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Long term variability of the annual hydrological regime and sensitivity to temperature phase shifts in Saxony/Germany

M. Renner and C. Bernhofer

Dresden University of Technology, Faculty of Forestry, Geosciences and Hydrosciences – Institute of Hydrology and Meteorology – Department of Meteorology

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Correspondence to: M. Renner (maik.renner@mailbox.tu-dresden.de)

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Abstract

The timing of the seasons strongly effects ecosystems and human activities. Recently, there is increasing evidence of changes in the timing of the seasons, such as earlier spring seasons detected in phenological records, advanced seasonal timing of surface temperature, earlier snow melt or streamflow timing. For water resources management there is a need to quantitatively describe the variability in the timing of hydrological regimes and to understand how climatic changes control the seasonal water budget of river basins on the regional scale.

In this study, changes of the annual cycle of hydrological variables are analysed for 27 river basins in Saxony/Germany. Thereby monthly series of basin runoff ratios, the ratio of runoff and basin precipitation are investigated for changes and variability of their annual periodicity over the period 1930–2009. Approximating the annual cycle by the means of harmonic functions gave acceptable results, while only two parameters, phase and amplitude, are required.

It has been found that the annual phase of runoff ratio, representing the timing of the hydrological regime, is subject to considerable year-to-year variability, being concurrent with basins in similar hydro-climatic conditions. Two distinct basin classes have been identified, whereby basin elevation has been found to be the delimiting factor. An increasing importance of snow on the basin water balance with elevation is apparent and mainly governs the temporal variability of the annual timing of hydrological regimes.

Further there is evidence of coincident changes in trend direction (change points in 1971 and 1988) in snow melt influenced basins. In these basins the timing of the runoff ratio is significantly correlated with the timing of temperature, and effects on runoff by temperature phase changes are even amplified. Interestingly, temperature effects may explain the low frequent variability of the second change point until today. However, the first change point can not be explained by temperature alone and other causes have to be considered.

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1 Introduction

1.1 Motivation

In nival and pluvial catchments and river basins of Central Europe (CE) we observe a variable but distinct seasonal hydrological regime. This hydrological regime is a result of several processes induced by meteorological forcing and the properties of the receiving catchments. Looking at the water balance of typical basins in CE, precipitation has a small seasonal cycle compared to its variation and would alone not account for the pronounced seasonality of runoff. This is mainly introduced by basin evapotranspiration, resulting in lower flows during summer and early autumn. Also at catchments at higher elevations, snow accumulation and snow melt produce higher flows in late winter and spring. Besides the local climate, catchment properties such as water storage in soils, evaporative demand of vegetation and human water management moderate the resulting hydrological regime.

Water resources management has to deal with the seasonality of hydrological regimes. Generally demand and water availability are out of phase, i.e. when the availability of water is lowest (summer) the demand is highest. This pressure on water delivery systems increases the need to correctly estimate the seasonality of hydrological processes (Loucks et al., 2005). Still, it is common practise, to assume stationarity of monthly statistics and thus stationarity of the whole annual cycle in design studies for water resources management, e.g. the use of the Thomas-Fiering simulation model (Maniak, 2005). However, there is increasing evidence of changes in the timing of the seasons from various disciplines. Earlier streamflow timing and snow melt has been reported e.g. by Stewart et al. (2005); Déry et al. (2009); Stahl et al. (2010). Further, phenological studies provide evidence of earlier spring season, e.g. Dose and Menzel (2004). Based on station and gridded data of surface temperature, Thomson (1995) and Stine et al. (2009) found tendencies of advanced seasonal timing.

Consequently, there is a need to estimate the timing of hydrological regimes, its variability and to check for long term changes, which could possibly violate the stationarity assumption of the annual cycle of hydrologic records. Furthermore, the sensitivity of changes of the phase in temperature needs to be assessed. This is especially important when considering the regional impacts of climate and land use change.

1.2 Seasonal changes in hydrologic records

Hydrological studies concerned with changes in streamflow within regions throughout CE usually analyse annual runoff and seasonal changes by monthly data (KLIWA, 2003; Fiala, 2008; Stahl et al., 2010). Generally annual flows do not show significant trends, but spatially coherent trends in separate months have been reported. Most pronounced are positive streamflow trends in winter months, followed by negative trends in spring (Stahl et al., 2010). Mostly it is concluded that these trends are a result of warmer winters which in turn lead to an earlier onset of snow melt. Moreover, in regional climate studies also changes in winter precipitation have been ascertained e.g. (Franke et al., 2009). The KLIWA (2003) report, concerned about long term changes in regional runoff in Southern Germany concludes with the recommendation to jointly analyse seasonal changes of runoff and precipitation.

There is a range of measures that can be used to directly estimate the timing of annual streamflow regimes, such as the timing of the annual maximum, the fraction of annual discharge in a given month or half flow dates (Court, 1962; Hodgkins et al., 2003; Regonda et al., 2005; Stewart et al., 2005). Even though these measures are relatively simply and easily understood, these metrics are only useful for very pronounced hydrological regimes such as those dominated by snowmelt. Déry et al. (2009) note that synoptic events may dominate such measures rather than long term changes in climate, e.g. warm spells in winter or late season precipitation.

Relatively few studies have studied the variability of the annual cycle, being the strongest signal in climate records at mid to high latitudes (Huybers and Curry, 2006). By using a harmonic representation of the annual cycle, this variability has been studied for long records of surface temperatures (Thomson, 1995; Stine et al., 2009) and precipitation (Thompson, 1999). The resulting annual phases and amplitudes describe the timing of the annual cycle and its range, based on the whole cycle instead of considering each month separately.

1.3 Regional climate in Saxony

The Free State of Saxony is situated at the southeastern border of Germany. It is densely populated with about 4.3 Mio. inhabitants and covers an area of 18 413 km^2. The climate is characterised by two main factors. First, there is a transition of the maritime western European climate to the continental climate of eastern Europe, which leads to a temperate warm and humid climate with cool winters and warm summers. Second, there is an orographic factor due to the mountain ranges at the Southern border with elevations gradually increasing from 100 m up to 1200 m. Recently, the climate and observed trends have been described in detail by Bernhofer et al. (2008) and summarised by Franke et al. (2009). From the observed changes they deduce that climate change effects are more pronounced in Saxony than in other regions in Germany. Generally a positive trend in the number of droughts during growing season is reported, combined with intensified heavy precipitation. These effects are partly compensated at the annual level by increased winter precipitation.

1.4 Objective and structure

The objectives of this paper are (1) to derive a climatology of the timing of the annual hydrological regimes for a range of river basins throughout Saxony; (2) to evaluate their interdecadal variability and (3) to determine the proximal processes governing the locally coherent patterns of the observed changes in timing.

To resolve these issues, a reliable measure for the timing, being valid over different hydrological regimes is needed. So instead using streamflow records directly, we employ the basin runoff ratio, the ratio of discharge and basin precipitation. The seasonal fluctuation of runoff ratio is a direct measure of seasonal water availability and as being a normalisation it makes different basins more comparable. The series of

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monthly runoff ratio are filtered for their annual periodicity by the harmonic method described in Stine et al. (2009) and the resulting annual phases represent a timing measure of the regime of the runoff ratio. The climatologic behaviour of the timing is then analysed by circular descriptive statistics. The interdecadal variability of the timing is being addressed by a qualitative method, namely cumulative departures of the average. Together with a correlation analysis to observed climate variables, such as timing of temperature, annual temperatures and monthly snow depths, we aim to identify the driving processes governing the changes in the timing of the runoff ratio.

2 Methods

2.1 Annual periodic signal extraction

The aim is to estimate the timing of the annual cycle from a geophysical time series without a subjective definition of the seasons. Therefore, methods are necessary to filter the annual cycle from the data and to gain a time variant parameter which defines the timing of the seasons.

In general there are two ways to accomplish this task. First, there are form free models, which use some seasonal factor to describe the periodic pattern. This yields a good approximation to the periodic signal, at the cost, however, of estimating many parameters. The second approach are Fourier form models which are based on harmonic functions. These are generally defined by only two parameters: the phase (ϕ) and amplitude. These parameters are a natural representation of the seasonal cycle and are economic in terms of parameter estimation (West and Harrison, 1997). Using several long temperature series, Paluš et al. (2005) compared four different methods for estimating the temporal evolution of the annual phase (sinusoidal model fitting, complex demodulation via Hilbert Transform, Singular Spectrum Analysis and the Wavelet transform). They found good agreement between these methods and concluded that the annual phase is a robust and objective way to estimate the onset of seasons.

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The method of complex demodulation (Bloomfield, 2000) has been used to analyse shift in seasons from long temperature records by Thomson (1995) and Thompson (1999) analysed the seasonality of precipitation over the British isles.

The idea behind complex demodulation is to write a time series X_t as follows:

$$X_t = A_t \cos(f_0 t + \phi_t) + \epsilon_t, \tag{1}$$

where A_t is the amplitude, ϕ_t is the phase at frequency f_0 and both are allowed to change over time. The term ϵ_t is a residual component which has no component at frequency f_0. Then for a given frequency f_0 the complex demodulate Y_t is derived by:

$$Y_t = x \exp^{2\pi i f_0 t}. \tag{2}$$

From Y_t, the amplitudes and phases over time are derived from the complex demodulate:

$$\phi = \tan^{-1}(Im(Y_t)/Re(Y_t)) \tag{3}$$

$$A_t = |Y_t|. \tag{4}$$

Bloomfield (2000) assumes that A_t and ϕ_t change slowly over time and remove the high frequency components from Y_t by using various filters. In this work a simple moving average filter has been chosen, whereby several filters have been tried to resemble a likely climatology of the phase estimates.

Recently, Stine et al. (2009) analysed trends in the phase of surface temperatures on a global perspective. They used the Fourier transform to compute annual phases and amplitudes:

$$Y_X = \frac{2}{12} \sum_{t=0.5}^{11.5} e^{2\pi i t/12} x(t + t_0), \tag{5}$$

where $x(t + t_0)$ are demeaned monthly observations, with an offset t_0 denoting the middle of the month. Phase ϕ and amplitude A_x are computed for each period from Y_X by applying Eqs. (3) and (4).

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Equation (5) is applied separately for each calendar year in the record to gain a series of annual phases and amplitudes. Because of the simplicity of this method and good agreement with more complex methods, this method has been used to estimate the annual phases of temperature and the runoff ratio in this analysis.

2.2 The runoff ratio and its annual phase

To apply the method of Stine et al. (2009) for hydrological quantities, a pronounced seasonal cycle suitable for a harmonic representation is needed. While the runoff regime in Central Europe has distinct seasonal features, it is not very balanced and quite different from an harmonic such as a cosine function.

A better quantity is the ratio of runoff and basin rainfall, the runoff ratio (RR). It represents the fraction of runoff observed at the basin outlet from the amount of precipitation for a certain period. The regime of the runoff ratio naturally reflects key processes of the basin water balance. Most important are the seasonal characteristics of precipitation, the actual basin evapotranspiration and the storage and release of water in soil or snow pack. Over the year these processes form a pronounced seasonal cycle.

Moreover the runoff ratio is a direct measure of water availability of a basin, and is therefore an important quantity in water management. Lastly, the runoff ratio is a normalisation which allows to compare quite different hydrological regimes.

To illustrate the procedure of deriving a timing measure for the annual cycle of the runoff ratio, Fig. 1 (top) depicts monthly rainfall and runoff sums over a period of 5 years for an example basin. In the bottom graph, the resulting monthly runoff ratio (Q/P) is shown as grey line. Due to snow melt the ratio is larger than 1 in late winter time, also heavy rain events in summer may induce spikes in the record. Therefore, three-monthly running runoff ratios $\mathrm{RR}_t = \frac{Q_{t-1}+Q_t+Q_{t+1}}{P_{t-1}+P_t+P_{t+1}}$ for each month t have been computed. These are depicted as bold blue line and are generally smoother and better suitable for estimation of the annual cycle.

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The estimated cosine functions are depicted as dash-dotted lines in the bottom graph of Fig. 1. The annual phases ϕ are in fact circular variables in the range of $[-\pi, \pi]$. For convenience the phase angles are transformed to represent the day of year (doy): $doy = 365\phi/2\pi$.

To compare the derived timing of runoff ratio, an independent metric, the half-flow dates (Q_{50}) have been chosen. Half-flow dates are for example used by Stewart et al. (2005) to analyse streamflow timing changes and their link to seasonal temperature changes in Northern America. To compute Q_{50} the streamflow is summed over a period, e.g. one hydrological year, starting at the 1 November. Then the half-flow date is computed, when 50% or more of the annual sum have passed the river gauge. The derived half-flow dates of the illustrative example are shown in the upper sub plot of Fig. 1. Already for this short period, it can be seen, that both timing measures can have large differences in some years, whereby the phase estimate shows less fluctuations than Q_{50}.

2.3 Descriptive circular statistics

Circular variables have certain properties, such as the arbitrary choice of origin and the coincidence of "beginning" and the "end". Therefore, linear statistics may be inappropriate and special treatment is needed to derive correct conclusions from the data (Jammalamadaka and Sengupta, 2001).

As any circular variable can be represented as point on the unit circle, the starting point to derive correct statistics is to transform the angular variable into polar coordinates

$$x = \cos(\alpha) \qquad y = \sin(\alpha). \tag{6}$$

Considering a set of angular observations $\alpha_1, \alpha_2, \ldots, \alpha_n$, the circular mean and variance are computed as follows:

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$$\bar{\alpha} = \arctan\left(\sum_{i=1}^{n} \sin(\alpha_i) / \sum_{i=1}^{n} \cos(\alpha_i)\right) \tag{7}$$

$$\sigma_{\alpha}^2 = 1 - \frac{1}{n}\sqrt{(\sum_{i=1}^{n} \sin(\alpha_i))^2 + (\sum_{i=1}^{n} \cos(\alpha_i))^2}. \tag{8}$$

Another metric used in this work is the circular correlation ρ_{cc} between two circular vectors α and β:

$$\rho_{cc} = \frac{\sum_{i=1}^{n} \sin(\alpha_i - \bar{\alpha}) \sin(\beta_i - \bar{\beta})}{\sqrt{\sum_{i=1}^{n} \sin(\alpha_i - \bar{\alpha})^2 \sum_{i=1}^{n} \sin(\beta_i - \bar{\beta})^2}}, \tag{9}$$

with $\bar{\alpha}$ and $\bar{\beta}$ being the respective circular averages. For a full treatment of circular statistics the reader is referred to Jammalamadaka and Sengupta (2001).

2.4 Detection of nonstationarities, trends and change points

To analyse the variability of the estimated annual phase angles it is necessary to check for nonstationarities such as trends or structural changes of the mean or variance. As decadal changes of the mean may be expected from climatic variables, simple linear trends and significance testing may be not useful here. Another drawback is the sensitivity of the linear trend to the estimation period. Therefore, it is necessary to look for more complex trend patterns and analyse the low frequency variability.

There are many simple graphical methods available for this purpose, with simple moving averages and cumulative sums of standardised variables (CUSUM) used in this paper. The CUSUM method is often used in econometric literature (Kleiber and Zeileis, 2008), where the focus is on the analysis of regression residuals and their stability over time. However, the method can also be directly applied to a time series, whereby its stationarity is assessed.

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It is also possible to test the stationarity of such CUSUM processes. Under the Null Hypothesis of no structural change the limiting fluctuation process is assumed to be the Standard Brownian Motion. Then the probability of crossing some boundary line can be used for significance testing (Zeileis et al., 2002). However, Brown et al. (1975) note that these boundary lines "should be regarded as yardsticks" and recommend to graphically check the CUSUM graphs.

3 Data

The analysis comprises discharge series of 27 river gauges throughout Saxony and climatic data series such as precipitation, temperature and snow depth records. The station data used within this study have first been subject to a homogeneity test procedure, which has been used to detect possible structural changes in the series and to exclude anomalous series from the analysis. Further, climatic data such as rainfall, temperature and snow depths have been spatially interpolated to be able to compute river basin average values. All procedures are based on monthly data, as the method to filter the annual periodic components of the time series does not need higher temporal resolution (Stine et al., 2009). This was verified by using several temporal aggregation levels, which resulted in no essential change for temperature. The annual cycle estimates of runoff ratio show to converge with increasing aggregation periods (half monthly to quarterly).

3.1 Runoff data

Due to extensive hydraulic engineering projects since the industrial revolution in the 19th century, a dense network of hydrologic gauging stations has been established in Saxony. We have chosen 27 river gauge stations, which all almost fully cover the period 1930–2009. The stations cover large parts of Saxony, with catchment areas ranging between 37 and 5442 km^2. Most stations are within the Mulde River basin (15)

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or are tributaries of the Upper Elbe (6). Detailed information may be found in Table 1 and the map in Fig. 2.

The discharge data have been converted to areal monthly runoff (mm/month) using the respective catchment area. Then the data have been subject to a homogenisation test procedure. As reference series a weighted mean of neighbouring catchments and secondly the catchments runoff ratio has been used. For 8 runoff series significant inhomogeneities (cf. Table 2) have been detected by the Alexandersson and Pettitt homogeneity tests (Alexandersson, 1986; Pettitt, 1979), based on the continuous series as well as for separate months. Whereby two series, Streckewalde and Neundorf, are most severe. Both inhomogeneities are related to reservoir constructions during the 1970s. Traditionally, river runoff in the Ore Mountains has been altered by water management facilities, such as dams for drinking water supply, weirs for hydro power, drainage systems for forestry and water management of the various mining activities, see e.g. IKSE (2005, chapter 4.6). For a few runoff series, no obvious reason has been found for the detected inhomogeneities. These are usually not as severe as compared to basins where a dam has been built, and are most probably related to measurement errors (for example changes in the rating curve due to cross section changes).

3.2 Precipitation data

The geographical domain (11.5°–16° E, 50°–52° N) has been chosen for the spatial interpolation and station data selection.

The station network density has changed dramatically throughout time. Currently there is one station available in the database since 1858, 12 stations since 1891 rising up to 111 in the 1930's. Due to World War II were only 20 stations available in 1945. From the 1950's the network has improved to 374 in 1951 to a maximum of 873 in 1990. Since 2000 the network density has been decreased to 354 in 2008.

To check for influences of the changing network, three data sets have been prepared. One set only with stations covering the full period without longer missing periods, another set which consists of all observations available at a time step and another set

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which has been used in the analysis. This last set is a compromise between the other two sets, meeting the requirement that the series used must cover at least 40 years, i.e. from 1950–1990. This set contains 368 stations.

Based on these stations a homogeneity test procedure has been conducted. Depending on the meta data available a part of these stations has been tested for known breakpoints using the Kruskal-Wallis rank sum test for changes in the location and the Bartlett test for changes in the variance and the Alexandersson homogeneity test. If all these three tests reject the hypothesis of no change at the $\alpha = 0.05$ level then the series has been flagged as suspect. Next an iterative homogeneity test procedure has been done using a weighted series of about 5 reference stations. Reference stations are selected according to 4 criteria: (i) not inhomogeneous from previous test (ii) best correlation of the differenced series (Peterson and Easterling, 1994) (iii) cover most of the record of the candidate station and (iv) are close to the candidate. Then both, the Alexandersson homogeneity test and the Pettitt test have been applied. If both tests fail the hypothesis of stationarity at the $\alpha = 0.01$ level then the series has been flagged as suspect. Finally, a set of 299 precipitation series have been left for spatial interpolation, i.e. without any suspect series. There are 83 stations during the 1930's, about 290 from 1950–1990 with 170 in the last decade.

Based on the station dataset a spatial interpolation for each month has been computed. First a linear height relationship using a robust median based regression (Theil, 1950) has been established. Then the residuals have been interpolated onto an aggregated SRTM grid (Jarvis et al., 2008) of 1500 m raster size using an automatic Ordinary Kriging (OK) procedure (Hiemstra et al., 2009). Monthly basin average precipitation is then computed by the average of the respective grid cells. The method of height regression and OK of the residuals has been chosen, as this method showed to have the lowest root-mean-square errors (RMSE) among other methods, in a cross-validation based on monthly station data sets.

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3.3 Temperature and snow depths

The network of climate stations in the domain has also changed during time. Since 1930 9 long temperature series have been available, this increased to 47 in 1961 and again reduced to 38 in 2008. A few snow depth observations are available from climate stations. Additionally, a dense network of snow depths has been established in the region since 1950. On average 163 series are available. For both elements the basin average have been computed similar to the methods described for precipitation in Sect. 3.2.

4 Results

4.1 Time series of basin runoff ratios

To gain some insight in the general spatial behaviour of the runoff ratio of the selected basins, a map of the long term average runoff ratio is presented in Fig. 2. There is generally a higher runoff ratio in mountainous basins in the South, having a runoff ratio up to 0.6, which is due to higher precipitation, lower evapotranspiration and limited soil depths. The basins in the hilly North have lower runoff ratios ranging between 0.3 and 0.4 and are characterised by lower precipitation, higher evapotranspiration and in contrast to the higher basins, larger bodies of groundwater due to unconsolidated rock.

A time series of the runoff ratio is shown, as an example, for the gauge at Lichtenwalde in Fig. 3. The three-monthly running runoff ratio shows a distinct seasonal pattern, while the 2-year running runoff ratio exhibits a low-frequent periodicity. Looking at the spectra of the runoff ratio series, two distinct peaks are generally found, one at an annual and the other at the semiannual frequency (not shown).

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4.2 Variability of the annual phase of runoff ratio

As a next step annual phases and amplitudes have been computed by applying the method of Stine et al. (2009). When using monthly runoff ratios, single events such as summer rain storms or low monthly rainfall and larger streamflow, may prevent a reliable timing estimation and thus increases the variability of the estimated phases. Here the standard deviation of the annual phases estimated for monthly runoff ratios decreased from 23.7 to 17.9 days, when using three-monthly moving runoff ratios. This is mainly due to less extreme years, while keeping the overall phase average (54.4 to 55.2). Descriptive statistics for each basin are given in Table 3, displaying circular average and standard deviations of the annual phase estimates in column $\bar{\phi}_{RR}$.

Similarly to the runoff ratio, its annual phase is also quite dependent on the basin elevation, which can be seen in Fig. 4. Naturally, lower basins appear to have an earlier timing than higher basins, which is due to earlier snow melt in winter/spring. We found a strong linear relation of 5.6 ± 0.3 days per 100 m elevation change.

In Table 3, column *bin*, the results of a basin classification, using the robust method of partitioning around medoids (Kaufman and Rousseeuw, 2005; Hennig, 2010) are provided. The classification is based on the time series of runoff ratios and yield an optimal number of clusters of two. Cluster 2 represents mountainous basins in the Ore Mountains with a snow melt regime and a distinct seasonal cycle peaking in March. The average phase in this group is 1 March (doy 60) and has a standard deviation of 14 days. Basins grouped into cluster 1 appear to have an earlier seasonal cycle with an group average phase at the 13 February. The basins belonging to this group are situated in the lower hilly ranges of Saxony, with a less dominating snow regime.

As independent comparison to the annual phases estimated from RR, the basin average half-flow date and its standard deviation are shown in column 4 of Table 3. In general the half flow dates are later on average, with about 44 days in lower basins and about 35 days in higher elevated basins. Also the differences in standard deviation tend to decrease with higher elevations. Larger differences are found for the circular

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correlation between both timing metrics. Here weak positive correlations are found for the highest basins linearly decreasing with decreasing elevation (R^2 0.2 to −0.5).

Figure 5 presents annual phase estimates of the runoff ratio of representative basins of cluster group 1 and 2. While the annual values have been estimated using the method of Stine et al. (2009), the smoothed phase estimates are derived by complex demodulation using Eq. (2) and a simple moving average filter. Both are converted into doy. The graphs exhibit the temporal evolution of the annual phase and a relatively large year-to-year variability, which is coherent with catchments of the same cluster group. In general a period of larger temporal variability is found in the period 1950–1980. In addition there is a larger difference in the phase between mountainous basins such as Lichtenwalde and lower basins such as Königsbrück. While the higher elevated basins of cluster group 2 appear to have a later than average phase, the annual phases of cluster 1 appear earlier. During other periods the phase difference is smaller combined with lower variability.

Independent from the cluster analysis, a Principal Component Analysis (PCA) (Wilks, 2006) has been performed based on the circular correlation matrix of the annual phase estimates from each basin. Two dominant modes have been found, with 68% and 13% explained variance for the first and second mode, respectively. Thereby the first mode and its principal component generally reflects the mountainous basins of cluster group 1, while the second mode explains most of the temporal evolution of the lower basins in cluster group 1. This result, especially the first dominant eigenmode, supports that mountainous basins have a common temporal evolution of their timing and decadal variability. The first principal component as well as the representative basin Lichtenwalde, show a period of late timing from about 1960–1970, followed by a period with earlier timing but large variability in the 1970s, which is followed by another period of late timing in the 1980s. Since then a trend towards earlier timing is noted. Causes of these trend features will be discussed in the next subsection.

Circular density plots (Lund and Agostinelli, 2010) are a way to depict the empirical frequency distribution of circular data. For two periods of 30 years length (1950–1979

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and 1980–2009) and all basins of cluster 1 and 2, the circular density is presented in Fig. 6. The annual phase of basins of cluster 2 (19 basins) is generally more pronounced, with less late years and its mean is moved towards earlier seasons. In contrast basins of cluster 1 (8 basins) moved towards later seasons and there is a tendency towards less very early years in the second period after 1980.

4.3 Correlation to the annual phase of temperature

Having analysed the phase of the runoff ratio, it is interesting to check for direct links between climatic variables, especially temperature. In Fig. 7 annual average temperatures and the annual phase of temperature for the climate station Dresden are shown. The annual average temperature for this period is about 8.9 °C, while there are increasing temperatures since the end of the 1980s.

The average phase of the annual cycle of temperature for the climate station Dresden is the 20 July (doy 200) and has a long term standard deviation of 4 days. The long term variability of the annual phase of temperature is relatively constant, however, in the 1990's there is a decline in the average of about 5 days, concluding with the most extreme years (2006 very late, and 2007 very early) observed.

To check for direct temperature timing dependence, early years have been classified having annual phases below the first quartile (approx. doy 198) and late years having phases in the last quartile (approx. later than doy 204). In Fig. 8 boxplots of the monthly runoff ratio are depicted for early and late years. One representative basin from cluster 1 and 2 has been selected. As can be seen for the river gauge Königsbrück of cluster 1, larger runoff ratios and larger variability in February to April are apparent in late years compared to early years. At the river gauge Lichtenwalde the differences between early and late years are much more pronounced, with significant differences for the months April till August, with late years having an higher runoff ratio than earlier ones. And the opposite is true for the months October till December. The average monthly temperature, which is superimposed in Fig. 8, reflects these differences. So late years

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are characterised by distinct colder winters, which lead to increased snow accumulation and delay snow melt towards March and April.

Next a direct comparison of the annual phases of temperature and runoff ratio has been done. To assess a direct, linear link between these angular variables, the circular correlation has been computed from the annual phase of the basin runoff ratio and the annual phase of the monthly basin average temperature. The results are detailed for each basin in Table 3, column 5, with significant correlations, i.e. being significantly different from 0 at the $\alpha = 0.05$ level, marked bold. There is significant positive correlation found in mountainous basins belonging to cluster group 2. The correlation decays towards the lower basins. This might be caused by several reasons, the first is of course the minor importance of snow, secondly indirect effects of temperature due to the timing of evapotranspiration and other reasons such as surface – ground water dynamics.

A simple linear regression allows to assess the average effect of a change in the phase of temperature on the timing of the runoff ratio. The slope of the regression line for each basin is reported in Table 3, column 6 and plotted against average basin elevation in Fig. 9. Again, there is a distinct height dependence, which is increasing with 0.37 ± 0.01 per 100 m basin height. For mountainous basins, we found a coefficient of about three in magnitude, which means that a decrease of the phase of temperature of 5 days, amounts to a decrease of 15 days in the timing of the annual hydrological regime.

It is interesting to note that the correlation of the phase of runoff ratio to the annual average basin temperatures is in opposite direction and about half in magnitude. There are also notable correlations with monthly average temperatures, but these change with time. However, the average March temperature shows to have a correlation to ϕ_{RR} of similar magnitude, than ϕ_T.

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4.4 Trend analysis in snow dominated basins

Most of the basins in cluster group 2 appear to have a positive and significant correlation to the annual phase of temperature and have larger parts of their catchment at higher altitudes. It is clear that temperature alone does not influence the runoff ratio, temperature instead acts as a trigger for snow precipitation and snow melt. Therefore, a detailed look at snow depth observations is interesting. Since 1950 a dense network of snow depth observations has been established in Saxony, which is dense enough to compute basin averages.

We checked some annual statistics, such as winter average snow depths and snow cover duration for links to the annual phase of runoff ratio. However, the correlation, e.g. for the basin Lichtenwalde and the corresponding snow depth observations at the station Fichtelberg, is not very high and also not significant different from 0 at the $\alpha = 0.05$ level ($R^2 = 0.19$ for winter average snow depths and $R^2 = 0.25$ for snow cover duration). Instead the average snow depth in March appears to have the largest correlation ($R^2 = 0.55$) and a similar temporal evolution, which can be seen exemplary in Fig. 10, where the phase of runoff ratio at Lichtenwalde/Zschopau is shown with the time series of average March snow depth measured at the station Fichtelberg situated at head of the basin.

Having identified the links of temperature, snow depth and runoff ratio in snow melt influenced basins, it is now investigated, if these variables might explain the trend patterns found in the phase of runoff ratio. As there are no linear trends found over the whole period, decadal variations and deviations from the mean are analysed using cumulative standardised anomalies.

Again the river gauge Lichtenwalde is used to show the results of this approach in Fig. 11. For a stationary variable without any temporal trend or change of the mean it is expected that the line will fluctuate around zero. Here we observe that the annual phase of runoff ratio reaches the significance level in 1971, so there is some evidence to reject the stationarity hypothesis. Beside significance testing the graph also reveals

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the temporal evolution of low frequent changes of the mean. So it can be seen, that there is another peak in 1988, followed by a decline. It is very interesting to note, that the trend pattern of the annual phase of temperature also reveals the 1988 peak, but not the peak in 1971. However, the graph for the snow depth in March reflects both peaks, even though the significance levels are not reached. The CUSUM graph of the annual basin temperatures reveals a significant structural change, also in 1988.

These trend patterns for all three observation variables are found at every basin in the Ore Mountains, belonging to cluster group 2. The results of structural change tests for ϕ_{RR} and determined change points, i.e. peaks in the CUSUM graphs are listed for each basin in Table 3. Thereby the time and the magnitude of departure (referring to the y-axis in Fig. 11) of the two largest peaks are noted. Significant change points based on the OLS-based CUSUM test, i.e. withdrawing the null hypothesis of no change at the $\alpha = 0.1$ level are marked bold. In 6 basins significant change points have been detected, all having their headwaters in the upper Ore mountains. In most of these basins the first peak in 1971 is larger than the following peak in 1988. However, in some basins the 1988 peak, coinciding with a change point in ϕ_T and annual temperatures is more pronounced (Aue, Streckewalde, Zwickau, and to a minor degree Dohna and Neundorf). Within the lower situated basins, there is not such a clear trend pattern, but there seems to be a common negative departure of ϕ_{RR} around 1960.

5 Discussion

5.1 Estimation of the timing of the runoff ratio

The estimation of the annual phase of runoff ratio using the Fourier decomposition provides an objective measure of the timing of the hydrological regime. The fitted harmonic function describes the seasonal fluctuations accurately enough and explains most of the variability of the runoff ratio series. Further the phase angle computed as day of year gives a natural understanding of the timing of the hydrograph/water

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availability during the year. The measure allows to compare the timing of the annual cycle between different hydrological regimes.

Phase estimates derived by the method of complex demodulation (cf. Fig. 5) yield a similar, but smoothed temporal evolution compared to the annual phase estimates using the method of Stine et al. (2009). Moreover, the phase angles have been compared with the simpler measure, the half-flow dates. There has been a good agreement for snowmelt dominated basins, with correlation between both measures increasing with altitude. However, for basins at lower altitude half-flow dates do not seem to be useful on an annual basis. These differences are due to three factors which are in accordance to the conclusions of Déry et al. (2009): (a) choice of the period when half-flow dates are being used, e.g. the hydrological year; (b) sensitivity of Q_{50} to interannual variations in streamflow magnitude; and (c) the less pronounced seasonal cycle at lower basins compared to higher basins leads to significant influences of single events on the half flow dates. Still, the low frequent variability reflected by half-flow dates is similar to the one of ϕ_{RR}.

So, finally with ϕ_{RR} a robust measure for changes in seasonality of hydrological regimes has been gained. This measure may have practical relevance for seasonal water availability and it can be regarded as an interesting regional hydro-climatic signal.

5.2 Variability of the annual phase of runoff ratio

Average runoff ratio, as well as its phase is very much dependent on basin elevation, cf. Table 3 and Fig. 4. It is clear that this height dependence is an effect of increasing relevance of snow storage and release. The natural storage of water in snow affects greater water volumes than any human made reservoir (Mote et al., 2005) and thus changes in snow pack directly effect river runoff. As there is a smooth transition of the properties with basin elevation it is hard to assign the correct group for a few basins. However, two independent methods, (i) cluster analysis based on monthly runoff ratios and (ii) PCA based on the circular correlation matrix of ϕ_{RR} yield two cluster groups/two dominant principal components. These refer to mountainous basins, dominated by

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snow and low land basins, whereby each group has its distinct temporal evolution of ϕ_{RR}. Especially for mountainous basins there is some evidence to believe that there are non-stationary signals in the timing estimate of runoff ratio, cf. Table 3 and Fig. 6,11.

The employed approach in this paper allows to examine shifts in the seasonal cycle, based on the whole annual cycle instead of looking at single months. For CE Stahl et al. (2010) and for the nearby Czech Republic Fiala (2008) found eye-catching trends of increasing streamflow in winter, especially in March, while decreasing in spring months from April to June. From the perspective of an annual cycle these trends imply a change in the phase of the cycle, towards earlier timing of streamflow. Considering the same period (1962–2004) as Stahl et al. (2010), indeed a decreasing trend in the phase of runoff ratio is found for mountainous basins of cluster 2, see Fig. 5.

5.3 Does temperature explain trends in seasonality of runoff ratio?

To answer this question a quantitative correlation analysis and a qualitative assessment of the low frequent variability has been done.

Again the basin elevation distinguishes the different sensitivities to temperature. Most interesting is the correlation of the annual phase of runoff ratio to the annual phase of temperature. Almost all basins appear to have a positive and significant circular correlation, which is again dependent on basin elevation.

It is clear that the sensitivity to temperature found at higher basins is due to snow storage and release in winter. This relationship has been often cited in literature. Barnett et al. (2005) state that rising temperatures possibly lead to earlier timing of the hydrologic regime. We found indeed, that the annual basin temperature is negatively correlated with the annual phase of runoff ratio. However, this correlation is only half in magnitude compared to the phase of temperature. Considering the whole annual cycle, the timing relationship between temperature and runoff ratio is stronger and more relevant than the one with annual temperatures.

In Fig. 11 the CUSUM is plotted for the series of the phase of runoff ratio, the phase of temperature, of March snow depths and the annual average temperature separately.

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The coincidence of peaks in the CUSUM graphs indicate that the respective elements undergo structural changes at the same time and might even explain such a trend. We found that the peak in the late 1980s of the runoff ratio coincides with the phase of temperature, annual average temperatures and the snow depth in March. These changes in temperature may be related to drastic changes towards less air pollution with aerosols over CE since 1980, see e.g. (Philipona et al., 2009).

The other peak in 1971, which might indicate significant nonstationary behaviour, only coincides with the CUSUM graph for snow depths in late winter. That means that the impact of the timing of temperature on the timing of hydrological regimes is one of several relevant processes which in combination explain the above average departure of the phase of runoff ratio. This peak in the CUSUM graph for ϕ_{RR} resulted from two periods with late timing during the 1960s (1961–1965, 1969–1971). During both periods remarkable negative winter time North Atlantic oscillation (NAO) indices (http://www.cgd.ucar.edu/cas/jhurrell/indices.data.html) have prevailed. Negative NAO indices are said to bring cold winter weather to Europe (Thompson and Wallace, 1998), which in effect may result in late snow melt. In addition, there is some evidence of a global climate shift during this period (Baines and Folland, 2007). However, the overall correlation to the winter time NAO index to the first principal component of the phase of runoff ratio is relatively weak (−0.2) and only very extreme and enduring NAO circulation patterns may effect streamflow timing in Saxony.

5.4 Uncertainty and significance of the results

When dealing with real data, uncertainties come into play. For the interpretation of the results it is necessary to list the sources of uncertainty and to examine their relevance.

First of all there may be measurement errors or inhomogeneities in the observed runoff and rainfall series. When we assume that these errors lead to an abrupt but constant change of the mean, the cyclic behaviour and thus the phase is unlikely to be affected.

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In two basins strong inhomogeneities have been detected which are most probably a result of anthropogenic influences such as dam construction and management. Also mining activities have impacted runoff characteristics. Without detailed information it is impossible to correct for such changes. Therefore, these records have been kept in the dataset without a correction. However, we note that the anomalies of the phase of the runoff ratio was similar to basins without such changes in the runoff ratio. That means that the periodic signal has not been altered strongly in most cases.

Another source of uncertainty is the estimation of basin precipitation. Beside the spatial interpolation error, which is ideally normal distributed, we had to face the problem of changes of the observation network over time. To check for effects of this inhomogeneity, three different sets of input stations have been prepared. One set with stations which covered the full period without longer missing periods, another set which consisted of all observations available at a time step and the final set which has been used in the analysis. This set has been a compromise between the other two sets, meeting the requirement that the series used must cover at least 40 years, i.e. from 1950–1990. When comparing the resulting annual phases of these different precipitation input sets, only marginal differences have been found. One reason for this small effect is that the errors of the spatial interpolation do not decrease significantly when the number of stations is increased e.g. from 30 to 290. This has been verified by a cross validation study of the spatial interpolation during periods of high observation network density.

Last there is some uncertainty in the estimation of the timing of the annual cycle using the approximation of a harmonic function to the data. One can see from Fig. 1 that the fitted sinusoids are not always optimal. And very few years completely deviate from a sinusoidal form. To check for the reliability of the detected trends, the method of complex demodulation has been used to estimate the temporal changes of the annual phase, which showed good agreement to the other method.

So we can conclude that beside a range of uncertainties in the various steps of the analysis, the method proved to be quite robust. Moreover, having this large set of more or less independent river basins, the main features are repeated in different catchments

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with similar climatic and hydrologic properties. This last argument underlines that the change points found in the phase of runoff ratio are not random or catchment specific, but a result of a changing climate.

6 Conclusions

A hydro-climatic dataset of monthly values of runoff, precipitation, snow depths and temperature from 1930–2009 has been established for 27 basins throughout Saxony/Germany. Long term seasonal changes of the runoff ratio have been analysed by filtering the annual period using a harmonic function. The variability of the timing of annual cycle of runoff ratio has been central in the analysis. This measure is shown to be applicable to the different hydrological regimes in Saxony.

The main findings are that (i) there are interdecadal changes of the mean and variance of the annual phase and (ii) the timing is directly sensitive to the phase of annual temperatures within snowmelt influenced basins. It has been discussed that the trend in the phase of runoff ratio explains the opposite linear trends in winter and spring found by previous studies on streamflow changes. Moreover these changes are almost at the significance level of rejecting the hypothesis of stationarity of the annual cycle, but the repetition of these trends in different basins may further reduce the probability of this hypothesis. Therefore, it is recommended to be careful when seasonal statistics or removal of seasonality are required. Also hydraulic design studies should acknowledge these structural changes of the annual cycle by using time variant, e.g. dynamic seasonal time series models.

The sensitivity to the timing of temperature interestingly only explains the long term departures of the mean annual timing since 1980. The structural change observed in the early 1970s is most probably related to changes in the late winter snow cover. It would be very interesting, but out of the scope of this study, to investigate how strong these changes relate to changes in atmospheric circulation, air pollution or effects of forest die-back expected in the 1970s trough 1990s. In the course of the discussion

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on climate impacts on hydrological systems, it is notable that an amplified effect of changes in the phase of temperature within snowmelt dominated basins has been found. Moreover the timing of the temperature cycle has more influence on the timing of the runoff ratio than the magnitude of the annual temperature. This study also confirms the previous finding, that impacts of climate change are stronger in mountainous basins. Future climate changes as expected due to continued release of greenhouse gases will probably continue to reduce snow depth in low mountain ranges in CE, with serious consequences in the intra-annual variations in runoff characteristics.

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Table 1. River stations analysed over the period 1930–2009. The column *elev* denotes the mean basin elevation in meters above sea level, *area* denotes catchment area in km^2, RR denotes the long term average runoff ratio and *miss* gives the number of missing months.

station/river	major basin	elev	area	RR	miss
Merzdorf/Döllnitz	Upper Elbe	168	211	0.21	24
Grossdittmannsdorf/Röder	Schwarze Elster	248	300	0.30	36
Koenigsbrueck/Pulsnitz	Schwarze Elster	274	92	0.34	26
Groeditz/Löb. Wasser	Spree	284	195	0.29	12
Elbersdorf/Wesenitz	Upper Elbe	317	227	0.37	0
Bautzen/Spree	Spree	357	276	0.37	24
Niederstriegis/Striegis	Mulde	374	283	0.36	13
Porschdorf/Lachsbach	Upper Elbe	378	267	0.43	0
Kirnitzschtal/Kirnitzsch	Upper Elbe	381	154	0.36	0
Goeritzhain/Chemnitz	Mulde	410	532	0.47	0
Golzern/Mulde	Mulde	481	5442	0.42	12
Nossen/Freib. Mulde	Mulde	485	585	0.43	0
Wechselburg/Zwick. Mulde	Mulde	491	2107	0.46	0
Neundorf/Gottleuba	Upper Elbe	493	133	0.42	0
Mylau/Göltzsch	Weiße Elster	518	155	0.46	12
Dohna/Müglitz	Upper Elbe	555	198	0.46	9
Adorf/Weiße Elster	Weiße Elster	599	171	0.36	35
Lichtenwalde/Zschopau	Mulde	618	1575	0.47	0
Wolfsgrund/Chemnitzbach	Mulde	629	37	0.60	2
Zwickau/Zwick. Mulde	Mulde	631	1030	0.46	12
Borstendorf/Flöha	Mulde	663	644	0.47	0
Pockau/Flöha	Mulde	688	385	0.50	0
Hopfgarten/Zschopau	Mulde	701	529	0.50	0
Niederschlema/Zwick. Mulde	Mulde	705	759	0.52	12
Aue/Schwarzwasser	Mulde	742	362	0.54	0
Streckewalde/Preßnitz	Mulde	744	206	0.47	0
Rothenthal/Natzschung	Mulde	770	75	0.58	0

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Table 2. Inhomogeneous discharge series.

station/river	reason	inhomogeneity
Streckewalde/Preßnitz	dam construction [a] 1973–1976, 55 hm^3	strong inhomogeneity 1980s
Neundorf/Gottleuba	dam construction [b] 1976, 14 hm^3	large in March 1960–1985 and June 1980s
Groeditz/Löbauer Wasser		May 60–70s
Goeritzhain/Chemnitz		change in mean about 1970
Pockau/Flöha	Rauschenbach dam construction 1967 [c], 15 hm^3	weak inhomogeneity
Merzdorf/Döllnitz	Döllnitz dam[d] 3 hm^3	winter 1960–1970, September 1980
Niederstriegis/Striegis		changes in mean 1950–1970
Wechselburg/Zw. Mulde	various dams upstream, largest: Eibenstock dam [e] 1974–1984 83 hm^3	weak, 1950–1970 mainly winter half-year

[a] http://de.wikipedia.org/wiki/Talsperre_Preßnitz
[b] http://en.wikipedia.org/wiki/Gottleuba_Dam
[c] http://de.wikipedia.org/wiki/Talsperre_Rauschenbach
[d] http://de.wikipedia.org/wiki/Talsperre_Dllnitzsee
[e] http://de.wikipedia.org/wiki/Talsperre_Eibenstock

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Table 3. Statistics of the river gauge stations analysed, sorted according to basin elevation. The column *bin* denotes the respective bin assigned by the cluster analysis, $\bar{\phi}_{RR}$ denotes the phase average as calender day and its respective circular standard deviation in days. The average half-flow date is given in column $\bar{Q}_{50}$ and ρ_{cc} is the circular correlation between the annual phases of runoff ratio and temperature. In column *Tcoef* the strength of the linear link between ϕ_{RR} and ϕ_T is denoted as the regression slope coefficient and its standard deviation. The years of the two largest peaks in cumulative standardised anomalies plots and their respective deviation is given in the last 4 columns. Further explanations are given in the text.

station	bin	$\bar{\phi}_{RR}$	$\bar{Q}_{50}$	ρ_{cc}	Tcoef	$peak_1$	σ_{p1}	$peak_2$	σ_{p2}
Merzdorf	1	7 Feb ± 15	28 Mar ± 17	0.11	0.5 ± 0.51	1985	−0.97	1944	0.33
Grossdittmannsdorf	1	12 Feb ± 16	26 Mar ± 23	**0.25**	**1.13 ± 0.54**	1981	−0.93	1960	−0.80
Koenigsbrueck	1	13 Feb ± 15	25 Mar ± 19	0.19	0.82 ± 0.48	1978	−0.84	1959	−0.78
Groeditz	1	11 Feb ± 16	29 Mar ± 24	**0.28**	**1.39 ± 0.54**	1941	−0.56	1960	0.50
Elbersdorf	1	14 Feb ± 16	30 Mar ± 19	0.15	0.76 ± 0.54	1985	−0.94	1960	−0.83
Bautzen	1	12 Feb ± 18	30 Mar ± 20	0.16	0.96 ± 0.64	1961	**−1.28**	1982	−1.14
Niederstriegis	2	14 Feb ± 15	15 Mar ± 22	**0.35**	**1.42 ± 0.48**	1940	−0.45	1973	0.43
Porschdorf	1	21 Feb ± 17	1 Apr ± 20	**0.23**	**1.13 ± 0.57**	1957	−1	1943	−0.86
Kirnitzschtal	1	11 Feb ± 14	1 Apr ± 17	0.21	0.91 ± 0.47	1961	−1.14	1941	−0.58
Goeritzhain	2	15 Feb ± 16	3 Apr ± 22	**0.4**	**1.86 ± 0.49**	1971	0.93	1988	0.69
Golzern	2	25 Feb ± 16	4 Apr ± 21	**0.46**	**2.12 ± 0.47**	1971	0.85	1988	0.76
Nossen	2	21 Feb ± 15	27 Mar ± 21	**0.51**	**2.13 ± 0.42**	1971	0.68	1988	0.60
Wechselburg	2	26 Feb ± 18	12 Apr ± 20	**0.45**	**2.11 ± 0.48**	1971	1.12	1988	0.78
Neundorf	2	25 Feb ± 15	31 Mar ± 24	**0.49**	**2.08 ± 0.43**	1988	0.85	1970	0.71
Mylau	2	20 Feb ± 20	4 Apr ± 24	**0.48**	**2.4 ± 0.55**	1992	0.57	1952	−0.44
Dohna	2	26 Feb ± 13	27 Mar ± 21	**0.53**	**1.93 ± 0.35**	1988	0.68	1971	0.50
Adorf	2	28 Feb ± 17	2 Apr ± 20	**0.46**	**3.18 ± 0.78**	1971	0.71	1988	0.57
Lichtenwalde	2	3 Mar ± 16	4 Apr ± 20	**0.48**	**2.08 ± 0.43**	1971	**1.37**	1988	1.07
Wolfsgrund	2	26 Feb ± 17	31 Mar ± 21	**0.47**	**2.05 ± 0.44**	1958	−0.91	1941	−0.62
Zwickau	2	6 Mar ± 20	12 Apr ± 21	**0.53**	**2.76 ± 0.51**	1988	1.09	1971	0.75
Borstendorf	2	2 Mar ± 16	5 Apr ± 20	**0.52**	**2.28 ± 0.44**	1971	1.16	1988	0.75
Pockau	2	5 Mar ± 17	5 Apr ± 20	**0.48**	**2.12 ± 0.45**	1971	**1.49**	1988	0.85
Hopfgarten	2	6 Mar ± 17	8 Apr ± 20	**0.48**	**2.34 ± 0.48**	1971	**1.46**	1988	1.24
Niederschlema	2	8 Mar ± 19	14 Apr ± 20	**0.51**	**2.53 ± 0.49**	1988	0.95	1971	0.93
Aue	2	14 Mar ± 18	14 Apr ± 19	**0.58**	**3.01 ± 0.47**	1988	**1.34**	1971	1.12
Streckewalde	2	12 Mar ± 18	9 Apr ± 20	**0.53**	**2.62 ± 0.48**	1988	**1.4**	1971	1.23
Rothenthal	2	8 Mar ± 16	7 Apr ± 18	**0.5**	**2.21 ± 0.45**	1971	**1.29**	1988	1.11

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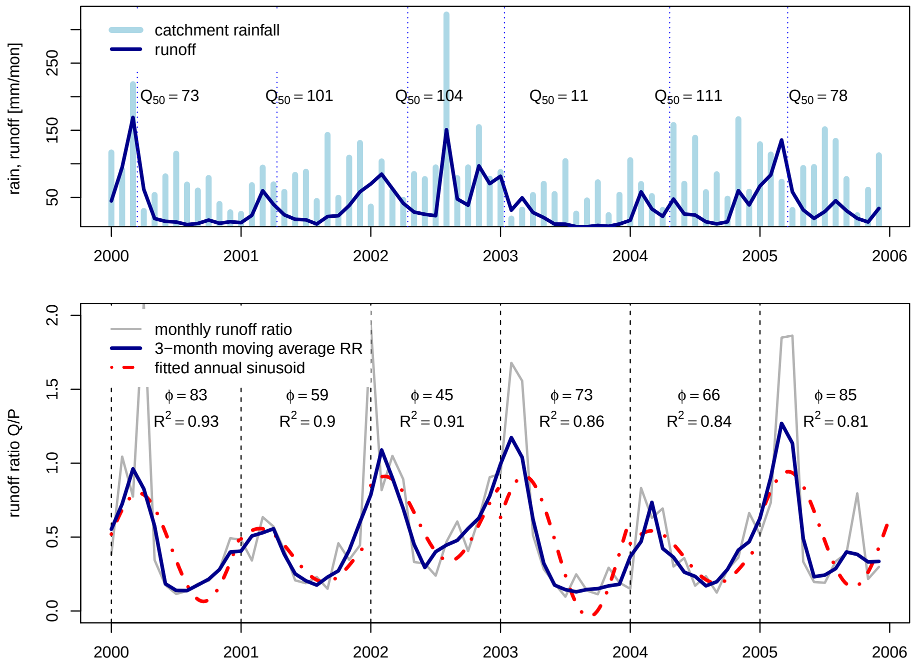

Fig. 1. Top: monthly data of precipitation and runoff of a sample period from the station at Lichtenwalde. The vertical dotted lines depict the half-flow date (Q_{50}) of the respective year and its value is denoted as doy. Bottom: monthly runoff ratio, three-monthly moving runoff ratio and the resulting annual sinusoidal fits. The annual phases ϕ_{RR} are computed as doy and the annual correlation of the fitted sinusoid's to the three-monthly running runoff ratios is given below.

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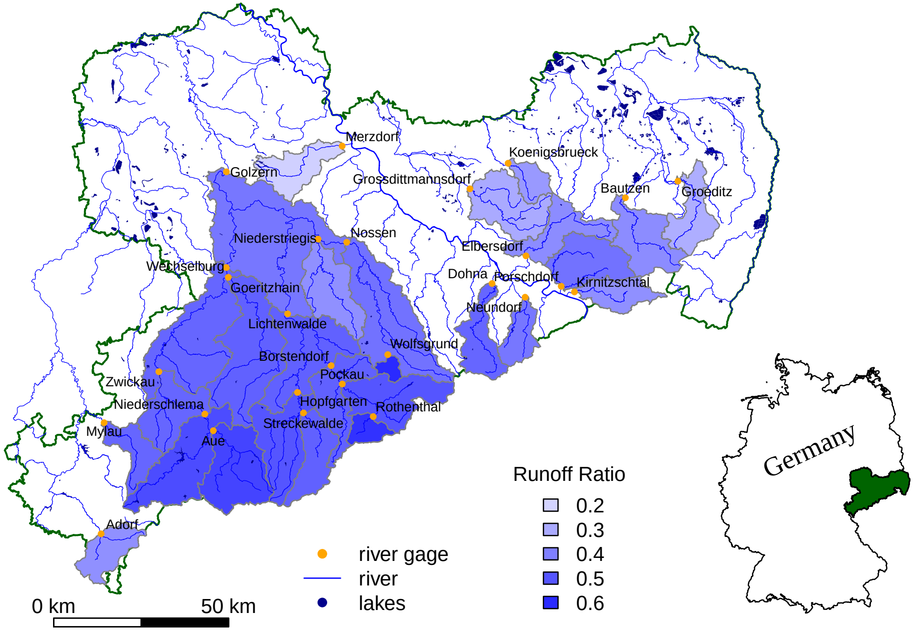

Fig. 2. Map of the study area and long term average basin runoff ratios of the basins investigated.

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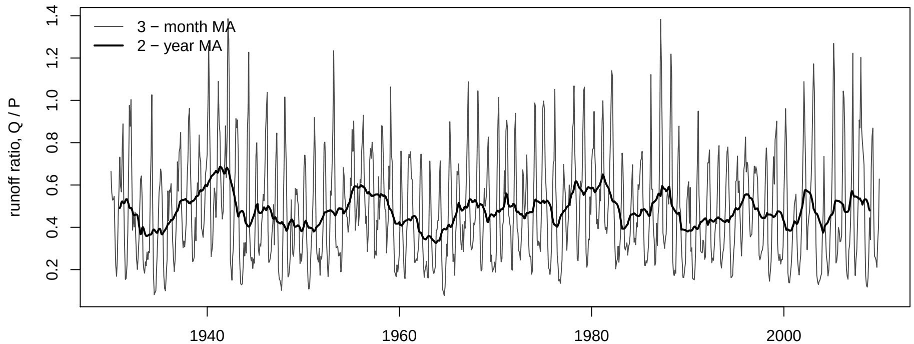

Fig. 3. Time series of the monthly runoff ratio at Lichtenwalde, Zschopau.

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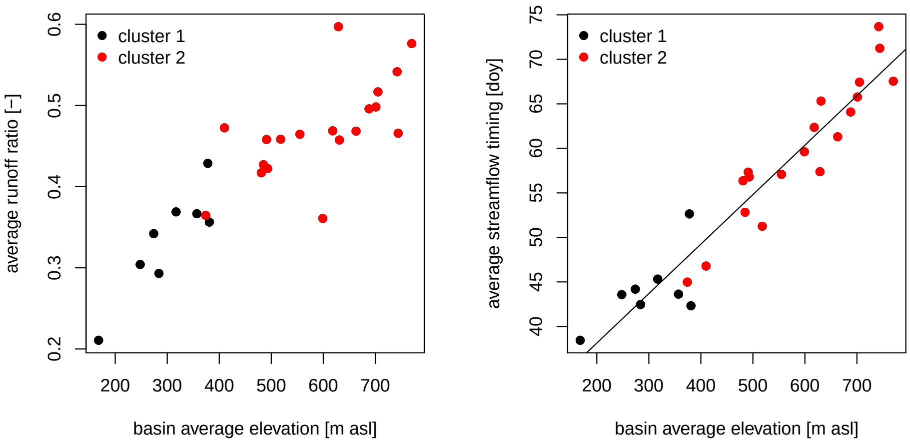

Fig. 4. Height dependence of long term average runoff ratio (left) and dependence of the average streamflow timing (right).

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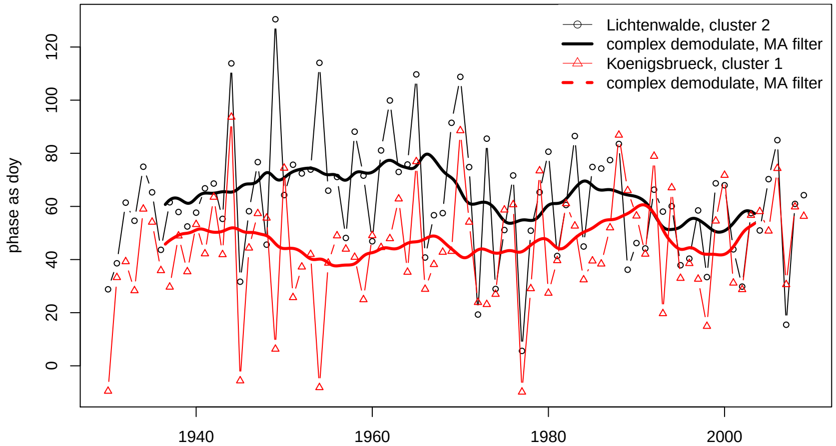

Fig. 5. Time series of the annual phase of runoff ratio at 2 selected representative gauging stations.

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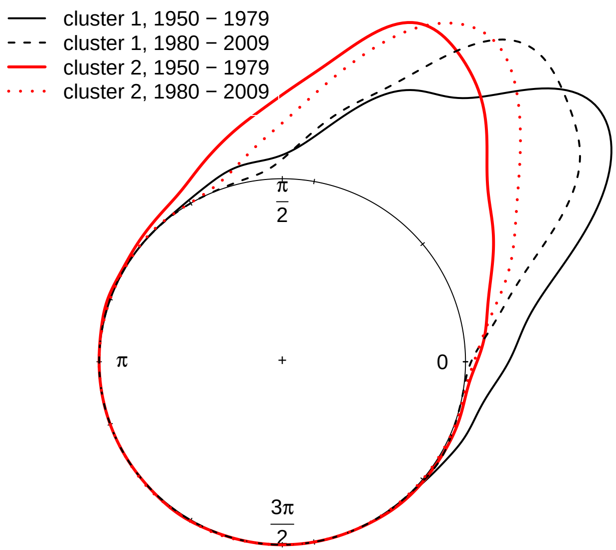

Fig. 6. Circular kernel densities of the phase of runoff ratio, estimated for the basins of cluster 1 and 2 and for two periods each. A bandwidth of 80 has been used to estimate the circular densities. Here angular units are being used, whereby $\pi/2$ equals doy 91 for example.

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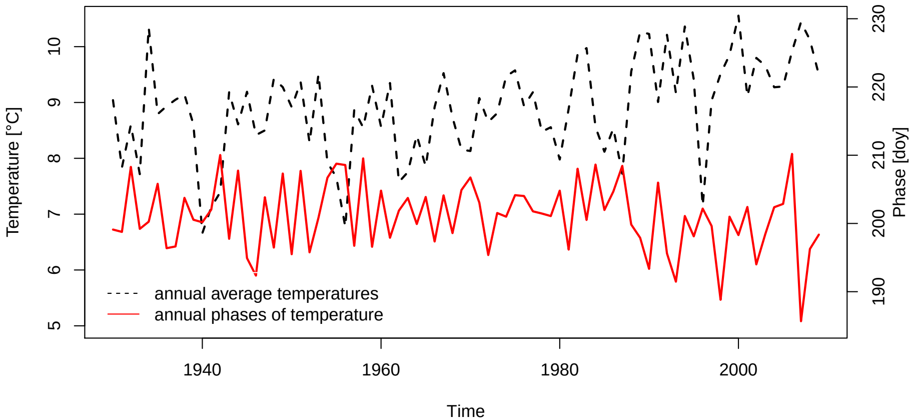

Fig. 7. Annual average temperatures (dashed) and annual phases of temperature ϕ_T of climate station at Dresden.

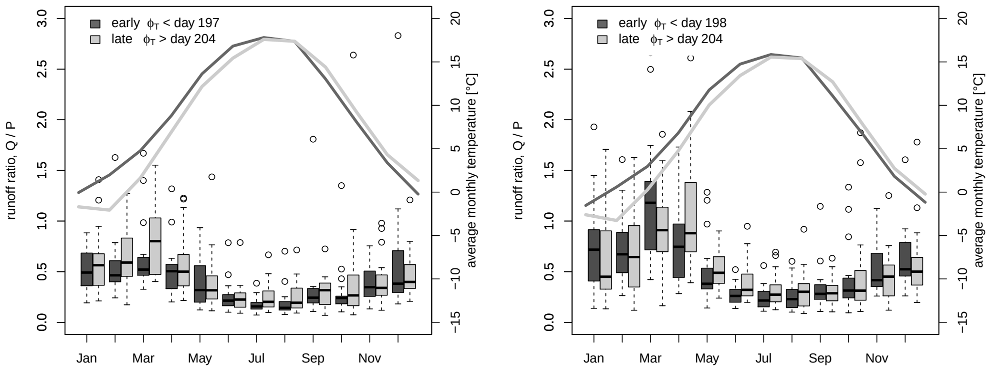

Fig. 8. Boxplots of the seasonal cycle of runoff ratio, depending on early years with the annual phase of temperature below the 1st quartile and late years beyond the 3rd quartile. The bold grey and black lines denote the average monthly temperature for late and early years, with the corresponding axis on the right. Left subplot: Königsbrück, Pulsnitz (cluster 1); Right: Lichtenwalde, Zschopau (cluster 2).

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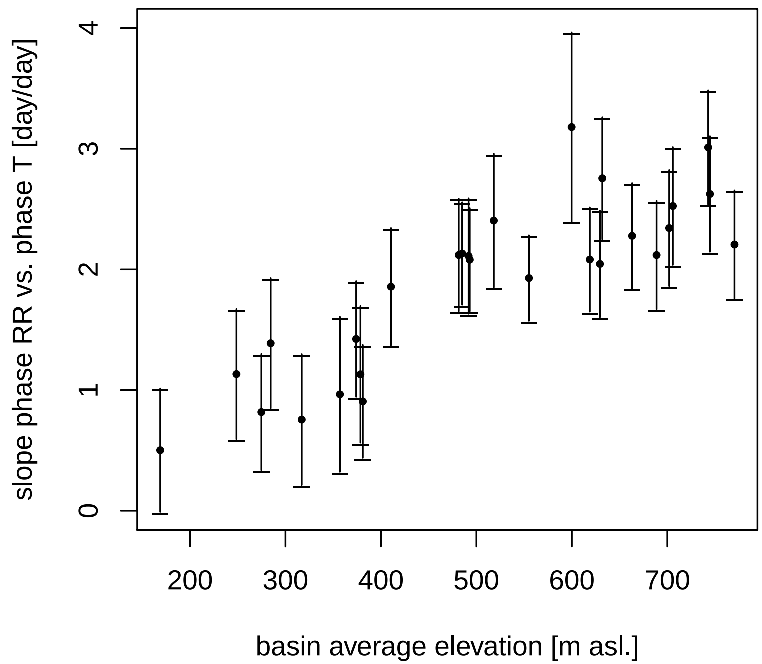

Fig. 9. Height dependence of the regression slope coefficient (± standard deviation) between annual phases of streamflow and temperature.

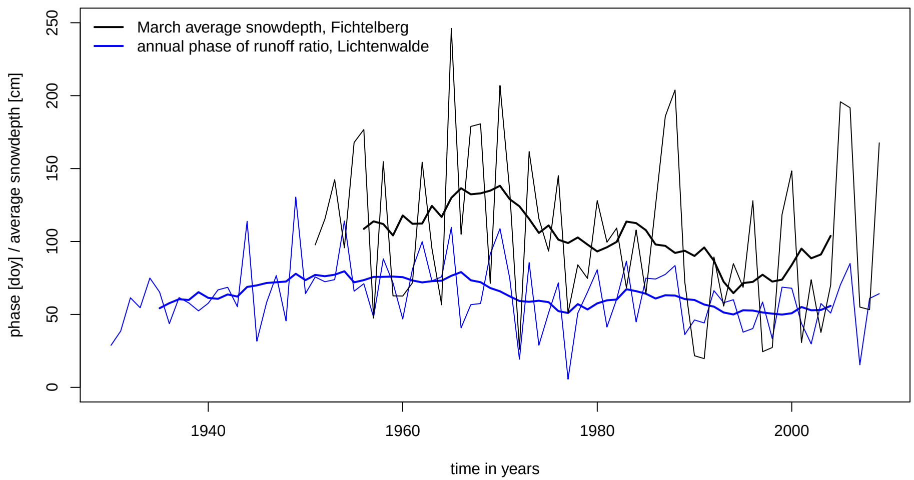

Fig. 10. Annual phase of runoff ratio at Lichtenwalde compared with the March average snowdepth series at the mountain climate station at Fichtelberg. The bold lines depict the 11-year moving averages.

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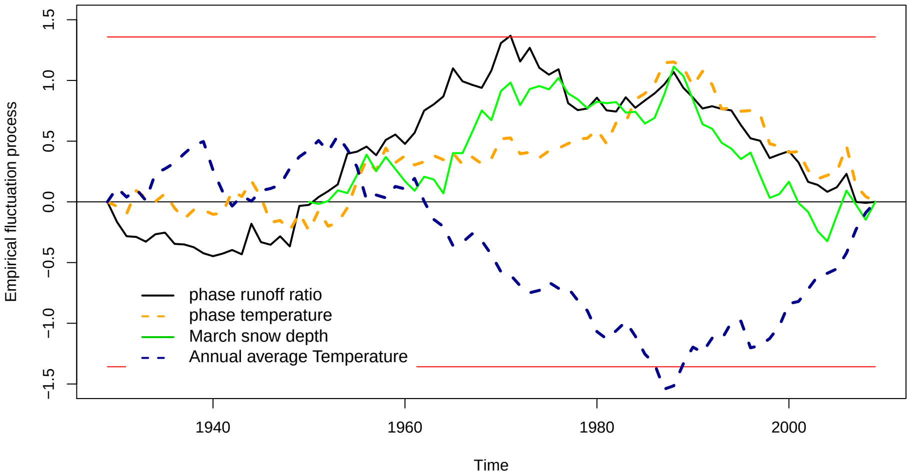

Fig. 11. CUSUM Analysis of the annual phase of runoff ratio at Lichtenwalde compared with the annual phase of temperature, annual basin temperatures and the March average snow depth. The significance levels ($\alpha = 0.05$) for a stationary process are denoted as horizontal lines at the top and bottom of the graph.

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