.

information about the most probable change in precipitation extremes. We consider scenarios of the small/moderate increase of mean precipitation that match the precipitation changes during the past 100 years over the countries under consideration (IPCC, 1996). We estimate parameters of model (3) for the period of the mass data availability and then use them to test the present and future tendencies in extreme precipitation.

The next section describes the spatial distribution of parameters of summer daily precipitation over the countries under consideration. It is followed by analyses of temporal and spatial stability of one of these parameters, η, and possible trends in precipitation frequency, P_{pr}. The final section presents major results of this study: the effects of changes in mean precipitation on the extreme daily precipitation values under the assumption that the shape parameter of the precipitation distribution and the frequency of the precipitation events do not change. Other scenarios for changes in the parameters of the precipitation distribution model are also considered. The scenario results are compared with direct estimates of trends in heavy precipitation during the past 100 years over the United States, Australia, and Norway.

4. Summer Daily Precipitation and Its Parameters

The results of model (3) for the summer (JJA, for Australia DJF) daily precipitation distribution are shown in Figure 4 for all eight countries under consideration. We selected a schematic presentation of the mean seasonal precipitation, μ, in Figure 4A because this quantity is well documented in climatological literature. We selected regions with very different precipitation regimes: Arctic tundra, deserts of central Asia, southern Australia, north-west Mexico, and western USA receive less than 1 mm per day, while precipitation over tropical regions of southern Mexico, northern Australia, and southern China exceed 10 mm per day. Over vast agricultural areas of the northern and southern extratropics, summer precipitation of 2 to 3 mm per day is common; eastern China, southeastern USA, the Gulf coast of Mexico, western Norway, and coastal regions of east Australia receive on average 5 mm per summer day.

Figure 4B presents the probability distribution of summer daily precipitation, P_{pr}. It shows that the frequency of rainy days varies widely over the study area. It is close to 0.5 over the wet tropics of southern Mexico, southern China, and northern Australia, over the North Atlantic region (Norway, Labrador Peninsula, Northwest USSR and Poland), and over coastal and mountainous regions along the Pacific rim (southern Alaska, northwest China, Sichuan and Yunnan provinces of PRC, and the Russian Far East). It is less than 0.05 over the deserts of North America, Australia, and Eurasia.

Figures 4C and 4D show the distribution of the scale and shape parameters of the gamma distribution of daily summer precipitation on the days with measurable

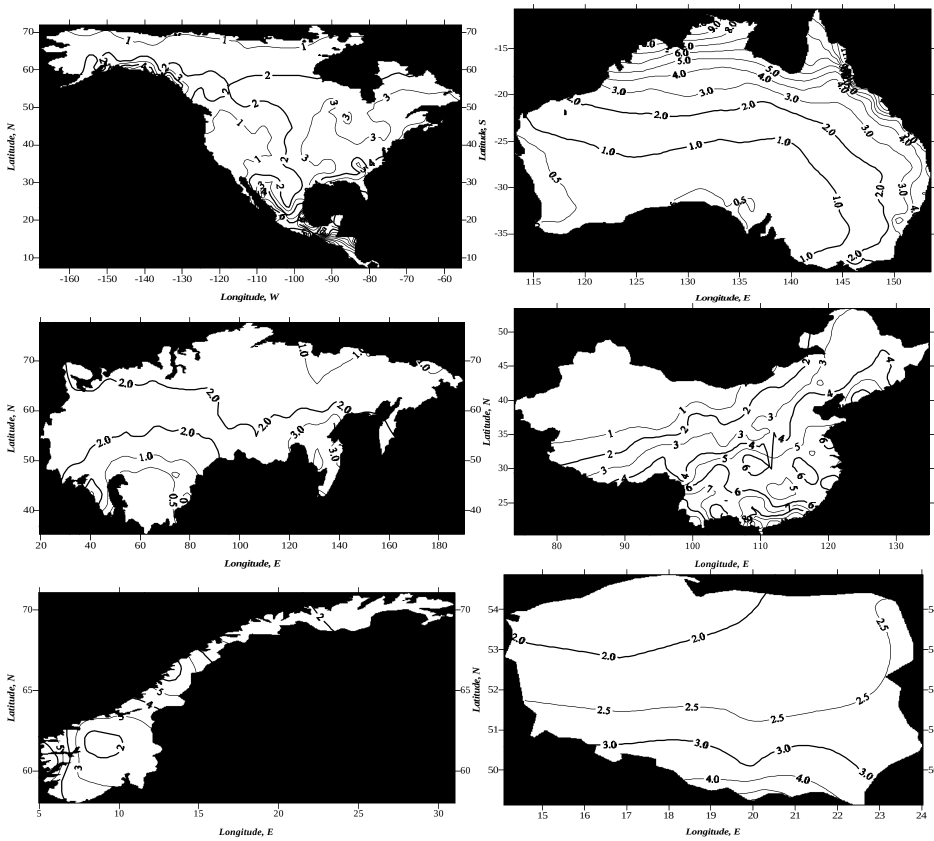


Figure 4 A. Mean daily summer (June-August) precipitation (mm day^{-1}) over North America (Canada, United States, and Mexico), Australia, the former Soviet Union, PRC, Norway, and Poland for the three parameter model of the precipitation distribution (3). In Australia, December through February are considered as summer months.

rain. The scale parameter, λ, has units of mm^{-1} for daily precipitation. Smaller values of λ indicate higher intensities of daily precipitation. It is a parameter that changes by an order of magnitude from subarctic regions and deserts (~0.30 mm^{-1}) to humid tropics (~0.03 mm^{-1}). However, the shape parameter, η, is dimensionless and has little spatial variation. The fact that the shape parameter is a spatially and temporally stable characteristic of regional precipitation has been shown at monthly and annual time scales (Shver, 1976; Groisman and Easterling, 1994). Figure 4D and the next section show that this is also true at the daily time scale. Over the eastern two thirds of the contiguous United States, Russia, and Canada with daily summer precipitation above 1 mm day^{-1}, this parameter varies by 10-15% around its mean value of approximately 0.8. Changes are very small over Poland, Norway, eastern Australia and eastern China. Over Australia, the η-values

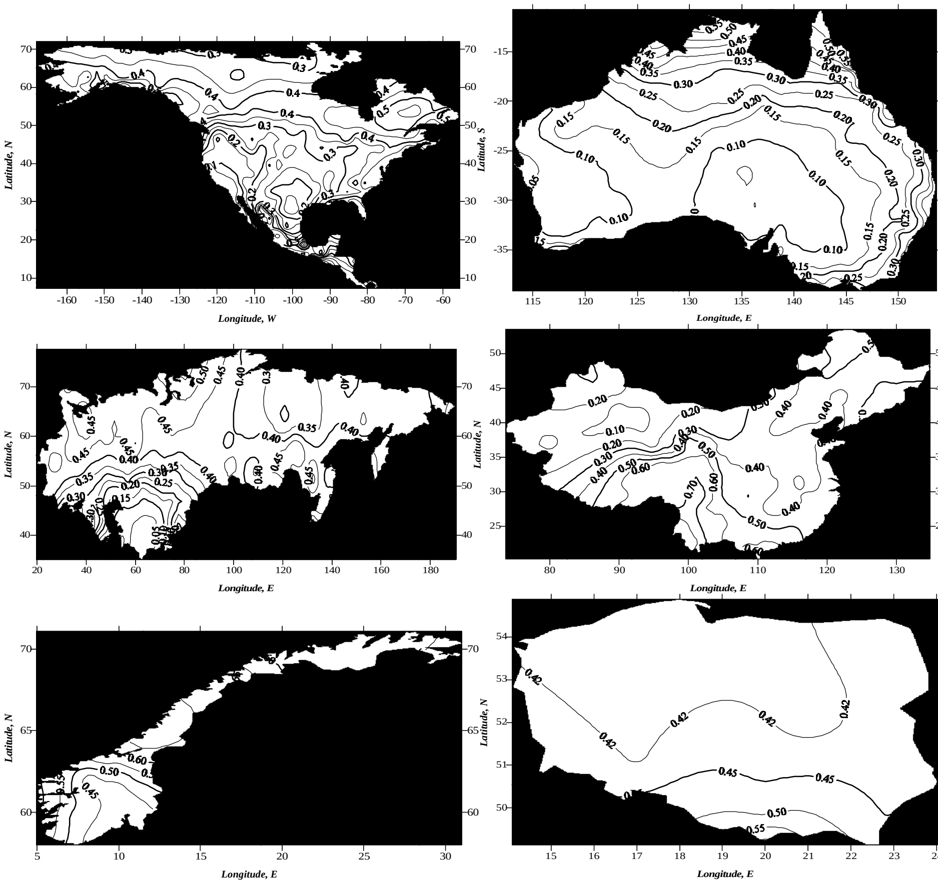


Figure 4 B. The probability of summer (June-August) daily precipitation over North America (Canada, United States, and Mexico), Australia, the former Soviet Union, PRC, Norway, and Poland for the three parameter model of the precipitation distribution (3). In Australia, December through February are considered as summer months.

for summer (DJF) are similar to those over North America and Eurasia and vary around a mean value of approximately 0.75. In monsoon regions of China and the Russian Far East the η-values for summer (JJA) are relatively low (varying from 0.5 to 0.6). Over regions with daily summer precipitation above 1 mm day^{-1}, the lowest values of η are estimated in eastern China (up to 0.45 on Shandong Peninsula) and the highest in the tropics of southern Mexico (up to 1.2 along the coast of the Gulf of Mexico). Generally, we found little spatial variation in our estimates of this parameter with the exception of mountainous and desert regions. This exception is further illustrated in Figure 5 where we single out the μ- and η-estimates over Mexico. Here, high gradients of mean precipitation (e.g., along the Gulf and Pacific coasts, it differs by more than an order of magnitude) are associated with a higher spatial variability of the η-parameter than over seven other countries: it varies from 0.7 to 1.2, i.e., by ±25%.

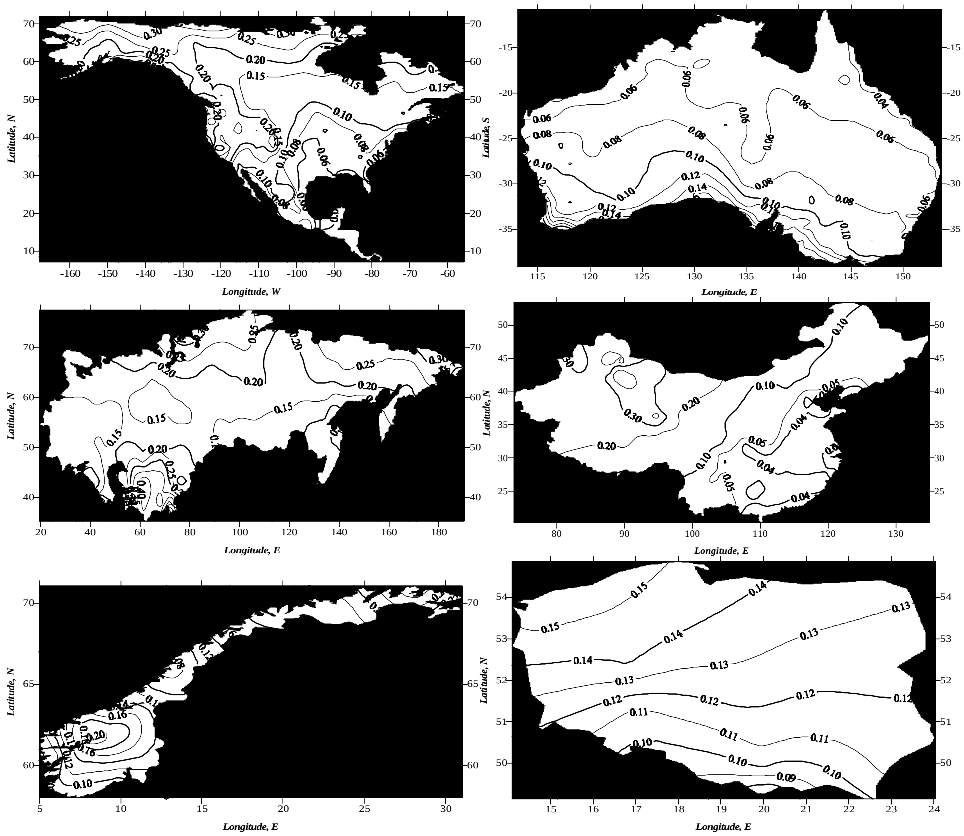


Figure 4 C. The scale parameter λ (mm^{-1}) over North America (Canada, United States, and Mexico), Australia, the former Soviet Union, PRC, Norway, and Poland for the three parameter model of the precipitation distribution (3). In Australia, December through February are considered as summer months.

5. Testing the Temporal Stability of the Shape Parameter

In the previous Section we have shown the spatial stability of the shape parameter of precipitation distributions. There are indications that an increase/decrease in mean precipitation at long-term stations is accompanied by an increase/decrease in precipitation variability (Bootsma, 1994). This leaves the coefficient of variation C_v of precipitation less affected by these changes and, according to equation (2'), the shape parameter, η, stays intact too. Now we test the temporal stability of the shape parameter in two ways:

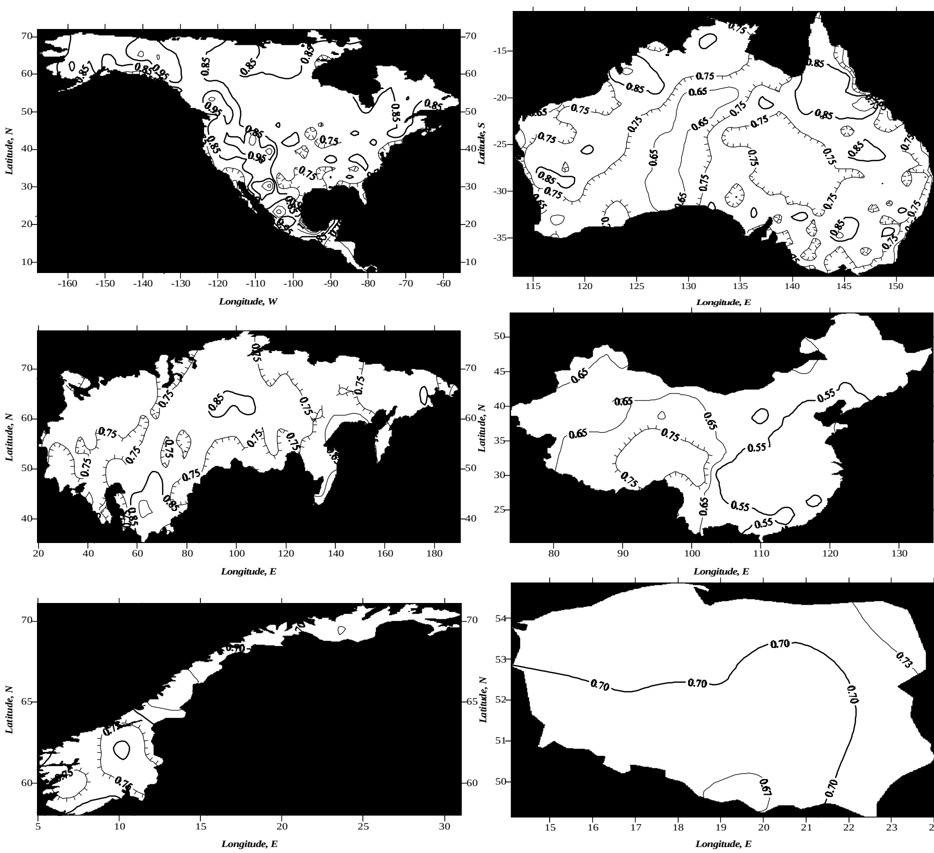


Figure 4 D. The shape parameter, η, over North America (Canada, United States, and Mexico), Australia, the former Soviet Union, PRC, Norway, and Poland for the three parameter model of the precipitation distribution (3). In Australia, December through February are considered as summer months.

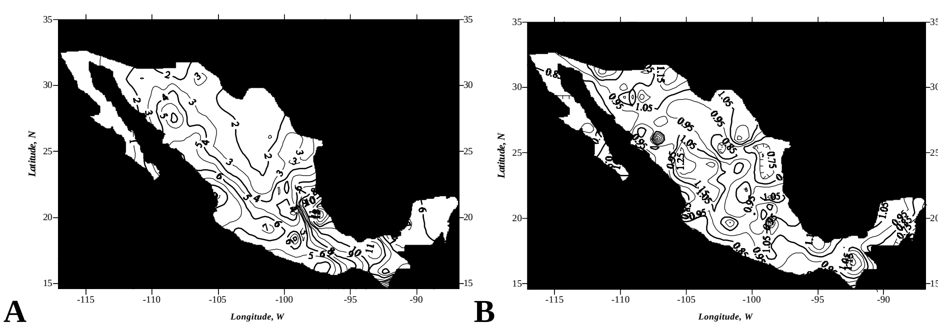


Figure 5. (A) Mean daily summer precipitation (mm day^{-1}) and (B) the shape parameter, η, over Mexico for the three parameter model of the precipitation distribution (3).

- changes in the seasonal cycle and

- changes between “wet” and “dry” summers.

For the latter purpose the entire period of homogeneous summer observations at each station was divided into two groups of summers, those that have seasonal total precipitation below the long-term mean value and those that have seasonal total precipitation above this mean. This dichotomy essentially changed the mean precipitation values in each group. The difference between “wet” and “dry” summers was usually on the order of magnitude of the mean precipitation in “dry” summers, thus this partition imitated a large “climatic” change in precipitation. We then compared the parameters of the model (3) for each of these two periods to find out which of the three parameters changes the most.

Table I

The country-wide percentage differences, Δ, in parameters of daily precipitation between “wet” and “dry“ summers for the United States, Australia, China, the former Soviet Union, Norway, and Poland over the regions with “dry” summer precipitation above 1 mm day^{-1}. Differences are presented in percent of the mean “wet” values [e.g., $\Delta\mu$ = 100% (μ(wet)- μ(dry))/ μ(wet)].

Country	$\Delta\mu$	ΔP_{pr}	$\Delta\eta$	$\Delta\lambda$
USA	50	31	-10	-52
Australia	52	27	-10	-71
China	36	12	-3	-43
former USSR	39	17	-6	-45
Poland	36	15	-8	-44
Norway	38	17	1	-33

Analysis of the behavior of the shape parameter in the seasonal cycle shows that it is also relatively stable during the march of the seasons (Figure 6). In this figure we present the seasonal cycle of monthly precipitation and the shape parameter of the distribution of daily precipitation for a broad variety of regions spanning from Subarctic Canada to the Gulf of Mexico, from the North Atlantic to the South-China Sea, and over many climatic zones of the Australian continent. For comparison, the mean monthly precipitation is plotted on the same graphs to illustrate the fact that the long-term mean precipitation, μ, is much more variable than the η-parameter. However, a noticeable exception occurs in regions with a very strong seasonal cycle of precipitation during the “dry” season, when daily precipitation is much less than 1 mm/day. This exception is further illustrated by the spatial pattern of summer daily precipitation and the η-parameter of its distribution over Mexico (Figure 5). We are mostly interested in the probability of heavy rains and changes associated with moderate changes in mean precipitation. This goal allows us to omit areas with the mean summer precipitation below 1 mm per day from further consideration in this study and focus on other regions which essentially include major agricultural areas in each country.

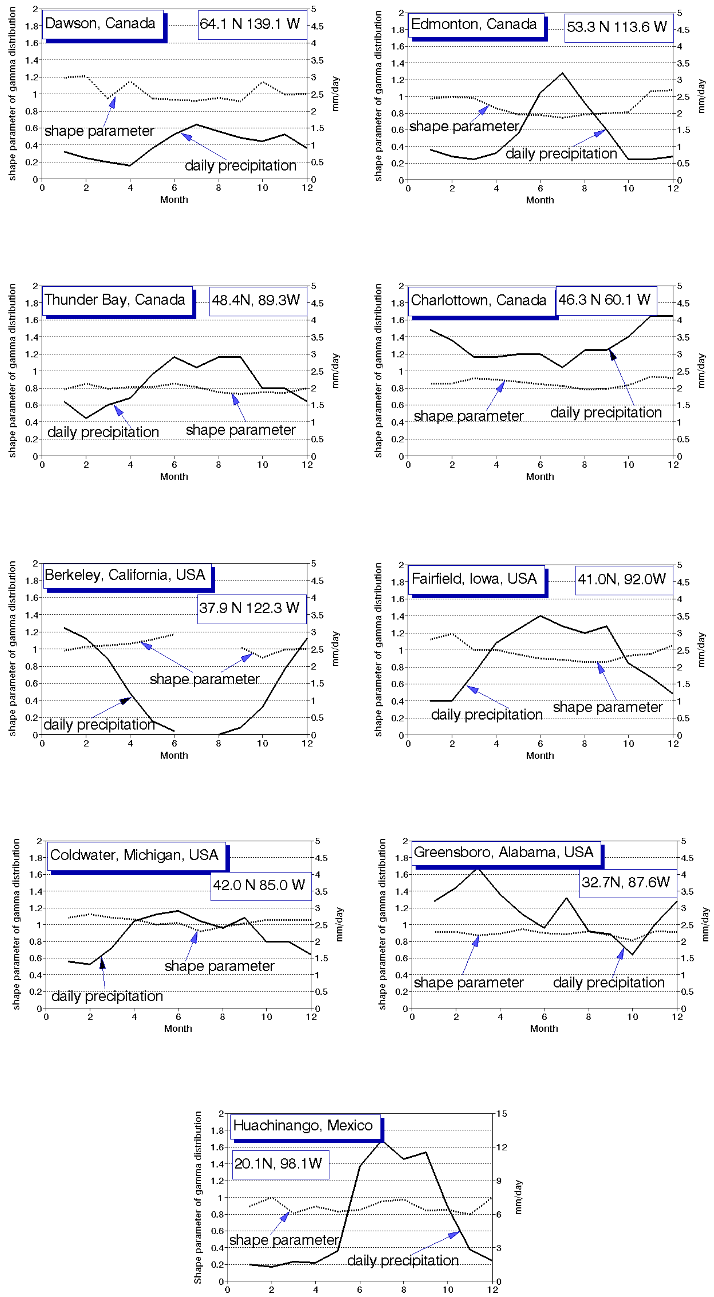
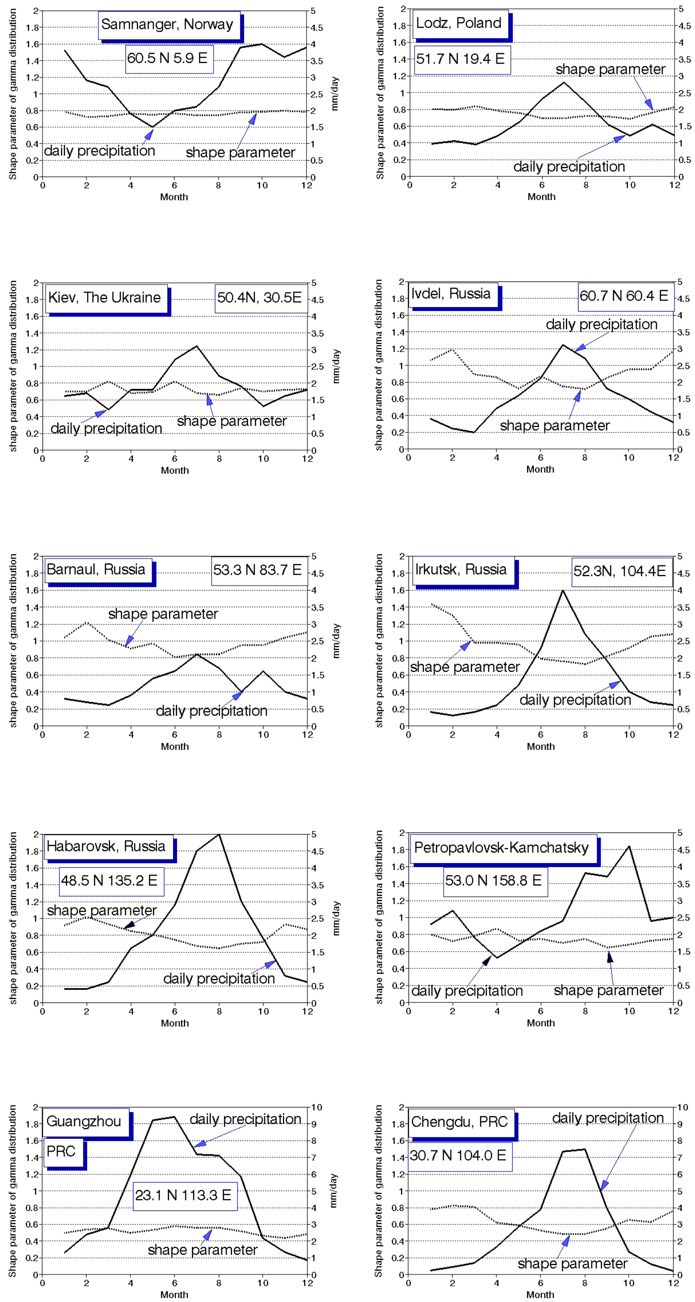


Figure 6 A. Seasonal cycle of the mean daily precipitation (mm day^{-1}), μ, and the shape parameter, η, for selected stations over North America for the three parameter model of the precipitation distribution (3). For Canada, only snowfall data were used instead of the gauge measurements to avoid the homogeneity problems with a cold season precipitation time series. The y-axis in the graph of Mexican daily precipitation was reduced threefold compared to other graphs in this Figure.


Samnanger, Norway
60.5 N 5.9 E
daily precipitation
shape parameter
Lodz, Poland
51.7 N 19.4 E
shape parameter
daily precipitation
Kiev, The Ukraine
50.4N, 30.5E
daily precipitation
shape parameter
Ivdel, Russia
60.7 N 60.4 E
daily precipitation
shape parameter
Barnaul, Russia
53.3 N 83.7 E
shape parameter
daily precipitation
Irkutsk, Russia
52.3N, 104.4E
shape parameter
daily precipitation
Habarovsk, Russia
48.5 N 135.2 E
shape parameter
daily precipitation
Petropavlovsk-Kamchatsky
53.0 N 158.8 E
daily precipitation
shape parameter
Guangzhou
PRC
23.1 N 113.3 E
daily precipitation
shape parameter
Chengdu, PRC
30.7 N 104.0 E
daily precipitation
shape parameter
Shape parameter of gamma distribution
mm/day
Month

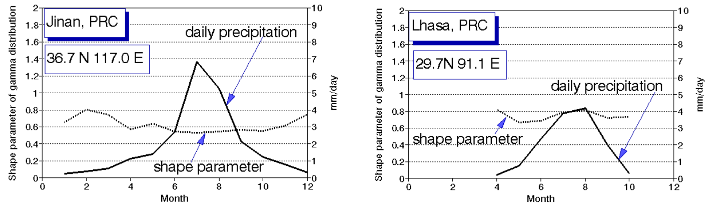


Figure 6 B. Seasonal cycle of the mean daily precipitation (mm day^{-1}), μ, and the shape parameter, η, for selected stations over Northern Eurasia for the three parameter model of the precipitation distribution (3). To derive the graphs in this figure for the former Soviet Union, only the data after 1967 were used to avoid the homogeneity problems with a cold season precipitation time series. The y-axes in the graphs of Chinese daily precipitation were reduced twofold compared to other graphs in this Figure.

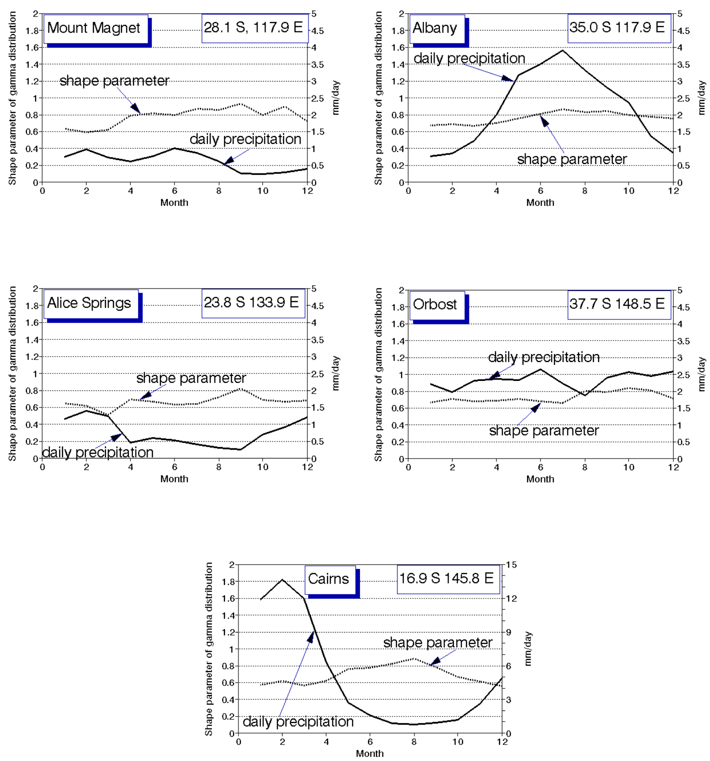
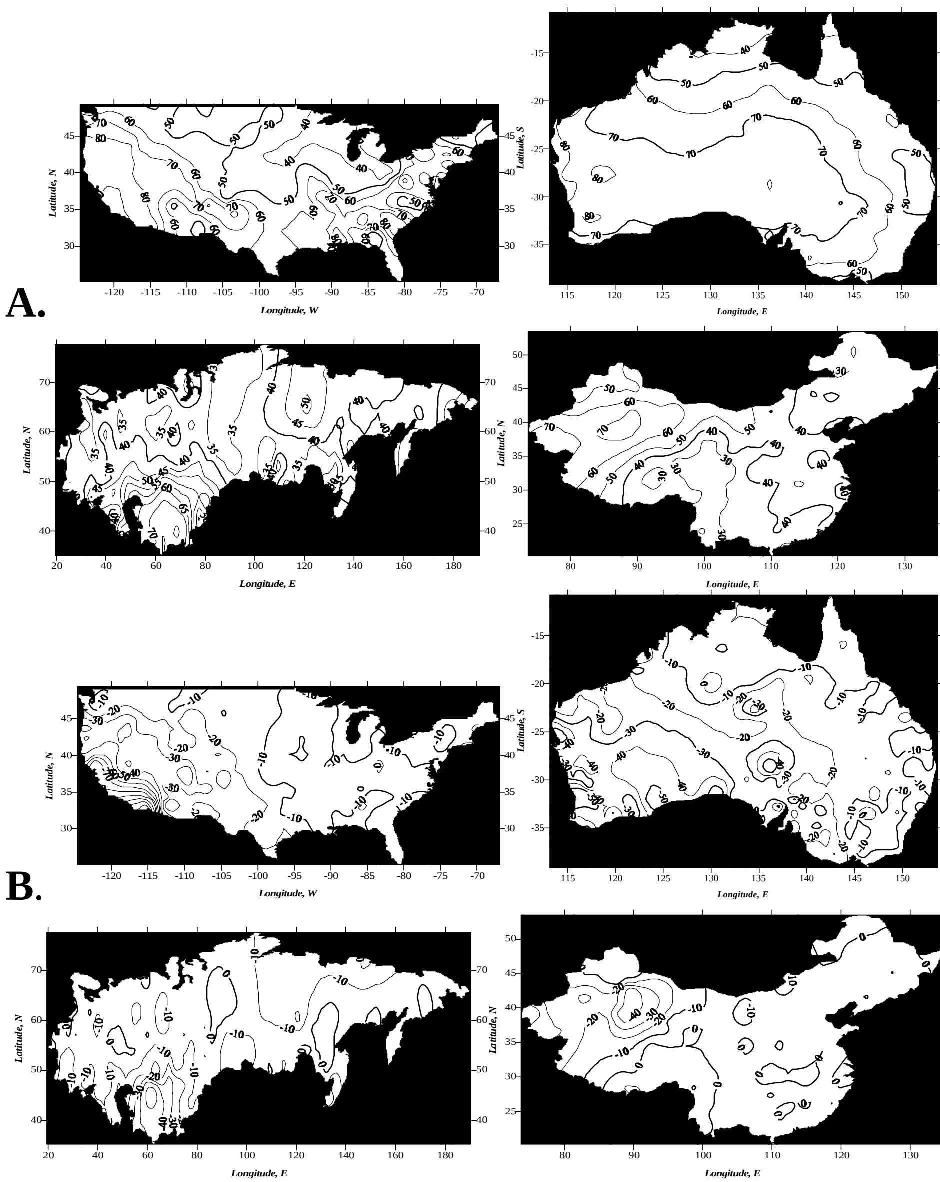


Figure 6 C. Seasonal cycle of the mean daily precipitation (mm day^{-1}), μ, and the shape parameter, η, for selected stations over Australia for the three parameter model of the precipitation distribution (3). The y-axis in the graph of daily precipitation in Cairns (northeastern Australia) was reduced threefold compared to other graphs in this Figure.

Figure 7 and Table I summarize our intercomparison of the parameters of daily precipitation distribution in “wet” and “dry” summers. This analysis was performed for six countries, excluding Canada and Mexico. The nature of our dichotomy forces the mean values, μ, to differ between “wet” and “dry” summers by 40 to 60% of the “wet” totals (Table I). The smallest variation between “wet” and “dry” summers was documented in the maritime climate of Norway (about 40% difference for a 100-year-long time series; in Poland, Russia, and China 40% differences were encountered for 30 to 60-year time series).   Over the continental



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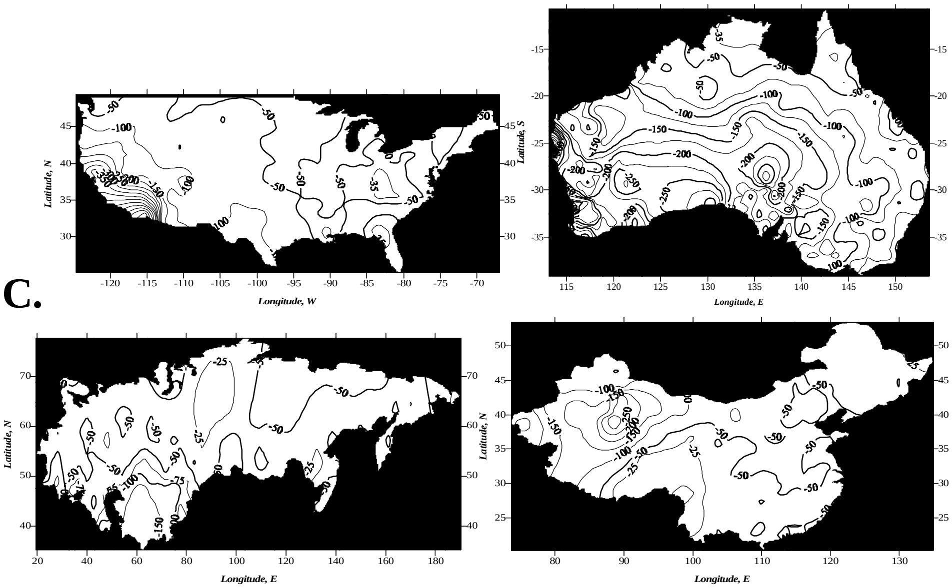

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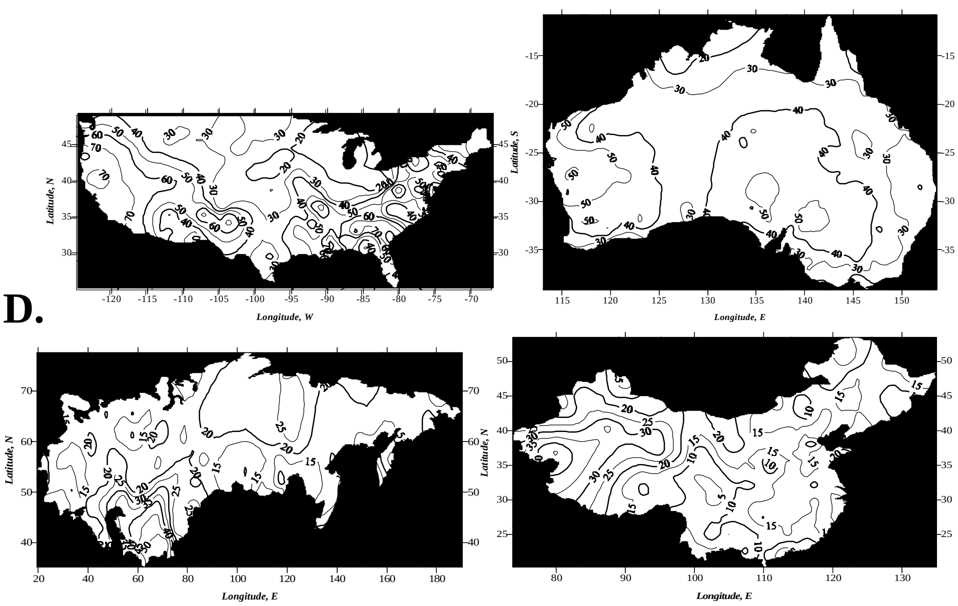

Figure 7. Differences (%) between “wet” and “dry” summers (JJA, DJF for Australia) in four countries (The United States, Australia, the former USSR, and China) as measured by (A) mean precipitation; (B) shape parameter of the distribution of precipitation totals; (C) scale parameter of the distribution; and (D) probability of a day with precipitation. “Wet” summers at a given station have the mean precipitation above the long-term mean value, while other summers are considered “dry”.

areas with the mean daily precipitation above 1 mm, these differences are also close to 40%. Over the contiguous United States and Australia the differences are higher than 40% because we have longer (~90 years) time series (e.g., for Russia only the last 27 years with homogeneous precipitation time series were used in this intercomparison). The mean precipitation is a product: $\mu = P_{pr}\eta/\lambda$, and its change is a result of changes in one or more of these three parameters. The average differences in Table I and the patterns of these differences in Figure 7 show that the most variable parameter, which contributes most to the difference between "wet" and "dry "summers, is the scale parameter which may change by 100% or more. The shape parameter is less variable and changes only slightly over eastern China, former Soviet Union, Poland, and Norway. High precipitation variability in Australia and the USA leads to twice as much precipitation in wet summers relative to dry summers in the past 100 years, yet there is only a modest 10% decrease in η which does not noticeably contribute to the change in μ over the regions with daily precipitation above 1 mm day^{-1}.

6. Testing Changes in Precipitation Probability

The probability of daily precipitation, P_{pr}, can be estimated from the available data sets even when we neglect the precipitation less than 1 mm (Bogdanova, 1987). The number of days with precipitation in this category is closely related to the lower threshold of the precipitation gauge measurements. This threshold was not constant for many precipitation networks throughout the world including, e.g., Russia, Australia, and Canada and may introduce artificial trends in the number of days *without* precipitation. To avoid the above mentioned inhomogeneities in Russian, Australian and Canadian precipitation data, and keeping in mind that the daily precipitation less than 1 mm usually contributes only a few percent to monthly totals and is not of practical importance, we analyze the probability of

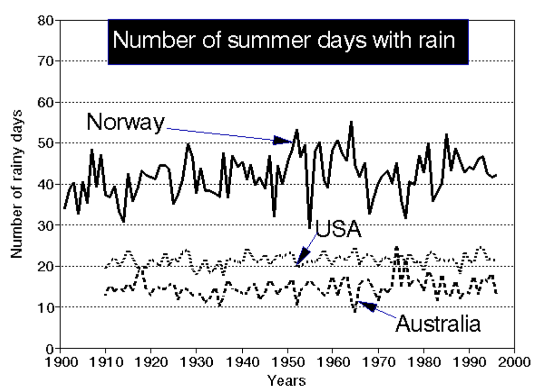

Figure 8. Average number of summer days with precipitation, area-averaged over the United States, Australia, and Norway. Only the days with precipitation above 0.2 mm and 1 mm have been counted at Norwegian and Australian stations respectively.

days with "measurable precipitation above 1 mm day^{-1}" for these three countries instead of P_{pr}.

For the United States, Australia, and Norway, we searched for century long trends in precipitation frequency (Figure 8, Tables II and III). For other countries our analyses are restricted to the post World War II period only. Before W.W.II, daily precipitation time series are unavailable for China and Mexico and there are inadequate data for the former Soviet Union, Poland, and Canada.

Table II

Linear trends in the number of precipitation days and mean precipitation for each season, area-weighted over the contiguous United States for period 1910-1996. Asterisk indicates a statistically significant difference from zero at the 0.05 significance level. The estimates are based on the century-long daily time series from the 187-station HCN data set.

Season	Mean number of days with precipitation	Linear trend (days/10yrs)	Correlation with seasonal precipitation totals
Winter	22	0.09	0.84
Spring	24	0.27*	0.91
Summer	22	0.12	0.90
Autumn	19	0.29*	0.93

Season	Long-term mean precipitation, mm	Linear trend, mm/10yrs	Linear trend, %/10yrs
Winter	170	0.0	0.0
Spring	200	2.0	1.0
Summer	205	1.5	0.7
Autumn	170	2.6*	1.5*

Table II shows the trends in area-weighted numbers of precipitation days per season and precipitation totals over the contiguous United States. Trends in summer and winter P_{pr} over the contiguous United States are not statistically significant at the 0.05 significance level. In spring and autumn upward trends in P_{pr} result in an annual increase of 5 to 6 precipitation days relative to the beginning of this century. The summer trend of P_{pr} over the U.S. is not statistically significant at the 0.05 level, but is significant at the 0.10 level[**]. In two regions of the Mid-West (The Upper Mississippi and Missouri River Basins) the increase of 4 to 5 summer days per 100 yrs is

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[**] Here and throughout this paper, we employ a two-tail student t-test for testing linear trends for statistical significance. But, when we applied the non-parametric test for nonrandomness of the ranks in this time series based on Kendall's τ statistic (Kendall and Stuart, 1979), the upward trend in summer P_{pr} was found statistically significant at the 0.05 level.

statistically significant at the 0.05 level. Trends in precipitation totals have been discussed by Karl and Knight (1998). The numbers presented here differ from those shown in the first line of Table I of Karl and Knight (1998) for the special HCN network due to several improvements in this data set (infilling of missing values, inclusion of few additional stations to cover the data sparse areas, and additional quality control that allow us to reveal and fix some erroneous values).

There is no indication of statistically significant changes in winter (JJA) frequency of rainy days in Australia. However, during the summer (DJF) season in the southeast of the continent, a century-long statistically significant 20% increase of precipitation frequency has occurred (Table III). In a related study, Hennessy et al. (1998) found significant increases in the number of rain days in all seasons except winter in Australia from 1910-1995.

Table III

Linear trend in the number of precipitation days and mean precipitation for the summer season (DJF) area-weighted over the Australian continent for the periods 1910-1996 (continent) and 1900-1996 (eastern coastal regions). Asterisk indicates a statistically significant difference from zero at the 0.05 significance level.

Region	Mean precipitation, P_1	Linear trend in P_1 (%/10yrs)	Number of days with precipitation above 1 mm, N_1	Linear trend in N_1 (%/10yrs)
Entire continent	210	0.9	15	1.1
Coastal regions of Queensland and Northern Territory	590	0.6	33	0.8
Coastal regions of New South Wales and Victoria	180	3.1*	17	2.0*

Over Norway a century-long trend in annual precipitation was reported by Hanssen-Bauer and Førland (1994). P_{pr} at Norwegian stations exceeds 50% and the increase in summer total precipitation was accompanied by a further increase in precipitation frequency. Five more summer days with precipitation above 0.2 mm (three with precipitation above 1 mm day^{-1}) are registered now compared to the beginning of the century (Figure 8). Analyses of the number of summer days with precipitation over Poland, China, Russia, Canada, and Mexico in the post W.W.II period show no indication of trends.

Tables IV and V and Figure 9 summarize our analyses of trends in the number of days with "measurable precipitation above 1 mm day^{-1}" for Russia and Canada.

Table IV

The annual number of days with precipitation above 1 mm averaged over southern Canada (south of 55°N) separately for liquid and frozen precipitation. There are no changes that are statistically significant different from zero at the 0.05 significance level. Asterisk indicates a statistically significant difference from zero at the 0.10 significance level.

Precipitation Type	Period 1943-1975	Period 1976-1995	Difference, days
Rainfall	89	90	1
Liquid equivalent of snowfall	46	44	-2*
Summer Precipitation	31	31	0

Table V
The average number of summer days with precipitation above 1 mm averaged over several regions of the former USSR. Asterisks indicate statistically significant differences/trends at the 0.05 significance level.

	Number of stations	Period 1943-1975	Period 1976-1986	Difference (days)	Linear trend 1936-1994 (day/50 yrs)
European part of the former USSR	89	25	26	1	0.4
Asian part of Russia	99	28	27	-1*	-2.2*
Kazakhstan and Central Asian States	35	7	6	-1*	0.2
Former USSR	223	23	23	0	-0.5

In southern Canada (Table IV) we found no trends in the number of days with precipitation above 1 mm, but registered a small redistribution between precipitation in frozen and solid form. On average, after the mid-1970s the average number of days with rainfall has increased by a day compared to the three previous decades, while the average number of days with snowfall decreased by approximately two days. There is no trend in summer P_{pr} for southern Canada.

Analysis of the number of days with precipitation above 1 mm for Russia (Table V) shows that since 1936 there has been an absence of systematic changes in P_{pr} over the European part of Russia but we found a statistically significant trend (19%/50yrs) in frequency of summer daily precipitation events above 20 mm. Over the Asian part of Russia (Siberia) we found a statistically significant *decreasing* trend in precipitation frequency.

In summary, over the United States, Norway, and Australia we found an increase in summer precipitation frequency over the past century. The increase in precipitation frequency in Norway (5 days per 100 years) is larger than in the other seven countries. The highest *relative* change in precipitation frequency (about 20%) has occurred in southeastern Australia. When the same analyses are repeated only for the post- W.W.II period, they do not show statistically significant trends in P_{pr} over these countries. For all other countries we could not find systematic changes in summer values of P_{pr}. Therefore, we conclude that there is no evidence that P_{pr} during summer months has substantially changed during the past five decades over the large-scale regions considered with the exception of the Asian part of Russia. Regional time series of P_{pr}, however, deserve a more thorough analysis, which is beyond the scope of this paper.

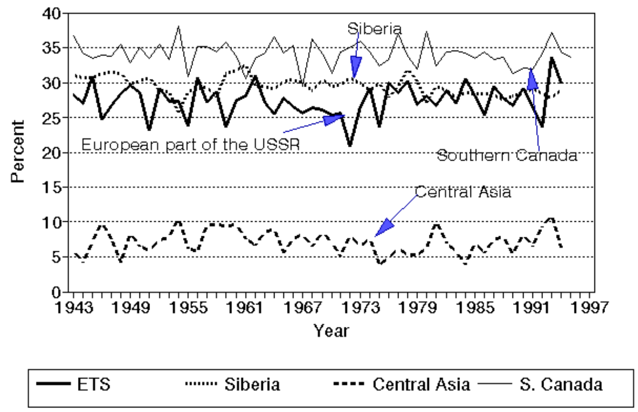

Figure 9. Percent of summer days with precipitation above 1 mm over the former Soviet Union and southern Canada. The numbers were arithmetically averaged over 58 first order stations in Southern Canada (south of 55°N), 89 first order stations in the European part of the former Soviet Union (ETS), 99 stations in Siberia, and 35 stations in Central Asian members of the Commonwealth of Independent States.

7. The Effect of Changes in Mean Precipitation on Heavy Daily Precipitation

Katz and Brown (1992) established that the probability of an extreme event (i.e., the probability of the meteorological variable exceeding an unusually high value threshold) becomes significantly larger for comparable increases in standard deviation compared to the mean. For daily summer precipitation during the post-W.W.II period, this general conclusion is superimposed over the spatial and temporal stability of the shape parameter (Figures 4 through 7 and Table I) and empirical evidence of the stability of the number of precipitation events (Tables IV and V and Figures 8 and 9). In moderately different climates (which can be associated with geographical shifts of climatic zones, the seasonal cycle of precipitation, cyclone tracks, etc.) we might expect that in each region the shape parameter will stay mostly intact. This, in turn, means that for summer precipitation over Australia, North Eurasia and America, changes in mean values will be approximately matched by changes in standard deviation (cf., Eqs. 2'), which will strongly affect the probability of extreme precipitation (cf., Figure 3).

In all eight countries considered (except China) at least a 5% increase in mean summer precipitation has been documented during the past 100 years (IPCC, 1996, 1998; Groisman and Easterling, 1994; Lettenmaier et al., 1994; Karl et al., 1993; Karl and Knight, 1998; Vinnikov et al., 1990; Groisman, 1991; Georgievsky et al., 1995, 1996; Lavery et al., 1992, 1997; Hanssen-Bauer and Førland, 1994; Hanssen-Bauer, 1994; Kozuchowski, 1985). In three countries (USA, Australia,

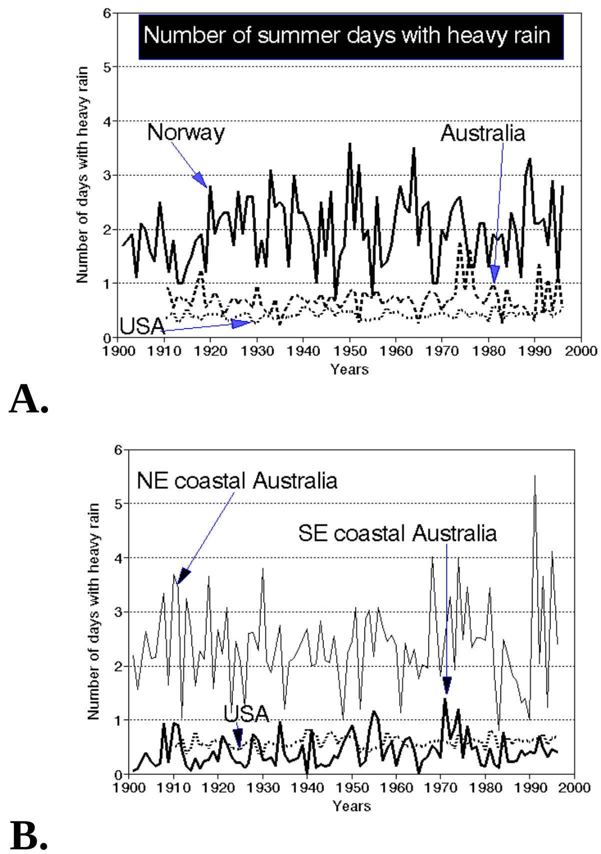

Figure 10. (A) Average number of summer days with precipitation exceeding 50.8 mm (25.4 mm for Norway), area-averaged over Australia, the United States, and Norway. (B) The same, but area-averaging was conducted only over the eastern two-thirds of the contiguous United States and over the northeastern (north of 30°S, east of 130°E) and southeastern (south of 30°S, east of 140°E) coastal regions of Australia less than 350 km away from the coast line.

and Norway) we found a century-long increase in heavy precipitation frequency and in P_{pr} (Karl et al., 1996; Karl and Knight, 1998; Suppiah and Hennessy, 1996; Tables II, III and VI; Figure 10). In other countries, where we have shorter and/or insufficient data, the direct detection of systematic changes in heavy precipitation using observational data is more difficult. Therefore, we exploit the findings of previous sections about the stability of η and P_{pr} (the temporal stability of P_{pr} is, however, dependent on the period of interest) and use various assumptions about how the mean precipitation may (or did) change to analyze the effect of these changes on extreme precipitation. As an example of this type of analysis, below we present the effect of a 5% increase in mean precipitation on the precipitation above selected thresholds assuming a scenario of no changes in η and P_{pr}. We apply this scenario to all eight countries although summer precipitation in some of them (Russia, Canada, Australia, Norway, Mexico) has increased at a higher rate during the past century, while over eastern China it decreased during the period from 1909 to 1993 (Ye et al., 1996).

Table VI

Country-wide linear trends of the number of summer days with heavy precipitation over the contiguous United States, Australia, and Norway. Asterisk indicates a statistically significant difference from zero at the 0.05 significance level.

Country	Period	Threshold used to define "heavy" rain	Average number of days with heavy rain	Linear trend, day/10years	Linear trend, %/10years
Contiguous USA	1910-1996	50.8 mm	0.4	0.007*	1.7*
Eastern two-thirds of the contiguous USA	1910-1996	50.8 mm	0.6	0.010*	1.7*
Australia	1910-1996	50.8 mm	0.7	0.018	1.1
Coastal regions of New S. Wales and Victoria	1900-1996	50.8 mm	0.4	0.019*	4.6*
Norway	1901-1996	25.4 mm	2.0	0.04	1.9

We define (somewhat arbitrarily) "heavy" precipitation, P_{heavy}, as a daily precipitation exceeding the 25.4 mm threshold in northern countries (Russia, Canada, Norway, and Poland) and exceeding the 50.8 mm threshold in mid-latitudes (the United States, Mexico, China, and Australia). Figures 11 and 12 provide the climatology of summer heavy precipitation estimated from model (3). Figures 13 through 15 summarize our estimates of the disproportionate increase in precipitation for heavy precipitation rates, compared to a 5% increase in mean precipitation, if the shape of the precipitation distribution and the probability of a precipitation event do not change.

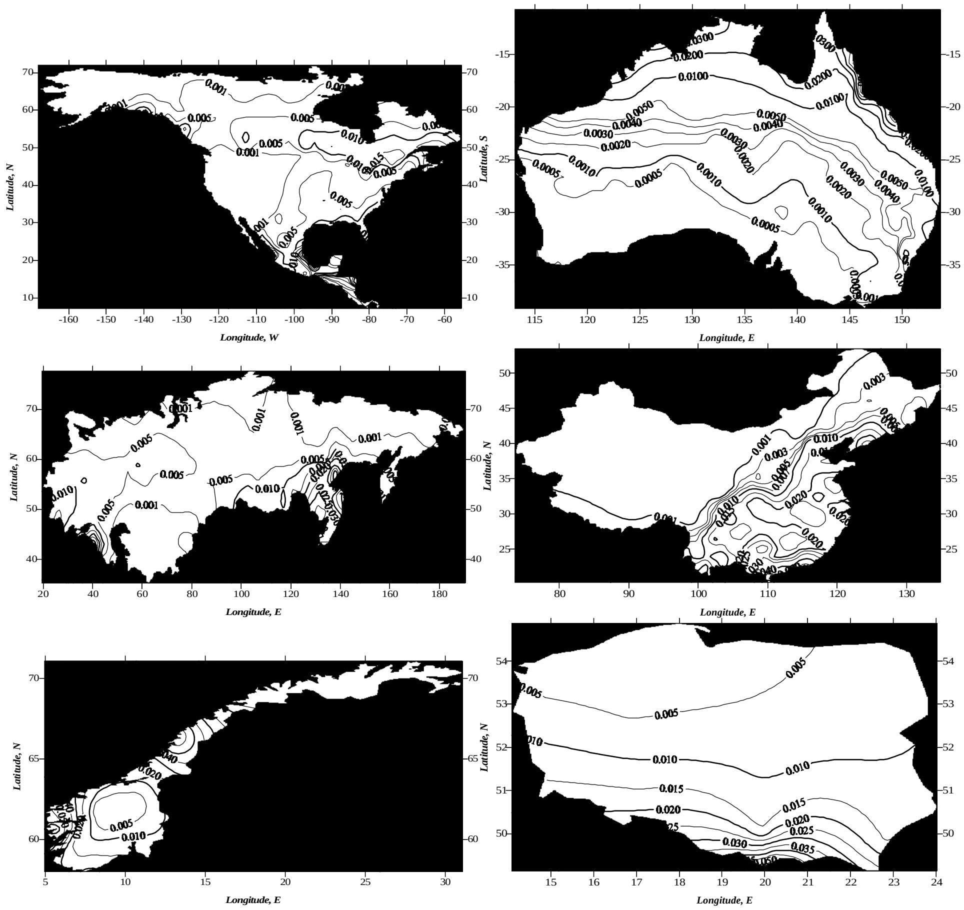

Figure 11. Probability of a day with precipitation exceeding 25.4 mm (Canada, the former Soviet Union, Poland, and Norway) and 50.8 mm (the United States, Mexico, PRC, and Australia). Estimates are based on model (3). Direct estimates of these probabilities based on century-long precipitation time series (e.g., Figure 2) resemble the pattern shown in this figure but can be produced only for a small part of the area under consideration (in regions with daily precipitation above 1 mm day^{-1} in contiguous United States, Australia, Norway, European Russia, and southeastern Canada).

Over all of southern Canada, the former Soviet Union, Poland, and Norway a 5% increase in mean summer precipitation manifests itself in a 20% increase of the probability of days with precipitation above 25.4 mm (Figure 13). These heavy precipitation events (which on Figure 12 contribute less than 5% of summer precipitation totals of Norway, Russia, Canada, and Kazakhstan and less than 10% of summer precipitation of Belarus, Poland, and The Ukraine) contribute up to 30% (Russia, Canada, northern Norway) and more than 40% (southern Norway, Belarus, Poland, and The Ukraine) of the increase of mean daily precipitation (Figure 14).

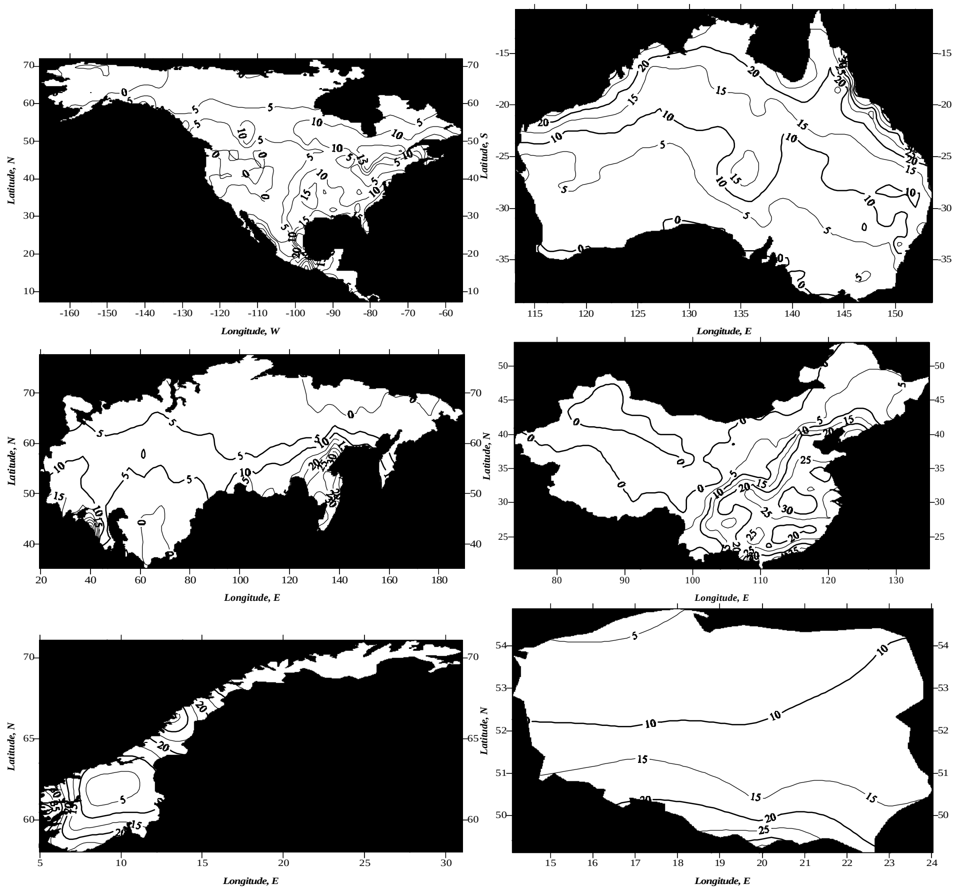

Figure 12. Percent of summer (JJA, DJF for Australia) precipitation that falls in "heavy" rains, i.e., with daily rates above 25.4 mm (Canada, the former Soviet Union, Poland, and Norway) and 50.8 mm (the United States, Mexico, PRC, and Australia). Estimates are based on model (3).

In the eastern United States, in regions with mean summer precipitation above 2 mm per day, an increase in mean daily precipitation by 5% yields an increase in the probability of daily precipitation above 50.8 mm (2 inches) by approximately 20%[***]. In the Mississippi River Basin up to half of the increase in mean summer precipitation is contributed by heavy rains (Figure 14). This helps explain why recent studies by Karl et al. (1995) and Karl and Knight (1998) were able to detect significant increases in extreme precipitation over the contiguous United States, while the century-long increases in summer precipitation totals over the same region were non-significant (Karl et al., 1993, Groisman and Easterling, 1994).

IPCC (1998) shows an increase in mean annual (summer) precipitation of 10 to 20% during the 20th century over most of Mexico. The scenario of a 5% increase

[***] By 15% to 20% in the Southeast and 20% to 30% in the Northern part of the country.

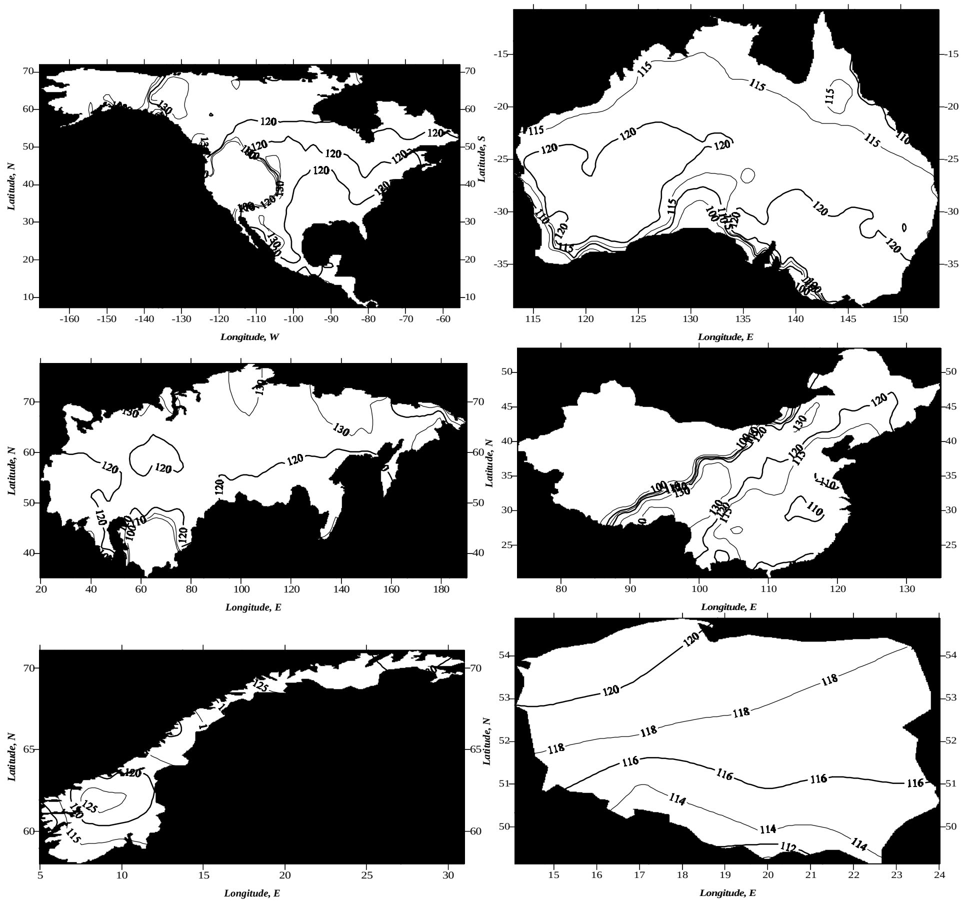

Figure 13. Percentage change of the probability of summer daily precipitation exceeding the heavy rainfall thresholds, P_{heavy}, (defined in text) when the mean daily precipitation increases by 5% assuming that P_{pr} and η do not change. The change is expressed as a ratio P_{heavy}(scenario)/P_{heavy}(climate).

in summer mean precipitation over Mexico yields a 20-30% increase in the probability of daily summer precipitation above 50.8 mm (2 inches) except in the desert regions of the country (Figure 13). In this scenario, more than 70% of the increase in mean summer precipitation over the tropical regions of Mexico is contributed by heavy rains (Figure 14). This is not a surprise because heavy precipitation is typical in the tropics (Figure 12).

For China, we considered a scenario of a 5% increase in mean summer precipitation although this scenario is not supported by real trends in mean precipitation during the past 50 years. For eastern China this scenario yields a 10 to 30% increase in the probability of summer precipitation exceeding 50.8 mm (Figure 13). This increase is less pronounced in the coastal areas with higher precipitation rates and more pronounced inland. Over the Tibetan Plateau and

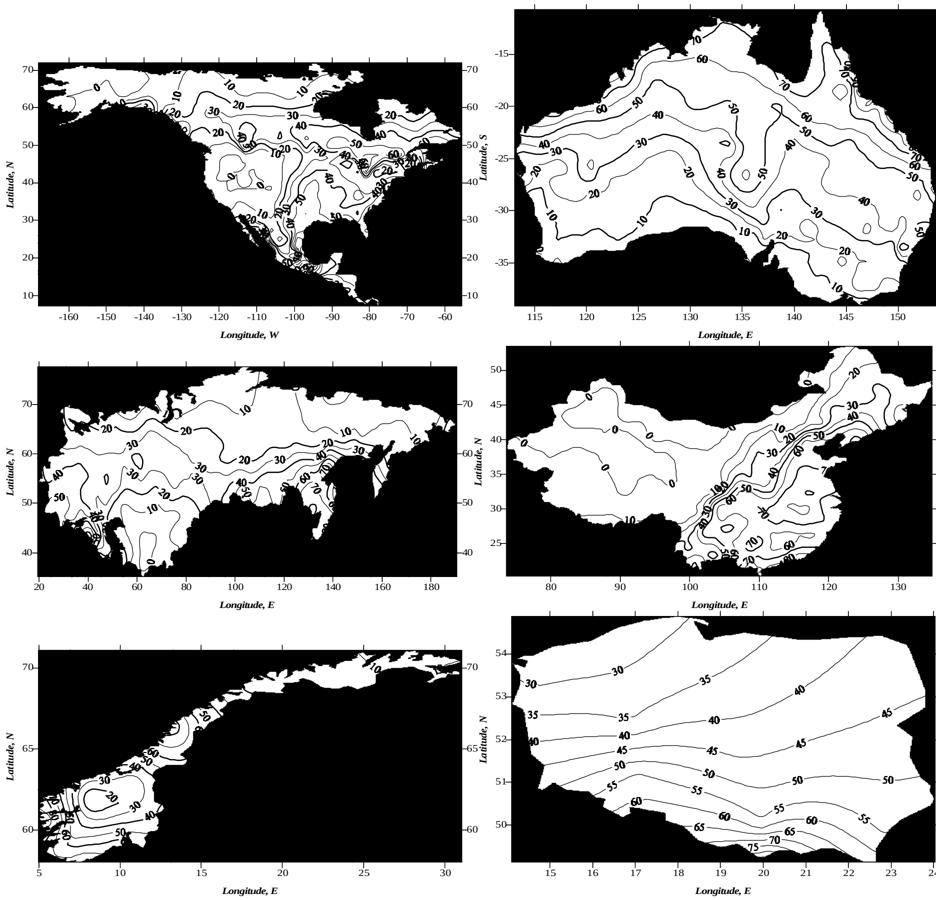

Figure 14. The contribution (percent of total increase) of summer daily precipitation exceeding the heavy rainfall thresholds (defined in text) when the mean daily precipitation increases by 5% assuming that P_{pr} and η do not change (same scenario and thresholds as in Figure 13).

Sinkiang Province it is too arid and there is no heavy precipitation. More than 50% of the scenario-increase in mean summer precipitation over eastern China is contributed by heavy rains (Figure 14). This contribution increases to 70% in tropical parts of southern China.

Hennessy et al. (1998) found a 9% increase in mean summer precipitation from 1910-1995 over Australia. Our scenario of a 5% increase in summer mean precipitation over Australia yields similar results to those for China. We found a 10 to 20% increase in the probability of summer precipitation exceeding 50.8 mm, more pronounced over relatively dry interior regions of the country and less pronounced over tropical coasts (north and northeastern Australia). Over the regions with low summer precipitation (south and southwestern Australia) 50.8 mm is not exceeded so no changes were found. The contribution of heavy rains to the 5% increase in mean precipitation gradually decreases from 70% in the north to 10% in the south of the continent (Figure 14).

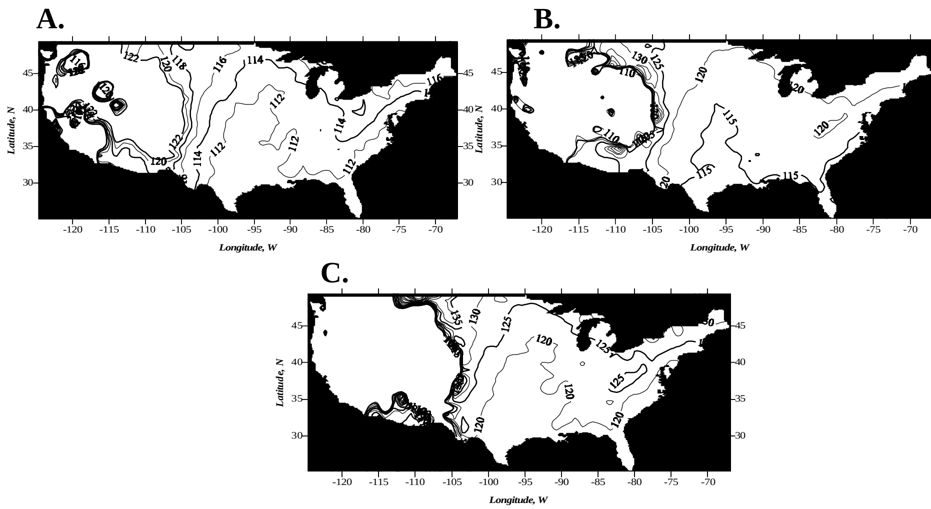

Figure 15. Percentage change of the probability of summer daily precipitation over the United States exceeding (a) 50.8, (b) 76.2, and (c) 101.6 mm thresholds when the mean daily precipitation increases by 7% assuming that P_{pr} is also increased by 5% (to match the estimate in Table II) and η does not change. The change is expressed as a ratio P_{heavy}(scenario)/P_{heavy}(climate).

For all countries, we calculated the change in the contribution of heavy rain events to total precipitation when mean precipitation increases by 5%. This involved calculating the ratio of the percentage of heavy rains associated with a 5% increase to the current percentage shown in Figure 12. The ratio indicates an increase in the contribution of heavy rains to the summer precipitation totals by a factor of 1.1 to 1.2 (i.e., a 10-20% increase). Karl et al. (1995) showed that changes in the proportion of precipitation contributed by heavy extreme precipitation events were increasing relative to the total annual precipitation received in the United States during the past century. Here we show that such a result is also consistent with an increase in mean daily precipitation with no change in the probability of precipitation and the shape parameter. Such a result is not intuitive and implies that not only does more precipitation occur in heavier rain events as precipitation increases, but that the increase is disproportionately larger than the change in the mean.

In order to better match the observed precipitation changes for the United States, Norway, and Australia we should split the observed increase in mean precipitation between increases in intensity and frequency. Below we present our best guess scenarios for extreme precipitation changes in these three countries with changing P_{pr}. In these scenarios we continue to fix η. The following scenarios are considered:

- the contiguous United States: a 7% increase in mean summer precipitation and a 5% increase in P_{pr} (scenario to match "century-long" trends shown in Table II).
- Norway: a 7% increase in mean summer precipitation in the "windward" part of the country (all northern Norway and coastal ocean-front part of southern Norway) and

a 5% increase in the number of rainy days; a 7% decrease in mean summer precipitation in a “leeward” part of the country (interior and southeastern part of southern Norway, south of 62.5°N and east of 6°E) without any changes in precipitation frequency; this scenario is based on the analysis of the summer precipitation changes over the past 60 years (Førland et al., 1996) and the trends in P_{pr}.

- Australia: a 9% increase in mean summer precipitation and a similar increase in P_{pr} (see Table III and Section 2 in Appendix).

Figure 15 shows that the scenario increase in the number of days with summer precipitation, P_{pr}, found in the United States (Table II) in conjunction with a higher increase of the mean summer precipitation does not significantly affect the results reported in Figure 13. We use this scenario to further investigate the changes in probability of extremely heavy rains (above 76.2 mm and 101.6 mm). This extrapolation can be used for practical purposes in hydrological calculations of the consequences of the contemporary climatic change.

Table VII

Scenario of the most-probable changes in the probability of summer heavy rainfall, P_{heavy} above 25.4 mm over Norway derived from the recent (past 60 years) variations in mean precipitation, μ, and frequency, P_{pr}. Average numbers of days with heavy rainfall and their linear trends estimated from the century-long homogeneous time series for period 1901-1996 are also shown. The partition of the country into two regions (windward and leeward) was accomplished by separation into a leeward part of the country the region south of 62.5°N and east of 6°E.

Region	$\Delta\mu$ (%)	ΔP_{pr} (%)	Δ P_{heavy} (%)	Average number of days with heavy rains	Average linear trend, %/10years
Windward part of the country	+7	+5	12	2.9	1.4
Leeward part of the country	-7	0	-26	1.3	-2.0

Table VII shows that when the scenario mean precipitation increase/decrease is accompanied by the change in frequency of precipitation events (as found in Norway), changes in P_{heavy} are less prominent but, nevertheless, still higher than the changes in mean precipitation. Empirical estimates of century-long trends in the number of days with heavy rainfall (the last column in Table VII) support this analysis and the order of magnitude of these trends is consistent with theoretical estimates of changes in P_{heavy}.

When a change in mean summer precipitation is solely due to P_{pr}, a proportional change in the probability of heavy precipitation is expected. Thus, a 9% increase in mean summer precipitation over Australia should give a 9% increase in the probability of heavy rains. Table VI and Figure 10 show that the increase in heavy rainfall over Australia observed during the past century was about 10% for the entire continent and about 45% in the southeast. However, unusually high precipitation variability during the past three decades[****] strongly affects all statistics computed from these time series and makes a linear trend of the all-Australian heavy rainfall statistically insignificant at the 0.05 level. For the coastal region of southeastern Australia (New South Wales and Victoria; Tables III and VI) we found statistically significant increases in μ (3.1%/10yrs), P_{pr} (2.0%/10yrs), and P_{heavy} (4.6%/10yrs). Each of these changes is statistically significant and the 45% increase in heavy rainfall in the southeast is particularly impressive and matches our model estimates of the P_{heavy} increase for the stations in this region.

In Figure 3B we sketch the changes in the probability of heavy rains, P_{heavy}, when the mean precipitation, μ, is changing without changes in the shape parameter, η, but with changes in P_{pr} and λ. It shows a variety of P_{heavy} changes depending upon changes of the ratio of these two parameters. But in our analyses, only *one* combination/realization has been observed in each region where we have sufficient homogeneous precipitation data on a century time-scale (the eastern two-thirds of the United States, coastal regions of southeast Australia, European part of the former USSR, and southern Norway): the changes in P_{pr} are of the same sign and less than the changes in μ by absolute value. This implies (according to our model) that the changes in P_{heavy} will be in the same direction as changes of μ with a higher than linear rate. This is exactly what we have observed in our empirical estimates of P_{heavy}.

The scenarios discussed above have been calculated using gamma distribution parameters of daily precipitation and observed trends in summer country-wide precipitation totals and frequency. There is no guarantee that these trends will continue in the future. However, the purpose of such exercises is twofold:

(1) We show how important parameters of daily precipitation events have contributed to the historical changes in mean seasonal precipitation without actual monitoring the changes in these parameters (which is otherwise arduous and often impossible due to data paucity and inhomogeneity problems); and

(2) If changes in mean precipitation can be predicted (e.g., by climate models), the revealed relationships between mean and extreme precipitation will assist us in the assessment of the hydrologic, ecological, and socio-economic consequences of these changes.

For the United States and eastern Australia, we possess sufficient century-long homogeneous time series of daily precipitation to evaluate the trends in heavy precipitation directly, i.e., without the help of model (3) (Figure 10, Table VI). These data support our conclusions about the century-long disproportionate

[****] The most humid northern part of the continent received a record high intensity and frequency of summer (DJF) heavy rains in 1991 and 1995 years and a record low intensity and frequency of heavy rains in 1983 and 1990 years.

increases in heavy precipitation over these two countries (cf., Karl et al., 1996; Karl and Knight, 1998; and Suppiah and Hennessy, 1996, 1998). To be conclusive, analyses of country-wide trends in extreme precipitation have to be based on a more dense network than similar analyses for mean precipitation, because of the low ratio of signal to noise in the data. For example, high variability due to the small number of stations used in the analysis of Norway and Australian continent heavy precipitation (cf., Figure 10) makes the trend estimates shown in Table VI statistically insignificant. However, the approach used in scenarios shown in Figures 13 through 15 can handle the data paucity problem.

8. Conclusions

A simple statistical model of daily precipitation applied to the data of eight countries shows that the shape parameter of the precipitation distribution remains regionally and temporally stable, the number of days with precipitation remains more or less stable, while the scale parameter is highly variable in time and space. This implies a likelihood that changes in mean monthly precipitation in these countries will be associated with disproportionately large changes in the extremes.

When mean summer precipitation increases by 5%, similar to what has occurred in several regions during the past century, with no change in the number of precipitation days and no change in the shape parameter of the precipitation distribution, there is a 20% increase in the probability of summer daily precipitation over a 25.4 mm threshold in northern countries (Russia, Canada, Norway, and Poland) or a 50.8 mm threshold in mid-latitudes, tropics, and subtropics (the United States, Mexico, China, and Australia). That is, the increase in the probability of “heavy” precipitation is four times the increase in mean precipitation.

Increases in atmospheric water vapor have been documented in North America, China, and a large portion of the tropical Oceans (IPCC, 1996; Ross and Elliot, 1996). Such increases in water vapor suggest an enhanced hydrological cycle. These changes have accompanied a widespread increase in temperature over the last century. Our results complement the above findings and indicate that in a warmer and wetter world, as projected by climate models driven by increasing greenhouse gases, increases in extreme precipitation are likely to be disproportionately large compared to any change in the total precipitation[*****].This is likely to have important socio-economic and ecological impacts. This feature of summer precipitation may already be manifested in

[*****] Various lines of evidence (climate models, observations and theory) indicate that future greenhouse warming will lead to increases in mean rainfall (IPCC, 1996), and our statistical model suggests that this will be accompanied by large increases in heavy rainfall. Moreover, future increases in heavy rainfall derived from our study are consistent with coarse resolution results from global climate models (Schaer et al., 1996; Jones et al., 1997; Hennessy et al., 1997).

recent increases in precipitation extremes over some regions, e.g., the United States and Australia.

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Appendix 1. Some Details Related to Quality Control, Pre-processing, and Parameter Estimation of the Gamma Distribution Parameters of Daily Precipitation Data.

1. Several quality control procedures had been performed with daily precipitation data before they were used in our analyses. The U.S. procedure is described in Hughes et al. (1992), the North American data set pre-processing is described by Easterling et al. (1998). The Australian procedure is described by Lavery et al. (1992, 1997) and Plummer et al. (1997, 1998). The Norwegian procedure is described by Hanssen-Bauer and Førland (1994), the former USSR method is described in Razuvaev et al. (1993), and the Polish procedure is described in Fortuniak (1996). We are not aware of inhomogeneity problems in precipitation time series for Mexico. The original data, however, have passed quality control procedures (Easterling and Peterson, 1995) and numerous small scale (i.e., on a station level) adjustments to the data have been made. Specifically, Mexican data were checked for extreme outliers by comparison with neighboring stations and existing climatologies.

There are no instrumental inhomogeneity problems with precipitation observations in Chinese and Polish stations of international exchange during the period of data availability. Careful analysis revealed some inhomogeneity problems at several Norwegian stations around 1900 which related to the introduction of wind shields on rain gauges or station relocations. Therefore, we use the precipitation time series from these stations only after the wind shields have been installed or after the date of the last relocation.

2. There is a very specific problem with century-long Australian precipitation time series. Sometimes rain gauges were not read for several days, so rainfall records appear as an accumulated total followed by flags indicating the accumulation period and the number of rain-days in that period. A significant number of Australian stations often reports accumulated precipitation values most of which coincide with Mondays and/or days after holidays. On days when accumulations are not reported, all-Australian average summer rainfall is 10% to 15% lower on Sundays than on any other day. This suggests that Sunday rainfall has not been consistently reported especially in the first half of the century. Our estimates show that this effect can affect (increase) the trends in P_{pr} and P_{heavy} by 5 to 10%. Since the 1970s, there has been an increase in the number of rainfall accumulations recorded at many stations. The National Climate Centre (Neil Plummer, Personal communication) believes this reflects a decline in recording of weekend rainfall. Plummer et al. (1997) show a marked decrease in frequently reporting rainfall stations around the early to mid 1970s. At many of the post office stations, this is because Australia Post went from a 5.5 day a week operation to 5 days from 23 Feb. 1974.

Keeping in mind the problems with accumulated rainfall and the need to analyze daily rainfall trends, we selected the following approaches: (a) we treated the accumulated totals as 24-hourly totals and (b) we split the accumulations into daily amounts determined by the reported number of rain-days in each accumulation period. Method (a) was tested by comparison with the trends in three- and seven-day precipitation totals (that should not be affected by accumulation). In this test, we found that the effect of our treatment of accumulation periods on daily trends in P_{pr} and P_{heavy} is minimal. Moreover, the comparison of P_{pr} and P_{heavy} for each day of the week shows that our estimates of P_{pr} and P_{heavy} in Tables III and VI can be inflated only by a few percent due to our treatment of accumulation reports in trend analysis. In method (b) each rain-day in the accumulation period was assigned the value of the accumulated total divided by the number of rain-days. For example, a total of 20 mm accumulated over three days including two rain-days would be replaced by two days of 10 mm and one day of 0 mm. Suppiah and Hennessy (1996) found that calculated rainfall trends were insensitive to the method of replacement of accumulated values. While there are problems and potential biases in both methods, the results presented in Tables III and VI are based on method (a) which is more likely to underestimate mean values of P_{heavy} and P_{pr} but keeps their trends intact.

3. Data for Canada and the former Soviet Union required special attention due to inhomogeneities (Goodison and Louie, 1986; Groisman et al., 1991; Groisman and Easterling, 1994; Metcalfe et al., 1997). In this paper we focus on summer precipitation. Thus, no efforts were made to homogenize cold season daily precipitation time series. An inhomogeneity problem that affects precipitation measurements in all seasons in Russia is the absence of a wetting correction in the data prior to 1966. This was mitigated by introducing this correction into the data prior to 1966 so they would be in agreement with the current observational practice (Struzer, 1975).

The daily Canadian precipitation data were quality controlled by the Canadian Atmospheric Environment Service. These data required an additional adjustment to homogenize the precipitation time series. We did not use the reported precipitation records because they contain an inhomogeneity related to a change in the technique of measuring frozen precipitation in the Canadian primary network during the early 1960s. Instead, to obtain mean daily precipitation, P, we used separate records of rainfall, R, and snowfall, S. The measurements were combined using the formula

$$P=R'+0.1*S, \text{ where}$$

$$R' = 1.02 * (R+0.2 \text{ mm}) \text{ for } R>0 \text{ before 1975 and } R'=R \text{ otherwise.}$$

The conversion R' was used to accommodate the results by Struzer (1975), Metcalfe et al. (1997), Mekis and Hogg (1997), Bogdanova and Mestcherskaya (1998), and our understanding of wetting rainfall losses from the old Canadian rain gauge before the mid-1970s. For further discussion of this issue see (Sevruk, 1982; Gray and Male, 1981; Groisman and Easterling, 1994, Metcalfe et al., 1997; and Groisman and Legates, 1995).

4. Difficulties related to estimation of the shape parameter of the gamma-distribution with limited data are reduced by use of the maximum likelihood estimators. These estimators are considered the best, but for small sample sizes they are biased (Crutcher and Joiner, 1980). Therefore, we considered our estimates valid only for the sample size, N, of more than one hundred rain events. This leaves us with a bias of less than 5% (Crutcher and Joiner, 1980). Usually in humid regions, however, we were able to accumulate samples ten times larger to prevent this problem from affecting our results and conclusions. The following expression for asymptotic variance of the maximum likelihood estimate of the shape parameter, $\hat{\eta}$, was used:

$$Var(\hat{\eta}) = \eta N^{-1} (\eta d^2\Gamma(\eta)/d\eta^2 - 1)^{-1},$$

where N is the sample size and Γ is the gamma function. It was derived from the general formulae for this variance (cf., Kendall and Stuart, 1979). For typical values of the shape parameters that we calculated, an asymptotic standard deviation of our estimates was much less than 4%.

The only difficulties with parameter estimation in our analyses resulted from problems in the low intensity precipitation data. The maximum likelihood estimators of the two parameter gamma distribution (in our case, a conditional distribution of daily precipitation) are functions of two sufficient statistics:

$$S_1 = \Sigma x_i \qquad \text{and} \qquad S_2 = \Pi x_i,$$

where the first is a sum and the second is a product of all sample values. While S_1 is relatively robust to the high and low precipitation values, S_2 is extremely sensitive to them. We are focused on extreme precipitation and the information carried by S_2 is of interest. However, changes in light precipitation also strongly affect this statistic, so if the number of light precipitation events changes due to factors unrelated to weather, this statistic will be severely contaminated. For example, the change in the threshold of measured precipitation with transition from imperial to metric measurements (Australia, Canada), introduction of wetting corrections to the data in an attempt to measure precipitation "up to the last drop" (Russia), and modification of the gauge to "make it more precise" (Canada, Russia) make the daily precipitation time series inhomogeneous.

By adjusting the Russian and Canadian data, we were able to create homogeneous time series of precipitation totals and some other statistics (e.g., number of days with precipitation above 1 mm). However, the time evolution of the shape parameter of the gamma distribution is strongly affected by the S_2- statistic and was not preserved by these adjustments. Therefore, in our estimates of the shape parameter for the former Soviet Union, we use the precipitation data only after 1967, when the last significant change in the instrumentation was introduced (Groisman et al. 1991). For Canada, we avoid assessment of the changes in the shape parameter after 1975, when the new rain gauge was introduced to the Canadian primary network (Metcalfe et al., 1997; Mekis and Hogg, 1997). In our assessment of the trends in the number of days with precipitation over Australia, Russia, and Canada, we considered only the days with precipitation above 1 mm to eliminate/reduce the contribution of the above mentioned inhomogeneities to our conclusions. For Norway, where precipitation has been consistently measured with a 0.1 mm accuracy, we nevertheless consider only the trends in the number of days with precipitation above 0.2 mm. This was done after we had discovered there a 100% increase in the number of days with precipitation equal to 0.1 mm in the 1930s and then found evidence (Bruun, 1949) that this was not a climate-related change but a result of improved observational diligence.

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