THE LANDSLIDE RESPONSE OF ALPINE BASINS TO POST- LITTLE ICE AGE
GLACIAL THINNING AND RETREAT

by

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ABSTRACT

The retreat of glaciers following the end of the Little Ice Age (since approximately 1850 AD) has been cited as a cause for increased landslide activity in alpine basins. The primary reasons for this are the combined effects of removal of ice buttressing from bedrock slopes and deposition of glacial drift in steep areas prone to instability.

Nineteen alpine basins along the upper Lillooet River valley, northwest of Pemberton, B.C., were studied to investigate the controls of post-Little Ice Age Neoglacial retreat on landslide activity. Terrain containing landslides was characterized by identifying how Neoglacial scouring and retreat have modified terrain, thus affecting slope stability. A decision-making flowchart was constructed to assist future identification of landslide hazards associated with ongoing glacial retreat. This work was based on field observation, GIS analysis, statistical associations between landslides and terrain attributes, and comparison of Neoglaciated and non-Neoglaciated terrain within each basin.

Examples of landslides influenced by Neoglacial retreat include debris slides and debris flows, rockfall, rockslides, rock avalanches, and, slow deep-seated slope movements. In bedrock, landslide response to post-Little Ice Age Neoglacial retreat ranges from severe in the Meager Creek volcanic area to low in many granitic rock basins. In general, the magnitude of landslide response depends on the intensity of glacial scour below the Neoglacial trimline. Basins underlain by weak volcanic rocks experienced significant oversteepening by Neoglacial scour, and active rockfall, deep-seated slope movements and large failures occur near glacial trimlines. Basins underlain by granitic rock rarely show increased bedrock instability resulting from Neoglacial retreat, except for shallow rockfall along some glacial trimlines and failures in previously unstable slopes.

In surficial materials, landslides associated with Neoglacial retreat occur in till and colluvium and are concentrated within the Neoglacial Limit along trimlines. Landslide processes in surficial materials are classified as primary and secondary Neoglacial effects. Primary effects involved evidence of direct Neoglacial influence on landslide activity, and included glacial undercutting of colluvial or drift embankments and deposition of glacial drift in areas prone to instability. Secondary effects involved deposition of drift material in locations entrainable by failures initiating upslope of the trimline.

The Neoglacial effects investigated in this thesis concern only one of many factors contributing to landslide activity in alpine basins. Nevertheless, the study results suggest that Neoglacial scour, retreat, and deposition are significant factors for increased landslide density in areas already prone to instability.
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CHAPTER 1. INTRODUCTION

The extent of alpine glaciers in southwestern British Columbia has fluctuated significantly since the end of the Pleistocene Epoch. Within the Holocene, at least three major ice advances have scoured valley bottoms and cirque basins over the last 6000 years, with the greatest advance occurring in the mid-nineteenth century during the Little Ice Age (Ryder and Thomson 1986). Since then, 20th Century climatic warming has caused substantial glacial retreat in most glaciated mountain areas (Grove 1988, Intergovernmental Panel on Climate Change 1990).

The objective of this thesis is to investigate effects of recent (post-Little Ice Age) glacial retreat on landslide activity in nineteen alpine basins near Pemberton, British Columbia (Figure 1). Several authors have cited recent glacial retreat as responsible for increased landslide activity in glaciated alpine basins, due to combined effects of the removal of ice buttressing from bedrock slopes and deposition of glacial drift in steep areas prone to instability (Evans and Clague 1993, 1994; Abele 1997; Berrisford and Matthews 1997; Ryder 1998). However, previous workers have only qualitatively cited examples of mass movements occurring in close proximity to recently deglaciated areas, or have conducted detailed studies of specific landslide events related to recent glacial retreat (Blown and Church 1985, Zimmermann and Haeberli 1992, Bovis and Stewart 1998, Bovis and Jakob 2000). No previous systematic investigation has ever been conducted in the Canadian Cordillera on the effects of recent glacial scouring and retreat on landslide activity.
Figure 1. Study basin locations and bedrock geology (Woodsworth 1977, Reed 1978).
1.1 Thesis Objectives

Landslide processes including rockfall, rockslides, rock avalanches, debris slides and deep-seated slope sagging features (sackung) were examined in seventeen glaciated alpine basins in the Lillooet River watershed and two sub-basins within the Rutherford Creek watershed. The objectives of this thesis are as follows:

1. Identify the terrain characteristics in which landslides most commonly initiate in recently deglaciated areas. Terrain characteristics are referred to as terrain attributes that are associated with landslide activity. This requires terrain mapping and a detailed landslide inventory using methods based on Howes and Kenk (1997). Air photo interpretation, field work, and GIS analysis are used to subdivide terrain based on terrain attributes that are statistically related to landslides.

2. Identify the glacial influence on landslide activity by identifying how recent glacial scour and retreat have modified terrain, thus affecting slope stability. This requires interpretation of field data and statistical results via a process-oriented view of slope instability in recently deglaciated areas.

3. Construct a decision-making flowchart to assist in the identification of landslide hazards associated with recent glacial retreat.

1.2 Study Limitations

In this thesis, research is limited to the effects of glacial retreat on landslide initiation. Landslide runout and secondary landslide processes will not be considered, such as debris flows that were triggered by upper basin slope failures. This project will also be limited to a study of landslide activity occurring at the present time. Temporal changes in landslide frequency, and possible landslide effects due to changes in hydrology following glacial retreat are also not examined. In addition, landslide magnitude is not studied in detail because individual landslide deposits were often eroded or obliterated, and because of difficulty in identifying whether landslide scarps were the result of sudden or slow retrogressive slope failure. Finally, no landslides with scarp lengths smaller than about
15 m were considered due to the difficulty of consistently identifying and cataloging such small failures.

1.3 Previous work on the effects of glacial retreat on landslide activity.

Several workers have identified connections between recent glacial recession and landslide processes such as rock slope deformation (sackung), rock avalanches, and debris flows. For example, glacial debuttressing has caused large scale rock slope deformation in many areas of the British Columbia Coast Mountains (Evans 1987a; Clague and Evans 1994b; Bovis and Evans 1995, 1996; Thompson et al. 1997); within the United States (Tabor 1971; Radbruch-Hall 1976, 1978; Varnes et al. 1989) and elsewhere (Zischinsky 1966, Beck 1968, Radbruch-Hall 1978, Savage and Varnes 1987, Dramis and Sorriso-Valvo 1994). Although the most active period of rock slope deformation probably occurred near the end of the last glaciation, gravitational creep of steep slopes apparently continues in recently deglaciated basins (Evans and Clague 1994). For example, Bovis (1982) identifies active sagging of glacially debuttressed bedrock slopes above Affliction Glacier, British Columbia, which has undergone at least 100 metres of downwasting in the last 150 years. Evans and Clague (1994) describe extensive sackung development above Melbern Glacier, St. Elias Mountains, British Columbia, which is a glacier that has thinned by 400-600 m in the last 200 years.

Several workers have also linked rock avalanche occurrences to recent glacial retreat. Evans and Clague (1988, 1993, 1994) note that 16 of 30 known historical (i.e. past 150 years) rock avalanches in the Canadian Cordillera occurred on slopes adjacent to glaciers, and some have detachment surfaces below Neoglacial trimlines. Examples include the Tim Williams Rock Avalanche near Stewart, British Columbia (Evans and Clague 1990) and the 1975 Devastation Glacier debris avalanche on Mt. Meager (Evans and Clague 1993). Plafker and Ericksen (1978) describe two catastrophic landslides from the North Peak of Huascaran that Evans and Clague (1993) argue were probably caused, in part, by recent glacial retreat. McSaveney (1992) describes a rock avalanche occurring above a glacier on Mt. Fletcher, New Zealand. Rupture occurred at the toe of an oversteepened
slope exposed by about 150 m of glacial thinning since the Little Ice Age maximum. In the past 100 years, at least three minor rock avalanches have occurred on slopes below Mt. Rainier, Washington, that were partially buttressed by glaciers during the Little Ice Age (O'Connor and Costa 1992). Post (1967) and McSaveney (1978) also cite examples of rock avalanches in Alaska that have occurred in close proximity to existing glaciers.

Workers have also examined connections between the recent exposure of glacial drift and debris flows (Eisbacher and Clague 1984, Haeberli et al. 1990, Jackson et al. 1989, Zimmermann and Haeberli 1992, Owen et al. 1995, Rebetez et al. 1997, Savoskul 1997, Benn and Evans 1998). For example, Jackson et al. (1989) documents debris flow activity in recently exposed glacial drift in the Canadian Rocky Mountains, and Zimmermann and Haeberli (1992) cite several similar examples from Switzerland. Benn and Evans (1998) suggest that the most active reworking of glacial sediment occurs immediately after deglaciation, when drift is unstable, unvegetated and there is an abundance of meltwater.

Previous work has also focused on debris flows and debris floods that were triggered by the failure of moraine dammed lakes (Costa and Schuster 1988). Catastrophic failure of these dams has caused large debris flows in many areas of the world, as in Peru (Lliboutry et al. 1977, Ericksen et al. 1970), United States (Nolf 1966, Laenen et al. 1987) and at least eight documented historical failures in the Canadian Cordillera (Ryder 1991, Evans 1987c, Clague and Mathews 1992, Clague and Evans 1992, 1994, 1995, Blown and Church 1985).

1.4 Previous work in the thesis study area.

Previous work in the thesis study area includes assessment of debris flow frequencies and magnitudes in lower valley areas and several studies of landslide activity in small watersheds surrounding Mt. Meager. Jordan (1987, 1994) and Jakob (1996) conducted debris flow studies on fans located below several of the study basins included in this thesis, including Devastation Creek, Capricorn Creek, Petersen Creek and "Rainy" Creek.
Jakob’s (1996) work included an assessment of debris flow frequencies and magnitudes and morphometric characterization of debris flow source areas, and Jordan (1994) focused on debris flow mobility. Although both authors cite glacial retreat as a likely factor influencing debris flow activity, neither author attempted to quantify this relationship.

The south side of the Mt. Meager area (Figure 1) has received more attention by researchers than any other location within the thesis study area. For example, Baumann Engineering and EBA (2000) described several locations within Devastation Creek involving large scale, catastrophic failure in glacially debuttressed areas, including a large rockslide/debris flow in 1975 that killed 4 BC Hydro workers. Bovis and Jakob (2000) conducted a detailed study of a recent large debris flow in Capricorn basin that most directly addresses the connection between glacial retreat and landslide activity. On July 30, 1998, a very large debris flow was triggered in upper Capricorn Creek when a debris slide initiated in thick colluvium in the upper cirque area. The initiation zone had been significantly oversteepened by glacial downwasting, and the authors cited glacial retreat and slope saturation from snowmelt as the two most likely factors causing failure.

Several previous studies have been conducted on non-catastrophic slope deformation in the Mount Meager area (Bovis 1982, Croft 1983, Smith and Patton 1984, Evans 1987b, Bovis 1990, Bovis and Evans 1996, Bovis and Stewart 1998). Bovis and Stewart (1998) discussed detailed field observations of rock slope failure and long-term (18 year) slope monitoring at Affliction Creek, and report results from a numerical model that tests the sensitivity of the slope to loading conditions such as glacial debuttressing and groundwater fluctuations. The authors found that glacial debuttressing induced long-term toppling and toppling-induced sliding controlled by a steeply inward-dipping joint set, and that groundwater fluctuations have produced bulk slope deformation that resembles slope creep.

On a larger scale, Holocene glacial processes can strongly influence fluvial processes. This has been termed the “paraglacial” influence of glaciers on fluvial systems (Church
and Ryder 1972). Lower valley flood plain aggradation can be a signature of paraglacial reworking of sediment in the upper part of a basin (Church and Slaymaker 1989, Ballantyne and Benn 1994). Recent sedimentation history in Lillooet River basin provides information on rates of paraglacial reworking of sediment (and, by inference, mass movement activity) since the Neoglacial maximum. Jordan and Slaymaker (1991) used a sediment budget approach to describe the sources, storage and yield of clastic sediment in the Lillooet River Watershed. They found that the Lillooet River delta began to advance much more rapidly after 1948, possibly due to paraglacial reworking of sediments exposed by Neoglacial retreat and routed through the alluvial valley. However, Kerr Wood Leidal (2001) suggests that artificial meander cuttoffs excavated in 1948-1950 were a more likely cause for delta aggradation, because of the resultant channel shortening (by 5.5 km) and associated increase in channel gradient that caused river downcutting and channel widening. Lillooet lake was also artificially lowered by 2 metres during this time, further steepening channel gradient by lowering the base-level. On a local scale, the alluvial fan at the confluence of Meager Creek and Lillooet river has aggraded in places by up to 2 metres in the past 6 years. Debris flows initiating in recently deglaciated terrain (particularly Capricorn Creek) may be the major factor for this observed aggradation (M. Jakob, pers. comm.).
CHAPTER 2. BACKGROUND

2.1 Landslide Hazard Mapping

The purpose of landslide hazard mapping is to subdivide terrain into areas with varying probabilities of landslide occurrence. Unstable terrain can be characterized by subjective judgement, statistical association between landslides and terrain characteristics, a physical understanding of factors influencing stability, or combinations of the three methods. In this study the investigated area is too large to permit detailed data collection for physically-based landslide hazard mapping. Consequently, I focus only on the statistical and subjective approaches in the following section.

2.1.1 Terrain stability mapping

Terrain stability mapping is the most common method of classifying terrain according to stability criteria in British Columbia (BC Ministry of Forests 1995). Terrain is subdivided into geomorphic units (polygons) that are thought to be homogenous with respect to characteristics that influence slope stability, such as slope gradient, surficial material type, and drainage characteristics. This is carried out by air photo interpretation and field checking and a slope stability rating is assigned to each polygon (Howes and Kenk 1997).

Terrain stability mapping is flexible and suitable for different scales of slope stability analysis. However, mapping accuracy is dependent on the skill and experience of the worker and results are therefore not easily replicable. For example, Carrara et al. (1992) and Carrara (1993) identified significant discrepancies in landslide hazard levels mapped by several workers for the same region in Northern Italy. Pack (1997) also highlighted similar problems in regional landslide hazard mapping in British Columbia, and I have experienced similar issues while conducting terrain stability mapping in southwestern British Columbia.
2.1.2 Probabilistic Landslide Hazard Mapping

Probabilistic landslide hazard mapping involves prediction of potentially unstable areas based on statistical associations between landslides and terrain characteristics in specific areas. The approach is based on the assumption that future landslides will occur in terrain similar to that in which landslides have already been documented. Production of landslide hazard maps based purely on statistical analysis is beyond the scope of this thesis, since some terrain data are relatively scarce (e.g. data on bedrock structure), and because some significant landslides appear to be singular events within the study area. For example, only one major failure occurred in thick colluvium (in Capricorn Creek). However, statistical methods are useful to obtain a ranking of terrain characteristics associated with landslides. This approach has been termed a terrain attribute study.

2.1.2.1 Selection of the terrain unit for analysis

One of the most important issues for terrain attribute studies (and landslide hazard mapping) is selection of the base-unit, or terrain unit, that is used to relate terrain attributes to landslide occurrence. Carrera et al. (1995) defines a terrain unit as “that portion of land surface which contains a set of ground conditions which differ from the adjacent units across definable boundaries.” Terrain units are the smallest level of resolution in a regional landslide study, and several approaches may be taken to characterize terrain into distinct units for analysis:

1. Manual techniques involve the use of air photo interpretation and fieldwork to subdivide the landscape into geomorphic terrain polygons. Skilled workers can subdivide landscapes into areas that have true geomorphic meaning. However, the choice of average terrain unit size and location is subjective, time consuming, and will rarely be replicated by two different workers. Indeed, it is challenging for a single practitioner to consistently delineate geomorphic unit boundaries throughout a large project area.
2. The grid cell approach involves subdivision of the landscape into a regularly spaced grid pattern (Carrara 1983, 1989). This method is convenient for raster GIS-based data analysis, but grid cells do not reflect geomorphically meaningful subdivisions of landscape. Using small grid cells (10 to 30 m) reduces this problem but significantly increases data storage requirements. In addition, some programs for statistical analysis may not be able to handle the large datasets that result from small grid cell sizes (Cararra 1995).

3. Unique condition units have been used by Chung et al. (1993) to subdivide terrain by overlaying all thematic layers in a GIS and produce analytical units that combine the attributes of each layer. These unique condition units are convenient for statistical analysis but may produce large numbers of small, meaningless slope units when many layers are overlayed (Rowbotham and Dudycha 1998, Chung et al. 1993).

4. Cararra et al. (1991, 1995) used a digital elevation model (DEM) to automatically subdivide terrain into slope units based on ridge and channel lines. This resulted in geomorphologically meaningful terrain units that consist of small basins and major slopes. Carrara successfully modeled landslide hazard with this approach, but Rowbotham and Dudycha (1998) point out that Carrara used characteristics of large slope units to model landslide activity occurring within only small parts of these areas. Alternatively, Rowbotham and Dudycha subdivided terrain into units based on not only ridge and channel networks, but also breaks in slope based on a model of surface water flow. This allowed subdivision of terrain into smaller slope facets that were homogenous with respect to slope gradient. Automated subdivision of terrain into geomorphic units is potentially very useful; however it is heavily dependent on the resolution and accuracy of the DEM.
2.1.2.2 Terrain attribute studies

In British Columbia, terrain attribute studies have focused on forestry applications (BC Ministry of Forests 1995). The most common approach has been to combine terrain stability mapping techniques with statistical analysis to identify associations between landslides and terrain attributes (Rollerson et al. 1986, 1992, 1997, 1998; Rollerson and Sondheim 1986; Howes 1987; Rood 1990, Jakob 1999, Roberts 2001). Terrain attribute studies in British Columbia have generally involved the following steps:

1. Data Collection:
   - Landslide inventory and geomorphic terrain stability mapping are completed using the conventions of Howes and Kenk (1997).
   - Terrain attributes, such as those in Table 1, are determined for each polygon.

2. Statistical analysis:
   - Terrain attributes are simplified by grouping each attribute into nominal or ordered classes within the attribute. Each polygon is assigned individual classes of attributes and defined as stable (contains landslides) or unstable (does not contain landslides).
   - Univariate and multivariate statistical methods are used to determine relations between the presence or absence of landslides and the terrain attributes.
Table 1. Terrain attributes that are often related to observed landslide activity in terrain attribute studies and landslide hazard mapping (Wadge et al. 1993, Rollerson et al. 1998).

Geographic Information Systems (GIS) are powerful tools for terrain attribute studies because they allow storage and processing of large volumes of data. Although most terrain attribute studies in British Columbia have involved simple cartographic output rather than actual analysis, several workers have employed GIS to evaluate landslide hazard in other mountainous areas (Gupta and Joshi 1990; Carrara et al. 1991; Guzzetti 1993; Wadge et al. 1993; Navarro et al. 1994; Carrara et al. 1995; Pack 1995, 1997; Dhakal 1999; Alzate et al. 1999). Most workers employed GIS to manage spatial data in terrain units, including terrain characteristics that were thought to be significant factors in landslide initiation. This information was then exported from the GIS for statistical analysis.

2.2 Statistical Methods for Terrain Attribute Studies.

The goal of statistical analysis in terrain attribute studies is to determine correlations between landslide events and terrain characteristics. The choice of method depends on the type of data and the purpose of the research. The principal methods fall into two groups:
2.2.1 Univariate Statistical methods

Univariate methods examine each terrain attribute separately in an analysis in order to provide relative hazard levels for different classes within an attribute (e.g. categorical slope classes within the attribute slope gradient). Univariate analyses are much simpler to perform than multivariate methods, but cannot be used to compare the importance of different attributes, or to determine how attribute combinations influence landslide activity.

"Failure Rate" (FR) analysis is a univariate statistical approach applied in this thesis (Section 6.1) that is useful for determining the relative contribution to landsliding of different classes within one attribute (Dhakal et al. 1999). A numerical score ("susceptibility score") is calculated that indicates the relative spatial frequency of landslides within each class of an attribute, normalized by the area of each class. The score is calculated as follows:

\[
\text{Susceptibility Score} = \frac{R(L)}{R(A)}
\]

Where \( R(L) \) is the ratio of the number of landslides in a class to the total number of slides in the study basin, and \( R(A) \) is the ratio of the area of terrain units in a particular class to the total area of terrain units in the study basin.

2.2.2 Multivariate Methods

Multivariate statistical methods allow consideration of combinations of terrain characteristics and identify both individual and combinations of variables that are associated with landslide activity. These methods can be used to either describe terrain on which landslides occur, or to predict areas of instability. In this thesis, the number of attributes potentially important for slope stability far exceeded the number that could be measured at a regional scale. Consequently, the purpose of statistical analysis in this
thesis is not to obtain a predictive model, but to obtain a ranking of the most important attributes that can be measured at a regional scale, and to use these results to strengthen a model for landslide hazard assessment that is based primarily on field observations.

Information collected for landslide hazard mapping commonly includes qualitative data and controlling variables that are not fully independent, thus limiting the types of appropriate multivariate methods. The two multivariate statistical approaches that were considered in this thesis include Wald forward stepwise logistic regression and classification trees.

2.2.2.1 Wald Forward Stepwise Logistic Regression

Wald forward stepwise, logistic regression (herein termed logistic regression) is a non-linear, non-parametric statistical method suitable for analyzing data where the sample size is limited, the predictor variables are not independent, the variable distribution is non-normal, and the response (dependent) variable is binary. Predictor variables can be either categorical or continuous. The method predicts an outcome of the dependent variable (e.g. landslide or no landslide) by fitting a non-linear equation to values of the predictor variables (terrain attributes):

\[ P = \frac{1}{1 + e^{-z}} \]

Where \( P \) = probability of a landslide
\( Z = C_0 + C_1X_1 + C_2X_2 + \ldots + C_nX_n \)
\( C = \) estimated coefficients
\( X = \) terrain attribute variable.

The program (e.g. SPSS Inc. 1997) builds the model by stepwise addition of terrain variables to the equation, and tests the fit of the model to the data at each addition. Redundant variables are also removed at each step in favor of better predictive
combinations. The outcome is a model that gives landslide probability values for
different combinations of predictor variables (see Section 6.2).

Logistic regression is potentially a useful tool for landslide hazard mapping because, in
theory, terrain characteristics could be entered into the model to determine a probability
of failure. However, success of the model is significantly limited by the quality of
predictor variables, and the model may be difficult to apply if it includes complex
combinations of many different terrain attributes.

2.2.2.2 Classification Trees

Like logistic regression, classification trees are a non-parametric multivariate analysis
tool that identifies the response of a dependant variable to several categorical or ordered
predictor variables. However, analysis is not conducted on the data as a whole; rather,
the program creates a branching “tree” that splits the data into subsets that classify the
data into landslide and non-landslide groups. Splits are accomplished as follows:

The first step is to determine the variable that is most correlated with the
presence/absence of landslides. The statistical method used to test correlation between
variables depends on the software package used; examples include Chi-Square Automatic
Interaction Detector (CHAID) (SPSS 1997) and discriminant-based univariate splits
(Statsoft 1999). In both methods, Chi-square analysis (for categorical data) or ANOVA
(for ordered or continuous data) are used to identify statistically significant associations
between controlling variables (terrain-attributes) and the presence/absence of a landslide.
Statistical significance is indicated by p-values\(^1\) for each controlling variable, and the
variable with the smallest p-value is selected to determine a split.

The next step is to determine how to split the data within the chosen attribute. The data
are split into subsets based on the combination of classes in the chosen attribute that best

\(^1\) A p-value measures statistical significance by indicating the probability that a correlation observed
between a predictor and dependant variable is simply a coincidence. A low p-value implies higher
confidence in the observed association.
distinguish stable and unstable terrain. For example, slope gradient might be selected as the most important controlling variable in step one (e.g. Section 6.3.1). Within slope gradient, if it is found that most landslides occur on terrain steeper than slope class “4”, the data would be split into two groups, slopes gentler than or equal to slope class 4 (lowest proportion of failed locations) and slopes steeper than slope class 4 (highest proportion of failed locations).

Steps one and two are performed repeatedly until no statistically significant controlling variables are found, or the sample is reduced to some minimum size. The final output is a tree diagram composed of “nodes” (predictor variables) and “branches” (splits) that identifies combinations of variables that occur on failed and stable terrain. This is a useful tool for landslide hazard assessment because such a diagram is visual, intuitive and readily applicable in decision making.

2.3 Rock Mass Strength Classification Systems

Glacial “oversteepening” of rock slopes is often cited as an important cause of rock slope instability following glacial scouring and retreat (e.g. Clague and Evans 1994). However, only a few previous workers have investigated this statement by evaluating the mechanical strength of glacially debuttressed rock masses (Bovis 1990, Bovis and Stewart 1996, Augustinus 1992, 1995).

A significant reason why little quantitative work has been done is lack of data: bedrock geologic maps typically show information on the type and age of rocks, not the engineering properties that affect slope stability. Several methods exist to make detailed engineering assessments of rock mass strength, such as the Hoek-Brown Criterion and the Geological Strength Index (Hoek 2000). Classifications include both field and laboratory analysis of bedrock characteristics such as joint orientation, roughness, width and continuity; uniaxial compressive strength of intact rock; and water pressure within discontinuities.
In this thesis, detailed estimates of rock mass strength were not practical due to the regional scale of work and because high rockfall hazard prevented access to many rock slopes. However, Selby (1980) developed an empirical rock mass strength (RMS) rating system that was suitable for limited use in this thesis (Table 2). The system incorporates seven factors that can be estimated in the field, including field estimates of intact rock strength (Table 3), degree of weathering, joint orientation with respect to the hillslope, joint spacing, width and continuity, and the presence of groundwater outflow. Each factor is given a numerical score according to its influence on slope stability, and the sum of scores at a particular rock exposure gives an estimation of rock mass strength.

2.4 Strength Equilibrium Rockslopes

An increase in landslide activity following glacial retreat can be thought of as a period of response as a hillslope re-equilibrates to different stress conditions. This response is generally a reduction in slope gradient to a more stable angle. Selby (1980, 1982) used the term “strength equilibrium rock slopes” to describe slopes that have eroded to a level where further mass movement is unlikely and weathering will cause retreat of the rock slope without any change in slope gradient. In general, strength-equilibrium slopes with high rock mass strength ratings tend to be steeper than slopes with low RMS ratings, assuming that there are no other outside controls on slope gradient (such as structural control).

Selby (1982) developed a tool for identification of strength equilibrium rock slopes by correlating rock mass strength with slope gradient for 98 strength equilibrium rock slopes in New Zealand, Antarctica and South Africa. He regressed RMS values on slope gradient and drew two lines parallel to the regression line that defined a “strength equilibrium envelope” for rock slopes in strength equilibrium. Further work by Moon (1984) and Abrahams and Parsons (1987) has refined the strength equilibrium envelope.
by addition of data points and replacement of the subjectively drawn envelope with 95% confidence intervals about the regression line (Figure 2):

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Very Strong</th>
<th>Strong</th>
<th>Moderate</th>
<th>Weak</th>
<th>Very Weak</th>
</tr>
</thead>
<tbody>
<tr>
<td>Field estimate of uniaxial compressive strength of intact rock</td>
<td>R6 (20)</td>
<td>R5 (18)</td>
<td>R4 (14)</td>
<td>R3 (10)</td>
<td>R2 (10)</td>
</tr>
<tr>
<td>Weathering</td>
<td>Unweathered (10)</td>
<td>Slightly weathered (9)</td>
<td>Moderately weathered (7)</td>
<td>Highly weathered (5)</td>
<td>Completely weathered (3)</td>
</tr>
<tr>
<td>Spacing of Joints</td>
<td>&gt; 3 m (30)</td>
<td>3-1 m (28)</td>
<td>1-0.3 m (21)</td>
<td>300-50 mm (15)</td>
<td>&lt; 50 mm (8)</td>
</tr>
<tr>
<td>Joint orientations</td>
<td>Very favorable.</td>
<td>Favorable, steep dips into slope (18)</td>
<td>Fair, horizontal dips or nearly vertical in hard rock (14)</td>
<td>Unfavorable.</td>
<td>Very unfavorable.</td>
</tr>
<tr>
<td>Width of joints</td>
<td>&lt; 0.1 mm (7)</td>
<td>0.1mm-1mm (6)</td>
<td>1-5mm (5)</td>
<td>5-20mm (4)</td>
<td>&gt; 20mm (2)</td>
</tr>
<tr>
<td>Continuity of joints</td>
<td>None continuous (7)</td>
<td>Few continuous no infill (6)</td>
<td>Continuous, thin infill (4)</td>
<td>Continuous, thick infill (1)</td>
<td></td>
</tr>
<tr>
<td>Outflow of groundwater</td>
<td>None (6)</td>
<td>Trace (5)</td>
<td>Slight, &lt; 25 l/min/10m² (4)</td>
<td>Moderate, 25-125 l/min/10 m² (3)</td>
<td>Great, &gt; 125 l/min/10 m² (1)</td>
</tr>
<tr>
<td>TOTAL RATING</td>
<td>100 - 91</td>
<td>90 - 71</td>
<td>70 - 51</td>
<td>50 - 26</td>
<td>&lt; 26</td>
</tr>
</tbody>
</table>

Table 2. Rock mass strength rating criteria, after Selby (1980).
### Table 3. Field estimation of the uniaxial compressive strength of intact rock.

<table>
<thead>
<tr>
<th>Grade</th>
<th>Term</th>
<th>Field Estimate of Strength</th>
<th>Examples</th>
</tr>
</thead>
<tbody>
<tr>
<td>R6</td>
<td>Extremely strong</td>
<td>Specimen can only be chipped with a geological hammer</td>
<td>Fresh basalt, chert, diabase, gneiss, granite, quartzite</td>
</tr>
<tr>
<td>R5</td>
<td>Very strong</td>
<td>Specimen requires many blows of a geological hammer to fracture it</td>
<td>Amphibolite, sandstone, basalt, gabbro, gneiss, granodiorite, limestone, marble, rhyolite, tuff</td>
</tr>
<tr>
<td>R4</td>
<td>Strong</td>
<td>Specimen requires more than one blow of a geological hammer to fracture it</td>
<td>Limestone, marble, phyllite, sandstone, schist, shale</td>
</tr>
<tr>
<td>R3</td>
<td>Medium strong</td>
<td>Cannot be scraped or peeled with a pocket knife, specimen can be fractured with a single blow from a geological hammer</td>
<td>Claystone, coal, concrete, schist, shale, siltstone</td>
</tr>
<tr>
<td>R2</td>
<td>Weak</td>
<td>Can be peeled with a pocket knife with difficulty, shallow indentation made by firm blow with point of geological hammer</td>
<td>Chalk, rocksalt, potash</td>
</tr>
<tr>
<td>R1</td>
<td>Very Weak</td>
<td>Crumbles under firm blows with point of a geological hammer, can be peeled by a pocket knife</td>
<td>Highly weathered or altered rock</td>
</tr>
<tr>
<td>R0</td>
<td>Extremely weak</td>
<td>Indented by thumbnail</td>
<td>Stiff fault gouge</td>
</tr>
</tbody>
</table>

If slopes scoured by recent glaciation are truly “oversteepened”, they should plot above the strength equilibrium envelope (have unusually high slope gradients for their RMS rating). Consequently, the RMS rating method is a useful tool in this study to determine whether rockslopes within the Neoglacial limit are oversteepened (see Section 5.3.4). This has been done previously by Augustinus (1995), who used the RMS rating method to assess oversteepening of recently deglaciated rock slopes in New Zealand. In his study, most slopes fell within the strength equilibrium envelope and he concluded that most glacially undercut rock slopes quickly re-equilibrate to a strength equilibrium condition following glacial retreat.
Figure 2. Strength Equilibrium Envelope of Abrahams and Parsons (1987). Terrain plotting in area “A” would have steep slope ratings for a given RMS value and would be considered as “oversteepened”. Terrain in area “B” is in strength equilibrium. Terrain plotting in area “C” is “understeepened” and slopes may be structurally controlled.
CHAPTER 3. STUDY AREA CHARACTERISTICS

The alpine basins examined in this study are located within a ~500 km$^2$ area of mountainous terrain west of the upper Lillooet River, British Columbia (Figure 1). Seventeen of these basins are within Lillooet River watershed, and two basins are within Rutherford Creek watershed. This chapter contains a general description of geology, geomorphology, and climate within this area. Detailed descriptions of each study basin are provided in Appendix A.

3.1 Bedrock Geology

Fourteen study basins are underlain primarily by granitic rock of the Coast Plutonic Complex, and five basins in the Mt. Meager area are underlain partly to completely by Quaternary volcanic rock (Figure 1) (Woodsworth 1977, Read 1978, BC Hydro and Power Authority 1982, Monger and Journeay 1994). Minor exposures of older, fractured volcanic and sedimentary rock of the Upper Cretaceous Gambier group also crop out in parts of lower Mosaic Creek and northeast Capricorn Creek. Highly fractured sedimentary rock is exposed along parts of one study basin on the south side of Ryan River (RY6).

Granitic bedrock in the study area is late Jurassic to early Cretaceous- aged granodiorite, quartz-monzonite and quartz-diorite. Exposures typically are light to medium grey coloured, medium to coarse grained, fresh to slightly weathered, massive, moderately to strongly jointed, and occasionally have gneissic layering. Although joint spacing ranges widely (<10 cm to >5 m) in the study area, fractures are generally planar except for areas of shattered bedrock along the contact of the Meager Creek Volcanic Complex.

The Meager Creek Volcanic Complex includes several dormant or extinct stratovolcanoes of Late Pliocene to Holocene age (Read 1978, Hickson et al. 1999, Green et al. 1988, Stasiuk et al. 1996). Bedrock consists of rhyodacite and andesitic flows
interbedded with tuff layers, and the entire assemblage overlies granitic basement rock. The volcanic assemblage is generally highly fractured, poorly consolidated and hydrothermically altered.

3.2 Surficial Geology and Geomorphology

Surficial geology and geomorphology in the granitic study basins are distinctly different from basins underlain by the Meager Creek Volcanic Complex. In granitic basins, multiple glacial advances during Pleistocene time were a dominant control on terrain development. Basin orientations tend to follow zones of structural weakness and repeated episodes of Pleistocene glacial advancement and retreat have steepened, widened and deepened upper valley (Mathews 1989). During Holocene time there were also at least two major Neoglacial episodes in the British Columbia Coast Mountains, the Tiedemann advance at about 2500 years BP and the Little Ice Age from about 1590 to 1850 AD (Ryder and Thompson 1986). The Little Ice Age reached its maximum extent (Neoglacial maxima) at about 1850 AD and appears to have overprinted any evidence of previous glacial episodes in the study basins. Since the Neoglacial maxima, there has been extensive glacial thinning and retreat in all of the study basins, reducing ice-covered areas by an average of about 50%.

In granitic basins, the cumulative effect of Neoglacial ice advances has caused minor steepening of some hillslopes below the Little Ice Age trimline. Upper hillslope and cirque areas are typically exposed bedrock, partially covered by thin deposits of rubbly to blocky colluvium. Lower valley slopes are partially mantled by gravelly silty sandy, slightly cohesive basal till, and coarse fluvial and debris flow deposits usually overlie glacial drift along valley bottoms.

The most common types of landslide activity in granitic basins are rockfall on upper hillslopes and debris slides and flows on lower valley slopes, particularly along lateral moraines. At least four large rock avalanche events have also occurred in the Ryan River study basins, including at least one event in the last 150 years (in RY4 study basin).
Deep-seated slope deformation features such as tension cracks and antislope scarp features also occur in several locations, particularly in the Ryan River area. One of these locations (in RY3 basin) shows evidence for active slope movement, including live roots stretched or broken across tension cracks (Figure 29).

In basins underlain by the Meager Creek Volcanic Complex, the surficial geology is more complicated because episodes of glacial advance occurred between periods of eruption in the late Pliocene and Early- to Mid-Holocene. Bands of till are interbedded with volcanic deposits, and surficial deposits in the upper basin areas are all younger than the most recent eruption at 2350 years BP. Recent (Little Ice Age) glacial scour and landslide activity have deeply incised the weak bedrock and produced the steepest average slope gradients in the study area at about 35 degrees. The upper slope areas are typically gullied bedrock, and lower hillslopes are dominantly covered in thick scarps of angular, rubbly colluvium or gravelly silty sandy glacial till. Thick debris flow deposits cover most valley bottoms.

The frequency and magnitude of both rockslope and surficial failures are higher in the vicinity of the Meager Creek Volcanic Complex than in the granitic study basins. Several exceedingly large Holocene landslides deposits fill Meager Valley to a cumulative thickness of over 100 metres in some areas, including rock avalanche, debris flow and lahar\(^1\) deposits (Baumann and EBA Engineering 2000). Frequent rockfall occurs on all steep rockslopes, and historical catastrophic failures have been documented in both volcanic rock and fractured granitic basement rock along the contact. This includes rock avalanche events in Devastation Creek, Capricorn Creek, and near the summit of Mt. Meager (Carter 1931, Mokievsky-Zubok 1977, Evans 1987b, Croft 1983).

The first documented major historical landslide in the Mt. Meager area occurred in 1931, when at least 3.1 million cubic metres of material travelled down Devastation Creek and continued over 15 km down Meager Creek (probably as a debris flood) to the Lillooet River (Carter 1931). Based on the texture of this deposit, the failure may have initiated

\(^1\) A Lahar is defined as a debris flow triggered from a volcanic source during a period of volcanic activity.
on the east side of Devastation Creek, although the precise headscarp location is unknown (Baumann and EBA Engineering 2000). In 1975, a $1.2 \times 10^7$ m$^3$ rockslide initiated on a glacially oversteepened slope in Devastation Creek and triggered a second large debris flow (see Section 5.3.2.1) (Evans and Clague 1988). This debris flow did not travel beyond the confluence of Devastation and Meager Creeks, but killed four BC Hydro workers awaiting helicopter pickup at the outlet of Devastation Creek. Since then, there has been active debris flow activity in several basins including debris flows in Canyon Creek in 1987 and Canyon and Capricorn Creeks in 1990. In Capricorn Creek, analysis of scarred trees indicates that debris flow events reach the confluence with Meager Creek about every 13 years, and that events of at least $10^5$ m$^3$ occurred in 1933, 1944, 1972 and 1998 (Jakob 1996). The 1933 event has been attributed to a large rock avalanche that initiated in granitic rock in the mid-valley area (see Section 5.3.2.1) (Jakob 1996). On July 29, 1998, a $1.2 \times 10^6$ m$^3$ debris slide/flow initiated in thick colluvium in upper Capricorn Creek (see Section 5.2.3) (Bovis and Jakob 2000). The resulting debris flow traveled the entire length of the creek and dammed Meager Creek, forming an 800 metre long landslide-dammed lake at the confluence that persisted for almost one year. This debris flow also destroyed the Meager Creek Forest Service Road and prevented access to Meager Creek Hotsprings, a popular tourist destination. The most recent major rock avalanche activity occurred in March or April 1986, when about $0.5 \times 10^6$ m$^3$ of Pleistocene rhyodacite detached from the north peak of Mt. Meager (Evans 1986). The rock avalanche travelled over a glacier in the upper transport zone, and some of the debris reached the Lillooet river almost 2000 m below.

Non-catastrophic rockslope deformation also occurs in parts of all study basins in the Mt. Meager Area, including deformation in both granitic and volcanic rock. The most intensively studied area of deep-seated slope deformation is the west side of Affliction Creek, where ongoing slope movement has occurred on a glacially debuttressed slope for about the last 3000 years (Bovis 1989). The 1975 failure in Devastation Creek also occurred in a location with ongoing, deep-seated slope deformation.
3.3 Climate

All study basins are located in a transition area between a wet maritime climate dominated by Pacific cyclones to the west, and a drier submaritime climate to the east. Generally, precipitation is heaviest in November, followed by a roughly linear decrease in precipitation until July and then a steep increase in autumn. The most severe floods tend to occur in fall, often associated with rain-on-snow events, although snowmelt caused by prolonged periods of high temperatures in the summer can also generate intense runoff (Bovis and Jakob 2000). Intense rainfall during periods of summer thundershower activity is also an important landslide triggering factor in this area, and has been a factor in debris flow initiation in several basins along the Lillooet River valley (M. Jakob, pers. comm.).
CHAPTER 4. METHODS

Seventeen glaciated alpine basins in the Lillooet River basin and two sub-basins within Rutherford Creek were examined to determine the relations between glacial retreat and landslide activity. Figure 3 shows the major steps in this work, including field descriptions, rock mass strength estimates, GIS analysis and statistical analysis of associations between landslide activity and terrain characteristics.

4.1 Scale of the investigation

One of the most important issues in this study was to determine an appropriate scale for investigation. On the one hand, detailed study could yield a good physical understanding of specific effects of glacial retreat on local slope stability. However, such an approach could not be used to make general statements about larger regions, and important influences of glacial retreat on landslide activity might be missed because they did not happen to occur in the detailed study area. On the other hand, a regional investigation could potentially identify many more interactions between glacial retreat and landslides but analysis results would necessarily be more qualitative.

In previous studies, workers have focused on either detailed engineering assessments of small areas (< 1:5000 scale) or qualitative studies of much larger areas (≥ 1:50,000 scale). In this study an intermediate scale (1:17,000 scale mapping in 5 km<sup>2</sup> – 20 km<sup>2</sup>)
Information Sources
- Airphoto interpretation (1:17 000)
- Field observations
- Digital TRIM map (1:20 000)
- Geology maps (1:20,000; 1:250,000)

Digitize and import into GIS
- Landslide locations
- Terrain mapping
- Geologic contacts

Vector GIS Layers
- Geologic units (polygons)
- Geomorphic terrain units with information on surficial materials, geomorphic processes and rock mass strength, based on fieldwork and airphoto interpretation.
- Landslide locations
- Location of the Neoglacial limit and glacial trimlines
- Subdivision of each basin into 3 zones: down-valley of the Neoglacial limit, mid-valley area and cirque.
- Study area boundary

Grid Layers
- All Vector layers converted to 25 m grids
- 25 m DEM from TRIM elevation points (gridded TIN)
- Layers derived from the DEM: slope gradient, upslope drainage area, topographic index, profile and plan curvatures, elevation and aspect.
- Layer derived from spatial analysis: proximity of terrain to the Neoglacial Limit, mid-basin area.

Analysis and Results
- Grid layers reclassified as simplified categorical nominal and ordered data.
- Univariate and multivariate statistics conducted to relate terrain characteristics to landslide occurrence.
- Qualitative assessments made of relations between landslide activity and glacier retreat, based on fieldwork and airphoto interpretation.
- Decision-making flowchart developed, based on both field observations and statistical results, that identifies terrain susceptible to failure following glacial retreat.

Figure 3. Flowchart for analysis
basins) was chosen to allow a relatively broad survey of basins experiencing glacial retreat while allowing detailed fieldwork and data collection for statistical analysis. This approach is facilitated by several factors. First, the lack of vegetation in alpine basins permits relatively detailed airphoto interpretation of terrain. Second, field access to some study basins was provided by helicopter which otherwise would have been very time consuming to reach on foot. Multi-day backpacking traverses were also conducted to access remote terrain. Finally, the use of GIS and a digital elevation model (DEM) for data analysis allowed processing of large amounts of information at a regional scale.

### 4.2 Selection of study basins

Seventeen alpine basins in the upper Lillooet River watershed and two basins in Rutherford watershed, British Columbia, were selected as suitable study areas using 1:30,000 and 1:17,000 scale aerial photographs, National Topographic System (NTS) maps (1:50,000 scale), Terrain Resource Information Management (TRIM) maps (1:20,000 scale) and bedrock information (Read 1978, Woodsworth 1977) (Figure 3). Basins with the following characteristics were selected:

- All basins are currently glaciated, have experienced glacial retreat, and contain landslides.
- All basins are at approximately similar elevations and are located in the same climatic region. These criteria minimize basin-to-basin differences in external influences on landslide activity such as rainfall.
- Each basin has similar slope characteristics (slope profiles) from the mid-basin area to some distance down-valley of the Neoglacial limit. This criterion facilitates comparisons between recently glaciated and non-recently glaciated parts of a basin without having to account for major differences in basin morphometry.
- For logistical reasons all basins are accessible by backpacking or a short helicopter trip from Pemberton or Whistler.
4.3 Data Selection

The purpose of data collection in this thesis was to obtain measures that characterized unstable terrain and identified any glacial influences on slope stability. Measures that characterize unstable terrain are similar in many landslide hazard-mapping studies and have been discussed in section 2.1: examples include slope gradient, the presence of gullies, and the type of surficial material. Measures that characterize glacial influences on slope stability are more difficult to obtain because any such influences on slope stability are a secondary effect; the glacier has modified the landscape in a way that affects slope stability. At a regional scale it is difficult to obtain a mechanical understanding of this influence, but the occurrence of instability in close proximity to glacial features may imply a causative relation. For example, a high spatial frequency of landslides near slopes steepened by glacial scouring might imply that glacial oversteeping was an influencing factor. Data collected included qualitative field observations of landslides in glaciated terrain, and development of measures in the GIS that measured the proximity of landslide activity to the glacial trimline.

4.4 Data Collection

4.4.1 Terrain Mapping and Landslide Inventory

A landslide inventory was completed from air photo interpretation, with reference to terrain stability studies by Baumann Engineering and EBA Engineering (2000) and Bovis and Jakob (2000). Landslide locations were mapped by delineating the failure scarp. The terrain was then divided into geomorphic terrain units (polygons) based on material type, slope angle and aspect, following the terrain stability mapping conventions of Howes and Kenk (1997). Each terrain unit was interpreted for surficial material types and observed geomorphologic processes (e.g. slope failures and gully erosion). Neoglacial trimlines and the location of the maximum Neoglacial extent were also delineated.
4.4.2 Fieldwork

Field traverses were conducted from June 20 to August 25, 2001 in twelve of the nineteen study basins. Areas selected for field traverses included basins with the most significant slope instability, any locations where airphoto interpretation was questionable, and areas judged to be representative of the basins not traversed on foot. The study areas that could not be visited on the ground were viewed during a helicopter overflight. The following work was completed during ground traverses:

- Verification of air photo interpretations,
- descriptions of surficial material characteristics such as texture,
- brief descriptions of bedrock structure and stratigraphy,
- estimations of rock mass strength,
- site visits to landslide initiation zones thought to be related to glacial retreat, and
- general field observations of possible interactions between glacial scouring, retreat and landslide activity

4.5 Data analysis and the role of GIS

The GIS package ArcView Spatial Analyst 3.2 (Environmental Systems Research Institute 1999) was used extensively for data management in this project, including queries of spatial data and creation of tabular information used for statistical analysis. The GIS work involved three major steps: 1) defining a research approach, including selecting the terrain unit used for analysis 2) creating the GIS database, 3) classifying information into a form suitable for statistical analysis and 3) conducting spatial queries and exporting tabular data for statistical analysis.
4.5.1 Selection of the Terrain Unit for Analysis

The terrain unit represents the smallest level of resolution in a study, and all attributes assigned to a terrain unit are considered homogenous throughout that unit. Selection of the appropriate designations for a unit depends on the type and resolution of data available, and Section 2.4.2 describes several possibilities including grid cells, manual delineation of terrain polygons (terrain mapping), or automated generation of terrain polygons based on a DEM.

Terrain polygons used in provincial terrain stability mapping (Howes and Kenk 1997) were initially considered because this terrain unit is meaningful in terms of slope stability for terrain attributes such as slope and surficial material type, and because terrain polygons are appropriate for mapping surficial geology. However, several DEM-derived attributes were not easily generalized over polygon areas because they vary significantly in value across each polygon. For example, attribute values for upslope drainage area\textsuperscript{1} vary continuously across polygons since a grid cell at the upslope edge of the polygon has a much lower drainage area value than a cell at the bottom. Consequently, grid cells were deemed the most appropriate unit for analysis. A grid cell size of 25 m was chosen, as this is the most detailed level of precision that is reasonable for a TRIM DEM (B. Klinkenberg, pers. comm.) Surficial geology was mapped using terrain polygons (Howes and Kenk 1997), and then subdivided into 25 m grid cells for further statistical analysis.

4.5.2 Creation of the GIS database

Airphoto interpretations, including failure locations, terrain polygons, glacial trimlines, and study basin boundaries, were digitized by Atticus Resource Consulting using

\textsuperscript{1} Upslope drainage area is defined as the area upslope of a particular grid cell that could deliver water to that grid cell.
monorestitution techniques\textsuperscript{1}. Digitized linework was then linked to attribute information in Arcview Spatial Analyst 3.2 (Environmental Systems Research Institute 1999). Slide scarps smaller than about 25 m in length were digitized as point locations. Larger slide scarps were originally digitized as lines along the apex of the scarp and then converted to point locations by digitizing a point at the centre of each scarp.

GIS layers for bedrock geology in the Mt. Meager area were generated from the digital geology map of Journeay (1999). All other basins were either 100% granitic rock, or mostly granitic rock with minor exposures of other lithologies; in the latter case the contacts were digitized onscreen, using the geologic map of Woodsworth (1977).

Except for landslide point locations, all vector layers were then converted to 25 m grid cells. Regions covered by glaciers were removed from the study areas because they did not contain any soil or rock failures.

Several new layers were then calculated from the TRIM DEM, using algorithms in Spatial Analyst and SINMAP (Pack et al. 1997). These include slope gradient, aspect, elevation, profile and plan curvature measures, upslope drainage area, and topographic index\textsuperscript{1}. A layer was also created for the mid-basin area that indicates the percent distance from the glacial trimline to the valley bottom and to the ridgetop.

4.5.3 Production of a database for statistical analysis

The purpose of statistical analysis was to correlate the occurrence of landslides with measured characteristics of the terrain. However, the original terrain attribute layers contained hundreds of detailed variable combinations that prevent determination of statistically significant correlation because many combinations are unique within the database. Also, general statements based on statistical results must be simple enough to

\textsuperscript{1} Monorestitution uses fiducial information and a DEM to correct for airphoto distortion without resorting to use of a full stereo model.

\textsuperscript{1} Topographic index is defined as the logarithm of drainage area divided by slope.
be applicable in practical situations. Consequently, all themes were reclassified into simplified categories for use in univariate statistical analysis (Table 4), and further simplified into lumped categories for multivariate statistical analysis (Table 5).

The most subjectively defined layer in Table 3 is the surficial material types layer. In general, surficial materials were grouped according to similarity of geotechnical properties. Only the material covering >60% of the polygon was considered. Where there was more than one layer of strata (up to two layers were mapped), the material judged more important for slope stability was picked as the representative material. Surficial materials were divided based on thickness (veneers < 1 m thick and blankets >1 m thick). Surficial materials were also differentiated based on whether the slope was material-controlled (slopes where the deposit is thick enough to mask the underlying topography) or rock controlled (underlying bedrock controls slope morphometry).

4.6 Process-based data analysis methods

4.6.1 Field observations of landslide activity

The effect of glacial retreat on landslide activity involves complex interactions between many geomorphic factors in an environment where information on terrain characteristics is usually scarce. In addition, statistical associations between terrain characteristics and slope failures are difficult to obtain for rare, but important, types of landslide activity. For example, Capricorn Creek contained the only large landslide in thick colluvium in the entire study area. Consequently, qualitative observations based on airphoto interpretation and fieldwork form an important first step in investigating relations between glacial retreat and landslide activity.

The following steps were completed to characterize unstable terrain to obtain a better understanding of the interactions between glacial retreat and landslide activity:
<table>
<thead>
<tr>
<th>Attribute</th>
<th>Class Code (25 metre grid cells)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Failures</td>
<td>rapid rock                                        rapid surficial                                    slow surficial                                    slow rock</td>
</tr>
<tr>
<td>Slope Gradient (degrees)</td>
<td>0-10                                               10-20                                            20-30                                            30-40                                            40-50                                            50-60                                            60-70                                            70-80                                            80-90</td>
</tr>
<tr>
<td>Slope Aspect</td>
<td>North (&gt;315-45) or flat                           East (&gt;45-135)                                     South (&gt;135-225)                                  West (&gt;225-315)</td>
</tr>
<tr>
<td>Elevation (m)</td>
<td>&lt;1200                                              1200-1400                                         1400-1600                                         1600-1800                                         1800-2000                                         2000-2200                                         2200-2400                                         2400-2600                                         &gt;2800</td>
</tr>
<tr>
<td>Geology</td>
<td>Mesozoic metamorphic rock                         Granitic Intrusives, mainly Mesozoic               Mesozoic volcanic and sedimentary rocks            Plio-Quaternary volcanic complexes</td>
</tr>
<tr>
<td>Upslope drainage area (log m^2)</td>
<td>0-2                                               2-3                                               3-4                                               4-5                                               5-6</td>
</tr>
<tr>
<td>profile curvature (1/100zunits)</td>
<td>concave (&lt;-0.25)                                  straight (-0.25 to +0.25)                          convex (+0.25)</td>
</tr>
<tr>
<td>plan curvature (1/100zunits)</td>
<td>concave (&lt;-0.25)                                  straight (-0.25 to +0.25)                          convex (+0.25)</td>
</tr>
<tr>
<td>Surficial Material</td>
<td>Rock                                               Till veneer (&lt;1 m thick mantle)                   Till blanket (&gt;1 m thick mantle)                   Till, material-controlled slope (till is thick enough to mask underlying topography)    Colluvial veneer                                 Colluvial blanket                                 Colluvium, material-controlled slope (colluvium is thick enough to mask underlying topography)    Fluvial                                          Lacustrine                                      Ice (not included in analysis)</td>
</tr>
<tr>
<td>Zones</td>
<td>Below NGL                                         Mid-Valley                                        Circule</td>
</tr>
<tr>
<td>Inside/Outside Neoglacial limit</td>
<td>in                                                out</td>
</tr>
<tr>
<td>Topographic Index (log (drainage area/slope))</td>
<td>0-1                                       1-2                                              2-3                                              3-4                                              4-5                                              5-6                                              6-7</td>
</tr>
<tr>
<td>Gullied</td>
<td>Not Gullied                                       Gullied</td>
</tr>
<tr>
<td>Mid valley area only: percent distance from glacial trimline to ridgetop and to valley bottom (vbott)</td>
<td>0-20% towards valley bottom                       20-40% towards valley bottom                       40-60% towards valley bottom                       60-80% towards valley bottom                       80-100% towards valley bottom                      0-20% towards ridgetop                            20-40% towards ridgetop                           40-60% towards ridgetop                           60-80% towards ridgetop                           80-100% towards ridgetop</td>
</tr>
</tbody>
</table>

Table 4. Terrain attributes and classes used for FR analysis.
<table>
<thead>
<tr>
<th>Attribute</th>
<th>Class Code (25 metre grid cells)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1</td>
</tr>
<tr>
<td>Failures</td>
<td>rapid rock</td>
</tr>
<tr>
<td>Slope Gradient (degrees)</td>
<td>n/a</td>
</tr>
<tr>
<td>Slope Aspect</td>
<td>North (&gt;315-45) or flat</td>
</tr>
<tr>
<td>Elevation (m)</td>
<td>&lt;1400</td>
</tr>
<tr>
<td>Geology</td>
<td>Mesozoic metamorphic rock</td>
</tr>
<tr>
<td>Upslope drainage area (log m²)</td>
<td>0-2</td>
</tr>
<tr>
<td>profile curvature (1/100 units)</td>
<td>concave (&lt;-0.25)</td>
</tr>
<tr>
<td>plan curvature (1/100 units)</td>
<td>concave (&lt;-0.25)</td>
</tr>
<tr>
<td>Surficial Material</td>
<td>Rock</td>
</tr>
<tr>
<td>Zones</td>
<td>Down-valley of the NGL</td>
</tr>
<tr>
<td>Inside/Outside Neoglacial limit</td>
<td>in</td>
</tr>
<tr>
<td>Topographic Index (log (drainage area/slope))</td>
<td>0-1</td>
</tr>
<tr>
<td>Gullied</td>
<td>Not Gullied</td>
</tr>
<tr>
<td>Mid valley area only: percent distance from glacial trimline to ridgetop and to valley bottom (vbert)</td>
<td>0-20% towards valley bottom</td>
</tr>
</tbody>
</table>

Table 5. Simplified terrain attribute classes for multivariate statistical analysis.
1) An inventory was made of unstable areas located close to recently deglaciated terrain.

2) Terrain containing landslides was grouped based on landslide type, terrain characteristics, and potential glacial influence.

3) A qualitative flowchart was developed that identified types of terrain in alpine basins with distinct landslide responses to glacial retreat.

4) Field and statistical analysis results were compared to test associations described in the flowchart.

4.6.2 Stability analysis based on rock mass strength estimates

A limited assessment of rock mass strength was conducted on recently ice-covered granitic bedrock slopes. The purpose of this preliminary work was to evaluate the suggestion that glacially scoured granitic rock slopes are typically "oversteepened".

Using the classification system of Selby (1980), rock mass strength (RMS) estimates were collected for 49 granitic rock slope polygons in seven of the study basins\(^1\). RMS measurements were only made for granitic rock because failures in other, weaker lithologies did not always occur along planar bedrock discontinuities, a major assumption in Selby's classification system.

Section 2.8 describes a "strength equilibrium envelope" for rockslopes. Rock slopes whose RMS rating plots above this envelope (have a high slope gradient for a given RMS value) are considered to be "oversteepened" and potentially prone to instability. RMS values were plotted with respect to average polygon slope gradients determined from the TRIM DEM to determine whether they fell within this strength equilibrium envelope.

\(^1\) R1, R3, RY1, RY3, RY4, RY6, and PE1
4.7 Statistical methods

The data available for analysis included categorical nominal or ordinal terrain attributes and the presence/absence of landslides at point locations. Univariate and multivariate statistical analyses were conducted to examine variations in landslide frequency within different classes of each terrain attribute, and to determine statistically significant relationships between terrain characteristics and landslide activity.

4.7.1 Univariate Statistical Methods

The purpose of univariate statistical analysis was to obtain a preliminary understanding of landslide spatial frequency in different types of terrain. Failure Rate (FR) analysis (Dhakal et al. 1999) was conducted to determine the relative contribution to landsliding of different classes within each attribute, as measured by Susceptibility Scores (Section 6.1). This involved the following steps:

1) Within each basin, each GIS terrain attribute layer was queried to determine the number of surficial or rockslope failures occurring within each class of an attribute and to determine the size of each class (number of grid cells).
2) Data from each basin were compiled into a single, larger database showing the numbers of failures in each class of an attribute for the entire study area.
3) Susceptibility scores were calculated separately for surficial and rockslope failures for each attribute in the compiled database (Section 6.1).

4.7.2 Multivariate statistical methods

The purpose of multivariate statistical analysis was to obtain a rank-ordering of terrain attributes and combinations of terrain attributes that were spatially associated with high spatial landslide frequency. This work involved two major steps: first, the creation of an
appropriate database for analysis, and second the use of multivariate statistical methods
to classify the data.

4.7.2.1 Database creation

Multivariate statistics in this study involve analyses of the relations between values of
several GIS attribute layers and the presence or absence of a slide in each terrain unit
(grid cell). Since GIS grid themes can only contain one value per cell, all grids were
converted to arrays of vector points located at the center of each cell. Vector points can
contain multiple attribute values, and values of all terrain attributes were combined to
produce one theme such that each point contains all attribute information at that location.
Attribute tables from each basin were exported from the GIS and combined. Both
logistic regression and discriminant-based classification trees are non-parametric (i.e. do
not require normally distributed data or fully independent controlling variables).
However, the algorithms are more stable when the dependent variable is not too skewed
in distribution (ideally there should be a similar number of points within failed or non-
failed terrain (A. Dhakal, pers. comm.). Furthermore, results may be more replicable and
easier to interpret if there is a small number of classes within each attribute. However,
the initial database contained a fairly large number of classes (up to 10 classes) for
several of the attributes, and the population of dependent variables was highly skewed:
there were 206,109 stable points versus 824 rockslope failures and 242 surficial failures
in the study area.

To correct for the imbalance in the dependent variable classes, samples of 1000 stable
points were randomly selected from the data to represent stable terrain. In order to test
whether the samples were representative of the entire stable population, ten different
random samples were analyzed; consistent results would then suggest that the random
samples were indeed representative of the larger population.
Although the level of resolution is lowered, simplifying some of the classes within each attribute increases the likelihood of replicable results between different sample sets. In addition, results may be more easily generalized to different areas with similar slope stability issues but minor differences in terrain characteristics. Using judgement and Failure Rate (FR) analysis results it was possible to identify overly detailed subdivisions within several attributes and reclassify this data into appropriate, simplified categories (Table 5).

4.7.2.2 Logistic Regression

Logistic Regression is a non-parametric statistical method that can be used to predict the occurrence of stable or unstable terrain based on a combination of predictor variables (or terrain attributes). Although this method has been used by previous landslide workers (Roberts 2001), it was recognized at the outset that a purely statistical approach to landslide hazard modeling in this thesis was unlikely to succeed because of scarcity of data. However, this approach was still attempted in order to determine whether any controlling variables were sufficiently important to produce an acceptable model of landslide prediction. A Wald stepwise forward logistic regression was run in SPSS on one of the random chosen samples to evaluate this method (Section 6.2).

4.7.2.3 Classification Trees

Classification trees provide more visually intuitive results than do the purely numerical output of methods such as logistical regression. In addition, the decision-tree format is convenient in this thesis because a stated objective was to produce a decision-making flowchart for landslide hazard assessment.

Classification trees were produced using Statistica for each of the randomly sampled datasets. “Splits,” or tree-bifurcation nodes, were set to occur until the best predictor variable had a p-value greater than 0.05 or the sample size of cells below a node was less
than ten. Classification trees were created separately for rockslope and surficial failures for each of the 10 randomly selected datasets (Section 6.3).
CHAPTER 5. FIELDWORK RESULTS AND DISCUSSION

Neoglacialation is only one of many factors that have modified terrain within the study basins. Other factors include earlier Pleistocene glacial episodes, fluvial erosion, weathering, periglacial processes and previous landslide events. Identifying how landscape modification by Neoglacial advance and retreat affects landslide initiation, when many other factors also influence slope stability, is a central goal of this thesis. Field observations of landslide processes near recently deglaciated terrain are a first step to meeting this objective.

This chapter presents a process-oriented characterization of terrain subject to rapid landslide activity in basins experiencing recent Neoglacial retreat. Landslide activity is separated into major groupings based on landslide location, general landslide type (bedrock or surficial material failures), material type (colluvium or glacial drift), terrain characteristics, and possible glacial influences.

5.1 Process-Based Grouping of Basins into Three Zones

Each study basin was subdivided into three main zones having different influences of recent glacial retreat on landslide activity:

1. Downvalley of the Neoglacial Limit (Figure 4A).
   - No likely influence of recent Neoglacial retreat on any landslide activity in this area.

2. Mid-basin (area from Neoglacial limit to the cirque) (Figure 4B)
   - Area where lower valley slopes have experienced the greatest amount of slope steepening from Neoglacial scour.
   - Area of strongest Neoglacial influence on slope stability
   - Landslide activity affected by recent Neoglacial retreat is in a recovery or slope-adjustment phase, as ice has retreated from most or all of the lower valley slopes.
Figure 4. Examples of the three valley zones. Insets show generalized basin zones, with a Neoglacial trimline also delineated. A) Zone 1 (RY3 basin), defined as the area down-valley of the Neoglacial limit. Note the Neoglacial Limit moraine in the foreground. B) Zone 2 (Job Creek), mid-valley area. C) Zone 3 (Job Creek), cirque area.
3. Cirque (Figure 4C)
   • Relatively lower magnitude of slope steepening by Neoglacial scour than in Zone 2.
   • Neoglacial retreat and ice thinning is ongoing. Further retreat may trigger slope collapse and landslide activity.
   • Landslide activity related to Neoglacial retreat is less common than in Zone 2.

5.2 Failures in Surficial Material

5.2.1 Primary Versus Secondary Neoglacial Effects

Neoglacial effects on landslide activity in surficial material can be divided into two categories: Primary Neoglacial Effects and Secondary Neoglacial Effects. Primary Neoglacial Effects include failures in material deposited where terrain characteristics (e.g. steep slopes) favour instability, and failures due to removal of support from thick deposits that have been de-buttressed by glacial retreat. Secondary Neoglacial Effects include deposition of drift material within the transport zone of an pre-existing failure. The drift material itself is stable but is susceptible to entrainment by a landslide process originating upslope, thereby increasing the resultant landslide volume. If the drift deposit is too large to be entrained it may instead cause blockage of failures originating upslope.

5.2.2 Primary Neoglacial Effects in glacial till.

Ongoing glacial retreat in all study basins has significantly increased the area of glacial drift material prone to landslide activity. Based on field observations, drift deposits with the following characteristics are most prone to failure:

1. Deposition of thick, material-controlled drift in areas with high upslope drainage area (Figure 5, Figure 6, Figure 9). Thick, material-controlled drift deposits are often prone to gully erosion in locations with high upslope drainage area, such as lateral
moraines below high valley slopes. Debris slides commonly initiate on gully
headwalls or sidewalls and are transformed into debris flows. For example, Figure 5
shows a lateral moraine on the south side of RY1 basin that extends from beneath the
main valley slope out onto a broad plain where RY1 merges with a small basin to the
south. The portion of this lateral moraine that lies directly below the steep valley
slope has a much higher density of landslides and gully erosion than the portion
down-valley that is not below the steep valley slope.

Figure 5. South side of RY1 basin. Immediately right (south) of the lake is a lateral
moraine extending from the south (right) side of the glacier to the broader valley area to
the east (upper left photo). Landslide density is much higher in the lateral moraine below
the steep rockslope.
2. Drift mantles on steep, rock-controlled lower valley slopes (Figure 7 and Figure 8).
Glacial drift material mantles benches and hollows on many rock slopes that are too steep to hold surficial material for long time periods. Many of these areas are failing to the drift-bedrock interface and will eventually lose most of their surficial material.
3. Deposition of thick drift at the edge of convex slope breaks. In Rainy Creek (Figure 10) and in a cirque glacier in RY6 basin (Figure 11), Neoglacial ice advance pushed thick drift material to slope breaks above steep channels prone to debris flow activity. Jakob (1996) cites unconsolidated drift material as a major source of material for debris flow activity in Rainy Creek, and indicates that recent glacial retreat has increased the area of sediment supply for debris flow activity by approximately a factor of two. In RY6 basin, a cirque glacier previously advanced to the edge of the main valley sideslope, forming lateral and end-moraines at this convex slope break and thick colluvial deposits on the valley slope below. The lateral moraines concentrate drainage from the existing cirque glacier into a single channel down the main valley slope, and thick, unconsolidated colluvium along the sides of the gully provide an almost unlimited material source for debris flow activity in this channel.
Figure 9. Debris flow gully in a lateral moraine on the lower east side of middle Mosaic Creek. Note person for scale (circled). This gully was likely formed by debris slides initiating in the gully headwall, in an area of high upslope drainage area.

Figure 10. Lateral moraine in upper “Rainy” Creek (right side of photo), providing source material to an active debris flow creek.
Figure 11A. Lateral moraines at the edge of a hanging valley on the west side of RY6 basin. These moraines concentrate drainage into a debris flow gully in thick colluvium and glacial drift on the main valley slope. B) Aerial view. C) Debris flow lobes and channels (snow filled) in the deposition area. RY4 creek is on the left side of the photo.
5.2.3 Primary Neoglacial Effects in colluvium.

Failures in colluvium were rare in the thesis study basins because most slopes covered in surficial material were drift-mantled. Failure initiation in colluvium seemed to be affected by the following factors associated with recent glacial retreat:

1. Deposition above Neoglacial trimlines. Several locations were noted in S1, S2, Capricorn and PE1 basins where colluvium had accumulated at its repose angle immediately above slopes subsequently steepened by recent glacial scour. Any material transported onto the steepened slope was subject to ravelling and minor debris avalanching.

2. Deposition in areas buttressed by glacial ice. Talus slopes were very common in most study basins in formerly ice-covered areas. Some of these deposits may have originally formed against ice margins and settled at the repose angle as the glacier thinned. Failures in this material are rare, but catastrophic collapse may occur where the colluvium is sufficiently cohesive to remain on steep slopes following glacial debuttressing. A notable example is the 1998 debris slide that occurred on the north side of Capricorn Creek (Figure 12). At this site, Neoglacial ice retreat removed buttress support from a thick surficial deposit to form a steep (30-40°) toe slope and 20-30° colluvial slope above. Reduction of toe support has resulted in extensive slope deformation, and numerous tension cracks and antislope scarps traverse the slope in the vicinity of the 1998 failure scarp (Figure 12). On July 29, 1998 a \(-1\times10^5\) m\(^3\) portion of this slope collapsed and formed a debris flow that traversed the entire length of Capricorn Creek and dammed Meager Creek at the confluence (Bovis and Jakob 2000). Several tension cracks observed during fieldwork show signs of recent movement, and this slope apparently has the potential for further catastrophic collapse.
Figure 12A. Debris slide initiation zone of the 1999 Debris flow in Capricorn Creek. Note the melting snow patch draining directly above the failure scarp, and the slope deformation features indicated to the right of the failure. Debris-covered Capricorn Glacier is in the lower portion of the photograph. B) Small tension crack in the vicinity of arrowed features in (A), with signs of active movement such as live roots under tension and freshly exposed soil.

In some instances colluvium may indirectly contribute to increased slope stability by covering and insulating ice, resulting in prolonged ice buttress-support of hillslopes. Figure 13 shows two examples of debris-covered ice mantling the toe of hillslopes in Mosaic and Devastation basins.
Figure 13. Examples of ice buttressing. The contact between debris-covered ice and the hillslope above is delineated by black lines. A) East side of middle Mosaic Creek. B) East side of middle Devastation Creek.
5.2.4 Secondary Neoglacial Effects in glacial till

The magnitude of slope failures initiating on upper hillslopes can be significantly increased through entrainment of glacial drift in the transport zone. This has occurred in several locations in the study area where failures initiated on the upper part of steep rockslopes and impacted drift material in the lower valley area. The most significant example of this occurred in a cirque glacier on the north side of RY4 basin (Figure 14).

Figure 14A. Site of a rock avalanche in RY4 basin that entrained an end-moraine in the transport zone. Local relief is about 650 m from valley bottom to ridgetop. B) Location of this failure in RY4 basin (arrowed). Also shown is the Neoglacial limit (dashed line) and currently glaciated areas (shaded).

Based on airphoto interpretation, field observation and a helicopter over-flight, it appears that a rock avalanche initiated in the upper cirque area, travelled across the cirque glacier, and impacted an end-moraine at the slope break above the main valley. The end moraine
was breached and entrained by the rock avalanche, and a large volume of debris traveled to the bottom of the main valley slope. This landslide deposit temporarily dammed RY4 creek, as evidenced by a floodplain up-valley of the deposit (Figure 15). Another example of this process exists on the west side of Job Creek, where debris flows initiating on upper hillslopes have entrained parts of a lateral moraine (Figure 16).

![Figure 15. Landslide deposit in RY4 basin.](image)

In several basins, including S1, S2, RY1, Capricorn, PE3, and RY3, the linear trough between a lateral moraine and the bedrock slope can act as a storage zone that impedes (blocks) transport of small rockslope failures initiating upslope (Figure 17).

Many of these troughs are almost filled to capacity, implying that in future there will be an increase in sediment delivered to the valley bottom. Debris flows initiating upslope of the lateral moraine did not appear to be blocked by the trough. However, a small basin that was not included in this study, Angel Creek (south of Pylon Peak, Mt. Meager area)
contains an end moraine that is acting as a catch-basin for debris flows initiating in the upper cirque area (Figure 18). This basin will also eventually reach capacity and allow debris flows to continue down the main valley stem.

Figure 16A. West side of Job Creek, showing debris flow gullies initiating on the upper valley slope that entrain glacial drift in a lateral moraine in the transport zone. B) View down the same rockslope from above, towards the gullied lateral moraine.
Figure 17. Lateral moraine on the east side of RY3 basin (trough is filled with snow). The trough prevents minor rockfall initiating upslope from travelling to the valley bottom. Debris flows initiating in gullies above the trimline were not blocked by the lateral moraine trough.

Figure 18. End moraine in Angel Creek (photos by M. Jakob). A) Aerial view looking north. B) Closer view of the moraine, looking northeast.

5.3 Failures in bedrock

The spatial frequency of bedrock failures is highly variable in the different study basins, ranging from an average of 1 failure per km² in R1 basin to 21 failures per km² in Job Creek basin. In general, basins underlain by rocks of the Meager Creek Volcanic Complex had much higher spatial frequencies of both rapid rockslope failures and deep-
seated bedrock instability. Almost all bedrock failures mapped were rockfall or minor, shallow rock avalanche scarps, commonly less than 20m in length. Major rock avalanches, defined here as events that have significantly altered basin topography or have significant landslide deposits, were mapped in eight locations in the study area, including Capricorn Creek (2 events), Devastation Creek (1 event), RY3 creek (3 events), and RY6 (1 event).

5.3.1 Minor, small rapid rockslope failures

A total of 819 minor rapid rockslope failures were mapped in the study area. Many of these features are shallow rockfall scarps that occur on steep slopes throughout the study basins, and reflect complex terrain conditions not necessarily related to recent glacial retreat. However, several common patterns of shallow rockfall activity were observed. In general, rockfall seemed to be concentrated along sharp convex slope breaks such as uppermost gully sidewalls or headwalls, ridgetops and glacial trimlines. Most failures initiated above (upslope of) the Neoglacial limit.

Neoglacial processes have either increased or decreased the spatial frequency of shallow rockfall scarps depending on the location. Immediately upslope of the Neoglacial limit, glacial scour may have increased shallow rockfall activity where it created a sharp, convex slope break along the trimline (Figure 19). Dilation of joint sets in some glacially debuttressed areas may also increase block toppling activity (Figure 20). Within the Neoglacial limit, glacial processes may have temporarily reduced shallow rockfall activity because slopes scoured by glacial ice have had insufficient time for subaerial weathering to loosen fresh rock.
5.3.2 Major rapid rockslope failures (rock avalanches and rock slides)

Few major rockslope failures were mapped in the study area. Field observations did provide some preliminary insight into the effects of glacial scour and retreat on some of
these events. This sub-section describes individual events and then discusses general
associations observed between failure initiation and glacial debuttressing.

5.3.2.1 Major rockslope failures in the Mount Meager Area.

The Mount Meager area has been subject to many large landslide events, and debris
avalanche, lahar and debris flow deposits fill the valley of Meager Creek below its
confluence with Devastation Creek (Carter 1931; Evans 1990, 1992; Mokievsky-Zubok
Unfortunately, rapid erosion of the weak, fractured volcanic rock makes identification of
landslide source areas difficult for all but the most recent events. Development of a
chronosequence for these major events is in its preliminary stages and comes primarily
from studies of debris flow deposits in the lower basin areas (Jordan 1994, Jakob 1996).

Three major rockslope failure scarps were identifiable in the Mt. Meager area in locations
near glacially debuttressed slopes, including two failures in Capricorn Creek, one failure
in Devastation Creek, and one failure in Affliction Creek. In Capricorn Creek, two major
rock avalanche scars are visible on the north side of the creek about 2.5 km above the
confluence with Meager Creek. (Figure 21). One or both of these may have occurred in
1933-34, as evidenced by dendrochronologic studies on Capricorn Creek fan (Jakob
1996), and may have been as large as 10 million m$^3$ (Croft 1983).

The eastern of the two rock avalanche scars is the largest and extends approximately 350
metres along the north rim of the valley and 800 vertical metres from valley bottom to
ridgetop (Figure 22). Bedrock exposed by the landslide scar is quartz diorite with minor
dacitic dykes, but it has distinctly different geotechnical characteristics on either side of
the scar. The east side exposes massive, jointed rock with the dominant joint orientation
dipping steeply parallel to the slope, and several other strong joint sets dipping into and
orthogonal to the slope.
Figure 21. Rock Avalanche Scars on North side of Capricorn Creek, with failure scarps indicated on the inset diagram (circled). The smaller western failure is only partly visible in the photo; detachment occurred where a small snowpatch exists in the photo, and the slide travelled southeast down into the gully. The approximate location of the Neoglacial trimline is also shown.

The west side of the gully exposes a highly fractured, altered rock band that dips southwest oblique to the slope at approximately $110^\circ/40^\circ$ orientation. This west side is an extremely weak, altered rock mass with clay infilling along joints and fractures, and is actively eroding due to frequent rockfall along most sections of the slope.

The lower 140 m of the slope has been steepened by recent glacial scour (Figure 21) and Croft (1983) cited glacial debuttressing as a major causative factor for collapse of this section of hillslope. However, field observations from a traverse around the perimeter of
this failure do not support this hypothesis, nor do they support Croft’s contention that the entire landslide scar was created from a single catastrophic event. In addition, there is no landslide deposit in Capricorn Creek or Meager Creek that is large enough to suggest the simultaneous failure of this entire rock mass. Although it is certainly possible that glacial scour along the toe of the slope destabilized portions of the lower gully area, my interpretation is that catastrophic failure did not simultaneously occur over the entire gully. Rather, it seems that non-catastrophic erosion and retrogressive failures are enlarging and deepening the west side of the gully. In the gully headwall area there is a fresh detachment zone visible in Figure 23 where approximately $3 \times 10^5$ m$^2$ of rock may have failed catastrophically. Ongoing enlargement of the gully’s west side would have removed support from this section of the gully headwall, and it is possible that failure of the uppermost gully area triggered the large 1934 debris flow event that is indicated by tree ring records (Jakob 1996).
Unlike the failure described above, glacial debuttressing may have been a causative factor for a second, smaller rock avalanche that initiated at 1800 m elevation about 500 m west of the eastern failure. The lower 150 m of this section of hillslope has been significantly steepened by recent glacial scour, and a series of apparently inactive (?) uphill facing scarps exist above a steep gully (Figure 24, Figure 25). Bedrock exposed in the area of uphill facing scarps includes strongly jointed quartz-diorite rock and a small, isolated pendant of fractured sedimentary rock (possibly Gambier Group shale). The west side of the gully sidewall exposes the same highly fractured rock described in the
gully to the east, and the orientation of the contact between fractured and intact rock implies that there is shattered rock at about 10-50m below the surface. The uphill-facing scarps are truncated by a rock avalanche scar above the gully and it appears that a portion of the sagging hillslope has failed catastrophically. It appears that glacial oversteepening has been a significant factor causing instability through the removal of slope toe-support, causing slope deformation and catastrophic failure near the contact between shattered and intact granitic bedrock.

Figure 24. View looking west at antislope scarps in Capricorn Creek, about 100 m southwest of the western rock avalanche scar. The failure scarp is partially visible in the upper right photo (indicated by arrow). Note the fractured rock on the gully sidewall. Granitic rock exposed at the antislope scarps is strongly jointed but less fractured than rock exposed in the gully.
Figure 25. Capricorn Creek slope profile intersecting the western rock avalanche failure scarp and area of slope deformation. The dashed line indicates an approximate slope profile prior to glacial scour, based on the present slope gradient above the trimline.
Devastation Creek fan and the upper Meager Creek valley contain evidence for at least two major, post-Little Ice Age landslide events in 1931 and 1975. In October 1931, a 70,000 m$^3$/s debris flow descended Devastation creek and continued along Meager Creek to the Lillooet River confluence as a debris flood (Carter 1931, Jordan 1994). Because of intensive weathering since the time of the failure, the location and nature of the initiation zone within Devastation Creek cannot be precisely delineated.

In 1975, a large debris avalanche descended Devastation Creek, killing four B.C. Hydro geologists working on Devastation Creek (Patton 1976, Mokievsky-Zubok 1977, Read 1978). The source of the 1975 debris flow event was a rockslide that initiated on the south side of a large, down-dropped block east of middle Devastation Creek (Figure 26). This area contains numerous tension cracks and antislape scarps (described in section 5.3.3.3), and occurs above a slope that has been significantly oversteepened and debuttressed by Neoglacial scour and retreat. About 1.2x10$^7$ m$^3$ of material appears to have detached from the south end of this block (Baumann and EBA Engineering 2000), and several tension cracks and antislape scarps are truncated along the north edge of the failure. The failure scarp exposes densely jointed, moderately fractured andesitic rock, interlayered with more highly fractured flow breccia and altered tuff beds. Detachment seems to have occurred along joint planes parallel to the sidescarp and headwall areas. The close proximity of this catastrophic failure, and the ongoing slope deformation to a glacially scoured and steepened slope, strongly suggest that Neoglacial debuttressing and retreat were major causative factors for slope instability in this area of Devastation Creek.
Figure 26. View looking south at the initiation area of the 1975 rock avalanche/debris flow. The inset diagram shows the location of the failure scarp in Devastation basin (circled). The detachment zone includes the area snow-covered area and the steep cliffs directly above. Note the person's head for scale (circled). The person is sitting in an uphill-facing scarp trench that was truncated by the failure.

In lower Affliction Creek, a ~200m long rockslide scarp exists immediately above the Neoglacial trimline, in highly fractured, volcanic felspar porphyry (Figure 27). Although the exact timing of the failure is not known, the deposit still exists, indicating that it post-dates glacial retreat in this section of the valley. The timing of failure (post-dating glacial retreat) and occurrence immediately above a glacially oversteepened slope imply that glacial debuttressing was a significant causative factor for slope failure at this location. The landslide deposit likely temporarily dammed Affliction Creek, evidenced by incised fluvial terraces and minor lacustrine deposits in a floodplain immediately up-valley of the deposit.
Figure 27. Rockslide failure scarp in lower Affliction Creek, showing the position of failure immediately above a glacially oversteepened slope. The dashed line indicates an approximate slope profile prior to glacial scour, based on the present slope gradient above the trimline.
5.3.2.2 Rock avalanches in RY3 and RY4 basins.

In general, catastrophic failures and deep-seated instability are rare on granitic bedrock slopes close to recently glacially debuttressed terrain. Only 4 significant rock avalanche scarps were mapped, two in RY3 basin and one in RY4 basin.

The rock avalanche scarp in RY4 basin (Figure 14) could not be examined because of its location on an extremely steep cirque headwall. The feature was mapped based on its freshly weathered appearance on aerial photographs. Its location suggests that glacial steepening of the cirque headwall area may have been a factor influencing slope instability, but this cannot be assessed without an additional site investigation. The runout and subsequent entrainment of an end-moraine by this failure have been described earlier in Section 5.2.3.1.

In RY3 basin, all three rock avalanches initiate on slopes in the lower valley area, immediately down-valley of the Neoglacial end-moraine (Sites 1, 2, and 3 on Figure 28). Site 1 is a poorly defined rock avalanche scarp that was mapped at 1800 m elevation near the north ridge crest. This failure was not examined due to its inaccessibility and because it was not associated with recent glacial processes. Sites 2 and 3 occur at 1600 m elevation on the southwest slope of RY3, about 300 m west of the Neoglacial trimline (Figure 29 and Figure 30). The Neoglacial limit moraine exists between the failure initiation zones and the deposits for both failures, indicating that they occurred prior to the Neoglacial maxima and were not associated with recent glacial retreat. However, these two failures are still interesting because they occur at a distinct slope break parallel to the hillslope on both sides of the valley, about 200 vertical metres above the Neoglacial trimline (Figure 31, Figure 32). This has been interpreted as an older (likely Pleistocene-aged) glacial trimline with a much greater magnitude of slope debuttressing than that occurring during Holocene time. Numerous uphill-facing scarps and tension cracks occur in the vicinity of this feature near failure sites 2 and 3, and a high concentration of minor rockfall scarps and zones of flexural toppling also exist along this "trimline". Although a detailed kinematic analysis of this slope is beyond the scope of
the thesis, this area would be an excellent location for further investigation of the mechanics of natural hillslope debuttressing in granitic rock.

Figure 28. Location of rock avalanche scarps (dotted, numbered line segments), slope deformation features (solid line segments) and the Neoglacial limit (dashed line) in RY3 Basin. The present location of glacier ice is shown in grey. Slope profile X- Y is shown in Figure 32. Areas of slope deformation labeled “A” and “B” are described in the text.
Figure 29. Site 1 failure scarp on the west side of lower RY3 basin (arrowed). The cone-shaped deposit from Site 2 is visible on the left side of the photo (covered in snow), and the Neoglacial limit moraine is visible in the lower right part of the photo.

Figure 30. View to south of Site 2.
Figure 31. A) View looking south up RY3 Basin, and (B) looking north down middle and lower RY3 basin, showing a distinct convex slope break that appears to be an older (Pleistocene) glacial trimline.

Slope Profile X-Y, RY3 Basin.

Vertical exaggeration 1.5 X

Figure 32. Slope profile in RY3 basin along the X-Y line indicated in Figure 28.
5.3.3 Deep-seated slope deformation

Several study basins showed evidence of slow deformation of hillslopes in both granitic and volcanic rock. This section is limited to a discussion of locations where slope deformation has occurred immediately above glacially debuttressed hillslopes, including areas in Capricorn, Devastation, Affliction, Job and RY3 basins. Although the regional scale of this study prevented detailed structural investigation of any of these sites, this section presents field observations at each site and discusses the possible relations between glacial retreat and the observed slope sagging.

5.3.3.1 Affliction Creek.

Numerous tension cracks, antislope and downhill-facing scarps, collapsed pits and graben features exist in a gentle area at about 1700 m elevation on the western side of Affliction Creek (Site “1” on Figure 33, Figure 34). Many of these features show signs of recent movement, including freshly exposed earth and root systems stretched across tension cracks.

Slope deformation features occur in a ~30 m thick unit of steeply jointed, Holocene-aged olivene basalt that overlies Miocene-aged Quartz Monzonite basement rock throughout the plateau area (Bovis 1990). A sharp slope break occurs along the Neoglacial trimline immediately below this area, and the terrain drops steeply for about 100 m to the valley bottom.
Figure 33. Location of slope deformation features (numbered groups of line segments) discussed in the text, Neoglacial limits (solid line) and present ice limits (grey) in the Meager Area study basins.
Figure 34. View of the extensive area of slope sagging above the Neoglacial trimline on the west side of Affliction Creek. A) View looking west. B) View looking north. C) Slope profile along the line X-Y shown in (B).
Detailed investigations and kinematic analysis by Bovis (1982, 1990) and Bovis and Stewart (1998) suggest that a large volume of material (~3x10^7 m^3) is subject to ongoing, deep-seated deformation throughout this area. Dating and analysis of surficial deposits by Bovis (1990) suggest that slope deformation initiated in the northern part of the plateau less than 3000 years ago and has propagated southward through to the present time, with several major tension cracks cutting through Neoglacial drift. This result is consistent with progressive debuttressing of the lower hillslope area by ongoing, southward ice retreat. In addition, kinematic analysis by Bovis (1990) indicates that steepening of the main rock face by 5-10 degrees would be sufficient to cause flexural toppling and development of a deep-seated shear surface along the contact between the basalt and granitic basement rock. This is within the range of slope steepening caused by recent glacial scour. Evans and Clague (1994) also analyzed the failure geometry of Bovis (1990), using the Morgenstern-Price method. Assuming that the slope is currently at limiting equilibrium (factor of safety = 1), they found that post-Little Ice Age glacial thinning has decreased the factor of safety at this site by 15%. Consequently, recent retreat of Affliction glacier has been cited as a major causative factor for ongoing slope deformation in this area.

5.3.3.2 Capricorn Creek.

Antislope scarp features exist on both sides of the gullies containing the “1934” rock avalanche and the smaller rock avalanche scarp to its west and are marked as “2”, “3” and “4” on Figure 33. All of these areas were traversed on foot, but extensive surficial material cover prevented observation of rock structure at sites 3 and 4. However, intact soil, vegetation and trees growing across the scarps at sites 3 and 4 imply that no significant movement has occurred for some time. Antislope scarps at site 3 were previously described in section 5.3.2.1 because of their association with the catastrophic rockslope failure in this area. The close proximity of these deformation features to a recently glacially debuttressed slope implies that slope movement is possible, but no clear signs of movement were observed.
5.3.3.3 Devastation Creek

Two areas containing slope deformation features were mapped in Devastation Creek near slopes scoured by recent glaciation and are labeled as sites 5 and 6 on Figure 33. Site 5 is a bench underlain by interbedded andesitic flows, flow breccia and tuff layers on the east side of middle Devastation Creek between 1500 m and 1600 m elevation (Figure 33). This is also the site of the 1975 rock avalanche described in Section 5.3.2.1, and represents one of the best examples of a glacially debuttressed slope in the thesis study area.

Thesis fieldwork and previous work by Baumann and EBA Engineering (2000) defined a ~150 m x 370 m, gently sloping area containing numerous tension crack and antislope scarp features above a steep (40-50°) glacially scoured slope (Figure 35, Figure 36). Debris-covered ice still exists along the lowermost part of this slope although it is not known whether it is still providing some degree of buttress support. Baumann and EBA Engineering (2000) suggest that this entire area represents a block displaced up to 80 m along a scarp located immediately upslope of the bench. Massive quartz diorite rock outcrops along the toe of this steep slope, and it is possible that slope deformation and displacement is occurring along the contact between competent basement rock and the weaker overlying lithologies. Evidence for active movement exists in antislope scarp features along the edge of the steep slope break (Figure 35), some of which are truncated by the 1975 rock avalanche detachment zone. This, combined with active shallow rockfall activity and the 1975 rock avalanche, indicate that further catastrophic rockslope failures are likely in this glacially debuttressed area.
Figure 35. One of several tension cracks exposed at Site 5.

Figure 36. Slope profiles W-X and Y-Z across a deformed, glacially oversteepened slope in the vicinity of Site 5, Devastation Creek.
Site 6 is located at the confluence of the west and east forks of Devastation Glacier, between 1750 and 1900 metres elevation (Figure 33). A large, 10-20m high vertical scarp extends for 690 metres between the two glaciers and exposes columnar jointed rhyodacite. Weaker volcanoclastic rock and ash layers outcrop below the vertical face, and several antislope scarps are visible in colluvium below the major detachment. Baumann and EBA Engineering (2000) estimate the volume of failing material at about $4 \times 10^6 \text{ m}^3$, assuming an average depth of 20 m below the scarp face. Glacial debuttressing of the east and west flanks of this slope may be a significant destabilizing factor for slopes in this area, and ongoing glacial retreat will further debuttress the toe of this slope.

![Bedrock slumping and deformation at Site 6, Devastation Creek (Photo by M. Jakob).](image)

Figure 37. Bedrock slumping and deformation at Site 6, Devastation Creek (Photo by M. Jakob).

5.3.3.4 Job Creek

A series of antislope scarps exists from 1700 to 1900 m elevation on the west side of Job Creek (Site 7 on Figure 33). The deformation features occur immediately above a sharp convex slope break along the Neoglacial trimline and continue upslope to near ridgetop height. This area could not be traversed on foot so it is not known conclusively whether there is any evidence for ongoing slope movement, but a remote view (from $\sim 200 \text{ m}$ distance with binoculars) did not reveal any obvious signs of recent activity. Antislope
scarps are truncated on their north edge by a series of steep gullies with active rockfall and debris flow activity.

5.3.3.5 RY3 Creek

RY3 Creek contains the most extensive areas of slope deformation of any of the granitic study basins. Two zones of deformation occur in close proximity to recently glaciated terrain, sites A and B in Figure 28. Zone A is located above a steep slope break immediately down-valley of the Neoglacial limit. This area has been previously discussed in section 5.3.2.2 in association with two rock avalanches that also occur in this area. As discussed in section 5.3.2.2, the slope break running parallel to the slope is not associated with recent glaciation but may represent an older (Pleistocene) glacial trimline. Bedrock along this slope is strongly jointed with steep cataclinal and anaclinal joint planes. Based on airphoto interpretation and a helicopter over-flight, it appears that deep-seated toppling may have occurred due to glacial debuttressing.

Zone B occurs on the north side of upper RY3. Although this area is within the Neoglacial limit, it occurs immediately above a main valley glacial trimline (Figure 38). Several antislope scarps extend for about 200-300 metres across the hillslope in bedrock with dense vertical joints striking parallel to the slope. Several of these features show signs of active movement, with live roots stretched tightly across fresh cracks in soil, and fresh tears across live moss (Figure 39). Although the antislope scarps are immediately above recently deglaciated terrain, below the trimline there is no significant slope break caused by glacial scour and in fact glacial deposition (a lateral moraine) occurs below the features. Consequently, it is inconclusive as to whether glacial debuttressing was a factor causing activation of slope movement in this area.
Figure 38. Antislope scarp features on the northeast side of RY3 basin.

Figure 39. Closer view of one of the antislope scarp features shown in Figure 38.
5.3.4 Rock mass strength estimates and strength equilibrium rockslopes

As described in section 3.6.2, rock mass strength estimates were collected for 49 rock slopes within the Neoglacial Limit of granitic study basins. The purpose of this work was to assess whether these slopes were “oversteepened” according to Selby’s (1982) equilibrium envelope. Each RMS value was considered to be representative of the terrain polygon in which the measurement was taken, and results were plotted with respect to average slope gradients in each polygon (measured from the TRIM DEM) (Figure 40). All measured RMS values fell within the strength equilibrium envelope described in Section 2.4, suggesting that either Neoglacial scour did not oversteepen the hillslopes or that slopes have quickly adjusted to a strength equilibrium angle through active failure following glacial retreat. Since few landslide initiation zones in granitic basins were mapped within the Neoglacial limit, and since most instability upslope of the trimline was limited to shallow rockfall activity, this result suggests that recent glacial scour did not significantly oversteepen these rock slopes.

Although this RMS rating system was the only practicable approach given the regional scale of this work, there are three main areas of uncertainty that limit confidence in these results. First, it has never been shown that the position of the strength equilibrium envelope is applicable to granitic rockslopes within the thesis study basins, or indeed elsewhere in the Canadian Cordillera. Second, there is subjectivity and judgement involved in determining RMS ratings, such as selection of the dominant joint plane whose orientation most strongly controls stability of the hillslope, and judgement of an RMS rating that is “representative” of a terrain polygon. There would likely be higher consistency of measurement if the same worker had collected RMS ratings for both strength equilibrium and glacially scoured rockslopes, perhaps along transects that crossed the glacial trimline and allowed comparison to non recently glaciated terrain immediately upslope.
Figure 40. Slope gradient versus rock mass strength estimates for 49 granitic rock slopes within the Neoglacial limit. Strength equilibrium envelopes are from Selby (1982).
CHAPTER 6. STATISTICAL ANALYSIS RESULTS AND DISCUSSION

This section presents results for Failure Rate univariate analysis, logistic regression, and classification tree analysis. FR Analysis was conducted using detailed terrain attribute classes (Table 2, Section 4.5.3) and data were compiled from all study basins. Logistic regression and classification tree analysis were conducted using simplified terrain attribute classes (Table 3, Section 4.5.3), all landslide locations, and random samples of stable terrain.

6.1 Univariate Statistical Results, Failure Rate Analysis.

In Failure Rate (FR) analysis, each class of an attribute is assigned a “susceptibility score” that indicates the relative contribution of each class of an attribute to landsliding (see Section 4.7.1). Note that Susceptibility Scores indicate only the relative contribution to landsliding for classes within individual attributes; the values cannot be used for comparison between different attributes.

Although FR Analysis cannot identify interactions between more than one attribute and landslide activity, it does provide a useful preliminary understanding of the data. In addition, this approach considers all of the available data, unlike the multivariate statistical approaches used which include a random sample of stable terrain and simplified terrain categories.

6.1.1 Surficial Material Failures

Figure 41 shows FR analysis results for surficial material failures. The highest landslide spatial frequency occurs on slopes with gradients between 30 and 40°. Failures are typically concentrated in gullied terrain, and occur most commonly on material-controlled till slopes with concave slope profiles and on north-facing slopes. The highest
failure susceptibility lies within the Neoglacial limit. Within Zone 2 (mid-basin area), the highest density of failures occurs within 20% of the distance between the trimline and valley bottom. Considered as a whole (both terrain within and outside the Neoglacial limit), terrain in Zones 1 and 2 show approximately similar landslide susceptibility, and both have higher Susceptibility Scores than the cirque area (Zone 3). This is reflected in the Susceptibility Scores for elevation, which indicates high values for elevations less than 1400 m and also from 1800 to 2000 m. The highest density of surficial failures occurred on terrain underlain by volcanic rock (the Meager Creek Volcanic Complex).

Field observations are generally consistent with most FR analysis results. The highest concentrations of failures were observed just below the glacial trimline, on lateral moraines that were often gullied and had slope gradients of 30-45 degrees. Cirques had few surficial material failures because little surficial material is currently exposed in these areas (although further glacial retreat may increase the area of potentially unstable surficial material). The Mt. Meager area basins had a higher overall landslide activity than the granitic basins, and this is reflected in the high Susceptibility Score for basins underlain by volcanic rock.

Several Susceptibility Scores disagree with qualitative conclusions drawn from field observations. For example, the surficial material class “colluvial veneer” has a slightly higher score than the “till veneer” class. In contrast, more failures were observed during fieldwork in till than in colluvium. The higher score for colluvium may reflect a limitation of simplified categorical terrain data, where surficial material classes can contain up to 40% “other material” (e.g. till) that contains landslides.

In addition, Susceptibility Scores for drainage classes do not reflect the field observation that most surficial material failures occurred on slopes with high upslope drainage area. This reflects a limitation of univariate statistical methods, which cannot identify combinations of attributes associated with landslide activity. Unstable slopes with high drainage also had relatively steep slope gradients, and this is not reflected in
Figure 41. Susceptibility Scores for failures in surficial material. Note that score values show relative differences in landslide susceptibility within each attribute and should not be used for comparison between different attributes.
Susceptibility Scores that include gentle, stable valley bottom locations with high upslope drainage area.

6.1.2 Rockslope failures

Figure 42 shows FR analysis results for rockslope failures. In general, the highest density of rockslope failures occurs in volcanic rock (Meager Creek Volcanic Complex). Most failures occurred between 2600m and 2800m. In Zone 2, most failures occurred near the ridgetop (80-100% of the distance from the Neoglacial trimline to the ridgetop). Most failures occurred on slope gradients above 40 degrees, and on slopes with small upslope drainage area (<100 m²). The highest density of failures occurred in Zone 2 (mid-valley area), on gullied slopes outside the Neoglacial trimline, and the highest failure susceptibility occurred on west-facing slopes. Susceptibility scores were also highest on slopes with convex curvatures in both plan (across-slope) and profile (down-slope) orientations.

In general, most FR analysis results agree with qualitative conclusions drawn from field observations. In most basins, rockfall activity was most active along convex slope breaks such as ridgetops, the top of gully sidewalls, and along glacial trimlines. The highest concentration of failures was observed in Zone 2, with variable activity between the ridgetop and the glacial trimline. There was also a marked decrease in rockfall activity below the trimline, because slopes were scoured clean by recent glaciation and because slope gradients were generally lower near the valley bottom.

However, some FR analysis results did not identify patterns indicated by field observations. For example, the concentrations of rockfall activity observed along glacial trimlines in Zone 2 were not identified by FR analysis. A possible explanation for this result is that FR analysis produces a score generalized over all trimlines, but concentrations of rockslope instability were observed only in locations where terrain steepened across the trimline.
Figure 42. Susceptibility Scores for rockslope failures. Note that score values show relative differences in landslide susceptibility within each attribute and should not be used for comparison between different attributes.
6.2 Logistic Regression

Wald Stepwise Logistic Regression was run using SPSS to determine whether a non-linear statistical model could describe the interaction between a small number of terrain attributes and landslide occurrences. If successful, this approach would provide a useful predictive tool whereby measurable attributes could be entered into the model to determine landslide probabilities. However, analysis results yielded a complex equation that could only successfully predict 56% of the failures (Appendix B). Consequently this approach was abandoned in favour of more visually intuitive classification trees that sort data into ranked categories based on terrain attributes that differentiate landsliding and non-landsliding terrain. Classification tree analysis also can be readily incorporated into a landslide hazard assessment flowchart that includes descriptive field evidence.

6.3 Classification Tree Analysis

This section presents results for classification tree analysis of ten data sets that consist of all failure locations and ten different random samples of locations from stable terrain. Results are presented for analysis of terrain data within all basin zones and analysis of data stratified into basin Zones 1, 2 and 3.

6.3.1 Rock Slope Failures

Analysis of ten data sets containing data for all basin zones yielded two groups of results, due to differences between random samples of stable terrain. Five decision trees indicate “slope” as the relatively most important (principal) attribute for discriminating landslide from non-landslide locations, and five decision trees indicate “gullied” as the best attribute distinguishing landslide and non-landslide locations (Appendix B). Within each group, trees are similar for the first few splits and then differ as the data is further split into smaller sub-populations. Between each of the two groups, overall tree structures are
not similar because initial division based on different principal nodes means that further analysis occurs on different data subsets.

Figure 43A shows the classification tree with “slope” as principal node, including only branches consistent between data sets. As shown in Figure 43, the following terrain is most associated with landslide locations, in rank order of importance:

1) Slopes steeper than 40°.
2) Gullied slopes on terrain that is gentler than 40°.
3) Non-gullied terrain with less than 40° slope gradient that is located in Zone 2, outside the Neoglacial Limit.

Figure 43B shows the classification tree with “gullied” as principal node, including only branches consistent between data sets. According to Figure 43B, the following terrain is most associated with landslide locations, in rank order of importance:

1. Gullied terrain above 1400 m elevation
2. Non-gullied terrain that is steeper than 40°
3. Non-gullied terrain with less than 40° slope gradient that is located in Zone 2, outside of the Neoglacial Limit.

Although results are different between these two groups of results, they are both geomorphologically plausible and generally agree with qualitative field observations and FR analysis. One surprising result was that the attribute “geology” was not an important part of any classification tree, since field observation and FR analysis results indicated that the highest spatial frequency of rockslope failures was in volcanic rock. It is possible that geology is indirectly an important attribute in the classification trees because there is a disproportionately high amount of gullied terrain in volcanic rock. In a random sample of 10,000 grid cells from the study area, about 22% of the gullied terrain occurs in volcanic rock, even though volcanic rock comprises only 12% of the sample.
Figure 43. Classification trees for rock slope failures. Histograms within each box indicate the relative proportion of landslide (bold line) and stable locations at each node. End-nodes discussed in the text are numbered to the right of the box. A) Slope as principal node. B) Gullied as principal node.
Figure 44 presents classification tree analysis results stratified by basin zones 1, 2 and 3. As shown in Figure 44A, the following terrain is most associated with landslide locations in Zone 1, in rank order of importance:

1. Non-granitic bedrock (volcanic, metamorphic or older volcanic/sedimentary rock).
2. Granitic bedrock above 1400 m elevation with a low topographic index (<1.2).

Terrain most associated with landslide locations in Zone 2 includes:
1. Terrain near ridgetop height (greater than 80% of the distance from the Neoglacial trimline to the ridge).
2. Terrain from 20% to 80% of the distance from the Neoglacial trimline to the ridge with slope gradients steeper than 40°.

Terrain most associated with landslide locations in Zone 3 includes:
1. Gullied terrain
2. Non-gullied terrain with slope gradients steeper than 40°.

In general, statistical results in each Zone are supported by field observations. In Zone 1, the strong association between landslide activity and non-granitic rock types is primarily due to the relatively lower rock strength of these lithologies. However, this result should be interpreted with some caution because distinct differences in slope morphometry in Zone 1 were observed between granitic and non-granitic study basins. The non-granitic basins tended to be more deeply incised in the lower valley areas and therefore were steeper and more prone to landslide activity.

Field observations showed that the highest density of failures occurs in gullied, steep terrain near ridgetop in both Zone 2 and 3. As with the FR analysis, a second concentration of landslide activity near the glacial trimline is not indicated in the analysis results. Likely this result is because increased landslide density near the trimline only
Figure 44. Classification tree analysis results for rockslope failures, stratified by basin zones. Histograms within each box indicate the relative proportion of landslide (bold line) and stable locations at each node. End-nodes discussed in the text are numbered to the right of the box. A.) Zone 1. B) Zone 2. C) Zone 3.
occurred in areas of increased glacial scouring, where there was a distinct convex slope break across the trimline.

6.3.2 Surficial Failures

In all ten data sets, "Surficial Material Type" is selected as the principal attribute (node) for discriminating landslide and non-landslide locations, and Figure 45 presents results that are most consistent between data sets. As shown in Figure 45, the terrain most associated with landslide activity includes terrain steeper than 30° with thick, material-controlled slopes of glacial drift, till mantles, or colluvial mantles. This is intuitive and is supported by field observations. Most failures occurred on lower hillslopes, particularly in lateral moraines and till mantles on rockslopes with slope gradients of at least slope class 4 (30-40°).

Figure 45. Classification trees for surficial material failures. Histograms within each box indicate the relative proportion of landslide (bold line) and stable locations at each node. End-nodes discussed in the text are numbered to the right of the box.

Figure 46 presents classification tree analysis results stratified by basin Zones 1, 2 and 3. As shown in Figure 46A, the following terrain is most associated with landslide locations in Zone 1, in rank order of importance:
Figure 46. Classification tree analysis results stratified by basin Zones. Histograms within each box indicate the relative proportion of landslide (bold line) and stable locations at each node. End-nodes discussed in the text are numbered to the right of the box. A.) Zone 1. B) Zone 2. C) Zone 3.
1. Gullied, north- or east-facing slopes
2. Non-gullied north- or east-facing slopes with slope gradients steeper than $30^\circ$

Terrain most associated with landslide locations in Zone 2 (Figure 46B) includes:
- Terrain steeper than $30^\circ$ that is covered in till mantles or material-controlled till slopes.

Terrain most associated with landslide locations in Zone 3 (Figure 46b) includes:
- Terrain below 2200 m that has material-controlled slopes of till or colluvium.

Statistical results for each basin Zone are generally consistent with qualitative field observations. Although no causative reason was found for the association between landslide activity and slope aspect in Zone 1, most surficial material failures in Zone 1 were mapped in Pleistocene-aged till or colluvium mantling dissected hillslopes. In Zone 2, statistical results are consistent with the field observation that most failures occurred near the crest of lateral moraines (material-controlled till slopes) or in pockets of till mantling steep rockslopes. In general, the statistical result indicating an association between landslide locations and elevations below 2200 m in Zone 3 is likely due to the upper cirque area containing relatively little surficial material. In most basins, failures in Zone 3 were observed primarily on material-controlled till slopes, except in Capricorn Creek where there is extensive slope instability in thick colluvium.
CHAPTER 7. CONCLUSION

In this thesis I examined relations between post-Little Ice Age Neoglacial retreat and landslide initiation within nineteen alpine basins in the Lillooet River and Rutherford Creek watersheds. This work included a landslide inventory, field and airphoto-based terrain mapping, and the use of GIS to examine statistical and process-based relations between terrain modified by Neoglacial processes and landslide activity.

Numerous examples of unstable terrain were documented near recently deglaciated areas, including debris slides (commonly transforming to debris flows), rockfall, rockslides, rock avalanches, and non-catastrophic, deep-seated slope movements. The magnitude of landslide response to post-Little Ice Age Neoglacial retreat ranged from high magnitude in the Mount Meager area to relatively low in many granitic basins.

Figure 47 and Figure 48 summarize the main findings of this thesis as flowcharts suitable for landslide hazard assessment in recently deglaciated areas. The purpose of these charts is to identify and categorize terrain susceptible to landslide activity in surficial material (Figure 47) and bedrock (Figure 48) that is related to Neoglacial retreat.

In each figure, basins are subdivided into three Zones: Zone 1 (down-valley of the Neoglacial Limit), Zone 2 (mid-valley area) and Zone 3 (cirque). Only Zones 2 and 3 are considered in the flowcharts, as they are the only areas directly subject to Neoglacial scouring and retreat.

For surficial material (Figure 47), the second subdivision defines two groups: Primary Neoglacial Effects and Secondary Neoglacial Effects. Primary Neoglacial Effects include Neoglacial processes that directly influence slope stability, such as deposition of material in areas prone to instability, or reduction of ice-buttress support to deposits following Neoglacial thinning or retreat. Secondary Neoglacial Effects include deposition of
Figure 47. Flowchart for identifying landslide hazards in surficial material associated with recent Neoglacial retreat.
Figure 47, Primary Effects continued:

- Landslide activity related to Neoglacial retreat occurs within the Neoglacial Limit.
- Landslide activity related to Neoglacial retreat occurs within or immediately above the Neoglacial Limit.

**Examples of hazardous locations**

- Slopes steeper than about 30°
- Deposits within about 20% of the distance from the Neoglacial Limit to the valley bottom
- Gullied slopes
- Deposits in locations with high upslope drainage areas
- Deposits on steep slopes underlain by smooth bedrock.

**Examples of hazardous locations**

- Deposits in locations with high upslope drainage areas, such as lateral moraines below gullied rock slopes.
- Deposits located at the edge of steep bedrock-controlled slope breaks, such as deposition at the lower edge of a hanging valley.

**Examples of hazardous locations**

- Deposits located at the edge of steep bedrock-controlled slope breaks, such as deposition immediately above a Neoglacial trimline or the lower edge of a hanging valley.

**Examples of hazardous locations**

- Cohesive material deposited against glacial ice that has remained at slope gradients above the friction angle following glacial thinning or retreat. Non-cohesive deposits (e.g. blocky talus slopes) are typically stable and settle non-catastrophically when subject to glacial debuttressing.
- Deposits in Zone 3 may be especially hazardous because slopes have had less time to re-equilibrate to a stable repose angle following ice retreat.
Figure 48. Flowchart for identifying landslide hazards on rockslopes associated with recent Neoglacial retreat

Subdivide basin into Zones

Zone A
Downvalley of Neoglacial limit

No Neoglacial Effects
(conventional hazard analysis)

Zone B
Mid-basin area between NGL and cirque

Is the terrain currently covered or buttressed by glacial ice?

No

Yes

Zone 3
Cirque

Map the pattern of ice retreat to identify terrain that will be exposed or debuttressed following further retreat.

Continued Next Page
Figure 48 continued

Bedrock Geology

Weak Rock
(e.g. volcanic suites, pervasively jointed sedimentary and metamorphic rock, finely jointed or altered igneous intrusives)

Strong Rock,
(e.g. massive granitic intrusives)

Terrain steepens (convex slope break) across the Neoglacial trimline?

Yes

No

Inside or Outside Neoglacial Limit

Outside

Inside

- High probability of ongoing shallow rockfall along the trimline.
- Tension crack and antilslope scarp features above the trimline indicate a stressed slope and the potential for catastrophic collapse.

Outside

- High probability of ongoing shallow rockfall on glacially steepened slopes.
- Potential for high magnitude detachments due to rupture of the toe of glacially oversteepened slopes.

Inside

- Low probability of increased landslide activity associated with Neoglacial retreat.

Outside

- Moderate probability of minor, shallow landslide activity immediately upslope of the Neoglacial trimline.
- High probability of catastrophic collapse if the slope was already prone to instability prior to Neoglacial scour and retreat.

Inside

- Slight probability of increased landslide activity from 0-20% of the distance between the Neoglacial trimline and valley bottom.
- Very low landslide probability in locations greater than 20% of the distance between the Neoglacial trimline and valley bottom.

- Very low probability of increased landslide activity associated with Neoglacial retreat.
- Landslide activity may be temporarily reduced where Neoglacial scour removed loose, weathered material below the Neoglacial trimline.
material in locations that are potentially subject to entrainment by failures initiating upslope, such as lateral moraines below unstable rockslopes. Potentially unstable types of terrain subject to Primary Neoglacial Effects are then categorized in the flowchart based on material type (glacial drift or colluvium) and whether the deposit is *rock-controlled* (mantles rockslopes without masking underlying topography) or *material-controlled* (deposit is thick enough to control slope morphometry).

For rockslopes (Figure 48), terrain subject to Neoglacial retreat is subdivided into two groups designated as *Weak Rock* (e.g. fractured or altered volcanic, sedimentary or metamorphic rock, or finely jointed or altered igneous intrusives) or *Strong Rock* (e.g. massive granitic rock). The distinction refers to the much higher ability of Neoglacial advance and retreat to scour, steepen and subsequently debuttress rockslopes composed of weak rock. The next division separates terrain into groups according to whether Neoglacial scour has steepened toe slopes, indicated by a convex slope break along the Neoglacial trimline. Terrain that does not have a convex slope break along the trimline is not subject to Neoglacial debuttressing and has little instability associated with post-Little Ice Age Neoglacial retreat. Landslide activity may be reduced in these areas where Neoglacial scour removed loose, weathered rock from lower hillslopes.

For terrain that steepens across the Neoglacial trimline, the level of landslide activity differs significantly above and below the trimline. In “strong” rock, most instability related to Neoglacial scour and retreat occurs as minor rockfall along the trimline, with little landslide activity inside the Neoglacial limit. Larger scale failures generally occur only where slopes were already prone to instability prior to Neoglacial scouring. In “weak” rock, slopes that steepen across the trimline have significant potential for deep-seated slope movement or large-scale, catastrophic failure. Often, evidence of deep-seated slope movement is a precursor to catastrophic failures such as the 1975 Devastation rockslide/debris flow. The rupture zone of such large events typically extends across the Neoglacial trimline, and active shallow rockfall is also likely along this line.
Work conducted in this thesis concerned only one of many factors that contribute to landslide activity in alpine basins. Nevertheless, the evidence presented strongly suggests that Neoglacial scour and retreat is a significant factor for increasing landslide density across Neoglacial trimlines, and for increasing the likelihood of catastrophic failures in areas already prone to instability. Perhaps most significantly, there is a significant spatial association between recent catastrophic failures and slopes that were oversteepened, then debuttressed, by Neoglacial erosion. All of the known catastrophic failures are associated with deep-seated gravitational slope deformations of the sackung type, implying that this type of deformation is a precursor to large-scale failures.

Several areas of future work would further increase the understanding of the relations between glacial retreat and landslide activity. First, more detailed slope stability analysis is needed to obtain a greater mechanical understanding of relations between glacial debuttressing and landslide activity. Most significantly, work is required to identify the conditions whereby slow deformation of glacially debuttressed slopes leads to catastrophic failure. Examples of locations that warrant further slope stability analysis include the July 30, 1998 Capricorn Creek debris slide, the 1975 Devastation Creek debris avalanche, and the two major rock slides in gravitationally deformed slopes on the southwest side of lower RY3 basin, within Ryan River basin.

Second, the high landslide density identified in recently deglaciated areas may be responsible for an increase in debris flow frequency and magnitude in the main valley stems of some basins. Further work is needed to determine the relations between glacial retreat rates, and debris flow frequencies and magnitudes, because glacial retreat is expected to be ongoing, and because debris flows can cause significant damage to people and structures at large distances from their source location. Although this work has been conducted previously in Switzerland (Zimmerman and Haeberli 1992) and in Capricorn Creek, BC (Bovis and Jakob 2000), the relation is still poorly understood at a regional scale.
Finally, work for this thesis was conducted in a relatively small region with approximately homogenous climatic conditions and a limited diversity of bedrock and surficial geology. Further work is required to test the applicability of results from this thesis under different climatic and geologic conditions.
REFERENCES


APPENDIX A. INDIVIDUAL STUDY BASIN DESCRIPTIONS

This section describes the geomorphology and surficial and bedrock geology for each study basin. Summary statistics are given in Table 1.

Table 6. Summary statistics.

<table>
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<tr>
<th>Study basin</th>
<th>Total area (km²)</th>
<th>Bedrock lithology</th>
<th>Percent glacial retreat (by area) since Neoglacial maximum</th>
<th>Number of rapid rockslope failures</th>
<th>Spatial frequency of rapid rockslope failures (# failures / km² ice-free terrain)</th>
<th>Number of rapid surficial failures</th>
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Capricorn Creek basin is one of the larger basins draining the south flank of the Meager Creek Volcanic Complex at the northern end of the volcanic Garibaldi Belt. The basin drains a glaciated area in its upper reaches, and encompasses over 2100 m of relief from the summit of Mt. Meager to the confluence of Capricorn and Meager Creeks. The upper part of Capricorn basin (above about 1700 m) is underlain by late Pliocene to Holocene aged volcanic rock, and highly fractured quartz diorite outcrops in the lower part of the basin (Read 1978, Woodsworth 1977).

Glacial processes and active landslide activity are the dominant controls on terrain development in Capricorn basin. A distinct glacial trimline indicates that Capricorn glacier covered about 6.3 km$^2$ of the basin at the time of the Neoglacial maximum (about 150 years B.P.), and has since receded by 3 km and downwasted by about 200m (Bovis and Jakob 2000), a total areal retreat of about 43%. This has resulted in widespread debuttressing and steepening of bedrock and drift-covered slopes.

Capricorn Creek is subject to frequent mass movement activity, since least 13 major landslide events may have occurred in the last 330 years, including those in 1998, 1972, 1944 and 1933 (Jakob 1996). The most recent 1998 event initiated as a $\sim$5x10$^5$ m$^3$ debris slide in thick colluvium in the upper basin area. The failure then transformed into a debris flow that traveled the entire length of Capricorn Creek, entraining material to reach a total volume of about 1.0x10$^6$ m$^3$. This landslide occurred in an area that was previously covered with glacial ice, and Bovis and Jakob (2000) conclude that glacial debuttressing of the colluvial scarp was a significant factor in this failure. Bovis and Jakob, as well as Croft (1983) also note two major rock avalanches that initiated on slopes undercut by Capricorn glacier during the Little Ice Age, although it is inconclusive as to whether glacial debuttressing was a factor in these events.
Devastation Creek lies at the southwestern corner of the Meager Creek Volcanic complex. Strongly jointed quartz diorite basement rock is exposed in the basin up to about 1500 metres elevation, and is overlain by Quaternary dacitic lavas. Minor areas of extremely weak, altered “rock” outcrop in several areas, including pyroclastic and sedimentary deposits that are likely Pleistocene till overridden by recent lava. Glaciers have experienced about 38% of areal retreat since the Neoglacial maximum.

Devastation Creek is an extremely geomorphically active basin with evidence of active slope instability that includes areas of slow rockslope deformation, major rock avalanche scarps, frequent debris slides and rockfall, and major debris flow deposits in the main valley stem. Rockslope instability is much more common than instability of surficial material, although debris flows triggered by rockslope failures are the largest hazard to humans and developments. Several major debris flows have been documented in this valley. In October 1931, a large debris flow initiated above Devastation Creek and travelled 15 km down Meager Creek to the Lillooet River confluence (Carter 1931). With an instantaneous discharge estimated at 70,000 m$^3$/s, this is the largest debris flow event ever documented in Canada (Jordan 1994). The headscarp of this major event has not been defined, but based on the texture of the deposit it may have initiated on the east side of the valley (Baumann and EBA Engineering 2000). In July 1975 a 1.2x10$^7$ m$^3$ rockslide initiated on the north side of the mid-basin area and triggered a large debris flow in the main valley stem. The debris flow travelled down the valley to form a large landslide dam at the Meager Creek confluence. Four BC Hydro employees working at the mouth of the stream were killed.

Down-valley of the Neoglacial limit, most slope instability is concentrated on the steeper west side of the valley, including areas of rockslope deformation with antislope scarps traversing the midslope area (Bovis and Evans 1996). Rockfall and debris flow activity are also common at the head of a large gully on the basin’s west flank. The lower west
slope exposes rusty weathered, densely jointed, fractured granitic basement rock that appears to be slightly altered.

The west slope of the mid-basin area is a uniform volcanic rock slope with weak horizontal bedding and numerous, minor, shallow rockslope failures. The lower east side of the mid-basin has been significantly oversteepened by Neoglacial scour. Extensive areas of antislope scarps and tension cracks exist upslope of the trimline, and this area was the site of the major, 1975 debris avalanche (see section 5.3.2.1, Site 5 in Section 5.3.3.3). A large, down-dropped block and antislope scarp features also occur in upper Devastation creek and are described in Section 5.3.3.3 (Site 6).

Slope failures on the mid- to upper- east slope of Devastation basin are strongly controlled by rock lithology. Volcanic rock in this area includes layers of altered and fractured volcanoclastic rock overlain by more competent dacitic rock, and there is a high concentration of failures along this contact. Numerous minor rockfall scarps also exist along the ridgetop at the head of the cirque.

A3 Mosaic Creek

Mosaic Creek is a basin underlain primarily by Mesozoic quartz monzonite rock of the Coast Plutonic Complex, contacting the Meager Creek Volcanic Complex along the eastern boundary of the study basin. Greenstones of the Lower Cretaceous Gambier Group are also exposed in the lowermost basin area. The granitic rock is medium to fine grained, and is generally massive and widely jointed along the valley bottom. On upper valley slopes, granitic rock is more densely jointed, fractured and slightly altered, especially near the contact with the Meager Creek Volcanic Complex. Rare dacitic dykes were also observed. The glacier in this study basin has experienced about 49% areal retreat since the Neoglacial maximum.

The western side of the mid-basin area contains several shallow rockslide initiation zones about 100-200 m above the glacial trimline. It appears that the scarps may have formed
through retrogressive failure initiating near the trimline, but a detailed field assessment would be required to test this hypothesis. Debris slides in till are extremely common near the main valley trimline in the mid-basin area. A cirque basin exists above the sharp slope break defined by the western trimline, and debris slides in till are initiating where small, first-order streams intersect this slope break. Downvalley of the Neoglacial limit, Mosaic Creek narrows to form a steep, V-shaped canyon. Rockfall is common on the upper part of these steep rockslopes but no surficial material failures or deep-seated rockslope instability were observed. Antislope scarp features were also mapped above the glacial trimline on the west side of the basin, although it is unknown whether glacial debuttressing is a factor in their development.

A4 Affliction Creek

Affliction Creek is located at the western edge of the Meager Creek Volcanic Complex. Quartz monzonite basement rock crops out along the southwestern ridge of the study basin and much of the lower slopes and valley bottom areas. This is overlain by basalt and andesite flows interbedded with highly friable breccia and ash layers. Affliction Glacier has diminished about 36% by area since the Neoglacial maximum.

Landslide activity is relatively frequent in Affliction Creek with active rockfall occurring on steep slopes throughout the basin. Neoglacial scour and retreat have significantly steepened the lower west slope and caused extensive areas of deep-seated slope instability (see Section 5.3.3.1). A rockslide scarp also exists immediately above the glacial trimline on the mid-lower west side of the valley (see Section 5.3.2.1). The Neoglacial readvance extended to within about 700 m of the mouth of Affliction valley. The small part of the study area downvalley of the Neoglacial limit contains areas of sporadic rockfall but no significant failures in either rock or surficial material.
A5  Job Creek

In Job Creek, the lower valley area and lowermost slopes of the mid-valley area are underlain by quartz monzonite and minor exposures of Cadwalder Group amphibolite. The upper slopes of the mid basin area and the cirque area are underlain by porphyritic andesite, flow breccia and ash layers of the Meager Creek Volcanic Complex. Much of the bedrock is highly altered, jointed and fractured. Job Glacier has experienced about 35% of areal reduction since the Neoglacial maximum.

Job Creek is an very geomorphically active basin with a higher concentration of failures than any other study basin. Most failures are shallow rockslope failures, primarily rockfall. The mid basin area and upper cirque contain extensive exposures of extremely fractured, altered volcanic rock and have a very high spatial frequency of rockfall and shallow rockslide scarps. Most failures begin at slope breaks, either at ridgetop level, along aretes or along the glacial trimline. An area of antislope scarps also exists above the trimline on the west side of the mid-basin, although no evidence for active movement along the antislope scarps was observed.

Very few areas of surficial material instability were observed in Job Creek. However, debris flows from several gullies on the upper west side of the mid basin area have impacted a lateral moraine along the lower valley slope, causing gully erosion and larger debris flow magnitudes as morainal material is entrained.

A6  Rainy Creek

Rainy Creek is located in Rutherford Creek basin and is entirely underlain by quartz diorite rock of the Coast Plutonic Complex. The glacier in the upper basin area has experienced about 54% of areal reduction since the Neoglacial maximum.

The mid- and upper- basin areas contain relatively moderate topographic relief and few failures. However, the Neoglacial limit extends to the edge of a sharp slope break where
Rainy Creek descends more steeply to Rutherford Creek, and glacial drift ravelling off the lateral moraines has apparently contributed to debris flow activity in the main valley stem (Jakob 1996). In the steeper area down-valley of the Neoglacial limit, slope instability is primarily rockfall at the ridgetop level above steep, rock-controlled gullies.

A7 “R1” and “R3” Creeks.

“R1” and “R3” basins are located in the upper Rutherford Creek Watershed, and contain alpine glaciers originating in the Pemberton Ice-Cap. R1 and R3 glaciers have retreated about 51% and 54% in area, respectively, since the Neoglacial maximum. Both basins are underlain by massive, grey-weathered quartz diorite of the Coast Plutonic Complex. In general, basin orientations follow the structural trend of the major joint discontinuities in the valley, which include joints dipping 30°-40° and 70°-90° to the west, and a near-vertical, southeast striking joint. Glacially scoured slopes are generally structurally controlled with slopes parallel to the dominant joint orientation, with a stepped profile along the steep joint sets.

In general, rockslopes and surficial material in R1 and R3 are relatively stable both above and below the Neoglacial limit. Most mass movement activity in the upper valley areas is minor rockfall and block toppling, although a small area of open joints and uphill facing scarps exists on the north side of R3. Rock at this site is biotite-rich, strongly foliated, steeply jointed rock and slope sagging has occurred immediately above one location at the glacial trimline. Surficial material failures are dominantly small debris slides on the crest of lateral moraines and in till pockets on recently ice-free rockslopes. Downvalley of the Neoglacial limit, slope instability is confined to rockfall on the south ridgetop and two minor debris slides in colluvium on the lower north slope of R1.

A8 “PE3” Creek

PE3 study basin is a small basin draining into Petersen Creek that has experienced about a 55% areal retreat since the Neoglacial maximum. The basin is underlain by quartz
diiorite and granodiorite of the Coast Plutonic Complex. In general, both rockslopes and surficial material are relatively stable in this basin. The mid-basin area is steepest on the north and east valley sides, and glacial scour has significantly steepened the lower hillslope. Most slope instability involves rockfall and is concentrated along the glacial trimline. Two small antislope scarps were also mapped on the upper hillslope area. The cirque area of PE3 is relatively gentle, and contains no mapped slope failures. Terrain downvalley of the Neoglacial limit is partially vegetated and contains generally moderate slope gradients except for the the upper north side of the valley. Four minor debris slides in thin colluvium and two minor rockfall scarps are visible in this area.

A9 "PE1" basin.

"PE1" study basin forms part of upper Petersen Creek in Ryan River watershed. The basin is entirely underlain by Coast Plutonic Complex quartz diorite and granodiorite. Joints are dominantly gentle or near-vertical with two major sets dipping steeply to the west and north. A band of highly fractured rock exists along the west side of upper PE1 basin, striking at roughly 100 Azimuth. PE1 Glacier has retreated about 47% by area since the Neoglacial maximum.

Most rockslopes and surficial material in PE1 are stable. In the mid and upper basin areas, minor rockslides and rockfall occur at ridgetop and near the Neoglacial trimline, but no major failures were observed. Southeast of the lake in the lower basin area the bedrock is relatively more fractured than in the rest of the basin, and rockfall scarps exist at the head of several bedrock gullies.

A10 "RY3" basin

"RY3" basin is underlain by massive, strongly jointed quartz diorite rock of the Coast Plutonic Complex, with localized gneissic layering. Some areas of dense jointing occur on the mid-basin area, where the dominant joints dip steeply to the north and to the
southwest, and moderately to the northeast. RY3 Glacier has retreated about 51% by area since the Neoglacial maximum.

RY3 basin contains some of the most significant rockslope instability found in any of the granitic study basins, including at least three major rock avalanches, several areas with uphill facing scarps, and numerous rockfall scarps. Landslide activity in the mid-basin area is distinctly different on the northeast and southwest sides. The upper northeast valley slope (upslope of the Neoglacial trimline) is relatively more vegetated and weathered. Most rockslope failures occur as minor rockfall in gullies. A poorly defined rock avalanche scarp and deposit were also mapped on the northeast valley side above a gully draining just below the Neoglacial limit end-moraine (see Section 5.3.2.2). Most surficial material failures occur in thick till scarps as debris slides at the crest of lateral moraines. Most of these debris slides transform into debris flows and some have traveled out over the surface of the glacier. Slope sagging features (uphill facing scarps) occur on the north side of the mid-upper basin, immediately above the Neoglacial trimline. These are described in Section 5.3.3.5.

The southwest side of the mid-basin contains several cirque glaciers. Rockslope instability (primarily rockfall) is concentrated along a distinct slope break about 150-200 metres above the glacial trimline, along what may be an older (Pleistocene) glacial trimline. Several possible rock avalanche scarps exist in this area but the lack of rock avalanche deposits at the slope base implies that failures occurred prior to the latest glacial advance. Joint orientations in this area are similar to those further down the valley (although more widely spaced), with one major joint set dipping very steeply out of the slope and a second set dipping steeply into the slope. Surficial material failures include debris slides in till at the at the crest of lateral moraines, and in remnant till pockets on recently ice-free rockslopes.

Downvalley of the Neoglacial limit, most slope instability occurs on rockslopes and is concentrated near a distinct slope break on the southwest side of the valley that appears to
be an older (Pleistocene?) glacial trimline. This area of instability is described in Section 5.3.2.2.

A11 “RY4” basin

“RY4” basin is tributary to Ryan River and is underlain by massive, strongly jointed quartz diorite of the Coast Plutonic Complex. Several major joint sets exist, dipping steeply to the northwest, northeast and southeast. RY4 glacier has diminished about 57% in area since the Neoglacial maximum.

In general, RY4 has a moderately high spatial frequency of rockslope failures in the mid-to lower-basin and very few surficial material failures. The most significant failure is a large rock and debris avalanche that initiated out of a small hanging valley on the west side of the main basin; this failure is described in Section 5.2.4. In the mid-basin area, the west side of RY4 is a very steep rockslope with minor Neoglacial scour. There appears to be a second, higher glacial trimline (?) that is at a similar elevation to the previously mentioned trimline in lower RY3. The dominant joint in this area dips steeply northeast downslope, and rockslope instability occurs primarily along this joint set near the older, higher “trimline”. The east side of RY4 contains much closer-spaced joints with several minor, shallow rockslide scarps on the upper slope.

Down-valley of the Neoglacial limit, the east slope of the basin contains moderate, treed slopes and no observed failures, but the west slope contains several debris slides in thin colluvium and several shallow rockfall scarps along the edge of a large block that appears to have slumped.

A12 “RY6” basin

“RY6” basin is tributary to Ryan River underlain by Coast Plutonic Complex quartz diorite and granodiorite on the western side of the basin, and older (Miocene?) sedimentary rock in parts of the eastern part of the basin (Woodsworth 1977). The
sedimentary rock is fine-grained, quartzitic sandstone, rusty-weathered and highly fractured, and numerous minor, shallow rockslide scarps exist on the upper part of the east side of the basin. RY6 glacier has diminished about 59% in area since the Neoglacial maximum.

RY6 contains numerous rockfall failure scarps, but relatively little surficial material instability except for one active debris flow channel near on the lower west side of the basin (see Section 5.2.2). Downvalley of the Neoglacial limit, the west side of the basin contains several rockslide initiation zones and active rockfall in highly fractured and jointed bedrock. Although the dominant joint set dips steeply into the slope in this area, a strong second set dips moderately downslope and appears to be a factor in reducing slope stability.

The east side of RY6, both in the mid-basin area and down-valley of the Neoglacial limit, contains numerous shallow rockslide initiation zones in highly fractured, rusty weathered sedimentary rock. Parts of this slope are deeply dissected and most failures form debris flows that travel to the bottom of the slope.

A13 “RY1” basin

“RY1” basin is one of the headwater drainages for the Ryan River basin. The upper main basin and Northwest part of the upper sub-basin are underlain by massive, well-jointed quartz diorite. The lowermost slopes north and south of the lake, and as far west as site KH91 expose areas of highly deformed gneissic rock of unknown age. Weakly layered, vertically jointed volcanic rock overlies the granitic and gneissic bedrock in the north-central part of the study area. Intrusive rock in the vicinity of the volcanics is rusty weathered and moderately fractured, with numerous dyke intrusions parallel to joints (e.g. site KH82). Several sets of joints exist, including two near-vertical sets striking southwest and northeast, and several major joints dipping moderately steeply to the southeast. Rockslopes are generally structurally controlled, and have stepped profiles
across the major joint orientations. The ice area in RY1 basin has been reduced by about 51% since the Neoglacial maximum.

Mass movement processes, mostly debris slides and resulting debris flows, are relatively common in the mid-basin area. Since 1991 (the airphoto flight year used for mapping), landslide deposition into a valley bottom lake has reduced the width of the lake by about one-third. Landslides originate from two main source areas on the north and south sides of the basin. On the south side of the mid-valley area, a lateral moraine contains numerous debris slides initiating from the crest of the moraine (west of site KH 79). Most of these debris slides transform into debris flows down gullies eroded into the moraine. West of the lake (west of site KH85), the lateral moraine was deposited below a rockslope that drains onto the moraine. Material thickness is much thinner in this area and large sections have failed down to bedrock. On the north side of the main RY1 basin, a second basin forms a hanging valley. A glacial trimline defines the slope break and slopes are composed of both rock and steep till scarps. The till scarps are actively failing along the trimline and form several active debris flow channels. On the east side of the trimline (near sites KH 97,98), till veneers cover the lower part of the valley slope. Debris slides initiate in several of these till pockets and have travelled as debris flows into the lake. Many of these debris slides have formed in areas of concentrated drainage from the slope above.

A14  “S5” basin

“S5” basin is located in upper South Creek, draining into the upper Lillooet River, and is underlain by quartz monzonite rock of the Coast Plutonic Complex. In general, this study basin has a relatively low spatial failure frequency and most slope instability occurs as rockfall at ridgetop level. The area of S5 glacier has been reduced by about 45% since the Neoglacial maximum.

In the mid-basin area, the north side of S5 is a uniform to shallowly dissected, steep rockslope. Although there is a very distinct glacial trimline (indicated by lighter colored,
freshly weathered rock exposures below the trimline), glacial scour does not appear to have significantly steepened the slope gradient. Most slope instability occurs as rockfall along the mid to upper slopes along the north side of the basin, especially in gullied areas. The south side of the basin contains relatively gentle terrain, and no slope instability was mapped except for four minor debris slides along lateral moraines.

Slopes down-valley of the Neoglacial limit are vegetated rockslopes partially covered in vegetation and a thin colluvial veneer. The south side of the valley is highly dissected with rock-controlled gullies and talus slopes. Slope instability is limited to this gullied area with active rockfall in the gully headwalls and debris flow activity in the channels.

A15 “S2” basin

“S2” basin is located in upper South Creek, draining into the upper Lillooet River, and is entirely underlain by quartz diorite and quartz monzonite of the Coast Plutonic Complex. The glaciated area in S2 has been reduced by about 73% since the Neoglacial maximum, one of the two highest magnitudes of areal retreat in this study. S2 basin contains a relatively low spatial frequency of slope failures (4 failures/ km²), and all occur in bedrock. The upper cirque area is gentle, heavily glaciated, and contains no mapped failures.

A16 “S1” basin

“S1” study area is located in upper South Creek, draining into the upper Lillooet River, and is underlain by quartz diorite and quartz monzonite of the Coast Plutonic Complex. Glaciated areas in this study basin have been reduced by about 26% since the Neoglacial maximum.

The north and south sides of S1 study area are quite different in character. The north side is a uniformly steep rockslope, gullied towards the lower (eastern) end of the basin. Most slope instability occurs on the north valley side occurs at the head of these gullies.
Colluvial slopes below the gullies have filled the upslope trench of a lateral moraine at the slope base. This moraine is beyond the angle of repose of colluvium and a sharp contact exists between the colluvial slope above and the lateral moraine below. Despite this sharp increase in slope break below the unconsolidated colluvium, no failures were mapped.

The south side of S1 is heavily glaciated with glaciers extending to the top of the cirque. Slope instability is minimal, except for two debris slides in till along the trimline and rockfall on gully headwalls on the dissected eastern side of the south slope.

**A17 “P3” basin**

“P3” study basin drains into the Lillooet River about 7 km downstream of the confluence with Meager Creek. It is entirely underlain by quartz diorite and granodiorite rock of the Coast Plutonic Complex. Glaciated areas in P3 have been reduced by about 75% since the Neoglacial maximum, the largest magnitude of areal retreat of any basin in this study. In general, PE3 contains relatively little slope instability and most involves rockfall and debris flow activity on the steep, gullied slopes on the west side of the middle and lower basin areas.

Slope instability in the mid-basin area occurs primarily as active rockfall at ridge-top level on the west side of the basin. Very little slope instability occurs near the Neoglacial trimline, except for one small debris slide on a till scarp and one rockfall scarp on the Neoglacial trimline. The only slope instability observed in the upper cirque area is rockfall activity at ridge-top level.

Downvalley of the Neoglacial limit, PE3 is a relatively steep-sided valley with vegetated, gullied rockslopes covered in a colluvial veneer. Based on airphoto interpretation, shallow debris slides are common in the gully headwall and sidewall areas and are triggering debris flow activity within the gullies.
“HA1” study area is entirely underlain by quartz diorite rock of the Coast Plutonic Complex. Glaciated areas in HA1 have been reduced by about 42% since the Neoglacial maximum.

HA1 basin contains a moderate spatial frequency of slope failures, and most failures involve rockfall on the north and east sides of the basin. Most slope instability in HA1 occurs as rockfall on the mid to uppermost slopes of the north and east valley sides. The relatively gentle south slopes contain little slope instability except for one small area in the western-most part of the study area where several small, closely spaced shallow rockslides occur in highly fractured granitic rock.
## APPENDIX B. LIST OF AIRPHOTOS.

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BCC94102 No.42-45  
BCC94117 No.5,6 |
APPENDIX C. STATISTICAL ANALYSIS RESULTS

B1 Wald Forward Stepwise Logistic Regression Analysis (SPSS)

Note: this analysis was conducted on random sample no. 3 of 10.

1. Rockslope Failures

Total number of cases: 1723 (Unweighted)
Number of selected cases: 1723
Number of unselected cases: 0

Number of selected cases: 1723
Number rejected because of missing data: 0
Number of cases included in the analysis: 1723

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Log Likelihood decreased by less than .01 percent.

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Goodness of Fit  1709.677
Cox & Snell - R^2  0.185
Nagelkerke - R^2  0.248

Chi-Square  df  Significance
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2. Surficial Material Failures

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Number of unselected cases: 0

Number of selected cases: 1280
Number rejected because of missing data: 0
Number of cases included in the analysis: 1280
Dependent Variable Encoding:

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#### Dependent Variable

**FAILTYPE**

#### Variable(s) Entered on Step Number

**8.** IONGL

#### Estimation Details

- **-2 Log Likelihood**: 1075.149
- **Goodness of Fit**: 1699.164

Estimation terminated at iteration number 6 because Log Likelihood decreased by less than .01 percent.
### Cox & Snell - $R^2$
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### Classification Table for FAILTYPE
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*Overall 81.95%*

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B2 Decision Tree Results (Statistica)

1. Random sample No. 1

Classification Tree for FAILTYPE
Number of splits = 5; Number of terminal nodes = 6

Rock tree all zones

Classification Tree for FAILTYPE
Number of splits = 5; Number of terminal nodes = 6

Rock tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 2; Number of terminal nodes = 3

Rock tree Zone 2

Classification Tree for FAILTYPE
Number of splits = 2; Number of terminal nodes = 3

Rock tree Zone 3
Classification Tree for FAILTYPE
Number of splits = 10; Number of terminal nodes = 11

Surficial tree all Zones
Classification Tree for FAILTYPE

Number of splits = 5; Number of terminal nodes = 6

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 11; Number of terminal nodes = 12

Surficial tree Zone 2
Classification Tree for FAILTYPE

Number of splits = 1; Number of terminal nodes = 2

Surficial tree Zone 3
2. Random sample No. 2

Classification Tree for FAILTYPE
Number of splits = 15; Number of terminal nodes = 16

Rock tree all Zones
Classification Tree for FAILTYPE

Number of splits = 12; Number of terminal nodes = 13

Rock tree Zone 1
Classification Tree for FAILTYPE

Number of splits = 3; Number of terminal nodes = 4

MVNGLDIS=1, 2, 3, 5

MVNGLDIS=4

DSLOPE < 3.7403

Rock tree Zone 2
Classification Tree for FAIL TYPE

Number of splits = 2; Number of terminal nodes = 3

GULLIED = 1

DSLOPE \leq 4.5051

Rock tree Zone 3
Classification Tree for

Number of splits = 23; Number of terminal nodes =
Classification Tree for FAILTYPE
Number of splits = 5; Number of terminal nodes = 6

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 12; Number of terminal nodes = 13

Surficial tree Zone 2
3. Random sample No. 3

Classification Tree for FAILTYPE
Number of splits = 9; Number of terminal nodes = 10

Rock tree all Zones
Classification Tree for FAILTYPE
Number of splits = 6; Number of terminal nodes = 7

Rock tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 3; Number of terminal nodes = 4

Rock tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 2; Number of terminal nodes = 3

Rock tree Zone 3
Classification Tree for FAILTYPE
Number of splits = 13; Number of terminal nodes = 14

Surficial tree all Zones
Classification Tree for FAULTYPE

Number of splits = 4; Number of terminal nodes = 5

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 5; Number of terminal nodes = 6

Surficial tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 2; Number of terminal nodes = 3

Surficial tree Zone 3
4. Random Sample No. 4

Rock tree all Zones
Classification Tree for FAULTYPE

Number of splits = 7; Number of terminal nodes = 8

Rock tree Zone 1
Classification Tree for FAULTYPE

Number of splits = 10; Number of terminal nodes = 11
Classification Tree for FAILTYPE

Number of splits = 2; Number of terminal nodes = 3

Rock tree Zone 3
Classification Tree for FAILTYPE
Number of splits = 28; Number of terminal nodes = 29

Surficial Tree all Zones
Classification Tree for FAILTYPE

Number of splits = 12; Number of terminal nodes = 13

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 11; Number of terminal nodes = 12

Surficial tree Zone 2
Classification Tree for FAILTYPE

Number of splits = 5; Number of terminal nodes = 6

Surficial tree Zone 3
5. Random Sample No. 5

Classification Tree for FAILTYPE
Number of splits = 6; Number of terminal nodes = 7

Rock tree all Zones
Classification Tree for FAILTYPE
Number of splits = 6; Number of terminal nodes = 7

Rock tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 2; Number of terminal nodes = 3

217
MVNGLDIS=1,2

126
DSLOPE≤3.623

2

0

3

1

696

570

4

0

5

1

Rock tree Zone 2
Classification Tree for FAILTYPE

Number of splits = 3; Number of terminal nodes = 4

Rock tree Zone 3
Classification Tree for FAILTYPE

Number of splits = 7; Number of terminal nodes = 8

Surficial tree all Zones
Classification Tree for FAILTYPE
Number of splits = 8; Number of terminal nodes = 9

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 4; Number of terminal nodes = 5

Surficial tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 11; Number of terminal nodes = 12

Surficial tree Zone 3
6. Random sample No. 6

Classification Tree for FAILTYPE
Number of splits = 12; Number of terminal nodes = 13

Rock tree all Zones
Classification Tree for FAILTYPE
Number of splits = 3; Number of terminal nodes = 4

Rock tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 5; Number of terminal nodes = 6

Rock tree Zone 3
Classification Tree for FAILTYPE
Number of splits = 9; Number of terminal nodes = 10

Surficial tree all Zones
Classification Tree for FAILTYPE
Number of splits = 4; Number of terminal nodes = 5

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 6; Number of terminal nodes = 7

Surficial tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 4; Number of terminal nodes = 5

Surficial tree Zone 3
7. Random sample No. 7

Classification Tree for FAILTYPE
Number of splits = 12; Number of terminal nodes = 13

Rock tree all Zones
Rock tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 3; Number of terminal nodes = 4

Rock tree Zone 2.
Classification Tree for FAILTYPE
Number of splits = 5; Number of terminal nodes = 6

Rock tree Zone 3
Classification Tree for FAILTYPE
Number of splits = 9; Number of terminal nodes = 10

Surficial tree all Zones
Classification Tree for FAILTYPE

Number of splits = 4; Number of terminal nodes = 5
Classification Tree for FAILTYPE
Number of splits = 6; Number of terminal nodes = 7

Surficial tree Zone 2
Classification Tree for FAILTYPE

Number of splits = 4; Number of terminal nodes = 5

Surficial tree Zone 3
8. Random sample No. 8

Classification Tree for FAILTYPE
Number of splits = 10; Number of terminal nodes = 11

Rock tree all Zones
Classification Tree for FALLTYPE

Number of splits = 6; Number of terminal nodes = 7

Rock tree Zone 1
Classification Tree for FAILTYPE

Number of splits = 3; Number of terminal nodes = 4

Rock tree Zone 2
Classification Tree for FAILTYPE

Number of splits = 2; Number of terminal nodes = 3

Rock tree Zone 3
Classification Tree for FAILTYPE

Number of splits = 18; Number of terminal nodes = 19

Surficial tree all Zones
Classification Tree for FAILTYPE

Number of splits = 1; Number of terminal nodes = 2

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 4; Number of terminal nodes = 5

Surficial tree Zone 2
Classification Tree for FAILTYPE

Number of splits = 7; Number of terminal nodes = 8

Surficial tree Zone 3
9. Random sample No. 9

Classification Tree for FAILTYPE

Number of splits = 7; Number of terminal nodes = 8

Rock tree all Zones
Classification Tree for FAULTYPE

Number of splits = 6; Number of terminal nodes = 7
Classification Tree for FAILTYPE
Number of splits = 6; Number of terminal nodes = 7

Rock tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 5; Number of terminal nodes = 6

Rock tree Zone 3
Classification Tree for FAILTYPE
Number of splits = 8; Number of terminal nodes = 9

Surficial tree all Zones
Classification Tree for FAILTYPE

Number of splits = 1; Number of terminal nodes = 2

Surficial tree Zone 1
Classification Tree for FAILTYPE
Number of splits = 6; Number of terminal nodes = 7

Surficial tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 16; Number of terminal nodes = 17

Surficial tree Zone 3
10. Random sample No. 10

Classification Tree for FAILTYPE
Number of splits = 25; Number of terminal nodes = 26

Rock tree all Zones
Classification Tree for Fault Type

Number of splits = 4; Number of terminal nodes = 5
Classification Tree for FAILTYPE
Number of splits = 8; Number of terminal nodes = 9

Rock tree Zone 2
Classification Tree for FAILTYPE
Number of splits = 2; Number of terminal nodes = 3

Rock tree Zone 3
Classification Tree for FAILTYPE

Number of splits = 7; Number of terminal nodes = 8

Surficial tree all Zones
Classification Tree for FAILTYPE
Number of splits = 11; Number of terminal nodes = 12

Surficial tree Zone 2
Surficial tree Zone 3

- **Stable**
- **Landslide**