TRENDS IN PRECIPITATION VARIABILITY: PRAGUE (THE CZECH REPUBLIC)

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Abstract. Variability in precipitation at scales from 1 to 10 days was investigated with the use of the time series measured at Prague-Sporilov (the Czech Republic) between 1994 and 2001. Variability was detected by the method of absolute difference in precipitation between two adjacent discrete time periods. The results indicated a general increase in precipitation variability at all investigated scales within the 8-yr observational period. The variability patterns also showed quasi-seasonal variations. The summer wetter season proved to be the most variable. The analysis was accomplished by the investigations of precipitation variability at a monthly scale based on a century-long historical time record. On a longer time perspective, precipitation variability exhibits a general increase interrupted by quasi-decadal oscillations. The range of quasi-decadal variability has become more pronounced after about 1950, the fact that hints the possibility of further intensification of the hydrologic cycle. An obtained significant correlation between the North Atlantic Oscillation (NAO) activity and precipitation variability implies that the NAO may account for a large fraction of precipitation variability. Higher NAO-index values tend to be associated with low variability. The variability investigations may have a certain implication for climate change assessments both at the local scales as well as associated with the build-up of greenhouse gases.

1. Introduction

Up to recent years the scientific attention regarding climate change has been oriented towards the effects of changes in climatic averages. However, the climate is variable on all time scales, and for better understanding of the nature of the climate changes an attention is to be paid not only to the course of the mean climate characteristics, but also to the changes in climate variability and climate extremes. The importance of including the variability characteristics into the climate change studies was demonstrated in several works (Katz and Brown, 1992; Rebetez, 1996; Wilks and Riha, 1996; and the references therein). For example, in the face of likely climate change produced by increased greenhouse gas concentration in the Earth's atmosphere, there is strong evidence that changes may occur not only in climatic mean state but also in their higher order moments. General Circulation Models (GCMs) associated with the build-up of greenhouse gases predict not only an increase in temperatures but also a possible decrease in temperature variability (McGuffie et al., 1999; Karl et al., 1999). Moreover, as shown in some recent works, an impact on climate change would result from changes in climate variability or

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from the extreme event occurrence rather than from an increase in the mean temperature itself (Houghton et al., 1996). Even relatively small changes in the means and in variations of climate variables can induce a considerable change in the variability and/or in the severity of extreme events (Hennessy and Pittock, 1995; Colombo et al., 1999).

Numerous recent investigations of the surface air temperature (SAT) variability in fact revealed definite decreasing variability trends in SAT records (Karl et al., 1995; Moberg et al., 2000; Rebetez, 2001; Bodri and Cermak, 2003). Thus, the investigations of variability likewise the investigations of warming trends can be further used for the validation of the simulated models for various scenarios of greenhouse-gas emission and land use. The situation is even more noticeable with additional associated climatic variables, such as precipitation, solar radiation, etc., that can also be predicted by the GCMs. Major changes in the hydrological cycle (e.g., in evaporation and precipitation rates) are likely to depend on the investigated spatial domain. Nevertheless, some conclusions are reasonably well established. Hulme et al. (1998) have investigated the sensitivity of the global precipitation to the global warming. The precipitation sensitivity (standard linear regression fit between unfiltered precipitation and temperature anomalies) for land areas simulated by a set of eight HadCM2 models with different forcing performed at the Hadley Centre (Mitchell and Jones, 1997) achieved 1.5–2.5%/K over the period 1900–1996. The observed precipitation sensitivity for global land areas over the same period was calculated to be 2.37%/K, thus coincide well with predicted values. Given the global warming in the last century between 0.3–0.6 K (Nicholls et al., 1996), we could expect an increase of observational global mean precipitation between 0.7% and 1.4%, such amount being small relative to interannual and interdecadal variability.

Houghton et al. (1992) summarised model results for conditions under doubled carbon dioxide concentration and concluded that there is some indication of increase of the precipitation rate in selected regions that could be accompanied by the simultaneous increase in daily variability of the precipitation. The simulations of the 21st century climate performed by Kattenberg et al. (1996) used GCMs forced with increasing atmospheric concentrations of greenhouse gases and indicated an increase in the intensity of the hydrological cycle as global temperature increases, implying increase in total precipitation, in the number of wet days and the days with extreme precipitation. These conclusions were further specified by Mearns et al. (1995). Investigations of $2 \times CO_2$ influence conducted for different regions of the United States with a regional climate model (RegCM) nested in a GCM indicated a significant increase of variability in daily precipitation intensity (with or without changes in median intensity). In vast regions there are larger and more significant changes in variability rather than in medians. The most recent model simulations (Wetherald and Manabe, 2002) studied the change in land surface hydrology that accompanies the global warming and tried to extract the forced response from the natural precipitation variability. The authors used combined ocean-atmosphere-land model forced by the increase of greenhouse gases and sulphate aerosols in the atmosphere, which are based upon the IS92a scenario. The experiment showed a systematic increase by 5.2% of global mean rates of precipitation above the pre-industrial level by the middle of the 21st century. The calculated signal is significant only on a global scale. Similar systematic changes were obtained also on local scales, e.g. an increase in precipitation rate in Central Europe. However, the large natural variability in local precipitation is usually comparable to or larger than the radiatively forced component. Thus, the forced systematic change of local precipitation generally could not be understood without simultaneous investigations of variability. The existing projections of climate change outputs produced by GCMs for the Czech Republic (area under investigation) generally consist of the estimates of the mean values of the variables. Together with the lack of accuracy at smaller scales, this fact creates difficulties in the way of direct assessment of the variability changes (e.g. Kalvová and Nemešová, 1997). However, the use of indirect variability related precipitation characteristics also hints the possibilities of the future change in precipitation variability.

There are only a few investigations of precipitation variability based on measured data. Karl and Knight (1998) analysed daily precipitation data from the U.S. for the period 1910–1996 and found an approx. 10% increase in annual precipitation. Such increase in most cases occurred because of a greater number of rainy days, with the greatest contribution arising from large number of extreme precipitation events, indicating increased precipitation variability. The most recent investigations of precipitation trends and variability of the 20th century by New et al. (2001) based on the grid data sets revealed secular increasing trends in different domains, for example 8.9 and/or 41.6 mm/100 yr for global data set and in the mid-latitudes (40–60°N) in the Northern Hemisphere, respectively, in many regions accompanied by the increasing wet spells frequency. The above authors concluded, that over much of the global land area (including also Europe) there was an increase in the intensity of precipitation events on scales from 1 day to 3 months. The possibility of detectable alterations in precipitation variability with no and/or imperceptible changes in its mean characteristics is particularly an interesting problem and hints the necessity to include the higher order moments in the precipitation change assessments. On the other hand, the potential response of the socio-economic fabrics of the global community to the changes in climate variability may be stronger than the response to the changes in climatic averages (Rebetez, 1996; Wilks and Riha, 1996). These changes may be completely obscured when examining only the evolution of mean characteristics.

In our paper we analysed an 8-yr long daily precipitation time series to understand the pattern of its variability variations. The variability was investigated for the time intervals within the range of 1 to 10 days. Obtained high frequency variability patterns were compared with the variability of historical precipitation time series on the monthly scale of aggregation. As shown in several works (e.g. Hurrell, 1995), the North Atlantic Oscillation (NAO) accounts for a large fraction of the climate changes in Europe, thus, particular attention has been paid to the potential effect of the NAO on the precipitation variability changes. Present work contributes to the existing previous investigations of daily precipitation variability changes (Karl et al., 1995; New et al., 2001) and completes the still insufficient information about recent changes in precipitation variability.

2. Daily Precipitation Time Series

The precise monitoring of climatic characteristics started at Prague-Sporilov (the Czech Republic) in 1994 (Cermak et al., 2000). The station is located on the top of a low hill in the campus of the Geophysical Institute of the Czech Academy of Science (50.04°N, 14.48°E, 275 m a.s.l.) on the rim of the large urban agglomeration. The experiment accentuates on the continuous air and soil/ground temperature monitoring at different depth levels, with some associated climatic variables such as radiation, precipitation rate, etc. The precipitation itself was measured by means of a weighting–recording gauge, produced by a certified Czech firm. Cross-section of the gauge cylinder is 200 mm^2. The gauge has a sharp rim, placed 100 cm above the surface. Inner surface of the gauge is matted black, outer surface is bright white. Surface surrounding the gauge is short grass. About 5 m high deciduous tree is located approximately 10 m to the south and a low building is several ten meters apart. Weighing double-balance switches over after 0.2 mm of rainfall. For further calculation we used daily sums of precipitation.

Figure 1 shows the results of the eight-year long daily precipitation time series. Since there were no changes in the observational procedure and/or in the equipment installation during the whole experiment, the only possible source of the inhomogeneity in the time series is the data incompleteness. As seen, the precipitation

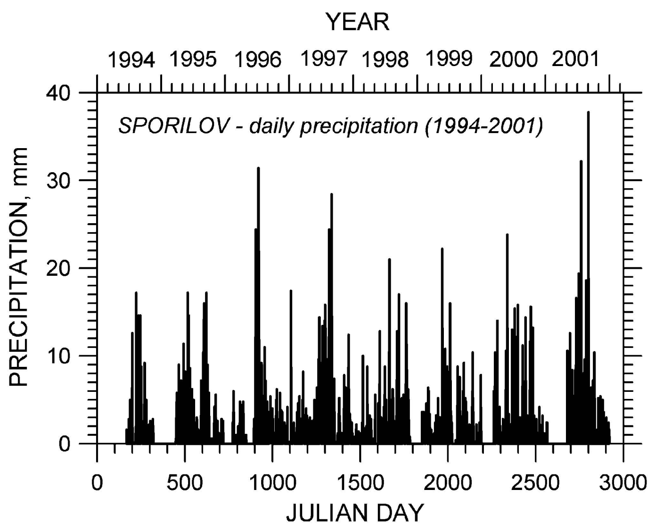

Figure 1. Results of 8-yr measurements of daily precipitation at Prague-Sporilov.

record contains several gaps, and the impact of missing data could bias the precipitation trend if assumed to be nil. In the present work the gaps, the length of which did not exceed ten days, were filled with the gamma distribution technique. An extensive literature on the use of the gamma distribution as a precipitation model has been developed over the past several decades (e.g. Haggard et al., 1973; Ropelewski et al., 1985; Bradley et al., 1987). It was shown, that gamma distribution provided a good fit to the precipitation data and enabled the precipitation amounts to be accurately expressed in terms of probabilities. In the present work we followed the method, described in details by Karl et al. (1995). We used the gamma distribution physically constrained by zero as a lower bound, and with unlimited upper bound. The shape and scale parameters of the gamma distribution were calculated, based on the distribution of daily precipitation for months, which did not contain gaps. The character of missed precipitation in the months with gaps was estimated by comparing the probability distribution obtained for such months with the "standard" shapes. Most of the missed data fell in the lowest range of the variations, and considerable part of them composed the days without precipitation. To determine the position of the rain-free days within the gap we applied a random-number generator, and the rest days of the given gap acquired the remaining precipitation values. The longest gaps occurred at the end of 1994 and in the beginning of 2001. These two and other four shorter gaps were not filled and for the further calculation the precipitation time series was regarded as consisting of six separate sections.

Daily precipitation has no significant linear trend for the whole 8-yr long observational period, but revealed a certain seasonal character (Figure 1); the wetter season occurs in May–August period and the precipitation minimum occurs in winter. This conclusion is confirmed by the meteorological observations of the 19–20th centuries on the monthly scale of aggregation (Figure 2). The increase in precipitation in "wet" years occurs mainly due to its significant growth in summer period, when the actual monthly amounts of precipitation can be a few times higher than the average. Prevalence of summer precipitation is a specific feature of the hydrologic cycle in the Czech Republic, while the annual distribution of precipitation may differ in surrounding areas (Kalvová and Nemešová, 1997; Bodri and Cermak, 2000, 2001). However, such bell-shaped distribution realises only in the long-time averages. Almost 30% of years of the 20th century have the bimodal distribution of precipitation, e.g., in the year 1960 the first precipitation maximum occurred in August, and the second similarly strong precipitation rise was in October. The spectral analysis of the precipitation record does not show any significant cycle.

Figure 3 shows the results of the spectral analysis performed on the longest continuous section of the precipitation record (June 1996–November 1998). A fast Fourier transform (FFT) algorithm was applied to the data. Since the data have no significant trend, the pre-processing of the data included only the tapering of the profile to handle the remainder problem. The tapering was done using a cosine-squared (Hanning) window, so that the beginning and the end of the profile begin

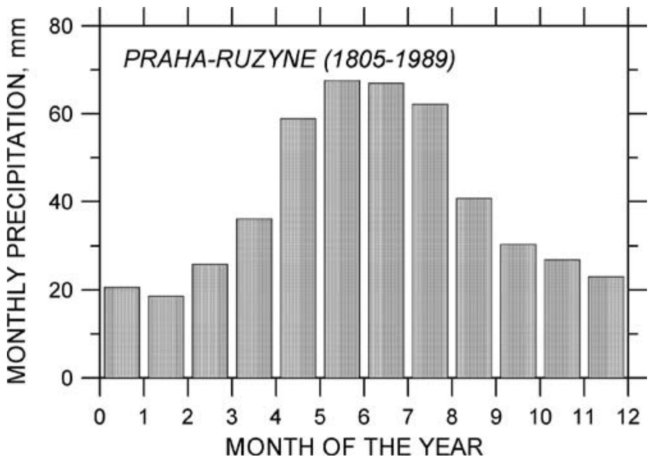

Figure 2. Averaged monthly precipitation at Prague-Ruzyne. The 2214 months between 1805 and 1989 were used for averaging (Data source: The Global Historical Climatology Network, GHCN 1; www.worldclimate.com).

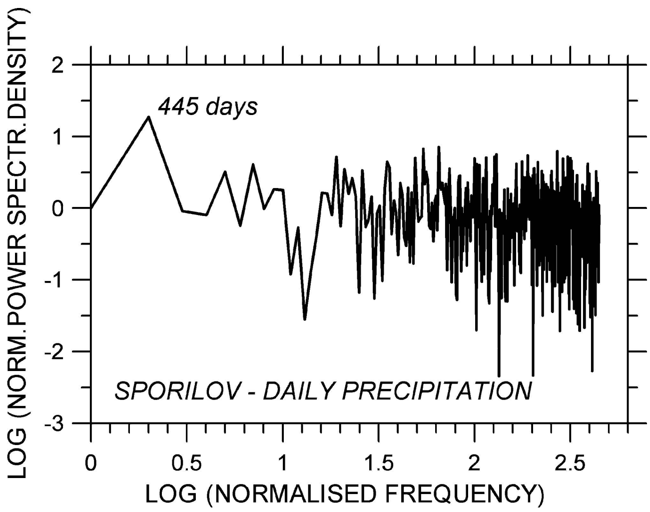

Figure 3. Power spectrum of the section of precipitation time series from Sporilov embracing the period of June 1996–November 1998 in a log–log plot. The values are relative: The frequencies are normalised to the lowest frequency in the spectrum, the power spectral density to that at the lowest frequency.

with a zero. The FFT algorithm describes the profile data as a sum of sine and cosine waves. By squaring the amplitude at each frequency and normalising it with the profile length we calculated the power spectral density. The spectral density versus frequency was plotted on a log–log plot. As seen, the power spectral density of the only rising wave with period of 445 days does not significantly exceed the background. The absence of either 1-yr or half-year wave can be explained by the relative shortness of the whole investigated time series.

3. Short-Term Precipitation Variability

There is a number of different ways to detect variability changes. In principle, the variability change presents a departure from a certain reference value, e.g. some more or less long-term mean or a departure from the previous discrete value. Variability change may also be distinguished by applying the chosen statistics, such as the precipitation difference between two subsequent periods, or as the standard deviation or the change in the probability density function parameters, etc. All measures have their own limitations. For example, the standard deviation method gives a highly averaged picture of the variability changes (see e.g. Schönwiese et al. (2003), who compared 50–50 yr periods) and is insensitive to the temporal position of the extremes. The choice of the method is, thus, stipulated by the data pattern and by the goal of the investigation.

The measure of variability used in the present study is based on work by Karl et al. (1995) and is defined by the absolute value of the difference in precipitation amounts between two adjacent periods of time. We have precipitation time series T_1, $T_2, \ldots, T_i$. The measure of variability, which we call N-point change, is calculated as the absolute difference between the average of a sequence of N successive values that begins at the measured point t and the similar average of N values that begins at point $t + \Delta t$:

$$\Delta T_N = \mathrm{abs}(\bar{T}_i - \bar{T}_{i-k}),$$

where

$$\bar{T}_i = \frac{1}{N}\sum_{l=i}^{i+N-1} T_l, \quad \bar{T}_{i-k} = \frac{1}{N}\sum_{l=i+N-k}^{i+2N-k-1} T_l.$$

This measure provides a continuous time series of variability, and it is its continuity that is important because of a strong frequency dependence of variability. On the other hand, it also clearly marks temporal position of the extreme variability changes. Karl et al. (1995) showed that this method is free of some disadvantages of other conventional methods, such as the standard deviation, e.g. it prevents from confounding of high- and low-frequency variability. Because of homogeneity of the data used in the present work, this method can be used without limitations stipulated in the work by Moberg et al. (2000). Generally, $\Delta t \leq N$, which implies the possibility of the overlapping. An overlapping is useful for the long intervals when the use of the strictly non-overlapping differences artificially constrains their number, and may lead to noisy seasonal estimates.

The measure of variability ΔT_N was calculated successively for the whole precipitation time series, and we obtained the time series of variability measures for

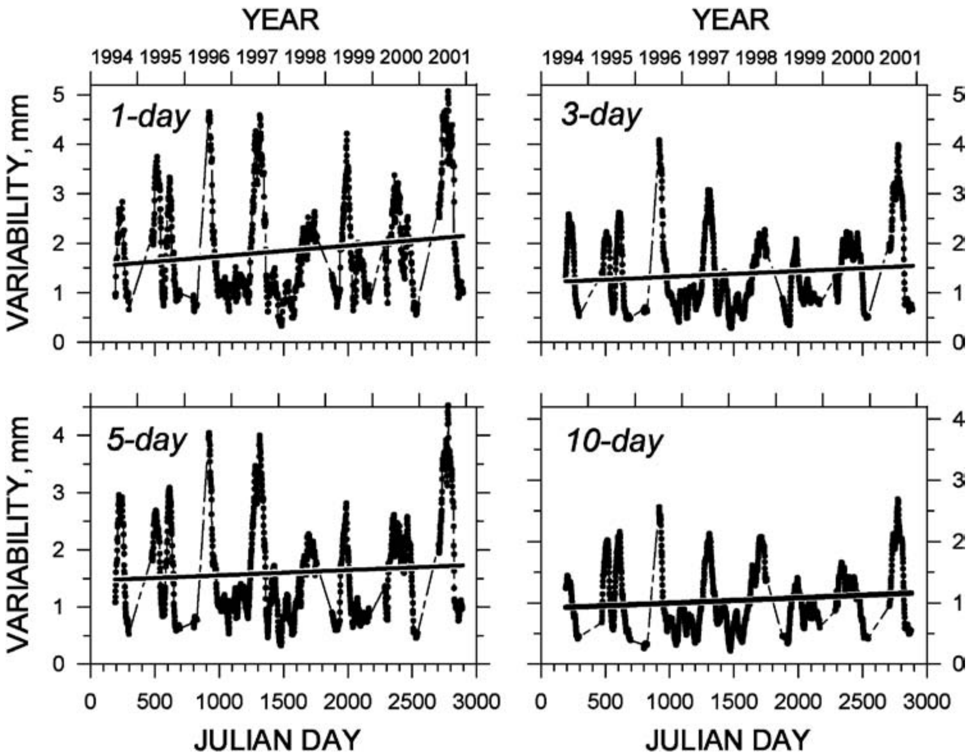

Figure 4. Variations and trends of the differences in average precipitation for various averaging intervals. Low frequency changes are highlighted by a Gaussian filter with $k = 15$ days corresponding roughly to 50-day moving averages. Straight lines represent the linear fit to the data.

different averaging intervals. This measure was then presented as "variability" in Figure 4 and following figures; Gaussian filter highlighted the low frequency changes. Transfer function of Gaussian filter is $S(t) = \exp(-k^2 f^2)$, where f is frequency and k is the scaling parameter. The cut-off threshold was used as $0.707(1/\sqrt{2})$, thus, cut-off frequency f_c can be obtained from the equation $f_c = 0.6/k$. Such cut-off frequency roughly corresponds to the $(k/0.3)$ moving averages (Meskó, 1984).

As seen in Figure 4, variability of the precipitation increases for all investigated averaging intervals (the slopes of the linear trend were calculated by conventional method). All are significant at the 0.05 level and achieve following values:

1-day 0.077 ± 0.011 mm/yr,
3-days 0.035 ± 0.008 mm/yr,
5-days 0.044 ± 0.007 mm/yr,
10-days 0.033 ± 0.007 mm/yr.

Contrary to this finding, no linear trend was detected in the daily precipitation of the 8-yr period of monitoring. The non-existence of a significant trend in the precipitation itself together with a noticeable change in its variability was reported by Mearns et al. (1995) in their "warmer world" model. It is not clear, however, whether observed variability increase in our case is due to the effect of the global warming, or it may be attributed to the natural provisional changes in the climate variability or even to the urbanisation effect of Prague City. Anthropogenic alteration of the urban precipitation, the so-called heat-island circulation (HIC) effect results from the difference in the sensible heat transport from the surface to the lower atmosphere caused by different land surface properties of urban and suburban areas. Practical studies have shown a certain influence of the HIC on the precipitation in major cities across the globe (Lo et al., 1997; Bornstein and Lin, 2000; Morris et al., 2001; Shepherd et al., 2002; Dixon and Mote, 2003). Field studies and numerical simulations (Bornstein and LeRoy, 1990; Ohashi and Kida, 2002) revealed, that the HIC scale and its intensity depended on many factors, among them on the urban area size, the atmospheric thermal stability, differences in the surface profiles of the urban and suburban areas, and confirmed their strong local dependence. In some definite cases the HIC can seriously affect the local hydroclimate (Mullen, 1999; Daly, 2000), while in other areas the urban-induced changes in the rainfall are subtle and less detectable than the changes in temperature, winds, etc. Results from the Metropolitan Meteorological Experiment (METROMEX) project showed that in some urban areas the HIC effect did not increase the number of new precipitation events, but rather stimulated the already occurring events (Changnon, 1981). In many cases the urban heat-island only modifies the rainfall character, and the precipitation events do not coincide with the highest HIC intensities (Dixon and Mote, 2003). The HIC effect exhibits also a strong areal influence, e.g. the field results (Shepherd et al., 2002) of the 3-yr (1998–2000) warm-season rainfall in the vicinity of several U.S. metropolises revealed an average 28% increase in the monthly rainfall rates within 30–60 km downwind of the metropolis, with a modest 5.6% increase over the metropolis. Portions of the downwind area exhibited an increase as high as 51%. The increase in the precipitation in the urban areas, even in the case of intensification of rainfall events only, hints a possible increase of the corresponding variability. However, since none of the mentioned studies dealt directly with the precipitation variability, the HIC influence on the variability is still a questionable one. It requires further investigation studies into the subject of HIC-induced precipitation by using a wide spectrum of both meteorological and environmental data. Such modeling, however, is far beyond the scope of this paper. It may be worth to note that so far the existence of the heat-island effect of the Prague-city was actually mentioned (Kyselý, 2002) but in the air temperature series and no such evidence has been detected in the precipitation data. Therefore, here we can only point out the possibility of such potential influence on the precipitation variability.

In all four averaging intervals the variability oscillations exhibit a quasi-seasonal character. Generally (but not compulsory) highly variable periods are very short and coincide with the summer wetter season. Especially in the shorter averaging intervals (day-by-day and 3-days) the increasing variability trend occurs above all due to increase in maximum variability, when the minimum variability remains on the same level during most of the monitoring period. As the maximum variability is typical for summers, more heavy rains are to be expected in this period and if this tendency will further continue, the risk of the summer floods will increase in future. It appears that the seasonal dependence of variability may be a generic feature of climatic records. Shabalova and Weber (1998), Datsenko et al. (2001), Shabalova and Van Engelen (2003) revealed the seasonality in the low-frequency temperature variability. While the summer temperature variability showed strong oscillations in the range of 60–80 yr, the winter variability was more monotonous and did not possess any characteristic time scale.

4. Long-Term Precipitation Variability

Variability patterns are sensitive to time period and to data aggregation. Therefore we have completed the above short-term variability analysis by the longer-term studies. For this purpose we used the historical time series of precipitation. Traditionally, long-term precipitation data in Europe have consisted mostly of monthly mean values. The most informative homogenised land precipitation series for the 20th century were offered by Hulme et al. (1998) (see www.cru.uea.ac.uk) for period 1900–1998 gridded at 2.5° latitude by 3.75° longitude resolution. Gridded data accumulate effects of the individual stations. Such time series appear more smoothed, thus, the range of variability variations may be lower than in the measured one-station data, and the "explosions" of variability due to local extreme events may be missed. On the other hand, general trends revealed in these data are of areal importance. When they coincide with the local data, they serve as a good verification of variability patterns obtained from original station data.

Figure 5 (top) shows data for the grid box with the centre at 50°N and 15°E (corresponding to the location of Prague) based on averaging the observations from 7 meteorological stations and consisting approx.1200 data points. These data (Hulme, 1992) were homogenised including screening for gross outliers and for typographical errors, station data were corrected to exclude gauge biases due to changing place and gauge design. On the other hand, there were no corrections for gauge undercatch, varying sensitivities to snowcatch of different gauge designs and no topographic weighting was applied to the interpolation scheme.

The calculated variability pattern corresponding to the time series of monthly precipitation (Figure 5, bottom) reveals a general increasing trend of 0.060 ± 0.003 mm/yr characteristic of the whole 20th century, while the precipitation data show no trend at all (Figure 5, top). Period between years 1980–1995 is characterised by a provisional variability decrease that significantly lowers the amplitude of the calculated linear trend (the value calculated for the period of 1900–1985 amounts to 0.11 mm/yr). The observed decrease in variability may represent the part of the

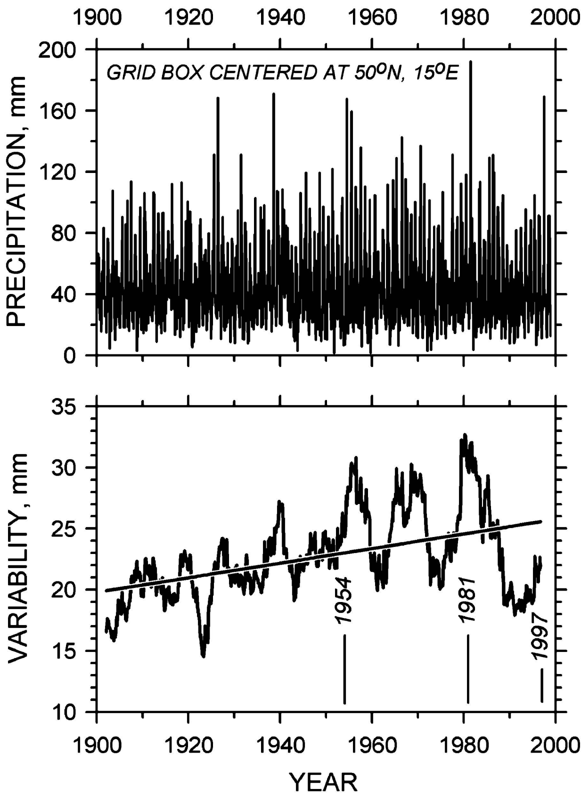

Figure 5. *Top*: Time series of monthly precipitation for the historical monthly precipitation data set grid box centred at 50°N, 15°E (Hulme et al., 1998). *Bottom*: Variability pattern for monthly precipitation. Low frequency changes are highlighted by a Gaussian filter with $k = 14$ month, corresponding roughly to 4-yr moving averages. Straight lines represent the linear fit to the data.

"near-decadal" variations (see below). Increasing tendency has been restored after year 1995. This result can be well confirmed by the investigations of Schönwiese et al. (2003) based on the 100-yr long gridded European monthly data set as well as by the similar data set from the German station-based network. Comparing the first and the second half of their time series, the authors demonstrated an increase of the number of extremely wet months, that was also reflected in the systematic increase in the second moments of data, such as variance and the Weibull probability density function parameters.

The reliability of the above trend can be illustrated with the similar calculations done for the neighbouring grid box 50°N, 11.25°E. This time series (Figure 6, top) is based on the averaging observations at five meteorological stations. As previously, data themselves do not contain linear trend for the observational period. The range of precipitation fluctuations as well as the range of precipitation variability is somewhat lower in comparison with the previous grid box. However, obtained pattern of variability is similar to that presented in Figure 5 (bottom) also

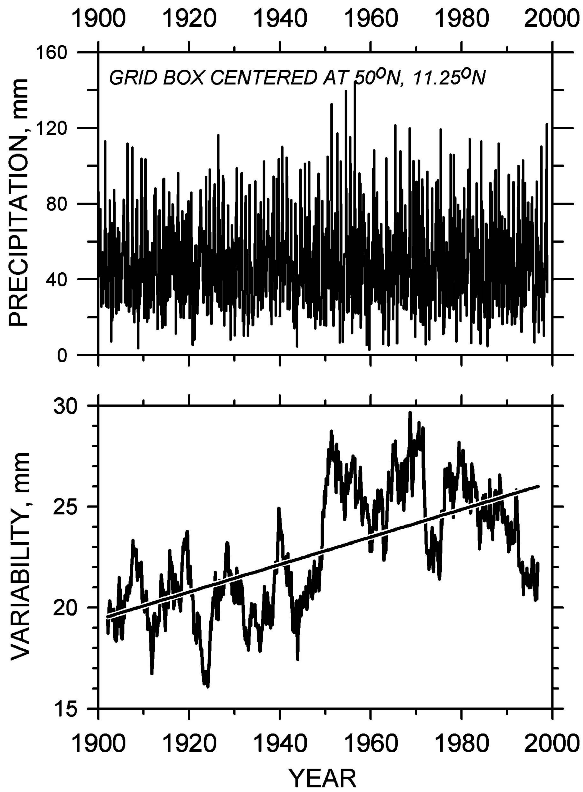

Figure 6. The same as Figure 5 for the grid box 50°N, 11.25°E.

revealing general increasing trend of 0.068 ± 0.003 mm/yr for all observational period.

The spectral analysis of the variability time series in addition to the annual periodicity revealed also a certain important "near-decadal" quasi-periodicity. This "near-decadal" periodicity is represented in Figure 7 as a series of local maxims labelled as "13.5 years" corresponding to the period of the strongest wave within the interval of periods from 10 to 15 years. Physical explanation of such quasi-decadal precipitation variability is so far restricted to the present insufficient knowledge of the regularities in the local atmospheric circulation pattern. Similar signal can be observed both in SAT records (Bodri, 2004) and in precipitation record for neighbouring grid box (see Figure 6 and subsequent discussion). Certain link between decadal trends in precipitation and the North Atlantic Oscillation were hypothesised by Hurrell (1995) (see below). Periods of high variability generally coincide with the enhanced monthly precipitation. Heino et al. (1999) reported a strong coupling of extreme and monthly precipitation and showed that in the Czech Lands the wettest day of the month contributes considerably to the total monthly sum. It may be interesting to mention that at least for the last 50 years a definite interdependence

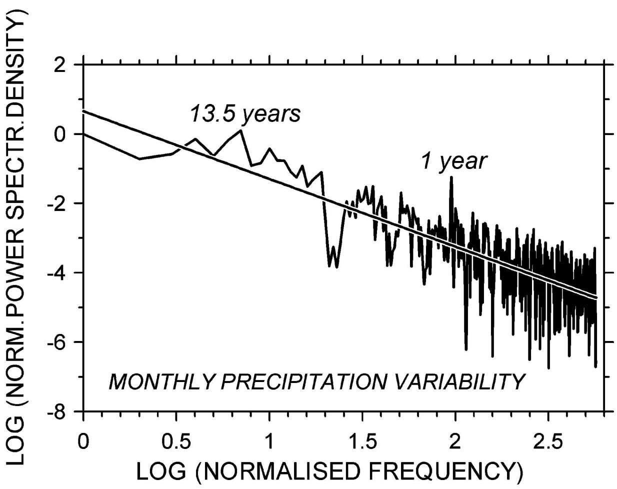

Figure 7. Power spectrum of the variability time series for the grid box centred at 50°N, 15°E.

between extreme precipitation, increased precipitation variability and the summer floods can be observed. E.g. summer floods of the Vltava river occurred in July of 1954, 1981 and 1997 (Kakos, 1983, 1997) and recently in August 2002. Since approx. year 1950 the range of variability variations increased significantly and if such tendency continues, the flood risk increases.

5. Precipitation Variability and the NAO

It is becoming increasingly evident that the large-scale modes of atmospheric variability, such as the North Atlantic Oscillation (NAO) and Arctic Oscillation (AO), strongly influence temperature and precipitation patterns at mid- and high-latitudes of Europe (Hurrell, 1995). The AO is associated with the deep zonal symmetric structures with pressure anomalies of opposite sign in the polar cap region and a zonal ring near 45°N. The positive polarity of the AO is characterised by a strengthening of the polar vortex from the surface to the lower stratosphere. Under this condition storms in the North Atlantic bring precipitation and mild air to NW Europe. During negative polarity of the AO cool air flows to W Europe and storms bring rainfall to the Mediterranean region. The AO index is defined by the standardised leading principal component time series of monthly mean surface level pressure (850 hPa geopotential height) fields polewards of 20°N derived from the NCEP reanalysis dataset (Thompson and Wallace, 2000; Thompson et al., 2000). The AO is highly correlated with land precipitation in high- and mid-latitudes of Europe. According to New et al. (2001) the correlation between AO indices and precipitation in the 40–60°N domain achieves −0.33 and −0.59 on the annual scale and/or for winter months, respectively. The strength of the NAO phases

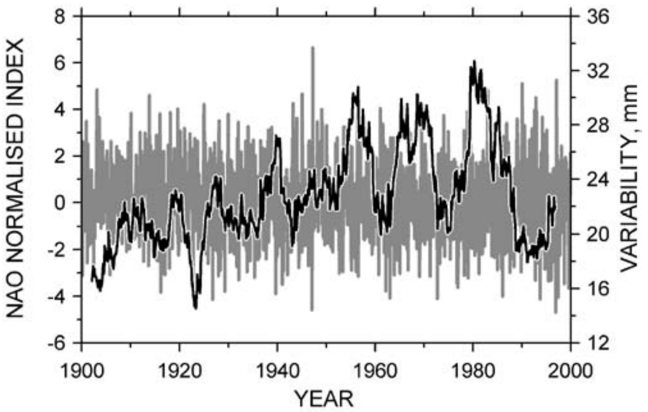

Figure 8. Variability pattern for monthly precipitation superimposed on the monthly variations of the NAO index.

is measured through the index, defined as the difference between the normalised pressure anomaly, at Gibraltar, and that at Reykjavik. The NAO index fluctuates on all time scales in a range from intra-annual to decadal (Kelly and Jones, 1999; see also Figure 8). The dependence of European temperatures and precipitation on the NAO is particularly evident during the winter period; the larger correlation between the NAO index and climatic signal is typically observed from December through March (Hurrell, 1995; Hurrell and van Loon, 1997). The NAO regulation of the European weather is realised through the westerlies regime. The positive phase of the NAO leads drier conditions over much of Central and Southern Europe and the Mediterranean, whereas wet anomalies occur in Iceland and Scandinavia. Opposite conditions are observed during low-NAO winters. Because of a rather short observational interval the Sporilov data are hardly suitable to investigate the correlation between precipitation variability and the NAO index and we have limited ourselves only on the monthly data for the Czech box. Figure 8 shows the superposition of the normalised monthly NAO indices (Jones et al., 1997, corrected and updated, see www.cru.uea.ac.uk) and the monthly variability pattern of precipitation at grid box centred at 50°N, 15°E. Definite interdependence between variability and the NAO can be observed, see e.g. intervals between 1945–1950 and 1985–1995 characterised by the frequent occurrence of high positive NAO indices and low variability, while generally negative NAO values of 1980–1985 are associated with high variability. The NAO control on the weather of the Northern Hemisphere is the strongest in winter, the correlation between winter (DJFM) NAO index and the variability averaged over this period amounts to $r = -0.39$ and is significant at 0.05 level (Figure 9). This value is close to the correlation coefficient $r = -0.29$ between the land precipitation and winter NAO indices for the latitude band of 40–60°N calculated by New et al. (2001). This fact implies similarly strong NAO-impact on precipitation variability as on the precipitation itself.

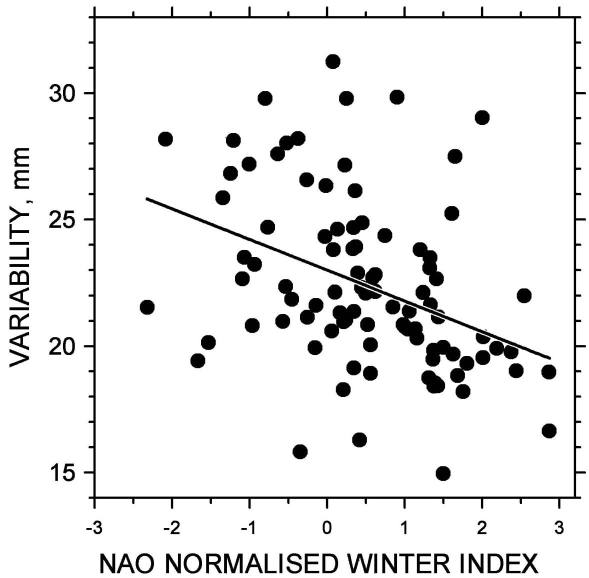

Figure 9. Variability of monthly precipitation as a function of a normalised NAO index. Both quantities are averaged over winter (DJFM) months. Thick line represents the best-fit linear regression.

The additional test of significance was performed with Fisher's z-transformation of correlation coefficient $z = \frac{1}{2}\log[(1+r)/(1-r)]$. The standard error of this transformation can be calculated as $\mathrm{var}(z) = \frac{1}{n-1}\{1 + \frac{4-r^2}{2(n-1)} + \frac{22-6r^2-3r^4}{6(n-1)^2} + \cdots\}$, where n is the number of samples (Kendall and Stuart, 1967). According to Kendall and Stuart (1967), the approximation represented by the latter formula yields satisfactory results already if $n > 10$. In our case $z = -0.41 \pm 0.10$, that implies the significance of obtained correlation. The linear regression fit of the NAO on variability values has the next coefficients: slope $= -1.21 \pm 0.30$ mm and intercept $= 23.00 \pm 0.36$ mm. The relative rms misfit $\{\sqrt{\sum[(v - v_{\mathrm{calc}})/v]^2}\}/(n-1)$, where v are the real values of variability, v_{calc} are the values of variability calculated from the NAO indices, and n is a number of samples, amounts to only 14%. This value also implies a good coincidence of variability and the NAO.

The signature of the NAO in climatic variables is strongly regional. Linking higher precipitation variability to negative NAO indices hints the situation similar to those in S Europe. As shown by Rodwell et al. (1999) stronger NAO values were linked to high pressure and stable weather situation, when both weather type and the temperature are changing only little, while during low (negative) NAO winters westerlies assume a more zonal trajectory bringing wetter and warmer weather and possibly increase their variability to S Europe. Our results may complete the earlier studies by Rebetez (2001), who described the negative correlation between SAT variability and the NAO index for two meteorological stations in Switzerland, when stronger NAO values are accompanied by reduction in day-to-day SAT variability.

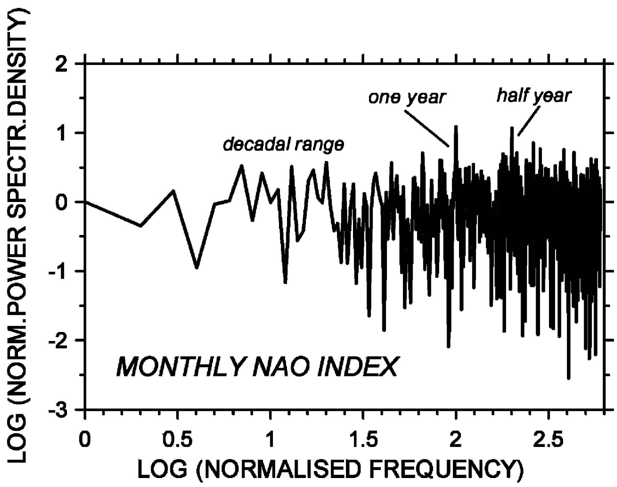

Figure 10. Power spectrum of monthly NAO index.

Quasi-decadal trends in precipitation variability time series discussed above are similar to those revealed by Hurrell (1995) in the regional temperature and precipitation time series in Europe. The author linked the former to the decadal changes in the atmospheric circulation inspired by the NAO. As previously, this means only quasi-periodicity, since the spectral analysis reveals only a relatively weak annual and six month periodicity in the NAO (Figure 10), and does not show any near decadal wave, even when several weak waves appear in the precipitation variability spectrogram. The causes for such variability in the Atlantic are not clear, certain relation of the NAO to greenhouse gas forcing and possible links to coherent variations in tropical Atlantic sea surface temperatures cannot be avoided.

6. Concluding Remarks

As was shown in several works the quantities that are to be used to describe climate change scenarios are not only the mean characteristics of the climate but also the higher order statistical moments, such as variability. The study of climate variability gives a more comprehensive summary of climate than conventional climate averages. Results of variability-related studies have many practical applications, since the impact of the variability changes on the environmental and associated socio-economic systems may be more perceptible, than the slow changes in the mean characteristics.

The GCM models, that have been applied to the problem of the effect of increasing CO_2 levels on climate, predict not only global increase of surface air temperature, but also changes in the precipitation rate. However, they do not give an unequivocal picture of how precipitation might be expected to change. While a global warming trend of 0.3–0.6 K during the last century has been widely recognised, there is some debate about the significance of the increasing trend in global land precipitation for the same period (Nicholls, 2001). The recognised global and land precipitation increase by about 9 mm/100 yr is small in comparison with the range of its variability. Numerical simulations as well as the real data studies revealed regional dependence of the precipitation characteristics and the possibility of significant increase in precipitation variability with relatively conservative or without changes in the mean quantities. The use of both characteristics can provide better validation of the models for the climate change and land use.

In our work variability patterns obtained for all investigated averaging intervals from day-by-day to 1 month have shown apparent increase in variability that (at least partly) can be attributed to the results predicted by the global warming models. The increasing trend exists since the beginning of the 20th century. Quasi-seasonal and quasi-decadal oscillations are superimposed on the increasing tendency. The oscillations of variability have become more pronounced after year 1950. Part of these oscillations can be attributed to the large scale forcing mechanisms, such as e.g. the NAO and AO. Our calculations imply similarly strong impact of the NAO on variability as was obtained in the earlier work by New et al. (2001) between the NAO and monthly precipitation rates. The variability increase as well as the associated increase in the frequency of the extreme events may have a serious impact with respect to water resources and/or flooding, soil erosion, agriculture, etc.

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