Thermal Regime of Two Talus Slopes in Northwestern Ontario

by

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Abstract

This thesis investigates the thermal regimes of two talus slopes near lakes Superior and Nipigon in Northwestern Ontario. Mean annual air temperatures in the region are periglacial, but little work has been done on whether permafrost or anomalously cold ground exist in blocky materials in the area. Air and internal talus temperatures were recorded, and the dimensions of the rocks and the temperature at the base of the snow were measured. Internal talus temperatures increased with elevation. The talus slopes at both Doghead Mountain and Ouimet Canyon experienced an unstable thermal regime dominated by convection during the winter, and a stable, stratified thermal regime dominated by conduction during the summer. The summer-time thermal regimes of the lower sections of the slopes were dominated by zero curtains. Thaw completion depended on the length of the zero curtain, and thus on the amount of ice in the talus. Permafrost was found in Ouimet Canyon.

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Table of Contents

Abstract	i
Acknowledgements	ii
Table of Contents	vi
List of Figures	xi
List of Tables	xviii
Permission to Reproduce	xx
1 Introduction	1
1.1 Introduction	1
1.2 Isolated permafrost incidence	5
1.3 Objectives	7
1.4 Hypotheses	8
1.5 Methods	8
1.6 Thesis structure	8
2 Background	10

2.1	Permafrost	10
2.1.1	Factors affecting the distribution of permafrost	11
2.1.2	Factors affecting the thermal regime of the ground	15
2.1.2.1	Heat movement through the ground	17
2.1.2.2	Latent heat and the zero-curtain	17
2.2	Blocky materials	21
2.2.1	The geomorphology of talus slopes	24
2.3	Thermal anomalies and permafrost in blocky materials	25
2.3.1	Observations from around the world	26
2.3.2	Thermal regimes in kurums	30
2.4	Heat movement in blocky materials	32
2.4.1	Changes to conductive heat transfer	33
2.4.2	Convective heat transfer	33
2.4.2.1	Free convection	33
2.4.2.2	The chimney effect	35
2.4.2.3	The Balch effect	37
2.4.2.4	Limits on convection	39
2.4.2.5	Continuous air exchange with the atmosphere	44
2.4.3	Movement and phase change of water	45
2.5	Summary	45

3 Study Sites and Methods

3.1	The Lake Superior and Lake Nipigon region, Northwestern Ontario	47
3.1.1	Climate	47
3.1.2	Geologic history of the Lake Superior and Lake Nipigon region	55
3.1.2.1	Mesoproterozoic rift activity	56
3.1.2.2	The country rock	56
3.1.2.3	The intrusive rock	57
3.1.3	Glaciation	59
3.2	Study sites	60
3.2.1	Doghead Mountain	60
3.2.1.1	Bedrock and surficial geology	65
3.2.1.2	Vegetation	67
3.2.1.3	Weather	67
3.2.2	Ouimet Canyon	69
3.2.2.1	Bedrock and surficial geology	76
3.2.2.2	Vegetation	76
3.2.2.3	Weather	77
3.3	Methods	80
3.3.1	Thermal regime of talus	80
3.3.2	Air temperature	88
3.3.3	Basal temperature of snow	90
3.3.4	Rock size measurements	92

3.3.4.1	Original method - Random direction, random distance	93
3.3.4.2	Modified method - Random transect, set distance	93
3.3.5	Geometry of the talus slopes	93
3.3.6	Timeline of field research	94
4	Results	96
4.1	Rock and pore sizes	96
4.2	Air temperature	102
4.2.1	Seasonality	108
4.3	Ground temperatures	110
4.3.1	Conditions at each site in talus	114
4.3.2	Ground conditions during the period of snow cover . .	121
4.3.2.1	Insulation of the surface from the atmosphere	125
4.3.2.2	Conditions in the ground	126
4.3.3	Effects of winter convection	134
4.3.4	Ground conditions during the summer	135
4.3.5	Effects of summer stratification	141
4.4	Ice accumulation and ablation in the talus	141
4.4.1	Melting, infiltration, and refreezing	142
4.4.2	Ablation	150
4.4.2.1	Interstitial ice and the zero curtain	150

4.4.2.2	Water accumulations	156
4.4.3	Effects of ice in the talus	156
4.5	Summary	159
5	Discussion	161
5.1	Research objectives and hypotheses	161
5.2	Thermal regime of the talus slopes	161
5.2.1	Does convective heat transfer exert significant influence on the thermal regime?	166
5.3	Permafrost in talus	171
5.3.1	Evidence for isolated patches of permafrost in talus in Northwestern Ontario	171
5.3.2	Field conditions that may indicate the presence of per- mafrost	172
6	Summary and Conclusions	174
6.1	Summary of results	174
6.1.1	Heat transfer mechanisms in the talus slopes	174
6.1.2	The effects of ice and water	175
6.1.3	Permafrost	176
6.2	Conclusions	176
6.3	Suggestions for future research	177

List of Figures

1.1	Zones of permafrost extent in northern Ontario	2
1.2	Talus slopes in Ouimet Canyon	3
2.1	Mean annual near surface ground thermal regime of permafrost	13
2.2	A generalized temperature envelope	18
2.3	An example of a ground thermal regime that experienced a zero-curtain	20
2.4	Mean monthly temperatures in adjacent loess and blocky ma- terials	31
2.5	Unicellular convection in an inclined rock layer	36
2.6	a) Rayleigh numbers and b) filtration Rayleigh numbers for varying ΔT from set parameters (Table 2.3).	40
3.1	Locations of Doghead Mountain and Ouimet Canyon, includ- ing the Cameron Falls and Thunder Bay weather stations . . .	48
3.2	Climographs for a) Cameron Falls station 1981-2010 and b) Thunder Bay A station 1971-2000	52

3.3	A section of the east side and bottom of Ouimet Canyon, showing the column-like structure of the rock that makes up the cliffs and the talus sheets that have formed from their disintegration. Photo by Jessica Abbott.	58
3.4	Topographic map of Doghead Mountain, near Nipigon, Northwestern Ontario, with the area of Fig. 3.5 indicated	61
3.5	Satellite photo of Doghead Mountain, with the area of Fig. 3.6 indicated by the rectangle	62
3.6	Satellite photo of the instrumented section of the talus sheet at Doghead Mountain	63
3.7	The westernmost section of south-facing talus slope of the Doghead ravine	64
3.8	A transect of the instrumented section of the Doghead talus sheet. Distances and slope angles are to scale	66
3.9	Mean monthly total precipitation at Cameron Falls, 1981-2010, and values measured during the study period.	68
3.10	Mean, mean maximum, and mean minimum daily temperatures plotted by month from Cameron Falls 1981-2010 and values measured during the study period at Cameron Falls.	70
3.11	Topographic map of Ouimet Canyon and surrounding area, with the area of Fig. 3.12 indicated	71

3.12	Satellite photo of Ouimet Canyon, with a) Fig. 3.13 and b) the section used for basal temperature of snow measurement outlined in white	72
3.13	Satellite photo of the instrumented part of Ouimet Canyon. See Fig. 3.15 for the slope transect	73
3.14	View north-northeast toward the study site at Ouimet Canyon.	74
3.15	A transect of the instrumented section of Ouimet Canyon. Distances and slope angles are to scale	75
3.16	Mean monthly total precipitation at Thunder Bay A, 1971-2000, and values measured at Thunder Bay CS during the study period.	78
3.17	Mean, mean maximum, and mean minimum daily temperatures plotted by month from Thunder Bay A 1971-2000 and values measured at Thunder Bay CS during the study period.	79
3.18	The layout of a) DHL and b) DHA	84
3.19	The layout of a) DHM and b) DHU near the middle and the top of the talus slope, respectively. See Fig. 3.6 for their relative locations.	85
3.20	The layout of OCA and OCL at the bottom of the canyon, near and in a mossy pit, respectively. See Fig. 3.13 for their relative locations.	86

3.21	The layout of a) OCM and b) OCU near the middle and top of the talus slope, respectively. See Fig. 3.13 for their relative locations.	87
4.1	The particle size distributions and Fuller curves of sections of the Doghead Mountain talus.	100
4.2	The pore size distribution of a) section 2-3, b) section 4, and c) section 4 without the outlier, of the Doghead Mountain talus.	103
4.3	Mean daily air temperature at Cameron Falls (T_{CF}) and Doghead Mountain (T_{DHA})	104
4.4	Mean daily air temperature at Thunder Bay CS (T_{TB}) and Ouimet Canyon (T_{OCA})	105
4.5	Mean daily air temperatures at Doghead Mountain (T_{DHA}) and Ouimet Canyon (T_{OCA})	107
4.6	Mean daily air temperatures at Doghead Mountain (DHA) and Ouimet Canyon (OCA) from 1 September 2016 to 31 August 2017.	109
4.7	Minimum and maximum daily mean temperatures for DHU, DHM, DHL, OCU, OCM, and OCL	120
4.8	Air and surface temperatures for the snow-covered season at Doghead Mountain.	123

4.9	Air and surface temperatures for the snow-covered season at Ouimet Canyon.	124
4.10	Examples of partial insulation from DHL-S and DHM-S	127
4.11	Temperatures at all DHU sensors in February 2017, showing the response of internal talus sensors to changes in temperature at the surface sensor.	128
4.12	Temperatures at a) all DHM sensors and b) all DHL sensors in January 2017, showing the response of internal talus sensors to changes in temperature at the surface sensor.	129
4.13	Temperatures at a) all OCU sensors and b) all OCM sensors in February 2017 showing the response of internal talus sensors to changes in temperature at the surface sensor.	130
4.14	Temperatures at all OCL sensors in January 2017 showing the response of internal talus sensors to changes in temperature at the surface sensor.	131
4.15	Responses of some internal talus sensors at a) Doghead Mountain and b) Ouimet Canyon to air temperatures in January 2017.	132
4.16	Air and internal talus temperatures at a) DHU and b) DHM for the first two weeks of July 2017	136
4.17	Air and internal talus temperatures at a) OCU and b) OCM for the first two weeks of July 2017	137
4.18	Air, surface, and internal talus temperatures at a) DHL and b) OCL for the first two weeks of July 2017	138

4.19	Surface and internal talus temperatures at DHL and precipitation at Cameron Falls for August 2017.	139
4.20	Surface and internal talus temperatures at OCL and precipitation at Thunder Bay CS for August 2017.	140
4.21	Temperatures for several sensors at Doghead Mountain from 16 February to 3 March 2017, showing possible meltwater infiltration	143
4.22	Temperatures for several sensors at Ouimet Canyon from 16 to 27 February 2017, showing possible meltwater infiltration .	145
4.23	Temperatures for several sensors at Doghead Mountain from 4 to 12 March 2017, showing possible meltwater infiltration . .	146
4.24	Temperatures for several sensors at Ouimet Canyon from 4 to 12 March 2017, showing possible meltwater infiltration	147
4.25	Temperatures at a) Ouimet Canyon and b) Doghead Mountain from 1 to 9 April 2017. Some upper sensors coupled with the atmosphere, while the lower sections showed possible meltwater infiltration	148
4.26	Temperatures for several sensors at Ouimet Canyon in late April and early May 2017, showing possible meltwater infiltration	149
4.27	Interstitial ice at Doghead Mountain	152
4.28	Interstitial ice at Ouimet Canyon	153

4.29	The temperature record for OCL-1.5 from 1 October 2016 until 17 August 2018	155
4.30	Interstitial ice at Ouimet Canyon on 11 October 2017	157
4.31	Standing water ponded in the talus in the southern part of Ouimet Canyon, 7 June 2017	158
5.1	Ice formation from the infiltration and refreezing of meltwater in the talus in spring	165
5.2	Wintertime air movements in the talus slopes	168

List of Tables

2.1	General morphology, slope angle, and sediment transfer processes of different kinds of kurums	23
2.2	Elevation, T_a and T_g of ice caves and blocky materials that contain permafrost	27
2.3	The parameters used to calculate the Rayleigh numbers in Fig. 2.6	41
3.1	Location and data available for the four climate stations used in this study.	49
3.2	Monthly climate normals for Cameron Falls, 1981-2010	51
3.3	Monthly Climate Normals for Thunder Bay A, 1971-2000 . . .	53
3.4	Names, sensor names, and installation date and time for the air and ground temperature loggers	81
3.5	Distances and slope angles between data logger sites	83
4.1	Means and standard deviations of the a , b and c axes, the nominal diameter, and the nominal volume of rocks in different sections of the Doghead Mountain talus slope	97

4.3	Annual, freezing season, and thawing season temperatures for air and lower, middle and upper talus sites	111
4.4	Mean freezing degree-days (FDD) and thawing degree-days (TDD) for air and talus temperature sensors	113
4.5	Annual mean temperatures for each logger	115
4.6	Freezing season mean temperatures for each sensor	116
4.7	Thawing season mean temperatures for each sensor	117
4.8	The estimated minimum period of snow cover, based on daily surface temperature ranges and trends at each logger	122
4.9	The estimated thawing zero curtain period for each sensor in the lower section of both talus slopes in 2016 and 2017	154

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Chapter 1

Introduction

1.1 Introduction

The purpose of this thesis is to investigate anomalously cold ground and occurrences of summertime ice in talus in Northwestern Ontario. Talus, sometimes called scree or debris, is composed of rock fragments that are derived from and lie at the bottom of a steep rock slope. The term may also refer to a heap or slope of such material (Bates & Jackson, 1980). Talus and other blocky materials are known to create ground thermal anomalies, and in some cases cause permafrost at elevations or latitudes where it is not present in fine-grained soils (Harris & Pedersen, 1998). The study sites are located approximately 400 km south of the established permafrost zones (Fig. 1.1), specifically near lakes Superior and Nipigon. In this region, talus can be found at the base of cliffs associated with diabase canyons, mesas, and cuestas (Fig. 1.2).

Local residents are aware of thermal anomalies and summer ground ice at

Fig. 1.1: Zones of permafrost extent in northern Ontario. The saturation of the permafrost classes is opposite to cartographic convention in order to make the small areas of isolated patches in Gaspé and Newfoundland visible. The study area of this thesis is outlined by the black square on the large scale map. Produced using data from <https://www.nrcan.gc.ca/earth-sciences/geography/topographic-information/free-data-geogratis/11042>.

Fig. 1.2: Talus slopes in Ouimet Canyon, near Thunder Bay, Northwestern Ontario. The view is to the southwest from the middle of the eastern slope.

locations across the region. “Ice caves” and ice-bound talus persisting well into summer in talus slopes next to mesas and cuestras have been described at Stewart Lake, Ruby Lake, Kama Ridge, Doghead Mountain, and the Nipigon Palisades (Chaboyer, H., 2011, pers. comm.). Wilkins (1994) noted that boulders at the bottom of Ouimet Canyon have a thick layer of year-round ice beneath them, which has been confirmed by hikers who entered the canyon before it was closed to the public in the 1970s (Bishop, J., 2009, pers. comm.). In 2011, cold thermal anomalies and interstitial ice were found in talus slopes in the region (Bishop-Legowski, 2011).

The presence of anomalously cold ground and ice in talus slopes in North-western Ontario is important because there are many existing and proposed mining projects in the region which are associated with accumulations of waste rock. Talus slopes and other blocky materials can be analogous to these waste rock piles. The stability of mine sites and their associated waste and byproducts are important for environmental and human health, so it is important to understand the processes that affect the temperature and drainage of blocky materials in the region. Furthermore, many sites with anomalously cold ground and ice are protected areas because they support cold-tolerant ecosystems and provincially rare species. These ecosystems exist only in places where boreal species cannot compete well, so the persistence of these cold environments is important to their survival.

1.2 Isolated permafrost incidence

The field portion of this study focuses on two locations where ice was found in talus in summer 2011 (Bishop-Legowski, 2011). The first site is a talus slope in an east-west ravine that rises to the top of Doghead Mountain, near Nipigon, Ontario. The talus slope is on the north side of the ravine, but is shaded during part of the day by a cliff opposite it. The bottom of the talus, where the ice was found, is covered in thick mats of moss and lichen and is further shaded by the forest that has grown along its base. The second site is Ouimet Canyon, a talus-filled canyon running mainly north-south. The talus undulates along the canyon length, with the deeper parts supporting thick moss and lichens and some stunted trees. Ice was found at one of these low points in the canyon in 2011.

Permafrost distribution in Canada is mostly confined to parts of the country where mean annual air temperature (T_a) is at or below -1°C (Brown, 1969). Typically, T_a must be less than 0°C for permafrost to exist because mean annual ground surface temperatures (T_s) are generally higher than T_a , a relation called the surface offset (Smith & Riseborough, 2002a).

There is permafrost outside Arctic and subarctic areas in places where high elevation decreases T_a . Permafrost is known or predicted to exist in the Western Cordillera as far south as Mexico (Pewe, 1983), in Gaspé's Chic-Choc Mountains at 48°N (Gray & Brown, 1981), in Newfoundland's Long Range Mountains at 50°N (Brown, 1979), and in the United States' White

Mountains as far south as 44°N (Walegur & Nelson, 2003). The study area in Northwestern Ontario is different from these regions because neither is it far enough north, nor are the mesas and cuevas sufficiently elevated to result in T_a less than 0°C. T_a in the region ranges between 2.5°C at Thunder Bay and 1.7°C at Cameron Falls (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html), values which are similar to or lower than locations where permafrost has been found in blocky materials (Delaloye & Reynard, 2001; Gude *et al.*, 2003; Gorbunov *et al.*, 2004; Holmgren *et al.*, 2017a).

In continuous permafrost, the low T_a ensures that in spite of the surface offset, T_s is negative and permafrost is found under 90% or more of the surface. In discontinuous permafrost, this is not always the case, so the distribution of permafrost in discontinuous zones is also driven by other factors: surface conditions, such as snow depth and vegetation, and thermal conductivity of the ground materials become critical influences on the presence of permafrost (Smith & Riseborough, 2002a). Blocky materials affect the temperature of the ground because they change the way heat is transferred through the ground; convection, rather than conduction, may exert significant influence on heat transfer in the void spaces of the blocky materials. Within the voids between rock fragments, air may move into, out of, and among the blocky materials. Ventilation of blocky materials in winter may lower ground temperatures sufficiently to preserve ice or frozen ground through the summer (Harris & Pedersen, 1998). Thus positive T_a

may be sufficiently overcome by the cooling effect of the blocky materials to create highly localized negative thermal anomalies sufficient to create permafrost. Thermal anomalies occur in blocky materials compared to adjacent rock or fine-grained soils (Balch, 1900; Harris & Pedersen, 1998; Juliussen & Humlum, 2008). Such environments include block fields, block streams, and talus slopes. In many mountainous regions the lowest permafrost is found in blocky materials, both as ice masses in talus and as rock glaciers, or in ice caves (Barsch, 1978; Gorbunov, 1978; Harris, 1979). Blocky materials also extend permafrost occurrence into non-alpine uplands with relatively low elevations (Gray & Brown, 1979, 1981; Gude *et al.*, 2003; Putnam & Putnam, 2009). This thesis project sought evidence of convection and anomalously cold ground in talus slopes in Northwestern Ontario, where the relief is less than 400 m, and the nearest known permafrost 400 km to the north.

1.3 Objectives

The objectives of this research were: 1) to determine the thermal regime of talus at two sites in Northwestern Ontario over at least one year; and 2) to determine the field conditions that indicate permafrost may be present in talus in Northwestern Ontario.

1.4 Hypotheses

This thesis has two principal hypotheses: 1) there are isolated patches of permafrost in talus in Northwestern Ontario; and 2) convective heat transfer exerts significant influence on the thermal regime within the talus.

1.5 Methods

In order to record the thermal regime of the talus, surface and internal temperature sensors were installed at the top, middle and bottom of two talus slopes. The basal temperature of the snow pack was measured above the talus to determine the spatial extent of thermal anomalies. Rock sizes in the talus were measured in order to describe the structure of the talus and relate it to the thermal regime.

1.6 Thesis structure

The thesis is presented in six chapters. Chapter two presents the factors affecting the presence of permafrost in areas where it is spatially and temporally discontinuous. The chapter also details the geomorphology of talus slopes, and the dynamics of convection in blocky materials which are thought to create thermal anomalies. The third chapter begins with an overview of the research locations' climate, geology, and geomorphology, and describes

the methods used for each stage of the research. Chapter four contains the findings related to the spatial distribution of the thermal anomalies and the observed talus thermal regime. The fifth chapter discusses the results, while the last chapter summarizes the results and concludes the thesis with a discussion of the implications of this research and further avenues for inquiry.

Chapter 2

Background

This chapter considers permafrost, the factors affecting the distribution of permafrost, and the thermal regime of the ground in permafrost terrain. Thereafter, it describes blocky materials and related landforms. Finally, it provides an introduction to heat movement in blocky materials and their thermal regimes.

2.1 Permafrost

The definition of permafrost used in Canada is ground “that remains at or below 0°C for at least two years” (Harris *et al.*, 1988, 63). Permafrost is overlain by a layer of ground that freezes and thaws seasonally, called the active layer (Harris *et al.*, 1988). In Canada, permafrost is found both at high latitudes and high elevations. About half of the country’s landmass lies in the permafrost regions (Fig. 1.1) (Heginbottom *et al.*, 1995).

Since the late 1960s, Canadian permafrost maps have been based on field data where available (Heginbottom, 1984); however, “mapping of per-

mafrost and ground ice requires considerable interpretation and interpolation” (Heginbottom, 1984, 81) due to the limited amount, uneven distribution, and mixed quality of field evidence (Heginbottom, 1984; Heginbottom *et al.*, 1993; Smith & Riseborough, 2002b). The 1995 Map of Permafrost in Canada (Fig. 1.1) assigned permafrost distribution based on known or assumed ground conditions (Heginbottom, 1984; Heginbottom *et al.*, 1995). These categories indicate the areal extent of permafrost, be it continuous (90-100%), extensive discontinuous (50-90%), sporadic discontinuous (10-50%), or isolated patches (0-10%) (Heginbottom *et al.*, 1995). The resulting zones of permafrost broadly correspond to the distribution of air temperature isotherms, with the southern limit of continuous permafrost between the -6°C and -8°C isotherms, and the southern limit of discontinuous permafrost at the -1°C isotherm (Smith & Riseborough, 2002a).

2.1.1 Factors affecting the distribution of permafrost

Mean annual air temperature (T_a) controls mean annual ground temperature (T_g) at a regional scale, so permafrost distribution across North America is a climatic phenomenon (Smith & Riseborough, 2002b; Burn, 2004). Ground surface conditions, such as slope, aspect, vegetation, moisture, and snow cover cause localized variation in the ground surface temperature in space and time (French, 2007). T_s , the mean annual ground surface temperature, and the mean annual temperature at the top of the permafrost (T_p) vary as a result. In continuous permafrost, variations in T_p cause changes to

the thickness of permafrost. In discontinuous permafrost, these near-surface characteristics also cause permafrost to exist in some places and times and not in others.

The difference between T_a and T_s is called the surface offset, and the difference between T_s and T_p is called the thermal offset (Fig. 2.1) (Romanovsky & Osterkamp, 1995; Smith & Riseborough, 2002a). T_s is usually higher than T_a , principally due to the thermal effects of snow and, where the plant canopy is thin, insolation (French, 2007). Under equilibrium conditions, T_s is also usually higher than T_p because of seasonal differences in the thermal conductivity of the active layer (Romanovsky & Osterkamp, 1995; Smith & Riseborough, 2002b). The shape of the temperature curve in Fig. 2.1 is applicable to many sites with permafrost.

The thermal conductivity (λ , $Wm^{-1}K^{-1}$) is the ability of a material to conduct heat (Williams & Smith, 1989). Where ground materials are granular or organic, the pore spaces between particles may be filled with air or water. Since the thermal conductivity of ice is four times that of water (Smith & Riseborough, 2002b), the active layer is more conductive in winter than summer in ground materials with significant water contents. Not all active layers exhibit the change in thermal conductivity that creates the characteristic shape of the curve in Fig. 2.1. Dry soil and solid rock contain little water. In dry materials or where there is little insulating snow cover, T_p and T_s are similar and the thermal offset is small, or may even be reversed (Smith & Riseborough, 2002a; Lin *et al.*, 2015).

Fig. 2.1: Mean annual near surface ground thermal regime of permafrost. Reprinted with permission from Smith & Riseborough (2002a), pg. 5, Fig. 3. ©2002 by John Wiley and Sons. In the notation used in this thesis, MAAT is T_a , MAGST is T_s , and TTOP is T_p .

The importance of T_p and the thermal conductivity of the ground materials for the thickness of permafrost is illustrated by equation 2.1. The thickness of permafrost (z , in meters) can be calculated where λ is the thermal conductivity, $|T_p|$ is the absolute value of the mean annual temperature at the top of the permafrost ($^{\circ}C$), and Q is the geothermal heat flux (Wm^{-2}) (Williams & Smith, 1989). Where $|T_p|$ or λ is larger, z will be larger.

$$z = \frac{\lambda|T_p|}{Q} \quad (2.1)$$

Equation 2.1 assumes heat flow in the ground is by conduction alone, but non-conductive heat transfer due to the movement of water or water vapour may also occur. Hinkel & Outcalt (1994), Hinkel *et al.* (1997), and Romanovsky & Osterkamp (2000) showed that gravity-driven infiltration of meltwater and rainwater may advect heat into the active layer. Hinkel & Outcalt (1994) also showed that movement of liquid water within soils may be driven by solute concentration and temperature gradients. Evaporation, vertical movement of water vapour, and condensation within the soil (“internal distillation”) may also be driven by similar gradients (Hinkel & Outcalt, 1994). These non-conductive heat flows may have significant short term effects on the rates of temperature change in the active layer (Hinkel & Outcalt, 1994; Hinkel *et al.*, 1997), but are commonly ignored at longer time scales (Romanovsky & Osterkamp, 2000).

Deming *et al.* (1992) showed that the movement of groundwater may also

be important for heat flow to the base of permafrost. In parts of a basin where groundwater recharge is taking place, heat is advected downward, depressing the geothermal gradient and surface heat flow. In places where basin geometry forces groundwater to move upward, heat is also carried upward. This exaggerates the geothermal gradient and surface heat flow, resulting in upward movement of the base of permafrost and a reduction in permafrost thickness.

Even considering these instances of non-conductive heat flow, the climate- and conduction-based models of permafrost distribution are useable at a continental scale (Henry & Smith, 2001). The study area of this thesis, however, is far outside the zones where permafrost is expected according to such methods (Fig. 1.1).

2.1.2 Factors affecting the thermal regime of the ground

The thermal regime concerns the distribution of heat fluxes and temperature in the ground over time (Harris *et al.*, 1988). The energy fluxes at the ground surface can be represented by the surface energy balance (equation 2.2) (Williams & Smith, 1989; Burn, 2004).

$$Q^* = Q_H + Q_E + Q_G \quad (2.2)$$

Q^* is the net radiation, representing incoming radiation, less the radiation emitted or reflected. Q_H is the sensible heat flux between the ground and

atmosphere, which depends on convection in the surface boundary layer for heat transfer (Williams & Smith, 1989). Q_E is the evaporative or latent heat flux, and depends on the availability of moisture at the ground surface for evaporation, or in vegetation for transpiration, and convection in the surface boundary layer (Burn, 2004). Q_G is the ground heat flux, and represents heat flow to or from the ground surface within the ground (Williams & Smith, 1989).

Ground surface conditions such as slope, aspect, moisture, snow cover, and vegetation type, height and density cause variation in Q^* , Q_H , and Q_E . Q_G varies depending on snow cover and the thermal conductivity of the ground materials. Variation of these ground surface conditions at a landscape scale leads to unique combinations of fluxes and localized ground surface temperatures (Burn, 2004), causing variation in permafrost depth and distribution. In the discontinuous permafrost zones, these variations may cause the presence or absence of permafrost at a site. Changes to fluxes occur on sub-hourly to seasonal time scales and lead to adjustments among the other fluxes in response (Williams & Smith, 1989).

At sites with low albedo and low Q_E due to lack of moisture, Q_H and the ground surface temperature (T_s) will be high. In locations with peat, Q_G will depend on soil moisture because the thermal conductivity of the peat layer changes with water content, depending on whether it is dry ($0.05 \text{ W m}^{-1} \text{ K}^{-1}$), saturated ($0.50 \text{ W m}^{-1} \text{ K}^{-1}$), or saturated and frozen ($2.00 \text{ W m}^{-1} \text{ K}^{-1}$) (Burn, 2004; French, 2007).

2.1.2.1 Heat movement through the ground

As discussed above, in many environments heat is transferred through the ground conductively. The heat capacity of the ground materials regulates changes in ground temperature derived from the heat flow. The specific heat capacity, C_s ($Jkg^{-1}K^{-1}$) describes the amount of heat required to raise the temperature of a unit mass of a material by 1 K (Williams & Smith, 1989). The ratio of thermal conductivity to volumetric heat capacity (λ/C_v) is the thermal diffusivity (κ , m^2s^{-1}), a measure of how well a temperature change may propagate through a material.

The effects of thermal diffusivity on the thermal regime may be seen with the temperature envelope (Fig. 2.2). The outside edges of the envelope represent the minimum and maximum ground temperatures at any given depth below the surface. The difference between the minimum and maximum temperatures represents the annual range of ground temperature. If the thermal diffusivity of the ground materials is high, the seasonal temperature variations at the surface propagate further into the ground, the temperature envelope is wide, and the depth of zero annual amplitude increases (Williams & Smith, 1989).

2.1.2.2 Latent heat and the zero-curtain

The latent heat of fusion is important to the ground thermal regime of freezing or thawing moist soils because it determines the change in energy required

Fig. 2.2: A generalized temperature envelope of an environment in a state of equilibrium (the ground and atmospheric climates are not perturbed or changing). The scales on each axis are for illustrative purposes only.

to accomplish the change of phase. While freezing is occurring in autumn, the temperature of the ground may change little, producing a period called the zero curtain. During the zero curtain, the ground temperature remains near the freezing point for days or weeks (Fig. 2.3). It is a common feature of the thermal regime of the active layer, but may also occur in seasonally frozen ground without permafrost (Harris *et al.*, 1988). The zero curtain is longer in soils with high water content or when the ground is covered by thick snow (Williams & Smith, 1989). Temperature change below and within the active layer resumes once the phase change is complete for most of the water (Williams & Smith, 1989). At this point, the heat flow is evident in changes to the sensible heat content of the ground. The zero curtain delays the ground temperature's response to seasonal air temperature changes (Romanovsky & Osterkamp, 2000).

While the zero curtain is especially pronounced during the freezing season (Williams & Smith, 1989; Romanovsky & Osterkamp, 2000), it may also occur during the thawing season. When freezing (in autumn), water molecules need to lose sufficient energy to build a lattice and solidify, so change in temperature is delayed until sufficient latent heat loss has occurred from the ground. When thawing (in spring), the molecules need to gain sufficient energy to break the lattice and liquify, so change in temperature is delayed until sufficient heat has been gained by the ground.

Fig. 2.3: An example of a ground thermal regime that experienced a zero-curtain at 0.6m in October and November, from Barrow, Alaska. Reprinted with permission from Williams & Smith (1989), pg. 106, Fig. 4.8. ©1989 by Cambridge University Press.

2.2 Blocky materials

Blocky or coarse materials are pieces of rock ranging from cobble to boulder size. The physical properties of such large materials in bulk are not well explored. The nearest analog with established physical properties is rock fill used in roads, dams, air convection embankments (ACE), and other engineering structures. There are limits to the analogy because large boulders (>50 cm) are typically removed from rock fill (Pusch, 2008).

The physical properties of rock fill such as porosity, compactability, and hydraulic conductivity depend on particle size and shape, which are determined by the structure of the parent rock (Maynard & Neill, 2008; Pusch, 2008). The particle size distribution of a non-engineered material is usually normal or log-normal (Bunte & Abt, 2001), and the cumulative frequency distribution can be described by a Fuller curve (equation 2.3),

$$p = 100(d/D)^m \quad (2.3)$$

where p is the percent of material by mass passing through a sieve of diameter d (mm), D is the maximum particle size (mm), and m is a variable that describes the range of particle sizes (Peronius & Sweeting, 1985; Latham *et al.*, 2002). The curve describes the influence of particle size on the density of a granular material. Materials with an m value of 0.5 are described as “ideal” because the particles pack most efficiently, minimizing porosity

(Peronius & Sweeting, 1985; Latham *et al.*, 2002). Such materials can be engineered. Graded riprap, which can contain pieces of rock up to 0.9 m, usually has a porosity of 25 to 35% (Maynard & Neill, 2008).

Models relating m to porosity have been developed for coarse materials, but agreement between them is poor (Latham *et al.*, 2002). Arya & Paris (1981) and Arya *et al.* (1999) relate particle size distribution to pore size distribution, which is important because the geometry and interconnectivity of the void spaces controls fluid transmission through the blocky materials (Latham *et al.*, 2002). It must be noted, however, that the Arya-Paris model was calibrated on soils, and assumes spherical particles and cylindrical pores.

The porosity and other physical properties of rock fill also depend on the extent of compaction (Pusch, 2008). Blocky materials may experience compaction under their own weight (self-compaction). As a result, porosity decreases with depth (Scapozza *et al.*, 2015). Periodic water flow may aid compaction by removing fine particles that prevent larger cobbles and boulders from coming into close contact (Pusch, 2008).

Landforms made up of blocky materials without a fine sediment matrix near the surface include blockfields (felsenmeer), block streams, block slopes, rock glaciers, talus cones, and talus sheets. These kinds of features, excluding rock glaciers, are collectively called *kurums* (Harris *et al.*, 1998; Kotlyakov & Komarova, 2007). Different types of kurums are classified based on morphology, slope angle, and sediment transfer processes (Table 2.1). Regardless of the distinctions between the them, they have all been found to harbour

Kurum type	Morphology	Slope angle	Sediment transfer process
Blockfield	Relatively flat fields of blocks	$< 5^\circ$	Frost heave; solifluction, frost creep, boulder flow or debris flow followed by fines removal by slope wash
Block stream	Narrow band of blocks trending downslope	$> 5^\circ$ and $< 10^\circ$	Rolling/sliding when lubricated by ice; solifluction followed by removal of fines
Block slope	Extensive fields of blocks on slopes	$> 5^\circ$ and $< 10^\circ$	Rolling/sliding when lubricated by ice; frost heave
Talus cone	Conical debris pile at the base of a cliff	$> 25^\circ$	Dry grain flow; frost-coated clast flow; debris flow; snow or slush avalanches; wind redistribution
Talus sheet	More-or-less planar debris pile along the base of a cliff	$> 25^\circ$	Dry grain flow; frost-coated clast flow; debris flow; snow or slush avalanches; wind redistribution

Table 2.1: General morphology, slope angle, and sediment transfer processes of different kinds of kurums from van Steijn *et al.* (2002); definitions of talus cones and sheets from Gerrard (1990).

permafrost or cold ground thermal anomalies. Frost shattering and thermal expansion are thought to be the most important weathering processes in breaking the parent material into blocks, although chemical weathering may also be effective (van Steijn *et al.*, 2002).

2.2.1 The geomorphology of talus slopes

Talus is both “rock fragments of any size or shape (usually coarse and angular) derived from and lying at the base of a cliff or very steep rock slope” and “the outward sloping and accumulated heap or mass of such loose broken rock, considered as a unit” (Bates & Jackson, 1980, 638). In this thesis “talus” or “blocks” will refer to the fragments, and “talus slope” to the mass unit.

Talus slopes are usually at the angle of rest for talus (31° to 33°) or oversteepened (van Steijn *et al.*, 2002). Slopes near the angle of rest are dominated by rockfall processes. Large rocks have the momentum to tumble to the bottom of the slope, while smaller clasts are caught among the larger blocks (Bones, 1973).

Steeper slopes may experience dry grain flow. Flow begins on the upper slopes where the angle is greater than 38° and ends in the part of the slope where the angle decreases to less than $34\text{--}35^\circ$. The resulting slope is made up of beds of fine gravel where the flows originate, coarser openwork materials below this, and finer matrix-rich beds near the bottom (van Steijn *et al.*, 2002). When temperatures are near 0°C and the clasts are coated with ice,

the clasts are able to continue moving on the sections of the slope with angles of 24-26°. Where this frost-coated flow dominates, channelized movement creates long, crudely stratified sections of the slope with angles of 25-30° (van Steijn *et al.*, 2002).

Van Steijn *et al.* (2002) note that in places with strong winds, long periods of subzero temperatures, and very friable rock, aeolian erosion and sediment transport can be important to talus formation. Talus slopes with extensive aeolian transport will have a concave base made up of finer sediments than the rest of the slope, with an angle below 25°. These forms may be found worldwide, including the Canadian Arctic and Gaspé, but are not characteristic of the sites investigated in Northwestern Ontario (van Steijn *et al.*, 2002).

2.3 Thermal anomalies and permafrost in blocky materials

A peculiarity of landforms made up of blocky materials is that many have been found to harbour comparatively cold air and cold ground temperatures relative to their surroundings. Such instances have been documented and quantified in rock glaciers and kurums of all kinds.

2.3.1 Observations from around the world

Several papers have indicated differences between the thermal regimes of blocky materials and conductive materials. These differences lead to the presence of permafrost at relatively low elevations in Canada, Switzerland, Germany, and the Czech Republic (Harris & Pedersen, 1998; Delaloye & Reynard, 2001; Gude *et al.*, 2003), and in places where T_a is positive (e.g., Sawada *et al.*, 2003; Zacharda *et al.*, 2007). The differences also lead to decreases in ground temperatures of 1-7°C where permafrost is already known to exist (Harris *et al.*, 1998; Juliussen & Humlum, 2008). In many regions, the activity and ice content of rock glaciers have been used to define past and present extent of discontinuous permafrost (Barsch, 1978; King & Kalisch, 1998; Aoyama, 2003; Hétu *et al.*, 2003; Putnam & Putnam, 2009; Boeckli *et al.*, 2012). In North America, cold thermal anomalies in blocky materials have been confirmed as far south as California (Millar *et al.*, 2014). Table 2.2 contains the elevation, T_a and T_g of examples of ice caves and blocky materials that contain permafrost from around the world.

These observations come not only from alpine environments where blocky materials and periglacial conditions are widespread, such as the Alps and the Western Cordillera, but also from more temperate regions with gentler relief. In eastern North America, permafrost and thermal anomalies have been found in Gaspé, Maine, and Pennsylvania, where conditions favouring active rock glaciers have not existed for at least two thousand years (McGerrigle,

Location	Type	Elevation (m a.s.l.)	T_a (°C)	T_g (°C)	Difference $ (^{\circ}\text{C}) $	Source
<i>Europe</i>						
Jura Mtns., Switzerland	<i>kurum</i>	1170-1290	5.5	0.5-2	3.5-5	Delaloye & Reynard (2001); Delaloye <i>et al.</i> (2003)
Valais Alps, Switzerland	active protalus ramparts	2600-2800	-1	$\approx$ -0.5*	$\approx$ 0.5*	Scapozza <i>et al.</i> (2011)
Western Switzerland, incl. Jura Mtns. and Valais Alps	<i>kurums</i> , relict rock glaciers	770-1900	3.5-9.5	-0.59-3.29	2.06-8.35	Morard <i>et al.</i> (2008, 2010)
Bever, eastern Swiss Alps	<i>kurums</i>	1790-1900	1	-2.1	3.1	Kneisel (2010)
Harz Mtns., Germany	<i>kurum</i>	600	6.2	1.6	4.6	Gude <i>et al.</i> (2003)

Table 2.2: Elevation, T_a and T_g of ice caves and blocky materials that contain permafrost. The height and depth, respectively, at which T_a and T_g were measured varied. Where more than one value of T_g was available, the shallowest measurement in blocky materials at the lowest elevation was selected. * indicates estimated values based on tables or figures.

Location	Type	Elevation (m a.s.l.)	T_a (°C)	T_g (°C)	Difference $ (^{\circ}\text{C}) $	Source
<i>Europe cont.</i>						
Lusatian Mtns., Czech Republic	<i>kurums</i> , ice cave	300-580	6-8	-1.2-0.1	7.1-7.3	Gude <i>et al.</i> (2003); Zacharda <i>et al.</i> (2007)
Apuseni Mtns., Romania	ice cave	1165	-	-0.9	≈ 8	Holmlund <i>et al.</i> (2005)
Elgahogna and Sølén, eastern Norway	<i>kurums</i>	1043-1420	-1.1-0.9	-0.9-2.2	0.2-3	Juliusen & Humlum (2008)
<i>North America</i>						
Plateau Mtn., Alberta	<i>kurum</i>	1950	-1.17	$\approx -5.5^*$	$\approx 4.3^*$	Harris & Pedersen (1998)
Sierra Nevada, California	<i>kurums</i>	3100-3300	-0.2-3.7	$\approx 3-7.6^*$	$\approx 2-4^*$	Millar <i>et al.</i> (2014)
White Mountains, New Hampshire	<i>kurum</i>	668-689	4.83	-3.75	8.58	Holmgren <i>et al.</i> (2017a)

Table 2.2 (cont.): Elevation, T_a and T_g of ice caves and blocky materials that contain permafrost. The height and depth, respectively, at which T_a and T_g were measured varied. Where more than one value of T_g was available, the shallowest measurement in blocky materials at the lowest elevation was selected. * indicates estimated values based on tables or figures.

Location	Type	Elevation (m a.s.l.)	T_a (°C)	T_g (°C)	Difference $ \text{°C} $	Source
<i>Asia</i>						
Kunlun Pass, Tibetan Plateau	<i>kurum</i>	4800	-6	-2.6	3.4	Harris (1996)
Mt. Fuji, Japan	ice cave	1120	8.4	<0*	>8*	Ohata <i>et al.</i> (1994)
Yari-Hotaka Mtns., Japan	inactive rock glaciers	2360-2890	-2.5	-1.2-1.8	1.3-4.3	Aoyama (2003)
Transili Alatau Range	<i>kurums</i>	1800-2550	1.38-5	eg.-1.38	2.5-4	Gorbunov <i>et al.</i> (2004)

Table 2.2 (cont.): Elevation, T_a and T_g of ice caves and blocky materials that contain permafrost. The height and depth, respectively, at which T_a and T_g were measured varied. Where more than one value of T_g was available, the shallowest measurement in blocky materials at the lowest elevation was selected. * indicates estimated values based on tables or figures.

1952; Gray & Brown, 1981; Hétu *et al.*, 2003; Lendemer *et al.*, 2009; Putnam & Putnam, 2009).

2.3.2 Thermal regimes in kurums

Harris & Pedersen (1998) compared the mean monthly temperatures of a soil and an adjacent block slope, illustrating thermal regime characteristics that can be observed in other conductive and non-conductive systems. In the soil (Fig. 2.4, left), the near-surface temperatures showed a marked seasonal pattern that closely follows the air temperature. With increasing depth, a lag of up to a few months developed in minimum and maximum temperatures. The range of ground temperatures was also dampened with depth. The dampening of annual amplitude and development of a lag in the annual cycle with depth are characteristics of a ground thermal regime dominated by conduction.

In the block slope (Fig. 2.4, right), the near surface temperatures also showed a marked seasonal trend, but the occurrence of minimum ground temperatures and warming of the ground in spring were coincident irrespective of depth. Some lag in maximum temperatures and autumn cooling was observed, but the lag time was reduced. Some dampening of the ground temperature amplitude was observed with depth, but the ground at 75 cm depth still experienced a range of about 6°C while the ground at 67 cm depth in the loess experienced a range of about 1°C. The reduced dampening and reduced or absent lag are characteristics of a ground thermal regime where

Fig. 2.4: Mean monthly temperatures in adjacent loess and blocky materials on Plateau Mountain, Alberta. Reprinted with permission from Harris & Pedersen (1998), pg. 112, Fig. 4. ©1998 by John Wiley and Sons.

the thermal diffusivity is high or non-conductive heat flow occurs.

It should also be noted that in spite of the sites being adjacent, ground temperatures year-round at the block slope were lower than at the soil site. At the block slope, the ground did not thaw from 27 cm downward, while the entire profile in the soil site had thawed by mid-summer. This is an example of a thermal anomaly commonly observed in kurums. Such anomalies also occur on shorter time scales; Harris & Pedersen (1998) show that the block slope responds throughout the profile to changes in air temperature almost immediately, but this did not occur at the soil site.

There is some variety in zero curtain behaviour in kurums. Some sites exhibit a zero curtain only in spring, in contrast to its usual appearance in autumn (Juliussen & Humlum, 2008); and others experience a zero curtain in both spring and autumn, with its duration in spring being longer (Sawada *et al.*, 2003; Zacharda *et al.*, 2007).

2.4 Heat movement in blocky materials

There are several mechanisms thought to contribute to the thermal anomalies observed in blocky materials. These include changes to heat conduction, free and forced convection, and thaw of ice within the talus.

2.4.1 Changes to conductive heat transfer

Blocky materials are characterized by large particles and correspondingly large void spaces between them. The surface area of the blocks exposed to air, which is a relatively poor thermal conductor, can be large, and the contact area between blocks is small. Gruber & Hoelzle (2008) found that as a result, conduction is reduced by an order of magnitude compared to solid rock. Not only does this restrict the penetration of heat into the ground in summer, but it also reduces the heat flux from the bottom of the blocky layer to the bottom of the snow in winter. This may account entirely for some thermal anomalies where snow cover prevents air exchange between the blocky materials and the atmosphere (Gruber & Hoelzle, 2008).

2.4.2 Convective heat transfer

In addition to reduced heat conduction, blocky materials may experience increased non-conductive heat flow because air and water may easily circulate in the voids among the blocks (Harris & Pedersen, 1998).

2.4.2.1 Free convection

Free convection of air is driven by differences in density. Heat from a warm surface is transferred by radiation and conduction to a thin layer of air adjacent to it (Doré & Zubeck, 2009). This air parcel's density decreases relative to the surrounding air and it rises. When a parcel of air loses heat, its density

increases and it sinks. When a parcel of air is displaced upward or downward, it is replaced by another one. In a closed, level system, with a warm surface beneath a cold surface, a circulation develops between the surfaces. Air that has gained heat at the warm surface rises and loses that heat to the cold surface. It then sinks back to the warm surface, producing one or more convection cells (Narasimhan, 1999). This is the principle used to cool air convection embankments in winter (Goering & Kumar, 1996; Goering, 1998; Cheng *et al.*, 2007). A block field completely covered with snow in early winter, where the warm surface is the bottom of the blocky layer or the top of permafrost and the cold surface the bottom of the snow pack, may be another example of such a closed, level system (Cheng *et al.*, 2007). Heat transfer by convection is more efficient than by conduction, so convective heat transfer may lead to substantial heat loss from the ground to the atmosphere during the winter.

If the warm surface is above the cold one (e.g. in the summer), the air at the bottom of the blocky layer is colder and denser than the air near the surface, so the air is stable and convection does not take place. Instead, the air within the blocky materials becomes thermally stratified (Goering & Kumar, 1996; Goering, 1998; Jiang, 2004; Cheng *et al.*, 2007). Heat transfer then takes place by conduction through the air and through the limited contact area between blocks (Cheng *et al.*, 2007).

This pattern of wintertime convective cooling and summertime thermal stratification has been numerically modelled (Goering & Kumar, 1996; Jiang,

2004; Cheng *et al.*, 2007), and observed in embankments made of different-sized rock fill (Goering, 1998; Cheng *et al.*, 2007). These air convection embankments are built over permafrost sensitive to thermal disturbance in order to stabilize it (Cheng & Li, 2003; Wu *et al.*, 2006; Cheng *et al.*, 2007; Mu *et al.*, 2012). It should be noted that Goering & Kumar (1996), Goering (1998), Jiang (2004) and Cheng *et al.* (2007) designed their models and experiments to incorporate impermeable barriers on both the top and sides of the porous materials, and assumed that the blocky layer sits on a level subsurface. This is consistent with some engineering applications (e.g., some ACE and crushed rock or block revetments), but not with many naturally occurring blocky materials. Many landforms made of blocky materials are sloped, and the surfaces of blocky materials are generally permeable, except where there is thick snow or vegetation cover.

2.4.2.2 The chimney effect

The chimney effect is a form of free convection that occurs in slopes made of blocky materials that are at least partially covered with snow or vegetation (Cheng *et al.*, 2007; Zacharda *et al.*, 2007). The chimney effect occurs when a single large convective cell forms in the entire slope (Cheng *et al.*, 2007) (Fig. 2.5). This convective cell extends from the bottom to the top, with air flowing in the direction of the change in elevation (Bories & Combarous, 1973; Cheng *et al.*, 2007). The convective cell forms parallel to the surfaces of the slope, so that the angle of movement of air is not only related to

Fig. 2.5: Unicellular convection in an inclined rock layer. Reprinted with permission from Cheng *et al.* (2007), pg. 155, Fig. 7b. ©2007 by John Wiley and Sons.

gravity but also to the angle of the blocky slope (Cheng *et al.*, 2007). The warmer, less dense air flows upward next to the warmer side of the blocky layer, while the cooler, denser air flows downward next to the cooler side of the blocky layer (Bories & Combarnous, 1973; Sawada *et al.*, 2003; Cheng *et al.*, 2007). In winter, warm internal air may be expelled from the top of the blocky slope if there are openings to the atmosphere. The resulting lower pressure inside the blocky layer is offset by the intake of cold external air at the bottom (Delaloye *et al.*, 2003). Holes through snow cover (chimneys) or snow-free conditions are not necessary for the chimney effect to occur; heat loss and air movement through the snow cover can be sufficient (Delaloye *et al.*, 2003). In summer, the air flow pattern may reverse, with cold expelled from the lowest part of the blocky slope and warm air drawn in higher up (Delaloye *et al.*, 2003). As a result of this pattern, the blocky slope warms from the top down and cools from the bottom up (Harris & Pedersen, 1998; Delaloye *et al.*, 2003). Engineered chimney effects occur in heat drains (e.g. pipes) built into embankments to accelerate replacement of warm air in the embankment with cold air from the atmosphere (Doré & Zubeck, 2009).

2.4.2.3 The Balch effect

The Balch effect is another form of free convection and describes air exchange between the blocky materials and the atmosphere. It is caused by a temperature-driven difference in the density of air inside and outside blocky materials or caves. When the external air becomes colder and denser than

the internal air (in autumn and winter), the external air moves into blocky low-lying materials or caves through void spaces or fissures and displaces the warmer internal air (Balch, 1900). When the internal air is colder than the external air (in spring and summer), displacement occurs in the opposite direction, with the cold air flowing out of the lowest parts of the blocks or cave, and warm air being drawn in to the higher sections to fill the vacuum (Balch, 1900). In Balch’s observation, the displacement of warm air tends to be much more rapid than that of cold air. Cold air remains longest where there is an obstacle to be overcome (a lip) in order for it to drain away (Balch, 1900; Holmgren *et al.*, 2017a). This creates pockets of long-lasting low temperature which, combined with a water supply, may create and preserve ice in the lowest parts of the cave or blocky materials (Balch, 1900).

Ice caves are found throughout North America, and particularly in the Western Cordillera and the northern Appalachian Mountains (Harris, 1979; Holmgren *et al.*, 2017b,a). While some are caves in permafrost, others are found well outside the permafrost limit. Harris (1979) describes gravity-driven cold air drainage through fissures at the Canyon Creek ice cave in Alberta, and credits this with the preservation of ice year-round despite the cave’s location 500 m below most permafrost. The coldest parts of the cave in winter and the warmest parts in summer are found closest to the fissures through which the air flow occurs (Harris, 1979).

2.4.2.4 Limits on convection

Convection does not always occur where air movement is possible. For free convection to occur the air must become unstable. The stability of a fluid is described by a dimensionless parameter called the Rayleigh number (Ra), given in equation 2.4,

$$Ra = \frac{g\beta\Delta TH^3}{\kappa\nu} \quad (2.4)$$

where g is the gravitational acceleration (ms^{-2}), β is the coefficient of thermal expansion of the fluid (K^{-1}), ΔT is the temperature difference between the warmest and coldest surfaces (K), H is the distance between the warm and cold surfaces (the thickness of the fluid) (m), κ is the thermal diffusivity of the fluid (m^2s^{-1}), and ν is the kinematic viscosity of the fluid (m^2s^{-1}). Equation 2.4 shows that in order for convection to occur, the temperature difference between the surfaces must be great enough that the air is unable to equilibrate it by conduction, and the buoyancy must overcome the air's viscosity (Narasimhan, 1999). Where the Rayleigh number exceeds a critical value (Ra_c), convection may take place (Narasimhan, 1999; Cushman-Roisin, 2014). For a system with rigid boundaries, the Ra_c is 1708 (Cushman-Roisin, 2014). For a system with a free upper boundary, Ra_c is approximately 1100 (Weisstein, 2007). The larger the Rayleigh number, the more unstable the fluid is and the more likely that convection will occur (Narasimhan, 1999).

Figure 2.6a shows calculated Ra values for a box of air 0.1 m high. The values of the terms used to calculate Ra (Table 2.3) are based on an air

Fig. 2.6: a) Rayleigh numbers and b) filtration Rayleigh numbers for varying ΔT from set parameters (Table 2.3).

Term	Value	Units	Source
g	-9.8	$m s^{-1}$	
β	4.08×10^{-3}	K^{-1}	(Engineering Toolbox, 2003a)
κ	1.654×10^{-5}	$m^2 s^{-1}$	(Engineering Toolbox, 2018)
ν	1.159×10^{-5}	$m^2 s^{-1}$	(Engineering Toolbox, 2003b)
H_{Ra}	0.1	m	
H_{Ra^*}	1	m	
k	6.32×10^{-7}	m^2	Porous layer of an ACE, rocks 3 cm in diameter (Goering & Kumar, 1996)
C_v	998	$J m^{-3} K^{-1}$	Calculated using C_s from (Engineering Toolbox, 2003a) and density from (Engineering Toolbox, 2003c)
λ^*	0.484	$W m^{-1} K^{-1}$	Calculated according to Johansen's formula using the λ of air (Engineering Toolbox, 2009) and gabbro (Eppelbaum <i>et al.</i> , 2014)
ΔT	variable	-	$T_{upper surface} - T_{lower surface}$

Table 2.3: The parameters used to calculate the Rayleigh numbers in Fig. 2.6. These values are based on an air temperature of $-20^\circ C$, but the values of all terms except for C_v at $20^\circ C$ are less than 3.06×10^{-3} different. C_v decreases to $864 J m^{-3} K^{-1}$.

temperature of -20°C because convection in blocky materials is limited during summer (Goering & Kumar, 1996; Goering, 1998; Jiang, 2004; Cheng *et al.*, 2007), and air temperatures within the talus can drop below -20°C in winter. From the graph it is evident that even a very slight temperature difference, less than $6 \times 10^{-3}^{\circ}\text{C}$ for a free upper boundary and less than $9 \times 10^{-3}^{\circ}\text{C}$ for rigid boundaries, will produce convection. This means that within cavities that are on the order of the pore spaces in kurums, the potential for convective heat transfer is large.

The filtration Rayleigh number (Ra^*) is used for convection in porous media (Bories & Combarnous, 1973) and incorporates the volumetric heat capacity of the fluid (C_v , $\text{Jm}^{-3}\text{K}^{-1}$), the permeability of the medium (k , m^2), and the thermal conductivity of the porous medium saturated with the motionless fluid (λ^* , $\text{Wm}^{-1}\text{K}^{-1}$).

$$Ra^* = g \frac{\beta C_v}{v} \frac{k}{\lambda^*} \Delta T H \quad (2.5)$$

The critical filtration Rayleigh number (Ra_c^*) is $4\pi^2$, or approximately 40 (Bories & Combarnous, 1973). It is again critical that the buoyancy of a heated air parcel should overcome the air's viscosity, but also that the ability of the fluid to move through the pores should overcome the thermal conductivity of the system as a whole.

Figure 2.6b shows calculated Ra^* values for a 1 m box of air and gabbro rocks with a porosity of 35%. This is a rock type common to the talus slopes

in the study area of this thesis (Heaman *et al.*, 2007). In this case, convection occurs when ΔT is roughly 9°C (Fig. 2.6b). Where $H = 1$, Ra^* increases about 4.5 for every 1 degree increase in ΔT .

In equations 2.4 and 2.5, $\frac{g\beta}{\kappa v}$ and $g\frac{\beta C_v}{v}\frac{k}{\lambda^*}$ can be treated as constants. This allows the relations between H , ΔT and Ra^* (Ra) to be explored. Ra^* increases with H or ΔT . As H increases, the magnitude of ΔT required to reach Ra_c^* decreases (Fig. 2.6b). This means that if the porous layer thickness varies, or there is temperature variation at its surface or base, free convection may take place in some parts of the porous layer but not others.

The degree of air instability and the slope angle affect the pattern of convection that forms in the system. Modelling using solid impermeable boundaries and fluid layers of uniform thickness has found that if the slope angle (ϕ) is less than 15° , convection cells develop as many polyhedral cells next to each other (Bories & Combarous, 1973). The buoyancy causes vertical air movement that is perpendicular to the surfaces of the system, so where ϕ is greater than 0° , the axes of the convection cells are related to ϕ (Cheng *et al.*, 2007). If the slope angle is greater than 15° but less than 80° , unicellular flow or longitudinal coils develop, depending on the relation between buoyancy and the angle. The more unstable the air or greater the slope angle, the more likely it is that coils develop. The longitudinal coil pattern has multiple convection coils running side-by-side up and down the slope. Air moves upward or downward in a spiraling pattern (Bories & Combarous, 1973). When the fluid is quite unstable, convective cells

continually form, collapse and reform (Bories & Combarnous, 1973).

In some cases, a forcing agent such as wind is responsible for destabilizing the fluid, allowing convection to occur before the critical difference in temperature and density is reached (Narasimhan, 1999). This is called forced convection. Air may move through even thick snow when forced (Delaloye *et al.*, 2003). Cheng *et al.* (2007) note that compared to blocky materials with impermeable barriers, long-term cooling is weaker in blocky materials where wind-forced convection may occur.

2.4.2.5 Continuous air exchange with the atmosphere

Harris & Pedersen (1998) propose that in kurums where snow cover is thin or absent, continuous air exchange may occur between the blocky materials and the atmosphere. This constant exchange means that any change in external air temperature can be transferred into the blocky layer “almost instantaneous[ly]” (Harris & Pedersen, 1998, 114). This may occur with free convection, but it is most effective on steep slopes where windy conditions are common (Harris & Pedersen, 1998), because air may be blown into and out of the blocky materials by the wind, if its speed and direction are favourable (Juliussen & Humlum, 2008). Juliussen & Humlum (2008) observed wind-forced convection in the upper 30 cm of some block slopes, which resulted in more efficient heat transfer and increases in apparent thermal diffusivity. Holmgren *et al.* (2017a) observed unusually rapid and large temperature fluctuations within a talus slope in response to a hurricane, but only when

the winds were blowing toward the slope.

2.4.3 Movement and phase change of water

In spring, refreezing of meltwater that percolates into still-frozen parts of blocky slopes causes abrupt increases in temperature (Sawada, 2003). The slow thaw of ice during the summer may be crucial to maintaining ground temperatures $< 0^{\circ}\text{C}$ (Delaloye *et al.*, 2003), and the refrozen ice may protect multi-year ice inside the blocky materials by buffering it from thaw for some time (Sawada, 2017).

2.5 Summary

Permafrost in Canada is typically found at high latitudes and high elevations, where T_a is less than -1°C . However, in other parts of the world permafrost has been observed in blocky materials far from where permafrost is expected. Heat movement through the ground is predominantly conductive, but blocky materials contain void spaces that dramatically increase the importance of convective heat transfer and reduce the effectiveness of conductive heat transfer within them. Convective mechanisms, including free and forced convection, the chimney effect, and the Balch effect, efficiently remove heat from the blocky materials but do not provide heat to the materials as effectively. This may result in persistent cold thermal anomalies within blocky materials. For blocky materials, convection occurs when the system reaches a critical Ra^* of

approximately 40. Ra^* is controlled by the thickness of the blocky materials and the temperature gradient from the bottom to the top of the blocky layer. For a blocky layer 1 m thick, the minimum temperature difference required for convection is 9°C. Ra^* increases by about 4.5 for every 1 degree increase in ΔT or 1 m increase in H .

Chapter 3

Study Sites and Methods

This chapter discusses the characteristics of the study area at a regional and local scale and then describes the methods used in this research. The study area is near Lake Superior and Lake Nipigon in Northwestern Ontario (Fig. 3.1). The area is located on the Canadian Shield, in the boreal forest. Two study sites were chosen: Doghead Mountain, near Nipigon, and Ouimet Canyon, between Nipigon and Thunder Bay.

3.1 The Lake Superior and Lake Nipigon region, Northwestern Ontario

3.1.1 Climate

The closest meteorological stations to Doghead Mountain are at the Cameron Falls generating station north of Nipigon, Ontario (Table 3.1). Environment Canada's air temperature and precipitation normals for 1981-2010

Fig. 3.1: Locations of Doghead Mountain and Ouimet Canyon, including the Cameron Falls and Thunder Bay weather stations. Produced using CanVec data from <https://www.nrcan.gc.ca/earth-sciences/geography/topographic-information/free-data-geogratis/11042>.

Station name	Location	Data range	Normals
Cameron Falls	49°09'38" N, 88°21'00" W	1924-1998	1961-1990, 1971-2000, 1981-2010
Cameron Falls (AUT)	49°09'14" N, 88°20'41" W	1998-2017	-
Thunder Bay A	48°22'10" N, 89°19'38" W	1953-2012	1961-1990, 1971-2000
Thunder Bay CS	48°22'10" N, 89°19'38" W	2000-2017	-

Table 3.1: Location and data available for the four climate stations used in this study.

(Table 3.2, Fig. 3.2,) at the Cameron Falls station were based on data collected between 1981 and 1998, with one (unspecified) year missing (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html). As a result the normals do not include the 12 years from 1999 to 2010, which, globally, were warmer than most of the preceding 18 years and were 0.25°C to 0.55°C warmer than the 1961-1990 average (Hansen *et al.*, 2010).

The closest meteorological stations to Ouimet Canyon with both long term and current records are at Thunder Bay International Airport (Table 3.1), where several different meteorological stations have operated since 1941. Of these, the Thunder Bay A station has climate normals for 1971-2000 (Fig. 3.2, Table 3.3). Climate normals for 1981-2010 are not available for any of the stations. The 1971-2000 normals were calculated using data from the 23 years between 1971 and 1993 (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html).

Northern Ontario has a “modified” continental climate, with continentality increasing toward the Manitoba border and away from lakes Superior and Nipigon (Chapman, 1953). The mean annual air temperatures (T_a) of Cameron Falls (elevation 228.6 m a.s.l.) and Thunder Bay A (elevation 199 m a.s.l.), 1.9°C and 2.5°C, respectively, indicate a periglacial environment (French, 2007). Assuming an environmental lapse rate of 6.5°C km⁻¹ (Oke, 1987), the mean annual temperatures at the base of Doghead Mountain (elevation ≈210 m a.s.l.) and Ouimet Canyon (elevation ≈335 m a.s.l.) are

Month	Mean air temp. (°C)	Precipitation (mm)	Snow fall (cm)
January	-16.2	55.4	55.4
February	-13.1	32.2	29.8
March	-6.5	30.2	23.4
April	1.7	43.6	13.9
May	8.7	62.9	1.2
June	14.0	86.3	0.0
July	17.1	91.3	0.0
August	16.8	94.3	0.1
September	11.3	99.4	0.9
October	4.7	91.8	11.6
November	-4.0	73.8	43.8
December	-11.9	63.7	58.1
Annual	1.9	824.8	238.2

Table 3.2: Monthly climate normals for Cameron Falls, 1981-2010 (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html).

Fig. 3.2: Climographs for a) Cameron Falls station 1981-2010 and b) Thunder Bay A station 1971-2000. T_a at Cameron Falls for the period is 1.9°C, and average total annual precipitation is 824.8 mm. T_a at Thunder Bay Airport for the period is 2.5°C, and average total annual precipitation is 711.6 mm (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html).

Month	Mean air temp. (°C)	Precipitation (mm)	Snow fall (cm)	Snow depth at month end (cm)
January	-14.8	31.3	41.2	34
February	-12.0	24.9	26.9	32
March	-5.5	41.6	26.8	13
April	2.9	41.5	12.4	0
May	9.5	66.5	1.7	0
June	14.0	85.7	0.0	0
July	17.6	89.0	0.0	0
August	16.6	87.5	0.0	0
September	11.0	88.0	0.5	0
October	5.0	62.6	6.1	0
November	-3.0	55.6	27.8	9
December	-11.6	37.5	44.1	24
Annual	2.5	711.6	187.6	9

Table 3.3: Monthly Climate Normals for Thunder Bay A, 1971-2000 (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html).

estimated to be 2.0°C and 1.6°C, respectively.

Freeze-thaw weathering is important to the break down of exposed rock outcrops and the creation of talus. While air temperatures above 0°C occur occasionally at both stations through the winter months, the typical window for diurnal freezing and thawing is from March to May in spring and October to November in the fall (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html). Spring warming takes place slowly in the study area because of the influence of Lakes Superior and Nipigon on regional climate. Until the ice melts, the lake surface is 0°C or below. Warming of the lakes begins only once there is open water (Austin & Colman, 2007). While the ice reflects incoming radiation, and while the ice or lake surface water remain near 0°C, air masses over, adjacent to, and downwind of the lakes will also be close to 0°C (Rouse, 1991), prolonging the springtime diurnal freeze-thaw window.

The precipitation at both sites has a seasonal pattern, with more occurring in warm months as rain than in winter as snow. Snow depth normals are available for Thunder Bay A, and indicate that deep snow does not accumulate at the airport (Table 3.3). The effect of the large lakes is visible in the amount of precipitation in late summer and autumn, especially at Cameron Falls; the heat and moisture from the lakes cause convection in the surface boundary layer, leading to precipitation (Scott & Huff, 1996).

T_a in the Lake Superior region increased by 1.5°C between 1906 and 2005 (Austin & Colman, 2008), and ground surface temperature (T_s) increased

1-2°C between 1800 and 2000, and at least 0.5°C between 1980 and 2000 (Gosselin & Mareschal, 2003). The energy available to the regional climate system has been increasing over the long term. However, the two winters preceding the study period (2013-2014 and 2014-2015) were several degrees colder than normal, and summer temperatures both years were near normal.

3.1.2 Geologic history of the Lake Superior and Lake Nipigon region

The geology of the study area is characterized by old rocks from no later than the Mesoproterozoic era (Ontario Geological Survey, 2005). The basement rock is the Superior craton, which is exposed as shield in some places and covered by sedimentary and igneous rocks in others. The southern edges of the craton, close to Lake Superior and Lake Nipigon, are the apex and northern margin of a 2000 km-long, horseshoe-shaped, aborted intracontinental rift (Heaman *et al.*, 2007). The rift's associated magmatism created intrusions that have, after 1 billion years of erosion, become mesas and cuevas (Pye, 1997). The intrusive formations between Lake Nipigon and Lake Superior and on the northwestern side of Lake Superior are of particular interest to this study, but the geology and geomorphology is similar across a wider area.

3.1.2.1 Mesoproterozoic rift activity

The continent of Laurentia, of which the Superior craton was part, experienced rifting and associated magmatism and volcanism in the Mesoproterozoic era from 1150 million years ago (Ma) to 1094 Ma (Cannon, 1992; Heaman *et al.*, 2007). Dyke swarms near the Ontario-Minnesota border and between Lake Superior and western Quebec, as well as diabase on the northeastern edge of Lake Nipigon, were intruded early during the rifting episode (1150-1130 Ma) (Heaman *et al.*, 2007). Most of the magmatism in the northern Lake Superior region and the Nipigon Embayment, an extension of the Midcontinent Rift centred under Lake Nipigon, occurred during the middle of the rifting episode (1115-1105 Ma) (Heaman *et al.*, 2007), forming the Logan Igneous Suite. The suite can be informally subdivided into the Nipigon sills north and east of Thunder Bay, and the Logan sills southwest of Thunder Bay (Hollings *et al.*, 2007). Intense rifting occurred from 1100 Ma to 1094 Ma, creating a deep central basin beneath what is now Lake Superior, which was subsequently filled with basalt (Cannon, 1992).

3.1.2.2 The country rock

The country rock into which the magma intruded was primarily Archean and early Proterozoic sedimentary rock which had been deposited to a thickness of a few hundred metres on top of the continental crust (Heaman *et al.*, 2007). The Archean deposits were fluvial and lacustrine sandstones while the

Proterozoic deposits, such as the Mesoproterozoic Sibley Group, ranged from iron-rich marine sediments to aeolian ones (Ontario Geological Survey, 2005; Heaman *et al.*, 2007). Some metamorphism occurred to the sedimentary rocks where they came into contact with the intruding magma (Pye, 1997). The area also features some igneous rock deposited during the Archean and Proterozoic eons, including volcanic granite-greenstone and felsic rocks (Ontario Geological Survey, 2005; Heaman *et al.*, 2007).

3.1.2.3 The intrusive rock

The intrusions took the form of sills, dykes, and lopoliths (saucer-shaped), and sometimes formed in groups called swarms. They are composed of mafic, ultramafic and alkaline rock, including diabase and gabbro (Heaman *et al.*, 2007). The placement of intrusions seems to have been controlled by north- and northwest-trending faults (Heaman *et al.*, 2007) and the bedding planes of the country rock (Pye, 1997).

Upon intrusion into the country rock, the magma began to cool. This occurred most rapidly at the contact between the magma and the country rock. Magma forms small crystals where it cools rapidly, so the smallest crystals in the intrusions are found at the margins between them and the country rock (Pye, 1997). The intrusive rock contracted as it cooled, forming cracks at crystal boundaries, perpendicular to the cooling front. When the intrusions were later exposed by erosion, these cracks meant that the rock of the sills and dykes was already friable, and could break apart into columns (Fig. 3.3)

Fig. 3.3: A section of the east side and bottom of Ouimet Canyon, showing the column-like structure of the rock that makes up the cliffs and the talus sheets that have formed from their disintegration. Photo by Jessica Abbott.

without the environmental conditions required to produce significant freeze-thaw or chemical weathering (Pye, 1997). The rock in thicker intrusions is characterized by larger columns because they cooled more slowly and formed larger crystals (Puumala, M., 2017, pers. comm.). As the pre-Cambrian sedimentary rocks weathered away preferentially due to their relative softness, the sills and dykes were left exposed as mesas and cuestras, where columns of rock might be dislodged from the cliff faces and create talus slopes (Pye, 1997). Over 3000 talus slopes have been identified across northern Ontario as part of the work conducted for this thesis.

The intrusions mostly range from a few tens of metres to nearly 200 m thick. In general, the coarseness of the diabase increases toward the middle of the intrusion, and the rock near the contact between the intrusion and the country rock is amygdaloidal (Heaman *et al.*, 2007).

3.1.3 Glaciation

Northwestern Ontario has been glaciated many times, most recently during the Wisconsinan glaciation. The Superior basin was the last of the Great Lakes basins to be freed from the ice sheet, and it is the only one thought to have been glaciated during the Younger Dryas (Algonquin) readvance (Larson & Schaetzl, 2001). Lake Superior and Lake Nipigon basins were finally deglaciated around 9.5 ka before present (BP) (Saarnisto, 1974; Larson & Schaetzl, 2001). These glaciations, by removing less resistant rock, were responsible for much erosion of the landscape (Larson & Schaetzl, 2001). The

ice sheet was not only an erosional force, however. Between 11 ka and 10.1 ka BP, it deposited several moraines across the region where it halted during its retreat (Larson & Schaetzl, 2001). This till has been eroded away in some places, particularly from the tops of exposed intrusions.

Before and after the Younger Dryas stadial, a succession of gigantic proglacial lakes formed in the region (Saarnisto, 1974; Larson & Schaetzl, 2001). These lakes and their sometimes catastrophic drainage have left relic beaches, outwash plains, deltas and spillways across the region (Saarnisto, 1974; Larson & Schaetzl, 2001). Lake Superior water levels finally dropped to their current levels around 2.2 ka BP (Larson & Schaetzl, 2001).

3.2 Study sites

3.2.1 Doghead Mountain

Doghead Mountain is a cuesta located on the east side of the Nipigon River where it enters Lake Superior (Fig. 3.4). It is located in Ruby Lake Provincial Park. Talus sheets can be found on the west and east sides of the cuesta (Fig. 3.5). The talus sheet chosen for study leads into and runs the length of a ravine on the west side of the cuesta (Figs. 3.6 and 3.7). The ravine trends east-south-east to west-north-west, and rises to the top of the cuesta. Most of the talus has a southwestern exposure, except within the ravine where a north-facing talus slope has also formed beneath the cliffs on the southern

Fig. 3.4: Topographic map of Doghead Mountain, near Nipigon, North-western Ontario, with the area of Fig. 3.5 indicated. Produced using CanVec data from <https://www.nrcan.gc.ca/earth-sciences/geography/topographic-information/free-data-geogratis/11042>.

Fig. 3.5: Satellite photo of Doghead Mountain, with the area of Fig. 3.6 indicated by the rectangle. Base image from Google Earth Pro (Google LCC, Menlo Park, CA).

Fig. 3.6: Satellite photo of the instrumented section of the talus sheet at Doghead Mountain. The solid line marks the slope transect in Fig. 3.8. The dotted lines mark the boundaries of the slope sections used for rock size sampling. Base image from Google Earth Pro (Google LLC, Menlo Park, CA).

Fig. 3.7: The westernmost section of south-facing talus slope of the Doghead ravine. The photo was taken on 25 February 2017 from the north-facing slope.

side. Figure 3.8 shows a transect of the instrumented section of the talus sheet.

3.2.1.1 Bedrock and surficial geology

Doghead Mountain is a diabase intrusion (Ontario Geological Survey, 2005) that has become an elevated bedrock plain (Ontario Geological Survey & Northeast Science and Information Section, 2005). It is one of the Nipigon sills, part of the Logan Igneous Suite. This sill intruded into the Mesoproterozoic sedimentary bedrock of the Sibley Group, and the sill's northern edge corresponds to the boundary between the Sibley Group and Archean metasedimentary bedrock (Ontario Geological Survey, 2005).

On some parts of Doghead Mountain, wet organic materials have accumulated on a clay glaciolacustrine plain; in others there are no significant surface materials covering the bedrock. Glacial erosion is responsible for the thin soil and exposed bedrock on the top of the intrusion. A low, dry, pitted or kettled glaciolacustrine delta made of sand and gravel covers the Archean bedrock immediately north of the mountain. To the northwest, there is a low glaciolacustrine plain made of clay and silt (Ontario Geological Survey & Northeast Science and Information Section, 2005). These glaciolacustrine deposits are adjacent to the talus slopes studied.

Fig. 3.8: A transect of the instrumented section of the Doghead talus sheet. Distances and slope angles are to scale. The angle of the measured section of the slope is 28° . * indicates the approximate locations of the loggers relative to the transect. See Table 3.5 for the distances and angles between loggers. Some sections of the talus, indicated by dashed lines, were too unstable to measure. Their length was estimated, and for the purposes of this diagram, their slope was assumed to be the same as the nearest measured section. Slope angle was measured with a Brunton Pocket Transit and distance along the slope with a 30 m measuring tape (see page 94 for methods).

3.2.1.2 Vegetation

There is a marked difference between the talus slope and the forested area below it. Balsam fir (*Abies balsamea*) and black spruce (*Picea mariana*) grow up to and onto the bottom edge of the talus, and provide shade to the lowest part of the talus. The talus itself has very few trees, except in its upper parts where it meets the cliff. Most trees on the talus are white birch (*Betula papyrifera*), black spruce, white cedar (*Thuja occidentalis*), or trembling aspen (*Populus tremuloides*) with the latter three more common near the cliff. Lichens are abundant on the talus except on the upper sections, and are particularly thick at the bottom of the talus. Thick blankets of moss have also developed in the lowest section of the slopes.

3.2.1.3 Weather

In 1998, the Cameron Falls meteorological station was replaced by an automated station called Cameron Falls (AUT), which continues to operate but was not included in the calculation of the 1981-2010 Cameron Falls normals (Environment Canada Climate Normals are available at http://climate.weather.gc.ca/climate_normals/index_e.html). Weather data for Doghead Mountain for 2016 and 2017 were taken from Cameron Falls (AUT).

The study period was relatively dry compared to the 1981-2010 Normals (Fig. 3.9). July through September 2016 were particularly dry, receiving roughly half of their normal precipitation. April 2017 was the only month

Fig. 3.9: Mean monthly total precipitation at Cameron Falls, 1981-2010, and values measured during the study period.

with an excess of precipitation, receiving about 25 mm more than usual, much of which fell during a 39.5 mm event on April 26.

The air temperatures during the study period ranged from normal to higher than normal (Fig. 3.10). The largest departures from normal occurred in the fall and winter of both years. In January 2016, and from November 2016 to February 2017, the minimum temperatures recorded were closer to the normal mean temperatures, and the mean temperatures recorded were closer to the normal maximum temperatures. Maximum temperatures during these periods were roughly 5°C above normal. This warm weather followed two consecutive winters of temperatures well below normal.

3.2.2 Ouimet Canyon

Ouimet Canyon is a 3.1 km-long canyon in a bedrock outcrop (Fig. 3.11). It is located in Ouimet Canyon Provincial Park. The canyon runs north from the edge of the bedrock outcrop for about 1.7 km, takes a bend to the northwest for about 0.6 km, and then turns east for about 0.7 km. The canyon becomes progressively narrower, starting at nearly 300 m wide at the canyon mouth but narrowing to only a few tens of metres at its northeastern end. For most of its length, it is 50-100 m wide. The canyon is 15-120 m deep, with the depth increasing southward toward the canyon mouth. Talus sheets line the cliffs on both sides of the canyon for most of its length (Fig. 3.12). The talus sheet chosen for study is on the east side of the canyon, slightly south of the canyon's northwestern bend (Figs. 3.13 and 3.14). Figure 3.15 shows a

Fig. 3.10: Mean, mean maximum, and mean minimum daily temperatures plotted by month from Cameron Falls 1981-2010 and values measured during the study period at Cameron Falls.

Fig. 3.11: Topographic map of Ouimet Canyon and surrounding area, with the area of Fig. 3.12 indicated. Produced using CanVec data from <https://www.nrcan.gc.ca/earth-sciences/geography/topographic-information/free-data-geogratis/11042>.

Fig. 3.12: Satellite photo of Ouimet Canyon, with a) Fig. 3.13 and b) the section used for basal temperature of snow (BTS) measurement outlined in white. BTS was measured in the southern part of the canyon because of difficulties accessing the rest of the canyon. Base image from Google Earth Pro (Google LLC, Menlo Park, CA).

Fig. 3.13: Satellite photo of the instrumented part of Ouimet Canyon. See Fig. 3.15 for the slope transect. Rock size sampling at this site was abandoned for safety reasons, so the slope sections were omitted. Base image from Google Earth Pro (Google LCC, Menlo Park, CA).

Fig. 3.14: View north-northeast toward the study site at Ouimet Canyon.

Fig. 3.15: A transect of the instrumented section of Ouimet Canyon. Distances and slope angles are to scale. The angle of the measured section western slope is 32° , and the angle of the eastern slope, including the level segment at the base of the cliff, is 29° . * indicates the approximate locations of the loggers relative to the transect. See Table 3.5 for the distances and angles between loggers. Some sections of the talus, indicated by dashed lines, were too unstable to measure. Their length was estimated, and for the purposes of this diagram, their slope was assumed to be the same as the nearest measured section. Slope angle was measured with a Brunton Pocket Transit and distance along the slope with a 30 m measuring tape (see page 94 for methods).

transect of this part of the canyon.

3.2.2.1 Bedrock and surficial geology

Ouimet Canyon runs mainly north-south from the centre to the edge of an elevated bedrock plain (Ontario Geological Survey & Northeast Science and Information Section, 2005). This bedrock plain is a diabase intrusion that is part of the Logan Igneous Suite. It appears to have intruded between Archean igneous rocks and sedimentary rocks of the Sibley Group (Ontario Geological Survey, 2005). The southern end of the canyon abuts Sibley Group rocks that are surrounded by the diabase intrusion. The canyon mouth opens onto the low, flat terrain of the Sibley Group rocks. Opposite and south of the canyon mouth is a sand and gravel kame (Ontario Geological Survey & Northeast Science and Information Section, 2005) lying on another arm of the intrusion (Ontario Geological Survey, 2005). In other places the intrusion is covered by a dry, sandy ground moraine veneer (Ontario Geological Survey & Northeast Science and Information Section, 2005).

3.2.2.2 Vegetation

There is no perceptible difference in vegetation between the eastern and western slopes. At the northern end of the study site, a band of trees (white cedar, white birch and black spruce) grows on both talus slopes (Fig. 3.14). On both sides of the canyon, a narrow band of trees (cedar, maple (*Acer saccharum*), black spruce) and occasional currant bushes (*Ribes oxycan-*

thoides) grow along the top of the talus where it meets the cliff. Below these bands, both slopes are almost entirely free of trees. Lichens are abundant except near the top of the talus, along with communities of ferns and occasional raspberries (*Rubus idaeus*) and currants. The lowest point of the talus is blanketed with thick moss and lichens, and supports communities of Labrador tea (*Rhododendron groenlandicum*) and occasional small maples, birches, and black spruce. Disjunct species more common to the Arctic and subarctic can be found as well (Kor & Teller, 1986; Wilkins, 1994).

3.2.2.3 Weather

Thunder Bay A was closed in 2012, so the weather record for this research was obtained from Thunder Bay CS. It has a long overlap with the Thunder Bay A data (12 years), and it has fewer missing values in 2016 and 2017 than other stations nearby.

During the study period, Thunder Bay CS was at times drier, and at other times considerably wetter, than normal (Fig. 3.16). June and November 2016 were exceptionally wet, receiving more than twice as much precipitation as usual due to two or three particularly heavy precipitation events. June 2017 also received over 30 mm more precipitation than normal due to a single event.

Air temperatures at Thunder Bay CS were generally close to normal or above normal during the study period (Fig. 3.17), although it should be noted that the winters of 2013-2014 and 2014-2015 were much colder than normal.

Fig. 3.16: Mean monthly total precipitation at Thunder Bay A, 1971-2000, and values measured at Thunder Bay CS during the study period.

Fig. 3.17: Mean, mean maximum, and mean minimum daily temperatures plotted by month from Thunder Bay A 1971-2000 and values measured at Thunder Bay CS during the study period.

Temperatures in July and August 2016 were higher than usual, and led into fall and winter temperatures that were also considerably higher than usual. Between December 2016 and February 2017, positive departures of nearly 5°C from the normal values occurred for the minimum, maximum and mean temperatures. Air temperatures for the rest of 2017 were near normal.

3.3 Methods

The thermal regimes of the two talus slopes were of primary interest. Thermistors were installed at the surface and within the talus for more than two years on the two slopes, and air temperature loggers were also installed so that the relation between the talus thermal regime and the atmosphere could be investigated. In order to determine the spatial extent of thermal anomalies that developed in the winter, basal temperature of the snow pack was measured in late winter 2016 and 2017. The size of the blocky materials in the talus and the geometry of the talus slopes were also measured.

3.3.1 Thermal regime of talus

In order to record the talus temperatures at Doghead and Ouimet, one ONSET HOBO Outdoor 4-Channel data logger (Onset Computer Corp., Bourne, MA) was installed at the lower, middle and upper section of the talus in mid-July 2016. Table 3.4 documents the installation configurations and dates and the logger code names. At Ouimet Canyon the installations were on a west-

Location	Logger	Sensors	Date	Time
Doghead Mountain				
Air	DHA	DHA	14-Jul-16	14:00
Lower	DHL	DHL-S	14-Jul-16	17:00
		DHL-0.7		
		DHL-1.35		
		DHL-1.75		
Middle	DHM	DHM-S*	16-Jul-16	14:30
		DHM-0.7*		
		DHM-1.4		
		DHM-1.45*		
Top	DHU	DHU-S	16-Jul-16	16:00
		DHU-0.4A		
		DHU-0.4B		
		DHU-0.5		
Ouimet Canyon				
Air	OCA	OCA	23-Jul-16	13:40
Lower	OCL	OCL-S	20-Jul-16	14:30
		OCL-0.8		
		OCL-1.2		
		OCL-1.5		
Middle	OCM	OCM-S	23-Jul-16	12:00
		OCM-0.5		
		OCM-0.65		
		OCM-0.66		
Top	OCU	OCU-S	23-Jul-16	13:00
		OCU-0.46		
		OCU-0.73		
		OCU-0.77		

Table 3.4: Names, sensor names, and installation date and time for the air and ground temperature loggers. The values in the sensor names represent the depth of the sensor in metres. Sensors with an asterisk (*) were vertically aligned. DHU-0.4A and DHU-0.4B were at the same depth but in different voids in the talus.

facing talus slope, and at Doghead on a south-facing talus slope. The location of the lower loggers, DHL and OCL, determined the section of the talus slopes used in the study. DHL and OCL were installed in pits in the talus where ice was found in August 2011 and again at the time of logger installation. The middle loggers (DHM and OCM) were installed about halfway up the slope, and the upper loggers (DHU and OCU) were installed as close to the cliff as possible safely. Due to the interference of the cliffs with GPS reception, the elevations of the logger sites are not known precisely. Table 3.5 shows the estimated distances and angles between the loggers at both sites, based on the transect data used to draw Figs. 3.8 and 3.15. There is approximately 20 m difference between the lowest and highest loggers at both sites. The loggers were programmed to record temperature every hour.

Each HOBO logger was connected to three TMC6-HD thermistors and one TMC20-HD thermistor which had all been calibrated in an ice bath. All the sensors returned a value of -0.16°C during the period of immersion. Temperatures were recorded in increments of 0.45°C (a resolution of $\pm 0.225^{\circ}\text{C}$), and the factory-specified accuracy of the thermistors was between $0.25\text{--}1^{\circ}\text{C}$ (Onset Computer Corp., 2013). Three thermistors were placed to measure the air temperature inside the talus. These were inserted as far as possible into the voids between the blocks; between 30 cm and 170 cm. The fourth was installed to measure the surface temperature (Figs. 3.18, 3.19, 3.20, and 3.21). The thermistors were labeled according to the distance they were inserted into the talus, with surface temperature thermistors labeled “S” (Ta-

Loggers	Horizontal distance (m)	Vertical distance (m)	Ground distance (m)	Average slope angle (°)
DHL - DHM	17	8	19	26
DHM - DHU	22	14	26	31
DHL - DHU	40	22	45	29
OCL - OCM	19	10	21	26
OCM - OCU	14	11	18	38
OCL - OCU	34	20	39	32

Table 3.5: Distances and slope angles between data logger sites, based on the field data used to for Figs. 3.8 and 3.15. The ground distances and slope angles were measured, and the vertical and horizontal distances were calculated.

Fig. 3.18: a) The layout of DHL in a pit at the bottom of the talus slope and b) DHL suspended in a birch tree roughly 2 m from the surface of a large boulder at the bottom of the talus slope. See Fig. 3.6 for their relative locations.

Fig. 3.19: The layout of a) DHM and b) DHU near the middle and the top of the talus slope, respectively. See Fig. 3.6 for their relative locations.

Fig. 3.20: The layout of OCA and OCL at the bottom of the canyon, near and in a mossy pit, respectively. See Fig. 3.13 for their relative locations.

Fig. 3.21: The layout of a) OCM and b) OCU near the middle and top of the talus slope, respectively. See Fig. 3.13 for their relative locations.

ble 3.4). Where humus was available, the surface temperature thermistor was placed under approximately 3 cm of humus. Where humus was not available, the surface thermistor was placed on a rock surface and covered with a 12 cm by 10 cm piece of reflective bubble wrap to protect it from insolation. The reflective bubble wrap, while protecting the surface sensor from heating up in direct sunlight, does not approximate the physical characteristics of the rock that the sensor was attached to, so the surface temperature measurements at the middle and upper sections during the snow-free season likely do not represent field conditions. The surface sensors do, however, provide a reasonable measurement of the conditions between the snow and the rock surface after the snow pack became established. Vertical alignment of the thermistors was not possible in most cases because of the tortuosity of the void spaces.

3.3.2 Air temperature

In order to record the air temperature at Doghead and Ouimet, a gill-shielded ONSET HOBO-TEMP data logger (Onset Computer Corp., Bourne, MA) was installed at the base of the talus. The loggers were suspended approximately 2 m above the talus surface (Figs. 3.18 and 3.20). The air temperature loggers had a precision of $\pm 0.13^{\circ}\text{C}$, however gill shields are known to introduce error of up to several degrees where they receive direct solar radiation (e.g. Anderson & Baumgartner, 1998). The loggers were partially shaded by trees and by the cliffs, which reduced but could not eliminate the potential

for this kind of bias to occur. The loggers were programmed to record air temperatures at three or six hour intervals, depending on the storage capacity of the logger and recording period. These loggers were installed in mid-July 2016. The installation configurations and dates can be found in Table 3.4.

The limited number of air temperature loggers necessitated the assumption that the air temperature approximately 2 m above the bottom of the talus slope was representative of the conditions at the site as a whole. This may not have been the case because of 1) shading resulting from aspect and terrain at both sites, and the forest at the bottom of the Doghead talus; 2) cold air exiting the bottom of the talus in warm weather, which has been observed at these talus slopes, and warm air rising out of upper sections of the talus during cold weather in winter, which has been described in the literature but has not been observed at these two sites. Nevertheless, it was important to record the air temperatures at the bottom of the talus slopes because that is where the conditions may lead to the presence of ice.

Trends in air temperature were used to define the four seasons. The freezing and thawing seasons were also defined using the daily mean air temperatures at each site. There was day-to-day fluctuation around 0°C during the transition between the freezing and thawing seasons and during periods of warm weather in January and February, so seasons were demarcated using a method similar to Karunaratne (2002). The end of a thawing season was defined as the point at which the number of consecutive days with a daily mean temperature $> 0^{\circ}\text{C}$ was less than the previous consecutive days with

a daily mean temperature $< 0^{\circ}\text{C}$. The reverse defined the end of the freezing season. Thawing days that occurred during the freezing season and *vice versa* did not contribute to the mean. Annual freezing degree-days (FDD) and thawing degree-days (TDD) were also calculated, but in this case all negative daily mean temperatures from 1 September 2016 to 31 August 2017 contributed to FDD, and all positive daily mean temperatures contributed to thawing degree-days TDD, whether or not they fell within the freezing or thawing season (Klene, 2000; Karunaratne, 2002).

3.3.3 Basal temperature of snow

Thermistor cables were built specifically for basal temperature of snow (BTS) measurements. The thermistors (BetaTHERM 2.2K3A1A) were individually calibrated in an ice bath. “The precision of the measurements when read with a multimeter to $\pm 1\Omega$ is $> \pm 0.01^{\circ}\text{C}$, and the accuracy $> \pm 0.05^{\circ}\text{C}$ ” (Burn & Zhang, 2009). Each thermistor was soldered to electrical cable 1.5 m long and enclosed in a plastic sheath and epoxy to protect it. Each cable was attached to a dowel marked in 1 cm intervals so that snow depth could be measured at the same time as the temperature measurement was taken.

The dowels were inserted through the snow to a maximum depth, which was assumed to locate the ground surface. The thermistors were left for at least 15 minutes to allow the thermistor to equilibrate. Thereafter, five readings of the resistance were taken at one-minute intervals using a Fluke multimeter, and recorded by hand. The snow depth was also recorded, and

notes about the surroundings such as proximity to trees, rocks that broke snow cover, holes in the snow cover, and local trampled snow. These methods were consistent with those used by Bishop-Legowski (2011) to ensure that the results would be comparable.

At Doghead Mountain, several sampling strategies were attempted. In 2016, sampling was conducted in two grids. The first grid was approximately 26 m by 90 m and located in the eastern section of the ravine. Measurements were taken every 2 m in rows that crossed the ravine 5 m apart. The second, smaller grid was a triangle with sides 40 m long on the talus west of the ravine. Measurements were taken every 5 m. In 2017, measurements 0 m, 10 m and 20 m from the bottom of the talus slope were taken every 10 m along the talus slope.

At Ouimet Canyon, logistical constraints enabled only two small square grids to be measured, each of nine thermistors spaced 10 m apart. These samples were taken at low points in the southern end of the canyon.

The BTS surveys in 2016 and 2017 were unsuccessful because several periods of warm weather occurred both before and during the campaigns, including freezing rain events. Ice layers within and at the bottom of the snow pack indicated that meltwater penetrated the snow pack, bringing heat into it and perturbing temperatures. This was confirmed in March 2016 when BTS measurements were repeated a few days later and found to have increased by several degrees. As a result, the BTS measurements were not indicative of ground thermal conditions. No further discussion of the BTS

surveys is presented in this thesis.

3.3.4 Rock size measurements

Rock size measurements were taken to determine how rock size varied with location on the talus slope and in order to estimate the size of the voids between rocks. The talus slope was divided into 10 m by 30 m sections, the edges of which were marked with ropes or flagging tape. The short sides of each section ran parallel to the slope, and the long sides of each section connected points of equal elevation. At Doghead Mountain, the talus slope 30 m on either side of DHL was used as the sampling area (Fig. 3.6). A total of ten sections were delimited, of which six sections were sampled, covering roughly the middle three-fifths of the talus. Slope instability made it too dangerous to sample in the highest parts of the talus, and thick moss and lichen cover at the bottom of the talus made it impossible to distinguish individual rocks or determine rock sizes without irreparably disturbing the area. Similar measurements were attempted at Ouimet, but instability throughout the slope forced sampling there to be abandoned.

Sixty-three samples were taken in each section, leading to a total sample size of 378. The method was adapted from Bunte & Abt (2001)'s method for sampling the size of particles in stream beds. The method was altered part-way through to reduce the destruction the sampling method was causing to the mosses and lichens. One section was sampled twice, once using each method, to see if the two methods yielded comparable results.

3.3.4.1 Original method - Random direction, random distance

A random starting point within a section was chosen. An eight-sided die was rolled to randomly determine a direction (1= north, 2= northeast, etc.). A ten-sided die was rolled to determine a distance in metres (d), where $d = x * 0.5$. Using a compass and a tape measure, a rock was selected for measurement. After measuring the length, width and depth of the rock, and its GPS location, the rock was marked so that it would not be sampled again. The process was then repeated.

3.3.4.2 Modified method - Random transect, set distance

A random starting point along an edge of the section was chosen. The eight-sided die was rolled to determine the direction of the transect. Rocks 1 m apart along the transect were measured. The 1 m distance was always measured from the edge of the previous rock to avoid including it within the 1 m distance. If a rock that had already been measured was selected, the distance from the previous rock was increased by increments of 1 m until it was avoided. Once another edge of the section was reached, the eight-sided die was rolled again to determine the direction of a new transect.

3.3.5 Geometry of the talus slopes

The geometry of a transect through the instrumented sections of the talus slopes was measured using a Brunton Pocket Transit and a 30 m measuring

tape. The transects were conducted at the centre line of the rock size measurement sections. Points where a discernible change in slope angle occurred were identified while moving across the slope. Standing on two such points, two people held the tape between them to measure the distance, and one person looked through the Brunton Pocket Transit at a landmarked point on the second person to measure the angle. There is some uncertainty in these measurements, particularly in the angles, because of the rough surface of the talus.

3.3.6 Timeline of field research

Installation of air and talus temperature loggers took place in July 2016. The data loggers remained in place through October 2018, with the exception of OCM which was removed on September 3, 2016, to replace DHL, which had been damaged. OCM was replaced on October 14, 2016. The loggers were downloaded several times during each summer, with the last download occurring in early to mid-October. During each download visit, the progress of thaw in places where interstitial ice or frozen soil had been observed was assessed visually.

After installation of the loggers in July 2016, rock size sampling was conducted until late August. Three of six sections were sampled using the original method between July 25 and August 6, 2016. The other three sections were sampled using the modified method between August 16 and August 23, 2016. On August 24 and 25, the lowest of the sections sampled with the

original method was resampled using the modified method.

Chapter 4

Results

This chapter discusses the sizes of the rocks in the talus slopes. It then describes the thermal regimes of the atmosphere and talus at Doghead Mountain and Ouimet Canyon from 1 September 2016 to 31 August 2017. The relations between temperatures in the atmosphere and in the talus are examined. The thermal behaviour of upper, middle and lower sections of the talus slopes is summarized and periods of convective heat transfer and thermal stratification are identified. Finally, evidence for ice accumulation and ablation within the talus is presented.

4.1 Rock and pore sizes

The mean dimensions of the rocks in the Doghead Mountain talus can be found in Table 4.1, where the a axis was the longest, the b axis was of intermediate length, and the c axis the shortest, of each rock's three dimensions. The rocks ranged between spherical and rod-like (Bunte & Abt, 2001). The standard deviations and the similarity of the means between most sections

Section	Method	n	a (cm)		b (cm)		c (cm)		d_n (cm)		V_n (cm ³)	
			$\bar{x}$	SD	$\bar{x}$	SD	$\bar{x}$	SD	$\bar{x}$	SD	$\bar{x}$	SD
2-SE	1	63	59	48	42	37	28	29	40	36	167 117	528 680
2-SE	2	63	70	80	50	66	33	48	48	62	969 152	6 499 646
3-SE	1	63	85	85	58	58	38	47	56	59	793 934	3 835 279
4-SE	1	63	53	36	37	26	26	22	36	26	90 129	369 140
2-NW	2	63	62	48	43	34	29	26	42	34	175 377	743 084
3-NW	2	63	60	53	41	36	27	25	40	35	179 740	753 569
4-NW	2	63	47	51	35	46	20	11	31	26	91 243	530 431
All*	1*	378	61	56	43	41	28	29	41	38	249 590	1 666 949

Table 4.1: Means ($\bar{x}$) and standard deviations (SD) of the a , b , and c axes, the nominal diameter (d_n), and the nominal volume (V_n) of rocks in different sections of the Doghead Mountain talus slope. DHU was in section 5-NW, DHM was located at the boundary of sections 2-SE and 3-SE, while DHL was in section 1-NW. *The 2-SE sample using method 2 was excluded because some rocks were measured using both methods.

indicate that there was more variability within sections than between them. The samples were all positively skewed and leptokurtic. Shapiro-Wilk tests on the a , b and c values from the two samples from 2-SE confirmed that they were not normally distributed ($p < 0.01$). The data were transformed with a natural logarithm (Bunte & Abt, 2001). The transformed data were confirmed to be normally distributed with a Shapiro test ($p < 0.001$), and then t-tests were conducted comparing the a , b and c values of the samples. The results indicated that the two samples were not significantly different ($p > 0.2$ in all cases). This was confirmed with Wilcoxon-Mann-Whitney rank sum tests on the untransformed data.

The a , b and c axes were used to approximate the size of each rock by calculating the nominal diameter ($d_n = (a \cdot b \cdot c)^{1/3}$) and nominal volume ($V_n = \frac{\pi}{6} d_n^3$) of a sphere (Bunte & Abt, 2001). Shapiro-Wilk tests indicated that the volumes were also not normally distributed, and transformations were not successful at normalizing the data from all the sections. As a result, Wilcoxon-Mann-Whitney tests were used to compare the rock sizes between the the different sections of the slope. The sample from 2-SE using method 2 was not included because some rocks were were measured using both methods.

Change in rock size with horizontal movement across the talus slope was mostly insignificant. There was no significant difference between 2-SE and 2-NW or 4-SE and 4-NW. Rock sizes were significantly greater in 3-SE than 3-NW ($p > 0.01$), but this was attributed to several very large boulders

(>3 m long) in 3-SE. Overall, the difference in rock sizes was insignificant when all the NW sections and SE sections were compared as units.

There was significant change in rock size with elevation. Overall, there was a significant reduction in rock size from section 3 to section 4 ($p = 0.03$). Rocks in both 4-SE and 2-SE were smaller than those in 3-SE (both $p < 0.01$) but not significantly different from each other. Rocks in 4-NW were smaller than those in both 3-NW and 2-NW ($p = 0.02$ and $p < 0.01$, respectively), but the difference between rocks in 2-NW and 3-NW was not significant.

Based on the findings above, the slope was divided into two parts: 2-SE, 2-NW, 3-SE and 3-NW combined (section 2-3), and 4-SE and 4-NW combined (section 4). The particle size distribution of section 2-3 and section 4 were determined from the b axis length (Bunte & Abt, 2001). The mass (in grams) of the rocks in each sieve size category was estimated according to Bunte & Abt (2001)'s equation, and the mass and sieve diameter (in millimetres) were used to create cumulative frequency distributions for both section 2-3 and section 4 (Fig. 4.1). A Fuller curve (see section 2.2) was fitted to each cumulative frequency distribution, including a second Fuller curve for section 4 that excluded an outlier boulder that was >8 standard deviations from the mean. The minimum porosity (n_{min}) and void ratio (e_{min}) of the talus were calculated from the exponent of the actual Fuller curve (m) and the difference between the actual and ideal Fuller curves (D_v) (Peronius & Sweeting, 1985; Latham *et al.*, 2002) (Table 4.2). The rocks

Fig. 4.1: The particle size distributions and Fuller curves of sections of the Doghead Mountain talus. a) Section 2-3, where the equation of the Fuller curve is $y = 98.4224(x/4096)^{1.4913}$. b) Section 4, where the equation of the Fuller curve is $y = 94.0069(x/4096)^{2.0392}$. c) Section 4 without the outlier, where the equation of the Fuller curve is $y = 102.9907(x/2048)^{1.1884}$.

Section	m	D_v	n_{min}	e_{min}
2-3	1.4913	0.3692	0.44	0.80
4	2.0392	0.4074	0.47	0.89
4 W.O.	1.1884	0.3307	0.42	0.73

Table 4.2: The m -value of the Fuller curve, the difference between ideal and actual Fuller curves (D_v), the minimum porosity (n_{min}), and the minimum void ratio (e_{min}) for sections 2-3, 4, and 4 without the outlier (W.O.) of the Doghead Mountain talus slope.

were very angular, so a roundness value (R) of 0.14 was used (Powers, 1953; Peronius & Sweeting, 1985). The value of the porosity correction M was maintained from Peronius & Sweeting (1985), as a dense packing structure would provide a minimum value for porosity. The calculated n_{min} and e_{min} values were within the range of the other models (Youd, 1973; Latham *et al.*, 2002). The effect of even a single extremely large rock on n_{min} and e_{min} was illustrated by the difference between section 4 with and without the outlier. The n_{min} and e_{min} values for section 4 with the outlier likely apply only to its immediate vicinity.

Using e_{min} , d_n and V_n , the pore size distribution was calculated using the Arya-Paris model, returning the mean pore radius associated with each segment of the particle size frequency distribution (Arya & Paris, 1981) (Fig. 4.2). An α value of 1.6 was used in order to calculate the minimum mean pore radius. As the model assumes cylindrical pores multiple particles in length, the mean pore radius represents the shorter axis of a pore. In every section, the model predicted a few very large pores (mean radius > 0.5 m), but the mean radius of most pores was roughly 3-20 cm.

4.2 Air temperature

The mean daily air temperatures at Doghead Mountain and Ouimet Canyon were compared with the mean daily air temperatures at Cameron Falls and Thunder Bay CS weather stations, respectively (Figs. 4.3 and 4.4). Data for

Fig. 4.2: The pore size distribution of a) section 2-3, b) section 4, and c) section 4 without the outlier, of the Doghead Mountain talus.

Fig. 4.3: Mean daily air temperature at Cameron Falls (T_{CF}) and Doghead Mountain (T_{DHA}) from 15 July 2016 to 5 October 2017. The equation for the least-squares linear regression line is $T_{DHA} = 0.974T_{CF} - 0.206$. For the line of best fit, $n = 418$, $R^2 = 0.990$, and $p < 0.001$ for both slope and intercept.

Fig. 4.4: Mean daily air temperature at Thunder Bay CS (T_{TB}) and Ouimet Canyon (T_{OCA}) from 21 July 2016 to 11 October 2017. The equation for the least-squares linear regression line is $T_{OCA} = 0.969T_{TB} - 2.524$. For the line of best fit, $n = 428$, $R^2 = 0.958$, and $p < 0.001$ for both slope and intercept.

these variables are available from 15 July 2016 to 5 October 2017 at Doghead Mountain, and from 21 July 2016 to 11 October 2017 at Ouimet Canyon. The entire period, rather than the year from 1 September 2016 to 31 August 2017, was used for comparison of the study sites to the weather stations. Linear models fit the data well, and show that there is no substantial difference between the air temperature at Doghead Mountain and Cameron Falls, and that air temperatures at Ouimet Canyon are about 2.5°C lower than at Thunder Bay CS. This is similar to the Geraldton A station, about 180 km northeast of Ouimet Canyon, and the Armstrong A station, about 170 km north of Ouimet Canyon (Environment Canada Climate Normals, available at http://climate.weather.gc.ca/climate_normals/index_e.html). This indicates that the atmospheric climate at Ouimet Canyon is generally cooler than its local region, and similar to locations 170 km to the north.

On occasion, the temperature at Ouimet Canyon was considerably lower than at Thunder Bay CS (Fig. 4.4). On these days, warm weather followed a period of colder conditions, and the warm air was not able to displace the cold air in the canyon. The canyon is narrow and runs north-south, so it is possible for easterly and westerly winds to blow over the top without mixing much with air in the canyon. Most of these events took place in winter, when shading was greatest.

Air temperatures at Ouimet Canyon were, on average, about 1.6°C lower than at Doghead Mountain, but the relationship between the two sites is strong (Fig. 4.5). Doghead Mountain often experienced higher temperatures

Fig. 4.5: Mean daily air temperatures at Doghead Mountain (T_{DHA}) and Ouimet Canyon (T_{OCA}) from 1 September 2016 to 31 August 2017. The equation for the least-squares linear regression line is $T_{OCA} = 0.954T_{DHA} - 1.744$. For the line of best fit, $n = 348$, $R^2 = 0.963$, and $p < 0.001$ for both slope and intercept.

than Ouimet Canyon, but both sites experienced similar low temperatures over most of the year (Fig. 4.6). The Doghead Mountain site is not as confined as Ouimet Canyon and opens westward onto a plain, so it does not experience a delayed response to warm weather. Furthermore, in summer the Ouimet Canyon site experiences shading both morning and evening from the canyon walls (Fig. 3.14), but only the lower, forested section of the instrumented Doghead talus slope is consistently shaded in summer. The slope faces southwest and the sun's elevation is greater than 60° during this season (Fig. 3.7).

4.2.1 Seasonality

Seasons were defined according to trends in air temperature (Fig. 4.6). Autumn was associated with cooling that was underway in September 2016. Winter was characterized by fluctuating, predominantly negative temperatures, a pattern that began on 17 November 2016 at both sites. Spring began with warming in March. The rate of warming decreased in the third week of June, marking the beginning of summer and leading to a period of sustained high temperatures. Autumn cooling began again during the last 10 days of August 2017.

Fig. 4.6: Mean daily air temperatures at Doghead Mountain (DHA) and Ouimet Canyon (OCA) from 1 September 2016 to 31 August 2017.

4.3 Ground temperatures

Annual mean temperatures were calculated for the air and the three internal talus sensors at each site (Table 4.3). Annual mean talus temperatures were not uniform along the slope. Both talus slopes had annual mean temperatures $< 0^{\circ}\text{C}$ in the lower sections, but in the middle and upper sections the values were $> 0^{\circ}\text{C}$. Temperatures in the middle sections were close to the annual mean air temperature, but the annual means in the upper sections were higher.

The differences between the three sections of the slope were also clear in the mean freezing season and thawing season temperatures (Table 4.3). The freezing season at Doghead Mountain in 2016-2017 was from 18 November 2016 to 30 March 2017, and at Ouimet Canyon from 18 November 2016 to 28 April 2017. The freezing season at Ouimet Canyon was longer than that at Doghead Mountain because the canyon's geometry and orientation caused it to experience more shading, and because of its delayed or dampened responses to warm weather events.

During the freezing season, OCU and DHU were considerably warmer than the other sites and the air. During the thawing season, the lower sections of both slopes were nearly 0°C , while the middle and upper sections of the slope were at or above the air temperature. It is clear that the thermal characteristics of the talus vary with the season. During the freezing season, the middle and upper slopes were sufficiently different from the air tempera-

Maximum	Air (°C)	Lower (°C)	Middle (°C)	Upper (°C)
DH annual	22.09	3.27	21.74	22.22
OC annual	18.94	2.11	20.37	23.63
DH freezing season	4.39	0.06	3.85	6.39
OC freezing season	3.66	-0.59	2.79	6.68
DH thawing season	22.09	3.27	21.74	22.22
OC thawing season	18.94	1.93	20.37	23.63
Mean	Air (°C)	Lower (°C)	Middle (°C)	Upper (°C)
DH annual	3.01	-3.02	3.61	6.18
OC annual	1.48	-5.24	2.63	6.33
DH freezing season	-8.62	-8.61	-8.98	-3.12
OC freezing season	-8.86	-11.89	-5.03	-1.20
DH thawing season	10.17	0.06	10.75	11.70
OC thawing season	9.74	0.10	10.28	12.20
Minimum	Air (°C)	Lower (°C)	Middle (°C)	Upper (°C)
DH annual	-24.85	-16.89	-22.84	-14.03
OC annual	-25.41	-19.26	-12.68	-10.70
DH freezing season	-24.85	-16.89	-22.84	-14.03
OC freezing season	-25.41	-19.26	-12.68	-10.70
DH thawing season	-5.40	-0.61	-0.93	-0.24
OC thawing season	0.32	-7.15	-1.41	0.18

Table 4.3: Annual, freezing season, and thawing season temperatures for air and lower, middle and upper talus sites from 1 September 2016 to 31 August 2017 at Doghead Mountain (DH) and Ouiment Canyon (OC). The temperatures were calculated from all sensors in the talus, i.e. surface sensors were excluded.

ture that, according to the values in Fig. 2.6, an unstable environment such as outlined by Cheng *et al.* (2007) could have developed. During the thawing season, the lack of temperature difference suggests that the environment would have been stratified and stable (Cheng *et al.*, 2007).

The annual, freezing season, and thawing season means all indicate that temperatures were not uniform across the talus. This would not be anticipated for most slopes with a uniform aspect and minor changes in angle and elevation. There was no difference in the amount of shading that different parts of the slope received at Ouimet Canyon, only in the daily timing of shading. At Doghead Mountain, the lowest part of the talus containing both the lower and air temperature sensors was forested and covered by mosses, and therefore shaded, but the rest of the slope was not. The similarity between the means for the air and lower sensors at Ouimet Canyon and Doghead Mountain indicates that the shading at Doghead Mountain was not as critical for talus temperature as other factors. In fact, the shaded talus at Doghead Mountain was warmer overall than the exposed talus at Ouimet Canyon. The lack of shade on the middle and upper sections of both talus slopes may be responsible for thawing season talus temperatures being higher than air temperatures because the rock has a relatively low albedo so the net radiation at the surface is high.

The FDD and TDD further illustrate that there are differences between sections of the talus (Table 4.4). At both Doghead Mountain and Ouimet Canyon, FDD decreased and TDD increased with elevation. Compared to

Site	Air (degree-days)	Lower (degree-days)	Middle (degree-days)	Upper (degree-days)
DH FDD	1170.49	1203.60	1094.56	394.97
DH TDD	2216.05	114.71	2408.36	2650.96
OC FDD	1463.9	1995.78	831.65	289.58
OC TDD	1957.87	88.46	1682.16	2558.87

Table 4.4: Mean freezing degree-days (FDD) and thawing degree-days (TDD) for air and talus temperature sensors from 1 September 2016 to 31 August 2017 at Doghead Mountain (DH) and Ouiment Canyon (OC). Lower, middle and upper sensors' surface temperatures were not included in calculating the cumulative degree-days.

the FDD and TDD of the air, the lower sections of the talus slopes were colder and the upper sections of the talus slopes were warmer. The upper sections of the slopes experienced little freezing, while the lower sections experienced more freezing than indicated by the air temperature and experienced little thawing. DHM had slightly fewer FDD and more TDD than the air, while OCM had considerably fewer FDD and fewer TDD than the air. DHM also experienced more thawing and freezing than OCM. The difference in ground thermal conditions between the two sites is not explained by air temperature, aspect, or shading.

4.3.1 Conditions at each site in talus

Annual mean temperatures for each sensor (Table 4.5) reveal that there does not appear to be a consistent relationship between depth and temperature in the talus; at some locations annual mean temperature increased with depth, and at others it decreased. The freezing season mean temperatures (Table 4.6) also showed no clear relationship between depth and temperature, except perhaps at OCU. The thawing season mean temperatures (Table 4.7) show no relationship between temperature and depth among the middle and upper sensors, but mean temperatures decreased with depth at the lower section sensors. These inconsistent relationships between depth and temperature are unusual, as an increase in depth should cause dampening of temperature change in a conductive environment, with deeper locations slightly warmer during the freezing season and slightly cooler in the thawing season.

Logger	Mean temp. A (°C)	Mean temp. B (°C)	Mean temp. C (°C)	Mean temp. D (°C)
DHA	3.01	-	-	-
DHU	7.90 (S)	6.22 (0.4A)	5.95 (0.4B)	6.37 (0.5)
DHM	6.65 (S)*	4.03 (0.7)*	3.58 (1.4)	3.22 (1.45)*
DHL	-2.62 (S)	-3.02 (0.7)	-3.02 (1.35)	-3.00 (1.75)
OCA	1.48	-	-	-
OCU	5.82 (S)	6.45 (0.46)	6.20 (0.73)	6.35 (0.77)
OCM	4.11 (S)	2.42 (0.5)	2.67 (0.65)	2.78 (0.66)
OCL	-4.37 (S)	-4.60 (0.8)	-6.10 (1.2)	-5.01 (1.5)

Table 4.5: Annual mean temperatures for each logger from 1 September 2016 to 31 August 2017. Sensor depths (m) are in brackets. * indicates sensors that are vertically aligned.

Logger	Mean temp. A (°C)	Mean temp. B (°C)	Mean temp. C (°C)	Mean temp. D (°C)
DHA	-10.09	-	-	-
DHU	-3.21 (S)	-1.58 (0.4A)	-3.63 (0.4B)	-4.14 (0.5)
DHM	-8.34 (S)*	-9.66 (0.7)*	-7.96 (1.4)	-9.32 (1.45)*
DHL	-8.07 (S)	-9.02 (0.7)	-8.65 (1.35)	-8.17 (1.75)
OCA	-9.89	-	-	-
OCU	-4.30 (S)	-1.86 (0.46)	-0.96 (0.73)	-1.55 (0.77)
OCM	-5.60 (S)	-5.59 (0.5)	-5.85 (0.65)	-5.14 (0.66)
OCL	-12.69 (S)	-11.76 (0.8)	-14.46 (1.2)	-10.92 (1.5)

Table 4.6: Freezing season mean temperatures for each sensor. Sensor depths (m) are in brackets. * indicates sensors that are vertically aligned.

Logger	Mean temp. A (°C)	Mean temp. B (°C)	Mean temp. C (°C)	Mean temp. D (°C)
DHA	10.65	-	-	-
DHU	(S)	10.95 (0.4A)	11.60 (0.4B)	12.54 (0.5)
DHM	(S)*	11.73 (0.7)*	10.19 (1.4)	10.34 (1.45)*
DHL	0.40 (S)	0.17 (0.7)	0.16 (1.35)	-0.16 (1.75)
OCA	9.74	-	-	-
OCU	(S)	12.66 (0.46)	11.69 (0.73)	12.24 (0.77)
OCM	(S)	9.94 (0.5)	10.66 (0.65)	10.24 (0.66)
OCL	1.94 (S)	0.71 (0.8)	0.10 (1.2)	-0.53 (1.5)

Table 4.7: Thawing season mean temperatures for each sensor. Sensor depths (m) are in brackets. * indicates sensors that are vertically aligned. Surface temperatures from sensors that were protected by the reflective bubble wrap were not included because they are unlikely to be representative of field conditions.

Annual mean temperatures from sensors in the lower sections of the talus slopes were uniformly negative and at least 5°C lower than the annual mean temperatures for the air and other sections of the talus (Table 4.5). This indicates that a substantial cold thermal anomaly occurred in the lower sections of both slopes over the course of the year. During the freezing season, the mean temperatures at the lower sections of the slope were similar to the air temperature, but quite different from the upper sections (Table 4.6). The thawing season mean temperatures at the DHL and OCL sensors were at least 8°C lower than all other sensors, indicating that the anomaly was most pronounced during the thawing season (Table 4.7). The sensors at the deepest parts of the lower section of the talus had thawing season mean temperatures $< 0^{\circ}\text{C}$.

Annual mean temperatures at depth at DHM approximated the annual mean air temperature (DHA), while shallower sensors at DHM, and all sensors at OCM, OCU, and DHU, had higher annual mean temperatures than the air (Table 4.5). As discussed above, such temperature variation with elevation is not commonly reported and cannot be explained by shading or slope geometry.

The thawing season mean temperatures at the middle and upper sections of the slopes were near or just above the air temperature (Table 4.7), providing further evidence of the thermal stability of the talus slopes during the thawing season.

The maximum and minimum daily mean temperatures for each sensor

(Fig. 4.7) indicate that even at roughly 1.5 m depth at several sites in the talus, the range between the maximum and minimum daily mean temperatures remained $> 15^{\circ}\text{C}$. Figure 4.7 also shows that maximum and minimum temperatures did not dampen with depth in a consistent way. At the lower sites, the surface temperatures were reliable throughout the year. The maximum temperatures showed dampening with depth, but minimum temperatures did not, leading some deeper sensors to experience a larger temperature range than shallower ones or the surface. The coherence of the maximum daily mean temperatures suggests that the outlying points among the minimum daily mean temperatures were not due to instrument error, as a systematic error should have displaced both data points. This indicates that at these sites, heat transfer was different in the freezing and thawing seasons, and suggests that the thawing season had stable conditions while the freezing season did not.

Consistency was observed in the behaviour of most sensors, however. If a sensor's maximum was less than its neighbouring sensors', in most cases its minimum would be greater, and *vice versa*. This suggests that conduction occurred in some places and not others, due to the heterogeneous ground materials. The lack of vertical alignment among most of the sensors may also account for some of this, but even the vertically aligned sensors at DHM (0 m, -0.7 m and -1.45 m) were ambiguous. The amplitudes at DHM were similar to that of the air temperature even at 1.45 m depth, indicating a close relationship to the atmosphere and little conductive flux.

Fig. 4.7: Minimum (blue circles) and maximum (red diamonds) daily mean temperatures for a) DHU, b) DHM, c) DHL, d) OCU, e) OCM, and f) OCL. Maximum daily mean surface temperatures at sites that used the reflective bubble wrap are shown by pink squares.

4.3.2 Ground conditions during the period of snow cover

Snow depths were recorded close to the instrumented slopes (but not on them) during basal temperature of snow measurements in late February and early March 2017. Mean snow depth was 61 cm deep at Doghead Mountain and 84 cm deep in the southern part of Ouimet Canyon. The surface temperature sensors were used to reconstruct the period of snow cover at the talus slopes (Table 4.8). Freezing season daily surface temperature range and trends indicated when the snow pack was thick enough to insulate the surface from the direct effects of the atmosphere. At the middle and upper slopes of Doghead Mountain, snow cover first developed on 19 November 2016 but was not completely established until 6 December (Fig. 4.8). The daily surface temperature ranges at DHL were small year-round, so it was not possible to determine when snow accumulated or ablated there.

OCU experienced a mix of a few days of low ranges followed by a day or two of medium ranges from 20 November to 27 December 2016, which may indicate snow accumulation followed by its removal by wind or melting (Fig. 4.9). OCL-S experienced zero curtains in both autumn and spring, which obscured when accumulation and ablation occurred. OCL-S was likely snow-covered by 23 November 2016 as it was steadily cooling to -5°C while at the other sensors the temperature was either near 0°C or rising. Snow lasted until at least 18 April 2017, when sub-zero diurnal fluctuation in tandem

Logger	Start date	End date	Length (days)
DHU	6 Dec 2016	22 Feb 2017	79
DHM	6 Dec 2016	22 Feb 2017	79
DHL	-	-	-
OCU	27 Dec 2016	1 Apr 2017	96
OCM	19 Nov 2016	7 Apr 2017	139
OCL	23 Nov 2016	18 Apr 2017*	147*

Table 4.8: The estimated minimum period of snow cover, based on daily surface temperature ranges and trends at each logger. DHL daily surface temperature ranges were small year-round, so no information about the snow cover could be gained from it. *OCL experienced a zero-curtain which made it difficult to determine when ablation occurred. This figure represents the earliest possible date.

Fig. 4.8: Air and surface temperatures for the snow-covered season at Doghead Mountain.

Fig. 4.9: Air and surface temperatures for the snow-covered season at Ouimet Canyon.

with, but a several degrees lower than, air temperature was observed over a period of warm weather. However, after these fluctuations, temperatures remained below zero with no response to air temperature until entering a zero curtain on 15 May. An alternate explanation for this behaviour will be discussed in section 4.4.1.

The timing of snow ablation was quite different at Doghead Mountain and Ouimet Canyon. The middle and upper sections of the Doghead Mountain talus slope lost their snow during a period of warm weather between 18-22 February, when daily mean air temperatures were $> 0^{\circ}\text{C}$ on all five days (Fig. 4.8). During this period, daily mean air temperatures at Ouimet Canyon were $< 0^{\circ}\text{C}$, and the snow cover persisted until April (Fig. 4.9). The longevity of the snow pack at Ouimet Canyon decreased with elevation. While a snow cover capable of providing insulation was likely gone by late April at both sites, there was granular snow between rocks, i.e., in the talus, at both DHL and OCL in early June 2017.

4.3.2.1 Insulation of the surface from the atmosphere

Comparison of the air temperature and surface temperatures at each site reveals that the talus surface at all locations became sufficiently insulated from the atmosphere that no diurnal air temperature fluctuations were recorded (Figs. 4.8 and 4.9), although temperature fluctuations less than 0.45°C would not have been registered by the sensors. Not all sections of the talus were equally insulated from the atmosphere, however. Only DHU-S (Fig. 4.8) and

OCL-S (Fig. 4.9) were fully insulated, while the other surface temperatures showed some dampened and lagging response to air temperatures (Fig. 4.10).

4.3.2.2 Conditions in the ground

Temperatures for all surface and internal sensors for January or February 2017 can be found in Figs. 4.11 and 4.12 for Doghead Mountain and Figs. 4.13 and 4.14 for Ouimet Canyon. Dampening and lagging of temperature with depth occurs in conductive systems, as at DHU-0.4A, OCL-0.8, and OCL-1.5, indicating that surface temperature was likely the mechanism responsible for temperature change at these sensors. However, most sensors did not show consistent lagging or dampening, and instead paralleled, anticipated, or overshoot surface temperature changes. This indicates that the surface temperature was not the only forcing mechanism at work in the talus. At DHL-0.7, DHM-0.7, and OCL-1.2, internal temperatures followed air temperatures more closely than surface temperatures, indicating an ongoing connection to the atmosphere, likely through chimneys or wind holes. These sensors closely responded to drops in air temperature, but responded to rises more gradually (Fig. 4.15). This shows that air rapidly entered some sections of the talus during cold weather, but that warm air was not always able to do the same, particularly in the lowest parts of the talus.

An examination of the graphs site-by-site reveals other non-conductive behaviour. At DHU, the deepest sensor was consistently the coldest, while the two sensors at 0.4 m depth acted very differently, such as from 22 to

Fig. 4.10: Examples of partial insulation from DHL-S and DHM-S. Both sensors showed a dampened and lagging response to air temperatures that smoothed out the diurnal fluctuation but retained the trends.

Fig. 4.11: Temperatures at all DHU sensors in February 2017, showing the response of internal talus sensors to changes in temperature at the surface sensor.

Fig. 4.12: Temperatures at a) all DHM sensors and b) all DHL sensors in January 2017, showing the response of internal talus sensors to changes in temperature at the surface sensor.

Fig. 4.13: Temperatures at a) all OCU sensors and b) all OCM sensors in February 2017 showing the response of internal talus sensors to changes in temperature at the surface sensor.

Fig. 4.14: Temperatures at all OCL sensors in January 2017 showing the response of internal talus sensors to changes in temperature at the surface sensor.

Fig. 4.15: Responses of some internal talus sensors at a) Doghead Mountain and b) Ouimet Canyon to air temperatures in January 2017. The sensors followed air temperatures but responded more rapidly to decreases in air temperature than increases.

28 February (Fig. 4.11). Meanwhile, the sensors at OCM behaved nearly identically through the snow-covered season in spite of the 0.5 m or more between the surface and the internal sensors (Fig. 4.13).

The site-by-site examination also reveals that temperatures at deeper internal locations were at times warmer than those at shallower locations or the surface, for example during the first week of January at DHL (Fig. 4.12), the last week of February at DHU (Fig. 4.11), all of January at OCL (Fig. 4.14), and most of February at OCU (Fig. 4.13). It is possible to determine whether the temperature differences in the talus were large enough to reach the critical filtration Rayleigh number (Ra_c^*), but there are two caveats. First, the majority of sensors were not vertically aligned, so the temperature gradients do not represent a column in which the top and bottom are parallel or even adjacent. Furthermore, the heterogeneous ground materials mean that heat movement is not the same from column to column. The measured temperatures are specific to their location. Second, even the deepest measurements may not be an accurate representation of conditions at the bottom of the talus layer because it is not known how far the sensors are from it, and the depth of the porous layer can dramatically change the temperature difference required to reach Ra_c^* (Fig. 2.6).

However, if $9^\circ\text{C}/1\text{ m}$ is used as a test value, none of the examples above would have had a high enough Ra_c^* for convection to occur between the sensors. If the talus surface temperatures are compared to the air temperatures, several periods in which the difference was $>9^\circ\text{C}$ were observed:

mid-November (DHL-S, DHU-S, and OCL-S), mid-December (DHL-S and DHU-S), early January (DHU-S, OCM-S and OCU-S), and early February (DHU-S and OCU-S) (Figs. 4.8 and 4.9). Thus it was possible for convection to occur between the slope and the atmosphere, as described by Balch (1900), which could have led to the development of whole-slope convection cells as described by Bories & Combarrous (1973) and Cheng *et al.* (2007). The existence of convection within the slopes is supported by the relatively high temperatures recorded by the sensors in the upper sections of the slopes and the lower temperatures recorded in the middle and lower sections of the slopes.

4.3.3 Effects of winter convection

The thermal behaviour of talus slopes was highly variable through the winter, as different parts of the slopes responded differently to atmospheric and surface temperatures. Connections to the atmosphere and whole-slope convection allowed cold air to accumulate in the lower part of the talus while warm air flowed upward, causing dramatically different mean temperatures, FDD and TDD in the lower, middle and upper sections of the talus slopes. The low temperatures in the lower sections of the talus slopes persisted through the spring and into the summer.

4.3.4 Ground conditions during the summer

In contrast to the variable behaviour observed during the snow-covered and snow melt seasons, the thermal behaviour of the different sections of the talus slope and the sensors within the talus was quite regular during the summer. Temperatures in the upper and middle sections of the slope were closely related to the air temperatures at both Doghead Mountain (Fig. 4.16) and Ouimet Canyon (Fig. 4.17). Diurnal fluctuations were observed at all sensors in these sections. Inconsistent dampening and lagging with depth, most obvious between DHU-0.4A and DHU-0.5 (Fig. 4.16) or OCU-0.73 and OCU-0.77 (Fig. 4.17), indicated heterogenous conductive heat transfer occurred within the talus, but there was no evidence of convection. This agrees with Cheng *et al.* (2007)'s model showing summertime heat transfer in an ACE as mostly conductive.

The similarity in temperature trends between the middle and upper sections of the talus slopes suggests that both sections experienced similar conditions. In contrast, the lower sections of the talus slopes were $<0^{\circ}\text{C}$ and unresponsive to air temperature for much or all of the summer, experiencing a prolonged zero-curtain (Fig. 4.18). Where the ground came out of zero curtain, two different behaviours were observed. In most cases, fluctuations in the temperature appeared to be sporadic and without a consistent response to diurnal trends (Figs. 4.19 and 4.20). In some cases, short-lived changes in temperature corresponded to days with precipitation, suggesting

Fig. 4.16: Air and internal talus temperatures at a) DHU and b) DHM for the first two weeks of July 2017. Internal talus temperatures followed air temperatures but were dampened and lagged with depth.

Fig. 4.17: Air and internal talus temperatures at a) OCU and b) OCM for the first two weeks of July 2017. Internal talus temperatures followed air temperatures but were dampened and lagged with depth.

Fig. 4.18: Air, surface, and internal talus temperatures at a) DHL and b) OCL for the first two weeks of July 2017. Most sensors were still in zero curtain and did not respond to fluctuations in air temperature. OCL fluctuated with air temperature but was colder.

Fig. 4.19: Surface and internal talus temperatures at DHL and precipitation at Cameron Falls for August 2017.

Fig. 4.20: Surface and internal talus temperatures at OCL and precipitation at Thunder Bay CS for August 2017.

they represent the arrival of infiltrating rain water at the sensor. At OCL-S and OCL-0.8, diurnal fluctuations appeared and there was lagging and dampening with depth, but temperatures remained considerably lower than the air and the other sections of the talus, and fluctuations were smaller in magnitude.

4.3.5 Effects of summer stratification

Taken together, the summer behaviour of the different slope sections at both sites suggests that air within the talus slopes was stable and stratified. This agrees with the summertime thermal stratification in air convection embankments observed and modelled by Goering & Kumar (1996), Goering (1998) and Cheng *et al.* (2007). This stratification prevented mixing of the air in the talus and allowed the lower sections of the talus slopes to remain cold and unresponsive to air temperatures through much or all of the thaw season.

4.4 Ice accumulation and ablation in the talus

The ground temperature records from the lower sections of the talus during the summer indicate that the zero curtain was very important to the thermal regime of this part of the talus. The existence of the zero curtain indicated the presence of at least one body of ice within both talus slopes. This ice melted back into the talus as the warm season progressed, but was still visible between blocks on 1 September 2016 and 24 August 2017 at Doghead

Mountain and 14 October 2016 and 11 October 2017 at Ouimet Canyon. The ice was likely composed of snow and refrozen snow melt from the vicinity of the ice and refrozen meltwater from other parts of the talus.

4.4.1 Melting, infiltration, and refreezing

Melting of snow and ice on the talus took place during several warm weather events in late winter and early spring. At the time, temperatures at the lower sections of the slopes were still $<0^{\circ}\text{C}$ and had yet to enter zero curtain, so temperature increases resulting from the arrival of water at the lower section of the talus could be observed.

In response to the warm weather 18-22 February, all sensors at DHU and all sensors but one at DHM approached 0°C for at least a day (Fig. 4.21). The surface sensors at both sites then began to respond to diurnal fluctuations of air temperatures. This represents a zero curtain for the duration of snow melt, followed by recoupling with the atmosphere once the snow cover was thin enough. At the same time, a brief spike in temperature to near 0°C was observed at DHL-S, representing the presence of water at the sensor. The sudden warming was also observed at DHL-1.35, suggesting that it was not a sensor error. The water at DHL-S quickly refroze, but it led to an increase in temperature at both DHL-S and DHL-1.35 that gradually dissipated as cold weather returned.

During the same period at Ouimet Canyon, temperatures at all OCU and OCM sensors stayed near 0°C for several days, indicating a period of melting

Fig. 4.21: Temperatures for several sensors at Doghead Mountain from 16 February to 3 March 2017. Temperatures at sensors higher on the slope neared 0°C, followed by sudden warming at sensors in the lower section of the slope, suggesting infiltration of meltwater. The melting event was followed by cooling and refreezing.

and refreezing to depths of > 0.5 m (Fig. 4.22). Temperatures at most of the sensors at OCL remained below -12°C during this time, but the temperature at OCL-1.5 spiked briefly on 20 February, indicating the possible infiltration of meltwater there.

Snow pits dug near the bottom of the talus at Doghead Mountain on 4 March and Ouimet Canyon on 5 March encountered layers of ice within and at the bottom of the snow pack, indicating that melting and infiltration had taken place on at least one prior occasion and that water had refrozen within and at the bottom of the snow pack.

Other examples of snow melt and water infiltration occurred during short periods of warm weather in early March (Figs. 4.23 and 4.24), and early April (Fig. 4.25). As with the incident described above, temperatures at most sensors in the upper and middle sections of the talus slopes neared 0°C . Temperatures in the lower sections and at some deeper sensors in the middle sections of the talus experienced spikes or step changes in temperature but remained $<0^{\circ}\text{C}$. Any water arriving there would have refrozen. On March 6, temperatures neared 0°C as deep as 1.45 m within the talus at DHM (Fig. 4.23). Shallower sensors warmer earlier, and this downward heat movement at DHM is evidence of liquid water.

From mid-April onward, melt and infiltration events at Doghead Mountain could not be discerned because the sensors were either fully coupled with the atmosphere or in zero curtain. Two other events may have occurred at Ouimet Canyon in late April and early May (Fig. 4.26), before it too re-

Fig. 4.22: Temperatures for several sensors at Ouimet Canyon from 16 to 27 February 2017. Temperatures at sensors higher on the slope approached 0°C, followed by sudden warming at sensors in the lower section of the slope, suggesting infiltration of meltwater. The melting event was followed by cooling and refreezing.

Fig. 4.23: Temperatures for several sensors at Doghead Mountain from 4 to 12 March 2017, showing warming in the upper sections of the slope followed by possible infiltration of meltwater deeper into the talus and into the lower section of the slope. The melting event was followed by cooling and refreezing.

Fig. 4.24: Temperatures for several sensors at Ouimet Canyon from 4 to 12 March 2017. Temperatures at OCU and OCM approached 0°C, followed by sudden warming at sensors at OCB, suggesting infiltration of meltwater. The melting event was followed by cooling and refreezing.

Fig. 4.25: Temperatures at a) Ouimet Canyon and b) Doghead Mountain from 1 to 9 April 2017. Temperatures at OCU-S, and at DHU and DHM were often $>0^{\circ}\text{C}$ and coupled to air temperatures. Temperatures in the lower sections of the slopes spiked or rose slightly, suggesting infiltration of meltwater. While the temperature increase at DHL-S is more ambiguous than those at OCL, it occurred at the same time that internal sensors in the upper and middle sections recoupled with the atmosphere, suggesting loss of insulating material, and the temperature increase suggests that a substantial amount of heat arrived at DHL-S.

Fig. 4.26: Temperatures for several sensors at Ouimet Canyon in late April and early May 2017, during snowmelt. Temperatures at sensors as deep as OCU-0.77 and OCM-0.65 were $\geq 0^\circ\text{C}$ much of the time, indicating thawing and freezing. Temperatures at OCL-S and OCL-1.2 spiked and rose gradually, suggesting infiltration of meltwater.

coupled and entered zero curtain. The fluctuations in OCL-S at this time may be interpreted as influxes of meltwater to the vicinity of the sensors during the warmest parts of the day. They may also have represented the loss of the insulating snow cover, as discussed in section 4.3.2, although OCL temperatures were considerably below the air temperature.

This kind of springtime melting and refreezing has been credited with the bulk of the ice formation in talus slopes and cellars attached to blocky materials (Ohata *et al.*, 1994; Tanaka *et al.*, 2000; Sawada, 2003; Zacharda *et al.*, 2007; Sawada, 2017). The long zero curtains observed at some depths suggest that enough ice accumulated in the talus to have a significant effect on the thermal regime.

4.4.2 Ablation

Sensors in the lower sections of both Doghead Mountain and Ouimet Canyon talus slopes experienced long zero curtains in 2016 and 2017. Interstitial ice was found in the lower sections of both talus slopes during both years, as was standing water. Due to the length of the thawing season, some information from before 1 September 2016 and after 31 August 2017 will be introduced below in order to provide context.

4.4.2.1 Interstitial ice and the zero curtain

When the DHL sensors were installed on 14 July 2016, DHL-1.35 and DHL-1.75 were placed near or below interstitial ice and returned temperatures

$<0^{\circ}\text{C}$. Ice was found in another pit (Pit 2) in the talus as well (Fig. 4.27a). When the OCL sensors were installed on 20 July 2016, ice was found in between blocks (Fig. 4.28a). OCL-0.8 and OCL-1.5 were placed near and on top of the ice, respectively, and OCL-1.2 was placed into a cold void between rocks. Only OCL-1.2 returned temperatures $<0^{\circ}\text{C}$.

As the summer progressed, the interstitial ice melted back into the talus as the ground thawed at depth. Ice was observed in Pit 2 on 1 September 2016 (Fig. 4.27d, e) but not on 13 October. Table 4.9 lists the known thaw dates for DHL and OCL sensors for 2016. Where there was variability around 0°C before the ground thawed completely, the thaw date was defined as the date on which the number of consecutive hours with a temperature $< 0^{\circ}\text{C}$ was less than the previous consecutive hours with a temperature $> 0^{\circ}\text{C}$. All sensors had thawed by mid-September, but some ice remained near OCL-1.5 on 14 October 2016 (Fig. 4.28b). The temperature record for OCL-1.5 unfortunately does not reflect the persistence of ice nearby because the ice melted away from the sensor. The sensor had been placed on top of an ice body on the assumption that it would stay in contact with the ice surface as it melted downward. Instead, the sensor became stranded on a rock. As a result, OCL-1.5 was moved twice to put it back in contact with the ice body, lastly on 14 October 2016. After 14 October 2016 the sensor's position was constant, and it stayed $<0^{\circ}\text{C}$ until the record ended on 17 August 2018 (Fig. 4.29). The combined evidence of the interstitial ice in the talus and the OCL-1.5 temperature record after October 2016 indicates that some parts of

Fig. 4.27: Interstitial ice at Doghead Mountain, in a 0.6 m-deep pit (Pit 2) at the bottom of the talus about 30 m ESE of the DHL and DHA loggers. a) shows the entire pit. The white dashed line indicates the extent of the ice visible in the pit. All subsequent photos were taken crouched inside Pit 2 looking left (b and d) and right (c and e). Blue dashed lines show the surface of standing water.

Fig. 4.28: Interstitial ice at Ouimet Canyon, at the site where sensor OCL-1.5 was installed. a) was taken from just outside the cavity that contained the ice that OCL-1.5 was placed on. b) was taken roughly from the location of the ice in a), looking left. OCL-1.5's cable is visible on the right, in its final position. The orange line outlines a reference log for comparing this photo with Fig. 4.30.

Logger	2016		2017		
	Thaw date	Start date	Thaw date	Length (days)	Change in thaw date (days)
DHL-S	<14 July	9 April	1 August	115	>+19
DHL-0.7	<14 July	14 April	23 August	132	>+41
DHL-1.35	18 July	16 April	24 August	131	+37
DHL-1.75	10 September	9 April	30 September	175	+20
OCL-S	<14 July	5 May	14 June	41	?
OCL-0.8	<14 July	2 May	28 June	58	?
OCL-1.2	26 July	3 May	10 August	100	+15
OCL-1.5	<14 July*	5 May**	n/a*	≈176*	n/a

Table 4.9: The estimated thawing zero curtain period for each sensor in the lower section of both talus slopes in 2016 and 2017. *OCL-1.5's 2016 thaw date reflects OCL-1.5's original position 1.1 m from the surface, and the 2017 data reflect its position as of 14 October 2016, 1.5 m from the surface. **Ground temperatures on this date were -2.44°C. Water at -1.5 m may not have been changing state yet, but from this time temperatures gradually increased.

Fig. 4.29: The temperature record for OCL-1.5 from 1 October 2016 until the end of the record on 17 August 2018. After the sensor was placed back in contact with interstitial ice on 14 October 2016, the temperature stayed $>0^{\circ}\text{C}$.

the Ouimet Canyon talus did not thaw in 2016 or 2017.

Thaw progressed more slowly in 2017. Table 4.9 compares the date that the ground at each sensor thawed in 2016 and 2017. The timing of thaw varied more between years at Doghead Mountain than at Ouimet Canyon. Ice, still bonded to the rocks, was confirmed in Pit 2 at Doghead Mountain on 24 August 2017, but had melted by the next visit on 5 October. Ice was confirmed at OCL-1.5 through 11 October 2017. The ice did not appear to have retreated as far into the talus as on 14 October 2016 (Fig. 4.30), and the end of the OCL-1.5 cable was still trapped in the ice.

4.4.2.2 Water accumulations

During the ablation period, water accumulated in some parts of the talus. In both 2016 and 2017, meltwater formed puddles around the ice in Pit 2 at Doghead Mountain (Fig. 4.27b-e), remaining well into October. Ponds of standing water were also observed in low points of the southern end of Ouimet Canyon on 7 June (Fig. 4.31). The water disappeared sometime before 15 August 2017. Water would typically drain easily through the void spaces in the talus. Standing water at low points in the talus suggests that drainage may be temporarily blocked by ice deeper within the talus.

4.4.3 Effects of ice in the talus

Taken together, the buildup and ablation of interstitial ice and lengthy zero curtains observed at DHL and OCL revealed a control on the thermal regime

Fig. 4.30: Interstitial ice at Ouimet Canyon on 11 October 2017, at the site where sensor OCL-1.5 was installed. The orange line outlines a reference log for comparing this photo with Fig. 4.28.

Fig. 4.31: Standing water ponded in the talus in the southern part of Ouimet Canyon, 7 June 2017. (Contrast and saturation were altered in order to distinguish the water from the shadow of the cliff.)

of the talus slopes. Parts of the talus stayed below 0°C into or through the summer because they were in zero curtain. They were in zero curtain because there was melting ice, and where the ice did not completely melt before freezeback began, the talus did not thaw. Thus, the persistence of cold thermal anomalies into the summer and presence of permafrost in the talus slopes depends on the accumulation of ice inside the talus and its rate of melting.

4.5 Summary

Annual mean internal talus temperatures increased with elevation. The upper sections of the talus were warmer, the middle sections were similar, and the lower sections of the talus were colder than the air temperature. During the winter, the lower sections and DHM followed air temperatures, while the upper sections and OCM were steadier and close to 0°C . This suggests that slope-scale convection occurred. In the summer, the slopes were thermally stratified, and there was evidence of conductive heat transfer. The lower sections of the slope stayed near 0°C and experienced long zero curtains.

Ice accumulated in the talus, and temperature records linked the formation of ice to meltwater infiltration. The long zero curtains in the lower sections of the talus were caused by melting ice, so ice was important to the existence of thermal anomalies in the talus. In Ouimet Canyon, the ice did not thaw completely between July 2016 and August 2018, indicating there

was permafrost in the talus.

Chapter 5

Discussion

5.1 Research objectives and hypotheses

This thesis focused on two objectives: 1) to determine the thermal regime of two talus slopes at Doghead Mountain and Ouimet Canyon over one year; and 2) to investigate field conditions that indicate presence of permafrost in talus. Additionally, there were two hypotheses: 1) there are isolated patches of permafrost in talus in Northwestern Ontario; and 2) that the thermal regime within the talus is significant influenced by convective heat transfer.

5.2 Thermal regime of the talus slopes

The thermal regimes of the Doghead Mountain and Ouimet Canyon talus slopes may be considered in terms of changes over space and time. For annual and seasonal mean temperatures, the upper sections of the talus were warmer and the lower sections of the talus were colder than the air temperature. FDD

and TDD revealed large differences in thermal state with elevation. The upper sections of the slopes experienced little freezing, while lower sections of the slope experienced little thawing.

The talus slopes had distinct thermal regimes in the freezing season and thawing seasons. During the freezing season, mean temperatures in the lower sections of the talus approximated air temperatures, while the upper sections of the talus stayed relatively warm. The freezing season mean temperatures for the middle and upper sections were higher than the air temperature. Talus temperatures were unstable in this season, especially in the lower sections. At Doghead Mountain, the middle section behaved more like the lower section, while at Ouimet Canyon the middle behaved more like the upper section, suggesting that either a continuum or two distinct regions existed up and down the slopes.

In the spring, temperatures at sensors in the middle and upper sections of the slopes began to approximate air temperatures as soon as they had thawed (Figs. 4.8, 4.9 and 4.25). After that, temperatures in the upper and middle sections of the talus followed fluctuations in air temperature. During this time, the lower sections of the talus warmed until they neared 0°C, at which point the temperature stabilized.

During the thawing season, talus temperatures were generally stable and the talus was thermally stratified. Mean temperatures in the middle and upper sections of the talus approximated and exceeded, respectively, boundary layer temperatures. The lower sections of the talus remained near or below

0°C, harboured interstitial ice, and experienced long zero curtains, suggesting that a considerable amount of ice accumulated in the talus. Thaw occurred first in the upper section and last in the lower section, and penetrated downward from the surface.

The thermal regimes of the Doghead Mountain and Ouimet Canyon talus slopes were similar to those described at other slopes that harbour cold thermal anomalies, perennial ice, or permafrost. Seasonal changes in the relationship between talus and boundary layer temperatures have been described in Switzerland (Delaloye *et al.*, 2003), Germany (Gude *et al.*, 2003), and the Czech Republic (Gude *et al.*, 2003; Zacharda *et al.*, 2007). Temperatures at some of these site were similar to Doghead Mountain and Ouimet Canyon. Gude *et al.* (2003) reported annual means of 8.5°C and 0.1°C at the top and bottom of their slope, respectively. Zacharda *et al.* (2007) reported minimum internal talus temperatures between -6°C and -17°C, with mean internal temperatures between -0.1°C and -1.6°C and maximum internal temperatures from 0.3°C to -8°C. Doghead Mountain and Ouimet Canyon are colder than some sites, however, such as Delaloye *et al.* (2003) and Sawada *et al.* (2003), who report temperatures $\geq 0^\circ\text{C}$ in the upper sections of the talus slopes during the winter.

The presence of interstitial ice in the talus was noted in nearly all previous reports on this topic, and was closely related to lengthy zero curtains (Sawada, 2003; Zacharda *et al.*, 2007). Delaloye *et al.* (2003) and Gude *et al.* (2003) proposed that wintertime freezing allowed more ice to accumulate in

the talus than could melt during summer. Delaloye *et al.* (2003) stated that this is why permafrost existed in their talus, and that freezing produced a “cold reservoir” (Delaloye *et al.*, 2003, 179) in the form of latent heat of fusion. The accumulation of ice inside the talus and its rate of melting is thought to be a control on the thermal regime of the talus slope (Sawada, 2017), and this appeared also to be the case at Doghead Mountain and Ouimet Canyon. The formation of ice has been mainly attributed to percolation and refreezing of meltwater during snowmelt (Balch, 1900; Ohata *et al.*, 1994; Sawada, 2003; Zacharda *et al.*, 2007) (Fig. 5.1), and there was evidence for this at both Doghead Mountain and Ouimet Canyon.

After thawing, sensors in the lowest parts of the talus continued to experience low temperatures. This may have been the result of the continued presence of ice near the sensors (Zacharda *et al.*, 2007), but cold air drainage from the talus may have also contributed. Cold air sometimes accumulated in depressions in and near the talus, including the pits where the DHL and OCL sensors were installed. The longevity and areal extent of these cold air pools may have been important to preventing heat from penetrating the lower parts of the talus. Given that most summertime heat transfer in the lowest section of the talus was conductive, cold air pools would have directly limited the energy available for thaw deeper in the talus by preventing the surface from warming to the boundary layer temperature. The thick, dry mosses and lichens in the bottom of the talus would have been an added barrier to the conduction of heat into the talus from the boundary layer (Stiegler

Fig. 5.1: Ice formation from the infiltration and refreezing of meltwater in the talus in spring.

et al., 2014; Kneisel *et al.*, 2015), and may also have restricted convection by filling or covering up holes between the blocks (Holmgren *et al.*, 2017a).

While dry moss is an excellent thermal insulator, wet moss has a much higher thermal conductivity (Kneisel *et al.*, 2015). Heat may also be carried into the talus by percolating rain water (Sawada, 2003), and the observations from Doghead Mountain and Ouimet Canyon indicated that temperatures in the lower sections of the talus were responsive to precipitation events. This suggests that a wet summer or autumn might increase thaw in the talus.

5.2.1 Does convective heat transfer exert significant influence on the thermal regime?

In short, it depends on scale and season. The rock size analysis indicates that many of the pores are greater than 10 cm diameter. Based on Figure 2.4a, at this scale free convection could have occurred within the pores as a response to even very small temperature gradients. It is likely that this pore-scale convection was important in transferring heat between rock surfaces. At the slope scale, the gradients across the whole talus layer were rarely large enough to provoke free convection, but there were occasions when the gradient between the atmosphere and the talus were sufficiently large, indicating that air exchange between the talus and the atmosphere was possible.

From the records at both slopes over the entire year, it is clear that heat does not move in a consistent way through the talus. There was evidence

of both convective and conductive heat transfer in all sections of the talus at different times. Convective heat transfer was responsible for the slope-wide temperature gradient that characterized the freezing season at both talus slopes. During many periods in winter, the temperature difference between the boundary layer and parts of the talus slopes could have provoked convection, as described by Gude *et al.* (2003). The increase in temperature with elevation in the slopes supports the existence of whole-slope convection cells (Bories & Combarrous, 1973; Cheng *et al.*, 2007). According to Bories & Combarrous (1973), the entire slope becomes a convection cell because the warm air must reach the top before it can exit the slope or lose enough heat to begin sinking, while the cold air continues to lose heat as it sinks. Cold air entered or accumulated in the lower part of the talus while warm air flowed upward, accumulating or exiting near the top (Fig. 5.2a). This resulted in dramatically different mean temperatures in the lower, middle and upper sections of the talus slopes (Table 4.3). The source of heat for the warm air may be the latent heat of freezing from any moisture in the talus (a much reduced version of the process described by Phillips *et al.* (2009)) and heat that penetrated blocks during the summer (Sawada, 2003). The cold air was a result of heat loss to the atmosphere through the snow pack (Delaloye *et al.*, 2003) and ingress through chimneys, where present (Gude *et al.*, 2003; Sawada *et al.*, 2003).

At a more local scale, convection contributed to the thermal regime at the sensors that paralleled, anticipated or overshoot surface temperature changes.

Fig. 5.2: Wintertime air movements in the talus slopes. a) shows the whole-slope convection cell that developed, with cold air entering and warm air leaving the slope through holes in the snow cover. b) shows the process at work when cold air drainage was observed. While some of the cold air mixes with the atmosphere, rising ground adjacent to the talus can trap the cold air in depressions. This kind of drainage was observed in summer as well.

In some places the influence of convection was so strong that sensors followed air temperatures more than surface temperatures, responding especially strongly to cold weather. This suggests that these sensors were near chimneys (Gude *et al.*, 2003). The Rayleigh number increases with the temperature gradient, meaning that the strongest potential for convection between the slope and the atmosphere is when there is cold weather, and the weakest in warm weather. This is likely why the lower parts of the talus responded so quickly to cold weather.

During the summer the thermal regime across the slope was thermally stratified and stable. This resembled the behaviour observed and modelled in air convection embankments (ACE) (Goering & Kumar, 1996; Goering, 1998; Cheng *et al.*, 2007) and in wind holes (Tanaka *et al.*, 2000). Cheng *et al.* (2007)'s model indicated that summertime heat transfer in an ACE was mostly conductive. The observations from both talus slopes corroborate this, though conductive heat transfer was inconsistent. The stable stratified environment does not mean that no air movement occurred within the talus, however. The stratification develops because air parcels move until they are in equilibrium with their surroundings.

Much of the literature links cold air drainage from the talus to warming of the upper parts of the slopes, proposing that the drainage of cold air from the bottom draws air from the boundary layer into the top of the talus (eg. Gude *et al.*, 2003; Zacharda *et al.*, 2007; Morard *et al.*, 2010). However, at Doghead Mountain and Ouimet Canyon, the sensors in the middle and upper

sections of the slopes began to approximate air temperatures as soon as they had thawed (Figs. 4.8, 4.9 and 4.25). This indicated that drainage of cold air from the bottom of the talus occurred more slowly than air exchange between the boundary layer and the upper and middle sections of the talus. Secondly, temperature gradients favourable to convection were rare. Localized convection could have occurred when and where the internal air was warmer than the boundary layer. The thawing season mean temperatures at the middle and upper sections of the slopes were similar to or warmer than the mean air temperatures (Tables 4.3 and 4.7), and the air in the boundary layer was often warmer than the air in the talus (Figs. 4.16 and 4.17). The boundary layer was only cooler than the middle and upper sections of the talus at night or on days of cool weather, but a temperature gradient $>9^{\circ}\text{C}/1\text{ m}$ was rare. The boundary layer was warmer than the lower sections of the talus during almost all of the thawing season (Fig. 4.18), so it seems unlikely that slope-scale convection occurred.

Nevertheless, cold air was observed draining and pooling in depressions at the bottom of the talus slopes, similar to Balch (1900), indicating some buoyancy-driven air movement occurred. The air that left the talus was replaced by air from the boundary layer near the elevation on the slope where the internal and external air were similar temperatures. While the entire process caused slow air movement, heat transfer occurred primarily by conduction because the air was stratified, and interacting with thermally similar air and surfaces. At no point in the boundary layer-talus interface

was warm air cooling and sinking, or cold air warming and rising, precluding the formation of convection cells and the mixing of air. As a result, the lower sections of the talus slopes were unresponsive to changes in the air temperature for most or all of the thawing season.

5.3 Permafrost in talus

5.3.1 Evidence for isolated patches of permafrost in talus in Northwestern Ontario

There is permafrost in Ouimet Canyon, based on the combined evidence of interstitial ice in the talus and the temperature record of OCL-1.5 from October 2016 through August 2018 (Fig. 4.29). The ice present in October 2016 had most likely formed in during a preceeding year. This is the first solid evidence of permafrost in this part of Ontario, and indicates that permafrost may exist in some talus slopes in Northwestern Ontario. Further field work is necessary to determine how widespread the phenomenon is. Talus slopes near the northern end of Lake Nipigon could be particularly good candidates because Ouimet Canyon's air temperatures were similar to the mean annual air temperature at Armstrong A station, which is 170 km to the north.

Permafrost was not confirmed at Doghead Mountain, but maximum temperatures at DHL-1.75 were close to 0°C: 2.03°C in 2016 and 0.73°C in 2017. Zacharda *et al.* (2007) assumed that internal temperatures stayed $\leq 4^\circ\text{C}$ in

their talus slopes “because deeper ground layers were still cryotic” (Zacharda *et al.*, 2007, 307). The difference in the thermal regime in 2016 and 2017 indicates that thaw penetration varies from year to year and suggests that permafrost may be temporally intermittent where it exists.

5.3.2 Field conditions that may indicate the presence of permafrost

The talus slopes and Doghead Mountain and Ouimet Canyon have many similarities to permafrost-bearing talus slopes in other studies: dense mosses on the lowest parts of the talus (Gray & Brown, 1979; Delaloye *et al.*, 2003; Sawada *et al.*, 2003; Gorbunov *et al.*, 2004; Zacharda *et al.*, 2005, 2007; Morard *et al.*, 2010; Stiegler *et al.*, 2014; Kneisel *et al.*, 2015); cold air drainage from the bottom of the talus (Delaloye *et al.*, 2003; Sawada *et al.*, 2003; Zacharda *et al.*, 2005; Morard *et al.*, 2008, 2010; Stiegler *et al.*, 2014); holes through the snow cover, some of which eject cold air (Sawada *et al.*, 2003; Morard *et al.*, 2010) and others which eject warm air or grow hoar frost (Delaloye *et al.*, 2003; Sawada *et al.*, 2003); and disjunct communities of alpine, sub-arctic and arctic species (Zacharda *et al.*, 2005; Morard *et al.*, 2008, 2010). They also meet the geomorphological preconditions for talus permafrost outlined by Gude *et al.* (2003).

Ouimet Canyon and Doghead Mountain also had standing water bodies that formed and disappeared over the course of the thawing season. Their

transience suggests that ice within the talus was preventing drainage until it melted. Such water accumulations could appear or persist in some years and not others, depending on the thickness and location of ice bodies. These phenomena may be useful indicators for discerning talus slopes that contain permafrost. If the water bodies are as large as those at Ouimet Canyon, they would be visible in sufficiently detailed satellite imagery, like the one on the right side of frame b) in Fig. 3.12. Smaller water bodies like those at Doghead could be assessed in the field.

Chapter 6

Summary and Conclusions

6.1 Summary of results

This thesis examined two talus slopes in Northwestern Ontario in order to determine their thermal regimes and to investigate the potential presence of permafrost. Analysis of the thermal regime of the talus slopes over one year revealed the types of heat transfer at work and the variables controlling talus temperatures.

6.1.1 Heat transfer mechanisms in the talus slopes

The talus slopes at both Doghead Mountain and Ouimet Canyon experienced an unstable thermal regime with convective heat transfer during the winter. Convection cells formed across the entirety of each talus slope, transporting warm air to the top of the slope and cold air to the bottom. As a result, the upper sections of the talus were relatively warm and temperatures changes were gradual. The lower sections of the slopes were cold and experienced

fluctuations in temperature related to the weather. The middle section of the Doghead Mountain slope behaved rather like the lower section, while the middle section of the Ouimet Canyon slope behaved more like the upper section.

Both talus slopes experienced a stable, stratified thermal regime with conductive heat transfer during the summer. The thermal regimes of the upper and middle sections of the talus were similar to the air temperature, while the thermal regimes of the lower sections of the slopes were characterised by zero curtains.

6.1.2 The effects of ice and water

The length of the zero curtains were the result of the amount of ice near the sensor and the rate of melting, so the amount of ice in the talus determined whether thaw would complete before freezeback began. The formation and ablation of interstitial ice was therefore an important control on both the thermal regime and the presence of permafrost in the talus. Interstitial ice was formed during warm weather events during the late winter and in the spring, when meltwater infiltrated the talus and refroze. Infiltration of rain water during the summer also affected the talus by briefly raising temperatures.

6.1.3 Permafrost

Permafrost was found in the talus slope at Ouimet Canyon. The temperatures records and interstitial ice observations from July 2016 to August 2018 showed that the talus did not thaw completely over 25 consecutive months.

6.2 Conclusions

Six conclusions were drawn from this study of talus slopes in Northwestern Ontario:

1. There is permafrost in Ouimet Canyon, some 400 km south of the regional limit of permafrost.
2. The thermal regimes at the Doghead Mountain and Ouimet Canyon talus slopes were similar to other talus slopes in the literature and to air convection embankments. Annual mean talus temperatures increased with elevation.
3. An unstable thermal regime significantly influenced by convection existed in the talus slopes at both Doghead Mountain and Ouimet Canyon during the winter. Talus temperatures increased with elevation.
4. A stable, stratified thermal regime characterized by conduction existed in the talus slopes at both the Doghead Mountain and Ouimet Canyon during the summer. Temperatures in the upper and middle sections

of the talus were similar to air temperature, while temperatures in the lower sections of the slopes were dominated by zero curtains.

5. A long zero curtain was critical to maintaining the cold thermal anomaly through the summer.
6. Precipitation events caused warming of the ground.

6.3 Suggestions for future research

Little work on permafrost and ground thermal anomalies has been conducted in this part of Ontario. This thesis focused on two talus slopes, but there are thousands of similar talus slopes in Northern Ontario. Future research should investigate how many of these slopes contain permafrost so that the extent of permafrost in the region can be understood. Candidate talus slopes in other regions, e.g. western Quebec and northern Minnesota, may be identified. Few papers frame talus permafrost in terms of horizontal distance from known permafrost. The extent of talus outside alpine areas but within a few hundred kilometres of the southern limit of permafrost in Canada should be investigated.

Cold air pools were observed at both Doghead Mountain and Ouimet Canyon. Ongoing monitoring of the sites could reveal the effect of cold air drainage on the thermal regimes of the talus slopes. The installation of additional air temperature sensors in depressions, recording the timing and

temperature of cold air drainage, could reveal the relationship of these variables to boundary layer temperature (as in Delaloye *et al.*, 2003; Gude *et al.*, 2003; Zacharda *et al.*, 2007), and whether they significantly affect ground surface temperatures. Infrared imaging would render cold pools visible and allow the volume of cold air expelled from the talus during these events to be calculated.

Interstitial ice was found at both Doghead Mountain and Ouimet Canyon. The ice at Ouimet Canyon persisted for two years. The ice at Doghead Mountain was presumed to have melted away, although it is possible that it simply melted out of sight. Very little is known about the ice in the talus. Geophysical techniques, such as electrical resistivity tomography, may reveal its size, depth, and position, providing information about degree to which the permafrost is spatially and temporally sporadic (as in Delaloye *et al.*, 2003; Gude *et al.*, 2003; Lambiel & Pieracci, 2008). Dating of organic material from within or under the ice could provide an estimate of its age, and an idea of how long conditions conducive to permafrost have existed in the talus (as in Sawada, 2006).

The thaw progression at Doghead Mountain was rather different in 2016 and 2017, but it is unclear why this was the case. Future research could examine the effect of climate variables such as air temperature, snow depth, and rainfall on aspects of the talus slope system, such as the thermal regime, the rate of thaw, the mechanism of heat transfer, or the accumulation and ablation of ice. This information could be used to predict the sensitivity of

these talus slopes to different climate change scenarios. Changes to these talus systems may have implications for the regionally unique ecosystems and rare species that are found on many talus slopes. The local hydrology may be affected by changes to the amount of water retained as ice in the talus. There also may be parallels between the effects of climate variables on talus slope systems and their effects on the crushed rock used in construction and on waste rock piles from mining.

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