# RAINFALL AND LANDSLIDES

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**Summary**

This review first considers examples of landslide disasters induced by rainfall. It continues with a discussion of the subject under four headings. 1. Empirical relationships, 2. Hydrological modelling, 3. Failure mechanisms, and 4. Engineering geology mapping. First time slides and pre-existing slides are considered separately. The review suggests a combination of factors leads to instability and that there is no index or threshold which can be applied generally. A multidisciplinary approach is thus required and it is concluded that the large spatially referenced datasets are ideally suited for storage and manipulation within a Geographic Information System (GIS)

## 1. Introduction

There can be little argument that rainfall has been responsible for some of the worlds most destructive landslides. Mountainous areas with weak, tectonised rocks, weathered soils, glacially deposited debris and man-made earthfill structures are especially susceptible to periods of prolonged or high intensity rainfall. An extensive literature has arisen, which approaches the problem in four ways, i) empirical relationships between rainfall and landsliding, ii) hydrological modelling and estimation of rainfall recharge, iii) the failure mechanisms of rainfall induced slides, and iv) engineering geology mapping of slopes susceptible to mass movement. This review attempts to broadly summarise these approaches in relation to the two basic groups of mass movement, ie:

1. First-time slides

2. Slides on pre-existing shear surfaces

### 1.1 Examples of First-Time Slides

A wide range of first-time failures, ranging from rockslides to debris flows have been triggered by rainfall. The latter are more common, and are found in low, to medium plasticity materials on steep, weathered slopes, usually with angles of between 25° and 45°, but occasionally up to 55° (Campbell, 1975).

Debris torrents, involve similar materials (Vandine, 1985), but are confined to temporary stream channels. Hack and Goodlett (1960) call these 'chutes' and suggest they are incipient water courses (formed by surface runoff) which allow debris to move downslope by both mass movement and mass transport processes.

Debris slides often start as shallow (1-2m depth) translational, or rotational movements from upper parts of hillsides and like debris torrents undergo post-failure displacements reaching tens of metres. This is the characteristic feature of rainfall induced first time failure which causes natural disasters. There are examples from most countries, although the better publicised have affected populated areas. For example, Eisbacher and Clague (1981) list 27 disastrous events which occurred in the Vancouver region, Canada between 1900 and 1979 . Landslides (mainly debris flows in Pleistocene sediments) were triggered by daily rainfalls of between 31 and

124mm. They also describe an extreme event in 1979 which caused widespread damage from flooding and landslides with 2 day totals reaching 300mm.

Debris torrents are also a problem in the Vancouver region and have been described by Vandine (1985). These occur in areas of high relative relief and have caused an estimated $ 100 million of damage since 1962.

In 1977, between 250 and 300mm of rain fell in a 9 hour period in the Appalachians, Pennsylvania, triggering debris slides causing $ 300 million of damage and making 50000 homeless (Pomeroy, 1980). Jacobson et al (1989) describe a lower intensity, but longer duration event, also in the Appalachians in 1985, which resulted in 70 deaths and $ 1.2 billion of damage. Here the greatest intensity recorded was 38mm/hour with 2 day storm totals reaching 160 to 240mm.

In 1967 586mm of rain fell in a 48 hour period in Rio de Janeiro, Brazil (Jones, 1973), which caused tens of thousands of shallow slides and an estimated 1000 deaths.

In California, a storm in 1982 was resposible for 33 deaths and over 18000 slides with 440mm of rain falling in 32 hours (Ellen and Wieczorek,1988).

Fukuoka (1980) describes a storm in the Boso peninsula, Japan where 559mm of rain with a maximum intensity of 122mm/hour fell in three days. Shimizu (1988) records a comparable disaster in the San-In district, Japan where a rainfall intensity of 50mm/hour lasted for 10 hours, caused 100 deaths and damage estimated at 300 billion Yen. Similar extremes have been recorded in Hong Kong, for example 525mm fell in 1966 over a 24 hour period resulting in 35 deaths by landsliding.

One of the highest rainfall figures recorded was in Fiji, (Vaughan, 1985) where 1500mm fell in 3 days (the average annual figure is 2000 to 3500mm). Landslides (debris flows) occurred, although Vaughan (1985) suggested that the island did not have a long history of rain-induced mass-movement.

European examples of rain induced mass movement can be found in the alpine parts of Switzerland, Austria, Italy, France, Greece and Spain. Similar rainfall figures are involved. Over 200mm was registered on November 7/8, 1982 in the Eastern Pyrenees, Spain which resulted in extensive structural damage and the loss of 30 lives (Corominas and Moreno, 1988). Capecchi and Focardi (1988) suggest slides in the Tuscany region of Italy are triggered by 24 hour rainfalls greater than 100mm. A figure which is similar to the threshold suggested by Moser and Hohensin (1982) for Austria.

Rain induced landslides are also known in the UK. For example on the 10th July, 1968 a daily rainfall total of 172mm was recorded near Bristol (Hawkins, 1973) with peak intensities of 37mm/hour. There were 8 deaths, 15 bridges lost and numerous landslides in periglacially disturbed Lower and Middle Jurassic rocks. However, perhaps the most well known storm disaster occurred in Lynmouth on the 15th August 1952 where 228mm of rain fell in 24 hours. There were 30 deaths, flooding, dam and embankment failures and extensive natural landsliding. (Green, 1955, Kidson and Gifford,1953).

Man made soil structures are also susceptible to rain induced failures, involving fill and sometimes the underlying foundation. The rainfall records for a number of colliery tip failures in South Wales were reviewed by Bleasdale (1969) and published as part of the Aberfan flowslide tribunal. He found that the 1960 Abercwmboi colliery slide could be directly related to a 24 hour rainfall of 137mm, whereas the Aberfan disaster (21st October, 1966) was only preceeded by 70mm over two days. However, the analysis of the yearly records for the latter suggested that average monthly figures were greatly in excess of the mean annual totals. Bishop (1973) suggested slow artesian build up of groundwater in underlying Coal Measures sandstones may have been responsible for the slide which liquefied and moved nearly 600m down a 12.5° slope killing 140 people, including 112 children.

Similar uncertainties between the role of the natural foundation and fill led Kawamura and Sano (1989) to develop a link between runoff and landsliding in Japan, after rainfall intensities of 30mm/hour and 20 day totals of 500mm triggered a number of embankment failures.

### 1.2 Examples of Slides on Pre-existing Shear Surfaces

Pre-existing shear surfaces which cause landslide problems usually consist of medium to high plasticity clays. They are found either at the base of ancient slides or within sedimentary sequences which have undergone tectonic deformation. Renewed movements can take place on lower, overall slope angles than those described above (generally 10 to 20°). Reactivation follows small changes in strength and rainfall may be one of several factors leading to periodic mass movement. Displacements often average no more than a metre a year, but these may result in a long history of engineering problems. Typical slide examples are known from i) East Pentwyn in South Wales (mean rainfall 710mm) where about 60m of movement have been recorded in a debris slide on a 12.5° slope since 1954 (Gostelow,1977). ii) The 5.3km long Drynoch slide in British Columbia where the mean annual rainfall is only 255mm. but seasonal movements have averaged between 1.5 and 4m on a slope of 8.7° (Vandine, 1980). iii) Iverson and Major (1987) have measured annual movements of 1.5m in an earthflow, 750m long on a 15° slope in California over a 3 year period (1982-1985). iv) Skempton et al (1989) summarise seasonal movements in an 850m long earthflow/slide at Mam Tor, Derbyshire, UK (mean annual rainfall 1250mm) where radiocarbon dating and historical records suggested that 320m of movement on a 12° slope occurred in 3200 years.

The 1979 Abbotsford slide near Dunedin, New Zealand is a good example of a large destructive slide caused by a bedding plane shear surface (Bell and Pettinga, 1988). A thin (50mm) montmorillonitic volcanic ash band dipping at between 6° and 10° and sandwiched between two sand beds at a depth of 22m was responsible for a 500m long translational block slide which destroyed 69 houses. The rate of displacement gradually increased from 15mm/day to 300mm/day over a 40 day period before a catastrophic movement of 50m occurred. A number of man made and natural triggering causes were considered during the subsequent investigation, and it was concluded that rainfall was an important factor. The large post failure displacement suggested there was a sudden loss of strength, either as a result of the stress-strain characteristics of the clay or an undrained porewater pressure increase. It is interesting to note however, that the previous slide examples all have a positive relief in relation to the surrounding ground level, while the Abbotsford slide includes a substantial, passive, 'first time' shear zone. The latter will influence the deformation response following a strength reduction from increased groundwater pressures.

## 2. Triggering Mechanism

### 2.1 General

The rainfall triggers mass movement in both slide categories by raising groundwater or air pressures ($u_w$) and ($u_a$) respectively, following infiltration. This reduces the effective shear strength s, which, for saturated soils is commonly expressed by

$$s = c' + (\sigma - u_w) \tan \varnothing'$$

and for unsaturated soils (following Fredlund, 1987) by

$$s = c' + (\sigma - u_a)\tan \varnothing' + (u_a - u_w)\tan \varnothing_b$$

where c' is the soil cohesion, $\sigma$ is the total normal stress, $\varnothing'$ is the friction angle $u_a$ is pore air pressure and $u_w$ is the pore water pressure. The friction angle $\varnothing_b$ is equal to the slope of a plot of matric suction ($u_a - u_w$) versus shear strength when ($\sigma - u_a$) is held constant. This triggering mechanism applies to both first-time and pre-existing slides, but the stress-strain behaviour of the landslipped materials differs and this controls the subsequent downslope displacements. The mechanism can be divided into four interconnected steps:

1. The storm, or rainfall event

2. Infiltration and increase in water pressure in the slope

3. Reduction in shear strength

4. Failure and displacement along the shear surface

Debris torrents can be mobilised by the flow of rainfall run off within a channel deposit. Here the height of water, combined with the slope angle may be sufficient to cause bedload transport and movement of individual rock particles. According to Vandine (1985) this process can take place at lower gradients than infinite slope type shear failures.

### 2.2 Rainfall

#### 2.2.1 General

Rainfall is usually classified as convective, orographic or cyclonic. The meteorological conditions which cause high rainfall events and mass movement are able to develop within each class. Measurement is comparatively easy, and research has been aimed at finding a suitable index or threshold. Typical approaches (NERC Flood Studies Report, 1975) are to determine the depths which may be reached or exceeded.

i) in a certain period of years

ii) as a probability of occurrence

iii) in a given return period or recurrence interval

The risk (probability P) of encountering a mass movement rainfall event within a given period of time (n) for different return periods, T, can be obtained from

$$P = 1 - (1 - 1/T)^n$$

For example, the probability of a landslide inducing rainfall with a return period of 10 years occuring in a three year period is

$$1 - (1 - 1/10)^3$$

which is 0.271, or 27.1 %.

A series of extreme rainfall events (R) can be ranked from m = 1 for highest, m = 2 etc in descending order and the recurrence interval T, calculated from one of several formula, for example

$$T \doteq n + 1/m$$

where m = ranking and n = number of events. The probability P of an event occurring in a return period T is

$$P = 100/T$$

so that R can be plotted against T or P on normal or log normal probability paper (if the series are normally distributed). However, there are other frequency distributions used in hydrological forecasting which could be applied to landslide studies. For example Gumbel's extreme value approach is often used for analysing river discharge figures for flood prediction (see detailed discussion in NERC, 1975). This assumes that a number of extreme rainfalls (R) are exponentially distributed. The cumulative probability P' that any of the n values will be less than a particular value in a return period T then approaches the value

$$P' = e^{-e^{-y}}$$

where e is the natural logarithm base and

$$y = -\log_e[-\log_e(1 - 1/T)]$$

The event R of return period T years is given by

$$R = R_{average} + \sigma\,(0.78y - 0.45)$$

where $R_{average}$ is the average of all rainfall events and σ is the standard deviation of the series. R and T plot as straight lines where the distance y (the reduced variate) is given equal spacing on probability graph paper.

A rainfall triggering parameter R is thus required for estimating likely return periods of landslide events (first-time or reactivated). Rainfall measurements are made in single gauges, or networks, each gauge being representative of a particular area. They are summarised as daily totals, monthly averages, seasonal averages, annual totals and as an intensity. The most appropriate landslide triggering parameters (R) for different levels of slope susceptibility and climates may differ, but must be selected from these measurements.

Figure 1 (and table 1) lists the greatest recorded rainfall/duration depths on a worldwide scale, for different periods, together (for comparison) with records from the UK. They suggest there is an approximate relationship, possibly related to local climate, between magnitude (M) and duration (D) of the form $M = D^{f}$.

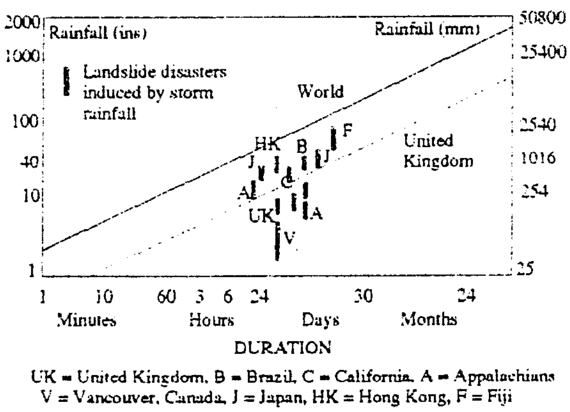



Figure 1. Plot of maximum rainfall and durations for the world and the UK, showing figures for landslide disasters

The figure shows that landslide disasters (section 1) tend to be triggered (or follow) high intensity rainfalls which last from between 9 and 72 hours, ie a certain duration and total must be exceeded. They are not however, always the most extreme for a given climate and the plot suggests there is a wide range of rainfall capable of inducing (triggering) landlides. The meteorological and geographical conditions which favour these intensities are generally found where orographic influences accentuate convectional or frontal/cyclonic rainfall. Unfortunately lack of space prevents a detailed discussion of the mechanisms and countries involved.

| Duration | Depth,mm | Location | Date |
|---|---|---|---|
| 1 min | 30.8 | Unionville, USA | 4.7.56 |
| 15 min | 195.8 | Plumb Pt, Jamaica | 12.5.16 |
| 12 hrs | 1324.3 | Belouve, La Reunion | 28.2.64 |
| 24 hrs | 1847.4 | Cilaos, La Reunion | 15/16. 3. 52 |
| 2 days | 2470.4 | Cilaos, La Reunion | 15/17.3.52 |
| 7 days | 4061.4 | Cilaos, La Reunion | 12/19.3.52 |
| 1 month | 9190.1 | Cherrapunji, India | 7.1861 |
| 4 months | 18561.3 | Cherrapunji, India | 5/7 1861 |
| 1 year | 26471.6 | Cherrapunji, India | 8/1860-1861 |
| 2 years | 40783.3 | Cherrapunji, India | 1860-1861 |

Table 1. Maximum recorded rainfalls and duration

Landslides are also known from the locations in table 1, for example Antoine et al (1988) provide an account of the catastrophic slides on the volcanic island of La Reunion.

### 2.2.2 First time Slides

Most of the detailed work on empirical relationships

between rainfall and first time landsliding has been carried out in America ( mainly the coastal zone of California) and in Hong Kong and this section draws heavily from these sources.

Brown (1988) recognises two different rainfall patterns in California. The first is associated with regional storms of several weeks duration. The second is a local high intensity rainfall lasting for a few hours or days. Regional storms cover larger areas than local storms and Brown (1988) suggests they are associated with deep seated, slower moving slides (probably pre-existing), while local storms trigger shallow first time debris slides.

Fig 2 from Campbell (1975) shows a typical cumulative rainfall plot for a local storm where 660mm of rain fell in 7 days. The curve is divided into rising and horizontal segments and illustrates how most rain falls within short periods of high intensity. In this example debris flows were initiated after the antecedent total reached 266mm and the intensity increased to more than 6mm/hour.

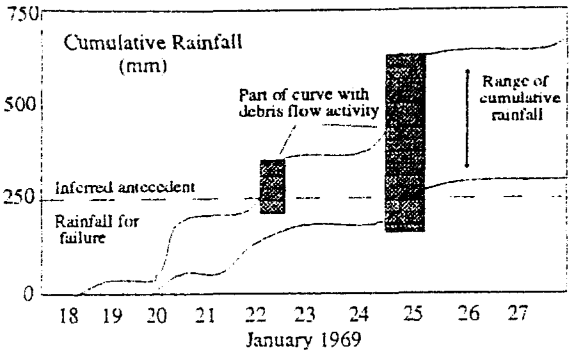



Figure 2. Cumulative rainfall plot for a storm in California showing times of debris slides (After Campbell, 1975)

This cumulative rainfall storm pattern is commonly associated with landsliding and has been recorded in New Zealand (Phillips,1988), Japan (Shimizu, 1988, Kawamura and Sarno,1989), Canada (Eisbacher and Clague, 1981) and Hong Kong (Brand et al, 1985). Figure 3 from the latter is an example which also shows how landslides are mainly associated with storms which contain 'rising intensity limbs' - known in Japan as J-type storms.

Brand (1985) points out that the majority of slides in Hong Kong are induced by local, short duration, spatially dependent rainfalls of high intensity. They provide evidence that storm rainfall totals can vary from as little as 10mm to over 200mm in a distance of only 3km.. They quote intensity rates of 70mm/hour as a threshold figure and suggest that 24 hour rainfall totals are the most useful warning of landslides because they include a sufficient time span to cover these events. They conclude that a) the antecedent rainfall (before the triggering event) is less useful for defining thresholds and b) that a 24 hour rainfall of less than 100mm is unlikely to result in major landslides. This latter figure has also been suggested by Heath and Saraso (1989) as a threshold for slides in Java, Indonesia.

Church and Miles (1987) have analysed rainstorms and debris slides in British Columbia, Canada and found that the 24 hour precipitation threshold of between 50 and 150mm was not exceptional for the climate. Expected return periods (section 2.2.1) were

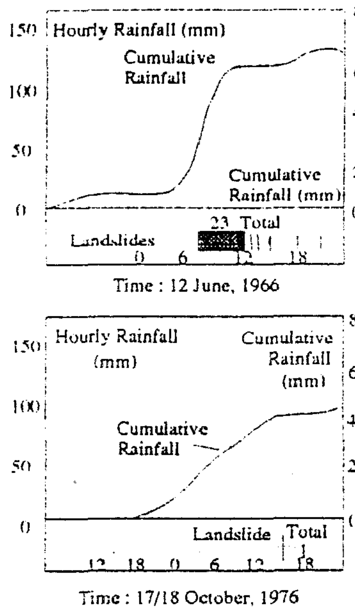



Figure 3. Cumulative rainfall curves for storms in Hong Kong showing the influence of intensity rates on landslide numbers (After Brand, 1985)

around 3 to 5 years. Maximum, measured intensities within storms were about 20mm/hour decreasing to 10mm/hour over a duration of 5 hours. They suggested, following Brand (1985) that general figures could be misleading and that debris slides were caused by locally intense, short periods, of precipitation which were not necessarily observed by a widespread rain gauge network. These were thought to be derived from small convection cells embedded within larger storms and they suggested that local topography in relation to the storm track was an important rainfall control.

Neary and Swift (1987) report similar threshold figures from the S. Appalachians, ie a 24 hour threshold of 125mm, but with peak debris slide rainfall intensities of between 90 and 100mm/hour.

Several attempts have been made to generalise intensity (i) - duration (D) relationships. For example Wieczorek and Sarmiento (1988) define a relationship,

$$i = 1.7 + 9.0/D$$

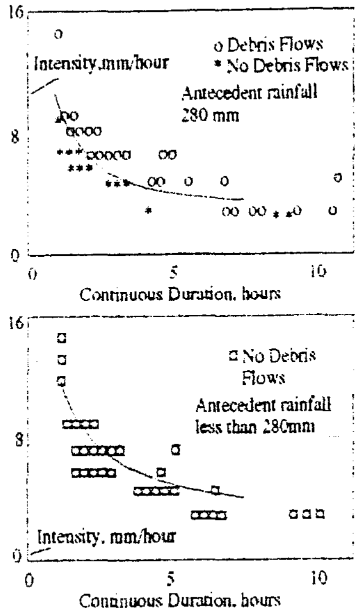



Figure 4. Landslides and intensity-duration thresholds in California showing the importance of antecedent rainfall (After Wieczorek and Sarmiento 1988)

in California (fig 4), but found that an antecedent rainfall of 280mm was required before the triggering threshold could be used.

Govi and Sorzana (1980) suggested that the normalised rainfall (the ratio of storm rainfall to mean annual precipitation, MAP) is a better way of presenting data. Cannon (1988) have used this approach in

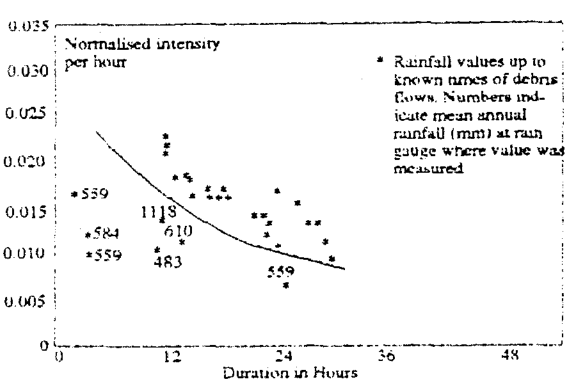


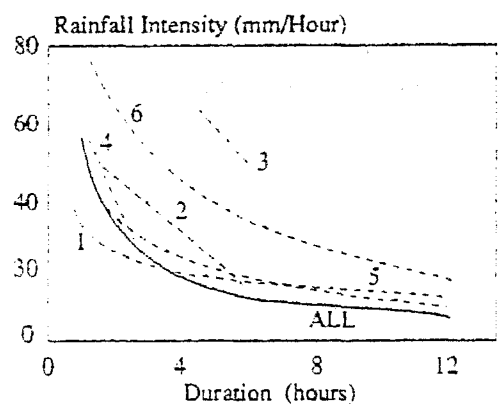

Figure 5. Normalised intensity-duration threshold for andslides in California showing influence of mean annual precipitation (MAP) (After Cannon, 1988)

California by dividing the rainfall intensity by the MAP at different recording raingauges. Fig 5 is a plot of the normalised intensity (In) against duration (D) which defines a threshold between sliding and not sliding (based on six storms) of

$$D = 46.1 - 3.6 \times 10^{-3} In + 7.4 \times 10^{-4}$$

The points below the curve are In-D values which triggered landslides in areas with a low MAP (values indicated) and Cannon suggests the approach is less well defined in areas where the MAP is less than 508mm.

Jibson (1989) has reviewed these different intensity-duration approaches using data from the countries listed on Figs 6 and 7. The results suggest the form and range of the curves are similar, regardless of mean annual rainfall, although the relationships differ in each country. According to his figures the lowest intensity threshold is in Hong Kong and the highest in Puerto Rico. Jibson (1989) suggests this is because landscapes with higher annual rainfalls adjust to




1 = California, 2 = China, 3 = Indonesia
4 = Brazil, 5 = Hong Kong, 6 = Puerto Rico
7 = California (ALL is a lower bound to all data sets) - numbers refer to figures 6 and 7

Figure 6. Rainfall intensity relationships for landsliding (After Jibson, 1989)

different states of geomorphological equilibrium than those with lower mean annual rainfall.

Caine (1980) has also plotted intensity (mm) against

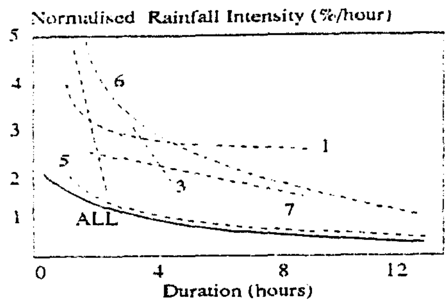



Figure 7 Normalised intensity-duration for landslides (After Jibson, 1989)

relationshipsduration (hours), using worldwide data and arrived at a lower bound to all results of,

$$i = 14.82 \times D^{-0.39}$$

Cannon (1988) points out that the role of pre-storm (antecedent) rainfall is unclear in these intensity/ duration relationships.

Fukuoka (1980) takes this into account by plotting rainfall intensity against preceeding total rainfall (PT) to derive thresholds in Japan. For example,with a PT of 300mm, an intensity of 35mm/hour was required compared to 50mm/hour for a PT of 100mm. A similar approach was used by Lumb (1975) in Hong Kong who preferred to use the 24 hour rainfall plotted against the 15 day antecedent rainfall. He found that severe and disastrous events ocurred when the latter exceeded 100mm and the 15 day total was greater than 200mm. Crozier (1985) has extended this approach by plotting the soil water status (EPa) for a 10 day period against 24 hour rainfall in New Zealand. EPa is the daily rainfall in excess of potential evaporation and storage requirements. Figure 8 is a self-explanatory example of the approach showing the landslide threshold for Wellington City.

Mark and Newman (1988) simply use cumulative pre-storm seasonal rainfall as a guide to landsliding in California. They suggest that 300 to 400mm is required, or 30% of the mean annual precipitation. Similar, preliminary studies in SW Cyprus (mean annual rainfall 443mm) suggest landslides take place

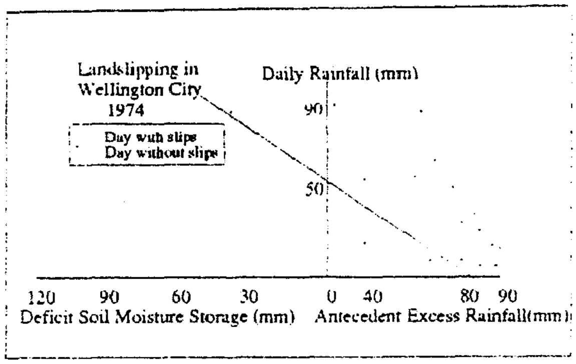



Figure 8. Landslide threshold for Wellington, New Zealand showing importance of soil moisture deficits and antecedent rainfall (After Crozier, 1985)

after a cumulative antecedent winter rainfall of at least 250mm, ie with more than 56% of the annual mean (Gostelow and Loucaides, 1988, Northmore et al, 1986)

## 2.2 Slides on Pre-existing Shear surfaces

Rainfall induced movements on pre-existing shear surfaces cause engineering problems, but these are generally less destructive than debris slides (this statement may not apply to movements on tectonic surfaces which include an element of first time failure). Landslide boundaries are usually known, leaving seasonal displacements and their control as problems for the engineer. Pre-existing slides are usually larger and deeper than the first time slides described above and may have a failure surface tens of metres below the ground surface. Slope deformation is thus generally controlled by groundwater recharge, ie by fluctuations in the permanent groundwater table (section 3). Rainfall thresholds may thus differ from those described in section 2.1.

There are many descriptions of slides and their remedial measures which refer to movements and rainfall in a general way, but few specific studies which link them over a long period of time. An exception is the recent analysis of the Mam Tor slide by Skempton et al (1989). A statistical review of climatic results (using figures from the NERC Flood Studies Report and movement records from 1915 to the present has shown there is a 50% probability of movement when total monthly rainfall reached 200mm/month. This figure is reached every 3 years with corresponding three day, six day and 10 day totals for the same period of 55, 80 and 105mm respectively. The results also suggested thatf movement was virtually inevitable when the 3 day rainfall reached 100mm, or monthly total 310mm: an event with a return period of 50 years.

### 2.3 Discussion

Empirical studies of this kind do not provide a link between rainfall and the water pressure increases which actually trigger slide movements. Neither do they take into account local variations in geology and geomorphology. Nevertheless they have proved to be a useful guide to planners and engineers. For example the relationships established for California and Hong Kong have enabled automatic rain gauge networks to be used as hazard warning systems. However, the different thresholds reviewed here show a) there is no agreement on the best method of relating rainfall figures to mass movement and b) there is no simple relationship which can be applied generally.

The four broad conclusions which can be drawn from the results are that

i) if an antecedent monthly rainfall total exceeds c. 250mm that any high intensity rainfall event within a storm can trigger shallow first-time slides in susceptible slopes - a conclusion deduced originally by Campbell (1975).

ii) high intensity falls greater than 70mm/hour within storms trigger first-time debris slides.

iii) 24 hour totals may be most useful for forecasting, with landslide (first time and pre-existing) thresholds of around 100mm.

iv) Pre-existing slides may have different threshold relationships.

The application of 'extreme value' statistics to 24 hour totals coupled with an analysis of their spatial distribution in relation to geology and geomorphology seems to be a useful approach in empirical studies, particularly where the boundaries of a landslide are known. Rainfall events can then be related to past ground movements and used for engineering prediction, as at Mam Tor (Skempton et al, 1989). The method is less useful for predicting where first time slides will take place. The latter also depends on the geological properties of the underlying soils and rocks and on the groundwater conditions.

## 3. Infiltration and Increase in Water Pressure

### 3.1 General

Figure 9 is a cross section through a typical soil/weathered rock horizon which shows the three major groundwater zones. Below the water table the ground is saturated and the positive pressure in the soil pores or rock joints increases hydrostatically with depth. The zone immediately above the water table is also saturated, but the hydrostatic pressure is negative (relative to atmospheric or zero gauge pressure) because of capillary action. The unsaturated zone continues to the ground surface. The water pressure in the pores is also negative, but they contain both air (usually assumed to be continuous with the atmosphere) and water. Large suctions, derived from surface tension forces at the curved air-water interface can develop between soil grains. The effective pressures

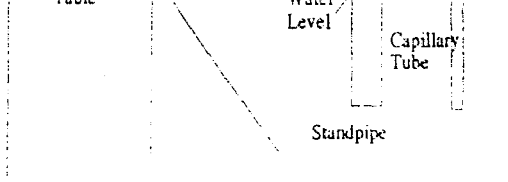



Figure 9. Section showing the three main groundwater zones

(and strengths) can be high in the two zones above the water table, and, without rainfall infiltration these may be capable of maintaining steep natural slope angles.

The figure also shows that depending on the combination of geomorphology and strength of the soil/rock layers that slides can be initiated in any of the three water pressure zones. A prediction of where, can only be made by measurement, or by using a physical model.

### 3.2 Regional Groundwater Flow

The elevation of the saturated zone, or water table depends on topography, geology and climate. Rainfall infiltration is largely controlled by the latter and can be defined, following Freeze (1969) as the movement of water from the ground surface towards the saturated zone. Recharge is water entry into the saturated zone, and discharge is the movement of water towards and across the water table surface.

Toth (1962) has identified three types of groundwater flow system which are driven by differences in topography, ie

1. A local system with a recharge area at a topographic high and a discharge area at an adjacent topographic low.

2. An intermediate system with one or more topographic highs between the recharge and discharge areas.

3. A regional system where the recharge area occupies the basin's water divide and the discharge area lies at the bottom of the basin.

Local geology and the three dimensional nature of groundwater flow within a drainage basin can cause variations in this pattern, but the water table always tends to be closer to the ground surface in discharge areas and these are often associated with landslips (Freeze and Cherry, 1979). A number of groundwater models have been developed which are capable of estimating piezometric heads (eg Freeze and Witherspoon, 1968, Freeze, 1971 and Freeze and Cherry, 1979, Beven, 1985, Calver, 1988). A recent application, linked to field mapping of recharge and discharge features is described by Ophori and Toth (1989). This approach, which combines theory with geological and geomorphological evaluation seems to be a promising method of identifying areas susceptible to landsliding.

### 3.3 Water Pressure Changes in the Unsaturated Zone and Groundwater Recharge

#### 3.3.1 General

There is a widespread literature on water movement in the unsaturated zone aimed principally towards hydrological and agricultural applications. Few of these studies have been adapted for slope stability purposes, although some hillslope hydrology texts have considered this (Kirkby, 1978 Anderson et al 1989).

Fig 10 is a typical moisture content-pressure relationship of a soil above the water table. It shows that suction depends on wetting and drying cycles and is very sensitive to increasing moisture contents. As the latter increases during rainfall infiltration, suction and hence strength decreases. The most comprehensive review of this process is given by

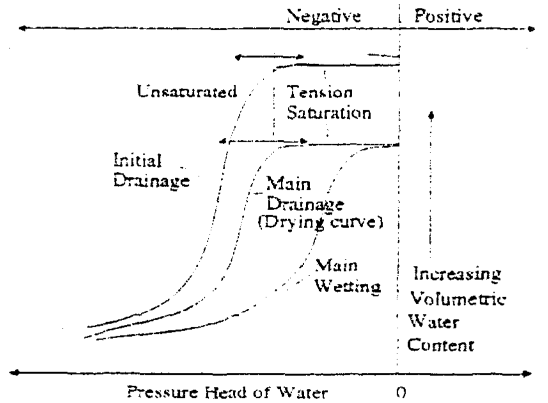



Figure 10. Diagrammatic water content/ pressure head relationships in the unsaturated zone

Freeze (1969). He suggests that groundwater recharge and discharge conditions in different climates create different soil moisture conditions which may vary across a drainage basin, even in homogeneous soils.

Water table fluctuations result when the rate of groundwater recharge or discharge is not matched by the unsaturated flow rate following infiltration or evaporation. Freeze's model shown in fig 11 assumes continuous moisture flow from the ground surface to a point just below the water table. The upper boundary condition is

$$dp/dz = R/K(p)$$

where R is rainfall rate per unit area, K(p) is hydraulic conductivity which depends on p, the pressure head, and z is depth below ground level. The lower boundary condition is given by

$$dp/dz = Q/K$$

where Q is positive and negative for recharge and discharge rates respectively and K is the saturated value of hydraulic conductivity. For flow in the unsaturated zone

$$\partial/\partial z\ [\ K(p)\ (\partial p/\partial z + 1)\ ] = C(p)\ \partial p/\partial t$$

and for flow in the saturated zone

$$\partial^2 p/\partial z^2 = 0$$

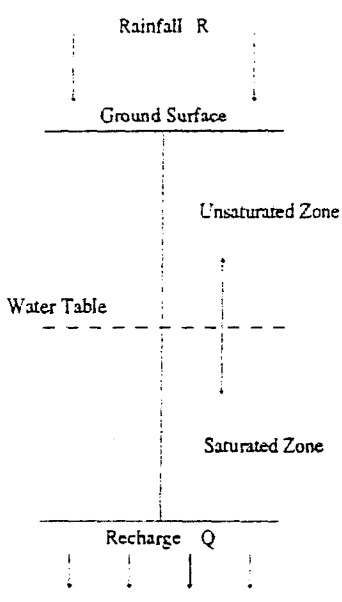



Figure 11. Infiltration model through the unsaturated zone and recharge to the water table (After Freeze, 1969)

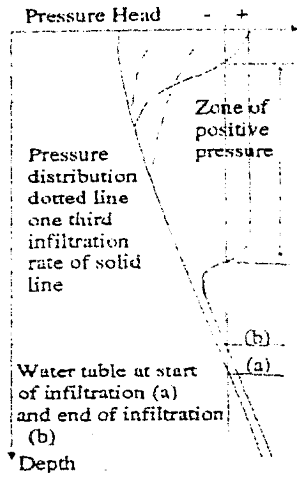



Figure 12. Pressure head variation with depth after infiltration into the unsaturated zone (After Freeze, 1969)

The parameter C is the specific water content which varies with pressure head, p. Fig 12 from Freeze (1969) shows the results for a constant recharge Q, a slow infiltration (dotted line) and fast infiltration (solid line). The higher intensity results in ponding and a water table rise while the slower rate does not (in this example the same quantities of water are introduced into the ground by maintaining the slower rate for a longer period). A small positive pressure develops in the unsaturated zone with the fast rate - ie the suction is completely lost. Figure 13 illustrates the redistribution of water after infiltration finishes. The solid and dotted lines represent infiltration at the same rate but for short and long periods respectively. Redistribution for the short period which did not result in ponding failed to cause a water table rise. However the longer infiltration period resulted in surface ponding and this led to a water table increase and a reduction in suction. Freeze (1969) also found that the higher the antecedent moisture content the

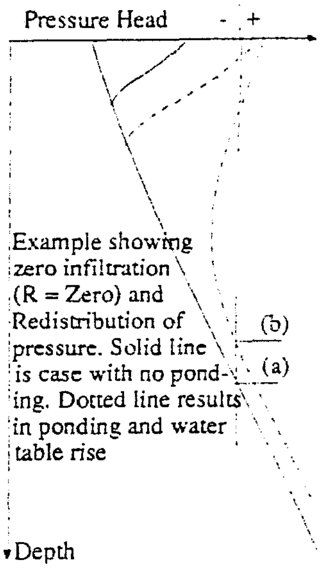



Figure 13. Redistribution of pressure head following cessation of rainfall (After Freeze, 1969)

greater the rate and height of water table rise. Of interest is the result that at certain recharge rates (Q) a lowering of the water table took place before the eventual rise. With (Q) acting as discharge, ie with a negative rate of flow the water table increase was also greater.

These results and brief discussion suggest there are two separate processes which affect shallow soil stability.

i) A loss of suction in the unsaturated zone, ie a comparatively large change in pore pressure which is accompanied by a loss of shear strength. This process seems to be responsible for most shallow first-time soil slides.

ii) A rise in the water table which takes place after a delay, and results in a smaller change in pore pressure (comparatively) and hence strength. This process seems to be responsible for both first time slides, and more commonly, slides on pre-existing shear surfaces.

Several, inter-related, physical parameters are involved in both processes.

1. The attainment of ponding conditions at the ground surface

2. The antecedent moisture content

3. The value of Q, ie either recharge or discharge

4. The depth of the saturated zone

5. Rainfall intensity

6. Hydraulic conductivity

It is interesting to note from figures 12 and 13 that in soils with a uniform (homogeeous) permeability the antecedent moisture content conditions affect the rate at which suction is lost in the saturated zone and the magnitude of the water table rise. This suggests that, if shallow slides (above the water table) are triggered by pressure changes in the unsaturated zone (from a high intensity event) the antecedent conditions affect the duration rather than the level of rainfall intensity required for mass movement.

**3.3.2 Pore water pressure changes in the unsaturated zone**

A simple approach often applied to slope stability problems in the unsaturated zone is to neglect the effects of partial saturation and assume that a fully saturated wetting front advances into an incompressible soil by gravity (Lumb, 1975, Vaughan, 1985). The velocity at which the front advances is given by:

$$V = -k/n \, \Delta S$$

where k = the coefficient of permeability of the soil, n = the porosity and $\Delta S$ = increase in saturation. It is assumed the front induces a complete loss of suction and an increase in pore pressure (or piezometric head) when it meets the water table. The thickness, H of the wetting zone after a rainfall time t is given by

$$H = kt/n \, \Delta S$$

The value of S is difficult to determine and in Hong Kong the head increase, H is often assumed to be 2m Brand (1985). .

Vaughan (1985) has extended this model (assuming a porous medium) by introducing a decreasing permeability with depth, on a sloping surface, ie where

$$k = k_o.e^{-A.z}$$

and $k_o$ = permeability at the surface, z = depth and A is a constant. He takes typical values from Hong Kong, ie where $k_o = 5x10^{-5}$ m/s and A = 0.5 $m^{-1}$. The model, shown in figure 14 assumes a vertical flow R per unit area passes into the soil under a head, h so that oh/oz = - R/k (the same boundary condition as Freeze, 1969). A solution, in the form of a typical transient flow net and pore pressures is also shown in figure 14. The plot shows that substantial positive values (approaching those with steady seepage parallel to the slope) are predicted as a wetting front passes through the soil layers. A perched water table is formed and transient pressures are greater than those predicted by Freeze (1969) (figs 12 and 13). A similar effect is obtained if a soil layer of lower permeability is included in the model. Vaughan (1985) suggests that when the front meets the water table the latter rises by the thickness of the zone (comparable to the assumption made by Brand, 1985). However, many soils have dual hydraulic properties which are derived from a) the intergranular medium and b) a secondary structure such as fractures, discontinuities, pipes etc. which are collectively known as macropores. The permeability in the latter is often greater, and the porosity less than in the surrounding soil grains and this can result in a rapid and even higher head response, which may be localised in part of a slope profile. These structures are difficult to include within a groundwater model, and most predictions of pore pressure change assume a porous medium

Fig 14 shows that as a saturated wetting zone travels downwards it must displace the pore air ahead of it. A pressure gradient is thus set up and it has been suggested, for example by Freeze (1969) and Headworth (1972) that this may cause a pressure increase in the underlying saturated zone. This affect has not been investigated in natural slopes, but may be significant. The increased pore air pressure also reduces the effective strength of the soils in the unsaturated zone according to the equation in section 2.1.

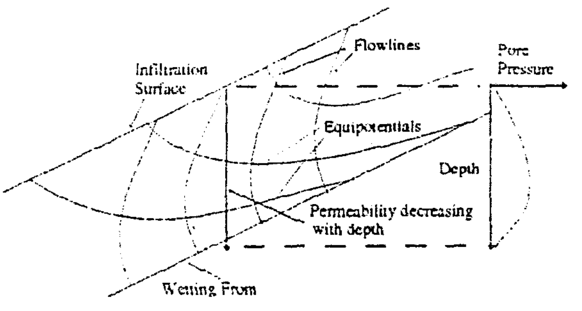



Figure 14. Equipotentials, flow lines and pore water pressures below a sloping surface with decreasing-permeability with depth (After Vaughan, 1985)

Very few, seasonal, field measurements of suctions in the unsaturated and tension saturated zone have been made for slope stability studies. Comparison with the models described here is therefore difficult. However, Fredlund (1987) has published some suction profiles from Hong Kong. Fig 15 from his paper shows measurements above a water table 30m below ground level. They vary considerably as a result of rainfall and evaporation (by nearly 40 kpa). According to Fredlund the profile of 27.10.81 was c. 10% to 20% of hydrostatic which maintained a factor of safety just above 1.0.

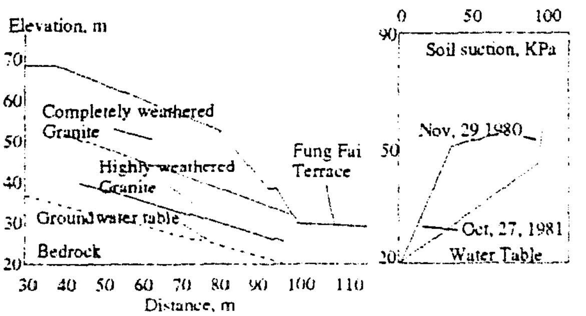



Figure 15. Variation of soil suction below a residual soil slope (After Fredlund, 1987)