

**Investigating Changes in Retrogressive Thaw Slumps in the
Richardson Mountains (Northwest Territories, Canada) based on
Tasseled Cap Trend Analysis of Landsat Image Stacks.**

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Table of Contents

List of figures	ii
List of tables	iv
Acknowledgements	vi
Abstract	vii
Résumé	vii
1. Introduction	1
1.1. General context	1
1.1. Objectives of thesis	2
2. Background	5
2.1. Permafrost	5
2.2. Mass wasting in permafrost environments	7
2.2.1. Slow Mass Wasting	8
2.2.2. Rapid Mass Wasting	8
2.2.3. Complex Mass Wasting	9
2.3. Life cycle of thaw slumps	9
2.3.1. Initiation of thaw slumps	11
2.3.2. Growth of thaw slumps	12
2.3.3. Stabilization of thaw slumps	14
3. Study area	16
3.1. Physiography	16
3.2. Regional climate	20
3.3. Glacial history of the Richardson Mountains	24
3.4. Holocene climate variation in the western Canadian Arctic	27
4. Methodology	29
4.1. Creation of the Landsat Image Stack	29
4.2. Creation of the Linear Trends map	34
4.3. Analysis of annual Tasseled Cap trends	35
5. Results	39
5.1. Identification of thaw slumps using the Tasseled Cap method	39
5.2. Distribution of thaw slumps	42
5.3. Annual Tasseled Cap trends	59

6. Discussion	82
6.1. Distribution of thaw slumps in the Richardson Mountains	82
6.2. Thaw Slump Life Cycle	89
6.2.1. Initiation of thaw slump	89
6.2.2. Growth of thaw slump	92
6.3.3 Stabilization of thaw slumps	101
7. Conclusions	102
References	103
Appendix A: Landsat scenes included in each of the three stacks.....	112
Appendix B: Individual thaw slump information.	116

List of figures

Figure 1. Map showing study area (red rectangle) in the Richardson Mountains, NWT, Canada.	4
Figure 2. Map showing permafrost zones in Canada (from Burn, 2011).	5
Figure 3. Map showing ground ice conditions of study area (modified from Heggibottom and Radburn, 1992; ice content classification from Brown et al., 1998).	7
Figure 4. Field photographs showing the various stages of a thaw slump "life-cycle" (Photos taken by Steve Kokelj and Alex Brooker).	10
Figure 5. Photograph showing the effect of debris flow on stream channel morphology and the subsequent initiation of new thaw slumps (Photo taken by Kamylle Poirier, summer 2012).	12
Figure 6. Surficial geology of the Richardson Mountains (Côté et al., 2013).	17
Figure 7. Names of watersheds along the eastern portion of the Richardson Mountains, NWT, Canada.	19
Figure 8. Mean Annual Air Temperature (MAAT), mean winter air temperature (MWAT) and mean summer air temperature (MSAT) for the three closest meteorological stations: Inuvik (NWT), Fort McPherson (NWT) and Old Crow (YT).	22
Figure 9. Number of warm days in Inuvik, NWT (1960-2011) and number of days with high-intensity precipitation events in Fort McPherson, NWT (1986-2010).	23
Figure 10. Total snowfall and rainfall for Fort McPherson, NWT, Canada (1986 to 2010). ..	23
Figure 11. Location map of study area (red rectangle) showing position of the maximum westward extent of the Laurentide Ice Sheet (last glacial maximum) and the Tutsieta recessional phase (glacial limits from Duk-Rodkin and Hughes, 1992).	26
Figure 12. Extent of the Landsat image stacks in the Richardson Mountains, NWT, Canada.	31
Figure 13. Summary of the Landsat Image Stack Trend Analysis method.	35
Figure 14. Example of pixel sampling of a thaw slump and the surrounding undisturbed tundra. Each point represents one Landsat pixel. Yearly (1985-2011) Tasseled Cap values were extracted for each point.	37
Figure 15. Tasseled Cap values (Brightness, Greenness and Wetness) of a transect of pixels from 1988 to 2011.	38
Figure 16. Comparison of different-single image Landsat views and Ikonos imagery with Tasseled Cap and Linear Tasseled Cap Trends maps.	40
Figure 17. Visualization of a stabilizing thaw slump from 1954 to 2011, using airphotos, Landsat natural colour, single year Tasseled Cap transformation and the Tasseled Cap linear trends map.	41
Figure 18. Distribution of 251 thaw slumps identified using Tasseled Cap trends between 1985 and 2011 in the Richardson Mountains, NWT, Canada. This number includes thaw slumps active before the year 1985.	43
Figure 19. Distribution of aspect, surface area and elevation of all identified thaw slumps in the Richardson Mountains, NWT, Canada.	44

Figure 20. Distribution of thaw slumps in the Big Fish River watershed	47
Figure 21. Distribution of thaw slumps in the Unnamed Watershed #1	48
Figure 22. Distribution of thaw slumps in the Willow River watershed.....	49
Figure 23. Distribution of thaw slumps in the Unnamed Watershed #2	50
Figure 24. Distribution of thaw slumps in the Rat River watershed	51
Figure 25. Distribution of thaw slumps in the Stony Creek watershed.....	52
Figure 26. Distribution of thaw slumps in the Vittrekwa River watershed	53
Figure 27. Distribution of thaw slumps in the Unnamed Watershed #3	54
Figure 28. Distribution of thaw slumps in the Road River watershed	55
Figure 29. Distribution of thaw slumps in the Unnamed Watershed #4	56
Figure 30. Distribution of thaw slumps in the Trail River watershed	57
Figure 31. Distribution of thaw slumps in the Caribou River watershed	58
Figure 32. Location of sampled slumps within the Richardson Mountains	61
Figure 33. Annual greenness trends for thaw slump #1 in the Rat River watershed (1/3). .	66
Figure 34. Annual Greenness trends for thaw slump #2 in the Rat River watershed (2/3)..	67
Figure 35. Annual Greenness trends for thaw slump #3 in the Rat River watershed (3/3)..	68
Figure 36. Annual Greenness trends for #4 "Charas" thaw slump in the Stony Creek watershed (1/3).....	69
Figure 37. Annual Greenness trends for #5 "Melanie" thaw slump in the Stony Creek watershed (2/3).....	70
Figure 38. Annual Greenness trends for thaw slump #6 in the Stony Creek watershed (3/3).	71
Figure 39. Annual Greenness trends for thaw slump # 7in the Vittrekwa River watershed (1/4).	72
Figure 40. Annual Greenness trends for thaw slump #8 "Dover" in the Vittrekwa River watershed (2/4).....	73
Figure 41. Annual Greenness trends for thaw slump #9 in the Vittrekwa River watershed (3/4).	74
Figure 42. Annual Greenness trends for thaw slump #10 in the Vittrekwa River watershed (4/4).	75
Figure 43. Annual Greenness trends for thaw slump #11 in the Road River watershed (1/3).	76
Figure 44. Annual Greenness trends for thaw slump #12 in the Road River watershed (2/3).	77
Figure 45. Annual Greenness trends for thaw slump #13 in the Road River watershed (3/3).	78
Figure 46. Annual Greenness trends for thaw slump #14 in the Trail River watershed (1/3).	79
Figure 47. Annual Greenness trends for thaw slump #15 in the Trail River watershed (2/3).	80
Figure 48. Annual Greenness trends for thaw slump #16 in the Trail River watershed (3/3).	81

Figure 49. Thaw slump distribution in relation to the limits of the Last Glacial Maximum and the Tuetsia phase, as well as the mapped ice content of the area (Hegginbottom and Radburn, 1992).....	86
Figure 50. Thaw slump distribution in relation to the regional surficial geology (Côté et al., 2013).....	87
Figure 51. Topographic cross-sections along Stony Creek, Stony Creek watershed. Cross-sections A and B show canyon-type valleys, while cross-sections C,D, E and F demonstrate valley asymmetry.....	88
Figure 52. Greenness values of thaw slump #11 that initiated following a forest fire in the area.....	91
Figure 53. Annual retreat rate of an unnamed thaw slump (thaw slump #6) in the Stony Creek watershed, as derived from the annual greenness trends.	94
Figure 54. Annual headwall retreat rate of the #5 "Melanie" thaw slump in the Stony Creek watershed, as derived from the annual greenness trends.....	95
Figure 55. Annual headwall retreat rate of the #4 "Chara" thaw slump in the Stony Creek watershed, as derived from the annual greenness trends.....	96
Figure 56. Annual growth rate of headwall for thaw slumps in a given watershed; as calculated from the greenness annual trends.....	98
Figure 57. Average annual headwall rate of thaw slumps in the Richardson Mountains, NWT, compared with mean summer air temperature, thaw degree-days and number of days with rainfall >10mm (climate data from the Fort McPherson meteorological station). .	99

List of tables

Table 1. Name and surface area of the watersheds in the Richardson Mountains, NWT...	18
Table 2. Slope value and statistical significance (p-value) of linear regression analysis of various climate parameters for Inuvik, NWT. Slope (p-value), bold values are statistically significant (p-value ≤ 0.05).....	24
Table 3. Synthesis of the regional Holocene climatic history reported by various studies. The radiocarbon age limits provide only rough dates for environmental change as these changes were gradual. (modified from Lauriol et al., 2002).....	28
Table 4. Spectral bands of the Landsat Thematic Mapper (TM) sensor on the Landsat 5 satellite (USGS, 2012).	30
Table 5. Spectral bands of the Landsat Enhanced Thematic Mapper Plus (ETM+) sensor on the Landsat 7 satellite (USGS, 2012).	30
Table 6. Equation used for each Tasseled Cap transformation and NDVI (Crist and Ciccone, 1984) (Weier and Herring, 2013).....	33
Table 7. Thematic Mapper Tasseled Cap Coefficients (Crist and Ciccone, 1984)	33
Table 8. Thaw slump distribution in the various watersheds in the Richardson Mountains, NWT, Canada.	42

Table 9. Slump ID, watershed name, surface area, orientation and figure number for all sampled thaw slumps.	60
Table 10. Distribution of permafrost by previously mapped ground ice content (Hegginbottom and Radburn, 1992; ground ice contents from Brown et al., 1998)	84
Table 11. Distribution of thaw slumps by surficial geology (adapted from Côté et al., 2013).	84

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Abstract

This thesis applies a novel method of change detection, the Landsat Image Stack Trend Analysis method to the monitoring of retrogressive thaw slumps in the Richardson Mountains, NWT. This method represents a significant improvement upon previous methods of thaw slump monitoring, which utilized air photos and high-resolution satellite imagery. This method applies Tasseled Cap brightness, wetness and greenness indices to Landsat TM/ETM images acquired between 1985 and 2011 and analyzes the temporal change of each pixel for the different indices values. This method is useful in retrogressive thaw slump monitoring in two ways. First, by creating a map showing the linear change over time from 1985 to 2011, retrogressive thaw slumps can be easily identified, as they are more dynamic than the surrounding tundra. In total, 251 thaw slumps were identified within an area of roughly 18 000km². Second, thaw slump activity, from initiation, growth and stabilization can be studied by plotting the annual vegetation index pixel values of adjacent pixels in a thaw slump. This method allows for the efficient extraction of annual thaw slump headwall retreat rates, provided the availability of cloud-free imagery. The retreat rates of 16 slumps were extracted, which were found to have an average annual retreat rate of 11.8 m yr⁻¹.

Résumé

Cette thèse applique une nouvelle méthode de détection de changement, l'analyse de tendances des piles d'image Landsat pour la surveillance des glissements de dégel régressif dans les monts Richardson, TNO. Cette méthode représente une amélioration significative sur les méthodes précédentes de suivi de crise dégel, qui a utilisé des photos aériennes et des images satellitaires haute résolution. Cette méthode applique les transformations Tasseled Cap (Brightness, Greenness et Wetness) aux images Landsat TM / ETM acquises entre 1985 et 2011 et analyse l'évolution temporelle de chaque pixel pour les différentes valeurs des indices. Cette méthode est utile pour la surveillance des glissement de dégel dans deux façons. Premièrement, en créant une carte montrant l'évolution linéaire dans le temps de 1985 à 2011, les glissements de dégel rétrogrades peuvent être facilement identifiés, car ils sont plus dynamiques que la toundra environnante. Au total, 251 glissements de dégel ont été identifiés dans une zone d'environ 18 000km². Deuxièmement, l'activité des glissements de dégels, de l'initiation, la croissance et la stabilisation peuvent être étudiés avec l'extraction des valeurs annuelles des indices de végétations des pixels adjacents dans un glissement de dégel. Cette méthode permet l'extraction efficace de la vitesse de croissance annuelle des glissements, à condition que la disponibilité d'images sans nuages. La vitesse de 16 glissement était extrait, qui ont été trouvés à avoir un taux annuel moyen de retraite de 11,8 m an⁻¹.

1. Introduction

1.1. General context

The Earth's mean annual air temperature has increased 0.7°C between 1906 and 2005, with the rate of temperature increase nearly doubling since 1950 (IPCC, 2007). This warming has not been uniformly distributed across the planet; with Arctic temperatures increasing at a rate nearly double that of the global average (IPCC, 2007). This warming has substantially impacted the cryosphere, with decreases in the extent of frozen ground, snow cover, sea ice and the mass of glaciers (IPCC, 2007).

The thermal regime of permafrost in the Arctic has already responded to increasing air temperatures with a northward migration of the southern boundary of discontinuous permafrost (ACIA, 2004; Osterkamp and Romanovsky, 1999), as well as a warming of ground temperatures, with the rate of increasing ground temperature being more rapid in the continuous permafrost zone (Osterkamp and Romanovsky, 1999; Smith et al., 2010; Lewkowicz et al., 2012). Warmer permafrost will most likely increase the rate of permafrost degradation, also referred to as thermokarst (Jorgenson et al., 2006). Thermokarst relates to the thaw of ice-bearing permafrost (Murton, 2001). Thermokarst features include land subsidence following melting of ground ice in permafrost (Schuur et al., 2008), the degradation of ice wedges and pingos (Jorgenson et al., 2006; 2008), the formation of thaw slumps and thaw lakes (Jorgenson et al., 2008), as well as coastal erosion along shorelines (Jones et al., 2009), all of which have been increasing in prevalence in the western Arctic.

Retrogressive thaw slumps are one of the largest thermokarst features found in permafrost environments, as they can reach tens of hectares in surface area. In response to a changing Arctic climate, thaw slumps are growing both in number and surface area (Lantz and Kokelj, 2008; Lantuit and Pollard, 2008; Lacelle et al., 2010; Lantuit et al., 2012). Recent studies have demonstrated the potential ecological impacts of thaw slumps as they release previously frozen materials into streams and

lakes, drastically affecting sediment load and water chemistry (Kokelj et al., 2009; 2013; Mesquita et al., 2010; Malone et al., 2013). Given the ecological impact of thaw slumps at the watershed scale, a better knowledge of their distribution across the landscape, as well as the factors affecting their initiation, growth and stabilization is required.

One approach that can be used to monitor changes in Arctic regions is remote sensing, which includes both aerial photographs and satellite images (Kääb, 2008). In Arctic regions, the temporal resolution of air photographs is very coarse (acquisition every 20-30 years), which limits its use in effectively monitoring temporal changes of landforms showing continuous changes. For example, thaw slumps in the western Canadian Arctic (Hershel Island, Richardson Mountains and Mackenzie Delta) have been previously mapped by Lantuit and Pollard (2008), Lacelle et al. (2010) and Lantz and Kokelj (2008) using airphotos acquired in the 1950s, 1970s and 2000s. Conversely, satellite images have a temporal resolution of a few days (i.e., acquisition interval for Landsat images is 16 days, provided there is sufficient daylight for image capture), which makes them better suited to monitor landscape changes, especially large features like thaw slumps.

1.1. Objectives of thesis

This thesis will use a novel remote sensing approach to study the 1985-2011 distribution and activity of hillslope thaw slumps in the Richardson Mountains, Northwest Territories, Canada (Figure 1). The approach consists of a change detection method developed by Fraser et al. (2011; 2012a): the Landsat Image Stack Trend Analysis. Previous landscape change detection in the Arctic utilized images from the 25-year Advanced Very High Resolution Radiometer (AVHRR) archive, which provided continental-scale coverage at a coarse spatial resolution (Fraser et al., 2011). The Landsat Image Stack Trend Analysis method was developed to detect vegetation change across landscapes using a series of Landsat 5 and 7 image stacks. This change detection method applies the three Tasseled

Cap transformation indices (brightness, greenness and wetness) to a temporally dense stack of Landsat images.

The objective of this thesis is to investigate the distribution and spatial changes of retrogressive thaw slumps in the Richardson Mountains based on long-term (1985-2011) analysis of Landsat image stacks by applying the Landsat Image Stack Trend Analysis method. This objective will be reached by: i) evaluating if the method can be employed to detect thaw slumps; and ii) subsequently applying the method to monitor the life-cycle of thaw slumps (from initiation, growth and stabilization). The results will be compared with previous approaches, which utilized higher spatial resolution imagery. Overall, this thesis demonstrates the usefulness of the Landsat Image Stack Trend Analysis to monitor thaw slumps across landscapes. Since the method can be automated, this approach could be useful in monitoring other large scale landscape changes, such as coastal erosion and desertification.

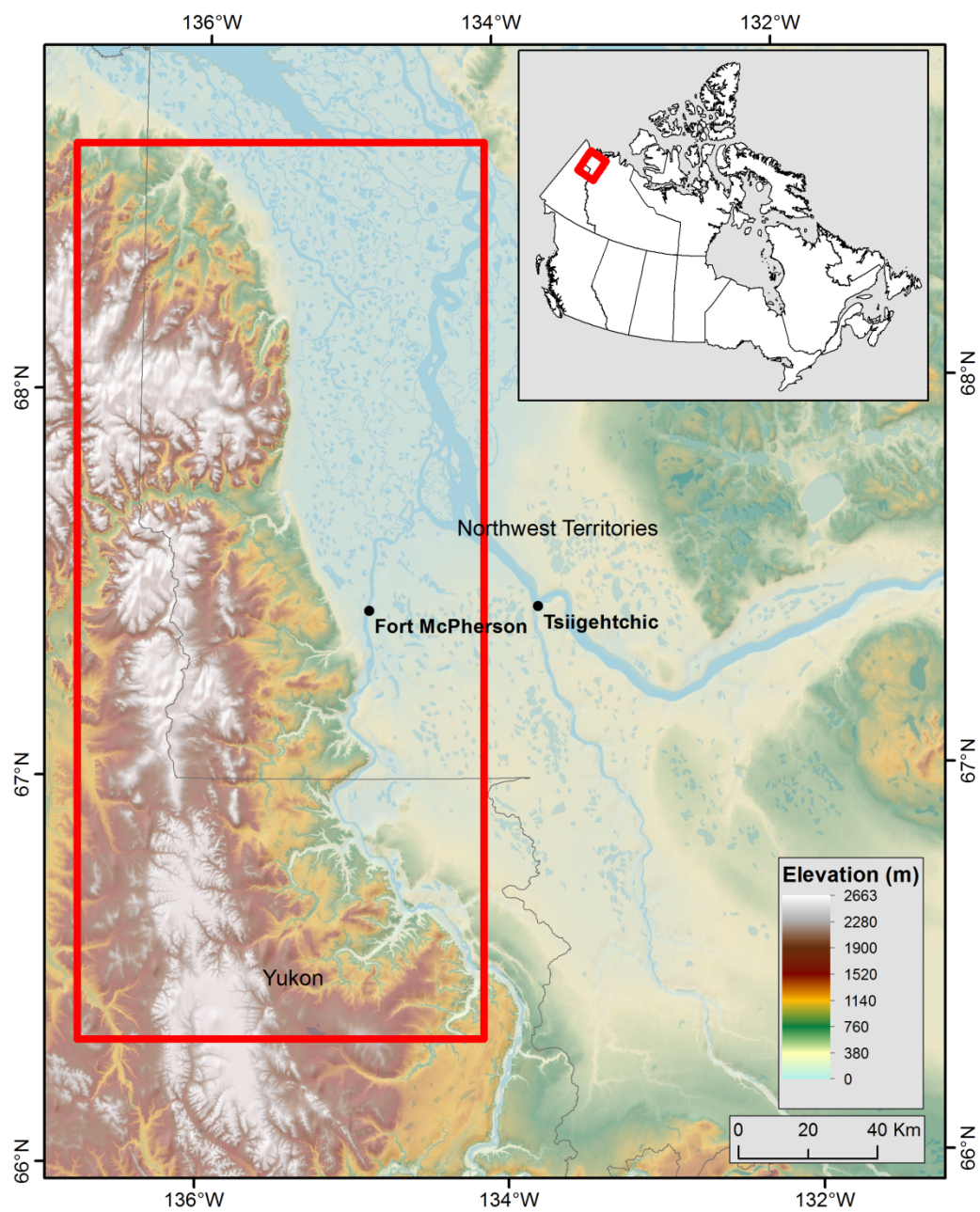


Figure 1. Map showing study area (red rectangle) in the Richardson Mountains, NWT, Canada.

2. Background

2.1. Permafrost

Permafrost is defined as ground that remains at or below 0°C for two or more consecutive years (van Everdingen, 1998). At a regional scale, this ground thermal condition is generally present when the mean annual air temperature (MAAT) is at or below -2°C (French, 2007). Permafrost can also be present as relict permafrost in environments that are no longer suitable for the formation of permafrost (e.g. MAAT > -2°C).

Permafrost regions occupy approximately a quarter of the Earth's terrestrial landscape and its distribution is broadly categorized into zones based on the percentage of the landscape it occupies (Figure 2). The continuous permafrost zone is the area in which permafrost underlies more than 90% of the landscape; the discontinuous permafrost zone is the area in which permafrost underlies between 10 - 90% of the landscape; and isolated permafrost, is found in areas underlain by less than 10% permafrost (Burn, 2011). The boundary between continuous and discontinuous permafrost corresponds roughly with the -8°C MAAT isotherm (Burn, 2011).

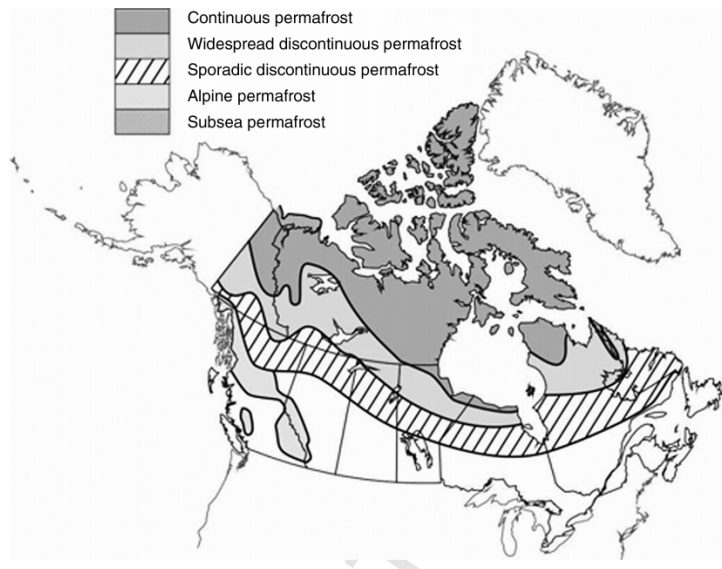


Figure 2. Map showing permafrost zones in Canada (from Burn, 2011).

Permafrost has variable ground ice content that may take various forms. Pore ice is ice found between pore space of sediments and acts to “cement” the grains together. Segregated ice forms when soil freezes and water is drawn to the freezing front to form ice lenses (French, 2007). The movement of water towards the freezing front is called cryosuction and is most commonly found in fine-grained soils where this process is easiest to sustain (Cheng, 1983; French, 2007). Intrusive ice forms when water is forced into the seasonally or perennially frozen zone and it commonly leads to the formation of ice laccolith (French, 2007). Vein ice is formed when water infiltrates an open fissure at the surface and freezes in the permafrost. Ice wedges, are an example of vein ice (French, 2007). Massive ice is a type of ground ice that is characterized by very high ice contents; gravimetric water content greater than 250% (van Everdingen, 1998). Broadly speaking, massive ice can be divided into two main types: buried ice and intrasedimental ice (Mackay and Dallimore, 1992). Buried ice is formed when a body of surface ice is buried by sediment, if the insulating sediment layer is thicker than the active layer in the area. The most common forms of buried surface ice are glacier ice or snowbank (i.e., Astakhov, 1986; St-Onge and McMartin, 1995; Lacelle et al., 2009). The other form of massive ice is intrasedimental ice, which forms *in situ* by freezing of groundwater (Rampton, 1988; Lacelle et al., 2004). The advance of an ice sheet causes pressure melting of shallow permafrost. Following deglaciation, this water re-freezes and is further supplemented by subglacial water drawn to the margins of the ice sheet (Rampton, 1988; French and Harry, 1990; Lacelle et al., 2004). Intrasedimental massive ice is then formed through the freezing processes of ice intrusion and ice segregation (Mackay and Dallimore, 1992). All types of ground ice, with the exception of pore ice, may produce thermokarst landscapes upon thawing of permafrost (French, 2007). In the study area, the ground ice content ranges from nil to moderate (Figure 3).

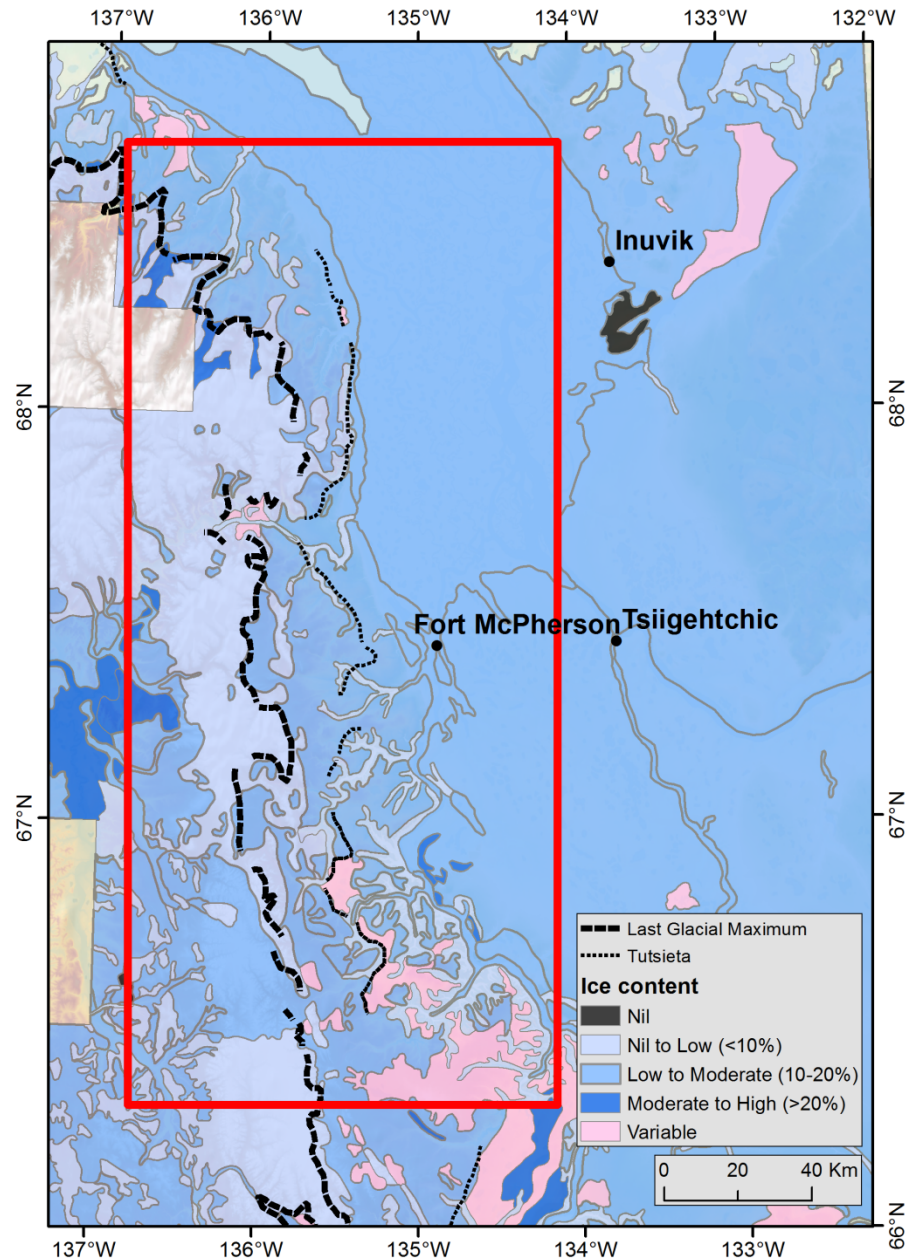


Figure 3. Map showing ground ice conditions of study area (modified from Heggibottom and Radburn, 1992; ice content classification from Brown et al., 1998).

2.2. Mass wasting in permafrost environments

Mass-wasting is defined as “the movement of soil downslope by gravity without the aid of a transporting medium.” (Tarbuck & Lutgens, 2005). Despite the cold climate conditions and presence of permafrost, mass-wasting processes in

periglacial regions may be highly effective in transporting material and can be broadly categorised based on the speed of debris movement downslope (French, 2007).

2.2.1. Slow Mass Wasting

The most common form of slow mass-wasting in permafrost environments is solifluction. This term refers to the downslope movement of water-saturated soils resulting from the combined effect of frost creep and gelifluction mechanisms (Harris, 1981; Smith, 1992; French, 2007). Frost-creep refers to the downslope movement of soil particles as a result of freeze-thaw processes as the soil expands and resettles (Benedict, 1970; Harris et al., 1993; French, 2007), while gelifluction is the slow, viscous flow of a water-saturated soil layer over a frozen ground, caused primarily by the seasonal melt of ice lenses (Benedict, 1970; Harris et al., 1993; French, 2007).

2.2.2. Rapid Mass Wasting

There are several rapid mass wasting processes that are unique to permafrost regions. One such process is an active layer detachment slide, which is a small-scale slope failure that is limited to the active layer. These slides occur when the shear strength of the soil is decreased following a gradual increase in basal active layer ice content and occur in mid to late-summer, when active layer depth is deepest (Lewkowicz, 2007; Harris and Lewkowicz, 2000; Lewkowicz and Harris, 2005).

Debris flows, which are defined as rapid movement of material downslope due to gravity, are also important processes in permafrost environment to varying degrees. This process is common in permafrost environments as the frozen ground restricts drainage and leads to a saturated sediment layer above the permafrost table (Huscroft et al., 2003).

Finally, rockfalls are another process of rapid mass wasting that is common in permafrost regions. Rockfalls occur following a reduction of shear strength of the

material from the melt of ice from within rock joints (Matsuoka and Sakai, 1999), as well as from the expansion of frost wedges during the freezing of water (Matsuoka, 2001; French, 2007).

2.2.3. Complex Mass Wasting

Retrogressive thaw slumps are found only in ice-rich permafrost environments and represent a unique mass wasting process (Burn and Lewkowicz, 1990). Retrogressive thaw slumps move material downslope when the exposure of a body of massive ice in the headwall causes it to begin to ablate, which results in the overlying sediment and vegetation to fall to the slump floor. This sediment and vegetation may then be transported downslope and away from the slump floor via the thaw slump's debris flow channels during periods of high rain events (i.e, Lacelle et al., 2010). Retrogressive thaw slumps move material downslope incrementally, through the recession of the ice-rich headwall, as well as rapidly, when debris is moved rapidly away from the headwall from high-intensity precipitation events (Lacelle et al., 2010). The melt of the ice-rich headwall creates a muddy mixture that is mostly liquid and easily flows downslope (Wang et al., 2005). The flow of this muddy mixture is slowed by the addition of material from the overlying ice-poor active-layer, which causes debris to build-up at the base of the headwall. However, as noted by Wang (2011), as the debris builds-up, its weight increases, increasing the likelihood of the debris flowing under its own weight.

2.3. Life cycle of thaw slumps

Retrogressive thaw slumps, which are among the most impressive thermokarst features, can be recognized in the landscape by horseshoe-shaped depressions (Mackay, 1966; Burn and Lewkowicz, 1990; Wolfe et al., 2001). Thaw slumps have been observed along the shorelines of lakes, rivers, along the coastlines of seas and oceans as well as along hillslopes on plateaus. The morphology of thaw slumps includes a near-vertical headwall and a slump floor of

low gradient ($2\text{--}10^\circ$). The slump floor is commonly wet and muddy as a result of thawing of the ice-rich headwall and contains little to no active vegetation. In plateau areas, the slump floor is connected to the lower fluvial system via a steeply sloping debris-flow channel ($15\text{--}25^\circ$) (Lacelle et al., 2010). Unlike landslides and slumps in non-permafrost soils or low ice content soils, thaw slumps remain active for many years until the headwall is covered by accumulating sediments. The following section describes the life-cycle of a thaw slump, from initiation, growth and stabilization and Figure 4 shows field photographs of the various stages of the life-cycle.

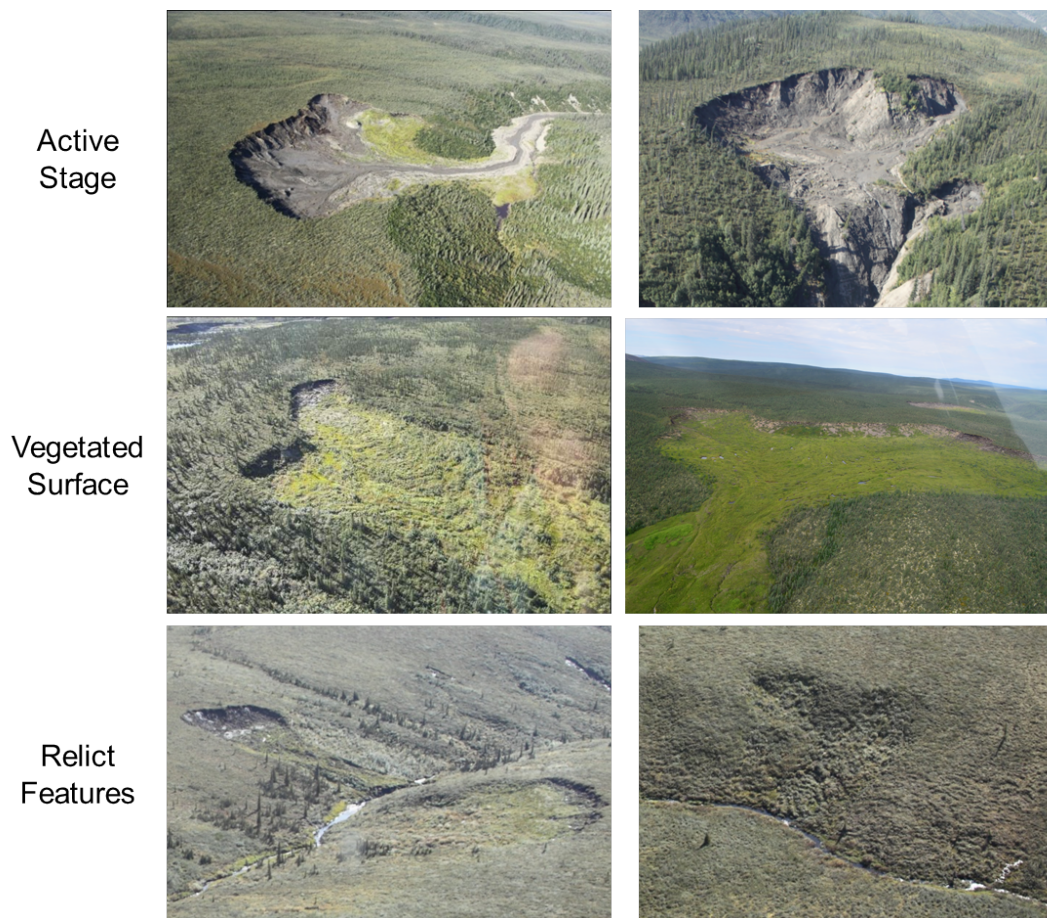


Figure 4. Field photographs showing the various stages of a thaw slump "life-cycle" (Photos taken by Steve Kokelj and Alex Brooker).

2.3.1. Initiation of thaw slumps

Retrogressive thaw slumps may initiate from any process that exposes ice-rich permafrost. One process of massive ice exposure that has been proposed is active layer detachments (Aylsworth et al., 2000). Active layer detachments are shallow landslides that occur following the melting of ice lenses at the base of the active layer, resulting in high-pore water pressure and low shear strength (Lewkowicz, 2007). Typically, these detachments occur when the active layer thickens to a deeper than normal level and if this active layer thickening occurs in regions where the top of permafrost is ice-rich, an active layer detachment may occur (Aylsworth et al., 2000). Retrogressive thaw slumps can result from these events if the detachment exposes underlying massive ice or icy sediment (Aylsworth et al., 2000).

Thaw slumps may also form through the melt of ice wedges (de Krom, 1990). Thaw slumps are thought to initiate in this manner when massive ice is exposed following the catastrophic drainage of pond water on the top of ice wedge troughs (Lantuit and Pollard, 2008). In coastal areas, wave erosion of the shoreline has been advanced as a mechanism of thaw slump initiation (Lantuit and Pollard, 2008).

The mechanism of thaw slump initiation in highland areas, such as those found in the Richardson Mountains, is much less understood (Lacelle et al., 2010). However, it has been proposed that initiation is related to precipitation events, as precipitation might cause active-layer detachments, or other mass-wasting events (Lacelle et al., 2010). One mechanism through which hillslope thaw slumps might initiate is through the damming of a stream by the debris flow of a thaw slump. This damming might result in a shift of a stream channel and rise of stream water levels, which can result in the undercutting of the surrounding hillslope and the potential to expose underlying massive ice (Figure 5).



Figure 5. Photograph showing the effect of debris flow on stream channel morphology and the subsequent initiation of new thaw slumps (Photo taken by Kamyille Poirier, summer 2012).

2.3.2. Growth of thaw slumps

Once initiated, thaw slumps may enlarge by several meters per year across ice-rich permafrost terrain and remain active as long as the tabular ground ice body in the headwall is exposed (i.e., Lantz et al., 2008). At the scale of individual slumps, it has been found that the rate of headwall retreat is affected by: i) amount of ground ice exposed in the headwall; ii) height of the headwall; and iii) net radiation received on the exposed headwall. The amount of ground ice exposed in the headwall influences the rate of headwall retreat (i.e., backwasting) and higher ground ice contents tend to be associated with higher backwasting rates (French, 1974). Thaw slumps with headwalls up to ca. 6-8 metres displayed a strong correlation between

growth rate and headwall height, but those with higher headwall do not seem to grow any quicker (Wang et al., 2009). Finally, the rates of headwall retreat are largely affected by the net short-wave radiation received at the ground ice surface, which are affected by cloud cover and the albedo of the ablating ice (Lewkowicz, 1986). Thaw slumps along shorelines of lakes are most commonly found with orientation facing south and west, as these orientations receive greater amounts of solar radiation (Kokelj et al., 2009).

According to Lewkowicz (1986), the melt rate of ground ice in the headwall of thaw slumps can be summarized as: $Q_m = Q^* + Q_h + Q_e + Q_g + Q_p + Q_d + Q_w$ where Q_m is the energy flux available for ablation; Q^* is the net radiation flux; Q_h is the sensible heat flux; Q_e is the latent heat flux; Q_g is the ground-heat flux; Q_p is the heat flux from precipitation; Q_d is the heat flux from debris falling from ice; and Q_w is the heat flux from water flow down the ice. The relative importance of these variables varies given different geographical locations and meteorological conditions. With sunny skies, the importance of Q^* (net radiation input) is increased, compared with cloudy skies, where its relative and absolute importance are diminished (Lewkowicz, 1986).

At a regional scale, thaw slump growth rates are controlled by summer air temperature and rainfall, and grow fastest on days that are warm, cloud-free and windy (Burn and Lewkowicz, 1990). The amount of rainfall received appears to play a much greater role in the growth rates of thaw slumps located in highland areas than those located along shorelines. The reason for this relation is believed to be due to the role that precipitation events play in the movement of debris downslope away from the exposed headwall (Lacelle et al., 2010). On the other hand, thaw slumps located along shorelines flow directly into a water body and are not as dependent on precipitation to clear debris away from the headwall. For these slumps, summer air temperature is the driving factor in thaw slump growth rate (Lantuit & Pollard, 2008; Lantz et al., 2008). In coastal areas, slumped debris is

removed through storm activity, which typically occurs at least once or twice a year (Burn and Lewkowicz, 1990).

Thaw slump growth rates on a regional scale have been studied in the Mackenzie Delta region (Lantz and Kokelj, 2008). Based on air photo analysis from 1950, 1973 and 2004, it was found that thaw slump growth rates between the years 1973-2004 was significantly higher than between 1950-1973. The increase in growth rate was attributed to a 1.3°C increase in mean summer air temperature and a 9.9 day increase in days with temperatures over 20°C at Inuvik between 1950-1973 and 1973-2004 (Lantz and Kokelj, 2008).

2.3.3. Stabilization of thaw slumps

Thaw slumps will continue to grow as long as the ice-rich sediments in the headwall remains exposed or upon exhaustion of ground ice supply (Burn, 2000). Stabilization often occurs when the rate of slumped debris accumulation at the base of the headwall is greater than the rate of debris removed (French, 1974). Slump stabilization is a gradual process, with different sections of the headwall stabilizing at different times (Burn, 2000). Slumps can be reactivated when the debris covering the headwall is removed, such as through rising water levels or wave erosion from storms (Aylsworth et al., 2000).

Once a thaw slump stabilizes, its effects on the ground thermal regime remain, as the thaw slump has removed the insulating organic layer and warmed the upper layers of the permafrost (Burn and Lewkowicz, 1990; Kokelj et al., 2009). Thaw slumps have been found to increase the mean near-surface ground temperature by roughly 3° to 4°C and the full effect of the disturbance created by thaw slumping is not completed until the permafrost depth nears the maximum active-layer thickness (Burn, 2000; Kokelj et al., 2009). In addition, once the active layer reaches its maximum depth, the slumped area will continue to be warmer than the surrounding area during the winter, due to the release of latent heat from the freezing of the wetter soils (Burn, 2000).

Permafrost aggradation in the slump floor only begins once the vegetation community begins to re-establish itself (Burn, 2000). The vegetation succession of the slump floor following the stabilization of thaw slumps has been studied by Bartleman et al. (2001). The slump floor will eventually re-vegetate and match the vegetation of the undisturbed land around it (Bartleman et al., 2001). However, active and stable thaw slumps contain many plant species that are rare or are undocumented in the adjacent undisturbed terrain (Lantz et al., 2009). The two greatest factors in vegetation succession of thaw slumps are the amount of time the area has been stabilized, and the relative ease of water drainage of that area. Areas with poor water drainage displayed slower succession rates (Bartleman et al., 2001).

3. Study area

3.1. Physiography

This study investigates the distribution and growth of thaw slumps in the Richardson Mountains, NWT, Canada (Figure 1). These mountains are situated west of the Mackenzie Delta and run north-south from the Beaufort Sea to the lower Peel River, with peak heights reaching 1600m. The Richardson Mountains, which measure ca. 300km long and nearly 50km wide, separate the unglaciated Yukon lowlands to the west from the glaciated Mackenzie Delta region to the east. A series of plateaus are situated at the foothills of the Richardson Mountains, these are the Aklavik and Peel plateaus. These plateaus are characterized by relatively smooth terrain, with low relief (<100m) and elevations ranging from 650 m a.s.l. in the west to 100-150m a.s.l. in the east (Catto, 1996).

The geology of the Richardson Mountains is composed of lower Cretaceous shale and siltstone; Permian clastic and carbonate rocks; and Cambrian to Silurian limestone (Norris, 1975). A distinct characteristic of the study area is its non-glacial mountainous relief (> 750 m) including cryoplanation terraces (Lamirande et al. 1999) adjoining a glaciated lowland (< 700 m) (Duk-Rodkin and Hughes 1992). Exposed bedrock and thin colluvial deposits dominates the high elevation unglaciated areas in the Richardson Mountains (Duk-Rodkin and Hughes, 1992). The glaciated region is dominated by till veneer, ranging from broad morainal plains less than 20m thick to hummocky and rolling moraines up to 50m thick (Duk-Rodkin and Hughes, 1992). The surficial geology of the Richardson Mountains is shown in Figure 6.

The fluvial system is characterized by narrow V-shaped and canyon-type valleys that have incised the landscape following the late Pleistocene deglaciation. The valleys are occupied by a mixture of colluvial and alluvial deposits. The study area contains 12 watersheds (Figure 7). There is quite a large range in catchment size among watersheds in the Richardson Mountains, with the smallest being an

unnamed watershed (#2) with an area of 99.36 km² and the largest being Big Fish River with an area of 2935.54 km² (Table 1).

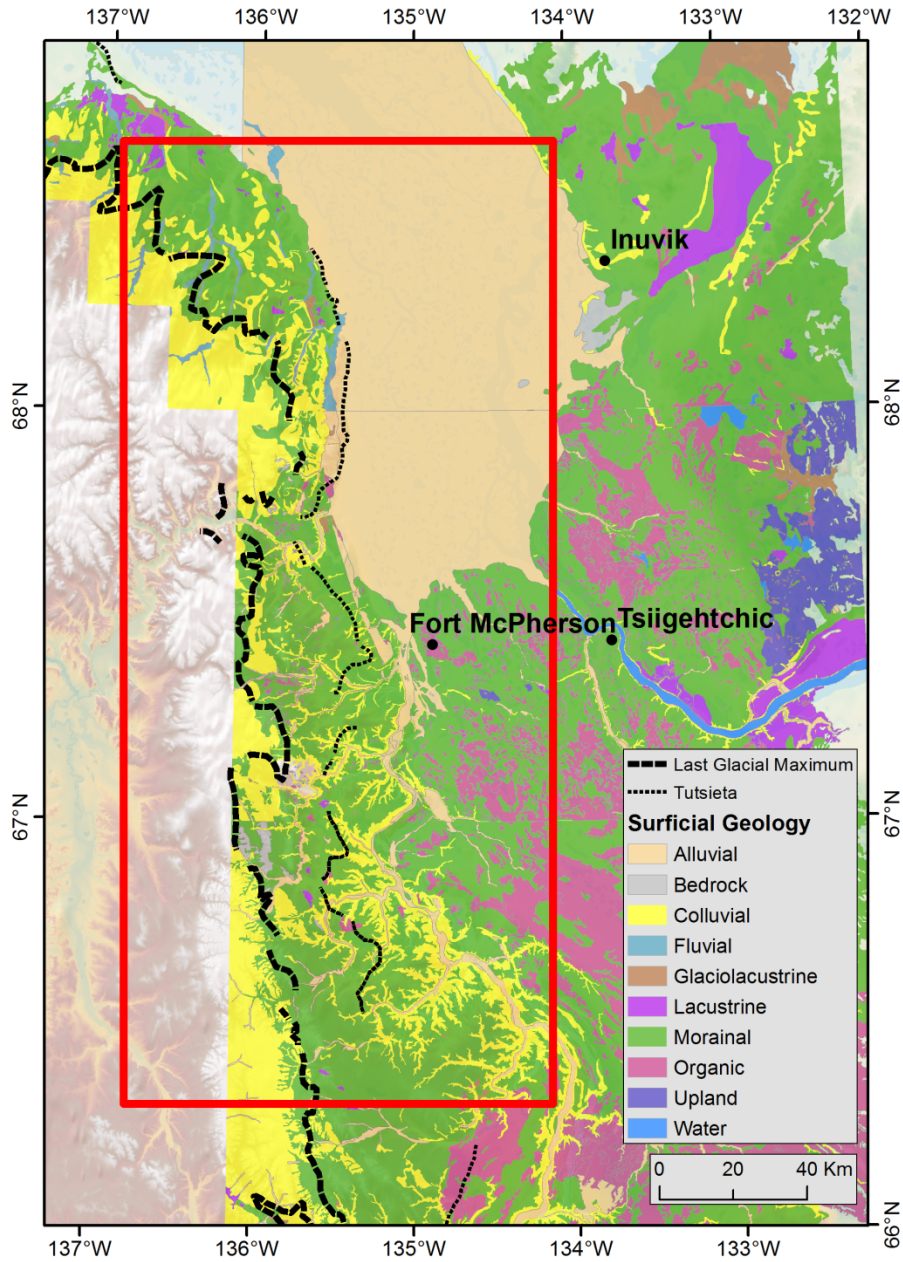


Figure 6. Surficial geology of the Richardson Mountains (Côté et al., 2013).

Table 1. Name and surface area of the watersheds in the Richardson Mountains, NWT.

Watershed name	Watershed Area (km²)
Big Fish River	2935.54
Unnamed watershed #1	404.8
Willow River	808.77
Unnamed watershed #2	99.36
Rat River	2692.94
Stony Creek	1224.46
Unnamed watershed #3	314.86
Vittrekwa River	1979.11
Road River	1545.54
Unnamed watershed #4	155.18
Trail River	1281.29
Caribou River	2227.77

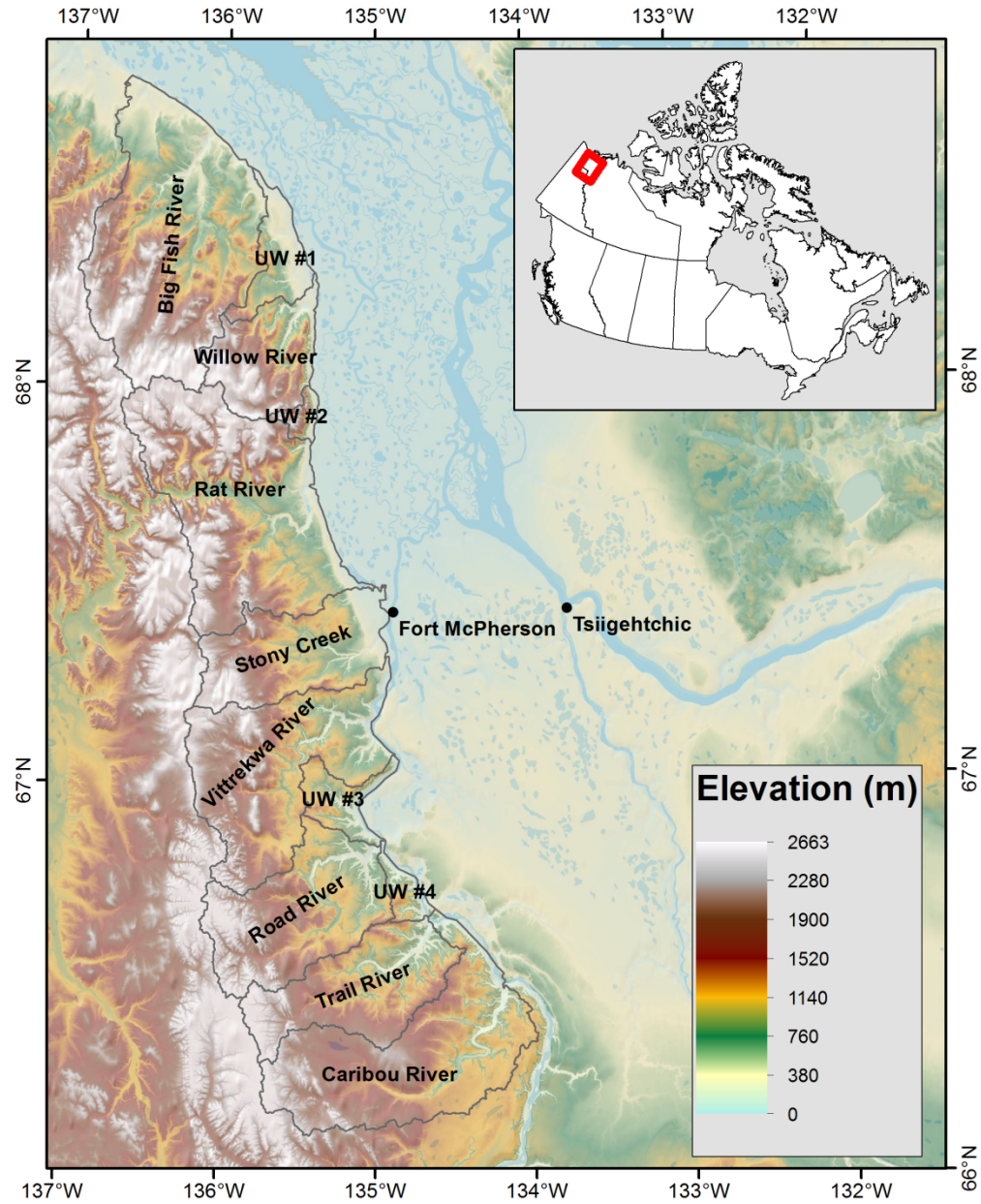


Figure 7. Names of watersheds along the eastern portion of the Richardson Mountains, NWT, Canada.

3.2. Regional climate

The climate type in the study area is classified as a polar continental. The closest communities with long-term meteorological stations are Fort McPherson, NWT, Inuvik, NWT (110 km north-east of Fort McPherson) and Old Crow, YT (210 km west of Fort McPherson, on the western side of the Richardson Mountains). Based on the climate records from these stations, the mean annual air temperature is -8.3°C (mean air temperature in January: -26.4°C ; mean air temperature in July: 15.1°C ; Environment Canada, 2012) and the total annual precipitations approaches 230 mm, with nearly half falling as rain (Environment Canada, 2012).

Figure 8 displays the mean annual air temperature (MAAT), mean summer air temperature between the months of June, July and August (MSAT) and the mean winter air temperature between the months of December, January and February (MWAT) for the three meteorological stations. The meteorological station closest to the study area is Fort McPherson where climate records are available from 1986-2011; while the climate record for Inuvik covers the 1960-2012 period. For this reason, figure 8 displays two plots for each MAAT, MSAT and MWAT with one showing air temperature between 1960-2010 and the other between 1986-2012. Table 2 shows the slope of the linear regression of various climate parameters and their statistical significance.

At the Inuvik station, the MAAT, MSAT and MWAT have increased at a statistically significant rate between 1960 and 2012, with the highest rate of temperature increase being observed in the winter (Table 2). In addition, the number of days with temperatures greater than 10°C and 20°C has slightly increased at a statistically significant rate (Figure 9). Conversely, while climate trends observed over the last 26 years (between 1986 and 2012) are not statistically significant, MSAT has shown a decreasing temperature trend, resulting in a slowing of the rate of increase in MAAT.

At the Fort McPherson station, the MAAT, MSAT and MWAT between 1986 and 2010 are showing similar trends to those observed at Inuvik. The total annual precipitation in Fort McPherson amounts to ca. 287mm/year, with nearly half falling as snow (Figure 10). Annual precipitation is highly variable, with a maximum of 407mm in 1987 and a minimum of 133mm in 2008. As shown in Table 2, precipitation trends measured at Fort McPherson are not statistically significant, except for snowfall, which has shown a strong, statistically significant decrease since 1986. However, since 1986, high-intensity daily rainfall events have been increasing, with the trend of daily rainfall events >25mm being statistically significant (Figure 9).

These climate conditions ensure that the area is underlain by continuous permafrost. The thickness of permafrost reaches almost 500 m thick along the Beaufort Sea and decreases southward, measuring 100 m in the Richardson Mountains (Smith et al., 2010). Permafrost temperature measurements from the nearby Mackenzie Valley region indicate that the mean annual ground temperature varies between -1°C and -5°C (Smith et al., 2010). Although situated in the discontinuous permafrost zone south of the study area, permafrost temperature measured at the Norman Wells station since 1945 showed an increase in MAGT from -8°C to -3°C (Smith et al., 2010).

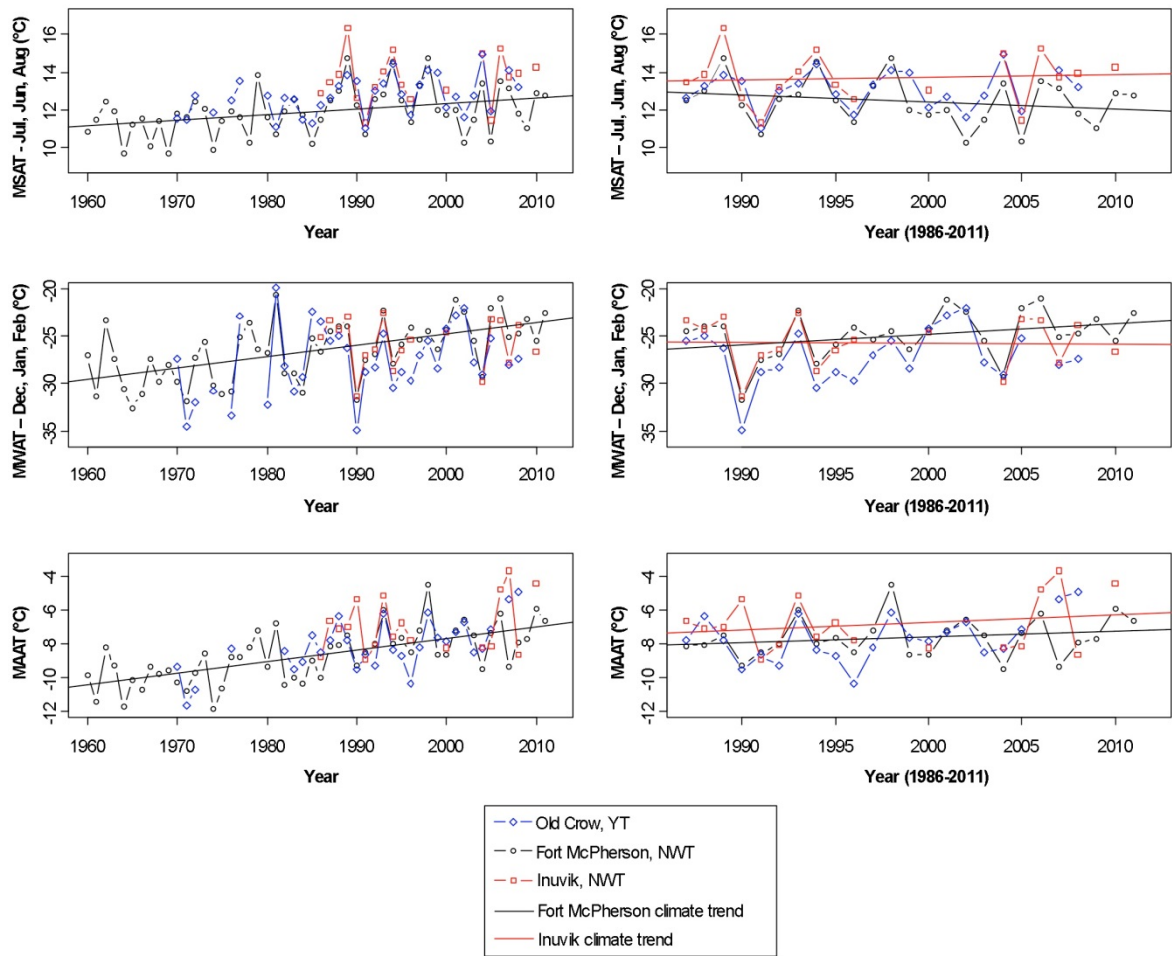


Figure 8. Mean Annual Air Temperature (MAAT), mean winter air temperature (MWAT) and mean summer air temperature (MSAT) for the three closest meteorological stations: Inuvik (NWT), Fort McPherson (NWT) and Old Crow (YT).

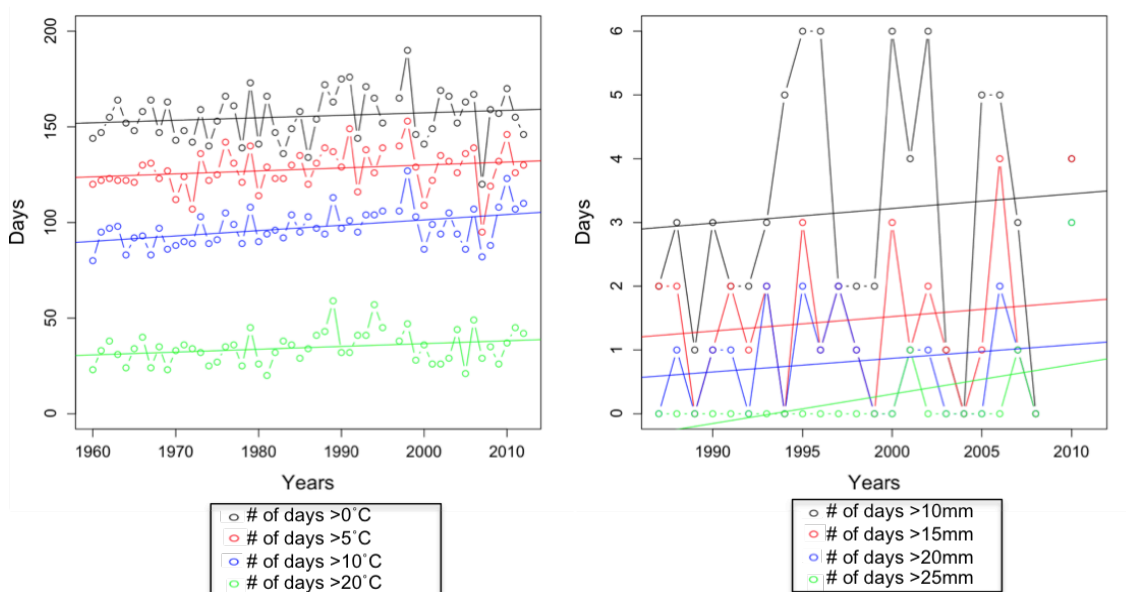


Figure 9. Number of warm days in Inuvik, NWT (1960-2011) and number of days with high-intensity precipitation events in Fort McPherson, NWT (1986-2010).

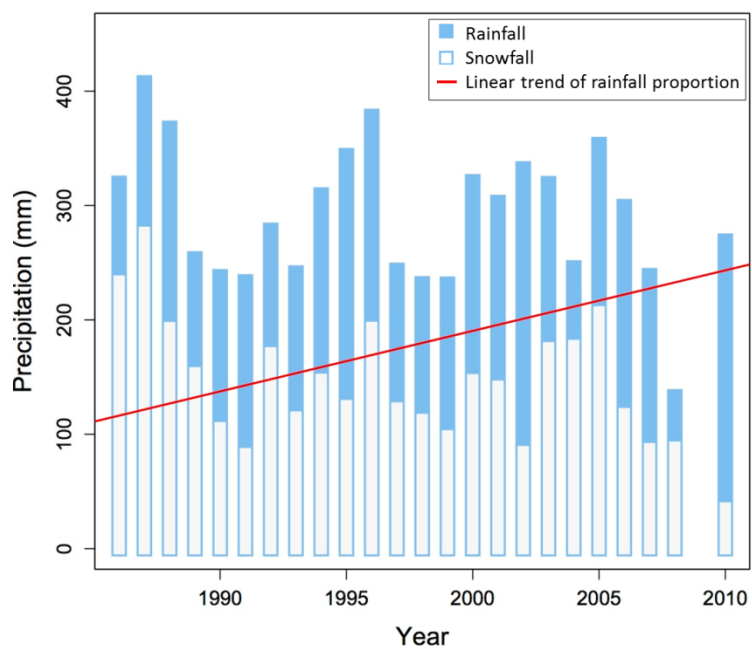


Figure 10. Total snowfall and rainfall for Fort McPherson, NWT, Canada (1986 to 2010).

Table 2. Slope value and statistical significance (p-value) of linear regression analysis of various climate parameters for Inuvik, NWT. Slope (p-value), bold values are statistically significant (p-value ≤ 0.05).

Parameter	1960-2011	1986-2011
MAAT	0.068 (0.001)	0.034 (0.31)
MSAT	0.029 (0.008)	-0.036 (0.29)
MWAT	0.119 (0.001)	0.113 (0.10)
Total precipitation*		-2.563 (0.196)
Rainfall*		0.726 (0.636)
Snowfall*		-3.290 (0.040)
# of days >0°C	0.131 (0.265)	-0.542(0.153)
# of days >5°C	0.155 (0.109)	-0.354(0.290)
# of days >10°C	0.279 (0.001)	0.077(0.792)
# of days >20°C	0.1481 (0.057)	-0.384(0.136)
# of days >10mm*		0.0417 (0.472)
# of days >15mm*		0.0247 (0.484)
# of days >20mm*		0.0270 (0.305)
# of days >25mm*		0.0428 (0.022)

* Precipitation data is from Fort McPherson, NWT, which has a more complete precipitation record

3.3. Glacial history of the Richardson Mountains

The study area is situated at the maximum westward extent of the Laurentide Ice Sheet (Figure 11). It was initially thought that the maximum extent of the Laurentide Ice Sheet was reached ca. 30 ka cal yr BP (Duk-Rodkin and Lemmen, 2000) but new U/Th ages indicate that the ice sheet reached the area at ca. 18 ka cal yr BP (Lacelle et al., 2013). During the last glacial maximum, the paleo-Porcupine River had its eastward flow impeded by the ice sheet at McDougall Pass, which resulted in the damming and raising of water levels of the glacial Lake Old Crow in the adjacent northern Yukon Territory. Eventually, water levels breached to the west at the Rampart of the Porcupine and joined the Yukon River in Alaska (Lemmen et al., 1994). The effect of the Laurentide Ice Sheet on the flow of the

Porcupine River was permanent, as it continues to drain into the Yukon River to this day (Duk-Rodkin and Lemmen, 2000).

The Richardson Mountains was affected by a standstill or re-advance of the Laurentide Ice Sheet at ca. 15 ka cal yr BP (Tutsieta Lake phase). The Tutsieta Lake phase is marked by the presence of moraines and ice-marginal melt channels which created new flow directions for rivers and streams. Parts of the ice melt channels that allowed meltwater to drain freely during the Tutsieta Lake phase still serve as the modern route of the Peel River (Figure 1) (Duk-Rodkin, and Lemmen, 2000). The Mackenzie River was also diverted northwards as a result of erosion and drainage diversion during the last glacial period. Temporary glacial lakes formed on the rolling moraine terrain of the Richardson Mountains due to the drainage blockage caused by glacial re-advance. The Richardson Mountains has been ice-free since ca. 13 ka cal yr BP (Zazula et al. 2009) and postglacial stream incision of up to 500m has produced the rivers and canyons observed in the Richardson Mountains (Duk-Rodkin, and Lemmen, 2000).

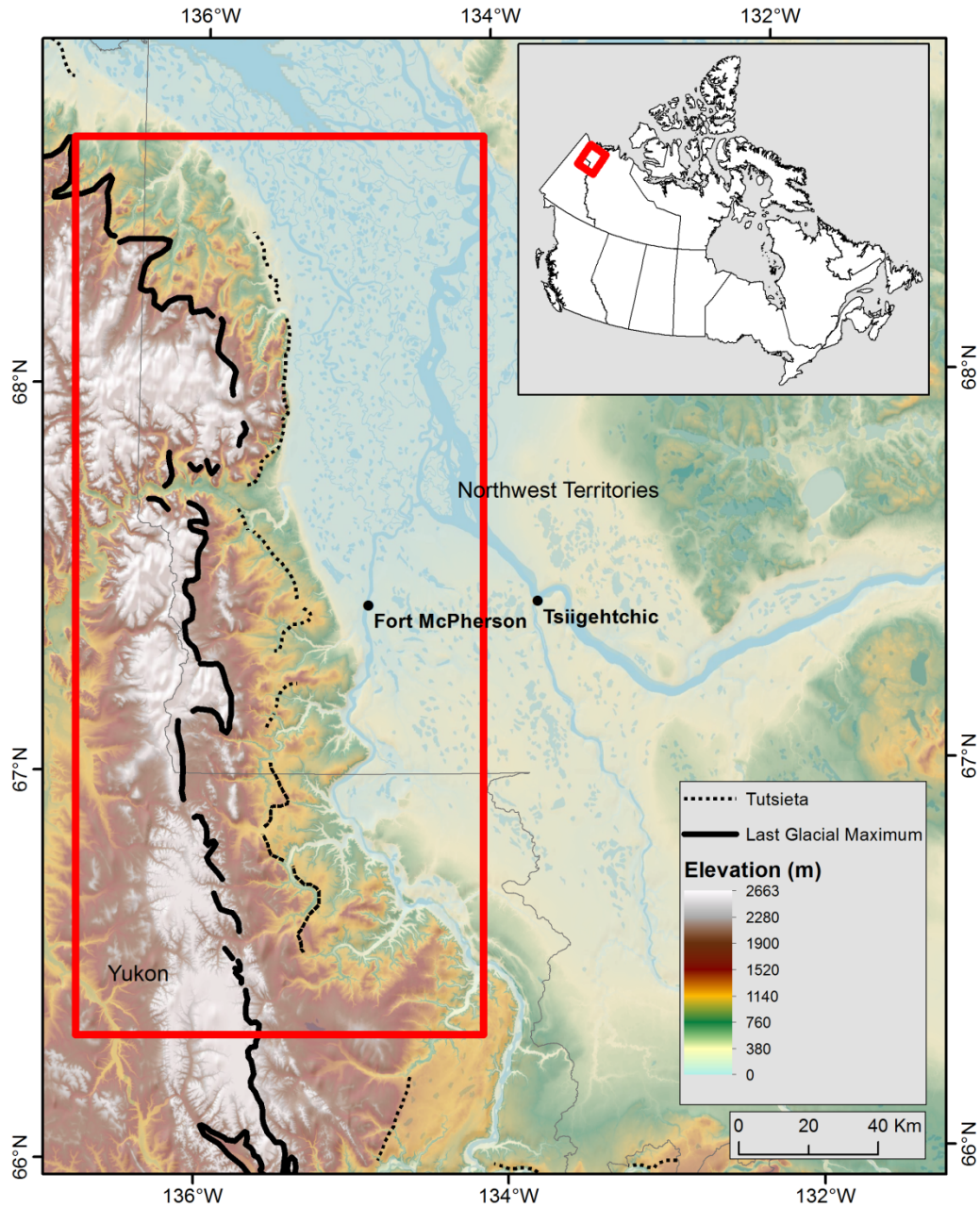


Figure 11. Location map of study area (red rectangle) showing position of the maximum westward extent of the Laurentide Ice Sheet (last glacial maximum) and the Tutsieta recessional phase (glacial limits from Duk-Rodkin and Hughes, 1992).

3.4. Holocene climate variation in the western Canadian Arctic

The regional climate history in the western Canadian Arctic and associated geomorphic changes is synthesized in Table 3. The regional climate during the early Holocene was warm and dry (Lauriol et al., 2002), as the Northern Hemisphere was receiving maximum solar insolation due to Milankovitch variations (Ritchie et al., 1983). Due to the presence of continental ice sheets at this time, sea level was considerably lower than today, resulting in the coastline in northern Yukon and adjacent NWT to be ca. 100 km north of today's shoreline (Hill et al., 1985). This increased distance from the coast gave the Tuktoyaktuk Peninsula and interior NWT a more continental climate. During the early Holocene, mean summer air temperatures were 3 – 8°C higher than present (Clark et al., 2004; Kauffmann et al., 2004). The warmer summer air temperatures resulted in active layer depths being roughly 1.5 to 2 times deeper than today and led to the development of thaw lakes (Mackay, 1992; Burn, 1997). Maximum active layer depths were reached around 8000 years ago (Burn, 1997), at which point climate conditions changed from warm and dry to warm and wet, due to a shift in atmospheric circulation (Lauriol et al., 2002; 2009). The dry climate of the early Holocene is thought to have been due strong anticyclonic activity from the northeast Pacific, resulting in northwestern Canada being dominated by westerly winds, while the warmth is due to increased solar insolation as a result of Milankovitch variations (Ritchie et al., 1983). The cooler and wetter climate of the mid-Holocene is attributed to increased sea-level from the recession of the ice sheets, giving this area a more maritime climate (Burn, 1997).

Table 3. *Synthesis of the regional Holocene climatic history reported by various studies. The radiocarbon age limits provide only rough dates for environmental change as these changes were gradual. (modified from Lauriol et al., 2002).*

Years BP	Regional Holocene climate history
0-4000	Climate similar to present with some minor fluctuations
4000-8000	Climatic cooling until temperatures reached values close to present. Retreat of tree-line and thinning of the thickness of the active layer.
8000-12,000	Summer temperatures were 3-7C higher than present. Climate was very dry between 10,000 and 9000 BP. Active layer was ~1.5 times thicker than today.
15,000-18,000	Gradual warming. Glacial retreat in the Richardson Mountains.

4. Methodology

4.1. Creation of the Landsat Image Stack

The Landsat Image Stack Trend analysis method was developed by Fraser et al. (2011; 2012a) in an effort to monitor temporal changes in vegetation across broad regions. The methodology of the Landsat Image Stack Trend Analysis included in this section is a simplified description of the steps required to create the Landsat Image Stack and is adapted from a protocol document (Fraser et al., 2012b). This technique uses a dense stack of Landsat images (derived from images acquired annually, when possible) for a given area, applies the Tasseled Cap transformations to each Landsat scene and then calculates the trends in the change over time of the three Tasseled Cap transformations for each pixel. The stack of Tasseled Cap Landsat images and the Tasseled Cap trends maps were created as part of the NWT Cumulative Impact Monitoring Program (CIMP) at the Canadian Centre for Remote Sensing.

The Landsat Image Stack Trend Analysis applies the Tasseled Cap transformations to Landsat TM (Landsat 5) and Landsat ETM+ (Landsat 7) imagery acquired between 1985 and 2011. Landsat imagery has a resolution of 30m, a swath width of roughly 185 km and a revisit frequency of 16 days (NASA, 2013). Since May 31, 2003, each Landsat 7 scene is missing roughly 22% of its imagery due to striping, following the failure of the Scan Line Corrector (SLC) (USGS, 2013). Landsat imagery from 1984 to present is available from various websites operated by the USGS, including Earth Explorer and Glovis. Tables 4 and 5 displays the wavelengths captured by the TM and ETM+ sensors onboard the Landsat 5 and 7 satellites, respectively.

Table 4. *Spectral bands of the Landsat Thematic Mapper (TM) sensor on the Landsat 5 satellite (USGS, 2012).*

Band	Wavelength (μm)	Resolution (m)
Band 1 - blue	0.45-0.52	30
Band 2 - green	0.52-0.60	30
Band 3 - red	0.63-0.69	30
Band 4 - Near Infrared	0.77-0.90	30
Band 5 - Short-wave Infrared	1.55-1.75	30
Band 6 - Thermal Infrared	10.40-12.50	120*30
Band 7 - Short-wave Infrared	2.09-2.35	30

Table 5. *Spectral bands of the Landsat Enhanced Thematic Mapper Plus (ETM+) sensor on the Landsat 7 satellite (USGS, 2012).*

Band	Wavelength (μm)	Resolution (m)
Band 1 - blue	0.45-0.52	30
Band 2 - green	0.52-0.60	30
Band 3 - red	0.63-0.69	30
Band 4 - Near Infrared	0.77-0.90	30
Band 5 - Short-wave Infrared	1.55-1.75	30
Band 6 - Thermal Infrared	10.40-12.50	60*30
Band 7 - Short-wave Infrared	2.09-2.35	30
Band 8 - Panchromatic	.52-.90	15

In this thesis, 122 Landsat TM and ETM+ summer scenes with minimal cloud cover (<10%) were acquired via Glovis (<http://glovis.usgs.gov>) between the years 1985 and 2011. Figure 12 displays the extent of the three image stacks and Appendix A lists all of the Landsat scenes used in the creation of the Tasseled Cap trend map which cover the extent of the Richardson Mountains.

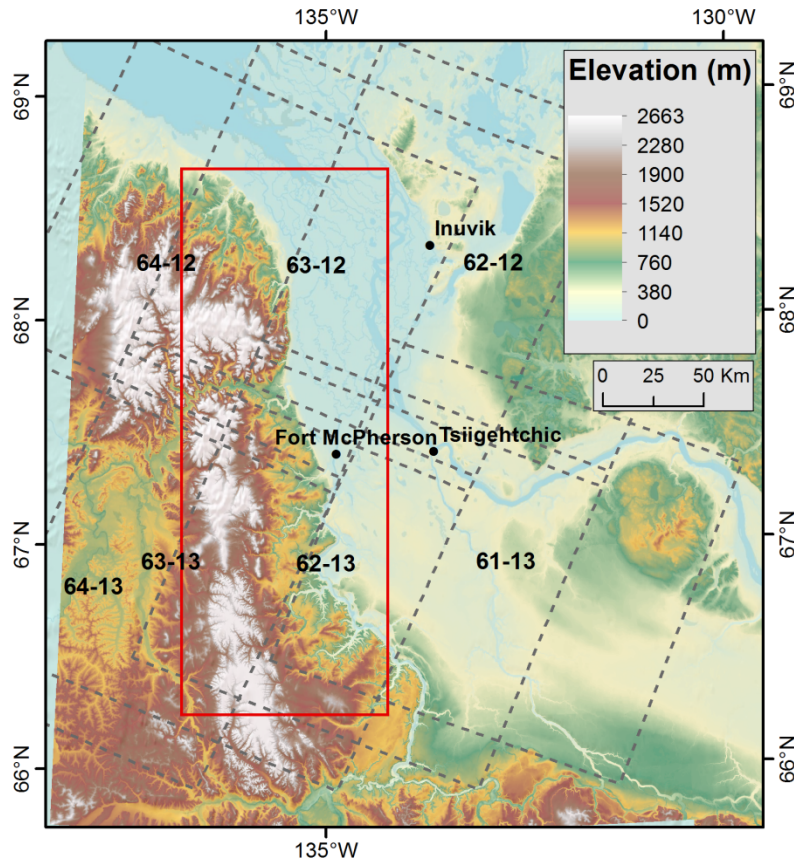


Figure 12. Extent of the Landsat image stacks in the Richardson Mountains, NWT, Canada.

The Landsat scenes acquired from Glovis were in six raw tif files for each Landsat channel. These files were combined into a single .pix file using the Data Merge Wizard tool in PCI Geomatica 10. The projection was set to UTM zone 8N (datum and spheroid = WGS 1984).

To ensure the images were displaying a common radiometric reflectance and to ensure compatibility between the Landsat TM and ETM+ images, the digital numbers were then converted to top of atmosphere reflectance. This was done by running an EASI script which merged the raw TIFF files and 6 new channels that have been variably scaled to a uniform 8-bit range (one for each Landsat band) .

Clouds were then masked out of each Landsat image, through both manual and automated means. Preliminary masking was performed with another EASI

script which writes a layer masking cloud extent to an 8-bit bitmap layer and another layer which masks the bands of no-data of Landsat 7 imagery, following the failure of its Scan Line Corrector in 2003 (USGS, 2013) to the .pix file of each Landsat scene. This script is not able to perfectly mask all clouds in the Landsat scene, so clouds were manually masked by editing the cloud mask bitmap layer in PCI Geomatica.

Following this, the three Tasseled Cap transformations were calculated (brightness, greenness and wetness) for each pixel of each image. This was accomplished using an EASI script that creates 3 empty 16-bit channels and writes the three Tasseled Cap transformations to these channels. Tables 6 and 7 provide the equation for each of these three Tasseled Cap transformations. The Tasseled Cap transformations were created as a way of summarizing the information contained in the various spectral bands into a more manageable and more easily interpretable number of features (Crist and Cicone, 1984). Brightness is a weighted sum of all six reflective bands, which means that this transformation is particularly sensitive to any changes in total reflectance, as well as any physical processes that might affect total reflectance (Crist and Cicone, 1984). The second Tasseled Cap feature, greenness is a measure of the contrast between the weighted sum of the visible bands and the near-infrared bands. As its name suggests, greenness is most capable at displaying green vegetation, which absorbs high amounts of the visible bands, while reflecting high amounts of infrared (Crist and Cicone, 1984). The third feature, wetness, contrasts the weighted sum of the visible and near-infrared bands with the sum of the longer-infrared bands. The longer infrared bands are particularly able to capture soil and plant moisture (Crist and Cicone, 1984).

The stack of Landsat imagery was used in two ways. First, a Tasseled Cap linear trends map was created to evaluate if this approach can identify thaw slumps, and from which the distribution of thaw slumps in the Richardson Mountains was mapped. Secondly, annual trends in the Tasseled Cap pixels were used to derive patterns in thaw slump activity (initiation, growth and stabilization).

Table 6. Equation used for each Tasseled Cap transformation and NDVI (Crist and Cicone, 1984) (Weier and Herring, 2013)

Vegetation Index	Equation
Tasseled Cap Brightness	$=((a)(\text{Band } 1)) + ((b)(\text{Band } 2)) + ((c)(\text{Band } 3)) + ((d)(\text{Band } 4)) + ((e)(\text{Band } 5)) + ((f)(\text{Band } 7))$
Tasseled Cap Greenness	$=((a)(\text{Band } 1)) + ((b)(\text{Band } 2)) + ((c)(\text{Band } 3)) + ((d)(\text{Band } 4)) + ((e)(\text{Band } 5)) + ((f)(\text{Band } 7))$
Tasseled Cap Wetness	$=((a)(\text{Band } 1)) + ((b)(\text{Band } 2)) + ((c)(\text{Band } 3)) + ((d)(\text{Band } 4)) + ((e)(\text{Band } 5)) + ((f)(\text{Band } 7))$

Where: the Bands are defined in Table 4 and 5 and a, b, c, d, e and f are weighing coefficients defined in Table 7.

Table 7. Thematic Mapper Tasseled Cap Coefficients (Crist and Cicone, 1984)

Tasseled Cap Feature	Landsat bands					
	1 (a)	2 (b)	3 (c)	4 (d)	5 (e)	7 (f)
Brightness	0.3037	0.2793	0.4743	0.5585	0.5082	0.1863
Greenness	-0.2848	-0.2435	-0.5436	0.7243	0.084	-0.18
Wetness	0.1509	0.1973	0.3279	0.3406	-0.7112	-0.4572

4.2. Creation of the Linear Trends map

Following the calculation of the three Tasseled Cap transformations for each Landsat scene, the linear trends map was created by combining all Landsat images into three image stacks. This was done using an EASI script which transfers the three Tasseled Cap channels of each Landsat scene to an empty database, which covers the extent of the image stack, created using CIMPRO.

Trends of the three indices (Tasseled Cap brightness, greenness and wetness) for each pixel in the stack were calculated using the Theil-Sen regression technique. The advantage of this regression technique is its relative inability to be affected by outliers that can skew trends, such as clouds that were missed during cloud-masking, snow or rain on the ground (Fraser et al., 2012). This step was performed using a program created by D. Pouliot of CCRS with the IDL programming language. This program calculates the trends of the three Tasseled Cap transformations of each pixel within a stack. An integer version of the Tasseled Cap trends was then calculated by multiplying the three rasters by 1000, making viewing of the trends easier. Figure 13 summarizes the creation of the Tasseled Cap linear trends map from the raw Landsat imagery.

To evaluate if the Landsat change-detection map is able to identify thaw slumps, the map was verified using high-resolution Ikonos (two images from 2000 and 2012) and Quickbird (one image from 2008) satellite imageries, a mosaic of normalized 20m SPOT imagery and air photographs.

Each thaw slump identifiable using the Tasseled Cap Linear trends map was traced in ArcGIS 10.0 and its surface area (in ha) was calculated using the Calculate Geometry tool in ArcMap. To simplify analysis, the thaw slump polygon shapefile was converted to a point shapefile using the Feature to Point function (Spatial Analyst) in the ArcToolbox. For each slump, their elevation and aspect was extracted from the Canadian Digital Elevation Data, which has a spatial resolution of 30m.

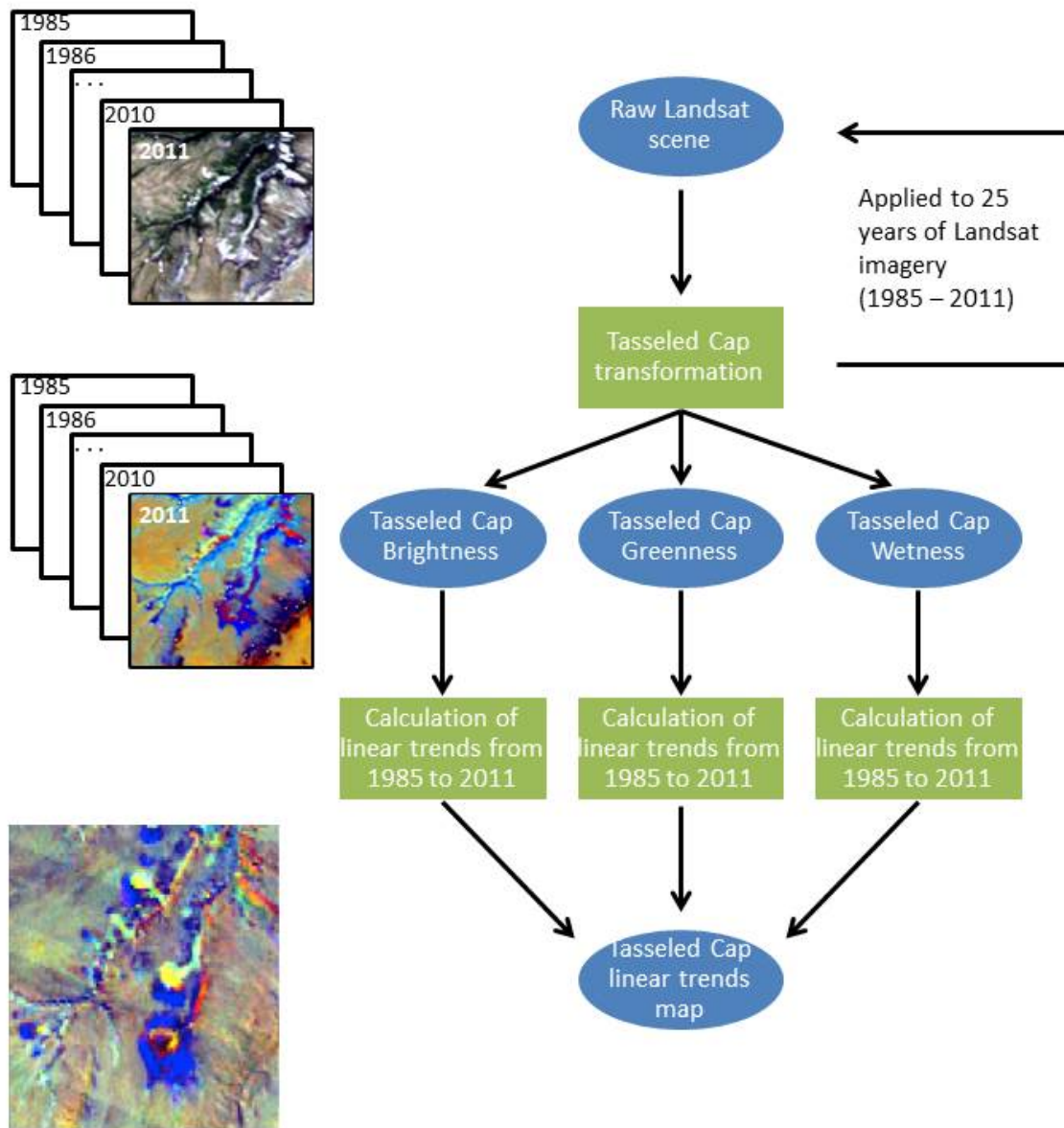


Figure 13. Summary of the Landsat Image Stack Trend Analysis method.

4.3. Analysis of annual Tasseled Cap trends

To derive the annual Tasseled Cap (Brightness, Greenness, Wetness) values for each pixel, each Landsat pixel in a thaw slump was assigned a point in a point

shapefile (Figure 14). The Tasseled Cap values for each point in the shapefile within the thaw slump were then extracted using the Sample function in the Spatial Analyst ArcToolbox. This function works by extracting the pixel values of all selected raster cells to a point or polygon shapefile. When a given point or area contains more than one pixel, this function uses spatial interpolation to calculate pixel values. Thus, to ensure accurate pixel values, each pixel within a thaw slump area was assigned a unique point. The output table contains the annual Tasseled Cap (Brightness, Greenness and Wetness) values from each Landsat scene of all pixels in a thaw slump. The annual Tasseled Cap values of pixels from successive years were plotted to determine if thaw slump activity could be extracted from these trends.

Figure 15 displays the Tasseled Cap values of a transect of pixels across a slump floor from 1988 to 2011. As can be seen in the Brightness and Greenness index values, the pixel values of all pixels remain relatively constant, ranging in values of 300 to 350 for Brightness and 100 to 150 for Greenness until roughly 1998. At this point, the Brightness and Greenness values drop to around 175 and below 50 for Brightness and Greenness respectively. This drop in value represents the time during which the headwall is retreating through the pixel and the corresponding change in ground type from vegetated tundra to wet mud. Upon pixel values dropping as a result of headwall retreat, Brightness values do not remain consistently low, but can show sudden spikes in Brightness values, while Greenness values remain consistently low. This is due to Brightness being an index of overall reflectivity of the ground surface, while Greenness only displays the reflectivity of vegetation. This means that Brightness values are more likely to vary, as a result of the ground conditions on any given day (dry or wet ground), while vegetation amounts do not vary from day to day. As shown in this figure, it is very hard to interpret any pattern in the Wetness index values. Upon comparing the trends for the three Tasseled Cap indices, it was decided to concentrate analysis solely upon the Greenness index.

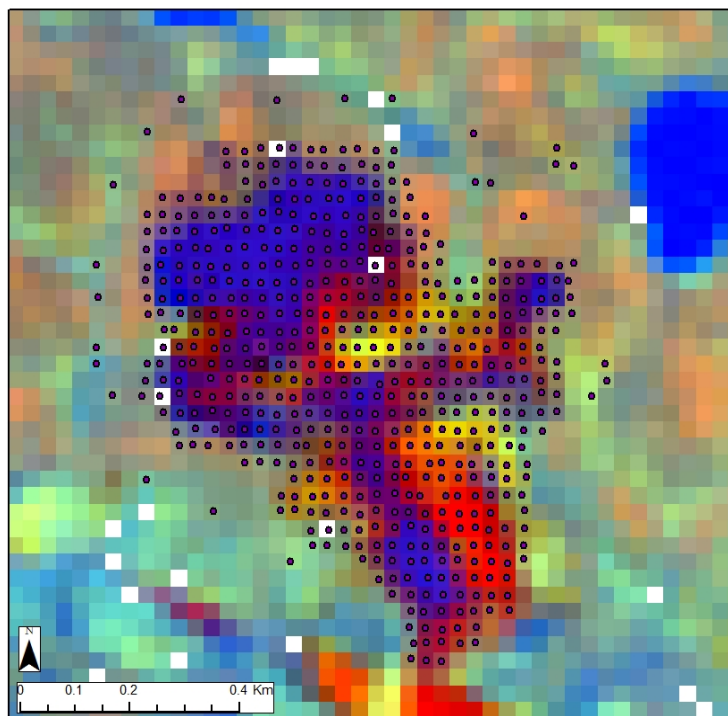
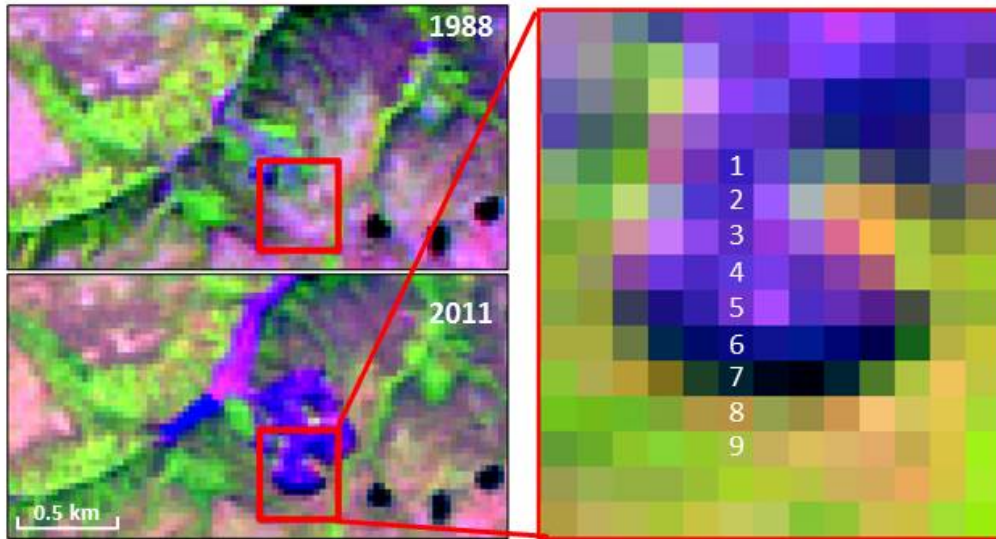


Figure 14. Example of pixel sampling of a thaw slump and the surrounding undisturbed tundra. Each point represents one Landsat pixel. Yearly (1985-2011) Tasseled Cap values were extracted for each point.



Legend	
—	Pixel 1
—	Pixel 2
—	Pixel 3
—	Pixel 4
—	Pixel 5
—	Pixel 6
—	Pixel 7
—	Pixel 8
—	Pixel 9

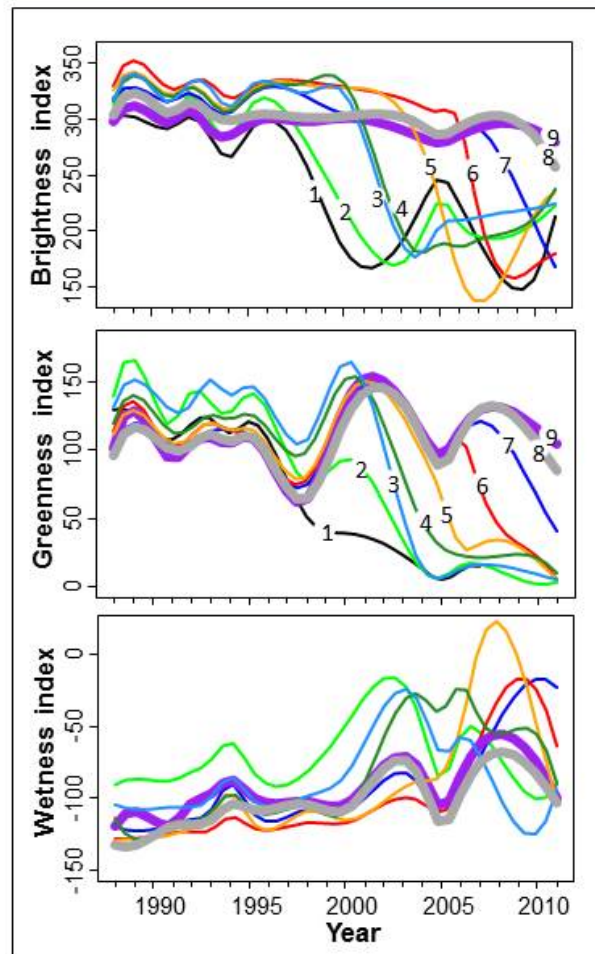


Figure 15. Tasseled Cap values (Brightness, Greenness and Wetness) of a transect of pixels from 1988 to 2011

5. Results

5.1. Identification of thaw slumps using the Tasseled Cap method

Figure 16 compares how different portions of a thaw slump are represented with an Ikonos image (4m resolution), Landsat natural colour (band combination 3,2,1; resolution 30m) and the Tasseled Cap linear trends map. The slump is divided into three parts, the active, wet slump floor (A), the stable and re-vegetating slump floor (B) and the debris flow (C). The linear trends map displays the wet portion of the slump floor in a distinctive blue and the stable, revegetating portion as orange, yellow and green. Using the Ikonos and Landsat natural colour views, it is hard to distinguish between these two areas of the slump floor. The stable, revegetating portion is easily distinguishable from the surrounding vegetated tundra on the Linear Trends map. This is because the re-vegetating portion of the thaw slump has been more dynamic between the years of 1985 to 2011, as the vegetation slowly changes to match the surrounding tundra. In comparison, the surrounding tundra has been rather static, as it has not changed from tundra vegetation over this period.

Figure 17 shows the same area as figure 16, but shows the ability of the Linear Trends map at identifying stabilized thaw slumps, when compared with different Landsat views. This figure shows the growth of a thaw slump in air photos from 1954 to 1971 and its near stabilization by 1988. When using 2011 natural colour and single year Tasseled Cap Landsat views, this stable thaw slump is nearly indistinguishable from the surrounding landscape, but is easily identifiable on the Linear Trends map. As discussed with the previous figure, this is because the re-vegetating thaw slump floor has been more dynamic between the years of 1985 to 2011 than the surrounding tundra. Due to the polycyclic nature of thaw slumps (Kokelj et al., 2009), stable thaw slumps can re-initiate from any process that re-exposes the headwall. This thaw slump was observed in the summer of 2013 to have completely re-initiated, with a headwall of several meters and its existence may have gone unnoticed if it had not been identified on the Linear Trends map.

This shows the importance of mapping stable thaw slumps for understanding the potential impact of thaw slump activity.

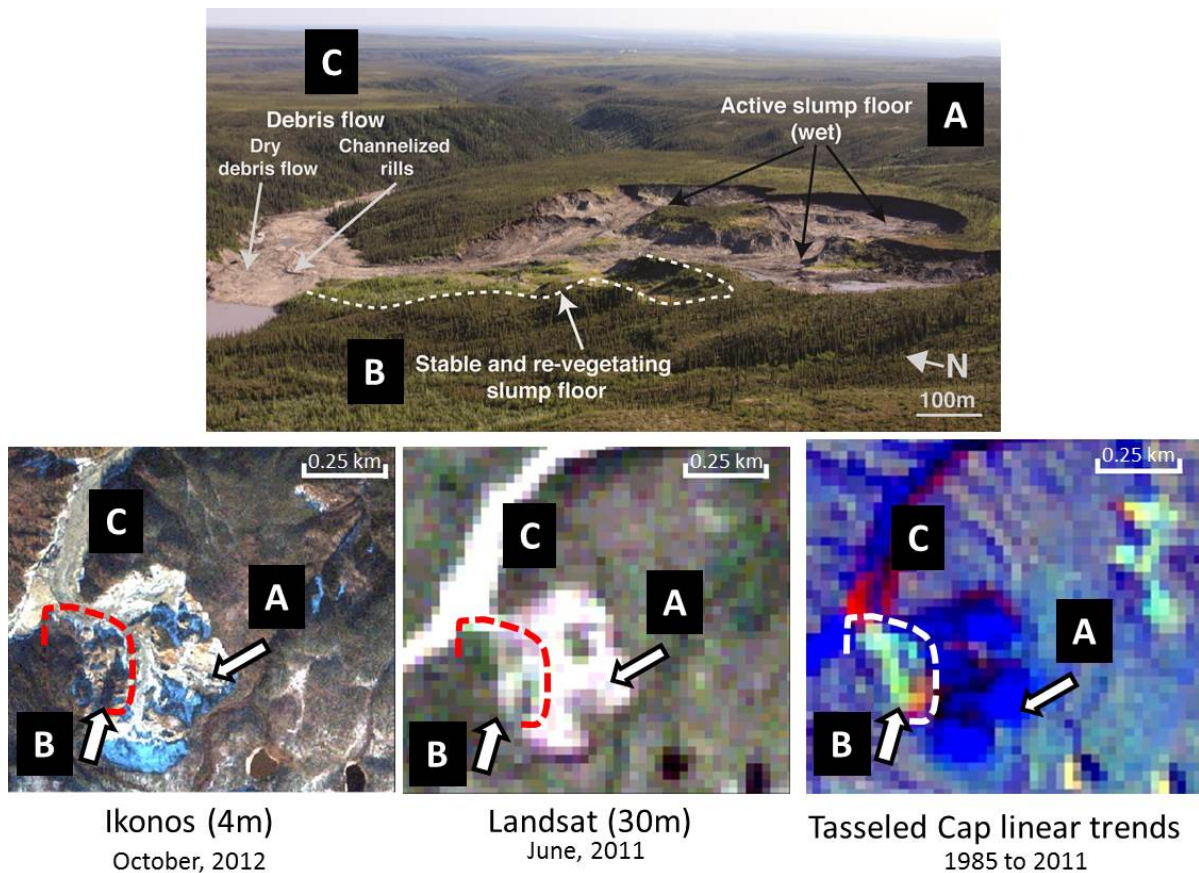


Figure 16. Comparison of different-single image Landsat views and Ikonos imagery with Tasseled Cap and Linear Tasseled Cap Trends maps.

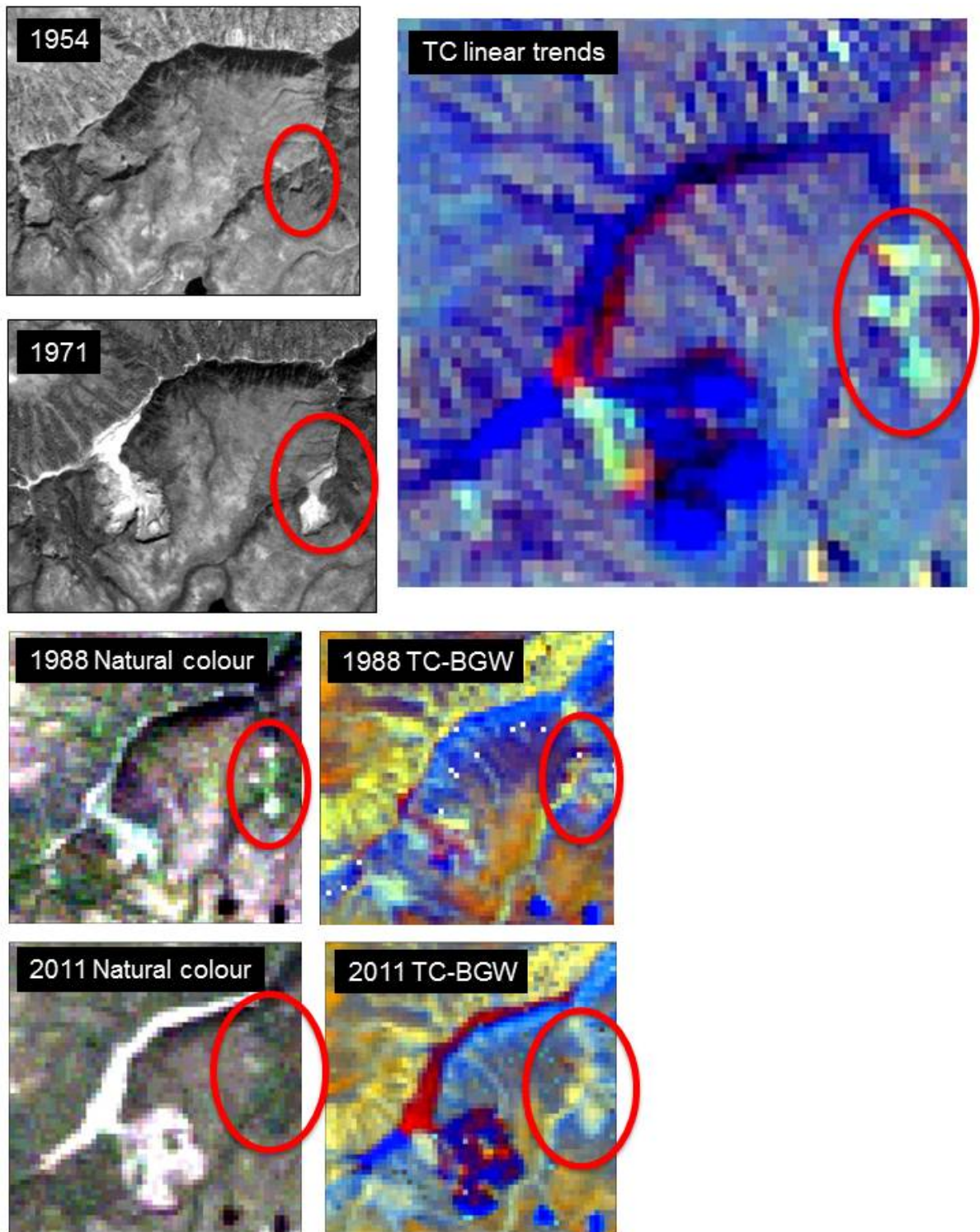


Figure 17. Visualization of a stabilizing thaw slump from 1954 to 2011, using airphotos, Landsat natural colour, single year Tasseled Cap transformation and the Tasseled Cap linear trends map.

5.2. Distribution of thaw slumps

Using the linear Tasseled Cap trends (between the years 1985 and 2011) map, the distribution of thaw slumps in the Richardson Mountains was mapped (Figure 18). More than 251 thaw slumps are found throughout the Richardson Mountains, with the Willow River, Vittrekwa River and Stony Creek watersheds containing the highest density of slumps and accounting for 60% of all slumps (Table 8). This number includes thaw slumps that had initiated before 1985. The surface area of slumps range between 0.2 and 52ha, with an average surface area of 5.4ha (Figure 19). The majority (220/251) of thaw slumps have a surface area less than 10 ha. Thaw slumps in the Richardson Mountains are found on slopes of all aspects, but are found predominantly on north-east facing slopes (average aspect of 60°)(Figure 19). In addition, these features are found on elevation, ranging from 0 to 765m.

Table 8. *Thaw slump distribution in the various watersheds in the Richardson Mountains, NWT, Canada.*

Watershed name	Watershed Area (km ²)	Number of slumps	Density of thaw slumps per 1000 km ²	Average surface area of slump (ha)	Average elevation of slump (m)
Big Fish River	2935.54	8	2.73	2.8	246.5
Unnamed watershed #1	404.8	4	9.88	2.2	293.5
Willow River	808.77	59	72.95	3.38	418.48
Unnamed watershed #2	99.36	1	10.06	1.5	566
Rat River	2692.94	30	11.14	4.5	428.8
Stony Creek	1224.46	40	32.67	6.75	429.41
Vittrekwa River	1979.11	52	26.27	1.5	449.37
Unnamed watershed #3	314.86	1	3.18	7.04	351
Road River	1545.54	13	8.41	5	417.54
Unnamed watershed #4	155.18	18	115.99	1.8	335.05
Trail River	1281.29	24	18.73	9.4	374.54
Caribou River	2227.77	3	1.35	18.4	410

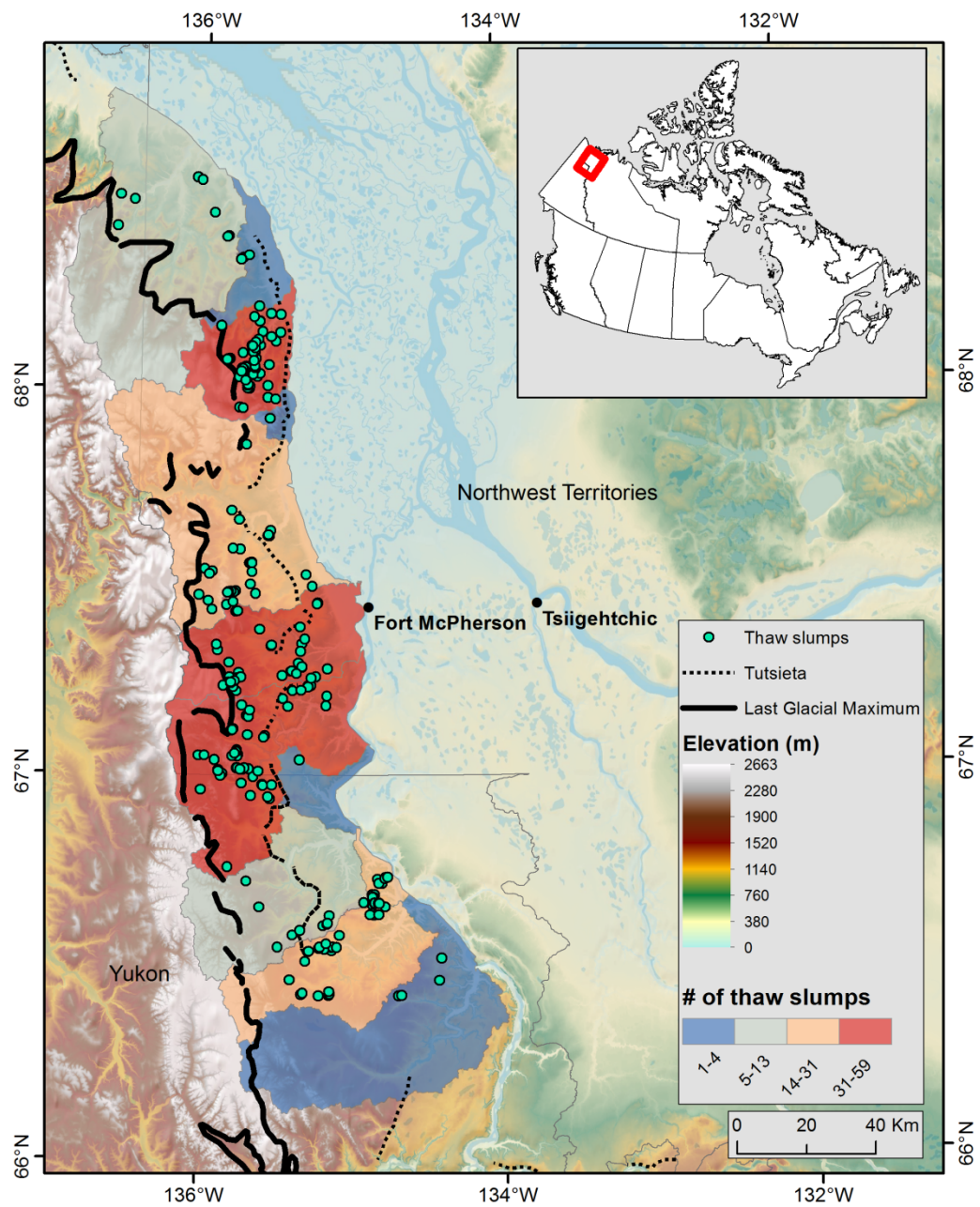


Figure 18. Distribution of 251 thaw slumps identified using Tasseled Cap trends between 1985 and 2011 in the Richardson Mountains, NWT, Canada. This number includes thaw slumps active before the year 1985.

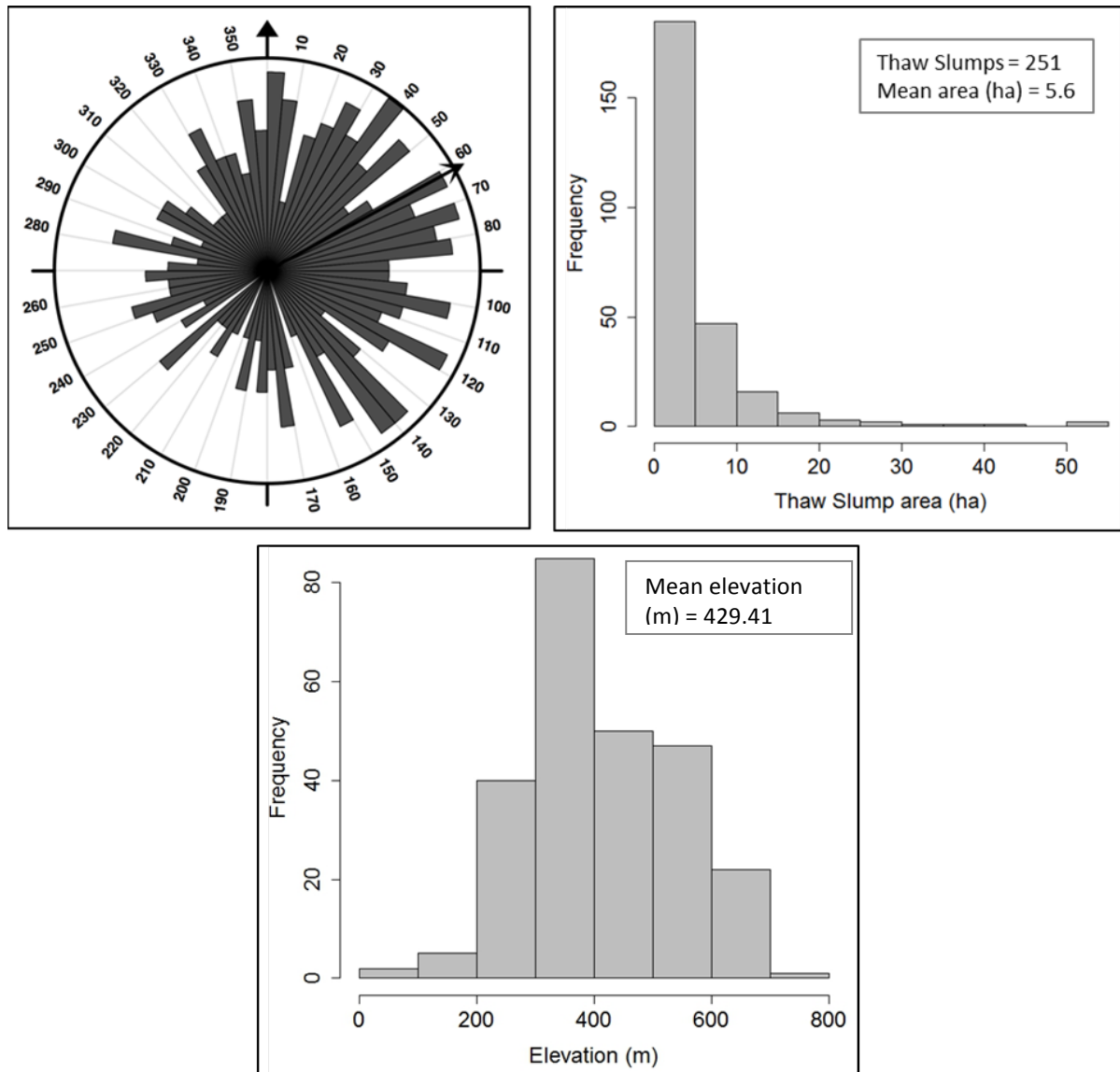


Figure 19. Distribution of aspect, surface area and elevation of all identified thaw slumps in the Richardson Mountains, NWT, Canada.

The following section will describe the distribution of thaw slumps within each watershed. The Big Fish River (Figure 20) watershed is the northernmost watershed on the eastern side of the Richardson Mountains. With a surface area of 2935.6 km², the Big Fish River watershed is the largest watershed in the Richardson Mountains yet is only minimally impacted by thaw slumps. This watershed contains

only 8 thaw slumps, which have an average surface area of 2.7 ha and the largest being 4.9 ha.

To the east of the Big Fish River watershed is the Unnamed Watershed #1 (Figure 21), with a surface area of 404.8 km². Thaw slump activity in this watershed is also very low, as there are only 4 thaw slumps with an average surface area of 2.2ha and the largest being 3.1ha.

Directly south of the Unnamed Watershed #1 is the Willow River watershed (Figure 22). This watershed has an area of 808.77 km² and has the largest number of thaw slumps of any watershed in the Richardson Mountains. This watershed has 59 thaw slumps, which have an average area of 3.4ha. The largest thaw slump has a surface area of 12ha.

The smallest watershed, Unnamed Watershed #2 (Figure 23), in the Richardson Mountains is found to the south of the Willow River, with an area of 99.36 km². This watershed contains only one thaw slump, which has a surface area of 1.5ha.

The second largest watershed, Rat River (Figure 24), has an area of 2692.94 km² and borders the Big Fish River, Willow River and Unnamed Watershed #2 watersheds to the north. This watershed has 30 thaw slumps, with all but one being located in the southern half of the watershed. Thaw slumps in this watershed have an average area of 4.5 ha, with the largest being 13.2 ha.

To the south of the Rat River watershed, is the Stony Creek watershed (Figure 25), which has an area of 1224.46 km². This watershed contains 41 thaw slumps, with an average area of 6.6 ha and contains two of the largest thaw slumps, which have surface areas of 48 and 52 ha.

The Vittrekwa River watershed (Figure 26) is found to the south of the Stony Creek watershed and has an area of 1979.11 km². After Willow River, this watershed has the second-most thaw slumps, with 52. These thaw slumps have an average area of 7.0 ha, with the largest having an area of 43.4 ha.

South-east of the Vittrekwa River watershed is Unnamed Watershed #3 (Figure 27). This watershed has an area of 314.86 km² and contains only one thaw slump, which has an area of 7.8 ha.

South of the Unnamed Watershed #3 is the Road River watershed (Figure 28), which has a surface area of 1545.54 km². This watershed contains 13 thaw slumps, which have a mean area of 5.0 ha and the largest thaw slump having an area of 13.2 ha.

The Unnamed Watershed #4 (Figure 29) is found to the east of the Road River watershed and, with an area of 155.18 km², is the second smallest watershed in the Richardson Mountains. Despite containing 18 thaw slumps, their impact is relatively minimal, as the average surface area of thaw slumps in this watershed is only 1.8 ha, with the largest thaw slump having an area of 6.7 ha.

South of the Unnamed Watershed #4 and Road River watershed is the Trail River watershed (Figure 30). This watershed has an area of 1281.29 km² and contains 24 thaw slumps. These thaw slumps have a mean area of 9.4 ha, with the largest thaw slump having an area of 39.4 ha.

The southernmost watershed in the Richardson Mountains is the Caribou River watershed (Figure 31). This is the third largest watershed, with an area of 2227.77 km² and contains only four thaw slumps. These thaw slumps have a mean area of 18.4, with the largest thaw slump having an area of 18.4 ha.

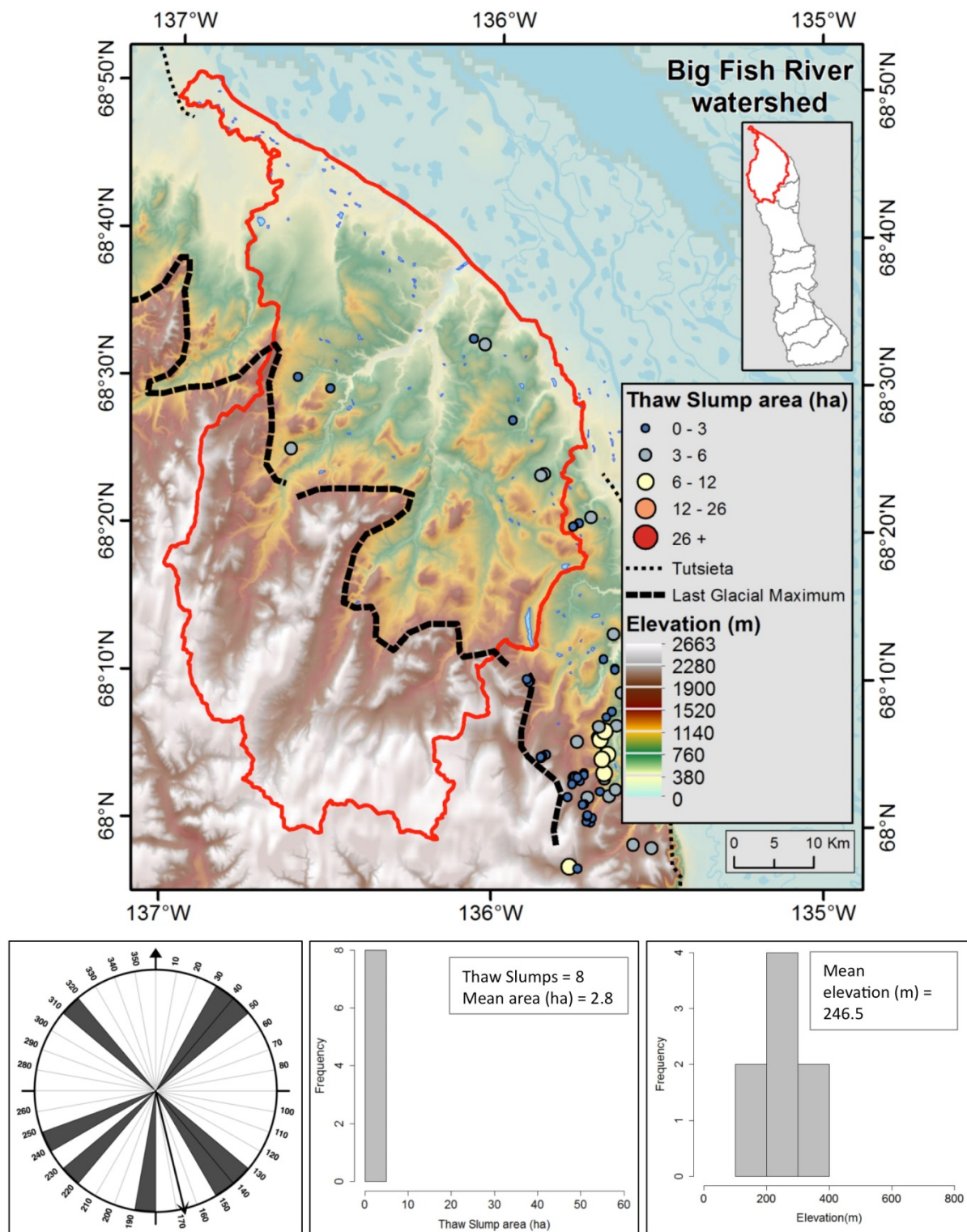


Figure 20. Distribution of thaw slumps in the Big Fish River watershed

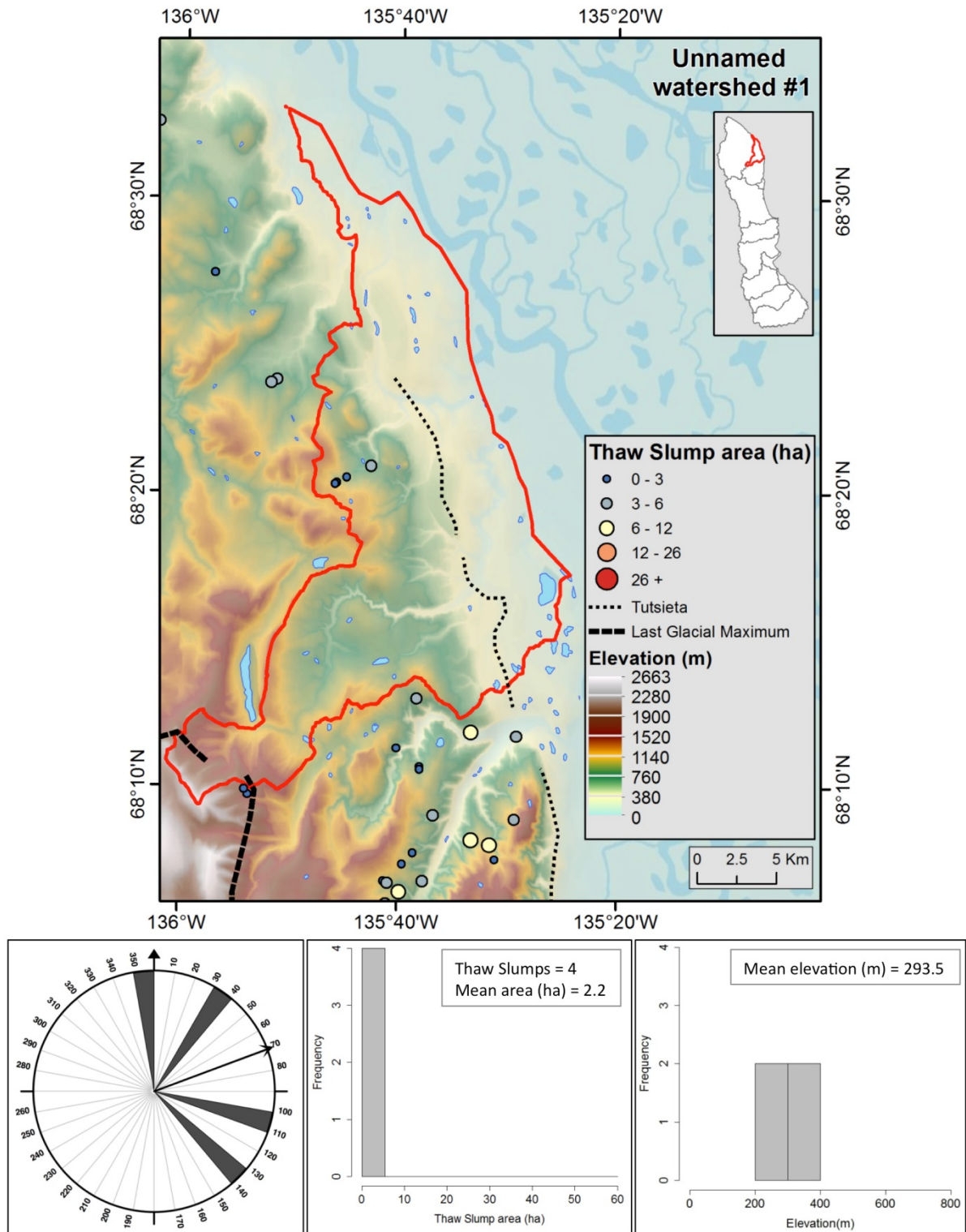


Figure 21. Distribution of thaw slumps in the Unnamed Watershed #1

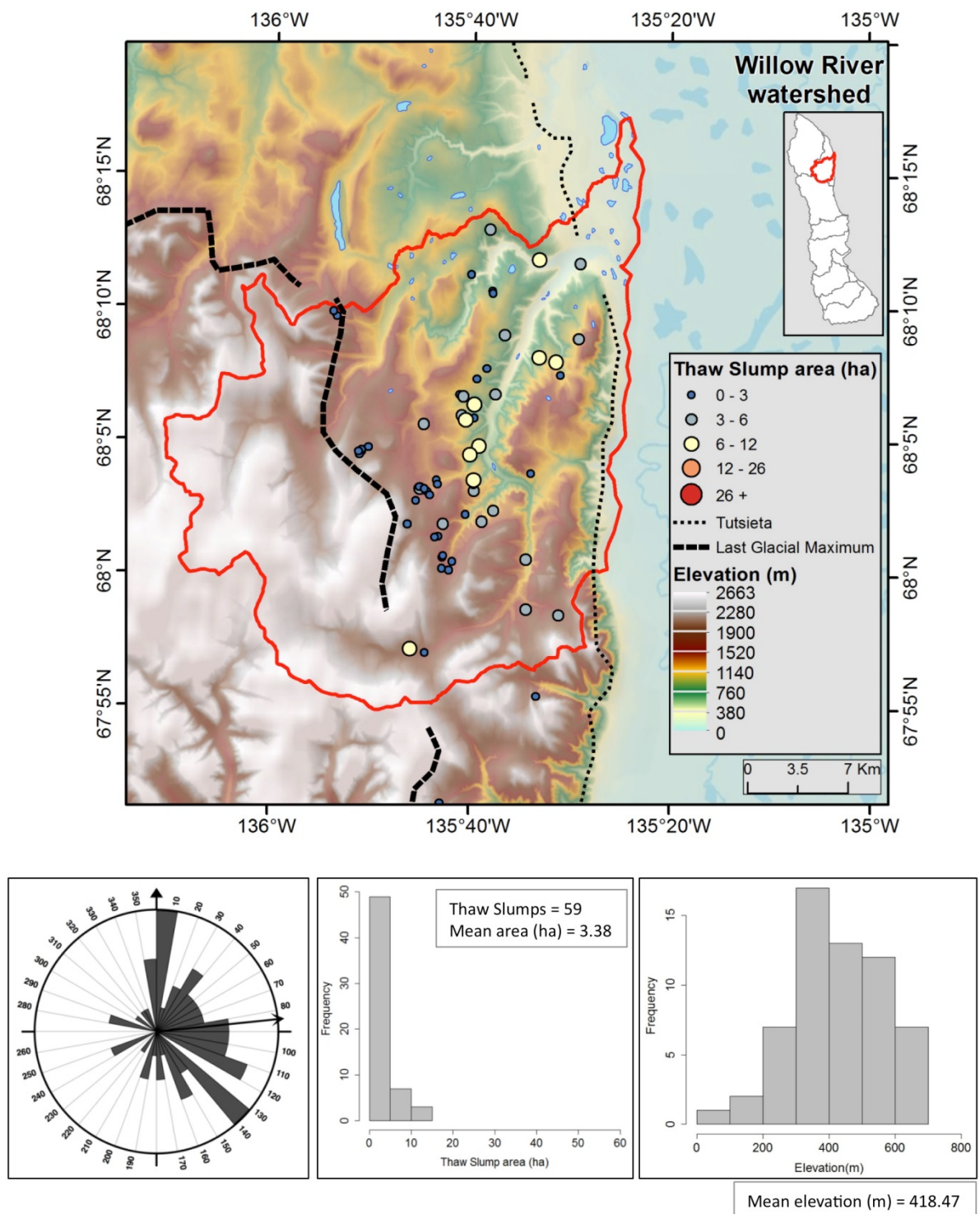


Figure 22. Distribution of thaw slumps in the Willow River watershed

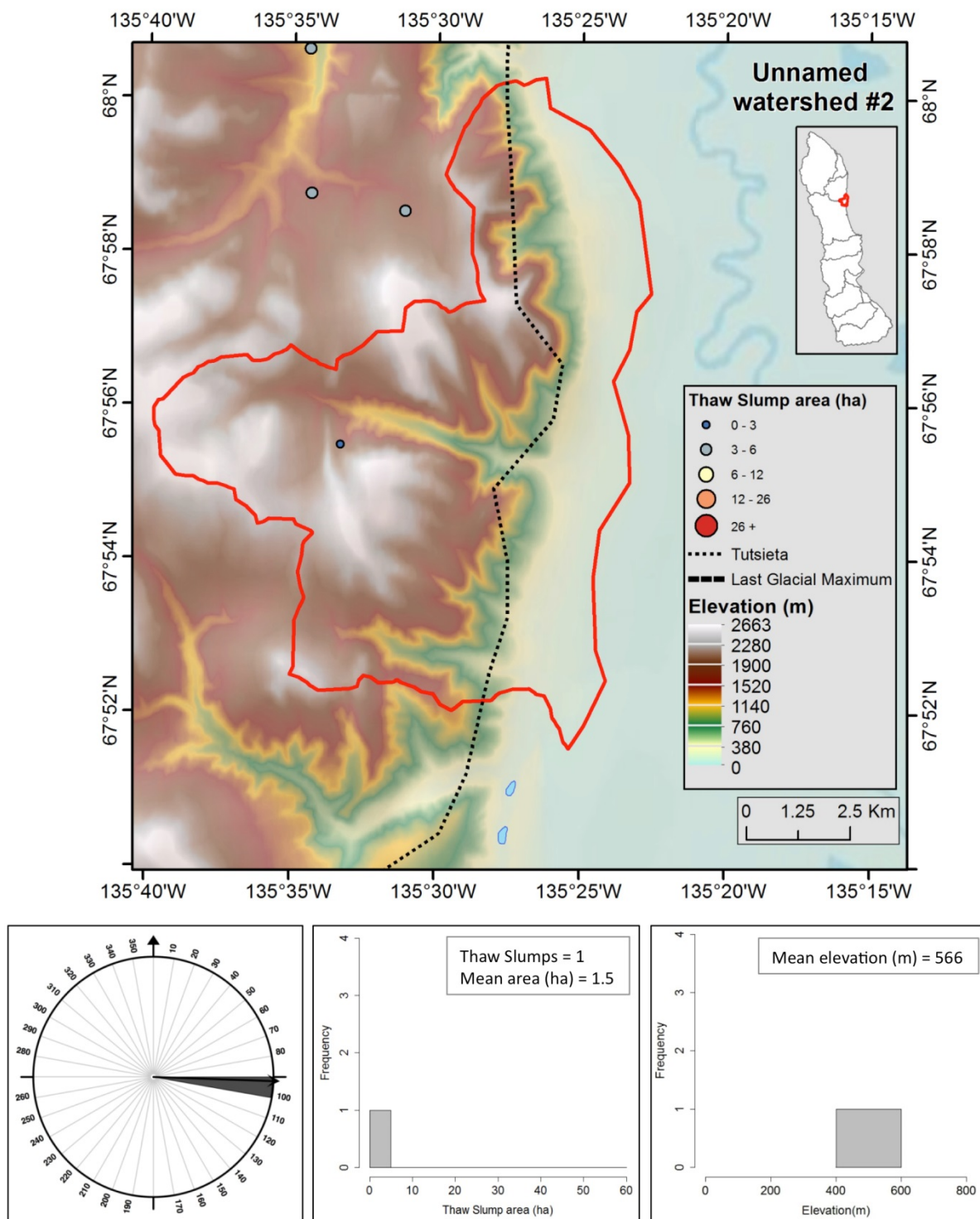


Figure 23. Distribution of thaw slumps in the Unnamed Watershed #2

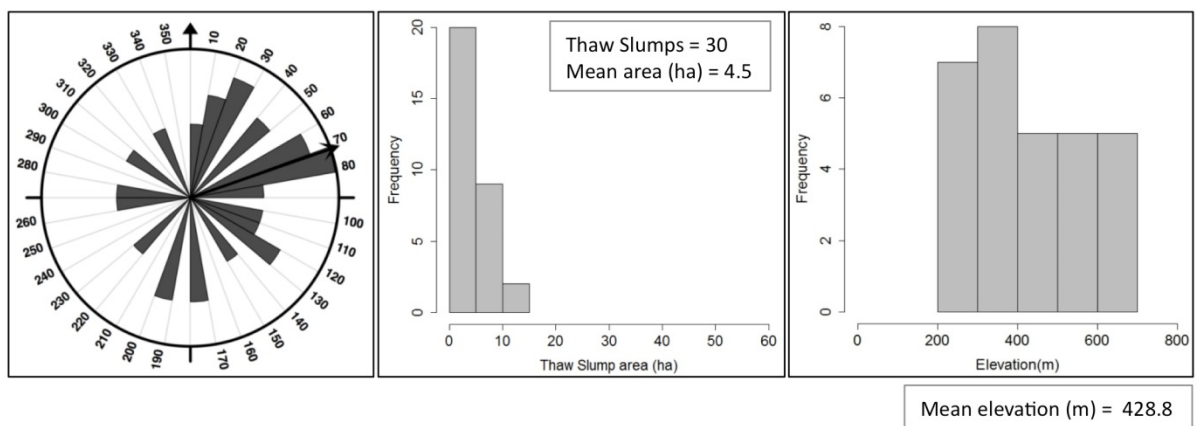
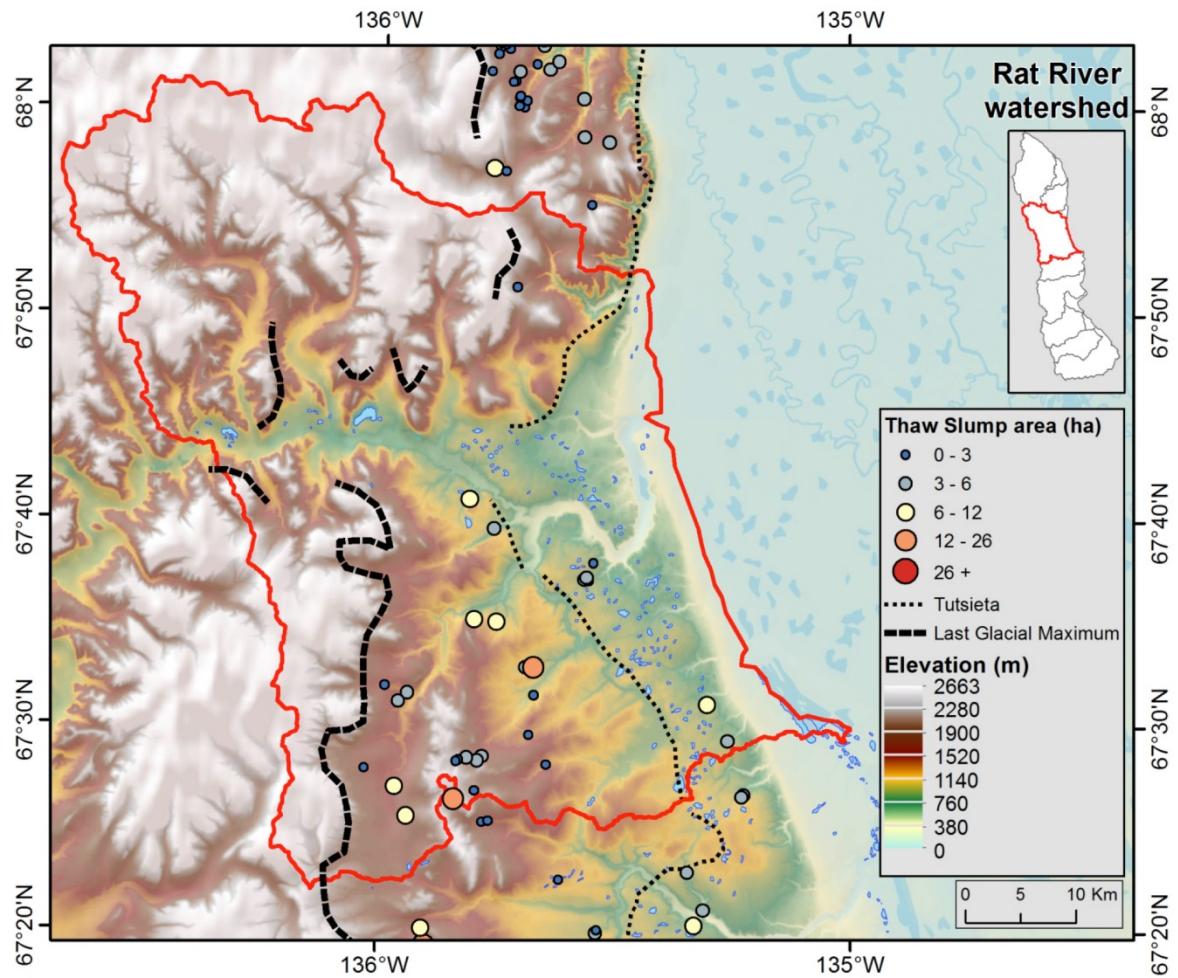


Figure 24. Distribution of thaw slumps in the Rat River watershed

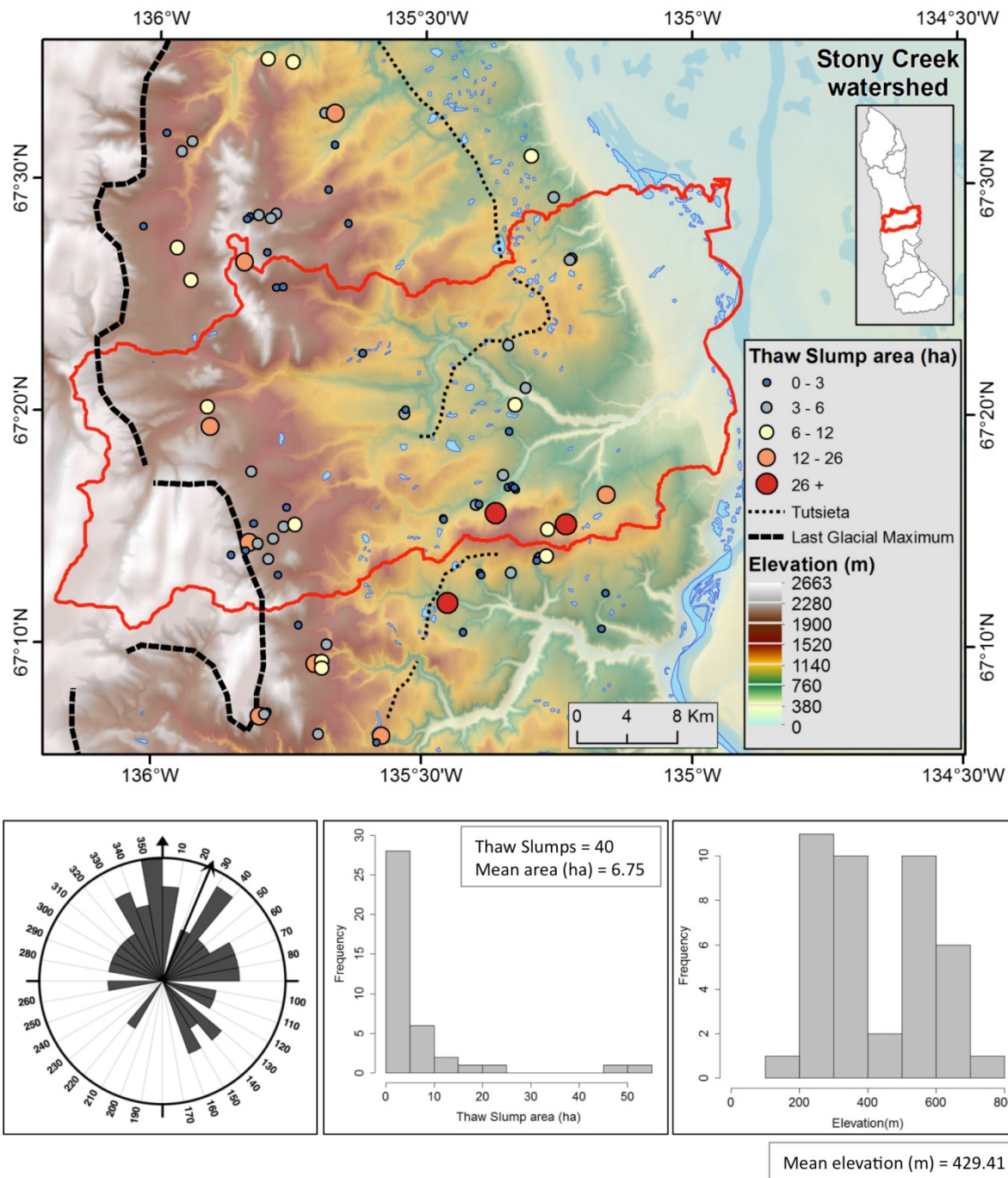


Figure 25. Distribution of thaw slumps in the Stony Creek watershed

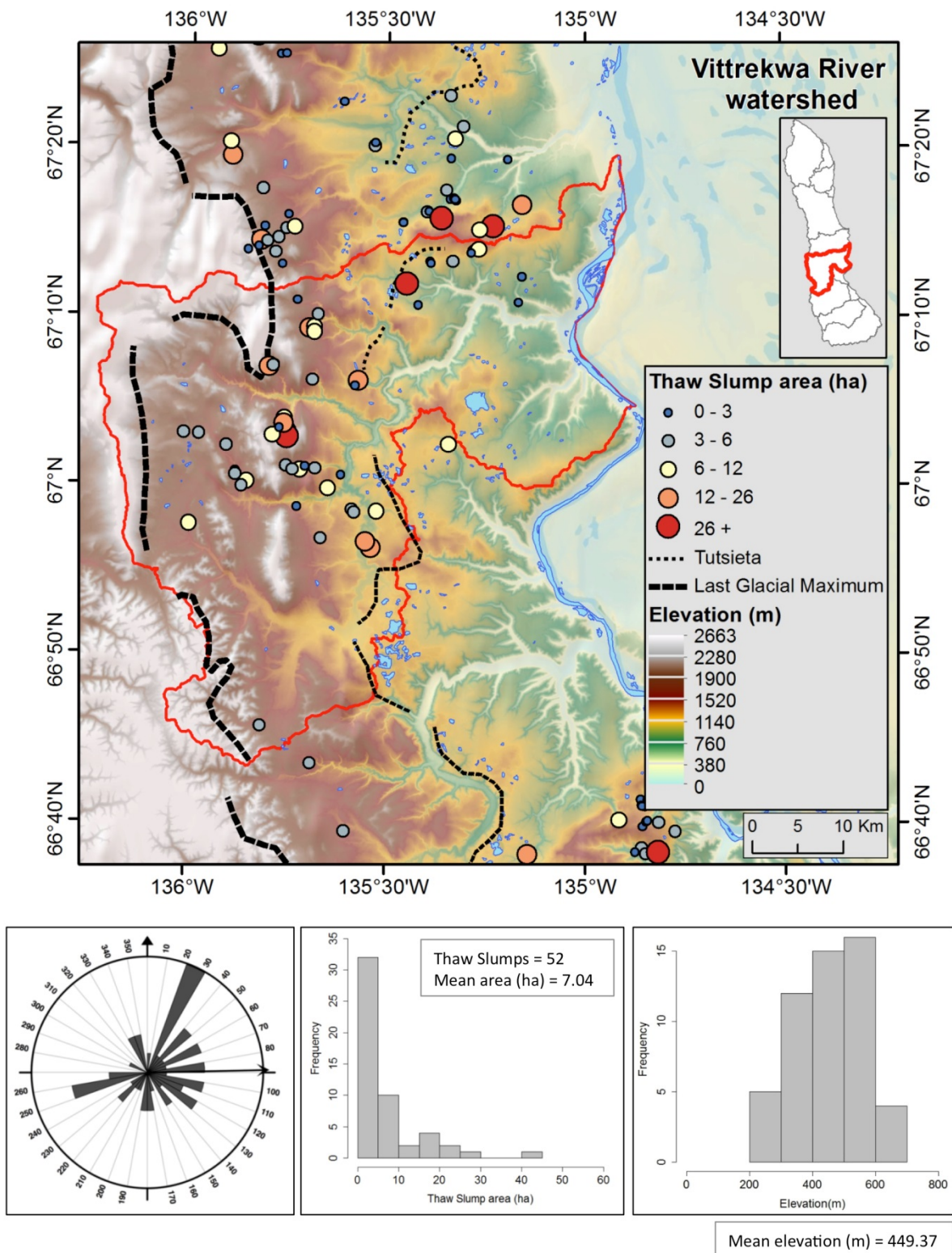


Figure 26. Distribution of thaw slumps in the Vittrekwa River watershed

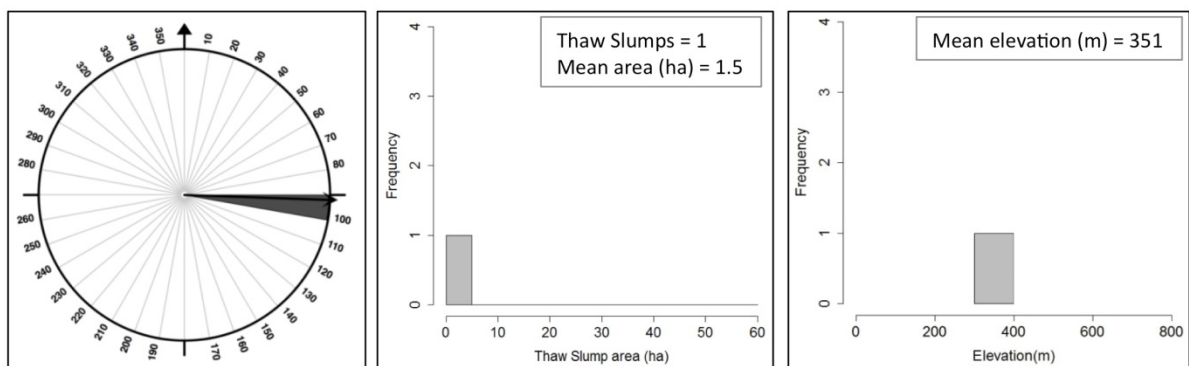
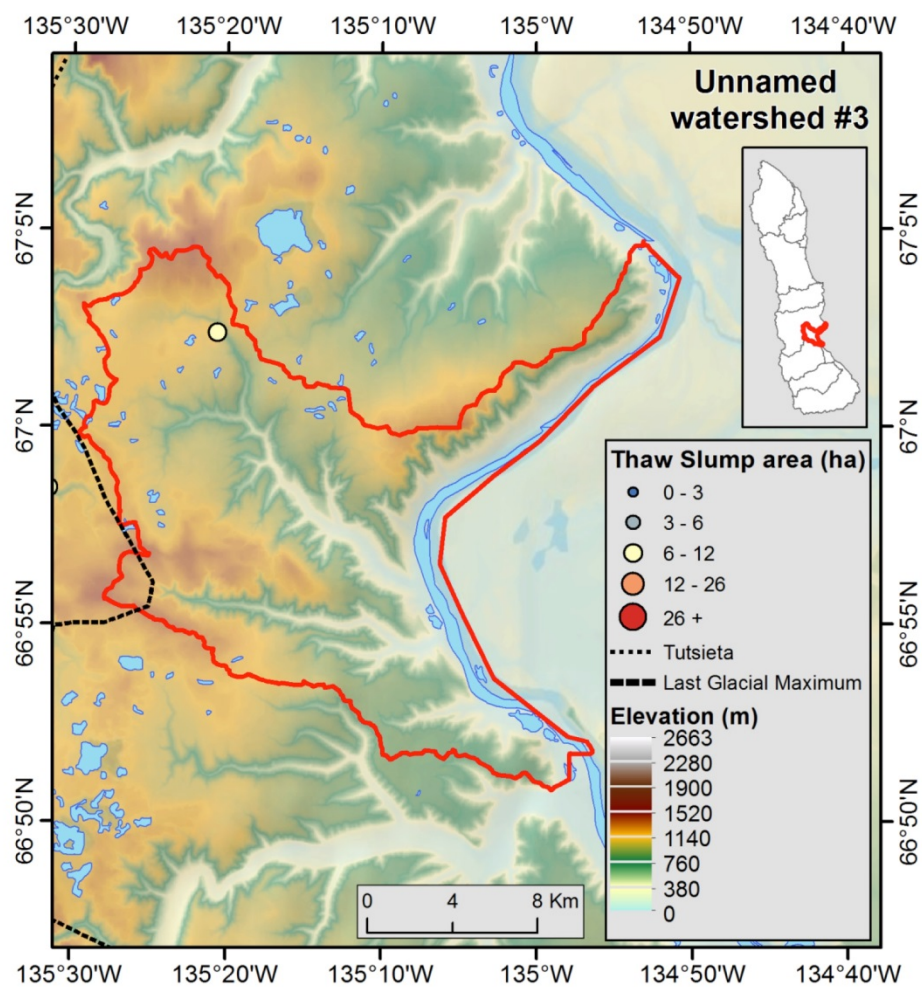


Figure 27. Distribution of thaw slumps in the Unnamed Watershed #3

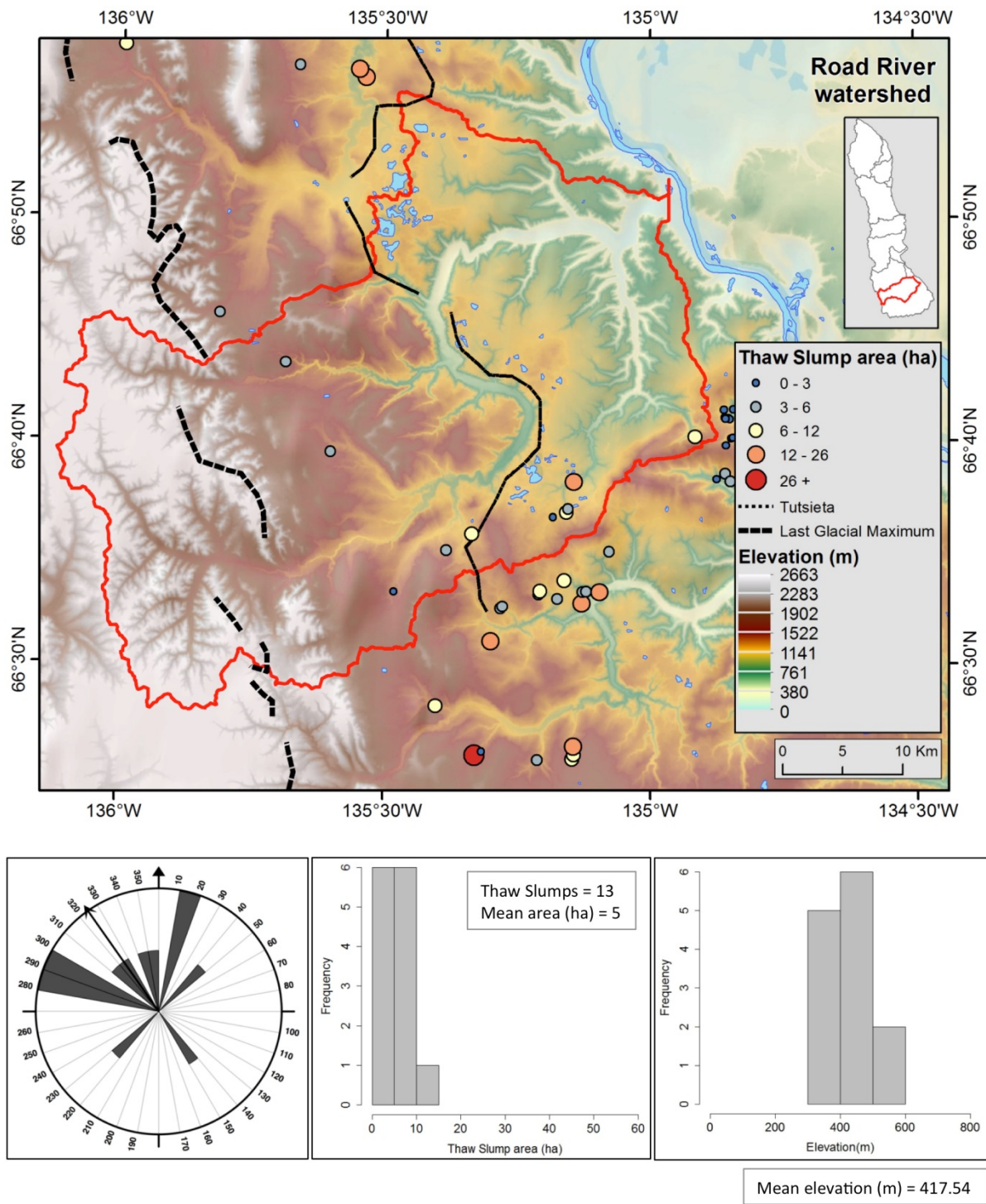


Figure 28. Distribution of thaw slumps in the Road River watershed

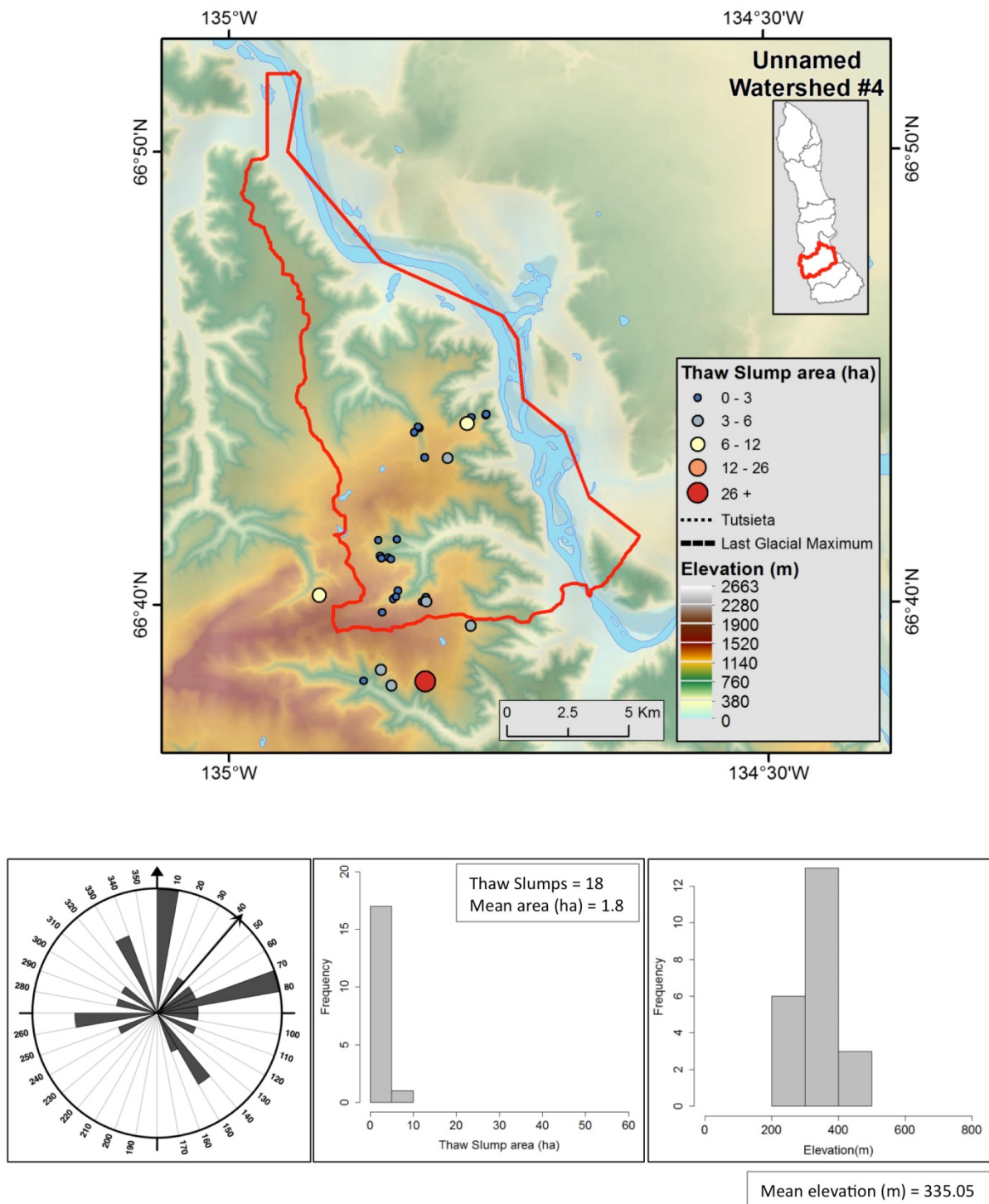


Figure 29. Distribution of thaw slumps in the Unnamed Watershed #4

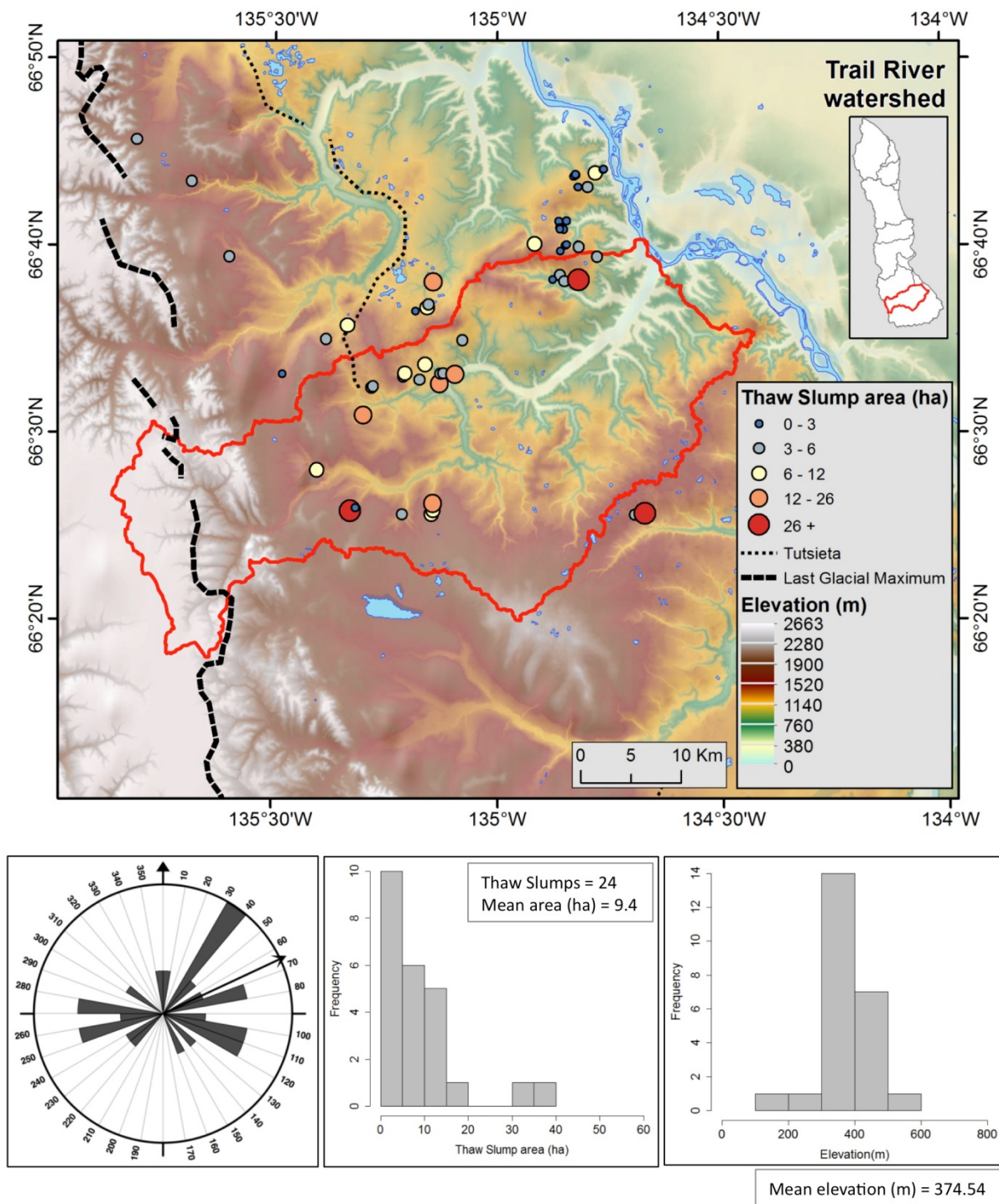


Figure 30. Distribution of thaw slumps in the Trail River watershed

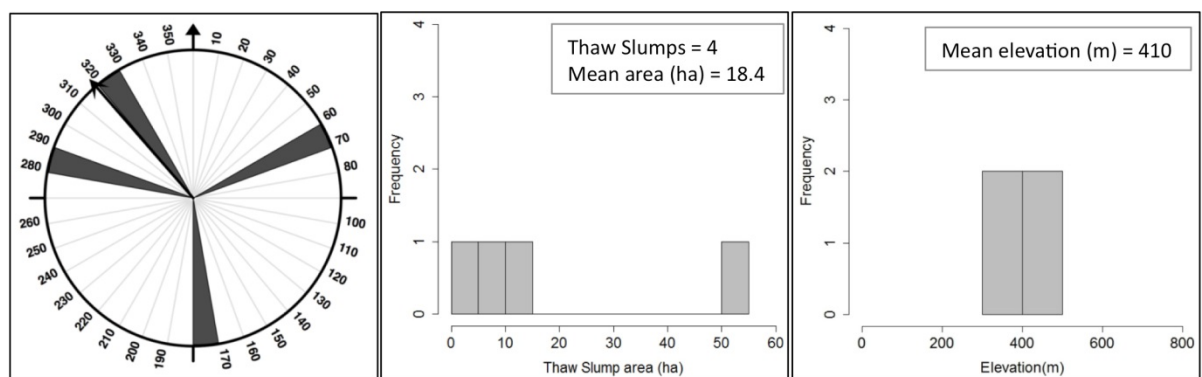
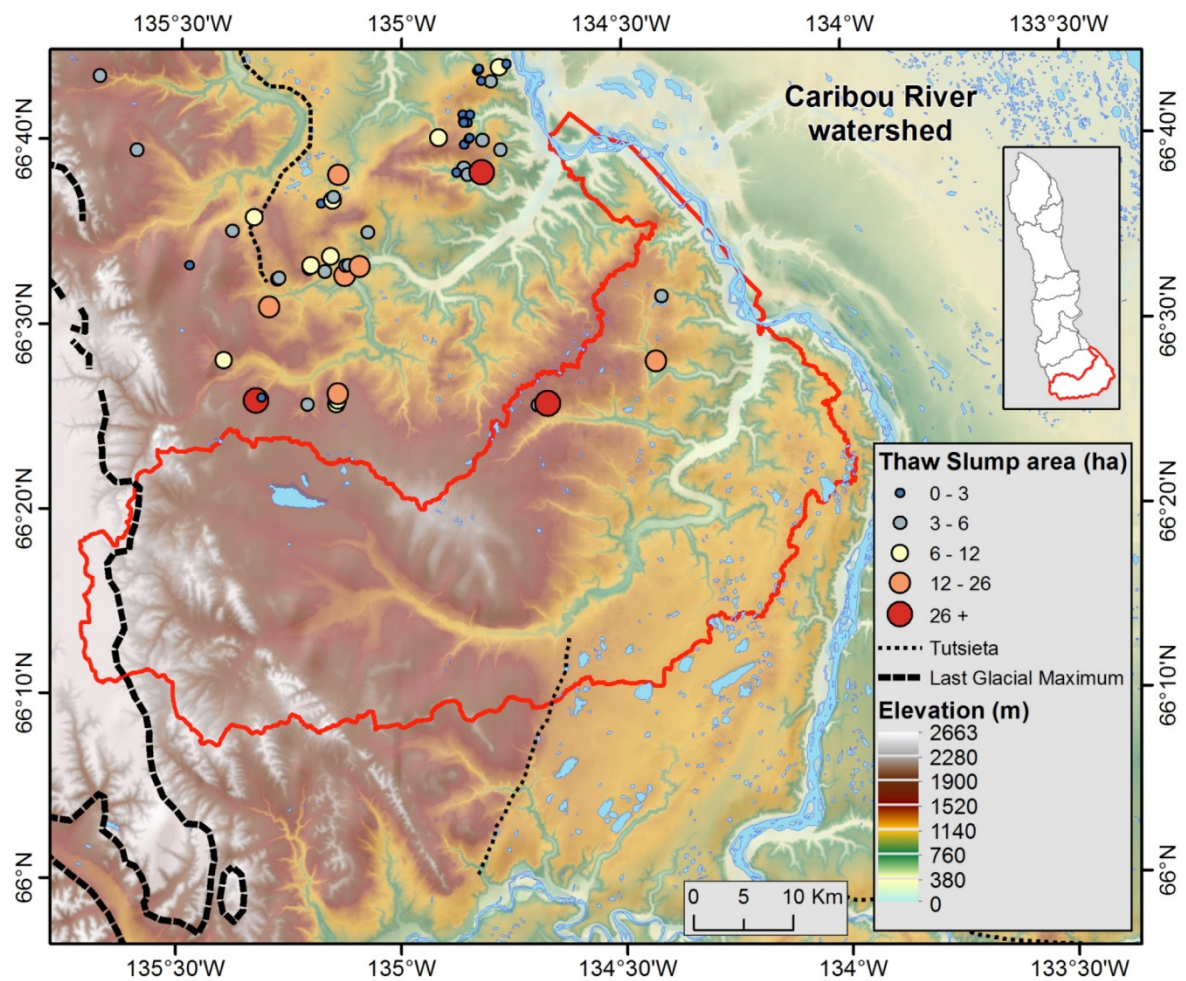


Figure 31. Distribution of thaw slumps in the Caribou River watershed

5.3. Annual Tasseled Cap trends

The Landsat Image Stack Trend Analysis method calculates three Tasseled Cap transformations (brightness, greenness and wetness) to a Landsat image stack, with images acquired between 1985 and 2011. The extraction of an index value, greenness for example, for the same pixel through each image in a dense stack of Landsat images offers the ability to monitor changes in surface cover at that location over time. As such, the application of the annual Tasseled Cap trend approach to a transect of adjoining pixels would be able to detect both temporal and spatial change in surface cover. Given the ability of the greenness index at representing vegetation, and a change in surface cover, this section will present the annual Tasseled Cap greenness values from 16 thaw slumps (figure 32) and verify if this approach is able to identify the growth of thaw slumps over time, from initiation to stabilization. These 16 thaw slumps were selected as being the best examples for the application of the Landsat Image Stack Trend Analysis approach. These thaw slumps were observed to be growing in a uniform direction and did not show a pattern of polyclicity (cycle of initiation and stabilization), which would have shown more complicated greenness patterns.

Three to four thaw slumps were selected for each watershed located within the Landsat WRS 13 row (Figure 12) as it contained the greatest number of cloud-free images, which allowed for a higher temporal resolution (denser image stack) of Landsat images (Appendix A). Table 9 shows the surface area and orientation of the 16 thaw slumps for which the greenness values of adjoining pixels was extracted. The sampled thaw slumps range in surface area from 5.2 ha to 43.4 ha, and are located on a wide range of slope orientations.

Following the analysis of greenness pixel values, undisturbed tundra surrounding thaw slumps has greenness values ranging between 110 and 150. By contrast, the greenness values decrease to near 20 on active slump floors where the ground is wet. Therefore, in the case of a growing thaw slump, the temporal pixel trends may represent the change in surface from unaffected tundra (with

greenness values near 130), to wet slump floor (with greenness values near 20). In the case that a thaw slump would stabilize, the greenness value would increase toward those of tundra values. Below, we examine if this pattern is observed for the 16 selected thaw slumps. To this end, a transect of 8 to 18 pixels was defined for each slumps, extending from the pixel above the position of the headwall in 2011 and extending into the slump floor. In addition, 1 or 2 pixels from the adjacent tundra were selected as control points.

Table 9. *Slump ID, watershed name, surface area, orientation and figure number for all sampled thaw slumps.*

Slump ID	Watershed	Surface area of slump (ha)	Orientation of slump (°)	Figure #
1	Rat River	7.8	88	33
2	Rat River	9.1	307	34
3	Rat River	10.7	48	35
4 (Charas)	Stony Creek	48	323	36
5 (Melanie)	Stony Creek	9.6	33	37
6	Stony Creek	20	38	38
7	Vittrekwa River	43.4	62	39
8 (Dover)	Vittrekwa River	20.6	70	40
9	Vittrekwa River	18	169	41
10	Vittrekwa River	11.4	221	42
11	Road River	13.2	142	43
12	Road River	9.2	342	44
13	Road River	8.6	299	45
14	Trail River	6.8	104	46
15	Trail River	5.2	105	47
16	Trail River	11.7	37	48

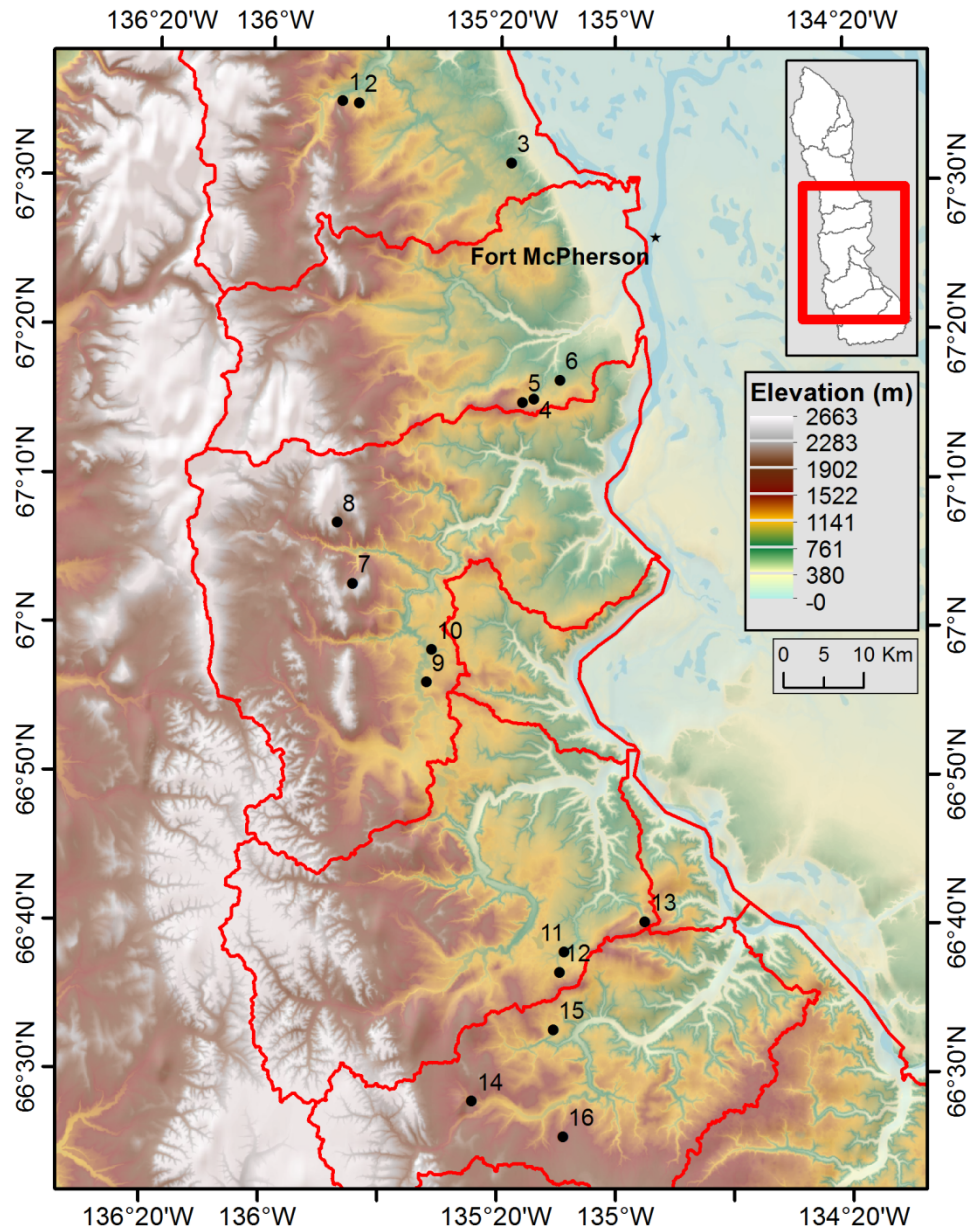


Figure 32. Location of sampled slumps within the Richardson Mountains

Pixel values for thaw slump #1 (Figure 33) shows the successive decrease in greenness values from values around 140 to values around 20. This decrease in greenness value represents the recession of the headwall through the pixels and the change in ground type from tundra to wet mud. Each successive decrease in greenness value is interpreted to represent 30m of headwall recession. As can be

seen in figure 33, greenness values in pixels 1 and 2 do not remain low after reaching their lowest value, but increase to values over 100 by 2004. This represents the drying out and subsequent re-vegetation of the ground in these pixels. By 2010, most pixels in this slump had undergone significant re-vegetation, with greenness values ranging from 72 to 101.

Thaw slump #2 (Figure 34) shows the successive decrease in greenness values from values around 140 to around 25. Unlike thaw slump #1, pixels in this slump do not remain wet for very long after the headwall recedes through the pixel. Instead, the ground begins to immediately dry out and revegetate, as can be seen by the immediate increase in greenness values after reaching its lowest greenness level.

The recession of the headwall can also be noted with thaw slump #3 (Figure 35) by the successive decrease in greenness values. This slump began to quickly stabilize and re-vegetate in 2005, as can be noted by the sudden increase in greenness values.

Figure 36, shows the trends of thaw slump #4, or “Charas Slump”. This figure shows the successive decrease in greenness values as the headwall recedes through the pixels. The slump floor of this slump remains wet following the recession of the headwall, limiting the growth of early successional vegetation, as pixel greenness values do not increase much after reaching their lowest value.

Figure 37 shows the greenness pixel values of thaw slump #5, or “Melanie Slump”. The recession of the headwall can be seen by the decrease in greenness values of 7 pixels since 1998. Only pixel 1 has shown significant amount of re-vegetation upon reaching its lowest value, while pixels 2 to 7 remaining wet ground.

The successive decrease in greenness values of 7 pixels since 1988 in thaw slump #6 can be seen figure 38, representing the recession of the headwall through these pixels. The slump floor of this thaw slump only began to slowly re-vegetate in 2005, as can be seen by the increase in pixel values approaching zero in 2005 to values around 25 in 2010.

The recession of the headwall through 17 pixels in thaw slump #7 since 1988 can be seen in figure 39. As can be seen by the number of pixels that have decreased in greenness value since 1988, this is one of the fastest growing thaw slump in the Richardson Mountains. Ground in this thaw slump does not remain wet and barren for very long after the recession of the headwall through the pixel, as can be seen by the strong increases in greenness values upon reaching their lowest greenness value. Since 2000, the rate of re-vegetation in most pixels appears to have increased.

Figure 40 shows the decrease in greenness values of 11 pixels in thaw slump #8 or “Dover slump” since 1995, representing the recession of the headwall through these pixels. The slump floor of “Dover” remains very wet and free of vegetation following the recession of the headwall through that pixel, with only minor increases in greenness values from 2005 to 2010, from values approaching zero to values between 25 and 50.

The recession of the headwall through 10 pixels in thaw slump #9 since 1988 can be seen in figure 41, as shown by the successive decrease in greenness values from values around 140 to values of around 10. The ground in pixels 1 and 2 quickly dried out and re-vegetated after the recession of the headwall through these pixels. This can be seen by the steady increase in greenness values from 15 in 1993 to 94 in 2004 for pixel 1 and from 9 in 1995 for pixel 2 to 60 in 2005. It also appears that pixels 3 to 7 have begun to re-vegetate beginning in 2005.

A more complicated pattern of pixel greenness is shown in thaw slump #10, (figure 42). Given the large range in pixel greenness values, there was likely also a large range in ground type that was affected by the growth of this thaw slump, possibly ranging from spruce forest to shrubby tundra. This figure shows the recession of the headwall through pixels 3 to 10, as demonstrated by the successive decrease in greenness values. Ground in this thaw slump does not remain wet and vegetation-free long after the recession of the headwall through the pixel, as can be

noted by the quick increase in greenness values upon reaching its lowest greenness value.

The initial greenness trend in thaw slump #11, shown in figure 43 differs from all other slump greenness trends. This figure shows the increase in greenness from values near zero in 1988 to values of around 100 in 1996. This trend is likely caused by the recovery and re-vegetation of the ground following a recent forest fire. In 1996, the headwall begins to recede through pixel 1 and the recession of the headwall through six pixels from 1996 to 2011 can be seen in this figure. Aside from pixel 2 which underwent a temporary increase in greenness values in 2009, all pixels affected by the recession of the headwall have remained wet and vegetation-free.

Figure 44 shows the successive decrease in greenness values of thaw slump #12 located 2.6 km south of thaw slump #11. This figure shows the recession of the headwall through 7 pixels from 1994 to 2011. In general, the slump floor remains wet and free of vegetation following the recession of the headwall through the pixel. There have only been temporary increases in vegetation in 2007 for pixels 1 and 2 and 2010 for pixel 1.

Figure 45 shows the recession of the headwall through 6 pixels from 1988 to 2011 in thaw slump #13. The notable aspect of this slump is the variability in recovery and re-vegetation of individual pixels after the recession of the headwall. For example, in 2010, the range in greenness values of pixels that have already been affected by the recession of the headwall ranged from 20 to 58. This suggests a patchy and heterogeneous pattern of ground recovery.

Figure 46 shows the expansion of the headwall through 7 pixels in thaw slump #14 from 1988 to 2011. It is worth noting that the time interval between pixel greenness decrease has increased over time, suggesting a slowing of the growth rate of this thaw slump. Until 2000, these pixels were remaining relatively wet and free of vegetation, at which point all disturbed pixels began to increase in greenness values, reaching values of nearly 100 by 2006. Given the increase in time interval

between greenness drops and the overall trend of increasing greenness, this thaw slump appears to be in the process of stabilization.

Thaw slump #15, shown in figure 44, shows a similar greenness pattern to that of figure 47. Figure 41 shows the decrease in greenness values of 6 pixels from 1988 to 2011. The time span between greenness value decreases has increased between the years of 1988 and 2011. In addition, pixels already affected by the growth of the headwall have begun to increase dramatically in greenness values since 2000 to values close to those of unaffected tundra.

Figure 48 shows the expansion of thaw slump #16 through 10 pixels from 1988 and 2011. This figure shows the successive re-vegetation of pixels affected by the expansion of the headwall, with 2011 greenness values of pixels affected earlier being higher than pixels affected later by the expansion of the headwall.

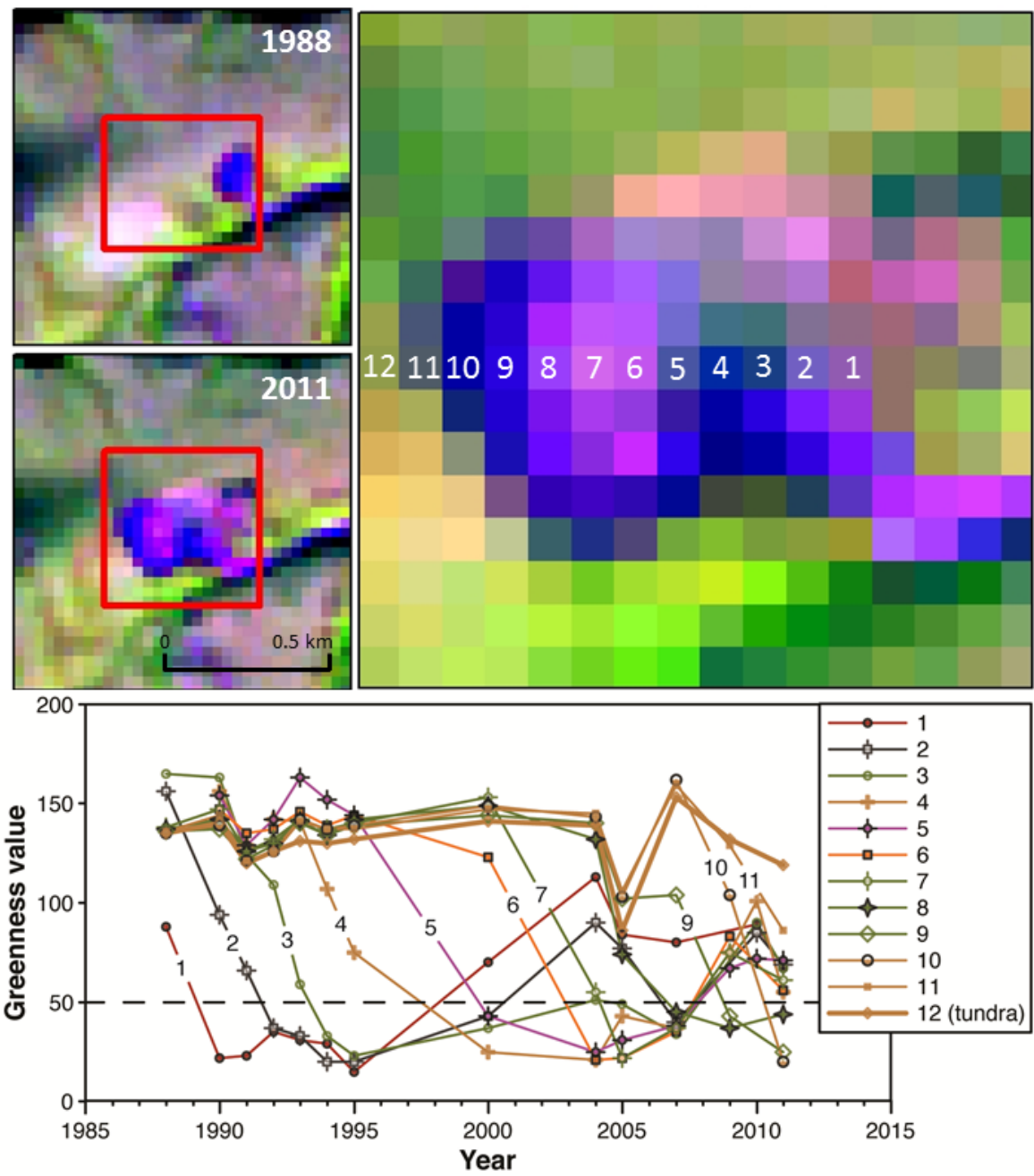


Figure 33. Annual greenness trends for thaw slump #1 in the Rat River watershed (1/3).

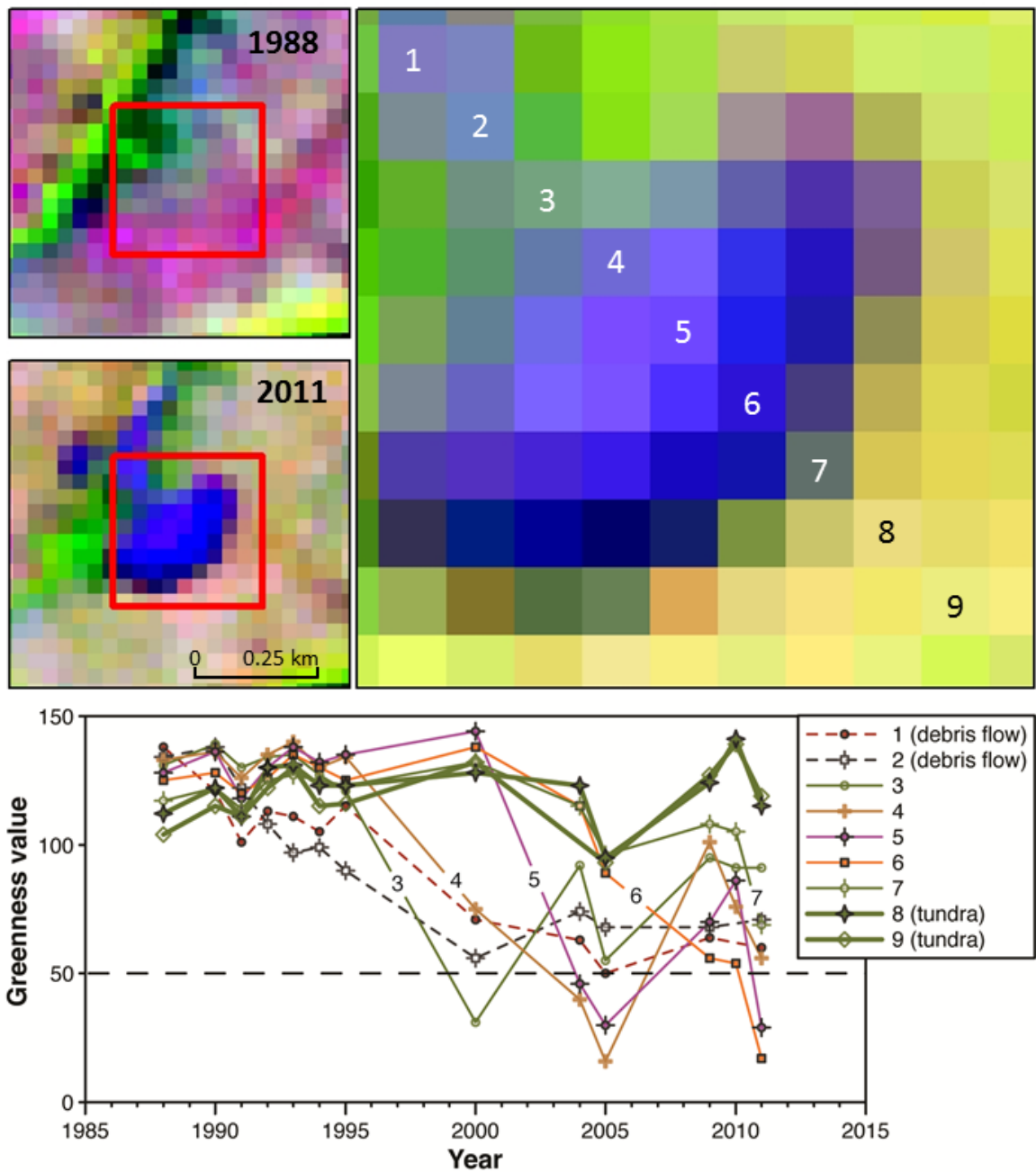


Figure 34. Annual Greenness trends for thaw slump #2 in the Rat River watershed (2/3).

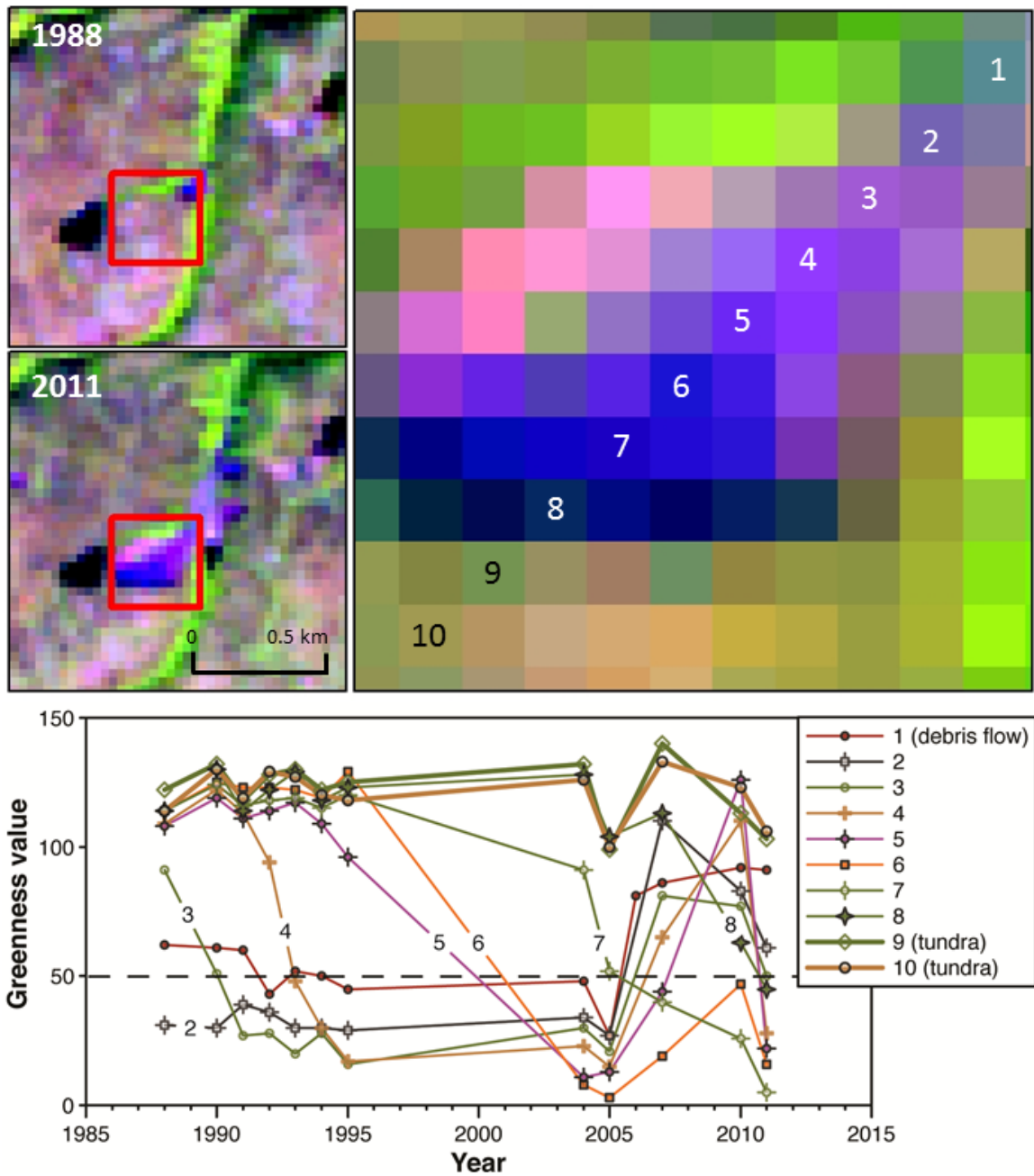


Figure 35. Annual Greenness trends for thaw slump #3 in the Rat River watershed (3/3).

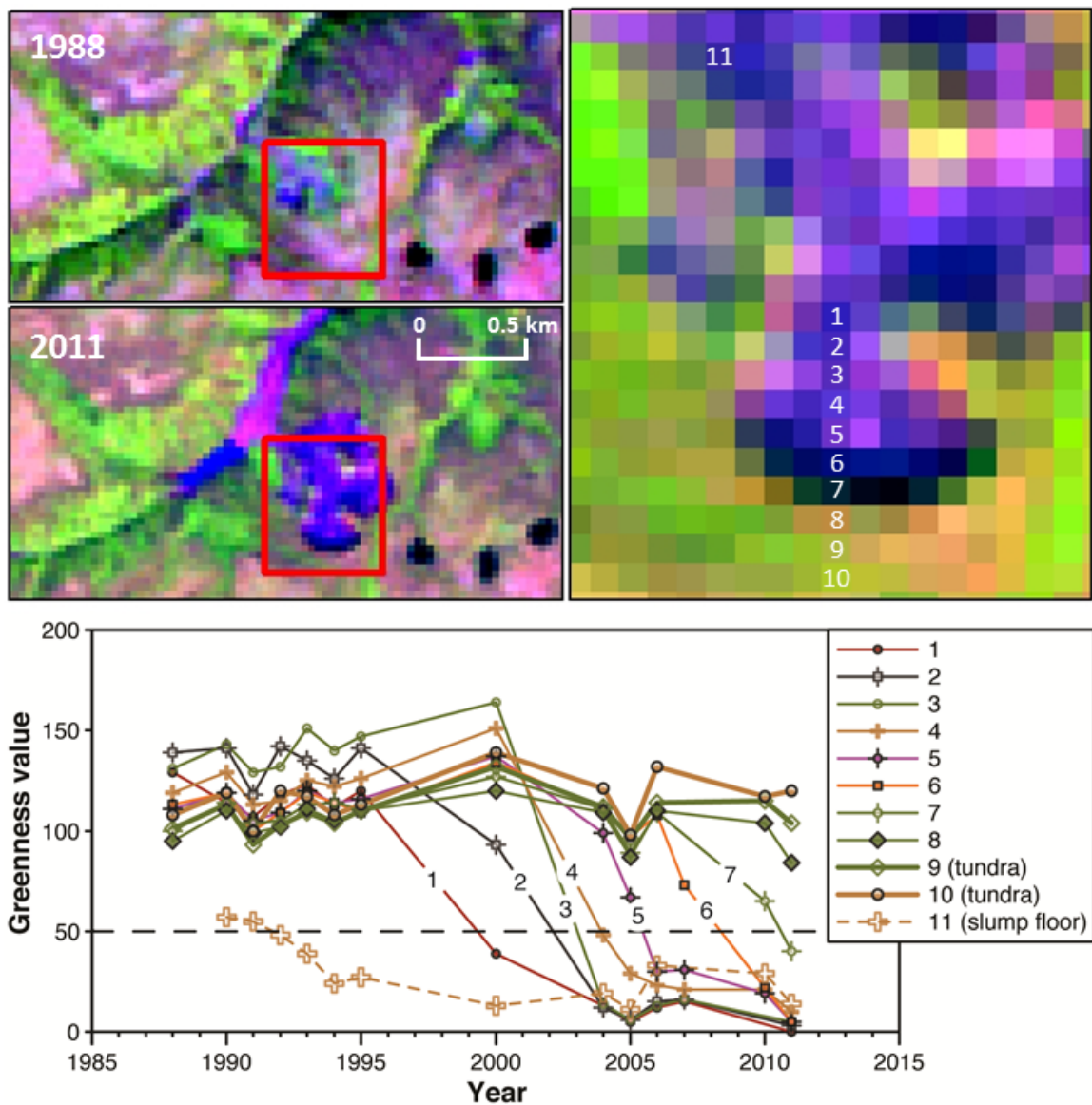


Figure 36. Annual Greenness trends for #4 "Charas" thaw slump in the Stony Creek watershed (1/3).

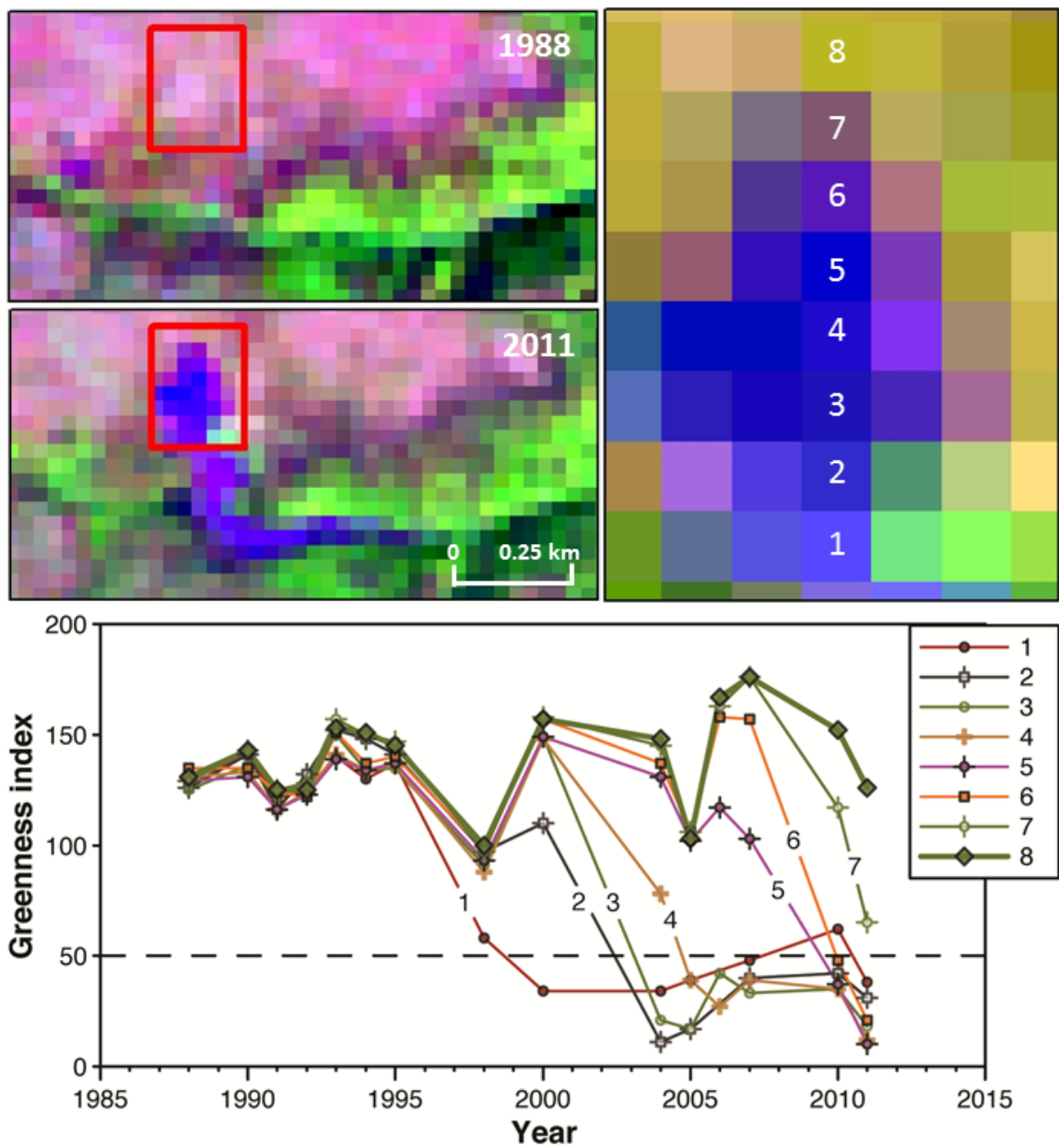


Figure 37. Annual Greenness trends for #5 "Melanie" thaw slump in the Stony Creek watershed (2/3).

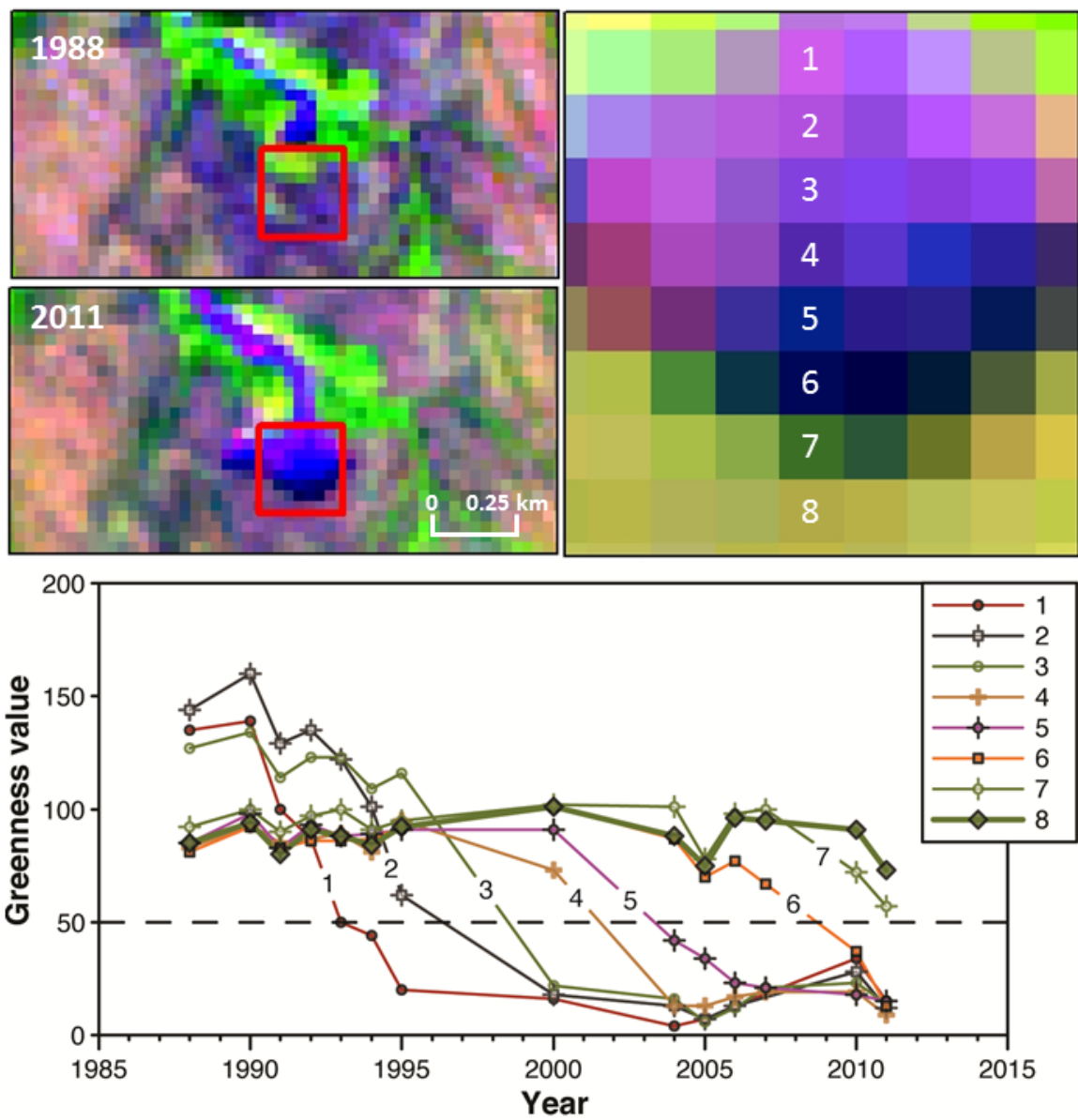


Figure 38. Annual Greenness trends for thaw slump #6 in the Stony Creek watershed (3/3).

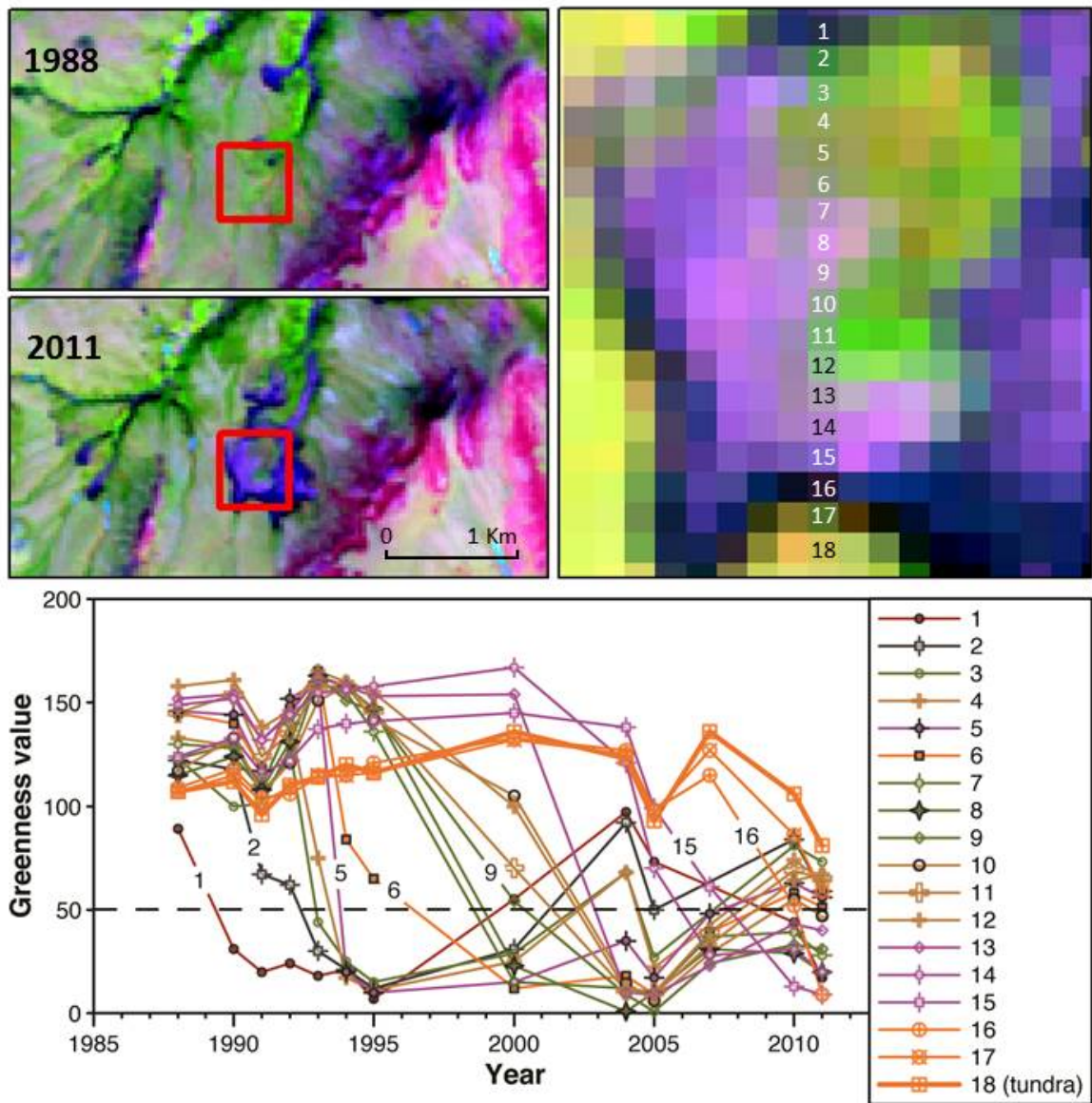


Figure 39. Annual Greenness trends for thaw slump # 7 in the Vittrekwa River watershed (1/4).

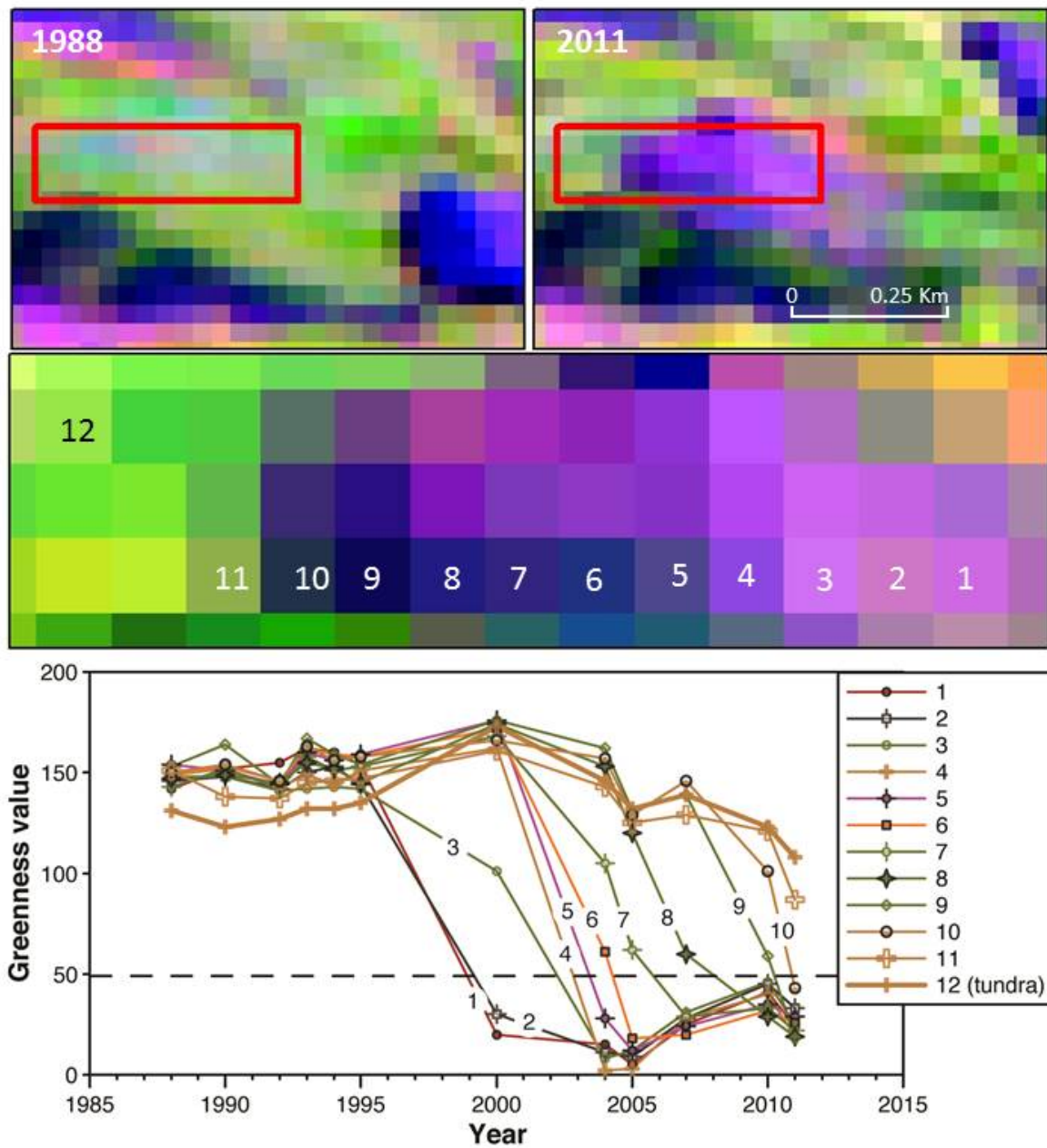


Figure 40. Annual Greenness trends for thaw slump #8 "Dover" in the Vittrekwa River watershed (2/4).

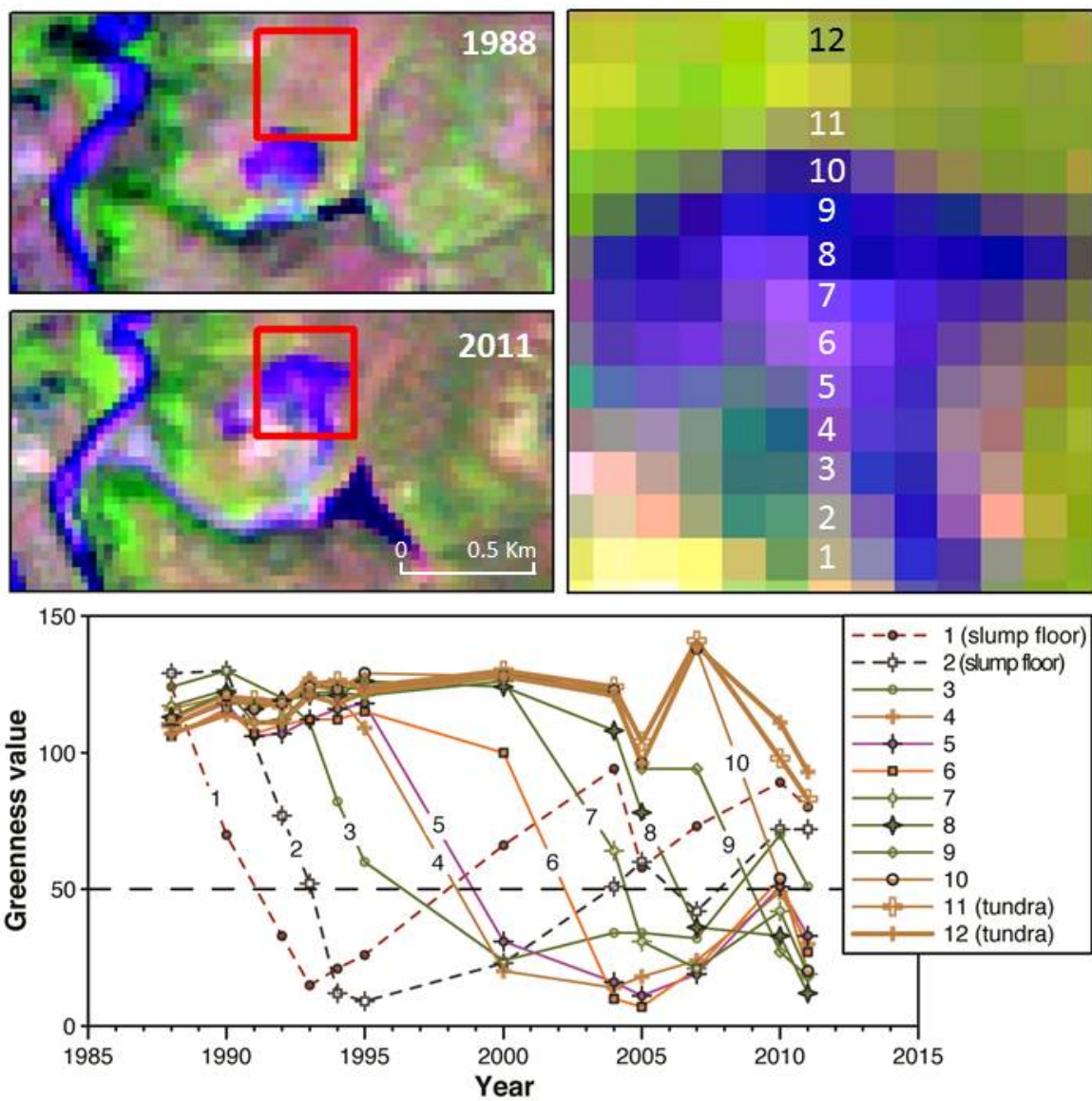


Figure 41. Annual Greenness trends for thaw slump #9 in the Vittrekwa River watershed (3/4).

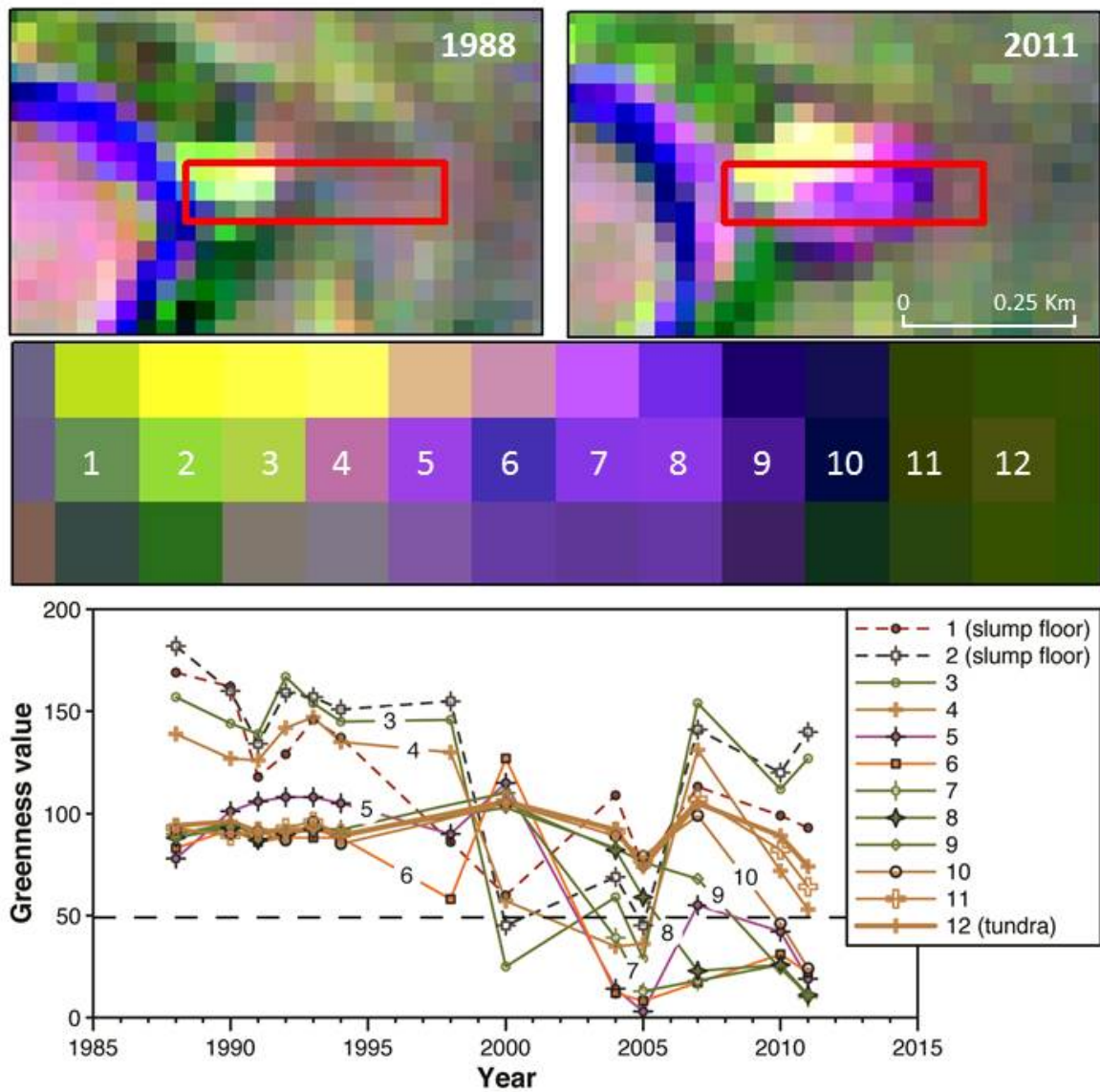


Figure 42. Annual Greenness trends for thaw slump #10 in the Vittrekwa River watershed (4/4).

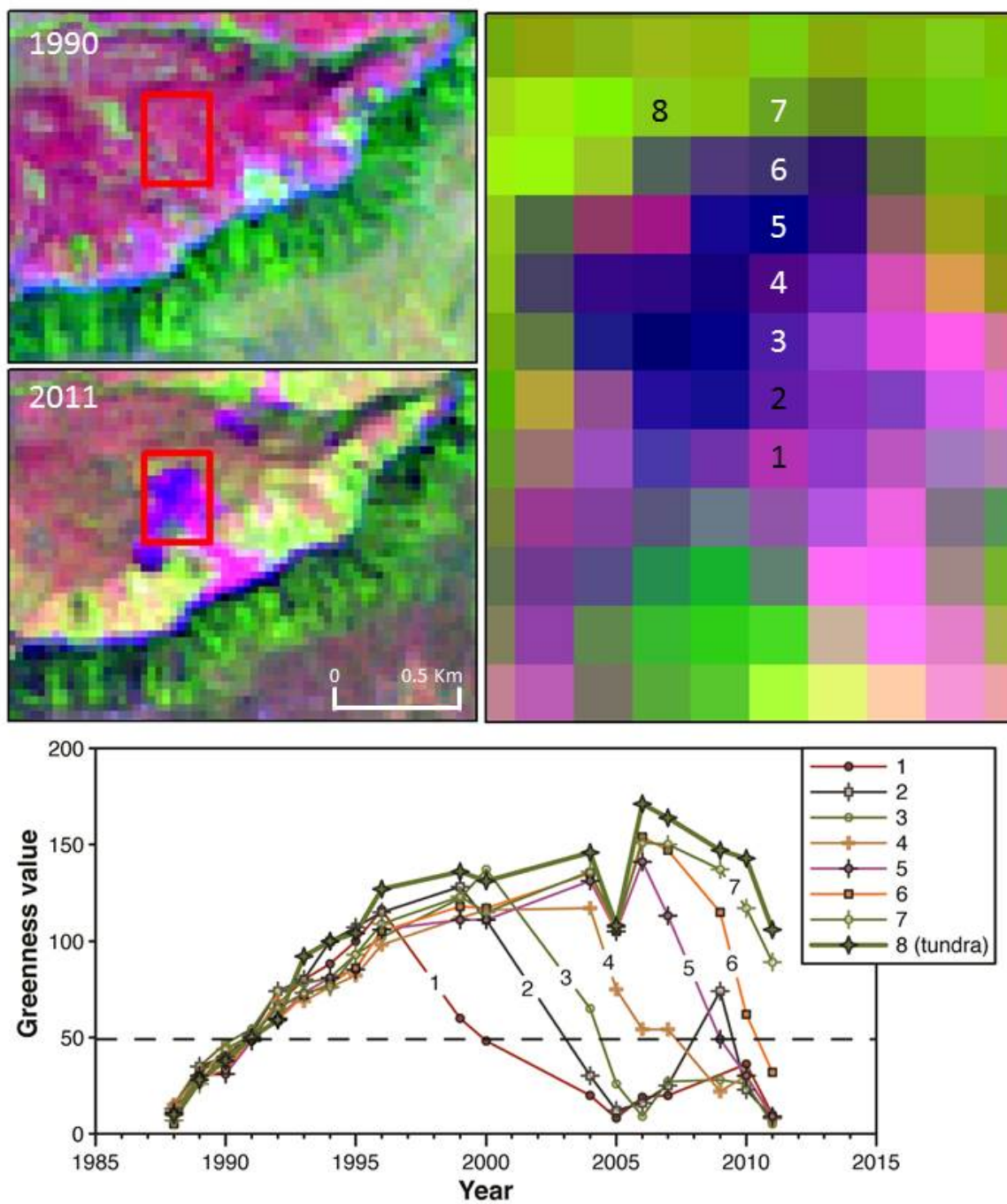


Figure 43. Annual Greenness trends for thaw slump #11 in the Road River watershed (1/3).

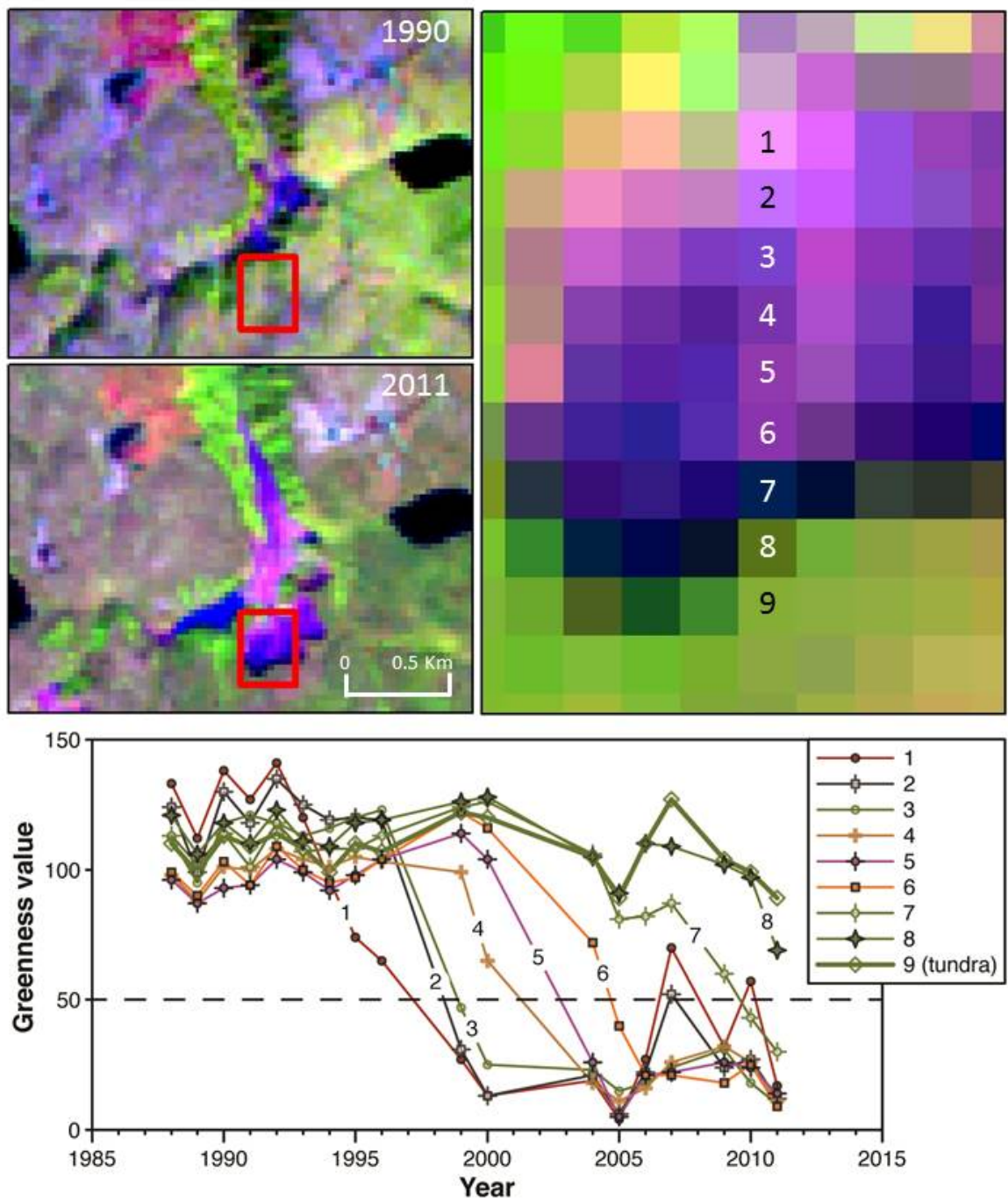


Figure 44. Annual Greenness trends for thaw slump #12 in the Road River watershed (2/3).

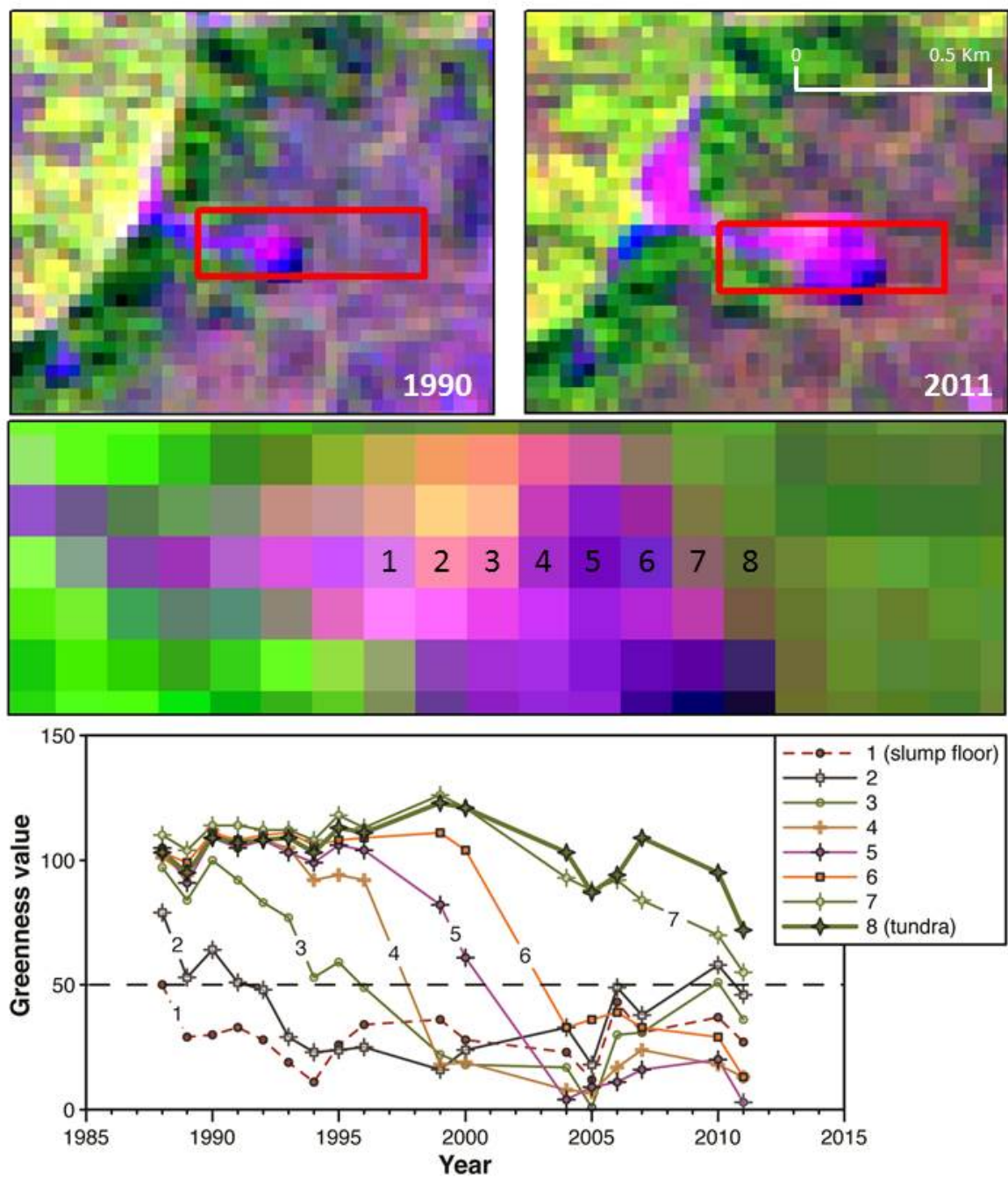


Figure 45. Annual Greenness trends for thaw slump #13 in the Road River watershed (3/3).

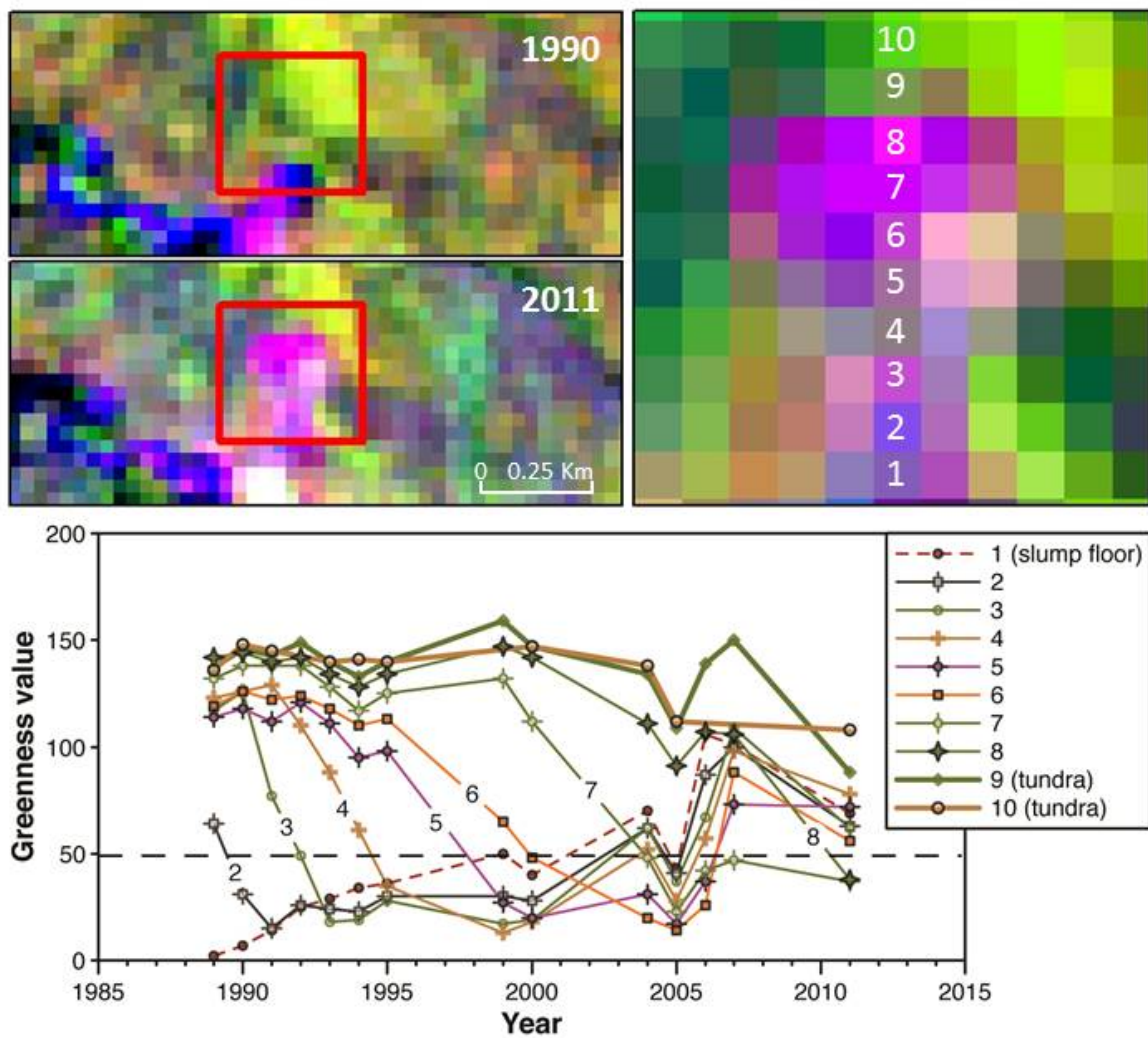


Figure 46. Annual Greenness trends for thaw slump #14 in the Trail River watershed (1/3).

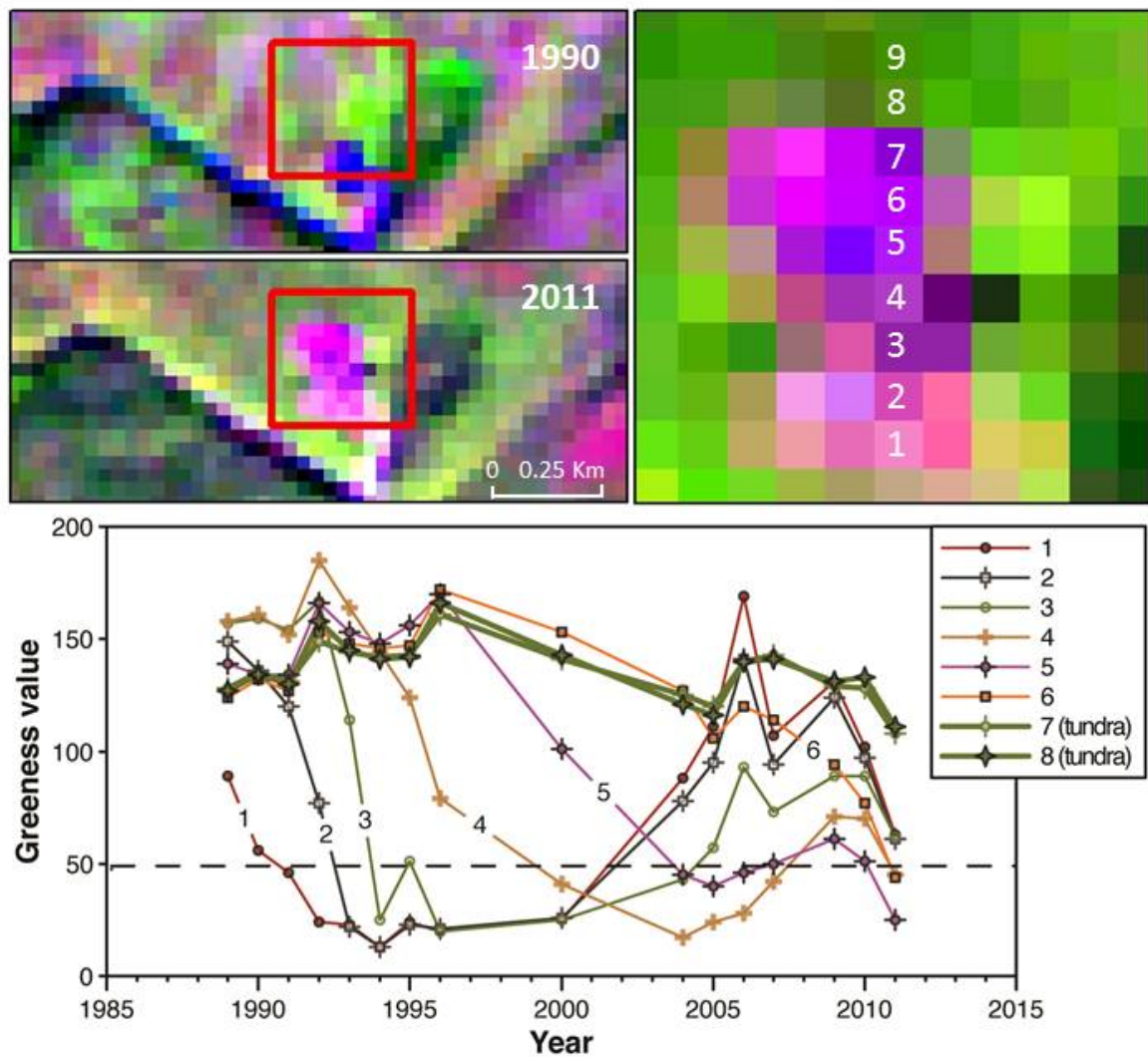


Figure 47. Annual Greenness trends for thaw slump #15 in the Trail River watershed (2/3).

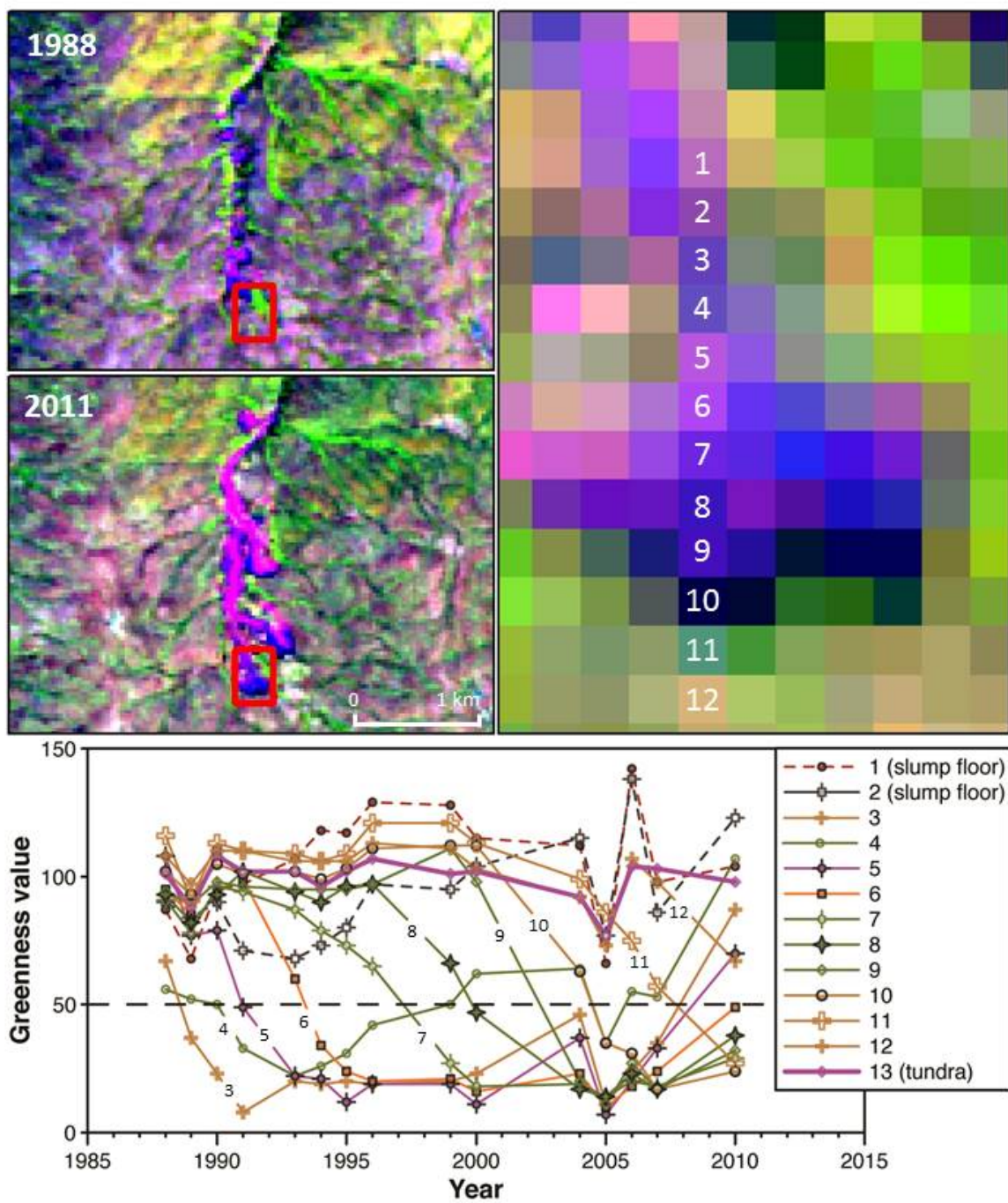


Figure 48. Annual Greenness trends for thaw slump #16 in the Trail River watershed (3/3).

6. Discussion

6.1. Distribution of thaw slumps in the Richardson Mountains

Retrogressive thaw slumps are amongst the most dramatic form of permafrost degradation in permafrost regions (Burn and Lewkowicz, 1990). Some of the highest densities of thaw slumps occur in the ice-rich permafrost terrain of the Mackenzie Delta region and Yukon Coastal Plain (Mackay, 1963; Lantz and Kokelj, 2008, Lantuit and Pollard, 2008). This ice-rich landscape is largely the result of late Pleistocene glaciation, with the origin of massive beds of ground ice tens of meters thick in the glaciated western Canadian Arctic attributed to either the burial of glacial ice during the late Pleistocene (Murton, 2005), or freezing of glacial meltwater in sediments following the retreat of the Laurentide Ice Sheet (Mackay, 1971; Rampton, 1988; Lacelle et al., 2004). This suggests that in this region the distribution of ground ice related to the late Pleistocene glaciation is likely the major influence on the nature of thermokast processes. This section examines the potential controls on the distribution and orientation of thaw slumps in the Richardson Mountains, NWT (Figure 18). These include relations with: i) limits of the Laurentide Ice Sheet; ii) ground ice content and terrain units; and iii) topography of the landscape.

Figure 18 shows the spatial distribution of thaw slumps in the Richardson Mountains in relation to the last glacial maximum limit of the Laurentide Ice Sheet (reached at 18ka cal yr BP; Lacelle et al., 2013) and its recessional position during the Tuetsia Lake phase (dated to 15ka cal yr BP; Duk-Rodkin and Hughes, 1995). All thaw slumps are situated within the last glacial maximum limit, and of the 251 thaw slumps identified in the Richardson Mountains, 172 (68.5%) are constrained between the limits of the last glacial maximum limit and the Tuetsia recessional

phase. However, it should be noted that a high number of slumps are also found in the Mackenzie Delta, east of the Tuetsia Lake phase (i.e., Lantz and Kokelj, 2008).

The distribution of thaw slumps in the Richardson Mountains is also closely tied to ground ice content in the region (Figure 49). The majority (69%) of the slumps are situated in areas where the ground ice content was mapped by Hegginbottom and Radburn (1992) as low to moderate (Table 10). Only 66 slumps (27%) are found in areas mapped to contain nil to low ice content. However, considering that thaw slumps are geoindicators of ice-rich permafrost (Mackay, 1963; Burn and Lewkowicz, 1990), the permafrost and ground ice content map of Hegginbottom and Radburn (1992) should most likely be modified to moderate to high ground ice content in areas where slumps are observed in the Richardson Mountains. This ground ice content map is largely used in assessing the response of permafrost to future climate change, as such having accurate information on the ground ice content in the Richardson Mountains would improve the predictions.

Thaw slumps in the Richardson Mountains are also found in three distinct terrain units (Table 11). These include: i) alluvial deposits in the form of coarse sand and gravel; ii) colluvial d; iii) alluvial deposits in the form of fine-grained sand and silt; and iv) stony clay diamicton from morainal and colluvial deposits. Most of the thaw slumps are found in fine sand to silt material and relatively few in coarse-grained material. This observation might be related to the conditions necessary to generate ice-rich permafrost of segregated-intrusive origin (i.e., French, 2007). In the western Canadian Arctic, segregated-intrusive ice is typically found within fine-grained material, such as silts and clays, that overly coarser sediments (Mackay and Dallimore, 1992). The higher abundance of thaw slumps in terrain dominated by fine-grained material is most likely related to the necessary conditions for the formation of segregated-intrusive ground ice, as documented by Lacelle et al (2004) in Willow River watershed, Richardson Mountains, NWT. The distribution of thaw slumps in relation to the surficial geology of the Richardson Mountains can be seen in figure 50.

Table 10. Distribution of permafrost by previously mapped ground ice content (Hegginbottom and Radburn, 1992; ground ice contents from Brown et al., 1998).

Extent of permafrost	Ground ice content	# of thaw slumps
Discontinuous permafrost	Low to moderate (10-20%)	106 (42%)
Continuous permafrost	Low to moderate (10-20%)	67 (27%)
Discontinuous permafrost	Nil to low (<10%)	40 (16%)
Continuous permafrost	Nil to low (<10%)	26 (11%)
Discontinuous permafrost	Highly variable	12 (5%)

Table 11. Distribution of thaw slumps by surficial geology (adapted from Côté et al., 2013).

Surficial geology type	Description	# of thaw slumps
Morainal	Glacial till (nonsorted silt, sand, and clay) with minor gravel	157 (63%)
Colluvial	Diamicton or rubble, derived from bedrock and surficial materials by a variety of colluvial and sheetwash processes	87 (35%)
Alluvial	Coarse sand and gravel with minor silt, fine sand and clayey silt	5 (2%)

*Surficial geology type not available for 2 thaw slumps

While thaw slumps in the Richardson Mountains are found on slopes of all aspects, they are found most commonly on slopes with an average aspect of 60° (north-east orientation). This observation is contradictory with the distribution of thaw slumps along shorelines of lakes in the Mackenzie Delta, which are most commonly found along south and west facing shorelines (Kokelj et al., 2009). The

preference for modern hillslope thaw slumps to initiate on north-east facing slopes might be a remnant of the early Holocene warm period, with summer temperatures ca. 3-7°C warmer than today (Clark et al., 2004; Kauffmann et al., 2004). This period experienced an increase in thermokarst activity, as evidenced by thicker active layers and the development of thaw lakes (Mackay, 1992; Burn, 1997). In mountainous region, such as in the study area, the thickness of the active layer is typically greater along south and west facing slopes due to higher amount of solar radiation received (French, 2007). Consequently, during the early Holocene warm period, the active layer thickness would have been thicker along south and west facing slopes. In ice-rich environments, the difference in active layer thickness due to slope orientation, would lead to the development of valley asymmetry, with shallower south and west facing slopes, due to the melting of the ground ice and subsidence of the land surface (i.e., French, 2007). Figure 51 shows topographic cross-sections of a few fluvial valleys in the Richardson Mountains where slumps developed along north or east facing slopes. For most of these valleys, valley asymmetry is observed. Therefore, because valley asymmetry is observed in the Richardson Mountains, the predominance of modern hillslope thaw slumps to develop along north and east facing slopes is likely due to the fact that ice-rich permafrost is found closer to the surface along these slopes, making it easier for a surface process (i.e., active layer detachment slide, forest fire) to expose the underlying ice-rich permafrost. In the Mackenzie Delta, this process would not occur; hence the predominance of modern thaw slumps to be exposed towards the west and south.

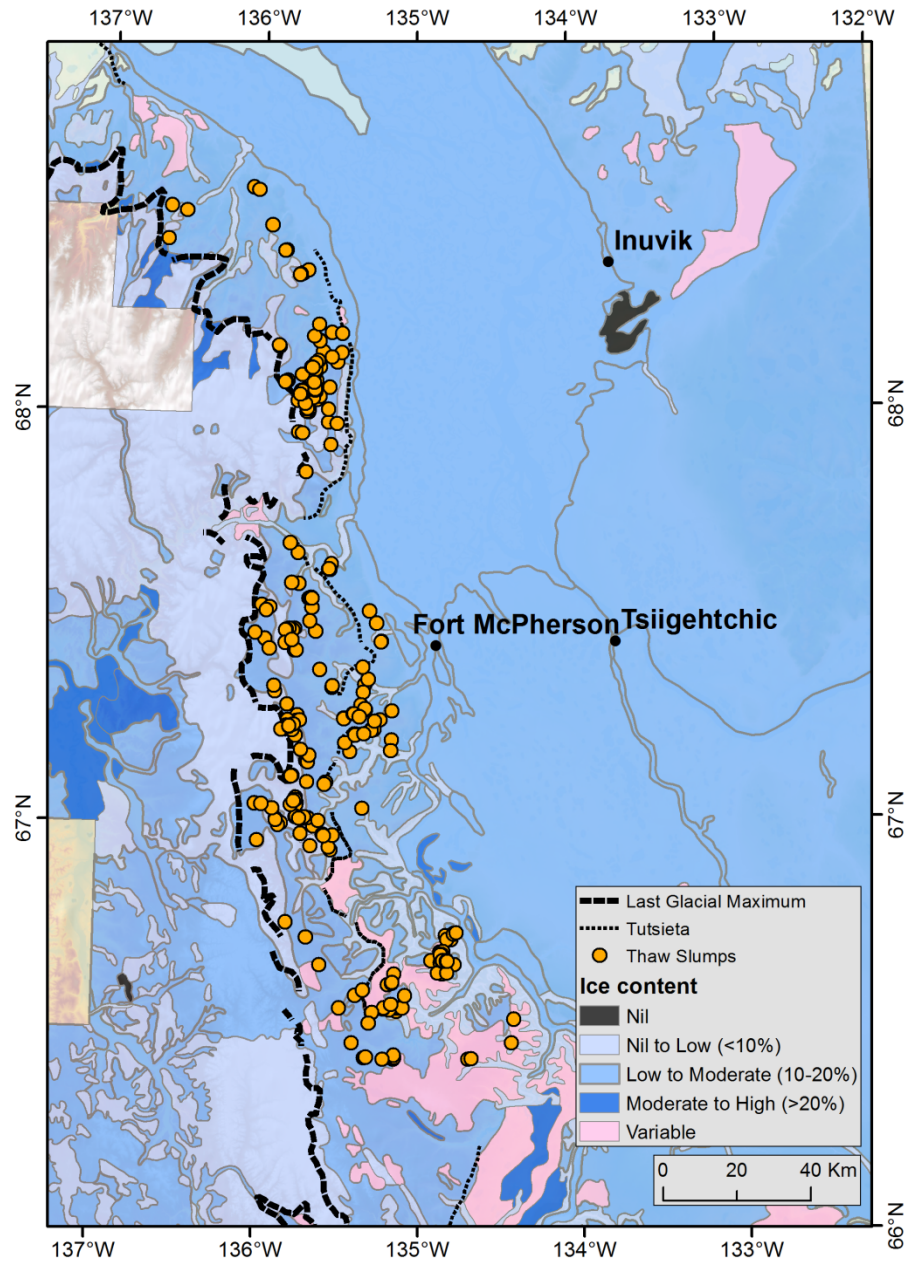


Figure 49. Thaw slump distribution in relation to the limits of the Last Glacial Maximum and the Tuetsia phase, as well as the mapped ice content of the area (Hegginbottom and Radburn, 1992).

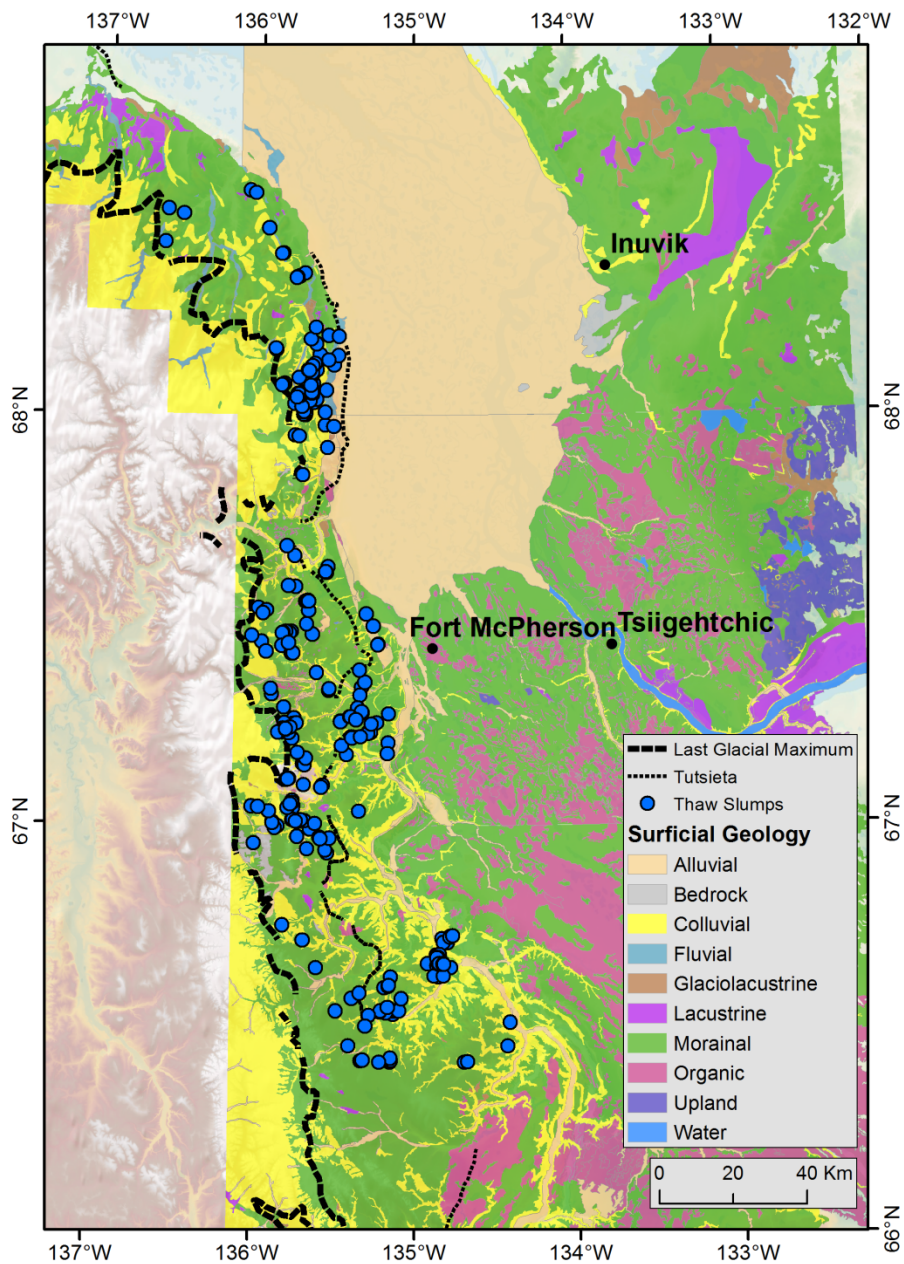


Figure 50. Thaw slump distribution in relation to the regional surficial geology (Côté et al., 2013).

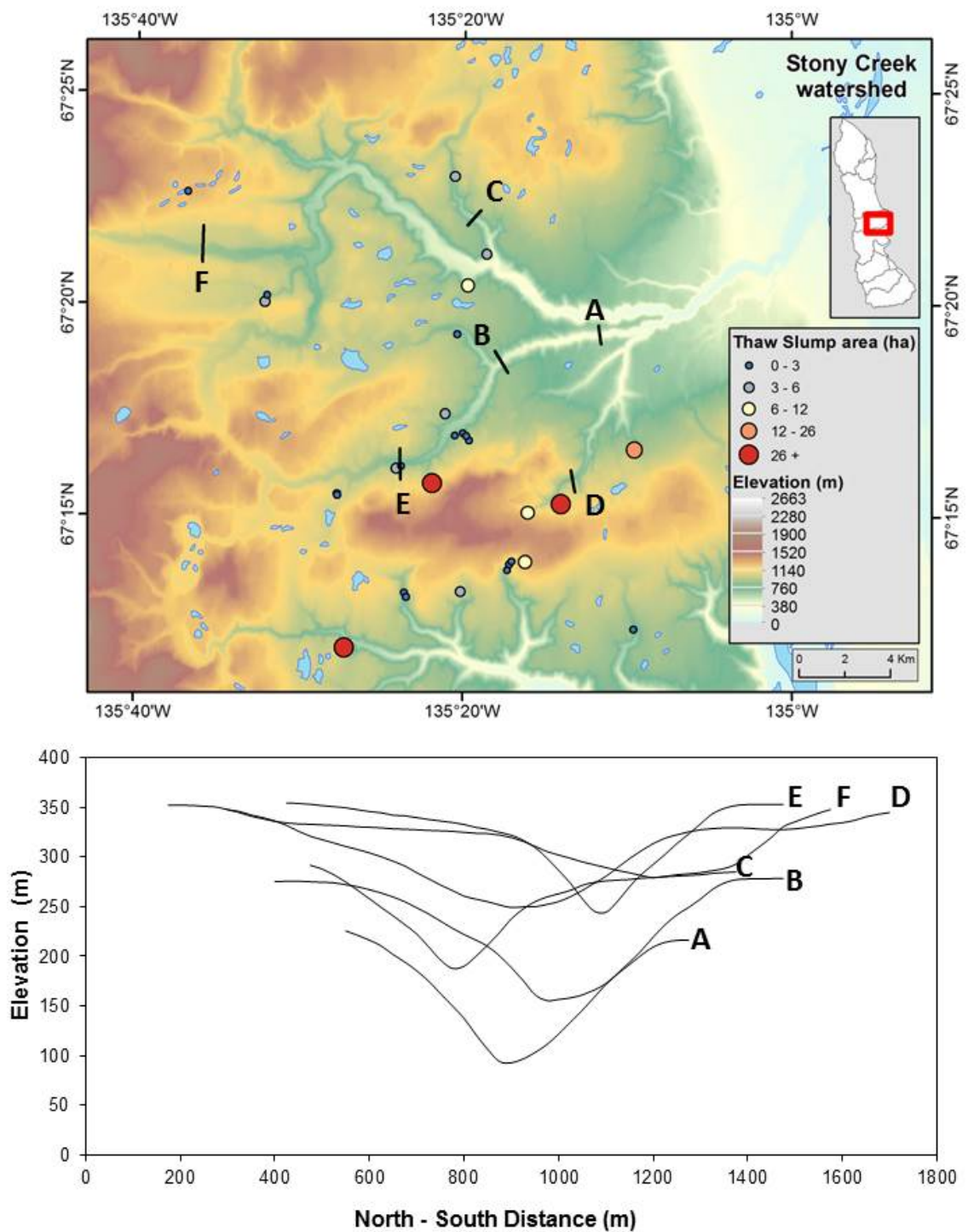


Figure 51. Topographic cross-sections along Stony Creek, Stony Creek watershed. Cross-sections A and B show canyon-type valleys, while cross-sections C, D, E and F demonstrate valley asymmetry.

6.2. Thaw Slump Life Cycle

As shown in Figures 33 to 48, the extraction of greenness values for the same pixel through each image in a dense stack of Landsat images and along a linear series of adjoining pixels offers the possibility to monitor both temporal and spatial change in surface cover. This section examines the possibility of the annual Tasseled Cap trends to identify the growth of thaw slumps over time, from initiation to stabilization.

Following the analysis of annual change in greenness values from 16 thaw slumps, the tundra surrounding thaw slumps typically has greenness values ranging between 110 and 150. By contrast, the greenness values decrease to near 20 on active slump floors where the ground is wet. Therefore, in the case of a growing thaw slump, the temporal pixel trends may represent the change in surface from unaffected tundra (with greenness values near 110-150), to wet slump floor (with greenness values near 20). In the case of a stabilizing thaw slump, the greenness value would slowly increase toward those of tundra values.

The following section will present examples of the annual Tasseled Cap trends analysis to monitor the initiation, growth and stabilization of thaw slumps. This section will also discuss how to derive annual growth rates using this approach, a significant improvement over the traditional method of multi-decadal growth rates of thaw slumps (i.e., Lantuit and Pollard, 2008; Lantz and Kokelj, 2008).

6.2.1. Initiation of thaw slump

The initiation of thaw slumps can be identified using the annual Tasseled Cap trend analysis approach. Thaw slump initiation is characterized by a sudden decrease in greenness values of a pixel from around 140 to about 20.

This approach is also capable of displaying possible causes for thaw slump initiation. In the Road River watershed, a forest fire occurred in 1987 and burned an area covering 14.1km² (Figure 52). The burned area is recognizable in the 1988 Landsat natural color image and by 1990, post-fire vegetation regrowth was well underway. On the 2011 Landsat image, a thaw slump is visible along the stream;

however, the exact year of its initiation is unknown. On the linear Tasseled Cap trends map, (Figure 52), the forest fire area is identifiable by a distinctive teal colour, while the active thaw slump is easily identified by its blue colour. The extraction of the greenness index values along a transect from the slump floor and extending above the position of the headwall in 2011 show that the initiation and growth of the thaw slump can be monitored. By plotting the annual changes in greenness index values, we first observe the vegetation re-growth in the burned area, with values increasing from near 10 to ca. 125. The latter value is similar to the greenness index values for the area outside the burned region, suggesting that vegetation has fully recovered. In 1999, greenness values in pixel 1 decrease from values of roughly 125 to less than 50, suggesting that the thaw slump initiated. The growth of the thaw slump in subsequent years can then be seen from the sequential decrease in pixel values along the transect until the position of the headwall in 2011. This trend in changes in greenness index values along the transect suggests that the thaw slump initiated in 1999 most likely following a landslide that occurred following the forest fire which resulted in newly exposed massive ice to begin to back waste and form a new thaw slump.

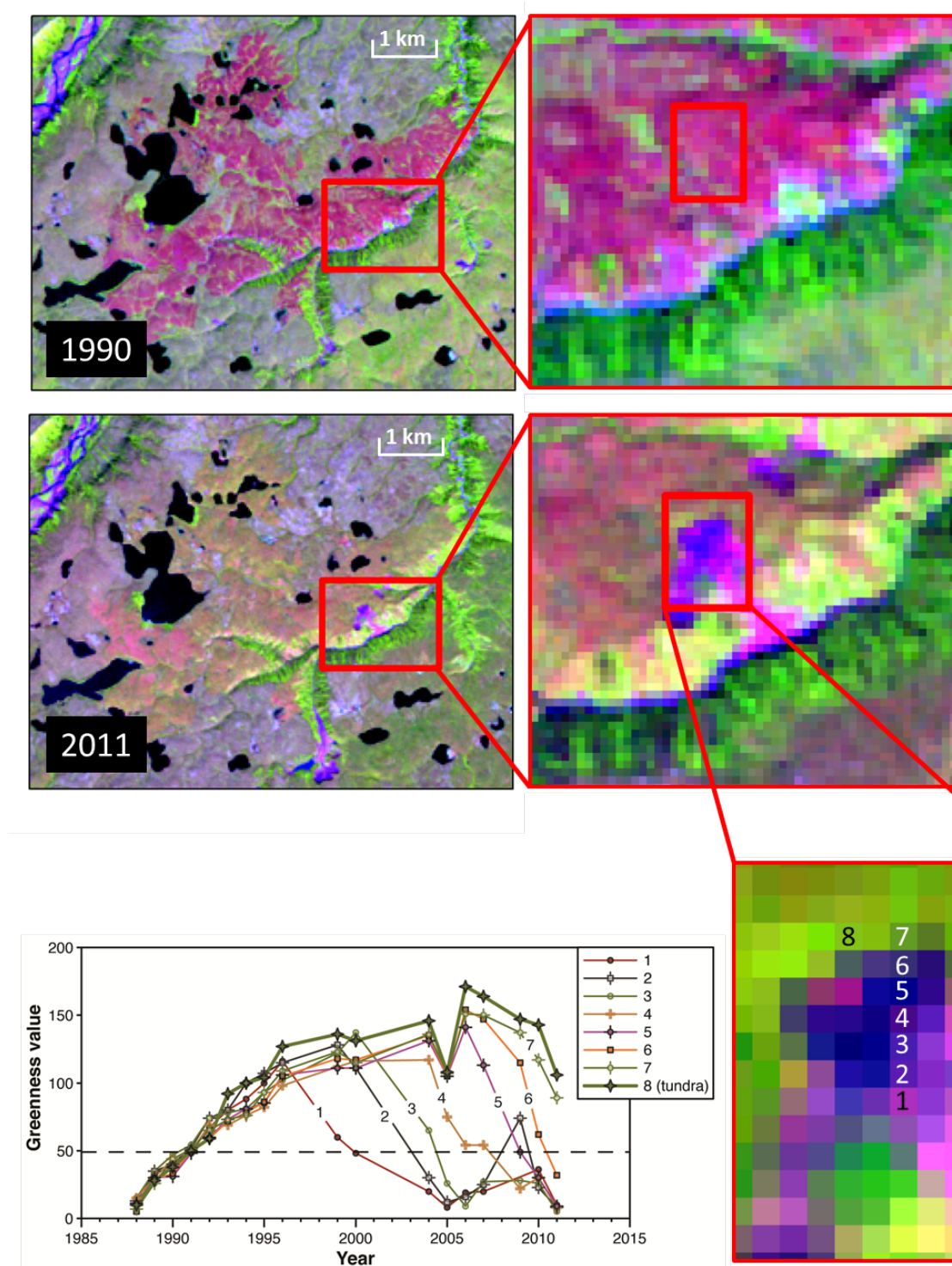


Figure 52. Greenness values of thaw slump #11 that initiated following a forest fire in the area.

6.2.2. Growth of thaw slump

6.2.2.1. Deriving headwall retreat rates from Greenness index trends

The extraction of greenness values from a dense stack of Landsat images along a transect of adjoining pixels that extends from the pixel above the headwall of a slump into the slump floor offers the possibility to monitor the annual growth of thaw slumps. Based on 33 to 48, as the headwall of a thaw slump retreats, the greenness values rapidly decrease from values near 150 to less than 20.

This section will verify if a record of annual slump growth rate may be extracted from the analysis of annual Tasseled Cap trends by examining the annual growth rates of 3 thaw slumps in the Stony Creek watershed. These slumps were selected they can be identified in Ikonos images (4m resolution) from 2000 and 2012. Annual growth rates from the 3 thaw slumps were extracted by assuming that the headwall had receded through the pixel once the greenness value reached a value of 50. This value was chosen as it is higher than the fluctuations of greenness values in the slump floor and lower than the fluctuations of greenness values in the undisturbed tundra. The years in which greenness values intersected the greenness value of 50 were then noted (i.e., dashed horizontal line in Figures 33 to 48). With a Landsat pixel resolution of 30m, a retreat rate can be calculated using the following equation:

$$R = \frac{L}{T_{i+1} - T_i}$$

Where R = thaw slump headwall retreat rate (m yr⁻¹), L = is the resolution of a Landsat pixel (30m), T is the year at which greenness values drop below the threshold line and i is the pixel number.

Figures 53 to 55 shows the annual retreat of 3 thaw slumps in Stony Creek watershed (1990 to 2010) as well as 2000 and 2012 Ikonos images. For thaw slump #6 (Figure 53), calculated annual retreat rate varies between 6 to 15 m yr⁻¹, with an average of 9.4 m yr⁻¹. To assess if these rates are reasonable, the total calculated

retreat rate was compared with the one measured from the Ikonos images. The headwall retreated by 150m between 1990 and 2010 (the sum of annual retreat rate between 2000 and 2010). However, over the time period of the Ikonos images, the headwall retreated by 91m (the sum of annual retreat rate between 2000 and 2010), which is very close to the total headwall retreat of 95.4m as determined using the Ikonos images. The total headwall retreat calculated using the Ikonos images will be higher than the one calculated using the Landsat images because: i) the Ikonos images covers until 2012; and ii) the calculated total headwall retreat from the Landsat images is considered a minimum value as it often does not include the later years of the 2000s unless a cloud-free Landsat image is available for 2011 and the headwall had retreated through this pixel.

Figure 54 shows the calculated annual retreat rate of Melanie slump in the Stony Creek watershed. For this slump, annual retreat rates vary between 7.5 and 30m yr⁻¹. The total retreat rate over the 1990-2010 period is 150m, and over the 2000-2010 period is 135m. In the time between the 2000 and 2012 Ikonos images, the headwall was measured to have retreated by 184m.

Finally, for the Charas thaw slump, one of the largest in the Richardson Mountains, the calculated annual retreat rates vary between 10 and 30m yr⁻¹ (Figure 55). The southern lobe headwall retreated a total of 180m between 1990-2010, and 170m between 2000-2010. Based on the Ikonos images, the headwall retreated a total of 210m, again consistent with the annual Tasseled Cap trend approach.

Overall, the extraction of greenness values from a dense stack of Landsat images along a transect of adjoining pixels that extends from the pixel above the headwall of a slump into the slump floor is able to determine annual retreat rates of thaw slumps. Calculating the annual headwall retreat rates of thaw slumps using the Landsat Image Stack Trend Analysis has several advantages over determining the change in headwall position between air photographs or images, acquired at multi-decadal interval (20 to 30 years interval), by tracing the position of the headwall over time (i.e., Lantuit and Pollard, 2008; Lantz et al., 2008). First, it offers the possibility

to examine annual headwall retreat rates over a 25 year period, something that is missing from previous studies. Second, it offers the possibility to develop an automated method to extract annual headwall retreat rates and explore differing trends in retreat rates over broad regions. For example, in the Richardson Mountains, headwall retreat rates could be compared between slumps situated in the lowlands versus those in the highlands. Finally, the novel approach would allow for a better understanding of the response of thaw slumps and their environmental impacts to changing climate.

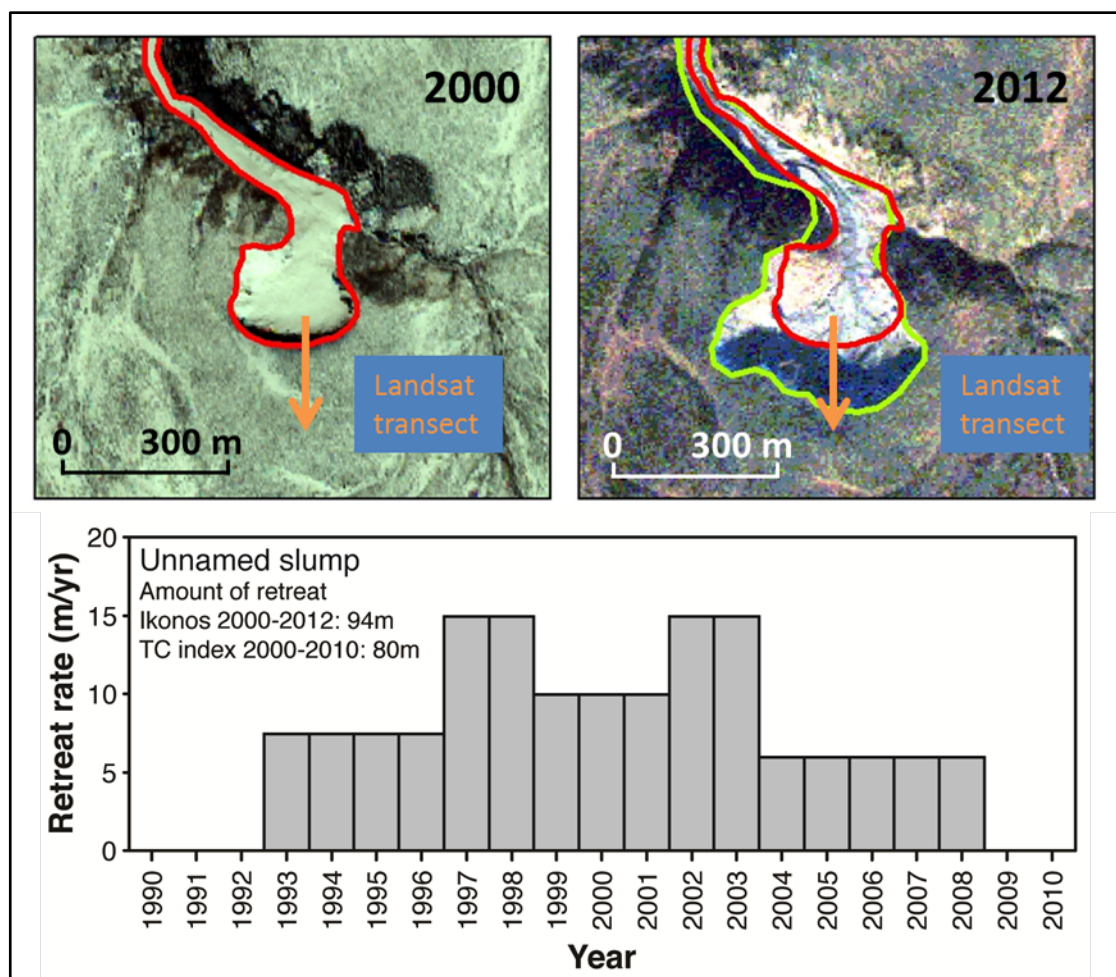


Figure 53. Annual retreat rate of an unnamed thaw slump (thaw slump #6) in the Stony Creek watershed, as derived from the annual greenness trends.

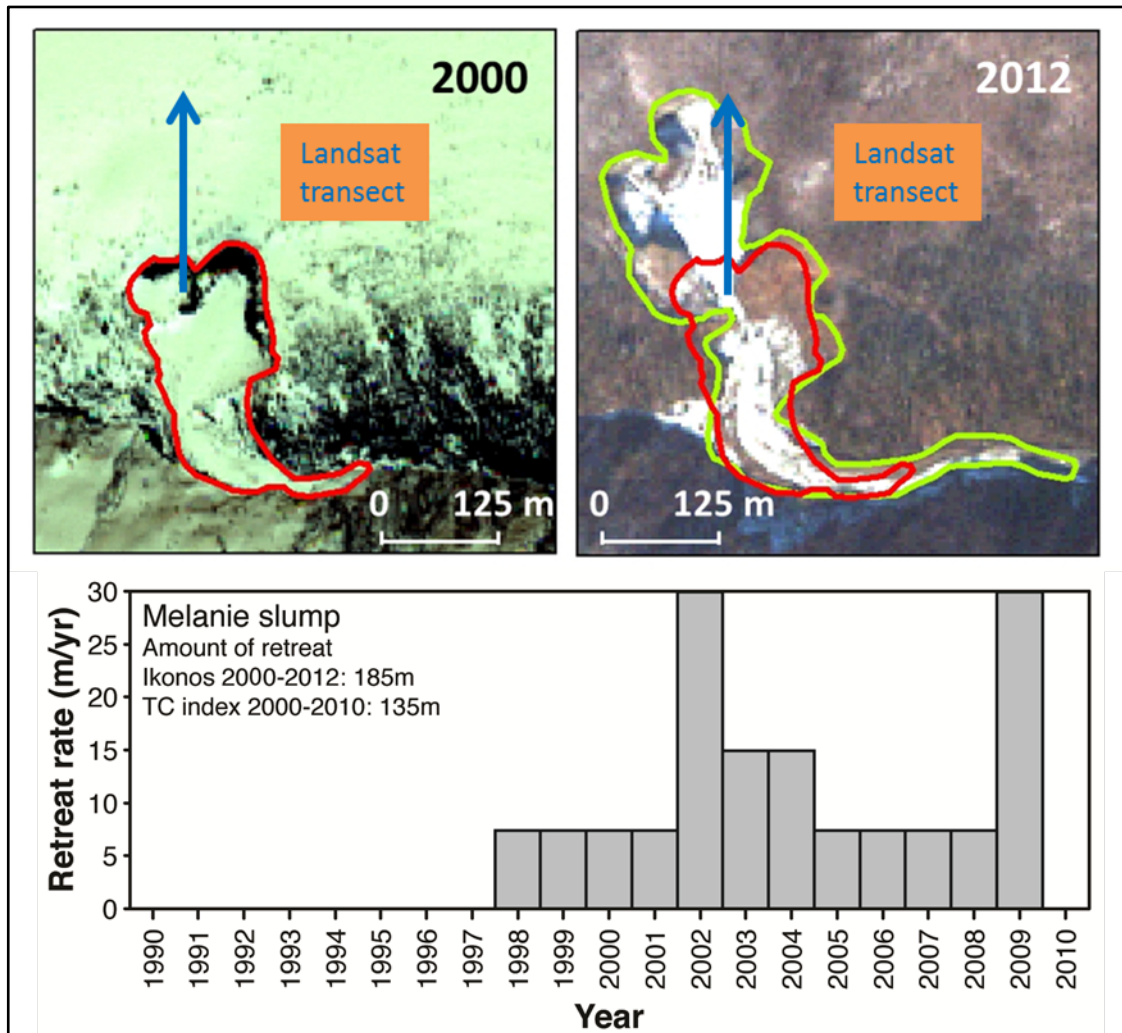


Figure 54. Annual headwall retreat rate of the #5 "Melanie" thaw slump in the Stony Creek watershed, as derived from the annual greenness trends.

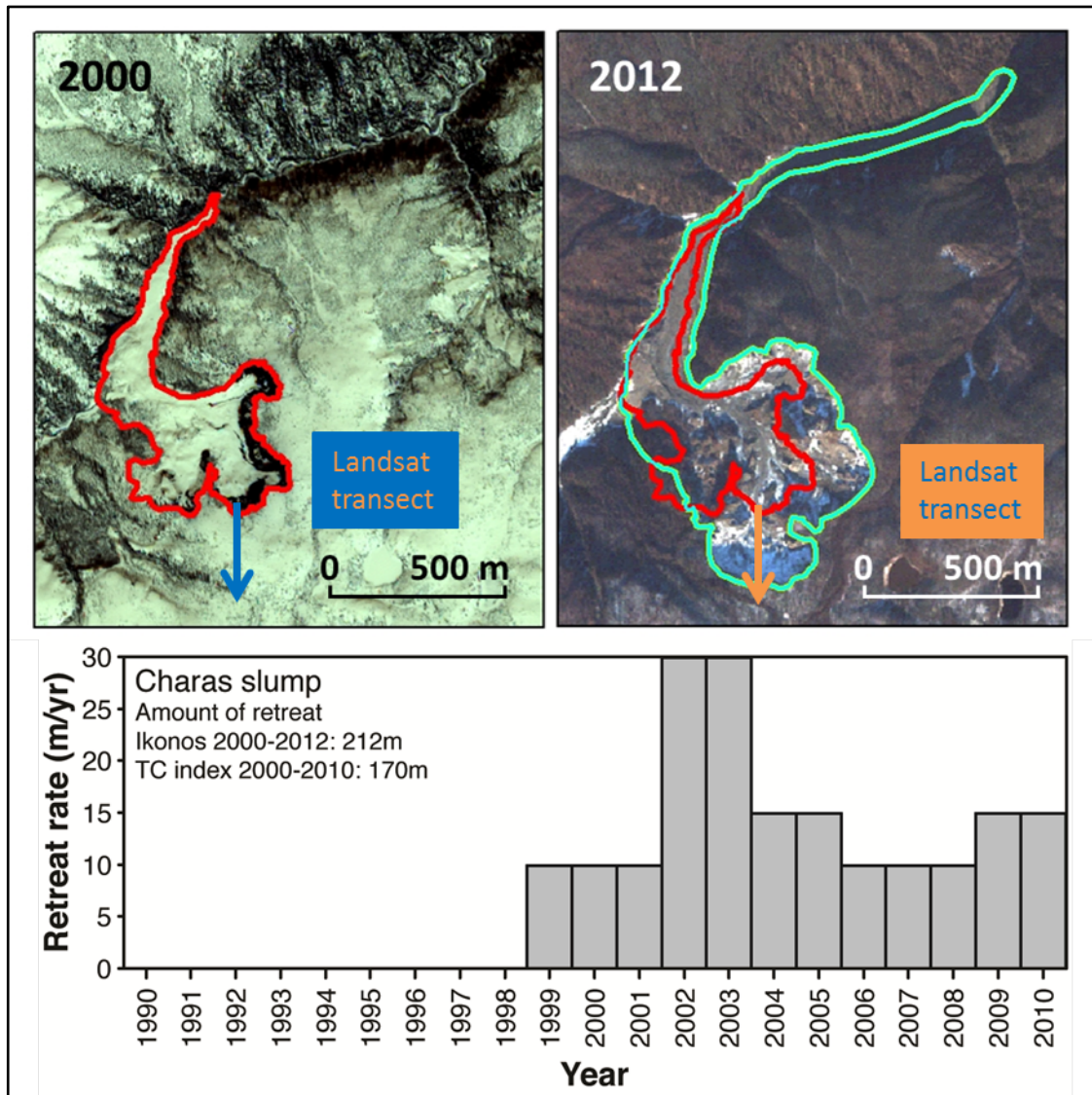


Figure 55. Annual headwall retreat rate of the #4 "Chara" thaw slump in the Stony Creek watershed, as derived from the annual greenness trends.

6.2.2.2. Annual headwall retreat rates in the Richardson Mountains: some preliminary results.

The previous section demonstrated the effectiveness of using the annual Tasseled Cap trend analysis approach in extracting annual headwall retreat rates for 3 thaw slumps in the Stony Creek watershed. This section presents the extracted growth of 11 additional thaw slumps situated in the Rat River, Vittrekwa River, Trail River and Road River watersheds. The location of these thaw slumps in the Richardson Mountains is shown in Figure 7 and their annual change in greenness values is shown in Figures 33 to 48.

Figure 56 shows the annual headwall retreat rates for 14 thaw slumps in the Richardson Mountains. The retreat rates vary between 5 and 60 m yr⁻¹, and the average headwall retreat rate over the 1990-2010 period is 11.8 ± 6.9 m yr⁻¹. This average headwall retreat rate is similar to average headwall retreat rate of 13m yr⁻¹ (1949-1987) and 9.6m yr⁻¹ (1970-2004) for thaw slumps near Mayo, Yukon (Burn and Friele, 1989) and along the coast of Herschel Island, Yukon (Lantuit et al., 2005), respectively. However, headwall retreat rates of lakeside thaw slumps in the Mackenzie Delta were measured to be 1m yr⁻¹ between the years of 1973 and 2004 (Lantz and Kokelj, 2008).

To assess potential relation between headwall retreat rate and climate in the Richardson Mountains, the annual headwall retreat rate of each thaw slumps was averaged (Figure 57) and compared to mean summer air temperature, summer thaw degree-days and number of days with rainfall greater than 10mm from the Fort McPherson meteorological station, the closest to the study area. There are no statistically significant relations between the average headwall retreat rates and the various climate variables.

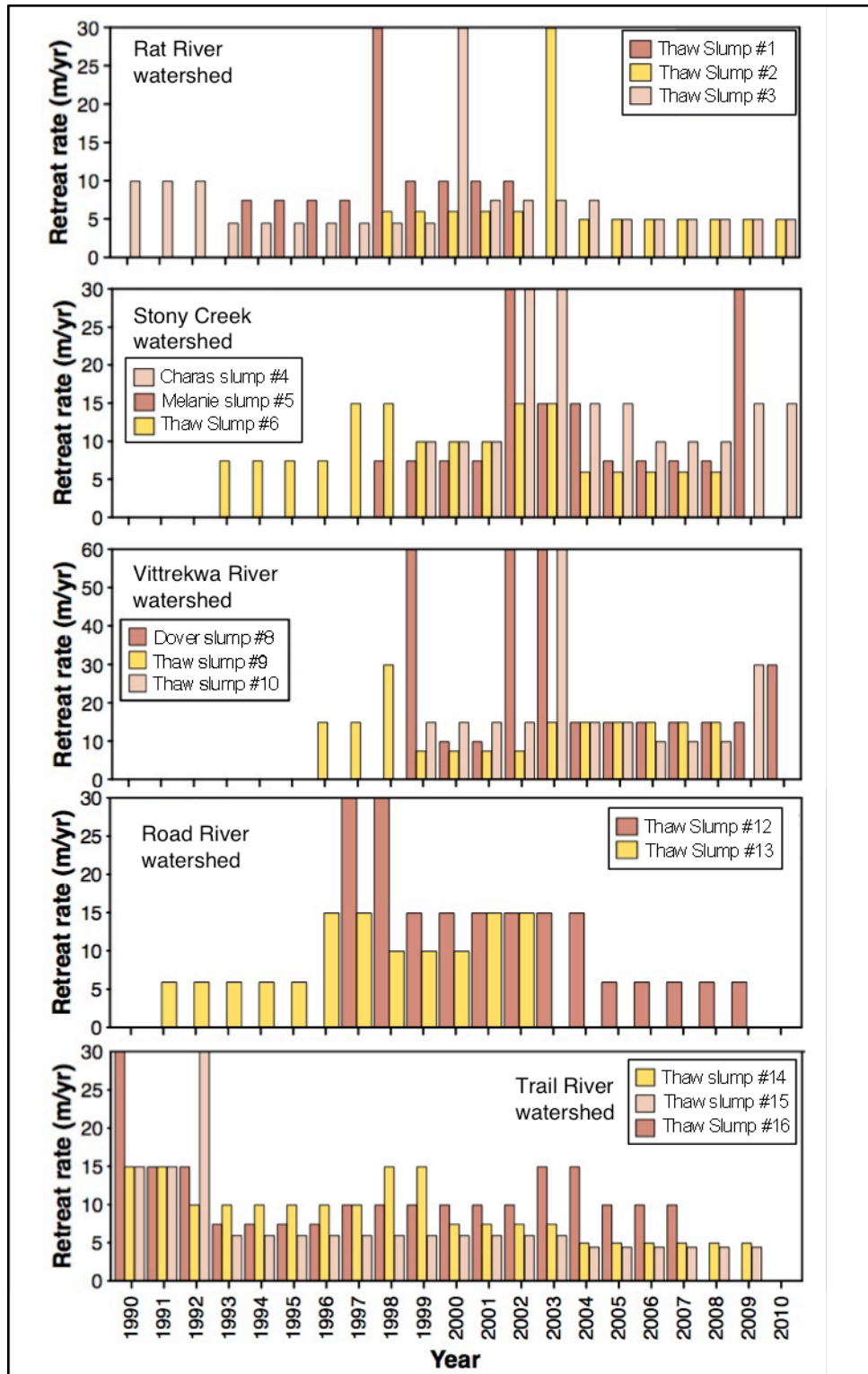


Figure 56. Annual growth rate of headwall for thaw slumps in a given watershed; as calculated from the greenness annual trends.

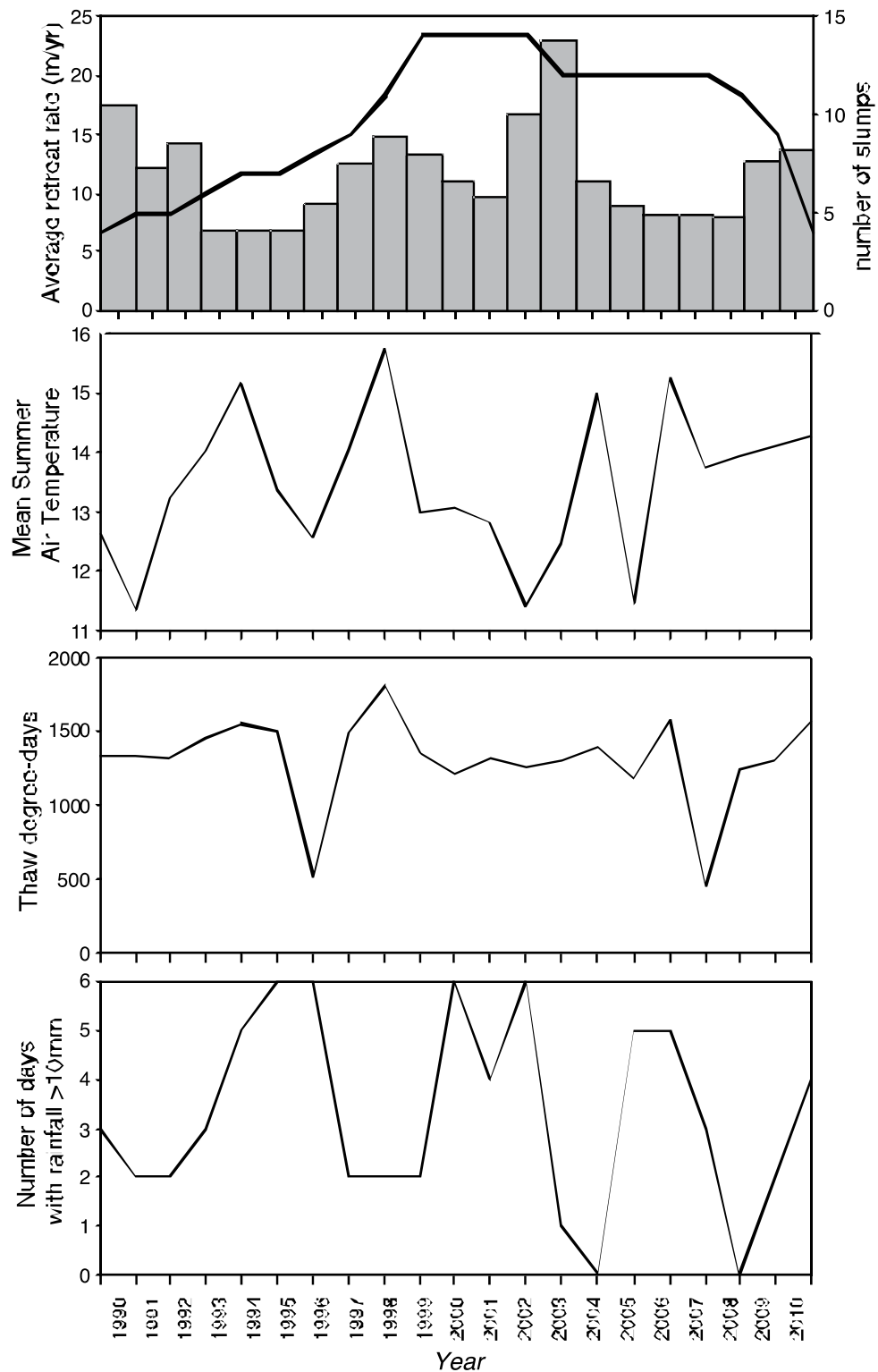


Figure 57. Average annual headwall rate of thaw slumps in the Richardson Mountains, NWT, compared with mean summer air temperature, thaw degree-days and number of days with rainfall >10mm (climate data from the Fort McPherson meteorological station).

6.2.2.3. Limitations of the Landsat Image Stack Trend analysis method in calculating growth rates

The lack of relation with the climate parameters highlights some of the limitations of the Landsat Image Stack Trend Analysis method. Firstly, growth rates were calculated by determining the years at which the pixel greenness values dropped below a threshold value of 50. This value was chosen as it is higher than the annual greenness fluctuations of the slump floor and lower than the annual greenness fluctuations of undisturbed tundra. The year at which greenness values drop below 50 is not necessarily the year at which the headwall has completely receded through the pixel, depending on the rate of headwall retreat. The difference in years between when greenness values drop below 50 and when they reach their lowest value could produce an offset in retreat rates for certain slumps. It might be possible that this produces an offset between the year in which the headwall retreats in the field and that derived from the annual Tasseled Cap trends.

Extracted annual growth rates are not 100% correct, as cloud-free images were not available for every year from 1985-2011, with cloud-free imagery being particularly sparse between the years of 1995 and 2004. To facilitate the interpretation of growth rates, lines were connected between yearly values using Aabel 3 statistics software. By connecting lines between values, patterns in headwall retreat can be easily viewed, despite the missing yearly values. Also, it's possible that the climate in the Richardson Mountains, which extends more than 300km in latitude and 50km in longitude, is different than those measured at Inuvik and Fort McPherson. In fact, with an elevation gradient of ~1000km, climate in the Richardson Mountains is most likely extremely variable, with warmer and longer thaw season in the lowlands compared to the higher elevations. Finally, it is possible that, once initiated, the growth of thaw slumps in the Richardson Mountains is not only dependent on climate, but also on local conditions. For example, the Charas thaw slump has been visible on air photographs since the late 1940s and has been

active since that time. By contrast, nearby thaw slumps have experienced stabilization and re-activation over the same time period.

6.3.3 Stabilization of thaw slumps

The stabilization of thaw slumps can be identified using the Landsat Image Stack Trend Analysis method. Stabilizing thaw slumps are characterized by a drying of the slump floor and re-vegetation (Burn and Lewkowicz, 1990). The stabilization of thaw slumps can be identified by analyzing temporal changes of the Tasseled Cap greenness transformation values at the pixel level. For example, Figure 46 shows the greenness index values along a transect for a thaw slump in the Trail River watershed. The slump was active in 1990 as a much smaller feature and had retreated 180m upslope by 2011. The greenness values of pixel 1, situated in the furthest portion of the slump floor from the current position of the headwall, increase slowly from values near 10 in 1990 to values near 100 in 2006 and subsequent years. This suggests vegetation re-growth on a stable slump floor. Therefore, the stabilization of thaw slumps and associated re-vegetation can be monitored by observing the greenness index values increasing from values near 20 to greater than 100.

Monitoring the stage of stabilization with the Landsat Image Stack Trend Analysis offers several advantages over using high-resolution imagery. Firstly, using high-resolution imagery, one can only make general observations on the amount of slump floor that has begun to regenerate. On the other hand, the Landsat Image Stack Trend Analysis quantifies the greenness value of each pixel allowing for the study of yearly fluctuations in thaw slump regeneration.

7. Conclusions

In conclusion, the Landsat Image Stack Trend Analysis method represents a significant improvement over previous remote sensing methods of thaw slump monitoring which used air photos and high-resolution satellite imagery (Lantuit and Pollard, 2005; Lantz et al., 2008; Lacelle et al., 2010). Given the limitation of using 30m resolution pixels, thaw slumps with surface areas greater than 3600m² are easily identified. This method also allows for the detailed analysis of all phases of the life-cycle of an individual thaw slump, from initiation, to growth, and stabilization. The accuracy of extracted growth rates is limited by the availability of cloud-free imagery, and by the offset in growth rates produced from considering the headwall to have retreated completely through the pixel upon reaching greenness values of 50.

Thaw slumps in the Richardson Mountains are found primarily on North-East facing slopes, in fine-grained morainal deposits. The growth rates of 14 thaw slumps was extracted using the Landsat Image Stack Trend Analysis method, with average growth rates being $11.8 \pm 6.9 \text{ m yr}^{-1}$, which is similar to rates of roughly 13 m yr^{-1} in Mayo, Yukon (Burn and Friele, 1989) and 9.6 m yr^{-1} Herschel Island, Yukon (Lantuit et al., 2005)..

In the future, this method can be improved through increased automation. The identification of thaw slumps could be automated, by selecting pixels which are undergoing the characteristic pattern of greenness value change, as shown in figures 33 to 48. The extraction of annual greenness values of pixels within a thaw slump can also be automated, as well as the subsequent extraction of thaw slump growth rates, as shown in section 6.2.2., which would allow for the analysis of a much greater number of thaw slumps. In addition, new Landsat images for years post-2011 can simply be added to the image stack, so thaw slump activity up to the present day can be studied.

References

References

- ACIA. (2004). *Impacts of a warming climate*. Cambridge University Press.
- Astakhov, V. I. (1986). Geological conditions for the burial of pleistocene glacier ice on the Yenisey. *Polar Geography and Geology*, 10, 286-295.
- ASTER - advanced spaceborne thermal emission and reflection radiometer. (2004). Retrieved 12/27, 2013, from <http://asterweb.jpl.nasa.gov>
- Aylsworth, J. M., Burgess, M. M., Desrochers, D. T., Duk-Rodkin, A., Robertson, T., & Traynor, J. A. (2000). Surficial geology, subsurface materials, and thaw sensitivity of sediments. *The Physical Environment of the Mackenzie Valley, Northwest Territories: A Base Line for the Assessment of Environmental Change*, 547, 41.
- Balser, A. W., Gooseff, M. N., Jones, J. B., & Bowden, W. B. (2009). *Thermokarst distribution and relationships to landscape characteristics in the Feniak Lake region, Noatak National Preserve, Alaska: Final report to the National Park Service, Arctic Network, Fairbanks, Alaska, December 31, 2009*
- Bartleman, A., Miyanishi, K., Burn, C. R., & Côté, M. M. (2001). Development of vegetation communities in a retrogressive thaw slump near Mayo, Yukon Territory: A 10-year assessment. *Arctic*, 54(2), 149-156.
- Benedict, J. B. (1970). Downslope soil movement in a Colorado alpine region: Rates, processes, and climatic significance. *Arctic and Alpine Research*, 2(3), 165-226.
- Brown, J. (1998). In Ferrians O. J., Hegginkbottom J. A. and Melnikov E. S. (Eds.), *Circum-arctic map of permafrost and ground ice conditions*. Boulder, CO: National Snow and Ice Data Center.
- Burn, C. R. (1997). Cryostratigraphy, paleogeography, and climate change during the early Holocene warm interval, western Arctic coast, Canada. *Canadian Journal of Earth Sciences*, 34(7), 912-925.
- Burn, C. R. (2000). The thermal regime of a retrogressive thaw slump near Mayo, Yukon Territory. *Canadian Journal of Earth Sciences*, 37, 967-981.
- Burn, C. R. (2011). Permafrost distribution and stability. In H. M. French, & O. Slaymaker (Eds.), *Changing cold environments: A Canadian perspective* (1st ed.,)

- Burn, C. R., & Lewkowicz, A. G. (1990). Canadian landform examples - retrogressive thaw slumps. *The Canadian Geographer*, 34(3), 273-276.
- Catto, N. R. (1996). Richardson mountains, yukon-northwest territories: The northern portal of the postulated 'ice-free' corridor. *Quaternary International*, 32, 3-19.
- Cheng, G. (1983). The mechanism of repeated-segregation for the formation of thick layered ground ice. *Cold Regions Science and Technology*, 8, 57-66.
- Clark, I. D., Lauriol, B., Marschner, M., Sabourin, N., Chauret, Y., & Desrochers, A. (2004). Endostromatolites from permafrost karst, yukon, canada: Paleoclimatic proxies for the holocene hysithermal. *Canadian Journal of Earth Sciences*, 41, 397-399.
- Cory, R. M., Crump, B. C., Dobkowski, J. A., & Kling, G. W. (2013). Surface exposure to sunlight stimulates CO₂ release from permafrost soil carbon in the arctic. *Proceedings of the National Academy of Sciences, Early Edition*
- Côté, M. M., Duchesne, C., Wright, J. F., & Ednie, M. (2013). *Digital compilation of the surficial sediments of the mackenzie valley corridor, yukon coastal plain, and the tuktoyaktuk peninsula* No. Open File 7289) Geological Survey of Canada.
- Crist, E. P., & Cicone, R. C. (1984). A physically-based transformation of thematic mapper data - the TM tasseled cap. *IEEE Transactions on Geoscience and Remote Sensing*, 22(3), 256.
- Currey, D. R. (1964). A preliminary study of valley asymmetry in the ogotoruk creek area, northwestern alaska. *Arctic*, 17(2), 84-98.
- de Krom, V. (1990). *Retrogressive thaw slumps and active layer slides on herschel island, yukon*. Unpublished McGill University, Montréal, Québec.
- Delorme, L. D., Zoltai, S. C., & Kalas, L. L. (1977). Freshwater shelled invertebrate indicators of paleoclimate in northwestern canada during late glacial times. *Canadian Journal of Earth Sciences*, 14, 2029-2046.
- Duk-Rodkin, A., & Hughes, O. L. (1992). *Surficial geology: Fort McPherson - bell river, yukon - northwest territories* (Map 1745A ed.) Geological Survey of Canada.

- Duk-Rodkin, A., & Lemmen, D. S. (2000). Glacial history of the mackenzie region. *The Physical Environment of the Mackenzie Valley, Northwest Territories: A Base Line for the Assessment of Environmental Change*, 547, 11.
- Dyke, L. D. (2000). Climate of the mackenzie river valley. *The Physical Environment of the Mackenzie Valley, Northwest Territories: A Base Line for the Assessment of Environmental Change*, 547
- Environment Canada. (2013). *Climate normals & averages*. Retrieved 06/05, 2013, from http://climate.weatheroffice.gc.ca/climate_normals/index_e.html
- Fraser, R., Olthof, I., Carrière, M., Deschamps, A., & Pouliot, D. (2012). A method for trend-based change analysis in arctic tundra using the 25-year landsat archive. *Polar Record*, 48, 83-93.
- Fraser, R. H., Olthof, I., Carrière, D., A., & Pouliot, D. (2011). Detecting long-term changes to vegetation in northern canada using the landsat satellite image archive. *Environmental Research Letters*, 6(4), 9.
- Fraser, R. H., Olthof, I., Deschamps, A., & Carrière, M. (2012). *Protocol for remote sensing based vegetation change detection in canadian arctic and subarctic national parks*. Ottawa, ON: Natural Resources Canada.
- French, H. M. (1974). Active thermokarst processes, eastern banks island, western canadian arctic. *Canadian Journal of Earth Sciences*, 11(6), 785-794.
- French, H. M. (2007). *The periglacial environment* (3rd ed.) John Wiley & Sons, Ltd.
- French, H. M., & Harry, D. G. (1990). Observations on buried glacier ice and massive segregated ice, western arctic coast, canada. *Permafrost and Periglacial Processes*, 1, 31.
- Harris, C. (1981). *Periglacial mass wasting: A review of research* Geo Books: Norwich.
- Harris, C., Gallop, M., & Coutard, J. -. (1993). Physical modelling of gelifluction and frost creep: Some results of a large scale laboratory experiment. *Earth Surface Processes and Landforms*, 18, 383-398.
- Harris, C., & Lewkowicz, A. G. (2000). An analysis of the stability of thawing slopes, ellesmere island, nunavut, canada. *Canadian Geotechnical Journal*, 37, 449-462.

- Heginbottom, J. A., & Radburn, L. K. (1992). *Permafrost and ground ice conditions of northwestern canada*. Boulder, CO: Geological Survey of Canada.
- Hill, P. R., Mudie, P. J., Moran, K., & Blasco, S. M. (1985). A sea-level curve for the canadian beaufort shelf. *Canadian Journal of Earth Sciences*, 22, 1383-1393.
- Huscroft, C. A., Lipovsky, P. S., & Bond, J. D. (2003). Permafrost and landslide activity: Case studies from southwestern yukon territory. *Yukon Exploration and Geology*, , 107-119.
- International Panel on Climate Change (IPCC). (2007). ***Climate change 2007: The physical science basis***. Cambridge, UK; New York, NY, USA: Cambridge University Press.
- Jones, B. M., Arp, C. D., Jorgenson, M. T., Hinkel, K. M., Schmutz, J. A., & Flint, P. L. (2009). Increase in the rate and uniformity of coastline erosion in arctic alaska. *Geophysical Research Letters*, 36(3)
- Jorgenson, M. T., Shur, Y. L., & Pullman, E. R. (2006). Abrupt increase in permafrost degradation in arctic alaska. *Geophysical Research Letters*, 33(2)
- Jorgenson, M. T., Shur, Y. L., & Osterkamp, T. E. (2008). Thermokarst in alaska. *Proceedings of the Ninth International Conference on Permafrost*, Fairbanks, AK.
- Kääb, A. (2008). Remote sensing of permafrost-related problems and hazards. *Permafrost and Periglacial Processes*, 19
- Kaplan, J. O., & New, M. (2006). Arctic climate change with a 2°C global warming: Timing, climate patterns and vegetation change. *Climatic Change*, 79, 213-241.
- Kauffman, D. S., Ager, T. A., Anderson, N. J., Anderson, P. M., Andrews, J. T., Bartlein, P. J., et al. (2004). Holocene thermal maximum in the western arctic (0-180°W). *Quaternary Science Reviews*, 23(5-6), 529-560.
- Kokelj, S. V., Jenkins, R. E., Milburn, D., Burn, C. R., & Snow, N. (2005). The influence of thermokarst disturbance on the water quality of small upland lakes, mackenzie delta region, northwest territories, canada. *Permafrost and Periglacial Processes*, 16, 343-353.
- Kokelj, S. V., Lacelle, D., Lantz, T. C., Tunnicliffe, J., Malone, L., Clark, I. D., et al. (2013). Thawing of massive ground ice in mega slumps drives increases in stream sediment and solute flux across a range of watershed scales. *Journal of Geophysical Research-Earth Surfaces*, 118, 1-12.

- Kokelj, S. V., Lantz, T. C., Kanigan, J., Smith, S. L., & Coutts, R. (2009). Origin and polycyclic behaviour of tundra thaw slumps, mackenzie delta region, northwest territories, canada. *Permafrost and Periglacial Processes*, 20, 173-184.
- Lacelle, D., Bjornson, J., & Lauriol, B. (2010). Climatic and geomorphic factors affecting contemporary (1950-2004) activity of retrogressive thaw slumps on the aklavik plateau, richardson mountains, N.W.T., canada. *Permafrost and Periglacial Processes*, 1, 1-15.
- Lacelle, D., Bjornson, J., Lauriol, B., Clark, I. D., & Troutet, Y. (2004). Segregated-intrusive ice of subglacial meltwater origin in retrogressive thaw flow headwalls, richardson mountains, NWT, canada. *Quaternary Science Reviews*, (23), 681.
- Lacelle, D., Lauriol, B., Zazula, G. D., Ghaleb, B., Utting, N., & Clark, I. D. (2013). Timing of advance and basal condition of the laurentide ice sheet during the last glacial maximum in the richardson mountains, NWT. *Quaternary Research*, 80(2), 274-283.
- Lamirande, I., Lauriol, B., Lalonde, B. E., & Clark, I. (1999). La production de limon sur des terrasses de cryoplanation dans les monts richardson, canada. *Canadian Journal of Earth Sciences*, 36, 1645-1654.
- Lantuit, H., & Pollard, W. H. (2008). Fifty years of coastal erosion and retrogressive thaw slump activity on herschel island, southern beaufort sea, yukon territory, canada. *Geomorphology*, 95
- Lantuit, H., Pollard, W. H., Couture, N., Fritz, M., Schirrmeister, L., Meyer, H., et al. (2012). Modern and late holocene retrogressive thaw slump activity on the yukon coastal plain and herschel island, yukon territory, canada. *Permafrost and Periglacial Processes*, 23, 39-51.
- Lantz, T. C., & Kokelj, S. V. (2008). Increasing rates of retrogressive thaw slump activity in the mackenzie delta region, N.W.T., canada. *Geophysical Research Letters*, 35
- Lantz, T. C., Kokelj, S. V., Gergel, S. E., & Henry, G. H. R. (2009). Relative impacts of disturbance and temperature: Persistent changes in microenvironment and vegetation in retrogressive thaw slumps. *Global Change Biology*, 15(7), 1664-1675.
- Lauriol, B., Duguay, C., & Riel, A. (2002). Responses of the porcupine and old crow rivers in northern yukon to holocene climate change. *The Holocene*, 12(1), 27-34.

- Lauriol, B., Lacelle, D., Labrecque, S., Duguay, C. R., & Telka, A. (2009). Holocene evolution of lakes in the bluefish basin, northern yukon, canada. *Arctic*, 62(2), 212-224.
- Lemmen, D. S., Duk-Rodkin, A., & Bednarski, J. M. (1994). Late glacial drainage systems along the northwestern margin of the laurentide ice sheet. *Quaternary Science Reviews*, 13, 805-828.
- Lewkowicz, A. G. (1985). Use of an ablatometer to measure short-term ablation of exposed ground ice. *Canadian Journal of Earth Sciences*, 22, 1767-1773.
- Lewkowicz, A. G. (1986). Rate of short-term ablation of exposed ground ice, banks island, northwest territories, canada. *Journal of Glaciology*, 32(112)
- Lewkowicz, A. G. (2007). Dynamics of active-layer detachment failures, fosheim peninsula, ellesmere island, nunavut, canada. *Permafrost and Periglacial Processes*, 18, 89-103.
- Lewkowicz, A. G., Bonnaventure, P. P., Smith, S. L., & Kuntz, Z. (2012). Spatial and thermal characteristics of mountain permafrost, northwest canada. *Geografiska Annaler: Series A, Physical Geography*,
- Lewkowicz, A. G., & Harris, C. (2005). Frequency and magnitude of active-layer detachment failures in discontinuous and continuous permafrost, northern canada. *Permafrost and Periglacial Processes*, 16, 115-130.
- Mackay, J. R. (1966). Segregated epigenetic ice and slumps in permafrost mackenzie delta area, N.W.T. *Geographical Bulletin*, 8, 89-80.
- Mackay, J. R., & Dallimore, S. R. (1992). Massive ice of the tuktoyaktuk area, western arctic coast, canada. *Canadian Journal of Earth Sciences*, 29, 1235.
- Malone, L., Lacelle, D., Kokelj, S. V., & Clark, I. D. (2013). Impacts of hillslope thaw slumps on the geochemistry of permafrost catchments (stony creek watershed, NWT, canada). *Chemical Geology*, 356, 38-49.
- Matsuoka, N. (2001). Direct observation of frost wedging in alpine bedrock. *Earth Surface Processes and Landforms*, 26, 601-614.
- Matsuoka, N., & Sakai, H. (1999). Rockfall activity from an alpine cliff during thawing periods. *Geomorphology*, 28(3-4), 309-328.

- Mesquita, P. S., Wrona, F. J., & Prowse, T. D. Effects of retrogressive permafrost thaw slumping on sediment chemistry and submerged macrophytes in arctic tundra lakes. *Freshwater Biology*, 55, 2347-2358.
- Murton, J. B. (2001). Thermokarst sediments and sedimentary structures, tuktoyaktuk coastlands, western arctic canada. *Global and Planetary Change*, 28, 175-192.
- NASA. (2013). *Landsat science*. Retrieved 3/5, 2013, from <http://landsat.gsfc.nasa.gov/about/science.html>
- Norris, D. K. (1975). *Geology, blow river and davidson mountains, yukon territory-district of mackenzie* (NRCAN-RNCAN ed.)
- Olfthof, I., Butson, C., Fernandes, R., Fraser, R., Latifovic, R., & Oraziotti, J. (2005). Landsat ETM+ mosaic of northern canada. *Canadian Journal of Remote Sensing*, 31(5), 412-419.
- Osterkamp, T. E., & Romanovsky, V. E. (1999). Evidence for warming and thawing of discontinuous permafrost in alaska. *Permafrost and Periglacial Processes*, 19(1), 17-37.
- Rampton, V. N. The influence of ground ice and thermokarst upon the geomorphology of the mackenzie-beaufort region. *Research in Polar and Alpine Geomorphology*, , 43.
- Rampton, V. N. (1988). *Quaternary geology of the tuktoyaktuk coastlands, northwest territories* No. Memoir 423) Geological Survey of Canada.
- Ritchie, J. C., Cwynar, L. C., & Spear, R. W. (1983). Evidence from north-west canada for an early holocene milankovitch thermal maximum. *Nature*, 305(8), 126.
- Schuur, E. A. G., Bockheim, J., Canadell, J. G., Euskirchen, E., Field, C. B., Goryachkin, S. V., et al. (2008). Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle. *Bioscience*, 58(8), 701-714.
- Shao, J. (2003). *Mathematical statistics* (2nd ed.) Springer Science+Media, LLC.
- Skinner, B. J., & Porter, S. C. (1995). *The dynamic earth: An introduction to physical geology* (3rd ed.) John Wiley & Sons Inc.
- Skinner, B. J., Porter, S. C., & Park, J. (2004). *Dynamic earth: An introduction to physical geology* (5th ed.) John Wiley & Sons. Inc.

- Smith, D. J. (1992). Long-term rates of contemporary solifluction in the canadian rocky mountains. *Periglacial Geomorphology*, 203-221.
- Smith, M. W., & Riseborough, D. W. (2002). Climate and the limits of permafrost: A zonal analysis. *Permafrost and Periglacial Processes*, 13
- Smith, S. L., Burgess, M. M., Riseborough, D., & Nixon, M. (2005). Recent trends from canadian permafrost thermal monitoring network sites. *Permafrost and Periglacial Processes*, 16
- Smith, S. L., Romanovsky, V. E., Lewkowicz, A. G., Burn, C. R., Allard, M., Clow, G. D., et al. (2010). Thermal state of permafrost in north america - A contribution to the international polar year. *Permafrost and Periglacial Processes*, 21, 117-135.
- Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., et al. (2007). ***Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change, 2007***. Cambridge, U.K and New York, NY, USA: Cambridge University Press.
- St-Onge, D. A., & McMartin, I. (1995). *Quaternary geology of the inman river area, northwest territories* No. Bulletin 446) Geological Survey of Canada.
- Tarbuck, E. J., & Lutgens, F. K. (2005). *Earth science* (11th ed.) Prentice Hall PTR.
- USGS. (2012). *Frequently asked questions about the landsat missions.*, 2012, from http://landsat.usgs.gov/band_designations_landsat_satellites.php
- USGS. (2013). *SLC-off products: Background*. http://landsat.usgs.gov/products_slcoffbackground.php
- van Everdingen, R. (1998). *Multi-language glossary of permafrost and related ground-ice terms*. Boulder, CO: National Snow and Ice Data Center.
- Wang, B. (2011). Retrogression rate of thaw slumps in permafrost - an update from the latest monitoring data. *2011 Pan-Am CGS Geotechnical Conference*, Toronto, ON.
- Wang, B., Nichol, S., & Su, X. (2005). Geotechnical field observations of landslides in fine. *Landslides* (pp. 203-212)
- Wang, B., Paudel, B., & Li, H. (2009). Retrogression characteristics of landslides in fine-grained permafrost soils, mackenzie valley, canada. *Landslides*, 6(121), 127.

- Weier, J., & Herring, D. (2013). *Measuring vegetation (NDVI & EVI)*.
http://earthobservatory.nasa.gov/Features/MeasuringVegetation/measuring_vegetation_2.php
- Wolfe, S. A., Kotler, E., & Dallimore, S. R. (2001). *Surficial characteristics and the distribution of thaw landforms (1970 - 1999), shingle point to kay point, yukon territory*. Geological Survey of Canada, Open File 4115.
- Zazula, G. D., Duk-Rodkin, A., Schweger, C. E., & Morlan, R. E. (2004). Late pleistocene chronology of glacial lake old crow and the north-west margin of the laurentide ice sheet. *Quaternary Glaciations - Extent and Chronology, Part II, 2*, 347.
- Zazula, G. D., Mackay, G., Andrews, T. D., Shapiro, B., Letts, B., & Brock, F. (2009). A late pleistocene steppe bison (*bison priscus*) partial carcass from tsiigehtchic, northwest territories, canada. *Quaternary Science Reviews*, 28, 25-26.
- Zhang, Y., Chen, W., & Riseborough, D. W. (2008). Transient projections of permafrost distribution in canada during the 21st century under scenarios of climate change. *Global and Planetary Change*, 60, 443-456.

Appendix A: Landsat scenes included in each of the three stacks.

Path-row	Year	Month	Day	Stack 1	Stack 2	Stack 3
60-13	1985	7	8		Y	Y
61-13	1985	7	15		Y	Y
61-12	1985	7	31		Y	
64-11	1985	8	5	Y	Y	
64-12	1985	8	5		Y	
62-11	1985	8	7	Y	Y	
62-12	1985	8	7		Y	
64-11	1986	7	23	Y	Y	
64-12	1986	8	8		Y	
62-11	1986	8	10	Y	Y	
62-12	1986	8	10		Y	
60-13	1987	7	14		Y	Y
60-13	1987	8	15		Y	Y
63-13	1988	7	5		Y	Y
62-11	1988	7	14	Y	Y	
64-11	1989	7	15	Y	Y	
64-12	1989	7	15		Y	
60-13	1989	7	19		Y	Y
62-11	1989	8	18	Y	Y	
62-12	1989	8	18		Y	
61-13	1989	8	27		Y	Y
62-11	1990	7	4	Y	Y	
62-12	1990	7	4		Y	
62-13	1990	7	4		Y	Y
61-12	1990	8	14		Y	
61-13	1990	8	14		Y	Y
63-13	1991	6	28		Y	Y
64-12	1991	7	5		Y	

61-13	1991	7	16		Y	Y
62-12	1992	7	9		Y	
62-13	1992	7	9		Y	Y
61-13	1992	7	18		Y	Y
62-11	1992	7	25	Y	Y	
63-12	1992	8	1		Y	
59-13	1992	8	21		Y	Y
64-11	1993	7	10	Y	Y	
62-11	1993	7	23	Y	Y	
62-12	1993	7	28		Y	
62-13	1993	7	28		Y	Y
60-13	1993	8	15		Y	Y
63-12	1993	8	20		Y	
59-13	1994	7	10		Y	Y
62-11	1994	7	26	Y	Y	
63-12	1994	8	7		Y	
63-13	1994	8	7		Y	Y
62-13	1995	7	2		Y	Y
61-11	1995	7	11	Y	Y	
61-12	1995	7	11		Y	
61-13	1995	7	11		Y	Y
63-11	1995	8	10	Y	Y	
61-13	1996	6	30		Y	Y
63-12	1996	7	11		Y	
61-13	1996	7	29		Y	Y
60-13	1996	8	23		Y	Y
63-11	1997	7	30	Y	Y	
63-12	1997	7	30		Y	
60-13	1997	7	31		Y	Y
60-13	1998	6	26		Y	Y
63-13	1998	7	1		Y	Y
59-13	1998	7	21		Y	Y
64-11	1998	7	24	Y	Y	

61-13	1998	8	20		Y	Y
61-13	1999	7	14		Y	Y
59-13	1999	7	16		Y	Y
64-12	1999	8	4		Y	
62-11	1999	8	6	Y	Y	
61-12	1999	8	7		Y	
62-13	2000	7	7		Y	Y
61-12	2000	7	16		Y	
61-13	2000	7	16		Y	Y
62-11	2000	7	23	Y	Y	
62-12	2000	7	23		Y	
63-12	2001	7	17		Y	
59-13	2002	7	8		Y	Y
62-11	2002	7	13	Y	Y	
61-13	2002	7	22		Y	Y
63-11	2002	7	30	Y	Y	
60-13	2003	7	18		Y	Y
62-12	2003	8	9		Y	
63-13	2004	7	1		Y	Y
62-13	2004	7	2		Y	Y
60-13	2004	7	20		Y	Y
62-13	2004	7	26		Y	Y
64-11	2004	8	9	Y	Y	
64-12	2004	8	9		Y	
62-13	2005	6	27		Y	Y
62-11	2005	8	14	Y	Y	
62-12	2005	8	14		Y	
61-13	2005	8	15		Y	Y
60-13	2005	8	16		Y	Y
61-13	2006	7	25		Y	Y
60-13	2006	7	26		Y	Y
62-11	2006	8	1	Y	Y	
62-13	2006	8	1		Y	Y

61-13	2006	8	10		Y	Y
59-13	2007	6	28		Y	Y
63-12	2007	7	10		Y	
63-11	2007	7	26	Y	Y	
62-11	2007	8	4	Y	Y	
62-12	2007	8	12		Y	
62-13	2007	8	12		Y	Y
62-11	2008	7	5	Y	Y	
61-12	2008	7	14		Y	
61-13	2008	7	14		Y	Y
62-12	2008	8	5		Y	
62-12	2009	7	8		Y	
60-13	2009	7	10		Y	Y
62-13	2009	7	16		Y	Y
61-13	2009	7	17		Y	Y
64-12	2009	8	23		Y	
62-11	2009	8	25	Y	Y	
60-13	2010	7	5		Y	Y
63-12	2010	7	10		Y	
62-11	2010	7	27	Y	Y	
62-13	2010	8	4		Y	Y
63-11	2010	8	19	Y	Y	
62-12	2011	6	28		Y	
62-13	2011	6	28		Y	Y
62-11	2011	7	14	Y	Y	
62-12	2011	7	14		Y	
64-12	2011	7	28		Y	
59-13	2011	8	26		Y	Y

Appendix B: Individual thaw slump information.

Surficial geology map unit designation (from Côté et al., 2013)

Map unit	Description
Ap	Alluvial plain
C	Colluvial
Cv	Colluvial veneer
Cx	Colluvial complex
M	Morainal
Mb	Moraine blanket
Mh	Hummocky moraine
Mm	Morainal, rolling
Mp	Moraine plain
Mv	Moraine veneer
P	Piedmont

Slump #	Latitude	Longitude	Area (ha)	Aspect (°)	Slope (°)	Elevation (m)	Surficial Geology	Watershed name
1	-135.86	68.40	3.2	225.0	2.4	233	M	Big Fish
2	-135.87	68.40	5.0	45.0	3.4	250	M	Big Fish
3	-135.96	68.46	1.2	140.5	10.4	235	M	Big Fish
4	-136.64	68.42	3.3	318.5	16.1	339	M	Big Fish
5	-136.52	68.49	1.3	139.1	13.9	294	M	Big Fish
6	-136.62	68.50	1.8	180.8	16.0	304	M	Big Fish
7	-136.08	68.55	2.8	38.7	3.1	153	M	Big Fish
8	-136.04	68.54	4.1	242.4	12.2	164	M	Big Fish
9	-135.75	68.34	2.6	110.0	5.6	282	M	Unnamed watershed #1
10	-135.71	68.35	3.1	130.8	9.1	210	M	Unnamed watershed #1
11	-135.76	68.34	0.6	35.9	8.5	332	M	Unnamed watershed #1
12	-135.77	68.34	2.5	353.7	6.5	350	M	Unnamed watershed #1
13	-135.57	68.06	2.9	150.9	4.9	382	M	Willow
14	-135.65	68.03	4.3	63.4	7.4	456	M	Willow
15	-135.71	68.03	3.9	181.5	8.8	475	M	Willow
16	-135.65	68.08	8.1	116.6	9.5	302	M	Willow
17	-135.68	68.10	4.9	85.6	9.3	344	M	Willow
18	-135.66	68.10	2.6	16.6	13.9	308	M	Willow
19	-135.63	68.11	4.6	139.8	19.2	193	C	Willow

20	-135.68	68.10	10.3	112.2	11.2	336	M	Willow
21	-135.63	68.04	4.7	57.2	12.6	435	M	Willow
22	-135.77	68.03	2.0	104.3	14.2	672	M	Willow
23	-135.53	68.13	11.2	65.2	6.8	336	M	Willow
24	-135.52	68.13	2.3	92.7	9.9	365	M	Willow
25	-135.49	68.15	4.3	281.3	6.1	378	M	Willow
26	-135.61	68.15	5.5	176.6	4.1	257	C	Willow
27	-135.63	68.18	0.9	49.7	19.8	246	M	Willow
28	-135.63	68.18	2.1	92.3	22.2	238	M	Willow
29	-135.67	68.19	3.0	138.6	5.4	232	C	Willow
30	-135.64	68.22	4.4	214.7	3.8	246	C	Willow
31	-135.56	68.20	7.0	170.9	22.9	94	C	Willow
32	-135.49	68.20	4.3	22.2	6.9	112	M	Willow
33	-135.70	68.00	1.8	79.3	16.3	517	M	Willow
34	-135.70	68.01	1.3	288.9	19.9	473	M	Willow
35	-135.72	68.02	0.9	350.4	18.0	520	M	Willow
36	-135.71	68.00	1.3	70.0	5.6	547	M	Willow
37	-135.66	68.05	4.0	352.6	10.9	387	M	Willow
38	-135.72	68.06	2.1	166.0	7.8	405	M	Willow
39	-135.72	68.06	1.8	4.6	8.8	442	M	Willow
40	-135.75	68.09	3.4	247.8	13.7	422	M	Willow
41	-135.66	68.11	8.3	9.5	8.6	322	M	Willow
42	-135.84	68.08	2.8	0.0	3.3	609	M	Willow
43	-135.85	68.08	0.7	327.5	12.3	630	M	Willow
44	-135.85	68.08	0.7	148.1	14.6	612	M	Willow
45	-135.85	68.08	0.9	309.0	12.7	632	M	Willow
46	-135.86	68.08	0.9	85.4	8.8	608	M	Willow
47	-135.89	68.16	1.6	80.0	8.2	524	C	Willow
48	-135.90	68.16	1.5	153.1	16.9	556	C	Willow
49	-135.57	67.98	3.3	41.4	5.4	514	P	Willow
50	-135.52	67.98	3.2	36.9	4.8	564	P	Willow
51	-135.57	68.01	4.3	59.0	4.2	390	C	Willow
52	-135.64	68.13	1.2	119.7	3.8	311	M	Willow
53	-135.66	68.12	3.0	190.1	13.3	325	M	Willow
54	-135.77	67.95	7.7	95.7	9.5	609	P	Willow
55	-135.74	67.95	1.5	140.4	7.1	536	P	Willow
56	-135.68	68.04	1.8	31.0	1.4	420	C	Willow
57	-135.71	68.01	1.1	116.6	3.2	505	M	Willow
58	-135.71	68.01	1.8	102.3	5.6	499	M	Willow
59	-135.73	68.02	1.2	6.3	8.6	510	M	Willow
60	-135.74	68.05	1.5	135.0	8.0	461	M	Willow
61	-135.74	68.05	2.8	3.4	8.1	492	M	Willow
62	-135.75	68.05	3.3	25.8	8.2	505	M	Willow
63	-135.75	68.06	0.4	151.4	6.0	499	M	Willow
64	-135.74	68.05	1.2	351.0	4.6	481	M	Willow

65	-135.76	68.05	0.6	35.5	10.2	558	M	Willow
66	-135.66	68.06	2.3	336.8	1.8	354	M	Willow
67	-135.66	68.06	7.8	131.6	11.3	341	M	Willow
68	-135.67	68.08	12.0	135.0	9.0	323	M	Willow
69	-135.69	68.11	1.1	106.7	9.9	331	M	Willow
70	-135.68	68.11	3.3	192.5	13.0	290	M	Willow
71	-135.55	68.14	6.2	248.2	7.7	259	C	Willow
72	-135.56	67.92	1.5	92.0	13.6	566	Cx	Unnamed watershed #2
73	-135.26	67.49	5.1	62.5	6.7	276	Cx	Rat
74	-135.30	67.52	10.7	47.9	13.3	262	Mm	Rat
75	-135.96	67.45	6.1	273.7	11.1	534	P	Rat
76	-135.99	67.53	3.0	199.4	8.5	467	P	Rat
77	-135.75	67.59	9.1	307.3	12.4	368	P	Rat
78	-135.80	67.59	7.8	88.3	7.8	383	Cx	Rat
79	-135.78	67.48	3.7	78.7	7.3	539	P	Rat
80	-135.79	67.47	5.8	67.4	6.2	586	P	Rat
81	-135.81	67.48	3.7	63.4	5.3	616	P	Rat
82	-135.83	67.47	1.8	70.0	11.0	659	P	Rat
83	-135.83	67.47	0.8	78.7	9.6	692	P	Rat
84	-135.64	67.47	2.5	144.9	7.8	411	Mv	Rat
85	-135.69	67.55	3.0	24.0	4.7	334	P	Rat
86	-135.76	67.66	5.2	121.6	14.3	267	Cx	Rat
87	-135.81	67.69	6.5	104.3	10.5	262	P	Rat
88	-135.55	67.63	2.1	10.7	16.3	303	Cx	Rat
89	-135.67	67.53	3.0	334.1	9.2	330	P	Rat
90	-135.68	67.49	2.3	171.9	3.4	436	P	Rat
91	-135.80	67.45	1.9	16.9	5.7	685	P	Rat
92	-135.94	67.43	8.4	265.7	15.6	534	P	Rat
93	-136.03	67.47	2.0	20.1	10.3	584		Rat
94	-135.94	67.53	5.1	225.0	4.4	425	Cx	Rat
95	-135.96	67.52	4.4	175.9	6.7	458	P	Rat
96	-135.71	67.86	2.6	73.9	11.0	622	P	Rat
97	-135.67	67.55	2.4	124.5	9.2	323	Cx	Rat
98	-135.68	67.55	1.3	115.0	3.9	322	Cx	Rat
99	-135.67	67.55	13.2	43.6	13.3	356	P	Rat
100	-135.57	67.62	4.9	25.9	9.2	287	Cx	Rat
101	-135.56	67.62	1.6	7.3	13.0	284	Mm	Rat
102	-135.56	67.62	3.7	191.3	3.6	259	Cx	Rat
103	-135.33	67.34	6.6	344.7	18.4	248	Cx	Stony Creek
104	-135.31	67.35	4.8	74.2	16.2	232	Cx	Stony Creek
105	-135.34	67.32	1.3	21.5	16.5	275	Cx	Stony Creek
106	-135.54	67.33	3.5	40.0	11.4	346	Cx	Stony Creek
107	-135.90	67.32	15.8	67.1	11.5	622	Cv	Stony Creek
108	-135.90	67.34	6.3	0.0	11.8	552	Cv	Stony Creek
109	-135.82	67.29	3.4	350.0	8.2	569	P	Stony Creek

110	-135.75	67.27	2.4	355.4	11.8	550	P	Stony Creek
111	-135.76	67.25	5.0	103.7	9.0	519	Ap	Stony Creek
112	-135.74	67.25	6.3	1.6	8.3	521	P	Stony Creek
113	-135.23	67.45	3.3	338.7	17.8	320	Mp	Stony Creek
114	-135.23	67.45	3.8	335.4	15.1	337	Mp	Stony Creek
115	-135.34	67.28	2.2	304.0	12.2	259	Cx	Stony Creek
116	-135.46	67.26	0.5	81.0	4.6	335	Mm	Stony Creek
117	-135.46	67.26	0.3	80.5	5.8	336	Mm	Stony Creek
118	-135.53	67.34	2.1	212.9	9.6	299	Cx	Stony Creek
119	-135.34	67.38	4.8	263.9	25.1	250	Mm	Stony Creek
120	-135.40	67.27	3.2	358.9	12.0	311	Cx	Stony Creek
121	-135.33	67.28	1.5	284.3	12.4	319	Mm	Stony Creek
122	-135.33	67.28	0.9	38.7	14.9	268	Mm	Stony Creek
123	-135.35	67.29	5.7	64.7	5.0	295	Cx	Stony Creek
124	-135.33	67.28	1.4	334.4	13.0	292	Mm	Stony Creek
125	-135.40	67.27	1.7	0.0	2.4	270	Ap	Stony Creek
126	-135.84	67.44	12.1	140.4	7.1	683	Cv	Stony Creek
127	-135.78	67.42	1.3	154.8	8.9	525	P	Stony Creek
128	-135.76	67.42	2.8	135.0	3.0	497	P	Stony Creek
129	-135.62	67.38	1.9	299.3	11.1	411	P	Stony Creek
130	-135.81	67.25	1.4	358.7	10.6	618	P	Stony Creek
131	-135.82	67.24	13.6	53.1	2.4	643	P	Stony Creek
132	-135.83	67.23	2.9	353.3	15.9	691	Cv	Stony Creek
133	-135.86	67.23	1.8	45.0	3.0	765	Cv	Stony Creek
134	-135.77	67.22	2.6	70.9	12.9	575	P	Stony Creek
135	-135.79	67.23	4.5	119.1	7.3	575	P	Stony Creek
136	-135.78	67.24	3.7	138.8	5.1	571	P	Stony Creek
137	-135.80	67.24	1.4	349.6	16.7	626	P	Stony Creek
138	-135.81	67.24	3.5	151.4	6.0	590	P	Stony Creek
139	-135.37	67.26	52.2	312.4	7.4	352	Cx	Stony Creek
140	-135.16	67.28	20.1	37.9	2.7	300	Mv	Stony Creek
141	-135.23	67.26	48.1	322.9	12.1	354	Mm	Stony Creek
142	-135.27	67.25	9.6	32.7	7.9	384	Cx	Stony Creek
143	-135.34	67.04	7.8	73.1	5.7	351	Cx	Unnamed watershed #3
144	-135.29	67.23	1.1	145.5	17.9	318	Cx	Vittrekwa
145	-135.28	67.23	2.2	174.8	17.9	339	Cx	Vittrekwa
146	-135.27	67.23	6.6	239.9	5.2	412	Cx	Vittrekwa
147	-135.16	67.21	0.5	66.4	8.3	258	Cx	Vittrekwa
148	-135.17	67.18	0.7	48.2	12.0	244	Cx	Vittrekwa
149	-135.42	67.18	0.7	342.0	18.0	310	Mp	Vittrekwa
150	-135.45	67.20	26.2	25.9	17.0	307	Cx	Vittrekwa
151	-135.39	67.22	1.5	264.4	12.1	261	Cx	Vittrekwa
152	-135.39	67.22	1.0	252.0	19.3	264	Cx	Vittrekwa
153	-135.70	67.15	17.0	95.8	11.6	522	P	Vittrekwa
154	-135.69	67.16	7.6	135.0	3.4	451	Cx	Vittrekwa

155	-135.68	67.15	9.7	22.6	6.2	520	Cx	Vittrekwa
156	-135.29	67.23	0.6	259.7	5.3	302	Cx	Vittrekwa
157	-135.75	67.05	14.6	59.6	8.0	571	Cx	Vittrekwa
158	-135.75	67.05	43.4	61.5	9.4	611	P	Vittrekwa
159	-135.76	67.07	8.0	27.5	6.7	518	P	Vittrekwa
160	-135.76	67.06	18.7	102.9	16.7	456	Cx	Vittrekwa
161	-135.80	67.12	20.6	69.4	4.1	672	Cv	Vittrekwa
162	-135.57	67.10	21.4	216.9	2.4	397	P	Vittrekwa
163	-135.65	67.00	6.6	27.3	8.8	426	C	Vittrekwa
164	-135.54	66.94	18.0	168.7	3.6	355	Mb	Vittrekwa
165	-135.55	66.94	17.5	227.7	7.1	299	Mb	Vittrekwa
166	-135.53	66.97	11.4	221.2	14.9	319	Mb	Vittrekwa
167	-135.68	67.02	3.1	71.6	0.8	431	Cx	Vittrekwa
168	-135.76	67.02	2.0	170.5	5.8	542	Cx	Vittrekwa
169	-135.76	67.02	4.0	118.1	15.8	518	Cx	Vittrekwa
170	-135.72	67.01	6.6	184.4	9.3	445	Cv	Vittrekwa
171	-135.79	67.05	9.7	80.5	5.8	621	P	Vittrekwa
172	-135.77	67.06	2.5	124.7	7.5	538	Cx	Vittrekwa
173	-135.73	66.98	2.8	31.0	8.3	453	Mh	Vittrekwa
174	-135.85	67.00	6.4	252.9	6.5	505	P	Vittrekwa
175	-135.87	67.00	3.4	111.8	10.2	449	C	Vittrekwa
176	-135.88	67.01	2.0	109.4	8.5	454	P	Vittrekwa
177	-135.88	67.01	3.5	29.1	18.9	487	Ap	Vittrekwa
178	-135.91	67.04	3.3	331.4	11.8	565	P	Vittrekwa
179	-136.01	67.05	3.4	205.0	3.9	580		Vittrekwa
180	-135.98	67.05	5.3	345.3	13.8	599	P	Vittrekwa
181	-135.68	67.17	3.2	129.5	10.4	481	Cx	Vittrekwa
182	-135.69	67.10	4.7	141.3	9.1	459	P	Vittrekwa
183	-135.58	67.10	2.4	27.4	14.2	406	Cx	Vittrekwa
184	-136.00	66.96	7.9	268.2	7.4	540	Mh	Vittrekwa
185	-135.81	66.76	3.2	334.3	19.6	562	Ap	Vittrekwa
186	-135.73	67.18	2.0	109.1	12.9	578	Cv	Vittrekwa
187	-135.79	67.12	1.9	252.1	16.6	575	Cv	Vittrekwa
188	-135.79	67.12	3.1	84.6	15.0	601	Cv	Vittrekwa
189	-135.59	66.98	4.1	124.9	9.5	364	Mh	Vittrekwa
190	-135.58	66.97	3.2	296.6	1.1	381	Mh	Vittrekwa
191	-135.67	66.95	4.4	8.7	6.3	468	Mh	Vittrekwa
192	-135.34	67.22	3.9	45.0	1.0	330	Mp	Vittrekwa
193	-135.71	67.02	2.2	188.7	12.4	451	Cv	Vittrekwa
194	-135.62	67.01	2.1	80.8	17.3	329	P	Vittrekwa
195	-135.74	67.01	3.4	40.4	12.3	523	Cx	Vittrekwa
196	-135.69	66.72	5.8	16.4	4.2	512	C	Road
197	-134.92	66.67	8.6	298.7	8.4	399	Mh	Road
198	-135.14	66.64	13.2	141.7	5.8	314	Mh	Road
199	-135.18	66.61	1.7	291.8	15.1	369	Mh	Road

200	-135.38	66.59	5.7	288.4	6.0	468	Mv	Road
201	-135.48	66.55	2.5	282.1	6.8	488	Mh	Road
202	-135.60	66.66	5.2	40.9	4.7	507	Mh	Road
203	-135.16	66.61	9.2	341.6	3.8	379	Mh	Road
204	-135.15	66.62	3.0	329.9	5.2	367	Mh	Road
205	-135.34	66.60	7.6	19.7	7.1	410	Mh	Road
206	-134.86	66.66	1.3	7.7	15.7	452	Mh	Unnamed watershed #4
207	-134.80	66.72	4.2	111.6	13.4	340	C	Unnamed watershed #4
208	-134.86	66.69	0.6	140.9	13.0	361	C	Unnamed watershed #4
209	-134.83	66.73	1.2	52.1	25.4	340	C	Unnamed watershed #4
210	-134.82	66.72	1.1	282.8	10.6	381	Mh	Unnamed watershed #4
211	-134.77	66.74	1.1	62.8	18.2	237	C	Unnamed watershed #4
212	-134.78	66.73	6.7	70.1	22.0	340	C	Unnamed watershed #4
213	-134.76	66.74	1.2	78.5	22.6	299	C	Unnamed watershed #4
214	-134.84	66.69	2.5	83.0	20.5	248	C	Unnamed watershed #4
215	-134.85	66.68	0.4	4.4	18.1	325	C	Unnamed watershed #4
216	-134.85	66.68	0.6	3.8	23.7	337	C	Unnamed watershed #4
217	-134.86	66.68	0.6	96.1	15.2	324	C	Unnamed watershed #4
218	-134.86	66.68	0.5	31.0	15.0	338	C	Unnamed watershed #4
219	-134.85	66.67	0.6	143.2	30.0	336	C	Unnamed watershed #4
220	-134.84	66.67	2.2	159.9	24.1	305	C	Unnamed watershed #4
221	-134.82	66.67	2.8	248.2	12.6	375	Mh	Unnamed watershed #4
222	-134.82	66.67	1.7	332.4	16.5	412	Mh	Unnamed watershed #4
223	-134.82	66.67	3.1	303.7	13.5	410	Mh	Unnamed watershed #4
224	-134.78	66.66	3.6	48.4	16.4	366	C	Trail
225	-134.86	66.64	3.2	276.3	6.5	322	Mh	Trail
226	-134.85	66.64	4.4	256.6	15.1	320	Mh	Trail
227	-134.88	66.64	1.8	115.0	22.5	203	C	Trail
228	-134.82	66.64	39.4	252.6	17.0	327	Mp	Trail
229	-135.13	66.55	12.0	99.2	14.7	314	C	Trail
230	-135.13	66.55	4.4	31.6	22.1	360	C	Trail
231	-135.12	66.55	3.4	5.4	21.7	325	C	Trail
232	-135.15	66.43	11.7	37.4	10.1	501	Mh	Trail
233	-135.14	66.43	11.9	270.0	3.8	491	Mh	Trail
234	-135.14	66.44	14.4	355.6	3.1	474	Mh	Trail
235	-135.21	66.43	4.7	76.3	9.0	420	C	Trail
236	-135.33	66.43	33.8	130.9	4.7	468	Mb	Trail
237	-135.40	66.47	6.8	104.0	2.0	439	Mv	Trail
238	-135.17	66.55	5.2	104.7	13.8	307	Mh	Trail
239	-135.21	66.55	5.3	37.6	7.8	419	Mh	Trail
240	-135.21	66.55	6.5	225.0	2.4	383	Mh	Trail
241	-135.28	66.54	4.4	60.1	17.6	399	Mh	Trail
242	-135.28	66.54	4.1	238.7	14.0	373	Mh	Trail
243	-135.30	66.52	18.6	116.6	3.2	396	Mv	Trail
244	-135.10	66.55	13.1	150.9	2.5	164	Ap	Trail

245	-135.16	66.56	7.3	75.5	6.6	394	C	Trail
246	-135.08	66.58	5.2	266.2	10.6	329	Mh	Trail
247	-135.32	66.43	1.4	302.2	7.6	471	Mb	Trail
248	-134.69	66.43	5.3	63.4	1.1	458	C	Caribou
249	-134.67	66.43	50.2	171.6	12.8	419	C	Caribou
250	-134.43	66.47	13.8	281.8	10.4	366	Mh	Caribou
251	-134.41	66.53	4.5	329.0	6.9	397	Mh	Caribou
