UNIVERSITÉ DE GENÈVE
FACULTÉ DES SCIENCES

Specialisation certificate in geological and climate related risk

CERG-C 2012

MODELING SHALLOW LANDSLIDES TRIGGERED BY RAINFALL IN TROPICAL AND MOUNTAINOUS TERRAINS

by

Edier Vicente Aristizábal Giraldo

under the supervision of : Michel Jaboyedoff

February 2013

Schweizerischer Erdbebendienst
Swiss Seismological Service

Schweizerische Eidgenossenschaft
Confédération suisse
Confederazione Svizzera
Confederaziun svizra
Direction du développement et de la coopération DDC

UNIVERSITÉ DE GENÈVE
INSTITUT DES SCIENCES DE L'ENVIRONNEMENT

UNIL | Université de Lausanne

International Strategy ISDR for Disaster Reduction

UNOSAT
UNITAR Operational Satellite Applications Programme

Avec le soutien de la FONDATION HANS WILSDORF

REPUBLIQUE ET CANTON DE GENEVE
POST TENEBRAS LUX

STATEMENT

I certify that all statements in this text that is not the result of my personal reflection is attributed to its source and that any passage copied from another source is also placed in quotes.

Table of Contents

ACKNOWLEDGMENTS

This thesis was possible thanks to the financial support from the Hans Wilsdorf Foundation to the author. Thanks a lot for making possible this extraordinary experience in Switzerland.

I would like to thanks to my supervisor, Prof. Michel Jaboyedoff, and his post-doctoral research student, Ivanna M. Penna, in the University of Laussane, for your support, valuable guidance and advice.

Also, I wish to express my deepest gratitude to Dr. Corine Frischknecht, CERG-C coordinator, Prof. Costanza Bonadonna, CERG-C director, Section of Earth and Environmental Sciences, University of Geneva, for your abundantly helpful and invaluable assistance. Thanks a lot.

Finally an honorable mention to the CERG-C staff and friends for providing me valuable knowledge, guidance and motivation. Hope to see all you again.

To my brother, you will always be with me, I love you so much.
For my son, thanks for giving back sense to my life.

INTRODUCTION

Landslides are one of the most common causes of fatalities and economic losses worldwide (Schuster, 1996), therefore the capacity to predict these movements has been a topic of great interest to scientific community (Caine, 1980; Montgomery & Dietrich, 1994; Finlay et al., 1997; Iverson, 2000; Aleotti, 2004; Crosta & Frattinni, 2008; Sidle & Ochiai, 2006).

Although landslides do represent changes in terrain morphology within the natural and continuous geomorphologic cycle (Scheidegger, 1998) its occurrence and losses associated in the last decades has been closely tied to world population growth and consequent urban expansion on susceptible slopes to landslides. The urban population of developing countries has increased by 5 in 40 years and continues increasing rapidly (UNFP, 2007). The greatest landslide losses occur in the Ring of Fire countries (Alcantara - Ayala, 2002). Estimation made by Varnes (1981) indicates that 89% of deaths, due to landslides, are located in those countries. Data presented by Sidle & Ochiai (2006) pointed to Nepal, Japan and China as the countries with the biggest number of landslide fatalities per year, with values between 190 and 150; in Latin America, Brazil lead the ranking with an average of 88 people killed per year. In economic terms, Japan is the country most affected by landslides, with an estimated loss of 4 billion dollars annually, followed by Italy, India and the United States with losses ranging from 1 to 2 billion dollars per year (Cruden et al., 1989; Schuster, 1996; Schuster & Highland, 2001; Sidle & Ochiai, 2006). Unfortunately Colombia has no accurate databases to estimate this statistics.

Numerous studies have been developed in recent years that have allowed an increasing understanding of the causes that involve these morphodynamic processes. However, because of the complexity involved in landslide occurrence, a great uncertainty still exists in predicting this kind of events.

Landslides have multiple and complex variables, such as soils mechanic and hydrological aspects, but just a single factor becomes the triggering element for landslides, generating an almost immediate response, which is to mobilize slope materials, either by the rapid increase efforts or by reducing shear strength (Wang & Sassa, 2006). This triggering factor is generally rainfall events, earthquakes, volcanic eruptions or human activities. In tropical and complex terrains a high percentage of these movements are triggered by heavy or prolonged rainfalls (Aristizábal & Gómez, 2007).

Rainfall as a triggering factor in landslide occurrence has been studied by many authors (e.g. Montgomery & Dietrich, 1994; Finlay et al., 1997; Crozier, 1999; Iverson, 2000). One of the most acceptable approaches is from a mathematical point of view, developing physical models based on geotechnical and hydrological elements that relate rainfall, pore pressure and slope stability. These models combined with rainfall forecasts and real-time monitoring has been implemented during the last years into early warning systems.

The present research project aims to set-up a model to assess shallow landslides trigger by rainfall in tropical and complex terrains, based on geotechnical and hydrological aspects. The overall goal is to develop a tool that could aid to predict shallow landslide occurrences and that could be incorporated into an early warning system in the Aburrá Valley (Colombia).

1. HAZARD, VULNERABILITY AND RISK CONDITIONS OF THE ABURRÁ VALLEY

1.1. The Aburrá Valley - Colombia

The Aburrá Valley, with an area of 1326 km^2 and a length of 65 km, is located in the northern Central Cordillera of the Colombian Andes. It extends approximately between latitudes 6° 00" N and 6° 30" N and longitudes 75 ° 15" W and 75 ° 45" W (Figure 1). Its climatic conditions are typical of tropical environments, with an average temperature of 22 ° C and relative humidity of 70%. Precipitation has a bimodal dsitribution, with peaks during May and October. The mean annual rainfall varies from 1400 mm in the central part and 2700 mm in the north and south of the valley.

Morphologically, the Aburrá valley is defined by Arias (2003) as a depression with north-south orientation and flat bottom. It is limited by steep rock slopes and bottom covered by debris and mud flow deposits. Geologically it is composed of Paleozoic metamorphic basement, ultra-basic igneous rocks, and a sequence volcano - sedimentary and intrusive granitic bodies. Large deposits derived from ancient landslides, classified as debris flows and landslides, cover the middle and lower slopes. The rocks and deposits have thick weathering profiles. The depths of profiles vary significantly depending on the parental rock and local conditions. In granitic rocks, the weathering profiles are characterized by reddish yellow exceeding 30 m depth. The weathering profile developed from ultra basic and metamorphic rocks is thin, varying between 10 and 20 m deep, generally red and orange (7.5 YR 7/6), with intensely weathered rock fragments (Aristizábal et al. 2005).

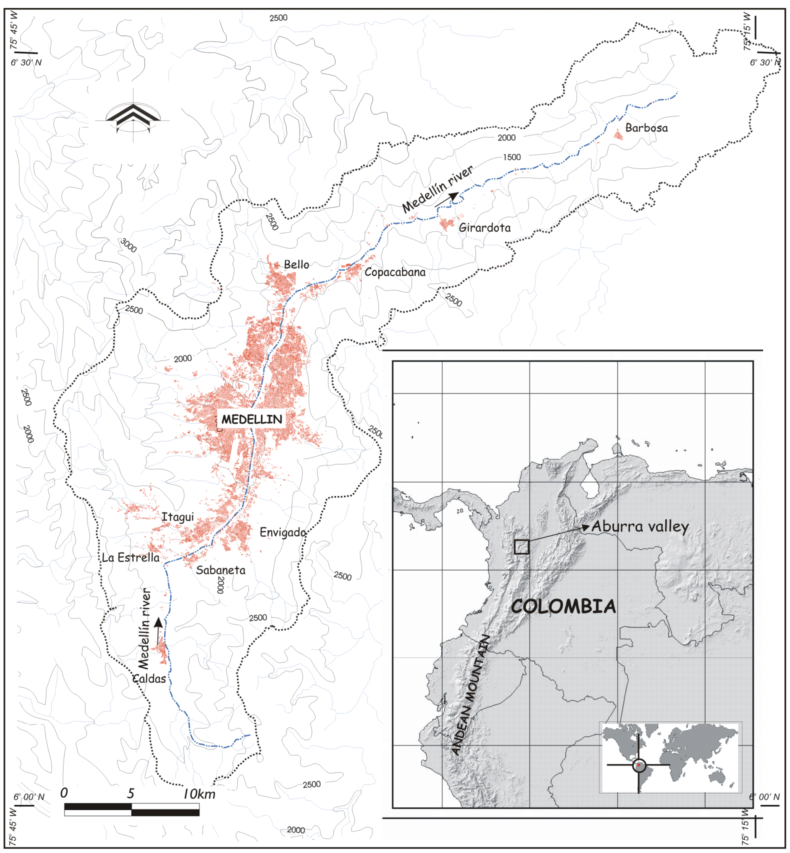

Figure 1. Location of the Aburrá Valley.

Currently, the Aburrá Valley has an estimated population of 3.3 million inhabitants, where 95% are urban population located in only 26% of the territory (340 km^2 of urban area) (Table 1). The most populated municipalities, Medellín (2.2 million), Bello (372 K), and Itagüí (230 K), concentrate much of its population over the valley slopes (DANE, 2005). This population growth has been extremely fast, in a century the valley's population has multiplied by 30, from 103,305 population in 1905 increased to 3'317.166 in 2005. Since 1950, thousands of immigrants have occupied areas exposed to natural disasters (Hermelin, 1984), and even over areas that had been affected by major events.

Table 1. Population and population growth in the Aburrá Valley (Data from DANE, 2005).

Municipalities	Population in 1951	Population in 1964	Population in 1973	Population in 1985	Population in 2005	Urban denstiy population, 1997 (Inh/km²)
Medellín	385.189	772.887	1.151.762	1.468.089	2.223.078*	15.639
Bello	34.307	93.207	129.173	212.861	371,973*	9.500
Envigado	28.797	61.546	73.057	91.391	175.240*	>10.000
Itagüí	20.151	68.086	103.898	137.623	230.272*	21.281
Sabaneta	----------*	----------*	16.518	20.491	44.820*	5.380
Barbosa	15.507	15.242	22.271	28.623	42.537*	7252
Caldas	12.431	25.081	33.630	42.158	67.372*	26.769
La Estrella	8.698	16.479	23.619	29.918	52.709*	9.763
Girardota	10.956	12.729	17.879	23.684	42.744*	3.838
Copacabana	10.720	19.403	29.997	40.309	61.421*	10.917
Total	526.756	1.084.660	1.601.804	2.095.147	3.317.166*	

1.2. Hazard and vulnerability in the Aburrá Valley

The Aburrá Valley has been affected by a large quantity of disasters, most of them with a magnitude between small (<10 deaths) and moderate (10 – 100 deaths). According to Aristizabal & Gomez (2007), during the period 1880 – 2007, it was registered 6750 disasters. From this period 42% of total events correspond to floods, 35% to landslides, and 15% to forest fires. The sum of these three types of natural phenomena account for 92% of all disasters. It means in 10 disasters that occurred in the Aburrá Valley, about 8 of them owe to floods or landslides, reflecting the close relationship between natural disasters and hidrometerological conditions of the valley. Man-made disasters are small, however its impact and recurrence has increased during the last two decades (Figure 2).

These geological and hydrological hazards have been recurrent in the Aburrá Valley associated to the origin and evolution of the valley (Aristizabal et al, 2005). However, in the last five decades the hazard dynamics have changed due to the human occupation on the valleys slopes and flooding plains, causing hundreds of deaths and millions in economic losses. Most populated cities in the valley are the most affected, Medellín with 72% of disasters, following by Itagui (5,4%), Envigado (4,9%), and Bello (4,8%). The city into the valley with the lowest number of disaster recorded is Barbosa with only 1.2% of the total.

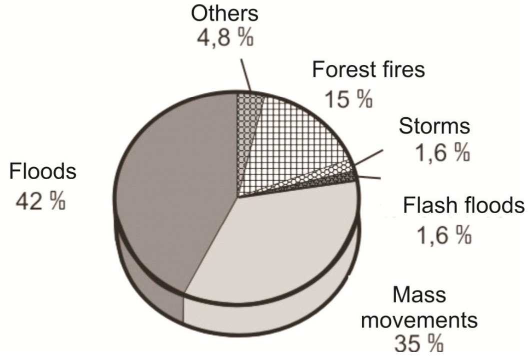

Figure 2. Percentage of records according to hazard, predominanting hydroemteorlogical hazards. Man-made disasters are within the item called others.

According to Aristizabal et al. (2005) disasters generated by floods are concentrated in Medellín (74%), followed by Itagüí (8.5%), Bello (5.5%) and Envigado (4%). Regarding landslides, 82.4% of them occurred in Medellín, followed by Caldas (3.6%) and Girardota (3%).

Nine moderate disasters, with fatalities between 11-100, have been recorded in the Aburrá Valley, and only two events classified as major disasters, with more than 100 deaths, which correspond to (1) Media Luna landslide, occurred on July 12, 1954 with a balance of more than 100 people dead, of which only recovered the bodies of 77 people and an undetermined number of missing, which is considered to be about 70 passes, and (2) Villatina landslide on September 27th, 1989, although never able to determine the exact number of victims is considered around 500 people, where only rescued about 200 bodies (Table 2).

Landslides recognized in the Aburrá Valley are mainly soils slips and debris/mud flows. Some of the landslides more remembered for its great impact are: Rosellon (1927), Media Luna (1954), Santo Domingo Savio (1974), Villatina (1987), La Cruz (2007), El Socorro (2008), El Poblado (2008), and La Gabriela (2010).

Table 2. Main landslides occurred in the Aburrá valley (Modified from Aristizábal & Yokota, 2006).

Landslide	Date	Place	Deaths
Mudflow	18 June 1927	Rosellón (Envigado)	22
Mudflow	12 July 1954	Media Luna (Santa Elena)	>100
Debris slide	25 June 1973	La Manguala (S.A. Prado)	13
Mudflow	29 September 1974	Santa Domingo (Medellín)	>70
Landslide	4 February 1975	Medellín	18
Debrisflow	20 October 1980	San Antonio (Medellín)	>18
Debris slide	23 November 1984	Santa Maria (Itagüí)	10
Mudflow	27 September 1987	Villatina (Medellín)	>500
Complex rotational slide	31 May 2008	El Socorro (Medellín)	27
Complex rotational slide	16 November 2008	El Poblado (Medellín)	12
Complex rotational slide - debris flow	5 december 2010	La Gabriela (Bello)	84

The annual disaster distribution shows a clear bimodal distribution, with peaks in the months on May and October, indicating a direct relationship between disaster occurrence and rainfall annual distribution.

1.3. Risk condition of the Aburrá Valley

In the Aburrá Valley there is still a lack of rigorous databases that consider the assessment of damages and economical losses by disasters. According to partial results showed by Aristizabal & Gomez (2007) using the DesInventar database, the disasters recorded have left a tragic toll of 1390 deaths during the last century, most of fatalities were generated by landslides (74%) and flash floods (13%) (Figure 3). Although floods are the most recurrent disaster, represent only 5% of the total fatalities. The number of homes affected is mainly associated to flooding (53%), followed by landslides (33%), and the highest proportion of people affected is in the city of La Estrella with 41% followed by Medellin with 28%.

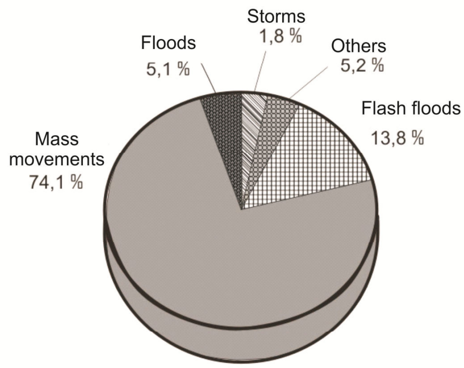

Figure 3. Porcentage of fatalities according to the hazard phenomenon.The most tragic events in terms of human lives islandslides.

Only landslides have caused 1030 fatalities and huge economic losses, but it has been not estimated in detail, Aristizábal & Gómez (2007) for the period 1880-2007 and using just the information available estimated a minimum value of US$10million.

Only in Medellín, a total of 29,174 households are located in areas of high risk, equivalent to 112,697 people, or 4.9% of total households of the municipality. For the rest of the Aburrá Valley recent studies have established that there are 16.847 homes in high risk areas, which corresponds to 62.057 inhabitants (UNal, 2009). These data together with the results obtained in Medellín in 2005 show a total of 174,384 people located in high risk areas equivalent to 5.1% of the total population (Table 3).

Table 3. Homes and population located in high risk areas

Municipalities	High Risk Population	High Risk Homes
Medellín	112.697	29.174
Barbosa	3.646	961
Bello	21.391	5.381
Caldas	2.357	614
Copacabana	4.345	1.172
Envigado	6.856	2.260
Girardota	457	123
Itagüí	17.954	4.942
La Estrella	4.489	1.199
Sabaneta	562	195
TOTAL	**174.754**	**46.021**

1.4. Background

Although the occurrence of landslide has affected the Aburrá Valley for a long time, studies on this subject arise only since 1980's (Shlemon 1979, Hormaza, 1991; Flórez et al., 1996 & 1997, Rendón & Vargas, 1998; SIMPAD, 1999; AMVA, 2002). These studies essentially evaluated the landslide susceptibility on the Aburrá Valley slopes, through methodologies based on morphological criteria (Chica, 1987; Ingeominas 1990, UNDP - 1998). Recent research projects have introduced new methodologies and statistical tools based on GIS or neural networks in the Aburrá Valley and similar regions in Colombia (UNal, 2009, Londoño, 2007, García, 2004, Ángel et al., 2002, van Westen & Terlien, 1996), some of them including rainfall as an element of analysis.

Although in Colombia, and specifically in the Aburrá Valley, most landslides have been induced or triggered by intense or prolonged rainfall, few studies considering rainfall as a triggering factor have been carried out (Paz & Torres, 1989; Gómez, 1990; Van Westen et al., 1994; Castellanos, 1996; Castellanos & Gonzalez 1997; Terlien, 1997; Mayorga, 2003; Echeverri & Valencia, 2004; Moreno et al., 2006, Suarez, 2008).

Aristizabal & Gómez (2007) compared the Aburrá valley disaster database with the rainy seasons in the period 1880-2007, finding a close relationship between precipitation and landslide occurrence, with a bimodal seasonal cycle with maxima during May and October, and minima during January and July. For every 10 events occurring in the Aburrá Valley, 8 were originated by hydro-meteorological phenomena, mainly landslides and flash floods (see Figure 4).

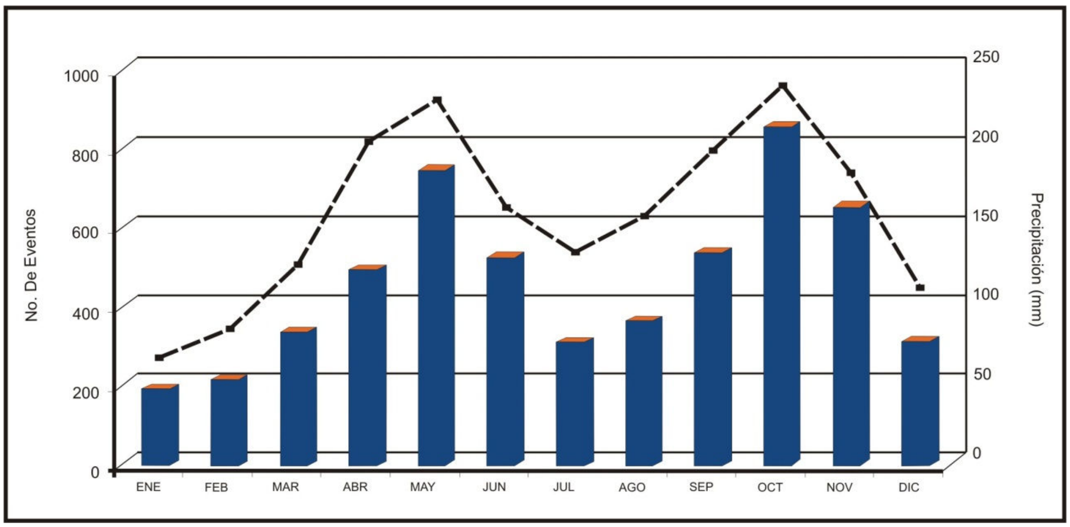

Figure 4. Average seasonal cycle of distribution of landslide occurrence in the period 1880 - 2007 versus average monthly precipitation for the Aburrá Valley (from Aristizábal & Gómez, 2007).

Aristizabal et al. (2011) analyzed critical rainfall thresholds for landslides forecasting in the Aburrá Valley by empirical procedure, using a database of landslides and precipitation. The results show that the major conditioning for the occurrence of landslides in the Aburrá Valley is the antecedent rainfall. The data indicate that landslides used in the analysis occurred for antecedent rainfall over 60 mm for 30 days, 160 mm for 60 days and 200 mm for 90 days.

2. FORMULATION OF THE PROBLEM

2.1. Objective

To develop a physically based model for predicting shallow landslides triggered by rainfall in tropical mountainous terrains.

2.2. Specific objectives

- To propose a conceptual and mathematical model that responds to the physical process involved in shallow landslide occurrence in tropical mountainous terrains.
- Based on the conceptual model, to develop a computational model supported on hydrological and geotechnical aspects for the prediction of landslides triggered by rainfall in tropical mountainous terrains.

2.3. Working hypothesis & scope

- Landslide occurrence is the result of a highly complex nonlinear system; however it is possible to fit a simplified model with acceptable confidence levels according to the degree of understanding reached on the variables and parameters.
- Shallow landslide occurrence is a function of rainfall infiltration as a triggering factor; besides a large number of variables. This function is complex and it is based on shear strength reduction by pore pressure increasing.
- Shallow landslides triggered by rainfall in tropical environments generally have irregular fault surface but planar trend, they can be classified as soil-slip or shallow translational slides rapidly changing to debris or mud flows.
- The analysis of this project will not include deep landslides with complex surface fault or displacement behavior after slope failure. The analysis will consider only so far the slope failure. The subsequent process is another key element to hazard and risk assessment, which involves a detail analysis of rheology, but it is out of the scope of this study.
- The methods used for developing and testing the model will be mathematical tools and computer simulation, the implementation and validation of the model in a real basins is out of the scope.
- The project aims to understand, explain and model the more *simple* process involved on landslide occurrence, which essentially considers natural terrain conditions. External factors such as human disturbances will not be incorporate into the analysis.

4. STATE OF THE ART

Landslides triggered by rainfall, usually called soils-slip, have a planar slip surface. They are characterized by their small thickness (0.3 - 2 m) much smaller than the flow length, slip surface fault parallel to the slope and escarpment small area (Anderson & Sitar 1995). Shallow landslides are generated during intense rainfall events by the rapid increase in pore pressure or by the loss of apparent cohesion component (Wang & Sassa, 2003; Terlien, 1998, Crosta, 1998; Crosta & Frattini, 2003). Subsequently, the displaced material, by processes of liquefaction and rapid reduction of shear strength in undrained conditions (Anderson & Sitar, 1995), becomes a flow that spreads down, transporting sediment eroded from the channel, increasing starting material displaced volume (Wang & Sassa, 2003).

Collins & Znidarcic (2004) proposed two different failure mechanisms generated by infiltration. In the first mechanism failure occurs due to positive pore pressure increasing caused by liquefaction of the material, while the second failure mechanism occurs in negative pore pressures where the material is still in unsaturated state and failure occurs due to reduced suction and mass behaves like a rigid body.

These considerations permit to evaluate pore pressure generated by rising saturated layer on a predefined critical failure surface, or conversely assess the development of pore pressure from a wetting front advance.

4.1. Variables of shallow landslides triggered by rainfall

The complexity in finding the probability of reaching a critical depth of saturation and therefore predict shallow landslides occurrence triggered by rainfall is a function of a large number of parameters involved and closely linked, such as mechanical, physical and hydraulic properties of soils, critical soil moisture content, thickness of weathering profile antecedent and rainfall intensity, local geomorphology, vegetation, which contribute to soil strength and conditions of subsurface flow, inducing varying conditions of instability in response to rainfall patterns (Crosta, 1998, Wang & Shibata, 2007; Rahardjo et al., 2007) (Figure 5).

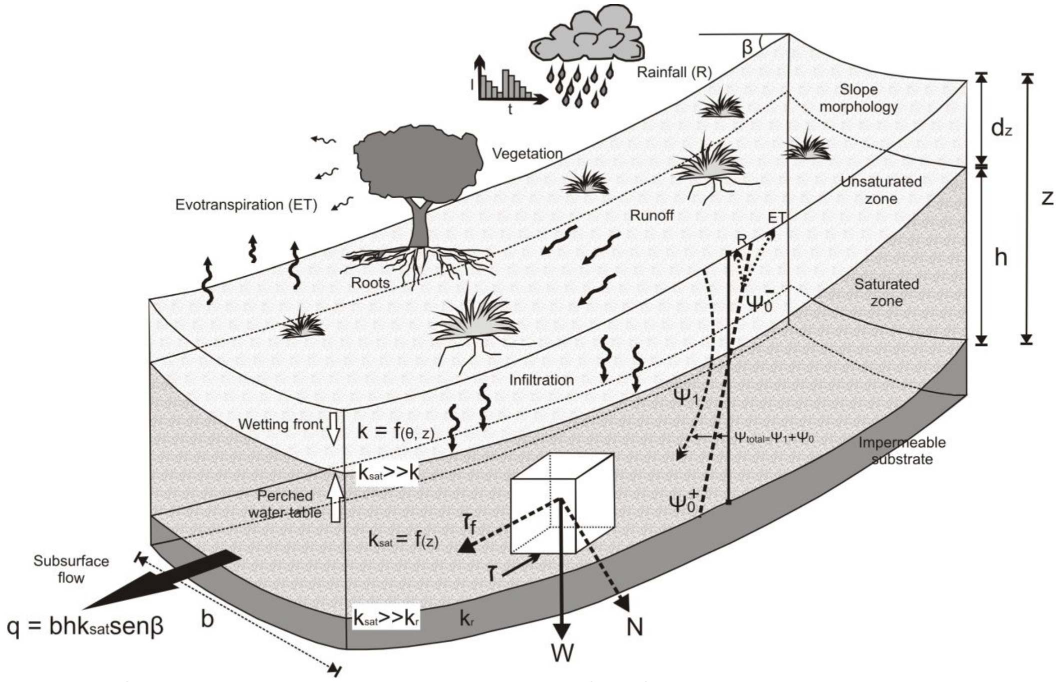

Figure 5. Schematic three-dimensional weathering profile of a convergent morphology slope under rainfall conditions.(Θ) volumetric water content (Ψ) pore pressure, (k) permeability, (W) weight.

4.2. Models for shallow landslide prediction

A variety of techniques have been developed for landslide susceptibility and hazard assessment (Varnes, 1984; Barredo et al., 2000, Dai & Lee, 2001, Guzzetti et al. 1999, and Hutchinson, 1995). In essence, these methodologies can be grouped into: (1) heuristic methods based on the understanding of geomorphological processes, (2) statistical methods based on statistical predictions for combination of variables generating landslides in the past, and (3) methods based on deterministic models of slope stability, which thanks to the development of geographical information systems have been used in regional zoning.

In general, neither of these methods take into account dynamical variables or triggering factors, nor the short and long term behavior of these variables (Crosta & Fratiini, 2003; van Beek & van Asch, 2004). Therefore, that those analysis only reflect the susceptibility of slopes to landslide occurrence. To characterize landslide hazard it is required to evaluate slope susceptibility to fail and the probability of mass movement in terms of time (Crosta & Fratiini, 2003). In the case of hazard, the triggering factors must be converted in terms of frequency and magnitude, in this case rainfall, which are very specific and dynamic to occurrence site (van Westen et al., 2006).

In practice, the incorporation of dynamic or triggering factors is done by statistical or physical methods. The methods based on statistical considerations define critical thresholds that usually relate rainfall intensity and magnitude with landslide occurrence. These studies depend largely on data quality in both the inventory of mass movements as record rainfall. On the other hand a large group of researchers have faced this problem from the mathematical point of view, developing physical models based on geotechnical and hydrological patterns that relate rainfall, pore pressure and slope stability, these models have the ability to assess the activity spatial and temporal instability of slopes but depend heavily on input variables and boundary conditions (Crosta & Frattini, 2003; Aleotti, 2004, van Beek & van Asch, 2004). Quantifying triggering mechanism is an essential step towards landslide hazard prediction, so currently the challenge focuses on quantifying physical processes related to infiltration of rainfall, recharge of sub flows -surface and consequently landslide occurrence (van Westen et al., 2006).

4.3. Physical methods to predict landslide occurrence

Physical methods generally explain shallow landslide occurrence combining geotechnical analysis to determine critical pore pressures and hydrologic analysis to assess rainfall amount that is required to raise such critical pore pressures (Terlien, 1998).

4.3.1 Hydrological analysis in physical methods. A variety of hydrological triggering systems for shallow landslides have been presented for several decades; there are many types based on a wetting front descending into the soil or a perched water table at the contact between soil and bedrock (van Asch et al. 2009) (Figure 6).

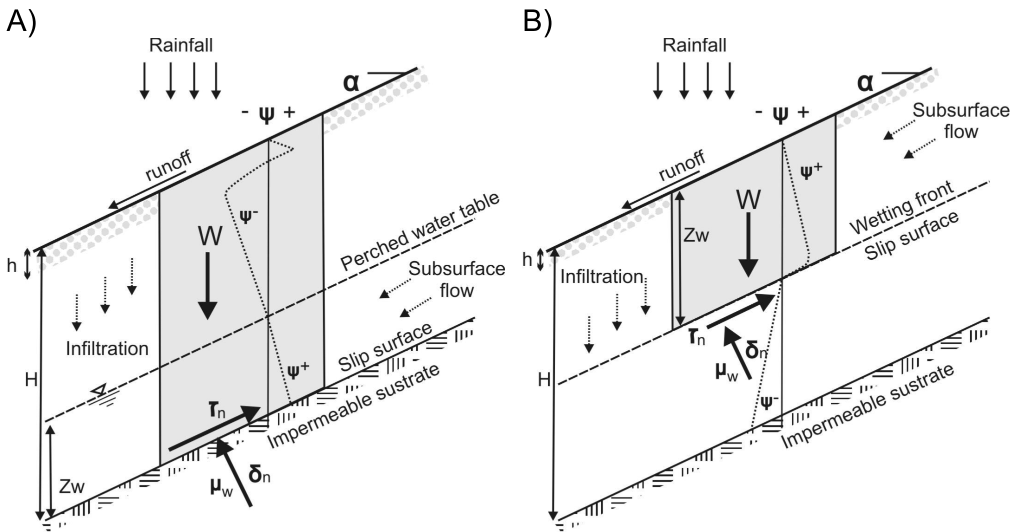

Figure 6. Two possible mechanisms for the saturation for shallow landslides triggered by rainfall: A) rising perched water table with parallel to slope seepage; B) wetting front advancing from the slope surface (modified from Xie et al., 2004)

One of the best known models is based on the idea of increasing density and reducing hydraulic conductivity of the regolith with depth, where rainfall rate exceeds the percolation rate in depth, creating a perched subsurface water flow in the regolith and assuming that saturated zone flow is parallel to the slope. In this model the most critical situation for slope stability is considered when a saturated zone reaches the slope surface and water pressure in soil pores is limited by its height (Anderson & Sitar 1995).

According to this view, there are simple considerations that propose hillside hydrology as a sub-surface flow in a static state and evaluate topographic control on pore pressure (Montgomery & Dietrich, 1994), which have a tendency to overestimate the hazard spatially depending on data quality (Crosta & Frattinni, 2008) and hydrological models for unsaturated slopes initially considered transient dynamic flows, assessing shallow landslide hazard for specific storms. Pore pressure that develops in soils, on these cases, occurs as a transient process according to vertical infiltration movement, additional shear strength depends on suction degree or negative pore pressure (Sharma & Nakagawa, 2005; Huat et al., 2006; Collins & Znidarcic, 2004).

Other models based on static state in kinematic wave hydrology have been used for saturated hillside slopes for a large number of researches (Troch et al., 2002; Paniconi et al., 2003; Rezzoug et al., 2005). They generally used the Boussinesq equation for Hillside Storage Capacity formulated in terms of Darcy's equation and continuity in terms of water storage capacity on the ground as the dependent variable (Troch et al., 2003).

Some authors consider that subsurface water flow concept in a static state is not appropriate to assess triggering shallow landslide causes due to short periods of response of pressure head in some soils (Matsushi et al., 2006, Chiang & Chang, 2009). In assuming rainfall in a static state is consequently removing the effect of the redistribution of water pressure on the floor perpendicular to the slope associated with transient rainfall infiltration, and so these models cannot predict temporal response of shallow landslides to rainfall variable patterns (Iverson, 2000). Borga et al. (2002) consider unrealistic assumption of a static state for wet indexes, due to these models assume that sub surface flow anywhere in a landscape depends on upstream drainage area, which is valid only if subsurface water flow recharge occurs by the time required for each point to reach equilibrium sub surface drainage and generate drainage from its entire upstream contributing area, but due to low sub surface flow velocity this assumption is very difficult to meet, usually only receives input from a small portion of total drainage area.

One of the dynamic and vertical hydrological models most widely known and used worldwide was developed by Iverson (2000), who considered a transitional flow regime and partially saturated soil from Richard equations, requiring as input rainfall intensity-duration and a characteristic of hydraulic diffusivity. Baum et al. (2002) developed a Fortran program called TRIGRS based on a model of transient one-dimensional vertical infiltration model with a simple slope stability, according to the developments of Iverson (2000), assuming saturated conditions or very close to saturation. Some discussed aspects of the model proposed by Iverson are based on the failure not considering slope flow direction, morphology, and infiltration levels (Montgomery & Dietrich, 1994), and the model is only significant for short duration rainfall (Frattini et al., 2009).

Another model of vertical infiltration widely known is the Green-Ampt model, which is defined as a simple infiltration model with very consistent results with Richard equations (Ekanayabe & Phillips, 1999, Xie et al., 2004, Qiu et al., 2007). Originally this model was developed for water infiltration on horizontal surfaces where ponding occurs. Therefore, to use it on slopes some modifications are needed (Setyo & Liao, 2008).

4.3.2 Geotechnical aspects of physical methods. Geotechnical models for rainfall triggered shallow landslides are generally used constant slope and infinite length, assuming failure surface parallel to ground surface and failure length much greater than the thickness of the displaced mass (Borga et al., 2002). These tests are based on shear (τf) on the slope should not exceed shear strength (τ) of the material, therefore the factor of safety (FS) of the slope can be defined in terms of effective efforts by the relationship between τ / τf (Brunsden & Prior, 1984).

Although the vast majority of models use soil properties to calculate the factor of safety based on stability analysis of infinite slopes, they differ in the method in which the pore pressure is calculated as discussed above. There are physical models as DSLAM (Wu & Sidle, 1995); SINMAP (Pack et al., 1998); SHALSTAB (Montgomery & Dietrich, 1994), LISA (Hammond et al., 1992), which assume a static state, saturated flow parallel to the slope and use Darcy's Law to estimate spatial distribution of pore pressure, except in LISA which requires only water table depth. In contrast, the stability model of Iverson (2000), in contrast, considers transient unsaturated flow to estimate pore pressure response at depth.

One of the first known methods was the DSLAM, it is a physical model of stability, dynamic, and distributed, which uses a water flow model and a kinematic wave model dynamic vegetation growth. The most recognized physical model was proposed by Montgomery & Dietrich (1994), called SHALSTAB. This model employs a TOPOG hydrological model (O'Loughlin, 1986) to estimate the height of saturated portion of profile, which assumes that the dominant control of spatial distribution of shallow landslides is given by topography, which define slope angle and subsurface flows converge. They define an index for analysis of soil saturation which is used to predict the water table in terms of water flow in soil and rainfall intensity.

SINMAP method corresponds to a computer program that predicts the potential for shallow landslide stability, similar numerically to SHALTAB, due to it uses the same factor of safety equation and Darcy's Law for saturated flow. The difference is that SHALTAB ignores cohesion.With respect to the LISA method (for Level I Stability Analysis) this was developed by the USDA, for soils with similar topography and geology. LISA is a development under a probabilistic analysis based on the factor of safety, considering the trees weight and vertical depth of saturated soil.

5. NEW MODEL PROPOSED BY TROPICAL AND MOUNTAINOUS TERRAINS

Numerous studies have demonstrated that shallow landslides triggered by rainfall have multiple factors, such as antecedent and rainfall pattern, hillslope morphology, mechanical and hydrological soil properties. Therefore, to model the positive pore pressure increasing and subsequently reduction of shear strength and slope stability, it is necessary an integral approach based on hydrology and geology elements.

In order to model shallow landslides triggered by rainfall in tropical and mountainous terrains and obtain consistent and coherent results, is needed a complete understanding of the conceptual model and assumptions. This chapter describes initially the conceptual model and assumptions considered to propose a computational program to forecast the potential occurrence of shallow landslide triggered by rainfall in tropical and mountainous terrains. The document describes the hydrological and geotechnical modules in detailed, to finally describe the model program proposed, and how it works.

The models available in literature differ on the detail to describe the hydrological and geomechanical mechanism in hillslopes. Most of the models consider the variation of the pressure head response, related to rainfall infiltration, and then it is used in an infinite slope approach to estimate slope stability. However, there is a still need for quantifying the hydrological processes on hillslopes and appropriate models for tropical and mountainous terrains.

Additionally, the spatial scale of available models can range from the regional scale that considers huge basins, to small area of few square meters for the analysis of single hillslopes or individual landslides. The quality and quantity of hydrologic, hydraulic and geotechnical available data, and detail of the model change considerably, and the consequence consistency and coherency of the model.

The aim of the model presented here is to explore and try to clarify the dynamics of the physical and mechanical processes leading to the development of shallow landslides triggered by rainfall in tropical and mountainous terrains, pointing out the factors and parameters that play the main roles in this occurrence. And consequently forecasting the landslide occurrence and produce maps of relative potential of shallow landsliding.

5.1 Conceptual model for landslide triggered by rainfall in tropical and mountainous terrains

Usually models show different conceptual models and assumptions. For the model proposed in this work, it is necessary to define different assumptions relating to spatial-temporal variability of hydraulic and mechanical parameters of the soil, water content into the soil, and subhorizontal flow formation, would allow permit to propose a coherent model.

Most of these considerations are based on field observation and the experience in tropical and mountainous terrains supported by hydrology and geotechnical theory for tropical residual soils.

5.2 Tropical weathering profile

To consider the mechanism for rainfall infiltration and subsurface flow in tropical environments, it is necessary to understand in detailed the processes forming tropical residual soils and weathering layers. For tropical environments hillslopes are covered with deep residual soils characterized by a wide range of physical and mechanical properties depending on their parental rocks and degree of weathering.

The susceptibility of rocks to chemical action is a function of mineralogical composition and texture as well as the presence of fractures. However, the dominant control of the mode of weathering are rainfall, mean temperature, and their variability for very short time periods (Curties, 1976; Ollier, 1988; Anon, 1995). All these variables present high values for tropical and complex terrains, explaining the presence of deep residual soils over these areas.

Weathering increase with depth causing rock material to become more porous, and hence permeability and shear strength are reduced as well. The weathering processes forming tropical residual soils include physical desintegration, chemical decomposition and biological intervention, which produce a succession of distinct horizons parallel to the slope surface called soil weathering profile (Fookes, 1997).

Numerous classifications for soil weathering profile have been adopted based on multiple purposes (Little, 1969; Deere & Patton, 1971; Anon, 1995). For this model, based on studies proposed by Little (1969) and Anon (1995) the tropical residual soil profile is divided into three basic units of particular qualities to which can be assigned engineering characteristics (Figure 7).

(i) **Residual soil**, in which all material has been converted to soil and mass structure and material fabric were destroyed, with significant change in volume increasing permeability and hydraulic conductivity. Into this unit is included the organic soil over the surface conform by just few centimeters of depth. Permeability and hydraulic conductivity is high at shallow depth. This upper soil horizon is influenced by roots vegetation, animal disturbance, and chemical processes generating macro pores structures, such as natural soil pipes or open relict joints.

(ii) **saprolite**, where the mass properties are still “soil-like”, correspond to highly to completely weathered material, in which original structure and fabric are preserve because of pseudomorphic replacement of clay minerals and a lack of subsequent disturbance or transportation. The saprolite formation is isovolumetric process and hence permeability is less than the upper soil layer,

(iii) and finally, slightly weathered to fresh **rock**, where rock-like characteristics begin to dominate and permeability is very low.

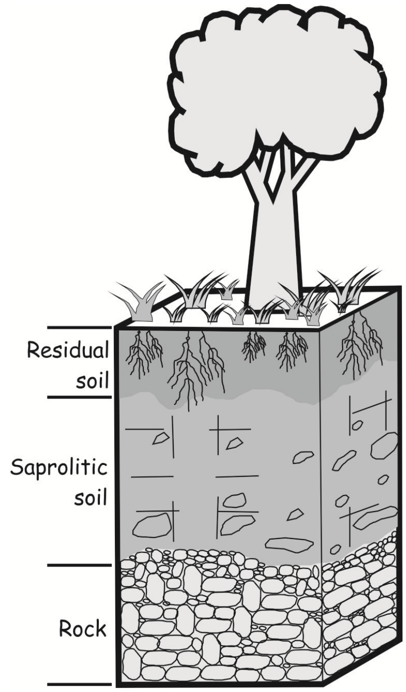

Figure 7. Typical weathering profile of tropical environments and complex terrains

Each of these basic units or layers present different and complex hydrological and geotechnical properties. However, the variability of the soil properties are not just in the weathering vertical profile. One of the most complex elements for tropical environments corresponds to the huge changes of the hydrological and mechanical properties of the soil along the hillslopes, as well as soil thickness, one of the most important factors controlling shallow landsliding. Topographic control over soil properties is particularly strong in tropical environments, reflecting the importance of lateral movement of water and soil material down-slope, as well as down-profile (Fookes, 1995). Actually some of the most important physically based models proposed by landslides triggered by rainfall used topographic indexes (Montgomery & Dietrich, 1994).

For this reason to consider lateral variability of soils along the hillslopes, it is important to define a catenas soil pattern. Milne (1935) used the term *catena* to describe the succession of soils down a slope and repeated in a pattern across the landscape. These profile differences are attributable either to downslope movement of fine soil particles and material in solution or to site differences related to slope angle and depth of water table (Fookes, 1995).

Multiple sequences of soil properties and profiles have been proposed by the authors (Ollier, 1976). Considering previous studies, which have described the typical weathering profile in tropical humid and complex terrains (Aristizábal et al., 2011), a simple catenas soil pattern has been selected to this model, shown in figure 8. It is characterized by deep profiles above and below hillslopes retaining shallow weathering profile, and soil eroded from the top of the slope tends to accumulate near the bottom (Figure 8).

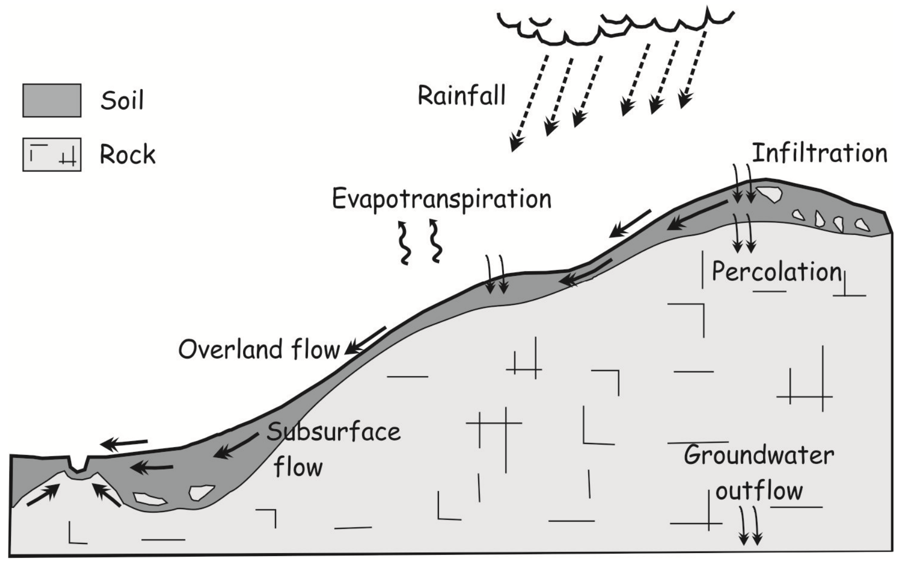

Figure 8. Catena and hillslope hydrological processes. Soil thickness distribution varies according to the slope inclination.

Under this catena soil pattern is reasonable to link soil thickness and slope angle in basin scale. The soil thickness on a hillslope, which coincides with the failure depth, is a critical parameter in performing a slope-instability analysis (Segoni et al., 2012; Ho et al., 2012). Relatively thin soils are more prone to saturated overland flows compared to thicker soils which have greater water storage potential.

Slope-derived soil thickness pattern are widely used because the easy and quick application over large areas without extensive field information available. In the catena soil pattern proposed, soil thickness values are inversely proportional to the slope gradient, according to a linear law derived from several calibration measures. In the same direction to the catenas, this soil thickness pattern relies on the assumption that on steeper slopes erosive processes are more intense. Thus, soils are shallower and erosion is weaker on flat surfaces so that deposition prevails and thicker soils are usually found (Segoni et al., 2012; Salciarini et al., 2006).

Much more complex methods that make use of multivariate statistical analyses or than employ process-based models exist for soil thickness patterns. However, they require huge effort to be correctly applied and calibrated (Segoni et al., 2012).Therefore it is necessary to keep in mind that any particular region shows different catenas, and hence they should be defined on the field and laboratory tests. Any catenas and soil thickness pattern could be integrated into the model proposed according to the information available of the area selected. For this initial approach, a simple catenas and slope-derived soil thickness pattern has been selected.

5.3 Subhorizontal flow formation

Shallow landsliding in tropical environments is related to the formation of a subsurface flow parallel to the slope.Subsurface flow, also known as interflow, lateral flow or soil water flow, is a saturated water flow phenomenon parallel to the slope inclination. According to the literature there are two recognized mechanisms for rainfall infiltration and subsurface flow formation: (i) by the rising of a perched water table from the potential failure surface or (ii) by the development of an advancing wetting front from the slope surface.

According to the soil profile formation, hillslopes for tropical environments are heterogeneous, then hydraulic conductivity and permeability varies with depth, controlling rainfall infiltration, subsurface flow formation and location of shear surface. This permeability and hydraulic conductivity contrasting with the underlying horizon forces the starting of a perched water table, and then positive pore water pressure caused by subsurface flow will occur near the ground surface. For tropical conditions in most cases the permeability contrasting is located to the weathering profile change of residual soils to saprolite.

According to literature and based on laboratory and field tests carried out by multiple investigations, the contrasting of physical-mechanical properties, where the perched water table is formed, corresponds at the same time to the failure surface. This assumption is imposed to the model.

5.4 Water storage capacity

The model needs to know the amount of water present in the soil and how much water could storage. Soil water storage capacity depends on the open spaces or pores found within the soil. It means porosity determines the total amount of water a soil will hold, and it varies from one soil to another.

When water falls on the land surface through precipitation, some of it runs off the land into pounds and streams, and some infiltrate into the ground. Water infiltrated first goes through an unsaturated zone in the soil, and some of the voids between the soil particles are filled with water. While the pull of gravity tends to draw the water downward, soils in the unsaturated zone are able to hold some water in the smaller voids because of surface tension effects as a film around the soil particles. This water is called hygroscopic water. If gravity exerts a force sufficient to exceed that of the surface tensions, the excedent water will flow downward. This water is called gravitational or free water.

Water availability is illustrated in the Figure 9 according to the soil water content. Free or gravitational water drains quickly from the soil after rainfall because of gravitational forces (saturation point to field capacity). Available water is retained in the soil after the excess has drained (field capacity to wilting point).

According to this, saturation (Ws) occurs when all the voids in the soil are completely filled with water.

Field capacity (Wfc) means the water content of the soil where all free water has been drained from the soil through gravity. It is the maximum value of water content that can be maintained without the water draining rapidly. Excess water over the field capacity usually drains, according to the soil permeability, within one or several days back to the field capacity.

Permanent wilting point ($Wpwp$) occurs when the volumetric water content is too low for the plant to remove water from the soil and will wilt and die. Some authors consider the permanent wilting point when the water content of the soil is at -1.5 MPa water potential. It is important to differentiate, that in this case soil is not dry, water is still be present in the soil, but plants are unable to access it.

The difference between field capacity and wilting point is termed the plant available water storage. And the difference between saturation and field capacity is called drainable porosity.

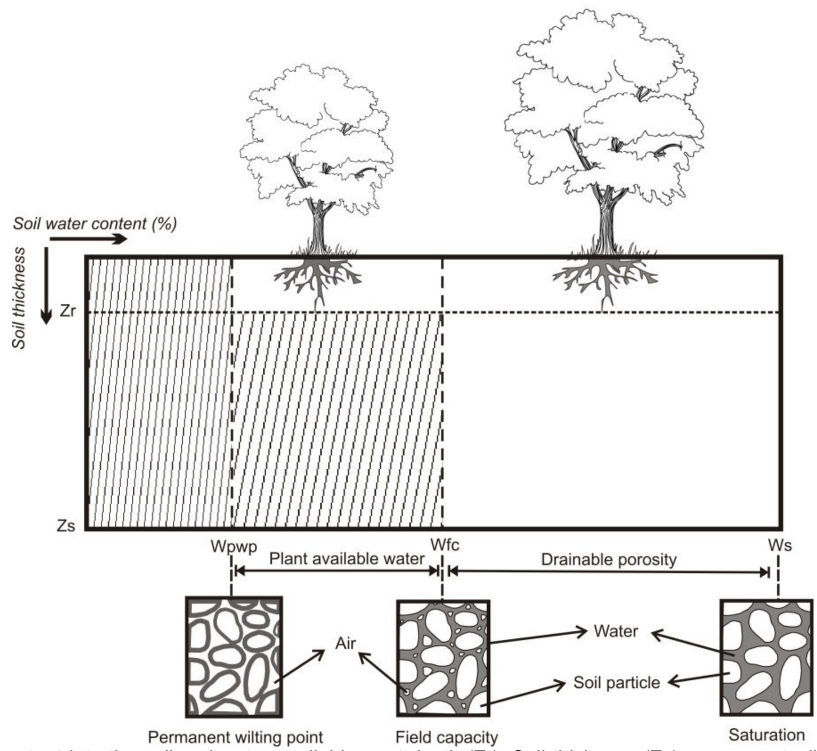

Figure 9. Water content into the soil and water available. root depth (Zr), Soil thickness (Zs), permanent wilting point (Wpwp), field capacity (Wfc), Saturation (Ws)

5.5 Water content in the soil

Relating to the water content in the soil, shallow landslide modeling considers the soil properties under saturated or partially saturated conditions.

During the last years, matrix suction and hydraulic conductivity as a function on saturation degree has been incorporated for modeling soils partially saturated. For unsaturated condition shallow landslides are caused by the corresponding suction reduction and the resulting slope movement is in the form of a relatively rigid slab.

To get this unsaturated soil condition, it is necessary previous long periods without rainfall, and the unsaturated condition is preserved just at the beginning of the rainfall event. During the initial rainfall period, unsaturated condition control the soil mechanic and shallow landslides occur by suction reduction, but after a short period of rainfall saturated conditions appear controlling shallow landslide occurrence.

For tropical climate conditions, rainfall is a common phenomenon, conserving most of the time the entire soil profile wet to field capacity and shallow landslides occur mostly under saturated or near saturated conditions during wet periods or after intense rainfall by the positive pore pressures. Reason for the resulting shallow landslides are in the form of a liquefied soil mass, such as soils slips or mud/debris flows, where the displaced material of shallow landslides changes suddenly to a flow.

To fulfill this consideration the model include an static storage tank, which corresponds to the plant available water, to receive the initial rainfall, and just when this tank is full, infiltration occurs under saturation condition of the upper soil layer.

5.6 Rainfall

Relating to rainfall conditions, it is well known antecedent rainfall play an important role for starting time to landslide occurrence, and during the storm the rainfall intensity and duration is a fundamental factor.

In this sense the model integrate both, the initial conditions and rainfall variation during the storm.

5.7 Vegetation

The effect of vegetation on slope stability may broadly be classified as either hydrological or mechanical, with positive and negative effects on slope stability. Mechanical effects, such as root reinforcement, increase soil properties of the soil but at the same time surcharge and wind loading increase forces transmitted to the soil reducing stability. On hydrological effects, evapotranspiration affect positively slope stability, but at the same time roots increase secondary hydraulic conductivity and infiltration.

Considering this complex influence on both sides of the slope stability equation, vegetation is not directly considered into the model. However the influence of vegetation on cohesion could be include indirectly by increasing the values of theses parameters. To evaluate hillslopes with similar geology and residual soil, but different soil cover, this procedure could improve the coherency and final results of the model.

5.8 Spatial and temporal scale

Another important element for the model is the spatial scale. The catchment inputs change in space, and traditional lumped models cannot reproduce any spatial variability (Frances et al., 2007).

Generally detailed models for slope scale has to include very complex and detailed physical process, which increase the running time and needs for a considerable number of soils and hydraulic parameters. In contrast, for extend areas the models have to oversimplify the natural process involved for landslide occurrence, loosing fundamental elements on this kind of processes.

Increasing this complexity, tropical soils are not just characterized by changes in function of space; they also change in function of time.

To consider all the soils variations under tropical conditions is necessary to propose a model that has to be supported on a robust hydrological component considering the spatial and temporal scale of the processes. As a consequence, the model to propose consider this important factor implementing distributed modeling that could consider this variation in space and time, dividing the catchment into uniform grid cells. Distributed models are the most suitable way to reproduce organization and randomness associated to spatial heterogeneity (Frances et al., 2007).

The model proposed is focused for middle scale catchment, where the most important processes are considered: antecedent rainfall and pattern, water infiltration, slope configuration, subsurface flow formation, and hydrological and soils properties.

6. THE MODEL: SHIA_LANDSLIDE

6.1 Hydrological module

The hydrological module to be implemented for the development of the model is based on the Open and Distributed Hydrological Simulation (SHIA), methodology developed by Velez (2001). It is formed by two sub-models representing the production and the translation of the overland runoff and subsurface flow. Initially the overland, subsurface and base flows are defined by a 3D mesh of connected tanks which drain toward the corresponding tank in the downstream cell, following the surface flow directions until it reaches the channel network.

This hydrological module due to its open source, was adjust to the needs and specific conditions for landslide triggered by rainfall in tropical environments. A more detailed description of the SHIA could be found in Velez (2001), Velez et al. (2004) and Frances et al. (2007).

For the computational program the catchment is divided into regular grid cells horizontally layered to control the infiltration and percolation processes. According to the tropical weathering profile assumed, each grid cells is formed by three layers with three different saturated hydraulic conductivities: (Ks) residual soil, (Kp) saprolite, and (Kpp) rock, where the residual soil is more permeable than the underlying saprolite soil and the impermeable rock.

Spatial interpolation of rainfall data is based on a Delaunay Triangulation method proposed by Velasquez (2011), using the incremental algorithm developed by Watson, where the rainfall stations are used as the vertices of the triangles that represent a three dimensional plane of the rainfall.

The centre of each grid cell forms the computational point. The hydrology module simulates flow over and through the discrete catchment by moving water between adjacent cells and soil layers in the horizontal and vertical direction respectively (Figure. 10).

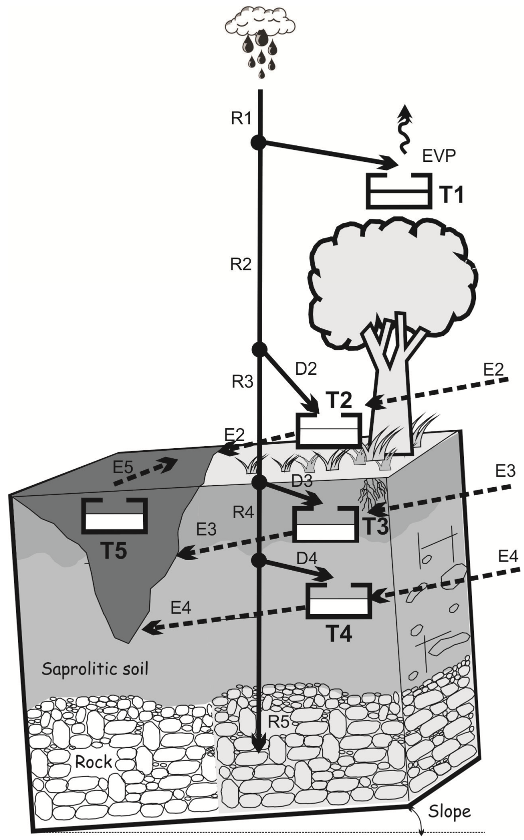

Figure 10. Hydrological conceptual model. Static storage (T1),surface storage (T2), Gravitational storage (T3), aquifer (T4), channel (T5), rainfall (R1), excedence (R2), Infiltration (R3), Percolation (R4), groundwater outflow (R5), overland flow (E2), subsurface flow (E3), base flow (E4), stream flow (E5), inflow to the tanks (D1:5), and evapotranspiration (EVP).

Each grid cell correspond to a system of five interconnected tanks and communicated with the respective tanks in the downstream cell, that represent the water flow and storage as a hydrological response unit, including the following hydrological processes: interception, detention, infiltration, evapotranspiration, overland runoff, percolation, subsurface flow, and return base flow in the channels of the drainage system.

Initially, the model determines the portion of the rainfall that is intercepted on the vegetation of the basin and as capillary water entering the ground. Then estimated the portion of the rainfall can infiltrate as gravitational and left as overland runoff. The model considers the gravitational water storage in the soil divided into two parts: one of them in the residual soil whit a higher permeability and the second one correspond to the saprolite with lower permeability and slower response.

The first four tanks represent the runoff production processes of the basin, while the last tank represents the transfer process runoff thereof, as follow:

The first tank (T1) is called *static storage* and represents the interception and water detention in puddles and the capillarity water storage in the upper part of the soil. Capillarity storage is the water retained by capillary forces in the soil rooting zone which is a function of the field capacity and effective root depth (Figure 11).

This tank models the water that passes through the catchment without participation in the process of horizontal transfer or runoff. According to the saturation assumptions discussed before, the rainfall (R1) is stored first in the static storage, until the maximum capacity is reached. The amount of water that gets into the static storage during a timestep depends on the maximum capacity of T1 (S_{1max}), type of soil and moisture content of background, in the following way:

$$D_1 = Min\left\{R_1\left[1-\left(\frac{S_1^*}{S_{1max}}\right)^2\right], S_{1max} - S_1^*\right\}$$

Where S^*_1 is the volume of water in T1 at the end of the previous timestep. When the volume of water in T1 is increasing, the content of water that could get into T1 is decreasing. The maximum volume of water that could get into T1 is when the tank is empty; it means the soil is completely dry. The excedence water that continues to the next tanks increase when T1 is almost full or wet.

The S_{1max} is equal to the sum of plant available water storagein the soil (Fig. 9). This value could be increase according to the ability of the surface coverage for storing water.

$$S_{1max} = (W_{fc} \quad W_{pwp})Z_r$$

The excedence water of the static storage that goes to the Tank 2 is:

$$R_2 = R_1 - D_1$$

The volume of water in T1 is updating for each timestep considering the maximum capacity, S_{1max}, in the following way:

$$S_1 = Min\,(S_1^* + R_1 - R_2, S_{1max})$$

The only outflow from this storage is the evapotranspiration (E_1). It has been included in the model as a function of the available water (S_{1max}), and the potential evapotranspiration:

$$E_1 = Min\left\{Evp * \left(\frac{S_1}{S_{1max}}\right)^{0.6}, S_1\right\}$$

Where Evp is the potential evapotranspiration defined by a parameter according to the local area.

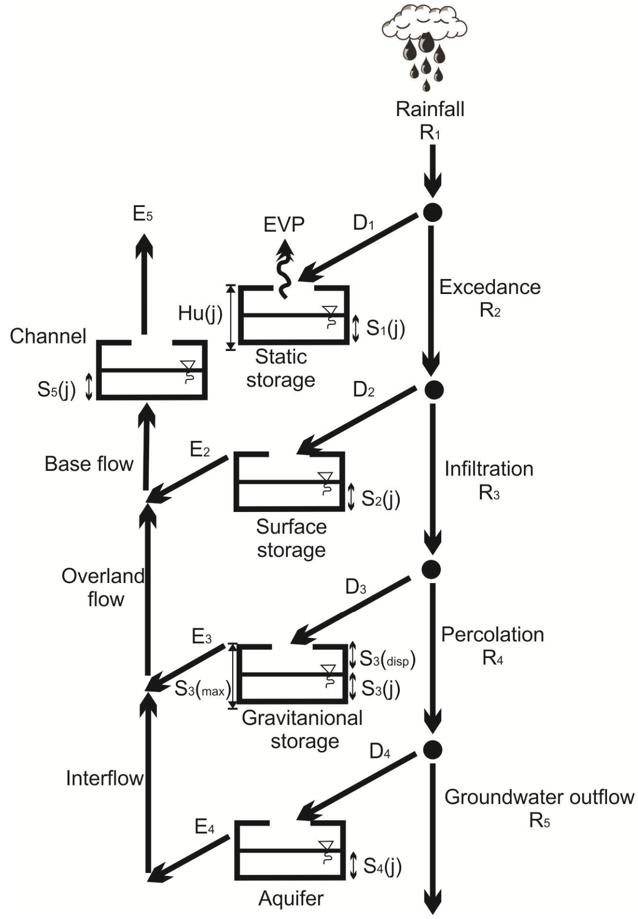

Figure 11. Hydrological module proposed

The second tank (T2) is called the *surface storage* and represents the water on the hillslope surface flowing over the slope, which it is not infiltrated. After ponding of the static storage tank (T1), the infiltration capacity can be approximated by the upper soil saturated hydraulic conductivity (Figure 11). Then the amount of water that continues and infiltrate into the soil is:

$$R_3 = Min\left(R_2,\ K_s,\ S_{3disp}\right)$$

Where Ks is the saturated permeability of the upper layer and, S_{3disp} is the available water volume of T3. It means infiltration is control by the saturated permeability and the capacity of the residual soil to receive more water. It is important to consider this saturated permeability should consider macro pore structures.

The volume of water that goes to surface storage during a time interval is:

$$D_2 = R_2 - R_3$$

The volume of water in T2 is updated for each timestep in the following way:

$$S_2 = S_2^* + D_2$$

Where S^*_2 is the volume of water in T2 at the end of the previous timestep.

The outflows of this tank, which are transferred to the neighbor cell, is function of the water velocity, the section area of the flow and the time interval. The section area is function of the storage water and the grid cell longitude, as well. The overland flow at each cell can be represented in different ways. For this case it is assume a not linear reservoir equation (Vélez, 2001):

$$V_2 = \frac{\xi A^{(2/3)e_1} \beta^{1/2}}{n}$$

Where ξ and e_1 are parameters associated to surface type. For flows over natural terrains Parsons et al. (1994) recommend values of 0.038 and 0.315, respectively. n is the manning coefficient and β the slope angle. This equation is a not linear equation due to the velocity is function of the transversal section area of the flow, and the transversal section area (A) change with the velocity and outflows.

$$A = \frac{S_2}{dx + v\,dt}$$

Then, it is assumed a value for the initial velocity, to calculate the area, and return to the equation to calculate the velocity, and obtain a mean velocity. This process is repeated three times for each interval of time, looking for a convergent velocity value.

$$v_{mean} = \frac{2v_{cal} + v_{initial}}{3}$$

Finally, the outflows of this tank to the downstream cell according to the overland flow velocity is:

$$E_2 = V_2 S_2 \frac{dt}{dx}$$

Finally, the volume of water in T2 is updating considering the outflows during this time interval:

$$S_2 = S_2 - E_2$$

The third tank (T3) represents the *gravitational water storage* in the residual soil between field capacity and saturation (Figure 11). It models the water column due to subsurface flow parallel to the slope surface, through the soil layer and into the drainage system. This tank corresponds to the residual soil, where the conductivity is considered by the model as saturated and vertically constant. One small portion of the water can percolate or flow toward the saprolite, according to the saprolite permeability (K_p), to feed the subsurface flow.

The volume of water that percolate to the saprolite is:

$$R_4 = Min\left(R_3, K_p\right)$$

The volume of water that goes into T3 during a time interval is:

$$D_3 = R_3 - R_4$$

The volume of water in T3 is updated for each timestep in the following way:

$$S_3 = S_3^* + D_3$$

Where S^*_3 is the volume of water in T3 at the end of the previous timestep.

Similar to the T2 the outflows from this storage to the neighbor grid cell is function of the section area of the storage water in the tank, the velocity and the time interval. In this case the subsurface flow velocity is estimated according to Kubota & Sivapalan (1995) as a lateral subsurface flow in mountains terrains covered by forests:

$$V_3 = \frac{K_s \sin \beta}{nA_{celda}(b+1)(S_{3max})^b}(S_3)^{b+1}$$

Where Ks is the saturated permeability, β is the slope angle, n the manning coefficient, and b is a parameter that depends of the soil type. Kubota & Sivapalan (1995) used b = 2 for a mountain basin covered by forests.

Similar to the T2, the velocity equation is not a linear because subsurface flow velocity is function of the storage water, and at the same time the storage water is function of the velocity and water that flow out for this tank. Therefore it is used the algorithm of the T2.
The volume of water that flow out from T3 to the downstream cell according to the subsurface flow velocity is:

$$E_3 = V_3 S_3 \frac{dt}{dx}$$

Finally, the volume of water in T3 is updating considering the flow out during this time interval:

$$S_3 = S_3 - E_3$$

And the volume of water that could get into T3 during the next time interval is:

$$S_{3disp} = S_{3max} - S_3$$

According to the figure 9, the maximum capacity of T3 is the drainable porosity:

$$S_{3max} = (W_s - W_{fC})Z_s$$

The fourth tank (T4) corresponds to the aquifer, where the vertical flow represents the system groundwater outflow and the horizontal flow is the base flow. This tank models the flow and storage in the aquifer (Figure 11). The model took into account that part of the water entering into the aquifer is not incorporated into the base flow of the basin, although in most basins this amount is very small and could be depreciate. The volume of groundwater outflow is:

$$R_5 = Min\left(R_4, K_{pp}\right)$$

Where Kpp is the groundwater outflow, which could be understand as water losses. The volume of water that goes into T3 during a time interval is:

$$D_4 = R_4 - R_5$$

The volume of water in T4 is updated for each timestep in the following way:

$$S_4 = S_4^* + D_4$$

The outflows from this storage to the downstream cell is estimated considering a linear equation in terms of the water level with a discharge coefficient that can be related to the aquifer saturated hydraulic conductivity:

$$E_4 = \left[1 - \frac{dx}{(K_{pp}dt + dx)}\right] S_4$$

This is the only flowouts from the tanks that is not consider in this model, because of this tank is not sensible to shallow landsliding and moves very low volume of water.

Finally, the volume of water in T4 is updating considering the ouflows during this time interval:

$$S_4 - S_4 - E_4$$

Finally, **the last tank (T5)** represents the streamflow channel at the cell, where each cell is connected to the downstream cell according to the drainage network, and models the flow of water in the drainage basin. Only the ephemeral and perennial channels grid cells content T5; slope grid cells do not have T5. It means the propagation to the outlet of the overland, subsurface flow and base flow is collected by the river channel network represented by T5 (Figure 11).

The routing along the channel network was carried out a non-stationary velocity using the Geomorphological Kinematic Wave (GCW) proposed by Velez (2001). The GCW is a simplification of the Saint Venant equations where inertial and pressure terms are neglected. Assuming prismatic canals with constant section along the reach, the discrete continuity equation and be expressed in terms of the two unknowns, the water velocity (V_t) and the cross section (A_t) as:

$$A_t \Delta x + V_t A_t \Delta t = I_t + S_{t-1}$$

Where *S* represents the volume of water in the channel reach and I_t is the total input flows from connected hillslopes (overland flow, subsurface and/or base flow) and/or upstream flow from river channels. The GCW simplification assumes the energy line slope is equal to the slope of the river bed (β). Then, flow velocity and flow cross section can be directly related by the Manning' s equation. The water velocity according to the Manning equation is expressed in terms the flow section top width (*W*), which is a function of the section (A_t).

$$v_4 = \frac{1}{n}\left(\frac{A}{W}\right)^{2/3} \beta^{1/2}$$

The velocity is controlled by the channel hydraulic characteristics (geometry and slope) at each reach and time step. The slope for each cell can be computed easily from a DEM. But unfortunately, in practice it is not economically feasible to measure the channel geometry for all cells. The GKW uses the Leopold & Maddock (1953) correlation, which relate the cross section geometry and velocity to the river discharge (Q) using potential equations.

In order to estimate the roughness the GKW propose a general equation in terms of the slope (β), accumulated area (Λ) and the height of the water (h):

$$n = \Omega \Lambda^{\sigma_1} h^{\sigma_2} \beta^{\sigma_3}$$

Where the coefficient Ω and the exponents σ_1, σ_2 and σ_3 are constant regional parameters.

In this way, finally the velocity of the water in the channel is function of the geometry of the channel and the geomorphology of the terrains:

$$V_4 = \frac{A^{w_2}\beta^{w_3}\Lambda^{w_4}}{(\Omega K)^{w_1}}$$

According to Frances et al. (2007) and Montoya (2008) the coefficient K and potential values w_1, w_2, w_3, and w_4, correspond to:

$$K = \left(c_1 k_0^{(\alpha_1-\alpha_2)}\right)^{(2/3-\sigma_2)}$$

$$w_1 = \frac{1}{1+\alpha_2\left({}^2/_3-\sigma_2\right)}$$

$$w_2 = (2/3-\sigma_2)(1-\alpha_2)w_1$$

$$w_3 = \left({}^1/_2-\sigma_3\right)w_1$$

$$w_4 = \{\varphi\left({}^2/_3-\sigma_2\right)(\alpha_2-\alpha_1)+\sigma_1\}w_1$$

For the sake of parsimony it is assumed that $\sigma_1 = 0.0$ and $\sigma_2 = \sigma_3$. The result is that the GKW needs six independent exponents and coefficients, which can be obtained with a geomorphologic regional study for hydrological homogeneous zones. However empirical studies have been carried out for multiple authors proposing different values according to the local conditions (Velez, 2001, Frances et al., 2007).

Table 4 shows the parameter values adopted for the modeling tropical and complex terrains.

Table 4. Geomorphological Kinematic Wave parameter ranges proposed for the model

Propagation parameter	Range
K	0.5 -0.75
φ	0.65 - 0.8
σ_1	0
σ_2	$\sigma_2 = \sigma_3$
σ_3	$\sigma_2 = \sigma_3$
α_1	0.34 - 0.55
α_2	0.05 - 0.2
C_1	.5 - 5.75
Ω	-

Similar to the previous tanks, the velocity is develop using a similar algorithm, because is function of the storage water in the tank.

The vertical connections between tanks describe the rainfall, evapotranspiration, infiltration and percolation processes. Simultaneously, the model considers the horizontal transfer of water between adjacent cells by using a sub-model to infer the direction of flow between them based on the topology of the basin. The horizontal connections describe the overland flow, interflow and base flow.

Tank interconnection depends of the grid cell type. There are three types of grid cell: (i) slopes, (ii) rill or ephemeral channels, and (iii) perennial channels (Figure 12).

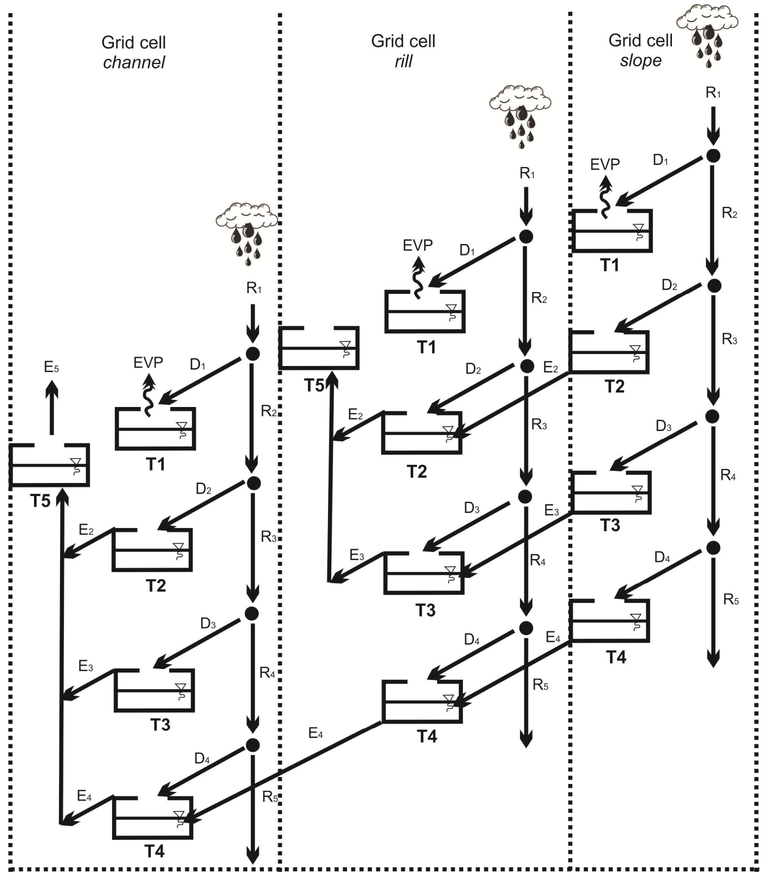

Figure 12. Interconnection tanks of the hydrological module

The grid cell type is assigned for the model according to the accumulated area or cells (threshold area). These thresholds are defined considering the field work observation and local studies. There are two threshold that should be defined and input to the models like parameters. The minimum accumulated area to form a rill or ephemeral channel, between overland runoff and subsurface flow, and the minimum area to be formed a perennial channel, between subsurface flow and base flow. The threshold for base flow is estimated based on the starting point of permanent flow in the channel network. According to these threshold area values the model defines the type of grid cell to the whole catchment.

For slope grid cells the horizontal flow of water between tanks becomes the same level, the water tank E2 passes to the tank E2 cell downstream, and similarly for the other tanks in which it is possible to transfer. Only T1 does not transfer water to the similar tanks because the only outflows from this tank is evapotranspiration. For ephemeral channel grid cells the horizontal flow of water occurs just between tanks T4, the outflows from T2 and T3 goes out to the tank 5. And finally for the perennial channel grid cells, the flow occurs just from T5 to T5.

A slope grid cell can drainage to any kind of grid cell, and rill grid cell can drainage to an rill or perennial channel grid cell, and a channel grid cell can only drainage just to a similar grid cell.

Finally, the model carries out a mass balance between rainfall input, infiltration, and runoff over the entire grid by allowing excess water to flow downslope cells, and the balance is performed to update the cumulative volume in each of the tanks.

6.2 Geotechnical module

The geotechnical module propose here is based on the idea that weathering soil profile increases density and decreasing hydraulic conductivity with depth and, when the rainfall rate exceeds the percolation rate between the residual soil and saprolite, a perched water table starts to be formed.

The implicit assumption is that soil is on saturated conditions and the subsurface flow in this saturated zone is roughly parallel to the slope. According to this, the stability conditions associated to the positive pore water pressure is constrained by the height of the perched water table.

On hillslopes mantled by tropical soils, the potential failure surface typically lies o near the contact between the relatively permeable residual soil and underling relatively impermeable saprolite. If the residual soil has limited thickness compared with the length of the slope, an infinite slope stability hypothesis can be assumed in the analysis.

The term infinite slope is used to designate a uniform slope of an extent large enough that a typical element can be considered representative of the slope as a whole, and irregularities at the toe and the crest of the slide can be ignored, where the soil properties and porewater pressures at any given distance below the ground surface are assumed constant (Graham, 1984).

The one-dimensional infinite-slope stability analysis is the most common approach in order to model the slope failure within a distributed catchment scale framework. It is based on a simplified landslide geometry that assumes a planar slip surface on an infinitely extended planar slope, both laterally and distally. The analysis assumes the slip surface is parallel to the ground surface and coincident with the impermeable substrate.

The soil is subject to two major opposition influences: the downslope component of soil weight, which acts to shear the soil along a potential failure plane parallel to the hillslope; and the resistance of the soil to shearing. The relationship between the two influences is expressed as a factor of safety. The factor of safety for an infinite soil slope can be expressed as:

$$FS = \frac{\text{Resisting Forces}}{\text{Driving Forces}}$$

The side forces for any vertical slice are equal and opposite, and the stress conditions are the same at any point of the failure surface. It is also assumed that the rigid perfectly plastic rheological model holds for the soil, that is, there is null strain until failure and shear strength is constant after the failure independently on strain. Typically the higher the pore water pressure, the lower are the frictional resistance and the shear strength; increased soil water content also increases the bulk weight of the soil (Figure 13).

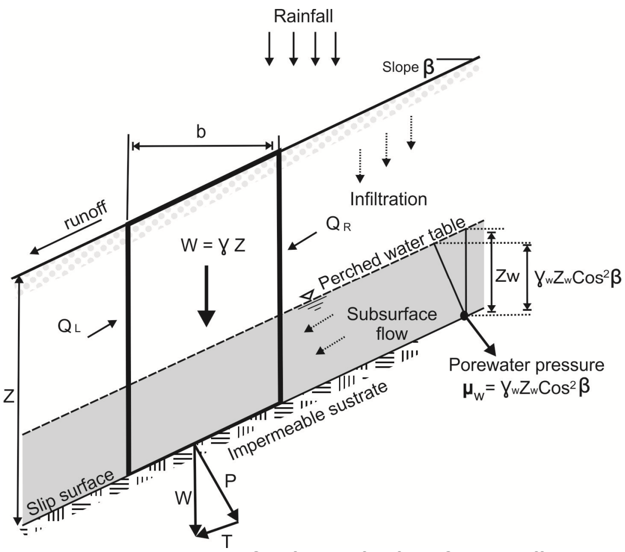

Figure 13. Geotechnical conceptual model proposed. FS = factor of safety, C is the effective cohesion, g is the gravitational acceleration, is the soil bulk density, is the water density, is the saturated soil thickness above the slip surface, Z, is the soil thickness measured vertically, β is the gradient of the hillslope.

From a consideration of the equilibrium of the slice by a vertical resolution of forces, the vertical force across the base of the slice must equal the weight (W). This can be resolved into its normal and tangential components P and T respectively.

$$W = \gamma b Z$$

$$P = \gamma b Z Cos\beta$$

$$T = \gamma b Z Sin\beta$$

The length of the slide surface is $bSec\alpha$, then the average normal and shear stresses produced by P and T are:

$$\sigma_n = \gamma Z\, Cos^2\beta$$

$$\tau = \gamma Z Sin\beta Cos\beta$$

The shear strength of the soil along the potential failure plane is given by the Mohr-Coulomb failure criterion, and the downslope shear stress (ι) must not exceed the shear strength (τ_f) of the clay.

$$\tau_f = C + \sigma_n Tan\phi$$

The safety factor in the slope can be defined in terms of effective stresses by $\frac{\tau_f}{\tau}$, that is:

$$FS = \frac{C + (\Upsilon Z Cos^2\beta - u)\tan\phi}{\Upsilon Z \sin\beta \cos\beta}$$

When a slope is subjected to pore pressure increase due to infiltration or rising perched water table, total stresses and shear stresses remain essentially constant but effective stresses, mean effective stress in particular, decrease. The effective stress principle states that the total stresses applied to soils are supported by the sum of effective interparticle stresses and neutral pore water pressure (Graham, 1984).

According to Graham (1984) in natural hillslopes with steady subsurface flow parallel to the slope, and the perched water level at distance Z_w above the slide surface, the pore water pressure is $\mu = \gamma_w Z_w Cos^2\beta$, and therefore:

$$FS = \frac{C + (\gamma - Z_w\gamma_w) Z Cos^2\beta Tan\emptyset}{\gamma Z Sin\beta Cos\beta}$$

Then, limit equilibrium condition for the slope occur when:

$$C + (\gamma - Z_w\gamma_w) Z Cos^2\beta Tan\emptyset = \gamma Z Sin\beta Cos\beta$$

This equation solved for Z_w provides the critical value of landslide-triggering saturated depth:

$$Z_{wcrit} = \frac{\gamma}{\gamma_w} Z\left(1 - \frac{Tan\beta}{Tan\emptyset}\right) + \frac{C}{\gamma_w Cos^2\beta Tan\emptyset}$$

To evaluate the slope stability for each grid cell in every timestep, it is necessary to get the perched water table height and compare with the critical value of landslide-triggering saturated depth (Z_{wcrit}). The hydrological component of the model provides the water content in the gravitational storage, this value has to be transform considering the water content of the soil. Then, according to the Figure 9, the perched water table height is:

$$Z_w = \frac{S_3}{(w_s - w_{fc})}$$

However to increase the velocity and efficiency of the model, it is previously evaluated the minimum and maximum residual soil thickness to define the unconditional stable grid cells and the unconditional unstable grid cells (Figure 14).

The residual soil thickness for which $Z_w - Z$ is defined as immunity depth :

$$Z_{min} = \frac{C}{\gamma_w Cos^2\beta Tan\emptyset + \gamma Cos^2\beta (Tan\beta - Tan\emptyset)}$$

Because saturated depth is necessarily smaller than residual soil thickness ($Z_w \leq Z$), when $Z < Z_{min}$ the deposit is always stable independently of rainfall.

And for a certain value of soil thickness, Z_{max}, the saturated depth necessary to trigger a landslide is zero and the soil is always unstable, regardless of rainfall occurrence (Iida, 1999). For the soil thickness higher than Z_{max} the soil is then always unstable. $Z_{max}$is determined by setting $Z_{min} = 0$

$$Z_{max} = \frac{C}{\gamma Cos^2\beta (Tan\beta - Tan\emptyset)}$$

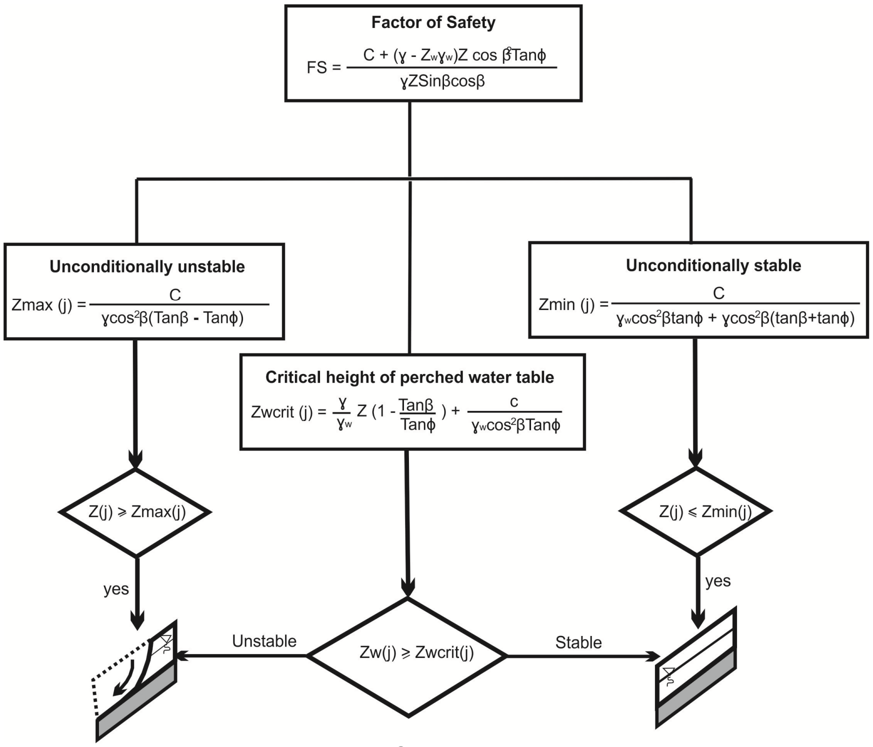

Figure 14. Geotechnical module

6.3. The coupled hydrologic – geotechnical model

SHIA_Landslide is a model program for computing positive pore pressure changes, and attendant changes in the factor of safety, due to rainfall infiltrations using an hydrological module coupled with an infinite stability slope geotechnical module.

The model is composed by a hydrological module, to analyze rainfall infiltration in saturated condition, and by a geotechnical module which, starting from limited equilibrium methods, evaluate slope stability. The model requires as input rainfall, which causes a rise of perched water table, and consequently a rise in pore pressures leading to instability conditions.

The model proposed have focused their attention on the topographic control on hydrological process, on the process that control subsurface flow at the hillslope scale and the effect of water infiltration on soil strength and slope stability.

Infiltration (R_3), gravitational storage (S_3) and subsurface flow (E3) are the most important process to care in the model. The water stored in the T3 from the hydrological module corresponds to the height of the perched water table to input in the geotechnical module.

Figure 15 shows the flow diagram showing the steps used by the model in hydrological and slope stability calculation.

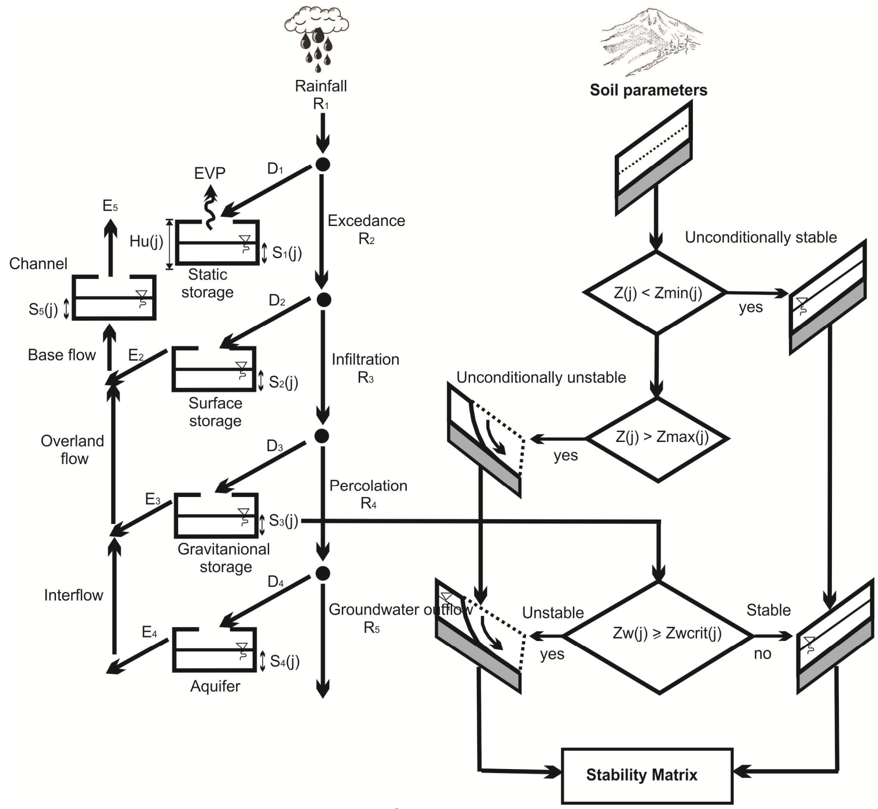

Figure 15. SHIA_Landslide model

6.4 Programming language

Relating to the language program, there are different languages to carry out this kind of model. However, due to this model correspond to a detail and distributed hydrological module combined with a geotechnical approach for the catchment scale, it was necessary to use a robust language and compiler. For this reason FORTRAN was selected as the programming language.

6.5 Subroutines

The program is divided into two sub models: SHIA_landslide_Pre and SHIA_Landslide. Each of these sub models is integrated by several subroutines. Figure 16 shows the program structure of SHIA_Landslide model.

The submodel SHIA_Landslide_Pre process input maps and parameters to prepare the information needed to runSHIA_Landslide.

SHIA_landslide_Pre creates a multiparameter matrix of the catchment and prepare an initial stability matrix classifying the grid cells into: unconditional stable, unconditional unstable and potential unstable grid cells. This step permits that SHIA_Landslide only tests the stability of the potential unstable grid cells.

SHIA_Landslide_Pre is integrated by five subroutines: Modules, Input Data, Basin, Rainfall and Matrix.

In the Module subroutine all the general variables are declared, and in the Input Data subroutine the geotechnical and hydrological parameters needed for the model are read. Subsequently, the Basin subroutine determines the position of the grid cells according to the flow direction map. The algorithm for defining flow direction is known as D8 (8 flow directions) which assign flow from each grid cell to only one of its eight possible neighbors, either adjacent or diagonally, in the direction with steepest downward slope. In the model each grid cell can receive flow from multiple neighbors but can discharge to only one neighbor grid cell.

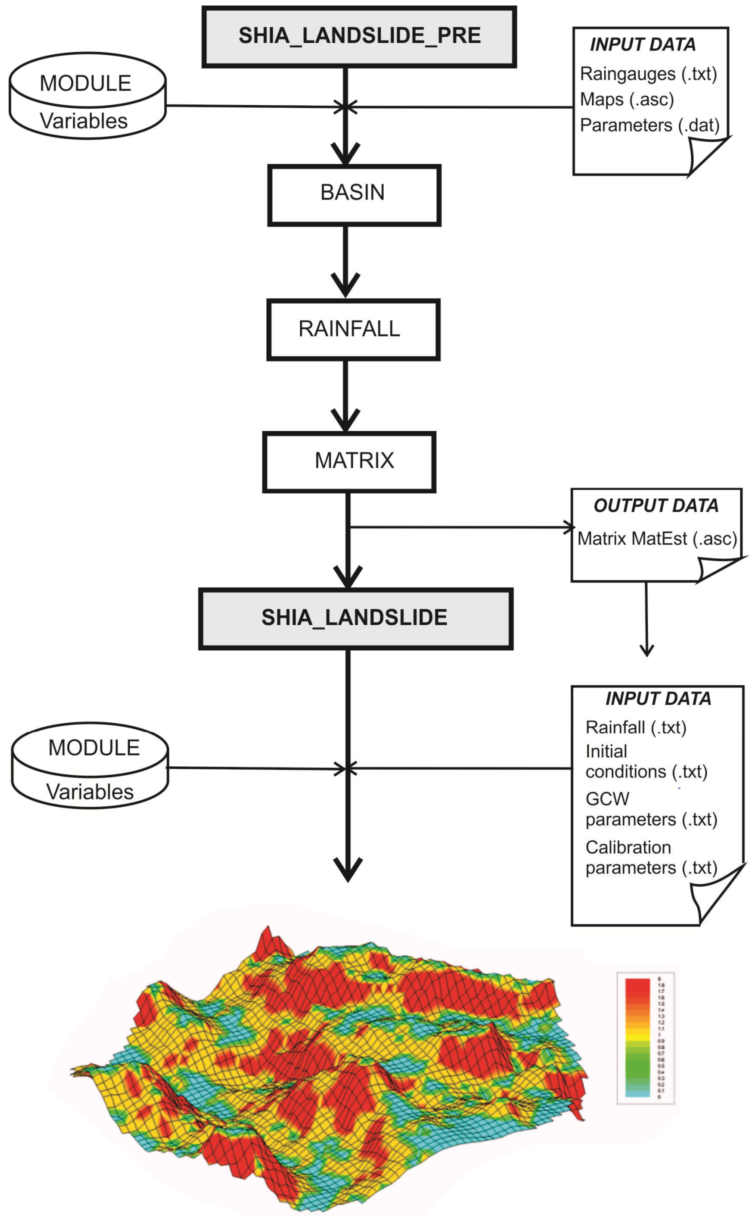

Figure 16. Flow chart of SHIA_Landslide program

The Basin subroutine use an algorithms, starting from the lowest grid cell in the catchment toward upstream, establishing the number and position of the grid cells that drain to each grid cell, and forming a vector from down to up.

And finally the Matrix subroutine conform a multiparameter matrix of the catchment to all the input parameters for each grid cell. In this step it is created an initial stability matrix. The minimum and maximum soil depth is calculated and the stability matrix is formed according to the unconditionally stable and unstable grid cells. In this subroutine also the critical height of the perched water table is calculated and included into the multiparameter matrix.

Relating to the SHIA_Landslide, it is integrated by three subroutines: Modules, Input Data, and SHIA Model.

Similar to the previous case, Modules and Input Data declares the general variables and input more parameters needed for the model during these steps. All these previous subroutines have the purpose to prepare the needed information for the final subroutine: SHIA model.

Finally the SHIA model subroutine runs the hydrological and the geotechnical components of the models.

In the hydrological component the subsurface flow is calculated according to the local drain direction imposed by the topographical relief, depending on GKW proposed by Velez (2001). Changes in height of the perched water table directly relates to a change in pore pressures, which are estimated for the stability module.

In the geotechnical component, for each time step the program evaluate the stability of the potential instable cells comparing the perched water table height (Z_w) with the critical water table height estimated previously.

6.6 Correction factors

In general, the parameter value which reproduces the average of the process in an area, called the effective parameter value, is not the mean value of the parameter within the area, and considering scale effects the models can be calibrated, but will not validate properly in a different scenario. According to this problem Frances et al. (2007) propose that the effective parameters for the model at each cell should be split in two components: the hydrological or geotechnical characteristic and a correction factor, common for all cells and taking into account all modeling errors including the temporal and spatial scale effects.

With the split-parameter structure proposed by Frances et al. (2007), the correction factors take into account the time and space scale effects and also the model and input errors, leaving the hydrological characteristics free of these problems while maintaining the physical meaning of the parameters.

The infiltration model and the flow channel routing model proposed include a few correction factors which correct globally for the different soil properties maps instead of each cell value of the calibration maps, thus reducing drastically the number of factors to be calibrated. This strategy allows for a fast and agile modification in different hydro- logical and geotechnical processes.

Table 5 shows the minimum and maximum values for each correction parameter according to Velez (2001).

Table 5. Correction parameters used for the model

PARAMETER	CORRECTION	MIN	MAX
Maximum static storage (S_{1max})	C1	0.1	1.5
Evapotranspiration (*EVP*)	C2	0.5	2.0
Infiltration capacity (K_s)	C3	0.0	1.0
Percolation capacity (K_p)	C4	0.0	2.0
Groundwater outflow capacity (K_{pp})	C5	0.0	10
Maximum gravitational storage (S_{3max})	C6	0.1	1.5
Surface runoff velocity (V_2)	C7	0.1	2.0
Subsurface velocity (V_3)	C8	1.0	1000
Base flow velocity (V_4)	C9	1.0	1000
Channel velocity (V_5)	C10	0.5	1.5
Cohesion (C)	C11	0.5	2
Friction angle ($\emptyset$)	C12	0.5	1.5
Soil thickness (Z_s)	C13	0.5	4

6.7 Input and output data

SHIA_Landslide derives its slope stability evaluation from inputs of topographic slope and specific catchment area from parameters quantifying geotechnical and hydrological properties.

The digital elevation model (DEM) constitutes a preliminary point of the modeling since direction drainage system, accumulated drainage area and slope angle of the catchment are derived from it. Every one of the parameters used is delineated on a numerical grid over the study catchment.

The program operates on a gridded elevation model of a map area and accepts the input parameters from a series of ASCII text files. Rainfall, hydraulic properties, and slope stability input parameters are allowed to vary over the grid are thus making it possible to analyze complex rainfall events over complex terrains.

The program allows the following input parameters to vary from cell to cell throughout the basin in SHIA_Landslide_Pre subroutine: cohesion (C), evapotranspiration (EVP), soil thickness (Z_s), slope inclination (β), friction angle ($\emptyset$), maximum static storage (S_{1max}), maximum gravitational water storage (S_{3max}), field capacity water content (W_{fc}), saturation water content (W_s), saturated hydraulic conductivity of the soil (K_s), saturated hydraulic conductivity of the saprolite (K_{pp}), saturated unit weight of soil (ρ_s), flow direction, and flow accumulation area. And the parameters input in *txt* files correspond to the rainfall event and several parameters. The rainfall event includes raingauges number and location, time and rainfall information. The parameters to input to the model in this step are: threshold area, the position of the lowest grid cell of the catchment and the position of the control points selected.

For the SHIA_Landslide subroutine input data are the multiparameter matrix of the catchment, the initial stability matrix created in SHIA_Landslide_Pre, the rainfall event, the parameters for the Geomorphological Kinematic Wave and the calibration parameters for the model.

Relating to the output data, the model simulates landslide activity in the form of maps where the factor of safety for conditional unstable grid cells is reported. The user can observe the change in the factor of security as the storm progresses.

The location of the sites where the landslides are triggered is therefore represented in raster maps in terms of potential unstable cells. The program saves output to a series of ASCII text files that can be imported to GIS software for display or further analysis.

Besides a spatial prediction, the dynamic section of the model also provides a temporal prediction, by means of the observation of landslide occurrence at a fixed time interval. Hence, the temporal scale of the model is defined by the duration of the simulated event, which consequently involves time increments of time.

7. CONCLUSIONS

Floods, torrential floods and floods have been a natural phenomenon often associated with the origin and evolution of the Aburrá Valley; however the affectations of these phenomena have increased substantially in recent decades due to the increasing occupation of land in vulnerable on land threatened by this phenomenon.

There is in the literature an important number of physical and statistical models to predict landslide triggered by rainfall; however most of these models do not consider the most important hydrological process and they were built for different condition to tropical environments.

The physical and conceptual model proposed here, aimed at full simulation and comprehension of the landslide triggered by rainfall in tropical and complex terrains under saturated conditions by the rising of a perched water table from the potential failure surface. The vertical soil profile is represented by two weathering layers and underlying impermeable rock. To incorporate the spatial – temporal soils variability the model proposed a spatially distributed model using a simple soil thickness pattern.

The model shows a very robust a conceptual structure, and although it is a physically based model, the numbers of parameters is low. The calibration process is carry out independently for the hydrological and the geotechnical sub model, obtaining a better performance and control of the model. Another important aspect for the calibration process is that the model permits to split the parameters into the hydrological o geotechnical characteristics and a correction factor which consider modeling errors and spatial-temporal scale effects.

The principal purpose to use the model is to understand shallow landslides triggered by rainfall in tropical and mountainous terrains and predict outputs. Although the model has been proved and runs properly, providing coherent values; it is necessary to implement the model for a real basin to evaluate their performance.

REFERENCES

Aleotti, P., 2004. A warning system for rainfall-induced shallow failures. Engineering Geology 73. Pág. 247-265.

Alcantara-Ayala, I. 2002. Geomorphology, natural hazards, vulnerability and prevention of natural disasters in developing countries. Geomorphology 47. Pág. 107-124.

Anderson S.A., Sitar N. 1995. Analysis of rainfall-induced debris flows. Ournal of Geotechnical Engineering. Pàg. 544-552.

Ángel, P., Botero, V., Sanín, M. V., 2002. Metodología para la evaluación de la susceptibilidad al deslizamiento utilizando sistemas de información geográfica. IX Congreso de Geotecnia, Medellín.

Anon. 1995. The description and classification of weathered rocks for engineering purposes. Geological Society Engineering Group Working Party Report. Quarterly Journal of Engineering Geology, 28, 207-242.

Area Metropolitana del Valle de Aburrá, AMVA. 2002. Microzonificación sísmica de los municipios del Valle de Aburrá y definición de zonas de riesgo por movimientos en masa e inundaciones en el Valle de Aburrá. Grupo de Sismología de Medellín. Informe interno.

Arias A., 2003. La diversidad del relieve y lo suelos en el altiplano de Santa Rosa de Osos (Antioquia): sus significados ambientales. Boletín Ciencias de la Tierra. No. 15. pp. 51-72.

Aristizabal, E. Gonzalez, T., Montoya, J.D., Velez, J.I., Martinez, H., Guerra, A. 2011. Analisis de umbrales empíricos de lluvia para el pronóstico de movimientos en masa en el Valle de Aburrá, Colombia. Revista Escuela de Ingenieria de Antioquia, No. 15, Pag. 95-111.

Aristizábal, E., Gómez, J., 2007. Inventario de emergencias y desastres en el Valle de Aburrá: originados por fenómenos naturales y antrópicos en el periodo 1880-2007. Revista Gestión y Ambiente, Vol. 10 No 2. Pág. 17-30.

Aristizábal, E., Roser, B., Yokota, S. 2005. Tropical chemical weathering of hillslope deposits and bedrock source in the Aburrá Valley, northern Colombian Andes.Enginnering Geology (ELSEVIER) No 81.Pág. 389-406.

Aristizabal, E., Yokota, S. 2006. Geomorfología aplicada a la ocurrencia de deslizamientos en el Valle de Aburrá. Revista Dyna 149. Pág. 5-16.

Baum R.L., Savage W.Z., Godt W. 2002. TRIGRS-a fortran program for transient rainfall infiltration and grid-based regional slope-stability analysis. Open file report 02-424 USGS.

Barredo J. I., Benavides A., Hervas H., van Westen C.J., 2000. Comparing heuristic landslide hazard assessment techniques using GIS in the Titajana basin, Gran Canaria Island, Spian. International Journal of Applied Earth Observation and Geoinformation 2, Issue 1, 9-23.

Borga, M., Dalla Fontana, G., Cazorzi F. 2002. Analysis of topographic and climatic control on rainfall-triggered shallow landsliding using a quasi-dynamic wetness index.Journal of Hydrology 268.Pág. 56-71.

Brunsden, D., Prior, D. B., 1984. Slope instability, John Wiley, New York.

Castellanos, R. 1996. Lluvias Críticas en la Evaluación de Amenazas de Eventos de Remoción de Masa. Tesis de Magister en Geotecnia - Universidad Nacional de Colombia - Departamento de Ingeniería Civil. Santafé de Bogotá. 270 p.

Castellanos, R. & González, A. 1997. Relaciones entre la Lluvia Anual y la Lluvia Crítica que Dispara Movimientos en Masa. IX Jornadas Geotécnicas de la Ingeniería Colombiana. Santa Fe de Bogotá. Pp. 4.62 – 4.70.

Caine, N., 1980. The rainfall intensity – duration control of shallow landslides and debris flows. GeografiskaAnnaler, 62A (1-2). Pág.23-27.

Chiang S-H., Chang K-T. 2009. Application of radar data to modeling rainfall-induced landslides. Geomorphology 103. Pág. 299-309.

Chica, A. 1987. Geología y geotecnia en terrenos inclinados, modelo para la elaboración de planos geologico-geotecnicos. Magazine Fac. De Minas, Universidad Nacional de Colombia 119, 27-32.

Collins B.D., Znidarcic D. 2004. Stability analyses of rainfall induced landslides. Journal of Geotechnical and Geoenvironmental Engineering.Pág. 362-371. Vol. 130, No. 4

Crosta, G., 1998. Regionalization of rainfall threshold: an aid for landslide susceptibility zonation. Enviromental Geology, 35, (2-3), 131-145.

Crosta, G., Frattini, P., 2003. Distributed modeling of shallow landslides triggered by intense rainfall. Natural Hazard and Earth System Sciences 3. Pág. 81-93.

Crosta, G., Frattini, P,. 2008. Rainfall-induced landslides and debris flows. Hydrological Processes 22, pág. 473-477.

Crozier, M., 1999. Prediction of rainfall-triggering landslides: a test of the antecedent water status model. Earth surface processes and landforms 24.Pág. 825-833.

Cruden, D.M.S., Thomson, B.D., Bomhold, J-Y., Locat, J., Evans, J.A., Heginbottom, K., Moran, D.J., Piper, R., Powell, R., Prior, D., Qugley, R.M. 1989. Landslides: extent and economic significance in Canada, en Landslides: extent and economic significance,editadoporBrabb E.E: y Harrod B.L., Proc. 28th Intl. Geol. Congr. Symp. On Landslides, Wash. D.C: Pág. 1-23.

Curtis, C.D., 1976. Chemistry of rock weathering, in E. Debyshire, editor, Geomorphology and climate: London, John Wiley & Sons, P. 25-57.

Dai, F.C., Lee C.F. 2001. Terrain-based mapping of landslide susceptibility using a geographical information system: a case study. Canadian GeotechnicalJournal 38, 911-923.

DANE – Departamento Administrativo Nacional de Estadística. 2007. Censo General de 2005. www.dane.gov.co.

Deere, D. U., Patton, F. D., 1971. Slope stability in residual soils, Fourth Panam. Conf. SMFE, San Juan, Puerto Rico, 1, 87–170, 1971.

Echeverri, O., Valencia, Y., 2004. Análisis de los deslizamientos en la Cuenca de la quebrada la Iguaná de la ciudad de Medellín a partir de la integración lluvia-pendiente-formación geológica. Dyna, Universidad Nacional de Colombia pág.33-45.

Ekanayake J.C., Phillips C.J., 1999. A model for determining thresholds for initiation of shallow landslides under near-saturated conditions in the East Coast region, New Zealand. Journal of Hydrology 38 (1): Pág. 1-28.

Finlay, P., J., Fell, R., Maguire, P., K., 1997. The relationship between the probability of landslide occurrence and rainfall.Can. Geotech. Journal 34. Pág. 811-824.

Florez, M., Molina M., Ramírez I. 1997.Método cualitativo para la determinación de los niveles de amenaza por movimientos en masa de la ciudad de Medellín, ladera occidental. Alcaldía de Medellín, 90 pp.

Florez, M., Molina M., Ramírez I. 1996. Método cualitativo para la determinación de los niveles de amenaza por movimientos en masa de la ciudad de Medellín, ladera occidental. Alcaldia de Medellín, 90 pp.

Fookes, P.G., 1997. Tropical residual soils. A geological society engineering group working party revised report. London: The Geological Society.

Francés, F. I. Vélez and J. Vélez, 2007. Split□parameter structure for the automatic calibration of distributed hydrological models. Journal of Hydrology, 332, 226□240.

Frattini, P., Crosta, G, Sosio, R. 2009. Approaches for defining thresholds and return periods for rainfall-triggered shallow landslides.HydrologicalProcesses, 23.Pag. 1444-1460.

García, H. 2004. Evaluación de la susceptibilidad a los movimientos de masa usando sistemas de información geogràfica y redes neuronales artificiales. Disertación de Maestrìa, Universidad de Brasilia, Brasilia.

Graham J., 1984. Methods of stability analysis In Slope Instability Edited by D. Brunsden and D.B. Prior. Jhon Wiley & Sons Ltda. Pag. 171-215.

Gómez, S., 1990. Predicción de niveles freáticos a partir de la precipitación y su influencia en la estabilidad de taludes (caso de aplicación en la meseta de Bucaramanga). Tesis de Maestría. Universidad Nacional de Colombia, Medellín.

Guzzetti, F., Carrara, A., Cardinali, M., Reichenbach, P., 1999. Landslide hazard evaluation: a review of current techniques and their application in a multi-scale study, Central Italy. Geomorphology 31, 181-216.

Hammond, C., D. Hall, S. Miller and P. Swetik, (1992), "Level I Stability Analysis (LISA) Documentation for Version 2.0," General Technical Report INT-285, USDA Forest Service Intermountain Research Station.

Hermelin, M., 1984. Riesgo geológico en el valle de Aburrá. I Conferencia geológica sobre Riesgos Geológicos del Valle de Aburrá. SociedadColombiana de Geología. Medellín.

Ho J.Y., Lee K.T., Chang T.C., Wang Z.Y., Liao Y.H., 2012. Influences of spatial distribution of soil thickness on shallow landslide prediction. Engineering Geology 124. Pag. 38-46.

Hormaza, M., 1991. Investigación preliminar de las causas probables de deslizamientos en las laderas de Medellín. Tesis pregrado, Universidad Nacional de Colombia, 529 pp.

Huat B.B.K., Ali F.H.J., Low T.H. 2006. Water infiltration characteristics of unsaturated soil slope and its effect on suction and stability. Geotechnical and Geological Engineering 24. Pág. 1293-1306.

Hutchinson, J.N. 1995. Keynote paper, Landslide hazard assessment. Proceeding of the 6th International Symposium on Landslides, Christchurch, New Zealand 3, 1805-1841.

Iida T., 1999. A stochastic hydro-geomorphological model for shallow landsliding due to rainstorm. Catena 34 (3-4). Pag. 293-313.

Iverson, R., 2000. Landslide triggering by rain infiltration. Water Resources Research, Vol. 36. No 7. Pág. 1897-1910.

Ingeominas, 1990. Zonificacion de aptitud del suelo para el uso urbano costado occidental de Medellín. INGEOMINAS. 91pp.

kubota, J., Sivapalan, M. (1995) Towards A Catchment-Scale Model Of Subsurface Runoff Generation Based On Synthesis Of Small-Scales Process-Based Modeling And Field Studies. En: Scale Issues in hydrological modeling. Editado por: M Sivapalan. John Wiley and sons. 297-310.

Leopold, L. B. and Maddock, T. J., 1953, Hydraulic geometry of stream channels and some physiographic implications. U. S. Geological Survey Professional Paper 252, 55 p.

Little, A.L., 1969. The engineering classification of residual soils. Seventh International Conference on Soil Mechanics and Foundation Engineering, ISSMFE. Mexico, vol. 1, pp. 1 – 10.

Londoño, J.P. 2007. Evaluación holística de riesgos frente a movimientos en masa en áreas urbanas andinas. Una propuesta metodológica. Boletín de Ciencias de la Tierra, Nº 20, Junio. Medellín.

Matsushi Y., Hattanji T., Matsukura Y. 2006. Mechanisms of shallow landslides on soil-mantled hillslopes with permeable and impermeable bedrocks in the Boso Peninsula, Japan.Geomorphology 76.Pág. 92-108.

Mayorga, M. R., 2003. Determinación de umbrales de lluvia detonantes de deslizamientos en Colombia. Tessi de posgrado Departamento de Geociencias, Universidad Nacional de Colombia, Bogotá. Pág 223.

Milne, G. 1935. Some suggested units of classification and mapping for East African soils. Soil Res. 4:183–198.

Montoya, J.J., 2008. Desarrollo de un modelo conceptual de producción, transporte y depósito de sedimentos. Phd Thesis. Universidad Politecnica de Valencia. Pag. 267.

Montgomery, D. R., Dietrich, W. E. 1994. A physically based model for the topographic control of shallow landsliding.Water Resource Research 30.Pág. 1153-1171.

Moreno, H., A., Vélez, M., V., Montoya, J., D., Rhenals, R., L., 2006. La lluvia y los deslizamientos de tierra en Antioquia: análisis de su ocurrencia en las escalas interanual, intranual y diaria. Revista Escuela de Ingeniería de Antioquia, No 5. Pág. 59-69.

O'Loughlin, E. M. 1986. Prediction of Surface Saturation Zones in Natural Catchments by Topographic Analysis, Water Resour. Res., 22, Pág. 794–804.

Ollier, C.D., 1988. Deep weathering, groundwater and climate. *Geografiska Annaler. Series A, Physical Geograph.* Vol. 70, No. 4 (1988), pp. 285-290.

Pack, R. T., D. G. Tarboton and C. N. Goodwin, (1998), "Terrain Stability Mapping with SINMAP, technical description and users guide for version 1.00," Report Number 4114-0, Terratech Consulting Ltd., Salmon Arm, B.C., Canada (www.tclbc.com).

Paniconi C., Troch P.A., van Loon E.E., Arno G., Hilberts J. 2003. Hillslope-storage Boussinesq model for subsurface flow and variable source areas along complex hillslopes: 2. Intercomparison with a three-dimensional Richards equation model. Water Resource Research 39.No 11, 1317.

Parsons, A.J., A.D. Abrahams, J. Wainwright. 1994. On determining resistance to interril overland flow. Water Resources Research, 30 (12), Pag. 3515-3521.

Paz, C. I., Torres A. M., 1989. Precipitación y su influencia sobre algunos deslizamientos ocurridos en las laderas del Valle de Aburrá. Tesis pregrado, Universidad Nacional de Colombia. 174 pp.

Qiu C., Esaki T., Xie M., Mitani Y., Wang C. 2007. Spatio-temporal estimation of shallow landslide hazard triggered by rainfall using a three-dimensional model.Environmental Geology 52.Pág. 1569-1579.

Rahardjo H., Ong T.H., Rezaur R.B., Leong E.C. 2007. Factors controlling instability of homogeneous soil slopes under rainfall.JOurna of Geotechnical and Geoenvironmental Engineering, Vol. 133, No. 12.Pág. 1532-1543.

Rendon, A., Vargas, R., 1998. Sistematización del proceso de evaluación de la amenaza por movimientos en masa, en zonas de ladera del Municipio de Medellín. Graduate Thesis, Universidad de Antioquia. 123 pp.

Rezzoug A., Schumann A., Chifflard P., Zepp H. 2005. Field measurements of soil moisture dynamics and numerical simulation using the kinematic wave approximation.Advances in Water resources 28.Pág. 917-926.

Salciarini D, Godt JW, Savage WZ, Conversini R, Baum RL, Michael JA (2006) Modelling regional initiation of rainfall-induced shallow landslides in the eastern Umbria Region of central Italy. Landslides 3:181–194.

Sharma, R., H., Nakagawa, H., 2005. Shallow landslide modeling for heavy rainfall events.Annuals of Disas. Prev. Res. Inst. Kyoto Univ. 48.

Scheidegger, A. E., 1998. Tectonic predesign of mass movements, with examples from the Chinese Himalaya.Geomorphology 26, 37-46.

Segoni, S., Rossi G., Catani F., 2012. Improving basin scale shallow landslide modelling using reliable soil thickness maps. Natural Hazards 61. Pag. 85-101.

Shlemon, R., 1979. Zonas de deslizamientos en los alrededores de Medellín, Antioquia (Colombia). Publicaciones Geológicas Especiales del Ingeominas. 45 pp.

Schuster R. L,. 1996. Socieconomic significance of landslides. In: A.K. Turner & R.L. Schuster (Eds) Landslides Investigation and Mitigation. Transportation Research Board, National Research Council, Special Report 247, National Academy Press, Washington, DC, ISA. 129-177 pp.

Schuster, R.L., Highland, M.L. 2001.Socieconomic and environmental impacts of landslides in the Western Hemisphere. Open file report 01-0276. USGS.

Setyo, A., Liao, H-J. 2008. Analysis of rainfall-induced infinite slope failure during typhoon using a hydrological-geotechnical model. Environ. Geol. DOI 10.1007/s00254-008-1215-2.

Sidle R.C., Ochiai H. 2006. Landslides: processes, prediction, and land use. Water Resources Monograph 18. American Geophysical Union, Washington D.C.

Sistema Municipal de Prevención y Atención de Desastres SIMPAD. 1999. Instrumentación y microzonificación sísmica del área urbana de Medellín. Municipio de Medellín – Sistema Municipal de Prevención y Atención de desastres, 135 pp.

Suarez D. J. 2008. Árbol de decisiones para la predicción y alera de deslizamientos activados por lluvias. XII Congreso Colombiano de Geotecnia. Bogotá. 6 pp.

Terlien, M.T.J., 1997. Hydrological landslide triggering in ash-covered slopes of Manizales (Colombia).Geomorphology 20.Pág. 165-175.

Terlien, M. T. J., 1998. The determination of statistical and deterministic hydrological landslide-triggering thresholds. Environmental Geology 35 (2-3). Pág. 124-130.

Troch P., van Loon E., Hilberts A. 2002. Analytical solutions to a hillslope-storage kinematic wave equation for subsurface flow. Advance Water Resource 25. Pág. 637-649.

Troch P.A., Paniconi C., van Loon E.E. 2003. Hillslope-storage Boussinesq model for subsurface flow and variable source areas along complex hillslopes: 1. Formulation and characteristic response. Water Res Res 39. No 11, 1316.

United Nations Program for Development, UNDP, 1998. La Amenaza y la vulnerabilidad en el análisis de riesgos. La microcuenca de la quebrada La Iguaná. Alcaldía de Medellín – PNUD. 175 pp.

UNFP (United Nation Population Fund), 2007.Estado de la Población Mundial 2007: liberar el potencial del crecimiento urbano. Informe de la UNPF, Pag. 108.

Universidad Nacional de Colombia, UNal, Escuela del Hábitat, 2009. Mapas de amenaza, vulnerabilidad y riesgo por movimientos en masa, inundaciones ya venidas torrenciales en el Valle de Aburrá. Área Metropolitana del Valle de Aburrá, Informe interno.

Van Asch Th.W.J., Van Beek L.P.H., Bogaard, T.A. 2009. The diversity in hydrological triggering systems of landslides. Proceedings of the 1st Italian Workshop on Landslides, Picarelli L., Tommasi P., Urciuoli G. & Versace P. (eds.) Rainfall-Induced Landslides: mechanisms, monitoring techniques and nowcasting models for early warning systems. Naples, 8-10 June 2009, Vol. 1.

Van Beek L.P.H., van Asch TH.W.J. 2004. Regional assessment of the effects of land-use change on landslide hazard by means of physically based modeling. Natural Hazards 31: pág. 289-304.

Van Westen, C.J., van Duren, I., Kruse, H.M.G., Terlien, M.T.J. 1994. GISSIZ: training package for Geographic Information System in slope instability zonation. ITC Publication, No. 5 Vol. 1, Pág. 359.

Van Westen, C.J. and Terlien, M.T.J. 1996. An approach towards deterministic landslide hazard analysis in GIS.A case study from Manizales (Colombia). Earth Surface Processes and landforms, 21: 853-868.

Van Westen C.J., van Asch T.W.J., Soeters R. 2006. Landslide hazard and risk zonation – why is it still so difficult?. Bull EngGeolEnv 65. Pág 167-184.

Varnes, D. J., 1984. Landslide hazard zonation: a review of principles and practice. Natural Hazards, 3, UNESCO Press.Paris, 64 pp.

Varnes, D.J. 1981. Slope stability problems of Circum-Pacific Region as related to mineral and energy rsources, Am. Assoc. Pet. Geol. Studies Geol., 12. Pág. 489-505.

Velasquez N., 2011. Simlacion de sedimentos aspartir de un modelo conceptual distribuido no lineal. MSc Thesis. Universidad Nacional de Colombia. Pag. 116.

Vélez, J. I., 2001. Desarrollo de un modelo hidrológico conceptual distribuido orientado a la simulación de crecidas. Valencia. Tesis Doctoral. Universidad Politécnica de Valencia.

Vélez, J., I., Villarraga, M., R., Álvarez, O., D., Alarcón, J., E., Quintero, F., 2004. Modelo distribuido para determinar la amenaza de deslizamiento superficial por efecto de tormentas intensas y sismos. XXI Congreso latinoamericano de hidráulica, Sao Pedro, Brasil.

Xie M., Esaki T., Cai M., 2004. A time-space based approach for mapping rainfall-induced shallow landslide hazard. Environmental Geology 46.Pág. 840-850.

Wang, G., Sassa, K., 2003. Pore pressure generation and movement of rainfall-induced landslides: effects of grain size and fine particle content. Engineering Geology Vol. 69, Pág.109-125.

Wang, F., Shibata, H. 2007. Influence of sol permeability on rainfall-induced flowslides in laboratory flume tests. Can. Geotech. Journal 44. Pág. 1128-1136.

Wieczorek G.F., Guzzetti F. 2000. A review of rainfall thresholds for triggering landslides, Mediterranean Storms, Proceeding of the EGS Plinius Conference, Maratea, Italia.

Wu, W., Sidle, R. C., 1995.a distributed slope stability model for steep forested basins. Water resources Research, vol. 31, No. 8. Pág. 2097-2110.