Model investigations of the effects of land-use changes and forest damage on erosion in mountainous environments

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Abstract An event-based distributed rainfall–runoff–erosion model is applied to investigate runoff and sediment yield in a small mountainous catchment in central Switzerland under hypothetical scenarios of forest change. Scenarios such as progressive afforestation and deforestation, and forest damage caused by windstorms, are developed in order to analyse the changes in streamflow, spatial and temporal distributions of hillslope and channel erosion/deposition, and total sediment load. Uncertainties in the predictions related to the choice of the hillslope sediment transport formula are also illustrated. The prediction of the spatial distribution of erosion and deposition patterns is potentially very useful for identifying sediment sources, locating areas for field measurements, as well as a first indicator for hillslope erosion prevention measures.

Key words erosion modelling; land-use changes; rainfall–runoff modelling

INTRODUCTION

Land-use changes are often considered a cause for increased flooding and enhanced erosion. In small mountainous catchments where deforestation or damage due to windstorms and snow avalanches often occur, the proportion of land exposed to erosion may lead to increased risk for downstream areas. Sustainable forest management, facing continuously higher costs and the need to minimize downstream impacts of erosion, requires a proper understanding of the catchment response to changes in the forest and vegetation cover. Distributed rainfall–runoff–erosion modelling plays an important role in this respect.

The modelling of long-term erosion rates in watersheds, such as the approaches based on USLE (Renard *et al.*, 1997) are able to quantify the susceptibility of the land surface to erosion, but are not good indicators of the amount of erosion or deposition in a single large storm event. Event-based models which directly solve the equations of motion for water and sediment are much more appropriate for this purpose (e.g. Mitas & Mitasova, 1998; Nord & Esteves, 2005).

This paper presents results of an application of a simple event-based distributed rainfall–runoff–erosion model to investigate runoff and sediment yield in a small mountainous catchment in central Switzerland under hypothetical scenarios of forest change and forest damage. Scenarios such as progressive afforestation/deforestation and forest damage caused by windstorms are developed in order to analyse the changes in streamflow, hillslope and channel erosion/deposition rates, and total sediment load. The purpose is to show the potential for quantifying the spatial distribution of erosion/deposition and hillslope/channel sediment yield under different land-use change scenarios. At the same time, we illustrate some uncertainties in the predictions, especially related to the choice of the hillslope sediment transport formula.

STUDY BASIN AND DATA

This study is conducted on the Vogelbach experimental basin in the Alptal watershed in central Switzerland. This basin is one of several in the area instrumented by the Federal Office for Forest, Snow and Landscape Research (WSL). The basin covers an area of 1.55 km^2; its altitude ranges from 1020 to 1550 m. Mean annual precipitation in the basin is about 2130 mm, runoff is 1590 mm and evapotranspiration is 540 mm (Burch, 1994). Streamflow and sediment transport are highest in the summer months due to high intensity rainfall events. There is a seasonal snow cover in the winter.

Over 60% of the basin area is forested; the remainder is covered by meadows and used partly for grazing. The average hillslope gradient is 37% with the forest floor covered by dense shrub growth and dead wood. The soil is shallow with a large clay content and low infiltration capacity. There is evidence of overland flow and soil erosion on the steep slopes, and soil creep and landslide processes deliver sediment to the streams locally. The main channel is incised into the hillslopes and consists of a step-pool morphology with occasional cascading sections and exposed bedrock. The median bed sediment particle size, D_{50}, is 120 mm; most bed grains range between 10 and 1000 mm in size. The average main stream channel gradient is 18% (Milzow *et al.*, 2006). Summer floods are responsible for most of the sediment transport and reworking of the channel bed.

The basin is instrumented with a continuously monitoring discharge and sediment transport gauging station at the outlet. Sediment transport is measured by geophones installed on the channel bed at the basin outlet, which record the impulse of bed particles larger than d = 10 mm. Total sediment load is computed by relations between the impulse count and total sediment yield developed from observations in the neighbouring Erlenbach basin (Rickenmann, 1997).

The hydroclimatic data used in this study are hourly precipitation and streamflow records for a selected number of extreme summer flood events (largest annual floods between 1986 and 1989). We focus here in particular on the 1986 flood which lasted approximately 6 hours, reached a peak of 4.7 m^3 s^{-1} and transported between 1100 and 1900 m^3 of sediment. The peak rainfall intensity was 7.5 mm in 10 minutes and total rainfall depth was 46 mm. Land surface data used in the modelling are a 25-m resolution digital elevation model (DHM25, Swisstopo), 100-m resolution land-use map (BONU1975, Geostat) and 25-m resolution soil map (BEK2000, 1:200 000, Geostat) expanded with the field data of Walthert *et al.* (2003).

RAINFALL–RUNOFF–EROSION MODEL

The model used in this study consists of two components: (a) a distributed event-based rainfall–runoff model, and (b) a distributed erosion model. The models are briefly

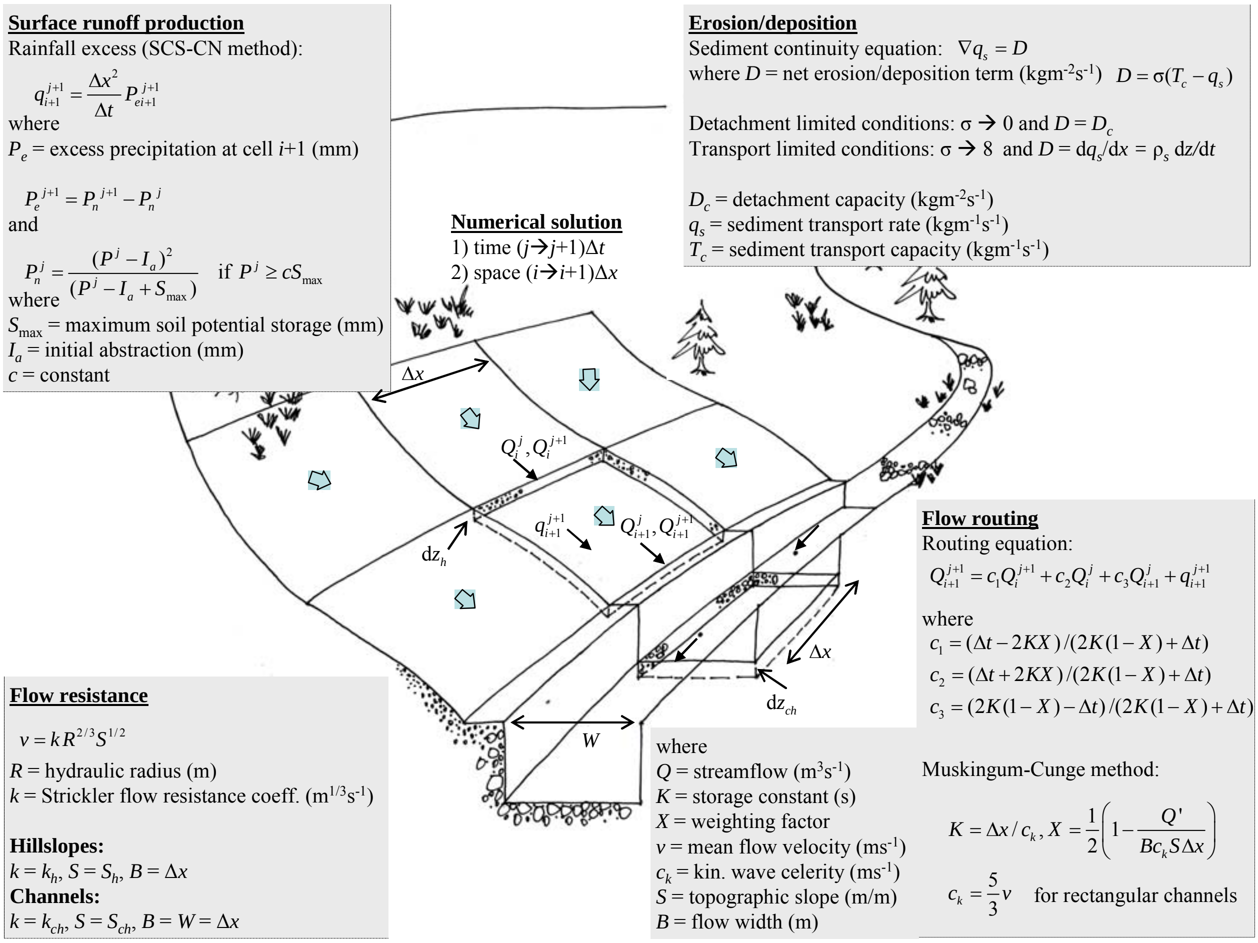

Fig. 1 Schematic of the different components of the rainfall–runoff–erosion model with a list of the main formulae and methods used.

reviewed in this section. Figure 1 illustrates the main mathematical abstractions and most relevant parameters; SI units are used throughout the following description.

Rainfall-runoff model

The distributed rainfall–runoff model used in this study is FEST (Flood Event Simulation Tool, Mancini *et al.*, 1998; Montaldo *et al.*, 2004). This model uses the curve-number (CN) method to determine infiltration losses and effective precipitation, and then routes the produced surface runoff with the variable-parameter Muskingum-Cunge (MC) method (Cunge, 1969) through hillslope and channel cells.

Surface runoff production is a function of soil and land surface properties through the CN number and the initial precipitation storage that needs to be satisfied before runoff occurs. The maximum soil potential retention and the CN number adjusted for Swiss soils were estimated from soil and land-use data (Kuntner, 2002). The CN method is applied in its differential form to each cell in the basin (Mancini & Rosso, 1989). The routing of surface runoff by the variable-parameter Muskingum-Cunge method is an approximation to the diffusive wave in which numerical and physical diffusion are used to arrive at relations between transfer parameters and channel and flow properties (Cunge, 1969; Ponce & Yevjevich, 1978). The routing is applied along flow paths derived by the steepest slope method. Hillslope and channel cells are separated by a constant drainage area threshold calibrated by the digitized river network from the 1:25 000 topographic map. The Manning-Strickler formula is used to describe resistance to flow on both hillslopes (k_h) and channels (k_{ch}). For the hillslope cells, flow is assumed to be shallow and flow width is equal to the DEM cell resolution ($B = \Delta x$), while channel cells have a prescribed channel width ($B \leq \Delta x$) which was measured in the field. The surface roughness coefficients for different hillslope and channel surfaces were taken from the literature (Chow, 1959) and adjusted by calibration.

Although subsurface flow is not directly relevant to surface hillslope erosion it is important for the calibration of the streamflow hydrograph. In the model, subsurface flow is treated as a single linear reservoir or a cascade of linear reservoirs for the basin as a whole (Kuntner, 2002).

Erosion model

The erosion model is based on sediment continuity and simple empirical sediment transport formulae. Assuming that sediment concentration does not vary significantly during the flood event, sediment continuity along the flow path x is:

$$\frac{\mathrm{d}q_s}{\mathrm{d}x} = D \tag{1}$$

where q_s is the sediment transport rate and D is the net erosion/deposition rate ($D < 0$ erosion, $D > 0$ deposition). In general, D may be assumed to be proportional to the difference between the sediment transport capacity T_c and the actual sediment transport rate q_s (e.g. Foster & Meyer, 1972):

$$D = \sigma(T_c - q_s) \tag{2}$$

where $\sigma = D_c/T_c$ and D_c is the detachment capacity. The reaction term σ is dependent on soil particle detachment and land cover properties. As $\sigma \to 0$ we have detachment limited conditions where $T_c >> q_s$ and $D = D_c$. In the case of $\sigma \to \infty$ we have transport limited conditions where $q_s \approx T_c$ and D can be computed from divergence of the sediment transport vector in equation (1) (e.g. Mitas & Mitasova, 1998).

Here we study the sediment transport capacity limiting case, which provides a potential maximum sediment transport and erosion/deposition for a flood event. This is likely the most interesting case to analyse the sensitivity of the basin response in terms of erosion potential. In applying the sediment balance equation to the watershed we separate between hillslopes and channels.

Sediment transport by shallow hillslope overland flow is modelled by two approaches in order to illustrate the non-negligible effect of the choice of the transport formula. Julien & Simons (1985) have shown by dimensional analysis that sediment transport by overland flow q_s can be expressed by a general equation of the form:

$$q_s = \alpha S^{\beta} q^{\gamma} i^{\delta} (1 - \frac{\tau_c}{\tau_0})^{\varepsilon} \qquad (3)$$

where S is surface slope (energy gradient), q is the specific water discharge, i is rainfall intensity, τ_0 is bed shear stress and τ_c is the critical shear stress for particle entrainment. Most existing transport relations can be written in the form of equation (3) with different exponents (e.g. Prosser & Rustomji, 2000). In most overland flow cases, the effect of rainfall intensity is minimal and it is assumed that $\delta = 0$.

The first approach tested here is a runoff-based approach which assumes laminar flow and assumes that $\tau_0 > \tau_c$. In this approach we use the empirical formula developed from flume experiments of Kilinc (1972) for the specific volumetric sediment discharge:

$$q_s = \frac{\alpha}{\rho_s} S^{\beta} q^{\gamma} \qquad (\beta = 1.66, \gamma = 2.03) \qquad (4)$$

where ρ_s is the soil/sediment mass density and S is approximated by the hillslope gradient.

The second approach is a shear stress-based approach typical for turbulent flow:

$$q_s = \frac{K}{\rho_s} (\tau_0 - \tau_c)^{\mu} \qquad (\mu = 1.5) \qquad (5)$$

where K is an effective transport capacity coefficient and μ is an exponent usually in the range $1 < \mu < 1.5$ (e.g. Foster & Meyer, 1972; Mitas & Mitasova, 1998).

Both approaches are very common in current hillslope erosion models. They are however fundamentally different in that equation (4) considers explicitly overland flow velocity in the formulation, while equation (5) does not. Furthermore, if equation (5) is written in terms of exponents β and γ, these take on values significantly different from equation (4); see Julien & Simons (1985) and Prosser & Rustomji (2000) for details.

Sediment transport in the channels is modelled here by the bedload relation for steep coarse-bed streams developed by Schoklitsch (1962):

$$q_s = \frac{2.5}{\rho_s / \rho} S^{2/3} (q - q_c) \qquad (6)$$

where q_s is a specific volumetric sediment discharge, ρ is the mass density of water, and q_c is the critical discharge for incipient motion (Bathurst *et al.*, 1987):

$$q_c = 0.21 \sqrt{g} S^{-1.12} d_{16}^{1.5} \qquad (7)$$

where d_{16} is the bed particle diameter for which 16% of the sediment is finer.

In the coupled rainfall–runoff–erosion model, the volumetric sediment transport capacities q_s are computed at every time step of the rainfall–runoff model by equations (4) or (5) and (6) and then the sediment mass conservation equation (1) is solved by a centred explicit finite difference scheme to arrive at the spatially distributed net erosion and deposition rate D and a vertical depth of erosion or deposition on the hillslopes and in the channels, $\Delta z = (1/\rho_s) D \Delta t$ (see Fig. 1).

Model calibration and validation

The rainfall–runoff model requires as main inputs a distributed map of the land use and land cover (forest, meadow, pasture) which is used to derive the CN number and parameterize the hillslope roughness coefficients, and a digital elevation map for determining flow directions. Channel width was taken to be constant based on field surveys (Milzow *et al.*, 2006). The timing of the flow hydrographs was fitted by adjusting the surface roughness coefficients, the initial abstraction parameters and the subsurface flow reservoir parameters. Calibration and validation was conducted for four large summer flood events and is reported in detail in Kuntner (2002) and Hinz (2004).

The erosion model requires additional data for the sediment transport relations, i.e. soil mass densities, grain-size distributions, transport capacity coefficients α and K, and the critical shear stress for particle entrainment τ_c. These parameters were either estimated from field data or taken from the literature (e.g. Foster & Meyer, 1972; Julien & Simons, 1985) and adjusted to produce the event total sediment yield estimated by a relation developed for the neighbouring Erlenbach basin instrumented with a sediment retention basin (Hinz, 2004).

An important part of this study is the division of the basin event sediment yield by volume into the fine sediment supply from the hillslopes into the channels V_h and the channel coarse bedload transport out of the basin V_{ch}. Hillslope sediment supply is computed as:

$$V_h = \int_T \int_L q_s \mathrm{d}l\mathrm{d}t \tag{8}$$

where q_s is computed from equations (4) or (5), T is the duration of the flood event, and L is the length of the channels in the basin, which is computed from the number n of hillslope cells contributing overland flow and sediment to the channel, i.e. $L = n\Delta x$. The event channel sediment transport V_{ch} is simply the bedload sediment transport volume at the basin outlet:

$$V_{ch} = W \int_T q_s \mathrm{d}t \tag{9}$$

where q_s is computed from equation (6). The total event sediment yield is the sum of the two volumes, $V_t = V_h + V_{ch}$. The model is run at a spatial resolution Δx = 25 m and time resolution Δt = 10 minutes. The most relevant calibrated parameters and their ranges are listed in Table 1.

RESULTS

Parameter sensitivity

An extensive parameter sensitivity was carried out by Kuntner (2002) for the rainfall–runoff model and by Hinz (2004) for the erosion model. For the routing component of the rainfall–runoff model the most sensitive parameters were the surface roughness

Table 1 An example of a calibrated parameter set for the rainfall–runoff model and the ranges of parameter variability for the erosion model for the 1986 flood event.

Rainfall–runoff model		
CN (forest/meadow/pasture)	77 / 74 / 76	
k_h (forest/meadow/pasture)	0.5 / 5 / 5	$m^{1/3}$ s^{-1}
k_{ch}	25	$m^{1/3}$ s^{-1}
λ	0.05	
Erosion model		
d_{16} (channel)	22	mm
ρ_s (channel/hillslope)	2000 / 800	kg m^{-3}
τ_c (forest/meadow)	60 – 100 / 15	Pa
α	6 (10^4) – 2 (10^5)	Eqn (4)
K	5 (10^{-5}) – 1 (10^{-4})	Eqn (5)
μ	1 – 1.5	Eqn (5)

coefficients. For the erosion model, the most sensitive parameters were the transport capacity coefficients α and K and the exponent μ, followed by the hillslope and channel sediment mass densities and the critical bed shear stress for particle detachment on the hillslopes. Several parameter sets within expected ranges of parameter variability were calibrated and used in further analyses (Table 1).

It is known that the absolute magnitude of sediment transport and therefore erosion and deposition are difficult to predict with confidence with empirical flume or field-derived formulae because they are strongly site and event specific (e.g. Julien & Simons, 1985; Prosser & Rustomji, 2000). For this reason our focus here is not on the absolute magnitude of sediment transport and erosion/deposition, but rather on the direction and relative magnitude of change, and on the spatial distribution of potential erosion/deposition when forest change and damage scenarios are implemented.

Lad-use change scenarios

We evaluated eight forest change scenarios in the Vogelbach which were motivated by land-use management, i.e. full/partial afforestation or deforestation, and by long-term climate change impacts, i.e. shifting of the treeline by predefined altitude ranges. Each scenario was implemented by changing the land cover dependent parameters in the models, e.g. forest-related parameters were changed to meadow-related ones, and the differences from the calibrated base scenario for every storm were analysed.

The spatial distribution of simulated erosion/deposition patterns (Fig. 2) shows the expected behaviour: deforestation would lead to widespread basin erosion due to increased erodibility of the land surface, while afforestation would lead to concentrated erosion only on steep hillslopes, mostly close to the channels. Although basin-wide erosion is the dominant process during a large flood, deposition may also occur locally on the hillslopes as well as in the channels. Also illustrated in Fig. 2 is the impact of the choice of hillslope sediment transport parameters. By decreasing τ_c for the forest

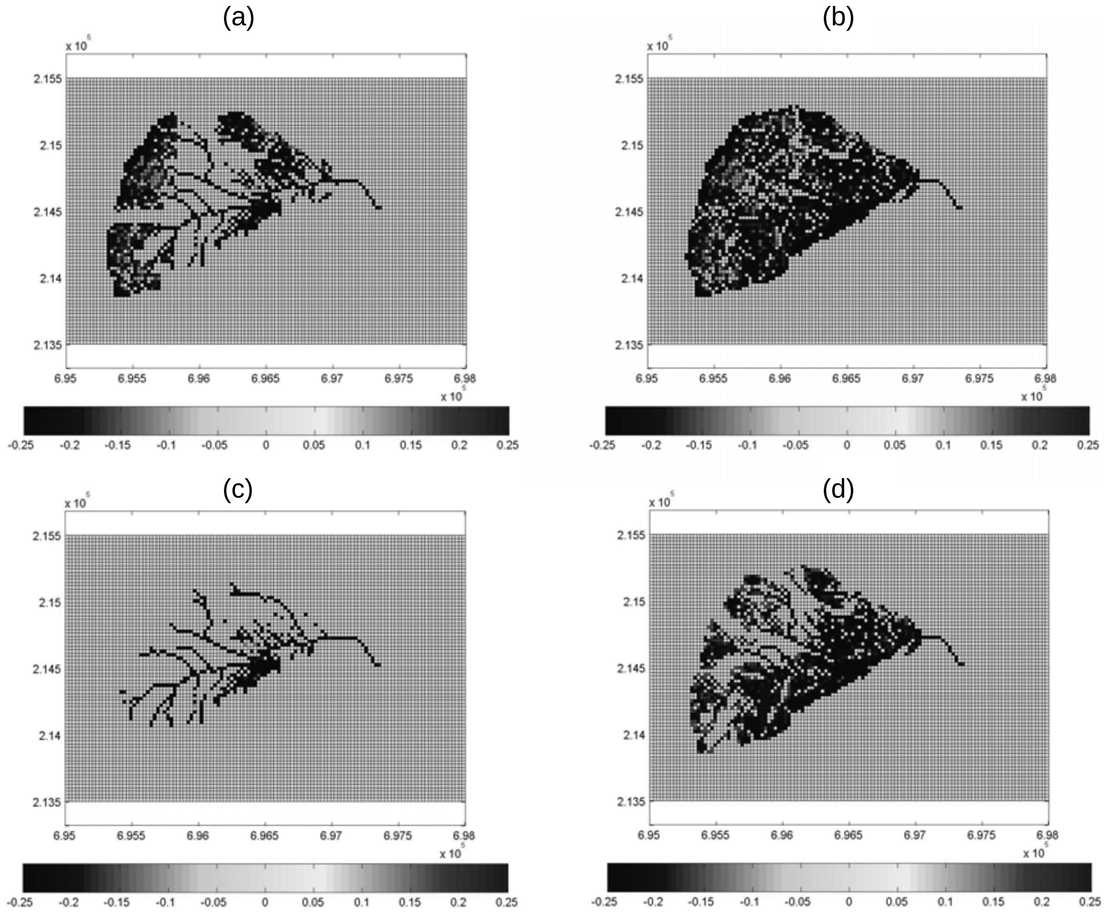

Fig. 2 Modelled spatial erosion and deposition patterns in the Vogelbach for the 1986 storm (units are in kg m^{-2}, negative values indicate erosion, positive values indicate deposition). The erosion parameter set used was τ_c = 100 Pa for the forested areas, K = 5 (10^{-5}), μ = 1.5. Simulations shown are (a) base scenario, (b) complete deforestation, and (c) complete afforestation. Simulation (d) shows complete afforestation with the parameter set with τ_c = 60 Pa for the forested areas, K = 1 (10^{-4}), μ = 1 for comparison.

cover (i.e. increasing the erodibility of the land surface) and linearizing the relation in equation (5) we get a more widespread erosion pattern even in the case of basin afforestation. In practice, this scenario may be representative of a different forest undergrowth with less dense vegetation and thus more prone to erosion.

The temporal effects of changes in the forest cover on streamflow and sediment yield are evident (Fig. 3). Deforestation leads to an increase in peak runoff and an earlier onset of the flood due to a decrease in the surface roughness. The volume of the flood wave is not affected substantially. Afforestation generally has the opposite effect on flood wave propagation. Bedload sediment transport out of the basin closely follows streamflow behaviour. When streamflow is close to the critical threshold for incipient motion, transport may be intermittent, as can be see in Fig. 3 for the deforestation case. The magnitude of hillslope fine sediment supply to the channels is significantly affected by the choice of the transport formula. Because sediment transport computed by equation (4) which has a highly nonlinear dependency on

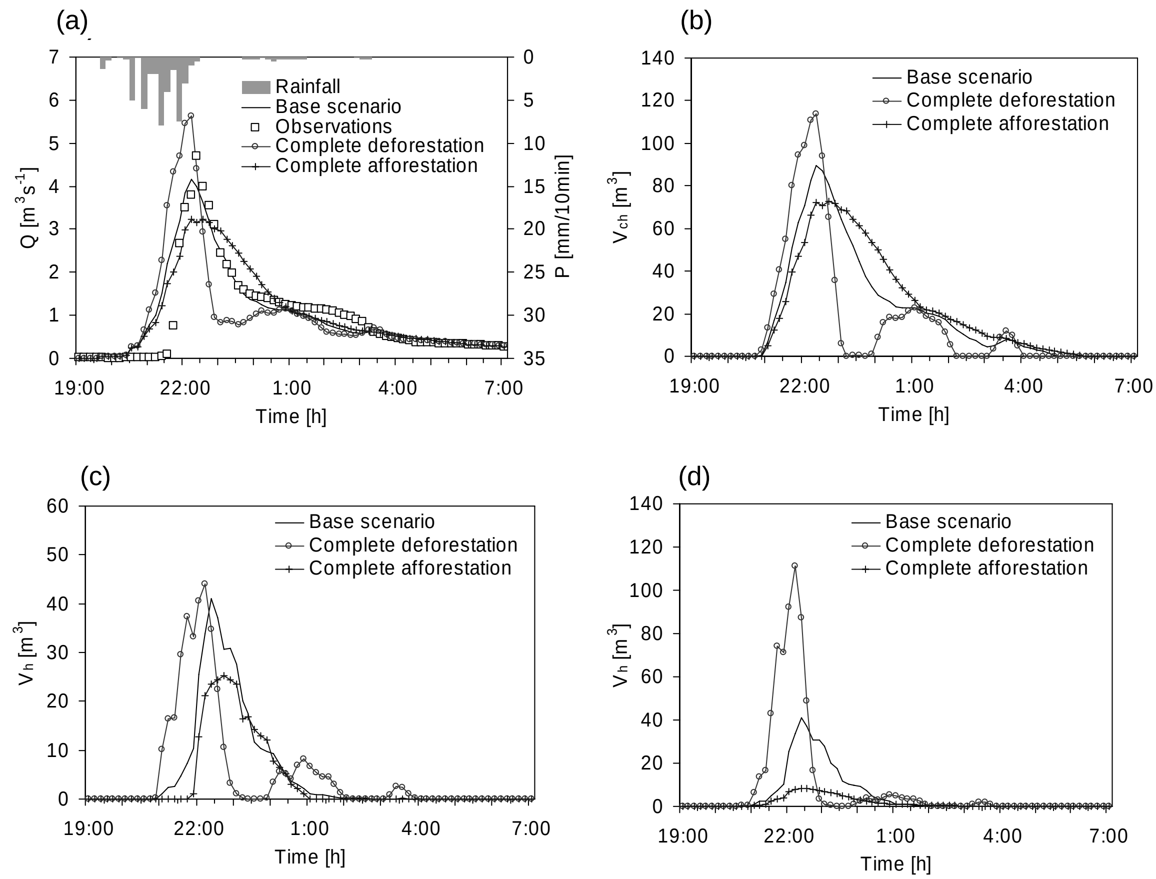

Fig. 3 Modelled time series of streamflow and sediment yield for the complete deforestation and afforestation scenarios compared to the base scenario for the 1986 storm: (a) streamflow, (b) channel bedload sediment transport out of the basin computed with equation (6), hillslope sediment supply to the channels for (c) sediment transport computed with the shear stress-based equation (5) and for (d) sediment transport computed with the runoff-based equation (4). The erosion parameter set for hillslope sediment transport was τ_c = 100 Pa for the forested areas, K = 5 (10^{-5}), μ = 1.5, α = 6 (10^4).

specific discharge in contrast to equation (5), deforestation leads to a much stronger response in sediment yield from the hillslopes using this formulation.

Forest damage scenarios

We evaluated several forest damage scenarios developed on the basis of experiences with the impacts of windstorms Vivian (1990) and Lothar (1999) in Switzerland. Light (20% of basin area) and heavy (80% of basin area) destruction of the forest cover was simulated with a spatial distribution centred on the most wind-prone slopes and valleys. Three hypothetical simulations were conducted: immediately after the storm, two and seven years later. In all cases we also considered whether fallen trees were removed or not from the damaged area, as this has consequences on the soil cover and storage capacity.

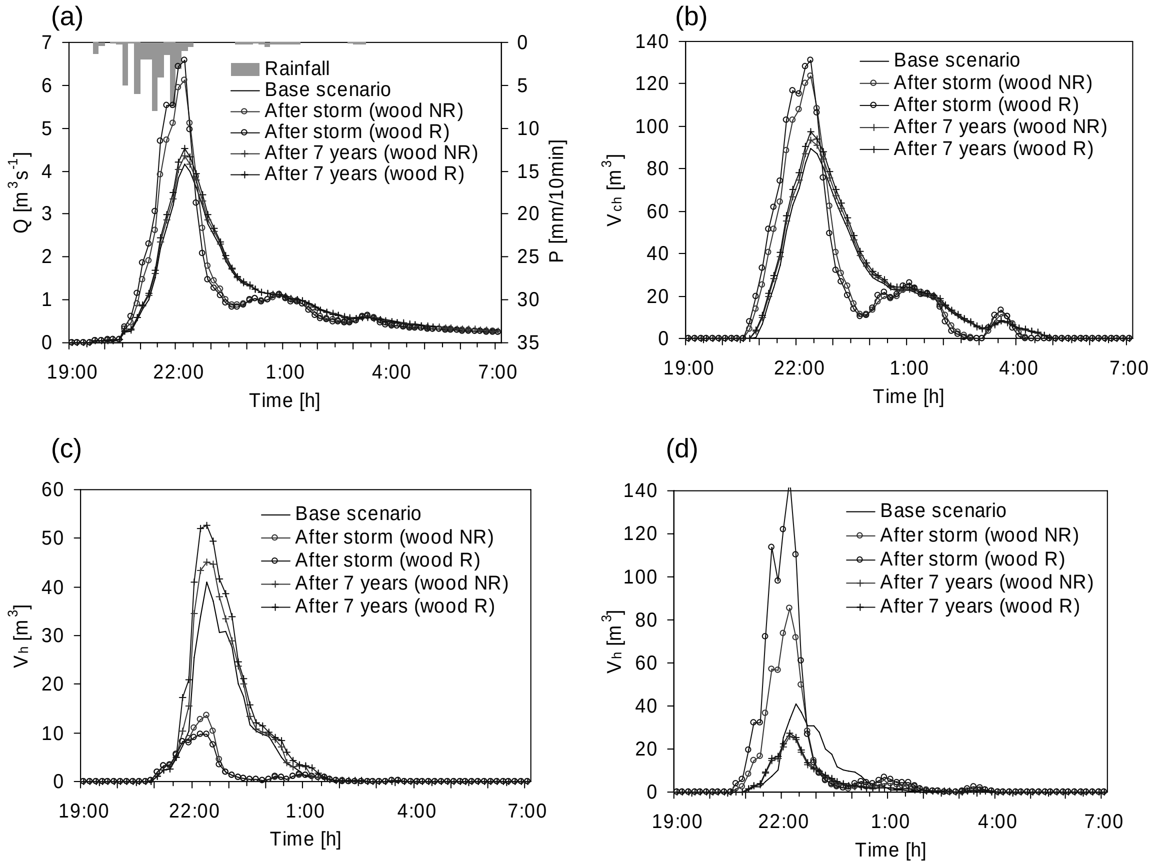

Fig. 4 Time series of modelling results for the 80% forest damage scenario for the 1986 storm immediately after the event and seven years later. Two cases are shown: with wood not removed (NR) and wood removed (R) for (a) streamflow, (b) channel bedload sediment transport out of the basin computed with equation (6), hillslope sediment supply to the channels for (c) sediment transport computed with the shear stress based equation (5) and for d) sediment transport computed with the runoff-based equation (4). The erosion parameter set for hillslope sediment transport was τ_c = 100 Pa for the forested areas, $K = 5\ (10^{-5})$, $\mu = 1.5$, $\alpha = 6\ (10^4)$.

We used the following argumentation for the parameterizations involved in each scenario (Kirsch & Burlando, 2005). Immediately following a storm, tree trunks, branches and leaves form an organic layer on the disturbed soil surface which leads to a decrease in the soil storage capacity S and to a decrease in surface flow resistance $1/k_h$. The disturbed soil surface is more prone to detachment and so the critical shear stress for particle detachment τ_c decreases. Subsequently, after two and seven years we hypothesize that these parameters gradually change towards their original values due to the removal of the organic layer and new tree undergrowth. In the case when fallen and damaged trees are removed from the affected areas, we hypothesize that there is also a decrease in S (partly due to soil compaction by the heavy machinery required to remove trees), a larger decrease in $1/k_h$, a small increase in τ_c, and a gradual recovery in time (Hinz, 2004).

Higher runoff production due to forest damage causes a general streamflow increase at the basin outlet. Figure 4 shows that this response is highest immediately after the storm and gradually returns to pre-storm conditions after seven years (the response after two years lies between these two conditions). The removal of fallen trees in the damaged areas leads to an increase in runoff and sediment yield, and from this point of view it appears not to be an advisable watershed management action after windstorms.

Hillslope sediment yield is again strongly dependent on the choice of the overland transport formula. In this case, the choice leads to opposing tendencies: the shear-stress based equation (5) indicates that sediment yield from the hillslopes would be lower immediately after the storm and would increase in time, while the runoff-based equation (4) predicts the opposite. This is due to the decrease in surface flow resistance, which leads to a reduction in flow depth and therefore shear stress for a large proportion of the hillslopes, in some cases below the threshold for particle detachment, and therefore leads to a reduction in hillslope sediment transport as predicted by equation (5). In the runoff-based formulation, the increase in specific discharge following forest damage directly leads to an increase in sediment transport predicted by equation (4). This illustrates the importance of understanding the driving mechanisms of erosion in relation to overland flow processes. Users of hillslope sediment transport formulae should accordingly be careful in the choice of the appropriate formula because that may substantially affect their results.

CONCLUSIONS

This paper presents a modelling study of hypothetical land-use change and forest damage impacts on runoff and on hillslope and channel sediment yield. The goal is to convey that modelling is a vital tool for predicting the spatial and temporal distribution of water and sediment fluxes in a watershed, especially in cases where it is difficult or impossible to perform field observations. However, the users of rainfall–runoff–erosion models have to be aware of the uncertainties involved in the modelling and the consequences they may have on the interpretation of the results, in particular in studies where land-use change scenarios are modelled by subjectively changing model parameterizations without the support of direct observations. This paper shows that the choice of the formula for hillslope sediment transport is a very important one.

In this study we focused on the "worst case scenario" of potential sediment transport driven by the transporting capacity of the flow. However it is known that sediment transport on hillslopes may be detachment limited in many cases. Detachment limitations are dependent on the relations between flow, land cover and soil erodibility, which are difficult to parameterize, and which include another layer of uncertainty in the modelling. Nevertheless where detachment limitations are important they should be included in the modelling.

Our results illustrate that although absolute magnitudes of erosion and deposition are difficult to predict because of their site and event specific character, relative changes with regard to a defined base scenario may be a good and useful indicator of the extent and direction of land-use change impacts. In our studied case, the loss of forest cover led to an increase in flood peaks and sediment yield which could be quantified. Furthermore, the prediction of the spatial distribution of erosion and deposition patterns is potentially very useful for identifying sediment sources, locating areas for field measurements, as well as a first indicator for hillslope erosion prevention measures.

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