Deterministic Landslide Stability Analysis:

an example from NW Oregon

Pago Lumban-Tobing

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Oregon State University

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LANDSLIDE STABILITY ANALYSIS
IN THE ASTORIA BASIN

Abstract

A slope stability map is created for the Astoria Basin, northwestern Oregon. The stability analysis is based upon a topographically based model, TOPOG model. The model predicts the occurrence of subsurface water-logging which reduces the shear-strength of a slope-profile. Simply stated, subsurface saturated zones occur wherever the local drainage flux exceeds the slope’s ability to transmit water. The model is coupled with the infinite slope stability model for analysing the effect of topography on slope stability. The coupled model divides the landscape into four classes based upon its slope stability. The model is simulated using three different rainfall events. The simulations provide a means to observe the behavior of each slope basin at a given rainfall rate. The result of the simulations creates a map that shows the distribution of the slope-stability classes. The map is tested against the active slide occurrences which are mapped in the field and from geological maps. The locations of the active slides seem to be consistent with the unstable slope-classes in the model.
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INTRODUCTION

Landslides occur frequently in the forested regions of the Oregon coastal mountain ranges. Increasing human activity within these areas have an important impact on the number of landslide occurrences and slope stability in general. For Clatsop County, as well as many parts of the Oregon, forests play a major role in its economic development. The growing need to utilise these forested areas causes an increased interest in slope stability problems. Unfortunately, a map showing the distribution of slope stability in the landscape was not available at this moment for Clatsop County. The topography, soil properties, hydrologic conditions, geology, and vegetation of this area control the relative stability of slopes. The purpose of this project, thus, is to create a slope stability map that can incorporate these factors.

Various approaches to examining the importance of topography in controlling the slope stability have been developed. The available data and model suitability limit these various approaches to few suitable models. The study area in this analysis is located within the Oregon Coast Range and covers an area of about 10,000 square kilometres. The landslide distribution, which is usually the most fundamental map requirement for most methods, is not available. The data that are currently available include some geologic maps, Digital Elevation Models, and hydrologic as well as soil maps. Thus, a model that can incorporate these parameters will be more attractive.
GENERAL TRENDS IN LANDSLIDES ANALYSIS

Several types of landslide hazard analysis techniques have been developed in the last 30 years; namely: landslide frequency & distribution analysis, qualitative analysis, statistical analysis, and deterministic analysis. To begin, each method was analysed to select the most appropriate approach in accordance with data and personnel availability as well as model suitability.

Landslide distribution analysis is the most straightforward stability analysis. The approach is, nonetheless, only a landslide inventory map. Mass movement features are directly mapped from aerial photo interpretation, ground survey, and/or a data base of historical occurrences of landslides. The landslide distribution can then be shown as a density map or by means of landslide isopleths. Although the easiness of this method makes it favorable, we cannot use this approach due to lack of a suitable data-base. Large scale aerial photographs of the study area, which provides critical information in creating the landslide distribution, are not available. A number of large-scale geological maps of the area include some information on historical mass movements; however, these maps cannot be utilised to create a historical data-base because of the inconsistency of the map-scale, variations in technique, as well as incomplete coverage.

The second approach is qualitative hazard analysis. In this method the degree of hazard is evaluated by an expert geomorphologist. The geomorphologist will create decision rules in classifying the landscape based on its stability to mass movements.
These decision rules can vary from place to place and from one person to another. Thus, this approach is considered subjective. In addition to its subjectivity, variations in personnel expertise make this approach less attractive.

A more objective method is to incorporate statistical techniques to classify the landscape based on relative stability. These statistical techniques have their advantages in forecasting slope instability at a regional scale (Carrara, 1993). At this scale, mass movements can be described as a result of an interplay of a large set of interrelated factors (Carrara, 1983). Many of these factors are unknown or their relationship cannot yet be identified. Statistical analysis provides the means to exclude the insignificant factors and rank the significant factors as promoting agents for shallow mass movements. In addition, the statistical measurements can also be utilized in hypothesis testing for a causal relationship (Gao, 1993).

Statistical landslide hazard analysis is performed in two steps. The first step involves with identifying those elements in the physical landscape which are important in accounting for variability in slope stability (Reading, 1993). In this process, various maps and a data base is created. These maps describe the landscape in terms of topographical, morphological, geological, and vegetational attributes. The data needed for this investigation are derived from existing topographic maps, aerial photographs, and field surveys (Carrara 1983, 1991; Reading, 1993, Gao, 1993). In addition to the landscape attribute maps, statistical hazard analysis requires a landslide distribution map of the area. The landslide distribution map is utilized in selecting the causal elements from the attribute maps.
The selection process can be performed subjectively (Reading 1993) or by applying statistical analysis. Statistical techniques usually involve stepwise discriminant or regression analysis (Carrara 1983, 1993; Gao, 1993). At this stage, parameters that show insignificant relationships with the landslide occurrences are discarded. As a result, a handful of remaining parameters are included in the discriminant function. These factors will be utilized to classify the landscape stability.

In order to test the discriminating factors against the study area, the study area is split into two parts. One part is then used to estimate the model, while the other is reserved for testing the goodness of fit (Carrara, 1993). In testing the fitness of the model, the predicted sites are compared with those active slides occurring in the area. The landslide distribution map is the most important data source in the statistical approach (van Asch et al., 1993). Unfortunately, for our study such a map is not available.

The last method is based upon a physical model. The model simulates the occurrence of a landslide in the landscape as the landscape is affected by various factors. For practical considerations in the simulation process, the models are simplified to include only a few parameters. The main criticism of this method is its high degree of oversimplification (van Asch et al., 1993); for a similar reason, it may not work properly for a regional study. However, the simplicity of the model as well as the smaller amount of data needed to perform the stability analysis makes this approach more favorable than the others. As we will see in the following sections, the simplicity of the model does not preclude the inclusion of various important factors of mass-movements.

*Deterministic Model:*
Several deterministic models have been developed for shallow landslides stability analysis, and three of these models will be discussed. The three models selected all utilize Digital Elevation Models (DEMs) in the model simulations; thus, their simulations can be automated. The first model was created by Okimura et al. (1988) who proposed a prediction method based upon the simulation of the groundwater level during a storm event. The second model uses DEMs to develop a geometric signature for each slope basin (Pike, 1984). The geometric signature classifies the landscape into various basin slope classes. Locations of active slides are then correlated with each slope-class, to determine whether the slope-class is stable or not. Lastly, the TOPOG model simulates the movement of rainfall water as it infiltrates, percolates, and flows within the soil-profile (O'Loughlin, 1986). This model predicts the occurrence of saturation zones where the throughflow will return to the surface. These saturated-zones are highly correlated with the unstable zones.

The first topographical model utilizes the DEMs to predict water table levels at any given time during a storm event (Okimura et al., 1988). In this model, the levels of the water table for a given time period can be determined by simulating the movements of groundwater. Rain water is assumed to immediately infiltrate vertically and form a groundwater table. If the rainfall intensity for a given time exceeds the rate of groundwater percolation, the excess water will raise the water table. The model simulates the percolating-water movement only in the vertical direction and ignores its lateral component. In heavily forested areas with deep soils, surface runoff seldom occurs (Dunne et al., 1975). Within this landscape, lateral soil-water movement becomes a
major contributor to hillslope runoff. Within topographic convergence, this water-flux accumulates and reduces the soil-strength against mass-movement (Crozier, 1986). In addition, the model relies heavily on infiltration rates derived from piezometric measurements in the field (Okimura, 1988). Such detailed measurements are lacking for this area, therefore, we cannot utilize the model for the Astoria Basin stability analysis.

Another topographical model that uses digital elevation models (DEM) to develop geometric signature for each slope basin is described by Pike (1984). A geometric signature is defined as a set of measurements (e.g., slope length, gradient, and slope reversals) that describes topographic form of a hillslope-profile (flat, shallow, hilly, steep, etc.). The topographic forms then distinguish the landscape based upon classes of geomorphic structures (valley, hilly terrain, foot-slope, hilltops). The geomorphic terrain is then correlated with the locations of active slides. Three classes of landscape terrain identified; namely, 'hard' terrain, characterized by steep side-hill valleys and side-hill edges. 'Soft' terrain lacking vales and ridges; and 'intermediate' terrain which is gradational between the two. Extensive field surveys have identified a strong correlation between the terrain types and erosion processes. 'Hard' terrain typically hosts debris flows and debris avalanches. 'Soft' terrain is associated with chiefly slump/earth flow, earth flow and earth slide. In addition to extensive field surveys, this model requires numerical data on lithology, structure, soils, vegetation, climate, and variables such as the most recent uplift (Pike, 1984).

The third model, (TOPOG), is created by simulating the hillslope flow of soil-water movement. The model has been selected for the following reasons. First, the
TOPOG model is derived from the variable area concept which is believed to fit best for hillslope runoff processes in the undisturbed forested area (Benson, 1964). In addition, the simplicity of the model is also favorable for the ease of automation; however, the model still can include the variations within topography, hydrology and soil-hydraulic properties (O’Loughlin, 1981). Finally, the model has been applied successfully within the Coast Range in Oregon and Washington States (Montgomery & Dietrich, 1993).

Within a heavily vegetated hillslope with deep soil, such as the Oregon Coast Range, surface runoff seldom occurs (Ragan, 1975). Infiltration capacities of the forested soils exceed the vast majority of measured rainfall intensities (Dunne et. al, 1975). In addition, steep slopes and high permeability of surface soils are conducive to rapid, shallow subsurface flow. However, when the soil storage capacity is exhausted, such as within swales and convergence zones, subsurface stormflow is unable to remove all the incoming rainwater. Consequently, an increase in the amount of water stored in the soil raises the water table to the soil surface. Subsurface water then emerges from the soil surface as return flow. Thus, within a heavily vegetated area, only a small part of a watershed produces storm runoff. This runoff model is commonly described as the variable source area concept (see Figure 1).
According to the variable source area model, only a small part of a watershed produces storm runoff (Benson, 1964). These source areas also appear to expand and contract, during and between rainstorms (Dunne & Black, 1970a). Their position and expansion can be related to geology, topography, soil and rainfall characteristics (Dunne & Black, 1970b). Saturation overland flow spreads first up the previously dry, low-order tributary channels, then up unchannelled swales and gentle foot slopes of hillsides, especially those covered with shallow, poorly-drained soils. These saturation zones are also believed to be correlated with the location of unstable zones. Thus, any stability analysis within such an area has to include this topographic variation.

Figure 1. Variable source area model (Kirkby & Carson, 1972).
STUDY SITES

The study area includes all of Clatsop county, located in Northwest Oregon (see Figure 2). The study area covers a 10,000 square kilometre area; thus, any landscape study has to be considered as a regional study (van Asch et al., 1993). The area has low terrain where the highest elevation does not exceed 1,000 meters and most of the elevations are below 800 meters. In addition, the terrain is considered relatively flat since none of the hillslope profiles has a slope greater than 50 percent (see Figure 3). Thus, a first look at topography alone will classify the landscape as stable.

Figure 4 shows the landuse and land cover classification. The map is created from downloaded-files from the State GIS Service Center (1996). It classifies the Clatsop county as mostly forested area, although the satellite image taken in 1989 (see Figure 5) showed that most of these forested areas have been clear-cut. Clear-cutting and logging road construction may increase runoff and slope instability.

Hydrologically, the study area is considered as one of the wettest parts of the State. It has a high annual rainfall with most of the rains falling during cyclonic storms where the prolonged rainfall may fill up the soil-storage and thus reduce the slope strength against water movement (Brooks and Richards, 1994). The general soil map in Figure 6 shows that the whole area is covered with deeply permeable soils. Deep soils and heavy vegetation cover provides high infiltration capacity. In such conditions, most of the hillslope water is transferred within the soil profile in the form of throughflow (Dunn and Black, 1970a). When conditions permit, this throughflow can act as a promoting agent for shallow mass-movements (Kirkby and Chorley, 1967).
Astoria Basin
Elevation model

- above 800m
- 600m - 800m
- 400m - 600m
- 200m - 400m
- 0 - 200m
- Water bodies
Land Use and Land Cover classification

Legend
- Urban or built-up land
- Agricultural land
- Rangeland
- Forest land
- Water body
- Wetland
- Barren land
Aerial View of the Astoria Basin
Legend

Soils on floodplains, terraces
- Coquille-Clatsop
- Grindbrook-Walliski-Hebo
- Waldport-Gearhart-Braillier

Soils on mountains in fog belts
- Skipanon-Necanicum-Ascar
- Klootchie-Necanicum-Ascar

Warm soils on floodplains & terraces
- Lacoda-Wauna
- Eilertsen-McNulty-Kirkendall

Warm soils on mountains
- Rinearson
- Hemcross-Klisan-Harslow
- Aistony-Scaponia-Braun

Cold soil on mountains
- Cater-Laderly-Murtip
Geologically, the Clatsop county is considered as tectonically active. The northwestern Oregon is included within an active tectonic belt which encircles the Pacific Ocean (Niem & Niem, 1985). A total of 47 earthquakes have been felt in the Portland area since 1841 (Tonoyer, 1985). Even though these earthquakes are considered as low intensity, they can trigger the movement of hillslope profiles. The soft Tertiary sedimentary rock units in the area have a low shear strength and are very prone to mass wasting (Tonoyer, 1985). Geomorphological studies in the area reveals that most of the active slides in the area are located on the sedimentary units. These sedimentary units appear to be the most susceptible to slumping because interbedded sandstones permit ground water to percolate between the mudstone beds (Penoyer, 1977). The geologic map in Figure 6 shows that many parts of the County are composed of these sedimentary rock units (Qal, Qat, Qt, Qmt).

In conclusion, hydrological, soil, and geological properties of the area promote the occurrence of shallow mass-movements. These factors working in combination may promote such a movement even though the topography classifies the area as stable. Hence, any studies predicting the occurrence of soil-movements should incorporate these factors. In the next sections, we will discuss the development of a deterministic model for predicting shallow mass-movements which incorporates hydrological and soil-hydraulic properties as well as topographical parameters.
Geology of the Astoria Basin

Legend

- Yellow: Qal
- Red: Tf
- Green: Ts
- Purple: Tn
- Orange: Tc
- Lilac: Tp
- Brown: Tt
- Light Green: Water
- Blue: Tg
- Green: Tac
- Green: Th
- Purple: Tay
- Purple: Taw
- Blue: Ts
- White: Outside
TOPOG MODEL

The TOPOG (Topographic) model predicts the occurrence of subsurface saturated zone areas by simulating the flow of rainfall as it percolates and fluxes within a soil profile (O’Loughlin, 1986). The analysis is based on the role of soil conductivity, hillslope gradient and catchment wetness in determining the size of a surface-runoff producing area. The soil conductivity and hillslope gradients control the slope’s ability to transmit water; likewise, the catchment area determines the volume/rate of soil-water drainage. A saturation zone will occur wherever the local drainage rate exceeds the local slope’s ability to carry water.

Hillslope capacity for drainage:

The flow of water within a soil profile is commonly described as a Darcy flow. According to the Darcy law/equation, soil-water movement is controlled by the hydraulic conductivity of the soil. This flux of water stays inside the soil-profile as long as two sets of conditions are satisfied. First, at any location on the hillslope, the direction of the subsurface flow has to be less than the gradient of the local slope. Secondly, the rate of flow must not exceed the soil’s ability to transmit the flux; otherwise, the excess water has to be transferred to the soil surface (O’Loughlin, 1981).
Figure 8 depicts an idealised hillslope soil-profile. Within a soil-profile, the hydraulic conductivity of the soil will decrease with depth below the surface. In this soil profile, the depth at which the soil hydraulic conductivity approaches zero is described as depth 1. According to the Darcy equation, the component of flow in the direction parallel to the slope will be:

\[ dq = K(z) \, dz \]  

(i)

where \( dq \) is equal to the flow rate, and \( K(z) \) is the change of the hydraulic conductivity in the \( z \)-direction (vertical). The total flux of water in the soil profile is equal to the integration of this equation to the depth at which the hydraulic conductivity approaches zero, i.e. :

\[ q = \int_{0}^{1} dq = \int_{0}^{1} K(z) \, dz \]  

(ii)

We can simplify the integration by evaluating the integral in the equation above. We are call this value the soil transmissivity, \( T \):

\[ T = \int_{0}^{1} K(z) \, dz \]  

(iii)

The second factor controlling the occurrence of subsurface flow is the slope gradient. In order for a continuous subsurface-flow to occur at any location within the hillslope, the flow direction cannot exceed the hillslope gradient. The diagram in Figure
8 illustrates that if the flow direction is greater than the slope, then the flux will start to move up to the soil surface. In this hillslope profile, the local slope gradient is described as $M (\sin \theta)$. Therefore, combined with the hydraulic conductivity factor, the maximum subsurface flux which can occur parallel to the hillslope equals the product of the hillslope gradient and the downslope water flux, or:

$$C = M \int_{0}^{1} K(z) \, dz = MT$$

(IV)

where $C$ is the hillslope capacity to transmit subsurface flux. Thus, we have defined the hillslope capacity in terms of the soil hydraulic property ($T$, soil transmissivity) and the hillslope soil profile ($M$, slope gradient). Next, we are going to compare this value to the local drainage flux that a hillslope profile has to transmit.

**Local drainage flow:**

In order to calculate the local drainage flow at any location within a watershed, the catchment area has to be subdivided into smaller cells within which the local drainage flow can be calculated. In this model, the watershed will be divided into smaller parcels according to the contour format. The flow of water in a hillslope-profile will follow the line of steepest gradient (Jensen & Domingue, 1988). Within a contour format, this line is perpendicular/orthogonal to the contour lines. And for this model, each parcel/cell will be defined as an area bounded by two adjacent contour-lines and two adjacent flow-lines.
At each cell, we can calculate the amount of rainfall that has to be drained within the soil profile, and this total local drainage flux will be compared against the capacity of the cell to transmit subsurface flow. Figure 9 describes a three-dimensional view of one of the cells. In this cell, the total amount of flux will be equal to the product of the surface area of the cell, \( a \), and the amount of discharge, \( q \). In this regards, the discharge \( q \) will be equal to the sum of precipitation \( p \) minus the loss to deep drainage \( r \) and evaporation \( e \) (or \( q = p - r - e \)). Therefore, the total amount of flux is equal to:

\[
Q = aq \quad (v)
\]

This total volume will be transmitted in a downslope direction passing the lower cross-section of the cell; in this cell, the lower cross-section is equal to length \( b \). Thus, the total discharge per unit width in any cell is equal to:

\[
Q = \frac{aq}{b} \quad (vi)
\]

Now we can compare the hillslope capacity \( C \) with the actual drainage flux \( Q \). The comparison yields a criterion for whether a zone of saturated soil occurs in the cell or not. Local saturation occurs wherever the drainage flux from upslope exceeds the capacity of a soil profile to conduct that flux, i.e.:

\[
Q \geq C \quad \text{or} \quad \frac{aq}{b} \geq TM \quad (vii)
\]
Later on in the stability analysis, we will describe the importance of the water table level in determining the slope-stability. At this stage, we are only going to derive the method to calculate the water-table level from the TOPOG model. In the following, we will calculate the water-table level at each cell based upon the local drainage flux. In Figure 9, the water-table level is described as height $h$. This saturation level represents the amount of input water to one particular cell; and, its height is equal to the volume of the flux divided by its surface area ($Q/a$). In comparison, the depth of the soil profile, $z$, represents the hillslope capacity to transmit soil-water; and, its value can be calculated by dividing the hillslope-capacity with its surface area ($C/a$). Thus, the ratio of the water-table level to the soil height becomes:

$$\frac{h}{z} = \frac{Q}{a} = \frac{C}{Q} = \frac{qa}{TMB} = W$$

This ratio of water table level to soil depth will be called the \textit{wetness value}, $W$. This value represents the percent saturation condition of the soil-profile. Wetness values greater than 1.0 indicate that the soil profiles are fully saturated and the soil water has to be transferred into the soil surface.

In conclusion, the TOPOG model predicts the occurrence of saturation zones by simulating the movement of rainfall water as it percolates and fluxes through the soil profile. Simply stated, the saturation condition will occur whenever the local drainage rate exceeds the hillslope capacity. This ratio of local drainage rate to hillslope capacity is referred to as the \textit{wetness value}, $W$. It is, nonetheless, the ratio of water-table level to
soil-depth. In the following section, we combine the TOPOG model with the infinite slope stability model to analyse the effect of topography on shallow mass-movements.
Stability analysis of shallow slides is commonly conducted using an infinite slope stability model. In this model, the failure surface is approximated as planar and parallel to the surface. It is also assumed that the end and side effects are ignored; the justification for this treatment is the great length of the slope relative to the depth. In the analysis the soil layer is considered as composed of an infinite layer of clay whose failure plane is assumed to be at a constant depth beneath the surface along the slope (Ritter, 1988). Therefore, with this method we need consider only the shear strength and stress at the base of one column of soil (one dimension instead of three).

For a system in equilibrium, the forces tending to promote movement are exactly balanced by the resistance opposing them. Force requires energy, and all energy in geomorphic systems is ultimately derived from either gravity or climate (Kirkby & Carson, 1972). The force provided by gravity is simply that of the weight of the soil. Climate, likewise, exerts force through its control on temperature and available water. In this model, we consider only rainfall water as a probable agent for mass-movements. Thus, the two probable forces for movements being considered are the weight of the soil and the movement of rainfall water within the soil-profile.

The force of gravity on any object acts vertically downwards and it is expressed in terms of the weight of that object. On a slope, the vertical force of weight can be resolved into two components; the first is a downslope force which tends to move the object downhill parallel to the surface; and the second is a force perpendicular to the
surface which acts generally to hold the material onto the slope. Figure 10 shows an
idealised hillslope model and the breakdown of gravity force into these components.

Because slope movements are generally parallel to the surface, the downslope
component of weight is the more important promoting agent. The downslope component
is equal to the weight multiplied by the sine of the slope gradient angle:

\[ \tau = W \sin \theta \]

The weight of the soil is equal to the product of the density of the
soil, \( \gamma_s \), and the soil depth, \( z \).

Because the soil is located on a slope, the depth of the soil can be
described as \( z = h \cos \theta \), where \( h \)
is the vertical distance between the surface and the slip plane. Thus, the two components of gravity can be described as:

**downslope component = shear stress** = \( \tau = W \sin \theta \)

\[ = \gamma_s z \sin \theta = \gamma_s h \sin \theta \cos \theta \]

**vertical component = normal stress** = \( \sigma = W \cos \theta \)

\[ = \gamma_s z \cos \theta = \gamma_s h \cos \theta \cos \theta \]

\[ = \gamma_s h \cos^2 \theta \]
This force of gravity is resisted against movement by the forces that keep the soil-profile intact, or its shear strength.

*Shear strength:*

Resistance that opposes the shear stresses imposed by gravity is termed shears strength. The most popular, and the simplest, interpretation of shear strength is provided by Coulomb in 1776 (Carson and Kirkby, 1972). According to Coulomb equation, this strength can be divided into three components: a) overall frictional characteristics, usually expressed as the angle of internal friction, b) effective normal stress, and c) cohesion. The Coulomb equation is written as:

\[ S = c + \sigma \tan \phi \]

where \( S \) is the shear strength, \( c \) is the cohesion, \( \sigma \) is the effective normal stress, and \( \phi \) is the angle of internal friction. As previously stated, the infinite slope model assumes that the slope-profile is composed of infinite clay soil. Clay soil is mostly regarded as a cohesionless material so that the cohesion factor of the Coulomb equation, \( c \), from now on can be omitted.

*Internal friction:*

Internal friction is composed of two separate types: plane friction, which is produced when one grain slides past another along a well-defined planar surface, and interlocking friction, which originates when particles are required to move upward and over one another (Ritter, 1988). Interlocking friction is commonly greater than plane friction, because extra energy must be used to move interlocked particles in an upward
direction. In loose particulate matter of any size, such as the soil profile, the angle of repose should approximate the angle of internal friction (Carson & Kirkby, 1972).

Normal stress:

The third element of the Coulomb equation is the normal stress, \( \sigma \). This is the perpendicular force that acts to push the particles together. Within a soil-profile, this force is none other than the weight of the soil-particles. Because soil profile is composed of solid particles as well as pore space, this normal stress would also be diverted to two components. Stress induced by solid-to-solid contact across the shear zone is called the effective normal stress, \( \sigma' \); likewise, the normal stress exerted on the pore space is known as the pore pressure, \( \mu \). Thus, the total normal stress is the sum of the effective normal stress and its pore pressure:

\[
\sigma = \sigma' + \mu \quad \text{or} \quad \sigma' = \sigma - \mu
\]

In calculating the stability analysis, it is the effective normal stress that is more important than the total normal stress because the total normal stress can be reduced or added according to value of the pore space.

Pore pressure:

Pore pressure has been defined as the portion of the normal stress that is being exerted within the soil pore space. Pore pressure can be exerted in two different forms: water pressure and tension force (Crozier, 1986). The occurrence of either force is dependent upon the saturation condition of the soil-profile. Under saturation condition,
the pore pressure is exerted as water pressure; whereas, within unsaturated conditions, the
tension forces make up the pore pressure. Both forces work in the opposite direction to
the stability of the slope profile; water pressure reduces the slope strength, whereas
tension forces tend to stabilize the slope profile.

The first component of pore pressure is exerted through capillary tension force. This
pulling force is created between soil particles and water molecules. Above the water
table, the soil is within an unsaturated condition. Within the unsaturated zone, water is
held in the soil pores under surface tension forces. This water will be prevented from
moving downward because it is attached to soil particles. Simply stated, this attached
moisture increases the weight of the soil; thus, it is added to the normal stress/shear
strength.

The second force, water pressure, is produced because of the fact that the volume
of water is only slightly altered when it is being pressured. As a result, water particles
exert a resistance when being pushed. This phenomenon causes any object immersed in
water to be subjected to a relief of weight. The force is equal to the weight of water being
displaced. Water pressure occurs only within a saturated condition because within an
unsaturated zone, water particles can still fill up the empty pore-space. Below the
water table, all open pores are saturated; thus, at a point below the water table, there is a
hydraulic upthrust or relief of weight equal to the water pressure at that point. The
upthrust is equal to the weight of the water being displaced by the soil articles; and this
would be equal to the weight of the water below the water table. The height of the water
below the water table, \( W_w \), can be written as the product of the water
density and the volume of water below the water table. Thus, the upthrust force is equal to:

\[ W_w = \gamma_w z_l \]

where \( z_l \) is the depth of the water table (Figure 9). Also from Figure 9, we can see that \( z_l \) is equal to the product of the vertical distance of the water table to the bottom surface, \( z \), and the slope gradient, \( \cos \theta \). Thus, the upthrust force is equal to \( W_w = \gamma_w z \cos \theta \).

Another factor that should be considered is that the upthrust is important only in the perpendicular direction, therefore, it has to be once again multiplied by the slope gradient.

\[ \text{Upthrust} = W_w \cos \theta \]

\[ = \gamma_w z_1 \cos \theta = \gamma_w z \cos^2 \theta \]

Combined with the pore pressure, now we can write the Coulomb equation in its full format: shear strength = \( S = \sigma' \tan \phi = (\sigma - \mu) \tan \phi \)

\[ = (\gamma_s z \cos^2 \theta - \gamma_w z_l \cos^2 \theta) \tan \phi \]

\[ = \cos^2 \theta (\gamma_s z - \gamma_w z_l) \tan \phi \]

In conclusion, topography controls the strength of a soil-profile through the location of the water table. The pore pressure at any point in the soil-profile is dependent upon the location of the water table (see Figure 11). Below the water table, there is a positive upthrust relieving the overburden weight of the soil. Above the water table there

![Figure 11. Pore pressure within the soil profile (Ritter, 1988).](image-url)
is a hydraulic force which increases the effective weight of the soil by capillary cohesion. Thus according to the infinite slope stability model, the location of the water table controls the shear strength of a soil-profile.

The TOPOG model provides a method for calculating the level of the water table (wetness value, $W$), given the topographic parameters and rainfall rate. In the following, we are going to combine the TOPOG model with the infinite slope model to calculate the stability of a slope profile against a shallow mass-movement.

**Factor of safety:**

The stability analysis within the infinite slope model is commonly described as the ratio between forces that resist the downwards movement, shear stress, and forces that cause that movement, shear strength. This ratio is commonly known as the factor of safety. A slope profile is defined as stable if the ratio is larger than 1.0 (therefore, the shear strength is greater than the shear stress). Previously, we have provided the mathematical equations to calculate both forces. Thus, factor of safety can be written as:

$$\text{factor of safety} = F = \frac{\text{shear strength}}{\text{shear stress}} \geq 1$$

$$\Leftrightarrow \left[ \cos^2 \theta (\gamma_s z - \gamma_w z_1) \tan \phi \right]/\gamma_s z \sin \theta \cos \theta \geq 1$$

$$\Leftrightarrow \left[ (\gamma_s z - \gamma_w z_1) \tan \phi \right]/(\gamma_s z \tan \theta) \geq 1$$

$$\Leftrightarrow \left[ (\gamma_s z - \gamma_w z_1) \tan \phi \right]/\gamma_s z \geq \tan \theta$$

$$\Leftrightarrow (1 - (\gamma_w/\gamma_s)(z/z)) \tan \phi \geq \tan \theta$$

or $$(\gamma_w/\gamma_s)[1 - \tan \phi/\tan \theta] \geq z/z$$
where \( \gamma_w \) is the density of water, \( \gamma_s \) is the density of the soil profile, \( z_1 \) is the depth of the water table, \( z \) is the depth of the soil profile, \( \phi \) is the angle of repose, and \( \theta \) is the hillslope gradient.

Previously, from the TOPOG model we have derived the ratio of the water table level to the depth of the soil from the topographic parameters. This ratio is called the wetness value:

\[
\frac{z_1}{z} = W = \frac{aq}{TMb}
\]

We can couple the two models together so that the stability analysis can be described using the topographic, soil-hydraulic and hydrologic parameters. Thus, the combined model of the TOPOG and the infinite slope stability equation becomes:

\[
\frac{aq}{TMb} \geq (\gamma_s/\gamma_w) [1 - \tan \phi / \tan \theta]
\]

The equation above defines the model in terms of three important factors: hydrologic parameters (\( T \)), soil parameters \((\gamma_s/\gamma_w) [1 - \tan \phi / \tan \theta]\), and topographic parameters \((\frac{a}{b} \text{ and } \sin \theta)\). If we keep two of the parameters steady, then we can simulate the effect of the third parameter on the stability of a hillslope.

The major goal of this study is to examine the control of topography on the occurrence of shallow landslides. Hence, in the simulation part, we will keep the hydrologic and soil parameters constant and vary only the topographic parameters. The justification is as follows. Infinite slope stability is based on the assumption that the
soil-profile is homogeneous; thus, the soil hydraulic characteristics are constant throughout the study area. Such a condition may not prevail in our study area; however, we have already accepted such a pre-condition in order to use the infinite slope model. We are also going to keep the hydrologic factor constant by running a steady-state rainfall event over the area. The TOPOG and infinite slope models have simplified the conditions so that including the variation in rainfall may not have much of an effect on the calculations. Thus, by keeping the two other parameters constant, we can simulate the effects of topographic parameters on the stability of the hillslope.
MODEL SIMULATIONS

The TOPOG model is coupled with the infinite slope model to simulate the effects of topography on slope-stability. The combined model consists of three classes of parameters: hydrologic, soil-hydraulic, and topographic properties. For the purpose of examining the effect of topography on shallow mass movements, the other two components will be held constant. Three simulations are performed by using different steady-state rainfall events. The rainfall rates are selected to represent three different return periods. At each rainfall event, the importance of topography is analysed. The effect of soil-water movement will be summarised in the form of the wetness value \( W = \frac{a_q b T \sin \theta}{b T \sin \theta} \). This wetness value represents the saturation condition of a slope profile, which in return affects the stability of the slope-profile.

Four stability classes are defined to describe the elements within a catchment; namely: unconditionally unstable, conditionally unstable, conditionally stable, and unconditionally stable (Dietrich & Montgomery, 1993). The unconditionally unstable class includes the area where the hillslope gradient is larger than the angle of repose \( \tan \theta \geq \tan \phi \); therefore, the stability is not affected by the hydrologic parameters. In contrast, the unconditionally stable describes the area that does not move at any rainfall rate \( \tan \phi [1 - (\gamma_w/\gamma_s)] \geq \tan \theta \). The other
two classes are affected by the rainfall event. The conditionally unstable area is defined as the area where the factor of safety is reduced to unstable for that particular rainfall event \((z_1/z > (\gamma_w/\gamma_s)[1 - \tan \phi /\tan \theta])\); and, the conditionally-stable area has a stable factor of safety \((z_1/z \leq (\gamma_w/\gamma_s)[1-\tan \phi /\tan \theta])\). Figure 12 shows the boundaries for the four stability classes. The stability of each topographic element is shown by plotting its slope-gradient against the calculated wetness value.

**Calculations of the wetness value:**

The TOPOG model reduces the calculations of saturation condition to only five parameters: surface area, cross-section length, slope-gradient, soil-transmissivity, and local drainage flow \((W = \frac{aq}{TM_b})\). Previously, it has been shown, for a steady-state rainfall event, that the local drainage flow is equal to precipitation minus losses to evaporation and deep-drainage \((q = p - e - r)\). These net precipitation values are reported in the Soil Survey as well as in the Hydrological Atlas produced by NOAA. For the simulation purposes, three rainfall rates are selected: 20mm/day, 100mm/day, and 200mm/day. These values are selected to represent the 2-year, 10-year, and 50-year return periods, respectively.

The soil-hydraulic properties in the coupled model are represented by the density of the soil \((\gamma_s)\) and the transmissivity value, \(T\). The density of the soil can be measured in the laboratory or it can be obtained from the Soil Survey. In this study, the soil-profile is considered as composed of clayey soil and its density is selected at 1.6 gr/cm\(^3\) as listed in
the Soil Survey (SCS, 1988). The soil transmissivity value ($T$), on the other hand, has to be measured in the field using piezometric measurements. Similar measurements have been performed from other studies in the same area (Dietrich & Montgomery, 1993; Schroeder & Alto, 1983). The two studies reported similar numbers; therefore, we are going to use those values in our simulations. For the simulation purposes, the transmissivity value is selected at 65 m/day. Thus, this leaves the calculation to only the topographic parameters.

The last parameters, topography, are calculated from the USGS Digital Elevation Model (DEM) at a scale of 1:100,000. The DEMs have a resolution of 90m x 90m and are available from the USGS Internet site (USGS, 1996). From these DEMs, the topographic parameters (surface area, slope gradient, and cross-sectional length) can be easily calculated. All of the data preparations and calculations are performed within the Arc-Info GIS.

Before the DEMs can be used for the topographic calculations, they have to be free of errors. These errors are usually in the form of isolated topographic depressions. Such a depression does not commonly occur in the field; but, rather is usually a result of digitising errors. The identification and removal of these topographic depressions is performed using the algorithm created by Jenson & Domingue (1988). Once the depressions have been filled, we can simulate the flow of rainfall within a watershed. The same algorithm provides the calculation of this flow-accumulation.

The procedure to calculate the flow-accumulation is illustrated in Figure 13. The first step involves the calculation of the slope gradient for each cell. The gradient is
defined as the largest drop from
the center cell to any of its eight
neighboring cells and it represents
the cell flow direction. Once the
flow direction of all the cells have
been determined, we can continue
to the flow accumulation. The
flow is calculated from the ridge
down to the stream. A ridge is
defined as the cell where none of
its neighbouring-cells flows into it. In the beginning of the calculation, all cells are given
a value of 1. At each step, the flow is accumulated following the flow direction. At the
end of the calculation, the flow accumulation gives each cell a value equal to the total
number of cells that drain into it. With this calculation, we have obtained all of the
parameters needed to run the model.

In the simulation, the wetness value
calculated using the topographic parameters is
tested against the factor of safety for a given
slope profile. The calculated-value then
categorises each cell according to the four
stability classes as described above. The result
of the first iteration of the model using a two-year return period is shown on Figure 14.
The graph shows the distribution of the wetness values against the topographic parameter (slope gradient). Instead of showing a random distribution, as would be expected, the graph shows a logarithmic dispersal. This discrepancy is believed to be related to the data format. In this simulation, the topographic data are stored as raster data, whereas, the model is developed using a contour-based model. In the TOPOG model, the flow is originated from a single point on a ridge and flows downslope to a cross section of a contour line (Figure 9). On the other hand, in a raster format, the flow is accumulated into a single point at the lowest point of the watershed. Thus, in the raster based format, the flow-accumulation values tend to grow in a logarithmic instead of a linear trend.

To solve this problem, we modify the raster data and transform them into a triangulated format. The triangulated format is selected because it is similar to the TOPOG model format. Within a triangulated model, the flow direction is also calculated from a single point, the highest point within the triangle. The sides of the triangle where outflow occurs become the cross section width. The following chapter will be devoted to the transformation of this data format.
TRIANGULATED-IRREGULAR-NETWORK (TIN)

The following steps are involved when transforming the regular grid elevation models into irregularly-distributed data points. In this model, the Triangulated Irregular-Network (TIN) is selected because it has several advantages over the regular grid structure; also, it is one of the most fundamental point data-structures (Tsai, 1993). In addition, the TIN format is supported by the Arc-Info GIS.

*Generation of TIN models:*

There are two principal phases in the generation of TIN models from a regular grid structure: the selection of the data points and their connection into triangular facets (Palacios and Cuevas, 1986). Selecting the data points from the Digital Elevation Models involves checking the importance of the point in describing the terrain. Data points that can be predicted from their eight neighboring points are considered less important; and not included in the triangulation. On the other hand, if the elevation interpretations from the eight-neighboring points produce an error that is larger than a threshold value then the point will be recognized as a very important point and it will be included in the final triangulation.

The procedure for checking the very important points is illustrated in Figure 15. In this method, A through G represents the neighboring points of P, the point being checked. The elevation of point P then is interpolated from horizontal, vertical, and two
diagonal cross-sections. The error, the vertical difference between the actual and the interpolated elevations, is then calculated. If the average of the error is larger than a threshold value (the threshold is set up as one percent of the range of the elevation value: \( \frac{\text{MAX} - \text{MIN}}{100} \)) then P will be considered as an important point because its elevation cannot be interpolated accurately from its eight neighboring points. The process proceeds until all the points in the DEM have been checked. After a set of points that can best describe the terrain has been selected, the triangulation can proceed. We also have to make sure that repeating the process will never provide a different result. Therefore, the triangulation has to create a unique set of triangles (Tsai, 1993a). Delaunay triangulation is generally selected because it gives one of the most fundamental data models in computational geometry (Guibas and Stolfi, 1985; Tsai, 1993b) and also because it allocates a unique set of triangles that can best describe the terrain (Tsai, 1993b).
Delaunay triangulation is considered as the dual of its Voronoi diagram (Guibas and Stolfi, 1985, Tsai, 1993a, 1993b) because Delaunay triangulation can only be created in its Voronoi/Thiessen polygons. A sample diagram of both Delaunay triangles and their Voronoi/Thiessen polygons is provided in Figure 16. A Voronoi diagram is defined as a union of contiguous polygons whose boundaries are made up of the perpendicular bisectors of lines joining neighboring points (Tsai, 1993b). With this definition in mind, the set of very important points that have been selected in the first stage becomes the centers of these polygons. Then the boundary of the polygon is created by drawing a perpendicular line that bisects the line joining two adjacent points. In so doing, every point within a polygon will be closer to the center of that polygon than to any other center points. This is the criteria for creating a Voronoi/Thiessen diagram. The Delaunay triangulation is constructed by connecting the point whose associated Voronoi polygons share a common edge (Tsai, 1993a).

**TIN data structure:**

There are at least two ways to define the structure of a TIN (Palacios and Cuevas, 1986). The first is to store the labels of all points (the coordinates are stored separately) which are linked with each other point, in a counter clockwise (or clockwise) direction. The second approach uses a storage by triangle method. Each data record contains the information describing all three adjacent triangles and the nodes that compose the triangle; and, the coordinate of each node is stored separately. With this...
data structure, the required topological relationships can be derived from each record. Thus, the triangle data structure provides more efficient topological data retrieving.

ARC-INFO stores its Triangular Irregular Network (TIN) data in the triangular method data structure (ESRI, 1993). The record for each triangle can be recorded in a polygon, arc or point data structure. For this project, both polygon and arc (line) data-structures will be created. The polygon data structure stores attribute as well as topological information. The attribute data includes information such as: aspect, slope gradient, planar and surface area. The topological information describes its left and right triangles. The TOPOG model, as described previously, requires topographical parameters about surface area (a), slope angle (sin θ), as well as the cross-section length of the runoff area (b). Besides the cross-section length, all other data have been provided in the polygon data structure.

Calculating the cross section width requires the creation of the data structure in the format of lines. The information stored within the arc TIN data structure includes

\[ lslope \text{ (left slope)}, \; rslope \text{ (right slope)}, \]

\[ length, \; slope \; gradient, \text{ and aspect of the arc, as well as the triangle’s ID number.} \]

The \( lslope \) and \( rslope \) describe the slope of the triangles that are bounded by each arc. The slope value is positive if it flows from a

<table>
<thead>
<tr>
<th>Left</th>
<th>Right</th>
<th>Flow side</th>
</tr>
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<tbody>
<tr>
<td>+</td>
<td>+</td>
<td>none</td>
</tr>
<tr>
<td>-</td>
<td>-</td>
<td>both</td>
</tr>
<tr>
<td>+</td>
<td>-</td>
<td>left</td>
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<tr>
<td>-</td>
<td>+</td>
<td>right</td>
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</table>

Figure 17. TIN outflow calculations.
lower elevation point than the arc; in contrast, it has a negative sign if it flows from a higher elevation point (ESRI, 1993). By using this information, we can simulate the flow of water through each triangle.

There are four possible combinations for \textit{lslope} and \textit{rslope} signs. These combinations can be used to determine the geomorphic shape of the triangle whether it is a valley, a ridge, or part of a slope. Figure 17 diagrams the four conditions based upon the combination of its \textit{lslope} and \textit{rslope}. An arc is defined as a ridge if there is no flow of water through that arc (or both \textit{lslope} and \textit{rslope} have positive signs; see Figure 18). On the other hand, if it is a valley, then both the left and right triangles would flow into that particular arc; or, both the left and right triangles have negative signs. The last possibility is that the arc belongs to part of a slope. In this matter, the \textit{lslope} and \textit{rslope} have different signs, and the arc would only be calculated as an outflow for the higher triangle.

Based upon the above conditions, the calculation has to be performed four times (four possible conditions). The result of each iteration is accumulated and its value is defined as the cross-section width for the outflow. Thus, the outflow calculation provides the last parameter needed for the simulation which will be described in the following section.

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure18}
\caption{An example of a TIN flow-side calculation.}
\end{figure}
RESULTS

The combined TOPOG and infinite slope stability model is run using three different rainfall rates. The rainfall events represent three different return periods: 2-year, 10-year, and 50-year. The simulation provides a means to observe each slope basin behavior for any given rainfall event. Based on this observation, the landscape is divided into four slope-stability classes. The landscape division is then tested against the occurrences of active slides. Active slides have been mapped in the field and from available geological maps. Overlaying both maps has shown that the occurrences of active slides seem to be positively correlated with the unstable classification of the landscape.

The coupled model is simulated using three different rainfall events. The first simulation is operated using a 2-year return period. The result of the simulation is presented on Figure 20. The landscape division classifies most of the landscape into saturated and stable areas. Similarly, Figure 19 diagrams the distribution of wetness values against topographic parameter. Most of the points fall into stable slope classes. This concludes that the rainfall rate is too low to promote shallow mass-movements.
Predicted instability at a rainfall of 20 mm/day

- Surface saturation
- Unconditionally unstable
- Conditionally unstable
- Conditionally stable
- Unconditionally stable
Predicted instability at a rainfall of 100 mm/day

- Surface saturation
- Unconditionally unstable
- Conditionally unstable
- Conditionally stable
- Unconditionally stable
The second simulation is run using 100mm/day rain or rainfall with 10-year return period. Figure 21 shows the distribution of slope stability classes. At this rainfall rate, saturated zones have moved to the unconditionally stable areas. These are the areas that are too flat for mass-movements to occur. In this situation, surface erosion is more prevalent. In this simulation, the conditionally unstable area grows at the expense of the conditionally stable zones. Figure 22 shows the distribution of wetness values against slope tangent. Many slope profiles described as unconditionally stable at 20 mm/day rainfall rate become only conditionally stable at this rate. More slope profiles are grouped as conditionally unstable at this rainfall rate than at the lower rainfall rate. This observation indicates that at this rainfall rate, subsurface flow begins to act as the promoting agent for shallow mass-movements.

The trends continue when the model is simulated using a higher rainfall rate. The last simulation is run using a 50-year return period rainfall and is presented on Figure 24. Most of the landscape is classified as saturated. The distribution of wetness values against slope tangent at this rainfall rate is diagrammed on Figure 23. The graph contains much fewer points than that at the lowest rainfall rate. This is the result of
surface saturation condition within most slope profiles. The diagram, however, illustrates that most of the conditionally stable zones have been altered into unstable condition. In conclusion, at this rainfall rate subsurface water flow is an important factor in controlling mass movement.

Based on the simulations, we classify the landscape according to its slope stability; namely: unconditionally unstable, conditionally unstable, conditionally stable, unconditionally stable, and surface erosion. Figure 28 shows the distribution of the overall slope stability classes. Unconditionally unstable is defined as the area where the slope instability is not affected by the hydrological events. This slope class includes those areas identified as unstable even at the lowest rainfall rate. The second slope-class, conditionally unstable, encompasses those zones where only higher rainfall rates cause the occurrence of mass movements. The rest of the landscape is divided into three slope classes based on the effect of rainfall rates on its stability. The conditionally stable area is defined as the area that becomes unstable only at the highest rainfall rate, whereas the conditionally stable area is defined as those zones that are still stable even at the largest rainfall rate. The remaining area is classified as saturated zone.
Predicted instability at a rainfall of 200 mm/day
The landscape division based upon its slope stability is then tested against the occurrence of active slides in the area. The landslide distribution map is shown on Figure 29. The map is created from field surveys and geological maps. Some of the large scale geological maps provide active slide distribution. Unfortunately, these maps do not provide a complete coverage of the whole study area. None of the data are available for the northeastern part of the Clatsop County. The field study, on the other hand, is concentrated on the road networks and major highways. Accessibility is the major limitation for the field study. More than fifty active slides are located into the landslide distribution map. The location of the active slides is then compared with slope-stability classes.

Comparison is performed by overlaying the landslide distribution map against the slope-stability map. Due to incomplete coverage of the landslide distribution map, no statistical analysis seems appropriate for the comparisons. Figure 25 diagrams the distribution of landslide pixels in comparisons with their slope stability classes. The majority of landslide pixels occur on the saturated slope class. The classification of most slope profiles into saturation category may explain this distribution. More than 60 percent of the

Figure 25. Overlaying landslide distribution map with predicted stability classes.

Figure 26. Adjusting the pixel counts by the number of pixel per class.
area falls into the saturated condition class. Therefore, the distribution tends to be bias towards this slope stability class. In order to reach a more justifiable distribution, data adjustment is necessary.

The adjustment is made by dividing the landslide pixel counts for each slope class with its percentage area. Figure 26 illustrates the distribution after being adjusted. Even though the distribution seems to be more equally spread, the saturated condition still dominates its distribution. The following adjustment is made by excluding the saturated areas altogether. The justification of this action is based on the fact that within this slope profile surface erosion is the more prevalent agent of movement than shallow landslides.

Figure 27 diagrams the distribution of landslide pixel counts for each slope stability class. The cell count distribution depicts that active slides may have an equal chance to occur within a stable or unstable slope profile. A more closer look on cell to cell analysis may give the explanation.

Figure 30 depicts an example of the occurrence of active slides within the landscape. Three slides located at the lower end of the map may give the explanation on the cell count distribution. The circles representing active slides are large enough to include both stable and unstable cells. This explains why active slides have the same chance of occurrence on both stable and unstable slope profiles.

Another plausible explanation is the inaccuracy in locating the active slides into the GIS. The location of active slides from geological maps is transferred onto a quadrangle topographic map by visually relating them with structures visible on both
Similarly, active slides identified in the field were only roughly located on the topographic map by means of visible structures on the map. Only a few of the slides were located in the field by using a GPS (Global Positioning System). Once the point is located on the topographic map, it was then digitized and enlarged into a circle. Thus, the inaccuracy in transforming the location of active slides may have offset the pixel count distribution.

A closer look at slide per slide basis reveals a positive correlation with the locations of the unstable slope profiles. More than eighty percent of the active slides (44 out of 50) include cells described as unstable slope profile. Based on this outcome, we conclude that the overlay analysis shows that most of the active slides occur within the unstable slope-classes.
Predicted instability for the Clatsop County, OR

- Surface saturation
- Unconditionally unstable
- Conditionally unstable
- Conditionally stable
- Unconditionally stable
Overlaying active slides on stability map

Legend
- surface saturation
- unconditionally unstable
- conditionally unstable
- conditionally stable
- unconditionally stable
- active slides
CONCLUSION

The goal of this project was to create a slope stability map for the Astoria Basin, northwestern Oregon. The stability analysis is based upon a topographically based model, TOPOG. The model predicts the occurrence of subsurface water logging which reduces the shear strength of a slope-profile. The model is coupled with the infinite slope stability model for analyzing the effect of topography on slope stability. The coupled model divides the landscape into four classes based upon its slope stability. The model is then simulated using three different rainfall events. The simulations provide a means to observe the behavior of each slope basin at a given rainfall rate. The result of the simulations creates a slope instability map for the Astoria Basin. The map is tested against the location of active slides which are mapped in the field and from geological maps. The analysis shows that the location of the active slides seem to be positively correlated with the unstable slope classes.

Several adjustments can be made to enhance the result of the analysis. The TOPOG model is designed to physically mimic groundwater flow within slope profiles. The use of DEMs at 90m by 90m resolution may not always capture slope breaklines in this type of terrain. A study has concluded that the appropriate resolution for DEMs for terrain studies should be below ten meters.

Even though the outcomes of the model seem to be positively correlated with the active slides, the models can be simulated more accurately with more accurate parameter values. One of the parameters of the TOPOG model is the soil transmissivity. This parameter represents the speed of groundwater flow. In this study, only a single value is
utilized to represent the soil transmissivity of an area as large as 10,000 square kilometres. In addition, the infinite slope stability model is based upon the assumption of a homogeneous slope profile, including a homogeneous depth. Such a precondition seldom prevails in this type of terrain where soil-depths can vary from few inches to several meters within a short distance. Thus, further adjustments can be made to incorporate soil depth variation.

The utilization of a physically based model such as the TOPOG model has its advantage in inclusion of parameters actually measured in the field. Thus, the phenomenon being modelled will replicate the actual occurrences within that specific terrain. In addition, the inclusion of Digital Elevation Models within the deterministic model makes it possible for automating the simulations. These two major factors can speed up the slope stability analysis on a regional basis. The resolution of the DEMs may not provide a detailed observation, however the study can serve as a preliminary research for a more detailed study. To conclude, we believe that the approach is best suited for a general instability analysis on a regional basis for a moderately variable terrain.
REFERENCES


State GIS Centre, [http://www.sscgis.state.or.us](http://www.sscgis.state.or.us)


