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Controls on Rockfall-Talus Process-Response Systems, Kananaskis, Canadian Rockies

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Abstract

Rockfall processes are important contributors to hillslope erosion in mountainous regions. Assessment of rockfall activity processes over regional scales has not been widely undertaken, and further assessment about controls of rockfall-talus processes is required. The present study investigates rockfall-talus processes over extended spatiotemporal scales in Kananaskis, Canadian Rockies utilizing aerial photographs, digital elevation models, geologic and paleoclimatic data. An inventory of talus deposits in fifth-order drainage basins was collected for Kananaskis, with rockfall erosion rates of 2.9 mm yr⁻¹. Results show a strong association of talus with faults and cirques. Although frost cracking is likely an important process leading to rockfall erosion, reconstructed temperature data suggests frost cracking alone is not the major determinant of locations of rockfall activity and talus deposition. Findings of this study provide important information on the contribution and controls of rockfall activity that can be utilized to better understand landscape evolution in mountainous regions.
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# Table of Contents

Abstract.......................................................................................................................................................... ii
Acknowledgements ........................................................................................................................................ iii
List of Tables ................................................................................................................................................ vii
List of Figures and Illustrations ........................................................................................................................ viii
List of Plates .................................................................................................................................................. x
List of Symbols ............................................................................................................................................. xi
List of Abbreviations ..................................................................................................................................... xii

CHAPTER 1: INTRODUCTION............................................................................................................................. 1
  1.1 Introduction.............................................................................................................................................. 1
  1.2 Research objectives ................................................................................................................................. 3
  1.3 Thesis organization ................................................................................................................................. 4

CHAPTER 2: LITERATURE REVIEW..................................................................................................................... 5
  2.1 Introduction.............................................................................................................................................. 5
  2.2 Talus slope characteristics ..................................................................................................................... 6
  2.3 Major factors controlling rockfall-talus process response systems ..................................................... 9
    2.3.1 Structural geology of the Canadian Rockies .................................................................................. 10
      2.3.1.1 The Front Ranges of the Canadian Rockies ...................................................................... 11
      2.3.1.2 Review of previous research about tectonic influence on rockfall .................................. 13
    2.3.2 Glacial imprint in the Front Ranges of the Canadian Rockies .................................................... 16
      2.3.2.1 Rockfalls and glacially steepened topography ................................................................. 17
    2.3.3 Influence of climatic and environmental factors in rockfall activity ........................................ 19
      2.3.3.1 Climate and rockfall activity ......................................................................................... 21

CHAPTER 3: STUDY AREA AND METHODOLOGY ............................................................................................ 24
  3.1 Study area .............................................................................................................................................. 24
  3.2 Data acquisition .................................................................................................................................... 27
    3.2.1 Aerial photographs and orthoimages ......................................................................................... 27
    3.2.2 Digital Elevation Model (DEM) ................................................................................................. 28
    3.2.3 Paleoclimate data ......................................................................................................................... 28
    3.2.4 Modern meteorological data ...................................................................................................... 29
  3.3 Method of analysis ................................................................................................................................ 30
    3.3.1 Talus identification for study basins .......................................................................................... 31
3.3.2 Talus surface area and volume calculation

3.3.3 Erosion rate calculation

3.3.4 Controlling factors of rockfall-talus activity

3.3.4.1 Structural Geology

3.3.4.2 Glacial Topography Analysis

3.3.4.3 Paleoclimate reconstruction

CHAPTER 4: RESULTS AND DISCUSSION

4.1 Talus slope inventory

4.2 Erosion rates due to rockfall

4.3 Control of structural geology and glaciation on rockfall-talus process

4.3.1 Results for control of structural geology and glaciation on talus deposits

4.3.2 Talus deposits associated with structural geology

4.3.3 Talus deposits and associated glacial topography

4.3.4 Statistical analysis of talus associated with structural geology and glacial topography

4.4 Frost-cracking and talus deposits

4.5 Significance of results to drainage basin development in Kananaskis, AB, Canadian Rockies

4.5.1 Glacial influence on topography

4.5.2 Postglacial erosion

4.5.3 Implications of results to drainage basin development

CHAPTER 5: CONCLUSION

5.1 Summary of findings

5.2 Summary of implication and future research

REFERENCES

APPENDIX
List of Tables

Table 3.1 Geological timescale with rock group and formation in fifth-order drainage basins, Kananaskis, AB .................................................. 25

Table 3.2 Temperature reconstruction of 12 ka BP for days (91 and 183) using model temperature ($T_{\Delta \Psi}$) and daily weather station temperature data ($T_{\Delta \Psi}$). $T_{\Delta \Psi}$ is the final reconstructed temperature........................................................................ 39

Table 4.1 Descriptive statistics for number of talus polygons in 11 study basins within Kananaskis watershed and its volume......................... 42

Table 4.2 Calculated erosion rates from talus surface area and volumes. ER (Actual) refers to the erosion rates calculated only from the area of steepland and ER (Normalized) represents erosion rates calculated by normalizing talus volume throughout basin area. Difference in these erosion rates displays the importance of considering only steepland as a rockfall source area............................................. 46

Table 4.3 Descriptive statistics for volume of talus polygons associated with different topography.............................................................. 65

Table 4.4 Nonparametric Kruskal-Wallis test (K-W) and Mann-Whitney U test (MW-U) among various topography. Topographical categories: (1) Cirque, (2) Fault/Folds in Cirque/U-shaped valley, (3) Fault, (4) Folds, (5) Headwater basin, (6) U-shaped valley......................... 66
List of Figures and Illustrations

Figure 1.1  Graphical representation of a talus slope  ..................................................  2

Figure 2.1  Geological provinces of the Southern Rocky Mountains.............  11

Figure 2.2  Cross-section showing the structure of the fold-and-thrust belt with indication of the main thrust faults. Simpson Pass draws a boarder on the western side between the Front Ranges and Main Ranges whereas McConnell thrust acts as an eastern border between the Front Ranges and Foothills.................................  13

Figure 2.3  Diagram illustrating cross-section view of fold hinge zone, limbs, and interlimb angle. The inflection points mark the limits of an individual fold which can be divided into a hinge zone and fold limbs. The interlimb angle is the angle between the lines tangent to the inflection points on the profile curve and this is used to indicate the tightness of a fold.........................................................  15

Figure 2.4  Idealization of freezing of cracked rock....................................................  20

Figure 2.5  Annual air-temperature variations based on daily temperature data for different elevations in central Southern Alps.........................  23

Figure 2.6  Annual temperature curve based on monthly temperature data in Little Lake, Oregon.................................................................  23

Figure 3.1  Kananaskis Watershed with eleven fifth order drainage Basins. The Kananaskis River flows from South to North eventually merging into the Bow River. A star sign adjacent to Barrier Lake represent the location of Kananaskis climate station 3053600 at University of Calgary’s Biogeoscience Institute. The inset shows the location of Kananaskis watershed with reference to Calgary.................................................................  26

Figure 3.2  Conceptual diagram of polygon volume tool in ArcMap.................  34

Figure 4.1  Talus slope delineation in fifth order drainage basins, Kananaskis Watershed. Each red polygon signifies talus deposits delineated based on aerial photographs, orthoimage, and hillshade DEM.....  41
Figure 4.2  Distribution of talus volume in drainage basins within Kananaskis watershed for A: Smith-Dorrien, B: Ribbon, C: Foche, D: Pocaterra, E: Three-Isle, F: Upper-Kananaskis, G: Evan-Thomas, H: Rocky, and I: Boulton. Note: Porcupine and Marmot basins are excluded due to the minimum number of talus polygons. All histograms represent a positively skewed distribution.  

Figure 4.3  Location of cirques (Cyan colour polygons) present within fifth order drainage basins of Kananaskis watershed, Front Ranges of the Canadian Rockies. The elevation greater than 2100 masl is represented in brown colour, whereas elevation less than 2100 masl is represented in grey. Red dashed lines across the watershed represent folds and maroon dashed lines represent faults. 

Figure 4.4  Talus slopes with associated topography such as folds (yellow), fault (green), cirque (cyan), U-shaped valley (orange), headwater basin (blue), and cirque/fault* (magenta) identified within eleven fifth order drainage basins on a hillshade DEM, Kananaskis. Cirque/fault* represents topography with fault/fold in cirque/U-shaped valley. Drainage basins include A: Ribbon, B: Rocky, C: Smith-Dorrien, D: Evan-Thomas, E: Three-Isle F: Boulton, G: Upper-Kananaskis, H: Foche, I: Pocaterra, J: Marmot, and K: Porcupine. 

Figure 4.5  Volume of talus slope associated with topographic features such as cirque, U-shaped valley, faults/folds present at cirque/U-shaped valley, fault, folds, and headwater basins. Each chart represent different fifth order drainage basin within Kananaskis watershed such as A: Ribbon, B: Rocky, C: Smith-Dorrien, D: Evan-Thomas, E: Three-Isle F: Boulton, G: Upper-Kananaskis, H: Foche, I: Pocaterra, J: Kananaskis watershed (All basins included). 

Figure 4.6  Variation in elevation of cirque and talus deposits present at eleven fifth order drainage basins in Kananaskis watershed. Solid black line represent mean cirque elevation and dotted black line shows mean talus slope elevation. Note relatively consistent elevation of talus slope slightly below the cirque elevation. 

Figure 4.7  Mean daily air-temperature variations at different elevations in Kananaskis watershed. Using the lapse rate of 4.4°C km⁻¹, temperature is produced at 500 m intervals i.e. 1391 m, 1891 m, 2391 m, 2891 m, and 3391 m. Temperature data is reconstructed for 12 (A), 10 (B), 8 (C), 6 (D), 4 (E), 2 (F), and 0 ka BP (G). Blue shaded area represents frost-cracking window (FCW) (-3 to -8 °C).
**List of Plates**

| Plate 2.1 | Talus cone developed below chutes that act to funnel the debris downwards in Evan-Thomas Drainage Basin, Kananaskis Watershed | 7 |
| Plate 2.2 | Compound talus slope alongside a rockwall formed by the lateral growth of talus cones in Pocaterra Drainage Basin, Kananaskis Watershed | 8 |
| Plate 2.3 | Talus sheets developed mainly by rockfall with uniform debris supply along rockwalls in Upper Kananaskis Drainage Basin, Kananaskis Watershed | 8 |
| Plate 3.1 | A sample of an orthophoto (0.2 m × 0.2 m resolution) showing talus slopes (red polygons). Black line indicates boundary of Ribbon drainage basin | 32 |
| Plate 3.2 | Cirque identification from orthophotographs and topographic maps at Smith-Dorrien drainage basin. Blue line represent the boundary of a cirque. Lake present at the cirque floor and talus slopes at the base of a steep rockwall makes it more obvious to identify | 37 |
| Plate 3.3 | U-shaped valley identification from an orthophotograph at Smith-Dorrien drainage basin. The orange dashed line represents the U-shape of a valley and red star is the location of another cirque. Also, notice talus slopes formed along the U-shaped valley bottom | 37 |
| Plate 4.1 | Talus slopes associated with fault and cirque/fault* in Pocaterra drainage basin. Cirque/fault* represents topography with fault/fold in cirque/U-shaped valley. Yellow dashed line shows the fault line. Red and yellow dashed line shows the outline of a cirque and fault | 57 |
### List of Symbols

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>°C</td>
<td>degree Celsius</td>
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<tr>
<td>θ</td>
<td>slope angle</td>
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<tr>
<td>°</td>
<td>degree</td>
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<tr>
<td>ρ_s</td>
<td>density of sediment</td>
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<tr>
<td>ρ_b</td>
<td>density of bedrock</td>
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<tr>
<td>~</td>
<td>approximately</td>
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<td>±</td>
<td>range</td>
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<td>%</td>
<td>percentage</td>
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<td>&gt;</td>
<td>greater than</td>
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<td>≥</td>
<td>greater than or equal to</td>
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<tr>
<td>α</td>
<td>level of significance</td>
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<tr>
<td>Σ</td>
<td>summation</td>
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<tr>
<td>T_{ΔΨ_i}</td>
<td>model temperature variation</td>
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<tr>
<td>T_{Ψ_i(t_0)}</td>
<td>paleoclimate model data for modern day</td>
</tr>
<tr>
<td>T_{Ψ_i(t_n)}</td>
<td>paleoclimate model data for n period of interest</td>
</tr>
<tr>
<td>T_{ΔW_{sj}}</td>
<td>daily temperature differences (reconstructed temperature)</td>
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<tr>
<td>T_{W_{dj}}</td>
<td>weather station temperature dataset</td>
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## List of Abbreviations

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<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>AEP</td>
<td>Alberta Environment and Parks</td>
</tr>
<tr>
<td>ANOVA</td>
<td>Analysis of Variance</td>
</tr>
<tr>
<td>BP</td>
<td>Before Present</td>
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<tr>
<td>CCSM3</td>
<td>Community Climate System Model Version 3</td>
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<tr>
<td>CHILD</td>
<td>Channel-Hilslope Integrated Landscape Development Model</td>
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<td>cm</td>
<td>Centimeter</td>
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<td>DEM</td>
<td>Digital Elevation Model</td>
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<tr>
<td>ER</td>
<td>Erosion Rate</td>
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<td>FCW</td>
<td>Frost Cracking Window</td>
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<td>GENESIS</td>
<td>Generic Enterprise Spatial Information Services</td>
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<tr>
<td>GIS</td>
<td>Geographical Information System</td>
</tr>
<tr>
<td>ka</td>
<td>Kilo-annum (thousand years)</td>
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<td>kg</td>
<td>Kilogram</td>
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<tr>
<td>km</td>
<td>Kilometer</td>
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<tr>
<td>K-W</td>
<td>Kruskal-Wallis</td>
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<tr>
<td>Ma</td>
<td>Mega-annum (million years)</td>
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<tr>
<td>Masl</td>
<td>Meters above sea level</td>
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<tr>
<td>MAT</td>
<td>Mean Annual Temperature</td>
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<td>m</td>
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<td>mm</td>
<td>Millimeter</td>
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<td>NAD</td>
<td>North American Datum</td>
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<td>NCAR</td>
<td>National Center for Atmospheric Research</td>
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<td>NTS</td>
<td>National Topographic System</td>
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<tr>
<td>SANDS</td>
<td>Spatial and Numeric Data Services</td>
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<tr>
<td>SPSS</td>
<td>Statistical Package for the Social Sciences</td>
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<tr>
<td>TIN</td>
<td>Triangular Irregular Networks</td>
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<tr>
<td>WMS</td>
<td>Web Map Service</td>
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xii
CHAPTER 1: INTRODUCTION

1.1 Introduction

Rockfall is one of the major erosional processes in the alpine environment (Ballantyne, 2002; Moore et al., 2009) and has a significant role in the total denudation of mountain slopes (Matsuoka & Sakai, 1999). The accumulation of rock debris falling from a cliff and depositing at its base is called a rockfall talus slope (Figure 1.1) (Luckman, 2007). This thesis considers rockfall-talus systems in Kananaskis AB in the Canadian Rockies. Rockfalls are recognized as being an important geomorphic process in mountain regions: (i) rockfalls are major contributors to hillslope erosion and drainage basin development; (ii) rockfalls represent a significant natural hazard; (iii) rockfalls and their locations of deposition affect sediment connectivity between hillslopes and channel networks (Hoffmann et al., 2013); and (iv) rockfalls have distinct morphological and sedimentological properties that affect the development of weathered layers in the landscape; (v) talus affects mountain hydrology, in particular the storage and transmission of water (Muir et al., 2011). For these reasons, it is important to obtain a better understanding of rockfall-talus process-response systems.

Although studies about present-day rockfall activity provide important information about process operation, rates of rockfall activity have likely not been uniform throughout the Holocene. Glacial activity in steep mountain ranges often generates oversteepened rock slopes that respond by initially having very high rates of rockfall activity resulting in the development of paraglacial talus accumulations below cliffs (Augustinus, 1992; Ballantyne, 2002). Several factors are also expected to have had a significant impact on
rockfall-talus processes. Such factors include the presence of steep rock slopes bedrock (related to the structural geology of the mountains as well as glacial sculpting) and the weathering and development of joints leading to processes such as frost cracking (Luckman & Fiske, 1997; Selby, 1982; Matsuoka & Sakai, 1999; Hales & Roering, 2005). Other hillslope processes such as rock slides, debris slides and debris flows have generally been the subject of more investigations (Anderson & Anderson, 2010; Allen, Cox, & Owens, 2011) compared to the relatively low-magnitude, high-frequency process of rockfall that produces talus slopes (Gardner et al., 1983; Hales & Roering, 2005).

Figure 1.1 Graphical representation of a talus slope. (Redrawn from Obanawa & Matsukura (2006))

Most previous studies of rockfall-talus have focused on local events (e.g. Gardner et al., 1983), with a particular focus on rockfalls occurring due to glacial unloading (Ballantyne & Eckford, 1984; Augustinus, 1995; Hinchliffe & Ballantyne, 1999; Curry & Morris, 2004). Rockfall studies have been undertaken in different locations, such as the Scottish Highlands (Hinchliffe & Ballantyne, 1999), Bavarian Alps (Sass & Wollny, 2001), and the Canadian Rockies (Luckman, 1981; Gardner et al., 1983). However, investigations of rockfall-talus process-response system studies over large spatial scales
(i.e. scale of mountain ranges) have not been widely undertaken. In one of the most comprehensive regional rockfall studies, Hales and Roering (2005) conducted a study of talus processes in the Southern Alps, New Zealand covering an area of 3200 km². Despite the abundance of talus in alpine environments of the Canadian Rockies, studies investigating rockfall-talus processes and the collection of large talus inventories over large spatial scales have not been undertaken in the Canadian Rockies. The current research seeks to address this gap in knowledge by investigating rockfall-talus processes-response systems over large spatial scales in the Canadian Rockies.

1.2 Research objectives

The purpose of this research is to collect and analyze talus inventories in fifth order drainage basins within the main Kananaskis River drainage basin using aerial photographs and digital elevation models (DEMs) to better understand rockfall-talus process-response systems in the Front Ranges of the Canadian Rockies. The primary objectives of this research include:

i) To collect large inventories of talus deposits using aerial photographs for the Front Ranges in the Canadian Rockies covering ~500 km².

ii) To investigate rockfall-talus processes by analyzing the association of these talus inventories with glacial topography and geological features.

iii) To analyze climatic control on frost cracking and its possible role in determining locations of rockfall-talus process in the Canadian Rockies.

iv) To estimate rockfall erosion rates and their contribution to drainage basin and mountain development in the Canadian Rockies.
1.3 Thesis organization

This thesis is divided into five chapters. Chapter 1 provides the significance of the study along with the research gaps and objectives. Chapter 2 presents a review of the literature on rockfall-talus processes. This includes a discussion about the role of structural geology, glaciation, and climatic conditions on rockfall activity and talus deposition. Chapter 3 introduces the study area and methodologies used for calculating rockfall erosion rates and analyzing the possible factors affecting rockfall-talus systems. Chapter 4 describes and discusses the research results and implications. Chapter 5 contains a final summary of the findings for this study.
CHAPTER 2: LITERATURE REVIEW

2.1 Introduction

The Canadian Rocky Mountains were formed by plate tectonic collision and uplift during the Laramide Orogeny (80 to 55 Ma) producing landscapes of high-elevation, yet fairly low relief (Gardner et al., 1983; English & Johnston, 2004; Osborn et al., 2006; Trenhaile, 2010; Hoffman et al., 2013). Substantial evidence shows that several kilometers of overburden have been removed from the Canadian Rockies since the end of the Laramide Orogeny (approximately 60 Ma) (Osborn et al., 2006; Trenhaile, 2010). The present-day topography of the Canadian Rockies represents ongoing erosion since the end of the Laramide Orogeny. During this time, the most resistant rocks have generally remained at higher altitudes, while the less resistant rocks have been displaced to lower-gradient portions of the landscape (Trenhaile, 2010). High-elevation, steep mountain slopes are a major source of sediment supply in the Canadian Rockies (Hoffman et al., 2013). In addition to tectonics, glacial activity has also had a significant role in generating oversteepened rockwalls. The Canadian Rockies have undergone several episodes of glacial advance. During the Late Wisconsinan, both Laurentide and Cordilleran ice sheets were extensive, and their margins overlapped along the Rocky Mountain Foothills in north central British Columbia (Bednarski & Smith, 2007; Menounos et al., 2009). Multiple advancements of local valley and cirque glaciers occurred between 15 and 11 ka (Liverman et al., 1989; Menounos et al., 2009). Similarly, glacier expansion during the Holocene (beginning after ~ 11 ka) was punctuated by several advances and retreats (Menounos et al., 2009).
Sediment flux rates are expected to have generally been higher in the immediate years following deglaciation of the Canadian Rocky Mountain beginning in the Late Pleistocene/Early Holocene, with the occurrence of considerable rock slides and rockfall events (e.g. Luckman, 1981; Gardner et al., 1983; Hoffmann et al., 2013). These rock slides and rockfall events are expected to have continued to operate throughout the Holocene, although it is generally thought that these processes would be most active directly after glaciation. The term ‘paraglacial’ was introduced by Ryder (1971) and refers to ‘nonglacial processes that are directly conditioned by glaciation’ (Ballantyne, 2002). In this case, the oversteepening of slopes by glaciation is expected to have led to high rates of rock slide and rockfall activity. The ongoing operation of these mass wasting processes has resulted in the formation of numerous talus slopes at the foot of steep slopes throughout the Canadian Rockies.

2.2 Talus slope characteristics

Talus slopes (or scree slopes) are developed by the weathering and subsequent fall of rock debris from a cliff that forms an accumulation (deposit). The deposits typically have slope gradients in the range of approximately 30-40° and a concave-upwards profile (Rapp, 1960; Kirkby & Statham, 1975; Church, et al., 1979). Talus slopes undergo sorting mechanisms during deposition whereby smaller particles are retained at the top of the deposit and larger boulders are found at the base. Typically, rockfall volumes do not generally exceed approximately 100 m³ (Dussauge et al., 2003; Hales & Roering, 2005). Talus deposits may display numerous possible shapes and morphology (Rapp, 1960). Talus cones (Plate 2.1) occur at the outlet of a rock chute or a mountain gully along major joints or other rock weaknesses that localize rock delivery (Rapp, 1960; Church et al., 1979).
Chutes restrict the movement of water and snow, as well as falling rock, such that the transfer mechanisms are more variable than for talus sheets (see below; Church et al., 1979). *Compound talus slopes* (Plate 2.2) form when a series of talus cone coalesce laterally, and *talus sheets* (Plate 2.3) form when falling rocks accumulate at the base of a continuous rock face usually having similar geology (Rapp, 1960; Evans, 1976).

Plate 2.1 Talus cone developed below chutes that act to funnel the debris downwards in Evan-Thomas Drainage Basin, Kananaskis Watershed (Image Source: AEP GENESIS (2013))
Plate 2.2 Compound talus slope alongside a rockwall formed by the lateral growth of talus cones in Pocaterra Drainage Basin, Kananaskis Watershed (Image Source: AEP GENESIS (2013))

Plate 2.3 Talus sheets developed mainly by rockfall with uniform debris supply along rockwalls in Upper Kananaskis Drainage Basin, Kananaskis Watershed (Image Source: AEP GENESIS (2013))
Rockfall-dominated features, such as talus slopes, provide information about the role of rock erosion in the landscape development of mountain ranges (Church et al., 1979; Curry & Morris, 2004). Rockfall-talus systems are expected to have been a particularly significant component of hillslope sediment transfer processes during the paraglacial and are expected to have maintained an important role throughout the Holocene up until the present day. While some previous studies have quantified rockwall retreat based on the talus volume (e.g. Andre, 1997; Curry & Morris, 2004; Hales & Roering, 2005; Moore et al., 2009; Hoffmann et al., 2013) and other studies have investigated the mechanical processes underlying rockfall events, only a limited number of studies have investigated factors controlling rockfall-talus processes at regional spatial scales and over longer temporal scales. Section 2.3 provides a review of possible factors that control rockfall erosion and associated talus formation at these scales.

2.3 Major factors controlling rockfall-talus process response systems

Hillslopes and channels are the basic landform constituents of morphological systems in drainage basins. The various components of a morphological system are linked by processes. Together, these landforms and processes interact as a process-response system (Charlton, 2008). Rockfall-talus process-response systems are typically associated with: (i) steep topography with abundant erodible and jointed bedrock (both tectonic activity and glaciation are important in creating these conditions); and (ii) climatic conditions that lead to further joint enlargement and frost cracking (Selby, 1982).
2.3.1 Structural geology of the Canadian Rockies

Tectonic activity not only increases the relief but as folds develop, hillslope gradients may also increase and faulting may produce growing scarps (Selby, 1982). Rockfall activity is expected to be significant in tectonically-active belts in rising mountain chains (Kingdom-Ward, 1955), but may also be significant in tectonically inactive mountain range. The Laramide Orogeny in Canada refers to thin-skinned fold-and-thrust deformation (Figure 2.1) that is Late-Cretaceous to Early-Tertiary in age, and involves cratonic/platformal rocks (Osborn et al., 2006). The Southern Canadian Rockies are commonly divided into the following sub-provinces: Front Ranges, Main Ranges, Western ranges, and Foothills (Bally et al., 1966; North & Henderson, 1954; Osborn et al., 2006). During the Laramide, relatively widely-spaced thrust faults propagated eastward, causing large displacement in what became the Main and Front Ranges (Figure 2.1).
2.3.1.1 The Front Ranges of the Canadian Rockies

For the most part, the Front Ranges comprise a series of steep, sub-parallel, westward-dipping thrust sheets, stacked in an imbricate trend between the Foothills and the Main Ranges of the Canadian Rockies (Figure 2.2) (North & Henderson, 1954; Bally et al., 1966). At the end of the Laramide orogeny (~60 Ma), the present Front Ranges were
almost or entirely covered with Mesozoic rock, with low-to-moderate local relief and the dominant vegetation resembling that of present Foothills. Due to erosion, the removal of Mesozoic rocks exposed the Paleozoic and Proterozoic rocks that could support the steep and high slopes (Osborn et al., 2006). Evidence from fission-track analysis shows that deformation within the Rocky Mountain Fold and Thrust Belt continued for a few million years, past the 60 Ma age (Kalkreuth & McMechan, 1996). Further deformation and continued incision of rivers into the resistant bedrock thus created high local relief thereafter (Osborn et al., 2006).

The present day Front Ranges display a combination of structural geology features. The most easterly portion of the Front Ranges tends to resemble the Foothills in structure. For example, the McConnell Fault resembles the features of the Foothills with highly-folded rocks, and is sinuous in outline. The most westerly portion of the Front Ranges has some structural and stratigraphic features in common with the Main Ranges; for example, folds dominate the structural style in the Bourgeau thrust sheet where Upper Devonian group strata have changed facies into shale, and thick sections of cleaved Ordovician strata are preserved (McMechan, 2012a). The major thrust faults in the Front Range include the Bourgeau, Sulphur Mountain, Rundle, and McConnell thrusts carrying Paleozoic carbonates, with Mesozoic clastic strata dominating the structural geology (Figure 2.2) (Vandeginste et al., 2012).
Figure 2.2 Cross-section showing the structure of the fold-and-thrust belt with indication of the main thrust faults. Simpson Pass draws a border on the western side between the Front Ranges and Main Ranges whereas McConnell thrust acts as an eastern border between the Front Ranges and Foothills (Price & Fermor, 1985, modified by Vanbegeinste et al., 2012)

Steep, exposed bedrock is a requirement for rockfall occurrence. During the formation of the Canadian Rockies, the development of steep local relief and increased bedrock exposure resulted from: (i) the exposure of carbonate rock, which is less favorable for tree growth than Mesozoic clastic rocks (Osborn et al., 2006), and (ii) deteriorating climate, resulting in a lowering of the treeline (Zachos et al., 2001). Although not focusing on the Canadian Rockies, previous research has investigated the role of tectonic activity on rockfall activity; this is discussed in the following section.

2.3.1.2 Review of previous research about tectonic influence on rockfall

The role of tectonics on rockfall susceptibility has been reviewed by several authors (Selby, 1982; Miller & Dunne, 1996; Coe & Harp, 2007). Bedrock fractures at Earth’s surface are abundant and range in scale from microcracks to local joint sets, to continent-transversing fault zones. Topography is one of several factors that affect the formation of fractures. Fracturing occurs when induced stress is greater than rock strength. There is greater potential for fracturing along ridges or valleys because topography induces
tensional and compressional stresses along valley floors and ridge tops (Miller & Dunne, 1996). Wise (1964) investigated microjointing in the Middle Rocky Mountains of Montana and Wyoming. Based on field observations, it was found that microjoints appear to be a late Laramide feature due to their continuous vertical development in almost all mountain ranges (Wise, 1964). Here, the author interpreted microjoints as expansion features in Laramide block mountains, created as the blocks were lifted free of the confines of the adjacent basin floors. Miller and Dunne (1996) demonstrated topographic effects on bedrock fracturing by using a simple elastic model including regional and gravity-induced stresses. The model indicated that topographic relief can cause stresses of sufficient magnitude to break rock. Elastic stresses were found to be directly proportional to relief. The higher the ridge or deeper the valley, the greater is the potential for fracturing, thus accelerating erosional rates. As a result, fracture sets have their spatial distribution and orientation governed by landform shape, and regional distributions of stress. For example, valleys in areas of compressional tectonics will have fractures that are nearly parallel to side slopes and that favor mass movement along slip surfaces. Tension induced along ridge crests results in locations favorable to opening-mode fractures oriented normal to the ground surface (Miller & Dunne, 1996).

The interaction between relief, fracturing, and erosion shows a negative feedback mechanism along ridges; once the relief threshold is reached, fracturing in the ridge promotes mass wasting that reduces total relief. On the other hand, the relationship between relief, fracturing, and erosion for valleys has a positive feedback mechanism because an increase in stress and incision promotes fracturing. This then accelerates incision rates and reaches the point that further valley deepening becomes transport limited. It is important
to recognize these feedbacks because they are important in drainage basin development and the evolution of longitudinal river profiles (Miller & Dunne, 1996).

A study of rockfall susceptibility in tectonically folded strata was conducted in the fold-thrust belt exposed in American Fork Canyon, north-central Utah (Coe & Harp, 2007). Analysis of large-scale geologic mapping (using aerial photographs), talus production (based on field observations), rock-mass field measurements, and a compilation of historical rockfall data, led to the conclusion that rockfall susceptibility depends on the dip of fold limbs and the curvature of fold hinge zones (Coe & Harp, 2007). Rockfall susceptibility is greater in fold hinge zones than on adjacent limbs, and it increases as fold curvature tightens, whereas susceptibility is proportional to dip on fold limbs (Figure 2.3). This result was supported by Cooke (1997) who articulated that joints are restricted to regions of high curvature within the outer arc of the fold, or for regions of interlayer slip along fold limbs.

![Figure 2.3 Diagram illustrating cross-section view of fold hinge zone, limbs, and interlimb angle. The inflection points mark the limits of an individual fold which can be divided into a hinge zone and fold limbs. The interlimb angle is the angle between the lines tangent to the inflection points on the profile curve and this is used to indicate the tightness of a fold (after Price & Cosgrove, 1990; Coe & Harp, 2007)](image)
Faults are another feature of structural geology that represent locations that are prone to rockfall occurrence. Butler et al. (1986) found that large rockfall avalanches were associated with thrust faults in Northeastern Glacier National Park. Melzner et al. (2013) found that in Eastern Alps of Austria, cliffs and scarps are associated with tectonic structures such as faults, and that these locations are also associated with high-volume rockfalls.

2.3.2 Glacial imprint in the Front Ranges of the Canadian Rockies

The historical legacy of climatic variation from the Pleistocene and Holocene resulted in episodes of glacial advance and retreat (Luckman, 2000; Ballantyne, 2002; Hoffman et al., 2013). The Canadian Rockies have been encroached upon three different ice advances during the Quaternary: (i) as a western limit to the Laurentide Ice Sheet (that advanced towards the Foothills during the Late Wisconsinan) (Liverman et al., 1989); (ii) as an eastern margin of the Cordilleran Ice Sheet (that covered the entire Main Ranges in British Columbia during the Wisconsinan); and (iii) as a center of local Montane ice (Bobrowsky & Rutter, 1992). More specifically, the Front Ranges were overridden by the Bow glacier (an outlet glacier of the Cordilleran Ice Sheet, CIS), the Kananaskis valley glacier, and local cirque glaciers during the last glacial maximum (Hoffman et. al., 2013).

Harrison (1976) reported that the last major advances in Kananaskis valley south of the Bow River predated 12 ka BP. This is because the lower Elk Valley, which shares a common ice source with the Kananaskis Valley, has been free of ice since 12 200 ± 160 yr BP. The paraglacial period (beginning after deglaciation ~12.5 ka BP) (Luckman, 1979) represents one of the most significant time periods in the geomorphological development
of the Canadian Rockies. An important geomorphological consequence of deglaciation in mountainous environments is the exposure of glacially oversteepened rockwalls, and the abundance of glacial deposits across great portions of the landscape. Ice retreat removes the support of rock walls, thereby altering the state of stress within the rock mass (Ballantyne, 2002). The adjustment processes in response to these conditions includes the collapse of steep upper rockwalls by rock slides and rockfalls. Furthermore, there is a notable increase in sediment transport rates by debris slides and debris flows given the large extent of unconsolidated, surficial glacial deposits (Dadson & Church, 2005).

Glacial features such as cirques and U-shaped valleys can preserve the fallen rock debris for long periods of time (i.e., the material goes into longer-term storage. Based on estimates of talus accumulation, Gray (1971) found that rockwall retreat rates are greater in cirques compared to undissected mountain walls. The Southern Canadian Cordillera has abundant cirques with strongest cirque development in the north and northeastern sides of mountains (Trenhaile, 1976). Further research is required to better understand the significance of glacially-oversteepened features in rockfall occurrence over both local and regional scales in the Canadian Cordillera.

2.3.2.1 Rockfalls and glacially steepened topography

Glacial erosion sculpted mountains during the Pleistocene that resulted in valley widening and overdeepening, as well as the formation of glacial cirques, U-shaped valleys, and widespread glacial deposition (e.g. Sauchyn & Gardner, 1983; Brocklehurst & Whipple, 2002; Moore et al., 2009; Hoffman et al., 2013). A question of interest remains
is whether rockfall occurrence following deglaciation is paraglacially or periglacially dominated.

Previous studies have documented the paraglacial occurrence of increased rockfall activity. Rockfall assessment on a French cirque was conducted by Gellatly and Parkinson (1994). They found that rockfalls within the study area were triggered directly by Late Holocene ice wastage and exposure of cliffs after deglaciation rather than periglacial environment. Hoffman et al. (2013) also documented paraglacial adjustment of rock slopes in Kananaskis, AB, Canadian Rockies (the region studied for the present thesis). Similarly, rapid rockwall degradation and the formation of supraglacial talus accumulations on adjacent glaciers have been recorded in recent retreat of glaciers associated with Mexican volcanoes (Palacios, 1998; Palacios & De Marcos, 1998). Hales and Roering (2005) reported that 10-20% of the Southern Alps, New Zealand are mantled with talus along U-shaped valleys, suggesting that rockfall activity is connected to glacial topographic features, and that these glacial valleys have high preservation potential. With reference to examples of rockfalls in response to deglaciation in a variety of different environments, none of which is subject to particularly severe periglacial conditions, Ballantyne (2002) suggested that paraglacial rock slope instability is a dominant influence on postglacial rockfall occurrence. The exposure of oversteepened slopes during deglaciation leads to increased rockfall activity that may last for thousands of years. That being said, frost-induced weathering may also be an important factor contributing to increased rockfall activity, and it is to this that we now turn our attention.
2.3.3 Influence of climatic and environmental factors in rockfall activity

Rapp (1960) analyzed air temperature, precipitation and frost growth patterns regarding mountain walls and talus slope formation in Spitsbergen, Norway. Luckman (1976) later supported climate as being a primary control of rockfall trigger mechanisms, due to its control over temperature as well as the availability of water. These factors are important in frost cracking, which has been attributed to having a key role in rockfall activity (Hales and Roering, 2005). Anderson (1998) defines frost cracking as “the slow growth of cracks within a rock mass by the crystallization of ice within existing crack”. Walder and Hallet (1985) theorized the analogy between the growth of cracks in freezing rocks and the generation of ice lenses in freezing soils (Figure 2.4). Water moves from the zone of unfrozen rock through a set of thin films (less than tens of nm) along mineral surfaces within the frozen fringe toward the existing cracks in a colder zone, where it favorably freezes (Walder and Hallet, 1985; Anderson, 1998). The water films separating ice from mineral surfaces exert an attractive force on water in pores to which it is hydraulically connected, as well as a "disjoining pressure" that tends to separate ice from the substrate (Gilpin, 1979; Walder & Hallet, 1985). The stresses induced by the presence of new water in the crack determine the growth rate of the crack. Such growth of cracks is very low at extreme minimum temperatures, because the viscosity of water increases and the thickness of the water films declines within the frozen fringe (Anderson, 1998).
Therefore, maximum frost cracking is generally thought to be at the temperature range of -3 to -8 °C, which is termed the “frost cracking window” by Anderson (1998). The rate of frost cracking depends on: (i) the proportion of time that rock is at temperatures between -3 and -8 °C; and (ii) the amount of available water (Hales & Roering, 2005). Water expands by ~9% to 11% of its volume when it freezes and, thereby, reduces the strength of the rock (Matsuoka, 2001; Anderson & Anderson, 2010). Frost damage to rocks results in the break-up of bedrock and large rock fragments due to the freezing of water in large pores. The widening or lengthening of pre-existing macrocracks that experience freezing of the crack-filling water is caused by macrogelivation. This process is significant in producing rock fragments on the cliff face (Rapp, 1960; Church et al., 1979; Matsuoka & Sakai, 1999).

Several authors have also noticed that frost cracking processes are dependent on environmental conditions such as aspect and seasons (Luckman, 1976; Church et al., 1979; Sass, 2005). Field observations in the Japanese Alps found that frost action was a primary
cause of the frost shattering, although the rate of shattering was higher in winter rather than in summer (Matsuoka, 1990b). This result contradicts the study conducted in the Bavarian Alps (Sass, 2005) where higher rockfall rates occurred in the spring season. The reasons for low rockfall rates in winter was also discussed. First, during the continuous winter frost in northerly aspects, jointed rock fragments are secured with ice. This stimulates the formation of segregation ice and the cracking of inherently sound rock. Second, moisture is essential for frost cracking to advance. During winter, there is limited water availability near the surface (Sass, 2005; Hales & Roering, 2005). Despite discrepancies regarding the seasonal effect on frost-cracking for different locations, researchers have found a strong correlation of higher rockfall activity on SE-facing rockwalls than for west and NE-facing rockwalls (Luckman, 1976; Matsuoka, 1990b; Sass, 2005). Other topographic attributes, such as elevation, also play an important role in rockfall activity. At higher elevations, the strength of freeze-thaw cycles and humidity from rain or snowfall causes rock disintegration (Gardner, 1970; Duarte & Marquínez, 2002).

2.3.3.1 Climate and rockfall activity

Since the early twentieth century, frost cracking has been visually recognized by the existence of in-situ fractured rocks and talus slopes (Lozinski, 1909 as cited in Matsuoka & Murton, 2008). More comprehensive studies were undertaken in the late twentieth century on rockfalls and talus slopes (Rapp, 1960), and frost crack growth in rock debris (Mackay, 1999). Tricart (1956) attempted to define microgelivation after recognizing the association between loosening of grains and wedging of ice in small cracks. Further laboratory experiments attempted to resolve the rates and products of frost weathering as a function of environmental conditions and rock properties (e.g. Battle, 1960;
Matsuoka, 1990a). The laboratory-derived criteria were applied to field-based observations of rock temperature, moisture and bedrock disintegration (e.g. Gardner, 1970; Thorn, 1979; Matsuoka, 1990b).

In addition, theoretical approaches have provided the basis for numerical modeling of bedrock weathering, and resulting landscape evolution. For example, Walder & Hallet (1985) used a mathematical model for the breakdown of porous rock by the growth of ice within cracks. They addressed the role of material parameters (grain size, shape, crack size), environmental conditions (temperature, temperature gradient, water pressure) in frost damage to rocks. Additionally, Anderson (1998) derived an analytic approach to determine the time spent by rocks within the frost cracking window as a function of depth in an alpine environment. This approach by Anderson for one-dimensional heat conduction with a sinusoidal surface temperature variation was also employed by Hales & Roering (2007).

Similarly, recent research in an unglaciated environment applied the concept of the analytic solution and found that frost-dominated erosional processes were prevalent during the Last Glacial Maximum (Marshall et al., 2015) (Figure 2.6).

Other studies have addressed climatic and environmental factors, including various meteorological conditions, such as seasonal (e.g. Matsuoka, 2001) or diurnal temperatures (e.g. Sass, 2005; Hales & Roering, 2005) and elevation-dependent weathering mechanisms (Hales & Roering, 2005) (Figure 2.5) that enhance the probability of rockfall occurrence.
Figure 2.5 Annual air-temperature variations based on daily temperature data for different elevations in central Southern Alps (after Hales & Roering, 2005)

Figure 2.6 Annual temperature curve based on monthly temperature data in Little Lake, Oregon (after Marshall et al., 2015)
CHAPTER 3: STUDY AREA AND METHODOLOGY

3.1 Study area

The present study is undertaken in Kananaskis Country, AB (Figure 3.1). The Canadian Rockies are divided into major sub-provinces, including the Front Ranges, Main Ranges, and Western Ranges. The Kananaskis River watershed falls within the Front Ranges (McMechan & Leech, 2011; McMechan, 2012a) and covers ~ 930 km². Topographical structures, which consist of northwest-southeast aligned ridges, range from ~1400 masl along the valleys, rising up to a peak of 3100 masl (Jackson, 1980; Hoffmann et al., 2013). On average, Kananaskis experiences 105 days of precipitation, with 350 mm falling as rain and 270 mm as snow (Whitfield, 2014). At the University of Calgary’s Biogeoscience Institute (adjacent to Barrier Lake) (see Figure 3.1), the monthly average temperature is -10 ºC in January and 14 ºC in July (Johnson & Larsen, 1991).

The Kananaskis watershed has a large number of tributary basins of different stream orders that enter the Kananaskis River. The structural geology and glacial imprint in this region results in many oversteepened rockwalls, and an abundance of talus features are preserved at the base of these rockwalls. For this study, eleven fifth-order drainage basins (~500 km² in total area) within the Kananaskis watershed were selected for the collection of the talus inventory and subsequent data analysis (Figure 3.1); this allowed for a large extent of spatial coverage with a variety of structural geology and glacial features in the various drainage basins. Most of the exposed rocks in the study basins are younger beds consisting of Paleozoic carbonates (North & Henderson, 1954; Fermor, 1999). The
formation and group of rocks associated with different period for the eleven study basins are listed in Table 3.1.

Table 3.1 Geological timescale with rock group and formation in fifth-order drainage basins, Kananaskis, AB (McMechan, 2013)

<table>
<thead>
<tr>
<th>Basin name</th>
<th>Period</th>
<th>Group/Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Porcupine</td>
<td>Upper Devonion</td>
<td>Exshaw Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Banff Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Palliser Formation</td>
</tr>
<tr>
<td>Evan-Thomas</td>
<td>Upper Devonian, Carboniferous, Permian</td>
<td>Rundle Group</td>
</tr>
<tr>
<td>Rocky</td>
<td>Upper Devonian Mississippian</td>
<td>Exshaw Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Banff Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rundle Group</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kootenay Group</td>
</tr>
<tr>
<td>Pocaterra</td>
<td>Jurassic</td>
<td>Rundle Group</td>
</tr>
<tr>
<td>Foche</td>
<td>Upper Devonion Quaternary</td>
<td>Palliser Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Quaternary</td>
</tr>
<tr>
<td>Three-Isle</td>
<td>Upper Devonion Quaternary</td>
<td>Palliser Formation</td>
</tr>
<tr>
<td>Upper-Kananaskis</td>
<td>Mississippian Quaternary</td>
<td>Rundle Group</td>
</tr>
<tr>
<td>Smith-Dorrien</td>
<td>Upper Devonian Triassic</td>
<td>Palliser Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Banff Formation</td>
</tr>
<tr>
<td>Ribbon</td>
<td>Mississippian Triassic</td>
<td>Rundle Group</td>
</tr>
<tr>
<td>Marmot</td>
<td>Jurassic</td>
<td>Kootenay Group</td>
</tr>
<tr>
<td>Boulton</td>
<td>Mississippian Jurassic</td>
<td>Rundle Group</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Banff Formation</td>
</tr>
</tbody>
</table>
Figure 3.1 Kananaskis Watershed with eleven fifth order drainage Basins. The Kananaskis River flows from South to North eventually merging into the Bow River. A star sign adjacent to Barrier Lake represent the location of Kananaskis climate station 3053600 at University of Calgary’s Biogeoscience Institute. The inset shows the location of Kananaskis watershed with reference to Calgary.
3.2 Data acquisition

3.2.1 Aerial photographs and orthoimages

This research utilizes aerial photographs to map talus in the eleven fifth-order drainage basins within the Kananaskis River drainage basin of the Canadian Rockies. This talus inventory provides the basis for an analysis of rockfall erosion rates and factors that control its operation, including structural geology and glacial topography. The aerial photographs used in this study were acquired from Spatial and Numeric Data Services (SANDS) archive at the University of Calgary (http://library.ucalgary.ca/sands). These colour stereo pairs have a scale of 1:30 000 (project number: 08-029, roll number: AS5449N and AS5450N) and were captured on August 15 and 18, 2008. In addition, a mosaicked high-resolution orthophoto (0.2 m × 0.2 m) was also utilized to aid in data collection. These orthophotos were captured on multiple dates between 18 July and 22 August 2013, after the major flooding that occurred earlier that year in Alberta. The orthophotos are true colour (red, green, and blue) and have a 30 cm ground sample distance. Orthorectification was performed using a combination of bare earth LiDAR (1 m pixel size, where available) and a DEM created from the stereo pairs (10 m point spacing) (E. Grass, personal communication, September 19, 2016). The server image is made available in ArcGIS through Alberta Environment and Park’s Generic Enterprise Spatial Information Services (GENESIS) (WMS: https://genesis.srd.alberta.ca/genesis_tokenauth/rest/services).
3.2.2 Digital Elevation Model (DEM)

The Alberta Digital Elevation Model (DEM) dataset utilized in this study is available at http://www.altalis.com/products/terrain/dem.html. This DEM was derived from 3033 1:20 000 NTS map sheets compiled from 1:60 000 aerial photography captured during 1980 to 1995 (AltaLIS, 2015). The 10 m × 10 m DEM (spatial reference of NAD_1983_Transverse_Mercator) was clipped to cover only the study area. Analysis of the talus deposits and other calculation were conducted using the DEM.

3.2.3 Paleoclimate data

In previous studies, it was found that rockfall activity was accelerated in the years following deglaciation, with lower rates occurring in present day (e.g. Luckman & Fiske, 1997; Hinchliffe & Ballantyne, 1999). This suggests that climate may be important in affecting rockfall rates because it has a direct impact on the rates of frost cracking, that might contribute to rockfall activity in alpine environments (see Chapter 2, section 2.3.3 for details). To analyze climatic patterns since deglaciation and to evaluate if frost cracking is a viable factor controlling rockfall activity, paleoclimate data were utilized in this study. The climatic data used in the frost cracking analysis were obtained from the paleoclimate simulations performed using the National Center for Atmospheric Research Community Climate System Model Version 3 (NCAR CCSM3) (Liu et al., 2009). Since this model stores decadal averages of seasonal climate variables from 22 ka BP to the present, the downscaled CCSM3 paleoclimate simulations for North America available at http://nelson.wisc.edu/CCR/resources/paleoclimate.php were used. Even though the downscaled CCSM3 data was available for as far back to 21 ka BP, the data for 12, 10, 8,
6, 4, 2, 0 ka BP were acquired (the time since deglaciation in Kananaskis is estimated at 12 500 years BP, Hoffmann et al. (2013)). For each millennium between 0 and 12 ka BP, decadal averages of seasonal climate variables from the first 100 years for each millennium of the simulation were extracted and averaged (e.g. 12 – 11.9 ka BP, 11 – 10.9 ka BP, etc.) (Veloz et al., 2012). A standard change-factor approach was followed to obtain the statistically downscaled and debiased climate data with a grid resolution of 0.5 × 0.5 degree, which is ~43 km × 43 km per pixel (Wilby et al., 2004). Therefore, 5 pixels of the raster file cover the area of interest defined for this study. The temperature data stored in every pixel within the study area was transferred in ArcGIS and mean was calculated for each season. Combining these data with modern weather station data, daily temperatures for all past years of interest were reconstructed (further details on methodological procedures are provided in section 3.3.4.3).

3.2.4 Modern meteorological data

Unlike seasonal temperature data calculated from a paleoclimate model (see section 3.2.2), the modern meteorological data provides daily temperature data along with elevation information. Herein, daily temperature data collected from Kananaskis climate station 3053600, located at University of Calgary’s Biogeoscience Institute at 51.03°N and -115.03°W (see Figure 3.1), was used to reconstruct daily temperature data for each paleotime of interest and analyze the influence of frost cracking in the study area.

At this weather station, daily maximum and minimum temperatures are collected for Environment Canada using manual thermometers. The max/min temperatures are recorded twice daily (8:00 am and 4:00 pm) (A. Cunnings, personal communication, November 12, 2014). Mean daily meteorological data was obtained between the years of
1940 and 2015 from the publicly available climate archive for 3053600 Kananaskis. Over this time period, the weather station has been relocated several times, which resulted in a changing elevation of the station (Whitfield, 2014). Since the change in station elevation is not significant, the current station elevation (i.e. 1391 m) was utilized for the study. Although other variables were available, this study focused on daily mean temperature, calculated as the average of the maximum and minimum temperatures. These temperature data were used in conjunction with the environmental lapse rate for the region to estimate temperature at different elevations. The environmental lapse rate is about 4.4°C km\(^{-1}\) in the Rocky Mountains and Foothills region of Southern Alberta, covering Mt. Evan Thomas and Moose Mountain within Kananaskis Country (Cullen & Marshall, 2011). Further details on methodological procedures are provided in paleoclimate reconstruction section 3.3.4.3.

### 3.3 Method of analysis

To fulfill the research objectives, the talus inventory was collected using aerial photographs, the orthoimage, and the DEM. The identified talus polygons were then analyzed to obtain the volumes for each talus deposit. This information was then used to calculate the rockfall erosion rates. Finally, the information about talus volume and erosion rates, as well as other topographic information, were then utilized to analyze the association of talus with factors related structural geology, glacial history, and climatic data (frost cracking).
3.3.1 Talus identification for study basins

The eleven fifth-order drainage basins were delineated by inputting the clipped DEM into Arc Hydro tools in ArcMap 10.2.2 (see Figure 3.1). A talus inventory was created for each of the study basins in the following manner. Each talus deposit was first identified from the stereoscopic viewing of aerial photographs. Talus identification and associated topographic characteristics (including the rockfall source area for each talus deposit) were identified on the basis of its overall appearance and specific factors including shape, size, pattern, tone, grain size texture, and association. Each talus deposit was then digitized onto the mosaicked high-resolution orthophoto. When possible, individual talus features (e.g., talus cone, talus sheet) were identified and recorded as one polygon. However, due to the many compound type talus slopes (see earlier definition) present in the study basins, separation of individual features often presented major difficulties. Compound talus features that were unable to be separated into smaller polygons were identified as one talus polygons (Plate 3.1).
Plate 3.1 a sample of an orthophoto (0.2 m × 0.2 m resolution) showing talus slopes (red polygons). Black line indicates boundary of Ribbon drainage basin (Image Source: AEP GENESIS (2013))

3.3.2 Talus surface area and volume calculation

Landscape area on the aerial photographs is viewed in terms of planimetric area, which does not account for difference in slope gradients for different locations. A 3D analyst tool in ArcMap 10.2.2 called ‘Polygon Volume’ (details discussed later) was used
to measure the surface area and volume of each polygon. Surface area is obtained from the equation (Berry, 2002; Jenness, 2004):

\[
Surface\ Area\ (m^2) = \frac{Planimetric\ Area\ (m^2)}{cosine\ (\theta)}\ \ \text{Equation 3.1}
\]

The planimetric area is a product of length and width of a polygon, but accounting for slope gradient provides the true surface area. When gradients are not flat, the surface area is greater than the planimetric area (Booth, 2000).

The Polygon Volume tool is utilized to obtain estimates of volume for each talus polygon. The boundary of each polygon was first intersected with the interpolation zone of the TIN (created for the landscape surface). This step identifies the common boundary. The surface area for each polygon is calculated (Jenness, 2013). The calculation for the volume of each polygon requires an additional parameter, height. Since the shape of a talus base is unknown, for this study, it was assumed that talus slopes have a flat (planar) base. The minimum elevation for each talus polygon was then extracted using the ‘zonal statistics as table’ tool, and set as a plane height (Figure 3.2; see blue line). Setting the reference plane to ‘ABOVE’, the Polygon Volume tool considers elevation difference for each pixel between the plane height and the underside of a TIN surface (Figure 3.2, dotted red area). By multiplying the elevation difference for each pixel with respective surface area, the volume for each talus polygon is computed. Volume calculations with this approach are likely to be an overestimate, however, error bars for volumes are difficult to quantify (Dussauge et al., 2003).
3.3.3 Erosion rate calculation

The volumes of talus polygons are used to calculate the bedrock erosion rate (m yr⁻¹) for each basin (equation 3.2).

\[ ER = \frac{\rho_s \sum_{i=1}^{j} V_i}{\rho_b \cdot A \cdot t} \]  

Equation 3.2

(Hales & Roering, 2005; Hoffmann et al., 2013)

where \( V \) is volume of talus polygon (m³), \( \rho_s \) is density of sediment (1600 kg m⁻³) (Hoffmann et al., 2013), \( \rho_b \) is density of bedrock (2600 kg m⁻³) (Sass & Wollny, 2001), \( j \) is the number of talus polygons in a basin, \( A \) is area being analyzed (m²) and \( t \) is time since deglaciation (12.5 ka BP). Erosion rates are often obtained by dividing the total volume by the total area under consideration (e.g. Hales & Roering, 2005; Hoffmann et al., 2013). However, this approach has an underlying assumption that the entire area under consideration is susceptible to erosion. The source area for talus are steep rock slopes, usually above 45-50° (Martin, 2000). Therefore, talus source areas in our study area were analyzed and it was determined that a threshold of approximately 45° is required for a slope to be source area for talus deposits (Moore et al., 2009; Frattini et al., 2008). For each basin,
the area of the basin having slope gradients > 45˚ (herein referred to as steepland) was determined and used in the calculation of erosion rates.

### 3.3.4 Controlling factors of rockfall-talus activity

The following section explains the underlying methodologies employed to analyze the potential controlling factors outlined in Chapter 2.

#### 3.3.4.1 Structural Geology

The geological maps for the Kananaskis Lakes area (NTS 82-J/11) and Spray Lakes Reservoir (NTS 82-J/14) cover most of the study area. These 1:50 000 scale maps (McMechan, 2012a, 2013) were obtained from Natural Resources Canada. These geological maps were created on the basis of field mapping, topographic sheets, aerial photographs (1:12 000 from the Federal Government and 1:20 000 from Provincial Government), and use of an altimeter and a compass (Maurel, 1987). Thrust faults with variably folded Paleozoic and Mesozoic carbonate and clastic strata dominate the structural geology of Front Ranges (McMechan & Macey, 2012b), and five major thrust faults (including Misty, Rundle, Lewis, Sulphur Mountain, and Bourgeau) fall within the Kananaskis watershed. With the aid of these geological maps, folds and faults were located within our study area, and examined in relation to the aerial photographs utilized in this study. Features such as normal faults, transverse faults, synclines, and anticlines were located. The geological maps also provided lithological information. This geological information was analyzed in conjunction with the talus inventory for each drainage basin and associations of talus polygons with particular structural geology features were noted and analyzed (see Chapter 4 for details).
3.3.4.2 Glacial Topography Analysis

Evans and Cox (1974) define a cirque in the following manner, “A cirque is a hollow, open downstream but bounded upstream by the crest of a steep slope ('headwall') which is arcuate in plan around a more gently-sloping floor”. In this study, all glacial cirques in each study basin were identified and characterized based on their location relative to major valleys. Mostly, cirques are present along the sides, or at the heads of valleys, occasionally coalescing at the valley heads to form large complexes. Occasionally, cirques are also present in isolated sites unrelated to major valleys (Trenhaile, 1976). A second criterion for examination is a qualitative assessment of other glacial landforms that are associated with talus polygons, including U-shaped valleys and their steep headwalls and sidewalls (Sauchyn and Gardner, 1983; Hoffman et al., 2013). Earlier studies in Kananaskis identified cirques from aerial photographs and field observations (e.g. Duford and Osborn, 1978; Smith et al., 1995; Hoffman et al., 2013). The glacial features outlined above were identified using the 1:30 000 aerial photographs previously used in this study and 1:50 000 Canadian National Topographic System maps (contour interval 100 feet) (Plate 3.2 and 3.3). The association of talus polygons with cirques and other glacial features was then analyzed (see Chapter 4 for details).
Plate 3.2 Cirque identification from orthophotographs and topographic maps at Smith-Dorrien drainage basin. Blue line represent the boundary of a cirque. Lake present at the cirque floor and talus slopes at the base of a steep rockwall makes it more obvious to identify (Image Source: AEP GENESIS (2013))

Plate 3.3 U-shaped valley identification from an orthophotograph at Smith-Dorrien drainage basin. The orange dashed line represents the U-shape of a valley and red star is the location of another cirque. Also, notice talus slopes formed along the U-shaped valley bottom (Image Source: AEP GENESIS (2013)).
3.3.4.3 Paleoclimate reconstruction

To analyze the possibility that frost cracking (related to temperature) is a factor affecting rockfall rates, paleotemperature data are required. This approach relies on paleotemperature data obtained from the model described earlier in this chapter. These paleotemperature data are available as seasonal values. The approach used to reconstruct daily temperatures from 0 to 12 ka BP for intervals of 2 ka BP is outlined below.

**Step 1.** The model data for 0, 2, 4, 6, 8, 10, 12 ka BP (herein referred to $T_{Psi(tn)}$) are utilized in the first step of this analysis. It should be noted that temperature estimates from this model are for the mean elevation of the Earth’s surface for the particular pixel under consideration (Kiehl et al., 1996). Model temperatures are used to calculate the differences between the model estimate for 0 ka BP and each date of interest (i.e., 2, 4, 6, 8, 10, 12 ka BP). The temperature difference ($T_{ΔPsi}$) for $i^{th}$ season calculated from model from 0 ka BP ($T_{Psi(t0)}$) and the period of interest ($T_{Psi(tn)}$):

$$T_{ΔPsi} = T_{Psi(tn)} - T_{Psi(t0)} \text{ Equation 3.3}$$

where, $i=$ Winter (Dec-Feb), Spring (Mar-May), Summer (Jun-Aug), and Fall (Sep-Nov) and $n$ is period of interest (e.g. $n=12$ ka BP, 10 ka BP, … 2 ka BP). Because the model only provides produced seasonal temperature data (only 4 values for each paleotime of interest), it was assumed that paleoclimate model temperature is consistent throughout a season. The model temperature variations ($T_{ΔPsi}$) calculated from equation 3.3 represents the degree to which the temperature for some previous time is colder or warmer than the model estimate for 0 ka BP.

**Step 2.** Next, the modern-day BGS weather station data (modern day temperatures at an elevation of 1391 m) are adjusted using the temperature variations for past times (2 ka BP,
4 ka BP etc.) obtained in step 1 above. This step provides estimates of the temperatures for each past time of interest at the elevation of the BGS weather station (1391 m). To obtain these results, daily temperature differences ($T_{\Delta Wsj}$) are calculated from the (modern-day) BGS weather station temperature data ($T_{Wdj}$) and $T_{\Delta Psi}$ (see above) using equation 3.4:

$$T_{\Delta Wsj} = T_{Wdj} + T_{\Delta Psi} \quad \text{Equation 3.4}$$

where, $j$ is year-day. An example of temperature reconstruction for Day 91 and Day 183 (Julian Day) is shown in Table 3.2.

Table 3.2 Temperature reconstruction of 12 ka BP for days 91 and 183 using model temperature ($T_{\Delta Psi}$) and daily weather station temperature data ($T_{Wdj}$). $T_{\Delta Wsj}$ is the final reconstructed temperature.

<table>
<thead>
<tr>
<th>Temperature in ºC</th>
<th>j=91</th>
<th>j=183</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T_{Psi(t0)}$</td>
<td>-11.23</td>
<td>-2.01</td>
</tr>
<tr>
<td>$T_{Psi(t12)}$</td>
<td>-15.29</td>
<td>-5.85</td>
</tr>
<tr>
<td>$T_{\Delta Psi}$</td>
<td>-4.06</td>
<td>-3.84</td>
</tr>
<tr>
<td>$T_{Wdj}$</td>
<td>-6.63</td>
<td>9.90</td>
</tr>
<tr>
<td>$T_{\Delta Wsj}$</td>
<td>-10.69</td>
<td>6.06</td>
</tr>
</tbody>
</table>

Step 3. Next, the number of days that a rock spends in a year in frost cracking window (-3 to -8 ºC) requires information about temperature not only at the valley bottom, but for various elevations throughout Kananaskis. The modern-day environmental lapse rate of 4.4ºC km$^{-1}$ (Cullen & Marshall, 2011) was applied to generate elevationally-adjusted daily temperatures at 500 m elevation intervals for the past times of interest. This information is then utilized to analyze the possible role of frost cracking on rockfall activity in our study area (see Chapter 4).
4.1 Talus slope inventory

Talus deposits were analyzed for 11 fifth-order drainage basins within the Kananaskis watershed (Figure 4.1). Areas for the study drainage basins range from ~12 km² to 102 km², with Smith-Dorrien being the largest drainage basin and Marmot being the smallest. Figure 4.1 shows the locations of all identified talus slopes within the study basins. Initial inspection of results reveals that talus slopes are mostly present at the headwater locations of the study basins. A total of 324 talus deposits covering a surface area of 28.51 km² is delineated within the study basins, with the number of talus polygons in the study basins ranging from 1 to 73. Table 4.1 provides a summary data table and some descriptive statistics about the talus polygons. Since one large talus polygon is digitized within Marmot basin, calculating the minimum and maximum volume for this basin is not applicable.

Mean volumes of talus polygons for the study basins range from 261.52 × 10⁻⁵ to 1723.79 × 10⁻⁵ km³. A minimum volume of 0.45 × 10⁻⁵ km³ is recorded for talus slopes in the Upper-Kananaskis drainage basin, with a maximum mean value of 12 495.55 × 10⁻⁵ km³ is found in Pocaterra drainage basin (Table 4.1).
Figure 4.1 Talus slope delineation in fifth order drainage basins, Kananaskis watershed. Each red polygon signifies talus deposits delineated based on aerial photographs, orthoimage, and hillshade DEM.
Table 4.1 Descriptive statistics for number of talus polygons in 11 study basins within Kananaskis watershed and its volume

<table>
<thead>
<tr>
<th>Basin name</th>
<th>No. of talus polygons</th>
<th>Mean volume ($\times 10^{-5}$ km$^3$)</th>
<th>Minimum volume ($\times 10^{-5}$ km$^3$)</th>
<th>Maximum volume ($\times 10^{-5}$ km$^3$)</th>
<th>Median volume ($\times 10^{-5}$ km$^3$)</th>
<th>Standard deviation ($\times 10^{-5}$ km$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulton</td>
<td>9</td>
<td>261.52</td>
<td>5.32</td>
<td>1004.98</td>
<td>102.65</td>
<td>372.98</td>
</tr>
<tr>
<td>Evan-Thomas</td>
<td>58</td>
<td>375.40</td>
<td>1.10</td>
<td>3155.14</td>
<td>245.36</td>
<td>531.44</td>
</tr>
<tr>
<td>Foche</td>
<td>28</td>
<td>822.46</td>
<td>30.28</td>
<td>6929.30</td>
<td>473.32</td>
<td>1349.06</td>
</tr>
<tr>
<td>Pocaterra</td>
<td>20</td>
<td>1328.92</td>
<td>1.14</td>
<td>12495.55</td>
<td>253.48</td>
<td>2788.45</td>
</tr>
<tr>
<td>Porcupine</td>
<td>5</td>
<td>898.33</td>
<td>89.11</td>
<td>2481.91</td>
<td>253.43</td>
<td>1082.86</td>
</tr>
<tr>
<td>Ribbon</td>
<td>41</td>
<td>1723.79</td>
<td>20.04</td>
<td>9256.21</td>
<td>562.09</td>
<td>2342.34</td>
</tr>
<tr>
<td>Rocky</td>
<td>19</td>
<td>585.67</td>
<td>1.61</td>
<td>2773.26</td>
<td>285.16</td>
<td>748.33</td>
</tr>
<tr>
<td>Smith-Dorrien</td>
<td>73</td>
<td>1109.42</td>
<td>2.20</td>
<td>11163.21</td>
<td>440.11</td>
<td>1891.97</td>
</tr>
<tr>
<td>Three-Isle</td>
<td>31</td>
<td>832.01</td>
<td>0.74</td>
<td>3502.47</td>
<td>148.84</td>
<td>1226.39</td>
</tr>
<tr>
<td>Upper-Kananaskis</td>
<td>39</td>
<td>290.98</td>
<td>0.45</td>
<td>1298.78</td>
<td>192.41</td>
<td>344.97</td>
</tr>
<tr>
<td>Marmot</td>
<td>1</td>
<td>266.93</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
</tbody>
</table>

When comparing the volumes of individual talus polygons within and between different drainage basins, it should be kept in mind that when compound talus deposits were identified, they were considered as a single talus polygon. The number and nature of compound talus polygons differ amongst the study basins, and will provide limitations in our examination of the distribution and mean values of talus volumes.
Figure 4.2 Distribution of talus volume in drainage basins within Kananaskis watershed for A: Smith-Dorrien, B: Ribbon, C: Foche, D: Pocaterra, E: Three-Isle, F: Upper-Kananaskis, G: Evan-Thomas, H: Rocky, and I: Boulton. Note: Porcupine and Marmot basins are excluded due to the minimum number of talus polygons. All histograms represent a positively skewed distribution.
Figure 4.2 shows the frequency of talus volumes for 9 of the 11 study basins. Porcupine basin and Marmot basin are not included in this figure due to the low number of talus polygons. In general, talus deposits show large frequencies at lower volumes, with the tail of the frequency distributions showing a rapid decline for larger volumes. The majority of talus deposits fall within the volume category ranging from 0-0.005 km$^3$ (Figure 4.2). For example, Evan-Thomas has 45 talus deposits out of 58 deposits falling within this range (Figure 4.2G). Similarly, 38 talus slopes out of 73 deposits in Smith-Dorrien, 29 deposits out of 39 in Upper-Kananaskis, 21 deposits out of 31 in Three-Isle, 20 deposits out of 41 in Ribbon, 15 deposits out of 28 in Foche, 12 deposits out of 20 in Pocaterra, 12 deposits out of 19 in Rocky, and 7 talus slopes out of 9 deposits in Boulton fall within the category 0 to 0.005 km$^3$. There are some drainage basins which contain some particularly large talus volumes as shown in the tails of the distributions; for example, Smith-Dorrien and Pocaterra basins consist of talus deposits with values of 0.112 and 0.125 km$^3$ (Figures 4.2A, D).

### 4.2 Erosion rates due to rockfall

The inventory of 324 talus deposits for all basins covers a total surface area of 28.51 km$^2$ with the total study area having a value of 499 km$^2$. It is assumed that the total time period for accumulation of talus is 12.5 ka BP years (see section 3.3.2). The total volume of rock fragments delivered from steeplands in Kananaskis watershed over 12.5 ka period is 2.77 km$^3$ (Table 4.2). To convert these values into annual erosion rates for each study basin, the total talus volume for each basin is divided by the total area of steepland (potential source area for rockfalls) and the time period over which talus accumulates (i.e. 12.5 ka). It is most reasonable to report erosion rates in relation to the area in each drainage
basin that has the potential to contribute to rockfall erosion and associated talus deposits. Based on topographic analysis of rockfall slopes in our study area, the minimum slope gradient for rockfall source area is \( \sim 45^\circ \) (see section 3.3.2); using this criterion the total area in each basin having a slope \( > 45^\circ \) is identified. Finally, erosion rates are adjusted to remove the effect of regolith porosity by following the procedure outlined by Hoffmann et al. (2013); volume is adjusted by multiplying this value by the ratio of the regolith density (1600 kg m\(^{-3}\)) to bedrock density (2600 kg m\(^{-3}\) for limestone bedrock as described in Sass & Wollny, 2001). Erosion rates calculated for the study basins range from 0.53 to 5.4 mm yr\(^{-1}\), with an average rate of 2.9 mm yr\(^{-1}\). The mean erosion rate suggests that, on average, steepland in the study area has eroded a value of \( \sim 36 \) m over the last 12 500 years. The maximum erosion rate measured at Pocaterra basin (5.4 mm yr\(^{-1}\)) translates to cliff erosion of nearly 68 m, while the minimum retreat rate at Boulton basin (0.53 mm yr\(^{-1}\)) involves only 6.6 m of rockwall retreat. Many reported erosion rates in the scientific literature do not normalize for only the possible source (contributing) area for mass movements (e.g., Sass & Wollny, 2001; Hales & Roering, 2005; Hoffmann et al., 2013), which would result in considerably lower estimates of erosion than if total area was utilized in the calculation. To allow for comparison with these other studies, erosion rates divided by the total area for each study area are also reported in Table 4.2; the erosion rates in these cases are much lower than when adjusted for the total area of steepland only. For example, Marmot basin has only 0.32% of its area greater than or equal to 45\(^\circ\), with the erosion rate for steepland areas having a value of 3.30 mm yr\(^{-1}\); this is a relatively high erosion rate. When the volume, in this case, is instead divided by total drainage area, the erosion rate is 0.01 mm yr\(^{-1}\), a value that is well below erosion rates reported in many previous studies. The erosion
rates normalized for total drainage area in this study range from 0.01 to 0.79 mm yr\(^{-1}\), a result that is consistent with erosion rates reported in Hoffmann et al. (2013) for Kananaskis (based on total drainage area).

Table 4.2 Calculated erosion rates from talus surface area and volumes. ER (Actual) refers to the erosion rates calculated only from the area of steepland and ER (Normalized) represents erosion rates calculated by normalizing talus volume throughout basin area. Difference in these erosion rates displays the importance of considering only steepland as a rockfall source area.

<table>
<thead>
<tr>
<th>Drainage basin</th>
<th>Total surface area of talus (km(^2))</th>
<th>Area of steepland (km(^2))</th>
<th>Total volume of talus (km(^3))</th>
<th>Basin Area (km(^2))</th>
<th>ER (mm yr(^{-1}))(^a)</th>
<th>ER (mm yr(^{-1}))(^b)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulton</td>
<td>0.319</td>
<td>2.182</td>
<td>0.0235</td>
<td>41</td>
<td>0.53</td>
<td>0.02</td>
</tr>
<tr>
<td>Evan-Thomas</td>
<td>3.139</td>
<td>4.468</td>
<td>0.217</td>
<td>75</td>
<td>2.41</td>
<td>0.15</td>
</tr>
<tr>
<td>Foche</td>
<td>2.594</td>
<td>3.927</td>
<td>0.230</td>
<td>34</td>
<td>3.0</td>
<td>0.36</td>
</tr>
<tr>
<td>Marmot</td>
<td>0.041</td>
<td>0.039</td>
<td>0.002</td>
<td>12</td>
<td>3.30</td>
<td>0.01</td>
</tr>
<tr>
<td>Pocaterra</td>
<td>2.271</td>
<td>2.430</td>
<td>0.265</td>
<td>68</td>
<td>5.40</td>
<td>0.20</td>
</tr>
<tr>
<td>Porcupine</td>
<td>0.444</td>
<td>1.383</td>
<td>0.044</td>
<td>29</td>
<td>1.60</td>
<td>0.07</td>
</tr>
<tr>
<td>Ribbon</td>
<td>6.035</td>
<td>8.331</td>
<td>0.706</td>
<td>50</td>
<td>4.20</td>
<td>0.79</td>
</tr>
<tr>
<td>Rocky</td>
<td>1.283</td>
<td>2.501</td>
<td>0.111</td>
<td>25</td>
<td>2.20</td>
<td>0.23</td>
</tr>
<tr>
<td>Smith-Dorrien</td>
<td>7.773</td>
<td>11.258</td>
<td>0.809</td>
<td>102</td>
<td>3.56</td>
<td>0.41</td>
</tr>
<tr>
<td>Three-Isle</td>
<td>2.769</td>
<td>2.978</td>
<td>0.257</td>
<td>23</td>
<td>4.30</td>
<td>0.6</td>
</tr>
<tr>
<td>Upper-Kananaskis</td>
<td>1.842</td>
<td>4.920</td>
<td>0.113</td>
<td>40</td>
<td>1.14</td>
<td>0.14</td>
</tr>
<tr>
<td>Total</td>
<td>28.657</td>
<td>44.421</td>
<td>2.784</td>
<td>499</td>
<td>2.88 (Average)</td>
<td>0.27 (Average)</td>
</tr>
</tbody>
</table>

\(^a\) This column represents the erosion rate based on steepland area only (>45°)

\(^b\) This column represents the erosion rate based on total drainage area
The erosion rates for steepland area presented in Table 4.2 indicate highly effective rockfall activity in the study basins. Similar rockwall erosion rates have been reported previously in the Norman Range area (District of Mackenzie, NW Canada) where rockwall retreat over the Holocene was up to 60 m, equivalent to an average retreat of 5.45 mm yr\(^{-1}\) (Smith, 1973). Similar to the rock slopes in our study area, talus deposits of the Norman Range area are dominated by limestone. Andre (1997) reported that overall rockwall erosion over the Holocene exceeds 10 m in dolomitic limestone. The order-of-magnitude erosion rates obtained in these previous studies are very similar to those found in the present study.

Interestingly, results of the present study (and the studies over the Holocene reported above) differ notably with the contemporary rockwall erosion rates reported in several studies, presumably due to the different temporal scales. For example, Luckman (1988) reported erosion rates ranging from 0.29 to 0.1 mm yr\(^{-1}\), specifically on limestone dominated rockwalls in Surprise Valley, Canadian Rockies, where only 13 years of recent rockfall activity was considered. A rockfall study at Mount Wilcox, Canadian Rockies (Luckman, 2008) covered a temporal scale of 40 years and erosion rates were estimated at between 0.1 and 0.55 mm yr\(^{-1}\). This previous study focused on an individual talus cone where the adjacent cliff is formed of calcareous and dolomitic siltstones, sandstones and shales, interbedded with limestones. The erosion rate of 2.9 mm yr\(^{-1}\) found in the present study is much greater than modern values of rockfall activity (e.g. Luckman, 1988; 2008). One suggestion for the present findings is that most of the erosion on rock slopes may have occurred over the several millennia following deglaciation during the paraglacial period, when hillslopes were oversteepened significantly by glacial activity, resulting in
accelerated rates of rock slope erosion; thereafter erosion rates would begin to lower eventually reaching the modern-day reported values.

Contemporary rock slope erosion rates have also been reported in other alpine environments including central Spitsbergen with reported values of 0.02-0.2 mm yr\(^{-1}\) for the period 1882-1954 but 0.35-0.5 mm yr\(^{-1}\) for the last 10 000 years (Rapp, 1960), Rockwall erosion rates for the French Alps ranged from 0.05-0.25 mm yr\(^{-1}\) for recent times but were estimated to be 1 mm yr\(^{-1}\) for the entire Holocene (Francou, 1988). Rock slope erosion rates in the Scottish Highlands were estimated to be 0.015 mm yr\(^{-1}\) in the present day (Ballantyne & Eckford, 1984). In summary, a comparison of Holocene rock slope erosion vs. erosion rates for contemporary times suggests that rockfall activity was greater in the distant past (probably being the most intense during paraglacial years) compared to the present day.

4.3 Control of structural geology and glaciation on rockfall-talus process

4.3.1 Results for control of structural geology and glaciation on talus deposits

Figure 4.3 shows the locations of glacial topography, including cirques and the geological features such as faults and folds within the Kananaskis watershed. A comparison of talus slope locations with glacial and geological topography is made easier with the maps for these same features shown individually for each drainage basin (Figure 4.4). Further details about the information shown in these figures is provided in section 4.3.2 and 4.3.3. Glacial topography includes cirques and U-shaped valleys; and geological features includes faults, folds and headwater basins. In this study, headwater basin refers to the topography that consists of steep slopes and is characterized by a mixture of glacial topography in the upper part of the basin and V-shaped characteristics in the lower part.
(Hoffmann et al., 2013). A combination of glacial topography and geological features is symbolized as cirque/fault* (Figure 4.4). Cirque/Fault* represents talus slopes that are associated with fault or folds in a cirque or U-shaped valley. Plate 4.1 is an example of an orthoimage showing the talus deposits in the Pocaterra drainage basin associated with fault and cirque/fault*. Additional pictures from the field visit of Pocaterra drainage basin is present in the Appendix. Further details are provided in section 4.3.2 and 4.3.3.
Figure 4.3 Location of cirques (Cyan colour polygons) present within fifth order drainage basins of Kananaskis watershed, Front Ranges of the Canadian Rockies. The elevation greater than 2100 masl is represented in brown colour, whereas elevation less than 2100 masl is represented in grey. Red dashed lines across the watershed represent folds and maroon dashed lines represent faults.
Figure 4.4 Talus slopes with associated topography such as folds (yellow), fault (green), cirque (cyan), U-shaped valley (orange), headwater basin (blue), and cirque/fault* (magenta) identified within eleven fifth order drainage basins on a hillshade DEM, Kananaskis. Cirque/fault* represents topography with fault/fold in cirque/U-shaped valley. Drainage basins include A: Ribbon, B: Rocky, C: Smith-Dorrien, D: Evan-Thomas, E: Three-Isle F: Boulton, G: Upper-Kananaskis, H: Foche, I: Pocaterra, J: Marmot, and K: Porcupine.

On the basis of the information about talus slopes and the associated topographic characteristics obtained from Figures 4.3 and 4.4, the percentage of the total volume of talus polygons associated with different topographic features is illustrated for all drainage basins (Figure 4.5). As previously mentioned, the topographic features in these graphs include faults, folds, headwater basins, cirques, U-shaped valleys, and faults/folds in cirques/U-shaped valleys. The results of these graphs are discussed in sections 4.3.2 and 4.3.3.
Plate 4.1 Talus slopes associated with fault and cirque/fault* in Pocaterra drainage basin. Cirque/fault* represents topography with fault/fold in cirque/U-shaped valley. Yellow dashed line shows the fault line. Red and yellow dashed line shows the outline of a cirque and fault (Image Source: AEP GENESIS (2013)). (See appendix for additional pictures)
Figure 4.5 Volume of talus slope associated with topographic features such as cirque, U-shaped valley, faults/folds present at cirque/U-shaped valley, fault, folds, and headwater basins. Each chart represent different fifth order drainage basin within Kananaskis watershed such as A: Ribbon, B: Rocky, C: Smith-Dorrien, D: Evan-Thomas, E: Three-Isle, F: Boulton, G: Upper-Kananaskis, H: Foche, I: Pocaterra, J: Kananaskis watershed (All basins included). Note: Porcupine and Marmot basins are not included (See text for description).
4.3.2 Talus deposits associated with structural geology

Delineation of geological features such as faults and folds are crucial for understanding rockfall processes. Steep cliffs, a necessary prerequisite for rockfalls in Kananaskis, are often associated with these features. The faults and folds located within the Kananaskis watershed are delineated on Figure 4.3 to allow for the association of talus deposits with these features.

The map of structural geology for Kananaskis shows that many faults and folds exist within the study area. For example, two major thrusts, Sulphur Mountain and Lewis along with other transverse faults are present in the Pocaterra drainage basin (McMechan, 2013; D. Spratt, personal communication, September 6, 2016). Similarly, the Rundle thrust passes through Rocky and Ribbon drainage basins, while the Bourgeau thrust is found within both Boulton and Smith-Dorrien basins. A small section of the Misty thrust is located within the Evan-Thomas drainage basin (McMechan, 2012a; McMechan, 2013; D. Spratt, personal communication, September 6, 2016). Normal and thrust faults are observed in the Foche basin, and it also shares a similar folded topography to Upper-Kananaskis drainage basin. Furthermore, a short-lived thrust fault is present at Three-Isle basin (not shown in Figure 4.3). Porcupine basin exhibits a complex structural geology because the two thrusts, Kananaskis and Bryant, present within this basin are also folded (McMechan, 2012a). Although no major thrust is present within the boundary of Marmot basin, it does exhibit a folded topography. Talus deposit location within the study basins display a consistent linear pattern along the folds and faults (D. Spratt, personal communication, September 6, 2016) (Figure 4.3). The association of talus slopes along these geological features is because folds increase the density and interconnectedness of discontinuities in
rock masses as a function of fold geometry (Lisle, 1994), stratigraphy (Cooke, 1997), and the presence or absence of preexisting discontinuities (Bergbauer and Pollard, 2004). On the other hand, faults induce regional disturbance in the fracturing density and modify the water table, thus affecting vegetation and generating bare rock slopes (Baillifard, Jaboyedoff, & Sartori, 2003; Burbank & Anderson, 2011). Furthermore, preservation of scarps on sides of a fault favors rockfall and talus activity (Burbank & Anderson, 2011). For example, a large volume of talus is associated on the cliff side of the Lewis thrust in Pocaterra drainage basin (Plate 4.1).

The percentage of talus deposits associated with faults and folds is shown in Figure 4.5A-J. Results demonstrate that talus slopes are more often associated with the presence of faults than folds. Certain basins that have major faults associated with them are expected to have a strong association between faults and talus deposits. For example, a substantial percentage of talus slope volume (63%) is associated with the fault line in Pocaterra basin (Figure 4.5I). This basin also has the highest erosion rates (5.42 mm yr⁻¹). Ribbon drainage basin has the second highest erosion rate, and about 20% of talus deposits are associated with faults and 11% of talus deposits are associated with folds. Furthermore, 24% of the talus volume in Rocky basin and 15% in Evan-Thomas basin display fault association.

These major faults and folds influence the pattern of outcrops of different lithologies at the surface. The lithology of the Kananaskis watershed is predominantly from the Devonian-Mississippian period. Firstly, tall resistant cliffs formed of the Palliser Formation (Upper Devonian) consist mainly of limestone and minor dolostone near the top; these cliffs are the location of many talus slope deposits in Evan-Thomas, Porcupine, Foche, Three-Isle, and Smith-Dorrien drainage basins. Similarly, the Rundle Group
(Mississippian) encompasses talus locations in Pocaterra, Rocky, Smith-Dorrien, Ribbon, Boulton, and Upper-Kananaskis drainage basins. The rocks in this group comprise massive limestone and dolostone that form cliffs due to differential erosion (Sauchyn & Gardner, 1983). There are some locations within drainage basins (i.e. Ribbon, Rocky, and Boulton) that do not coincide with the location of talus, and that consist of the Kootenay Group (Jurassic- Lower Cretaceous) (Figure 4.4A, B, and F). This Group has a resistant base that includes sandstones and rocks such as siltstone, mudstone, while thin coal seams are found in the upper part. Rockfall is uncommon in these areas because this lithology is found mainly in the valley bottoms where faults and folds are not generally found. Furthermore, talus slopes are not generally found within the lithology of the Exshaw and Banff Formations.

4.3.3 Talus deposits and associated glacial topography

Glacial topography, including cirques and U-shaped valleys, are often locations where abundant talus is found because of the oversteepened slopes and cliffs associated with these features. Cirques and U-shaped valleys in the study basins are delineated in the orthoimage to study the association of talus slopes with these features (Plate 4.1).

Cirques cover a total of 58 km² area within the 500 km² area of the study basins (~12% of total area). Analysis of the DEM for the study area shows that cirques are generally present at elevations greater than 2100 masl (Figure 4.3; Figure 4.4A-K). When all basins are considered together the mean elevation of cirques is ~ 2500 masl. Each pixel value obtained from DEM within a cirque polygon was accounted for and averaged to get the mean elevation of a cirque. Similar to our results, researchers found that in the southern Canadian Rockies, mean cirque elevation is ~ 2300 masl (e.g. Trenhaile, 1976; Hoffmann
et al., 2013). The mean elevation of cirques for individual basins ranges from ~ 2300 masl (at Boulton) to ~ 2650 masl (at Foche). Talus deposits in this study have a mean minimum elevation (i.e., the lowermost elevation of the cliff associated with the talus) of approximately 2100 masl and a mean maximum elevation (i.e., the uppermost elevation of the cliff associated with the talus) of 2450 masl. Figure 4.6 shows a graph of the mean elevation of cirques compared to the mean elevation of the talus (calculated using the same method as mean cirque elevation) for each study basin (note: this is the elevation for all talus in the basin, not just the talus associated with cirques). Talus deposits are present within a limited elevation zone that is similar to elevations of cirques. This graph suggests that the presence of cirques has a strong association with the elevation of talus in Kananaskis and that cirques may be an important control (in addition to faults and folds) on location of talus in Kananaskis.

Figure 4.6 Variation in elevation of cirque and talus deposits present at eleven fifth order drainage basins in Kananaskis watershed. Solid black line represent mean cirque elevation and dotted black line shows mean talus slope elevation. Note relatively consistent elevation of talus slope slightly below the cirque elevation.
Figure 4.5A-J also demonstrates a strong association of talus deposits with glacial features. Talus deposits are often associated with cirques and U-shaped valleys on the western side of the Kananaskis River (Figure 4.3; Figure 4.4A, C). Some of these basins show cirque coalescence, which generates a large number of oversteepened rockwalls and associated talus deposits. For example, 72% of the talus volume in Smith-Dorrien basin is associated with cirques (Figure 4.5C). Similarly, 58% and 55% of talus volume is associated with cirques for Foche and Boulton drainage basins (Figures 4.5H, F). In Three-Isle and Upper Kananaskis basins, 67% and 56% of talus deposits are associated with U-shaped valleys (Figures 4.5E, G). Marmot basin (not shown in figure 4.5) consists of a single talus polygon associated with both cirque and folded topography.

Talus deposits located on the eastern side of Kananaskis River most often show an association with headwater basins and faults (Figure 4.3; Figure 4.4B, D, and K). For example, Evan-Thomas basin has 82% of its talus volume in headwater basins, while Rocky basin has 46% of its talus volume in headwater basins (Figures 4.5D, B). Almost 100% of the talus in Porcupine basin is associated with headwater basins (Figure 4.4K). Overall, talus erosion rates are not remarkable (1.60-2.41 mm yr\(^{-1}\)) for basins where talus is associated with headwater basins (e.g., Porcupine, Rocky, and Evan-Thomas) when compared to erosion rates of 5.40 and 4.20 mm yr\(^{-1}\) for Pocaterra and Ribbon basins. Ribbon and Pocaterra basins show a strong association of talus with faults and cirques (Figure 4.4A, I). Figure 4.5J shows that 70% of talus deposits are associated with glacial topography when all basins are considered together. Headwater basins show the minimum contribution to talus volume, only accounting for 15% of the total talus volume.
4.3.4 Statistical analysis of talus associated with structural geology and glacial topography

Table 4.3 shows the volume of talus polygons associated with different glacial topographic and geological features. The mean talus volume is highest ($1763.86 \times 10^{-5}$ km$^3$) in topography that has faults and/or folds in cirques and/or U-shaped valleys and the lowest mean talus volume is found in headwater basin ($451.78 \times 10^{-5}$ km$^3$) (Table 4.3). Statistical analysis (Kruskal-Wallis, Mann-Whitney U test) is now undertaken to evaluate if the volume of talus polygon differs between talus associated with different features. In these tests, cirques and U-shaped valleys are considered under glacial topography, with faults, folds, and headwater basins as geological features. (Table 4.4).

Table 4.3 Descriptive statistics for volume of talus polygons associated with different topography

<table>
<thead>
<tr>
<th>Topography</th>
<th>Mean volume ($\times 10^{-5}$ km$^3$)</th>
<th>Median volume ($\times 10^{-5}$ km$^3$)</th>
<th>Minimum volume ($\times 10^{-5}$ km$^3$)</th>
<th>Maximum volume ($\times 10^{-5}$ km$^3$)</th>
<th>Standard deviation ($\times 10^{-5}$ km$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cirque</td>
<td>1138.38</td>
<td>406.10</td>
<td>1.83</td>
<td>11163.21</td>
<td>1905.12</td>
</tr>
<tr>
<td>U-shaped valley</td>
<td>723.90</td>
<td>252.54</td>
<td>0.74</td>
<td>5347.17</td>
<td>1033.44</td>
</tr>
<tr>
<td>Faults</td>
<td>1316.22</td>
<td>342.38</td>
<td>1.14</td>
<td>12495.55</td>
<td>2944.03</td>
</tr>
<tr>
<td>Fold</td>
<td>584.99</td>
<td>181.63</td>
<td>0.45</td>
<td>3166.02</td>
<td>929.68</td>
</tr>
<tr>
<td>Fold/fault in cirque/U-shaped valley</td>
<td>1763.86</td>
<td>601.60</td>
<td>7.44</td>
<td>7260.89</td>
<td>2354.50</td>
</tr>
<tr>
<td>Headwater basin</td>
<td>451.78</td>
<td>166.50</td>
<td>1.10</td>
<td>3778.58</td>
<td>729.42</td>
</tr>
</tbody>
</table>
Table 4.4 Nonparametric Kruskal-Wallis test (K-W) and Mann-Whitney U test (MW-U) among various topography. Topographical categories: (1) Cirque, (2) Fault/Folds in Cirque/U-shaped valley, (3) Fault, (4) Folds, (5) Headwater basin, (6) U-shaped valley.

<table>
<thead>
<tr>
<th>Topography</th>
<th>Test</th>
<th>p-value $(\alpha=0.05)$</th>
<th>Null Hypothesis</th>
</tr>
</thead>
<tbody>
<tr>
<td>Overall</td>
<td>K-W</td>
<td>0.023</td>
<td>Reject</td>
</tr>
<tr>
<td>Cirque (1) vs. Fault/Folds in Cirque/U-shaped valley (2)</td>
<td>MW-U</td>
<td>0.294</td>
<td>Accept</td>
</tr>
<tr>
<td>Cirque (1) vs. Fault (3)</td>
<td>MW-U</td>
<td>0.205</td>
<td>Accept</td>
</tr>
<tr>
<td>Cirque (1) vs. Folds (4)</td>
<td>MW-U</td>
<td>0.027</td>
<td>Reject</td>
</tr>
<tr>
<td>Cirque (1) vs. Headwater basin (5)</td>
<td>MW-U</td>
<td>0.001</td>
<td>Reject</td>
</tr>
<tr>
<td>Cirque (1) vs. U-shaped valley (6)</td>
<td>MW-U</td>
<td>0.05</td>
<td>Marginal</td>
</tr>
</tbody>
</table>

The Kruskal-Wallis test reveals a statistically significant difference in the volume of talus deposits between the different groups ($p = 0.023$) with a mean rank size score of 181 for cirques (group 1), 189 for faults/folds in cirques/U-shaped valleys (group 2), 167 for faults (group 3), 128 for folds (group 4), 140 for headwater basins (group 5), and 157 for U-shaped valleys (group 6). Despite confirmation that talus deposits do vary in volume depending on their association, it does not explain the extent of difference or similarity. Therefore, pairwise comparisons among the six groups are conducted to evaluate which of the six factors has the greatest impact on rockfall volume.

Table 4.4 demonstrates the result of one-tailed Mann-Whitney U tests where the null hypothesis is that there is no difference in talus volume between pairs of the various groups. Conversely, the alternative hypothesis states that talus volume is greater in cirques (group 1) compared to other topography. The Mann-Whitney test indicates that the volume
of talus deposits is greater for cirques (median value of $406.09 \times 10^{-5}$ km$^3$) than for folds (median value of $181.63 \times 10^{-5}$ km$^3$) and headwater basins (median value of $166.50 \times 10^{-5}$ km$^3$). The p-value of 0.027 for cirque vs. folds and 0.001 for cirque vs. headwater basin is lower than the significance level ($\alpha$) of 0.05. Thus, with 95% confidence, it can be said that talus volumes associated with cirques are significantly greater compared to talus associated with folds and headwater basins.

Talus volumes associated with faults/folds in cirques/U-shaped valleys and faults have a p-value of 0.294 and 0.205, thus providing weak evidence against the null hypothesis. This strongly suggests that the volume of talus deposits associated with cirques, faults/folds in cirques/U-shaped valleys, and faults do not have significantly different volumes. Therefore, it can be suggested that talus volumes are significantly larger for cirques and faults.

4.4 Frost-cracking and talus deposits

Results of the present study suggest a strong association of talus deposits with topographic features related to structural geology and glaciation. However, several previous studies have suggested that the elevation of talus deposits is strongly associated with locations of the most intense frost cracking activity (Hales & Roering, 2005; Hales & Roering, 2007; Savi et al., 2015). Analysis is undertaken to evaluate the role of frost cracking activity in influencing the locations of talus deposits in our study area. Based on the methodology outlined in section 3.2, graphs for seven different times (12, 10, 8, 6, 4, 2, and 0 ka BP) are constructed (Figure 4.7A-G). The number of days in a year is shown as Julian day on the x-axis and daily mean temperature (in °C) is shown on the y-axis. These graphs demonstrate the variation in daily mean temperatures in Kananaskis at an
interval of 500 m, with elevations ranging from 1391 masl up to 3391 masl (see earlier section on the methodology for paleoclimate reconstruction, section 3.3.4.3). In addition, the graphs also show the frost cracking window (FCW) (blue shaded area in Figure 4.7A-G) (Anderson, 1998; Hales & Roering, 2005). The number of days spent in the FCW at different elevations does not vary much for different elevations in Kananaskis at 12 ka BP; values range from 54 days at an elevation of 1391 masl and up to 67 days at elevation 3391 masl. This suggests that the FCW was (approximately) equally viable at most elevations at this point in time; the mean elevation of talus deposits (2100 to 2450 masl) was not an elevation that was more susceptible to frost cracking than other elevations. From 10 ka BP and up until the present day, the elevations that are most susceptible to frost cracking are much lower than the mean elevation of talus in the study basins. Results show that for most of the past 12 ka, the optimum climatic condition for the greatest frost cracking activity is near the valley bottoms, and not in the higher elevations in which talus is found. The valley bottom (1391 m) most often spends > 90 days in the FCW. We suggest, therefore, that although frost cracking is an important process in driving rockfall activity associated with talus deposits, intense frost cracking activity is alone not the major determinant of where rockfall-talus occurs in the landscape. Topographically-controlled factors related to structural geology and glacial activity are essential in determining rockfall activity and associated talus deposits.
Figure 4.7 Mean daily air-temperature variations at different elevations in Kananaskis watershed. Using the lapse rate of 4.4°C km⁻¹, temperature is produced at 500 m intervals i.e. 1391 m, 1891 m, 2391 m, 2891 m, and 3391 m. Temperature data is reconstructed for 12 (A), 10 (B), 8 (C), 6 (D), 4 (E), 2 (F), and 0 ka BP (G). Blue shaded area represents frost-cracking window (FCW) (-3 to -8 °C).
4.5 Significance of results to drainage basin development in Kananaskis, AB, Canadian Rockies

Sediment routing, sediment connectivity and drainage basin development are topics that have been a focus of much geomorphological investigation in the past several decades (e.g., Chorley & Kennedy, 1971; Benda & Dunne, 1997; Heckmann & Schwanghart, 2013). Modern technologies and methodologies, including dating methods, advances in imagery (e.g., LiDAR, satellite imagery) and numerical modeling techniques, have allowed for rapid and exciting advances in our understanding of drainage basin development (e.g., James et al., 2007; MacGregor et al., 2009). The particular combination and rates of geomorphic processes (weathering, hillslopes, fluvial, glacial) in operation vary depending on the region being considered. Weathering, hillslope and fluvial processes are processes that have been occurring for many millennia in the Canadian Rockies since the retreat of the last major glaciers. However, glaciation provided an important legacy for these landscapes through glacial conditioning of the topography, which in turn has affected subsequent geomorphic process operation over millennial time scales (Ballantyne, 2002).

4.5.1 Glacial influence on topography

A central question related to mountainous topography is the nature of valley development. Do valleys widen more quickly than they deepen or vice versa? What is the role of glaciation in influencing valley formation? How do valleys and associated landscapes change in the years following deglaciation? What processes were most important during the paraglacial period?

A large proportion of the landscape in Kananaskis consists of steep rock-faced slopes, due to the topographic effects of processes related to structural geology and
glaciation. Valley morphology and sedimentology are affected by both glacial erosional and depositional processes (which may vary considerably between main valleys and smaller tributary valleys; see Hoffman et al., 2013). Various studies have documented the effects of glacial erosion on valley form (e.g., Harbor et al., 2002; Montgomery, 2002; Osborn et al., 2006). Glacial erosion generally causes valleys to become significantly wider than fluvial valleys. For example, Montgomery (2002) found that glacial valleys are about twice as wide as fluvial valleys (Montgomery, 2002). In terms of cirque topography, Delmas et al. (2015) reported that glacial topography, particularly cirque widening and lengthening, outpaced deepening; in other words, headwall recession dominated over subglacial deepening, resulting in the formation of wide but shallow cirques. In addition to the sculpting of rock by glacial erosional processes, other parts of the landscape may be mantled in various glacial deposits that have notable effects on landscape characteristics (e.g., Osborn & Luckman, 1988; Glasser & Bennett, 2004). This is the landscape topography upon which various geomorphic processes operate, including rockfall processes that are the focus of the present study.

4.5.2 Postglacial erosion

Following deglaciation, mountain landscapes continue to be modified by weathering and erosion (Ballantyne, 2002; Hattestrand et al., 2008). It is important to understand these landscape processes because they may modify the altitudinal range of drainage basins and reduce the gradient of mountainous terrain (Barr & Spagnolo, 2015). After glaciation, the exposed landscape is oversteepened, leaving the steep rock slopes particularly susceptible to intense rockfall and rock slide activity for many thousands of years. In addition, portions of the landscape may be covered in glacial deposits that will
subject to erosional activity. Weathering processes will lead to regolith development and soil cover will begin to form over portions of the landscape. The regolith- or soil- mantled portions of landscape may undergo a period of more intense debris slide activity due to the overall glacial steepening of the landscape. Over time, the rock and regolith-mantled slopes are expected to become less steep, and rates of rockfall and debris slide activity are expected to decrease. Depending on the topographic characteristics of the landscape being examined, some of these deposits may go into long-term storage and remain disconnected to lower parts of the drainage basin (Hoffman et al., 2013). In other cases, rock and sediments may be deposited into upper reaches of channel networks (for example, steep gullies). Debris flows and/or fluvial processes may then episodically mobilize these sediments stored in gullies (Schrott et al., 2003; Dadson and Church, 2005).

4.5.3 Implications of results to drainage basin development

This study provides estimates of rockfall erosion rates and information about factors controlling the spatial distribution of rockfall erosion and associated talus deposits. It is now discussed how the results of this study can influence our ideas about drainage basin development as described above.

(i) An improved understanding of sediment routing and drainage basin development requires investigations that document operation rates for the range of geomorphic processes occurring in different regional settings. Such information will improve our conceptual and quantitative understanding of landscape change, and provides information that can be used in the development, calibration and testing of numerical models of drainage basin development. For example, SIBERIA and Channel-Hillslope Integrated Landscape Development Model (CHILD) (Willgoose et al., 1991; Tucker &
Bras, 2000). For portions of the landscape susceptible to rockfall (i.e., slopes > 45°), this study finds rockfall erosion rates of ~ 2.9 mm yr\(^{-1}\) (note that this value represents the average erosion rate over the past 12.5 ka). Given that Luckman (1988, 2008) reported modern rockfall erosion rates in the Canadian Rockies that are considerably lower than these values, it suggests that rockfall erosion rates were probably considerably higher during the paraglacial period. The rockfall erosion rates found in this study is also greater than mean erosion rate for shallow landsliding (0.10 mm yr\(^{-1}\)) observed in Pacific Northwest and coastal British Columbia (Martin et al., 2002) and average soil creep rate (1.9 mm yr\(^{-1}\)) in Coast Range of Oregon (Benda, 1990).

(ii) This study provides insights regarding the strong association of rockfall/talus to various topographic features, including those associated with structural geology (e.g., faults, folds) and glaciation (e.g., cirques). This information will contribute to an improved conceptual understanding of topographic change in such landscapes and can be incorporated into the framework of numerical models of drainage basin development. In particular, these results have implications to the nature of sediment routing and sediment connectivity in Kananaskis and other mountainous landscapes (Brardinoni et al., 2009; Hoffman et al., 2013). For example, talus is often found on the relatively low-gradient floors of cirques, which are not generally well connected to the fluvial network. This may provide a disconnection in the sediment routing regime. In addition, talus associated with faults may or may not be directly connected to fluvial systems.

(iii) Rockfall and talus are notable contributors to drainage basin development, with topographic form itself having notable effects on hydrological functioning, including storage and transmission of water (Muir et al., 2011). Talus deposits associated with
rockfalls are frequent occurrences in Kananaskis and other parts of the Canadian Rockies. Until recently, their role in the storage and transmission of groundwater was poorly understood; recent research has demonstrated that talus can have a significant role in determining the runoff processes in mountainous regions (Clow et al., 2003; Barnett et al., 2005; Muir et al., 2011). Finally, the results of this study also have significance to plant ecology, as topographic form affects moisture availability that is important for plant functioning (Chase, et al., 2012).
CHAPTER 5: CONCLUSION

Rockfall-talus processes are among the principal agents of landscape modification in mountainous terrain. Among rockfall-talus studies published in the literature, most are site-specific and have focused on local events. This study presents an inventory of talus deposits over a period of 12.5 ka BP within ~500 km² eleven fifth-order drainage basins in Kananaskis watershed using aerial photographs and a digital elevation model. To explain the distributions shown in this talus inventory, possible controlling factors, including structural geology, glacial activity, and climate-related weathering processes, are analyzed. The analysis of structural geology was completed using structural geology maps, aerial photographs were used to identify glacial topography, and paleoclimate model data was used to reconstruct daily temperature data back to 12 ka BP.

5.1 Summary of findings

A total of 324 talus deposits covering a surface area of 28.51 km² were delineated within the study basins. The erosion rates observed within eleven fifth order drainage basins ranges from 0.53 to 5.40 mm yr⁻¹, with an average rate of ~ 2.9 mm yr⁻¹. The mean erosion rate suggests that, on average, steepland in the study area has eroded ~ 36 m over the last 12.5 ka. This result is relatively high in comparison with other rates reported for the contemporary rockwall erosion. This suggests that rockfall activity was greater in past years compared to the present day. Based on these relative erosion rates, it can be suggested that most of the erosion on rock slopes may have occurred for several millennia following deglaciation during the paraglacial period when hillslopes were oversteepened significantly by glacial activity and accelerated rock slope erosion rates.
In the Kananaskis watershed, the talus accumulations display a consistently linear pattern along folds and faults. Moreover, cirques cover an area of 58 km², 12% of the 500 km² area of the study basins, generating an ideal topography for rockfall occurrence. The size of talus deposits varies significantly between different topographic characteristics, including faults, folds, headwater basins, cirques, and U-shaped valleys. Talus deposits are found to have a strong association with faults. In addition, a statistical test confirmed that talus volume is significantly greater in cirques (39%) compared to other topographic features such as folds and headwater basins. The concentration of talus deposits within a limited elevation zone (2 100 to 2 450 masl) slightly below the mean elevation of cirques suggests the influence of glacial topography on the distribution of rockfall activity in the study area.

Several previous studies have suggested that the elevation of talus deposits is strongly associated with locations of the most intense frost cracking activity. Since frost-induced rockfall is most effective at the frost-cracking window (FCW) (-3 to -8 °C), climatic control was analyzed using a paleoclimate model to reconstruct climate at 12, 10, 8, 6, 4, 2, 0 ka BP. Reconstructed daily temperature data indicate that this process may not consistently be the major factor affecting the distribution of talus deposits. At 12 ka BP, high elevations (3391 m) spent a high number of days (up to 67) in the FCW. Results suggested that the FCW was (approximately) equally viable at most elevations at this point in time. However, from 10 ka BP and up until the present day, the elevations that are most susceptible to frost cracking are lower than the mean elevation (2 100 to 2 450 masl) of talus in the study basins. Because the longest FCW is at low elevations and talus is found at higher elevations, intense frost cracking activity alone is not alone the major determinant
of where rockfall-talus occurs in the landscape. Therefore, we suggest that although frost cracking is an important process driving rockfall activity associated with talus deposits, intense frost cracking activity alone is not the major determinant of rockfall-talus distribution across the landscape. Topographically-controlled factors related to structural geology and glacial activity are suggested to be significant in determining rockfall and talus activity.

5.2 Summary of implication and future research

Rockwall erosion reshapes mountains and distributes deposits to lower slopes and elevations. Our study quantified the rockwall erosion rate as well as explained the distribution of talus deposits and their association with different topographic features. This talus inventory covers an extended spatiotemporal scale and provides important information about geomorphic process operation in Kananaskis, Canadian Rockies. This information can contribute to other studies seeking to understand hillslope geomorphology, fluvial geomorphology, sediment routing and storage, drainage basin development, as well as groundwater hydrology and plant ecology.

Several suggestions are now made for future research. It is recommended that talus inventories be collected for different regional settings, to evaluate erosion rates and factors influencing talus distribution over large spatiotemporal scales. Furthermore, additional research is required that considers the smaller-scale processes associated with rockfall activity. For example, the network of cracks and its orientation at several spatial scales, lithology (including rock mass strength), aspect (that controls microclimate), and precipitation are other factors related to rockfall processes that require additional investigation.
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APPENDIX

Field photographs from Pocaterra drainage basin, Kananaskis, AB.