A Linear System Approach To Convert Long-Term Stochastic Precipitation Into Streamflow

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EXTENDED ABSTRACT

Understanding and simulating the conversion of rainfall (precipitation) to streamflow (discharge) is one of the major tasks in general water resources engineering, be it on the short-scale (hours and days), for the purpose of rainfall-runoff modeling, i.e. flood prediction modeling, or on the long-scale (months and years), to understand seasonal and inter-annual water availability (drought) and, eventually, to look for evidences of climate change. The stream response to precipitation variability results from the passage of infiltrated water through several subsurface filters that, depending on the scale-range, have the effect to either amplify or to attenuate the precipitation's variability modes. Consequently, the output signal (streamflow) can have significantly different spectral characteristics than the input (precipitation). As both the precipitation and the streamflow time-series are usually stochastic in nature, methods of stochastic time-series theory or, more particularly, a stochastic filter approach must be used to quantify the precipitation-streamflow generation process.

The major idea of the stochastic filter is to transform the corresponding time-series into the frequency-domain, with the groundwater aquifer acting as the transfer function whose spectral properties depend on measurable properties of the aquifer, whereas those of the input (rainfall) and of the output (streamflow) time-series can be related in the reciprocal time-domain to their scaling, self-similar or fractal properties. The last are then shedding light on the so-called Hurst-phenomenon which describes the fact that many geophysical times-series possess some amount of auto-correlation, persistence or memory. These properties are usually quantified by the Hurst parameter *H* which ranges, for a fractal (persistent) time-series, $0.5 < H < 1.5$, though for precipitation and river discharge time-series values $0.5 < H < 1$ are most likely encountered. For $0 < H < 0.5$ a time-series is said to be anti-persistent, whereas the case $H = 0.5$ corresponds to a pure random (white-noise) process.

In the present paper we will analyze this phenomenon quantitatively through deterministic, integrated surface- and groundwater modeling of the relevant processes that lead to streamflow generation. The water movement through the hydrological cycle which, starting with precipitation and, after going through soil water infiltration and groundwater recharge, ending its "journey" as baseflow in the main stream, will be simulated with the SWAT model. The long-term groundwater fluctuations and their effects on the baseflow are modeled with MODFLOW whereby the recharging portion of the precipitation as computed by SWAT is used as input into this model. The primary goal is to follow the journey of the low-frequency climate signal through the hydrological cycle and to study its role in the scaling of the affected hydrological time-series. As such we will examine the role of the groundwater aquifer in the transport of the low-frequency climate signal through the groundwater aquifer towards the connected stream. Particular questions of interest: (1) Are the physically based models able to simulate the low-frequency basin's response to the climate forcing found in the river flow observations? (2) How important is the role of the groundwater aquifer in shaping the basin's spectral answer to that of the precipitation input signal? (3) What is the connection between the long-range memory phenomena (H>0.5) of the river flow and the processes involved in the transformation of the input basin signal (precipitation) to the output signal (streamflow)?

The results show that a power shift towards lower frequencies and, consequently, an increase in the estimated Hurst parameter *H* takes place for the modeled heads as the aquifer's lateral dimensions increase and its hydraulic conductivity decreases, all of which appear to be well in agreement with theoretical predictions of stochastic groundwater theory. For very large and low-pervious aquifers the limiting value of 1.5 for the estimated *H* of the groundwater head could be reached which corresponds to a non-stationary long-range memory process called brown noise.

1. INTRODUCTION

Stream response to precipitation variability results from the passage of infiltrated water through several subsurface filters that, depending on the scale range, have the effect to either amplify or attenuate the precipitation variability modes. Consequently, the output signal (streamflow) can have significantly different spectral characteristics than the input signal (precipitation). Particular stages of the hydrological cycle such as the interception and the evapotranspiration may lead to a reduction of the precipitation's high frequency variability while snow melting produces the opposite effect. At the single-event scales which are usually up to several days, the runoff response to the precipitation variability can be quantified from the estimation of the ratio between the minimum wavelet scales of the event hydrograph and hyetograph. At a larger time-scale, like months to decades, the calculation of this ratio would not be very useful because of the noise-induced amplification of the low-frequency signals.

In particular, the climate noise and the noise effects due to the groundwater aquifer's heterogeneity act as a low frequency amplifier. For example, Shun and Duffy (1999) indicated a strengthening of the climate signal at the inter-annual and decadal scales for groundwater-dominated streamflow time-series across the Wasatch Front (Utah) and concluded that groundwater aquifers represent a low pass filter for the precipitation input signal. In the case of a Floridian surficial aquifer, Koch and Cekirge (1996) observed a groundwater amplification of long-period precipitation variations with corresponding effects on effluent river flow that has to be taken into account for the long-term management of the available water resources. It turns out that the reduction of the high frequency-variability is a consequence of the conversion of soil water excess to groundwater recharge, head increases and, eventually, to baseflow.

Quantitative estimations of the baseflow are of particular importance for the long-term management of the possibly available streamwater supply, particular at locations and at times most likely affected by climate change, like parts of Germany (e.g. Marković and Koch, 2005). The analysis of the spectral properties through the determination of the variability scales of the baseflow should be one of the first steps in the estimation of the length of the potential prediction interval. In account of the aforementioned filtering of the high-frequency precipitation components, the baseflow is characterized more likely by a colored than by a flat, white-noise spectrum, since, unlike the latter, the power balance of the former is shifted towards lower frequencies. Zhang and Li (2005) showed that for a white-noise groundwater recharge, the baseflow spectra for homogenous and heterogeneous aquifers are fractal with a power law proportional to f^{β}, whereby β=1.22 and β =1.02, respectively. This indicates that a homogeneous aquifer acts as a stronger low-pass filter than a heterogeneous one. As the parameter β is related to the Hurst parameter H as $\beta = 2H - 1$ (Beran, 1994) these results imply H~ 1.1 for a homogenous and H~1 for a heterogeneous aquifer.

The Hurst parameter is commonly used as an indicator of the time-series' long-range memory, i.e. persistence (Beran, 1994). A value $0<H<0.5$ corresponds to antipersistent, H=0.5 to white, and $0.5<H<1$ to correlated or persistent noise. Moreover, a time-series of H=1 is called pink noise, of H=1.5 brown noise, and of H>1.5 black noise. For precipitation and river discharge time-series $0.5<H<1$ are most likely encountered.

Obviously, the filtering of the high-frequency components in the precipitation signal leads to an increase of the spectral decay parameter β, i.e. a power shift towards lower frequencies and, consequently, an increase of the long-range memory behavior. The remaining part of the precipitation spectrum is then often characterized by peaks at characteristic scales that may be related to large-scale climate patterns like the El Nino Southern Oscillation (ENSO) or, more likely for Germany, the North Atlantic Oscillation (NAO) (e.g. Marković and Koch, 2005; 2007; Marcović et al., 2007).

The aim of this paper is to quantify by deterministic modeling the stochastic and spectral properties found by Marković et al. (2007), for the long-term time-series of both precipitation and discharge within the Elbe river basin. The authors were able to show from the results of Singular Spectrum Analysis (SSA) that, although the major low-frequency modes for the precipitation in the Elbe river basin coincide with those detected for the discharge time-series, the percentage of the variance explained by the annual, inter-annual and inter-decadal cycles is clearly larger for the latter than for the former. This intriguing behavior manifested itself also through higher Hurst parameter estimates H for discharge than for precipitation. It was conjectured that these differences in *H* are due the fact that the high-frequency variability of the precipitation is mostly filtered out during the conversion of the effective rainfall to the soil water excess and subsequent conversion of the latter to the river discharge.

2. DETERMINISTIC AND STOCHASTIC THEORY OF BASEFLOW GENERATION

Short summer rain events usually do not cause groundwater recharge implying that the piezometric heads are not affected by such triggering surface events that act on a short time-scale. On the other hand, groundwater heads establish fingerprints to annual and, of course, inter-annual weather pattern changes with periods of "wet" or "dry" years, triggered, for example, by ENSO (e.g. Koch and Cerkirge, 1996), and more so to inter-decadal climatic variations of the precipitation input. This phenomenon is similar to the effect that long-term climatic temperature changes have on the temperature distribution in the shallow subsurface and it is, indeed, the foundation of the reconstruction of ancient paleoclimatic cycles from measurements of borehole temperatures and, not to the least, the reason for the detrimental consequences the controversially discussed "global change" of the recent past has on the permafrost regions of the Arctic.

Theoretically, all these phenomena are a manifestation of the so-called "skin-effect", which pertains to the low-frequency filtering of the external time-series (precipitation or temperature) by the subsurface. Particularly, short-period variations only propagate into the upper "skin" of the earth while the long-period oscillations penetrate deeper into the earth's crust, leaving a long-term record of the surface signal there that can be used to infer subsurface properties.

Mathematically, the skin-effect is a consequence of the parabolic nature of the diffusion-like PDE that governs the two physical applications above. In groundwater hydrology the underlying PDE is the Richards equation (Bear, 1979) which describes the soil moisture, or water content, $\Theta(z,t)$ as a function of (vertical) space z and time t and needs to be solved with the appropriate boundary condition (BC) $\Theta(0,t)$ at the soil surface. For example, assuming a periodic input signal for the precipitation, the BC becomes $\Theta(0,t)$ =A cos(ωt), with A, the amplitude; $\omega= 2\pi f$, the angular frequency and assuming, to first order, D=*const* and K=*const*., where D is the soil water diffusivity, and K the soil water conductivity, an analytical solution for $\Theta(z,t)$ is

$$\theta(z,t) = A\exp(-\sqrt{\omega/2D}z)\cos(-\sqrt{\omega/2D}z+\omega t) \quad (1)$$

This equation shows that the amplitude of the input signal decreases exponentially with depth whereby the decay factor is proportional to the signal frequency ω. Therefore, the cut-off-frequency (the highest signal frequency) decreases with depth, implying that the groundwater recharge (water that reaches the groundwater level) will most probably have a colored spectrum.

Spectral properties of groundwater head fluctuations for a known recharge spectrum have been studied by Gelhar (1993) and Zhang and Li (2005) for a stream-connected, phreatic aquifer. For such an aquifer, the rate of storage change can be related to the net recharge and the stream-aquifer inflow/outflow as

$$S\frac{\partial h}{\partial t} = R - a(h - h_0) \qquad (2)$$

where S is the specific yield; h *[L]*, the average groundwater head over the entire aquifer; h_0 *[L]*, the water level in the stream; R *[L/T]*, the recharge rate; and a, the outflow constant defined as $a=3T/L^2$; with T $[L^2/T]$, the transmissivity; and L, an appropriate length scale of the aquifer (Gelhar, 1993). A characteristic aquifer time scale t_c can be defined $t_c = S/a$. Assuming a constant river stage h_0 and a homogenous aquifer S, a=*const*, Gelhar (1993) derived the following relationship between the power spectra of the groundwater head fluctuations S_{hh} and those of the recharge S_{RR}:

$$S_{hh} = \frac{S_{RR}}{a^2(1+t_c{}^2\omega^2)} \qquad (3)$$

This equation states that the spectrum of the head is a result of a competitive relation between the geometric and the hydraulic properties of the aquifer (a, S) and of the signal frequency $\omega=2\pi f$. Since $t_c = S/a = SL^2/3T$, one has for a white noise recharge (S_{RR} = *const*), that S_{hh} moves from a white (β=0) towards a colored spectrum (β > 0) when $t_c > 1/f$. In particular, for large L and small T, $a^2 \rightarrow 0$, implying that for a white-noise recharge spectrum $S_{RR} \sim f^0$, the spectrum of the head S_{hh} will be proportional to f^{-2}. This is a special kind of noise called "brown noise" obtained by integrating a white-noise process. The "brown noise" corresponds to a random-walk process in physics called Brownian motion (e.g. Feder, 1989) Because the waveform of the "brown noise"-pattern remains the same, independently of the view-scale, this kind of noise is said to be a fractal or self-similar process. Zhang and Li (2005) confirmed that a white-noise recharge indeed causes fractal hydraulic heads in both, homogenous and heterogeneous aquifers. Since the groundwater aquifer is connected with a stream via the baseflow, the primary objective of the present paper is to understand the connection between the long- range memory of the streamflow data and the spectral properties of the baseflow.

3. ANALYSIS METHODS

3.1 Surface- and groundwater modeling

The examination of the components of the hydrological cycle requires the application of a physically-based, deterministic hydrological model, such as the well-known Soil Water Assessment Tool (SWAT) (Neitsch et al., 2002). SWAT is a spatially distributed model that needs specific information on weather, topography, vegetation, soil properties and land management practices in the basin. It is a continuous time model and enables the simulation of the long-term basin response to arbitrary weather- and soil management conditions. Primary model outputs are the actual evapotranspiration, soil water changes, groundwater recharge, the direct flow and the baseflow. For the automatic calibration of selected parameters SWAT has been coupled with the parameter estimation tool (PEST).

The well-known 3D-MODFLOW96 finite-differences model (Harbaugh and MacDonald, 1996) is applied for the simulation of the groundwater heads and the baseflow for a homogenous stream-connected hypothetical aquifer. Using Darcy's law, the groundwater flow Q_{gw} to the stream can be estimated by calculating the hydraulic gradient between the river stage and the simulated hydraulic head in the first cell next to the river and the river bed conductance.

3.2 Stochastic time-series analysis methods

For the time-series decomposition and the study of the time-series' spectral properties, the Singular Spectrum Analysis (SSA) method and the wavelet tool (Torrence and Compo, 1998) are used, both described in detail in Marković and Koch (2005) and Marković et al. (2007). The same holds for the description of the Detrended Fluctuation Analysis (DFA) method (Peng et al., 1994) employed for the practical estimation of the Hurst parameter describing the long-range memory.

4. STUDY AREA AND DATA

The overall study area is the German part of the Elbe river basin (Fig.1) with a drainage area of 96400 km^2. The discharge data base consists of 38 daily streamflow time-series all over the Elbe river basin with lengths that are greater than 50 yr and cover the time interval 1951-2000. The long-term (1951-2000) average annual basin precipitation is about 715 mm while the long-term discharge in most sub-basins is between 100 and 300mm/yr, with higher values only in the basins of the low mountain ranges and in the lower Elbe River valley. Other interesting details of the flow characteristics of the Elbe River pertaining to some of the objectives of the present study are discussed in Marković et al. (2007).

Daily climate data for use as input in SWAT, namely, precipitation, temperature humidity, wind speed, sunshine hours, etc. for numerous stations (Fig. 1) were provided by the German Weather Service (DWD) for the 1951-2000 time range. The soil, the land use and the DEM data were obtained from other national and regional sources.

5. RESULTS

5.1 SWAT modeling of Striegis River flows

The Striegis River is the right tributary of the Freiberger Mulde River and is situated west of Dresden in the state of Saxony. Its catchment area is about 283 km^2, with a maximum elevation difference of 400m. At the gage Niederstriegis, located 200 m upstream of the Striegis River mouth to the Freiberger Mulde River, the average long-term (1951-2000) daily flow measured is 2.8 m^3 /s. The average annual catchment precipitation is 780 mm and the average annual runoff at the basin outlet is 330 mm. More than 75% of the catchment is agricultural, about 10% is covered by evergreen forest and about 6.3% by med/low-density residential areas. For more than 80% of the catchment, the soil conductivity is classified as very low to low (K=10^{-9} -10^{-5}m/s)

In Fig. 2 (top panel) the mean monthly flows for simulated and measured data for the time interval 1960 to 1980 are illustrated. Except for an overestimation of the largest discharge values, the SWAT-PEST modeled data follow the pattern of the measured data rather well. Furthermore, Fig.2

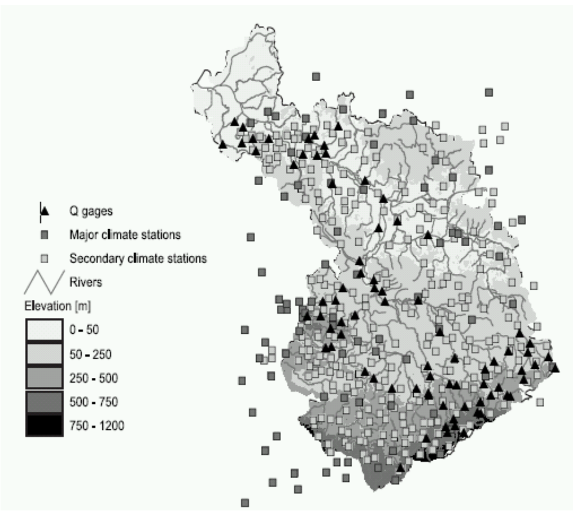

Figure 1. German part of Elbe river basin with climate stations and river gages used.

(bottom panel) shows the normalized global wavelet spectra (nGWS) (Torrence and Compo, 1998) of the simulated hydrological cycle components for time-scales s>2 yr. One notes that the precipitation spectrum is characterized by peaks at the ~7 yr and ~10.4 yr scale which are, however, not statistically significant at the 95% significance level. All the other spectra studied, namely, those pertaining to subsurface variables - obviously not the actual evapotranspiration- do show quite a bit of low-frequency variability.

The apparent and most striking feature of Fig. 2 is that the "journey" of water through the hydrological cycle, from the surface, through the underground, back to the stream, is accompanied by an increase of power at low-frequency scales. Likewise, except for a slight power-increase at scales s>8 yr, both the spectra of the direct flow component and of the percolation have almost the same shape as that of the basin precipitation. The spectrum of the soil water is characterized by a power increase at shorter inter-annual scales (2-4 yr), while the base flow components amplify the

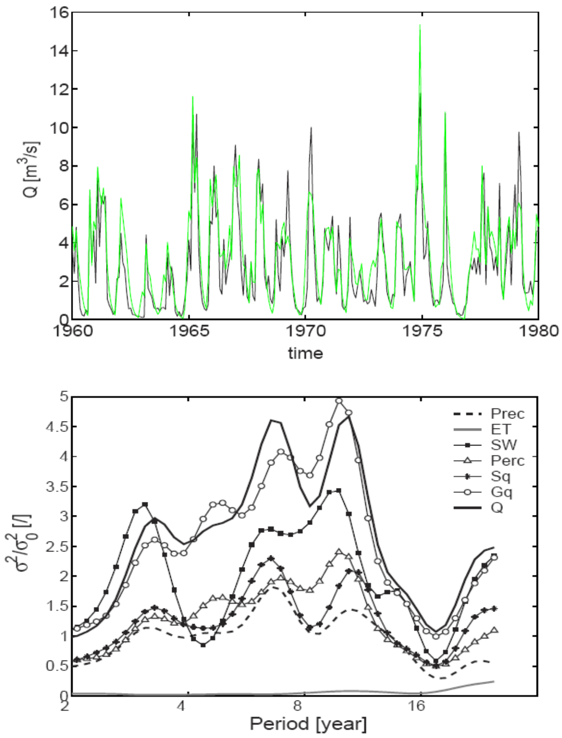

Figure 2. Top panel: Mean monthly Striegis river flows at gage Niederstriegis: measured flows (black), simulated flows (green); right panel: Global wavelet spectra of the hydrological cycle components for the Striegis river basin: precipitation (Prec), actual evapotraspiration (ET), soil water content (SW), groundwater percolation (Perc), direct flow (Sq), base flow (Gq), total flow at the basin outlet-Niederstriegis (Q).

precipitation's low-frequency components and, so, shape the river flow spectrum at the outlet of the basin. Also, the contribution of the direct flow to the low-frequency part of the flow spectrum is small compared to that of the base flow.

5.2 The role of the groundwater aquifer during the transport of low-frequency climate signals

Using the MODFLOW FD-model, transient groundwater flow in a one-layer phreatic aquifer of varying thickness and horizontal size is simulated using a grid size of dx=1m, dy=1m and a time step of one month. As input recharge R, the percolation (PERC) time-series (1953-2000) obtained with the SWAT model of the Striegis river basin is used. The aquifer has a steep gradient towards the river (10%), a specific yield of S = 0.25 and an average water table of 15 m. The left boundary is assumed to be the groundwater divide and the right boundary is constant and equal to the gage elevation in the stream. For each of the three aquifer types, sand, loamy sand and silt, with hydraulic conductivities of 7.1 m/d, 3.5 m/d and 0.06 m/d, respectively, three different lateral dimensions for the aquifer's extensions are simulated: L = 200m, 400m and 800m. The study of Zhang and Li (2005) has shown that the heterogeneity has little effect on the scaling properties of the hydraulic head fluctuation, wherefore all aquifers are assumed to be homogenous.

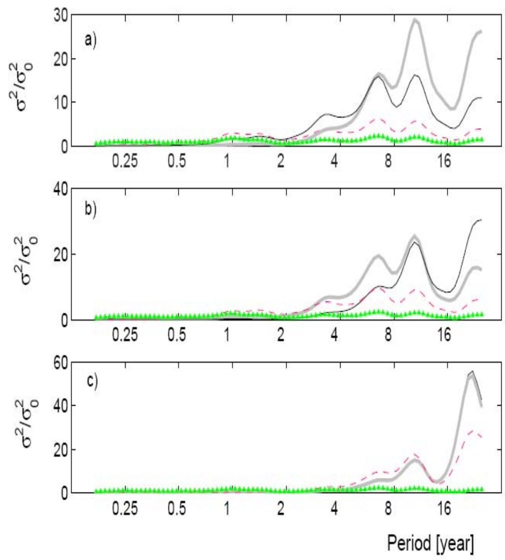

Figure 3. Normalized global wavelet spectra of the estimated groundwater recharge for the period Jan 1953-Dec 2000 (triangle) and the simulated groundwater levels 150m away from the groundwater divide for the sand- (a), loamy sand- (b) and silt-aquifer (c). Aquifer length is set to L=200m (dashed line), L=400m (thin solid line) and L=800m (gray line).

In Fig. 3 the normalized GWS's of the calculated groundwater head fluctuations, 150 m away from the stream, are shown. The apparent property of all spectra is a significant amplification of the low-frequency power of the input signal (recharge) and a negligible power at the annual scale within the output (head) signal. Moreover, the qualitative agreement with the theoretical formula (Eq. 3) is obvious, as an increase of the aquifer's lateral dimension L leads, by virtue of the definition of the characteristic time t_c also to an increase of t_c which, in turn, induces a power shift towards lower frequencies. A similar behavior is expected when the hydraulic conductivity $K \sim T$ is reduced. This is clearly visible from Fig. 3 which shows a shift to lower frequency-scales, when going from the sand, over the loamy sand, to the silt aquifer.

The scaling properties of the MODFLOW simulated head fluctuations are calculated (1) in the spectral domain, using the periodogram and (2) in the time domain using the DFA method.

As the periodogram is the power spectrum $S_{hh}(\omega)$, i.e. the FT of the head's autocorrelation function $\rho_h(\tau)$ (e.g. Priestly, 1981), there is a one-to-one correspondence between the frequency-dependency of $S_{hh}(\omega) \sim |\omega|^{-\beta}$ for $\omega \rightarrow 0$ and the decay law of $\rho_h(\tau) \sim \tau^{-\alpha}$, whereby $\beta = 1 - \alpha = 2H - 1$, i.e. $H = (\beta + 1)/2$ (Beran, 1994). Fig.4a shows the peridogram for a particular simulation of the "groundwater recharge → head fluctuation" process with a set of aquifer properties as specified (for which $t_c >> 1$). One notes that $S_{hh}(\omega) \sim |\omega|^{-\beta}$, with $\beta = 2$, i.e. a Hurst parameter of $H = 1.5$, which corresponds to brown noise. This agrees well with the theoretical prediction of Gelhar's Eq. (3).

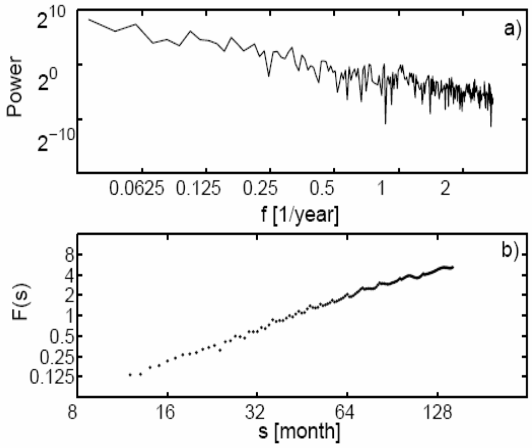

Figure 4. Scaling properties of the simulated ground water levels in a silt aquifer (L=800m), 150m away from the groundwater divide: a) power spectrum as a function of frequency f ($\beta \approx 2$); b) DFA fluctuation function $F(s)$ as a function of scale s ($H \approx 1.5$).

In the (time-domain) DFA method (Peng et al., 1994), which is nowadays the most widely used technique for looking at long-range scaling properties of time-series and for the estimation of the Hurst parameter H, one calculates a so-called fluctuation function $F(s)$ as a function of the scale s ($\sim 1/\omega$) from the detrended variance of the time-series for several non-overlapping segments of length s. One can then show that $F(s)$ scales as $F(s) \sim s^H$, for large s (e.g. Marković and Koch, 2005), which means that H can directly be derived from the slope of a log-log plot of $F(s)$. This is shown exemplarily (for the same set of parameters as before) in Fig. 4b, from which a slope of H ~1.5 is obtained, a result that agrees well with the one deduced from the periodogram in Fig. 4a.

Using this approach in the DFA method, Hurst parameters H_{est} of the monthly hydraulic head fluctuations are estimated at various positions along the topographic gradient that extends from the groundwater divide to the stream. The results of numerous simulations are shown in Fig. 5. Similar to the behavior of the power spectra, there is an increase of H_{est} when L is increased and K decreased, respectively. This figure confirms also that the limiting H_{est} of the groundwater head fluctuations corresponds to brown noise (H=1.5) and such a large value can be found in large, medium-to-low permeable aquifers (see Fig. 5c). For all three simulated aquifer types, H_{est} for the head decreases with decreasing distance L of the groundwater divide from the stream. The average H_{est} (for all L) of the baseflow for sand-, loamy sand- and silt aquifers is 0.80, 0.91 and 1.05 (pink noise), respectively. The "long-range memory" property (H >0.5) of the river flow is then described by the spectra's color of precipitation water on its "journey" through the subsurface compartments of the hydrological cycle.

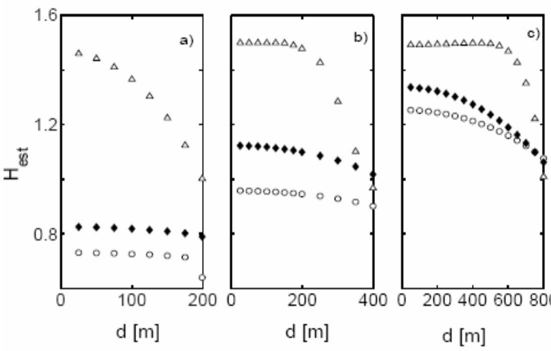

Figure 5. DFA H parameter estimates of the simulated monthly groundwater levels at locations d meters away from the groundwater divide for sand- (circles), loamy sand- (diamonds) and silt aquifer (triangles). The distance L between the groundwater divide and the stream is a) 200, b) 400m and c) 800m.

6. CONCLUSIONS

The present paper addressed the question of how important the role of the groundwater aquifer is in shaping the basin's spectral answer to the spectrum of the precipitation signal? This phenomenon is analyzed by means of deterministic groundwater flow modeling using the MODFLOW model whereby the SWAT-computed recharge in the Striegis river basin serves as input to a hypothetic phreatic stream-connected sand-, loamy sand- and silt aquifer.

The computational results show that there is an increase of the estimated Hurst parameter H of the groundwater head fluctuations with both an increase of the aquifer's lateral dimension L and an decrease of the hydraulic conductivity (K), until a limiting value for H of 1.5, corresponding to brown noise, is reached in large, medium to low permeable aquifers. The results agree well with predictions of stochastic analytical theory.

We find a connection between the long-term persistence ($0.5 < H < 1$) of the river flow and the processes involved in the transformation of the input basin signal- the precipitation- to the output signal- the streamflow. In fact the average H (for all L) of the baseflow is 0.80, 0.91 and 1.05 for sand-, loamy sand- and silt-aquifer, respectively. Obviously, the groundwater aquifer acts as a low pass filter of the precipitation signal resulting in power shifts towards low frequencies and an increase of the groundwater's and, eventually, the streamflow's persistence.

The above analysis qualitatively links amplified low-frequency oscillations to large-scale groundwater and base flow relationships. It is our contention that such subsurface controls are important for the understanding and detection of long-term hydro-climatic change and offer evidence for an underlying explanation of low-frequency variability of streamflow.

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