Journal of Hydrology (2007) **337**, 258–268

available at www.sciencedirect.com

ScienceDirect

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

Modeling the hydrological impact of land-use change in West Africa

K.Y. Li [a], M.T. Coe [b,*], N. Ramankutty [c], R. De Jong [d]

[a] *Center for Sustainability and the Global Environment, University of Wisconsin-Madison, WI, United States*
[b] *The Woods Hole Research Center, 149 Woods Hole Rd, Falmouth, MA 02540-1644, United States*
[c] *McGill University, Montreal, Canada*
[d] *Eastern Cereal and Oilseed Research Center, AAFC, Ottawa, Canada*

Received 20 March 2006; received in revised form 23 January 2007; accepted 25 January 2007

KEYWORDS
Sahel;
Runoff;
Discharge;
Model studies;
Vegetation

Summary Numerical simulations of idealized deforestation and overgrazing are performed for the Niger and Lake Chad basins of West Africa with a terrestrial ecosystem model IBIS (integrated biosphere simulator) and an aquatic transport model THMB (terrestrial hydrology model with biogeochemistry). The study reveals how land use changes affect hydrological regimes at the watershed scale. The results show that tropical forests, due to being situated in the regions of highest rainfall and exerting strong influence on evapotranspiration, have a disproportionately large impact on the water balance of the entire basin. Total deforestation (clearcutting) increases the simulated runoff ratio from 0.15 to 0.44, and the annual streamflow by 35–65%, depending on location in the basin, although forests occupy only a small portion (<5%) of the total basin area. Complete removal of grassland and savanna, which occupy much greater areas of the basins, result in an increase in simulated annual streamflow by 33–91%. The numerical simulations indicate that the hydrological response to progressive land cover change is non-linear and exhibits a threshold effect. There is no significant impact on the water yield and river discharge when the deforestation (thinning) percentage is below 50% or the overgrazing percentage below 70% for savanna and 80% for grassland areas; however, the water yield is increased dramatically when land cover change exceeds these thresholds. This threshold effect is a combined result of the non-linearity of the separate response of transpiration and soil and canopy evaporation to the imposed land cover changes.

Introduction

In the past 50 years, West Africa has experienced large land-use changes including deforestation, overgrazing and

* Corresponding author. Tel.: +1 508 540 9900; fax: +1 508 540 9700.
E-mail address: mtcoe@whrc.org (M.T. Coe).

0022-1694/$ - see front matter © 2007 Elsevier B.V. All rights reserved.
doi:10.1016/j.jhydrol.2007.01.038

reclamation (Ramankutty, 2004; Brunner et al., 1995; Stephenne and Lambin, 2001), and a persistent drought since the 1960s (Wang and Eltahir, 2000). Land-use and climate changes may have both immediate and long-lasting impacts on terrestrial hydrology, altering the balance between rainfall and evapotranspiration and the resultant runoff. In the short-term, destructive land use change may disrupt the hydrological cycle either through increasing the water yield or through diminishing, or even eliminating the low flow in some circumstances (Croke et al., 2004; Pereira, 1992; Bruijnzeel, 1990). In the long-term, the reductions in evapotranspiration and water recycling arising from land-use changes may initiate a feedback mechanism that results in reduced rainfall (Savenije, 1995). It has been argued that the persistent Sahelian drought since the 1960s may be attributable, at least in part, to vegetation changes (e.g., Wang and Eltahir, 2000; Zeng et al., 1999; Charney, 1975).

Climate change and variability directly alter the hydrological cycle as well as the type and abundance of vegetation, which may change the behavior of lakes and watercourses (Sircoulon et al., 1999). In countries of the Sahel, the aforementioned persistent drought has greatly reduced discharge rates of the rivers and has reduced the area of Lake Chad from 23,500 km^2 to about 1500 km^2 (Coe and Foley, 2001; Sircoulon et al., 1999; Olivry et al., 1996; Pouyaud and Colombani, 1989).

Extensive alterations in land cover and land use, which in turn impact the hydrological system both at basin and regional scales, have occurred throughout West Africa in the last 50 years. These changes are a result of multiple factors including climate change and variability, demographic growth, macroeconomic activities and development policies (Legesse et al., 2003). Physical understanding of the interactions between hydrology, climate, and land-use change is important not only for after-the-fact analyses, but also for understanding and perhaps predicting the potential hydrological consequences of existing land-use practices and climate trends in West Africa.

While field experiments can conclusively demonstrate the consequences of land use change, modeling studies often provide more insight into the mechanisms. Therefore, physically based and spatially distributed ecosystem, land-surface, and hydrological models are increasingly used to address the hydrological impacts of land-use changes (e.g., Bathurst et al., 2004; Legesse et al., 2003; VanShaar et al., 2002; Lorup et al., 1998; Calder et al., 1995; Bathurst and O'Connell, 1992; Refsgaard, 1987; Abbott et al., 1986). Modeling studies have advantages over basin experimental studies in being more flexible and rigorous in experimental design, and enabling mechanistic interpretation. In addition, models are able to provide results immediately with much less cost in staffing and operation. Numerical simulations depend on field experiments and observations for their construction, calibration, and validation and therefore cannot replace field experiments. However, numerical simulations can, in some cases, extend the scope and overcome the limitations of traditional field experiments, which is particularly true when addressing regional and global issues including land use and climate change impacts. In this respect, numerical simulations are a complement to and an extension of field experiments.

This study employs an integrated modeling approach aimed at mechanistic interpretation, to address the hydrological sensitivity to land cover changes in West Africa, including: (1) deforestation by means of clearcutting and progressive thinning and (2) overgrazing as progressive thinning of grassland and savanna vegetation.

Methodology

Model description

A terrestrial ecosystem model — IBIS (Kucharik et al., 2000; Foley et al., 1996) and an aquatic transport model — THMB (formerly know as HYDRA) (Coe et al., 2002; Coe, 2000, 1998) are jointly used to investigate the hydrological sensitivity to land cover changes. As an ecosystem model, IBIS provides a physically-sound framework for modeling land surface hydrology under different natural vegetation covers and crops. In terms of the hydrology, IBIS simulates canopy interception of rainfall, surface and sub-surface runoff, soil moisture, soil and canopy evaporation, and plant transpiration. However, IBIS does not simulate river discharge, which is an observable, practical indicator of land cover changes. Therefore, THMB is used to simulate the river discharge from the runoff estimated by IBIS.

IBIS terrestrial ecosystem model

IBIS is a physically-based model that integrates a variety of terrestrial ecosystem processes within a single, mechanistic model to simultaneously calculate a wide range of processes, including the land surface water and energy balances. The model has two vegetation canopies with an upper layer of trees and a lower layer of shrubs, grasses and crops, and 15 types of natural vegetation cover comprised of a combination of 12 plant functional types including woody and herbaceous plants. The soil module has six soil layers (with a total of 4-m depth in this study). The dynamics of soil volumetric water content are simulated for each layer. The soil moisture simulation is based on Richards' flow equation, where the soil moisture change in time and space is a function of soil hydraulic conductivity, soil water retention curve, plant water uptake and upper and lower boundary conditions. Plant transpiration is a mechanistic process governed by stomatal physiology, in IBIS it is tightly coupled to photosynthesis through the Ball-Berry formulation (Ball et al., 1986). The plant root-water uptake is a complex function of atmospheric demand, soil physical properties, root distribution, and soil moisture profile (Li et al., 2005; Kucharik et al., 2000). IBIS explicitly simulates surface and sub-runoff on a grid cell basis as a function of the soil, vegetation, and climate characteristics. Horizontal runoff transport between grid cells is subsequently simulated by the THMB hydrological routing model. Since IBIS has already been described by Foley et al. (1996) and Kucharik et al. (2000), and a thorough introduction of the hydrological modules of IBIS over West Africa was made by Li et al. (2005), further details are omitted here.

THMB hydrological routing model

THMB transports local surface and subsurface runoff across the land surface to oceans or inland basins, estimating flows and storage at all points in the basin at 5-min horizontal resolution (Coe et al., 2002). The model is based on a linear reservoir model that simulates water transport in terms of river routing directions derived from water head, residences times within a grid cell, and effective flow velocities. The total water influent into each grid cell is the sum of the land surface runoff (R_s), subsurface runoff (R_d), precipitation (P_w) and evaporation (E_w) over the surface waters, and the water flow from upstream and to downstream grid cells ($m^3\ s^{-1}$). The water transport is represented by the time-dependent change of three water reservoirs: river water reservoir (W_r), surface runoff pool (W_s) and subsurface drainage pool (W_d). W_r contains the sum of upstream and local water in excess of that required to fill a local surface water depression. W_s represents the water that has run off the surface locally, while W_d is the water that has drained through the local soil column. All three reservoirs are represented in m^3 and flow is governed by the following differential equations:

$$\begin{aligned} d(W_s)/dt &= R_s - W_s/T_s \\ d(W_d)/dt &= R_d - W_d/T_d \\ d(W_r)/dt &= (W_s/T_s + W_d/T_d) \times (1 - A_w) + (P_w - E_w) \\ &\quad \times A_w - (W_r/T_r) + \sum F_{in} \end{aligned} \tag{1}$$

where A_w is the fractional water area in the grid cell that is simulated by THMB, ranging from 0 (no water present) to 1 (the grid cell completely covered by a lake, wetland, or reservoir); T_s, T_d and T_r are the residence times (s) of the water in each of the reservoirs; P_w and E_w are the precipitation and evaporation rates ($m^3\ s^{-1}$) over the surface water, respectively, and $\sum F_{in}$ is the net water flux (m^3/s^{-1}) into the cell (+ or −) and is the sum of the fluxes in from upstream grid cells minus the flux out to downstream cells. For details of the model, readers are referred to Coe and Foley (2001), Coe et al. (2002) and Li et al. (2005).

Experimental design

Two large basins, the Lake Chad Basin (hereinafter referred to as LCB) and the Niger River Basin (referred to as NRB), were selected for the simulations (Fig. 1). The former is located in northern central Africa, spreading over seven countries. The latter, the second largest basin in Africa located in western Africa, is shared by 11 countries. Both basins, each with an area of over 2 million km^2, straddle the boundary between the Sahara desert and moist tropical forest to the south and are characterized by high precipitation variability at various time and space scales (Nicholson, 2000, 1988). The annual rainfall varies from less than 100 mm in the north to more than 1300 mm in the south, with an average of 373 mm/yr for the LCB and 425 mm/yr for the NRB in the simulation area. There is little vegetation in the arid northern part of the basins, while in the south the vegetation is dominated by savanna and grassland, with a small portion of tropical forest distributed along the south edge

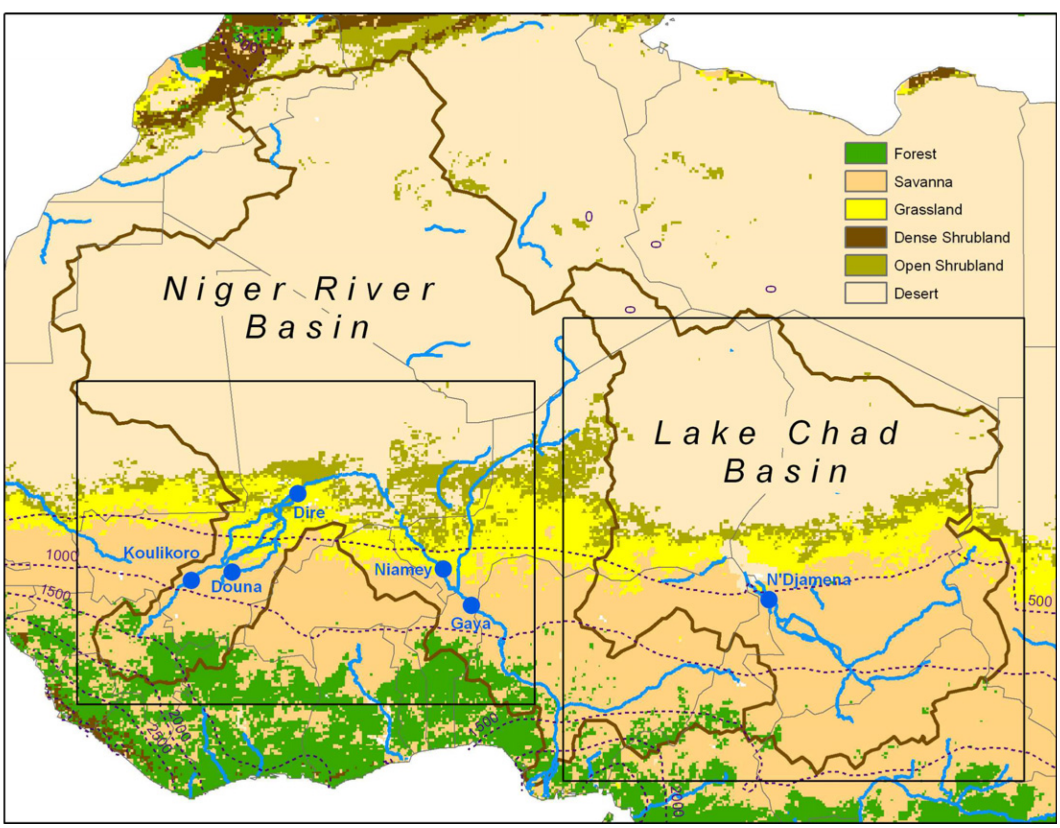

Figure 1 The Lake Chad Basin (LCB) and the Niger River Basin (NRB) (outlined in dark blue) and the simulation domains (black boxes). The potential vegetation (Ramankutty and Foley, 1999) is shown at 1 km resolution. Precipitation contours are overlaid as dashed black lines and discharge stations reported in text are shown in light blue.

of the NRB. This strong north-south gradient of precipitation and vegetation is ideal for simulations aimed at mechanistic interpretation of the land cover change impact. IBIS was calibrated with the vegetation fixed to the potential natural types by Li et al. (2005), and the experimental simulations, in this study, were conducted based on the same potential vegetation types (derived from Ramankutty and Foley, 1999).

The idealized land cover change experiments in this study represent deforestation (or thinning) and overgrazing. Forest thinning and overgrazing of grasslands were simulated across a wide range of degrees, including 25, 50, 60, 70, 80, 90 and 100%, where 100% thinning was regarded as clearcutting or complete removal of grasses. Within IBIS, the implementation of forest thinning and overgrazing by various degrees involved proportionally decreasing parameter values of the potential leaf area index and canopy fractional cover for each vegetation type in question.

The experiments include a control simulation (with potential natural vegetation and modern climate for the period 1935–1995) and a series of land cover simulations as indicated above. For all experiments IBIS was run at the resolution of $0.5° \times 0.5°$ and a time step of 1 h, and THMB at the resolution of $5' \times 5'$ and a time step of 1 h. The control simulation used monthly mean climate data for a period of 61 years from 1935 to 1995 (New et al., 2000), potential natural vegetation type derived from Ramankutty and Foley (1999) and soil type from IGBP-DIS (1998) as used by Coe and Foley (2001). IBIS generates daily and hourly weather from monthly weather data using a stochastic weather generator. The land cover change scenarios used the same climate and vegetation data over the same period of time, with our vegetation perturbations superimposed on them.

Model calibration and validation

The basin integrated mean annual runoff from IBIS was calibrated and validated by Li et al. (2005) using long-term river discharge data. The control (potential vegetation) experiment used in this study is the same as that calibrated and validated by Li et al. (2005). The validation by Li et al. (2005) showed good agreement between simulated annual mean runoff and river discharge, with the normalized root mean square error for all station-years not used in the calibration being 16% and the relative error being 3%. Therefore, the water budget as a function of the climate and land surface characteristics is well simulated by IBIS in these basins. For details of the calibration and validation of IBIS for the study basins, readers are referred to Li et al. (2005).

Results and discussion

In both LCB and NRB, about one half of the simulated area is covered by desert in the north and the other half by vegetation in the south: desert plants, open shrub, grass/steppe, savanna and tropical broadleaf evergreen green trees are successively distributed from north to south (Fig. 1). The different vegetation types correspond to the rainfall regimes (Fig. 2), with observed average annual precipitation (calculated from the New et al., 2000 dataset for the period 1935–1995) being 230 mm (in LCB) and 236 mm (in NRB) for the open shrub, 336 mm (in LCB) and 412 mm (in NRB) for the grassland, 808 mm (in LCB) and 976 mm (in NRB) for the savanna, and 1322 mm (in NRB) for the tropical evergreen forest.

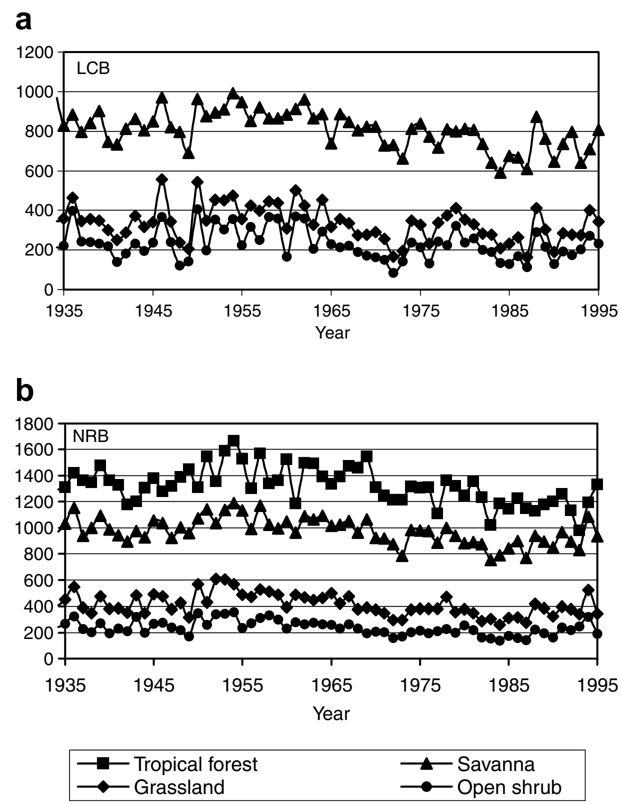

Figure 2 Annual rainfall (mm/yr) from 1935 to 1995 for the different potential vegetation zones in: (a) Lake Chad Basin (LCB); and (b) Niger River Basin (NRB). Rainfall values from New et al. (2000).

Deforestation

In this section, we discuss the experimental results for deforestation of the tropical broadleaf evergreen forest vegetation type, first for 100% removal of all vegetation and then for progressive fractional thinning of the forest. Although the mode of deforestation applied in this study is unrealistically simple (complete removal of all vegetation), the results are instructive because they provide a clearer understanding of the non-linear responses of the hydrological cycle to progressive vegetation changes. Tropical broadleaf evergreen forest constitutes a very small portion of the potential vegetation of this region. In the LCB there is no tropical evergreen forest in our experiments, while in the NRB it covers less than 5% of the basin area. However, as discussed below, that small portion of the NRB, i.e., the tropical broadleaf evergreen forest, represents a very important part of the NRB water budget.

With 100% removal of the tropical forest vegetation the runoff rate in the forest area averaged for the entire 61-year period increases by 378 mm/yr, from 203 mm/yr (with potential vegetation) to 581 mm/yr (with no vegetation) (Fig. 3), which is equivalent to an increase of 186%. The runoff ratio (the ratio of annual runoff to annual rainfall)

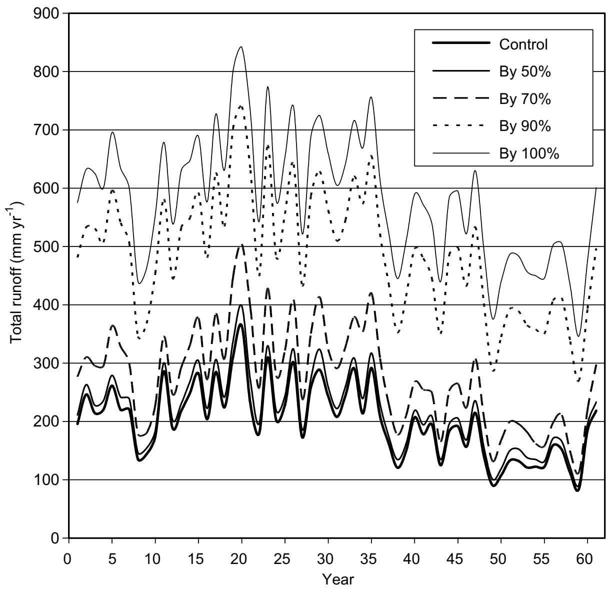

Figure 3 Comparison of total runoff (mm/yr) summed over the tropical forest region of NRB for the potential vegetation (control) experiment (thick black line) and deforestation experiments. Shown are results for 50%, 70%, and 90% and 100% removal of vegetation. The *x*-axis shows the year since 1935.

increases from 0.15 in the control experiment to 0.44 (with 100% deforestation).

Because of the relatively high precipitation rates in the tropical forest area of the NRB, this region provides a large portion of the water in the Niger River system despite occupying less than 5% of the total NRB area. The runoff generated from the forest region before deforestation accounts for 21% of all runoff generated in the entire basin, and this increases to 44% when 100% deforestation is applied. As a result, the impact of the clearcutting experiment extends to the downstream unforested portions of the basin. For example, the runoff increase in the forested area results in a large increase in streamflow in all downstream stations (some 1000s of km downstream, Fig. 4). The long-term average increase in annual river discharge is about 35% at Koulikora, Dire, Niamey, and Gaya on the mainstem of the river, and 65% at Douna, which is a heavily-forested sub-watershed (see Fig. 1 for locations).

The very large increase in total runoff (378 mm/yr) in our 100% deforestation experiment is due to an equally large decrease in total evapotanspiration (the sum of the plant transpiration and evaporation from the soil surface and of rainfall intercepted by the canopy leaves and stems). In our idealized clearcutting experiment all vegetation is removed in the forested region (both upper and lower canopy) and as a result, there is no transpiration or evaporation of rainfall intercepted by the canopy leaves and stems. Therefore, evaporation is limited to that from the bare soil. Evaporation from the bare soil is substantially increased in the deforestation experiment (to about 738 mm/yr, Fig. 5), because more soil is exposed, more moisture and energy reach the land surface, and the land surface temperature increases. However, this increase in surface evaporation is much less than the total evapotranspiration (1121 mm/yr) that occurred with intact vegetation because surface evaporation cannot access water in the deeper soil layers and because evaporation of intercepted water from the forest canopy is eliminated.

In our deforestation experiment both surface and subsurface runoff increase. About 20% of the 378 mm/yr increase in total runoff is due to surface runoff increase because some of the water that was formerly vegetation-intercepted evaporation became overland flow (Fig. 5). About 80% of the increase is due to an increase in sub-surface drainage of soil moisture that would have been transpired by plants in the control experiment. In the control experiment, only about 18% of the total runoff is from sub-surface drainage. In the deforestation experiment sub-surface drainage becomes about 60% of the total, which indicates that in the control experiment, most of the water that enters the soil column is eventually transpired by the forest.

The magnitude and individual proportions of the simulated runoff changes are unrealistic to the extent that we have maximized the vegetation change and not changed the soil compaction characteristics. However, they do clearly illustrate the pathways to changing hydrology as a result of vegetation changes. As natural vegetation is removed, decreased root water uptake results in increased soil moisture content and thus increased baseflow, even when total soil water infiltration may decrease. While evaporation at the surface is increased, due to higher temperatures and less vegetation cover, greater soil saturation and surface water ponding result in greater overland flow during rain events. This is consistent with observed increases in both overland runoff and baseflow following deforestation (Costa et al., 2003; Bruijnzeel, 1990; Bosch and Hewlett, 1982).

In order to better understand the response of hydrology to progressive forest removal we made a series of simulations with progressive thinning of the forest in NRB, including 25, 50, 60, 70, 80, and 90%. The results show a non-linear response of the water balance to vegetation removal: runoff and discharge did not markedly increase until more than 50% of the vegetation was removed (Fig. 3). This non-linear runoff change with the progressive thinning can be explained, as discussed below, by the non-linear response of transpiration and evaporation.

To illustrate the hydrological response to land cover change, an exponential equation was developed to fit the data for evapotranspiration or transpiration changes with deforestation and overgrazing, expressed as

$$W = W_0 - aRe^{(bR)} \qquad (2)$$

where W can be either evapotranspiration (ET, mm/yr) or transpiration (T, mm/yr); W_0 is either baseline evapotranspiration (ET_0, mm/yr) or baseline transpiration (T_0, mm/yr) without deforestation or overgrazing; R is either the deforestation (R_d, %) or overgrazing percentage (R_g, %); and a and b are fitting parameters. Similarly, another exponential equation was developed for evaporation (excluding plant transpiration), represented as

$$E = (E_0 - c) + ce^{(dR)} \qquad (3)$$

where E is evaporation (mm/yr), E_0 is baseline evaporation without deforestation or overgrazing (mm/yr); R is the

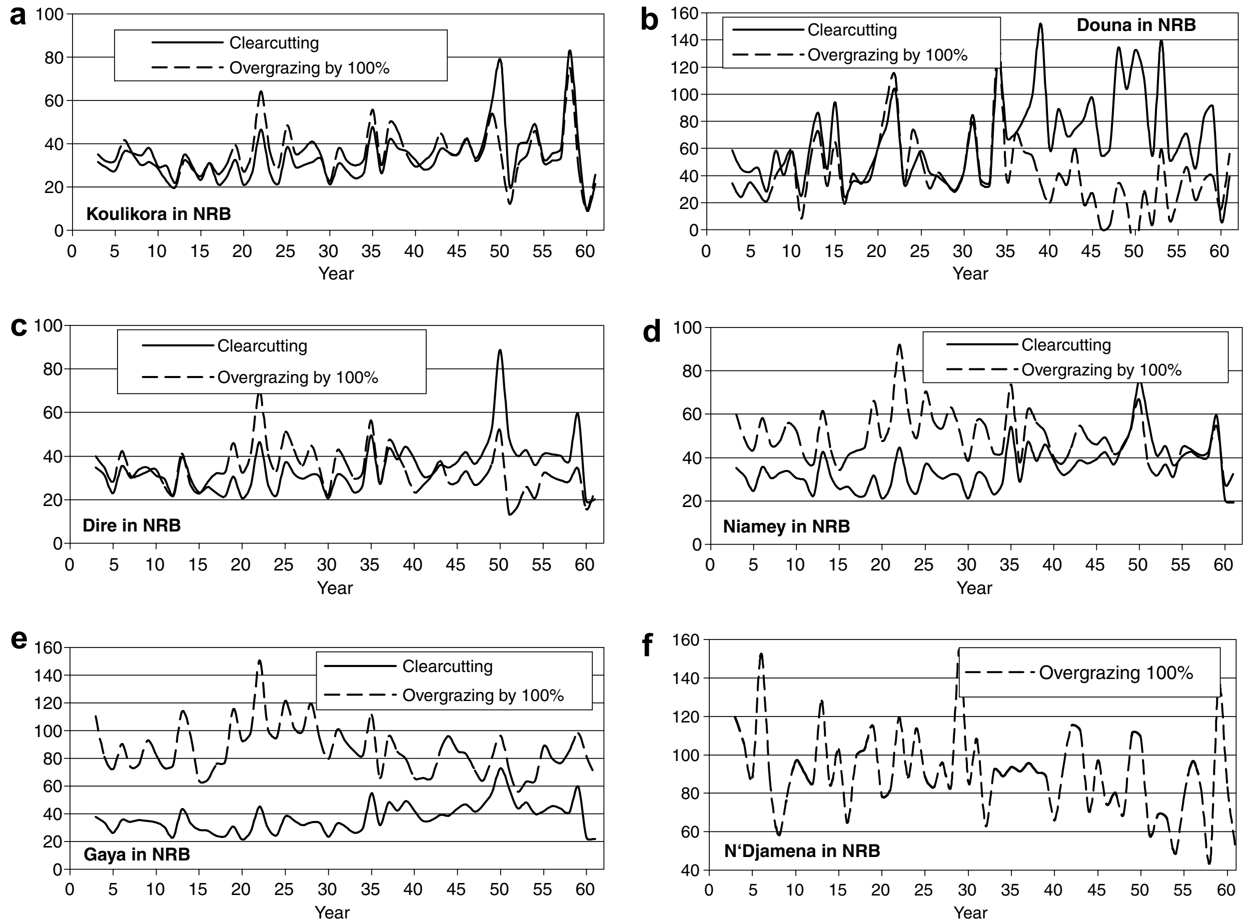

Figure 4 The percent increase in annual mean river discharge compared to the control due to 100% removal of the tropical forest (clearcutting, solid line) and 100% lower canopy vegetation removal in savanna and grassland areas (overgrazing, dotted line) in Niger River Basin (NRB) and Lake Chad Basin (LCB). The time series a–f refer to individual stations, a–e within the Niger River system moving from upstream to downstream, f on the Chari River in the Lake Chad Basin (see Fig. 1 for station locations).

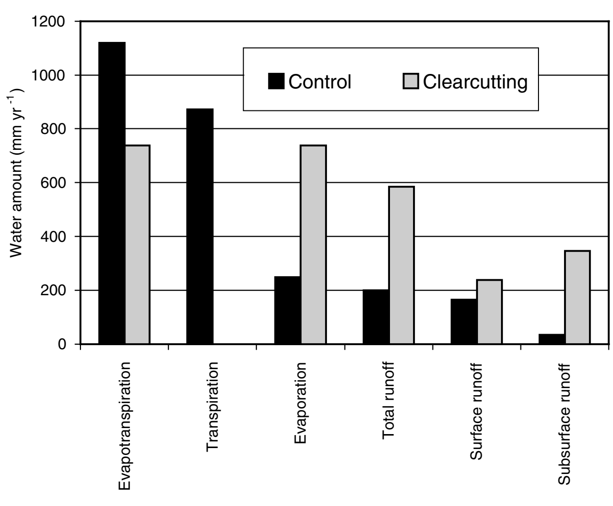

Figure 5 Simulated water balance components for control (potential vegetation) and deforestation (clearcutting) experiments in the NRB summed over the forest region (see Fig. 1).

deforestation or overgrazing percentage (%); and *c* and *d* are the fitting parameters.

With 60% of the vegetation removed in the tropical forest region the simulated total evapotranspiration has decreased by only a few percent (Fig. 6). However, with greater than 60% deforestation evapotranspiration decreases rapidly to about 16% less evapotranspiration at 80% deforestation, 26% less at 90% and 34% less at 100% deforestation. The non-linear response is a result of the competition between increasing evaporation and decreasing transpiration (Fig. 6). The evaporation term increases non-linearly with increasing deforestation but the exponent is relatively small. Less water is intercepted by vegetation (not shown) and evaporated as deforestation progresses but more water reaches the ground and the thinning forest steadily exposes more bare ground for evaporation. The transpiration term, however, responds more slowly to initial forest reduction. For example, the experiment with 50% forest cover reduction has a transpiration decrease that is 23% of the decrease that occurs in the 100% forest removal experiment (Fig. 6). The slower transpiration response is because initially forest thinning

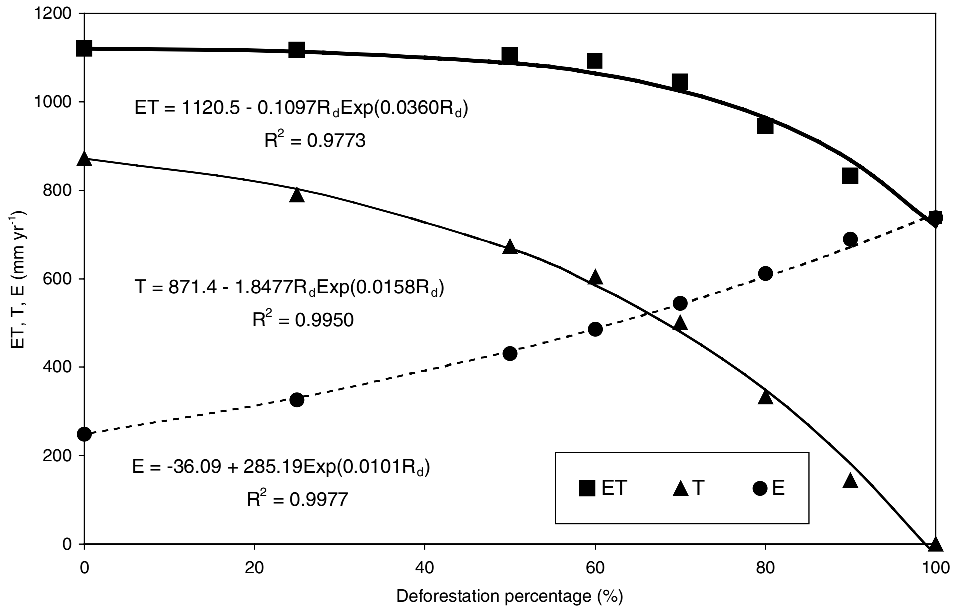

Figure 6 The relationships between the percentage removal of tropical evergreen forest (R_d) in the Niger River basin (NRB) and evapotranspiration (ET), transpiration (T) and evaporation (E).

is largely compensated by an increase in the transpiration rate per unit leaf area and the concomitant increase in extraction of water from the soil per unit root mass. There is less competition between individual leaves (or stomata) and roots for the relatively limited water resource and thus the total amount of water transpired does not greatly decrease. Additionally, more water reaches the surface and likely infiltrates the soil, since the soil characteristics are not changed in the simulation. However, after about 60% of the vegetation is removed the vegetation water stress is greatly reduced, the evaporative demands are more frequently met (not shown), and any reduction in vegetation leads directly to reduction in total transpiration. The runoff variations are inversely consistent with the evapotranspiration changes and therefore, there is no marked increase in runoff when the deforestation percentage is below 60% (not shown).

To test how precipitation amount influenced the structure and location of the non-linear response we re-ran the suite of deforestation simulations with the monthly precipitation increased by 33% over the forest region. We limited our sensitivity analysis to this one region and precipitation change because of the large number of simulations required (8 simulations for each vegetation and precipitation combination). The results indicate a modest shift in the threshold at which evapotranspiration begins to decrease greatly but no change in the basic shape of the response curve. For example at 60% forest removal, there is a 10% decrease in evapotranspiration and at 80% removal a 25% decrease, compared to a few percent and 16% decrease for our initial experiments, respectively (not shown). At 100% removal there is a 41% decrease compared with 34% in the original experiment. This shift towards earlier onset of the threshold and greater total decrease in evapotranspiration is consistent with there being more moisture in the soil with the higher precipitation rate, reduced plant water stress under these conditions, and therefore less ability at any given deforestation amount to compensate for decreased LAI with increased water use per leaf area compared to the lower precipitation rate.

Our numerical simulation results on forest thinning are consistent with a field experiment made in Finland (Heikurainen, 1967), for which there were four treatments of forest thinning with stand removal of 31, 33, 60, and 86%, respectively. The groundwater levels were monitored for the experiment, and the results showed that the treatments of 31% and 33% did not markedly increase the groundwater levels, while the treatment of 60% slightly and the treatment of 86% greatly increased the groundwater levels. Our simulation results on the impact of forest clearcutting and thinning are also consistent with a field experimental study in China, reported in a review by Shi and Li (2001), wherein, clearcutting induced a runoff increase of 14.4%, while thinning, that reduced the canopy cover from 95% to 60%, caused only a 0.9% increase.

Overgrazing

Overgrazing in this study was simulated by progressively thinning the lower canopy vegetation of both grassland and savanna vegetation types. The savanna, also known as tropical grassland, is a grassland with scattered shrubs and isolated trees, and is found in a wide band on either side of the equator poleward of tropical rainforests. In NRB and LCB, savanna and grassland are the dominant vegetation types, distributed to the south of the Sahara desert and to the north of the tropical rainforest. Overgrazing in this region has become a major environmental concern (Nicholson et al., 1998; Xue and Shukla, 1993; Charney, 1975), and an evaluation of its impact on the hydrological cycle is useful.

Similar to the forest thinning experiments, overgrazing was simulated at 25, 50, 60, 70, 80, 90 and 100% of the grassland and savanna regions but applied to the lower canopy vegetation only. There is no noticeable increase in simulated runoff with overgrazing less than 70% on savanna or 80% on grassland, and the runoff increases rapidly when the overgrazing percentage is greater than these thresholds (not shown). Eq. (2) was fitted to the evapotranspiration and transpiration data, respectively, and Eq. (3) to the evaporation data. For reasons similar to the deforestation experiment, there is no significant reduction in evapotranspiration and hence no significant increase in runoff, provided the overgrazing is below the threshold (Fig. 7a and b). The threshold for grassland (80%, not shown) and for savanna (70%, Fig. 7a and b) is greater than that for the tropical forest (50%). This suggests that water yield is less sensitive to grassland change than savanna change, and far more sensitive to tropical forest change. The differences in the response among vegetation types can be attributed to the different rainfall regimes they are associated with (Fig. 2): the higher the precipitation rate, the more sensitive the hydrology is to land cover change (Bosch and Hewlett, 1982), and to the changes in canopy structure and surface roughness. At low precipitation rates the very large difference between the relatively small water supply and large evaporative demand means that increased transpiration per unit leaf area and evaporation from the soil and ephemeral pools can compensate for the progressive loss of vegetation (Fig. 7a and b). Only when vegetation cover becomes small (<30% for savannah and <20% for grassland) does the available water begin to exceed evaporative demand and the soil moisture becomes cut off from surface evaporation so that runoff increases.

While precipitation in savanna and grassland areas is much lower than in tropical rainforest areas, savanna and grasslands cover a much larger portion of the basins. Hence, large-scale overgrazing can cause substantial increases in streamflow in the river systems within both basins (Fig. 4).

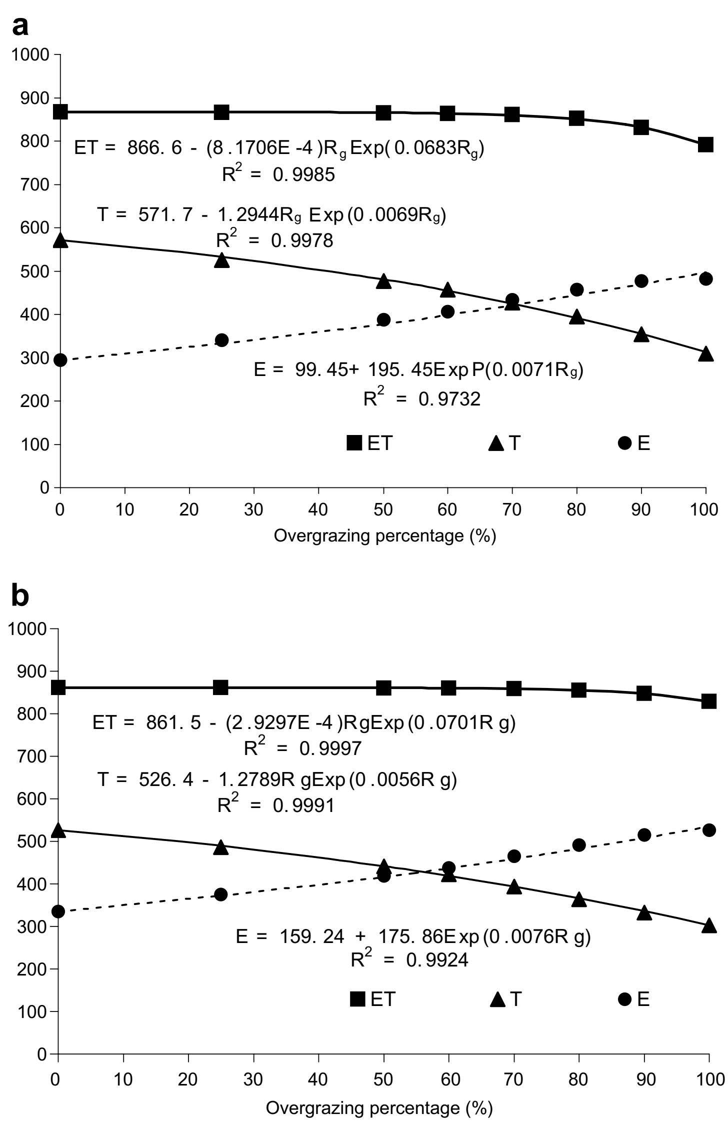

Figure 7 The relationships between the overgrazing percentage (R_g) and evapotranspiration (ET), transpiration (T) and evaporation (E) rates (mm/yr) for savanna in: (a) Niger River Basin (NRB); and (b) Lake Chad Basin (LCB).

The long-term average increase in streamflow with 100% overgrazing is 91% for N'Djamena in LCB, and in NRB are 33% for Dire, 35% for Koulikora, 41% for Douna, 49% for Niamey and 86% for Gaya. Similar to deforestation, the large increase in streamflow is mainly due to enhanced baseflow (data not shown). Fig. 4 shows the extreme case (i.e., overgrazing is 100%); when the overgrazing percentage is decreased, there is very little impact of vegetation thinning on runoff below the threshold of 70—80% removal of savannas and grasslands, respectively (not shown). It should, however, be noted that in the long term overgrazing may lead to other ecological and hydrological problems such as desertification and climate change (Zeng et al., 1999; Wang and Eltahir, 2000) and that these effects can be more persistent in drier areas because of the slow recovery of vegetation.

Runoff ratios of deforestation and overgrazing

The observed runoff ratio (the ratio of average annual runoff to average annual rainfall) is higher in moist areas than in dry areas. Observed values in Africa range from much less than 5% for precipitation <500 mm/yr to greater than 50% for precipitation greater than 2000 mm/yr (Ashton, 2002). Within a region, the runoff ratio is a useful measure to evaluate the effect of land cover change because the effect on runoff is normalized by the precipitation regime. Similar to the function used to fit simulated evaporation (Eq. (3)), the following exponential equation was used to fit the simulated runoff ratio data:

$$r = (r_0 - g) + ge^{hR} \quad (4)$$

where r is the simulated runoff ratio, and r_0 simulated base runoff ratio without deforestation or overgrazing, R deforestation or overgrazing percentage (%), and g and h are empirical fitting parameters. Different vegetation types have different baseline runoff ratios (Fig. 8): the baseline runoff ratio for tropical forest (0.154) is larger than for savanna (0.113 for NRB and 0.048 for LCB), and that for savanna is greater than for grassland (0.016 for NRB and 0.001 for LCB). The simulations indicate that the precipitation regime has a large influence on the runoff ratio (Fig. 8). However the vegetation cover does have an influence: when deforestation or overgrazing is greater than the threshold the runoff ratio rises dramatically.

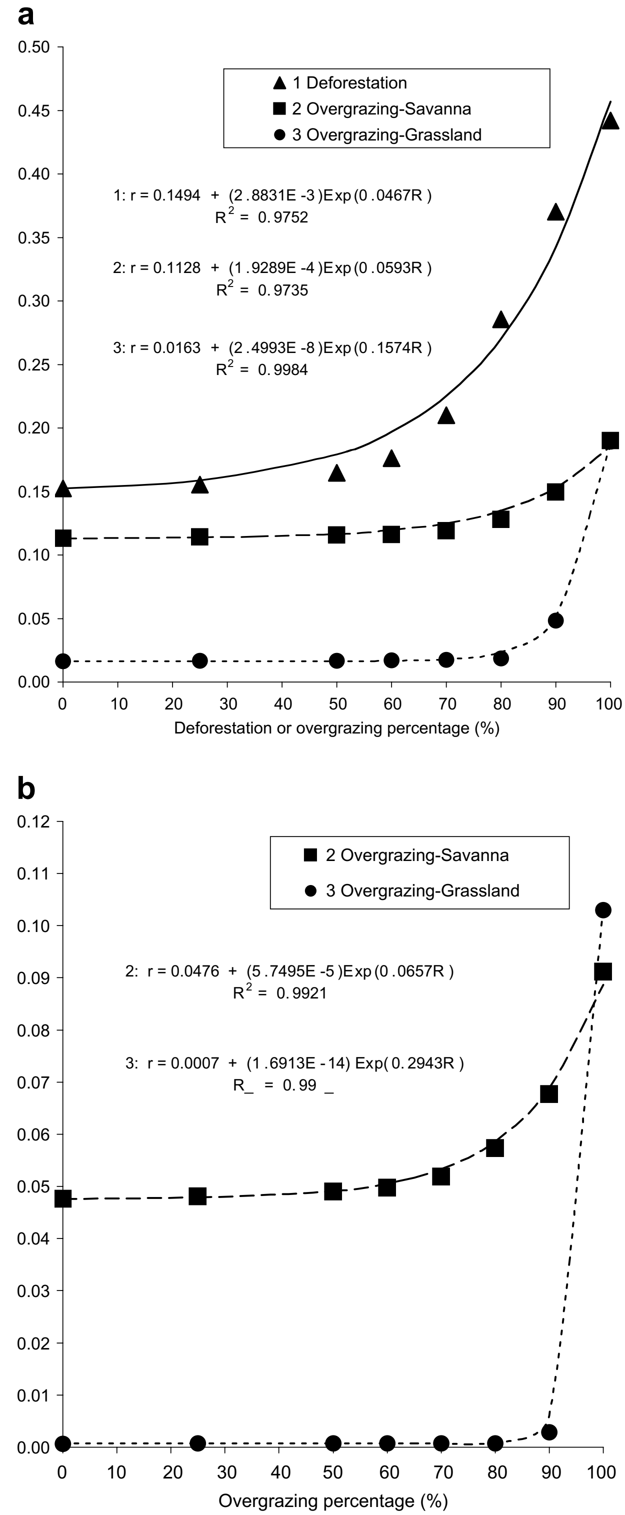

Figure 8 The relationship between runoff ratio (r) and the deforestation thinning percentage or overgrazing percentage (R) in: (a) Niger River Basin (NRB); and (b) Lake Chad Basin (LCB).

Summary and conclusions

We used a terrestrial ecosystem model (IBIS) and an aquatic transport model (THMB) to investigate the sensitivity of the hydrological cycle and water resources to land cover changes in West Africa. The land cover changes involved progressive removal from 25—100% of forest, savanna, and grassland.

The numerical simulation results show that although tropical forests represent only a very small portion of this region (none in the LCB and <5% of the NRB), total deforestation results in a substantial increase in simulated streamflow of the Niger River (by 35% to 65%, depending on the location). Overgrazing also has a considerable impact on the water yield, with simulated streamflow increases (for 100% overgrazing) ranging from 33% to 91%. Importantly, the numerical results also indicate that the hydrological response to the land cover change is non-linear with thresholds: there is little impact on the simulated water yield when the deforestation (thinning) is below 50% and the overgrazing is below 70% for savanna and 80% for grassland; however, water yield rapidly increases when above these thresholds.

This threshold effect is a function of the way in which transpiration and evaporation separately respond to the land cover changes. Initially with clearing, transpiration decreases more slowly than evaporation increases, because the plants not cleared are able to increase their transpiration per unit leaf area and root water uptake per unit root mass to meet plant water demands. As a result, runoff does not markedly increase until deforestation or overgrazing reaches a point at which the plant water stress becomes very small such that any further reduction of vegetation results in a direct reduction of the net transpiration. After this threshold is reached the increase in soil evaporation cannot compete with the decrease in plant transpiration, thus total evapotranspiration decreases and runoff increases rapidly.

Although, the particular method we chose in this study to investigate the response of the hydrology to land cover change is likely unrealistic (complete conversion of very large regions to bare soil) the change is similar to well-documented regional-scale degradation of vegetation due to overgrazing, conversion to crops, and fuelwood extraction (Dregne et al., 1992; Lambin and Erlich, 1997; Favreau et al., 2002; Ramankutty, 2004). Further, observations have confirmed large increases in water yield with destruction of natural vegetation. For example, observational analysis in the Niger basin has documented, decreased evapotranspiration, an order of magnitude increase in ground water recharge, and the formation of surface water ponds in regions where natural savannah has been replaced with less water demanding millet in the last 50 years (Leduc et al., 2001; Favreau et al., 2002).

The exact nature of these results is, of course, experiment dependent. We did not include changes to soil infiltration rates or water holding capacity with vegetation changes. These factors have an impact on the location of the non-linear water yield threshold through their controls on generation of surface runoff and available soil moisture. Soil compaction and decreased infiltration generally accompany deforestation and overgrazing and can significantly increase surface runoff (Leduc et al., 2001; Favreau et al., 2002). Therefore, it is likely that our simulated threshold is conservative: including realistic changes in soil properties may lower the threshold at which runoff increases drastically with vegetation change.

The exact response of the model to vegetation changes is also a function of the model itself: how physical processes and parameterizations are represented and the structure of the model. Processes of particular importance (e.g., surface water ponding, soil evaporation, canopy water stress, and root water uptake) have been carefully addressed and numerous improvements have been made to IBIS by Li et al. (2005, 2006) to better represent them. Therefore, it is most likely that the representation of these processes and the overall response of the model (if not the exact magnitude) is reasonable. However, the model does not currently take into account sub-grid spatial variability in climate forcing, soil physical properties, and vegetation type, which may introduce some bias. For example, the structure of the model dictates that any vegetation change is uniform within each grid cell: 50% deforestation or overgrazing means that the entire grid cell has a 50% reduction in biomass (above and below ground). Whereas, in the real world changes can be more heterogeneous; a 50% reduction could be 100% reduction over 50% of the area. This is important because in this particular example, the total runoff from the real world would be significantly increased over 50% of the area, while in the model, for a comparable area, the threshold would not yet be crossed and the runoff would not measurably increase. Therefore, again, the simulated water yield threshold may be conservative.

This study does not consider the ecological consequences of vegetation changes, which are likely to be large and may outweigh the gains from increased water availability. Nor does it consider atmospheric feedbacks that may be important when vegetation changes occur across large spatial extent (100,000 s km^2). For example, Wang and Eltahir (2000) showed with a numerical model that when feedbacks between the land surface changes and atmosphere are included, removal of as little as 20% of the vegetation of the entire Sahel can cause a significant reduction in simulated regional rainfall, thus counter-balancing runoff gains that may be incurred from decreased transpiration. Net changes to runoff when all aspects are considered therefore, remain uncertain. The results of this study are an exploration of the sensitivity of hydrology to a particular aspect of the larger ecosystem dynamics. They are likely to be representative of the response of the hydrology to vegetation changes over small regions but atmospheric feedbacks must be considered when the total area affected becomes large. Nevertheless, this study, which provides insight into how hydrology responds to land use change, will be useful in the analyses of interactions and feedbacks between climate, hydrology and land-use change.

Acknowledgements

We would like to thank Dr. Christine Delire and two anonymous reviewers for helpful comments on this manuscript. This work was supported by the National Aeronautics and Space Administration Office of Earth Science (Land Cover and Land Use Change).

References

Abbott, M.B., Bathurst, J.C., Cunge, J.A., O'Connell, P.E., Rasmussen, J., 1986. An introduction to the European Hydrological System — the Système Hydrologique Européen, 'SHE', 1: history and philosophy of a physically-based, distributed modeling system. J. Hydrol. 87, 45—59.

Ashton, P.J., 2002. Avoiding conflicts over Africa's water resources. Ambio 31 (3), 236—242.

Ball, J.T., Woodrow, I.E., Berry, J.A., 1986. A Model Predicting Stomatal Conductance and its Contribution to the Control of Photosynthesis Under Different Environmental ConditionsProgress in Photosynthesis Research, vol. 4. Martinus-Nijhoff, Dordrecht, Netherlands, pp. 221—224.

Bathurst, J.C., O'Connell, P.E., 1992. Future of distributed modeling — the Système Hydrologique Européen. Hydrol. Process. 6, 262—277.

Bathurst, J.C., Ewen, J., Parkin, G., O'Connell, P.E., Cooper, J.D., 2004. Validation of catchment models for predicting land-use and climate change impacts. 3. Blind validation for internal and outlet responses. J. Hydrol. 287, 74—94.

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.

Bruijnzeel, L.A., 1990. Hydrology of moist tropical forests and effects of conversion: a state of knowledge review. In: UNESCO International Hydrological Program, 224 pp.

Brunner, J., Henninger, N., Deichmann, U., Ninnin, B. 1995., West Africa long-term perspective study (WALTPS). Database and Users Guide, World Resources Institute and Club du Sahel, Washington, DC and Paris, pp. 26.

Calder, I.R., Hall, R.L., Bastable, H.G., Gunston, H.M., Shela, O., Chirwa, A., Kafundu, R., 1995. The impact of land use change on water resources in sub-Saharan Africa: a modeling study of Lake Malawi. J. Hydrol 170, 123—135.

Charney, J.G., 1975. Dynamics of deserts and drought in the Sahel. Quart. J.R. Met. Soc 101, 193—202.

Coe, M.T., 1998. A linked global model of terrestrial hydrologic processes: simulation of modern rivers, lakes, and wetlands. J. Geophys. Res 103, 8885—8899.

Coe, M.T., 2000. Modeling terrestrial hydrologic system at the continental scale: Testing the accuracy of an atmospheric GCM. J. Clim 13, 686—704.

Coe, M.T., Foley, J.A., 2001. Human and natural impacts on the water resources of the Lake Chad basin. J. Geophys. Res. 106, 3349—3356.

Coe, M.T., Costa, M.H., Botta, A., Birkett, C.M., 2002. Long-term simulations of discharge and floods in the Amazon basin. J. Geophys. Res. 107 (D20). doi:10.1029/2001JD00074. Art. No. 8044.

Costa, M.H., Botta, A., Cardille, J., 2003. Effects of large-scale change in land cover on the discharge of the Tocantins River. Amazonia. J. Hydrol 283, 206—217.

Croke, B.F.W., Merritt, W.S., Jakeman, A.J., 2004. A dynamic model for predicting hydrologic response to land cover changes in gauged and ungauged catchments. J. Hydrol 291, 115—131.

Dregne, H.E., Chou, N.T., 1992. Global desertification dimensions and costs. In: Degradation and Restoration of Arid Lands. Texas Tech. University, pp. 249—282.

Favreau, G., Leduc, C., Marlin, C., Dray, M., Taupin, J.D., Massault, M., Le Gal La Salle, C., Babic, M., 2002. Estimate of recharge of a rising water table in semiarid Niger from 3H and 14C modeling. Ground Water 40 (2), 144—151.

Foley, J.A., Prentice, I.C., Ramankutty, N., Levis, S., Pollard, D., Sitch, S., Haxeltine, A., 1996. An integrated biosphere model of land surface processes, terrestrial carbon balance, and vegetation dynamics. Global Biogeochem. Cycl. 10, 603—628.

Heikurainen, L., 1967. Effect of cutting on the ground-water level on drained peatlands. In: Scopper, W.E., Lull, H.W. (Eds.), Forestry Hydrology. Pergamon, Oxford, pp. 345—354.

IGBP-DIS, 1998. SoilData (V.0): a program for creating global soil-property databases, IGBP Global Soils Data Task, France.

Kucharik, C.J., Foley, J.A., Delire, C., Fisher, V.A., Coe, M.T., Lenters, J.D., Young-Molling, C., Ramankutty, N., 2000. Testing the performance of a dynamic global ecosystem model: water balance, carbon balance and vegetation structure. Global Biogeochem. Cycl. 14, 795—825.

Lambin, E.F., Erlich, D., 1997. Land-cover changes in sub-Saharan Africa (1982—1991): application of a change index based on remotely sensed surface temperature and vegetation indices at a continental scale. Remote Sens. Environ. 61, 181—200.

Leduc, C., Favreau, G., Schroeter, P., 2001. Long-term rise in Sahelian water-table: the continental terminal in south-west Niger. J. Hydrol. 243, 43—54.

Legesse, D., Vallet-Coulomb, C., Gasse, F., 2003. Hydrological response of a catchment to climate and land use changes in Tropical Africa: case study South Central Ethiopia. J. Hydrol. 275, 67—85.

Li, K.Y., Coe, M.T., Ramankutty, N., 2005. Investigation of hydrological variability in West Africa using land surface models. J. Clim. 18, 3173—3188.

Li, K.Y., De Jong, R., Coe, M.T., Ramankutty, N., 2006. Root-water-uptake based upon a new water stress reduction and an asymptotic root distribution function. Earth Interact. 10 (014), 22.

Lorup, J.K., Refsgaard, J.C., Mazvimavi, D., 1998. Assessing the effect of land use change on catchment runoff by combined use of statistical tests and hydrological modeling: case studies from Zimbabwe. J. Hydrol. 205, 147—163.

New, M., Hulme, M., Jones, P., 2000. Representing twentieth-century space-time climate variability. Part II: Development of 1901—96 monthly grids of terrestrial surface climate. J. Climate 13 (13), 2217—2238.

Nicholson, S.E., 1988. Land surface—atmosphere interaction: physical processes and surface changes and their impacts. Prog. Phys. Geogr. 12, 36—65.

Nicholson, S.E., 2000. Land surface processes and Sahel climate. Rev. Geophys. 38, 117—139.

Nicholson, S.E., Tucker, C.J., Ba, M.B., 1998. Desertification, drought and surface vegetation: an example from the West African Sahel. Bull. Am. Meteor. Soc. 79, 815—829.

Olivry, J-C., Chouret, A., Vuillaume, G., Lemoalle, J., Bricquet, J-P., 1996. Hydrologie du Lac Tchad, ORSTOM. monographie hydrologique 12, 266.

Pereira, H.C., 1992. Keynote paper. In: 10th World Forestry Congress, Proc. 3, Paris, 1991, pp. 139—150.

Pouyaud, B., Colombani, J., 1989. Les variations extrêmes du lac Tchad: l'asséchement est-il possible? Ann. Géographie, Paris 98, 1—23.

Ramankutty, N., 2004. Croplands in West Africa: a geographically-explicit data set for use in models. Earth Interact. 8 (23), 1—22.

Ramankutty, N., Foley, J.A., 1999. Estimating historical changes in global land cover: croplands from 1700 to 1992. Global Biogeochem. Cycl. 13, 997—1027.

Refsgaard, J.C., 1987. A methodology for distinguishing between the effects of human influence and climate variability on the hydrologic cycle. In: Proceedings of the Vancouver Symposium 'The Influence of Climate Change Variability on the Hydrologic Regime and Water Resources', IAHS, Publ. No. 168, Vancouver, August 1987 pp. 557—570.

Savenije, H.H.G., 1995. New definitions for moisture recycling and the relation with land-use changes in the Sahel. H. Hydrol. 167, 57—78.

Shi, P.L., Li, W.H., 2001. Influence of forest cover change on hydrological process and watershed runoff. J. Natural Resour. 16, 481—487, in Chinese.

Sircoulon, J., Lebel, T., Arnell, N.W., 1999. Assessment of the impacts of climate variability and change on the hydrology of Africa. In: Van Dam, J.C. (Ed.), Impacts of Climate Change and Climate Variability on Hydrological Regimes. Cambridge University Press, Cambridge, UK, pp. 67—84.

Stephenne, N., Lambin, E.F., 2001. A dynamic simulation model of land-use changes in Sudano-sahelian countries of Africa (SALU). Agric. Ecosyst. Environ. 85, 145—161.

VanShaar, J.R., Haddeland, I., Lettenmaier, D.P., 2002. Effects of land cover changes on the hydrologic response of interior Columbia River Basin forested catchments. Hydrol. Process. 16, 2499—2520.

Wang, G., Eltahir, E.A.B., 2000. Ecosystem dynamics and the Sahel drought. Geophys. Res. Lett. 27, 795—798.

Xue, Y., Shukla, J., 1993. The influence of land surface properties on Sahel climate. J. Climate 6, 2232—2245.

Zeng, N., Neelin, J.D., Lau, K.M., Tucker, C.J., 1999. Enhancement of interdecadal climate variability in the Sahel by vegetation interaction. Science. 286, 1537—1540.