Shallow Landslide Hazard Mapping Based on a Quasi-Dynamic Wetness Index

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Abstract: Shallow landslide hazard mapping based on steady-state wetness indices does not account for the relation between landslide occurrence and frequency of the triggering meteorological event. The quasi-dynamic wetness index (QDI) allows this limitation to be overcome while keeping the simplicity of the index approach. QDI estimates the local wetness conditions by accounting for the character of the upslope topography and the time of the lateral soil moisture distribution, assuming kinematic subsurface flow with constant recharge rate. Using this approach, different rainfall intensity and duration can be considered and a critical rainfall duration can be identified in an objective way. This study presents an effective method for calculating QDI in complex digital terrain models employing a multiple-flow-direction algorithm and suggests a procedure for using QDI in the analysis of the spatial occurrence of shallow landslides. The methods were employed to assess the role of the lateral subsurface flow in setting the triggering conditions during the hydrological disaster that occurred the 5th May 1998 in Campania, South Italy.

Keywords: Wetness index; Terrain analysis; Shallow landslide

1. INTRODUCTION

Hundreds of villages located at the toe of steep limestone ranges around the volcano Vesuvius, in the Campania Region, South Italy, are threatened by the risk of destructive debris-flows. These are originated by soil slips occurring in the shallow pyroclastic mantle covering the limestone bedrock of the steep slopes. Rainfall events of different characteristics and with different hydrological antecedent conditions have triggered several debris-flow episodes in the last 80 years.

On the 5th of May 1998 over one hundred mudflows occurred on Pizzo d'Alvano slopes, 40km east of Naples, causing deaths and damage in 5 villages. Preliminary investigations showed that lateral flow in the pyroclastic mantle and groundwater outlets both contributed to the triggering hydrological conditions [Cascini et al., 1998]. The relative importance of lateral flow and bedrock outflows in defining the triggering conditions depends on the intensity and the duration of the rainfall, as well as on the hydrologic conditions preceding the rainfall event [Chirico et al., 2000]. After the disaster, the Italian Civil Defence Department raised several issues with the scientific community [Cascini et al., 1998]. The first issue was the prediction of the temporal occurrence of landslides for a risk-warning system. This task was addressed by developing lumped statistical-conceptual models based on the analysis of the relation between rainfall and debris flow occurrence in the past [Chirico et al., 2000]. The second issue was the landslide hazard mapping. Some investigations were carried out to assess and develop computer-based methods, enabling the role of the topographically controlled lateral flow in landslide spatial occurrence to be defined.

Several methods employing digital terrain analysis have been proposed in the past. These methods couple a distributed model, assessing the spatial wetness patterns, with an infinite-slope stability model. The distributed model is generally structured on a network of elements defined by digital terrain analysis, either contour or grid based. The elements are characterised by specific attributes structured in a digital terrain model (DTM). The connectivity among the elements is defined by one-dimensional flow-path patterns.

Some terrain analysis methods adopt single-flow-direction algorithms, connecting each element with a single down-slope element, while other terrain analysis methods adopt more realistic multiple-flow-direction algorithms, connecting each element with one or more down-slope elements [e.g. Gallant and Wilson, 2000]. In the latter case the algorithms also define the proportion of outgoing flow allocated to each down-slope element.

Montgomery and Dietrich [1994] calculated the wetness patterns with a contour-based steady-state wetness index. Borga et al. [1998] extended this approach to grids with a multiple-flow-direction algorithm. Pack et al. [1998] included a parameter uncertainty analysis based on uniform probability distributions in order to assess the stability classes. Wu and Sidle [1995] applied a contour-based distributed event model. Borga et al. [1998] first suggested to apply the quasi-dynamic wetness index (QDI) [Barling et al., 1994] to overcome the limitations of the steady-state approach. Duan and Grant [2000] evaluated the capability of the QDI in predicting landslide location with a Monte Carlo simulation scheme, including rainfall intensity among the random variables. The QDI was calculated on a grid with a single-flow direction algorithm. Steady-state methods do not account for the relation between rainfall intensity and duration, and upslope topographic attributes in predicting the spatial location of the landslides. On the other hand, the distributed dynamic models, although able to account for complex rainfall patterns, are computationally complex and lose the practical advantages of an index-based approach. The QDI accounts for duration and intensity of the rainfall without losing the simplicity of the index approach.

This study presents a practical method for the computation of the quasi-dynamic wetness index on computational networks with multiple-flow direction algorithms and suggests a new approach for landslide spatial analysis employing the QDI. The QDI was used in a preliminary application, screening the spatial location of the landslides that occurred in May 1998 on Pizzo d'Alvano slopes.

2. AN ALTERNATIVE METHOD FOR COMPUTING QDI IN COMPLEX DTMs

The QDI computes the local wetness condition in the hypothesis of linear-kinematic subsurface flow with constant recharge rate (r). The one-dimensional form of the linear-kinematic wave equation along the generic flow-tube s of unit width is [Beven, 1981]:

$$\begin{aligned} &\eta \frac{\partial h}{\partial t} + \frac{\partial q}{\partial s} = r \\ &q = Khsin\beta \end{aligned} \qquad (1)$$

where η is the local storage capacity (porosity minus field capacity), that Barling et al. [1994] assumed equal to the effective porosity; K is the local lateral conductivity; β the local slope; h the local saturated depth defining the wetness condition; q is the flow per unit width; r is the recharge rate. The wetness index (w) can be expressed in a dimensionless form, as the ratio of the saturated depth (h) to the soil depth vertical to the slope (D). The integral solution of equation (1) for a generic point of the catchment is equivalent to the convolution of the recharge rate with the kinematic unit hydrograph of the upslope contributing area [Barling et al., 1994]. Thus, for constant recharge rate, the wetness index can be expressed as follows:

$$w(r,t) = \frac{h(r,t)}{D} = \frac{q(r,t)}{KDsin\beta} = \frac{r \cdot a(t)}{KDsin\beta} \qquad (2)$$

where $a(t)$ is the specific dynamic contributing area, i.e. the portion of the upslope contributing area per unit flow width, including all the points located at a distance that can be covered by the kinematic flow in a time less or equal to t. Given an element network, $a(t)$ can be computed for each element by accounting geometrically for all the possible upstream flow-paths [Wilson and Gallant, 2000]. The procedure consists of accumulating the areas of the elements along all the possible upstream flow-paths that can be covered by the kinematic flow in a time less or equal than t. The flow travel-time along each element is calculated on the basis of the local flow celerity (c):

$$c = \frac{Ksin\beta}{\eta} \qquad (3)$$

The most upstream element of each flow-path can partially contribute to the downstream flow. The element contributing-portion is estimated by assuming a relation between length of the flow-path within the element and corresponding element partial area. This relation is linear if trapezoidal elements are assumed. Equations (2) and (3) differ from those reported in Barling et al. [1994] in having the sine and not the tangent of the slope. This detail is significant only for very steep slopes. The geometric procedure used for the analysis of single flow-paths is computationally impracticable when the more efficient multiple-flow-direction algorithms are employed [Wilson and Gallant, 2000].

An alternative approach is to integrate numerically the kinematic wave equation on the element network. Moore and Foster [1990] suggested the following implicit four-point approximation as being the most stable finite difference scheme, with least numerical dispersion:

$$Q_4\left(\frac{2\Delta t}{\Delta S}+\frac{1}{c}\right)=\left(Q_1+Q_3\right)\frac{1}{c}+Q_2\left(\frac{2\Delta t}{\Delta S}-\frac{1}{c}\right)+A_s\frac{2r\Delta t}{\Delta S} \qquad (4)$$

where A_s is the plan area of the element; ΔS is the element length along the slope; Q is the cross-sectional hydraulic area and the discharge at the computational nodes, represented by the extremes of each element assumed as trapezoidal. Subscripts 1 and 3 refer to the time t at the positions s (upstream side of the element) and $s+\Delta S$ (downstream side of the element); subscripts 2 and 4 refer to the time $t+\Delta t$ at positions s and $s+\Delta S$. Equation (4) must be solved first for the most upstream elements with computations then proceeding downstream. Q_2 is calculated by adding the contributing discharge of the upstream elements. The only difficulty in this scheme and solution method is when the flow depth at a node on the space-time grid changes from a zero to a non-zero value and Δt is small. To overcome this problem, an explicit solution is used when this occurs.

Setting the initial and boundary conditions to zero, and r equal to the unit rate, at the generic computational time t, Q_4 divided by the width of the element represents the specific dynamic contributing area $a(t)$. There is no strong limitation in choosing a computational time-step small enough to limit the computational errors. In fact, in order to compute the whole range of possible $a(t)$ patterns, the numerical scheme has to be run only once for t up to the time of concentration of the study area, i.e. the longest flow-path travel-time. Then the actual wetness index $w(r,t)$ for a given constant recharge rate r can be computed by applying equation (2).

The numerical scheme was tested by comparing the results with those obtained by accounting geometrically for the flow-path length on a grid with single-flow-direction algorithm. Figure 1 shows a sample case. The dynamic contributing area was calculated with the two methods at the outlet of a 0.27km^2 catchment, represented with a 10m-grid DEM, assuming an arbitrary lateral conductivity. This is a representative test case, as the catchment topography is very complex, including large valleys (slope ~0.1%) and steep slope (up to 30%), with a second-order channel network. The largest errors (~11%) occur for small values of the contributing area. These are mainly due to the fact that the numerical scheme, unlike the geometrical approach, does not account for partially contributing elements. For larger values of the contributing area the errors (<6%) are mainly numerical, in particular larger errors occur for rapid change in slope at the head of the channels. The tests can also be considered representative for multiple-flow-direction algorithms. The numerical scheme presented above is applied at the extremes of each element and it is not affected by the type of flow-direction algorithm.

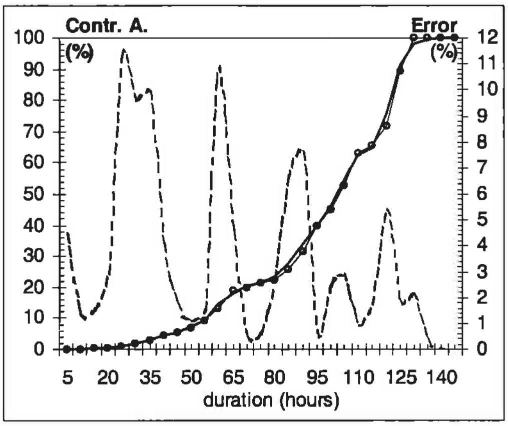

Figure 1. Dynamic contributing area computed with the geometrical (circles) and the numerical method. The broken line is the relative error.

3. LANDSLIDE HAZARD MAPPING BASED ON QDI

Different infinite-slope stability models have been applied to landslide hazard mapping based on digital terrain analysis. These are expressed in terms of a safety factor (*FS*) given by the ratio between the resistance of the soil to failure (shear strength) to the forces promoting failure. *FS* is a monotonically decreasing function of the wetness index (*w*). The QDI represents an estimate of the local wetness conditions induced by rectangular rainfall-input in the following hypothesis. First, the local celerity is assumed independent from the wetness conditions. This means in particular that no saturation should occur in any of the upstream elements, since the actual lateral celerity rapidly changes as the element is saturated. Furthermore, it is assumed that there is no limiting infiltrationcapacity, no delay induced by the vertical distribution of the soil moisture or due to wetting the profile to the point where significant lateral flow begins and no vertical losses. While the steady-state wetness index accounts only for the integral value of the upslope contributing area, the quasi-dynamic wetness index accounts for the character of the upslope contributing area and the time of the lateral distribution of soil moisture [Barling et al., 1994]. Thus it allows the estimation of spatial wetness maps for rainfall inputs of different duration and intensity. Similarly in many hydrological applications, rectangular rainfall inputs, with duration and intensity varying according to a specific intensity-duration curve ($I(d)$), can be adopted as a set of design hyetographs to assess the spatial occurrence of the landslides. For each elemental area, a critical duration (d_c) of the rainfall can be defined as the duration corresponding to the maximum wetness index. Given equation (2), d_c can be formally expressed as:

$$q(d_c) = \max_d \{I(d)a(d)\} \tag{5}$$

$a(d)$ increases with the duration d, while $I(d)$ generally decreases with d. For a given rainfall return period (T_r), rectangular rainfall can be varied according to the local intensity-duration-frequency (IDF) curve. It is widely used to represent IDF curves scaled by an index curve $\hat{I}(d)$ and isolating the probabilistic component in a single factor ($\Lambda(T_r)$) [e.g. Rossi and Villani, 1994], as follows:

$$I(d,T_r)=\Lambda(T_r)\hat{I}(d) \tag{6}$$

The critical duration of the rainfall for each element can be identified employing equation (5) with $I(d)=\hat{I}(d)$ and the maximum wetness index value (w_m) for a given T_r can be defined as:

$$w_m = \frac{\Lambda(T_r)\hat{I}(d_c)a(d_c)}{KDsin\beta} \tag{7}$$

One of the big issues in the landslide hazard mapping is the uncertainty in the soil-vegetation parameters affecting the slope stability. Monte Carlo simulations have been used in order to account for this uncertainty [e.g. Duan and Grant, 2000]. The expression (7) can be used in a Monte Carlo simulation scheme sampling $\Lambda(T_r)$ and all the other uncertain parameters according to their corresponding probability distributions. The critical duration d_c has to be computed only for the range of values assumed for the lateral celerity.

4. CASE STUDY

160 mm of rain occurred between the afternoon of the 4[th] and midnight on the 5[th] of May 1998 in the area of Pizzo d'Alvano range. Over a hundred landslides were triggered between the early afternoon of the 5[th] and the end of the rainfall. The maximum rainfall intensity was 15mm/h. The intensity-duration curve of the event shows intensities around 10mm/h for durations ranging from 2 to 4 hours, and between 7mm/h to 4mm/h for durations from 6 hours to 32 hours. The 4[th] of May was preceded by a wet period, characterised by ~60mm of rainfall unevenly distributed in 7 consecutive rainy days. Landslides occurred within the zero-order basins, on extremely steep slopes (30^o-40^o). These are covered by unconsolidated, extremely porous pyroclastic soils, a few metres thick and composed of pumice and ash, alternating with paleosoil horizons. Given the rainfall and soil characteristics, it can be assumed that lateral distribution of soil moisture was controlled by sub-surface lateral flow, with no infiltration-excess runoff.

A two-dimensional geotechnical model applied to a plain slope showed that lateral flow alone could not trigger the landslides. Therefore local outflows from the bedrock were thought to be the primary trigger for landslides [Cascini et al., 1998]. However, doubts remained concerning the effect of the topographic convergence on lateral flow distributions, as Cascini et al.'s study [1998] was based on a plain slope. Therefore we tested the topographic influence using two different approaches. One was fully dynamic and consisted of an event analysis employing the numerical scheme of equation (4) with r variable in time according to the observed hyetograph. The maximum wetness condition reached in each element during the event were stored and used for the stability analysis. The other approach consisted of the procedure presented in the previous section. The stability analysis was conducted with respect to the critical quasi-dynamic wetness index calculated with the intensity-duration curve of the event. The opportunity to employ a contour-based element network was first tested. Figure 2 shows the contour-based element network of the west-south side of Pizzo d'Alvano range, with contours spaced 5m in elevation. This was generated with CUTLASS [Maunder, 1999]. Unlike earlier methods, which are designed to produce element networks for single catchments, CUTLASS is able to develop element networks on arbitrary pieces of topography, so it is more suited to studies over large areas that include many sub-catchments. The contour-based element network, however, did not represent the convergent valley bottoms in adequate detail. Higher topographic resolution was needed and this could not be efficiently reached by improving the contour-DEM resolution. Thus a 5m-grid DEM was employed, with flow-path pattern defined using the $D\infty$ algorithm [Tarboton, 1997]. $D\infty$ was replaced with a single-flow-direction algorithm for contributing areas greater than 1ha in order to reduce the flow dispersion in the valley [Gallant and Wilson, 2000]. The threshold value was chosen after comparing the contributing area maps resulting from different threshold values against the actual channel network. Figure 3 shows the corresponding steady-state contributing area map.

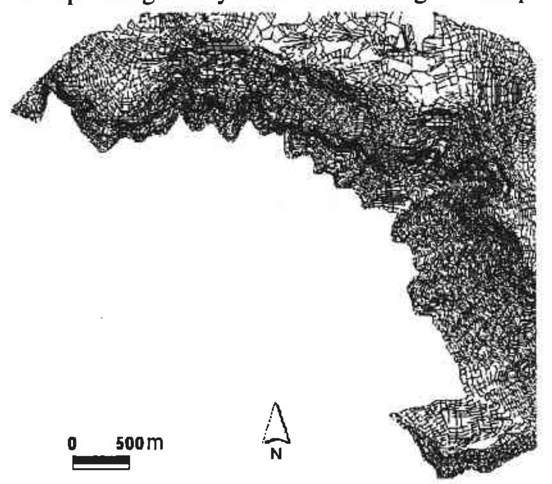

Figure 2. South-west side of Pizzo d'Alvano range. Contour-based element network generated with CUTLASS. Contours are spaced 5m in elevation.

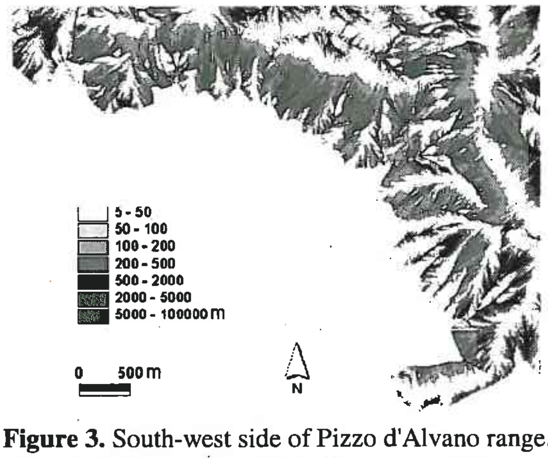

Figure 3. South-west side of Pizzo d'Alvano range. Steady-state specific contributing computed on a 5m-grid DEM using the $D\infty$ algorithm.

While a map of the depth of soil mantle was available, no specific surveys were conducted in order to assess the spatial variability of the parameters affecting the stability and the lateral flow. Given this uncertainty, a simplified stability model assuming equal dry and wet soil density was used [Pack et al., 1998]. *FS* is expressed as:

$$FS(w) = \frac{C}{D \sin\beta} + \frac{\tan\phi}{\tan\beta}\left(1 - w\rho\right) \tag{8}$$

where C [L] is the soil cohesion relative to the soil density; ρ is the water to soil density ratio; ϕ is the internal friction angle. The parameters were chosen equal to the most critical values for the slope stability, within the range of expected variability [Cascini et al., 1998]: $r=0.7$; $C=0.33$m; $\phi=32°$; $\eta=0.3$v/v; $K=1000$mm/h. The critical duration of the rainfall is always equal to the storm duration. This is due to the limited variation of the rainfall intensity with duration during the event. Larger increases in rainfall intensity for decreasing duration are needed to overcome the corresponding reduction in contributing area.

The two approaches produced results with no significant differences, confirming the validity of assumption in using rectangular rainfall as design hyetographs. Figure 4 shows the distribution of *FS* computed for all the elements with slope above ~15°, this being the threshold for points that are unconditionally stable (even for $w=1$). Also excluded are the elements with slopes above ~40°, corresponding to bedrock outcrops which are unconditionally unstable (even for $w=0$). The same figure shows the distribution of the predicted *FS* where the landslides occurred.

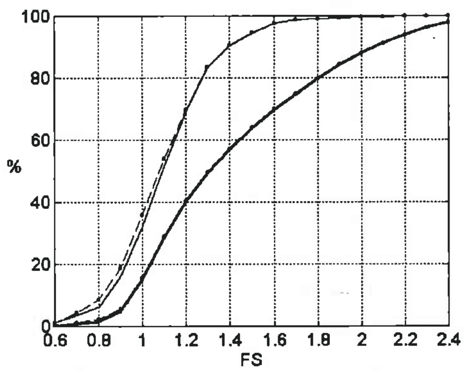

Figure 4. *FS* distributions for the whole study area (thick line) and the landslide locations (thin line). No significant differences resulted using the fully dynamic approach (broken line with dots).

Only 35% of the observed landslides occur in the area predicted to be unstable, corresponding to 15% of the total area. These results show that lateral flow by itself has limited capability in explaining the spatial occurrence of the landslides, confirming the results for a plain slope [Cascini et al., 1998].

5. CONCLUSIONS

The quasi-dynamic wetness index (QDI) is appealing for investigating the relation between the spatial occurrence of landslides and characteristics of the triggering meteorological event, as it estimates the local wetness conditions by accounting for the character of the upslope topography and the time of the lateral soil moisture distribution. This study showed that the QDI can be computed either geometrically or numerically, integrating the linear-kinematic wave equation. The numerical approach allows calculation of QDI in digital terrain models employing multiple-flow-direction algorithms. A procedure coupling the QDI and intensity-duration-frequency curves (IDF) was suggested in order to assess landslide hazard maps for a given frequency of rainfall occurrence. This procedure is based on the idea of using rectangular hyetographs with intensity and duration varying according to the local IDF curve as set of design hyetographs. This study presented an application for an event where more than one hundred landslides occurred. While the method using IDF curves was shown to be computationally tractable and equivalent to using the actual hyetograph, both failed to predict the location of the observed landslides, as processes other than lateral subsurface flow were dominant. More investigations are needed in order to test the capability of the procedure for landslide hazard mapping.

6. ACKNOWLEDGEMENTS

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