CHARACTERIZATION OF THE ZYMOETZ RIVER ROCK AVALANCHE

by

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Simon Fraser University, 2002

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Earth Sciences

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March 2005

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ABSTRACT

On June 8, 2002, the Pacific Northern Gas pipeline in the Zymoetz River valley was severed over a distance of tens of meters by a large debris flow. The event initiated as a rock avalanche in Glen Falls Creek, a tributary of the Zymoetz River. The rock avalanche involved $1 \times 10^6 \text{ m}^3$ of volcaniclastic bedrock, and travelled through a complex flow path, to finally deposit a large fan in the main Zymoetz River. Approximately half of the debris volume was deposited in the cirque basin at the head of the valley, with the rest deposited in the channel, and the fan. Photoanalysis software has offered insights into the grain size distributions throughout the deposit. Application of the modelling programs DAN-W /DAN3D has significantly increased our understanding of the dynamics of the event.
DEDICATION

I would like to dedicate this work to my family for their unfaltering support throughout everything that I do. Their pride in me made this work possible.
Thank you to Dr. Doug Stead for his constant support throughout this entire project, I don't know of a professor more dedicated to his students. Thank you also to Marten Geertsema and Jim Schwab of the B.C. Ministry of Forests for introducing me to this project, and for their many valuable conversations throughout. I wish to thank Dr. Peter Jordan, also of the Ministry of Forests, for introducing me to the world of geomorphology and landslide studies. Finally I thank Derek Kinakin and Marc-Andre Brideau for their valuable insights and technical help throughout my time at SFU.
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LIST OF SYMBOLS

$H_i$ height of boundary block
$P$ tangential internal pressure resultant
$T$ basal resisting force
$ds$ nominal length of a boundary block
$\alpha$ slope angle
$\delta$ bulk unit weight
$a_c$ centripetal acceleration $= \frac{v_i^2}{R}$
$r_u$ pore-pressure coefficient
$\Phi$ friction angle
$v_i$ velocity
$g$ gravity acceleration
$\xi$ turbulence coefficient
$R$ vertical curvature radius of the path
$C_U$ coefficient of uniformity
$D_{60}$ diameter of particle that 60% of the deposit is finer than
$D_{10}$ diameter of particle that 10% of the deposit is finer than
$C_C$ coefficient of curvature
$D_{30}$ diameter of particle that 30% of the deposit is finer than
$W_B$ explosive energy consumed (kWh/ton)
$K$ fragmentability
$S_b$ size of block before a blast
$S_a$ size of block after a blast
$W_C$ work required to crush material (sWh/short ton)
$W_i$ Bonds work index
$d_{50}$ final diameter after crushing
$D_{50}$ initial diameter after crushing
CHAPTER 1 - Introduction and debris flow / avalanche theory

1.1 Introduction

At approximately 1:30 am on June 8, 2002 the Pacific Northern Gas (PNG) pipeline in the Zymoetz River Valley was severed by a large debris flow (Cavers, 2003; Schwab et al., 2003; Boultbee et al., in press) (Figure 1-1). The event initiated as a rockslide when a $1 \times 10^6$ m$^3$ block of bedrock detached from the eastern wall of the cirque basin at the head of the valley. The principal objective of this thesis is to provide a detailed characterization of this event. Characterization follows the methodology described by Couture et al. (1999) for rock avalanche analysis, modified to accommodate advances in technology (Figure 1-2).

Long-runout failures such as the Zymoetz River rock avalanche (ZRRA) represent an increasing hazard in British Columbia. They have the potential to be a very high risk, if they occur in populated areas or transportation corridors. Where failures occur in unpopulated areas, they may still pose a risk to fisheries and forestry operations. This chapter presents a literature review describing the classification of landslides, mechanics of debris flows and rock avalanches, and climatic variables involved in the initiation of debris flows / avalanches. Chapter 2 provides a detailed description of the ZRRA and a comparison to similar recent events in British Columbia. The initiation, transport and depositional zones of the ZRRA are described. Chapter 3 includes an engineering geological characterization of the failure deposit and rock mass. Chapter 4 investigates the use of a photoanalysis program to determine grain size distributions and comminution for the event. Chapter 5 uses the rheological models DANW and DAN-3D to further characterize the event. Finally Chapter 6 includes a review of the current work and further discussion.
Figure 1.1 – Map showing the location of the Zymoetz River rock avalanche (ZRRA) and five other recent landslides in British Columbia.
Figure 1-2. Modified methodology used in the characterization of the ZRRA (based on Couture et al., 1999).
1.2 Classification of landslides

Varnes (1978) developed a classification of landslides based on both the type of material and movement. The materials defined are rock, debris and earth. The type of movements are fall, topple, slide, spread and flow. Debris is defined by Cruden and Varnes (1996) as containing a significant amount of coarse material, with 20 to 80% of particles >2mm. Following Cruden and Varnes (1996), flow is defined as a spatially continuous movement in which surfaces of shear are short-lived, closely spaced, and usually not preserved. The distribution of velocities in the displacing mass is typical of a viscous liquid. Debris flow is defined by Jordan (1994, p. 1) as "the rapid flow of a mixture of debris and water which is sufficiently saturated, that the entire weight of debris can be borne by pore pressure, and which has a sufficiently high sediment concentration that it behaves as a single phase slurry, in which fluid and large clasts do not separate upon deposition." There are different types of debris flows, classified according to location. Open-slope debris flows form their own path, while channellized debris flows follow an existing channel (Cruden and Varnes, 1996).

Avalanches are generally associated with snow, but debris and rock avalanches are also common. An avalanche is defined by Trenhaille (1990) as a rapid flow of material downslope in a mountainous area. Rock avalanche was first used to describe the Frank slide by McConnell and Brock (1904), and is a simplification of the "rockslide-debris avalanche" proposed by Varnes (1978) (Hungr et al., 2001; Hungr and Evans, 2004a). Hungr and Evans (2004a) recently proposed a new classification system for rockslides based on the role of the rock structure and the mechanical properties of the rock mass (Table 1-1). Their classification labels rockslides according to the type of movement that led to failure. Hungr and Evans (2004a) state that large rock avalanches (>20M m³) are rare, with a frequency of 1/500 to
1/5000 per 10,000 km² of mountain terrain. Frequencies reported for smaller events (>2M m³) are much higher at 1/80 to 1/800 per year per 10,000 km².

Table 1-1. Proposed classification relating the failure mechanics of large rockslides to rock mass structure and strength (From Hungr and Evans (2004a), by permission).

<table>
<thead>
<tr>
<th>Structural Control</th>
<th>No systematic structural control</th>
<th>Systematic structural control</th>
<th>Toppling</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kinematics Constraint</td>
<td>Translational Unconstrained</td>
<td>Sliding</td>
<td>Compound</td>
</tr>
<tr>
<td>Dominant Mechanism</td>
<td></td>
<td>Sliding</td>
<td>Compound</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Unconstrained</td>
<td>Constrained</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mechanism Type</td>
<td>A Rock slump</td>
<td>C Block slide</td>
<td>D Structurally defined compound slide</td>
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<td>Typical behaviour in weak rock</td>
<td>Rock collapse</td>
<td>Slow, rotational movement (Plateau d'Assy)</td>
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<td>Typical behaviour in strong rock</td>
<td>Catastrophic collapse, steep slopes, (Elm)</td>
<td>Catastrophic, limited pre-failure deformation (Goldau)</td>
<td>Catastrophic, large pre-failure deformation (Valont)</td>
</tr>
</tbody>
</table>

1.3 Flow mechanics

1.3.1 Rheological classification

When discussing debris flows and rock / debris avalanches, it is important to consider rheology, the study of deformation and flow. Rheology provides information on operative processes within flows / avalanches, and a means to analyze them according to pre-existing methods, including computer modeling. Pierson and Costa (1987) propose a classification that reflects the rheologic behaviour of movement involving water and earth materials. They define flow as "irreversible deformation of a physical substance, as a result of an applied stress of sufficient magnitude" (Pierson and Costa, 1987, p. 2). In
the case of landslides the applied stress is gravity, which usually acts as a disturbing or shear stress.

The behaviour of a flow is a function of (1) the relative proportions of the multi-phase components, (2) the grain-size distribution of the solids, and (3) the physical and chemical properties of the solids (Pierson and Costa, 1987). Pierson and Costa (1987) base their classification of flow type on mean flow velocity and sediment concentration (Figure 1-3). They state that mean flow velocity is a surrogate for rate of shear, and flow processes move at characteristic rates that are possible to measure or estimate, thus making it a logical axis of their classification matrix. Sediment concentration is the other axis of their classification matrix because rheologic response of a flow is primarily a function of sediment concentration. The boundaries defined in the classification of Pierson and Costa (1987) are approximate, as field measurements of flow events were limited.

The following types of flows are defined in the classification scheme of Pierson and Costa (1987). Apparent liquid flow is divided into streamflow and hyperconcentrated flow. Streamflow is flowing water with a sufficiently small amount of sediment, so that the flow behaviour is unaffected by the presence of the sediment. Streamflow is considered to be a Newtonian fluid, unless the special case of flocculated clays is encountered, where the fluid obtains a yield strength and is then considered a non-Newtonian fluid. Hyperconcentrated flow is defined as a flowing mixture of water and sediment possessing measurable yield strength yet still flowing like a liquid.

The category of plastic fluids considers streamflow, slurry flow and granular flow (Figure 1-4). As sediment concentration increases in a fluid so does the yield strength, and eventually an abrupt increase in the yield strength signals the onset of internal friction. Slurry flow is defined as a flow having sufficient yield strength to exhibit plastic flow behaviour, and yet becomes partially
liquefied as it is remoulded. A slurry flow is a saturated mixture, with pore water trapped within the framework of the grains, and a sufficient pore water pressure to partially support the solid phase. A viscous slurry flow occurs where the silt-clay content is very high, or the shear rate, mean grain size, grain density and water content are relatively low. Under these conditions viscous forces control the flow behaviour. Inertial slurry flows occur where momentum is transferred through particle collisions, the viscosity of the pore fluid is relatively low, and the shear rates, grain size, density and water content are high.

Granular flow occurs when the amount of pore water is so small that there is very little pore water pressure, and thus the entire weight of the mass is borne by grain-to-grain contact. A granular flow may still be saturated if it has a grain size distribution and a low enough shear rate to allow water to escape during remoulding. Inertial granular flow occurs when grain collisions transfer momentum between particles, resulting in a dispersive stress. Rapid inertial granular flows were defined by Pierson and Costa (1987) as flows in which momentum transfer by energetic inter-particle collisions primarily determine flow behaviour. Frictional effects are minimal for such flows.

Pierson and Costa (1987) applied their classification to geomorphic processes such as debris flows and debris avalanches (Figure 1-4) by superimposing rheologically based boundaries. In their classification they applied the name debris flow to the inertial and viscous slurry flows, and debris avalanche to rapid inertial granular flow.

Iverson and Vallance (2001) provide further insight into the manner in which rheology evolves during granular mass flows. They suggest that the behaviour of a granular mass depends on the evolution of granular temperature (a scalar measure of grain vibration, i.e. mixture agitation) and pore-fluid pressure. This evolution precludes assessment of flow-rheology
from steady-state experiments or static deposits. Iverson and Vallance (2001) state that intergranular shear stresses obey the Coulomb equation and depend linearly on flow depth, grain concentration and pore-fluid pressure. In attempting to predict and interpret granular mass flows, they suggest efforts should be concentrated on trying to determine the evolution of fluid pressures, rather than on speculative interpretations of rheology.
SEDIMENT CONCENTRATION (VOL%)

Fast Inertial Forces Dominant

Normal Streamflow

Hyperconcentrated Streamflow

Inertial Slurry Flow

Viscous Granular Flow

Fluidized Granular Flow

Sediment Velocities never measured or estimated

No mechanism to suspend sediment

Interstitial Fluid

Flow Category

Flow Behaviour

<table>
<thead>
<tr>
<th>Fluid Type</th>
<th>Newtonian</th>
<th>Non-Newtonian</th>
</tr>
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<tbody>
<tr>
<td>Interstitial Fluid</td>
<td>Water</td>
<td>Water + fines</td>
</tr>
<tr>
<td>Water + Air + Fines</td>
<td>Streamflow</td>
<td>Slurry Flow</td>
</tr>
<tr>
<td>Granular Flow</td>
<td>Liquid</td>
<td>Plastic</td>
</tr>
</tbody>
</table>

Figure 1-3. Rheological classification of sediment water flows (from Pierson and Costa (1987), by permission).
Figure 1-4. Fitting existing flow names into the rheological classification (from Pierson and Costa (1987), by permission).
1.3.2 Debris flow mechanics

A debris flow is made up of 3 different components: the snout, the body and the tail (Figure 1-5). The front part is generally composed of cobbles and boulders with little matrix, while the trailing parts become progressively finer grained with higher water contents. The tail resembles muddy streamflow or hyperconcentrated flow (Jordan, 1994). The dynamics of a debris flow depend on the sediment texture, and debris flows behave according to one of the fluid models proposed by Pierson and Costa (1987). Different parts of a debris flow can act differently, depending on the sediment texture. For example, the coarse grained snout may behave as granular flow, while the finer grained, wetter tail may resemble a more viscous flow model (Jordan, 1994).

According to Takahashi (1991, p.63), "a debris flow will be produced when enough water is supplied to saturate the voids between particles, and it is mixed with a mass of earth and rock, which starts to move." He presents three ways in which this can occur: (1) a landslide block transforms into a debris flow while in motion; (2) the collapse of a debris dam generates a debris flow; and (3) the gully bed becomes unstable and produces a debris flow when it reaches saturation. Takahashi (1991) states that the coarse snout of a debris flow will move downstream with a constant velocity, while maintaining its bulbous shape. The trailing uniform part moves with a rolling motion, where the upper parts are moving faster than the lower parts. Large boulders are retained at the front of the flow by either entrainment or migration toward the snout by a kinematic sieve mechanism (Iverson, 1997). Inverse grading is commonly observed in debris flow deposits, and different mechanisms have been proposed for this process, including: Bagnold's dispersive pressure, lift force, a kinematic sieve mechanism, and buoyancy in the grain mixture (Takahashi, 1991; Iverson, 1997). Iverson (1997) notes that granular temperature (agitation of the grains) is involved in the kinetic sieving mechanism and has an influence on the bulk density of the debris.
Takahashi (1991) demonstrated that in the Bingham fluid model, the yield strength is strong enough to produce a plug near the surface, allowing large boulders to be supported by the combined strength and buoyancy of the plug, rather than by turbulence or dispersive pressures. However a debris flow may contain sufficient water so as to act as a slurry only capable of supporting particles up to 3 mm in size. To support significantly larger particles, as observed in many debris flows, a yield strength must be produced by the effect of grain interactions, which is improbable in thin debris flows. Takahashi (1991) suggests that it is not the yield strength that increases in the debris flow, but the buoyancy. In theory, a very high fluid pressure should be required to support a clast, therefore all of the particle loads are transmitted directly onto the bed by particle interactions. If a debris flow has a large concentration of finer material, and therefore a greater density of interstitial fluid, the flow is able to transport much larger particles.

Pulses in a debris flow can occur in laminar flow, but not in turbulent flow. Surges or pulses in a flow have frequently been observed, and it has been suggested that they are associated with hydraulic instability (Davies, 1986, 1988; Iverson, 1997). This instability is inferred from observations that surges resemble roll-waves observed in shallow water flow, for example, shallow water flow in culverts, or on pavement after heavy rainfall (Jordan, 1994). Observations in United States Geological Survey (USGS) experimental flumes show that surges can develop spontaneously without any perturbations of the flow (Iverson, 1997).

Takahashi (1991) states that when a debris flow enters an area where the channel has a shallower grade, it decelerates, thickens and eventually stops. Deposition may also occur where the channel widens and becomes unconfined (Jordan, 1994). The denser the interstitial fluids and the lower the concentration of coarser debris, the flatter the slope must be for the debris flow to stop, or in other words, the longer the runout will be (Takahashi,
Field observations indicate that the snout of a debris flow will stop, while some of the latter parts continue to flow over top of the halted snout, and deposit further downstream.

Major and Iverson (1999) conducted flume experiments to examine the processes involved in debris flow deposition. They discovered that deposition resulted from grain-contact friction and bed friction concentrated at flow margins. This is contradictory to other models that explain deposition as a result of a reduction in excess pore water pressures. Their experiments showed that deposition of liquefied debris can occur when the debris flow movement is impeded by grain friction at the margins of the flow, where the sediment is coarsest and fluid pressures are lowest. While the edges of the debris remain stationary, the deposit interior can continue to be liquefied and weak until the higher pore pressures dissipate during consolidation (Major and Iverson, 1999).

Figure 1-5. Components of a debris flow.
1.3.3 Rock / debris avalanche mechanics

Different mechanisms have been suggested for flow in debris avalanches, but the exact mechanics remain uncertain (Melosh, 1987). Varnes (1978) suggested that the 1970 Huascaran debris avalanche was cushioned by air and steam. Contributions due to ice and snow may also have resulted in its long runout and high velocity. Compressed air, dense dust clouds, vaporized interstitial water, and mud have all been proposed as possible cushions for debris in these avalanches (Melosh, 1987). Pierson and Costa (1987) summarize different ideas proposed for the flow of debris avalanches including:

- fluidization involving dust and air;
- mechanical dilation under shear and separation of particles through collision and rebound;
- vacuum-induced upward flow of air from the airfoil shape of the upper flow surface;
- steam from frictional heating of entrained moisture;
- grain separation due to acoustical energy from the noise of the flow; and
- frictional melting of a thin layer of rock debris at the base of the flow.

Davies et al. (1999) also summarize proposed theories and divide them into those that reduce the basal friction and those that reduce internal friction. Through simple experiments with beads on glass, they noted that reduced basal friction increases translation of the mass, but not spreading. This is contrary to field observations, where large debris avalanches deposit close to the toe of the slope, and their long runout is due to spreading and not translation. Davies et al. (1999) claim that in order to produce increased spreading, without increased translation, you would need reduced internal friction. The reduced internal friction models seem to agree more closely with field observations. These models include air and acoustic fluidization;
However, physical evidence has yet to be observed (Davies et al., 1999). A
momentum-transfer model was proposed by Van Gassen and Cruden (1989)
that involves rockslide debris being driven forward by rock fall to the rear.
However, it is unclear why this mechanism should only apply to large volumes
of material, as in debris avalanches (Davies et al, 1999).

Davies et al. (1999) recently proposed another mechanism in which
fragmentation of the rock contributes to the long runout of debris avalanches.
They suggest that fragmentation processes generate an isotropic dispersive
stress within the moving mass causing the proximal part of the translating
mass to decelerate more rapidly and deposit near the toe of the slope, and
the distal part to decelerate more slowly and thus deposit further away. They
observe that large rock avalanches usually carry debris much farther relative
to their fall height, compared to smaller rock avalanches. Davies et al. (1999,
p.1103) describe the fragmentation process as:

“A rock avalanche begins as the detachment of a relatively
coherent rock mass from a mountain side; the mass
immediately begins to collapse into successively smaller and
more numerous joint-determined fragments as it moves down
the mountain side...field evidence suggests strongly that the
fragmentation process continues throughout the whole runout.”

Locat et al. (2003) further explain the fragmentation process, suggesting that
after a certain degree of initial crushing and joint-determined collapse, the
material becomes finer, and breakage occurs along mineralogical
imperfections and micro-heterogeneities. Fragmentation leads to an increase
in the amount of energy required to further fragment the rock (Figure 1-6).
Evidence of fragmentation is observed where the bulk of the subsurface
deposit material in long-runout rock avalanches is thoroughly broken /
crushed parent material (Davies et al., 1999). Rocks can also be found that
have been shattered, but are undisaggregated, showing that fragmentation
does actually occur. Davies et al. (1999) claim that the reverse grading seen
in many deposits suggests that fragmentation requires increased overburden pressures and intergranular contact stress with depth.

During fragmentation, the applied stress instantaneously exceeds the strength of the rock, so fragmentation occurs explosively, causing the accumulated strain energy to be released as very high fragment velocities in all directions (Davies et al., 1999). McCarr (1997) reports fragment velocities of 10 m/s and more when rock is shattered in this manner. The effect of these high velocity particles interacting within the slide mass is an isotropic dispersive stress. These fragmentation-induced dispersive forces would result in slow deceleration of the front part of the slide mass and more rapid deceleration of the rear part (Davies et al., 1999). The lateral component of the dispersive forces will cause lateral spreading, if the mass is not confined. According to Lucchitta (1979), spreading has been observed in many Martian deposits, where there is little lateral confinement.

Locat et al. (2003) have adapted techniques developed in the mining industry to assess the fragmentation energy of a rock avalanche. Their technique directly links the initial block size and the grain size distribution of the debris. After producing grain size distributions for the two zones using photographic analysis techniques, the distributions are compared using the following equations for $W_B$ (the explosive energy consumed) and $W_C$ (the work required to crush material from an initial diameter $D_{50}$ to a final diameter $d_{50}$), where $K$ is a measure of fragmentability, $S_b$ and $S_a$ are the block size before and after the blast and $W_i$ is Bonds' work index, which depends on the rock type:

$$W_B = K \left( \frac{S_b}{S_a} \right)^{1/5}$$

(Eqn. 1-1)

$$W_C = 10W_i \left( d_{50}^{-1/2} - D_{50}^{-1/2} \right)$$

(Eqn. 1-2)
Their study of seven rock avalanches showed that fragmentation energy ranged from 1% to 33% of the potential energy of a rock avalanche. They indicate that further work is required to fully understand the role of fragmentation in rock avalanche motion.

Figure 1-6. Graph showing the increase in fragmentation energy for decreasing particle size (from Locat et al. (2003), by permission).

Hungr and Evans (2004b) and Hungr et al. (2005) discuss another long runout theory involving entrainment of saturated substrate material. This theory had been previously proposed to explain the influence of a saturated substrate at the Frank slide where most of the damage was caused by the impact of a lateral outflow of mud from the debris as it overran the floodplain of the Old Man River at the base of the slope (McConnell and Brock, 1904; cf. Hungr and Evans 2004b). Similar observations have been made at the Elm slide in Switzerland (Heim, 1932; cf. Davies et al., 1999) and at numerous other slides around the world. Hungr and Evans (2004b) state that initial fragmentation of the debris in the first stages of the failure, causes a
volumetric bulking of ~25%, after which the volume is increased further by
entrainment of debris that is liquefied by rapid undrained loading (Hutchinson
and Bhandari, 1971). Numerous observations have been made that relate
the volume of a landslide to the total runout distance. Hungr and Evans
(2004b) claim that the runout distance is only weakly related to volume, and
the main control is the availability of liquefiable sediments that larger
landslides are more likely to encounter. They propose use of Varnes’ (1978),
“rockslide – debris avalanche” for landslides that initiate with the failure of a
rock slope but entrain significant amounts of debris, yielding an entrainment
ratio (ER) (Eqn. 1-3) of at least 0.25.

(Eqn. 1-3) \[ ER = \frac{V_{ENTRAINED}}{V_{FRAGMENTED}} \]

### 1.4 Climatic conditions related to initiation

Weather is important in triggering slope failures, and can be divided into two
components, short-term and long-term events (Crosta, 2004). Short-duration,
intense storms trigger shallow landslides, whereas long-duration, less intense
storms more commonly are associated with deep-seated failures and failures
of clayey soils (Crosta, 2004). Takahashi (1991) states that rainfall intensity
is more important than total rainfall amount in triggering debris flows.
Antecedent soil moisture conditions can predispose slopes to fail, and
bedrock close to the surface (within 1.2 m) is also a significant factor (Crosta,
pressure generation in rainfall-induced landslides. They discovered that the
presence of fines in the material led to development of flow slides, whereas a
lack of fines is associated with retrogressive failures.
Given the importance of rainfall in landslide analysis, different approaches have been presented to explain the relation between the two. One approach is the development of rainfall thresholds (Crosta, 2004). Thresholds can be defined on empirical (Jakob and Weatherly, 2002; Aleotti, 2004) or physical bases (Borga et al., 1998; Wilkinson et al., 1999; Collins and Znidarcic, 2004). Empirical thresholds are established by collecting data on meteorological events that initiate landslides and those that don’t initiate landslides. Physical thresholds are based on numerical models that incorporate the relationship between rainfall, pore pressure and slope stability (Aleotti, 2004).

Jakob and Weatherly (2002) propose an empirical model that incorporates antecedent rainfall and streamflow data to develop a landslide initiation threshold for the North Shore Mountains near Vancouver, BC. They collected data for storms that did and did not initiate landslides, and developed discriminant functions to provide a measure of landslide susceptibility. The landslide susceptibility is translated into a three-step Landslide Warning Threshold that allows warnings to be issued when landslides are expected (Jakob and Weatherly, 2002).

The climatic conditions that lead to the initiation of large rockslides and rock avalanches involve more than simple antecedent soil moisture and rainfall intensity, as is the case for debris flows. In the past five years there have been a number of large rock avalanches in British Columbia that may be a result of climate change (Schwab et al., 2003) (Figure 1-1). Climatic warming has resulted in thinning of mountain glaciers, resulting in debuttressing of slopes, degradation of mountain permafrost; and an increase in precipitation (Schwab et al., 2003). While a rock avalanche may occur after a single climatic event, such as a rain storm or a freeze thaw cycle, that event may have only been the final trigger of the failure and not the cause, as a rock slope failure is the result of progressive long-term degradation of the rock mass.
1.5 Summary

The ZRRA, according to terminology of Varnes (1978) and Hungr and Evans (2004b), is rockslide – debris avalanche. In this thesis, the precedent of Schwab et al. (2003) is adopted and the term rock avalanche is used. This recognizes not only the emphasis of this thesis on the debris avalanche, but also the significant control on risk due to the extensive runout.

Rheology was discussed in this chapter as an introduction to dynamic modelling summarized in Chapter 5. In the sediment classification matrix (Figure 1-3), the ZRRA would fall into the fluidized granular flow area, and thus the sturzstrom or debris avalanche category in Figure 1-4.

The ZRRA involved an initial rockslide that transformed into a debris avalanche. Very large boulders were transported by the ZRRA, indicating some grain support mechanism, possibly buoyancy associated with a high silt content. Deposition occurred when the flow exited the confined channel of Glen Falls Creek and entered the unconfined Zymoetz River channel. The long runout of the landslide may be associated with the mechanism discussed by Hungr and Evans (2004b), entrainment of debris that is liquefied by rapid undrained loading. This mechanism will be further investigated in Chapter 5.

2.1 Introduction

This chapter presents a characterization of the Zymoetz River rock avalanche (ZRRA), including a summary of analytical techniques employed, a debris avalanche path description, preliminary dynamic appraisal of the event and, finally, a brief comparison with other large landslides. The characterization of the event will follow the format used by Couture et al. (1999), and Evans et al. (1989) who divide the flow path into the detachment, transport and deposition zones (Figure 2-1).

The term rock avalanche is used to describe this complex landslide, although it has characteristics of a rockslide and debris flow / avalanche at different stages during the failure (Cruden and Varnes, 1996). The term rock avalanche has been retained based on the prior usage of Schwab et al. (2003), and Geertsema et al. (2003) and the naming of similar events such as the Pandemonium Creek rock avalanche (Evans et al., 1989) and the Mount Cayley rock avalanche (Evans et al., 2001). It is a shortened version of Varnes' (1978) rockslide – debris avalanche.
British Columbia, adapted by permission from Table 2-1. (Aerial Photograph © Province of BC.) Figure 2.1. Aerial photograph and longitudinal profile of the ZRRA. The long profile is divided into the detachment, transport, and deposition zones. The detached mass is outlined by the dotted line in the detachment zone. The bands used for velocity calculations are marked on the aerial photograph (see Figure 2-8). The detachment zone is outlined by the dotted line.
2.2 Event description

At approximately 1:30 am on June 8, 2002 the Pacific Northern Gas (PNG) pipeline in the Zymoetz River Valley was severed over an estimated length of several tens of meters at the mouth of Glen Falls Creek by a large debris avalanche (Schwab et al., 2003; Boultbee et al., in press) (Figure 2-2). The landslide initiated when a $1 \times 10^6$ m$^3$ block of bedrock detached from the east wall of the cirque basin at the head of the valley. The rockslide travelled over snow in the cirque basin for approximately 650 m and then dropped out of the basin and proceeded down Glen Falls Creek. Appendix 1 documents the ZRRA event in interactive poster format.

The ZRRA travelled a vertical distance of 1245 m, from el. 1390 m to el. 145 m, over a horizontal distance of 3.5 km, and a path length of 4.2 km (Figure 2-1). The ZRRA path is complex; the event initiated as a rockslide into the cirque basin, then transformed into a debris avalanche as it melted snow and incorporated water, trees and other debris. The landslide behaved fluidly as it progressed down the valley, with superelevated curves and mud splashes observed on trees up to 60 m above the base of the creek channel (Figure 2-2). The ZRRA then turned an almost $90^\circ$ corner and traveled through a constricted bedrock canyon, and finally over a 10-m-high waterfall into the main Zymoetz River valley (Figure 2-4). Erismann and Abele (2001) describe similar path shapes of landslides in Europe and North America.

Readings from the Water Survey of Canada (WSC) station (Station No. 08EF005), located 3 km downstream on the Zymoetz River, indicate a large turbidity spike and a sharp decrease in water level associated with the event (Figure 2-3). The gauge shows a smaller second event that occurred at approximately 9:30 am on June 8. This event was witnessed by PNG personnel who were investigating damage to the pipeline. The WSC station also collects water samples with an ISCO water sampler when a change in
the turbidity of the water is measured (Jim Schwab, pers. comm. 2004). The turbidity spikes were correlated to a suspended sediment load of approximately 3600 to 12000 mg/L.

Investigations into possible triggers for the landslide revealed that there was no seismic activity near the time of failure. The two weeks prior to the landslide included significant precipitation and colder temperatures than normal for that time of year. The days before the event were cool with snow at higher elevations (Schwab et al., 2003) (Figure 2-5). Failure was probably a result of progressive, long-term degradation of the tectonically deformed and altered volcaniclastic rock mass. The influence of freeze-thaw activity and changes in pore water pressures cannot be discounted in the gradual reduction of the shear strength along the eventual failure surface.

Figure 2-2. Glen Falls Creek channel above the canyon. Note the superelevations in the channel. The headscarp can be seen at the top centre of the photograph (arrow).
Figure 2-3. Zymoetz River flow and turbidity data from Water Survey of Canada Station No. 08EF005 on June 8, 2002. Two separate events can clearly be seen in the drops in water level and large increases in turbidity. (from Schwab et al., 2003, by permission).

Figure 2-4. The large fan deposited in the Zymoetz River by the ZRRA, and the constricted bedrock canyon that the debris flowed through. Note the dark patch of trees around the mouth of the canyon. These trees were burnt by the fire that started as a result of the gas pipeline burst.
Figure 2-5. Graph of temperature and precipitation in the days before and after the ZRRA. The days before the event had colder temperatures. Significant precipitation fell during the two week period prior to the event. (adapted by permission from Schwab et al., 2003).

2.2.1 GIS analysis

A variety of data collection and analysis methods were used to characterize the ZRRA, including a Geographic Information System (GIS) and the block size distribution program, WIPFRAG, which is discussed in Chapter 4 (WipWare Inc., 2003). On June 12, 2002, four days after the event, high-resolution air photographs were flown of the valley. Using both post-failure and 1975 air photographs, contour and image files were created that are suitable for use in a GIS. Fieldwork was completed in the summer of 2003, during which time photographs were taken specifically for use in the program WIPFRAG. Discontinuity surveys were also conducted for stereographic analysis (Chapter 3).

The GIS was created using ArcGIS (ESRI, 2002). The data were displayed in 2D, and distance and volume measurements made. A digital elevation model
(DEM) was created using the contour files. The data were displayed in 3D and rotated for different views in Arcscene. A digital image was draped over the DEM, providing a detailed 3D view of the area. With high-quality data, and a 3D model that could be rotated, repeated observations at different view points and elevations were possible. The 3D viewing capabilities improved observation of joint sets on inaccessible parts of the failure plane, and allowed for estimates of the volume of the failed block to be made.

2.2.2 The rockslide initiation zone

The rockslide initiated when a large mass of volcaniclastic bedrock detached from the northeast wall of the cirque basin at an approximate elevation of 1390 m. The headscarp is 125 m long at the top of the rupture surface, with a maximum width of 300 m. The failure involved a main sliding surface with a rear release plane (Figure 2-6). The main sliding surface is a very persistent (~200 m long), slightly curving surface dipping 43° - 45°, towards 300°, while the backscarp is sub-vertical and appears highly sheared in places (Figure 2-6). Immediately next to the fresh failure there is an old failure scar that can be seen in older airphotos.

Access to the headscarp was by helicopter. The failure plane itself was both dangerous and difficult to access, as the fresh surface still exhibited ravelling of loose material. A large block and considerable failure debris remain perched on the failure surface. An estimate of the volume of the failed mass of approximately $1 \times 10^6$ m$^3$ was obtained using the Geographic Information System code ArcGIS. The main failure plane can be traced under the backscarp to a small depression on the surface above the headscarp. An examination of the ZRRA in November 2003 by helicopter showed evidence of seepage onto the failure plane, from under the backscarp. Further discussion of the rock mass in the area of the headscarp is presented in Chapter 3.
Two views of the headscarp of the ZRRA. A is the main sliding surface. B is the backscarp (rear release surface). C is debris remaining on the slope. D are large blocks of intact bedrock that remain attached to the slope. E is the old failure surface. The dotted sections indicate vegetated areas that were not involved in the failure.
2.2.3 The cirque basin

After detachment, the failed mass impacted the cirque basin at an elevation of approximately 900 m, and travelled across a 3 m thick deposit of snow (Figure 2-7). The cirque basin varies in elevation from el. 940 m – el. 830 m over a distance of 600 m. Examination of air photographs shows streak marks on the snow on the west side of the basin, where the debris ‘skated’ over the snow and accelerated down valley (Figure 2-1). From the top of the basin the debris travelled over a moraine and then subdivided and funnelled down both sides of a forested island. The forested island itself is formed by a resistant volcanic bedrock knob that forms a knick point in the channel. The elevation drop out of the basin is approximately 150 m over a distance of 250 m (Figure 2-1).

A large volume of material was deposited in the cirque basin. Measurements from the GIS provide a volume estimate of 500,000 m$^3$, which is approximately half the volume of the failed mass. This estimate was obtained by measuring the basin from the far south end of the debris to the top of the forested island, and the width from the west to east wall. A debris depth of 3 m was assumed based on both the average deposit depth visible on top of the snow, and the size of some of the very large boulders found in the basin.

The failure debris deposited in the basin consists entirely of fragmented bedrock from the detached mass. The blocks are angular, with the main block size being boulders. A number of very large blocks measure up to 14 m in length. Frequently the blocks exhibit calcite, epidote, and possible zeolite on discontinuity surfaces, or contain veins of these minerals. The larger blocks typically exhibit joint bounded surfaces with coatings of calcite and epidote up to 0.5 cm thick.
Estimates of the snow pack in the basin prior to the slide are unavailable, but as approximately 3 m of snow remained under the debris after the event, there was probably little entrainment of snow and no entrainment of debris at the far south end of the basin. A contribution of the snow at the south end of the basin to the debris acceleration would have been a reduction in the surface frictional resistance. As the event moved further north in the basin, and encountered the moraine, saturated sediment would have been entrained, in addition to some snow. The debris towards the north end of the basin is considerably thicker, and estimates of any remaining snow under that portion of the basin could not be made. Snow pillow data from the closest station (Tsai Creek, 1360 m el. 54° 39' 127° 40') indicated a snow depth of approximately 2 m, therefore it is likely that the snow depth in Glen Falls creek was not significantly greater than the 3m that remained after the event.

Samples of debris collected immediately below the forested island had plasticity index values that plot below the A-line on a plasticity chart, and D_{50} values of ~15 mm, indicating entrainment of fines did occur. (Soil engineering characteristics are further discussed in Chapter 3.)
2.2.4 The main valley transport stage

Upon exiting the cirque basin at el. 690 m, the debris flowed on sides of the forested island shown in Figure 2-1. The majority of the rockslide debris emerged on the north side of the forested island and continued down valley to the Zymoetz River at el. 160 m. The approximate travel distance between the cirque basin and the Zymoetz River was 2.6 km. The extreme mobility of the debris is seen in superelevations along the length of the channel, some up to 40 m above the valley floor (Figure 2-8). The width of the main transport zone is 45 m – 90 m, compared to a pre-ZRRA channel width of approximately 10 m, obtained from the 1975 air photographs. This indicates that the landslide removed a significant volume of timber as it progressed down the valley.
It is unclear how much material was entrained in the main channel during the event, as very little evidence of the main component of the valley fill, a reddish till, was found in the debris. Minimal amounts of debris may have been entrained, due to the cohesive nature of the valley fill. While some till rip-up clasts were discovered in the deposit, the deposit was predominantly composed of sand to cobble sized volcanic bedrock, with few boulders, a moderate amount of wood fragments, and moderate fines content. The material that was entrained consisted of the surficial soil cover however, with the colluvial cover that now exists it is very difficult to estimate the exact amount of erosion that occurred. Similar difficulty in estimating entrainment is noted by Hungr and Evans (2004b).

Deposition in the valley was evident as distinct lobes/zones of debris. The lobes were defined by a length and width and an estimated average depth (Figure 2-9). The deposits in the channel ranged in depth from a thin veneer to 4 m deep, with an average of approximately 1.5 m. Using the GIS the estimate of material deposited in the channel is approximately 200,000 m$^3$, agreeing closely with an estimate of 190,000 m$^3$ made from field measurements of debris lobes.

The debris deposited in the channel is generally more compact and coherent than that found in the basin, possibly due to a higher concentration of finer-than-boulder material. In addition the channel deposits have been washed by Glen Falls Creek as the creek attempted to re-establish itself after the complete destruction of the channel. The main grain size of the channel deposits is cobble, though there are a few large boulder size blocks (up to 6 m long) throughout the channel.
Figure 2-8. Cross-sections in Glen Falls Creek valley, showing the superelevations of the debris. Locations of cross-sections are shown in Figure 2-1.
2.2.5 The landslide depositional zone

Deposition of debris occurred along the entire length of Glen Falls Creek, from the basin to the channel, with the majority of debris deposited in the form of a large fan in the Zymoetz River (Figure 2-4). Prior to this event, photographic evidence shows that there was no fan in the Zymoetz River at the mouth of Glen Falls Creek, however this section of the river has always been a sediment storage area (Figure 2-10). The newly formed fan spans the river, with a width and length of approximately 250 m. The fan initially blocked the river, as evidenced by the severe drop in water level measured by the WSC station (Figure 2-3), but was quickly overtopped. Water upstream of the fan has remained ponded however, and the forest service road upstream had to be raised up to 3 m. The depth of the fan is difficult to estimate, but may be up to 10-12 m in places. Large boulders, up to 10 m wide, were carried the full length of the channel to the fan. Estimates of the dimensions from the GIS provide a volume of material deposited in the fan as approximately 625,000 m³, assuming an average depth of 10 m.
Examination of the deposits on the fan shows that the debris is composed of fragmented volcanic bedrock of sand, gravel and cobble size. The dominant grain size in the fan is cobbles, and the majority of grains are subangular to subrounded. There are some limestone grains (<10%), as well as a small amount of granite (<2%). While the majority of material in the fan is cobble sized and smaller, there are a few very large boulders that were transported the entire length of the event. The debris blocks observed in the main river are up to 10 m long, similar in size to many of the large boulders found in the basin.

From estimates of the material deposited during the event, the total deposit has a volume of approximately $1.3 \times 10^6 \text{ m}^3$, while the calculated volume of the displaced mass is estimated at $1.0 \times 10^6 \text{ m}^3$. From Hungr and Evans (2004b), the entrainment ratio (Eqn. 1-3) is the ratio of the volume of material entrained to the volume of material fragmented in the failure. From this equation it is seen that the entrainment ratio for the ZRRA is 0.3. This entrainment ratio qualifies the ZRRA as a rockslide-debris avalanche (rock avalanche), as the value is greater than 0.25, indicating significant entrainment occurred during the event, possibly contributing to the long runout (Hungr and Evans, 2004b).
Figure 2-10. Series of air photographs showing that the mouth of Glen Falls Creek has always been a sediment storage area in the Zymoetz River, but no fan existed prior to the ZRRA.

Air Photograph Numbers:
- 1949 – BC 1024:40
- 1975 – BC 7728:006
- 1991 – TMC036-8325
- 2002 – Zymoetz orthophoto

© Province of British Columbia, adapted by permission.
2.3 Landslide velocity estimates

Velocity calculations for the ZRRA were undertaken using both run-up and superelevation data and the velocity potential energy equation (Evans et al, 1989; Jordan, 1994)

(Eqn. 2-1) \[ v^2 = 2gh \]

where \( h \) is the maximum vertical run-up and \( g \) is acceleration due to gravity (Table 2-1). This formula is used where debris impacts a surface that is almost perpendicular to the flow direction. For this event, equation (2-1) can be used in the basin, and in the channel at the canyon, where the debris turns two almost right angle corners (Figure 2-1). It is recognised that there may be errors associated in maximum velocity determinations associated with the derivation of the maximum vertical run-up, \( h \), (Erismann and Abele, 2002). The maximum velocity is calculated using the same technique as Evans et al. (1989) in order to allow comparison with the Pandemonium Creek event. This yielded velocities of 20 m/s to 34 m/s (Table 2-1).

Velocity was also estimated using superelevation data in equations (2-2) and (2-3).

(Eqn. 2-2) \[ v^2 = r_c g \tan \theta \cos \alpha \]

where \( r_c \) is the radius of curvature, \( \theta \) is the transverse slope and \( \alpha \) is the longitudinal slope (Hungr et al. 1984).

(Eqn. 2-3) \[ v^2 = \frac{ghr_c}{b} \]

where \( b \) is the channel width. Cross sections were drawn to obtain the superelevation data (Figure 2-10), and the radius of curvature was measured for curves from the GIS and printed maps. The velocities obtained from
equation (2) ranged from 14 m/s to 26 m/s (Table 2-1), while equation (3) yielded velocities of 19-26 m/s. Velocity calculations were also undertaken for the run-up in the basin using equation (2), which yielded a velocity of 26 m/s, while equation (1) gave 34 m/s. Measurement error associated with equation (2) exists as determining the radius of a curve from an air photograph is subject to limitations. Although the velocity of the landslide would have varied considerably within the transport zone (owing to the varying lateral confinement and funnelling of the flow), a range from 15 m/s to 25 m/s is probable.

Table 2-1. Run-up and superelevation data used in the velocity calculations at different locations in the channel, shown in Figure 2-1.

<table>
<thead>
<tr>
<th>Equation</th>
<th>Location</th>
<th>h (m)</th>
<th>g (m/s)</th>
<th>rc (m)</th>
<th>β'</th>
<th>tan θ</th>
<th>α'</th>
<th>cos α</th>
<th>b (m)</th>
<th>V (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1) Run Up</td>
<td>Basin 1</td>
<td>60</td>
<td>9.8</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>34</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>Basin 1</td>
<td>60</td>
<td>9.8</td>
<td>204</td>
<td>19</td>
<td>0.34</td>
<td>8</td>
<td>0.99</td>
<td>-</td>
<td>26</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>C-S 2</td>
<td>25</td>
<td>9.8</td>
<td>156</td>
<td>8</td>
<td>0.14</td>
<td>5</td>
<td>0.99</td>
<td>-</td>
<td>14</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>C-S 2</td>
<td>25</td>
<td>9.8</td>
<td>156</td>
<td>8</td>
<td>-</td>
<td>-</td>
<td>104</td>
<td>-</td>
<td>19</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>C-S 3</td>
<td>30</td>
<td>9.8</td>
<td>155</td>
<td>10</td>
<td>0.18</td>
<td>8</td>
<td>0.99</td>
<td>-</td>
<td>16</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>C-S 3</td>
<td>30</td>
<td>9.8</td>
<td>155</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>122</td>
<td>-</td>
<td>19</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>C-S 7</td>
<td>46</td>
<td>9.8</td>
<td>150</td>
<td>18</td>
<td>0.32</td>
<td>15</td>
<td>0.96</td>
<td>-</td>
<td>21</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>C-S 7</td>
<td>46</td>
<td>9.8</td>
<td>150</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>100</td>
<td>-</td>
<td>26</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>C-S 8</td>
<td>38</td>
<td>9.8</td>
<td>125</td>
<td>12</td>
<td>0.21</td>
<td>15</td>
<td>0.96</td>
<td>-</td>
<td>15</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>C-S 8</td>
<td>38</td>
<td>9.8</td>
<td>125</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>80</td>
<td>-</td>
<td>24</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>C-S 9</td>
<td>44</td>
<td>9.8</td>
<td>130</td>
<td>18</td>
<td>0.32</td>
<td>12</td>
<td>0.98</td>
<td>-</td>
<td>20</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>C-S 9</td>
<td>44</td>
<td>9.8</td>
<td>130</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>96</td>
<td>-</td>
<td>24</td>
</tr>
<tr>
<td>(1) Run Up</td>
<td>Corner 1</td>
<td>22</td>
<td>9.8</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>20</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>Corner 1</td>
<td>22</td>
<td>9.8</td>
<td>156</td>
<td>10</td>
<td>0.18</td>
<td>12</td>
<td>0.98</td>
<td>-</td>
<td>16</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>Corner 1</td>
<td>22</td>
<td>9.8</td>
<td>156</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>100</td>
<td>-</td>
<td>18</td>
</tr>
<tr>
<td>(1) Run Up</td>
<td>Corner 2</td>
<td>36</td>
<td>9.8</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>26</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>Corner 2</td>
<td>36</td>
<td>9.8</td>
<td>44</td>
<td>45</td>
<td>1</td>
<td>9</td>
<td>0.99</td>
<td>-</td>
<td>20</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>Corner 2</td>
<td>36</td>
<td>9.8</td>
<td>44</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>50</td>
<td>-</td>
<td>17</td>
</tr>
<tr>
<td>(2) Superelev.</td>
<td>Corner 3</td>
<td>40</td>
<td>9.8</td>
<td>52</td>
<td>33.7</td>
<td>0.67</td>
<td>9</td>
<td>0.99</td>
<td>-</td>
<td>18</td>
</tr>
<tr>
<td>(3) Superelev.</td>
<td>Corner 3</td>
<td>40</td>
<td>9.8</td>
<td>52</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>80</td>
<td>-</td>
<td>16</td>
</tr>
</tbody>
</table>
2.4 Case studies of long-runout rock avalanches

Long-runout events are relatively common in British Columbia and elsewhere around the world. Some have achieved notoriety, as they resulted in significant damage and/or loss of life (eg. Huascaran, Mount St Helens). There is a growing body of literature examining events that have occurred in remote areas, as they may provide further insight into why the events exhibit their high mobility and eventually lead to improved runout prediction methods.

The case studies presented here are:

- the Pandemonium Creek event, which exhibited extreme runout associated with flow over a glacier;
- the Mount Cayley event that ran-out long distances because of the rock type;
- the Nomash Creek landslide that entrained significant amounts of saturated debris.

These case studies were selected as they have been the subject of published dynamic analysis using the program DAN-W (Hungr, 1995) and thus provide useful comparisons for rheologic modelling of the ZRRA in Chapter 5. This chapter provides a preliminary comparison of these three events with the ZRRA emphasizing geology, morphology and mobility.

2.4.1 Pandemonium Creek rock avalanche

The Pandemonium Creek rock avalanche (Evans et al., 1989), occurred in 1959 in a remote area of Tweedsmuir Provincial Park, in central-western B.C. The Pandemonium Creek drainage is composed of layered gneiss, gneissic quartz diorite and diorite, and metavolcanics (Evans et al., 1989). The avalanche had a volume of approximately $5 \times 10^6$ m$^3$, and exhibited spectacular mobility.
Observations from air photographs pre- and post-dating the event suggest that the landslide initiated when a rock spur detached from a cirque headwall. The mass disintegrated as it fell and then spread out as it flowed over a glacier below. The debris was then constricted by a pair of lateral moraines below the glacier, which resulted in rapid acceleration of the landslide. The debris impacted the opposite valley wall and ran up approximately 335 m, one of the highest run-ups ever recorded. The material then proceeded down Pandemonium Creek, with superelevation in each corner of 35 to 70 m. When the debris ran-out onto a fan at the mouth of the creek, much of the debris was deposited, however some continued to run up 30 m on the east valley wall and entered Knot Lakes. The total travel distance was 9 km, with an elevation drop of 2000 m and an estimated velocity of up to 100 m/s. Evans et al. (1989) suggest that the high mobility was a consequence of the steep confined channel, flow over the glacier, followed by the funnelling effects of the lateral moraines below the glacier.

It was noted by Evans et al. (1989) that, had the fan of Pandemonium Creek been inhabited, there would have been a total loss of life, and that the geological and topographical conditions at Pandemonium Creek are common in the Coast Mountains.

2.4.2 Mount Cayley rock avalanche

Lu and Cruden (1996) present a description and analysis of the rockslide and debris flows from Mount Cayley in 1963 and 1984. A fresh debris flow was observed in 1964 air photos, and on June 28, 1984, a debris flow destroyed a bridge and deposited millions of tonnes of sediment in the Squamish River. The 1963 debris flow occurred in Dusty Creek and started as a rockslide involving $5 \times 10^6$ m$^3$ of columnar-jointed dacite (with minor breccia, volcanic tuffs and lapilli tuffs), transforming into a dry debris flow as the material fragmented. Velocity calculations from superelevation evidence suggest a
velocity of 26 m/s. The flow travelled down Dusty Creek, flowing into and subsequently damming Turbid Creek. Observations that Turbid Creek incised a narrow channel in the debris indicate that the dam was overtopped and eroded. Part of the dam remains as a deposit today, indicating that there was no catastrophic release of water. The debris left in the deposits is dense and coarse, relative to the 1984 event, giving the dam an increased resistance to erosion. The 1963 deposits preserved the original stratigraphy of the rocks, as seen in bedrock outcrops, indicating little mixing within the material, and laminar flow during this event.

The 1984 event was significantly different from the 1963 event. The 1984 failure occurred mainly in soft tuffs, in Avalanche Creek, with an initial volume of \(3.2 \times 10^6\) m\(^3\). The debris eroded snow and ice that was in the creek, entraining a further 850,000 m\(^3\) of debris and ice, then proceeded to the confluence with Turbid Creek where a debris dam consisting of 85% rock and 15% ice formed. The velocity was determined from run-up calculations to be 35 m/s. The higher velocity of the 1984 event is assumed to be associated with the snow cover in Avalanche Creek. Deposition in the Turbid Creek valley again formed a dam which lasted for 1-2 days, before bursting. The deposit was very loose, therefore lacking resistance to erosion. The ice within the dam could also have weakened it. The deposits from the 1984 event did not preserve original bedding, therefore flow was assumed to be turbulent.

Evans et al. (2001) provide a reinterpretation of the 1984 Mount Cayley rock avalanche. They calculated an initial failure volume of 740,000 m\(^3\), with further entrainment of 200,000 m\(^3\) of surficial deposits, a volume that is approximately 75% lower than the published estimates of Cruden and Lu (1992). The rock avalanche travelled 3.46 km horizontally over a vertical drop of 1.18 km. The distal debris flow travelled a further 2.6 km. Velocity estimates by Evans et al. (2001) are 42-70 m/s significantly higher than the
estimates of Cruden and Lu (1992). Evans et al. (2001) also determined through field investigation and dynamic modelling that there was no landslide dam associated with the event and suggest it occurred as one uninterrupted movement. The 1984 event was an order of magnitude smaller than the 1963 event, however it travelled considerably farther. The longer runout is thought to be associated with the porous nature of the failed pyroclastic rock mass and thus greater saturation, in comparison to the dacitic rock involved in the 1963 failure (Evans et al., 2001).

2.4.3 Nomash River slide

The Nomash River slide is described by Favero (2000) and Hungr and Evans (2004b). The Nomash River is located near the town of Zeballos on the west coast of Vancouver Island. The bedrock in the area consists of limestone and basalt of the Karmutsen Group. On April 25 or 26, 1999, the slide initiated on a 430 m high wall of a U-shaped valley. The slide moved across the valley, turned a right-angle corner and continued down valley for 1.2 km travelling at speeds of up to 22 m/s. The volume of the detached block (ignoring bulking) was approximately 300,000 m³. In its descent the rockslide overrode colluvial deposits on the valley floor, scouring a channel 100-150 m wide and 8 m deep. Field estimates indicate that 360,000 m³ of surficial materials were entrained by the flow. With a total flow distance of 2270 m and a vertical drop of 560 m, the fahrböschung is 13.8°.

Hungr and Evans (2004b) explain the sequence of events that occurred during the Nomash River Slide. A moderate sized rockslide of this nature would normally come to rest at the base of the slope, and not runout long distances; however in this event saturated materials were present at the base of the slope. When the rockslide impacted these materials, they experienced rapid undrained loading and liquefied, allowing for the entrainment of the sediments and a doubling in volume.
2.4.4 Comparison with ZRRA

A review of selected large rock avalanches in the literature illustrates two general types, [1] those that travel long distances in confined mountain valleys, exhibiting high fluidity, and [2] those that run out into relatively open valleys, are not confined, and travel shorter distances. The Zymoetz River event is of type [1], along with Pandemonium Creek (Evans et al., 1989), Mount Cayley (Evans et al., 2001), Huascaran (Plafker and Ericksen, 1978), and Little Tahoma Peak (Fahnestock, 1978), among others. These type [1] events are comparable to the A type situation discussed by Nicoletti and Sorriso-Valvo (1991), involving a channellized mass. Type [2] includes events like the Hope (Mathews and McTaggart, 1978) and Frank slides (Cruden and Hungr, 1986), and is equivalent to the types B (unobstructed spreading, or moderate energy dissipative) and C (right-angle impact against an opposite slope, or high-energy-dissipative) defined by Nicoletti and Sorriso-Valvo (1991).

The estimated volume, velocity and fahrböschung ($f$) angle of the ZRRA are compared with other morphologically similar British Columbia rock avalanches in Table 2-2. The fahrböschung angle is the angle between the headscarp and the extreme limit of the deposit of a landslide and has been used as a measure of mobility by numerous authors, including Evans et al., 1989 (Figure 2-12). In order to allow a direct comparison, estimates of the fahrböschung angle were measured in the same manner as in Evans et al. (1989): from the highest point of the detachment to the lowest point of the deposit, for the vertical elevation difference, and along the path length, for the horizontal difference measurement, thus accounting for major bends along the path. From this comparison it can be seen that the ZRRA exhibits high mobility, similar to the Mount Cayley and Pandemonium Creek rock avalanches and other BC landslides in Figure 2-11.
While the Nomash event involved a smaller failure volume and shorter runout, it is interesting to compare its flow path to that of PCRA and ZRRA (Figure 2-12). These BC landslides have very similar flow paths, all showing failure onto open slopes, impact against the opposite valley wall, and subsequent right-angle turns and confinement into creek channels. While the PCRA and ZRRA ran out to main river channels (for PCRA the South Atarko River, and for the ZRRA the Zymoetz River), the Nomash event does not travel as far as the junction with the Zeballos River; rather, it remains confined to the Nomash Valley. The shape of these channels seems to be a combination of Nicoletti and Sorriso-Valvo (1991) types A (channelling of the debris mass) and C (right-angle impact against opposite slope). They found that the type A shape seemed to have the longest runout, while type C had the shortest. Further work to investigate the frequency of occurrence of this particular channel shape, and to evaluate the run-out distances relative to the other channel shapes presented in the literature may be of interest.

![Figure 2-11. Graph showing the relationship between fahrböschung and landslide volume. Adapted by permission from Evans et al., 1989.](image-url)
Figure 2-12. Comparison of the similarities in flow path between (A) Pandemonium Creek rock avalanche (Evans et al., 1989), (B) Zymoetz River rock avalanche, and (C) Nomash River landslide (Favero, 2000).

Table 2-2. Mobility comparison between the ZRRA and other similar events. Adapted by permission from Evans et al., 1989.

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Est.Volume (x10^6 m^3)</th>
<th>H(m)</th>
<th>L(m)</th>
<th>F</th>
<th>Velocity (m/s)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pandemonium</td>
<td>1959</td>
<td>5 - 6</td>
<td>2000</td>
<td>8600</td>
<td>0.23</td>
<td>81-100 (peak)</td>
<td>Evans et al. (1989)</td>
</tr>
<tr>
<td>Zymoetz River</td>
<td>2002</td>
<td>1</td>
<td>1200</td>
<td>4200</td>
<td>0.28</td>
<td>30 (mean)</td>
<td>This study</td>
</tr>
<tr>
<td>Mount Cayley</td>
<td>1984</td>
<td>0.94</td>
<td>1180</td>
<td>3460</td>
<td>0.34</td>
<td>42-70 (peak)</td>
<td>Evans et al. (2001)</td>
</tr>
</tbody>
</table>


2.5 Summary

This chapter describes the Zymoetz River rock avalanche, a large natural landslide that resulted in significant direct and indirect damage. The direct damage associated with this event include the severed PNG pipeline, the Forest Service road, and lost timber, with an estimated value of $5.9 million. The indirect costs increase significantly to an estimated $27.5 million, when the lost time and revenue from the logging road and pipeline are taken into account (Schwab et al., 2003).

Initiation of the event is interpreted to be associated with progressive degradation of a damaged and altered rock mass over thousands of years. Although the event initiated as a rockslide, it exhibited high fluidity, evident from superelevations in bends and mud splashes on trees. The mobility is attributed to significant entrainment of snow and saturated debris in the basin. This entrainment, according to Hungr and Evans (2004b), characterizes this event a rockslide-debris avalanche.

Investigation of the event was aided by the use of a GIS that allowed for closer examination of the difficult-to-access headscarp, in addition to calculation of failure and deposition volumes. The volume of the failed mass was approximately $1.0 \times 10^6$ m$^3$. The total calculated deposit volume is $1.3 \times 10^6$ m$^3$, indicating entrainment of 300,000 m$^3$, and giving an entrainment ratio of 0.3 (Hungr and Evans (2004b). Velocity estimates for the ZRRA are 14-34 m/s, a similar range to the Nomash and Pandemonium slides. The time of year that the event occurred is significant, as there was still snow at higher elevations, including the cirque basin. If there had been no snow in the basin the rockslide may not have travelled far beyond the base of the slope, and may not have runout over 4 km.
Six large landslides occurred in British Columbia between 1999 and 2003, with five occurring in the spring and summer of 2003. Several authors have suggested that these events may be associated with retreat of glaciers in alpine areas, leading to de-buttressing of slopes and, eventually failure (Evans et al., 1989; Schwab et al., 2003). It is clear that there is considerable potential for remote alpine slope failures to develop into hazardous long-runout events.

The ZRRA exhibits similarities to other large landslides that have occurred relatively recently in BC, including the Pandemonium Creek rock avalanche, the Mount Cayley rock avalanche and the Nomash River slide. Considering the number of long runout landslides that has occurred in the last five years, engineers and geoscientists need be aware of this issue when developing an area. Detailed investigations should be undertaken of remote upstream areas, even as far as 9 km from the proposed development.
CHAPTER 3 – Engineering geological soil and rock classification

3.1 Introduction

This chapter presents an engineering characterization of the materials involved in the ZRRA including soil classification and rock mass description. The soil analysis includes both debris from the ZRRA and surficial deposits in the valley. The rock mass characterization includes a description of the local geology, intact rock strength testing, discontinuity mapping and kinematic analysis.

3.2 Soil classification methodology

Soil samples were collected for grain size analysis in order to determine the coarse and fine components, with the fine portion (that smaller than 0.075 mm) being used for modified Atterberg limit determinations. Hydrometer analysis on the silt and clay portion was not undertaken in this study. Determining the grain size characteristics of the debris at different locations may provide an indication of material entrained by the landslide as it travelled down the channel. It may in addition allow interpretation of different pulses involved in the event. Characterization of the deposits of the previous valley fill, including old debris flow deposits and till, provide for a more detailed interpretation of these deposits. Soil classification for engineering purposes was undertaken using the Unified Classification System (Das, 1997), and engineering names were assigned to the samples collected from Glen Falls Creek.
3.2.1 Sample collection

Samples were collected at various locations along Glen Falls Creek and on the fan of the ZRRA (Figure 3-1). Within Glen Falls Creek, sediment was collected from recent flow deposit, as well as material that is interpreted as old debris flow deposits. Samples were also collected of material interpreted as till at several locations in both the channel and basin.

Figure 3-1. Diagram showing the locations of the samples collected in the Glen Falls Creek valley (Aerial photograph © Province of British Columbia, adapted by permission).

3.2.2 Particle size analysis

A detailed description of the methodology used in sieve testing is presented in Appendix 2. In this analysis several minor departures from the engineering standards were made. According to ASTM standards (Das, 1997) for particle size analysis, only material passing the #4 sieve should be used. It was decided that all material should be accounted for, even that larger than the #4 sieve, in order to have samples that were as representative as possible. Wet sieving was also done on a portion of each of the 18 samples. After the wet sieving was complete, organics were removed from the sample by floating.
The organics that were found in the soils included sticks, pine needles, and wood slivers entrained from the forest floor or from trees broken up during the flow. When the data were plotted to derive a particle size distribution, graphs were made for each of the four individual samples, in addition to a single plot for a combination of the four samples for each site.

The fine material passing the #200 sieve was retained after sieving so that modified Atterberg limits could be determined. The procedure followed for the modified Atterberg limit testing is provided in Appendix 2.

The effective grain size is used routinely in the engineering characterization of soils, and is defined as the maximum particle size of the smallest 10% of a sample, $D_{10}$ (10% passes, 90% retained). The effective size, $D_{10}$ is used in the calculation of the coefficients of uniformity $C_U$ and curvature, $C_c$ where:

\[
C_U = \frac{D_{60}}{D_{10}} \quad \text{and} \quad C_c = \frac{D_{30}^2}{D_{60}D_{10}}
\]

The parameters $C_U$ and $C_c$ are used in the classification of soils as they indicate the slope/shape characteristics of the grading curve. The higher the coefficient of uniformity ($C_U$) the larger the range of particle sizes, or the more well graded the soil is. A well graded soil will have a coefficient of curvature of between 1 and 3 (Bell, 2000).

After the samples were sieved the results were graphed as grain size distribution curves. From these curves, the values of $D_{10}$, $D_{30}$, and $D_{60}$ could be determined. The D-values were subsequently used to calculate the coefficient of uniformity and curvature for each sample. All soil classification results and graphs can be found in Appendix 2. The coefficients of uniformity and curvature were also used to name the soils according to the Unified Classification System (Das, 1997) (Table 3-1).
Of the 18 samples collected, modified Atterberg limits could only be determined for 11 samples. Sample 1 and 3 did not have enough sediment passing the #200 sieve to be tested (Table 3-1), while samples 7, 11, 12, 13, and 17 were too sandy to test (a consistent paste could not be formed to test the liquid limit). With the exception of sample 7 (a sandy overflow material from the fan), all of the non-tested samples were identified in the field as till.

### 3.3 Soil testing results

#### 3.3.1 Particle size analysis

Particle size analysis predominantly produced grain size curves indicating well-graded soils, according to a calculated Cu > 5 (Figure 3-2). There were also some gap-graded (poorly graded) curves (Figure 3-2). As well-graded corresponds to a poorly sorted soil, this is to be expected, as poor sorting is a characteristic feature of diamicts such as colluvial and till deposits (Clague, 1984). According to the UCS however, some of the debris avalanche soils were named poorly graded, as the system names are based on the Cu and Cc (Table 3-1). This may be a limitation of the UCS in using the Cc and Cu as well-graded soils could be named poorly graded if they have curves that are gap-graded. Visual inspection of the soils clearly indicates that they are poorly sorted soils, and thus should be named well-graded.

When the curves for the soil samples are compared it is evident that the majority of the ZRRA samples (75%) are grouped relatively close together, with a $D_{60}$ ranging from 5-17 mm. There are 2 outliers in the deposits, one with a $D_{60}$ of 1.4 mm and the other 36 mm. When the samples collected in the channel are compared with those collected on the fan (except sample 7, as it is not believed to be part of the ZRRA deposit), it is apparent that the grain sizes in the channel are generally larger than those on the fan (Table 3-2).
The old debris avalanche deposits are similar to the majority of the ZRRA deposits, with $D_{60}$ values of 11-17 mm. The till deposits have similar $D_{60}$ values with a range from 1.4 – 21 mm. Table 3-3 summarizes the grain size percentage information and shows that the ZRRA deposits have a large range in each of the grain size categories, much larger than the other deposit types. The deposits in the channel of the ZRRA contain 24%-74% sand, compared to a much narrower range of 36%-42% sand in the old debris deposits. The tills were found to contain 32%-63.5% sand (Table 3-2).
Table 3-1. Table showing the percentages of grain size classes, Cu, and Cc for each soil. These values were used to name the soils according to the Unified Classification System.

<table>
<thead>
<tr>
<th>#</th>
<th>Sample Name</th>
<th>%Fines (&lt;0.75mm)</th>
<th>%Sand (0.75-2mm)</th>
<th>%Gravel (2-38.1mm)</th>
<th>Cu</th>
<th>Cc</th>
<th>Soil Classification</th>
<th>Group Symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Matrix normal fan</td>
<td>0.8</td>
<td>51.5</td>
<td>47.7</td>
<td>17.9</td>
<td>1.0</td>
<td>well graded sand with gravel</td>
<td>SW</td>
</tr>
<tr>
<td>2</td>
<td>Muddy matrix fan</td>
<td>4.7</td>
<td>53.0</td>
<td>42.3</td>
<td>32.4</td>
<td>0.8</td>
<td>poorly graded sand with gravel</td>
<td>SP</td>
</tr>
<tr>
<td>3</td>
<td>Below mud cap at tip of berm</td>
<td>0.9</td>
<td>23.9</td>
<td>75.2</td>
<td>67.9</td>
<td>3.0</td>
<td>well graded gravel</td>
<td>GW</td>
</tr>
<tr>
<td>4</td>
<td>Debris west side of island</td>
<td>5.0</td>
<td>34.4</td>
<td>60.6</td>
<td>85.0</td>
<td>1.0</td>
<td>well graded gravel with silt and sand</td>
<td>GW-GM</td>
</tr>
<tr>
<td>5</td>
<td>Debris east Side of island</td>
<td>4.3</td>
<td>37.2</td>
<td>58.5</td>
<td>65.2</td>
<td>1.0</td>
<td>well graded gravel with sand</td>
<td>GW</td>
</tr>
<tr>
<td>6</td>
<td>Mud cap from tip of berm</td>
<td>4.0</td>
<td>37.1</td>
<td>58.9</td>
<td>73.9</td>
<td>0.7</td>
<td>poorly graded gravel with sand</td>
<td>GP</td>
</tr>
<tr>
<td>7</td>
<td>Sandy overflow material on fan</td>
<td>44.7</td>
<td>53.7</td>
<td>1.6</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>Debris above berm</td>
<td>7.8</td>
<td>74.2</td>
<td>18.0</td>
<td>14.7</td>
<td>0.9</td>
<td>well graded sand with clay and gravel</td>
<td>SW-SC</td>
</tr>
<tr>
<td>9</td>
<td>Between island and bedrock in channel</td>
<td>5.6</td>
<td>54.0</td>
<td>40.4</td>
<td>34.3</td>
<td>1.0</td>
<td>well graded sand with clay and gravel</td>
<td>SW-SC</td>
</tr>
<tr>
<td>10</td>
<td>Old debris flow on berm</td>
<td>1.1</td>
<td>41.9</td>
<td>57.0</td>
<td>33.3</td>
<td>0.8</td>
<td>poorly graded gravel with sand</td>
<td>GP</td>
</tr>
<tr>
<td>11</td>
<td>Till west side bridge</td>
<td>6.8</td>
<td>51.9</td>
<td>41.3</td>
<td>44.2</td>
<td>0.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>Moraine in basin</td>
<td>7.6</td>
<td>61.2</td>
<td>31.2</td>
<td>24.7</td>
<td>0.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>Curve north of bridge</td>
<td>4.6</td>
<td>44.9</td>
<td>50.5</td>
<td>50.0</td>
<td>1.0</td>
<td>well graded gravel with sand</td>
<td>GW</td>
</tr>
<tr>
<td>14</td>
<td>Till sample cut-block rd</td>
<td>2.6</td>
<td>47.5</td>
<td>49.9</td>
<td>25.0</td>
<td>1.1</td>
<td>well graded gravel with sand</td>
<td>GW</td>
</tr>
<tr>
<td>15</td>
<td>Calcareous deposit - 3 trees top</td>
<td>2.4</td>
<td>35.7</td>
<td>61.9</td>
<td>58.6</td>
<td>1.3</td>
<td>well graded gravel with sand</td>
<td>GW</td>
</tr>
<tr>
<td>16</td>
<td>Calcareous deposit - 3 trees bottom</td>
<td>4.2</td>
<td>41.5</td>
<td>54.3</td>
<td>55.0</td>
<td>1.0</td>
<td>well graded gravel with sand</td>
<td>GW</td>
</tr>
<tr>
<td>17</td>
<td>Compacted soil basin wall</td>
<td>3.1</td>
<td>31.8</td>
<td>65.1</td>
<td>75.0</td>
<td>1.6</td>
<td>well graded gravel with sand</td>
<td>GW</td>
</tr>
<tr>
<td>18</td>
<td>Red till</td>
<td>7.0</td>
<td>63.5</td>
<td>29.5</td>
<td>18.0</td>
<td>0.8</td>
<td>poorly graded sand with clay and gravel</td>
<td>SP-SC</td>
</tr>
</tbody>
</table>
Figure 3-2. (a) Grain size distribution curves for the old and new debris flow deposits. Red lines: ZRRA deposit. Green: old debris flow deposits. The one anomalous line is sample 7 – the sandy overflow material collected from the fan.
Figure 3-2. (b) Grain size distribution curves for the till samples.
Table 3-2. D-values for the ZRRA deposits and the old debris flow deposits at different points along the flow path.

<table>
<thead>
<tr>
<th>Zone</th>
<th>D_{10}</th>
<th>D_{30}</th>
<th>D_{60}</th>
<th>Cu</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transport (Channel)</td>
<td>0.095-0.53</td>
<td>0.35-7.6</td>
<td>1.4-36</td>
<td>15-85</td>
</tr>
<tr>
<td>Deposition (Fan)</td>
<td>0.17-0.38</td>
<td>0.85-1.6</td>
<td>5.5-6.8</td>
<td>18-32</td>
</tr>
<tr>
<td>Old DF (Channel)</td>
<td>0.2-0.42</td>
<td>1.5-2.5</td>
<td>11-17</td>
<td>33-59</td>
</tr>
</tbody>
</table>

Table 3-3. Grain sizes in each of the three types of deposits.

<table>
<thead>
<tr>
<th></th>
<th>% Fines</th>
<th>% Sand</th>
<th>% Gravel</th>
</tr>
</thead>
<tbody>
<tr>
<td>ZRRA Dep (Channel)</td>
<td>0.9-7.8</td>
<td>23.9-74</td>
<td>18-75.2</td>
</tr>
<tr>
<td>ZRRA Dep (Fan)</td>
<td>0.8-4.7</td>
<td>51.5-53</td>
<td>42.3-47.7</td>
</tr>
<tr>
<td>Old DF Dep</td>
<td>1.1-4.2</td>
<td>35.7-42</td>
<td>54.3-62</td>
</tr>
<tr>
<td>Till</td>
<td>3.1-7.6</td>
<td>32-63.5</td>
<td>29.5-50.5</td>
</tr>
</tbody>
</table>

3.4 Interpretation of soil engineering data

The only sample that had an anomalous particle size distribution curve was sample 7, sandy overflow material from the ZRRA fan. This sample was probably deposited when the ponded river overtopped the temporary dam formed by the fan (Figure 3-2), and thus is not part of the ZRRA deposit.

There were two outliers in the grain size distribution curves, a relatively finer sample (sample 8, D_{60} = 1.4 mm) and a coarser sample (sample 3, D_{60} = 36 mm). These outliers are interpreted to be related to the location of the samples. The finer sample was located at the top of the berm in the channel (Figure 3-1), which is approximately 20m above the base of the channel. The coarser sample was located within the deposit below a mud capping on the channel floor. All other debris samples were located within the channel floor. Grain size analysis on the deposits from the 1984 Mount Cayley event showed D_{60} values of 4.1 mm, 55 mm, and 16 mm for subsurface, surface
deposits, and the rockslide deposits respectively (Cruden and Lu, 1992). Sample 8 may have been deposited by the finer mud splashing at the edge of the debris flow event. Sample 3 could be part of the coarser frontal part of the flow, capped by the finer tail of the debris flow, or even creek overflow as the creek re-established itself after the event. Sample 3 may be a similar deposit to the surface deposits observed by Cruden and Lu (1992).

Modified Atterberg tests on the fine fraction of the samples show that sample 4 had a very high modified liquid limit - 63%, and also had the highest organic content when the wood was floated out of the samples (50-80% greater than the other samples). This sample was located in the Glen Falls Creek channel on the west side of the island. It can be seen on the north-west corner of the basin, that a large section of forest was overrun (Figure 3-3), immediately above the area where the sample was collected. This high organic content may have partially contributed to the high liquid limit of the sample.

Sample 9 was located approximately half way down the channel, near a deposit of glaciolacustrine silts and clays. Entrainment of this material could account for the slightly higher clay content in the sample. The old debris flow deposits found in the valley all had a higher clay content than the ZRRA deposits, similar to sample 9. It is possible that the old events also entrained some of the glaciolacustrine deposit.

According to Clague (1984) there are two types of tills in the Terrace region, (1) tills with a matrix rich in clay, and (2) sandy tills. The clay rich tills generally occur in the main valleys of the Interior system of the Canadian Cordillera, or east of the Kitsumkalum-Kitimat trough (Figure 1-1), while the sandy tills occur in the western system, (west of the Kitsumkalum-Kitimat trough) and in the mountain valleys of the Interior system (Clague, 1984). The four till samples (11, 12, 13, and 17) for which modified Atterberg limits could not be determined (no paste could be formed) are probably the Type 2
sandy tills, that Clague (1984) notes are cohesionless. The modified Atterberg limits were tested on the material passing the #200 sieve, which has an opening of 0.075 mm. The lower limit for sand sized particles is 0.062 mm, which means that very fine sand can pass through the #200 sieve. In the case of the cohesionless tills, it is possible that they may contain significant rock flour, which would pass the #200 sieve and lead to a very low plasticity fines portion of the sample. The other two tills (samples 14 and 18) are Type 1, clayey tills that occur more frequently in the main valleys. Sample 18 could be a regional till, as it has a very distinctive red colour sourced from a red volcanic tuff in the Hazelton Group, rocks which were not found in the Glen Falls Creek drainage. Sample 14 was located along the forest service road through the cut-block on the east side of the valley. This sample could represent a more weathered version of the till in Sample 18, as it possessed a similar red colour, dried texture and clast appearance (See Appendix 2 for all sample pictures).

Figure 3-3. Section of forest overrun by the rockslide, on the northwest corner of the basin. This photograph was taken southwest of the island (See Figure 2-1).
3.5 Rock mass characterization

The rocks in the Glen Falls Creek area were examined in the field and laboratory in order to characterize the rock mass and intact rock strength properties. Samples were collected for thin section identification and point load testing. Discontinuity surveys were carried out in the vicinity of the headscarp for use in a kinematic analysis of the failure mode.

3.5.1 The geology of the ZRRA area

The ZRRA is located near the western edge of the Intermontane belt, within the Stikinia Terrane. The basement of Stikinia is composed of Devonian and Permian arc volcanics with platform carbonates. These rocks are overlain by Triassic and Lower Jurassic arc volcanics, volcaniclastics, chert and arc-derived clastics that are intruded by plutonic rocks (GSC, 1992; Mihalynuk, 1987).

The main rock Group in the Terrace area, and thus the Copper River Valley, is the Hazelton Group, with the largest portion being comprised of the Telkwa Formation which is Triassic in age and of non-marine origin. The Telkwa formation consists of reddish, maroon, purple, grey and green pyroclastic and flow rocks. Zeolite alteration is widespread in the formation with locally extensive veining and formation of zeolite cemented breccias. Other metamorphic minerals found in the formation include epidote, calcite and quartz. The alteration and consequent reduction in strength of the rock mass may have been a factor in the location of the initial rockslide.

Rock types observed in the Glen Falls Creek valley include green and purple volcaniclastics, and buff-coloured limestone (Figure 3-4). Field investigation showed that the initiation zone occurred in volcaniclastic rocks on the east side of the valley, and that the limestone only crops out on the west side. (See
Appendix 3 for rock descriptions) The geology in the valley is complex and highly variable over short distances, related to a complex depositional and alteration history. Within the valley, Glen Falls Creek has taken advantage of the weakness offered by the main contact between the volcanics and the limestone, and flows along this contact in the middle of the valley.

While the carbonate bedrock in the valley is intact (Geological Strength Index (GSI) 70-90), the volcanic bedrock varies from blocky (GSI 55-75) to disintegrated (GSI 25-35) Marinos and Hoek (2000), Hoek et al., (2002) (Figure 3-4) (See Appendix 3 for GSI Chart). The bedrock in the vicinity of the headscarp is also highly fractured. Several prominent joint sets are present including the joint sets that form the main failure surface, and the backscarp (rear-release) of the failure. (Figure 2-6). There are indications of tectonic deformation in the vicinity of the failure plane, including moderate shearing in the backscarp, slickensides and alteration. The rock mass in this area is sheared and disintegrated with an estimated GSI of 10-20 (Figure 3-5). The British Columbia Geological Survey (BCGS) geology map indicates numerous faults within the Zymoetz River valley, including both thrust and normal faults. The majority of faults are shown as unidentified (Figure 3-6) (BCGS, 2004). Though not indicated on the map, it is highly probable that the Zymoetz River follows major structure or faults, due to the distinctly linear course of the river (Figure 1-1). The BCGS map also shows a persistent fault running along the east side of Glen Falls Creek that appears to coincide with the strike of the main failure surface for the initial rockslide.
Figure 3-4. Photograph showing the two main rock types in the Glen Falls Creek valley, the green volcanic rock, in contact with the white limestone. The volcanic rock is highly fractured (GSI 25-35) while the limestone is more intact (GSI 70-90).

Figure 3-5. Photograph of the sheared rock mass in the backscarp.
Figure 3-6. BCGS hillshaded DEM showing the large fault in the Glen Falls Creek drainage (dashed box) passing through the area of the headscarp (H) of the ZRRA. Arrows outline the debris flow path. This fault is unidentified with respect to movement direction. DEM © British Columbia Geological Survey (2004), adapted by permission.

3.5.2 Point load strength testing

Rock samples were obtained from the backscarp of the failure surface and the slide deposit on the fan for thin section identification and strength testing. Estimates of compressive strength were made according to field geological hammer and point load tests on hand samples. The samples that were tested included rock with calcite veins, and rock without. Point load tests on 12 block samples of the purple meta-basaltic rock from the debris, conducted according to the ISRM suggested methods (1985), indicate an average point load index ($I_{50}$) of 6.7 MPa for samples that broke along a vein, and 15.7 MPa for samples that broke through the rock (Table 3-4). Most of the samples tested with the calcite veins, broke along the veins (Figure 3-7). It is recognized that breakage along a vein during a point load test invalidates the
test (Wyllie and Mah, 2004 p. 106), however in this analysis the relative strength values and the influence of the veins are what was sought. Rock samples from the backscarp were much less competent, and disintegrated during transport to the laboratory. From the thin sections that were prepared, micro-fractures can be seen within the backscarp rock at a spacing of 2-5 mm (See Appendix 3 for all rock mass data and thin section photos).

Figure 3-7. Point load test rock samples; (a) sample with a calcite vein and (b) without calcite vein.

Table 3-4. Point load index values (I_s50). Averages are calculated discarding the highest and lowest values in each group.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>I_s(50) MPa</th>
</tr>
</thead>
<tbody>
<tr>
<td>Broke along a vein</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>7.95</td>
</tr>
<tr>
<td>3</td>
<td>6.79</td>
</tr>
<tr>
<td>4</td>
<td>6.25</td>
</tr>
<tr>
<td>5</td>
<td>12.61</td>
</tr>
<tr>
<td>8</td>
<td>6.65</td>
</tr>
<tr>
<td>9</td>
<td>3.88</td>
</tr>
<tr>
<td>11</td>
<td>5.96</td>
</tr>
<tr>
<td>Avg</td>
<td>6.72</td>
</tr>
<tr>
<td>No vein</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>14.98</td>
</tr>
<tr>
<td>6</td>
<td>16.16</td>
</tr>
<tr>
<td>7</td>
<td>7.53</td>
</tr>
<tr>
<td>10</td>
<td>16.06</td>
</tr>
<tr>
<td>12</td>
<td>20.21</td>
</tr>
<tr>
<td>Avg</td>
<td>15.73</td>
</tr>
</tbody>
</table>
3.6 Discontinuity surveys

The BCGS geology map indicates a major, persistent fault of unknown relative displacement located along the east side of Glen Falls Creek (BCGS, 2004). The location of this fault closely coincides with the initiation zone (Figure 3-6). In addition to this linear north-northwest trending feature, westerly trending linears can also be identified in Figure 3-8 (F). These west-trending features appear to be of the same orientation as the creek that flows along the side of the headscarp (Figure 2-6 and Figure 3-8). Closer examination also shows northerly trending cliff faces below the failure surface that parallel the failure plane / fault (Figure 3-8G).

Discontinuity surveys were conducted in the area of the headscarp, both along the backscarp, and a small portion of the main sliding surface, and along outcrops around the headscarp to obtain data for subsequent kinematic analysis. Over 380 discontinuities were measured and the data collected included dip and dip-direction, persistence, spacing, and any infill or mineralization on the surfaces, including slickensides. Figure 3-8b(i) shows the strong concentration of discontinuities that formed the main failure plane. Surveys of other outcrops above the headscarp indicate joint sets with a general north-northeast dip-direction, which could have acted as lateral release surfaces for the failed block (Figure 3-8(ii)). The discontinuities measured had variable persistence (0.3 m-4 m) and spacing (<0.1 m-1 m), and commonly exhibited calcite or epidote mineralization on surfaces (Figure 3-9). There were also a number of slickensided surfaces throughout the headscarp area (~10% of surfaces measured) (Figure 3-10).

Immediately to the south of the headscarp is a westerly flowing creek. Discontinuity surveys in the creek show that the creek follows a contact between different volcaniclastic rocks, and probably is located along a fault. Very different structural domains are represented on either side of the creek.
lending some support to the presence of a fault (Figure. 3-8(iii) + (iv) and 3-11). Approximately halfway down the creek there is another contact that could be the continuation of the fault that coincides with the failure plane (Figure 3-12). It was noted that there was evidence of shearing in this area, including numerous slickensided surfaces.

The intact block size is difficult to interpret in the vicinity of the headscarp, as there is such variability in the joint spacing, from <0.1 m to > 1 m (Appendix 3). Large boulders (up to 14 m in length) were deposited in the cirque basin. These blocks contained joint sets with spacing between the joints of 5-30 cm (Figure 3-13). The joint spacing and rock mass quality in the sheared rock outcrops in the vicinity of the failure surface/fault does not reflect the size of the failed rock blocks in the cirque basin. The block size in the failure material remaining on the slope appears to reflect more closely the large boulders involved in the rockslide (Figure 2-6). Most of the faces of the blocks in the basin were noted to have infill on the joint surfaces. This appears to agree with the importance of rock breakage along veins as shown in the laboratory point load testing data. It is probable that the faulting has an important influence on the rock mass quality with increasing GSI values with distance away from the fault surfaces.
Figure 3-8a. Three-dimensional plot derived from the 2m contour lines of the ZRRA headscarp area and basin. D is the mass that failed during the event. E is the old failure surface (see also Figure 2-6). F shows the westerly trending linears that parallel the joints forming the lateral release for the headscarp. G shows the northerly trending cliff faces that parallel the main failure surface.

Figure 3-8b. Three-dimensional plot of the post-event 2m contours. Also seen on this plot are contoured stereonets, clockwise from i-iv. i) shows the high concentration of surfaces making up the main failure plane and backscarp; ii) shows the joint orientation in outcrops above the failure surface, which show the possibility of toppling failure; iii) shows the joint orientations in the outcrop on the north side of the creek (closest to the failure surface), the orientations are similar to those in the other outcrops above the headscarp, and could be associated with a lateral release surface for the failure; iv) the joint orientations in the outcrop on the south side of the creek, also potentially providing a lateral release surface.
Figure 3-9. Calcite on discontinuity surfaces.

Figure 3-10. Discontinuity surface with epidote crystallization and slickenlines (location is above the old failure plane).
Figure 3-11. Photograph looking east up the creek that runs alongside the headscarp. Different structural domains can be seen on either side of the creek in the discontinuity sets.

Figure 3-12. Photograph of the southern creek wall that runs alongside the headscarp. The separation between the two rock types can be seen, this is the possible extension of the fault that coincides with the main sliding surface.
3.7 Kinematic and limit equilibrium analysis

A preliminary kinematic analysis was completed on selected areas around the headscarp (Figure 3-8) using the stereographic analysis program DIPS (Rocscience, 2004). The pre-slide slope topography was obtained from the GIS, and used as the slope face. The slope and aspect surfaces created from the DEM gave a range of values: local slopes calculated ranged from 35° to 78°, with a northwest aspect that varied from 292° - 337°. A sensitivity approach to kinematic analysis was adopted using DIPS and seventeen of the sampled aspect and slope combinations (Table 3-5). Local slope face orientations were plotted along with daylight envelopes to determine the spatial variations of potential planar failure instability (Lisle, 2004). Translational failure is considered the primary failure mechanism, due to lack of detail on the subsurface 3D failure geometry and the presence of a major planar translational surface (fault) (Figure 2-6).
Based on an analysis of rock strength using RocLab (a program that allows for the calculation of rock strength properties based on user inputs) a friction angle of 30° was estimated (Hoek et al., 2002; Rocscience, 2004). RocLab inputs include an estimate of Uniaxial Compressive Strength for tuff of 100-250 MPa, and a GSI of 50 for very blocky rock (Rocscience, 2004). The 30° friction angle is a dry friction angle. When water pressures are considered, the mobilized effective friction angle may be significantly reduced.

In the DIPS analysis, the kinematic feasibility of failure was examined for planar, toppling, and wedge failure modes, using all 384 joint measurements taken at the headscarp (see Appendix 3 for all joint data). The pre-slide slope orientation used for the planar analysis was 71°/328°. This orientation accounted for almost all of the joint surfaces that were measured along the main planar sliding surface (the fault) and was derived from the results of the GIS based kinematic sensitivity analysis (Figure 3-14). This slope orientation was also assumed for the toppling (Figure 3-15) and wedge failure (Figure 3-16) analyses. The friction angles assumed were 20° (accounting for water pressures) and 30° (dry). The wedge analysis involved definition of five joint sets in DIPS based on contouring of the poles (Table 3-6). The mean dip and dip direction for each joint set was used to examine if potentially unstable, daylighting wedges were present within the rock slope. (Table 3-6; Figure 3-16).

The stereographic analysis indicated that approximately 10% of the discontinuities were oriented such that failure could occur by toppling (Figure 3-15). Stereographic analysis with a friction angle of 30° showed 4 potential wedge intersections between sets 1*,2,3, and 4 (joint set 1* is sub-parallel to the fault). With a friction angle of 20°, 7 potential wedge intersections are present involving all five joint sets (Figure 3-16) (Table 3-7). It is recognized that these joint set intersections will produce wedges
of greatly varying shape and risk, dependent upon discontinuity orientation and geometry (ie. persistence and spacing), as well, no information regarding factor of safety is provided. Accordingly, these wedge intersections were further examined with the Rocscience program SWEDGE (RocScience, 2004). The SWEDGE analysis was used primarily to illustrate the geometry of the wedges and also to determine a factor of safety (Table 3-8). For the SWEDGE analysis a 20° friction angle was assumed. Cohesion was assumed at 0, and the persistence of the joint sets 2-6 was limited to reflect the average size of the largest blocks observed in the failure debris (10 m). The persistence of joint set *1 was not limited, as this was the main planar sliding surface (a fault) that had an observed persistence in excess of 250 m on the slope. The wedge volumes and intersection trace length reported in Table 3-8 are for (1) the maximum possible values, calculated without limiting any of the joint sets, and (2) the user limited values. These values are reported so that the relative volumes can be compared for the wedges and to illustrate the need to incorporate joint geometry observations. The calculated factor of safety ranged from 0 – 1.5, and the wedge shape varied from obtuse wedges to unrealistic slivers (Table 3-8).

The planar failure situation was also investigated further, using the program RocPlane (RocScience, 2004). RocPlane is similar to SWEDGE, simply differing in the failure mode investigated. In RocPlane the slope dimensions and failure plane are input, with the option of including a tension crack. A sensitivity analysis approach was adopted varying the slope, failure plane and friction angles in addition to the cohesion and depth of water in an assumed tension crack. Table 3-9 and Figure 3-17 show the sensitivity analysis input and the influence of the percentage change in parameters on the calculated factor of safety for planar sliding.
Figure 3-14. The pre-slide slope orientation of 71°/328°, with a friction angle of 30° that encompassed the majority of joint surfaces measured on the main planar sliding surface (*fault).

Figure 3-15. Kinematic analysis for toppling instability using DIPS.
Figure 3-16. Kinematic analysis for potential wedge failure; (a) assuming a 20' friction angle, (b) assuming a 30' friction angle. Joint set 1 is the fault.
Table 3-5. Slope and corresponding aspect values obtained from the DEM using GIS, for use as the original slope orientation in the DIPS analysis.

<table>
<thead>
<tr>
<th>Slope (Dip)</th>
<th>Aspect (Dip Direction)</th>
</tr>
</thead>
<tbody>
<tr>
<td>40</td>
<td>340</td>
</tr>
<tr>
<td>39</td>
<td>328</td>
</tr>
<tr>
<td>36</td>
<td>337</td>
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<tr>
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</tr>
<tr>
<td>64</td>
<td>349</td>
</tr>
<tr>
<td>61</td>
<td>348</td>
</tr>
<tr>
<td>51</td>
<td>326</td>
</tr>
<tr>
<td>50</td>
<td>266</td>
</tr>
<tr>
<td>53</td>
<td>342</td>
</tr>
<tr>
<td>53</td>
<td>314</td>
</tr>
<tr>
<td>68</td>
<td>340</td>
</tr>
<tr>
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<td>342</td>
</tr>
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<td>306</td>
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<tr>
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<td>345</td>
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<tr>
<td>48</td>
<td>294</td>
</tr>
<tr>
<td>71</td>
<td>328</td>
</tr>
<tr>
<td>59</td>
<td>306</td>
</tr>
</tbody>
</table>

Table 3-6. Orientations of the joint sets defined in DIPS and used in the SWEDGE analysis.

<table>
<thead>
<tr>
<th>Joint Set</th>
<th>Dip</th>
<th>Dip Direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>1*</td>
<td>48</td>
<td>309</td>
</tr>
<tr>
<td>2</td>
<td>33</td>
<td>273</td>
</tr>
<tr>
<td>3</td>
<td>51</td>
<td>195</td>
</tr>
<tr>
<td>4</td>
<td>86</td>
<td>100</td>
</tr>
<tr>
<td>5</td>
<td>25</td>
<td>2</td>
</tr>
</tbody>
</table>

Table 3-7. Table illustrating the intersections between joint sets that form wedges the x's mark wedge intersections. With the 30° friction angle there are four potential wedges, whereas with the 20° friction angle there are seven wedges.

<table>
<thead>
<tr>
<th>20°</th>
<th>1*</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>30°</th>
<th>1*</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td>1*</td>
<td>-</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td>1*</td>
<td>-</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>x</td>
<td>-</td>
<td>x</td>
<td>-</td>
<td>x</td>
<td>2</td>
<td>x</td>
<td>-</td>
<td>x</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>3</td>
<td>x</td>
<td>x</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>3</td>
<td>x</td>
<td>x</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
<td>x</td>
<td>-</td>
<td>-</td>
<td>x</td>
<td>-</td>
<td>4</td>
<td>x</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>5</td>
<td>x</td>
<td>x</td>
<td>-</td>
<td>-</td>
<td>x</td>
<td>5</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
Table 3-8. Wedge geometry and factor of safety calculations using the program SWEDGE for the potential wedges indicated in the DIPS kinematic analysis.

<table>
<thead>
<tr>
<th>Joint Sets</th>
<th>Factor of Safety</th>
<th>Intersection Trace Length</th>
<th>Plunge of Intersection</th>
<th>Wedge Volume</th>
<th>Wedge Shape</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\phi=20^\circ$</td>
<td>$\phi=30^\circ$</td>
<td>Max</td>
<td>Limited</td>
<td>Max</td>
</tr>
<tr>
<td>1*-2</td>
<td>0</td>
<td>0</td>
<td>195 m</td>
<td>10 m</td>
<td>31*</td>
</tr>
<tr>
<td>1*-3</td>
<td>0.33</td>
<td>0.52</td>
<td>190 m</td>
<td>9.7 m</td>
<td>33*</td>
</tr>
<tr>
<td>1*-4</td>
<td>1.5</td>
<td>2.38</td>
<td>295 m</td>
<td>15.1 m</td>
<td>27*</td>
</tr>
<tr>
<td>1*-5</td>
<td>0.78</td>
<td>1.24</td>
<td>329 m</td>
<td>16.9 m</td>
<td>24*</td>
</tr>
<tr>
<td>2*-3</td>
<td>0.56</td>
<td>0.89</td>
<td>196 m</td>
<td>10 m</td>
<td>32*</td>
</tr>
<tr>
<td>2*-5</td>
<td>1.0</td>
<td>1.6</td>
<td>586 m</td>
<td>30 m</td>
<td>21*</td>
</tr>
<tr>
<td>5*-4</td>
<td>0.78</td>
<td>1.24</td>
<td>333 m</td>
<td>9.55 m</td>
<td>25*</td>
</tr>
</tbody>
</table>
Table 3-9. RocPlane input values. The highlighted values were assumed as the approximate slope conditions. The factor of safety is given for each parameter as they changed in the sensitivity analysis.

<table>
<thead>
<tr>
<th>Slope Angle</th>
<th>Failure Plane Angle</th>
<th>Friction Angle</th>
<th>Cohesion</th>
<th>Water</th>
</tr>
</thead>
<tbody>
<tr>
<td>Angle</td>
<td>FS</td>
<td>Angle</td>
<td>FS</td>
<td>MPa</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>FS</td>
<td></td>
</tr>
<tr>
<td>0</td>
<td>1.67</td>
<td>10</td>
<td>1.66</td>
<td></td>
</tr>
<tr>
<td>0</td>
<td></td>
<td>20</td>
<td>1.62</td>
<td></td>
</tr>
<tr>
<td>0</td>
<td></td>
<td>30</td>
<td>1.55</td>
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<td>55</td>
<td>2.37</td>
<td>30</td>
<td>0.82</td>
<td>1.75</td>
</tr>
<tr>
<td>60</td>
<td>1.72</td>
<td>35</td>
<td>1.14</td>
<td>2.00</td>
</tr>
<tr>
<td>65</td>
<td>1.32</td>
<td>40</td>
<td>1.05</td>
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<td>70</td>
<td>1.00</td>
<td>45</td>
<td>1.00</td>
<td>2.50</td>
</tr>
<tr>
<td>75</td>
<td>1.00</td>
<td>50</td>
<td>1.03</td>
<td>2.75</td>
</tr>
<tr>
<td>55</td>
<td>1.21</td>
<td>55</td>
<td>1.18</td>
<td>3.00</td>
</tr>
<tr>
<td>60</td>
<td>1.72</td>
<td>60</td>
<td>1.08</td>
<td>2.75</td>
</tr>
<tr>
<td>70</td>
<td>1.00</td>
<td>70</td>
<td>1.06</td>
<td>1.06</td>
</tr>
<tr>
<td>55</td>
<td>1.21</td>
<td>90</td>
<td>0.90</td>
<td></td>
</tr>
<tr>
<td>60</td>
<td>1.72</td>
<td>100</td>
<td>0.75</td>
<td></td>
</tr>
</tbody>
</table>

Figure 3-17. RocPlane sensitivity analysis. The planar failure was most sensitive to the slope angle, as evidenced by the steepest line. The slope was also relatively sensitive to the depth of water in the tension crack after a threshold value was reached.
3.8 Rock mass data interpretation

The volcaniclastic rock in the Glen Falls Creek valley contains calcite veins that affect the strength of the rock mass. The point load tests showed that blocks with veins broke along those veins, and had an $I_{50}$ that was less than half that of blocks that did not contain veins (Table 3-3). Hoek and Brown (1997) report $I_{50}$ values of 4-10 MPa for tuff, and >10 MPa for rocks such as basalt, chert, granite or quartzite. For the blocks that didn't break on veins, the $I_{50}$ values are in the range of the extremely strong rocks (Avg = 15.73 MPa) (Hoek and Brown, 1997). These values are very high for a volcaniclastic rock, however according to Hoek and Brown (1997) when testing very hard brittle rocks, laboratory tests tend to overestimate the in situ rock strength. It is also recognized that the rocks tested were potentially the strongest rocks at the site, as they were collected on the fan, and survived over 4 km of transport.

In the cirque basin, numerous blocks were found that had broken along veins (Figure 3-18) in addition to blocks that contained intact joints (Figure 3-13). This indicates that while the joints represent a weakness in the rock mass, perhaps their lack of persistence prevents significant fragmentation. Breakage would appear to have to occur through intact rock, which was shown to have a much higher strength than that of the veined rock. The significantly lower strength due to calcite veins enhances breakage along these surfaces, as opposed to through the solid rock.

A kinematic analysis of the failure headscarp indicates that all three failure modes, planar, toppling and wedge are kinematically possible in the slope. The main failure mode is considered to be planar failure, as an extensive planar failure surface is exposed in the headscarp. Four of the potential wedges that formed were intersections with joint set 1*, which is the main
planar sliding surface (the fault). The intersections between joint sets 1*/2, 1*/3, 1*/4 and 1*/5 reflect potential lateral and rear release surfaces. The potential wedge formed by joint sets 2 and 5 forms a major wedge, with an intersection that falls within the daylight envelope of the slope. The factor of safety of this wedge is 1.6 (Table 3-8). This major wedge was not readily apparent, although smaller wedges were present.

The RocPlane analysis indicated that the presence of water was important in the initiation of the slide. With the other slope parameters remaining constant, a 76% filled tension crack resulted in a factor of safety less than 1.0. The slope was also sensitive to the angle of the slope face, with the factor of safety decreasing sharply with changes in the slope angle (Figure 3-17).

This rock slope comprised a highly fractured rock mass that appeared to have failed rapidly along joint set 1*. Factors that contributed to the initial failure include:

- The orientation of the joint sets in the rock mass, including those that facilitate planar, potential toppling and wedge failures (Figures 3-14, 3-15, 3-16). It appears however that planar sliding was the dominant mechanism with lateral and rear release provided by 4 joint sets.
- Alteration associated with shearing, as evidenced by the closely jointed rock mass, slickensides and epidote found on joint surfaces (Figures 3-5, 3-10).
- Alteration associated with the intrusion of veins, mainly calcite providing a lower rock mass strength and increased potential for fragmentation (Figure 3-9).
- The presence of a major fault, potentially coinciding with the main planar failure surface (Figure 3-6, 2-6).
Progressive degradation of the rock mass by varied weathering/stress induced processes.

Figure 3-18. Joint-bounded blocks in the slide debris with white calcite on surfaces.

3.9 Summary

This chapter has included an engineering geological characterization of the deposits and rock mass involved in the ZRRA. This characterization has involved sample collection and lab testing, as well as the collection of field data, in the form of deposit observation and joint surveys.

The particle size analysis demonstrated that the deposits of the ZRRA are consistent with what that reported by previous authors, including Cruden and Lu (1992) for the grain size of deposits. The particle size analysis also indicated that significant entrainment of debris occurred during the
landslide, both in the basin, and throughout the channel, as evidenced by the presence of fines in the debris.

The characterization of the headscarp illustrated that the heavily altered rock mass was kinematically predisposed to failure, through the presence of a major fault and the orientation of the joint sets. Both wedge and translational failure were probable on this slope. The limit equilibrium analysis with the program RocPlane indicated that the presence of high water pressures could have contributed to the initiation of the slide.
CHAPTER 4 – Comminution analysis using WipFrag

4.1 Introduction

A variety of data collection and analysis methods were used to characterize the ZRRA. The techniques employed included the block size distribution program, WIPFRAG (WipWare Inc., 2003). WipFrag is an image analysis software tool that allows estimation of preliminary particle size distribution curves using a photograph with a user-defined scale. The program can determine the size of particles in the photograph, and hence output a block/grain size distribution curve. This program has been utilised extensively in blasting to determine the size of fragments that remain after a blast, and thus control blast efficiency. During fieldwork, photographs were taken with a 1 m square for scale, to be used in WipFrag.

4.2 Previous studies using photoanalysis techniques

Photoanalysis has been used by other authors, for rock avalanche analysis, as well as discontinuity studies. Tsoutrelis et al. (1990) determined that photoanalysis provided a rapid method for the geometric characterization of discontinuities in a rock mass. They describe a discontinuity study at a mine in Greece, where photographs were taken of the rock mass, and then digitized. They were able to demonstrate how important discontinuity parameters such as orientation, trace length, spacing and the joint roughness coefficient (JRC) can be statistically analyzed and presented in appropriate diagrams. The main advantage of the photoanalysis technique was the large number of measurements of any parameter that could be taken, resulting in more accurate estimates of the parameters (Tsoutrelis et al., 1990).
In an examination of the Frank slide, Cruden and Hungr (1986) used a photoanalysis technique to determine the percent of large particles in the debris at four different depths. At a cut excavated through slide debris for replacement of the CPR line, photographs were taken of the scarp at the base, middle and top of the debris, with one horizontal photograph of the surface of the debris, using a 4 m² reference frame. Samples of the material finer than 5 cm were also collected for sieve analysis, to complement the photographic analysis of the coarse fraction. The photoanalysis was undertaken by placing a clear sheet with a grid superimposed over the photographs. The photographs were then examined with a magnifying glass to count the number of large particles in each grid square, and make visual estimates of the percentage of matrix. The photograph and sieving curves were then combined to provide a complete distribution curve. Examination of the particle size distribution provided the authors with further insight into the rockslide dynamics, and suggested that any theories of rockslide motion dealing with gas pressures are inappropriate. The reverse grading observed was considered to be associated with dispersive forces during shearing and motion-induced vibration (Cruden and Hungr, 1986).

Couture et al., 1999 also discuss photographic sampling of debris, with a reference frame for determining the grain size distribution of the deposit. In their method blocks are traced onto translucent paper, and then measured manually. The computer analysis technique that they describe involves hanging the tracing on a wall, taking a photograph, and then digitizing. Couture et al. (1999) state that comparing the block size in the detachment zone with that in the deposition zone aids in an evaluation of the fragmentation process during the rock avalanche.

Hadjigeorgiou et al. (1996) use a photoanalysis technique to compare the grain size in the transition and deposition zones with the block size in the
detachment zone. They used the photoanalysis program Stereoblock to characterize the block size distribution in the detachment zone of the Charmonetier rockslide in the French Alps. The technique used for the debris characterization was similar to that described by Cruden and Hungr (1986), but with a 1 m² reference frame. They determine a Coefficient of Uniformity, which refers to the shape of the grain size distribution curve, and show that there is a clear reduction in volumetric size from the detachment area to the transition zone and the deposition zone (Hadjigeorgiou et al., 1996). Their work provides an improved understanding of the energy dissipation and comminution of materials during flow; however, they emphasize the difficulty in developing a methodology to quantify the impact of the energy components during comminution (Hadjigeorgiou et al., 1996).

Locat et al., (2003) attempt to quantify the energy transfer during a rock avalanche, to assess the effects of fragmentation on mobility (as discussed in Chapter 1). As part of their methodology, photographic analysis of both the rock mass and the debris is employed. They used Stereoblock to determine the initial block size in the failed mass from photographs of joint sets. An image analysis system was then used to determine the grain size distribution of the debris. Based on the differences in the $D_{50}$ values for the initiation and deposition zones, the fragmentation energy was determined.

Split engineering has developed a fragmentation analysis system, Split-Online, that provides fragmentation and texture classification analysis, with outputs of material size, shape, surface area and colour (SplitEng, 2005). Split-Online has been used in the mining, aggregate and fertilizer industries. Split Engineering has also developed software, SplitFX, which allows for the complete characterization of a rock mass, including fracture orientation and distributions of fracture size, shape and roughness (SplitEng, 2005).

Sirovision is another state of the art photoanalysis program that allows the user to collect data by taking stereo pairs of a rock or terrain surface and
resolving the pair into an accurate three-dimensional digital image (Surpac, 2005). The main benefit these programs offer is that of safety, they allow collection of data, from otherwise inaccessible locations. Another important benefit is the savings in time required to obtain a large quantity of measurements.

4.3 WipFrag theory

4.3.1 Sources of error

Maerz and Zhou (1998) describe the numerous advantages to using optical systems for grain size analysis, as opposed to sieving. These advantages include:

- complete automation, eliminating the expense of human hours;
- more measurements can be made, increasing statistical reliability, and reducing sampling error;
- results are available quickly, allowing for adjustments to production methods;
- and screening is too prohibitive in the case of very large blocks.

There are also inherent limitations to optical methods, which reflect the accuracy, precision and reproducibility of the results (Maerz and Zhou, 1998). These limitations stem from the fact that there are numerous variables to consider, which affect the outcome of measurements. The main categories of errors related to optical systems are:

- those associated with the method of analysis,
- the sample presentation,
- the imaging process, and
- errors related to the sample processes (Maerz and Zhou, 1998).
Errors related to the method of analysis include block misidentification due to colour and texture characteristics, and errors associated with using the wrong unfolding model (the process of transforming two-dimensional measurements into three dimensions) (Maerz and Zhou, 1998). Unfolding models are used in optical systems to convert the 2D photograph image into a representative 3D image. As block delineation is generally based on delineating the shadows between blocks, mixed colour assemblages and textured surfaces can be problematic. Wet fragments can also prove difficult, as water highlights the colour differences, and can result in false edges.

Errors associated with sample presentation include those related to fragment lay, or the difference between measurements determined by optical imaging, and sieving (Maerz and Zhou, 1998). The assumption that blocks will lay flat on the ground, states that the image will measure the major and intermediate axis, while sieving will measure the minor and intermediate axis. Therefore it can be expected the imaging processes will overestimate the sizes when compared to sieving, with the degree of overestimation dependent on the shape of the particles. An additional sample presentation error involves overlapping fragments. Overlapping fragments were found to result in an error of ~28%; however, with the use of an unfolding function the error was reduced to only ~6% (Maerz and Zhou, 1998). Errors related to the imaging process, include those associated with the size of the sampling window, camera variability, lighting variability, and perspective.

4.4 Procedure adopted at the ZRRA

4.4.1 Location and methodology for debris photographs

Photographs were taken with a digital camera during the 2003 field season of different sections of the deposits within Glen Falls Creek, and on the fan in the Zymoetz River. The camera used to take the photographs was a Kodak
2.1 mega pixel DC290 zoom digital camera and was set to take photographs at the highest resolution.

On the fan, the photographs were taken along a grid comprising seven lines, with photograph locations marked every 20 m (Figure 4-1). The photographs for the channel were taken along a separate grid, defined at locations of cross-sections measured across the channel (Figure 2-11). The cross-sections were defined wherever there was a superelevation and a high point in the channel. They were measured using a 30 m tape, a range finder and a clinometer, and extended from edge of debris to edge of debris, thus in some areas, including very steep banks. The photographs were taken at 10 m intervals.

Photographs with scale could not be taken in the basin, as the grain size was too large, with blocks up to 14 m long. During field work, measurements of some of the largest blocks that stood out from the deposit were taken. Potentially, the measurements of these blocks could be used to define scale in the aerial photograph, for use in WipFrag. However due to the small scale of the aerial photograph, when you zoom in on the photograph close enough to see the larger blocks, it is hard to delineate any blocks smaller than approximately 5 m. From the aerial photograph it is possible to define zones of larger blocks (Figure 4-2).

When the photographs were taken, a 1m x 1m square made of PVC pipe was placed over the center of the location flag (Figure 4-3). This square was used as scale for the photograph, which is required by the WipFrag program. The added benefit of using a square scale is that it allows for tilt in the photograph to be accounted for, as it was not possible to take the pictures from directly above, they were taken at slight angles. The 1 m scale was selected, as the fan deposits contained very few large boulders, and with the 1 m scale an adequate range of particle sizes could be identified in the photographs, from
granules to small boulders. Smaller scale photographs could have been taken with a higher resolution digital camera allowing the smaller grains to be clearly identified. Field notes were taken for each station, including the general appearance, the clast roundness and average sizes, as well as estimates of volcanic, carbonate, and other clast percentages (See Appendix 4 for descriptions).

Figure 4-1. Aerial photograph of the fan showing the grid of photograph locations for the WipFrag analysis. The grid lines are numbered, 1-7, and the photograph locations are marked every 20 m by the white dots. The photographs are numbered according to line and photograph location on the line, starting at the apex of the fan, i.e. the last picture on line 4 is 4-12. Aerial photograph © Province of British Columbia, adapted by permission.
Figure 4-2. Aerial photograph of the basin, with some of the largest blocks outlined (6-11m). Areas of higher block concentration can be identified along the sides of the flow, with few large blocks in the northern portion of the basin. Aerial photograph © Province of British Columbia, adapted by permission.

Figure 4-3. Sample photograph used in the WipFrag analysis. The white square is the 1m x 1m PVC pipe that was used as scale.
4.4.2 Computer analysis using WipFrag

The photographs were sorted to determine suitability for the WipFrag analysis. Suitability was established based on the ease of picking out grains, and the decision of whether the photograph was representative of the deposit. Photographs that included parts of the stream were not included.

Photographs were digitally compressed to 50% of their original size. The colours were enhanced with the freeware program Irfanview by increasing the contrast of the pictures was increased from 0 to 50 and the saturation from 0 to the maximum of 255. Although the photographs default to black and white in WipFrag, these changes in the colour allowed for increased contrast between the blocks, and enhanced edge detection in WipFrag. The resizing of the photograph was required to allow processing on the available PC.

The initial step in the WipFrag analysis is to set the scale and the tilt. This involves matching the scale bars to the upper and lower edges of the 1 m square (Figure 4-4). The next step is to use the program to draw a net on the photographs, based on edge detection parameters that are set by the user. There are predefined sets of parameters that can be used, or they can be totally user defined. The parameters that are used are window size, threshold, valley threshold, and search lengths 1, 2 and 3 (WipWare Inc, 2003). The edge detection parameters used are defined as:

- **Window size** - the length in pixels of the side of a square window used for thresholding. Decreasing the window size, will cause fusion of finer particles, whereas increasing leads to breakdown of larger particles and loss of information in shadows.
- **Threshold** - the difference in intensity (grey tone level, range between a pixel and its window average). Increasing the threshold results in fewer blocks, and increasing the threshold results in more blocks.
- **Valley Threshold** - specifies a minimum level of grey tone slope to trigger. Increasing the valley threshold gives fewer blocks, and decreasing it produces more blocks.
• **Search Lengths** – are the radial search lengths in pixels measured from a given vertex that the segment operator searches to find and join. SL 1 should be greater than SL 2, which should be greater than SL 3. Increasing these parameters results in more blocks, whereas decreasing them results in fewer blocks. (WipWare Inc, 2003).

The edge detection parameters that produced the best results with the enhanced photographs were those defined in Preset #4. Preset 1 gives the most edges, while Preset 9 results in the least number of edges.

The photographs were mainly of the area inside the 1m square; however there was generally some area around the edge of the square that lay within the photograph. For the analysis, only the area inside the square was included, unless a large grain extended outside the boundary of the square, when the whole grain was included. This procedure was adopted to ensure the same area, 1 m², was examined each time. When the net was drawn on the photograph, the net sections outside of the 1m square were erased, removing this area from the analysis (Figure 4-5). The nets were examined to determine how well they defined the edges of the blocks. Approximately 30 minutes to an hour of editing time was required for each net. The editing involved erasing lines that split blocks, as well as drawing lines on edges that were not detected. It was useful to have the original colour photograph open in another program, so that the blocks could be seen more clearly, if some areas were hard to see in WipFrag.

There is the option to mark areas in the net as either fines, or areas to ignore. Only if a large stick or log were in the photograph, were those areas ignored. Due to both the computer memory limitations and the difficulty in outlining very small grains, ability to resolve finer particles was reduced. Consequently, areas that were relatively fine, generally granules and smaller, were defined as fines.
The output from WipFrag includes a block size distribution histogram and cumulative curve, as well as a chart of the cumulative sizes in ISO standard sieve sizes (Figure 4-6a-c). From these curves, the $D_{10}$, $D_{25}$, $D_{50}$, and $D_{75}$ are defined by WipFrag. Grain size distribution curves were saved for each photograph analyzed, and the D-values recorded. Graphs were also made of the D-values against the station location, either across the channel for each cross section, or longitudinally down the channel (Figure 4-7).

For the longitudinal graphs, the photographs were graphed based on their location relative to the centre of the channel, i.e. a graph was made for the photographs from each cross section that were 10 m to the west of the centre of the channel and another graph for 10 m to the east of the channel centre. Graphing the results longitudinally could potentially help identify different 'flow zones' involved in the event. (All graphs can be found in Appendix 4).

In order to obtain a 3D representation of changes in grain size both across, and longitudinally down the channel, the program Surfer was used. Surfer is a contouring and surface-mapping program (Surfer, 2002). The photograph location and $D_{90}$ data were entered into Surfer, with the cross section number used as the Y-axis, the distance along the cross section line as the X-axis, and the $D_{90}$ value as the z-axis. The final 3D plot shows the changes in grain size across, and down the channel (Figure 4-8).

The coefficient of uniformity, $C_u$, Eqn. 3-1, was calculated from the $D_{75}$ and $D_{25}$ values given by WipFrag, in place of the $D_{60}$ and $D_{10}$ values. Hadjigeorgiou, et al (1996) used a similar approach, using photographs with a reference frame; they considered volume instead of grain size diameter, and also calculated the $C_u$ using the $V_{75}$ and $V_{25}$ values.
Figure 4-4. Setting the scale and tilt on a photograph in WipFrag. The red lines are the scale bars that are aligned with the scale used in the photograph.

Figure 4-5. Unedited net on a WipFrag photograph, with the net outside of the 1m square removed from the analysis.
Figure 4-6. Outputs for Sample 2-2 on the fan from WipFrag. (a) data values for the distribution curve (b) grain size distribution histogram (c) grain size distribution curve with D-values listed.
4.5 Results of WipFrag analysis

The graphs of $D_{90}$ values on the fan show a general decrease in grain size from the apex of the fan, near the mouth of the waterfall, to the fan edge (Figure 4-9). Some lines however show a large spike in the grain size representing a grain size increase of 32 to 84%, or 20 to 40 cm. The $D_{90}$ graphs for each line were plotted on the same graph and the minimum and maximum values were compared (Table 4-1).

In the channel, when the graphs of D-values against location along the cross-section line are examined, there are 2 different situations represented:

1) The grain size reaches a low near the centre of the channel, with peaks on either side of the low – cross-sections 2, 3, 4, and 5 (Figure 4-7a).

2) There is a high in the centre of the channel, flanked by lows – 11 and 6 (Figure 4-7b).

Examination of the graphs of each cross-section, show changes in the grain size distribution, relative to the centre of the channel. In an attempt to investigate these patterns further, longitudinal graphs down the channel were created. There are five graphs on either side of the channel, numbered from west 1 and east 1 at the centre of the channel, to west 5 and east 5 at the channel edges (Figure 4-7b). From the longitudinal channel graphs it is apparent that there is a general grain size decrease as you travel north down the channel, with spikes and troughs in the graphs at specific locations. The graphs are relatively consistent, as to location of the variations, and there are consistencies between graphs at equal distances from the centre of the channel, ie. graph east-4 has a spike at cross section 4, while west-4 has a trough at cross section 4 (Figure 4-10a-b). All WipFrag data graphs can be found in Appendix 4.
Using Surfer, the WipFrag data both across and along the channel could be graphed on one surface (Figure 4-8). A 3-dimensional surface enabled comparison of the relative locations of the peaks and troughs in the D₉₀ values. Peaks in the data occurring on either side of the approximate channel centerline on the cross channel 2D Excel graphs follow relatively straight lines down the channel (Figure 4-8). The 3D WipFrag data surface clearly identifies two relatively coarse zones within the channel. There is also a change in general grain size that becomes apparent, at the location of cross-section 6 (Figure 4-8).

Figure 4-7 (a)
Figure 4-7 (b)

West of Creek 5

Figure 4-7 (c)

Figure 4-7. (a + b) D-values plotted across the channel, with the location of the present day creek indicated. (c) D-values plotted longitudinally along the channel. Graph west of creek 5, all photographs 50 m east of the centre line of the channel.
Figure 4-8. Graph of the $D_{90}$ data with reference to station location.  
Y axis: cross-section number 20-110 (x10)  
X axis: photograph location along each cross section,  
Z axis: $D_{90}$ values
Figure 4-9: Graph showing the general decrease in grain size from the apex of the fan to the edge.

Figure 4-10 (a)
Figure 4-10 (b)

Figure 4-10 – Graph (a) Longitudinal channel graph East-4 (b) Longitudinal channel graph West-4. The graphs show consistency in the variations in the channel, where west-4 has a trough, east-4 has a spike in the grain size.

Table 4-1. Table showing the % difference associated with the grain size increase along the lines on the fan. The increase is seen near the edges of the fan, and may be associated with erosion of finer material at higher flows.

<table>
<thead>
<tr>
<th>Line #</th>
<th>Min (m)</th>
<th>Max (m)</th>
<th>% Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.04</td>
<td>0.16</td>
<td>72.7</td>
</tr>
<tr>
<td>2</td>
<td>0.06</td>
<td>0.14</td>
<td>58.2</td>
</tr>
<tr>
<td>3</td>
<td>0.08</td>
<td>0.25</td>
<td>67.5</td>
</tr>
<tr>
<td>4</td>
<td>0.07</td>
<td>0.44</td>
<td>83.4</td>
</tr>
<tr>
<td>5</td>
<td>0.09</td>
<td>0.13</td>
<td>32.8</td>
</tr>
<tr>
<td>6</td>
<td>0.06</td>
<td>0.33</td>
<td>83.0</td>
</tr>
<tr>
<td>7</td>
<td>0.04</td>
<td>0.25</td>
<td>84.3</td>
</tr>
</tbody>
</table>
4.6 Discussion

4.6.1 Deposits

In the 1991 aerial photograph of the mouth of Glen Falls Creek, it is clear that there is a large amount of sediment deposited in the form of a channel bar. The increase in grain size towards the edge of the fan as seen in the $D_{90}$ graphs could be associated with the old channel deposit seen in the 1991 photographs, where the grain size increase marks the boundary between the ZRRA deposit, and the old channel bar deposit. The grain size characteristics of the old channel deposit are unknown.

The increase in grain size may also be a function of the finer material being eroded from the edge of the fan at higher river flows. As the grain size spikes are seen at the edges of the fan, it is probable that high flows may overrun the fan edges and carry away the finer material, leaving only the coarser grains. Upon examination of the WipFrag photographs where the grain size spikes are indicated, it was observed that the majority of material is subangular, and appears to be part of the debris flow deposit (Figure 4-11). As the material is not rounded cobbles as found in river deposits, this indicates that the increase in grain size is more likely associated with erosion of fines, rather than an old channel bar deposit.

The grain size distribution in the channel was clearly evident when represented in 3D as a WipFrag surface (Figure 4-8). The 3D graph clearly shows where the coarse grains are located in two main zones on either side of the approximate centreline of the channel. The surface also shows the general decrease in grain size that occurs at cross section 6. The location of cross section 6 corresponds closely with a change in the channel gradient from approximately 15° to 7.5° (Figure 2-1). This change in gradient would have resulted in a drop in energy, and thus the carrying power of the event. Therefore the sediment deposited after this point would have been relatively
finer. It may also be expected that there was thicker deposition in this area. Upon examination of the deposition lobes in the channel (Figure 2-9), it can be seen that after cross section 6 the deposition is a consistent 1-2 m thick, whereas above that, the deposit was mostly <1m, though thicker in localized areas.

The presence of the zones of larger grain size could be explained by erosion along the centreline of the channel by subsequent pulses of the debris flow during the event, or perhaps by erosion associated with the reestablishment of the creek channel after the event. This would mean that there was a larger grain size zone along the centre of the channel that was simply dissected by erosion. In terms of conventional river dynamics, the fastest flow is found in the centre of a flowing mass, with the flow velocity decreasing outwards from the centre, as there is more friction associated with the contact with the channel boundary (Knighton, 1998). If the debris flow can be considered a viscous Bingham fluid, then similar principles can be applied. With decreased flow velocity away from the centre, there would also be a decreased carrying capacity, which could result in deposition of the coarser material along the edges of the high velocity flow. In response to decreasing velocity away from the centre, along with the decreased carrying capacity, the deposits would fine out towards the edges.

Experiments by Iverson (2003) with the USGS debris flow flume show that as the material leaves the flume, where it is no longer confined, and the slope decreases, it exhibits a clear interaction between the coarse head of the debris flow and the more fluid tail as it deposits. As the more fluid debris exits the flume, it pushes aside some of the coarser, already deposited debris, forming lateral levees. These levees confine the flow, decreasing lateral spreading, and increasing the runout. This process observed by Iverson (2003) may resemble processes operative in Glen Falls Creek, but on a smaller scale. As the debris flows into the section of the channel with a lower
slope, it slows, depositing some of the coarser debris, while the more fluid body and tail of the debris flow continue flowing, pushing aside some of the already deposited coarser debris, to form the coarse lobes on either side of the channel centreline.

![Image](image.png)

**Figure 4-11.** Photograph 4-12 from the fan, showing that the majority of sediment is still characteristic of a subangular colluvial deposit, and not rounded like fluvial gravels. While some rounded river cobbles can be seen in the photograph, the majority of grains are subangular.

### 4.6.2 Fragmentation

Locat et al. (2003) used a photoanalysis method to examine the comminution and fragmentation energy in seven rock avalanches, as discussed in section 1.3.3. They observed that there was a decrease in the mean size ($d_{50}$) of the deposits associated with an increase in excessive travel distance. The deposits of the ZRRA seem to follow this general trend (Figure 4-12). They also observed that with an increase in the volume of the failed mass, there was a decrease in the reduction ratio or degree of fragmentation. The reduction ratio, which is defined as the relation between the mean block size in the initiation zone and the mean block size in the deposit, is difficult to
determine for the ZRRA. Though the spacing of joints in the headscarp area ranged from 0.001 m – 1 m, large blocks up to 14 m long with intact joints are present in the cirque basin debris. Initial block size, based on the joint spacing, does not have a strong relationship to fragmentation at this site. A reduction ratio or degree of fragmentation could not be accurately calculated, and related to the volume of the ZRRA as undertaken by Locat et al. (2003). The presence of faulting in the vicinity of the ZRRA headscarp may provide an additional explanation for the lack of a relationship between joint spacing in the rock remaining in the headscarp area and the block size in the basin. The rock mass in the area of the headscarp is closely jointed due to tectonic damage associated with the faulting. However the rock in the mass that failed may be characterized by completely different joint set characteristics such as, spacing and persistence, leading to the larger blocks left on the slope and in the cirque basin. More work needs to be done in the failed area in front of the headscarp in order to characterize the initiation zone in more detail.
Figure 4-12 – Relation between the excessive travel distance (Le) as defined by Hsu (1975) and the \( d_{50} \) of the debris. The symbols are defined as follows: CL – Clap du Lac rockslide avalanche, French Alps; Ch – Charmonetier rockslide avalanche, French Alps; SM – Slide Mountain rockslide avalanche, Canadian Rockies; QE – Queen Elizabeth rockslide avalanche; JCN – Jonas Creek rockslide avalanches, Canadian Rockies; Fr – Frank rockslide avalanche, Canadian Rockies; LM – La Madeleine rockslide avalanche, French Alps (Adapted by permission from Locat et al, 2003).

4.7 Summary

Photoanalysis programs are increasingly being used for rock slope design in mines and road cuts, and with recent developments in photoanalysis software this technique is becoming more recognized as a valuable tool for both efficiency and safety concerns in industry. They allow for hundreds of measurements to be taken in a fraction of the time of conventional methods, as well as allowing for the analysis of dangerous slopes by taking photographs up to 700 m away (Surpac, 2005).

This chapter demonstrates the potential use of the photoanalysis program WipFrag in characterizing the block size distribution:
1) In the fan deposit
2) Both across and along the debris channel

This analysis allowed for the quantification of the reduction in grain size with increasing distance from the source area of the landslide.

The characterization of the deposit on the fan showed a general decrease in grain size from the apex of the fan towards the rim (Figure 4-9), some areas showed a rise in grain size near the edge of the fan. The grain size increase near the edge is interpreted to be associated with erosion of fines from the area during high river flows. The debris size characterization in the channel was along lines across the channel. WipFrag indicated changes in the debris across and longitudinally down the channel. Across the channel, coarser debris lobes were seen on either side of an approximate channel centerline, while the longitudinal changes in grain size showed a general decrease from the top to the bottom of the channel, with minor variations (Figure 4-7b, 4-10a-b). The changes in grain size across the channel may be associated with deposition of the snout of the flow, followed by erosion of the center of the deposit by the remaining portion of the flow. The decrease in grain size longitudinally down the channel may be associated with comminution of the debris during flow, or perhaps can be attributed to reduced energy levels in the flow resulting in deposition of the largest material first, followed by subsequently smaller grains further down the channel.

Further work is required in the headscarp area to fully characterize fragmentation during the rock slope failure process. The use of a program that allows for a detailed characterization of the remaining rock mass would allow for a better understanding of the comminution of the debris during the failure. Low-level aerial photography by means of a helicopter would provide an opportunity to use WipFrag to investigate the block size distribution in the cirque basin.
CHAPTER 5 – Dynamic analysis of the ZRRA using DAN

5.1 Introduction

In landslide runout prediction several important parameters need to be determined including the maximum distance reached; flow velocity; thickness and distribution of deposits; and behaviour in bends and at obstacles in the flow path (Hungr, 1995). Modelling can be used to predict these parameters. Both empirical and dynamic models have been used to investigate the motion and runout distance observed in debris flows / avalanches. Empirical models are based on limiting criteria or statistical relationships, while dynamic models use rigid-body analysis, such as mass or energy-based approaches (Fannin and Wise, 2001). Dynamic models may also use a deformable-body approach, employing the principles of continuum mechanics and material rheology.

This chapter outlines available empirical and dynamic models. The theory used by the DAN-W and DAN-3D models is introduced. A detailed review of previous studies using DAN-W is undertaken to provide an important constraint for modelling of the ZRRA. The results of modelling the ZRRA using the DAN-W and DAN-3D codes are presented and compared.

5.1.1 Empirical models

Numerous empirical models have been described within the literature (Benda and Cundy, 1990; Fannin and Rollerson, 1993; Fannin and Wise, 2001). A major difference between empirical and dynamic models is that empirical models do not consider material rheology or movement mechanics. Fannin and Wise (2001) suggest that empirical approaches offer a practical means of runout prediction
when little is known about the material or flow path of the debris flow. However, empirical models are limited, as they are dependent on an existing database of field observations. Fannin and Wise (2001) presented an empirical-statistical model called UBCDFLOW used in the analysis of debris flows in the Queen Charlotte Islands. Their model is based on field observations and post event measurements of landslides in clear-cuts. They examined a database of 449 landslides, and developed a volume-based model, where travel distance is based on cumulative flow volume. In their model, initiation occurs for a known, or assumed, initial volume. Reach morphology determines the flow behaviour, and reach slope angle determines whether the flow is erosional or depositional. The model then assigns a volume change based on reach morphology (Figure 5-1). The event is routed into the next reach if, after the flow volume is recalculated, the total volume is greater than zero. The debris flow stops when the total volume is calculated at zero, and total travel distance is given by the sum of all reach lengths the event traveled through (Figure 5-2). Volumes of entrainment and / or deposition are calculated through regression analysis, based on less entrainment occurring on steeper slopes. The results of the analysis give a probability of distance exceeded.

A back analysis was performed with events from the database, and the observed results agreed closely with those calculated by their model. Fannin and Wise (2001) claim that, based on the quality of results obtained in back-analysis, their model should be applicable to assessment of debris flow hazard in areas with similar characteristics to the Queen Charlotte Islands. As this is an empirical analysis, based on field observations in one specific area, this model is limited to areas with similar soil and terrain characteristics as the Queen Charlotte Islands.

Eliadorani et al. (2003) also used UBCDFLOW to model the runout of a small debris flow at Blueberry Creek, in the Interior of BC. They found very good agreement between model results and field observations. Their analyses showed that for open slope reaches below the initiation there was excellent
agreement between the predicted and observed cumulative flow volumes. For confined reaches, the results were less reliable, as only entrainment could be predicted. The modelled travel distance was accurately determined. Overall the net volume change was almost identical in most of the reaches, but the predicted cumulative volume was as much as 40% greater than the observed volume (Figure 5-3). This shows that the net prediction of entrainment of volume is less accurate than that of deposition. (Eliadorani et al., 2003).

Figure 5-1. Schematic plan view of a debris flow path, with reaches defined as used in the empirical model UBCDFLOW (from Fannin and Wise (2001), by permission).
Figure 5-2. Volumetric relationships produced by the empirical modelling (from Fannin and Wise (2001), by permission).

Figure 5-3. Observed and modelled volume of the Blueberry Creek event. (from Eliadorani et el. (2003), by permission).
5.1.2 The dynamic model (DAN-W)

According to Hungr (1995), dynamic models can be divided into two categories. The first category, lumped mass models, idealizes the motion of a slide based on a single point within the slide mass. The second category is models based on continuum mechanics that are each associated with a specific rheological formula. Evans et al. (1994) emphasize that, while the lumped mass models adequately present the movement of the center of gravity of the slide mass, they cannot simulate the motion of the flow front, which is the variable of interest in runout analysis. Lumped mass models are also unable to account for internal deformation of the slide mass (Hungr, 1995).

There is a well developed group of continuum mechanics models based on the Bingham rheology, as well as numerous lumped mass models based on frictional rheology (Hungr, 1995). Most models utilize a fixed Eulerian reference grid (Sassa, 1988; O'Brien et al., 1993), which, given the unsteady nature of landslide motion, may not be appropriate (Hungr, 1995). It is far more advantageous to use a moving Lagrangian reference grid, as do previously developed models by Savage and Hutter (1989) and Norem et al. (1990).

DAN ("dynamic analysis") is a numerical model of unsteady flow, whose purpose is to serve as a versatile tool for modelling post-failure motion (Hungr, 1995). In this model, the moving mass is replaced by an equivalent fluid, whose bulk properties will approximate the behaviour of the prototype. The properties of the prototype must be determined by back analysis of real events. Hungr (1995) lists

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1 “When a decision is made to analyze a problem numerically, an inherent decision exists to represent the physics discretely. Thus, computations are performed in a discretization of both time and space. The time discretization is dependent on the relative speeds of the phenomena and their coupled interactions. The space discretization is a grid, and the choice of grid is dependent on the geometry and topology of the problem being solved. The Eulerian formulation is one in which the grid is fixed and the dynamic quantities pass through it. The Lagrangian formulation is one in which the grid moves with the dynamic variables.” (Wiley, 2000)
the criteria of a model that are required to accurately represent an unsteady landslide:

(1) The model should have a general rheologic base.
(2) The model should take lateral confinement into account.
(3) The material properties should be capable of being changed at different points along the flow path, to account for overriding of liquefiable soil, or changes in pore water pressure.
(4) The model should allow for internal stiffness of the mass.
(5) The model should be easy to use and efficient.

Hungr (1995) states that the DAN-W model meets these criteria. The main limitation of DAN-W is that it is only an approximation, as it involves reducing a complex 3D problem into a simplified 2D formulation (Hungr, 1995).

5.2 Theory of dynamic analysis of landslides

5.2.1 The DAN-W model

DAN-W is an extension of a lumped mass model, where the slide mass is represented by a number of blocks in contact with each other, free to deform and retaining fixed volumes, while descending down a curvilinear path (Hungr, 1995) (Figure 5-4). The model formula contains a term for basal flow resistance (T), which depends on the material rheology. Each rheological model is characterized by a function relating the resisting force (T) acting at the base of an elementary column to the various parameters of the flow, such as thickness, mean velocity, and assumed rheological properties (Figure 5-5) (Evans et al., 2001). Seven different material rheologies are available within DAN-W:

a) Plastic flow, which is controlled by a constant shear strength;
b) Frictional flow, which occurs when T is a function of the effective normal stress at the base of the flow;
c) Newtonian laminar flow, when $T$ is a linear function of velocity with a dynamic viscosity;
d) Turbulent flow, when $T$ is a function of velocity;
e) Bingham flow, when the resisting force is a function of flow depth, velocity, constant yield strength, and Bingham viscosity;
f) Coulomb viscous flow, where the Bingham yield strength can be made dependent on the normal stress; and the
g) Voellmy fluid, which has a friction and turbulence term.

Rheology can be varied for different parts of the slide path, and at different time stages within the model. The rheologies that have been found to represent recorded events most accurately are the frictional and Voellmy rheologies. For a more detailed description of the rheologies, including equations, used by DAN-W see Hungr (1995).

The frictional rheology as used in DAN-W requires only one input parameter, the bulk friction angle ($\Phi$) (Hungr and Evans, 1996). In the frictional rheology the basal resistance force ($T$) is a function of only the effective normal stress on the base of the flow, it is independent of velocity (Hungr and Evans, 1996). Low friction values reflect high mobility, where:

(Eqn. 5-1) \[ T = A_i \delta H_i \left( \cos \alpha + \frac{a_e}{g} \right) (1-r_u) \tan \Phi \]  
(Friction Equation)

(See List of Symbols, p. xii) The Voellmy rheology contains a friction coefficient ($\Phi$) and a turbulence term ($\xi$). An increase in the friction coefficient brings the Voellmy model closer to the frictional model. The turbulence term represents all rate effects, which may include viscosity, and grain interactions, as well as turbulent effects. The turbulence term ($\xi$) is an inverse parameter, therefore an increase in the turbulence coefficient actually translates to less turbulence in the flow (Ayotte and Hungr, 2000) (See Hungr, 1995, for all equations).
(Eqn. 5-2) \[ T = A_i \left[ \delta H_i \left( \cos \alpha + \frac{a_g}{g} \right) \tan \Phi + \delta H_i \frac{v_i^2}{\xi} \right] \] (Voellmy Equation)

The DAN-W model also allows for channel shape, and provides a reasonable approximation of 3-D confinement effects in all but very narrow channels. Energy losses due to sudden constrictions or abrupt changes in channel shape, or flow direction are not accounted for. Deposition and entrainment are simulated by changing the volume of individual blocks in each time step by a prescribed amount, proportional to the distance traveled, using a momentum flux term. The model must be calibrated, where the appropriate constitutive relationships are selected empirically by back analysis of properly selected prototype events. Model fit is done by trial and error, and is based on 1) total runout distance 2) velocities and 3) distribution and thickness of the deposit (Ayotte and Hungr, 2000).

Figure 5-4. The DAN-W representation of a flow mass in a Lagrangian mesh as a series of blocks in contact with each other, descending down a curvilinear path (from Hungr (1995), by permission).
5.2.2 The three-dimensional rheologic model, DAN-3D

DAN-3D is a new continuum model developed at UBC that can accommodate flow over complex 3D terrain (McDougall and Hungr, 2004). DAN-3D is based on the same principles of hydrodynamics as DAN-W, but now incorporates "Smoothed Particle Hydrodynamics". The new model includes the following features required for modelling mobile landslides:

1) non-hydrostatic internal stress distribution, as appropriate for internal flow deformation of a frictional material,
2) an open rheological kernel, allowing simulation of flow in a variety of materials,
3) the ability to entrain material,
4) a flexible mesh-less solution method, suitable for complex 3-D path geometry. (McDougall and Hungr, 2004)

With DAN-3D there is no need to input a flow path, as was required in DAN-W. The model still treats the slide mass as an equivalent fluid, but with a predefined internal frictional rheology, and a user defined basal rheology. The basal
rheology is defined as in DAN-W, by trial and error back analysis of previous events. Erosion of the flow path is also user defined, as in DAN-W.

One of the most important changes from the DAN-W model is the incorporation of Smoothed Particle Hydrodynamics. This allows the model to be unlimited by a mesh. Therefore the particles are not tied together, and the fluid can flow around obstacles, and rejoin on the other side (Figure 5-6).

Figure 5-6. Hypothetical simulation of sand flowing around an obstacle and rejoining (from McDougall and Hungr (2004), by permission).
5.3 Previous applications of the DAN model

Numerous events have been modelled during the development and calibration of the program DAN-W (Hungr, 1995), including the Pandemonium Creek Rock avalanche (Hungr and Evans, 1996) and the Nomash River debris flow (Ayotte and Hungr, 2000). In addition, extensive studies have been carried out in Hong Kong (Ayotte et al., 1999) and Italy (Revellino et al., 2004) in order to calibrate specific models for these areas. Testing of the DAN-W model also provided satisfactory results for modelling of coalmine waste slides (Hungr et al., 1997).

Large rock avalanches that can be categorized as being similar to the ZRRA (as discussed in Chapter 2) are the Pandemonium Creek (Evans et al., 1989; Hungr and Evans, 1996), Nomash River (Ayotte and Hungr, 1999; Favero, 2000), and Mount Cayley (Evans et al., 2001) rock avalanches. All of these events were successfully modelled with DAN-W. Hungr and Evans (1996) present back analyses of 23 different rock avalanches, including the Pandemonium Creek event.

Analysis of landslides in Hong Kong was also undertaken with the DAN-W model, in order to compare slides of different volumes. The slides in Hong Kong were all of relatively small volume (< 5000 m$^3$), while those modelled in BC have had volumes of up to 100,000 m$^3$ (Ayotte et al., 1999). Models of different slides were successfully created through back analysis of known events. In this study debris avalanches were modelled using the frictional model, and long-runout channellized debris flows were modelled with the Voellmy model. In some debris flow events it was found that a combined frictional-Voellmy model was more accurate in cases where the event began as a slide then entered a channel and became a debris flow. The results of the study in Hong Kong indicate that the model has excellent potential for prediction purposes (Ayotte et al., 1999).
Coalmine waste slides can be very mobile and destructive, with runouts of up to 3 km. They have been likened to natural debris avalanches by Hutchinson (1995). Coal waste flow slides have been successfully modelled using the frictional rheology, combined with the Voellmy rheology where the flow slides enter a channel or gully and debris entrainment is likely to occur (Hungr et al., 1997). Successful modelling with these rheologies indicates that the long runout of coal waste flow slides is due entrainment of partially or fully liquefied saturated soil at the base of the slide. Slide mobility seems to be strongly influenced by the character and pore water pressure of the substrate.

When modelling these events it was discovered that the combination of the frictional model for the initial motion and the Voellmy model for the runout provided the most accurate results (Ayotte et al., 1999). When the frictional model alone was used the debris distribution consisted of thin fronts, and thick proximal sections, and produced exaggerated velocities (Ayotte et al., 1999; Ayotte and Hungr, 2000). Using the Voellmy model alone, the results produced lower velocities and uniformly distributed deposits over fairly short areas, with accumulation on the flatter parts of the slopes (Ayotte et al., 1999; Ayotte and Hungr, 2000). Overall the Voellmy model produced more accurate results, but the combination of the two models resulted in the most realistic representations of some of the events as observed in the field.

DAN-3D has been tested on laboratory experiments, as well as with real events (McDougall and Hungr, 2004). A debris avalanche near the village of Cervinara, Italy was accurately modelled (McDougall and Hungr, 2004). This was a complex event, that involved a portion of the initial landslide running up on the opposite valley wall and over a ridge into a smaller gulley, while the remaining material mobilized in a creek channel and traveled almost 2 km to the village. The model accurately simulated the run-up, and both the total travel distance and volumes. The frictional and Voellmy rheologies were used to model the event, and the input parameters were based on previous analyses using DAN-W of
similar events in the area. The fact that the same inputs were successful in both models suggests that DAN-W can be used for preliminary calibration in more complicated cases. (McDougall and Hungr, 2004).

5.4 Procedure for rheologic modelling

5.4.1 Methodology for DAN-W modelling

Prior to the analysis of the ZRRA an extensive literature review of events modelled using DAN-W was completed, and the model parameters used for each event were recorded in a spreadsheet (the complete spreadsheet is included in Appendix 5). Figure 5-7 shows the values used with the Voellmy model for the different events, $\xi=500$ and $\Phi=0.1$ are the values found to produce the most accurate results. The parameters chosen for modelling the ZRRA were based on those used when modelling similar events, including Pandemonium Creek, Nomash River, and Mount Cayley. As DAN-W allows the assignment of different rheological models to different locations along the flow path, the local channel and event characteristics were taken into consideration when assigning the various rheologies.

To input the ZRRA slope cross-section into DAN-W a profile was drawn of the pre- and post-slide valley in the GIS. This profile was imported directly into DAN-W. The width of the slide path was also determined from the GIS. The failed block was assigned an average depth of approximately 75 m based on the volumes calculated from the GIS, and the volume produced in DAN-W based on the slope configuration.

For the first stage of the ZRRA, the rockslide into the basin, a frictional rheology was used. As the first stage comprised a rockslide, there would have been no
overrunning of saturated and liquefiable sediments, thus the Voellmy model is not appropriate at this stage. The frictional model was used up to the midpoint of the basin, after which the Voellmy model was incorporated (Table 5-1). The Voellmy model was used for the remainder of the ZRRA event in two sections, with 0.2 m of erosion specified in the second section in the basin and 2 m of erosion specified below the basin in section 3 (Figure 5-8).

![Graph showing the distribution of values used in the DAN-W modelling for different landslides](image)

Figure 5-7. Graph showing the distribution of values used in the DAN-W modelling for different landslides: 2-Frank, 3-Flims, 4-Goldau, 6-ZRRA, 7-Pandemonium, 8-Nomash, 9-Hummingbird, 11-Shum Wan Rd., 12-Lantau Island, 17-Avalanche Lake S, 19-Rubble Creek, 21-Kennedy River, 22-Mystery Creek, 24-Mt St Helens, 26-Sherman Glacier, 27-Gros Ventre, 28-Val Pola, 30-Diablerets, 31-Elm, 32-Mount Ontake, 33-Mayunmarca, 34-Mount Cayley.)
5.4.2 Methodology for DAN-3D modelling

The DAN-3D modelling was undertaken in collaboration with Scott McDougall and Oldrich Hungr of the University of British Columbia. Digital elevation data, as well as event characteristics and a complete photographic documentation of the event (Appendix 1) were supplied to UBC for the modelling. The DAN-W modelling was also provided so that similar inputs could be used in both models (Table 5-1).

The frictional model was assigned for elevations above 900 m, with the Voellmy model used below that. For the 3D model, an erosion rate is defined, instead of a depth as in DAN-W. Erosion was assigned at a rate of 0.0001 m eroded per m of flow depth per m traveled, which resulted in a maximum erosion depth of 1.5 m in the channel. The initial volume in the 3D model was 920,000 m$^3$, with the final volume at 1,100,000 m$^3$. 

Figure 5-8. Sketch illustrating the assignment of rheologies to the DAN-W model. Section 1-frictional rheology; Section 2 – Voellmy with 0.2 m of erosion; Section 3 – Voellmy with 2 m erosion. The black section is the failed mass.
5.5 Results of rheologic modelling

5.5.1 DAN-W modelling of the ZRRA

Numerous variations of the final model were tested with different locations assigned for the changes in the rheologies used, as well as various material properties. Table 5-1 shows the final material input parameters that were found to provide the most accurate results when compared to observations in the field. The parameters that were used to constrain the models against actual field observations were: velocity, deposit depth, total runout, and final volume. The output from DAN-W provides the maximum velocity reached, and where that velocity was in the flow, final volume, total runout, and Fahrböschung (Table 5-2). The results also include graphs of velocity and deposit depth with distance (Figure 5-9). The velocity graph shows that a maximum velocity of 49.2 m/s was reached at the end of the rockslide section of the event. The simulated velocity through section 2 is approximately 20 m/s, while the average simulated velocity in section 3 is 15 m/s (Figure 5-9). The deposit profile along the length of the event shows a 10 m thick deposit in the basin, behind the moraine and island. The deposit thickness in the main channel is between 1 and 4 m, with a large thickening before the canyon, and some material deposited in the canyon. These values agree closely to the field observations (Figure 2-9).
Figure 5-9. Output from DAN-W. (a) Velocity vs. Distance (b) Thickness vs Distance.

Table 5-1. The final properties used in the DAN models.

<table>
<thead>
<tr>
<th>Final Properties</th>
<th>Material 1</th>
<th>Material 2</th>
<th>Material 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rheology</td>
<td>Frictional</td>
<td>Voellmy</td>
<td>Voellmy</td>
</tr>
<tr>
<td>Friction Angle</td>
<td>30</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Friction Coefficient</td>
<td>-</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>Turbulence Coefficient</td>
<td>-</td>
<td>500</td>
<td>500</td>
</tr>
<tr>
<td>Erosion Depth</td>
<td>-</td>
<td>0.2 m</td>
<td>2 m</td>
</tr>
<tr>
<td>Internal Friction Angle</td>
<td>20</td>
<td>15</td>
<td>10</td>
</tr>
</tbody>
</table>

Table 5-2. Results of the DAN-W and Dan-3D models.

<table>
<thead>
<tr>
<th>Model Results</th>
<th>2D</th>
<th>3D</th>
<th>Field Estimates</th>
</tr>
</thead>
<tbody>
<tr>
<td>Initial Volume (m³)</td>
<td>1,003,754</td>
<td>920,000</td>
<td>1,000,000</td>
</tr>
<tr>
<td>Final Volume (m³)</td>
<td>1,295,246</td>
<td>1,100,000</td>
<td>1,300,000</td>
</tr>
<tr>
<td>Max Velocity (m/s)</td>
<td>49.2</td>
<td>42</td>
<td>34.3</td>
</tr>
<tr>
<td>Travel Distance (m)</td>
<td>4326</td>
<td>-</td>
<td>4500</td>
</tr>
<tr>
<td>Fahrböschung</td>
<td>16.27</td>
<td>-</td>
<td>15.12</td>
</tr>
</tbody>
</table>
5.5.2 DAN-3D modelling of the ZRRA

The DAN-3D modelling was completed with the same inputs that were used in the 2D model. The results were slightly different however. The 3D model shows in detail the progression of the flow down the channel in time steps, as well as thickness of deposits (Figure 5-10).

The initial results of the 3D model indicate that the elevation of 900 m specified for the location of the change from the frictional to Voellmy rheologies may be too low, as the run-up in the basin is not high enough with the current transition at 900 m. This may be improved by moving the transition up to 1000 m, so that the Voellmy fluid was used as soon as the rockslide encountered the snow in the basin. The deposits in the channel are concentrated in similar locations as indicated in the DAN-W model. There is a thick deposit in the basin at the base of the slope, however relatively little is deposited behind the moraine and the island. This may be related to the frictional rheology being used until 900 m. As seen in Figure 5-10 the bulk of the deposit is in the frictional zone, therefore the material may have been retarded too much in this area to an extent that it failed to reach further into the basin. The thick deposit above the canyon also appears in the 3D model, with some material remaining in the canyon as well.
Figure 5-10. Model output from DAN-3D. Each frame is a snapshot at time steps of 100s, up to $t=600$s.

Legend
Blue – Voellmy rheology
Green – Frictional rheology
Dashed line – trim line
Solid contours – 2m contours of deposit depth
5.6 Discussion

DAN-W accurately represents the dynamics of large landslides (Hungr and Evans 1996; Ayotte and Hungr, 2000; Evans et al., 2001). It has been used in this study to model the ZRRA adopting combined frictional and Voellmy rheologic models, and producing realistic results. The model inputs were derived from a database of properties used in the modelling of other landslides (Figure 5-8 and Appendix 4). The values that were most commonly used with the Voellmy model $\xi=500$ and $\Phi=0.1$ also proved accurate for this analysis.

The DAN-W output included graphs of velocity and deposit depth with distance. The DAN-W model results indicate high velocities for the rockslide, up to 49 m/s. The velocity calculated for this section of the event, based on the run-up in the corner of the basin is only 34 m/s (Table 3-1). The higher velocity simulated in DAN-W is probably a result of the fact that the right angle corner could not be properly represented in 2-dimensions. Turning the almost 90° corner in the basin could potentially have slowed the rockslide, which would not have been accounted for in the model, where the right-angle corner could not be included. The calculated velocity estimates made in the valley based on superelevation correspond closely with the DAN-W results (Table 5-3).

The deposit depth graph showed a thickening in the deposit before the moraine in the basin, and a more extensive thickening in the channel immediately before the canyon. The deposit in the basin seems to have dammed behind the moraine in the model. These results are consistent with field observations. It was not possible to measure the basin deposit thickness in the field, as it was completely composed of large boulders; however it was noted that there was a thicker deposit in this area. The thickening of the deposit in the channel before the canyon is apparent on a smaller scale in the field. The deposit thickness in this area was noted as up to 4m (Figure 2-9). The model however predicts a
thickness of up to 15 m. As mentioned in Chapter 2, a second debris flow occurred at 9:30 the next morning. It is possible that the model is correct in leaving a thick deposit before the canyon, and this material in reality was later flushed out by the second debris flow, leaving only the 3-4 m thick deposit now present in the field.

The results of the 3D modelling are very useful for comparison with the 2D results. The 2D model is necessarily greatly simplified as the right-angle bends in the basin and the canyon must be ignored. Given the recognized complexity of the channel geometry a 3D model is more appropriate. Results showed that the parameters used in the 2D model were less appropriate for the 3D model, as the 3D version can actually incorporate the complexities of the channel. From the comparison of the models, the following interpretations / observations were made:

- The snow in the basin is a highly significant factor in the long runout of this event. If the slide had failed into a snow free basin, it may not have even run out of the basin.
- The Voellmy model should be used as soon as the rockslide would have encountered the snow in the basin, as the 30° friction angle defined in the frictional model is much too high to simulate flow over snow.
- The friction coefficient of 0.1 used in the Voellmy model correlates to a basal friction angle of 5.7°, which is lower than the slopes in the basin, allowing for the material to flow out of the basin. If the frictional model is assumed over too great a distance with a 30° friction angle, the material does past the top of the basin, where the slope is less than 30°.
- The lower basal friction angle associated with the Voellmy rheology may also extend the flow sufficiently to cause it to split around the island and rejoin on the downstream side.
The 2D and 3D DAN models have shown that they can accurately predict the runout of a large, complex rock avalanche event. The modelling has provided further insight into the second debris flow pulse that occurred the next morning, as prior to the modelling, the general source area of the second event was unclear. The back analysis of the ZRRA could be useful for determining the runout of potential events in other valleys in this region, and for future modelling of similar complex rockslide-debris avalanche events.

Table 5-3 – Velocity comparison between the calculated and modelled velocity results.

<table>
<thead>
<tr>
<th>Velocity</th>
<th>Calculated</th>
<th>Modelled-2D</th>
</tr>
</thead>
<tbody>
<tr>
<td>C-S 2</td>
<td>15</td>
<td>15</td>
</tr>
<tr>
<td>C-S 3</td>
<td>16</td>
<td>15</td>
</tr>
<tr>
<td>C-S 7</td>
<td>22</td>
<td>17</td>
</tr>
<tr>
<td>C-S 8</td>
<td>16</td>
<td>21</td>
</tr>
<tr>
<td>C-S 9</td>
<td>20</td>
<td>22</td>
</tr>
<tr>
<td>Basin</td>
<td>34</td>
<td>49</td>
</tr>
</tbody>
</table>

5.7 Summary

Modelling, both empirical and dynamic, has proven useful in the interpretation of landslide events as indicated by numerous authors (Sections 5.1 and 5.3). The DAN-W model has been used successfully to model rock avalanches in B.C. and around the world (Section 5.3). The recent development of DAN-3D should prove to be a valuable addition to the toolbox used in landslide runout prediction.

This chapter has demonstrated that the ZRRA exhibited similar characteristics to other modelled rock avalanches. This is shown in the fact that a similar methodology used in modelling of other events provided realistic results for the ZRRA. The methodology included using the frictional, and then Voellmy models, as well as similar input values for the Voellmy model (Table 5-4).
The DAN-W model was shown to be a useful first step for the 3D modelling. As the 3D version is more complex, and requires a significant amount of time to run, estimates of input values can first be determined in DAN-W, and then used in DAN-3D. This procedure could save considerable time in the modelling process.

The results determined in the modelling correspond closely to the observations made in the field for deposit depths, as well as the calculated velocities at different locations along the channel. The modelling may also have provided evidence indicating the source location of the second debris pulse that occurred the morning after the main event.
CHAPTER 6 – Summary and recommendations for future work

6.1 Summary and Discussion

This thesis has presented a detailed characterization of the Zymoetz River rock avalanche. The characterization followed a version of the methodology proposed by Couture et al. (1999) modified to accommodate advances in technology (Figure 1-2). The methodology adopted included studies of the rock mass and debris through data collection, photo documentation, field and laboratory work (Chapter 2 and 3) and photographic and dynamic analysis and interpretation (Chapters 4 and 5).

The ZRRA was a natural landslide that initiated as a result of progressive degradation of an altered rock mass, on the glacially over-steepened walls of a cirque basin. As the rockslide impacted the cirque basin it was accelerated down the valley, aided by reduced basal friction on the snow at the south end of the basin, and entrainment of some snow and saturated debris towards the north end. The debris avalanche exhibited spectacular mobility as it flowed down the valley, evidenced by high superelevations, and mud splashes on trees 60 m above the valley floor. A large amount of debris remained in the basin, some was deposited in the valley, replacing the thin layer of entrained soil, while the majority traveled over 4 km to the Zymoetz River, and formed a large fan. The damaged PNG pipeline had to be rerouted to the top of the bedrock canyon, so as not to be impacted again by future events initiating in Glen Falls Creek.

Difficulty was encountered in the naming of the ZRRA as discussed in Chapter 1. According to basic definitions this event had the components of a rockslide, and a debris flow. With the definition of Varnes (1978), this event is
classified as a rockslide-debris avalanche, with the term rock avalanche as a shortened version of that name. Hungr and Evans (2004b) also would classify this event as a rockslide-debris avalanche based on the amount of debris that was entrained, and an entrainment ratio of > 0.25. The term debris avalanche is generally reserved for debris flows that occur on open slopes, while the term debris flow is used for channelized events. In this case it would seem that according to the definition of debris flow, this term is more appropriate for the second stage of the event. However similar events in the literature, following the system of Varnes (1978), have been named rock avalanches, and as such, so is this event.

Soil sample collection and rock mass observations aided in the interpretation of the event. The grain size analysis, through sieving and Atterburg limit testing, provided details on the material entrained during the event (i.e. higher plasticity values were seen where significant organics had been incorporated, as well as higher clay contents where silt and clay deposits had been overrun). The BCGS geology map (Figure 3-6) indicated the presence of a large fault in the headscarp area, which was supported by field observations of shearing in the backscarp, epidote mineralization and slickensided surfaces. Joint survey data above the headscarp does not correspond with the block sizes seen in the debris, indicating that the rock mass that failed is potentially much different from the rock remaining at the headscarp. This further supports the interpretation that the fault coincides directly with the failure plane.

The utility of the program WipFrag for the analysis of this type of event has been proven useful and shows considerable potential in the analysis of long runout channelized flows / avalanches. The program allows for the analysis of a large collection of photographs in a relatively short time. The results obtained from this analysis fit with observations of other authors in terms of
decreasing grain size with distance from the source, and coarse lobes along the margins of the deposit (Locat et al, 2003; Iverson, 2001).

The DAN model also proved to be very useful in the characterization of the ZRRA. The integration of the DAN-W and DAN-3D models with field observations and the GIS techniques allowed for additional interpretations of the dynamics of the event (ie. potentially indicating the source location of the second debris flow). Because the 3D model is more complex, calibration of the model must be undertaken by running a model, and then field checking the results or correlating the observations with a detailed GIS database, as was attempted here. Further development of DAN-3D will lead to significant advances in the understanding of the mechanics of rock avalanches and debris flows, and provide a more accurate runout prediction tool.

As there is still unstable material remaining in the initiation zone of the ZRRA, there is the potential for another large failure. As mentioned in Chapter 5, the snow had a significant impact on the runout, as the reduced basal friction allowed the debris to leave the basin. DAN-3D modelling indicates that had the snow not been there, the debris may never have exited the basin. With \( \sim 500,000 \text{ m}^3 \) of rock in the basin, including blocks that are up to 5-7 m high, it is suggested that a significant snow pack would be required to reach a condition where the basal friction would be reduced again to the point where another failure could be mobilised enough to exit the basin. While the potential remains high for another large rock failure, the chances of the event exiting the basin are probably much lower.

In British Columbia the low population density rarely necessitates extensive monitoring of isolated high mountain slopes. As population, logging and recreation industries continue to spread into remote areas there is a consequent incremental increase in risk posed by rockslides. While it is already common practice to investigate the potential for debris flows or other
small mass wasting events in populated or environmentally sensitive locations, it may become increasingly necessary in the future to investigate the potential for long runout large landslides when considering recreational developments in remote areas or areas proposed for timber harvesting. This threat was clearly evident after the Pandemonium Creek event that devastated a fan 9 km away from the source (Evans et al., 1989). The Zymoetz River rock avalanche, which caused extensive damage to both a gas pipeline and forest road, further emphasizes the need to consider the risks posed by such long runout events.

6.2 Recommendations for Future Work

Based on this work, the following recommendations are made for further studies at the site and for the study of these types of events in general.

Further work on the ZRRA:

- **Initiation Zone**
  - More detailed characterization of this area will provide an improved understanding of the deformation and alteration of the rock mass and the influence of the fault.
  - Collection of more rock samples from the headscarp area and from the blocks remaining on the slope for strength testing.
  - Estimates of the volume of material remaining on the failure plane and evaluation of the stability of that material.
  - More detailed estimates of the initial block size, involving a traverse of the debris remaining on the slope, which would allow for further examination of the fragmentation of the debris.
Transport / Deposition Zone

- Detailed mapping of the deposit in the cirque basin to examine the different flow lobes that appear in the deposit.
- Further field characterization of the deposit to constrain the results of 3D modelling.
- Low-level flights to take high-quality photographs of the deposit in the basin for further WipFrag analysis.
- Collection and analysis of samples in the channel to more accurately determine the source of entrained material.
- Collection of higher resolution photographs and the use of a more powerful computer to process WipFrag pictures so that the 'finer' portion of the debris can be characterized more accurately.
- Continued DAN-3D modelling to examine the effect of the existing failure debris in the basin on the runout of another event.

Further study of these large events is required to learn more about their spectacular fluidity and associated long runouts. The need to recognize the potential for events of this magnitude is becoming greater as development and recreation extend farther into mountainous areas.
REFERENCES


APPENDICES

Appendix 1:
This appendix includes two digital posters. The introduction poster includes the orthophotograph of the landslide with information and photographs attached at different locations. The poster of the whole channel includes photographs of the entire west and east walls of the channel.

Appendix 2:
This appendix includes the procedures and results for the particle size analysis discussed in Chapter 2. Photographs of the samples from the sieve analysis are also included.

Appendix 3:
Appendix 3 includes the data collected in the discontinuity surveys of the Glen Falls Creek area, as well as descriptions of rock samples collected and thin sections.

Appendix 4:
Appendix 4 includes the photographs used in the WipFrag analysis. The Excel data files used to create the graphs in Chapter 4 are also located in this appendix.

Appendix 5:
This appendix includes the data gathered from a literature review of other studies that have used the DAN program. The data are compiled in an Excel spreadsheet.