Relative contribution of land use change and climate variability on discharge of upper Mara River, Kenya

McDonald, M.A.; Mwangi, H.M.; Julich, S.; Patil, S.D.; McDonald, M.S.; Feger, K.H.

Journal of Hydrology: Regional Studies

DOI:
10.1016/j.ejrh.2015.12.059

Published: 12/02/2016

Journal of Hydrology: Regional Studies 5 (2016) 244–260

Contents lists available at ScienceDirect

Journal of Hydrology: Regional Studies

journal homepage: www.elsevier.com/locate/ejrh

Relative contribution of land use change and climate variability on discharge of upper Mara River, Kenya

Hosea M. Mwangi[a,b,c,*], Stefan Julich[a], Sopan D. Patil[b], Morag A. McDonald[b], Karl-Heinz Feger[a]

[a] Institute of Soil Science and Site Ecology, Technische Universität Dresden, Pienner Str. 19, 01737 Tharandt, Germany
[b] School of Environment, Natural Resources and Geography, Bangor University, United Kingdom
[c] Soil, Water and Environmental Engineering Department, Jomo Kenyatta University of Agriculture and Technology (JKUAT), Nairobi, Kenya

ARTICLE INFO

Article history:
Received 1 June 2015
Received in revised form
16 December 2015
Accepted 17 December 2015
Available online 12 February 2016

Keywords:
Climate variability
Land use change
Hydrology
Streamflow
Water security
Budyko framework

ABSTRACT

Study region: Nyangores River watershed, headwater catchment of Mara River basin in Kenya.
Study focus: Climate variability and human activities are the main drivers of change of watershed hydrology. The contribution of climate variability and land use change to change in streamflow of Nyangores River, was investigated. Mann Kendall and sequential Mann Kendall tests were used to investigate the presence and breakpoint of a trend in discharge data (1965–2007) respectively. The Budyko framework was used to separate the respective contribution of drivers to change in discharge. Future response of the watershed to climate change was predicted using the runoff sensitivity equation developed.
New hydrological insights for the region: There was a significant increasing trend in the discharge with a breakpoint in 1977. Land use change was found to be the main driver of change in discharge accounting for 97.5% of the change. Climate variability only caused a net increase of the remaining 2.5% of the change; which was caused by counter impacts on discharge of increase in rainfall (increased discharge by 24%) and increase in potential evapotranspiration (decreased discharge by 21.5%). Climate change was predicted to cause a moderate 16% and 15% increase in streamflow in the next 20 and 50 years respectively. Change in discharge was specifically attributed to deforestation at the headwaters of the watershed.

1. Introduction

Changes in watershed hydrology may have far reaching impacts on a catchment water balance. The changes may be observed through change in water input (precipitation), water distribution into evapotranspiration and runoff, and in the short term, change in catchment water storage (i.e., soil storage and groundwater recharge). Climate variability and human activities are the main drivers of changes in watershed hydrology (Tomer and Shilling, 2009; Ye et al., 2013). At a local scale, change in precipitation may only be caused by changes in climate, while changes in streamflow, evapotranspiration and

* Corresponding author at: Institute of Soil Science and Site Ecology, Technische Universität Dresden Pienner Str. 19, 01737 Tharandt, Germany.
E-mail addresses: hosea.mwangi@mailbox.tu-dresden.de, pathosea2002@yahoo.com (H.M. Mwangi), Stefan.Julich@tu-dresden.de (S. Julich), s.d.patil@bangor.ac.uk (S.D. Patil), m.mcdonald@bangor.ac.uk (M.A. McDonald), karl-heinz.feger@tu-dresden.de (K.-H. Feger).

http://dx.doi.org/10.1016/j.ejrh.2015.12.059

watershed storage may be caused either by climate variability, human activities or both. Changes in streamflow (either total water yield or seasonal discharge) have a major implication on water resources management and especially water supply (Döll and Schmied, 2012; Farley et al., 2011; Charlton and Arnell, 2011). Human activities can alter streamflow through changes in land use, reservoir operation and direct abstraction of surface water or groundwater (Carpenter et al., 2011; Biemans et al., 2011). In absence of reservoirs and inconsiderable water abstractions, land use change and climate variability are the main drivers of change in streamflow (Carpenter et al., 2011). Separation of the impacts of the drivers is helpful in better understanding of the watershed hydrology as well as in developing sound water resources management strategies (DeFries and Eshleman, 2004; Arnell and Delaney, 2006). However, separation and quantification of the drivers' impact is challenging (Zhang et al., 2014; Li et al., 2009; Tomer and Schilling, 2009) because of the complex linkage between climate, human activities and the individual hydrological processes (Falkenmark and Rockström, 2004).

A number of studies have proposed approaches to separate the impacts of land use change and climate variability on streamflow (Li et al., 2012; Wang, 2014). The approaches can be broadly categorized as empirically-based and process-based. Proposed empirical methods are based on climate elasticity (Schaake, 1990) and test the sensitivity of streamflow to changes in climatic factors (Ma et al., 2010). Elasticity-based methods can further be categorized into non-parametric and water balance based methods (Sun et al., 2014). Non-parametric elasticity-based methods are empirical approaches that use linear relationships derived from long-term historical data (Schaake, 1990; Sankarasubramanian et al., 2001; Zheng et al., 2009; Ma et al., 2010). Most of the water balance-based elasticity methods (Dooge et al., 1999; Arora, 2002; Wang and Hejazi, 2011; Roderick and Farquhar, 2011) are based on the concept of the Budyko framework (Budyko, 1974) of catchment water-energy budget (Sun et al., 2014). Process-based methods use distributed physically-based hydrological models where separation is done by alternatively varying and fixing (holding constant) the meteorological inputs and land use/cover conditions (Xu et al., 2014). Process-based methods are more sophisticated, require more data as input and have high uncertainty in parameter estimation whereas non-parametric elasticity methods have weak or no physical meaning (Xu et al., 2014; Wang and Hejazi, 2011). Approaches based on catchment water-energy budgets are easier to use and also have better physical background (Sun et al., 2014; Roderick and Farquhar, 2011).

In this study, we used the catchment water-energy budget approach to separate the contribution of climate variability and land use change on discharge of Nyangores River; the river is a tributary of the trans-boundary Mara River in East Africa. Over the watershed of Mara River, competing land uses and socio-economic activities in the headwaters have been blamed for changes in its hydrological regime (Gereta et al., 2009; Mati et al., 2005, 2008; Dessu and Melesse, 2012). There has been significant deforestation and conversion to agriculture in the upstream regions of the Mara River basin (Mutie et al., 2006). Other studies have also linked observed high level of sediment yield and sedimentation in the Mara River to land degradation following deforestation (Kiragu, 2009; Defersha and Melesse, 2012). A land use change analysis study by Mati et al. (2008) found that the forest cover of 1973 in the Mara basin progressively decreased by 11% and 32% in 1986 and 2000 respectively. For the same periods, open forest increased by 73% and 213% respectively based on the 1973 land cover—a clear indication of the massive deforestation that was taking place in the area immediately after Kenya's independence in 1963. At independence, almost the entire upstream area of the Mara River basin including the Nyangores watershed was covered by dense natural forest and pockets of montane grassland (Government of Kenya—GoK, 1969). Cultivation was limited and strictly controlled by the colonial government (Kanogo, 1987). Mati et al. (2008) used the 1973 and 2000 land use maps to simulate the effect of land use change on hydrology of the Mara River. They found an increase in peak flow during the long rainfall season (March–May) between 1973 and 2000 which they attributed to deforestation in the basin. Mango et al. (2011) simulated deforestation in Nyangores watershed and likewise reported that further deforestation in the watershed may increase peak flows and reduce dry season flows. Based on the findings of these two studies, it can be deduced that deforestation (past or future) lead to increase in peak flows in the watershed. Change in streamflow, however, is not only caused by human activities (particularly land use change) but also by climate variability. Information on how much of observed change in streamflow is separately caused by land use change and climate variability is important for water resources management planning including simulation of informed future land use and climate change scenarios. Analysis of measured historical streamflow data gives valuable evidence-based information of watershed response to past changes in land use and climate variability either individually or in combination. Such information is however lacking for the Mara River basin.

Separation of the contribution from drivers of change in observed streamflow i.e., land use and climate variability is important for integrated watershed management in the Mara River basin. Herein, we focus on Nyangores watershed, one of the headwater catchments of Mara River basin where there has been a major competition between forest conservation and agriculture. The objectives of the study are: (i) to statistically test the presence of a trend in measured streamflow data, (ii) to empirically separate hydrological impacts caused by changes in land use and climate variability from historical streamflow data, (iii) to further partition the contribution of climate variability into that caused by changes in rainfall and potential evapotranspiration respectively, and (iv) to predict the future relative contribution of climate change to streamflow.

2. Materials and methods

2.1. The study area

The Mara River has a unique watershed that is characterized by several spatially-varied land uses: forest conservation and smallholder agriculture in the headwaters, wildlife conservation, pastoralism and large-scale agriculture in the mid-catchment, and mining and smallholder agriculture downstream. The watershed, therefore, is a major contributor to the economy of the region, especially through the wildlife-based tourism in the two national game reserves the watershed hosts (i.e., the Maasai Mara National Reserve and the Serengeti National Reserve). The headwater catchments (Nyangores and Amala) are the lifeline of the Mara River especially in dry weather season when they contribute more than 50% of streamflow (McClain et al., 2014; Dessu et al., 2014).

Nyangores River is a tributary of the Mara River which originates from Mau Forest in Kenya, flows through the Masaai Mara and Serengeti National Reserves in Kenya and Tanzania respectively and finally drains into Lake Victoria (Fig. 1a). Nyangores watershed covers an area of 690 km^2 and is located in the upper part of the trans-boundary Mara River basin (Fig. 1a). Lying at an altitude range of 1900–2970 m above sea level, the watershed main land uses are forest (Mau) and (cropland) agriculture. The main soils are Andosols and Nitisols (World Reference Base—Food and Agriculture Organization of the United Nations classification). The region receives bimodal rainfall pattern with long rains between March and May, and short rains between October and November. The mean annual rainfall is about 1370 mm.

2.2. Data

Daily discharge of River Nyangores recorded over the period 1965–2007 from the gauging station (1LA03) at Bomet town was granted for this study by Kenya Water Resources Management Authority. The meteorological data was obtained from Kenya Meteorological Department. Daily rainfall data was obtained for Bomet water supply, Tenwek mission hospital, Olenguruone District Officer's office and Baraget forest stations (Table 1; Fig. 1a). Monthly average data for temperature (T_{max}, T_{min}) (Fig. 3), wind speed, solar radiation and relative humidity was obtained for Kericho Hail research station (Fig. 1a). Potential evapotranspiration (PET) (Fig. 3) was calculated using Food and Agriculture Organization of United Nations (FAO) Penman–Monteith method (Allen et al., 1998). Several methods for estimation of PET are available in literature, some based on temperature (e.g., Hargreaves and Thornthwaite) and others based on radiation (e.g., Priestley–Taylor) (Tegos et al., 2015; Lu et al., 2005). FAO Penmann–Monteith method is a hybrid method that incorporates all climatic and biological factors affecting evapotranspiration. It has been widely applied in range of climatic conditions and found to give better estimates of PET compared to other methods (Garcia et al., 2004; Cai et al., 2007; Gavilán et al., 2006; Jabloun and Sahli,

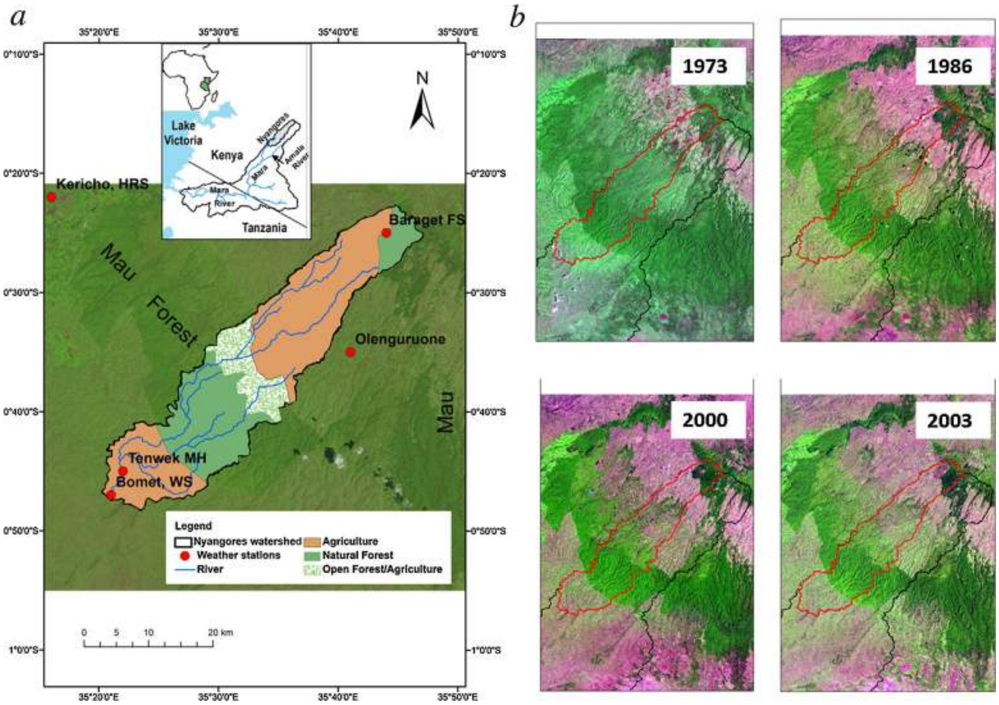

Fig. 1. (a) Nyagores River watershed; (b) Landsat satellite images showing the forest decline in Nyangores watershed. Dark green show the natural forest; light (faded) green and pink show cleared forest and cultivated land respectively.

Table 1
Overview of rainfall data.

Station name	station ID	From	To	% Complete	Annual mean (mm)	SD[a]	CV[a]
Bomet water supply	9035265	1967	2009	88	1363	226	0.17
Olenguruone district officer office	9035085	1960	2002	83	1520	406	0.27
Baraget forest station	9035241	1961	1998	95	1138	235	0.21
Tenwek mission hospital	9035079	1960	2008	93	1448	172	0.12

[a] SD is the standard deviation; CV is the coefficient of variation (SD/Mean).

2008; Ngongondo et al., 2013; Tegos et al., 2015). Though it requires more climatic data than most of the other methods, Allen et al. (1998) outlined a procedure for estimation of PET using FAO Penman–Monteith equation with limited data thus making it applicable in a wide range of conditions (Jabloun and Sahli, 2008; Garcia et al., 2004). Short gaps in the daily discharge data (ca. 5 days) were filled using linear interpolation and inference method using the hydrograph of the adjacent topographically similar Amala River watershed (Fig. 1a) (Rees, 2008); years with long continuous gaps (e.g., 1993–1995) were excluded from the time series analyses. Missing daily rainfall data was filled by arithmetic mean of rainfall recorded for the particular day in the neighboring stations. Average annual areal rainfall for the watershed was estimated by Thiessen polygon method (Szcześniak and Piniewski, 2015; Thiessen, 1911). The daily streamflow data was aggregated into mean annual discharge and was expressed as depth (mm) using Eq. (1) so as to conform to the units (mm) of rainfall (Fig. 2) and PET.

$$\text{Discharge}\left(\frac{\text{mm}}{\text{day}}\right) = \frac{\text{Discharge}\left(\text{m}^3/\text{s}\right) \times (3600 \times 24)}{\text{Watershed area}\left(\text{m}^2\right) \times 1000} \tag{1}$$

2.3. Trend analysis and breakpoint test

2.3.1. Mann Kendall test

The Mann Kendall test (Mann, 1945; Kendall, 1975) was used for trend analysis of the streamflow data. The method has been widely used for trend analyses in hydro-climatic studies (e.g., Zhang et al., 2015; Ye et al., 2013; Ongoma et al., 2013; Xu et al., 2014; Sun et al., 2014). This test is a rank based non-parametric method used for change detection in a time series. It accommodates missing values and outliers, and data with skewed distributions (Partal and Kahya, 2006; Hirsch and Slack, 1984). However, it has been shown that the results of the original version of Mann Kendall method are affected by serial correlations (von Storch, 1995) which may increase the probability of detecting trends when they do not exist and vice versa

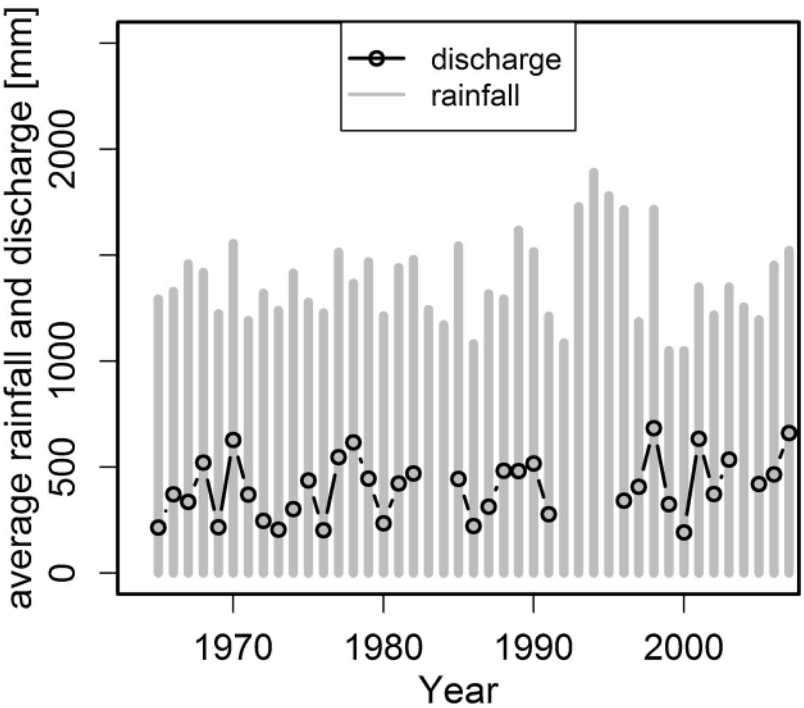

Fig. 2. Annual discharge for Nyangores River (at Bomet town gauging station) and average annual rainfall.

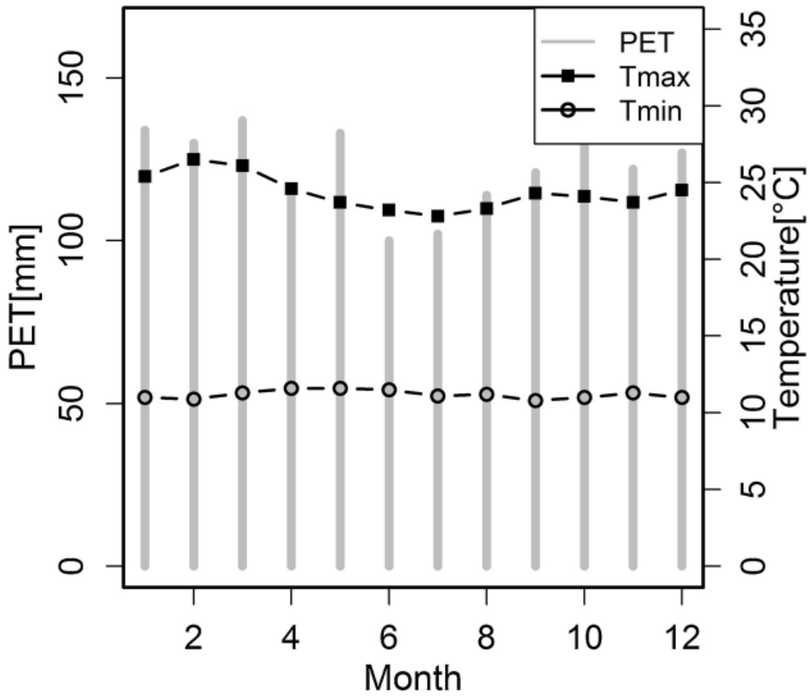

Fig. 3. Monthly temperature (maximum and minimum) and potential evapotranspiration.

(Yue et al., 2002; Hamed and Rao, 1998). Several modifications of the Mann Kendall method have been proposed to limit the influence of autocorrelation in trend analysis of hydro-climatological data (e.g., von Storch, 1995; Hamed and Rao, 1998; Yue et al., 2002; Yue and Wang, 2004; Hamed, 2009). The modifications mainly involve prewhitening (transformation of an autocorrelated series into an uncorrelated one before trend test) or modification of variance (Hamed, 2009; Yue et al., 2002). Each of these approaches have associated strengths and weaknesses as shown by several studies (e.g., Sang et al., 2014; Aissia et al., 2014; Zhang and Zwiers, 2004; Yue and Wang, 2002; Yue et al., 2002; Hamed and Rao, 1998) that have explored their robustness in dealing with autocorrelation. In this study, the method proposed by Hamed and Rao (1998) was used. Hamed and Rao (1998) modified the variance of the original Mann Kendall method based on effective sample size. The results were further verified by the method proposed by Yue and Wang (2004) that is also based on effective sample size but computed from the sample serial correlation estimated from a detrended series. The slope of the trend was estimated using Sen's method (Sen, 1968). The Hamed and Rao (1998) method (just like other versions of Mann Kendall) tests a null hypothesis of no trend in the time series. The time series (herein: annual discharge data) is arranged sequentially in order of (the year) measurement. The magnitude of the discharge for each year x_j ($j = 1, 2, \ldots n$) is compared with the magnitude of discharge of each of the preceding years x_k ($k = 1, 2, \ldots, j-1$), ($j > k$). The sign (sgn), given by Eq. (2), is used to count the difference between the two values (x_j and x_k) from the time series.

$$\operatorname{sgn}\left(x_j - x_k\right) = \begin{cases} 1 & \text{if } x_j > x_k \\ 0 & \text{if } x_j = x_k \\ -1 & \text{if } x_j < x_k \end{cases} \tag{2}$$

The test statitsic S, which is defined as the total sgn of the whole time series is calculated as:

$$S = \sum_{k=1}^{n-1} \sum_{j=k+1}^{n} \operatorname{sgn}\left(x_j - x_k\right) \tag{3}$$

For large series (number of observations, $n \geq 8$), the statistic S is approximately normally distributed with mean and modified variance (Hamed and Rao, 1998) calculated using Eqs. (4) and (5) respectively.

$$E(S) = 0 \tag{4}$$

$$V^*(S) = V(S) \times \frac{n}{n_s^*} = \frac{n(n-1)(2n+5) - \Sigma_{m=1}^{n} t_m m(m-1)(2m+5)}{18} \times \frac{n}{n_s^*} \tag{5}$$

Where *V(S)* is the variance of the original Mann Kendall, t_m is the number of data in a tied group (there is a tie when $x_j = x_k$), m is the number of tied groups, n^*_s is the effective sample size and n/n^*_s is the correction factor due to autocorrelation in the data which is calculated as:

$$\frac{n}{n_s^*} = 1 + \frac{2}{n(n-1)(n-2)} \sum_{i=1}^{n-1} (n-1)(n-i-1)\,\rho_s(i) \tag{6}$$

where n is the actual number of observations and $\rho_s(i)$ is the autocorelation function of the ranks of the observations.

The standardized statistic Z follows a standard normal distribution and is given by:

$$Z = \begin{cases} \dfrac{S-1}{\sqrt{V*(S)}} & \text{if } S > 0 \\ 0 & \text{if } S = 0 \\ \dfrac{S+1}{\sqrt{V*(S)}} & \text{if } S < 0 \end{cases} \tag{7}$$

The null hypothesis of no trend is rejected if the absolute value of Z is bigger than the theoretical value of $Z_{(1-\alpha/2)}$ at α level of significance. A positive value of S indicates an upward trend while a negative value indicates a downward trend.

2.3.2. Sequential Mann Kendall test

The sequential Mann Kendall test (Modarres and Sarhadi, 2009; Sneyers, 1990) was used to detect the occurrence of a breakpoint in discharge. The sequential Mann Kendall test is a graphical technique used to approximate the beginning of a change in a time series based on progressive and retrogressive analysis of the Mann Kendall statistic. Just like in Mann Kendall test, the annual discharge time series is arranged sequentially in order of measurement. The magnitude of the discharge for each year x_j ($j = 1, 2, \ldots, n$) is compared with the magnitude of discharge of each of the preceding years x_k ($k = 1, 2, \ldots, j-1$), ($j > k$). For each time step (year), the number of cases where $x_j > x_k$ is counted. Then the normally distributed statistic t_j is calculated using Eq. (8) where n_j denotes the number of cases where, $x_j > x_k$.

$$t_j = \sum_{i}^{j} n_j \tag{8}$$

The mean and variance of t_j are calculated using Eqs. (9) and (10) respectively and then the progressive variable statistic UF(t_j) (forward sequence) is calculated using Eq. (11). The retrogressive variable statistic UB(t_j) (backward sequence) is calculated with the same Eq. (11) but with a reversed series of the data.

$$E\left(t_j\right) = \frac{j(j-1)}{4} \tag{9}$$

$$\mathrm{Vat}\left(t_j\right) = \frac{j(j-1)(2j+5)}{72} \tag{10}$$

$$\mathrm{UF}\left(t_j\right) = \frac{t_j - E\left(t_j\right)}{\sqrt{\mathrm{Var}\left(t_j\right)}} \tag{11}$$

The intersection of the forward and the backward curves represented by the graphs of statistics UF(t_j) and UB(t_j) respectively indicates the beginning of the step change point (Partal and Kahya, 2006; Ye et al., 2013; Wang, 2014).

2.4. Separating the impacts of land use change and climate variability in runoff

The Budyko framework (Budyko, 1974) was used as the basis to quantify the relative contribution of climate and land use changes to the changes in the watershed hydrology. It is a water and energy balance method that is used to separate the component of precipitation (P) that contribute to evapotranspiration (E) and streamflow (Q). The Budyko hypothesis assumes steady-state water balance conditions of the watershed which require a time scale where change in watershed storage is negligible (e.g., annual basis) (Roderick and Farquhar, 2011). The Budyko curve represents the long-term watershed average evaporative index (i.e., ratio of actual evapotranspiration to precipitation (E/P)) and the aridity index, i.e., ratio of potential evapotranspiration to precipitation (E_0/P) (Donohue et al., 2011). A particular curve has the same catchment property (n) at all point along the curve but with different aridity indices (E_0/P) i.e., different climatic conditions (Sun et al., 2014). Thus, the Budyko hypothesis postulates that under stationary watershed conditions, a watershed will fall on Budyko curve while under non-stationary conditions (with effect of land use changes, i.e., change in catchment property-n) the watershed will deviate from the curve in a predictable manner. The steady state assumption of Budyko hypothesis requires use of long-term average (at least 1 year) of water balance in a watershed (Roderick and Farquhar, 2011; Donohue et al., 2011; Choudhury, 1999). In this study, the water balance was based on average values (P, E and Q) for time period spanning over 44 years separated into two periods based on the year when the change point in the streamflow time series is identified using the sequential

Mann Kendall test. The start of the calendar year coincides with the dry season (January and February) in the watershed thus minimizing the inter-annual change in water storage. The region has minimal 'loss' of water to deep groundwater storage (Dagg and Blackie, 1965; Krhoda, 1988). Water abstraction in the Nyangores River is less than 1% of mean daily discharge (Juston et al., 2014) and there are no significant storage dams on the river (McClain et al., 2014).

This study utilized an empirical model developed by Roderick and Farquhar (2011) to quantify the relative impacts of rainfall, potential evapotranspiration and land use change on change in runoff (discharge). The model is based on empirical Eq. (12) derived from Budyko hypothesis and proposed by Yang et al. (2008) and Choudhury (1999).

$$E = \frac{PE_0}{(P^n + E_0^n)^{1/n}} \tag{12}$$

E is the actual evapotranspiration, P is the precipitation, E_0 is the potential evapotranspiration and n is an empirical catchment characteristic that represent catchment properties.

The Roderick and Farquhar (2011) equation is expressed as:

$$\mathrm{d}Q = \left(1 - \frac{\partial E}{\partial P}\right)\mathrm{d}P - \frac{\partial E}{\partial E_0}\mathrm{d}E_0 - \frac{\partial E}{\partial n}\mathrm{d}n \tag{13}$$

where

$$\frac{\partial E}{\partial P} = \frac{E}{P}\left(\frac{{E_0}^n}{P^n + {E_0}^n}\right) \tag{14}$$

$$\frac{\partial E}{\partial E_0} = \frac{E}{E_0}\left(\frac{P^n}{P^n + {E_0}^n}\right) \tag{15}$$

$$\frac{\partial E}{\partial n} = \frac{E}{n}\left(\frac{\ln\left(P^n + {E_0}^n\right)}{n} - \frac{\left(P^n \ln P + {E_0}^n \ln E_0\right)}{P^n + {E_0}^n}\right) \tag{16}$$

$\mathrm{d}Q$, $\mathrm{d}P$, $\mathrm{d}E_0$ and $\mathrm{d}n$ are the changes in runoff, precipitation, evapotranspiration and catchment properties respectively.

The differential Eq. (13) indicates that change in runoff is a function of climate variability and changes in catchment properties. The change in runoff caused by climate variability ($\mathrm{d}Q^c$) is separated to that caused by change in precipitation and that caused by change in potential evapotranspiration. The last term in Eq. (13) represent the changes in runoff caused by changes in catchment properties. Thus, from Eq. (13) change in runoff caused by change in climate can be estimated as:

$$\mathrm{d}Q^c = \left(1 - \frac{\partial E}{\partial P}\right)\mathrm{d}P - \frac{\partial E}{\partial E_0}\mathrm{d}E_0 \tag{17}$$

Sun et al. (2014) considered the residual change in runoff (dQ^R) to be the difference between the observed change in runoff ($\mathrm{d}Q_{obs}$) and the estimated change in runoff caused by change in climate ($\mathrm{d}Q^c$), and is equivalent to runoff change caused by change in catchment properties (Eq. (18)). The residual change in runoff also includes short-term change in climate variability (i.e., intra-annual climatic effects such as precipitation intensity and temporal distribution of precipitation and potential evapotranspiration) (Sun et al., 2014; Roderick and Farquhar, 2011). Catchment property n cannot be easily measured and its value is usually estimated by fitting it in Eq. (12) using the observed precipitation, potential evapotranspiration and runoff (Donouhe et al., 2011). Thus, changes in runoff caused by changes in catchment properties can be best estimated by Eq. (18) (Sun et al., 2014).

$$\mathrm{d}Q^R = \mathrm{d}Q_{obs} - \mathrm{d}Q^c \tag{18}$$

Eqs. (17) and (18) and were used to calculate the changes in runoff caused by changes in precipitation, evapotranspiration and catchment properties. The relative contribution of each was calculated as a percentage of the observed (total) change in runoff.

2.5. Runoff sensitivity and prediction of future changes in runoff using IPCC projections

The sensitivity of the runoff to climate variability was estimated using Eq. (19), also proposed by Roderick and Farquhar (2011). Eq. (19) predicts the relative change in runoff as a result of unit percent change in precipitation and potential evapotranspiration.

$$\frac{\mathrm{d}Q}{Q} = \left[\frac{P}{Q}\left(1 - \frac{\partial E}{\partial P}\right)\right]\frac{\mathrm{d}P}{P} - \left[\frac{E_0}{Q}\frac{\partial E}{\partial E_0}\right]\frac{\mathrm{d}E_0}{E_0} \tag{19}$$

Table 2
IPCC projected monthly increase[a] in temperature (°C) for the watershed.

Period	2016–2035	2046–2065
December–February	1	1.5
March–May	1	1.5
June–July	1	2
September–November	1	1.5

[a] Based on Representative Concentration Pathway (RCP4.5)—median (50%) of the distribution of Coupled Model Inter-comparison Project Phase 5 (CMIP5)—IPCC, 2013a).

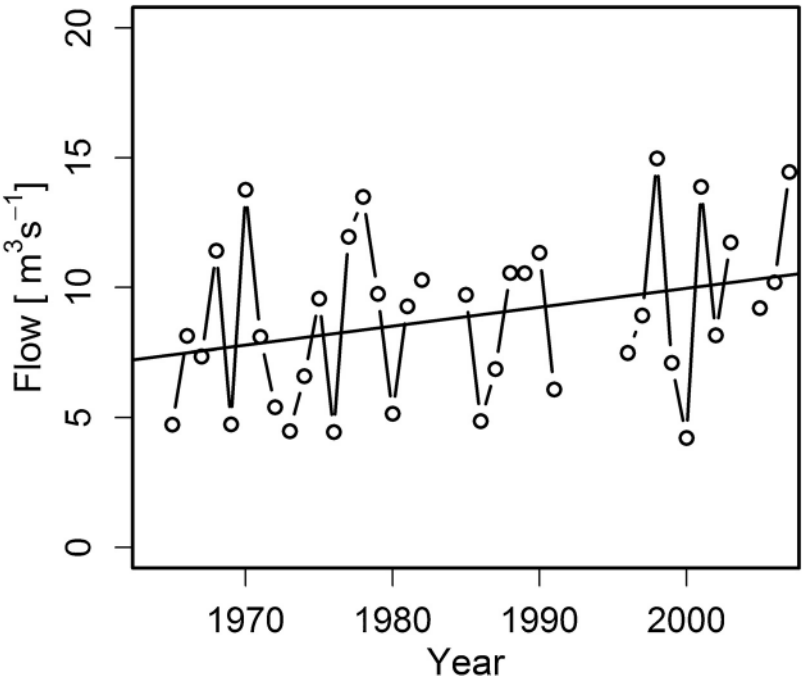

Fig. 4. Annual discharge of the Nyangores River.

Eq. (20) was adapted for the watershed, based on Eq. (19), to predict the sensitivity of runoff to climate change. The equation predicts the expected relative change in runoff based on unit percent change in precipitation, potential evapotranspiration or both.

$$\frac{dQ}{Q} = 2.07\frac{dP}{P} - 1.08\frac{dE_0}{E_0} \tag{20}$$

The Intergovernmental Panel on Climate Change (IPCC, 2013a) projected changes in monthly temperature for the region (Table 2) were then used to calculate the estimated potential evapotranspiration for the watershed in the near-term (2016–2035) and medium-term (2046–2065) periods using FAO Penman–Monteith method (Allen et al., 1998). The calculated changes in potential evapotranspiration and IPCC (2013a) projected changes in precipitation were then applied to Eq. (19) to predict the expected future changes in runoff due to climate change.

The IPCC fifth assessment report (AR5) (IPCC, 2013b) gives patterns of climate change computed from global climate model output gathered as part of the Coupled Model Inter-comparison Project Phase 5 (CMIP5). The climate change projections are made under the Representative Concentration Pathway (RCP) scenarios which are based on more consistent short-lived gases and land use changes. The scenarios specify emissions and are not based on socio-economic driven (SRES) scenarios used in fourth assessment (AR4) which considered future demographic and economic development, regionalization, energy production and use, technology, agriculture, forestry and land use (IPCC, 2013b). The new scenarios for AR5 are based on Radiative Forcing (RF) which quantifies the change in energy fluxes caused by changes in drivers of climate change. RCP4.5 is one of the four RCP scenarios and aims at stabilization of RF at 4.5 W/m^2. The values given in Table 2 are the estimates of the median (50% percentile) of the mean distribution of the 42 models used in CMIP5. More details about the future IPCC climate change projections can be found in the IPCC fifth assessment report (IPCC, 2013b)

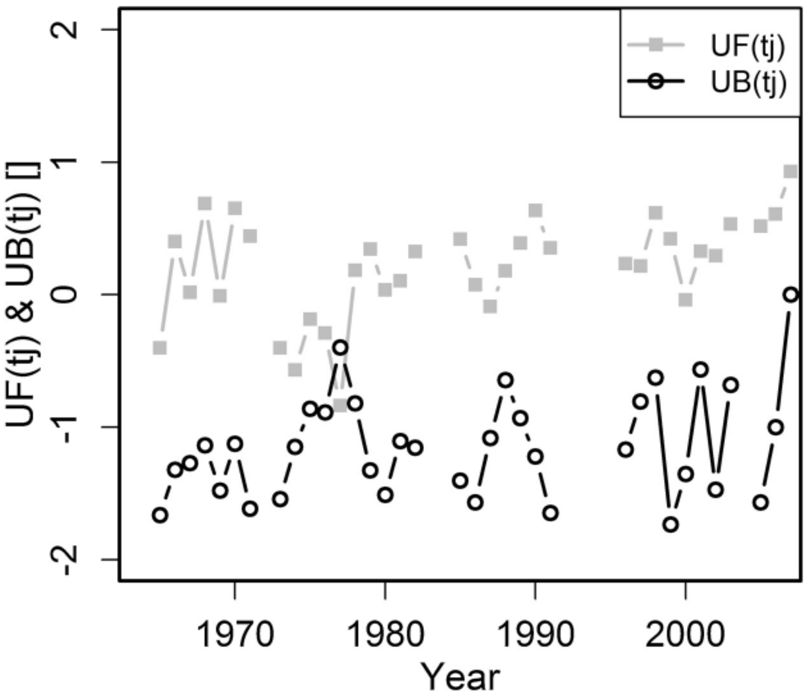

Fig. 5. Sequential Mann Kendall change point test for discharge data. The intersection of forward sequence statistic $UF_{(tj)}$ and backward statistic $UB_{(tj)}$ is the change point in the time series—in this case: 1977.

Table 3
Mean annual values of water balance components (P, Q, E, E_0) for the period before change point, period after change point and the entire (long-term) period, and catchment parameter (n).

	Period before change point(1965–1977)	Period after change point (1978–2007)	Long-term(1965–2007)
Precipitation (P) (mm)	1342	1382	1373
Potential evapotranspiration (E_0) (mm)	1517	1595	1556
Runoff (Q) (mm)	338	439	405
Actual evapotranspiration ($E=P-Q$) (mm)	1004	943	968
Catchment parameter (n)	1.99	1.54	1.75

3. Results

3.1. Changes in measured streamflow

Results from trend analysis of discharge data using the modified Mann Kendall tests (both approaches by Hamed and Rao (1998) and Yue and Wang (2004)) showed an increasing trend (with a slope of 4.75 mm/year) significant at 5% level (Fig. 4). The change point of the discharge data was identified as the year 1977 (Fig. 5) using the sequential Mann Kendall test. Based on the identified breakpoint, the precipitation, potential evapotranspiration and discharge data were split into the period before change point (1965–1977) and the period after change point (1978–2007) as shown in Table 3. This Table also shows the average annual values of potential evapotranspiration calculated using FAO Penman–Monteith equation for the two periods respectively. All the three input parameters to the water balance Eq. (12) were found to have increased between the period before change point and the period after change point. This implies an increase of both the water input (precipitation) and atmospheric demand (potential evapotranspiration) in the catchment. Actual evapotranspiration values were calculated for the two periods as the difference between the averages (averaged over the respective time periods) of measured precipitation and runoff (streamflow) (Table 3). Also shown in Table 3, are the long-term average annual values of the precipitation, potential evapotranspiration, runoff and the actual evapotranspiration covering the entire period (1965–2007) of the study. The long-term values represent the average measured or calculated estimates of the water balance parameters in the catchment.

3.2. Catchment properties parameter (n)

The catchment property (n) for the watershed—estimated by fitting it in Eq. (12) using the long-term mean annual values of precipitation, potential evapotranspiration and streamflow—was found to be 1.75 (Table 3). As reflected in Table 3

Table 4
contribution of climate variability and land use change to change in streamflow.

Driver of change in runoff	Contribution (mm)	Contribution (%)
Precipitation (dQ^P)	+24.4	+24.2
Potential evapotranspiration (dQ^{Eo})	−21.8	−21.6
Climate ($dQ^c) = dQ^P + dQ^{Eo}$)	+2.6	+2.5
Land use (Residual) dQ^R	+98.4	+97.5
Total change (observed) (dQ_{obs})	+101	

dQ^P and dQ^{Eo} are changes in runoff caused by precipitation and potential evapotranspiration respectively.

Table 5
Calculated PET and predicted change in runoff for near-term and medium term periods.

Period	1965–2007	2016–2035	2046–2065
PET (mm)	1556	1621	1638
Change in PET (%) (reference 1965–2007 period)		4.18	5.27
IPCC projected[a] change in precipitation (%)		10	10
Predicted change in runoff (%)—based on Eq. (20)		16	15

[a] Based on Representative Concentration Pathway (RCP4.5)—median (50% percentile) of the distribution of Coupled Model Inter-comparison Project Phase 5 (CMIP5)—(IPCC, 2013a).

and explained in Sections 3.3 and 4.2, the watershed has undergone through major changes in catchment properties and particularly land use changes.

3.3. Hydrological impact of land use change and climate variability

The estimated relative contributions of land use change and climate variability to the observed change in runoff are given in Table 4. The results indicate that the observed increase in precipitation (Table 3) caused a 24% increase in runoff while on the contrary the estimated increase in potential evapotranspiration caused a 21.6% decline in runoff. Therefore, the net change in runoff caused by the climate variability was only an increase of 2.5%. The rest of the observed change in runoff ($dQ^R = 97.5\%$), denoted as the residual change, was caused by changes in catchment properties which is mainly attributed to land use change as discussed in Section 4.2. From the results, we conclude that land use change is the main driver of change of the watershed discharge.

3.3.1. Runoff sensitivity to climate change

Runoff sensitivity Eq. (20) was developed for the watershed. The equation can be used to predict the expected relative change in runoff as a function of change in precipitation and potential evapotranspiration. The equation, for example, predicts that a 10% increase in rainfall would increase runoff by 20.7% while a 10% increase in potential evapotranspiration would reduce the runoff by 10.8%. Thus, it predicts that gain in runoff due to possible increase in rainfall would be minimized by possible increase in potential evapotranspiration.

3.3.2. Expected future response of runoff due climate change

Table 5 shows the calculated future estimates of potential evapotranspiration calculated using the IPCC projected change in temperature (Table 2) for the near-term (2016–2035) and medium-term (2046–2065) periods. The calculated values represent 4.2% and 5.3% increase in potential evapotranspiration for the near-term and medium-term periods respectively. The percentages were calculated based on the average potential evapotranspiration for the 1965–2007 period (Table 5). IPCC (2013a) projected an increase of 10% rainfall in the watershed region for both near-term and medium-term periods as shown in Table 5. The calculated percent change in PET and IPCC projected percent change in rainfall were applied in Eq. (20) to predicted future response of runoff due to climate change, and the results are also shown in Table 5. The results indicate that the streamflow will increase by 16% and 15% for the near-term and medium-term periods due to climate change.

4. Discussion

4.1. Change in streamflow

It was concluded that land use change was the main driver of change in streamflow. The increasing trend in streamflow can be attributed to deforestation and conversion into agriculture in the Mau Forest and particularly the Eastern, South-western and Transmara blocks of the forest (Nkako et al., 2005). The forest blocks are at the headwaters of Nyangores River. Major deforestation and encroachment have been reported in this region. Mati et al. (2008) found that the forest cover in the Mara River basin was reduced by 32% between the years 1973 and 2000 while agriculture doubled over the same period. The Government of Kenya (GoK, 2009) estimated that in the larger Mau Forest complex block (Fig. 1a), the closed canopy declined by 31% between 1973 and 2003 while the area under combined settlements and agriculture increased 5 times over

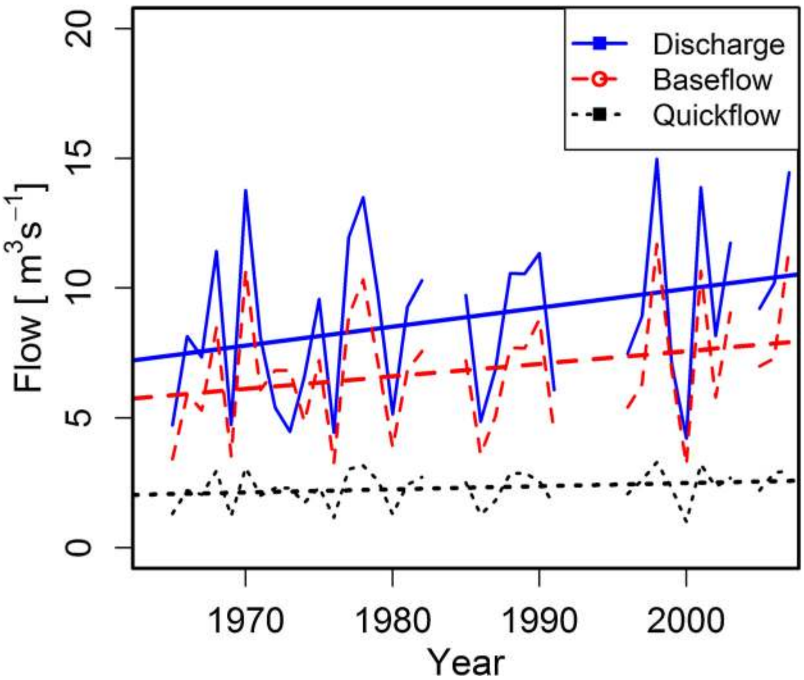

Fig. 6. Temporal trends of annual baseflow, quickflow and total discharge of Nyangores River.

the same period. Catchment water yield is likely to increase upon deforestation and conversion to agriculture although the extent depends on the scale, site and the level of degradation after conversion (Bruijnzeel, 2004; Calder, 2005). Other studies on paired catchment experiments have reported an increase in water yield after deforestation (Bosch and Hewlett, 1982; Mumeka, 1986; Sahin and Hall, 1996; Lal, 1997; Brown et al., 2005, 2013). Our results are also consistent with findings of a paired catchment experimental study by Recha et al. (2012). Their study catchment (Kapchorwa) under tropical rainforest of Nandi and Kakamega is also located within the Lake Victoria Basin in Western Kenya. They reported higher discharge for catchments that were deforested and converted to agriculture; the discharge also increased with time since deforestation.

The observed increase in discharge can be attributed to reduced evapotranspiration after deforestation (Bruijnzeel, 2004). This is because trees are generally known to have higher evapotranspiration than many other land uses, including agriculture (Calder, 2005). Comparatively, forests have higher interception 'losses', greater aerodynamic roughness and deeper roots—all which favour higher water use. The greater canopies of forests enable them to intercept and evaporate more rainfall while the extensive and deeper root network enhances their capacity to extract water from soil and groundwater storages (Bruijnzeel, 2004; Calder, 2005; FAO, 2006). In dry seasons, the tree roots, which are generally deeper than for most vegetation, act as 'pumps' that remove groundwater for transpiration (Bruijnzeel, 2004). Therefore, deforestation generally reduces vegetation water use in a watershed. The reduced 'pumping' of groundwater, particularly in dry seasons, make the water available for discharge inform of baseflow.

In Nyangores watershed, the observed increase in discharge was mainly contributed by increase in baseflow as shown in Fig. 6 where the baseflow, separated using Web-based Hydrograph Analysis Tool (WHAT) recursive digital filter method (Eckhardt, 2012), followed a similar trend to the total discharge. This implies that at the annual level, the reduced evapotranspiration – showing as increased baseflow – is responsible for increased discharge.

The breakpoint of the total annual discharge trend was found to be in 1977. Our findings are supported by Mati et al. (2005) who reported the increase in peak flows in Nyangores watershed starting in the same year, 1977. As shown in Fig. 1, deforestation has been going on progressively in the watershed since the 1970's when there was massive land adjudication of the former communal trust lands in Kenya following the enactment of the Land Adjudication Act of 1968. The residents of Olenguruone section (Fig. 1a) (formerly referred as Olenguruone settlement scheme) applied for land adjudication in 1976 (i.e., Land Adjudication Order, 1976 (Nakuru District)). The Olenguruone area, which is now under intensive cultivation, was formerly under dense natural forest and small pockets of montane open grassland (GoK, 1969; Muiru, 2012); grasslands, just like forest, have higher water infiltration capacities as compared with land under continuous cultivation (Gerla, 2007; Mao and Cherkauer, 2009; Heimann, 2009; Schilling et al., 2014; Everson, 2001). The colonial government that created the Olengurone settlement scheme in 1941 controlled the size and the location of land that the residents cultivated (Kanogo, 1987; Ochieng, 2009; Maxon and Ofcansky, 2014). After independence, in 1963, the restrictions were 'no more' and the locals abandoned the watershed conservation measures, put by colonial masters, which they deemed oppressive. At Olenguruone and the surrounding areas, increased acreages of land, including the hilly slopes, were put under cultivation which further increased with the land adjudication in the 1970s. The dense natural forest cover and the montane grassland in the area were cleared for cultivation and encroachment in the forest reserve started; all of which may have contributed to increase in discharge. Today, the area is under intensive subsistence agricultural cultivation and land ownership is a source of conflict among the ethnic communities living there. Indeed, Mati et al. (2008) found that the forest and grassland in the larger Mara River was basin reduced by 11% and 34% respectively between 1973 and 1986 while the area under open forest and cultivation increased by 73% and 96% respectively during the same period.

4.2. Attribution of changes in streamflow to changes in land use and climate variability

Climate variability was found to have only a minimal (2.5%) contribution to the observed change in discharge (Table 4). This can be attributed to the balance of the water input and atmospheric demand in the watershed. Both the water input (in form of precipitation) and the atmospheric water demand (in form of potential evapotranspiration) increased between the two periods. Thus, the total gain in discharge (24.2%) that would have been made by increased rainfall was reduced (by 21.6%) by the extra atmospheric water demand. On an annual basis, Nyangores can be classified as a water limited watershed (dryness index = 1.1). This implies that the available water (rainfall) does not fully satisfy the atmospheric water demand. The increase in rainfall between the two periods was also accompanied by a relatively higher increase in potential evapotranspiration (due to higher mean temperatures) which further raised the atmospheric water demand (i.e., further increasing the dryness index). Therefore, most of the extra rainfall was used up as evapotranspiration. Taking the effect of climate variability solely, actual evapotranspiration would have been expected to increase in the period after change point. However, as it can be seen in Table 3, the actual evapotranspiration decreased in the period after change point. The reduction in the estimated evapotranspiration between the two periods would then be attributed to change in catchment property (n). The change in catchment properties, occurring concurrently with climate variability, reduced the 'would be' gains in evapotranspiration in favour of increased runoff.

Change in catchment properties was found to be the main driver of the observed changes in runoff accounting for 97.5% of the change. Catchment properties that affect discharge are soil properties, vegetation and topography (Ward and Trimble, 2003; Yang et al., 2008; Price, 2011). Land use change affects these catchment properties and especially the former two in the case of deforestation. Therefore, the change in discharge caused by changes in catchment properties is equivalent to the changes caused by land use in this case. As highlighted in Section 4.1, the major land use changes in Mara River basin is deforestation and conversion to farmland which implies change of vegetation from natural tree vegetation to agricultural crops (mainly maize, beans and potatoes). Other than reduced water use, deforestation also exposes the land to degradation where soil properties are negatively affected eventually leading to reduction in water infiltration and increase in quick runoff. Soil-related factors that lead to decline in infiltration after deforestation include: compaction of top soil (increase bulk density), decrease in soil organic matter (reduce soil aggregation), decline in micro-faunal activity (reduces soil micro-pores), decrease in soil water holding capacities (Giertz et al., 2005; Celik, 2005; Recha et al., 2012).

The future watershed response of low flows to rainfall after deforestation depends on the balance between reduced evapotranspiration and the expected decrease in water infiltration due to degradation. If land degradation reaches a point where water infiltration is reduced to the extent that the quick flows exceeds the gain in baseflow, associated with reduced evapotranspiration after forest removal, then the dry season flows would decline. On the other hand, if the catchment properties do not change, i.e., no or minimal land degradation after forest removal and the original surface infiltration is maintained as before, then the effect of the reduced evapotranspiration may continue to be seen in high baseflow (Bruijnzeel, 2004; Brown et al., 2005). Thus, the observed increase in discharge and baseflow in Nyangores watershed may be short-lived depending on the future level of land degradation. There are already some signs of degradation in the cultivated areas of the watershed that were converted from the forests, as observed by runoff plot experiments by Defersha and Melesse (2012); they reported that cultivated lands in Nyangores watershed yielded higher sediment loads than other watersheds and land uses in the upper Mara River basin. It is also important to recognize that deforestation in the Mau Forest region has been progressive over time with more areas, illegally or legally, being carved out of the natural forest (Akotsi and Gachanja, 2004; Nkako et al., 2005; Akotsi et al., 2006; Mati et al., 2008; GoK, 2009; NEMA, 2013). Therefore, whereas the continued increase in discharge and baseflow may be due to accompanied decline in evapotranspiration, there may be some cultivated areas in the watershed facing high degradation, as observed by Defersha and Melesse (2012), whose response to rainfall may be quite opposite but their effect on baseflow being subdued. It is important therefore that efforts be made to arrest further deforestation and encroachment of the natural forests and more importantly to minimize degradation of the already deforested areas under cultivation.

The residual change in streamflow (dQ^R) may also contain, to a limited extent, change caused by intra-annual climate variability (Roderick and Farquhar, 2011). This is because the catchment property n encodes all factors that change the separation of P into E and Q under constant climate. Hence, other than change in land use discussed in this section, the changes in n over time may also be affected by factors such as changes in precipitation intensity or seasonal changes in precipitation and evapotranspiration (Roderick and Farquhar, 2011; Cuo et al., 2014; Zhang et al., 2015). For example, whereas an increase in dry season rainfall accompanied by an equal decrease in cold season rainfall may have no net change in annual rainfall (Onyutha et al., 2015), it may affect the separation of rainfall into runoff and evapotranspiration (Roderick and Farquhar, 2011). This is because the dry season generally has higher potential evapotranspiration than cold season and thus the change in evapotranspiration (occasioned by change in seasonal rainfall) for the two seasons may not completely balance at an annual scale. Seasonal variability in rainfall can be assessed by, for example, changes in quantiles (Ntegeka and Willems, 2008) or aggregation of rescaled series (Onyutha, 2015). However, since the change in streamflow caused by intra-annual variability is not separated from the residual change in streamflow dQ^R by the current version of Roderick and Farquhar, (2011) method used for this study, the seasonal changes in climate variability was not assessed; the qualitative description of its effect on n provided herein was considered sufficient and useful for further studies. We recommend use of more detailed hydrological models to compare the results obtained in this study.

In unregulated rivers like Nyangores, streamflow seasonality and persistence is more important measure of water availability than the total annual water yield (Döll and Schmied, 2012; Hoekstra et al., 2012; Bruijnzeel, 2004). Change in total water yield may also be accompanied by a change or shift in the seasonal streamflow (Brown et al., 2005; Zhang and Schilling, 2006). Although the study of streamflow seasonality is outside the scope of our paper, recent studies have reported that most downstream sections of the Mara River basin, which heavily rely on flow from the Nyangores River in dry seasons (McClain et al., 2014), are already facing water stress in dry months of the year (Dessu et al., 2014). Thus, further research on the effect of land use change on seasonal streamflow is highly recommended. Change in streamflow seasonality may be assessed by use of monthly/seasonal coefficient of variation (e.g., Zheng et al., 2007; Yang et al., 2009; Patil and Stieglitz, 2011) or non-uniformity coefficient (e.g., Li et al., 2014) and estimated by changes in seasonal/monthly flow duration indices (e.g., Li et al., 2014; Yang et al., 2009; Khaliq et al., 2008; Zheng et al., 2007).

4.3. Future change in runoff due to climate change

The runoff sensitivity Eq. (20) calibrated for the watershed predicts that runoff is more sensitive to changes in precipitation than changes in potential evapotranspiration. Using the projected future climate change scenarios (Tables 2 and 5), the equation predicted that climate change would have a net increase in mean annual streamflow of 16% and 15% in the next 20 and 50 years, respectively (Table 5). The expected gains in discharge due to projected increase in rainfall would be reduced by the predicted increase in evaporative atmospheric water demand (Eq. (20)). The IPCC projected increase in temperature would essentially raise the atmospheric water demand (potential evapotranspiration), which would then buffer the 'expected' gain in runoff due to projected increase in rainfall. The predicted climate change-induced relative change in runoff for the next 50 years is slightly lower than for the next 20 years (Table 5). This is because whereas the IPCC projected an increase of mean monthly temperatures of about 0.5°C between the two periods (Table 2), the rainfall increase remains constant at 10% (Table 5). Thus, the medium-term period would have a relatively higher PET and consequently less climate change-induced change in runoff as compared to the near-term period. The results indicate that direct climate change-induced change in streamflow is relatively moderate (i.e., 15% increase in 50 years). However, climate change may also have an impact in land use and human activities as people to adapt to the changes in climate. As already discussed, land use change has a major impact on both water yield and temporal pattern of streamflow and thus the effect may be greater than predicted.

We used the regional climate change projections based on the distribution of all the 42 models used in CMIP5. The purpose was to roughly show the sensitivity of runoff in the Nyangores based on general future projections. As already discussed in Section 4.2, the runoff sensitivity model developed does not account for the intra-annual variability in climate which may also affect the predictions of runoff (Roderick and Farquhar, 2011). The predictions are thus approximate based on average values. We therefore did not select outputs from any specific GCM nor did we downscale the outputs of the 42 GCMs used in this study. The regional projections in temperature and rainfall used in this study, however, compare well with the values downscaled for the same study area by Dessu and Melesse (2013), and Akurut et al. (2014). Runoff predictions by this simple model are similar to that of the more detailed hydrological model implemented in SWAT by Mango et al. (2011). They reported that a future increase of about 10% in rainfall in the study area will have a modest increase in runoff due to increase in evapotranspiration, driven by accompanying rise in temperature. Unlike the complex hydrological models that demand much effort, data and time, the simple runoff sensitivity equation developed in this study can be easily used by water resources managers in the watershed.

It is also important to recognize the effect the uncertainties arising from the used IPCC future climate projections (Tables 2 and 5) would have on the results obtained in this study. The future temperature values used are based on projections of RCP4.5 scenario. RCP scenarios are based on predicted future forcing (RF) of the climate system by natural and anthropogenic forcing agents such as greenhouse gases, aerosols, solar forcing and land use change (IPCC, 2013b). The RCP4.5 scenario is based on estimated RF of 4.5 W/m^2. However, the RF could fall outside this estimate depending on actual future emissions resulting from forcing agents. IPCC (2013b) gives different projections of temperature and rainfall for other estimates of RF (i.e., RCP2.5, RCP6.0 and RCP8.5) depending on the potential emissions from human activities and/or natural causes (e.g., volcano eruptions). To estimate the range of potential future change in streamflow, based on potential range of change in temperature and rainfall, future runoff prediction was carried out using the projections of the extreme climate change scenarios of RCP2.5 and RCP8.5 for medium-term period. For short-term period projection, the changes in temperature (i.e., 1 °C) and rainfall (i.e., 10%) are uniform across all the three RCP scenarios for the study area and therefore there would be no difference in the predicted change in streamflow (i.e., remains the same as for RCP4.5 (Table 5). As shown in Table 6 and compared with RCP4.5, lower future emissions (RCP2.5) will cause a slight increase in streamflow (to 16%) while higher emissions (RCP8.5) will reduce the potential gain of streamflow to 12.7 %. Thus, the predicted potential increase in

Table 6
Predicted change in runoff based on different IPCC emission projection scenarios.

RCP scenario (for the period 2046–2065)	RCP2.5	RCP4.5	RCP8.5
PET (mm)	1624	1638	1671
Change in PET (%) (reference period: 1965–2007, PET = 1556 mm)	4.4	5.3	7.4
IPCC projected change in precipitation (%)	10	10	10
Predicted change in runoff (%)—based on Eq. (20)	16	15	12.7

runoff of 15% for the 2036–2065 period could fall anywhere in the range between 12.7% and 16.0% depending on the actual future emissions.

5. Summary of results and conclusions

The relative impact of land use change and climate variability on streamflow at the Nyangores watershed in Kenya was investigated. The climate variability impact on streamflow was further partitioned into effects caused by changes in precipitation and those caused by changes in potential evapotranspiration. Future impact of climate change on streamflow was then projected. Quantification of the contributions of the observed change in streamflow of River Nyangores caused separately by land use change and climate variability is one of the main contributions of this study. Though there have been previous studies that have attributed change in hydrology of larger Mara River basin to land use change, information on how much of the observed change in historical streamflow record was caused by either land use change or climate variability has been lacking. Another unique contribution of this study is development of a simple runoff sensitivity equation that can easily be used by water resources managers in the watershed to estimate change in streamflow as a function of change in rainfall and potential evapotranspiration. Main findings and conclusions of the study are:

1. There is an increasing trend in the annual streamflow at the Nyangores watershed. Trend analysis using the Mann Kendall tests detected a significant increasing trend in annual streamflow. The breakpoint for the time series trend was found to be 1977 using the sequential Mann Kendall test.
2. Land use change is the main driver of the change in streamflow accounting for about 97.5% of the change. This can be attributed to the deforestation in the Mau Forest complex at the headwaters of the river. Forest removal and conversion to cropland agriculture caused the increase in streamflow due to reduced water use of crops as compared to forest. We recommend further study on the effect of land use change on seasonal flow regime of the river and its impact on the downstream water availability.
3. Climate variability contributed only a small percentage (2.5%) of the change of streamflow. There was an increase in both precipitation and potential evapotranspiration whose individual effect on streamflow change counters each other (increase in both water input and evaporative demand) resulting to a slight net change in runoff.
4. Streamflow change solely caused by climate change was predicted to increase by 16% and 15% for the next 20 and 50 years respectively. The effect of the predicted increase in rainfall on runoff would be offset, to some extent, by the expected increase in evaporative water demand due to projected increase in temperature. Judging from our findings of the last decades, land use change may still be the major driver of future change in streamflow and may overshadow the predicted impacts of climate change.
5. Deforestation is majorly responsible for change in Nyangores River hydrology. Thus, management measures that control further loss of natural forest and reduce degradation of farmland are required. Thus, the promotion of tree vegetation (e.g., as buffer strips or as integral part of agroforestry systems) may be helpful to mitigate the formation of surface runoff and associated soil erosion.

Conflicts of interest

The authors declare that there are no conflicts of interest.

Acknowledgements

This research was funded by Forest and Nature for Society (FONASO) under the Erasmus Mundus Joint Doctorate Programme (EMJDP). We are grateful to Jomo Kenyatta University of Agriculture and Technology for granting study leave to the corresponding author. We also thank the editor and the anonymous reviewer for their valuable comments.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.ejrh.2015.12.059.

References

Aissia, M.A.B., Chebana, F., Ouarda, T.B.M.J., Roy, L., Bruneau, P., Barbet, M., 2014. Dependence evolution of hydrological characteristics, applied to floods in a climate change context in Quebec. J. Hydrol. 519, 148–163.

Akotsi, E.F.N., Gachanja, M., 2004. Changes in Forest Cover in Kenya's Five Water Towers 2000–2003. Project Report. Department of Resource Surveys and Remote sensing (DRSRS) and Kenya Forest Working Group (KFWG), Nairobi, Kenya.

Akotsi, E.F.N., Gachanja, M., Ndirangu, J.K., 2006. Changes in Forest Cover in Kenya's Five Water Towers 2003–2005. Project Report. Department of Resource Surveys and Remote sensing (DRSRS) and Kenya Forest Working Group (KFWG), Nairobi, Kenya.

Akurut, M., Willems, P., Niwagaba, C.B., 2014. Potential impacts of climate change on precipitation over Lake Victoria, East Africa, in the 21st century. Water 6, 2634–2659.

Allen, R.G, Pereira, L.S., Raes, D., Smith, M., 1998. Crop evapotranspiration—Guidelines for computing crop water requirements. FAO Irrigation and Drainage Paper 56. FAO, Rome.

Arnell, N.W., Delaney, E.K., 2006. Adapting to climate change: public water supply in England and Wales. Clim. Change 78, 227–255.

Arora, V.K., 2002. The use of the aridity index to assess climate change effect on annual runoff. J. Hydrol. 265, 164–177.

Biemans, H., Haddeland, I., Kabat, P., Ludwig, F., Hutjes, R.W.A., Heinke, J., von Bloh, W., Gerten, D., 2011. Impact of reservoirs on river discharge and irrigation water supply during the 20th century. Water Resour. Res. 47, W03509.

Bosch, J.M., Hewlett, J.D., 1982. A review of catchment experiments to determine the effect of vegetation changes on water yield and evapotranspiration. J. Hydrol. 55, 3–23.

Brown, A.E., Zhang, L., McMahon, T.A., Western, A.W., Vertessy, R.A., 2005. A review of paired catchment studies for determining changes in water yield resulting from alterations in vegetation. J. Hydrol. 310, 28–61.

Brown, A.E., Western, A.W., McMahon, T.A., Zhang, L., 2013. Impact of forest cover changes on annual streamflow and flow duration curves. J. Hydrol. 483, 39–50.

Bruijnzeel, L.A., 2004. Hydrological functions of tropical forests: not seeing the soil for the trees? Agric. Ecosyst. Environ. 104, 185–228.

Budyko, M.I., 1974. Climate and Life. Academic, San Diego, CA.

Cai, J., Liu, Y., Lei, T., Pereira, L.S., 2007. Estimating reference evapotranspiration with the FAO Penman–Monteith equation using daily weather forecast messages. Agric. Forest Meteorol. 145, 22–35.

Calder, I.R., 2005. Blue Revolution—Integrated Land and Water Resources Management, 2nd ed. London, Earthscan.

Carpenter, S.R., Stanley, E.H., Zanden, M.J.V., 2011. State of the world's freshwater ecosystems: physical, chemical, and biological changes. Annu. Rev. Environ. Resour. 36, 75–99.

Celik, I., 2005. Land-use effects on organic matter and physical properties of soil in a Southern Mediterranean highland of Turkey. Soil Tillage Res. 83, 270–277.

Charlton, M.B., Arnell, N.W., 2011. Adapting to climate change impacts on water resources in England—an assessment of draft water resources management plans. Glob. Environ. Change 21, 238–248.

Choudhury, B.J., 1999. Evaluation of an empirical equation for annual evaporation using field observations and results from a biophysical model. J. Hydrol. 216, 99–110.

Cuo, L., Zhang, Y., Zhu, F., Liang, L., 2014. Characteristics and changes of streamflow on the Tibetan Plateau: a review. J. Hydrol.: Reg. Stud. 2, 49–68.

Dagg, M., Blackie, J.R., 1965. Studies of the effects of changes in land use on the hydrological cycle in East Africa by means of experimental catchment areas. Int. Assoc. Sci. Hydrol. Bull. 10 (4), 63–75.

Defersha, M.B., Melesse, A.M., 2012. Field-scale investigation of the effects of land use on sediment yield and runoff using runoff plot data and models in the Mara River basin, Kenya. Catena 89, 54–64.

DeFries, R., Eshleman, K.N., 2004. Land-use change and hydrologic processes: a major focus for the future. Hydrol. Process. 18, 2183–2186.

Dessu, S.B., Melesse, A.M., 2012. Modelling the rainfall–runoff process of the Mara River basin using the soil and water assessment tool. Hydrol. Process. 26, 4038–4049.

Dessu, S.B., Melesse, A.M., 2013. Impact and uncertainties of climate change on the hydrology of the Mara River basin, Kenya/Tanzania. Hydrol. Process. 27, 2973–2986.

Dessu, S.B., Melesse, A.M., Bhat, M.G., McClain, M.E., 2014. Assessment of water resources availability and demand in the Mara River Basin. Catena 115, 104–114.

Döll, P., Schmied, H.M., 2012. How is the impact of climate change on river flow regimes related to the impact on mean annual runoff? A global-scale analysis. Environ. Res. Lett. 7, 014037.

Donohue, R.J., Roderick, M.L., McVicar, T.R., 2011. Assessing the differences in sensitivities of runoff to changes in climatic conditions across a large basin. J. Hydrol. 406, 234–244.

Dooge, J.C.I., Bruen, M., Parmentier, B., 1999. A simple model for estimating the sensitivity of runoff to long-term changes in precipitation without a change in vegetation. Adv. Water Resour. 23, 153–163.

Eckhardt, K., 2012. Technical note: analytical sensitivity analysis of a two parameter recursive digital baseflow separation filter. Hydrol. Earth Syst. Sci. 16, 451–455.

Everson, C.S., 2001. The water balance of a first order catchment in the montane grasslands of South Africa. J. Hydrol. 241, 110–123.

Falkenmark, M., Rockström, J., 2004. Balancing Water for Humans and Nature. The New Approach in Ecohydrology. London, Earthscan.

FAO, 2006. New Generation of Watershed Programmes and Projects. FAO Forestry Paper 150. Food and Agriculture Organization of the United Nations (FAO), Rome.

Farley, K.A., Tague, C., Grant, G.E., 2011. Vulnerability of water supply from the Oregon Cascades to changing climate: linking science to users and policy. Glob. Environ. Change 21, 110–122.

Garcia, M., Raes, D., Allen, R., Herbas, C., 2004. Dynamics of reference evapotranspiration in the Bolivian highlands (Altiplano). Agric. Forest Meteorol. 125, 67–82.

Gavilán, P., Lorite, I.J., Tornero, S., Berengena, J., 2006. Regional calibration of Hargreaves equation for estimating reference ET in a semiarid environment. Agric. Water Manag. 81, 257–281.

Gereta, E., Mwangomo, E., Wolanski, E., 2009. Ecohydrology as a tool for the survival of the threatened Serengeti ecosystem. Ecohydrol. Hydrobiol. 9, 115–124.

Gerla, P.J., 2007. Estimating the effect of cropland to prairie conversion on peak storm run-off. Restor. Ecol. 15, 720–730.

Giertz, S., Junge, B., Diekkrüger, B., 2005. Assessing the effects of land use change on soil physical properties and hydrological processes in the sub-humid tropical environment of West Africa. Phys. Chem. Earth 30, 485–496.

GoK, 1969. Vegetation Map (1:250,000)—Sheet No. 3—prepared by British Government's Ministry of overseas Development (Directorate of Overseas Surveys—D.O.S. (L.R.) 3006). Government of Kenya (GoK), Survey of Kenya, Nairobi.

GoK, 2009. Report of the government's task force on the conservation of the Mau forest complex. Government of Kenya (GoK), Nairobi.

Hamed, K.H., 2009. Enhancing the effectiveness of prewhitening in trend analysis of hydrologic data. J. Hydrol. 368, 143–155.

Hamed, K.H., Rao, A.R., 1998. A modified Mann–Kendall trend test for autocorrelated data. J. Hydrol. 204, 182–196.

Heimann, D.C., 2009. Comparison of Hydrologic and Water-Quality Characteristics of Two Native Tallgrass Prairie Streams with Agricultural Streams in Missouri and Kansas: U.S. Geological Survey Scientific Investigations Report 2009–5213, 39 p.

Hirsch, R.M., Slack, J.R., 1984. A non-parametric trend test for seasonal data with serial dependence. Water Resour. Res. 20, 727–732.

Hoekstra, A.Y., Mekonnen, M.M., Chapagain, A.K., Mathews, R.E., Richter, B.D., 2012. Global monthly water scarcity: blue water footprints versus blue water availability. PLoS One 7, e32688, http://dx.doi.org/10.1371/journal.pone.0032688.
IPCC, 2013. Annex I. Atlas of Global and Regional Climate Projections Supplementary Material RCP4.5 [van Oldenborgh, G.J., Collins, M., Arblaster, J., Christensen, J.H., Marotzke, J., Power, S.B., Rummukainen M., Zhou T., (eds.)]. in: Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., Midgley P.M. (eds.), Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. Available from www.climatechange2013.org and www.ipcc.ch.
IPCC, 2013. Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., Midgley P.M. (eds.)]. Cambridge University Press, Cambridge and New York.
Jabloun, M., Sahli, A., 2008. Evaluation of FAO-56 methodology for estimating reference evapotranspiration using limited climatic data: application to Tunisia. Agric. Water Manag. 95, 707–715.
Juston, J., Jansson, P.E., Gustafsson, D., 2014. Rating curve uncertainty and change detection in discharge time series: case study of 44-year historic data from the Nyangores River, Kenya. Hydrol. Process. 28, 2509–2523.
Kanogo, T., 1987. Squatters and the Roots of Mau Mau. East Africa Publishing House, Nairobi.
Kendall, M.G., 1975. Rank Correlation Measures. Charles Griffin, London.
Khaliq, M.N., Ouarda, T.B.M.J., Gachon, P., Sushama, L., 2008. Temporal evolution of low flow regimes in Canadian rivers. Water Resour. Res. 44, W08436.
Kiragu, G.M., 2009. Assessment of sediment loadings and their impact on the environmental flows of the upper Transboundary Mara River, Kenya. MSc. thesis. Jomo Kenyatta University of Agriculture and Technology, Kenya.
Krhoda, G.O., 1988. The impact of resource utilization on the hydrology of the Mau Hills forest in Kenya. Mt. Res. Dev. 8, 193–200.
Lal, R., 1997. Deforestation effects on soil degradation and rehabilitation in Western Nigeria. IV Hydrology and water quality. Land Degrad. Dev. 8, 95–126.
Li, Z., Liu, W.Z., Zhang, X., Zheng, F., 2009. Impacts of land use change and climate variability on hydrology in an agricultural catchment on the Loess Plateau of China. J. Hydrol. 377, 35–42.
Li, H., Zhang, Y., Vaze, J., Wang, B., 2012. Separating effects of vegetation change and climate variability using hydrological modelling and sensitivity-based approaches. J. Hydrol. 420–421, 403–418.
Li, F., Zhang, G., Xu, Y.J., 2014. Spatiotemporal variability of climate and streamflow in the Songhua River Basin, northeast China. J. Hydrol. 514, 53–64.
Lu, J., Sun, G., McNulty, S.G., Amatya, D.M., 2005. A comparison of six potential evapotranspiration methods for regional use in the Southeastern United States. J. Am. Water Resour. Assoc. (JAWRA) 41, 621–633.
Ma, H., Yang, D., Tan, S.K., Gao, B., Hu, Q., 2010. Impact of climate variability and human activity on streamflow decrease in the Miyun Reservoir catchment. J. Hydrol. 389, 317–324.
Mango, L.M., Melesse, A.M., McClain, M.E., Gann, D., Setegn, S.G., 2011. Land use and climate change impacts on the hydrology of the upper Mara River Basin, Kenya: results of a modeling study to support better resource management. Hydrol. Earth Syst. Sci. 15, 2245–2258.
Mann, H.B., 1945. Non-parametric tests against trend. Econometrica 13, 245–259.
Mao, D., Cherkauer, K.A., 2009. Impacts of land use change on hydrologic responses in the Great Lakes region. J. Hydrol. 374, 71–82.
Mati, B.M., Mutie, S., Home, P., Mtalo, F., Gadain, H., 2005. Land Use Changes in the Transboundary Mara Basin: A Threat to Pristine Wildlife Sanctuaries in East Africa. 8th International River Symposium, Brisbane, Australia, September 6–9.
Mati, B.M., Mutie, S., Gadain, H., Home, P., Mtalo, F., 2008. Impacts of land-use/cover change on the hydrology of the transboundary Mara River, Kenya/Tanzania. Lakes Reserv.: Res. Manag. 13, 169–177.
Maxon, R.M., Ofcansky, T.P., 2014. Historical Dictionary of Kenya, 3rd ed. Rawman and Littlefield, Lanham, Maryland.
McClain, M.E., Subalusky, A.L., Anderson, E.P., Dessu, S.B., Melesse, A.M., Ndomba, P.M., Mtamba, J.O.D., Tamatamah, R.A., Mligo, C., 2014. Comparing flow regime, channel hydraulics, and biological communities to infer flow-ecology relationships in the Mara River of Kenya and Tanzania. Hydrol. Sci. J. 59, 801–819.
Modarres, R., Sarhadi, A., 2009. Rainfall trends analysis of Iran in the last half of the twentieth century. J. Geophys. Res. 114.
Muiru, P.N., 2012. The Social, cultural and economic impact of ethnic violence in molo division, 1969–2008. In: MSc. Thesis. Kenyatta University, Kenya.
Mumeka, A., 1986. Effect of deforestation and subsistence agriculture on runoff of the Kafue River headwaters, Zambia. Hydrol. Sci. J. 31, 543–554.
Mutie, S.M., Mati, B., Home, P., Gadain, H., Gathenya, J., 2006. Evaluating land use change effects on river flow using USGS geospatial stream flow model in Mara River basin, Kenya. In: Proceedings of the 2nd Workshop of the EARSeL SIG on Land Use and Land Cover, September 28–30, 2006. Bonn, pp. 141–148.
National Environment Management Authority-NEMA, 2013. Mau Complex at a glance-Report. National Environment Management Authority, Nairobi.
Ngongondo, C., Xu, C.Y., Tallaksen, L.M., Alemaw, B., 2013. Evaluation of the FAO Penman–Montheith, Priestly–Taylor and Hargreaves models for estimating reference evapotranspiration in southern Malawi. Hydrol. Res. 44, 706–722.
Nkako, F.M., Lambrechts, C., Gachanja, M., Woodley, B., 2005. Maasai Mau Forest Status Report 2005. Ewaso Ngiro South Development Authority, Narok, Kenya.
Ntegeka, V., Willems, P., 2008. Trends and multidecadal oscillations in rainfall extremes, based on a more than 100 year time series of 10min rainfall intensities at Uccle, Belgium. Water Resour. Res. 44, W07402.
Ochieng, R.M., 2009. The Mau Forest Complex, Kenya: A Review of Degradation Status and Possible Remedial Measures. GRIN Verlag, Norderstedt, Germany.
Ongoma, V., Muthama, J.N., Gitau, W., 2013. Evaluation of urbanization influences on urban temperature of Nairobi City, Kenya. Glob. Meteorol. 2, e1.
Onyutha, C., 2015. Variability of seasonal and annual rainfall in the River Nile riparian countries and possible linkages to ocean-atmosphere interactions. Hydrol. Res., http://dx.doi.org/10.2166/nh.2015.164 (in press).
Onyutha, C., Tabari, H., Taye, M.T., Nyandwaro, G.N., Willems, P., 2015. Analyses of rainfall trends in the Nile River Basin. J. Hydro-environ. Res., http://dx.doi.org/10.1016/j.jher.2015.09.002 (in press).
Partal, T., Kahya, E., 2006. Trend analysis in Turkish precipitation data. Hydrol. Process. 20, 2011–2026.
Patil, S., Stieglitz, M., 2011. Hydrologic similarity among catchments under variable flow conditions. Hydrol. Earth Syst. Sci. 15, 989–997.
Price, K., 2011. Effects of watershed topography, soils, land use and climate on baseflow in humid regions: a review. Progress Phys. Geogr. 35 (4), 465–492.
Recha, J.W., Lehmann, J., Walter, M.T., Pell, A., Verchot, L., Johnson, M., 2012. Stream discharge in tropical headwater catchments as a result of forest clearing and soil degradation. Earth Interact. 16, 1–18, Paper 13.
Rees, G., 2008. Hydrological data. in: Gustard, A., Demuth, S., (Eds). Manual on Low-flow Estimation and Prediction. Operational Hydrology Report No. 50, WMO-No. 1029. World Meteorological Organization. Geneva.
Roderick, M.L., Farquhar, G.D., 2011. A simple framework for relating variations in runoff to variations in climatic conditions and catchment properties. Water Resour. Res. 47, W00G07.
Sahin, V., Hall, M.J., 1996. The effects of afforestation and deforestation on water yields. J. Hydrol. 178, 293–309.
Sang, Y.F., Wang, Z., Liu, C., 2014. Comparison of the MK test and EMD method for trend identification in hydrological time series. J. Hydrol. 510, 293–298.
Sankarasubramanian, A., Vogel, R.M., Limbrunner, J.F., 2001. Climate elasticity of streamflow in the United States. Water Resour. Res. 37 (6), 1771–1781.
Schaake, J.C., 1990. From climate to flow. In: Waggoner, P.E. (Ed.), Climate Change and U.S. Water Resources. John Wiley, New York.
Schilling, K.E., Gassman, P.W., Kling, C.L., Campbell, T., Jha, M.K., Wolter, C.F., Arnold, J.G., 2014. The potential for agricultural land use change to reduce flood risk in a large watershed. Hydrol. Process. 28, 3314–3325.
Sen, P.K., 1968. Estimates of the regression coefficient based on Kendall's tau. J. Am. Stat. Assoc. 63, 1379–1389.
Sneyers, R., 1990. On statistical analysis of series of observation. Technical note No.143, WMO No. 415. World Meteorological Organization, Geneva.

Sun, Y., Tian, F., Yang, L., Hu, H., 2014. Exploring the spatial variability of contributions from climate variation and change in catchment properties to streamflow decrease in a mesoscale basin by three different methods. J. Hydrol. 508, 170–180.
Szcześniak, M., Piniewski, M., 2015. Improvement of hydrological simulations by applying daily precipitation interpolation schemes in meso-scale catchments. Water 7, 747–779.
Tegos, A., Malamos, N., Koutsoyiannis, D., 2015. A parsimonious regional parametric evapotranspiration model based on a simplification of the Penman–Monteith formula. J. Hydrol. 524, 708–717.
Thiessen, A.H., 1911. Precipitation averages for large areas. Mon. Weather Rev. 39, 1082–1089.
Tomer, M.D., Schilling, K.E., 2009. A simple approach to distinguish land-use and climate-change effects on watershed hydrology. J. Hydrol. 376, 24–33.
von Storch, H., 1995. Misuses of statistical analysis in climate research. In: von Storch, H., Navarra, A. (Eds.), Analysis of Climate Variability: Applications of Statistical Techniques. Springer, Berlin, Heidelberg.
Wang, X., 2014. Advances in separating effects of climate variability and human activity on stream discharge: an overview. Adv. Water Resour. 71, 209–218.
Wang, D., Hejazi, M., 2011. Quantifying the relative contribution of the climate and direct human impacts on mean annual streamflow in the contiguous United States. Water Resour. Res. 47, W00J12.
Ward, A.D., Trimble, S.A., 2003. Environmental Hydrology, 2nd ed. CRC press, Boca Raton, Florida.
Xu, X., Yang, D., Yang, H., Lei, H., 2014. Attribution analysis based on the Budyko hypothesis for detecting the dominant cause of runoff decline in Haihe basin. J. Hydrol. 510, 530–540.
Yang, H., Yang, D., Lei, Z., Sun, F., 2008. New analytical derivation of the mean annual water-energy balance equation. Water Resour. Res. 44, W03410.
Yang, T., Chen, X., Xu, C.Y., Zhang, Z.C., 2009. Spatio-temporal changes in hydrological processes and underlying driving forces in Guizhou region, Southwest China. Stoch. Environ. Res. Risk Assess. 23, 1071–1087.
Ye, X., Zhang, Q., Liu, J., Li, X., Xu, C., 2013. Distinguishing the relative impacts of climate change and human activities on variation of streamflow in the Poyang Lake catchment, China. J. Hydrol. 494, 83–95.
Yue, S., Wang, C.Y., 2002. Applicability of prewhitening to eliminate the influence of serial correlation on the Mann–Kendall test. Water Resour. Res. 38 (6), 1068.
Yue, S., Wang, C., 2004. The Mann–Kendall test modified by effective sample size to detect trend in serially correlated hydrological series. Water Resour. Manag. 18, 201–218.
Yue, S., Pilon, P., Phinney, B., Cavadias, G., 2002. The influence of autocorrelation on the ability to detect trend in hydrological series. Hydrol. Process. 16, 1807–1829.
Zhang, Y.K., Schilling, K.E., 2006. Increasing streamflow and baseflow in Mississippi River since 1940: effect of land use change. J. Hydrol. 324, 412–422.
Zhang, X., Zwiers, F.W., 2004. Comment on "Applicability of prewhitening to eliminate the influence of serial correlation on the Mann–Kendall test" by Sheng Yue and Chun Yuan Wang. Water Resour. Res. 40, 5.
Zhang, L., Podlasly, C., Ren, Y., Feger, K.H., Wang, Y., Schwärzel, K., 2014. Separating the effects of changes in land management and climate conditions on long-term streamflow trends analyzed for a small catchment in the Loess Plateau region, NW China. Hydrol. Process. 28, 1284–1293.
Zhang, A., Zheng, C., Wang, S., Yao, Y., 2015. Analysis of streamflow variations in the Heihe River Basin, northwest China: trends, abrupt changes and ecological influences. J. Hydrol.: Reg. Stud. 3, 106–124.
Zheng, H., Zhang, L., Zhu, R., Liu, C., Sato, Y., Fukushima, Y., 2009. Responses of streamflow to climate and land surface change in the headwaters of the Yellow River Basin. Water Resour. Res. 45, W00A19.
Zheng, H., Zhang, L., Liu, C., Shao, Q., Fukushima, Y., 2007. Changes in stream flow regime in headwater catchments of the yellow River basin since the 1950. Hydrol. Process. 21, 886–893.