

SOME ASPECTS OF LANDSLIDES RISK EVALUATION TAKING INTO ACCOUNT THEIR DISTRIBUTION AND PROPERTIES

Ancuța Rotaru¹, Ana Nicuță²

¹ Faculty of Civil Engineering, “Gheorghe Asachi” Technical University of Iasi, 43 Bv.Dimitre Mangeron, 700050, Romania. E-mail: arotaru@ce.tuiasi.ro

² Faculty of Civil Engineering, “Gheorghe Asachi” Technical University of Iasi, 43 Bv.Dimitre Mangeron, 700050, Romania

Abstract

Landslide events display large variations in landslide types, size, distributions, patterns and triggering mechanisms. Landslides are triggered by many different causes, including intense or prolonged rainfall, snowmelt and earthquakes, and they can occur individually or in groups of thousands simultaneously. In nature, landslide areas have been documented to span more than eight orders of magnitude and landslide volumes more than ten. Landslide velocities extend over fourteen orders of magnitude, from millimeters per year to hundreds of kilometers per hours. Landslide risk evaluation must take into account variable factors that affects the soil and also must admit the importance of landslide distribution.

Key words: *risk evaluation, fresh and historical landslides, frequency–area distribution, landslide magnitude sale, one-meter depth temperature, weak zones of a saturated clay landslide*

1. INTRODUCTION

Landslides or slope movements are complex natural phenomena that constitute a serious natural hazard in many countries (Brabb and Harrod, 1989) and play a major role in the evolution of landforms. These phenomena are influenced by many factors, such as: soil or rock types, bedding planes, topography, and moisture content.

Landslides cause damages, injuries and death and affect a great variety of resources. For example, water availability, quantity, quality and supplies, forests, dams and roadways can be affected for years after a slide event.

The negative economic effects of landslides include the cost to repair structures, loss of property value, disruption of transportation routes, medical costs in the event of injury, and indirect costs. Geotechnical studies and projects to assess and stabilize potentially dangerous sites are costly.

Landslide risk can be reduced with engineering and geo-science investigations.

By geological mapping, detecting slope hazards and determining the likelihood of landslide occurrence, geoscientists can assist engineers, developers, planners and building inspectors in avoiding high-risk areas. Through this process, structures such as homes, schools, hospitals, power-lines, fire stations and roads can be safely located away from potential landslide risk areas.

2. CAUSES OF LANDSLIDES

Some slopes are susceptible to landslides whereas others are more stable. Many factors contribute to the instability of slopes, but the most important factors are the nature of the underlying bedrock and soil, the configuration of the slope, the geometry of the slope, and ground-water conditions.

Three distinct physical events occur during a landslide: the initial slope failure, the subsequent transport, and the final deposition of the slide materials.

Landslides can be triggered by gradual processes such as weathering, or by external mechanisms including:

- Undercutting of a slope by stream erosion, wave action, glaciers, or human activity,
- Intense or prolonged rainfall, rapid snowmelt, or sharp fluctuations in ground-water levels,
- Shocks or vibrations caused by earthquakes or construction activity,
- Loading on upper slopes, or
- A combination of these and other factors.

Geologists and geotechnical engineers use a variety of classification schemes to describe causes of landslides. Because of wide variety of causes, more than a single scheme has been developed that address or describe all types of landslides (source: The Royal Academy of Engineering, 1995, Landslides Hazard Mitigation, Westminster, London):

External:	Geometrical change:	Gradient, Height, Slope length
	Unloading:	Natural, Human-induced
	Loading:	Natural, Human-induced
	Shocks and Vibrations:	Single, Multiple/continuous
Internal:	Progressive failure:	Expansion, Fissuring, Strain softening, Stress concentration
	Weathering:	Physical property changes, Swelling, Chemical changes
	Seepage Erosion:	Removal of cements, Removal of fines
	Water Regime Change:	Saturation, Rise in water table, Excess pressures, Drawdown

3. LANDSLIDE TYPES

Geological structure determines types and subtypes of landslides and the intensity of the landslide process. For each slope exist individual peculiarities in the structure, but the general features of big groups of slopes are considered, in order to determine the characteristics of the landslide process. The following peculiarities of geological structure of slopes are most important for landslide development: form and size of the bodies and conditions of their location with regard to the slope; physical and mechanical characteristics of the rocks comprising the slope; form of contact between the rocks, contact orientation with regard to the slope, presence of fissures and other surfaces of weakening.

Landslides can be classified in many ways using a lot of attributes as criteria, such as:

- Rate of movement: This ranges from very slow creep (millimeters/year) to extremely rapid (meters/second).
- Type of material: Landslides are composed of bedrock, unconsolidated sediment and/or organic debris.
- Nature of movement: The moving debris can slide, slump, flow or fall.
- Types of weakness planes and associated landslides: (1) slope failure in glacial sediment resulting in slumps; (2) parallel bedding in rock causing slides; and (3) fracturing of rock promoting falls.

Poorly planned forest clearing may increase rates of surface water run-off or ground-water infiltration. Inefficient irrigation may result in increased ground-water pressures, which in turn can reduce the stability of rock and sediment.

Landslides may result directly or indirectly from the activities of people. Slope failures can be triggered by construction activity. Many people, unaware of their exposure to landslide risks, modify the landscape, building on unstable slopes, or redirect the flow of surface or ground-water.

4. LANDSLIDE RISK EVALUATION

Considering the hazard and the probability of landslide occurrence as coincident, doesn't seem correct, since the term "hazard" corresponds to the real danger situation in a site for the advancing phenomenon (Canuti & Casagli, 1996; Crescenti, 1998, 1999). Instead, the probability doesn't show the real danger state, because not all the landslide phenomena are dangerous for the elements at risk. The coincidence hazard-probability is not appropriate because the term "hazard" is connected to the real danger, which is related not only to the occurrence probability of a landslide phenomenon, but also to its intensity. It is therefore more correct to define hazard on the basis both of probability of occurrence, and of the intensity of the landslide phenomenon, agreeing with Fell, 1994. With this definition of hazard, the term specific risk disappears and corresponds to the new definition of vulnerability. The new definitions of the elements of the risk are (those that differ from the UNESCO terminology appearing in *italics*):

- Intensity (I): severity of the landslide phenomenon. It can be assessed in relation to dimensions, velocity and energy of the phenomenon. The assessment of the relative intensity of landslide events is of interest for understanding the geomorphological evolution of a landscape shaped by mass-wasting process, and for landslide hazard-assessment studies.
- Probability (P): probability of occurrence of the landslide phenomenon. In its assessment, in the case of active landslides, the return time must also be taken into account.
- *Hazard (H): the degree of expected hazard for the elements at risk following the occurrence of a landslide phenomenon. It is a function of intensity and probability.*
- Elements at Risk (E): they represent by the population, property and facilities exposed to landslide risk.
- *Vulnerability [$V = F(H, C)$]: degree of loss (damage) produced or expected to elements at risk assessed according to their characteristics (C) for the occurrence of a landslide phenomenon of a given level of hazard.*
- Risk [$R = f(V, W)$]: value of expected damage to elements at risk from the occurrence of a landslide event of a certain level of hazard. It is expressed in terms of annual cost and/or number of units (people) lost per year.

Intensity and probability are the parameters of hazard; hazard and the value of the elements at risk are the parameters of risk.

The probability of occurrence (P) of landslide phenomena is deduced from the stability map, which divide the areas in: stable areas ($P = 0$), unstable areas ($P = 1$) and medium stable areas ($0 < P < 1$).

It follows that the stability map becomes the probability map. For the unstable areas, especially for the active landslides, data regarding the return time of the landslides can be inserted in the probability map.

The vulnerability map is obtained showing on the hazard map the damage produced or predicted to the elements at risk following landslide phenomena. This is achieved by keeping human vulnerability distinct from that of property and facilities, thus obtaining two distinct vulnerability maps.

The quantification of the predicted damage (the value of goods and services subject to damage and the predicted number of human losses), makes it possible to transform the vulnerability map into a risk map.

Since it is difficult to assess the value of damage and human losses in an absolute sense, the risk map can be drawn up with "relative" criteria, in the sense that the areas are divided in relation to their different degrees of risk (the value of the elements at risk subject to damage).

This methodology makes it possible to establish, on a geological-technical basis, the programmes of management and mitigation of risk.

The value of the elements at risk can be assessed both on the basis of their economic and social value, and on the basis of elements of judgment varying from case to case.

5. "FRESH" AND HISTORICAL LANDSLIDES

Landslide events are generally divided into two classes: (1) landslide-events that are associated with a trigger; and (2) historical (geomorphological) landslides, which are the sum of one or many landslide events over time in a region.

Immediately after a landslide event, individual landslides are "fresh" and clearly recognizable. The boundaries between the failure areas (depletion, transport and depositional areas) and the unaffected zones are usually distinct, making easy to identify the landslide. This is particularly true for small, shallow landslides, such as soil slips or debris flows.

For large, complex slope movements, the boundary between the stable area and the failed mass is transitional, particularly at the toe. The limit may also be transitional along the sides. For large deep-seated landslides, identifying the exact limit of the failed mass may not be easy even for fresh failures, particularly in urban or forest areas. Areas affected by the largest and most catastrophic landslide events are investigated exhaustively, because investigators concentrate in the areas where landslides were more abundant or where the damage was most severe.

"Fresh" landslides can be complete plotted if detailed mapping has been carried out shortly after the landslide event.

Landslide boundaries become indistinct with the age of the landslide. This is caused by various factors, including local adjustments of the landslide to the new morphological setting, new landslides, and erosion.

Historical landslide inventories are the total sum of single to many landslide events that occurred over time (tens, hundreds or even thousands of years) in a region. In many landscapes, landslides tend to occur where they have occurred in the past. A characteristic of historical landslides is that evidence of the existence of many of the smaller landslides has been lost due to various degrees of modification by subsequent landslides, erosional processes, human activities, and vegetation growth. All evidence of smaller landslides is likely to be lost, and the landslide boundaries become indistinct, making the landslides themselves much harder to identify.

A formal definition requires that a landslide inventory includes all landslides associated with a landslide event (a single trigger) or multiple landslide events over time (historical). This definition assumes that all landslides are visible and recognizable, and that the entire study area affected by the trigger(s), even marginally, is investigated. But, in practice, this criterion is never met.

A functional definition requires that the landslide inventory includes a substantial fraction of all landslides at all scales. An important attribute of this definition is that a complete inventory must include a substantial fraction of the smallest landslides. This definition may be applicable to landslide-event inventories, but not to historical inventories, because in this case many smaller-size and intermediate-size landslides have been erased by erosion and human actions.

6. THE DISTRIBUTION OF LANDSLIDES

6.1. The frequency–area (or volume) distribution of a landslide

Landslides are generally associated with a trigger, such as an earthquake, a rapid snowmelt or a large storm. The landslide event can include a single landslide or many thousands. Slope failures generally occur within minutes after an earthquake trigger, hours to days after a snowmelt trigger, and days to weeks after an intense rainfall trigger.

Landslides exhibit power-law frequency-area statistics. A simple model has been proposed to explain the behaviour of a landslide. The model is the “sand pile” model. Noever (1993) and other authors have associated the power-law frequency–area statistics of large landslides with the sand-pile model. In this model there is a square grid of boxes and at each time step a particle is dropped into a randomly selected box. When a box accumulates four particles, they are redistributed to the four adjacent boxes, or in the case of edge boxes they are lost from the grid. Redistributions can lead to further instabilities and avalanches of particles in which many particles may be lost from the edges of the grid. Each of the multiple redistributions during a time step contributes to the size of the model “avalanche”.

Having specified the size of the grid, the number of avalanches N_L with area A_L is determined. The area A_L is defined to be the number of boxes that participate in an avalanche.

There is a large extrapolation from the “sand-pile” model to actual landslides, especially for the medium and largest landslides. Considering the simplicity of the model and the three-dimensional nature of actual landslides, a revision of the rules for the “sand pile” model are needed for a realistic landslides nature. Nevertheless, the relatively simple “sand-pile” model may provide the basis for understanding the power-law statistics of large landslides. In the “sand-pile” model, the region over which an avalanche will spread is well defined prior to the avalanche. Similarly, the area over which a landslide will spread can be defined before landslide is triggered. In both, there are metastable areas. As particles are added, the metastable avalanche areas grow. As mountains grow, metastable landslide areas also grow. This growth is dominated by the coalescence of smaller metastable regions (Turcotte *et al.*, 1999; Gabrielov *et al.*, 1999; Turcotte, 1999). In addition, the coalescence cross-sections lead directly to a power-law distribution of metastable areas. It is expected that a similar coalescence and growth process would be applicable to metastable landslide areas and can explain the observed power-law frequency–area distribution of large landslides.

Other explanations for the power-law dependence have been given by Pelletier *et al.* (1997) which combined a slope stability analysis with a soil-moisture analysis and obtained a power-law distribution, and by Hergarten and Neugebauer (1998, 2000) which proposed two alternative models that give power-law distributions.

In many historical and fresh-event landslides, the frequency–area distribution of medium and large landslides decays as an inverse power of the landslide area (Whitehouse and Griffiths, 1983; Ohmori and Hirano, 1988; Sasaki, 1991; Noever, 1993; Sugai *et al.*, 1994; Hovius *et al.*, 1997).

The frequency–area (or volume) distribution of a landslide event quantifies the number of landslides that occur at different sizes. The mean area of landslides in the distribution is independent of the size of the event.

The quantification of landslides is useful in estimating the role of landslides in erosional processes and in assessing landslide hazards. The landslide-event hazard can be quantified only in function of a general landslide distribution. Based on this distribution a landslide-event magnitude scale was introduced to quantify landslide events.

An implication of having a general landslide probability distribution is the ability to create a landslide event magnitude scale. Measures of event sizes are useful for natural hazards. For example, the Richter scale for earthquakes is universally used. Other magnitude scales are available for other natural hazards, including the Saffir-Simpson scale for hurricanes, the Fujita scale for tornadoes, and

the Volcanic Explosivity Index. Malamud et.al. (2004) proposed a magnitude scale m_L for a landslide event based on the logarithm to the base 10 of the total number of landslides associated with the event:

$$m_L = \log(N_{LT}),$$

where: N_{LT} represents the total number of landslides associated with a trigger.

When examining landslides associated with individual landslide events, we are interested in knowing A_{LT} , the total area of all landslides that occurred as part of the event. Using the mean landslide area A_L , and the number of landslides that occur during the event, we calculate a predicted total area A_{LT} of landslides in the inventory.

The total measured area of landslides associated with a landslide event A_{LT} can also be used to determine the landslide magnitude m_L .

Using the distribution, we can extrapolate the largest landslide areas associated with a landslide event, to give a rapid estimate of the landslide-event magnitude, along with the total number and area of landslides in the event.

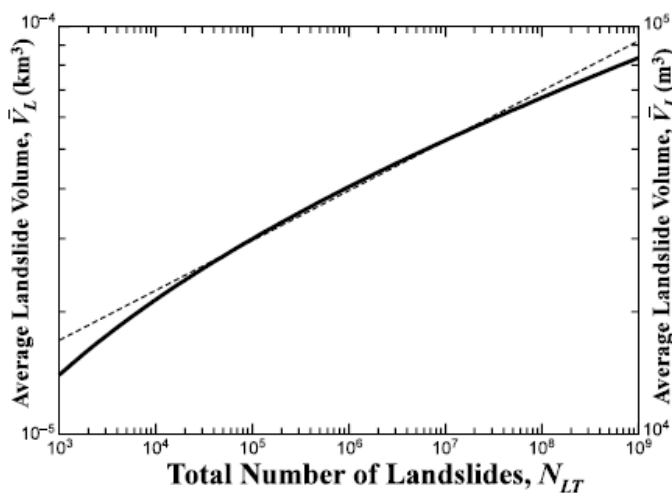


Fig.1. Dependence of the mean landslide volume on the total number of the landslide in the event (after Malamud *et al.*, 2004)

The frequency-area statistics of landslides can be presented using either cumulative or non-cumulative statistics (Malamud and Turcotte, 2000). The cumulative number of landslides N_{CL} with areas greater than or equal to landslide area A_L is plotted as a function of A_L .

$$N_{CL} = C(A_L)^{-\alpha}$$

where α is the power-law exponent and C a constant.

6.2. The general landslide distribution

We will focus on the general probability distribution of landslide areas for triggered landslide events, and the several implications of having a general distribution, including, first, a magnitude scale for landslides events, and second, the incomplete inventories of landslide-events in the case of historical landslides.

The direct determination of the landslide-event magnitude requires that the landslide inventory be substantially complete. However, the general landslide distribution can be used to determine a landslide-event magnitude from a partial inventory, where the inventory is complete only for landslide areas greater than a specified value.

It can also be used for historical inventories, which include the sum of landslide events over time. Using the general landslide distribution, the total area and volume of associated landslides in the event or sum of events over time, as well as the area and volume of the maximum landslides, can be directly related to the landslide-event magnitude m_L . These can be used to estimate regional erosion due to landslides.

6.2.1. Landslide magnitude scale

Keefer (1984) and Rodriguez *et al.* (1999) used a scale similar to the scale of Malamud *et al.* (2004) to quantify the number of landslides in earthquake-triggered landslide events: 100-1,000 landslides were classified as two, 1,000-10,000 landslides three etc. Some important landslide event magnitudes considered by Malamud *et al.* (2004) were: Northridge earthquake triggered event, $m_L = 4.05$; Umbria snow-melt triggered event, $m_L = 3.63$; Guatemala rainfall triggered event, $m_L = 3.98$.

While observed earthquakes span a wide range on the Richter scale, available landslide-event inventories are often restricted to a relatively narrow magnitude range. The reasons for this are: accurate inventories are restricted to populated areas and they haven been carried out only during the last years. Thus, in regions where studies have been carried out have occurred only few large landslide events with $m_L = > 4.0$.

Given several hundred landslide events and their magnitudes in a given region and time period, we hypothesize that there will be many smaller magnitude events compared to the larger ones. In analogy to earthquakes, we further hypothesize that these will follow a relationship such

$$\log N_C = -bm_L + a$$

where N_C is the number of landslide events with magnitudes greater than or equal to m_L , and b and a are constants.

6.2.2. Historical and incomplete inventories

Assuming that our landslide probability distribution is applicable to all landslide events, the sum of events over time (the historical inventory) will also satisfy this distribution (Malamud *et al.* 2004). However, in historical inventories, the evidence for the existence of many smaller and medium landslides will have been lost due to wasting processes over time. Therefore, for historical inventories, we attribute the deviation from our general landslide distribution. We can extrapolate an inventory, which contains just the largest landslides to give the total number and total volume of all landslides in the region.

The same method can be used for a single incomplete landslide event inventory where only the largest landslides have been measured and the medium and smaller landslides sizes are not know. Using the general landslide distribution we can extrapolate the frequency densities of the largest landslides in the event, and estimate of the equivalent landslide event magnitude.

Malamud *et al.* (2004) related the landslide-event magnitude for individual events to the total area and volume of the maximum landslides. They used the historical landslide inventories and made estimates of total area and volumes of landslides involved over time for each of the regions, and from these a lower bound estimate on regional erosion rates due to landslides.

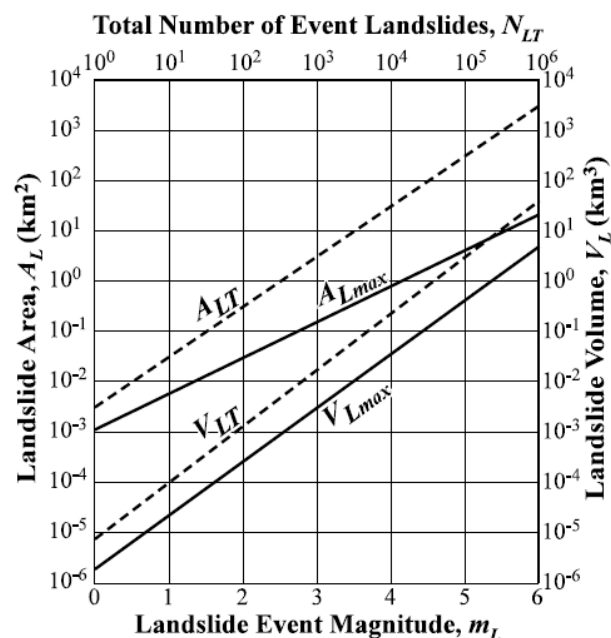


Fig.2. Predicted landslide-event areas and volumes associated with a given landslide-event magnitude (after Malamud *et al.*, 2004)

7. IMPORTANT FACTORS THAT INFLUENCE A LANDSLIDE EVENT

7.1. The presence of ground water in landslide

Most of the slope failure disasters in a humid area are caused by heavy rain. The topography of the surface of underground water does not follow the topography of ground surface. The topography of ground water surface in moved masses of a landslide is complex and forms a series of vein-streams. Detecting the veins of the ground water on landslide mass is the first necessary stage for planning drainage network and any other prevention works.

One-meter depth temperature measuring is a reliable method for conducting ground water detection on landslide masses.

Measurement of temperature of spring-water and wells in landslide area shows that the variation of temperature of ground water in winter and summer is negligible. However, the annual variation of temperature at one-meter depth is appreciable in landslide masses. This could be due to the non-uniform distribution of ground water. Based on this fact, it is possible to detect the vein-streams of ground water in displaced masses by measuring one-meter depth temperature (Takeuchi, 1980). In this method, the temperature of ground at one-meter depth and on a network of landslide map is measured.

The temperature around the vein-streams is lower during summer and autumn and higher during winter and spring, but it is nearly stable during winter and summer. As an example, the measured temperature at one-meter depth in Japan for one year is shown in Figure 3. It is about 16.5°C for all seasons (θ_w). The surface temperature is highly affected by ground water vein-streams. In Figure 3, the measured temperature at one-meter depth for the same point is shown (θ_u). It changes from 7.5°C in January to maximum 24.9°C in August. This phenomenon is the basic of the method for detecting vein-streams in landslide masses (Takeuchi, 1980).

This method could be used when the value of $|\theta_u - \theta_w|$ is greater than 2.5°C . Therefore, as shown in Fig. 3, late spring, early summer, late autumn and early winter are not appropriate for investigation.

The procedure of this method is as follows:

1. First the investigation area should be set up by mesh with the dimensions of $15\text{ m} \times 30\text{ m}$.
2. Non-conductive rods that thermal sensors are fixed to their tips were put at the bottom of holes. The thermal sensors were connected to the thermometer with the precision of $\pm 0.1^{\circ}\text{C}$.
3. One-meter depth temperature is measured after being constant (experimentally it was found that the waiting time should be about 10 minutes).

The measured temperature is affected by some factors. That is why correction of the influences of variation of measured value by a series of sensors, the effect of annual variation in temperature, geological condition, micro-topographical condition, existence of water in holes and the effect of ground condition should be done. All sensors must be calibrated before field survey by using high precision thermometer. Correction of annual variation is necessary when the fieldwork was done for a long time; a sensor must be set on a hole at the middle part mass and the temperature have to be measured three times a day. Experimentally, it was found that more than 350m difference in elevation could affect the results.

The calibrated values will be plotted on the map of landslide.

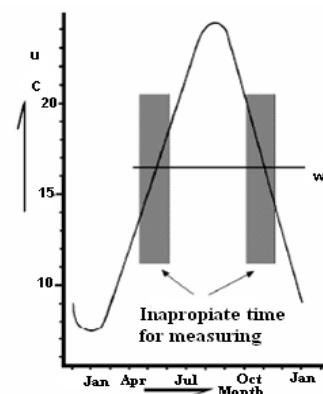


Fig.3. Annual time series of temperature changes of one-meter depth under ground, in Japan (after Malamud *et al.*, 2004)

7.2. The identification of weak zones of a saturated clay landslide

Once a landslide is triggered, material is transported by various mechanisms including sliding, flowing and falling. Landslides often occur along planes of weakness that may parallel the hill slope. In bedrock, planes of weakness are usually beds, joints or fractures. Soils such as silt and clay are weaker than rock and commonly have complex (or multiple) planes of weakness.

In the case of the saturated clays the shearing displacements lead to relative motions of the structural aggregates still uncrushed along discontinuity surfaces together with their opening. These relative movements bring about mechanical swelling, and consequently, the increase of water, the shear strength of soil is drastically diminished to the residual value, provoking local failure.

This happens in clays which are very near from the base of a loessian stratum and in which there is free water source (Tulucești, Dobrogea, Romania). In this way, due to dilatancy, a weak area, right under the loess appeared. This led to the detachment of a new plot of soil from the plateau, triggering in detrusive way the progressive landslide.

Landslides phenomena are very likely to occur when permeable strata lying on structured clay deposits facilitate the free water access towards mechanically swollen zones due to dilatancy.

It is very important to define the weak areas in which dilatancy appears, according to the stress values, because they are parts of the sliding surface. To simulate the behaviour of the soil in the plane state strain the hyperbolic model is used (Duncan, 1980).

For determining the parameters of the material (Izumi *et al.*, 1976), the tangent modulus E_t , bulk modulus B and dilatancy modulus H are calculated with relations:

$$E = \{1 - [R_{F1}(1 - \sin\varphi) (\sigma_1 - \sigma_3) / 2c \cos\varphi + 2\sigma_3 \sin\varphi]\}^2 K p_a (\sigma_1 / p_a)^{n_E}$$

$$B = K p_a (\sigma_3 / p_a)^{n_B}$$

$$H = \{1 - [3\tau_{net} R_{F2}(1 - \sin\varphi) / 2\sqrt{2}]\}^2 K p_a (\sigma_1 / p_a)^{n_H}$$

where n_E , K , n_B , n_H , R_{F1} , R_{F2} are hyperbolic models parameters and p_a is the atmospheric pressure.

In those areas of the slope an extra volumetric strain will appear.

Assuming that dilatancy appears as isotropic phenomena (Frank, Gullec, 1974) the correction of stresses is considered to be a state of initial stresses and a new vector of nodal loadings corresponding to it. In this way we can establish the deformations occurring with the corrected stresses, for all the finite elements, and we can identify the dilatancy areas along the slope.

Consequently, the landslides phenomena in fissured clays can be prevented by draining the critical areas or by modifying the soil mechanical properties using chemical or thermal methods.

8. CONCLUSIONS

Based on the results given in this paper and by numerous other authors, landslides satisfy power-law frequency–area statistics. Certainly landslide distribution is only approximate, but is it useful? The quantification of landslides is useful in terms of estimating the role of landslides in erosional processes and in assessing landslide hazards. It is only in terms of a general landslide distribution that the landslide-event hazard can be quantified.

It is based on this distribution that a landslide-event magnitude scale to quantify landslide events was introduced. Using the distribution, we can extrapolate the largest landslide areas associated with a landslide event. This approach should be useful to hazard managers who do not have the time and resources to derive substantially complete inventories.

With one-meter depth temperature measuring method we can conduct ground water detection on landslide masses, and we agree that solving the problem starting from the identification of weak zones, using dilatancy criterion, is more rational than using models with many parameters, usually more than 7, which are hard to determine.

REFERENCES

1. Brab E. E., Harrod B. L. (1989) Landslides: Extent and Economic significance. Rotterdam.
2. Canuti, P., and Casagli N. (1996) Considerazioni sulla valutazione del rischio da frana *CNR-GNDCI*, Tipografia Risma, Firenze, Italy.
3. Chirica A., Stănculescu I. (1994) Some aspects concerning the macrostructured soils of Romania Proceedings of XIII ICSMFE, New Delhi, Vol.1.
4. Crescenti U. (1998) Il rischio di frana: appunti per la valutazione *Quaderni di Geologia Applicata*, Vol.5(2), pp. 87-100, Pitagora Ed., Bologna.
5. Crescenti U., Rainone M.L., Sciarra N. (1999) An example of landslide risk assessment in a historical centre of central Italy *Proceeding of International Symposium on Engineering Geology, Hydrogeology and Natural Disasters with Emphasis on Asia*. Journal of Nepal Geological Society.
6. Duncan J.M. (1980) Strength, stress-strain and bulk modulus parameters for finite elements analysis of stresses and movements in soil masses, Berkley.
7. Frank R., Gullec P. (1974) Méthode de calcul de la dilatance Bulletin de Liaison de Laboratoires de Pont et Chaussées
8. Gabrielov A., Newman W.I., Turcotte D.L. (1999) Exactly soluble hierarchical clustering model: Inverse cascades, self-similarity, and scaling *Phys. Rev. E* 60, 5293 – 5300. Vol.5 – November 1999.
9. Guzzetti F, Malamud B., Turcotte D., Reichenbach P. (2002) Power-law correlations of landslide areas in central Italy *Earth and Planetary Science Letters*. Vol.195, pp.169–183.
10. Guzzetti F., Reichenbach P., Cardinali M., Ardizzone F., Galli M. (2003) Impact of landslides in the Umbria Region, Central Italy *Natural Hazards and Earth System Sciences*. Vol.3, pp.469–486.
11. Guzzetti F., Reichenbach P., Wieczorek G. (2003) Rockfall hazard and risk assessment in the Yosemite Valley, California, USA *Natural Hazards and Earth System Sciences*. Vol.3, pp.491–503.
12. Hergarten S., Neugebauer H.J. (1998) Self-organized criticality in a landslide model *Geophys. Res. Lett.*, Vol.25(6), pp.801-804.
13. Hergarten S., Neugebauer H.J. (2000) Self-organized criticality in two-variable models *Phys. Rev. E*, Vol.61(3), pp.2382-2385.
14. Hovius N., Stark C.P., Allen P.A. (1997) Sediment flux from a mountain belt derived by landslide mapping *Geology*.
15. Izumi H. et al. (1976) Finite element analysis of stresses and movements in excavations. Numerical methods in Geomechanics. *Second International Conference* Virginia Polytechnic Institute, USA, pp.701-711.
16. Keefer D.K. (1984) Landslides caused by earthquakes *Bulletin of the Geological Society of America*, Vol.95, pp.406-421.

- 17.Malamud B., Turcotte D., Guzzetti F., Reichenbach P. (2004) Landslide inventories and their statistical properties *Earth Surface Processes and Landforms*. Vol.29, pp.687–711.
- 18.Malamud B., Turcotte D. (2000) Cellular-automata models applied to natural hazards *IEEE Computing in Science and Engineering*. Vol.2, pp.42–51.
- 19.Noever D.A. (1993) Himalayan sand piles *Phys. Rev. E*, Vol.47, pp.724-725.
- 20.Ohori H., Hirano M. (1988) Magnitude, frequency, and geomorphological significance of rocky mud flows, land-creep, and the collapse of steep slopes *Z. Geomorph. N.F.*, Vol.67, pp.55-65.
- 21.Pelletier J., Malamud B., Blodgett T., Turcotte D. (1997) Scale-invariance of soil moisture variability and its implications for the frequency-size distribution of landslides *Engineering Geology*. Vol.48, pp.255–268.
- 22.Reichenbach P., Guzzetti F., Malamud B., Turcotte D. (2001) Comparison of two landslides triggering events using frequency-area statistics *Mediterranean Storms - Proceedings of the 3rd EGS Plinius Conference held at Baha Sardinia, Italy, October 2001*.
- 23.Rodriguez C.E., Bommer J.J., Chandler R.J. (1999) Earthquake-induced landslides: 1980-1997 *Soil Dynamics and Earthquake Engineering*, Vol,18, pp.325-346.
- 24.Sasaki Y. (1991) Fractals of slope failure number-size distribution. *J. Japan. Soc. Engineering Geol.*, Vol.32, pp.1-11.
- 25.Stanculescu I. (1960) Some problems concerning stability of clay slopes *SC Bul. Of Civil Engineering*, Institute of Bucharest.
- 26.Sugai T., Ohmori H., Hirano M. (1994) Rock control on magnitude-frequency distributions of landslides *Trans. Japan. Geomorph. Union*, Vol.15, pp.233-251.
- 27.Turcotte D., Malamud B., Guzzetti F., Reichenbach P. (2005) A general landslide distribution: further examination *Landslide Risk Management - Hungr, Fell. Couture & Eberhardt (eds) Taylor & Francis Group, London. ISBN 04 1538 043 X*.
- 28.Whitehouse I. E., Griffiths G. A. (1983) Frequency and hazard of large rock avalanches in the central Southern Alps, New Zealand *Geology* Vol.11, pp. 331-334.