Hydrological and vegetation response to climate change in a forested mountainous catchment

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Abstract Vegetal cover and the water cycle are closely linked. The climate controls the distribution and productivity of terrestrial vegetation, and the vegetal cover type is a key determinant for evapotranspiration and global runoff. In this study, a dynamic vegetation model (LPJ) has been coupled with a 3D hydrogeological model (MODFLOW) to estimate, for the first time, the impact of climate change on a small forested temperate watershed (Strengbach, Vosges, France). The model, calibrated with monthly hydrological and climate data, is able to globally reproduce the observed vegetal cover distribution and the water cycle over the 1987–2009 period. The discrepancies between calculated and observed intra-annual discharge variations highlight the importance of processes such as snow formation, snowmelt, and catchment recharge after drought periods, as well as the impact of evapotranspiration. Long-term simulations extending up to the year 2100 have been performed with climatic output from the Meteo-France climate model ARPEGE/Climate (IPCC, 2007 scenario A1B). With a predicted increase in temperature of 2.6 °C and a rise in atmospheric CO₂ concentration of 80%, the mean annual precipitation decreases by 4.5% in the

Strengbach watershed over the course of the century. The models cascade predicts a limited decrease of evapotranspiration (by 2.5%), an impact on discharge (11% decrease) at the watershed outlet over the 21st century, and a significant change in vegetation distribution starting in approximately 2085. The response of land plants to climate change in the future seems to only slightly affect the water resources in the Strengbach catchment. This study also highlights existing shortcomings and limitations of simulations for measuring the impacts of climate change.

Keywords Hydrology · Evapotranspiration · Vegetation · Climate change · Modeling · Strengbach catchment

Introduction

Vegetal cover and the water cycle are intrinsically coupled. The water cycle controls the distribution and productivity of terrestrial vegetation (Stephenson 1990; Churkina et al. 1999), and in turn, the vegetal cover type is a key determinant for evapotranspiration and global runoff (Dunn and Mackay 1995). The runoff is constrained by vegetation through stomatal conductance (Field et al. 1995; Morison and Gifford 1983), root pattern, leaf area (Milly 1997; Betts et al. 1997), interception and transpiration (Leipprand and Gerten 2006; Beaulieu et al. 2010).

The impact of climate change on water resources and runoff is thought to be dependent on the responses of vegetation to temperature, atmospheric CO₂ concentrations and precipitation variations. For the Northeastern region of France, global circulation models predict an increase in air temperature up to 4 °C and changes in precipitation patterns under the worst scenario (atmospheric CO₂ doubling before the end of the 21st century) (IPCC 2014). These

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climate changes could influence ecosystems and alter the vegetation dynamics and consequently the water resources and ecosystems services in mountain regions (Björnsen Gurung 2007).

The response of vegetation to elevated CO₂ concentrations could be an increase in carbon uptake by plants (fertilization effect). Previous studies based on modelling or experimental methods showed that stomatal conductance may decrease by as little as 25% (Field et al. 1995) or as much as 40% (Morison and Gifford 1983; Luo et al. 2013), since plants tend to close their stomata to maintain an optimal balance between water loss and CO₂ absorption (Van de Geijn and Goudriaan 1997; Woodward 2002). Under a doubled CO₂ concentration, global evapotranspiration is predicted to decrease by 7%, while runoff and soil moisture are predicted to increase by 5 and 1%, respectively (Leipprand and Gerten 2006).

At the global scale, the geographical distribution of vegetation plays a major role in the variability of annual runoff (Peel et al. 2001). Moreover, a reduction in forest cover leads to a decrease of evapotranspiration and thus an increase of runoff. In addition, global climate changes include changes in precipitation and temperature patterns, which can modify water quantity and quality at the watershed scale (Luo et al. 2013; Morán-Tejeda et al. 2014; Fan and Shibata 2015). Under constant precipitation, Bocchiola (2014) highlighted a decrease in discharge in Alpine catchments over the past decades due to increased temperatures and evapotranspiration. In mountainous regions, the decrease of snowfall and the concomitant increase of rainfall during the winter (Arnell 1999; Arnell and Gosling 2013), accompanied by an earlier melting of the snowpack in spring (Bavay et al. 2013), will increase the flood risk (Fan and Shibata 2015) and will modify groundwater recharge processes, water storage, and therefore, water availability.

The hydrological processes in mountains are thus highly sensitive to changes in climate, and a comprehensive understanding of these processes that control the water balance in mountainous regions is crucial to predict the impact of climate change on these fragile environments.

The aim of this study is to quantify the impact of future climatic changes on water resources and vegetation dynamics in the Strengbach catchment (located in the Vosges Mountains in France). This catchment has been monitored for more than 30 years (Observatoire Hydro-Géochimique de l'Environnement—ohge.unistra.fr). Its hydrological and geochemical functioning is well known, allowing for model calibration. We use a models cascade combining a dynamic vegetation model (LPJ) and a hydrogeological model (MODFLOW). We first calibrate the models cascade with monthly hydrological and climate data available over the 1987–2009 period. The originality

of our approach lies in the direct comparison of the results of the models cascade with a long-term data series. Then, we use a simulation of the Météo-France ARPEGE climate model (A1B IPCC scenario) to project our models cascade over the years 2009–2098.

Materials and methods

Model description

Cascade of two models: LPJ model and MODFLOW model

Mean climate data (air temperature, precipitation, cloud cover, amount of wet days) measured monthly over the 1950–2009 period at the Strengbach watershed and additional data from Harris et al. (2014) are used to force the LPJ model to estimate water and carbon exchanges between atmosphere, vegetation and soil and to simulate the vegetal cover dynamic (Sitch et al. 2003). The vegetal cover and associated carbon fluxes (such as respiration or net primary production) reach a state of equilibrium after thirty years (1950–1980). The groundwater recharge for the 3D hydrogeological model on each grid cell is computed using precipitation (Prec) transpiration, interception (Int), and soil evaporation (Es), as calculated by the LPJ model (Fig. 1) for the whole watershed. The 3D hydrogeological model is run over the 1987–2009 period to estimate the monthly and annual discharge at the outlet of the Strengbach watershed.

LPJ model

The Lund-Potsdam-Jena Dynamic Global Vegetation Model (LPJ) is derived from the BIOME family of models, and combines a mechanistic representation of terrestrial vegetation dynamics with the calculation of the water and carbon exchanges between soil and the atmosphere (Haxeltine and Prentice 1996; Kaplan 2001). LPJ includes a numerical description of the vegetal cover structure and dynamic competition between plant functional types and soil biogeochemistry. Ten plant functional types (PFTs) are represented: eight forests and two herbaceous environments. The presence or absence as well as the fractional coverage of each PFT depend on individual bioclimatic limits as well as physiological and morphological parameters.

The LPJ model includes a box model representation of the soil (2 layers: a superficial layer from 0.0 to 0.5 m, and a deep layer from 0.5 to 1.5 m). The available water content and the water flux incorporated by roots are calculated for both soil layers. The PFT's transpiration depends on the photosynthetic activity, the soil water content and the

Fig. 1 Schematic representation of water fluxes computed on Strengbach catchment by LPJ and MODFLOW models. The abbreviations of these water fluxes are defined in the text

rooting pattern of the PFTs. The water content in the superficial soil layer is determined by the difference between input water fluxes (precipitation and/or snowmelt) and output water fluxes [interception, transpiration (T_1), percolation (P , Fig. 1) between two layers, and superficial runoff]. The precipitation falls as rain or snow, depending on the daily mean air temperature. All precipitation for days displaying negative temperatures is added to the snowpack. The snowpack begins to melt above 0°C (Gerten et al. 2004). The water content in the deep soil layer is calculated as the difference between water percolation (depending on soil texture and layer thickness) and the water uptake by roots (T_2). The transpiration rate is related to the snow processes because the water available in the two layers depends on snowpack evolution, and the transpiration is calculated from the water content in the soil. Superficial and deep runoffs are calculated when the content exceeds the field capacity in layers 1 and 2, respectively. The soil evaporation is calculated on the simulated fraction of the grid cell not covered by vegetation (Gerten et al. 2004).

The water fluxes depend on the characteristics of the PFTs established on site. The photosynthesis and soil water dynamics are simulated with a daily time step, while vegetal cover structure and PFT population densities are updated annually. The LPJ model is fully described in (Sitch et al. 2003). The LPJ model has been extensively validated against observations of vegetation structure, phenology, water and carbon fluxes as well as their seasonal variability (Lucht et al. 2002; Sitch et al. 2003; Gerten et al. 2004).

Conceptual hydrogeological model

Several studies have already used models to simulate the hydrological and geochemical behavior of the Strengbach catchment system (Goddéris et al. 2006; Viville et al. 2006), but no hydrogeological model has been applied until

now. In this work, a groundwater model has been developed using the existing MODFLOW code to simulate the outlet discharge variations over 23 years. Previous studies (Ladouche et al. 2001; Viville et al. 2006) have shown that discharge variations are explained by the existence of two hydrological compartments with different hydrodynamic characteristics, with the higher compartment being more permeable than the lower one, which is also thicker and which contributes to the base flow. Therefore, the underground has been discretized into two layers of unconfined aquifers. At this stage of our understanding of the hydrogeological processes, the contribution of the vadose zone is neglected because its thickness is not relevant compared to the thickness of the lower hydrological compartments (see “Hydrogeological model calibration”) (Ladouche et al. 2001; Viville et al. 2006). The Strengbach river is not explicitly accounted for because superficial runoff in the catchment is only significant in some areas near the outlet where the soil can be saturated. The total extension of those areas represents less than 3% of the total area (Ladouche et al. 2001).

The MODFLOW model is a numerical model based on Darcy’s law and the mass conservation concept developed by McDonald and Harbaugh (1988). This model solves the three-dimensional groundwater flow equation for a porous medium using a finite-difference method:

$$\frac{\partial}{\partial x} \left(K_{xx} \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{yy} \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_{zz} \frac{\partial h}{\partial z} \right) - W = S_s \frac{\partial h}{\partial t},$$

where K_{xx} , K_{yy} and K_{zz} are values of principal hydraulic conductivity along the x , y and z coordinate axes, respectively; h represents the piezometric head, and W is a volumetric flux per unit volume representing sources and/or sinks of water. S_s is the specific storage of the porous material, and t is time. This is one of the most widely used models to simulate groundwater flow (e.g., Xu et al. 2012; Cho et al. 2009).

Study site description

The Strengbach watershed (80 ha) is a small granitic catchment located on the eastern side of the Vosges Mountains (Northeast France) where hydro-geochemical processes have been monitored since 1986 (Riotte and Chabaux 1999; Chabaux et al. 2005; Rihs et al. 2011; Gangloff et al. 2014; Pierret et al. 2014). The topography ranges from 883 m at the outlet to 1146 m at the catchment divide (Fig. 2a). The mean slope of the Strengbach catchment is 15°. The climate is oceanic mountainous, with monthly mean temperatures ranging from -2 to 14 °C (Viville et al. 2012). The mean air temperature for the Strengbach catchment calculated for the years 1986–2012 is 6 °C (Hydro-geochemistry Observatory of Environment, OHGE data). The coldest month is January, with a mean temperature of -1.5 °C, whereas August is the warmest month, with a mean temperature of 14 °C. The mean annual precipitation is 1370 mm over the last three decades with large inter-annual variations (from 893 to 1713 mm). However, the Strengbach catchment is characterized by low intra-monthly variability in precipitation due to the West-dominant wind regime. The mean annual discharge is 814 mm, also displaying strong inter-annual variations ranging from 494 to 1132 mm/year. The highest flow rates occur during the cold period, and the lowest water discharges occur at the end of the summer season. The outlet discharge fluctuates from a few l/s to several hundred l/s during high flow discharge, corresponding generally to snowmelt (OHGE data). The forest covers 90% of the watershed area and consists of 80% spruces and 20%

beeches. The bedrock is mainly composed of a hydrothermally altered Hercynian base-poor granite (332 ± 2 Ma) (Boutin et al. 1995), and of gneiss on the northern slope (Fig. 2b).

Data acquisition

Climate data

Climate data (atmospheric CO_2 concentration, monthly air temperature, precipitation, amount of wet days, cloud cover) for the Strengbach watershed are provided by OHGE (for the years 1986–2009), and the CRU-TS 3.1 spatial climate data for the simulation over the twentieth century prior to 1986, are set to a spatial resolution of 0.5° latitude versus 0.5° longitude (Harris et al. 2014). As in previous studies (François et al. 1998), we calculate the monthly mean anomalies to correct the monthly climate data from the CRU-TS database (details in supporting information). The same procedure is applied for future projections. The monthly precipitation, amount of wet days, and temperature are calculated using the Meteo-France climate model ARPEGE/Climate (scenario A1B, middle scenario; Déqué 2007) for the years 1951–2098.

Soil properties

Soil information, including textural fraction (clay, silt, sand), bulk density, and porosity, was measured at the site. The mean compositions of the superficial layer (0.0–0.5 m) and deep layer (0.5–1.5 m) were 15% clay,

Fig. 2 Land numerical model representing the altitudes distribution (a) and map showing the bedrock types (b), of the Strengbach catchment with the boreholes (B1–B6)/piezometers (P1, P2, P3, P4) location

Table 1 Hydraulic conductivity (m/s) and storage coefficient (–) values of gneiss and granite bedrock as well as the references for these two parameters

Rock type	Hydraulic conductivity (m/s)	Storage coefficient (–)	References
Gneiss	3.10^{-12} to 3.10^{-4}	10^{-4} to 3.10^{-3}	Heath (1983), Stober (1997), Maréchal and Perrochet (2003), Sieber and Uhlenbrook (2005), Abdelaziz and Merkel (2015)
Granite	6.10^{-11} to 4.10^{-5}	6.10^{-8} to 3.10^{-2}	Rasmussen (1964), Heath and Durrance (1985), Morrice et al. (1997), Stober (1997), Maréchal et al. (2003)

19% silt, 66% sand, and 9% clay, 19% silt, 72% sand, respectively. They were determined from field values averaged over 35 samples from 4 weathering profiles that were geographically distributed on the two slopes of the watershed. The soil texture is used in the LPJ model to determine a soil type (Gerakis and Baer 1999). Parameters are assigned to each soil type, allowing for the calculation of the percolation.

The groundwater model MODFLOW was calibrated by assuming that the hydrodynamic parameters have constant values over defined zones. The geometry of the zones for the first layer were estimated using proton Magnetic Resonance Sounding (MRS) (Legchenko et al. 2011; Vouillamoz et al. 2015) campaigns in 2013 and 2015. MRS provides an estimate of the water content over approximately 10 m depths, and we assume that the water content variability is linked to the underground heterogeneity (Boucher et al. 2015). Five zones have been defined. The initial parameter values have been taken from the literature, accounting for the deep saprolitic layer type (Table 1). These parameter values have been changed during the groundwater model calibration (Fig. 4; “Hydrogeological model calibration”). For the second layer, for which MRS data give no information, the zones number has been reduced to four.

Hydrogeological model calibration

The boundary conditions of the model are a no-flow boundary set at the catchment limits, and a constant head at the outlet. If the lateral limits are known, the aquifer depth remains poorly defined. An arbitrary constant depth of 60 m was chosen, consistent with a previous study (Viville et al. 2006), which suggests that the depth of the fractured granite extends over 45 m.

A steady-state calibration of MODFLOW was first conducted over the Strengbach basin by using piezometers and water levels in numerous boreholes (Fig. 3). Two layers are considered (0–10 and 10–60 m depths; Fig. 4). Each layer is discretized into 5476 cells (20 m × 10 m). The hydraulic conductivities of the two

Fig. 3 Comparison between observed (in 2014) and calculated heads (m) for the available boreholes (B1–B6)/piezometers (P1, P2, P3, P4) in steady state

layers were calibrated independently. Hydraulic conductivity generally decreases with increasing depth (Carlsson and Olsson 1977; Stober 1997). Therefore, conductivity values are smaller than the first layer values, except for the northern zone, where the presence of gneissic rocks in the first layer can make it less permeable (Stober 1997).

Next, a transient simulation of the model was run to determine the discharge at the outlet. The run starts from the steady-state simulation, and the starting date is set at 1950. Storage coefficients were adjusted (Fig. 4) from the literature (Table 1). The storage coefficient of the second layer had to be set at very small values, particularly in the zone near the catchment outlet. Small values can be partially justified by the fact that the underlying granite is fractured and that these fractures tend to decrease storage dramatically (Spitz and Moreno 1996; Lucas and Vaute 2011). In the specific case of the outlet area, this means that the aquifer depth is probably smaller than the value assumed in the model (60 m). This is consistent with the fact that the zone closest to the outlet is a wetland.

Fig. 4 Hydrodynamic calibrated parameters: hydraulic conductivities (K; m/s) and storage coefficients (S; –) of the two layers

Results and discussion

Vegetation dynamic over the last two decades

The simulated vegetal cover stabilizes after 30 simulated years. The LPJ model predicts that the forest covers 95% of the catchment area, consisting of 70% boreal evergreen PFT, and 30% boreal summergreen PFT (deciduous forest). The temperate herbaceous PFT represents 5% of the watershed area over the 1987–2009 period. The distribution of plant species estimated by the LPJ model corresponds to that observed in the Strengbach watershed. The calculated mean biomass at steady state (4 t/ha) matches the 3.7 t/ha measured for the 85-year-old forest (3.68 t/ha; Le Goaster et al. 1991).

The transpiration (water flux taken up by roots in the two layers of the LPJ model) and the water flux intercepted by the vegetal cover represent 28 and 18%, respectively, of the 1987–2009 average precipitation (Fig. 5). The evapotranspiration rate (approximately

3.5 mm/day) is lower than that estimated by Viville et al. (1993) during the summer of 1989 (1.5 mm/day) on the Strengbach catchment but is of the same magnitude as those found at similar stands (Knight et al. 1981; Amiro et al. 2006; Zha et al. 2010; Clark et al. 2012). The low physiological functioning level of the trees and the low density of the stand (Viville et al. 1993) can explain this discrepancy. Moreover, this discrepancy can be explained considering that the simulated vegetal cover corresponds to an ideal healthy forest. However, the trees on the Strengbach watershed, especially spruces, are characterized by defoliation and signs of dieback and decline. Previous studies have reported a decreased rate of evapotranspiration caused by defoliation (Dennison et al. 2009; Clark et al. 2012).

Monthly discharge at watershed outlet

As mentioned above, the 3D hydrogeological model estimates the discharge at the outlet of the Strengbach

Fig. 5 Interception and transpiration calculated by LPJ model over 1987–2009 period on Strengbach catchment. Precipitation is measured on watershed since 1987

Fig. 6 Intra-annual variations in precipitation and discharge calculated by models cascade and observed on Strengbach watershed over 1987–2009 period

catchment using calculated water fluxes from the LPJ model (Fig. 6).

The Nash–Sutcliffe coefficient (NSE) and the modified index of agreement (d_1) were used to estimate the quality of discharge predictions by models cascade (Legates and McCabe 1999; McGuire et al. 2002). The NSE and d_1 values between the simulated and observed monthly discharge are equal to 0.48 and 0.68, respectively. Strong (NSE superior to 0.7; Fig. 7a) or moderately strong (NSE superior to 0.5) correlations between calculated discharge and observed discharge emerged in 50% of the simulated period. The difference between simulated and measured outflow during some periods can be explained by the following:

1. an underestimation of the recharge of the watershed in autumn after a drought during spring and summer periods (Fig. 7b);
2. a snow formation during the winter period and a snowmelt in April (Fig. 7c); and
3. an overestimation of calculated evapotranspiration (Fig. 7d).

Figure 7b illustrates the first explanation, using the 2003/2004 period. The mean monthly precipitation was 61 mm over the February–September period of 2003 (the standard deviation is 20.1 mm). The strong increase in precipitation (by 106 mm) in October leads to an increase of 86 mm of simulated discharge. The observed discharge slightly increases (by 20 mm) because of the catchment

Fig. 7 Comparison of calculated (red line) and observed (dotted line) discharge during four hydrological years: 1989/1990 (a), 2003/2004 (b), 2008/2009 (c), and 1995/1996 (d). The precipitation (black line)

and evapotranspiration (green line) are represented on a, b, c and on d respectively

recharge. The recharge of the Strengbach watershed is observed during autumn after a drought period that lasted several years (for example: 1992/1993, 2005/2006, and 2008/2009). Otherwise, these differences between observed and simulated discharge could highlight a too-simplified representation of the Strengbach watershed in the hydrogeological model as the vertical discretization and the spatialization of hydraulic parameters.

The snow accumulation and snow thaw partly control the water fluxes from the autumn to the spring (explanation 2). Figure 7c highlights the importance of the two processes for the 2008/2009 period. The snow accumulation occurred in February. The measured discharge is stable, while the precipitation rate and the calculated discharge increase. Then, the snowmelt occurred in April, leading to an important increase in discharge at the outlet of the watershed. In this simulation, the snow processes are taken into account through the transpiration calculation. Indeed, the transpiration rate is calculated from the water available

in the soil and the snow processes control the soil water content (see “LPJ model”).

We performed a complementary simulation (called SWithS) in which the snowmelt module of LPJ model was used to take into account accumulation and melt processes in the calculated water fluxes. Snowmelt (M) is computed using a degree-day method (Choudhury et al. 1998):

$$M = (1.5 + 0.007Pr)T_a,$$

where Pr is the precipitation amount (mm day^{-1}), and T_a is the daily air temperature above 0°C . The water flux from snow processes was integrated in the recharge hydrogeological equation (“Cascade of two models: LPJ model and MODFLOW model”). Figure 8 shows the comparison between the observed discharge and calculated discharge, with and without the snow module. The simulated average monthly discharge is strongly underestimated by 48% in the winter period (December to March) and is overestimated by 59% the rest of the year (Fig. 8) in the SWithS

Fig. 8 Mean monthly discharge observed (Obs), calculated by models cascade without (SWithoutS) and with (SWithS) snow module on Strengbach watershed over 1987–2009 period

Fig. 9 **a** Comparison of simulated and observed discharge during the period 1987–2009. **b** Interannual variations in simulated and observed discharge

simulation, compared to the observed discharge. The LPJ model predicts snow formation from December to March and snowmelt mainly over the spring period (April to May). These results are consistent with previous studies (Gerten et al. 2004). The underestimation of discharge in winter and the overestimation in spring are probably related to the linear interpolation of the monthly air temperature to daily values in the LPJ model. Moreover, a simplified representation of snow processes could lead to an overestimation of accumulated snowpack and thus to failure in simulating short-term snowmelt events. Therefore, the snowmelt module will not be used in the following sections.

Finally, an overestimation of the calculated evapotranspiration by the LPJ model (explanation 3) could lead to an underestimation of the calculated discharge for June to August, and there was a low index of agreement between the calculated and observed discharge (Fig. 7d). This potential overestimation of evapotranspiration is discussed in “[Vegetation dynamic over the last two decades](#)”.

Annual discharge at watershed outlet

The NSE and d_1 values rise to 0.56 and 0.85, respectively, if we compare mean annual discharges, illustrating a better

model performance at the annual scale than at the monthly scale.

Figure 9 compares the observed and simulated inter-annual variability of discharge over the period 1987–2009. The simulated and observed trends in river flow are significantly correlated. The simulated annual mean discharge (770.2 mm/year) is close to the measured value (817.6 mm/year). The mean low water discharges (from May to October) are well reproduced, with the model-data difference remaining below 3%. The mean high water discharges (from November to April) are slightly underestimated (by 7%) in this simulation.

In summary, the LPJ-MODFLOW models cascade is able to reproduce the water budget of the Strengbach catchment at the annual scale over two decades. At the intra-annual scale, the performance of the model declines. Therefore, in the following section, devoted to the evolution of vegetation and the water cycle over the next century, the discussion of the results will be based on annual mean values only.

The hydrological response to climate change over the next century

The A1B IPCC scenario predicts an increase in the mean annual air temperature (Fig. 10) by 2.6 °C between 2010

Fig. 10 Temperature and precipitation predicted over 2010–2098 period on Strengbach watershed

Fig. 11 Evapotranspiration and discharge estimated by models cascade over 21st century on Strengbach catchment

and 2098 over the area surrounding the Strengbach watershed, and a decrease in mean annual precipitation by approximately 4.5% (from 1346 to 1285 mm/year; Fig. 10).

The models cascade predicts a decrease in discharge by 11% over the 21st century (Fig. 11). The change in hydrologic behavior partly depends on the response of the biome to climate change. The transpiration and interception calculated by the LPJ model increase slightly by 2.6 and 2.2%, respectively. The mean evapotranspiration over the last 10 years of the 20th and 21st centuries are, respectively, 609 and 638 mm/year, corresponding to a 5% increase. The increase in atmospheric CO_2 concentration and temperature change the rate of plant evapotranspiration (water uptake by roots and evaporation through the stomata). On the one hand, the higher temperatures could increase transpiration by changing the vapor pressure demand (Van de Geijn and Goudriaan 1997; Apple et al. 2000). On the other hand, the higher atmospheric CO_2 levels could decrease the transpiration by reduction of

stomatal aperture (Kreuzwieser et al. 2006; Houshmandfar et al. 2015). Indeed, plants tend to close their stomata to reduce their transpiration, leading to an increase in the intercellular CO_2 pressure and eventually a more efficient use of soil moisture (Friend and Cox 1995; Van de Geijn and Goudriaan 1997).

Moreover, the simulated evapotranspiration displays two significant trends: before 2080, and after 2080.

(a) *Before 2080* The decrease of discharge by 14% (Fig. 11) is related to the decrease of precipitation by 6% (Fig. 10) and an increase of evapotranspiration from 629 to 651 mm and then up to 2080 (Fig. 11). We calculated the precision of the evapotranspiration trend to a 90% confidence interval. The wide 90% confidence intervals of this trend [0.01;0.59] suggest that the calculated trend is uncertain. The evolution of evapotranspiration is not related to vegetal cover change because the LPJ model predicts a rather stable vegetal cover distribution until approximately 2080. The evapotranspiration increases by 4% because the air temperature rise promotes atmospheric

Fig. 12 Cover percentage of plant functional type (PFT) calculated by LPJ model as a function of time on Strengbach watershed. The temperate broad-leaved summergreen and needle-leaved evergreen are merged

evaporative demand and consequently vegetation transpiration.

(b) *After 2080* The model predicts a significant decrease of discharge by -17 ± 11 mm/year over the last 10 years of the 21st century. This decreasing trend corresponds to a decrease in precipitation (by -24 ± 17 mm/year) and evapotranspiration (by 5%). However, the wide 90% confidence intervals ($[-7.32; 1.54]$) and the fact that the evapotranspiration trend spans zero suggest that the calculated trend is uncertain. The decrease in evapotranspiration over the last 10 years of 21st century is related to the change in vegetation cover. Indeed, the model predicts a rapid decrease of the boreal evergreen forest and an increase of both temperate deciduous forests and herbaceous environments after 2080 (Fig. 12). In agreement with previous studies, the boreal forests are partly replaced by temperate forests or herbaceous environments due to the increasing occurrence of warmer winter, favoring temperate tree development. At the same time, hot summers lead to boreal forest decline (Joos et al. 2001; Lucht et al. 2006). Moreover, deciduous forests and herbaceous biomes are characterized by a lower evapotranspiration compared to evergreen forests (Zhang et al. 2001; Peters et al. 2013).

Two sensitivity tests have been performed to quantify the individual effects of increasing atmospheric CO_2 concentrations and temperatures on evapotranspiration. In the first test, predictions of evapotranspiration based solely on the scenario of an increase in atmospheric CO_2 concentrations were computed with a constant air temperature and precipitation rate (in 2010). The second test simulates the evapotranspiration variation with an increase in both air temperature and precipitation rate, with CO_2 concentration held constant (at 390 ppm).

In agreement with previous studies (Friend and Cox 1995; Van de Geijn and Goudriaan 1997; Beaulieu et al. 2010), the atmospheric CO_2 concentration increase (up to approximately 700 ppm) leads to the decrease in

evapotranspiration by approximately 8% because of stomatal closure.

The vegetal cover distribution is stabilized until the 2080s. After 2080, a part of the evergreen forest cover is replaced by temperate herbaceous environment. The annual mean transpired water is equal to 408 mm/year at the beginning of the 21st century and decreases to 391 mm/year by the end of the 21st century, which represents a 4% decrease (Fig. 13a). This trend is in agreement with Misson et al. (2002), who found a transpiration decrease in two deciduous forest species in Belgium under doubled CO_2 concentrations. The evapotranspiration decrease leads to the increase of discharge by approximately 6% (from 808 to 853 mm/year).

In the second test, the annual transpiration displays a significant increasing trend from 2010 to 2098, with a rate of 0.7 ± 0.15 mm/year. The LPJ model predicts a rather stable vegetal cover distribution and no significant change from leaf interception during the next century (a decrease by 0.002 mm/year; Fig. 13b). The transpiration increase predicted by the model is consistent with the results of previous studies, which reported that the simulated annual transpiration increased due to the increase of atmospheric evaporative demand and stomatal opening (Misson et al. 2002).

The change in simulated discharge due to the transpiration change is -0.70 mm/year from 2010 to 2098.

Conclusions and perspectives

The models cascade that combines a dynamic vegetation model and a hydrogeological model is able to reproduce the vegetal cover distribution and the annual discharge at the outlet of the Strengbach watershed over the last two decades. Based on the ability of the model to reproduce recent past conditions, we explore the response of the

Fig. 13 Simulation of transpiration, interception and discharge for the 21st century in the Strengbach watershed for **a** increased atmospheric CO₂ concentration, and **b** increased temperature

Strengbach water budget to the future climate (21st century).

The salient results are as follows:

1. Climate change (A1B IPCC scenario) over future decades leads to a decrease in discharge by 11%, principally controlled by a reduction in precipitation. The evapotranspiration variations due to plant physiology and vegetal cover changes are not significant.
2. The impact of increased air temperature on land plants offsets the impact of increased atmospheric CO₂ concentration. Therefore, the simultaneous increase of these two climatic parameters seems to minimize the effects of climate change on discharge evolution. This conclusion is in agreement with previous studies, which demonstrated that manipulation of single factors often lead to stronger response of land plants than multi-factor experiments caused by complex and nonlinear process interactions (Zhou et al. 2006; Luo et al. 2008). The predicted variations remain very close to the uncertainties of the model used.
3. The simultaneous decrease of evapotranspiration and precipitation leads to a significant decrease of discharge (26%) over the later years of the 21st century.

Based on our simulations, it seems that the response of land plants to climate change in the future may not strongly affect the water resources in the Strengbach catchment. These simulations represent preliminary results in the study of regional impacts of climate change. However, the models still suffer from some shortcomings, and lack (1) a better representation of snow storage and melt, (2) data related to the calcium and magnesium nutrient cycle, which was included in the biospheric model to simulate forest decline, and (3) more detailed knowledge of hydraulic parameters in heterogeneous soils.

Future research should include further modelling of the Strengbach catchment using different scenarios of climate change, including modelling with low (A1/B1 scenario) or high (A2/B2 scenario) carbon dioxide emissions as well as simulations at shorter time scales, such as daily or weekly, rather than monthly simulations.

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