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FINDING THE MOST SUITABLE SLOPE STABILITY MODEL FOR THE ASSESSMENT OF THE IMPACT OF CLIMATE CHANGE ON A LANDSLIDE IN SOUTHEAST FRANCE

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ABSTRACT

The response of a landslide near Barcelonnette (southeast France) to climatic factors was simulated with three slope stability models: a fully empirical gross precipitation threshold, a semi-empirical threshold model for net precipitation, and a fully conceptual slope stability model. The three models performed with similar levels in reproducing the present-day temporal pattern of landslide reactivation, using dendrogeomorphological information as test data. The semi-empirical and conceptual models were found to be overparameterized, because more than one parameter setting matching the test data was identified. In the case of the conceptual model, this resulted in strongly divergent scenarios of future landslide activity, using downscaled climate scenarios as inputs to the model. The uncertainty of the landslide scenarios obtained with the semi-empirical model was much lower. In addition, the simulation of strongly different scenarios by the fully empirical threshold was attributed to its incomplete representation of the site-specific landslide reactivation mechanism. It is concluded that the semi-empirical model constitutes the best compromise between conceptual representation and model robustness.

KEY WORDS: hydrology; slope stability; climate change; numerical model; model uncertainty

INTRODUCTION

Physically based or conceptual slope stability models are potentially valuable tools to assess the impact of climate change on landsliding (Anderson and Kemp, 1988). Their ability to translate changing input conditions to changing impacts in a logical, consistent way is a potential advantage over more empirical approaches. A disadvantage of physically based models is their large data requirement. The more detailed the model, the more input data are necessary to make the detailed representation useful. This makes these models not readily applicable in practical assessments. This can be a particular problem in scenario studies, when input conditions change beyond the range for which the model was validated (Collison, 1996). Furthermore, the level of detail of these models complicates their use in regional (policy planning) studies. For these reasons, empirical models are still widely used.

This paper examines how the impact of climate change on a small landslide in southeast France can best be modelled. For this purpose, three slope models were tested, all involving thresholds for landslide reactivation, but with different levels of complexity:

- a fully empirical climatic threshold, which relates the temporal variation of landslide activity to gross precipitation without considering the physical link between the two;
- a semi-empirical climatic threshold, which relates the temporal variation of landslide activity to a climatological variable which was identified as a key variable from examination of field data;

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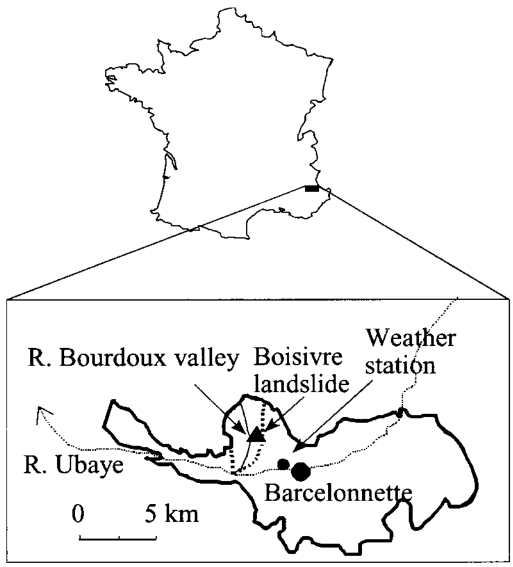


Figure 1. Regional setting of the Boisivre landslide in the Barcelonnette basin (bold line)

- a conceptual slope model, consisting of a hydrological model simulating groundwater levels in the slope, and a model linking groundwater levels to slope stability.

The models were tested against datings of eccentricities in the rings of trees growing on the landslide, considered a proxy for periods of landslide activity. Subsequently they were fed with scenarios of precipitation and temperature.

STUDY AREA

The Boisivre landslide developed on the eastern slope of the Riou Bourdoux valley in the basin of Barcelonnette, Alpes de Haute Provence, SE France (Figure 1). It has been investigated in detail by Mulder (1991) and Caris and van Asch (1991). The length of the landslide is 170 m, its width is 30 to 45 m, and its elevation is 1300–1400 m. The slope angle is 20°. The landslide is composed of seven larger and numerous smaller subunits. Its geology is characterized by a top layer of 1–2 m thickness, consisting of morainic colluvium, underlain by weathered 'Terres Noires' black marls. This in turn overlies unweathered 'Terres Noires' bedrock. Depth to bedrock varies between 4 and 9 m, and averages 7 m. The slide surface runs roughly parallel to the slope surface, at an average depth of about 7 m.

The hydrology is governed by a transition in permeability between the colluvium and the underlying *in situ* marls. This transition favours the development of perched groundwater tables. Tensiometric data indicated that the weathered marls are unsaturated, which further reduces their actual permeability, but regain moisture towards the bedrock. Caris and van Asch (1991) proposed that the slide surface coincides with the top of the bedrock, and that the height of the water table above the bedrock controls the stability of the landslide. They theorized that short-term heavy rainfall would not be the landslide trigger. Although infiltrating precipitation quickly finds its way through the root zone, further percolation is slowed down by the low permeability of the unsaturated marls. The latter process, which supplies the slide surface with water, works on a longer timescale and is probably controlled by the prolonged presence of a perched water table in the root zone. Caris and van Asch (1991) conclude that a persisting perched water table can only occur after several months of high precipitation and low evapotranspiration, when supply outstrips root water uptake. Therefore, net (or effective) precipitation is regarded as a key factor in determining the activity of the landslide.

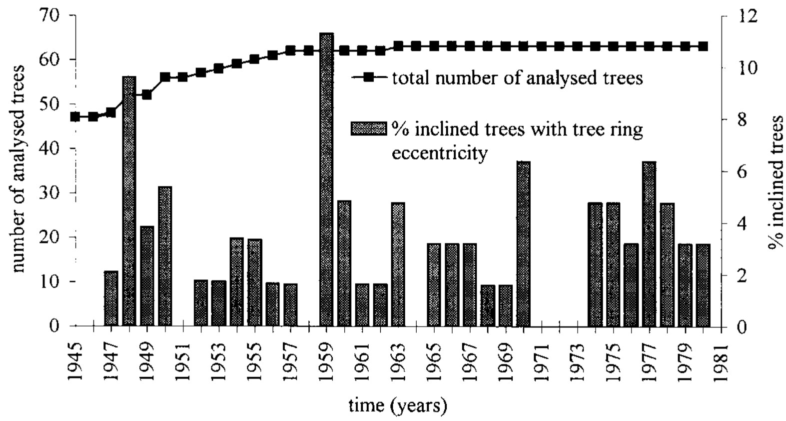


Figure 2. Temporal pattern of ring eccentricities of trees on the Boisivre landslide (after Verhaagen 1988)

Datings of eccentricities of tree rings, assumed to result from slanting, can be used as indicators for years with mass movements (Alestalo, 1971). Verhaagen (1988) analysed ring eccentricities of 63 trees on the Boisivre landslide, using an automated method for processing tree ring data (Braam *et al.*, 1987). Figure 2 shows the temporal variation of tree slanting expressed as a percentage of the total number of analysed trees. The significance of the identification of 'landslide years' is questionable, because it is based on rather few slanted trees (five or six at most). In addition, trees slanted in the same year can often be associated with the same subunit of the landslide, and are located close to each other. In these cases they could indicate a local soil slip rather than activity of the entire landslide. On the other hand, more trees could have slanted, but they could have collapsed and died afterwards, so that they are not available for analysis. Considering only years with slanted trees on more than one subunit, the highest percentages (>4 per cent) of tree ring eccentricities occurred in 1948, 1950, 1959, 1960, 1963, 1970, 1974, 1975, 1977 and 1978. Owing to missing or excessive 'false' tree rings, tree slanting due to other processes like wind action, and counting errors, it seems appropriate to consider a one-year uncertainty margin around these years, i.e. to consider also the preceding and succeeding years.

THREE SLOPE MODELS FOR THE BOISIVRE LANDSLIDE

Fully empirical climatic threshold (R-model)

Model description. The simplest way to model the response of landslide activity to precipitation is to plot the historical precipitation time series against the temporal pattern of landslide activity, and to find a precipitation threshold, which effectively discriminates between periods with and without landslide activity. This was done for the Boisivre landslide and is called the 'R-model'.

The daily precipitation record (1928–1994) of Barcelonnette 'Le Verger' meteorological station was obtained from Météofrance. This station is located 4 km southeast of the landslide. The field data presented in the previous section suggest that landslide activity is triggered by climatic factors that are working on a timescale of one or several months; therefore precipitation amounts summed over one or several months were considered. Given the crude (annual) timescale of the dendrogeomorphological test data, the calibration of the model proceeded as follows.

- A monthly time series of precipitation was established.
- For each year of the precipitation record, the maximum of 12 monthly values was compared with an *a priori* defined threshold value; if the maximum value exceeded the threshold, the year was considered 'unstable', else 'stable'.

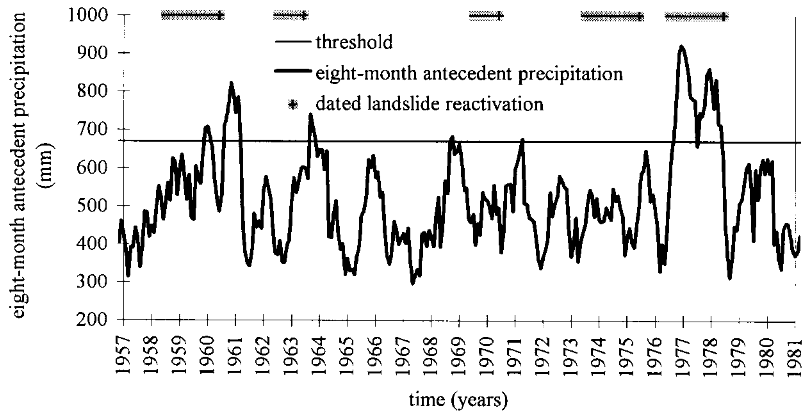


Figure 3. Eight-month antecedent precipitation sums against dated periods of landslide activity

- This series with 'unstable' or 'stable' years was compared with the dendrogeomorphological record, also consisting of 'unstable' and 'stable' years. The one-year uncertainty margin around the years identified in the previous section was thereby taken into account, so that, for example, a record of instability in 1971 of the dendrogeomorphological series may be matched by a record of instability in 1970, 1971 or 1972 of the precipitation series.
- The best fit was determined by minimizing the number of stable–unstable mismatches between the series.
- The optimal precipitation threshold was found by trial and error.

The calibration period is 1956–1980. Note that a 'year' does not mean a calendar year, but pertains to the growing season of trees, roughly April–September. A precipitation event exceeding the threshold in, for example, November, will give rise to a tree ring eccentricity in the growing season of the following year.

Results. Single-monthly precipitation was poorly correlated to landslide activity (not shown). The best results were obtained with eight-month running sums of antecedent precipitation (R8), where a threshold of 670 mm was found to be an adequate discriminator between 'stable' and 'unstable' years (Figure 3). Discrepancies occurred only in 1974 and 1975, which were dry years while high tree ring eccentricity rates were found (see Figure 2).

Semi-empirical climatic threshold (RN-model)

Model description. Although the precipitation threshold of the previous section provided satisfactory results, it does not use any field information obtained by Caris and van Asch (1991) and Mulder (1991), which indicated that not gross but net precipitation is important for landslide activity. Net precipitation in this context is defined as the amount of supply to the perched groundwater table in the root zone. Therefore the R-model was expanded to the 'RN-model' by first calculating the amount of net precipitation on a monthly scale, and using the monthly series thus obtained, in the procedure outlined above. Net precipitation was calculated from monthly precipitation and temperature, using a simple water balance model of the root zone:

$$R + M = AET + \Delta S + RN \tag{1}$$

where $R + M =$ *liquid* precipitation including snowmelt [L T^{-1}], $AET =$ actual evapotranspiration from the root zone [L T^{-1}], $\Delta S =$ change of water storage in the root zone [L T^{-1}], $RN =$ net precipitation or supply to perched groundwater table [L T^{-1}].

Snowmelt (M) must be included because the elevation of the landslide is over 1300 m. It is simulated with a temperature-index formula (e.g. Bergstrm, 1976; Martinec *et al.*, 1983):

$$M = \min[SWE, \xi(T - T_0)] \tag{2}$$

where M = snowmelt rate [L T^{-1}], SWE = snow water equivalent–the amount of precipitation accumulated on days with average temperatures below zero [L T^{-1}], ξ = snowmelt rate [L T^{-1}°C^{-1}], T = mean daily air temperature [°C], T_0 = threshold temperature for snowmelt [°C].

Actual evapotranspiration (AET) is defined in a similar way as by Thornthwaite and Mather (1957):

$$AET = PETf\left(\frac{W_{RZ}}{W_{RZ,MAX}}\right) \tag{3}$$

where PET = potential evapotranspiration [L T^{-1}], W_{RZ} = water available for root water uptake [L], $W_{RZ,MAX}$ = maximum water capacity available for root water uptake [L].

Potential evapotranspiration (PET) was calculated from mean monthly temperature only, according to the model of Thornthwaite (1948), which will not be discussed in this paper.

W_{RZ} and $W_{RZ,MAX}$ are respectively:

$$W_{RZ} = (\theta_{actual} - \theta_{wilt})D_{RZ} \tag{4a}$$

$$W_{RZ,MAX} = (\theta_{field} - \theta_{wilt})/D_{RZ} \tag{4b}$$

where θ_{field} = volumetric water content at field capacity (pF 2–2·6) [-], θ_{wilt} = volumetric water content at wilting point (pF 4–4·2) [-], D_{RZ} = thickness of the root zone [L].

In the present study, a simple linear relation between W_{RZ} and $W_{RZ,MAX}$ is applied (f = 1).

ΔS equals dW_{RZ}/dt. ΔS is negative if AET is greater than $R + M$, in which case RN is zero. ΔS is positive if $R + M$ is greater than AET, *and* if $W_{RZ} < W_{RZ,MAX}$. If W_{RZ} equals $W_{RZ,MAX}$ again, the root zone is at field capacity and drainage towards the perched water table reassumes (so $RN > 0$).

Assignment of values to the model parameters. Four model parameters require (constant) values: D_{RZ}, $W_{RZ,MAX}$, T_0 and ξ. Values for T_0 and ξ were derived from literature (Allewijn, 1990; Kwadijk, 1993). An average D_{RZ} of about 1·50 m was reported by Caris and van Asch (1991), which was used in the model. $W_{RZ,MAX}$ was calculated according to Equation 4b from D_{RZ} and 24 suction-moisture curves measured on root zone samples (Oosterwegel and van Veen, (1988). The parameter values are shown in Table I (shown after the next section). Calibration proceeded as described in the previous section.

Results. Monthly net precipitation was poorly correlated to landslide activity (not shown). Better results were obtained with three-month net precipitation sums (RN3), with a threshold of 270 mm (Figure 4). Again, discrepancies only occurred in 1974 and 1975.

Conceptual slope model (EPL-SLIDE)

Overview. In the previous two models, a link was established between climate and landslide activity, but the exact nature of this link was not considered. van Asch and Buma (1997) attempted to model the activity of the landslide using a conceptual slope model describing this link. A simplified version of this model is described here. The model consists of a hydrological and a stability component. The hydrological component (called EPL) uses net precipitation, as it is simulated in the previous section, as well as geometrical and hydrological characteristics of the weathered marls and the slide surface, for the calculation of a groundwater level. In fact the model is an expansion of the net precipitation model presented in the previous section. The (lumped) groundwater height above the slide surface is assumed to correspond to the pore pressure at the slide

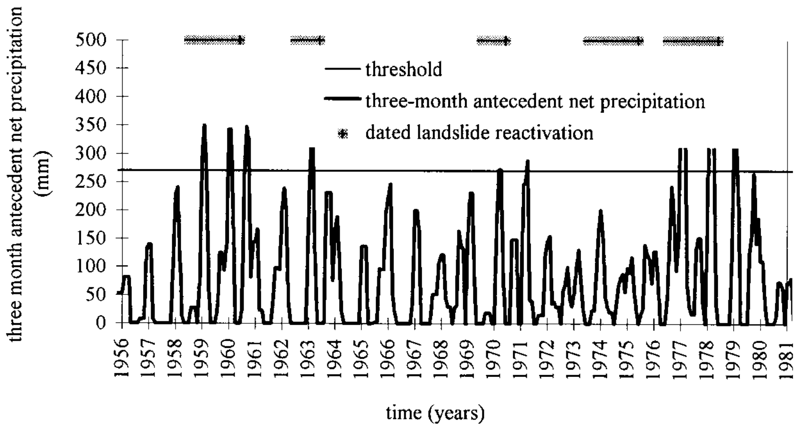


Figure 4. Three-month sums of net precipitation simulated with the Thornthwaite (1948) and Thornthwaite and Mather (1957) models, against dated periods of landslide activity

surface. The stability component (called SLIDE) calculates a groundwater threshold using information on the (lumped) pore pressure and mechanical strength of the material. By comparing the calculated monthly groundwater level with the critical level, the timing of landslide activity is obtained, and this can be compared with the dendrogeomorphological data.

Hydrological component (EPL). The model Estimation of Piezometric Levels (EPL) (Hendriks, 1992) is a linear reservoir that represents the weathered marls, and uses the amount of net precipitation as an input from the perched water table above. The principle behind linear reservoirs is the linear proportionality between water storage and discharge of the reservoir (Figure 5).

$$Q = \frac{S}{k} \tag{5}$$

where Q = discharge [L T^{-1}], S = storage[L], k = discharge coefficient[T].

Also

$$dS = (RN - Q)dt \tag{6}$$

where RN is the input into the reservoir, [L/T], and

$$dH = \frac{dS}{\varepsilon} \tag{7}$$

where H is the groundwater level [L] and ϵ is a storage coefficient [−]. In the present context, ϵ is best defined as the effective porosity (volume of pore space between saturation and field capacity) of the weathered marls.

Although k is an empirical parameter, the model is termed 'fully conceptual' to indicate that all the relevant processes are conceptualized, and to distinguish it from the semi-empirical RN-model presented in the previous section.

Stability component (SLIDE). The stability of a slope mass, bounded by a potential or actual slide surface and the slope surface, is defined by its safety factor F:

$$F = \frac{S}{T} \tag{8}$$

where S is the total of resisting forces and T the total of driving forces. When $F < 1$ the slope is unstable. T is

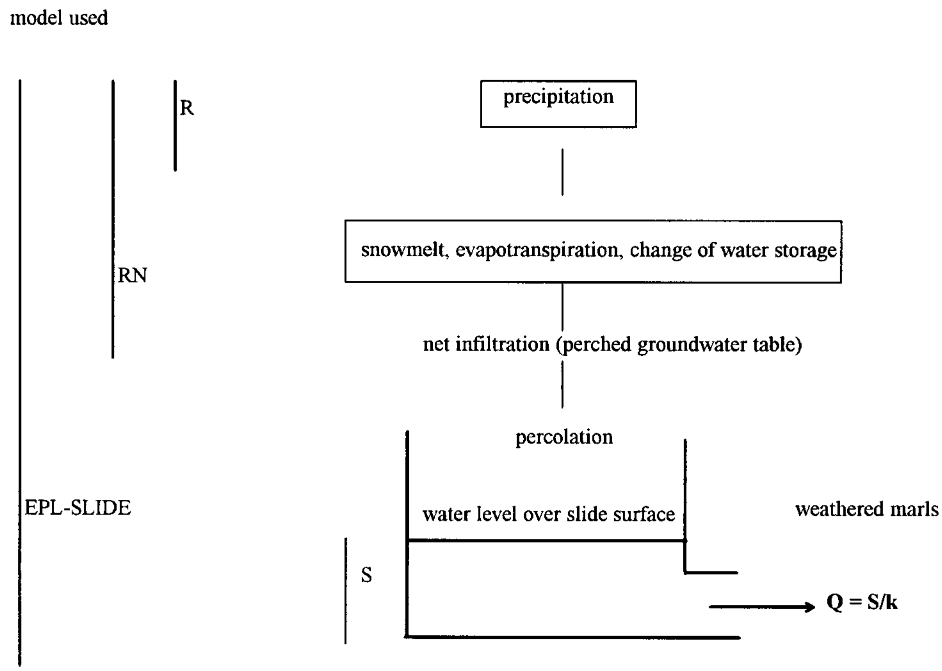


Figure 5. The three slope stability models for the Boisivre landslide. R = fully empirical climatic threshold, RN = semi-empirical climatic threshold, EPL-SLIDE = conceptual slope model

generally determined by the slope-parallel component of the weight of the slope mass. S is described by Coulomb's law:

$$S = C + (\sigma - u)\tan\varphi \tag{9}$$

with S = critical strength [F L^{-2}], C = cohesion [F L^{-2}], σ = total stress [F L^{-2}], u = pore pressure [F L^{-2}], φ = angle of internal friction.

The higher the pore pressure, the lower the intergranular force and resistance at the shear surface, and the lower the stability (safety factor) of the slope.

Several methods exist to analyse the stability of a slope using the theory outlined above. All of these methods divide the slope mass into vertical slices, because generally the forces are not uniformly distributed throughout the soil mass. For each slice, the stability is calculated. The stability of the total slope mass follows from the sum of the results for the individual slices. In addition, assumptions are made about the slide surface; in most cases it is assumed to be circular. In the simplified Janbu method (Janbu, 1957), however, no assumptions are made about the slide surface. Because the slide surface runs roughly parallel to the slope surface, the simplified Janbu method was chosen for the stability analysis. On the other hand, the interslice forces are neglected in this method (this is the simplification compared to Janbu's rigorous method). To account for this, a correction factor is introduced, which depends on the depth/length ratio of the landslide.

The simplified Janbu equation for the stability is as follows:

$$F = \frac{f\Sigma(\{S/\cos^2\theta_i\}\Delta x_i}{\Sigma W_i \tan\theta_i} \tag{10}$$

where S is now given by

$$S = \frac{C' + (\{W_i/\Delta x_i\} - u_i)\tan\phi'}{(1 + \tan\theta_i \tan\varphi'')F} \tag{11}$$

where C' = effective cohesion [F L^{-2}]1, W_i = weight of slice i [F], Δx_i = horizontal width of slice i [L], u_i = pore pressure at the base of slice i [L], θi = angle of the slide surface in slice i, φ' = effective angle of internal friction, f = correction factor (to account for the neglection of interslice forces).

This set of equations must be solved iteratively (since F exists on both sides). If the geometry and geotechnical properties of the slope are known, a threshold pore pressure / groundwater level (corresponding to u) can be back-calculated while $F = 1$. The simplified Janbu method was implemented in the numerical model SLIDE (De Vos, (1990). This model was used in the Boisivre study.

Assignment of values to the model parameters. EPL. The mean thickness of the landslide body (D_{ss}) was derived from geophysical data by Caris and van Asch (1991). Only one suction-moisture curve was available for estimation of the effective porosity ϵ of the weathered marls. The resulting poor degree of physical realism of the model value of this parameter introduces a large amount of uncertainty. A more detailed treatment of this problem is given later in the paper. Finally, the discharge coefficient k is purely empirical. It was used to tune the model to the observations.

SLIDE. The slide surface runs roughly parallel to the slope surface, therefore the angle of the slide surface equals the slope surface in each slice. The landslide geometry (length) was obtained from a DEM, whereas its depth was derived from geophysical data (Caris and van Asch, 1991). The correction factor f was calculated from the depth / length ratio. Geotechnical data on the Terres Noires in general (Antoine *et al.*, 1988), and on the Boisivre landslide in particular (Hazeu, 1988; van Asch *et al.*, 1989) were used. The data comprised saturated and dry weight (to determine the weight of the slope mass), peak and residual values of the angle of internal friction (φ') and cohesion (C') for weathered marls and bedrock. Because the landslide has already failed, residual strength values were used for the weathered marls above the slide surface.

The assigned parameter values are shown in Table I. The value for k is the optimized value giving the best fit (see below). The calibration procedure was similar to the other two models, the difference being that not an optimized precipitation threshold, but the groundwater threshold, back-calculated with SLIDE from the geotechnical data, was used as discriminator between 'unstable' and 'stable' years.

Results. A threshold pore pressure corresponding to a groundwater level of 3·6 m below the surface was calculated. The best fit, shown in Figure 6, is slightly worse than the fits obtained with the threshold models. In particular the years 1974, 1975 and 1980 do not successfully simulate unstable conditions.

Quantitative measures of landslide activity derived from the model results

The frequency of exceedance (or its reciprocal, the return interval) of the respective thresholds of the three slope models was defined as the frequency of landslide reactivation. It was quantified by means of a linear regression according to Gumbel (1958), which relates the absolute values of a set of (in the present case annual) extremes to a double-logarithmic function of their rankings. In the present case, the highest of 12 monthly values (gross precipitation, net precipitation, groundwater) per annum were ranked, giving rise to a series of 30 annual maxima for the period 1960–1989. The rankings are a measure for their return intervals; for example, the highest of the 30 values is considered to be exceeded only once every 31 years, while the lowest value is exceeded 30 times over the same period. If the values of the annual maxima and the double-logarithmic function of their rankings are linearly related, the annual maxima are said to obey a Gumbel distribution. The linear relationship enables return intervals to be calculated for any annual maximum value, such as in the present case a threshold for landslide reactivation. For a more rigorous, mathematical treatment of the Gumbel distribution, the reader is referred to Gumbel (1958) or, for example, standard books on the statistics of meteorology.

In considering only the annual maxima, the duration of critical conditions is discarded, and hence the *degree* of landslide activity: it only indicates *whether* there is landslide activity. There is, however, little point in considering this duration when the mechanics of landslide movement are poorly known. The frequency of

Table I. EPL-SLIDE parameter values yielding the best fit (BEST) for 1956–1980

Parameter	Value	Model
T_0	0°C	EPL-SLIDE, RN
ξ	1·3mm d^{-1}°C^{-1}	EPL-SLIDE, RN
PET	Thornthwaite (1948)	EPL-SLIDE, RN
θ_{RZ}	180 mm	EPL-SLIDE, RN
ρ_w*	2140kg m^{-3}	EPL-SLIDE
ρ_d*	2113kg m^{-3}	EPL-SLIDE
φ'	23·9°	EPL-SLIDE
K	11 d	EPL-SLIDE
D_{SS}	7 m	EPL-SLIDE
ϵ	0·11(−)	EPL-SLIDE

* ρ_w = saturated unit weight, ρ_d = dry unit weight

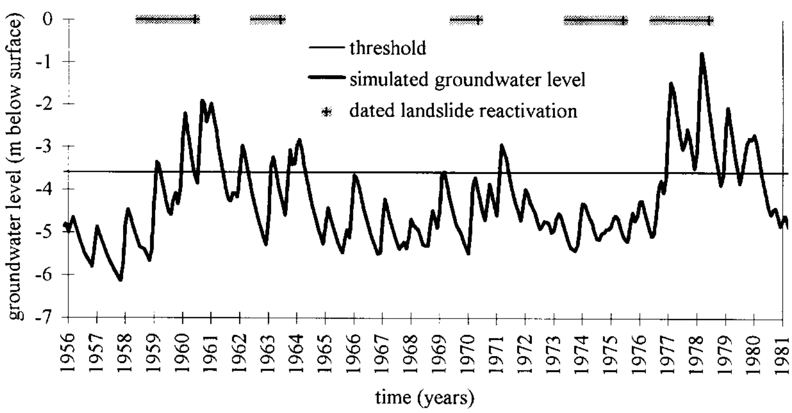


Figure 6. Groundwater levels simulated with EPL (best fit, or BEST), against dated periods of landslide activity

threshold exceedance Φ is therefore regarded as an adequate indicator for landslide activity on the present crude scale.

In all cases, the regression provided a good fit, with r^2 values invariably exceeding 0·90. The regression yielded a Φ of 0·45 a^{-1}. Note that, for example, '1960′ is in fact the period from October 1959 to September 1960, owing to the growing season of the trees. Likewise, Φ values were obtained for both threshold models. For the RN-model a Φ value of 0·27 a^{-1} was obtained. For the R-model, the Φ value was 0·25 a^{-1}, more closely resembling the RN-model Φ value than the value obtained with EPL-SLIDE. The figures do not give clues, however, as to which of the models is better or worse.

Discussion of the results obtained with the three slope models

In general, the three models perform equally well. The years 1974 and 1975 are consistently simulated as 'stable' while the dendrogeomorphological data indicate landslide activity. The slanting of trees in these years involves five trees on three different subunits. The cause of this slanting can only be speculated on, owing to the absence of independent data.

The inclusion of additional parameters in the RN-model and EPL-SLIDE does not give rise to improved model results. Meanwhile, their additional parameters decrease the significance of these models, in particular of EPL-SLIDE. A simplified sensitivity analysis was carried out by changing the individual input parameters by fixed percentages, and calculating the relative change in Φ. The analysis showed that Φ is most sensitive to ϵ, k and φ^*, which are present in EPL-SLIDE only. On the other hand, parameters which are also present in the RN-model are less influential (ξ, T_0, D_{RZ}, $\theta_{RZ,MAX}$). Moreover, they could be estimated more precisely, either from literature values or from field measurements, while this was not the case for k and ϵ. The most interesting results of the sensitivity analysis are shown in Figure 7.

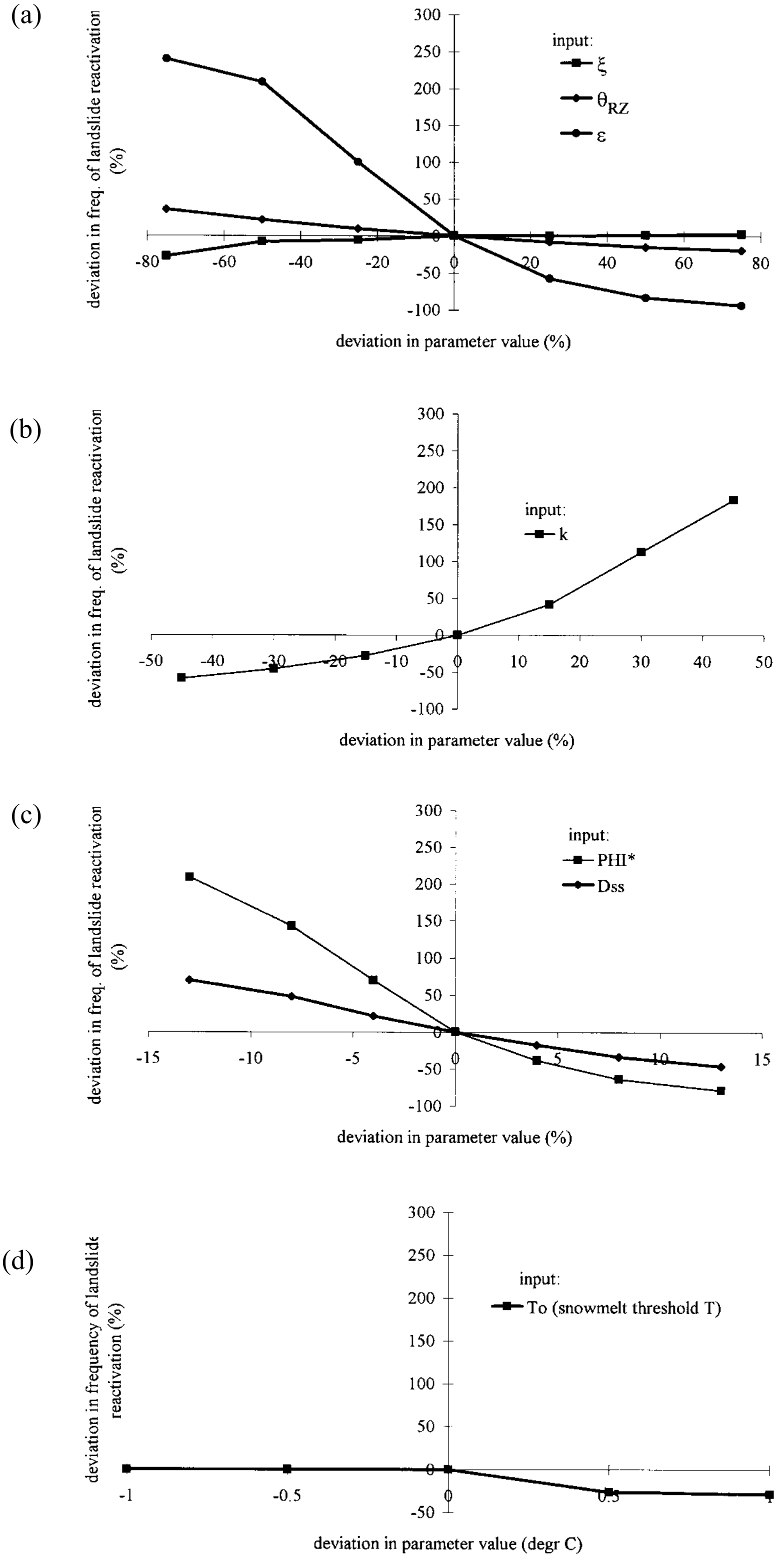


Figure 7. Sensitivity analysis of EPL.

Based on the results presented above, the R-model should be given most credit for its robustness. However, its omission of temperature could have consequences if the model is applied to a perturbed-climate scenario. The influence of temperature in the RN-model and EPL-SLIDE is illustrated by the fact that critical conditions are mostly simulated in March, April and May (not shown). This firstly illustrates the influence of the buildup of a root zone water deficit over summer: almost no critical conditions occur in October and November, which are the wettest months in terms of gross precipitation. Secondly, it confirms the importance of snowmelt for the stability of the slope. The model results therefore confirm that temperature-related processes are important to landslide activity.

In the next sections, an attempt is presented to find the ideal trade-off between conceptual description and model robustness. This is done first by assessing the significance of the parameter settings of the two more complex models yielding the 'good' fits. Secondly, the three models are fed with precipitation and temperature scenarios, while taking into account the significance determined previously.

UNCERTAINTY ANALYSIS OF THE THREE SLOPE MODELS

Procedure

The uncertainty of model results can be assessed by means of advanced error propagation techniques. Such procedures were considered inappropriate in the present study, because only few model parameters were known sufficiently well to establish probability distributions that are suitable for use in, for example, a Monte Carlo simulation procedure. Therefore a cruder alternative was chosen, comprising the following steps:

1. crude estimation of the uncertainty ranges of each model parameter
2. derivation of two 'extreme' parameter settings, within the estimated uncertainty ranges
3. matching of the extreme parameter settings to the dendrogeomorphological data in order to obtain the same result as presented in the previous section.

The procedure was carried out for the RN-model and EPL-SLIDE. There are no parameters in the R-model, except the threshold itself. The uncertainty of this threshold is not assessed here for the sake of the length of this paper, but it was shown to be of minor importance relative to the uncertainty sources discussed later in this paper (Buma, 1998).

Estimation of uncertainty ranges. The uncertainty range was defined by the standard error of the average when there were sufficient field data available. This was the case with D_{ss}, c, $\varphi^{(*)}$ and $\theta_{RZ,MAX}$. In other cases the uncertainty ranges were estimated from literature (ξ, T_0, D_{RZ}). A general underestimation of *PET* by the Thornthwaite model, owing to the discarding of generally low air humidity in Mediterranean regions, was accounted for using data of Palutikof *et al.* (1994). The uncertainty range for ε was made very large, because the effective porosity at greater depth is likely to be strongly influenced by the degree of weathering of the marls, involving an unknown degree of fracturing. Strong fracturing could greatly increase the effective porosity, whereas a slight degree of fracturing could greatly decrease it.

Derivation of extreme parameter settings. Extreme parameter settings were derived as follows. Each parameter was assigned one (of two) values binding their uncertainty ranges. This was done in such a way that the ensemble of parameter values would amplify a (de)stabilizing effect on the slope. In less abstract terms: by increasing the shear strength parameter values of SLIDE, and meanwhile increasing the drainage parameter values of EPL, a parameter setting was obtained yielding lower groundwater levels combined with a higher critical groundwater level for landslide activity, so that the threshold is exceeded less frequently. The opposite was done to obtain the least stable setting. Thus, three parameter settings were obtained: the 'best-fit' setting of Table I (BEST), the most stable one (STAB), and the least stable one (UNSTAB). Table II displays the STAB and UNSTAB settings, corresponding to the estimated uncertainty ranges for each model parameter, and how they were derived.

Matching the extreme parameter settings to the dendrogeomorphological data. In order to maintain a good fit to the observations, the UNSTAB and STAB configurations of EPL-SLIDE were matched by retuning k.

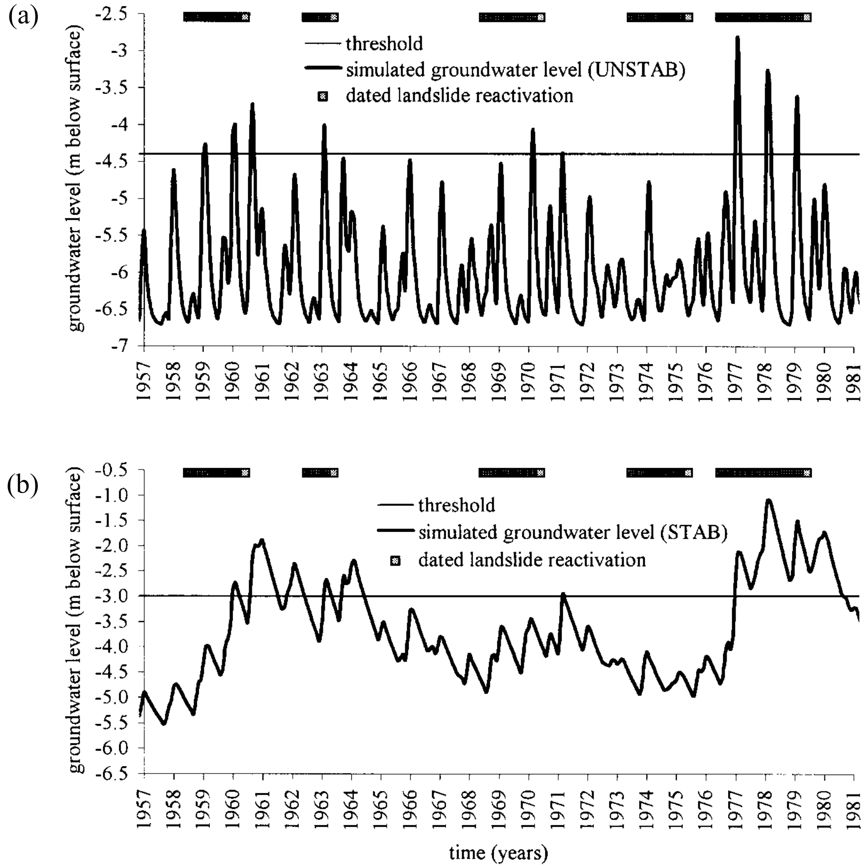


Figure 8. (a) UNSTAB fit simulated with EPL-SLIDE, against dated periods of landslide activity. (b) STAB fit simulated with EPL-SLIDE, against dated periods of landslide activity

Table II. EPL-SLIDE parameter values, falling within estimated uncertainty ranges, yielding STAB and UNSTAB fits for 1956–1980

	Value in UNSTAB-parameter setting	Value in STAB-parameter setting	Ranges derived/estimated from
φ^*	22·7°	25·1°	Van Asch *et al.* (1989)
T_0	1·0 °C	−1·0 °C	Allewijn (1990); Kwadijk (1993)
ξ	3·2 mm d^{-1} °C^{-1}	0·9 mm d^{-1} °C^{-1}	Allewijn (1990); Kwadijk (1993)
PET	—	1·25* [Thornthwaite *PET*]	Palutikof *et al.* (1984)
θ_{RZ}	100 mm	200 mm	Caris and Van Asch (1991)
k	1·45 d	25 d	Calibration
D_{ss}	6·7 m	7·3 m	Caris & Van Asch (1991)
ϵ	0·06 (–)	0·16 (–)	—

* residual angle of internal friction

The adjusted k-values are also shown in Table II. In the RN-model, no matching was possible because all the parameters are fixed at their extreme STAB and UNSTAB values, and no empirical parameter is present for tuning.

Results

Equally good results could be obtained with the three parameter settings of both the RN-model and EPL-SLIDE. Unstable conditions are simulated in exactly the same years. The results for UNSTAB and STAB, compared to BEST, are shown in Figure 8a and b for EPL-SLIDE.

Discussion

It was shown that the temporal variation of landslide activity between 1956 and 1980 could be reproduced with more than one parameter setting in the RN-model and EPL-SLIDE. This indicates overparameterization of both models. This is most severe in EPL-SLIDE where the conceptualization of processes in the weathered marl layer may be correct, but could not be quantified with field measurements. This increases the uncertainty range of the parameters ε and k to which the output of EPL-SLIDE is very sensitive. It is therefore supposed that the additional parameters of EPL-SLIDE, relative to the RN-model, will have only negative effects. The following section shows the consequences of overparameterization of the models for the simulation of scenarios of landslide activity.

APPLICATION OF THE MODELS IN A CLIMATE CHANGE SCENARIO STUDY

Climate scenario

The climate scenario comprised a temperature and a precipitation scenario derived from simulations of large-scale climate by a General Circulation Model (GCM). GCM simulations of the ECHAM4/OPYC3 coupled ocean / atmosphere experiment (Max Planck Institut für Meteorologie, Hamburg) were used. The temperature scenario was obtained by interpolation of GCM temperatures to the Barcelonnette coordinates, followed by a correction for altitude. The precipitation scenario was obtained with a statistical downscaling technique based on canonical correlation analysis (von Storch *et al.*, 1993), using GCM-simulated North Atlantic sea level pressure as a predictor for Barcelonnette precipitation. The downscaling procedure was carried out in cooperation with the Department of Geography of the University of Bonn. The downscaling procedure involved the simulation of 1000 precipitation scenarios in a Monte Carlo experiment. These 1000 scenarios cover the range of possible scenarios of local Barcelonnette precipitation, given a large-scale climatic situation defined by the GCM time series of sea level pressure. For detailed descriptions of the downscaling method and its application to the Barcelonnette area see Dehn and Buma (1998).

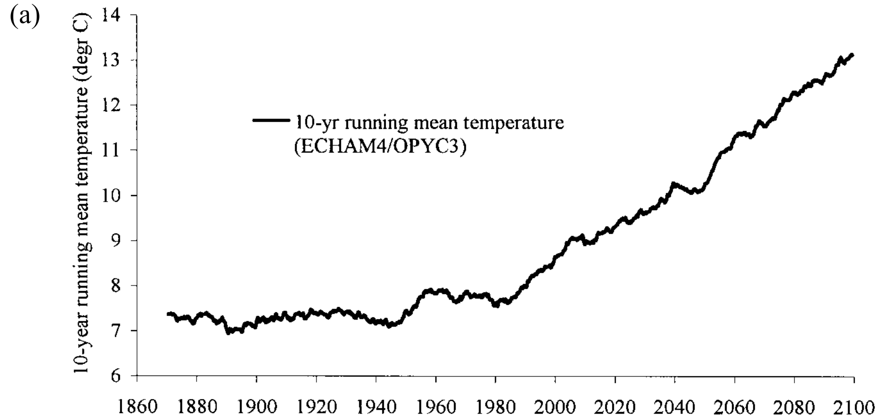
Figure 9a and b show the precipitation and temperature scenarios. A steady upward temperature trend is visible, as well as a general decrease in precipitation, notably in autumn (September–November).

The landslide models were fed with the temperature and precipitation scenarios. Because of the 1000 precipitation scenarios, the slope models were run 1000 times as well, each model yielding 1000 scenarios of landslide activity expressed in Φ values. Φ-values were calculated for running windows of 30 years, shifted at ten year intervals over the scenario period (Oct. 1869 –Sept. 1899, Oct. 1879 –Sept. 1909, . . ., Oct. 2069 – Sept. 2099). In this study, the 0·05, 0·50 and 0·95 quantiles of the 1000 Φ-values are considered for each of these periods.

Results

Figure 10 shows the results of the scenario calculations with the three models using the BEST parameter configuration (0·50 quantile values). The three models all simulate a general decrease in landslide activity in the next century. The most interesting differences occur towards the end of the scenario, where the R-model simulates a less pronounced decrease than the other two models.

To evaluate the influence of model overparameterization, the climate scenarios were fed into the RN and EPL-SLIDE models under the three settings BEST, STAB and UNSTAB. Figure 11a and b shows the resulting scenarios of Φ (0·50 quantile values). The uncertainty of landslide activity, marked by the difference between the STAB and UNSTAB curves, is much larger for EPL-SLIDE than for RN3, in particular if not only the 0·50 but also the 0·05 and 0·95 quantiles are considered, as shown in Figure 12.



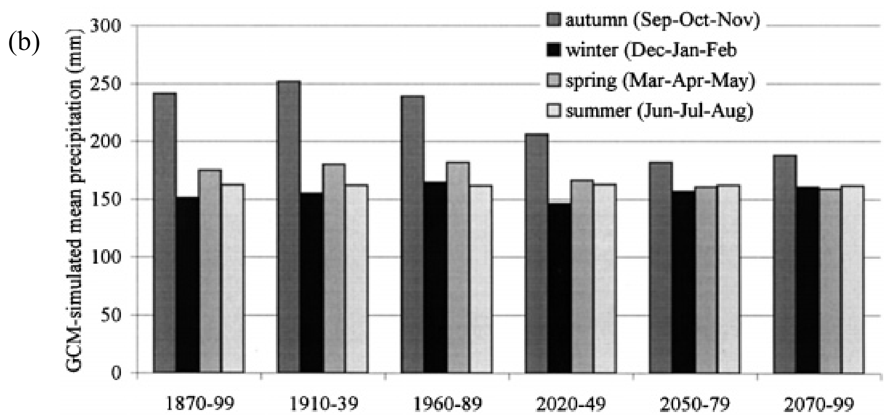


Figure 9. Climate scenarios for Barcelonnette. (a) Temperature (10-year running mean), derived by interpolation from ECHAM4/OPYC3 scenario. (b) Mean seasonal precipitation, derived by statistical downscaling from ECHAM4/OPYC3 GCM-scenario

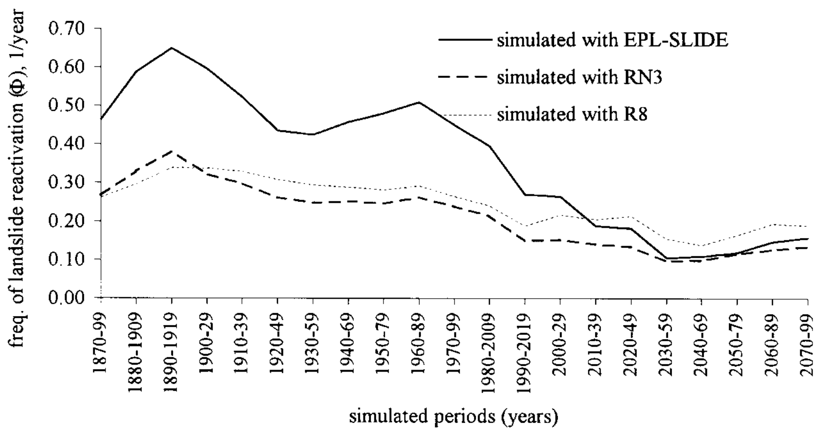


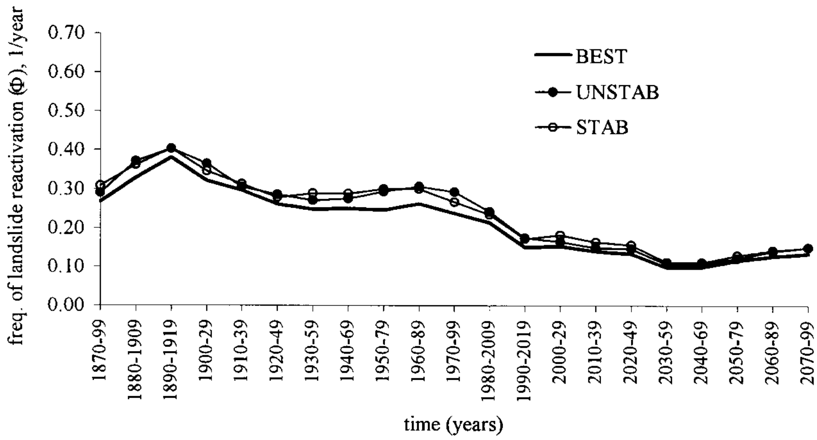
Figure 10. Scenario of frequency of landslide reactivation Φ, simulated with BEST parameter settings (median Φ values of the Monte Carlo simulation)

Discussion

There are two important issues to be addressed.

Firstly, the relatively modest decrease in landslide activity simulated with the R-model can be attributed to the neglect of evapotranspiration. The additional effect of temperature rise on Φ, simulated with the RN-

(a)



(b)

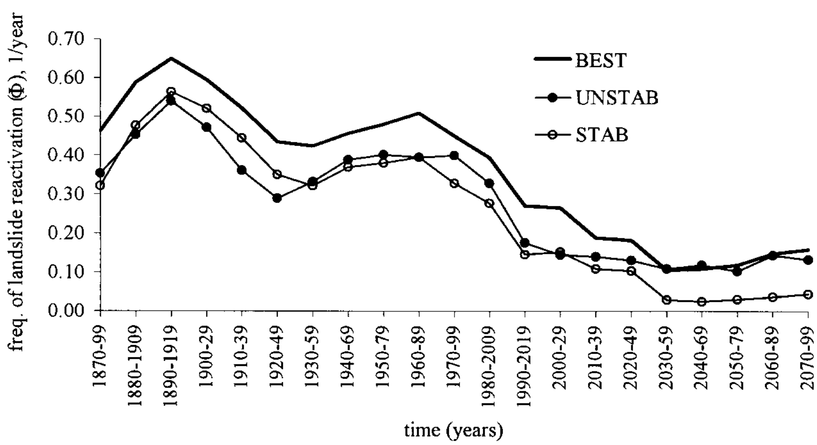


Figure 11. Scenarios of Φ, simulated with BEST, UNSTAB and STAB parameter settings for (a) the RN3-model (0·50 quantiles of Φ) and (b) EPL-SLIDE (0·50 quantiles of)

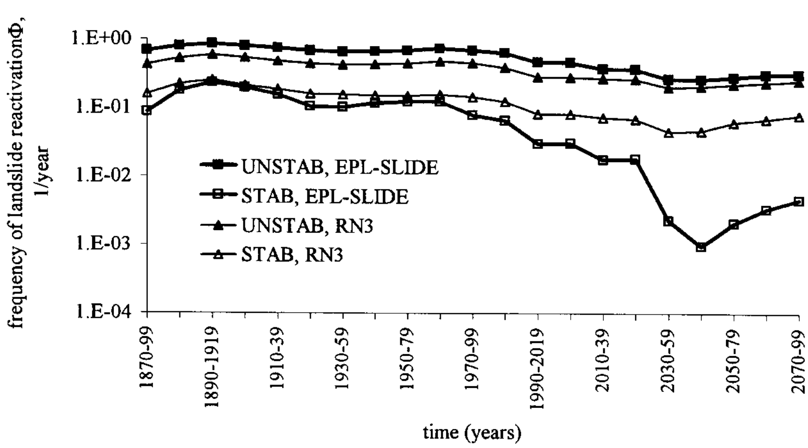


Figure 12. Uncertainty ranges of scenarios of Φ, defined by the 0·05 (STAB) and 0·95 (UNSTAB) quantiles (EPL-SLIDE and RN3). Note the ^{10}log scale

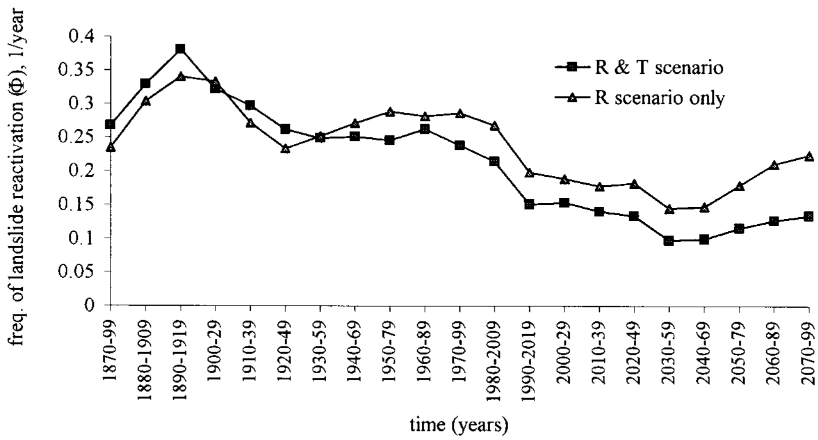


Figure 13. Isolated effect of the temperature scenario on the scenario of Φ. RN3 model, 0·50 quantiles

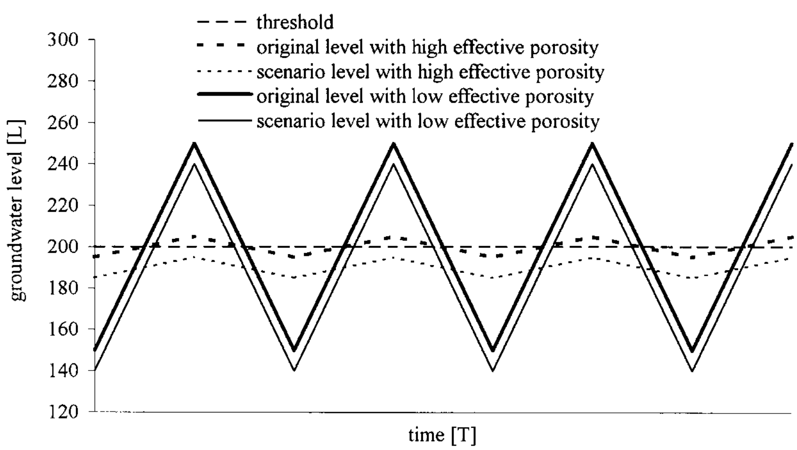


Figure 14. Effect of the ε-parameter of EPL-SLIDE on a groundwater regime subject to an absolute decrease of 10 cm (example). The decrease has little effect on the frequency of threshold exceedance of the high-amplitude series (→lowε; continuous line), but the effect is extreme for the low-amplitude series (high ε; dotted line)

model, is clearly perceivable and is illustrated in Figure 13. This results in significantly different Φ probability distributions for all periods from 1989 to 2019 onwards (Kolmogorov–Smirnov test, $\alpha = 0{\cdot}001$, R-model *versus* RN-model). Since the inference of temperature influence on landslide activity from the field data of Caris and van Asch (1991) is plausible, it is argued that the diverging scenario results disqualify the R-model despite its low data requirement and small number of parameters.

Secondly, the uncertainty of Φ simulated with EPL-SLIDE is much larger than that simulated with the RN-model. Effective porosity ε is a particularly tricky EPL-SLIDE parameter because its value does not influence the simulated *absolute groundwater level* so much, but rather the *amplitude of fluctuations*. This means that for the present-day situation, the frequency of landslide activity is not influenced (see Figure 8a and b), but if the climate becomes drier, a general lowering of the groundwater has different consequences, depending on the effective porosity. A large effective porosity means little fluctuation, while the opposite is true for a low effective porosity. In the (drier) scenario climate a general groundwater decline is simulated. If the amplitude of fluctuations is high compared to the overall decline, threshold exceedances still occur, albeit less often than in the present-day situation. However, if the amplitude is low compared to the overall decrease, the threshold is hardly exceeded anymore. It is these situations that cause the extremely low Φ values in STAB (see Figure 12). The 'ε-effect' is illustrated in Figure 14, and can also be interpreted from Figure 8a and b.

CONCLUSIONS

This study has shown that the performance of a potentially useful conceptual slope model (EPL-SLIDE) is reduced by test data that either are poorly physically based (for example, ε), or have a low temporal resolution (for example, the dendrogeomorphological test data). This performance loss derived not so much from the inapplicability of the model but rather to its uncertainty, and it turned out to have serious consequences in a climate change impact study. The problem could be (partly) solved by monitoring groundwater fluctuations, thus obtaining indications of ε. However, even then it is questionable if the model can be used in practical applications, especially because the piezometer installation and groundwater monitoring require quite large time and capital investments. A better option might be to apply a simpler model using the data already present, which may not be abundant, but should be sufficient to obtain a rough idea of the landslide reactivation mechanism. In regional studies in which much information has to be collected over a large area in a relatively short time, it is better to focus on regionalization with a simple model, rather than to go into detail at one single site.

Two such simple models were evaluated in this study. The fully empirical climatic threshold (R-model) discarded some essential processes and consequently yielded disputable scenario results. The semi-empirical net precipitation threshold (RN-model) turned out to be the best compromise between conceptual description and model robustness. This model is therefore considered most suitable for the presented study of climate change impact on landsliding. It should, however, be realized that even with this model the presented scenarios may be meaningless if the sensitivity of landslide triggering to climate change is small compared to other factors such as material availability, self-stabilization and vegetation–climate feedback mechanisms (Dehn and Buma, 1998). This was not within the scope of this paper, but should be tested in a sensitivity analysis.

The presented conclusions have implications for the regionalization of scenario studies like the one presented here, because it shows that conceptual knowledge about site-specific landslide reactivation mechanisms remains necessary even when simple models are used. The number of required pilot studies depends on the size and complexity of the area under study.

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