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Extreme precipitation and extreme streamflow in the Dongjiang River Basin in southern China

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Abstract

Extreme hydro-meteorological events have become the focus of more and more studies in the last decade. Due to the complexity of the spatial pattern of changes in precipitation processes, it is still hard to establish a clear view of how precipitation has changed and how it will change in the future. In the present study, changes in extreme precipitation and streamflow processes in the Dongjiang River Basin in southern China are investigated. It was shown that little change is observed in annual extreme precipitation in terms of various indices, but some significant changes are found in the precipitation processes on a monthly basis. The result indicates that when detecting climate changes, besides annual indices, seasonal variations in extreme events should be considered as well. Despite of little change in annual extreme precipitation series, significant changes are detected in several annual extreme flood flow and low-flow series, mainly at the stations along the main channel of Dongjiang River, which are affected significantly by the operation of several major reservoirs. The result highlights the importance of evaluating the impacts of human activities in assessing the changes of extreme streamflows. In addition, three non-parametric methods that are not-commonly used by hydro-meteorology community, i.e., Kolmogorov–Smirnov test, Levene's test and quantile test, are introduced and assessed by Monte Carlo simulation in the present study to test for changes in the distribution, variance and the shift of tails of different groups of dataset. Monte Carlo simulation result shows that, while all three methods work well for detecting changes in two groups of data with large data size (e.g., over 200 points in each group) and big difference in distribution parameters (e.g., over 100% increase of scale parameter in Gamma distribution), none of them are powerful enough for small data sets (e.g., less than 100 points) and small distribution parameter difference (e.g., 50% increase of scale parameter in Gamma distribution).

1 Introduction

Extreme meteorological and hydrological events may have huge impacts on human society. With significant global warming, it seems that the occurrence of extreme events gets more frequent, and therefore more and more efforts are put on the research of extreme events in various relevant fields in the last decade.

It is widely conceived that with the increase of temperature, the water cycling process will be speeded up, which in consequence will possibly result in the increase of precipitation amount and intensity. Many outputs from Global climate models (GCMs) indicate the possibility of substantial increases in the frequency and magnitude of extreme daily precipitation (e.g., Gordon et al., 1992Fowler and Hennessy, 1995; Hennessy et al., 1997; McGuffie, 1999). The increase also shows itself in observed data. Karl et al. (1995) found that the contribution to total annual precipitation of 1-day precipitation events exceeding 50.8 mm (2.0 in.) increased from about 9% in the 1910s to about 11% in the 1980s and 1990s. Further on, Karl and Knight (1998) found that the 8% increase in precipitation across the contiguous United States since 1910 is reflected primarily in heavy and extreme daily precipitation events. The results of Kunkel et al. (1999) confirm that the national trend in short duration (1–7 days) extreme precipitation events for the United States is upward at a rate of 3% decade^{-1} for the period 1931–1996. In Australia, much of the country has experienced increases in heavy precipitation events, except in southwestern Australia where there has been a decrease in both rain days and heavy precipitation events (Suppiah and Hennessy 1998; Haylock and Nicholls, 2000). In the United Kingdom increases in heavy wintertime events and decreases in heavy summertime events have been found (Osborn et al., 1999). New et al. (2001) show that, on the basis of gridded observed monthly precipitation data, global land precipitation (excluding Antarctica) has increased by about 9 mm over the twentieth century, and data from a number of countries provide evidence of increased intensity of daily precipitation, generally manifested through increased frequency of wet days and an increased proportion of total precipitation occurring during the heaviest events.

Roy and Balling (2004) found that, in general, evidence exists for an increase in the frequency of extreme precipitation events in India over the period 1910 to 2000. According to the observed data over half of the land area of the globe, there is a widespread increase in the frequency of very heavy precipitation in the mid-latitudes during the past 50 to 100 yr (Groisman et al., 2005). The results of Zhai et al. (2005) indicate that while there is little trend in total precipitation for China as a whole, significant increases in extreme precipitation have been found in western China, the mid-lower reaches of the Yangtze River, and parts of the southwest and southern China coastal areas.

While in many areas the increased intensity of heavy rainfall is observed, in quite a number of other areas and other studies little significant increase is observed. For instance, Nicholls et al. (2000) calculated various indices for monitoring variations in Australian climate extremes, and showed that, most of the trends in the various indices of climate extremes investigated were relatively weak and lacked statistical significance, and no clear trend has emerged in the percentage of Australia in extreme rainfall (drought or wet) conditions, since 1910. Zhang et al. (2001) showed that there has been no long-term trend in the frequency or intensity of extreme precipitation events in Canada during the 20th century. Koning and Franses (2005) show that no statistically significant shift is found in the annual largest values of daily rainfall in the Netherlands over the course of a century, which suggests that the probability of extremely high levels has not changed over time. Zhang et al. (2005) showed that the trends in precipitation indices, including the number of rainy days, the average precipitation intensity, and maximum daily precipitation events in Middle East, are weak in general. Su et al. (2006) analyzed the observed extreme temperature and precipitation trends over Yangtze from 1960 to 2002 on the basis of the daily data from 108 meteorological stations, and found no statistically significant change in heavy rain intensity from a basin-wide point of view, although a significant positive trend was found for the number of days with heavy rainfall (daily rainfall >=50 mm). Klein Tank et al. (2006) find that most regional indices of precipitation extremes show little change between 1961 and 2000 in central and south Asia. Moberg et al. (2006) show that, while winter precipitation totals, averaged over 121 European stations north of 40° N, have increased significantly by 12% per 100 years, trends in 90th, 95th and 98th percentiles of daily winter precipitation have been similar. New et al. (2006), in their study of trends in daily extremes over mainly southern Africa for the period 1961 to 2000, concluded that there are few consistent and statistically significant trends in the precipitation indices that they calculated.

While the evidence for increasing trends appears in many regions, statistically significant decreasing trends in extreme rainfall events have also been found in some areas, including the Sahel region of Nigeria (Tarhule and Woo 1998), southwestern and western Australia (Suppiah and Hennessy 1998; Haylock and Nicholls, 2000), Southeast Asia and parts of the central Pacific (Manton et al., 2001; Griffiths et al., 2003), northern and eastern New Zealand (Salinger and Griffiths, 2001), the UK in summer (Osborn et al., 2000), Poland (Bielec, 2001), and some parts in India (Roy and Balling, 2004). Therefore, the spatial pattern of changes in precipitation is complex and varied over the world.

On the other hand, in the context of significant global changes in many regions, whether or not the streamflow processes has changed is of great concern because streamflow processes are mainly driven by meteorological processes, and possibly more extreme weather may result in higher flood and drought risks. For example, when investigating the relationship of changes in the probability of heavy precipitation and high streamflow over the contiguous United States, Groisman et al. (2001) showed that the variations of high and very high streamflow and heavy and very heavy precipitation are similar. In a recent study, Zhang et al. (2005) evaluated the relations between the temperature, the precipitation and the streamflow during 1951–2002 of the Yangtze River basin, suggesting that the present global warming will intensify the flood hazards in the basin. However, the link between excessive precipitation and hydrologic flooding is affected by several factors, including meteorological factors (such as antecedent precipitation amount and the intensity, duration and spatial pattern of precipitation events), human activities (such as land-use change and dam construction), and basin characteristics (such as the size, topography, control structures, and drainage network of the basin). These factors vary from event to event, from season to season, and from region to region.

The objective of this study is to determine whether the precipitation process, especially the extreme precipitation, in the Dongjiang River Basin in southeastern China has changed in the context of global warming, and whether streamflows, including high flows and low-flows, in the basin have changed as well with the intensified climate change and human intervention. Furthermore, we want to introduce several techniques not commonly used by the hydrology and meteorology community for detecting changes in precipitation and evaluate the reliability of the methods. In Sect. 2, we will briefly describe our study area and the data used. Description of the change detection methods used in the study will be given in Sect. 3. Results for detecting changes in extreme precipitation and streamflow are reported in Sect. 4, followed by some discussions and conclusions in Sects. 5.

2 Study area and data used

Dongjiang River originates in Jiangxi Province in southern China and flows through eastern Guangdong Province, converged into the Pearl River. It has a 562 km long mainstream with a drainage area of 35 240 km^2. The streamflow process of Dongjiang River demonstrates strong seasonality due to a sub-tropical monsoon climate. The Dongjiang River is important for not only local region but also for Hong Kong because about 80% of Hong Kong's water supply comes from Dongjiang River through cross-basin water transfer in recent years. Three major reservoirs (see Fig. 1) were built in the basin, inlcuding Xinfengjiang Reservoir (started to operate in 1959), Fengshuba Reservoir (started to operate in 1973) and Baipenzhu Reservoir (started to operate in 1984).

In the present study, daily precipitation data at 4 meteorological stations and daily streamflow data at 6 hydrological stations are used for the analysis (Fig. 1). The descriptions of all the 10 gauging stations are listed in Table 1. Annual precipitation in the center of the basin (at station 59293) is about 1932 mm, with nearly 80% falling in spring and summer from March to August. As shown in Fig. 1, three (Longchuan, Heyuan and Boluo) of the 6 streamflow gauging stations are significant impacted by reservoir operation, whereas another three are little impacted by any major hydraulic works. Daily discharges were available for 45 to 50 years, whereas the daily precipitation mainly for 52 years. Very few data are missing in these series, and missing data are filled with linear interpolation. For the period 1956 to 2004, the basin daily areal rainfall is estimated from the daily precipitation observed at 4 meteorological stations by using the classical Thiessen polygon method.

3 Methods for detecting climate change

Statistical tools for assessing changes in extremes are varied, and the community has not generally agreed to a "best" approach. In the present study, four non-parametric methods will be applied, including the Mann-Kendall trend test, Kolmogorov–Smirnov distribution test, Levene's variance homogeneity test and quantile test. Descriptions of these methods are given in this section.

3.1 Mann-Kendall trend test

An important task in hydrological modeling is to determine if any trend exists in the data, not only for the purpose of modeling because many models have assumptions of stationarity, but also for detecting possible links between hydrological processes and environment changes. Many methods are available for detecting trend in mean values. Non-parametric trend detection methods are less sensitive to outliers (extremes) than are parametric statistics such as Pearson's correlation coefficient. In addition, nonparametric test can test for a trend in a time series without specifying whether the trend is linear or nonlinear. Therefore, a rank-based nonparametric method, the Mann-

Kendall's test (Kendall, 1938; Mann, 1945), referred to as MK test hereafter, is applied in this study.

Under the null hypothesis H_0, that a series $\{x_1, \ldots, x_N\}$ come from a population where the random variables are independent and identically distributed, the MK test statistic is

$$S = \sum_{i=1}^{N-1} \sum_{j=i+1}^{N} sgn(x_j - x_i) \text{ where } sgn(x) = \begin{cases} +1, & x > 0 \\ 0, & x = 0 \\ -1, & x < 0 \end{cases} \tag{1}$$

And *tau* is estimated as:

$$\tau = \frac{2S}{N(N-1)}. \tag{2}$$

Kendall (1975) showed that the variance of S, $Var(S)$, for the situation where there may be ties (i.e., equal values) in the x values, is given by

$$\sigma_s^2 = \frac{1}{18}\left[N(N-1)(2N+5) - \sum_{i=1}^{m} t_i(t_i-1)(2t_i+5)\right], \tag{3}$$

where, m is the number of tied groups in the data set and t_i is the number of data points in the ith tied group.

Under the null hypothesis, the quantity z defined in the following equation is approximately standard normally distributed:

$$z = \begin{cases} (S-1)/\sigma_s & \text{if } S > 0 \\ 0 & \text{if} S=0 \\ (S+1)/\sigma_s & \text{if } S < 0 \end{cases} \tag{4}$$

At a 0.05 significance level, the null hypothesis of no trend is rejected if $|z|>1.96$.

3.2 Kolmogorov–Smirnov test

The two-sample Kolmogorov–Smirnov (KS) test is one of the most useful and general nonparametric methods for comparing two samples to determine whether two samples follow the same distribution. The KS test is a distribution-free test, which is based on looking at the maximum vertical distance between the empirical distribution functions of two samples. Let n_1 and n_2 be the sizes of the two samples, $n_1 \geq n_2$. The value of the test statistic for the two-sided two-sample Kolmogorov–Smirnov test is

$$T = \sup_x |F_1(x) - F_2(x)| \tag{5}$$

where F_1 and F_2 are the empirical distribution functions based on the two samples. The asymptotic p value for this statistic as $n_1, n_2 \rightarrow \infty$ is given by

$$p = Q\left(T\sqrt{\frac{n_1 n_2}{n_1 + n_2}}\right) \tag{6}$$

where $Q(z)=2\sum_{k=1}^{\infty}(-1)^{k-1}e^{-2k^2z^2}$. Because the above series converges rapidly, Q(z) can be approximated using $Q(z)\approx 2e^{-2z^2}$, or for even greater accuracy, using $Q(z)\approx 2(e^{-2z^2}-e^{-8z^2})$ (Greenwella and Finchb, 2004). Massey (1951) calculated the exact value of p as an alternative to the use of symptotic formula given above when the two sample sizes are equal. Kim and Jennrich (1973) developed a more general algorithm for any two sample sizes, and created tables for various values of n_1 and n_2.

3.3 Test for homogeneity of variance

KS test is designed to detect a shift in the whole distribution of group 1 relative to the distribution of group 2, and it tends to be more sensitive near the center of the distribution than at the tails (Filliben and Heckert, 2006), whereas when detecting changes in extreme events, we are very interested in the variance and the tails of the data, because the variance difference and tail fatness may indicate the difference of the occurrence of extreme events. Therefore, in addition to the KS test, we would apply Levene's test, a test for the homogeneity of variances between different groups, and the quantile test, a test for the shift of the upper tail.

The F-test is most popular to test if the standard deviations of two populations are equal. But F-test is extremely sensitive to the non-normality, so is another commonly used test method Bartlett's test (Bartlett, 1937), while precipitation data almost always violate the normality assumption. Thus, in the present study, we use Levene's test (referred to as L-test hereafter), which is less sensitive than the Bartlett test to departures from normality (Conover et al., 1981; Snedecor and Cochran, 1989, p. 252), to detect whether the variances of k groups are identical.

L-test is based on computing absolute deviations from the group mean within each group. Given a variable Y with sample of size N divided into k subgroups, the L-test statistic is defined as:

$$W = \frac{(N-k)}{(K-1)} \frac{\sum_{i=1}^{k} N_i \left(\bar{Z}_i - \bar{Z}\right)^2}{\sum_{i=1}^{k} \sum_{j=1}^{N_i} \left(z_{ij} - \bar{Z}_i\right)^2} \tag{7}$$

where N_i is the sample size of the ith subgroup; the within-group absolute deviations $z_{ij}=|x_{ij}-\bar{x}_i|$, $i=1,2,\ldots, k$, $j=1,2,\ldots,N_i$, $\bar{x}_i$ is the mean of the ith sub-group; $\bar{Z}_i$ is the group mean of z_{ij}; and $\bar{Z}$ is the overall mean of the z_{ij}.

The L-test rejects the hypothesis of equal variances if

$$W > F(\alpha, k-1, N-k) \tag{8}$$

where $F(\alpha, k-1, N-k)$ is the upper critical value of the F distribution with k–1 and $N-k$ degrees of freedom at a significance level of α.

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3.4 Quantile test

For detecting the changes in extreme events, we are also interested in detecting a difference between two distributions where only a portion (especially the lower tail or upper tail) of the distribution of group 1 is shifted relative to the distribution of group 2. The quantile test (referred to as Q-test hereafter) is a two-sample rank test to detect such a shift (Johnson et al., 1987) based on permuting the ranks of the observations in the tail. The mathematical notation for this kind of shift is

$$F_1(t) = F_2(t) + \varepsilon[F_3(t) - F_2(t)), -\infty < t < \infty \tag{9}$$

where F_1 andF_2 denote cdfs of group 1 and 2; ε denotes a fraction between 0 and 1. Under the null hypothesis, F_1 and F_2 are the same. If the distribution of group 1 is partially shifted to the right of the distribution of group 2, F_3 denotes a cdf such that $F_3(t) \le F_2(t)$, $-\infty < t < \infty$, with a strict inequality for at least one value of t. If the distribution of group 1 is partially shifted to the left of the distribution of group 2, F_3 denotes a cdf such that $F_3(t) \ge F_2(t)$, $-\infty < t < \infty$, with a strict inequality for at least one value of t.

If the alternative hypothesis is that the distribution of group 1 is partially shifted to the right of the distribution of group 2, the test combines the observations, ranks them, and computes k, which is the number of observations from group 1 out of the r largest observations. The test rejects the null hypothesis if k is too large. The p-value is computed as

$$p = \sum_{i=k}^{r} \binom{N-r}{n_1-i} \binom{r}{i} \Big/ \binom{N}{n_1} \tag{10}$$

where n_1 and n_2 are the size of group 1 and group 2, and $N = n_1 + n_2$. The value of r is the smallest rank determined by $r/(N+1) > q$, where q is the target quantile.

4 Results for the precipitation processes and streamflow processes in Dongjiang River Basin

4.1 Extreme hydro-meteorological indices

Considerable efforts have been put on defining the index for evaluating the changes in extreme climate. For instance, Karl et al. (1996) proposed a Climate Extremes Index (CEI) based on an aggregate set of conventional climate indicators which, after two notable modifications in 2003 (http://www.ncdc.noaa.gov/oa/climate/research/cei/cei.html), include the following types of data: 1) monthly maximum and minimum temperature; 2) daily precipitation; 3) monthly Palmer Drought Severity Index (PDSI); 4) landfalling tropical storm and hurricane wind velocity. The Expert Team on Climate Change Detection, Monitoring and Indices (ETCCDMI), which was jointly established by the WMO Commission for Climatology and the Research Programme on Climate Variability and Predictability (CLIVAR), developed 27 climate change indices (Peterson et al., 2001), many of which are widely used in evaluating extreme temperature and precipitation in many studies for Middle East, central Asia, etc. (e.g., Zhang et al., 2005; Klein Tank et al., 2006; Alexander et al., 2006). Similar definitions for extreme climate events are also seen in many other studies (e.g., Nicholls et al., 2000; Frichet al., 2002; STARDEX Project, 2005). In the EMULATE (European and North Atlantic daily to multi-decadal climate variability) project a more detailed 64 climate indices are defined (Moberg et al., 2006).

In the present study, 9 indices defined by ETCCDMI are used. The indices reported on are listed in Table 2. RClimDex, which is developed at the Climate Research Branch of Meteorological Service of Canada, and available from the ETCCDMI Web site (http://cccma.seos.uvic.ca/ETCCDMI), was used for calculating these indices except for CDD. Because RClimDex calculate all indices based on calendar year without considering actual seasonality, which is not suitable to calculate CDD for cases where the dry season crosses two years, therefore CDD was calculated separately based on hydrological year starting from 1 October and ending on 30 September.

The analysis of extreme flood flow events can be conducted with the annual maximum flood (AMF) approach, or the peaks-over-threshold (POT) approach, also called partial duration series approach (PDS) (see Lang et al., 1999). An AMF sample is constructed by extracting from a series of flows the maximum value of each year (annual flood), i.e. only one event per year is retained. Due to its simplicity, the AMF approach is adopted in the present study for analyzing extreme flood events.

In the minimum low-flow analysis, the minimum 7-day low flow is used. The 7-day low-flow index was chosen for three reasons (Chen et al., 2006): (a) The 7-day low-flow is the most widely used index in the USA, UK and many other countries; (b) Previous studies have shown that, compared with 1-day low flow, an analysis based on a time series of 7-day average flows is less sensitive to measurement errors; (c) Dongjiang basin is dominated by a humid sub-tropical monsoon climate, the 7-day low flow better represents the drought conditions of concern and can be used more effectively in water management.

In addition, the timing of annual maximum daily discharge and minimum 7-day average discharge is also analyzed.

4.2 Trend test for annual hydro-meteorological series

The MK test was applied to all the annual precipitation and streamflow series, including annual total/average series and annual extremal series.

It has been found that the positive serial correlation inflates the variance of the MK statistic S and hence increases the possibility of rejecting the null hypothesis of no trend (von Storch, 1995). In order to reduce the impact of serial correlations, it is common to prewhiten the time series by removing the serial correlation from the series through $y_t = x_t - \phi\, x_{t-1}$, where y_t is the prewhitened series value, x_t is the original time series value, and ϕ is the estimated serial correlation coefficient at lag one. However, in our case, none of the data series for detecting trend has significant serial correlation at 5% level, except the minimum 7-day low-flow series at Boluo with a 0.599 lag-one autocorrelation. Therefore, prewhitening is not applied in this study. The results of trend test are listed in Table 3 and Table 4. From Tables 3 and 4 we see that:

1. There is no significant change in either annual total precipitation (PRCPTOT) processes or annual average discharge.

2. It is shown that no trend is present in annual extreme precipitation series in general at 0.05 significance level, except consecutive wet days (CWD) at station 59102.

3. Significant trends are detected in several annual streamflow processes, including: three annual daily maximum flow series at three stations (Longchuan, Heyuan, and Boluo) along the main channel and one at a station (Yuecheng) along a tributary which exhibit significant negative trends; two annual 7-day minimum flow series at two stations (Heyuan and Boluo) along the main channel and another two at stations (Jiuzhou and Yuecheng) along tributaries which exhibit significant positive trends. In addition, the timing of the occurrence of low-flow at the two stations (Heyuan and Boluo) along the main channel gets significantly earlier. For the stations along the main channel, the changes could be explained by the regulation of three major reservoirs. The reason of significant changes in the extreme flows at Yuecheng and Jiuzhou may be a combined effect of land-use/land-cover change and the impacts of small reservoirs.

4.3 Testing changes in precipitation for the periods before and after 1979

As shown by the trend test for various annual indices in Sect. 4.2, no significant trend is present in the annual extreme precipitation series when taking the period from 1950s to early 2000s as a whole. However, it is possible that significant changes occurred in different seasons. On the other hand, it has been found that the climate in China experienced a significant decadal change in the late 1970s (Wang, 1994), which is related to the abrupt change in the large-scale boreal winter circulation pattern over the North Pacific during the late 1970s (Graham, 1994). Dyurgerov and Meier (2000) showed that the time series of change in global glacier volume suggest a significant shift during the late 1970s. Yu and Lin (2002) showed that there is significant difference before and after the late 1970s in terms of the Northern Hemisphere sea level pressure, 500 hPa height and North Pacific sea surface temperature, and such a jump affected the climate of China significantly. Gong and Ho (2002) noticed a significant regime shift in the summer rainfall over the whole eastern China in about 1979. The existence of such a climate shift is also shown in many other research results (e.g., Xu et al., 2005; Li et al., 2006). Therefore, we will investigate evidences of changes in daily precipitation on a monthly basis in two periods, i.e., the period before 31 December 1979, and another after 1 January 1980.

A commonly used technique for checking if the distributions of two data sets are different is to draw a Quantile-Quantile (Q-Q) plot. A Q-Q plot is a plot of the quantiles of the first data set against the quantiles of the second data set. If the two sets come from a population with the same distribution, the points should fall approximately along a 45-degree reference line. The greater the departure from this reference line, the greater the evidence for the conclusion that the two data sets have come from populations with different distributions. To save space, only the Q-Q plots for the daily basin average precipitation for each month in the period 1956–1979 versus 1980–2004 are shown here in Fig. 2.

Q-Q plots give us graphical evidences indicating significant changes in the upper part of the probability distribution in many months, e.g., increase in heavy rainfall in January, February, March, April and July, decrease in heavy rainfall in June and October. Q-Q plots also indicate changes in many months for the precipitation observed at all the 4 meteorological stations, but the results are not in good agreement with each other.

To verify the heuristic results from Q-Q plots, we need some formal statistical tests. One way to do so is to fit a probability distribution model to the data set of each period, and then compare the fitted models to examine if anything has changed. Groisman et al. (1999) used a gamma distribution based model of daily precipitation to investigate the changes in the probability of summer heavy precipitation from eight countries.

They showed that the shape parameter of this distribution remains relatively stable, while the scale parameter is most variable spatially and temporally. Tromel and Schonwiese (2007) detected probability change of extreme precipitation by comparing the stationary Gumbel distribution of different periods of time fitted to observed 100-year monthly total precipitation.

However, on one hand, uncertainty in estimating the parameters of a distribution function may considerably affect results for assessing changes, especially for short data sets, although some methods, for instance, the L-moments (Hosking and Wallis, 1997), may improve the estimation for small dataset; on the other hand, statistical tests are not applicable for testing the significance of differences among estimated parameters based on two groups of dataset. Therefore, instead of detecting the changes in the parameters of fitted probability distribution models, we apply statistical tests, including the KS test, L-test and Q-test (for the upper tail with q=0.95), directly to the observed data for the data sets of the two periods. The calculation is conducted with software package *EnvironmentalStats for S-PLUS* (Millard, and Neerchal, 2001), which is an add-on module to the statistical software package S-PLUS. The results are reported in Table 5.

According to the test results, only the overall distribution of the rainfall in September at station 59096 and October at 59102 changed significantly. There are significant changes in variances for several months at each station, but the months are in not in good agreement among the stations. And, the rightward shift of upper tail is detected in several months at two stations (59096 and 59293), but the months are not in good agreement either between the two stations. As for the mean areal precipitation, significant changes in variance are detected for rainfall in March, June, July, August and October, but significant right-ward shift of upper tails is only found in March.

The statistical test results seem to be more or less different from what we see from the Q-Q plots. For instance, while the QQ-plot for areal precipitation in October shown in Fig. 2 indicates significant change in the distribution, the test result in Table 5 indicates no change by KS test. Are the statistical test results reliable? We therefore make a Monte Carlo simulation to evaluate the effectiveness of these test methods.

4.4 Evaluation of test methods for detecting changes in precipitation

To evaluate the methods for detecting changes in precipitation, we only consider Gamma distributed variables because rainfall processes are normally considered to follow Gamma distribution (e.g., Groisman et al., 1999; Liao et al., 2004) with a probability density function in the form of

$$f(x) = \frac{1}{\Gamma(\alpha)\beta^{\alpha}} x^{\alpha-1} e^{-x/\beta} \tag{11}$$

where α is the shape parameter and β the scale parameter. Groisman et al. (1999) estimated that the scale parameter β changes by an order of magnitude from subarctic regions and deserts (1/0.3) to humid tropics (~1/0.03), and the shape parameter α has little spatial variation, which may vary from 0.5 up to 1.2. Liao et al. (2004) showed that for rainfall data in most areas of China, $\alpha \in (0.3, 0.5)$, $\beta \in (2, 40)$. Therefore, in our simulation experiment, we concentrate on α=0.5 and β=10~40. The plots of distribution functions with α=0.5 and β=10, 40 are shown in Fig. 3, and the 0.99 and 0.999 quantiles for probability for each distribution are listed in Table 6. Obviously, the larger the value of β, the more extreme of the distribution, and the quantile corresponding to a given probability increases in a rate equal to the rate of increase in the scale parameter.

Now we investigate the robustness and power of the three test methods applied in our study in detecting the changes when the variable changes from a distribution with a low value of β to a distribution with a higher value of β, namely, a more extreme distribution. By robustness, we mean the ability of the test to not falsely detect changes when the underlying data are in fact distributed equally. By power, we mean the ability of the test to detect changes when the distribution indeed changes.

We make 10 000 simulations for Gamma distributed samples with fixed value of α = 0.5, varied values of β=10, 15, 20, 30, 40, and varied length of data size L=50, 100, 200, and 300. In each simulation two groups of data are generated, with one group of length L generated with low value of β=b1, and another group of the same length L but higher value of β=b2 (b2≥b1). The Monte Carlo simulations are repeated for b1=10 and 20, b2=10, 15, 20, 30 and 40 (b2≥b1), and L=50, 100, 200, 300. The results are reported in Table 7. The simulation results in Table 7 show that:

1. While the robustness of the tests has little dependence on data size, the power of all the tests is closely related to the data size and depends on the magnitude of change (in terms of variance ratio).

2. The KS test and especially Q-test are quite robust, with a wrong rejection rate less than 0.05 at 5% significance level mostly. But the L-test is not robust, with a wrong rejection rate of around 18% at 5% significance level for all cases where variance ratio = 1. Therefore, a rejection by L-test alone does not give a reliable evidence of change, whereas a rejection by KS test or Q-test is a good evidence of the presence of change.

3. In case of a sharp 100% increase of the scale parameter β changing from 10 to 20 or 20 to 40, while it is not possible to detect the change between two groups of data set with 50 points each, generally, all the three tests are powerful enough to correctly detect the change for large data sets (200 or 300 points) (over 98% correct rejection of null hypothesis). But for smaller changes, such as a 50% increase of changing from 20 to 30, the rate of correct rejection of null hypothesis is low, even for large data sets (e.g., 70.31% for data set with 300 points). Unfortunately, In real world, we often lack enough data, and as shown by Groisman et al. (1999), the change percentage in scale parameter is normally less than 50%, and seldom over 100% for most areas.

From the above analysis, we know that, the good news is, if the null hypothesis of no change is rejected, it is a good indication of change, whereas the bad news is, if the null hypothesis is accepted, we are still not sure if or not there is any significant change present. By revisiting the analysis in the Sect. 4.3, we see that the rainy days for the months from March to September are mostly over 300, while for the months from October to February over 90. Thus the test results for months from March to September should be reliable in general, especially if sharp changes occur, whereas less reliable for the months from October to February. Therefore, from Table 5 we know that, changes indeed occur in some months in the Dongjiang Basin, but the changes are not uniform at different locations in the region. On the other hand, because all the test methods for mediate changes, such as a 50% change in scale parameter, are not powerful enough, we cannot conclude that no change occurs in other months which seem to have apparently experienced change according to QQ-plots in Fig. 2. Consequently, we suggest combining the use of QQ-plot method and statistical test methods to detect changes in extreme events, but the combined use of these methods are still not conclusive.

5 Discussions and conclusions

1. Little change is observed in various annual extreme precipitation indices. Although it is widely believed that it is physically reasonable to expect increases in extreme precipitation if and when significant warming occurs, it is virtually certain that such changes will not be uniform across the globe (Kunkel, 2003), and, according to the literature, it seems that, from the year's point of view, there are no less regions (including the present study area) all over the world seen little changes in annual extreme precipitation events than regions experienced significant changes.

2. On the other hand, significant changes are observed in the precipitation processes on a monthly basis, although the seasonal variations are not uniform even in a medium-size basin such as the Dongjiang River Basin. This is probably because extreme events at a specific location depend not just on the moisture availability and thermodynamic instability, but also on other factors, primarily the fre-

quency and intensity of precipitation-producing meteorological systems (Kunkel, 2003), and the activeness of these systems is seasonally varied. To get statistically significant results in detecting changes, we need more robust statistical test methods, and may need some indices that take the changes in seasonality into account as well. In fact, seasonality has been considered in the calculation of the 64 climate indices in the EMULATE project (Moberg et al., 2006).

3. Despite of little change in extreme precipitation, significant changes are detected at all the three stations along the main river channel, i.e., Longchuan, Heyuan and Boluo. All of the three show significant negative trends in the annual maximum flow, and two of them (Heyuan and Boluo) exhibit significant positive trend in minimum 7-day low-flow. Among three streamflow series observed at tributary stations with medium-size drainage areas and no intervention by major reservoirs, one (Yuecheng) shows significant negative trend in annual maximum flows, and two (Jiuzhou and Yuecheng) show significant positive trend in minimum 7-day low-flows. The changes in annual extremal streamflows at the three stations along the main river channel are obviously due to the operation of several major reservoirs in the basin, whereas the changes at tributary stations are possibly due to land use change and/or operation of small reservoirs. The results indicate that, in the case of little precipitation changes, the operation of major reservoirs is most influential on the extreme streamflow events, whereas land-use/land-cover changes may have secondary impacts. It is common in many studies to examine if extreme high or low flows are associated with climate change or land-use/land-cover change (e.g., Tu et al., 2005; George, 2007; de Wit et al., 2007). But when there are major reservoirs present, in assessing the impacts of environment changes on streamflow processes, especially flood events, how the reservoirs are operated should be considered first.

4. It is expected that with the global warming, the water cycling will be sped up, and the enhanced water cycling will include the increase of evaporation and precip-

itation. But in fact, in many regions of the world the observed pan evaporation is decreasing (e.g., Peterson and Groisman, 1995; Chattopadhyay and Hulme, 1997; Roderick and Farquhar, 2002; Liu et al., 2004), which is considered as a "paradox". A significant decreasing trend is also observed in the pan evaporation processes in most parts (including the present study area) of China (Liu et al., 2004; Ren and Guo, 2006). It has been demonstrated by some researchers that the actual evaporation is negatively related to pan evaporation (Brutsaert and Parlange, 1998; Lawrimore and Peterson, 2000; Golubev et al., 2001). Whether the actual evaporation has increased with the decrease of pan evaporation for the case of China, specifically for the case of Dongjiang River Basin, is an open question. If it is true, still we have a problem that, with increased evaporation, no significant change is detected in annual total precipitation and annual runoff, and the amplitude of extreme precipitation has not changed much either. Runoff may be affected by the changes of water abstraction for industry and agriculture use (especially irrigation), and the increase/decrease of forest coverage which leads to increased/decreased plant transpiration, because the establishment of forest cover would result in increased transpiration and therefore decreases water yield (e.g., Hibbert, 1967; Bosch and Hewlett, 1982). Therefore, how the land cover has changed and how human activities affects the streamflow process in this area will be the subjects of a future study.

5. In detecting changes in extreme hydro-meteorological events, two approaches are commonly seen in literature, i.e., testing trend in annual mean or extremal series, and comparing probability distribution parameters for data observed during different periods. However, by using trend test, we cannot find changes in overall statistical property, while by comparing distribution parameters, the result is subject to parameter estimation uncertainties and statistical tests are hard to be applied to test for the difference in parameters. Therefore, in the present study, besides the Mann-Kendall trend test which has been widely used in the hydrology community, three more non-parametric methods, i.e., Kolmogorov-Smirnov test,

Levene's test and quantile test, are introduced to test for changes in the distribution, variance and the shift of tails of different groups of data. While all three methods work well for detecting changes in two groups of data with large data size (e.g., over 200 points in each group) and big difference in distribution parameters (e.g., over 100% increase of scale parameter in Gamma distribution), none of them are powerful enough for small data sets (e.g., less than 100) and small distribution parameter difference (e.g., 50% increase of scale parameter in Gamma distribution). Unfortunately, small dataset sizes and small distribution parameter changes are common in real world applications. Therefore, statistical testing methods of better performance are needed in detecting changes in extreme hydro-meteorological events, and graphical exploratory methods, such as Quantile-Qantile plots, are recommended to be used in combination with present statistical test methods.

6. Caution must be taken when prewhitening a series before conducting Mann-Kendall trend test, because removal of positive AR(1) from time series by prewhitening will remove a portion of trend and hence reduces the possibility of rejecting the null hypothesis while it might be false (Yue and Wang, 2002); on the other hand, when the change in a real-world process has its physical background, the detected trend cannot be ignored even if it is possibly resulted from a significant serial correlation. For instance, in the case of minimum 7-day low-flow series at Jiuzhou, there is a weak autocorrelation 0.223 at lag one which is not significant at 0.05 significance level. When the series is not prewhitened, a positive trend could be detected at a 0.05 significance level, but no trend would be detected after prewhitening. Similar is the case of annual maximum flow at Yuecheng. In another case of minimum 7-day low-flow series at Boluo, the autocorrelation at lag one is 0.599. If the series is prewhitened, the positive trend is not significant at a 5% level. But the positive trend, we believe, has its physical basis because three major reservoirs, whose major effects are lowering peak flows and increasing low flows, were built in the end of 1950s', the beginning of 1970s' and

early 1980s', which regulated streamflow significantly. Therefore, when there is a sound physical basis for the changes in a natural process, we suggest that the original series, rather than the prewhitened series, should be used for detecting the trend.

7. Before fitting distribution models to a sample precipitation or streamflow data series, it would be wise to investigate the stationarity first, not only the trending behaviour in mean value but also the behaviour in variance and even higher moments. The regulation of reservoir outflows and impacts of land-use/land-cover changes have made many streamflow processes exhibit significant changes, which make the flood frequency analysis more tricky than for stationary cases. If no consideration is given to the nonstationary situations in the flood and low-flow frequency analysis (e.g., Chen et al., 2006), the results may be biased. Techniques of food frequency analysis for nonstationary situations (see Khaliq et al., 2006) will get more and more popularity in the future research.

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Table 1. Meteorological and hydrological Gauging stations.

Station type	Station	Latitude	longitude	Elevation (m)	Drainage area (km^2)	Period	Length (year)
Meteorology	59096	114°29′	24°22′	214.5	–	1953–2004	52
	59102	115°39	24°57	303.9	–	1956–2004	49
	59293	114°41′	23°44′	40.8	–	1953–2004	52
	59298	114°25′	23°05′	22	–	1953–2004	52
Hydrology	Jiuzhou	114°59′	23°07′	–	385	1959–2005	47
	Yuecheng	114°16′	24°06′	–	531	1960–2005	46
	Lantang	114°56′	23°26′	–	1080	1958–2005	48
	Longchuan	115°15′	24°07′	–	7699	1952–2002	51
	Heyuan	114°42′	23°44′	–	15 750	1951–2002	52
	Boluo	114°18′	23°10′	–	25 325	1953–2002	50

Table 2. Extreme precipitation indices used in this study.

Index	Description	Unit
CDD	Annual maximum number of consecutive days with RR <1 mm	Days
CWD	Annual maximum number of consecutive days with RR ≥1 mm	Days
R20mm	Annual count of days when RR≥ 20mm	Days
R50mm	Annual number of days when RR ≥50 mm	Days
RX1day	Annual maximum precipitation in 1 day	mm
RX5day	Annual maximum precipitation in 5 consecutive days	mm
PRCPTOT	Annual total precipitation from wet days (RR≥ 1 mm)	mm
SDII	Simple pricipitation intensity index, average daily precipitation amount on wet days with RR ≥1 mm	mm/day

Note: RR denotes daily precipitation amount.

Table 3. Mann-Kendall trend tests on annual precipitation series.

Annual series	MK test	59096	59102	59293	59298	Areal
RX1day	tau	0.0287	–0.1108	–0.0309	–0.0136	–0.048
	p-value	0.7703	0.2704	0.7523	0.8933	0.635
RX5day	tau	0.0747	–0.0940	–0.0641	–0.0762	–0.146
	p-value	0.4393	0.3507	0.5074	0.4300	0.140
CDD	tau	–0.0706	0.0195	0.0157	–0.0847	–0.045
	p-value	0.4690	0.8516	0.8773	0.3842	0.656
CWD	tau	–0.1704	–0.2961	–0.1342	–0.0694	–0.167
	p-value	0.0739	**0.0026**	0.1580	0.4666	0.091
PRCPTOT	tau	–0.0136	–0.0656	0.0271	–0.0106	–0.021
	p-value	0.8933	0.5165	0.7824	0.9183	0.836
R20mm	tau	–0.0890	–0.0745	0.0920	0.1109	–0.052
	p-value	0.3548	0.4593	0.3387	0.2479	0.604
R50mm	tau	0.0837	–0.1507	0.0845	–0.0739	–0.050
	p-value	0.3813	0.1267	0.3785	0.4387	0.614
SDII	tau	0.0543	–0.0895	0.0551	0.0173	–0.020
	p-value	0.5749	0.3741	0.5698	0.8621	0.849

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Table 4. Mann-Kendall trend tests on annual discharge series.

Series	MK test	Longchuan	Heyuan	Boluo	Jiuzhou	Lantang	Yuecheng
Annual average discharge	tau	0.0573	0.0905	0.0748	–0.0361	–0.0559	–0.0667
	p-value	0.5587	0.3477	0.4533	0.7274	0.5815	0.5196
Annual maximum	tau	–0.3780	–0.4472	–0.2449	–0.1203	–0.1605	–0.2271
	p-value	**0.0001**	**0.0000**	**0.0133**	0.2368	0.1096	**0.0267**
Annual 7-day minimum	tau	0.0918	0.3394	0.4405	0.2081	–0.1118	0.3527
	p-value	0.3461	**0.0004**	**0.0000**	**0.0400**	0.2553	**0.0006**
Timing of annual maximum	tau	–0.0243	–0.0400	0.1216	–0.0130	-0.0399	0.1092
	p-value	0.8074	0.6815	0.2207	0.9051	0.6957	0.2888
Timing of annual 7-day minimum	tau	–0.1671	–0.3167	–0.3010	–0.0740	0.0743	–0.1691
	p-value	0.0850	**0.0009**	**0.0023**	0.4686	0.4515	0.0993

Table 5. P-values for changes in statistical properties of daily rainfall in each month for the periods before and after 1979.

Station	Test	Jan	Feb	March	April	May	June	July	Aug	Sep	Oct	Nov	Dec
59096	KS-test	0.354	0.168	0.680	0.792	0.658	0.128	0.092	0.196	**0.042**	0.076	0.262	0.377
	L-test	0.130	**0.005**	**0.005**	**0.039**	0.446	0.142	0.215	0.365	**0.001**	**0.001**	0.833	0.996
	Q-test	0.121	**0.008**	0.100	**0.010**	0.790	0.770	0.464	0.164	**0.027**	0.989	0.882	0.434
59102	KS-test	0.338	0.126	0.377	0.830	0.059	**0.027**	0.591	0.992	0.426	**0.024**	0.905	0.331
	L-test	0.160	0.251	**0.001**	0.059	0.298	**0.009**	0.135	0.694	0.601	0.155	0.474	0.181
	Q-test	0.131	0.623	0.086	0.230	0.557	0.895	0.446	0.086	0.561	0.916	0.442	0.744
59293	KS-test	0.988	0.963	0.359	0.916	0.608	0.644	0.169	0.167	0.198	0.336	0.723	0.718
	L-test	0.142	0.535	**0.000**	0.538	0.795	0.160	0.011	**0.000**	0.166	**0.037**	0.280	0.950
	Q-test	0.225	0.136	**0.002**	0.286	0.511	0.495	0.171	**0.012**	0.231	0.960	0.271	0.361
59298	KS-test	0.558	0.155	0.332	0.767	0.692	0.999	0.334	0.373	0.169	0.679	0.155	0.373
	L-test	0.422	**0.033**	0.090	0.141	0.330	0.458	0.964	0.098	0.862	**0.037**	0.202	0.968
	Q-test	0.558	0.185	0.148	0.290	0.946	0.722	0.596	0.431	0.616	0.973	0.938	0.211
Areal	KS-test	0.999	0.677	0.274	0.265	0.781	0.074	0.530	**0.010**	0.813	0.130	0.108	0.854
	L-test	0.118	0.098	**0.000**	0.416	0.647	**0.033**	**0.024**	**0.041**	0.497	**0.000**	0.827	0.717
	Q-test	0.521	0.302	**0.022**	0.361	0.802	0.915	0.081	0.019	0.192	1.000	0.716	0.405

Note: Significance level = 0.05. The alternative hypothesis of Q-test is that the distribution of data after 1980 is partially shifted to the right of the distribution of data before 1979, and the target quantile is q=0.95.

Table 6. Quantiles of Gamma distributions with different values of β.

Probability	β=10	β=15	β=20	β=30	β=40
0.99	33.2	49.8	66.3	99.5	132.7
0.999	54.1	81.2	108.3	162.4	216.6

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Table 7. Rejection rate (in percentage) of null hypothesis for testing changes in two groups of data with three test methods.

Size of each group	B1	B2	Variance ratio	KS test	L-test	Q-test (q=0.95)
50	10	10	1	4.08	18.51	3.16
		15	2.25	13.93	48.88	18.65
		20	4	36.55	80.77	42.52
		30	9	76.64	98.21	77.96
		40	16	93.01	99.86	92.82
	20	20	1	3.96	17.86	2.82
		30	2.25	13.9	49.17	18.71
		40	4	36.95	81.15	42.79
100	10	10	1	3.78	17.58	0.85
		15	2.25	25.05	69.22	17.7
		20	4	65.73	96.48	53.97
		30	9	97.38	99.99	92.77
		40	16	99.9	100	99.19
	20	20	1	3.59	17.98	0.82
		30	2.25	24.65	69.6	18.5
		40	4	65.7	96.55	53.53
200	10	10	1	5.4	17.51	1.95
		15	2.25	54.8	90.32	50.57
		20	4	95.85	99.9	92.58
		30	9	100	100	99.97
		40	16	100	100	100
	20	20	1	5.44	17.43	1.68
		30	2.25	55.66	90.37	50.83
		40	4	95.71	99.9	92.56
300	10	10	1	5.12	17.36	4.39
		15	2.25	73.75	96.82	81.39
		20	4	99.47	100	99.73
		30	9	100	100	100
		40	16	100	100	100
	20	20	1	5.12	17.97	4.52
		30	2.25	72.9	96.9	81.75
		40	4	99.52	100	99.62

Note: The null hypotheses of all three methods are no change. Significance level 0.05. B1 and B2 are respectively the scale parameters of the first and second group of simulated gamma distributed data with the same shape parameter 0.5. The variance ratio is the ratio of the variance of the first dataset over the second dataset.

Fig. 1. Location of the study area (left) and locations of gauging stations (right).

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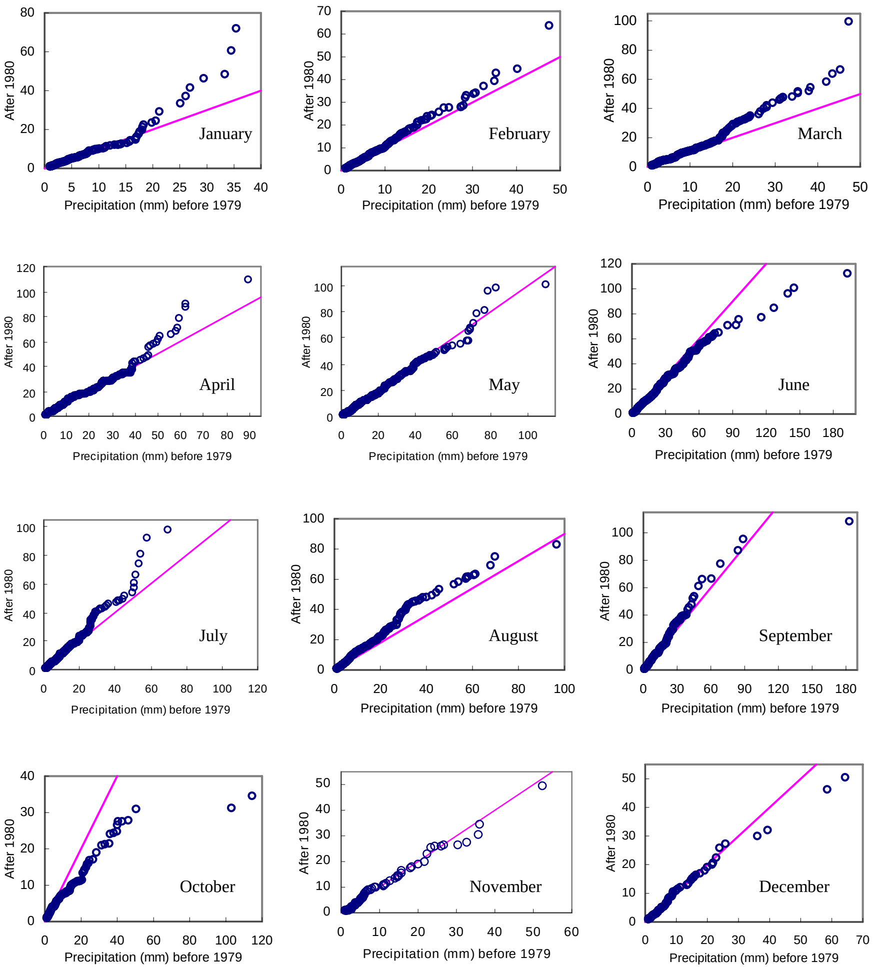

Fig. 2. Q-Q plots for the daily basin average precipitation for each month for periods before 31 December 1979 and after 1 January 1980.

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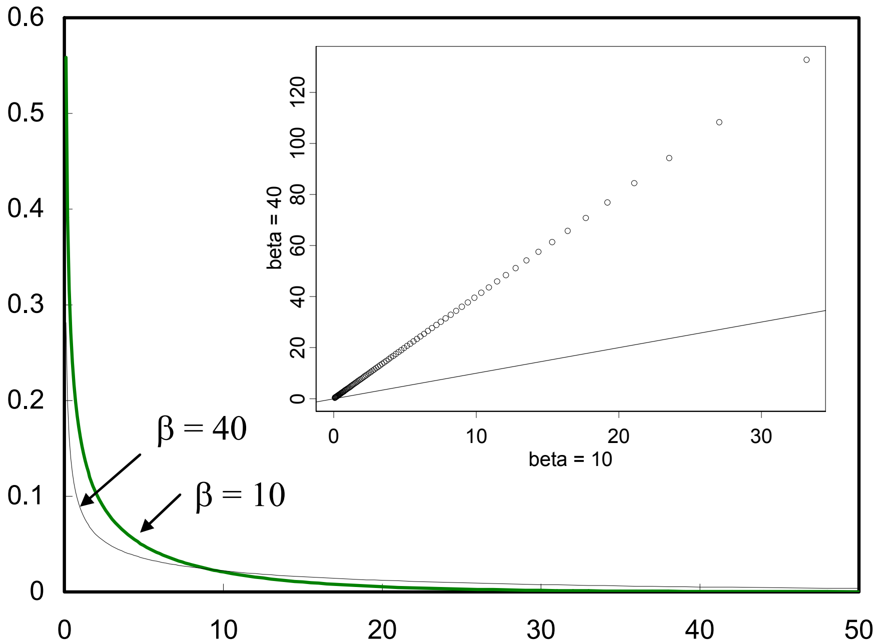

Fig. 3. Two Gamma densities with scale parameter β=10 and 40, and shape parameter α=0.5. (Note: The embedded figure is the Q-Q plot for the two Gamma distributions)

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