Search

Main Menu

CATASTROPHIC ROCK AVALANCHES, WEST-CENTRAL BRITISH COLUMBIA

James W. Schwab, BC Forest Service, Smithers, B.C.
Marten Geertsema, BC Forest Service, Prince George, B.C.
Stephen G. Evans, Geological Survey of Canada, Ottawa, Ont.

ABSTRACT

The Bulkley and Babine Ranges of the Hazelton-Skeena Mountains, west-central British Columbia, have experienced three recent large catastrophic rock avalanches: Howson, 19 September 1999; Zymoetz, 8 June 2002; Harold Price, 22 or 23 June 2002. These landslides are large relative to contemporary landslides, although many large prehistoric landslides are evident within the Hazelton-Skeena Mountains. The recent landslides may be a result of climate change. Climate warming has resulted in: a pronounced thinning and recession of mountain glaciers causing debutressing of unstable rock slopes; possible degradation of mountain permafrost; and, an apparent increase in precipitation (rain and snow) over the past few years. Large rock avalanches place utilities and transportation routes in the mountain valleys at a significant risk. In addition, risks are increased to forest workers and recreation users in the valleys. The tremendous down valley travel distance of these landslides suggest some communities may also be at risk.

RESUME

Les chaînes Bulkley et Babine situées dans les montagnes Hazelton-Skeena, au centre-ouest de la Colombie-Britannique ont récemment éprouvé trois grandes avalanches rocheuses catastrophiques, soient à Howson, le 19

002 et à Harold Price, le 22 ou 23 juin, 2002. On considère que ces glissements de terrain sont plus importants que d'autres avalanches rocheuses contemporaines. Par contre, il y a plusieurs glissements de terrain préhistoriques qui sont comparables. Il e

terrain soient provoqués par le changement climatique. Or, depuis quelques années, le réchauffement climatique a produit un amincissement prononcé et un retrait des glaciers de montagnes, ce qui ensuite a cré

contraintes des pentes rocheuses, une dégradation du pergélisol en milieu montagneux et une augmentation évidente de précipitation (pluie et neige). Les avalanches rocheuses portent un danger important aux services publics et aux routes de transport dans les vallées. De plus, les dangers sont aussi importants pour les travailleurs forestiers et les amateurs de plein-air voyageant dans ces vallées. Les énormes distances parcourues par ces glissements de terrain

que les communautés situées dans ces vallées sont davantage en danger.

1. INTRODUCTION

This paper presents preliminary findings on three large rock avalanches in west central British Columbia (Figure 1): Howson, September 19, 1999, located 50 km west of Smithers B.C. and a like distance east of Terrace B.C. (53° 31' N; 127° 46' W); Zymoetz, June 8, 2002, located 18 km south east of Terrace B.C. (54° 26' N; 128° 18' W); and, Harold Price, June 22 or 23, 2002, located 35 km north east of Smithers B.C. (55° 04' N; 126° 57' W).

These landslides are large compared to contemporary landslides in the Hazelton-Skeena Mountains. They are significant in that they add to the growing list of large rock avalanches involving steep rock slopes adjacent to glaciers or cirque basins. Other recent rock avalanches in B.C. include: Mount Munday, Waddington Range, spring of 1997 (Evans and Clague 1998); Nomash River, April 1999, 16 km north of Zeballos Inlet; Goat Mountain, July 1999, 45 km NW of McBride on Kendall Glacier; Glacier Bay, September 1999, east of Cascade point, Knight Inlet; McAuley Creek, June 2002, 15 km south of Lumby; Pink Mountain, June 2002, 15 km west of mile post 150 on the Alaska highway. Most landslides have occurred in relatively remote mountain valleys. However, these recent landslides in west central BC have had a major economic impact through the destruction of timber, severing of a natural gas pipeline, a year long closure of an important resource road and a short term industrial shutdown of facilities in Terrace, Kitimat and Prince Rupert.

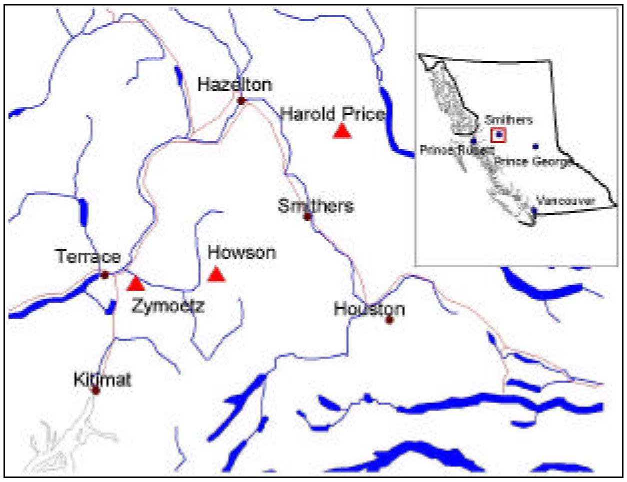

Figure 1. West central, British Columbia and the location of Howson, Zymoetz, and Harold Price rock avalanches.

"Rock avalanche" is a generic term used to describe extremely rapid and highly mobile landslides typically resulting from a rock fall or slide. Rock avalanches in the Canadian Cordillera have been described by Eisbacher, 1979; Eisbacher and Clague, 1984; Evans et al. 1989; Cruden and Lu, 1992; Hungr et al. 2001; and others.

1.1 Physiographic Setting

The Bulkley Ranges are situated east of the coastal mountains within the Hazelton Mountains commencing at Zymoetz River east of Terrace and extend through to the Bulkley Valley at Smithers (Figure 1). East of the Bulkley Valley is the Babine Range within the southern extent of the Skeena Mountains (Holland 1976). Bulkley and Babine Ranges are sculpted by glacial ice with glaciers remaining along the crest of higher mountains and in cirque basins. Glacial features from the Little Ice Age are evident throughout the mountains.

Bedrock is predominately volcanic in the Bulkley and Babine Ranges (Tipper 1976; Richards 1990) with a central stock of plutonic rocks. Sedimentary rocks are more prevalent in northern parts of Bulkley and Babine Ranges. These volcanic and sedimentary rocks are Jurassic or Cretaceous in age.

1.2 General precipitation and snowmelt

Precipitation in the Babine and Bulkley Ranges is in the order of 1,500 to 2,000 mm per year, respectively, with 50 to 60 % as snow. Snow accumulates from the end of September through to mid May. Snow can occur at higher elevations every month of the year. The snow pack in mid May ranges on average from 500 to 1200 mm, snow water equivalent. Maximum values have reached 1400 to 1900 mm. Run-off from snowmelt peaks early to mid June and is generally complete by the end of June.

2. HOWSON ROCK AVALANCHE

At 03:00 hours PDT on Saturday 11 September 1999, Pacific Northern Gas (PNG) pipeline through the Telkwa Pass was severed by a rock avalanche (Figure 2).

A series of mountain glaciers extend northward from Fubar Glacier in the Howson Range into Limonite Creek valley (Telkwa Pass). Ridges between the glaciers extend up to 300 m in height. Rock cliffs also rise directly above Limonite valley.

2.1 Howson Description

The rock avalanche originated as a topple of about 0.9x10^{6} m^{3} from a bedrock ridge at 1923 m.a.s.l. The rock toppled and slid on a 48° slope for 150 m on to glacial ice. The rock avalanche expanded to cover the glacier to a width of 300 m spreading out on a 10° slope (Figure 3). Rubble hurtled over the ice along an ever-increasing slope gradient, dropping into Limonite Creek valley over a slope of close to 40°. Travel through the forest was along a slope gradient of 22°. Deposition occurred on slopes of 5° to 10°. The rock avalanche travelled a distance of 2.7 km, dropping 1,300 m in elevation[1].

Figure 2. Howson rock avalanche. Note cliffs, pipeline, power line and new lake (September 16, 1999).

Figure 3. Howson rock avalanche, view toward head scarp. Note near vertical joints (September 16, 1999).

---

[1] Slope gradient and distance estimates as presented are derived from 1:20,000 TRIM topographic maps.

The slide path through mature forest covered an area 1,200 m long and up to 400 m wide (Figure 4). Scour along the slide path through the forest, incorporated colluvial and morainal material. Trees were blown over by the resultant air blast along the sides and front margin. Landslide debris dammed Limonite Creek, creating a lake that filled within a few days. The avalanche was contained along the west and east boundary, below the glacier, by lateral moraines and a gully. Flow features appear in the zone of accumulation, issuing from the gully along the east boundary of the avalanche—these flow features possibly indicate that a debris flow, from the gully, occurred shortly after the initial avalanche. The gully was filled with landslide debris, which remobilized as a debris flow. Despite the large forest area contained in the avalanche path, few trees are visible in the landslide debris. A few large boulders, some up to 10 m diameter, were strewn along the landslide path. Howson rock avalanche appears to have involved an estimated 1.5×10^6 m^3 of material.

Figure 4. Howson rock avalanche, view upslope toward glacier ice. Note temporary gas line in fore ground.

2.2 Geology

Steep granodiorite cliffs rise above the glacier. The bedrock is steeply jointed and highly fractured. Preliminary analysis shows no clear relationship between discontinuities and the apparent failure.

2.3 Velocity estimate

The Howson rock avalanche appears to have moved at an ever-increasing speed down the glacier into Limonite valley. Maximum velocity was probably reached as the avalanche dropped into Limonite valley. At about 500 m below the drop, uprooted and snapped trees were observed. Velocity estimates necessary to uproot and snap trees are in the order of 18 to 30 m/s, as discussed by Cruden and Lu (1992).

2.4 Climatic conditions

Precipitation and temperature at the Terrace airport prior to the Howson 1999 rock avalanche are presented in Figure 5. Precipitation during the months of May to September 1999, was 2 times normal for the Terrace airport at 217 m.a.s.l. Temperature was below normal. Similar conditions, with even cooler temperatures, were reflected at the airport in Smithers at 522 m.a.s.l. Conditions were generally clear and cool at the time with freezing temperatures at the higher elevations in Howson Range.

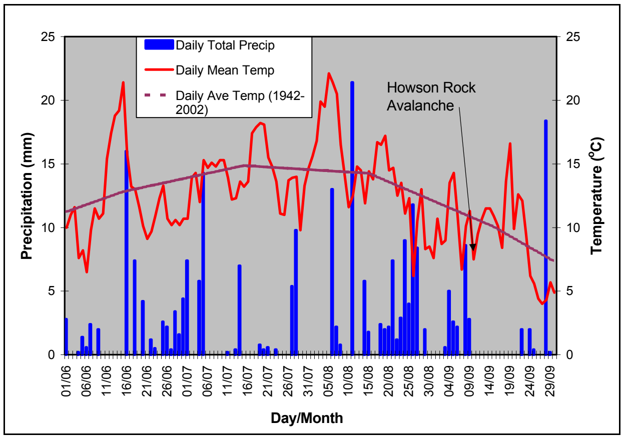

Figure 5. Precipitation and temperature, Terrace airport August-September 1999.

3. ZYMOETZ ROCK AVALANCHE

On Saturday morning approximately 01:15 to 01:30 hours PDT June 8, 2002, Pacific Northern Gas (PNG) pipeline was severed at 15.5 km on Copper River Road. An intense fire at the break could be seen by residents along Highway 16 near the bridge crossing of the Zymoetz River. Water Survey of Canada station (08EF005) situated 3 km down stream recorded the blocking of the river, over topping of the dam and a smaller debris flow observed by PNG personnel at 10:15 hours PDT (Figure 6).

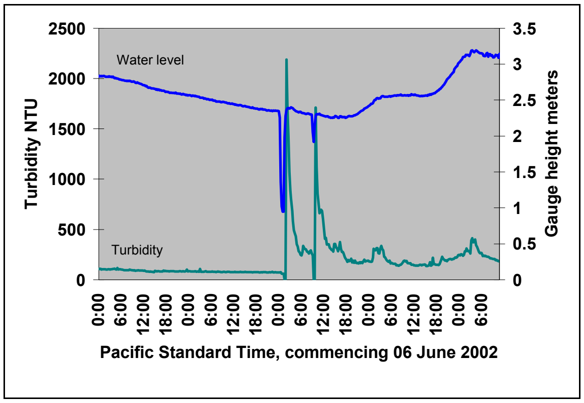

Figure 6. Water level and turbidity recorded for Zymoetz River at Water Survey of Canada station, 3km down stream of the Zymoetz rock avalanche.

3.1 Zymoetz Description

The Zymoetz rock avalanche originated in a bedrock promontory at 1390 m.a.s.l., mid slope in a steep cirque basin (Figure 7). About $1.4x10^6$ m^3 of rock was displaced along a sliding surface of roughly 42°. Displacement occurred across a front of about 255 m wide and up to 60 m deep. The initial landslide dropped 450 m over a travel distance of 600 m, into the cirque basin[2]. Run up on the far side of the basin reached 60 m. Rapidly moving debris spread out over the snow covered basin[3] to a width of about 370 m. Debris streamed along an average slope gradient of 8° for 450 m, prior to dropping out of the cirque basin into a steep confined channel. A considerable volume of rubble remained on the snow within the basin—some blocks are up to 10 m in size. Rubble entering the channel induced a debris flow. It maintained a width of about 100 to 120 m and manifested tremendous speed, evident by: super-elevated curves, run up, transport of airborne debris and mud splatter high on trees (Figure 8).

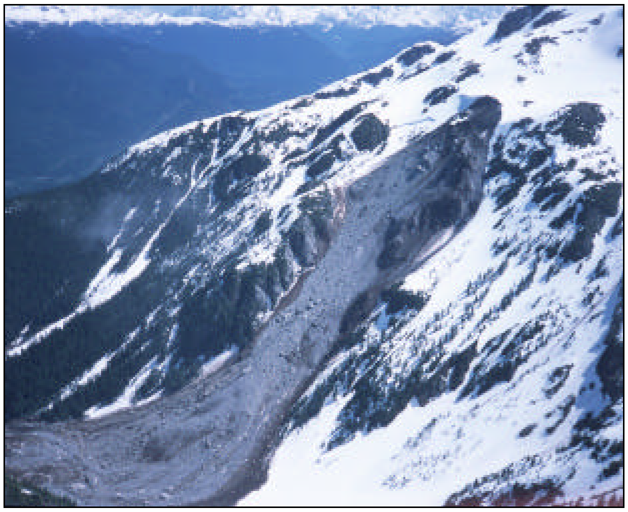

Figure 7. Zymoetz rock avalanche. Landslide length in the picture is about 1km. Note, snow in basin (June 8, 2002).

Along the lower reach, the debris flow travelled at high speed through a narrow canyon 300 m long (Figure 9). Debris exploded onto the Zymoetz floodplain, extending across the river a distance of 250 m, to a depth of 10 to 12 m. An estimated $0.5x10^6$ m^3 of debris, some rocks up to 7 m in diameter, dammed Zymoetz River. Water was backed up on the floodplain 1.5 km. An unknown volume of debris was transported down river. The dam was over-topped within 30 to 45 minutes. The river down cut a 5 x 60 m channel through the dam during high waters of June and July 2002.

The rock avalanche-debris flow travelled a distance of 4.3 km. The drop in elevation from the headscarp to the river is 1,255 m. The total volume entrained, transported and deposited through various zones approaches $1.6x10^6$ m^3. About 30% of the total volume is contained within the dam on the Zymoetz River (Figure 9).

Figure 8. Super-elevated curve, lower section, above the canyon (June, 2002).

Figure 9. Debris fan in Zymoetz River. Note, the narrow canyon and large boulders in river (October, 2002).

3.2 Geology

Bedrock near the crown is heavily jointed volcanic rock. Steeply dipping joints and active sliding surfaces were observed along the south east wall of the cirque basin. Dominant joint dip ranges from 40 to 60° degrees. Active movement of a toppling-sliding block within the crown area of the detachment zone was observed on a joint

---

[2] Slope, distance and volume estimates are derived by GIS analysis from a prepared pre and post 2 m grid.

[3] We estimated the snow depth in the basin at 3-5 m. This estimate is based on GIS measurements of pre and post surface, snow pillow data Tsai Creek station and our observations June 8 and 12, 2002. In addition, deep compact snow remained under debris all summer and was observed in October 2002.

dipping 45°. Joint dip, of 42 to 45° appears to coincide with the surface of rupture measured at 42°. The surface of rupture also appears to align with a polished joint surface 200 m down slope and with an older rupture surface immediately adjacent to the scarp. Rubble and blocks deposited in the basin are all volcanic in origin. The debris flow below the cirque basin collected large limestone boulders found in the stream channel that originated from the north west basin wall. The basin and stream channel down to the Zymoetz River appears to follow the contact between the volcanic rock and limestone.

3.3 Velocity estimate

Velocity calculations used the formulas as present by Cruden and Lu (1992). Run up observed in the upper basin was estimated by $(2gh)^{1/2}$ =34 m/s, where g is equal to the acceleration of gravity, and h is maximum vertical run up. Velocity was estimated at super elevated curves by $(RgtanP)^{½}$, where R is the radius of the curve in the bend and P is the tilt of flow surface at the bend (Figure 8). Preliminary velocity calculations provide estimates of 18 to 26 m/s along the debris flow. Mud splatter was observed at one location to have reached 13 m above the trim line, a total distance of about 33 m above the base of the stream channel.

3.4 Climatic conditions

The snow pack in the Bulkley and Babine Mountains in June 2002 was about 200% of normal. This was reflected in above normal precipitation and slightly below normal temperature in the preceding months. The Tsai Creek snow pillow station situated at 1360 m.a.s.l in the Telkwa River watershed, Bulkley Range, west of Smithers, recorded a snow water equivalent of 1909 mm on May 15. Precipitation and temperature conditions leading up to the Zymoetz landslide are presented in Figure 10. The days prior to the landslide were cool with snow falling at higher elevations. On the morning after the Zymoetz landslide, a rapid temperature rise is reflected in a rise in the Zymoetz River hydrograph (Figure 6).

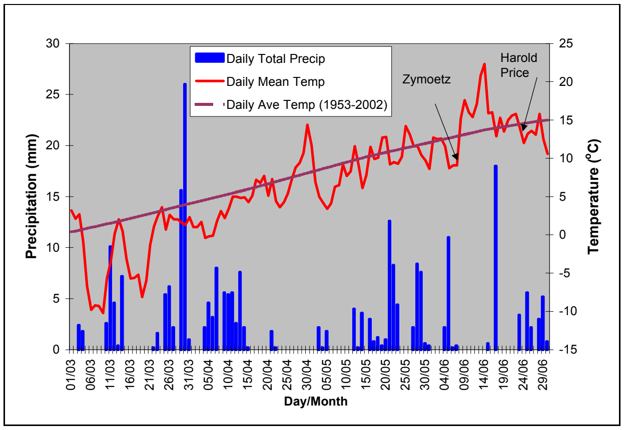

Figure 10. Precipitation and temperature, Terrace airport, March-June 2002.

4. HAROLD PRICE ROCK AVALANCHE

The Harold Price rock avalanche is believed to have occurred some time between June 22 to 24, 2002. It was first observed on the morning of 24 June, 2002 by a forestry crew doing helicopter reconnaissance in the area (Figure 11). It was raining hard at the time and high water flow was observed in streams on the morning of June 24. The rock avalanche originated at 1723 m.a.s.l. on the lip of a south west facing cirque occupied by a rock glacier. Interstitial ice was observed in the scarp face. The hillslope below the cirque basin is concave shaped—a result of valley glaciation. A deep till blanket covers the Harold Price valley below 1450 m.a.s.l.

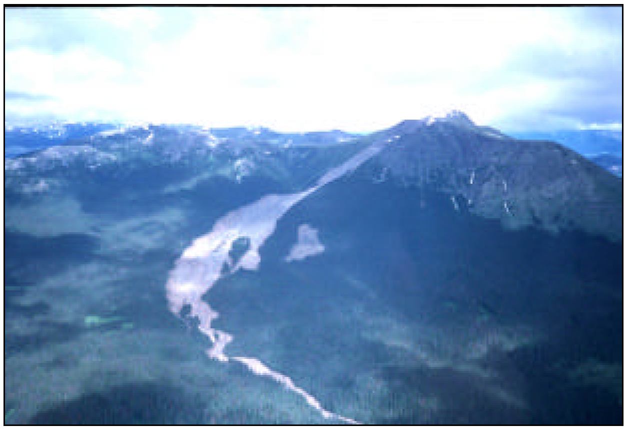

Figure 11. Harold Price rock avalanche-debris flow (June 28, 2002).

4.1 Description

Catastrophic failure occurred in deeply weathered and highly jointed volcanic bedrock/rubble and ice of the rock glacier above a steep concave slope. Volume[2] from the upper displacement zone is estimated at 0.7×10^6 m^3 across a scarp face of 175 m wide by 30 m deep (Figure 12). In addition, an 8 ha area of rubble within the basin exhibited post slide tension cracks and showed downward and lateral displacement. The estimated displaced volume of rock and ice that remained in the basin is in the order of 2×10^6 m^3. The rock avalanche dropped 300 m onto the open valley expanding to a width of about 360 m while moving at a rapid speed across a basal till surface. Unique features composed of disintegrating rock "molards" are present in the main landslide track extending out from the base of the mountain for about 850 m (Figure 13). The landslide, at 1.3 km, transformed into a debris flow transporting mainly forest debris. Debris entered an incised stream channel at 2.2 km and carried down to Harold Price Creek a further 1.8 km. Debris jammed over a 150 m stretch of Harold Price Creek and resulted in re-routing of the stream. A debris flood carried logs an additional 3.5 km down stream to a sharp corner leaving a debris pile inundation about 100 m onto the floodplain. Only a minor amount of rock and debris from the initial landslide appears to have been transported to Harold Price Creek. The main constituents of the channelized debris flow entering Harold Price Creek were trees and forest debris. The distance travelled by the rock avalanche-debris flow was 4 km to Harold Price creek with an elevation drop of 700 m—an overall slope gradient of 10°. Slope gradient in the depletion zone is about 34°, dropping to 17° in the middle zone occupied by “molards”, and gradually down to 7° in the zone of the channelized debris flow. The total displaced volume is about $1.6x10^6$ m^3.

4.2 Velocity Estimates

Preliminary calculations using the run-up and super-elevated curve equations as presented in Section 3.3 indicate velocity of 28 to 35 m/s as the rock avalanche reached the base of the mountain. Debris flow velocity dropped to less than 7 m/s in the channelized zone prior to entering Harold Price Creek.

4.3 Geology

Bedrock involved in the landslide is composed of highly weathered and heavily jointed volcanic rocks. The rocks show considerable hydrothermal alteration. Smaller rocks left along the main track were easily crumbled by hand or shattered with a rock hammer. We found at the base of the zone of depletion a hard volcanic rock, striated and polished by valley glaciation. We were only able to chip small fragments after repeated blows with a rock hammer. A deep till blanket in the valley is characteristic of Bulkley Valley till (Clague, 1984). Till samples collected contained 41.8% sand; 31.0% silt; and, 27.2% clay. Atterberg limits were: liquid limit 33.5; plastic limit 12.1; and plasticity index of 21.4. Bulk density for the till samples averaged 1.85 Mg/m^3.

Figure 12. Rock glacier, above the displacement zone.

4.4 Climatic conditions

A lesser amount of snow is generally present in the Babine Range than the Bulkley Ranges. However, snow depths were still on the order of 150 to 200% of normal for June (700 mm snow water equivalent, Chapman Lake snow survey station, 1460 m.a.s.l.). High discharge in local streams followed the rise to above normal temperatures recorded at the airport in Smithers. The same trend is evident at the Terrace airport (Figure 10). However, temperatures appear to have cooled prior to the landslide. Rain squalls were observed in the vicinity of Harold Price valley during the weekend of June 22-23, 2002.

Figure 13. Disintegrated rock, "molards" (August, 2002).

5. CONTRIBUTING FACTORS

No earthquakes were recorded or could have been felt in the Bulkley and Babine Ranges before or at the time of the Howson, Zymoetz and Harold Price rock avalanches (Personnel communication, Geological Survey of Canada, Sidney B.C., 2002).

5.1 Howson landslide

The thinning of the Howson glaciers has likely resulted in debuttressing and destabilization of the steeply jointed and near vertical rock slopes. Toppling features observed along ridge crests provide evidence of destabilization and indicate the most likely mechanism for the initial stage of landslide movement.

Above normal precipitation, recorded during the preceding summer months, may have created excessive joint water pressures. In addition, a greater incidence of freezing night temperatures—frost wedging, may have been the trigger.

5.2 Zymoetz landslide

Glacial erosion in the cirque appears to have resulted in the day lighting of a 42 to 45° joint surface below the rock promontory. The bedrock promontory situated at mid slope in the steep cirque basin is visible on pre-landslide air photographs. In addition, an older smaller landslide is evident, adjacent to the June 2002 landslide. Movement has probably been occurring along the joint for a considerable period of time in that the joint appears coincidental with the adjacent landslide surface.

The crown area was covered in a deeper than normal blanket of snow at the time of the landslide. Temperature was on a general rise, but it was cold and snowing on the days immediately preceding the landslide. How the above-normal snow pack, spring melt, warming and cooling temperatures contributed to failure is unknown. However, there may be a connection between the higher than normal precipitation, increased hydraulic pressure in joints and freeze-thaw at the site. We suspect the dense snow pack in the basin facilitated the rapid transport of debris through the basin into the incised channel. We also suspect, based on preliminary volume estimates, that there was probably a remobilization of old landslide debris from within the basin.

5.3 Harold Price

Rock glacier movement, with debris streaming over the lip of the cirque basin, is visible on the pre-landslide aerial photographs. In addition, debris from a previous landslide is evident on the adjacent slope. Initial movement may have occurred below the lip of the cirque, on a portion of the over steepened convex slope. This area of active movement is visible on air photographs. Degradation of interstitial ice may have played a role in catastrophic failure of the slope.

5.4 Climate warming

Periglacial features are evident in the crowns of these rock avalanches—ice in cracks, joints, fractures and interstitial ice in rubble. Mild winter temperatures and cooler summer temperatures, as experienced over the last few years in west Central B.C., may promote increased freeze-thaw cycles within the active layer of mountain permafrost. Slow, subtle changes may lead to catastrophic failure.

6. ECONOMIC COST

A dollar value is often not commonly attached to rock avalanches in remote mountain valleys. However, along the east-west utility and transportation corridor of west central B.C., costs attached to the landslides are significant. Direct and indirect costs are experienced by local industry and business. Direct costs pertain to lost production, lost revenue, and the repair to infrastructure. Indirect costs are more related to delays in production. Costs are easily determined for large industrial energy users. However, with small business, costs are not recognized in dollars but as an inconvenience. It therefore becomes extremely difficult to place an actual dollar value on the cost of a landslide. We were able to obtain general estimates from the gas utility and large industrial users of natural gas and from the forest industry. Conservative estimates place the direct cost attributed to Howson landslide at 10.3 million dollars and the indirect cost from lost production at 5 million. We estimate the direct costs connected to Zymoetz to be 5.9 million dollars and indirect costs, 27.5 million dollars. The timber volume, normally transported over the closed forest access road, dramatically increases the indirect costs. Two large plants, however, were not in production at the time of the Zymoetz landslide, hence direct costs could have been much higher. The Harold Price landslide resulted in the indirect loss of 1.6 million dollars of timber from an area proposed for harvest. In addition, forest site productivity lost within the landslide track would equal at least an equivalent dollars loss in timber production over one forest rotation.

7. DISCUSSION

Rock avalanches in mountainous terrain of west central British Columbia or in the Canadian Cordillera are not well documented. However, over the past few years the apparent number of events show a dramatic increase. In addition to the landslides we have described, three other high-elevation landslides have occurred in northern British Columbia since 1999 (Geertsema et al. this volume). Landslides in mountainous terrain are heavily influenced by climatic factors, including precipitation and temperature (Evans and Clague 1997). Catastrophic landslides triggered at high elevations may be particularly responsive to climate change, through warmer temperatures and increased precipitation. Evans and Clague (1994) suggest recent melting of mountain glaciers as an important process in debuttressing of rock slopes adjacent to glaciers. Mountain permafrost may be particularly sensitive to climate change in mid latitudes, as demonstrated in the European Alps (Etzemüller et al. 2001; Harris et al. 2001). Melting of alpine permafrost decreases the stability of mountain slopes (Davies et al. 2001; Harris et al. 2001). Recent large rock avalanches in the European Alps have been attributed to the degradation of mountain permafrost (Dramis et al. 1995; Bottino et al. 2002). While we have not studied permafrost parameters, degradation of mountain permafrost may also play an important role in initiating landslides in the Hazelton and Skeena mountains.

Rock avalanches from mountain/glacial environments pose a considerable hazard (Evans and Clague, 1999). In particular, they can involve large volumes of material, sustain high velocities, and travel distances of many kilometres. Rock avalanches, as have occurred recently, place utilities and transportation routes in the mountain valleys at risk. In addition, risks are now more apparent to forest workers and recreation users in the valleys. Land use planning in west central B.C. does not usually take into consideration hazards in mountain environments. The hazard is often a long way from development, and not recognized as a hazard or a risk. The tremendous down valley travel distance of these events suggests even communities may be potentially at risk. The Gitksan people of Hazelton B.C. in oral history describe "…some gigantic force coming down the valley…a giant grizzly bear...capable of uprooting trees…snapping of giant trunks…". The catastrophic event as described, is probably a landslide, ca 3500 BP (Gottesfeld et al. 1991). Oral history also describes "the bear" to have moved through the ancient village of Temlaham on the banks of the Skeena River, "…killing all who opposed him". This legend vividly describes what was likely a rock avalanche induced debris flow that originated high in the Chicago Creek basin near Hazelton B.C. It travelled a considerable distance to the Skeena River—possibly, 5 to 10 km.

The 1999 and 2002 rock avalanches in west central British Columbia occurred in remote mountain valleys, but, with the severing of roads and utilities, the economic impact to the local economy is enormous. We need to conduct detailed case studies to advance our understanding of damaging landslides, processes, and rates. In addition, we need to improve site identification to enhance our abilities to predict, mitigate and undertake quantitative risk analysis for potential damaging landslides. With a projected warmer and wetter future climate we can expect mountain environments to experience greater permafrost degradation, glacier thinning and more catastrophic rock avalanches. In short, landslide response to climate change also needs study.

8 ACKNOWLEDGEMENTS

Thanks to Drs. D.M. Cruden and J.K. Torrance for the informative in field discussions. Direct and indirect costs attributed to the landslides were provided by D. Towriss (PNG), G. Worbet (Methanex), S. Christansen (Eurocan), R. Ziegler (SCI) and D. Botten (BCFS). Matt Sakals helped with the landslide volume derivations.

9 REFERENCES

Bottino, G., Chiarle, M., Joly, A., & Mortara. G. 2002. Modelling rock avalanches and their relation to permafrost degradation in glacial environments. *Permafrost and Periglacial Processes* 13: 283-288.

Clague, J.J. 1984. *Quatenary geology and geomorphology, Smithers-Terrace-Prince Rupert area, British Columbia*. Geological Survey of Canada, memoir 413.

Cruden, D.M., & Lu, Z.Y. 1992. The rockslide and debris flow from Mount Cayley, B.C. in June 1984. *Canadian. Geotechnical Journal.* 29. 614-626.

Davies, M.C.R., Hamza, O., & Harris, C. 2001. The effect of rise in mean annual temperature on the stability of rock slopes containing ice-filled discontinuities. *Permafrost and Periglacial Processes* 12: 137-144.

Dramis, F., Govi, M., Guglielmin, M., & Mortara, G. 1995. Mountain permafrost and slope stability in the Italian Alps: the Val Pola landslide. *Permafrost and Periglacial Processes* 6: 73-82.

Eisbacher, G.H. 1979. Cliff collapse and rock avalanches (sturzstroms) in the Mackenzie mountains of northwestern Canada. *Canadian. Geotechnical Journal*.16,309-334

Eisbacher G.H. & Clague, J.J. 1984. *Destructive mass movements in high mountains: hazards and management*. Geological Survey of Canada, paper 84-16.

Etzemüller, B., Ødegård, R.S., Berthling, I, & Sollid, J.L. 2001. Terrain parameters and remote sensing data in the analysis of permafrost distribution and periglacial processes: principles and examples from southern Norway. *Permafrost and Periglacial Processes* 12: 79-92.

Evans, S.G. & Clague, J.J. 1994. Recent climatic change and catastrophic geomorphic processes in mountain environments. *Geomorphology* 10: 107-128.

Evans, S.G. & Clague, J.J. 1997. The impacts of climate change on catastrophic geomorphic processes in the mountains of British Columbia, Yukon and Alberta, in *Responding to Global Climate Change in British Columbia and Yukon, Vol. 1, Canada Country Study: Climate Impacts and Adaptation.* E. Taylor & B. Taylor (eds.). British Columbia Ministry of Environment, Lands and Parks and Environment Canada, Vancouver, BC pp 7-1 – 7-13.

Evans, S.G. & Clague, J.J. 1998. Rock avalanche from Mount Munday, Waddington Range, British Columbia, Canada. *Landslide News* 11: 23-25.

Evans, S.G. & Clague, J.J. 1999. Rock avalanches on glaciers in the Coast and St. Elias Mountains, British Columbia. *In Slope stability and landslides, Proceedings, 13th Annual Vancouver Geotechnical Society Symposium*, p. 115-123.

Evans, S.G. Clague, J.J., Wortsworth, G.J. & Hungr, O. 1989. The Pandemonium Creek rock avalanche, British Columbia. *Canadian Geotechnical Journal.* 26: 427-446

Geertsema, M., Evans, S.G., Cruden, D.M., Hungr, O., & Lu, Z. (this volume). *An overview of recent large landslides in northeastern British Columbia*.

Gottesfeld, A.S., Mathewes, R.W., & Johnson-Gottesfeld, L.M. 1991. Holecene debris flows and enviromental history, Hazelton area, British Columbia. *Canadian Journal of Earth Science*. 8:1583-1593

Harris, C., Davies, M.C.R., & Etzelmüller, B. 2001 The assessment of potential geotechnical hazards associated with mountain permafrost in a warming global climate. *Permafrost and Periglacial Processes* 12: 145-156.

Holland, S. S. 1976. *Landforms of British Columbia. A Physiographic Outline*. British Columbia Department of Mines and Petroleum Resources. Bulletin No. 48 Victoria, B.C., 138 pages.

Hungr, O., Evans, S.G., Bovis, M.J., & Hutchinson, J.N. 2001. A Review of the Classification of Landslides of the Flow Type. *Environmental and Engineering Geoscience,* Vol.VII, N0. 3 pp,221-238.