MODELLING GROUNDWATER FLUCTUATIONS AND THE FREQUENCY OF MOVEMENT OF A LANDSLIDE IN THE TERRES NOIRES REGION OF BARCELONNETTE (FRANCE)

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ABSTRACT

On the basis of measurements of hydrological parameters and field monitoring of a landslide in the Terres Noires in the basin of Barcelonnette (France), a hydrological model was developed, describing groundwater fluctuations in relation to precipitation. These groundwater fluctuations can be used as input to a stability model in order to assess the temporal frequency of instability of the landslide. The calculated groundwater fluctuations, which can forecast years with landslide incidents, were roughly calibrated against dated movements obtained by dendrochronological research.

The hydrological system of the landslide can be understood through a three-layer sequence: a rather permeable colluvial top layer underlain by a less permeable colluvial second layer, both overlying the nearly impermeable *in situ* non-weathered black marls (Terres Noires). The mean Ksat value for the matrix flow in the top layer is 15·7 cm/day and in the underlying layer 0·7 cm/day. However, water fluxes in these layers occur by two types of groundwater flow: matrix flow obeying Darcy's law, and more rapid gravitational flow through preferential flow paths, increasing the conductivity by a factor of 10 to 100, as cube method Ksat measurements revealed.

The model shows long-term yearly fluctuations of the phreatic surface, with peaks at the end of winter, as well as at the beginning of spring, and minimum values during the dry summer period. These long-term fluctuations are explained by the high drainage capacity of the top colluvial layer and the relatively low vertical water fluxes within the underlying colluvial layer. The model shows that maximum critical peak height conditions of the groundwater, causing instability, occur in wet seasons, with at least six consecutive months with high amounts (more than 60 mm) of precipitation. © 1997 by John Wiley & Sons, Ltd.

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INTRODUCTION

The temporal frequency of landslide movements is an important aspect of landslide hazard assessment. The frequency of landslide movements can be analysed on different scales using different methodologies. The most common approach is statistical modelling of meteorological parameters in combination with landslide incidents. The aim is to find, for a certain area and/or landslide type, meteorological threshold values which are correlated with landslide incidents. Time series analysis gives information about the recurrence of these critical meteorological thresholds and the related landslide incidents (Versace, 1988).

One can also opt for a deterministic approach to understanding and predicting frequency patterns of landslide incidents. The basic concept behind this approach is the relationship of movement frequency to critical groundwater levels within the landslide. Therefore a combination of a more or less detailed deterministic hydrological model and a slope stability model may describe and predict movement frequency for a landslide or a group of landslides (Anderson and Kemp, 1988). This approach requires more detailed information about the hydrological and geotechnical parameters of the landslide material. The study presented here is an example of such an approach. On the basis of measurements of hydrological parameters and

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hydrological field monitoring, a model was developed, describing groundwater fluctuations in relation to precipitation. These groundwater fluctuations can be used as input to a stability model in order to assess the frequency of exceeding the threshold value of the safety factor (F = 1) within a certain time period. The calculated landslide incidents which are related to fluctuations of the groundwater can be compared to measured landslide incidents obtained by dendrochronological research. The use of such combined hydrological and slope stability models for different types of landslides can explain the frequency pattern of landslide incidents, elucidate the most important factors controlling the hydrological triggering system of the movement and predict future activities. The paper will also show that knowledge of the hydrological system is important for the design of remedial measures.

The study was carried out in the basin of Barcelonette (France) where many shallow landslides (4–8 m) have developed in the morainic or colluvial material which covers the slopes of the Jurassic marls (Terres Noires) (Mulder and Van Asch, 1988). Dendrochronological research on tilted trees revealed landslide incidents with a mean frequency of 10–12 years (Van Asch and Van Steijn, 1991).

SITE DESCRIPTION

The landslide in this study lies in the Riou Bourdoux Valley, a tributary of the Ubaye river, situated about 4 km northwest of Barcelonnette. This slope has been reforested with pine trees. In contrast to the trees on the stable part of the slope, most trees on the landslide body are no longer vertical and many trees have fallen down. The bedrock consists of highly erodible Jurassic clayey black marls (Terres Noires) on the lower part of the slopes, and chalky Flysch (Eocene nappe with dark marls at its base) on the higher parts. The surficial landslides in these areas developed on Terres Noires slopes varying from 25° to 35°. During the Weichselian glaciation, parts of the Terres Noires and Flysch were covered with morainic material.

Auger drillings and soil profile pits have revealed that the investigated landslide mass (see Figure 1) is moving down in three blocks consisting of a mixture of Terres Noires marls which are weathered to varying

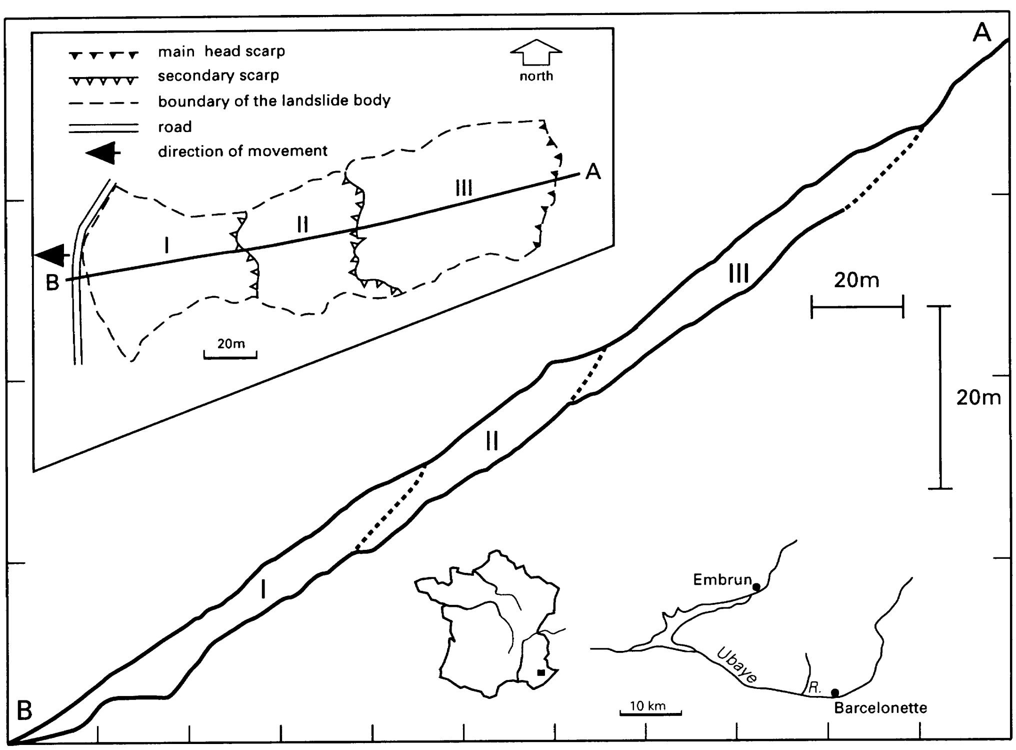


Figure 1. Cross-section and plan of the investigated landslide in Terres Noires colluvial material

degrees, over a slip surface which has developed on top of the more or less unweathered *in situ* Terres Noires. The geophysical investigations (resistivity and seismic refraction measurements) carried out during spring in 1988 and 1989 (Caris and Van Asch, 1991) indicate that the mean depth to the impervious Terres Noires marls is about 7 m (see Figure 1). Furthermore the geophysical data show that there exists an upper layer with many cracks of about 2 m depth and that there is a relatively moist layer close to the bedrock, while the rest of the landslide body consists mainly of unsaturated material.

HYDROLOGICAL INVESTIGATIONS

The inverse borehole method (IILRI, 1974) was used to measure the saturated hydraulic conductivity (Ksat) at different depth intervals in the colluvial material during a relatively dry period without water flowing into the boreholes. Two layers can be distinguished on the basis of these measurements: a relatively permeable top layer (layer 1), at about 1·5 m below the surface, gradually changing into a relatively impermeable layer (layer 2). The measured saturated conductivities vary considerably: the mean Ksat value down to a depth of 1·5 m was 15·7 cm/day (σ = 11·46 cm/day; n = 12), while in layer 2 (from 1·5 down to 7 m below the surface) mean Ksat is 1·5 cm/day (σ = 2·30 cm/day; n = 31). In four cases extremely high flow velocities were observed in borehole measurements carried out in the top layer for which no Ksat values could be determined. It was assumed that these boreholes cross one of the preferential flow paths of a fissure system.

From the hydrological field investigations, it appears that preferential flow paths provide a means for rapid water transport in the upper 1·5 m of the landslide. To investigate the effect of these preferential flow paths on the saturated conductivity of the top layer, the cube method (Bouma and Dekker, 1981) was used (see Caris and Van Asch, 1991). Cubic samples were taken at a depth between 0·25 and 0·50 m. An advantage of this method is that the size of these cubic samples (25 × 25 × 25 cm) increases the change of incorporating preferential flow paths in the sample. Moreover, with the cube method horizontal and vertical Ksat are measured separately, whereas with the inverse borehole method an undefined mixture of horizontal and vertical flow components is measured.

The results of the cubic tests illustrate a much higher vertical permeability of the top layer, due to preferential flow paths caused by vertical cracks (order of magnitude: 10^3 cm/day), compared with the horizontal permeability (about 200 cm/day) (Caris and Van Asch, 1991). Flow takes place mainly through shrinkage cracks and fissures caused by landslide movement, going down to a maximum depth of 1·5 m, and to a lesser extent by pores formed by soil fauna and plant roots. According to Kirkby (1988), these non-capillary voids may behave as a dendritic network, carrying a high proportion of the total hillslope flow towards a relatively small number of discrete seepage lines, which are difficult to recognize. Indeed a few seepage lines were observed downslope (Caris and Van Asch, 1991).

It may be argued that due to the sample size, the cube method gives a better idea of the effect of preferential flow paths on total hydraulic conductivity of the top layer, whereas the inverse borehole method yields correct values of Ksat for the soil matrix.

The main conclusion is that a significant decrease in permeability occurs in the upper 1·5 m and that the permeability of this top layer is mainly controlled by a preferential flow path system. The rapid decrease in vertical permeability points to the development and persistence of a perched water table (Weyman, 1973), inhibiting quick percolation deeper into the landslide body.

To monitor soil water fluxes in the soil, tensiometers were installed to a maximum depth of 6 m. Measurements were carried out in May 1989 and from May to July 1991. These were relatively dry years. Depending on the amount of rain, the tensiometer measurements were made at intervals of several days.

The results of the tensiometer measurements show a relatively wet top layer to a depth of 1·5 m, which dries slowly. The top layer is underlain by relatively dry soil which is only slowly wetted by downward percolation. Vertical drainage of water is slow due to the low unsaturated hydraulic conductivity of the dry underlying layer, despite the large hydraulic gradient between the top layer and the underlying layer. At a depth of 6 m the soil becomes moister again towards the bedrock, which was also shown by the resistivity measurements. If the soil at 6 m is in equilibrium with a water table then this water table is found at a depth of about 7·5 m. This depth coincides with the geophysical data (Caris and van Asch, 1991).

ASSESSMENT OF CRITICAL GROUNDWATER HEIGHTS FOR FAILURE WITH THE STABILITY MODEL OF JANBU

A stability model was used to calculate the critical water height at which the landslide becomes unstable. Because of the flat configuration of the slide (see Figure 1) the Janbu model was used. It appeared that the difference in safety factor values between the simplified Janbu method and the rigorous method was very small. Therefore the simplified method was used to save calculation time. The stability model needs the configuration of the slope and the slip surface. It was assumed that the form of the slip surface follows the boundary between weathered and unweathered material surveyed by refraction seismics (Caris and Van Asch, 1991; Mulder, 1991). For the stability calculations it was assumed that the slip surface has residual strength characteristics. The residual strength values ($\phi^{res} = 23{\cdot}9° \pm 2{\cdot}8°$) of the colluvium were estimated by the assessment of an apparent threshold value for 24 creep tests (Van Asch *et al*., 1989). Given the distribution of these strength values the probability of failure was determined according to the method given by Lee *et al*. (1983) assuming a normal distribution for the strength values. It appeared from the calculations that the landslide has a probability of failure of 50 per cent or higher if the groundwater level rises above a level of –4·2 m below the topographical surface. This level was arbitrarily chosen as the critical level for which the hydrological model was calibrated (see Discussion and Conclusion)

HYDROLOGICAL MODELLING

A simple one-dimensional hydrological three-layer slope hydrology model for transient modelling (GW-Fluct) was developed, based on hydrological theory derived from the observations and measurements given above: a highly permeable fissured top layer of about 1·5 m; a less permeable second layer of 4–7 m; and a third layer consisting of unweathered, nearly impermeable Terres Noires. The model is able to simulate bypass flow caused by a preferential flow path system in layers 1 and 2, as well as percolation from layer 1 to layer 2. Outputs are groundwater levels in layers 1 and 2. So far, the model has only been employed on a monthly basis. In the following, the calculations by GW-Fluct within one timestep will be shown.

(1) The model starts by subtracting a bypass flow discharge from the initial flux *Reff* entering the soil. *Reff* is read from a prepared data file, and represents net precipitation (precipitation minus interception, evapotranspiration, soil moisture storage). Bypass flow in layer 1 is calculated as a fixed percentage of *Reff*.

(2) The remainder *Reff** will reach the saturated zone of layer 1. If the new calculated groundwater level in this layer is greater than the layer thickness, then the excess will be lost as overland flow.

(3) The next step is to calculate the percolation flux from layer 1 to layer 2. The calculation is based on Darcy's law applied to a two-layer system (Figure 2). On the boundary (point B) between layers 1 and 2, the law of continuity implies that percolation equals the flux between points A and B (Q_2). To simulate percolation in a more realistic way, the total amount of S arriving in layer 2 is divided by the observed maximum daily rainfall, to obtain the number of 'rain days' (N) in that month. Percolation is calculated for a rain day, and then multiplied by the number of rain days.

Darcy layer 1:

$$Q_1 = -KSAT1\,\frac{(z_b - z_a) + (P_b - P_a)}{z_b - z_a} \tag{1}$$

where $z_b = 0$, $z_a = R$, Pb
= unknown (x), $p_a = 0$.

R represents the perched water height in layer 1:

$$R = H_1 + \frac{Reff^*}{P1\,N} \tag{2}$$

where N is the number of 'rain days' and $P1$ is the pore space between field capacity and saturation in layer 1.

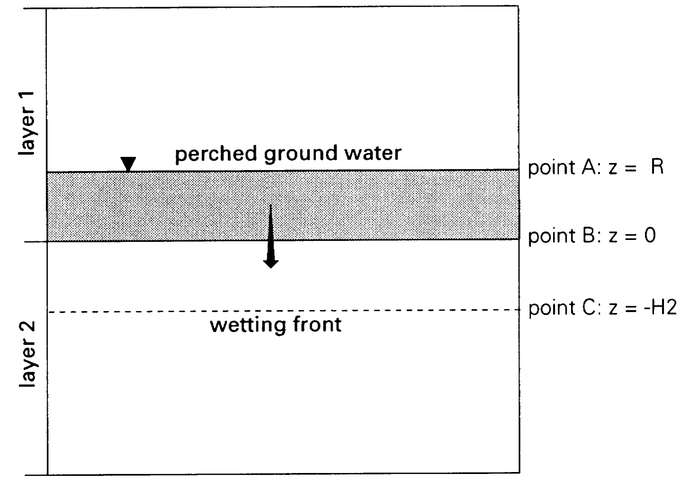


Figure 2. Schematic characterization of the percolation problem in the hydrological model

Darcy layer 2:

$$Q_2 = KSAT2 \frac{(z_c - z_b) + (p_c - p_b)}{z_c - z_b} \tag{3}$$

where $z_b = 0$, $z_c = -H_2$, p_b = unknown (x), $p_c = \omega$ (matric suction at wetting front). H_2 is the distance to the wetting front caused by percolation in the preceding timestep.

Combining the two equations gives:

$$KSAT1 \frac{R - x}{R} = KSAT2 \frac{H_2 - \omega + x}{H_2} \tag{4}$$

or

$$x = \frac{R(k_1 H_2 - k_2 H_2 + k_2 \omega)}{k_2 R + k_1 H_2} \tag{5}$$

When the pressure potential x at B is known, Q_p is easily calculated:

$$Q_1 = N\, KSAT1 \, (-\frac{x}{R} + 1) \tag{6}$$

(4) This amount is subsequently subtracted from the groundwater height in Layer 1, and in the next timestep will be added to the groundwater height in layer 2. This means that in timestep i, the percolation calculated in timestep i–1 will reach the saturated zone in layer 2.

(5) The next step is to calculate lateral hill flow L_i for both layers, again using Darcy:

$$L_i = KSATi\, H_i \frac{\Delta DZ}{\Delta DX} \tag{7}$$

where $\Delta DZ/\Delta DX$ represents the mean hydraulic gradient (hillslope height/hillslope length). In the case the layer 2 groundwater table extends into layer 1, the model considers the groundwater as one unit and calculates L_i with $KSAT = KSAT1\, H_1 + KSAT2\, H_2)/H_1 + H_2)$, subsequently subtracts L_i from H_1, and when $L_i > H_1$ subtracts the surplus from H_2.

When the depth of layer 2 is large compared with the groundwater height in layer 2, a 'delay time' can be calculated to account for the time lag between percolation and arrival at the layer 2 groundwater table. This merely means that in the layer 2 output file, the previously calculated groundwater levels are moved up along the time axis by:

$$t_{delay} = \frac{P2\; dz}{30\; KSAT2} \tag{8}$$

where dz is the distance the percolating water has to travel downward to the layer 2 groundwater table. This distance can be a user-defined constant value, or the distance from the layer 1/layer 2 boundary to the mean layer 2 groundwater depth calculated during the simulation period.

GW-Fluct has a few drawbacks.

- Since the model can only handle one fixed slope angle and one material type per layer, the model is not able to simulate groundwater in hillslopes with more complex lithology and geometry.
- Because unsaturated flow has been neglected in the model so far, timesteps shorter than the time needed for the water to reach the saturated zone cannot be chosen. Therefore in slopes with low permeability, very long timesteps must be chosen.
- GW-Fluct seems suitable for modelling simple landslides with relatively few and low-resolution input data, as long as precipitation, geometry and some dating or hydraulic conductivities are measured.

MODELLING TEMPORAL INSTABILITY OF THE RIOU BOURDOUX LANDSLIDE WITH GW-FLUCT

Model input and assumptions

Table I gives an overview of the mean values and estimated ranges of parameters used in the hydrological calculations and for the sensitivity analysis (see below).

It was assumed that the mean Ksat values measured with the inverse borehole method in the two layers (*KSAT1, KSAT2*) estimated reasonable values for the matrix flow. The mean values were used to calculate the percolation from layer 1 into layer 2 (Equations 4 and 5).

The mean effective porosities between field capacity and saturation (*P1, P2*) are based on pF curves measured on 25 soil samples. The slope height and length were obtained from profiles and maps (see Figure 1). The estimation of thickness of total colluvial material is based on geophysical research (Caris and Van Asch, 1991) while the thicknesses of layer 1 (*D1*) and layer 2 (*D2*) are based on the vertical distribution of Ksat values and tensiometer measurements (see above). The percolation parameter (*PCi*) which is needed to calculate the initial percolation in the first timestep (see previous section) is set on an arbitrary value.

Bypass flows in both layers (*BPF1* and *BPF2*) are parameters which can be used to calibrate the model. As a first attempt the *BPF1* was set at 0·20 (20 per cent). This value is based on a rough estimate of the volume fraction of macropores and fissures in the field. The maximum amount of rain per day is based on daily rainfall data, measured in the basin of Barcelonnette from 1990 to 1994.

Table I. Mean measured values and estimated ranges of parameters used in GW-Fluct

Parameter	Measured value	Estimated range	Model input value
KSAT1 (cm/day)	15·7	[4·2,27·2]	15·7
KSAT2 (cm/day)	0·7	[0·1,2·2]	0·7
P1	0·12	[0·09,0·25]	0·22
P2	0·11	[0·08,0·15]	0·11
ΔD*X* (m)	165	[150,180]	165
ΔD*Z* (m)	80	[70,90]	80
D1 (m)	1·65	[1·50,1·80]	1·65
D2 (m)	5·35	[5·15,5·55]	5·35
GW i1 (m)	–	–	–1·65
GW i2 (m)	–	[–4·30,–7·00]	–5·00
PCi (cm/day)	–	[0·1,10]	0·28
BPF1	0·20	[0·05,0·50]	0·20
BPF2	–	[0·01,0·15]	0·04
maxP (mm/day)	20 (winter mean)	[15,30]	20
Reff (mm)	– (Thornwaite)	[–20%,+20%]	–
ω (cm)	–160	[–180,–120]	–160
interception	–	–	0·25

The modelling of precipitation loss (evapotranspiration, interception and soil moisture storage) represents a delicate problem, because it determines the amount of water passing the root zone (R_{eff}), while on the other hand few data on this have been available until now. Therefore a very simple modelling approach has been adopted for the time being, potential evapotranspiration being modelled with Thornthwaite's (1948) model, soil moisture storage with Thornthwaite and Mather's (1957) model, and pine forest interception very roughly set on 25 per cent of total precipitation. Monthly temperatures were obtained from the Barcelonnette meteorological station. However, especially in the case of pine forest these models may not be very accurate. Effort is being made currently to obtain data and develop a more sophisticated approach to this.

The bedrock was assumed impermeable in relation to the other measured values.

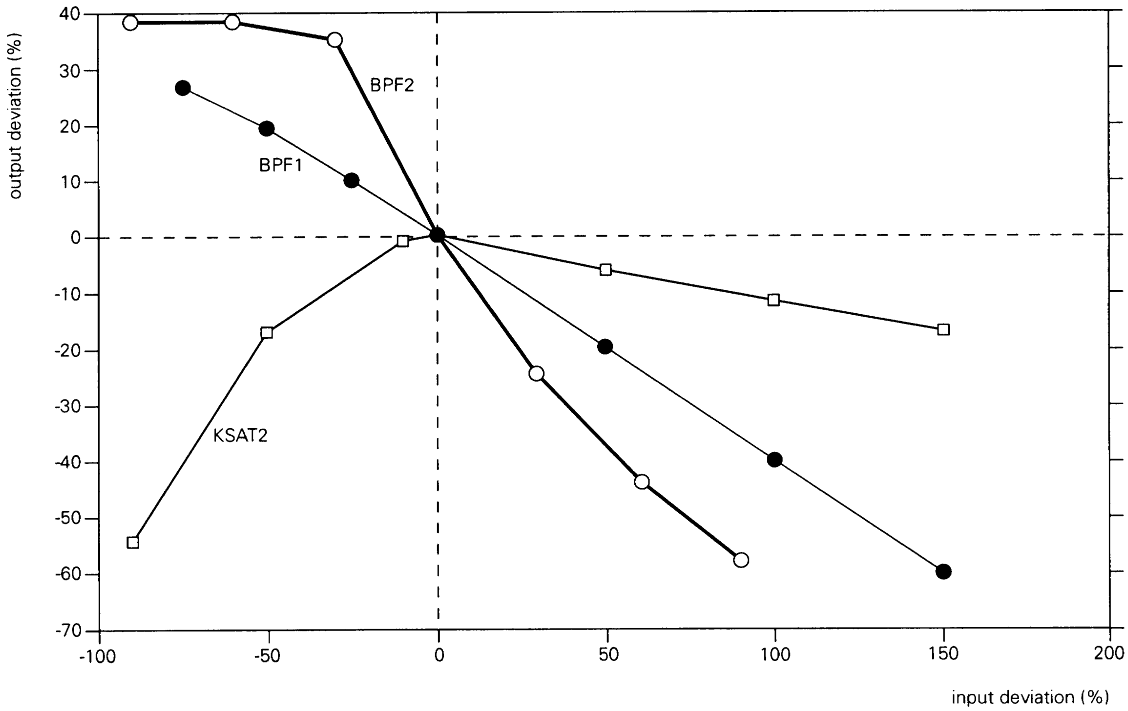
Sensitivity analysis

A sensitivity analysis was carried out to assess crucial parameters for the model. The model was run for a period of 25 years (1956–1980) using monthly precipitation data. For the sensitivity analysis the estimated range of the parametric values given in Table I was used. In Figure 3a and 3b this range is expressed on the *x*-axis as a percentage of input deviation from the mean value. On the *y*-axis, the relative difference in model output for mean and deviating parameter values is shown.

Figure 3a shows that bypass flows in both layers (*BPF1* and *BPF2*) are very sensitive parameters. These parameters are difficult to obtain from field measurements and were used as calibration parameters. The sensitivity analysis shows the importance of the installation of interception drains in the first layer: a 100 per cent increase in drainage effectivity in the first layer reduces the height of the groundwater peaks by 40 per cent. Figure 3b shows the high sensitivity of the layer 2 porosity and the effective precipitation (*Reff*). The latter indicates the importance of evapotranspiration and interception on groundwater fluctuations and hence the stability of the slopes.

It appears that initial groundwater in the two layers (*GWi1*, *Gwi2*), matric suction in layer 2 (ω) and initial percolation (*PCi*) are not sensitive. Layer 1 conductivity and porosity (*KSAT1* and *P1*) and the maximum amount of rain per day (maxP) appeared to be slightly sensitive: no deviation was observed when these input parameters were set higher (*maxP* lower) than the mean value, and a deviation of 10 per cent (deeper groundwater) was calculated when they varied from 0–50 per cent below (*maxP* above) the mean value. This can be explained by the development of a perched water table during more than 1 month at lower (*maxP* higher) input values. Because of this, lateral discharge in the more permeable layer 1 (*L*) is increased, and less water will enter layer 2. Layer 2 Ksat (*KSAT2*) seems to be very sensitive in two ways: at values above the mean there is a rapid decrease of the groundwater level because of an increased lateral discharge from the second layer; at

(a)



(b)

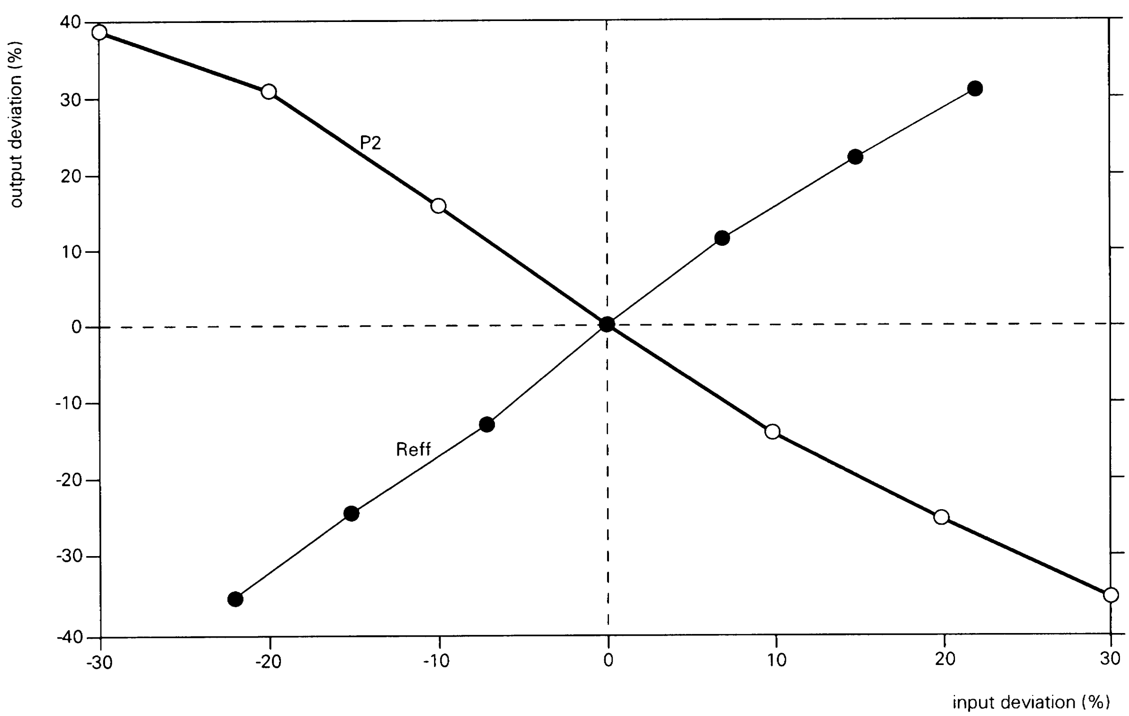


Figure 3. Output deviations for sensitive parameters of the model

values below the mean, groundwater levels decrease as well, because a perched groundwater table develops, as explained above.

Model calibration

Stability analysis of the landslide revealed that a probability of failure of 50 per cent or higher occurs when the groundwater level becomes higher than 4·2 m below the topographical surface (see above). Groundwater peak levels were calibrated in such a way that in a year of landslide activity, at least one groundwater peak exceeds −4·2 m. The years of activity were determined by dendrochronological dating. According to this research, the periods in which movements may have occurred are 1960–1962, 1969–1971 and 1973–1980, where 1960, 1962, 1970 and 1978 seem to be the years of major activity (Van Asch and Van Steijn, 1991). Also the model has to simulate total drainage of the second layer in prolonged dry periods because only a few decimetres of groundwater were observed in these periods (Caris and Van Asch, 1991).

The model was calibrated giving layer 1 bypass flow (*BPF1*) a value of 0·20 which was based on volumetric estimations of fissures and macropores in the field. In that case layer 2 bypass flow (*BPF2*) appeared to be 0·035 in order to meet the requirements which are defined above. No other absolute or rational values of (*BPF1* and *BPF2*) gave better results.

Modelling results

Figure 4 shows the graph of groundwater fluctuation, which is obtained after calibration of the drainage factors of the two layers. The graph shows yearly fluctuations from 1956 to 1980, with differences in maxima over the years. The height of these maxima seems to be controlled by the amount of yearly precipitation and even antecedent yearly precipitation as can be seen from the cumulative effective precipitation curve which is also given in Figure 4. The steeper parts in the cumulative curve are related to maximum peaks of the groundwater level. A comparison of the groundwater table fluctuation with the monthly precipitation shows that the highest groundwater peaks are generated mainly during the wet period by a number of consecutive months with high precipitation. Maximum peaks occur when the amount of rainfall is relatively high over a period of about six consecutive months during the wet season (period October to May). A series of consecutive years with rather high precipitation in the wet season may also lead to maximum levels. This can be seen, for example, in 1962 which did not have an excessively wet season. However, there were three antecedent years with high precipitation during the wet season. The highest peaks occur at the end of the observation period with the largest number of months with excessive rain (years 1977 and 1978).

The results of this modelling study show that the trigger system of the groundwater level is not controlled by single rain storms of high intensity and long duration or even by high weekly or monthly precipitation periods but by a long period of many months or even several years. This is explained by the hydrological system which could be observed in the field. The rather impermeable subsoil of clay material needs a permanent hanging groundwater table in the topsoil in order to let the groundwater level rise in the subsoil. In order to keep this relatively permeable top layer saturated, high precipitation figures are needed for a period of at least six consecutive months. It is not realistic to use shorter time intervals in the model in view of the low Ksat values, especially in the secondary layer, and the consequent long travelling time of the percolating water.

DISCUSSION AND CONCLUSIONS

The hydrological system of the landslide under study can be explained by a three-layer concept: a rather permeable top layer and more impermeable secondary layer, with both lying on top of the nearly impermeable *in situ* Terres Noires.

Water fluxes in these layers occur by two types of groundwater flow: matrix flow obeying Darcy's law, and a more rapid gravitational flow through preferential flow paths.

The residence time of a perched groundwater table in the first layer is crucial for the amount of water supply to the lower groundwater body, which develops on top of the *in situ* black marls by a vertical flux through rather impermeable clays.

Calibration of the model could not be done against measured data of groundwater levels. A tentative calibration was carried out using stability calculations, to define a critical groundwater level. The groundwater

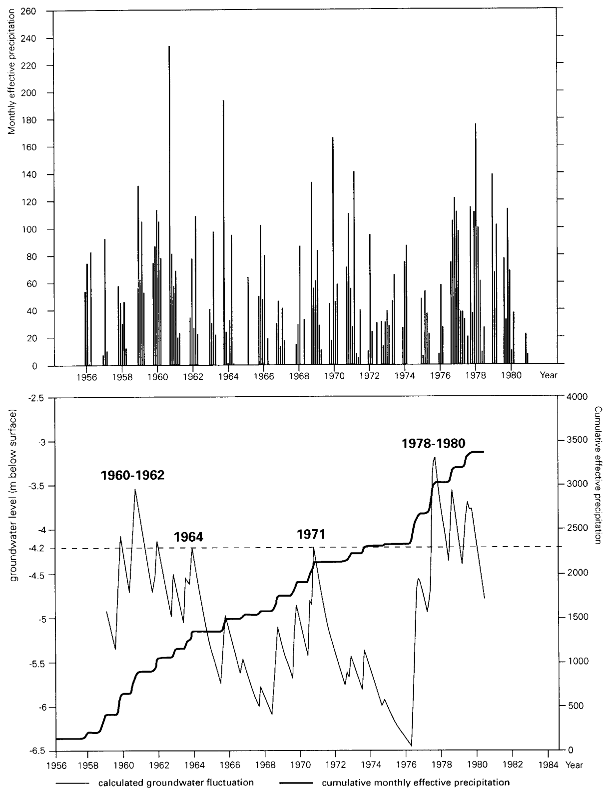


Figure 4. Monthly rainfall in relation to calculated groundwater fluctuations over a period of 25 years, and dated moving incidents of a landslide in Terres Noires colluvial material

model was calibrated in such a way that all the simulated groundwater peaks equal or exceed the critical level in the years where movements were indicated by dendrochronological research. Calibration was done by changing the very sensitive drainage factors of the two layers, assuming that all other parameters have been properly estimated by field measurements.

The calibration performed here, based on a critical groundwater level and dated movements on a yearly basis, does not give absolute guarantees with respect to all the parametric values and the performance of the model. However, some conclusions can already be made:

1. The model has improved our knowledge of the rainfall scenarios which might be expected as critical for instability. Short-term heavy rains are not effective because most of the water will be rapidly transported downslope through the upper layer. Even extreme precipitation during one month seems not to be critical for failure conditions. The modelling shows that maximum peak heights can be reached only in the wet season (colder seasons with lower evapotranspiration) with high precipitation during a period of at least six consecutive months.
2. Snow precipitation can be considered in this model as an equivalent of water precipitation. In general the amount of snow precipitation is more effective if it melts slowly over a longer period, contributing more effectively to the maintenance of a sufficiently high pore pressure groundwater head in the top layer. This is required to maintain the vertical flux to the subsoil. Rapid snow melt leads to rapid drainage and disappearance of water in the top layer. A simple snowmelt model, based on degree-day theory, could be combined with GW-Fluct to assess snowmelt impact on the groundwater level.

3. The model gives a better description of the hydrological system than a simple black box meteorological model, because it can forecast changes in stability conditions by external effects. For example, effects of changes in effective precipitation, which is a sensitive parameter, on the frequency of landslide movements can be forecast with this model in combination with a stability model. Also, the effect of an increase in lateral drainage in the top layer, by the installation of interception drains, can be assessed. The sensitivity analysis already shows that an artificially induced increase of the drainage system will drastically reduce the groundwater peaks.
4. The model is able to predict future instabilities on a yearly basis on the basis of monthly precipitation and hence may be used to predict the effect of climatic change.

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